TUNNEL VALLEY GENESIS AND SUBGLACIAL DYNAMICS IN SOUTH-CENTRAL ONTARIO

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TITLE: **TUNNEL VALLEY GENESIS AND SUBGLACIAL DYNAMICS IN SOUTH-CENTRAL ONTARIO**

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ABSTRACT

Glacial sediments are found across formerly glaciated regions across the world and host a variety of important resources, ranging from groundwater to hydrocarbons, aggregate material, and mineral deposits. In southern Ontario, Canada, thick successions (up to 200 m) of Quaternary glacial sediments are truncated by large valleys (>30km long, 2 to >8.5 km wide, and up to 200m deep) that formed subglacially and have characteristic morphology and infill stratigraphy. These valleys are interpreted as (a new class of) tunnel valleys and strongly affect groundwater resources and flow systems at local and regional scales.

The overall context of the valleys is evaluated through an introduction to the study area, objectives, and background information on subglacial systems and geologic history of south-central Ontario (Chapter 1). Interpretation of valley genesis in Simcoe County is provided through an integrated, multi-faceted approach, involving: description of the morphology and sediment infill succession within the valleys from surficial mapping, sedimentological logging of continuously-cored boreholes, and geophysical surveys (Chapter 2); delineation and characterization of seismic architecture from high-resolution lake-based sub-bottom profiles in one of the valleys (Chapter 3); detailed site-scale field description of the internal characteristics of the regional Late Wisconsin till sheet in various subglacial settings (Niagara Escarpment, uplands, lowlands; Chapter 4); comparison of the characteristics of the subglacial bed within the study area to adjacent regions in southern Ontario (Chapter 5); and a synthesis of the major findings from all the different components of this investigation and suggestions for future work to shed further light on several questions that arise from this study (Chapter 6). Together, key data from these studies of tunnel valleys and related deposits – a near-continuous till sheet on the surface of uplands and along the flanks and floors of the tunnel valleys, multi-stage drumlinization of the till sheet following development of the tunnel valleys, variations in internal facies and physical properties within the till sheet in different subglacial settings, localized distribution of coarse-grained tunnel valley in-fill sediments, and gradational upward transitions from tunnel valley in-fills to fossiliferous proglacial lacustrine sediments – indicate multiple phases of subglacial meltwater, and direct subglacial, erosion and deformation contributed to the development of the valleys over a protracted time period during the Late Wisconsin. Landform and sediment associations within the valleys in Simcoe County and surrounding parts of the bed of the former Laurentide ice sheet in south-central Ontario, are inconsistent with previous conceptualizations involving the presence of large (>1000 km^2) subglacial lakes and the storage and discharge of regional-scale subglacial meltwater sheetfloods followed by ice stagnation. This study provides new data and insight to help refine reconstructions and better understand the evolution of past ice dynamics and subglacial processes, evaluate competing theories of regional landscape evolution, and provide new conceptual and (hydro)stratigraphic frameworks for future hydrogeological investigations related to groundwater exploration and use.

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The ride is over!

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into channels/canals/conduits. These progressively sap sediment and meltwater from adjacent areas in the bed. Continued deformation of sediment into channels and flushing towards the margin promotes tunnel valley growth as the channel migrates. B) meltwater excesses are drained by channels eroded into the bed, which may be subsequently infilled with sediment during waning flow stages. Ice creeps into the channel during periods of low flow. Subsequent flows create new channels that enhance and/or truncate previous deposits. The final valley form is the sum of erosion and deposition from all discharge events. C) Catastrophic erosion of tunnel channels is most likely when there is potential connection to a significant meltwater reservoir, and when a physical barrier (such as bedrock or frozen sediment) is present, enabling growth of the reservoir and increase in hydrostatic pressures. Slow groundwater sapping initiated at the ice margin erode headwards beneath the margin and can progressively increase the pressure gradient between reservoir and proglacial systems, triggering catastrophic failure of the reservoir seal. Note that the panels for catastrophic drainage show a longitudinal section, whereas cross-sections are shown for steady state and time-transgressive.

Fig. 2.1. a) topography and physical setting, b) drift thickness and c) annotated classified map of the study area (black rectangle) in south-central Ontario showing the network of valleys previously interpreted as tunnel channels (simplified from Brennand et al., 2006). The central axes of tunnel valleys and inferred paleoflow direction are highlighted with arrows. Elongate lakes and bays fill several of the larger valleys in the region. Inset map in a) shows location of figure within Great Lakes region of North America. PDF = Peterborough Drumlin Field; ORM = Oak Ridges Moraine; OM = Oro Moraine; WCV = Willow Creek valley; BV = Barrie valley; AE = Alliston Embayment; CV = Cookstown valley; HMV = Holland Marsh valley; ABV=Aurora buried valley. Location of Figs 2.2 and 2.7a shown in black lines, Fig. 2.2c labelled.

Fig. 2.2. a) bedrock geology and b) topography of the study area (black rectangle) and Laurentian Valley (shaded with long dashed outline). Short dashed lines denote major breaks in bedrock lithology from crystalline shield rocks in the northeast, to limestone, shale, and dolostone further southwest. See Fig. 2.1 for location, Fig 2.2c shown in thick black line; c) cross-sectional profile showing the bedrock topography and lithology as well as Quaternary sediment thickness across the Laurentian Valley in the central part of the study area. Modern hydrological features added for reference and valley names (see Fig. 2.1) shown to highlight broad-scale sediment architecture and upland-lowland physiography (see Fig. 2,4). Note the significant thinning of Early-Middle Wisconsin sediments beneath lowlands compared to intervening uplands. Vertical exaggeration ~30x; bedrock unit dips are not to scale, see legend in a).

Fig. 2.8. a) morphology of the flanks of tunnel valleys overlain on shaded relief map (5m cell size). Note the common drumlinization of valley flanks (green), particularly in upflow areas, the perched head of the CV and downflow end of the WCV (brown), and abrupt start of the WCV (orange) in Wisconsin age sediment. Many areas with eroded flanks (red) have been emphasized by nearshore erosion in deglacial lakes. Locations of Figs b and c outlined in black; b) close up view of the morphology of the flanks along the upland separating the BV and WCV. Note the large bar and spit features and abundance of aggregate pits, which indicate significant erosion of the flanks of the uplands following deglaciation, but white dashed lines show extent of surficial exposures of Newmarket Till below the former shorelines; c) close up view of the morphology of the central part of the HMV and ABV. The flat surface within the valleys is due to the thick infill of glaciolacustrine sediments. Steep slopes at the transition between the flat floors and the valley flanks are partially due to intense wave erosion in glacial Lake Algonquin; a subtle raised delta surface is visible where the E. Holland River meets the flat valley floor. Note the significant variation in valley width over a short distance downflow. The morphology of valley flanks alternates from drumlinized to eroded, and several large groundwater piping scars are visible, especially along the southeastern flank. The relationship between the HMV and ABV is incompletely understood (but see Gerber et al., 2018).

Fig 2.9. a,b) High-resolution CHIRP sub-bottom profile across the eastern part of Kempenfelt Bay (submerged eastern extension of the BV; see Figs. 2.1, 2.4-2.8). A strong undulating reflector rises towards the northern and southern shorelines of the modern bay and is characterized by abundant smaller-scale (few m) ridges (drumlins/flutes?) on local highs beneath the valley. The drumlinized reflector is locally overlain by transparent or chaotic facies interpreted as coarse-grained sediments. An anticlinal reflector (esker?) is draped by horizontal strata (subaquatic fan?) in a). Well-laminated strata comprise the bulk of the sediment infill (modified from Mulligan, in prep); c) uninterpreted and d) interpreted S-wave seismic reflection profile (data from Pugin et al., 2018) with borehole logs and surficial mapping as geologic constraint (Figs. 6, 7). SS-11-07 is 200 m south of profile and SS-12-06 is projected from 2.7 km south for reference. A mound capped by strong amplitude reflectors shows similar 2D geometry to drumlins on uplands and is interpreted as a buried drumlin capped by NT (SU1). See Fig. 2.6 for additional information.

Fig. 2.10. SU1 facies. a) cobble-rich, dense, massive diamict; b) clast with sub-parallel striae and gouges; c) clast with chaotic striae, likely indicating transport in a mobile deforming bed; d,e) monolithic, angular clasts immediately above shale bedrock; f) silty upper facies sharply overlain by fine-grained rhythmites with dropstone layers (SU4). See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.

Fig. 2.11. SU2 facies. a) cobble-rich SU2 facies showing an interbedded contact with overlying SU3 deposits. Matrix material was poorly recovered, but there is apparent stratification of pebble- and cobble-rich facies; b) close-up of clasts in SU2 showing large, sub-rounded clasts of diverse lithologies; c) openwork pea gravel at the top of SU2 directly overlain by silty very fine-grained sand (SU3); d) poorly sorted gravel with coarse-grained sand matrix near the base of SS-12-07. See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.

deposition produces a mosaic of substrate materials at the IBI during Late Wisconsin ice cover. Thicker sand bodies act as well-drained, resistant cores, resulting in ice deforming around them contribution to higher ice fluxes and erosional capabilities in intervening areas. Newmarket Till (SU1) not drawn; c) the predominantly fine-grained substrate is unable to drain meltwater excesses and a more efficient drainage system develops (canals or discrete conduits), enhancing vertical erosion in swales. Predominantly N-S drainage developed due to regional ice gradients, but meltwater systems could have been captured by re-entrant valleys in the Niagara Escarpment to the west. Breaching of cohesive, fine-grained units and till provides connection between pressurized water in confined subglacial aquifers and drainage systems at the IBI. This increases meltwater discharges and sediment erosion by piping and/or turbulent flow. Inset shows the influence of gradients within the confined aquifers contributing to flow and sediment evacuation during breaches, which could have undercut valley flanks, leading to sediment collapse and significant widening of the valley during reoccupation by ice (dashed black line); d) during periods of reduced meltwater production, ice fills the valleys and till deposition (SU1) is resumed. Changing ice sheet configuration during late phases of glaciation changes ice surface gradients towards the southwest, and a second phase of valley systems oriented parallel to ice flow cross-cut older N-S systems. Following shut/slowdown of meltwater production, till deposition resumes; e) changes in ice dynamics and/or basal thermal regime shifts basal ice conditions from till accretion to till sheet drumlinization across both uplands and valleys; patches of till may have also been removed within valleys by solely glacial erosion; f) localised deposition of SU2 and SU3 as ice-contact subaquatic fans, primarily within the valleys, in ice-dammed lakes during deglaciation. SU4 is deposited as fine-grained glaciolacustrine sediments in more ice distal locations during progressive ice retreat. Diagrams are for illustrative purposes - no horizontal or vertical scale is implied.

Fig. 3.3. Line 4 profile showing the overall architecture of seismic stratigraphic successions in the survey. SSS I has an undulating topography that forms the base of the valley and deepens towards the center of the basin, marked by local ridges (interpreted here as drumlins). SSS II has a wedge or lenticular geometry infilling local lows and a transparent of chaotic internal fill. SSS III forms the bulk of the valley infill, consisting of well-layered parallel draped reflectors affected by near-vertical normal faulting, but is locally highly deformed (compressional folding and reverse faulting) high up within the succession and along the flanks of the valley (SSS IIIb). SSS IV locally occurs as wedges along valley flanks or local lows on the upper surface of SSS I or III. SSS V is nearly transparent and forms a drape of fairly consistent thickness across the base of the valley. Note numerous small gas chimneys and possible gas pockets trapped within the sediment fill, as well as some larger areas of fluid (gas and porewater?) escape along fault planes. In general, the upper surfaces of SSS I and SSS III form stronger multiples than the lake bottom (SSS V). X axis in meters, y axis in seconds.

Fig. 3.4. a) Line 2 profile showing sediment architecture in the eastern part of Kempenfelt bay. Note numerous drumlins/flutes on surface of SSS I, anticlinal reflector within SSS II at 0.04s in the low on SSS I in the north draped by flat-lying reflectors. Numerous small slumps occur within SSS III and upward fluid escape structures are visible through the deepest (poorly imaged) part of the basin; b) central part of Line 4 (see Figs. 3.2, 3.3 for location). Individual ridges on SSS I are clear on the overall undulating surface, which forms the oldest acoustic basement of the survey. SSS I is not imaged in the central part of the profile where SSS II is thickest. Note the transparent seismic facies of SSSII in the deepest part of the profile, compared to the chaotic facies in the local low in the northwest (0.0625s).

Fig. 3.5. a) Line 8 profile showing facies and architecture of SSS III. Note near-vertical normal faulting along the base and northwestern flank of the valley, with downwarping of adjacent reflectors in SSS III. Reflectors are largely parallel and conformable, high amplitude events occur along the flanks of highs and taper out towards central parts of lows. Aggradation in the deepest parts of the valley, likely due to traction current activity, creates a swell consisting of a broad anticlinal feature with concordant internal reflectors. A large zone of deformed sediments occurs along the southeastern flank of the valley and is shown in greatr detail in b; b) close-up view of deformed zone (SSS IIIb) showing it is composed of (minimum) two mass movement events (shown by dashed pink lines). Truncation of underlying reflectors points to localized erosion at the base of the slumps. Note stacked reverse faults and rotated strata indicating movement toward the basin center. Local downwarping of reflectors within SSS V may indicate fluid escape through the upper sediment packages; c) Line 6 profile showing a wide zone of deformed SSS IIIb sediments in the central part of the valley. Radial changes in fault planes are consistent with a flower structure indicating transverse movement beneath the visualized succession, but additional data is required in order to confirm relation of the deformation to underlying features. Strong reflectors in the lower part of the profile may be consistent with SSS I, but data quality is insufficient for confident identification.

Fig. 3.6. a) Line 13 profile showing significant topographic variation in the central part of the valley. The low on the lake bed (SSS V) coincides with numerous graben features penetrating SSS III into SSS V. The lower part of SSS V contains higher amplitude reflector that onlap the flanks of grabens, while the upper part is more transparent, with concordant reflectors draping the entire valley fill succession; b) Line 9 profile showing SSS IV infilling local lows and forming thin wedge-like bodies that taper out towards the basin center. A large slump scar within

SSS III is entirely infilled with SSS V muds. A small ridge of SSS V in the northerwesternmost part of the profile (.011-0.12 s, approx. 8 m depth) may correspond to a beach deposit formed during lowered lake levels.

Fig. 3.7. Line 14 profile showing thick successions of SSS III that are largely undisturbed except for minor normal faulting. SSS V has a positive relief in the central part of the valley, downlapping onto SSS III-IV near the valley margin in the southeast. Fluid escape (likely including porewaters) is evident along the base of a fault plane. Pitting of the lake bed indicates recent flow, either due to recent activation of the fault, or through sustained fluid flow due to strong upward gradients within the valleys.

Fig. 3.8. Conceptual model for valley infill; a) deposition of coarse-grained sediments (SSS II) along lows within the valley. Early sedimentation was likely in conduits confined by the glacier (eskers) which were later draped by more laterally extensive sediments (subaquatic fans). Ice blocks are locally buried by sediment; b) ice retreat removes the coarse-grained sediment supply, and fine-grained rhythmite deposition form suspension dominates in glacial lakes Schomberg and Early Algonquin. Melting of buried ice blocks leads to the development of non-penetrative faulting along the base of SSS III. Reflector amplitude typically increases upward due to falling water levels and re-working of unconsolidated and unvegetated valley walls by wave erosion; c) falling water levels to the Kirkfield low phase may have triggered instability along the valley walls, leading to the extensive mass movement deposits in the upper part of SSS III. Sedimentation continues during rise to main Lake Algonquin; d) following the Penetang phase of post-Algonquin lake levels, a climatic-induced lowstand during the early-mid Holocene (e.g. Lewis et al., 2008) triggered mass movement events and debris flow deposition (SSS IV) and during transgression to modern levels, mud and gyttja accumulated throughout the basin. Recent groundwater flow reaches the lake floor along fault planes.

Fig. 4.1. geography and topography of southern Ontario. Overdeepened Great Lakes basins are separated by intervening highs, where interlobate moraine complexes (short dashed lines) were locally deposited. Location of Fig. 4.2 outlined in black rectangle.

Fig. 4.2. A) topography and glacial landforms in south-central Ontario. Streamlined bedforms in black, recessional/end moraines in bold black, eskers in dashed blue lines, and axes of Late Wisconsin tunnel valleys in red, with arrows denoting paleoflow direction. Large end moraines labelled in white; B) major bedrock lithological divisions (Armstrong and Carter, 2010) and locations of re-entrant valleys (Straw, 1968) overlain on bedrock topography (Gao et al., 2006). Inset profile in upper right (see thin black line for location) shows geology of the Niagara Escarpment cuesta. The Laurentian Valley/trough is shaded and stippled; C) drift thickness map of south-central Ontario (Gao et al., 2006). Little to no sediment occurs on shield terrain, whereas thick sediment successions underlie the Laurentian Valley/trough (dashed black lines) and the Oak Ridges Moraine; D) previously-mapped distribution of Late Wisconsin till sheets in south-central Ontario (Ontario Geological Survey, 2003); E) simplified surficial geology of the study area (modified from Mulligan and Bajc, 2012; Mulligan 2017b, c). Red dashed lines show the boundary of physiographic regions (modified from Chapman and Putnam, 1984). NE=Niagara Escarpment, U=Uplands, L=Lowlands, ORM=Oak Ridges Moraine, OM=Oro Moraine.

Fig. 4.3. Glacial history of the study area. A) extensive proglacial lakes fronting the advancing LIS towards the end of the Middle Wisconsin Episode; B) regional-scale southwestward flow

during the last glacial maximum (LGM); C) thinning of ice sheets during deglaciation promotes the development of topographically-controlled ice lobes/streams flowing out of the Great Lakes basins; D) late-stage ice margins and ice flow directions in southern Ontario. Ice flows radially from within lake basins where the thickest residual ice masses remain. Thick and voluminous sediments are deposited within the Oak Ridges Moraine (orm) in southern Ontario.

Fig. 4.4. Topography of the study area and location of key exposures of the NT in outcrops (stars) and boreholes (circles). Previously-mapped till plain distributions and newly-mapped ice-marginal features correspond to major moraine systems (Barnett, 1992). Thick black lines outline 3D mapping study areas in southern (lower) and central (upper) Simcoe County.

Fig. 4.5. High-resolution digital elevation models displaying the morphology of the NT within selected parts of the study area (see Fig. 4.4 for locations). A) the northern Beaver Valley contains abundant drumlins displaying converging long-axis orientations into the re-entrant valley, consistent with strong topographic funneling; B) At least three generations of streamlined bedforms exist on the NT-Elma-Catfish Creek Till southeast of the Beaver Valley above the crest of the Niagara Escarpment, the youngest terminating at the Singhampton Moraine; C) streamlined uplands and intervening tunnel valleys west of Lake Simcoe. Irregular north-south oriented ridges may represent remnant drumlins from an earlier phase of ice flow; D) southeastward streamlining of the NT-Allenwood Till southeast of Georgian Bay indicating late-stage ice flow switching. Ice flow during the LGM is recorded by remnant drumlins oriented to the southwest; E) possible multiple generations of bedforms developed on the NT northwest of Lake Simcoe indicating progressive shift in ice flow towards the west. Bedforms occurring along the flanks of the Willow Creek (WCV) and Barrie (BV) tunnel valleys and below Lake Simcoe indicate they post-date the excavation of the valleys.

Fig. 4.6. Generalized stratigraphic architecture throughout the Laurentian Valley/trough (dashed red lines on inset map). Key landform-sediment successions are highlighted in arrows and italics. Radiocarbon dates are provided in ^{14}C kyr BP (Mulligan and Bajc, 2018). Acronyms below logs denote the environment/physiographic region within which the borehole was drilled and acronyms above the logs denote specific features: MB=Minesing Basin, BV=Barrie Valley, CV=Cookstown Valley; HMV=Holland Marsh Valley, AE=Alliston Embayment. Boreholes beginning with 11-, 12-, 13- are from Bajc et al., 2015. Boreholes beginning with 15-, 16-, 17- are from Mulligan (2016; 2017a). Sources for other cored borheoles provided below the logs.

Fig. 4.7. A) surficial exposure of the NT and the underlying TF which is deformed in the uppermost 1.5 m, above a 20-40 cm thick silt bed. Red box denotes location of close-up image in B). Modified from Mulligan and Bajc (2018); B) vertical facies gradation within the NT from stratified and deformed diamict and gravelly sediments upwards into highly consolidated massive and fissile silty sand diamict; C) striated and faceted boulder at the base of the NT where it overlies coarse sand of the TF. Ice flow was from the upper right (northeast); D) striated and faceted boulder at the base of the NT where it overlies thick silt and clay deposits. Ice flow from the bottom of the photo (towards the WSW), note termination of striae at plucked surface in the upper part of the photo; E) loaded contact (see flame developed in dark brown silt between cobbles) between very fine sand and silt of the TF and interstratified diamict and sands forming the base of the NT, approximately 700 m east of photo in D). See Fig. 4.4 for photo locations.

Fig. 4.8. Stratigraphic log, particle size analysis from sampled intervals, geochemical analysis of diamict units, and recovery statistics from borehole 13-07 drilled within a Late Wisconsin (MIS

2) tunnel valley (Figs. 4.4-4.6). Thick diamict successions between 48 and 86 m can be correlated as three distinct till units based on matrix texture, carbonate/dolomite ratios, heavy mineral abundances, and core recovery rates (strongly related to matrix consolidation).

Fig. 4.9 A) finely laminated, thick-bedded (20-40 cm) silt-clay rhythmites comprising the TF exposed in the core of a small drumlin along the eastern margin of the Holland Marsh Valley. The rhythmites are deformed by reverse faulting (highlighted by orange oxidation from modern groundwater flow). Shovel handle is 50 cm long; B) highly consolidated, fissile, silty sand NT directly overlying laminated fine to medium sand of the TF. Photo is approximately 20 cm across; C) multiple diamict beds and lenses occurring in direct contact or separated by thin, deformed sand stringers. Scale card is 9 cm wide; D) highly consolidated NT occurring as stacked massive beds 2-4 m thick exposed in a sand pit restricts pit wall excavation; E) winnowing of matrix by modern fluvial processes highlights moderately developed fabric (long axes oriented to the SW) along the flank of a large drumlin on the floor of a late Wisconsin tunnel valley; F) clast stacking along the margin of a tunnel valley. Clasts dip 80° E (solid arrow in inset), nearly parallel to drumlin long axes in the vicinity (dashed lines in inset). Knife blade at top is 3 cm wide; G) parallel fault planes accentuated by modern fluvial erosion strike NW-SE, orthogonal to drumlin long-axes. See Fig. 4.4 for photo locations.

Fig. 4.10 A) gradational vertical transition from silty sand diamict (greybrown colour below 9.1 m) to slightly sandy silt till (reddish gray above 8.6 m), typically referred to as Allenwood Till. Reddish colour derived from incorporation of red Queenston Formation shale outcropping to the west; B,C) variation in till facies observed in successive outcrops along the Nottawasaga River. Note clast size and abundance in (B) compared to (C). Grubhoe in (B) is 1.1 m long, knife blade in (C) is 3 cm wide; D) slickensided clay seam within sand-rich NT exposed in a possible remnant drumlin. Core is approximately 8.5 cm wide; E-G) selected intervals of a thick glaciotectonite succession grading from deformed silt and clay with rare grit and IRD, to deformed silt-rich diamict and interstratified silt, into clast-rich and highly consolidated silty sand till. Cores are approximately 8.5 cm wide, see Fig. 4.4 for photo locations.

Fig. 4.11 A) stratified sand and local gravelly beds deposited in subaquatic fans comprising the upper Thorncliffe Formation (TF). The NT was either not deposited or eroded during extensive littoral modification in postglacial lakes. Section is approximately 7 m high; B) stratified TF sediments incised by a steep-walled, asymmetrical channel infilled with interstratified and deformed diamict and gravel; C) sands in the TF within 20-30 cm of the channel are deformed by fluidization, injection structures and minor faulting (see black box in (B) for location); D) Overconsolidated, blocky silty sand facies of the NT overlying deformed fine to medium deformed sand of the TF; E) part of a time-section of a S-wave seismic reflection profile (data from Pugin et al., 2018; Figs. 4.4 and 4.5). Borehole 15-03 was drilled to intersect a channelized feature in the subsurface (from approx. 2100-2700 m on the profile). Approximately 10 m of poorly- to moderately consolidated, interstratified diamict and fine-grained sand comprise the NT. See Fig 4.6 for legend. Locations marked with '?' denote areas where reflector continuity limits confidence in unit correlations.

Fig. 4.12 A) stratified and strongly colour banded NT immediately downflow from a lithologic boundary between grey shale and sandstone beds and red shales; B,C) monolithic clast-rich and angular breccia facies comprising the basal part of the NT above black shale bedrock; D) striated and faceted limestone boulder forming part of a crudely-developed pavement; E) parallel striae (arrowed) ornamenting a limestone cobble near the base of the NT; F) glaciotectonized red shales overlain by thin fine-grained NT on a bedrock outlier along the Niagara Escarpment. Multiple scales of folding are present. Shovel at bottom right is 0.5 m long. Cores are approximately 8.5 cm wide, see Fig. 4.4 for photo locations.

Fig. 4.13 A) Kame and kettle topography within the Gibraltar Moraine (fence posts in foreground are 1.5 m tall); B) cobble-rich sandy diamict comprising a small moraine ridge upflow of the Gibraltar Moraine east of the Beaver Valley. Cobbles and boulder are derived almost entirely from subcropping beige Guelph Formation dolostone. Section is approximately 3.5 m high.; C) limestone and dolostone cobble-rich diamict directly overlying red, stone-poor diamict derived from the Queenston Formation shale. Section is approximately 4 m high.; D) sand, gravel and lenses of diamicton comprising a moraine ridge east of the Beaver Valley. Section is approximately 2.5 m high. See Fig. 4.4 for photo locations.

Fig. 4.14. A) 8-10 m of the NT (formerly Elma Till) exposed along an abandoned meander of the Beaver River. Till is overlain by postglacial lake and fluvio-lacustrine sediments; B) small boulder exposed in the outcrop shown in A), showing an older set of gouges and grooves (double-ended arrows) cross-cut by parallel striae parallel to long axis. Ice flow was toward the bottom of the photo, note streamlined upper edge of the clast. Scale on compass is in cm; C) finely laminated sand-rich diamict and fine- to medium-grained sand lenses; D) stratified, interbedded and deformed grey diamict and beige-brown very fine sand (colour changes enhanced by oxidation from enhanced groundwater flow) near the base of the NT; E) basal contact of the NT with the underlying faulted (arrows show sense of displacement) and deformed TF sediments. Photo is approximately 1.5 m below (D). See Fig. 4.4 for photo locations.

Fig. 4.15. Location map of surficial grain size samples collected from the NT in central Simcoe County (black line) overlain on physiography (modified from Chapman and Putnam, 1984). Matrix composition is displayed in pie charts, and a clear trend of decreasing sand content southward is evident across all three regions, except above the Niagara Escarpment in the southwesternmost corner of the study area, where high sand content is prevalent.

Fig. 4.16. Ternary diagrams showing matrix grain size data from borehole samples collected from southern Simcoe County. The NT and older tills (OT1 and OT2) share many textural characteristics, but OT1 has a higher silt and clay proportion and OT2 has higher silt and less sand in the matrix. Numerous additional parameters are required for stratigraphic assessment.

Fig. 4.17 Thickness of the NT in all Simcoe County boreholes, grouped by physiographic region. Average thicknesses for all borehole intersections, and adjusted averages removing effects of post-depositional erosion are displayed. The NT is generally <12 m thick, except in glaciotectonite complexes, drumlins, and ice-marginal areas.

Fig. 4.18. Summary diagram illustrating the stratigraphic architecture (see Fig. 4.6 for legend) and facies variability within the NT across the physiographic regions/ subglacial environments in the study area. No scale implied but typical thickness ranges are provided at the left of each log. Arrows denoting paleo-ice flow directions are interpreted from clast provenance, till sheet morphology, clast faceting/striae orientations and/or substrate deformation features. TV=tunnel valley.

shown in black, eskers in dashed blue, and moraine crests in solid orange lines. Note the generally parallel orientation of streamlined features on the Shield, and their relationship with features to the south. See Fig. 1 for location. B. 25-m elevation model showing orientation and distribution of streamlined bedforms (black lines) in the Ottawa- St. Lawrence lowlands. Note the curvilinear track of bedforms, parallel to the overall contours of the eastern part of the Algonquin Highlands. Location of Fig. 7E, F shown in black, see Fig. 5.1 for location. C. Satellite image (top; Google Earth) and 2-m elevation model of MSGL on a streamlined till residual in the southern Algonquin Highlands (see Fig. 5.3A). The feature is bisected along the central axis, creating two 'tails'. D, E. 2-m elevation model and surficial geology overlain on hillshade (2-m cell size) showing morphology of streamlined features and their relation to mapped sediment cover (Ontario Geological Survey 2010).

Fig. 5.4. A. 5 m hillshade of PDF east of Lake Simcoe (see Fig. 5.1 for location). Streamlined bedrock escarpment noses eroded into Ordovician limestones shown by black arrows (pointing in direction of former ice flow). B. Digital elevation model and glacial landform distributions for area shown in (A). Note the local curvilinear paths of bedforms in the northeast, and their correspondence to the location and orientations of valleys, lake basins, and gaps between streamlined escarpment noses. At least three generations of bedforms are visible, labelled 1-3, and are identified by strong obliquity of bedform orientations locally (center of image) and/or the overprinting of large (older) drumlins (shown with thicker black lines) with more subtle flutings (northeast corner, see Fig. 5.5C). The downflow extent of each flow set is marked by terminal moraines.

Fig. 5.5. A. 5 m hillshade of PDF in the Sturgeon Lake area (see Fig. 5.1 for location). Streamlined bedrock escarpment noses eroded into Ordovician limestones shown by black arrows (pointing in direction of former ice flow). Multiple classes of streamlined bedforms are visible including comma-form (c), and parabolic (p). B. Digital elevation model and glacial landform distributions for area shown in (A). Note the general parallelism of large, broad (older) drumlins (bold black lines) compared to the radial fanning of younger low-relief flutings (thin black lines) that commonly terminate at end moraines. C. Detailed view of morphology west of Sturgeon Lake. Apparently parabolic forms shown in (A) can be explained through multiple discrete streamlining phases: an early drumlin-forming phase from ice flowing SSW, followed by a later phase by SW-flowing ice. Two sinuous channels cross-cut the upland and are marked with flat-topped lobate features at their eastern ends. These are interpreted as glaciofluvial or glacial lake spillways that would have developed following deglaciation, while ice still blocked lower outlets to the north. D, E. close-up views (see (A) for location) of bedform orientations and relationships to end moraines. Note the radial fanning and orthogonal orientation of younger flutings to end moraines in (D) and the clear termination of a flow set identified by flutings at a set of small moraine ridges.

Fig. 5.6. A. Bedform distribution and orientation within the Rice Lake Corridor in the eastern part of the PDF (lateral boundaries shown by transparent grey lines, see Fig. 5.1 for location). This area was previously interpreted to mark the location of a lateglacial ice stream (Maclachlan & Eyles 2013) and has the densest distribution of streamlined forms within the PDF. Large, broad (older) drumlins are consistently oriented NNE-SSW and are best preserved outside the margins of the Rice Lake Corridor. A few older drumlins are preserved within the Rice Lake Corridor, particularly in the southwestern part, where it appears ice was thinner, given the

deflection of bedforms around the upland near the southwest end of Rice Lake. A few large drumlins (bold lines) are visible among the otherwise low-relief MSGL in the central part of the Rice Lake Corridor. B. Close-up of the central part of the Rice Lake Corridor, where a wide range of streamlined bedform morphologies are visible. Spindle forms predominate, but comma-form (c), asymmetric (as), parabolic (p), multi-tailed (m), grooved/bisected (g), and features with crescentic scours around their stoss ends (cs). Many of these more complex forms originate downflow of large, high streamlined hills, which likely represent re-moulded older drumlins. Black hashed lines denote areas where Palaeozoic bedrock crops out at surface. C. Enlarged view of subsection of the area shown in (B) demonstrating the common occurrence of linear to slightly curved grooves (arrowed) that appear to entirely truncate numerous drumlin forms and possibly create the multi-tailed drumlins. Black hashed polygons show areas of dense coniferous vegetation has led to ground surface interpolation issues.

Fig. 5.7. A,B. 2 m cell size hillshade and interpreted elevation model for the area east of the ORM and PDF. Older PDF drumlins trace a broad, curvilinear path form the NNE toward the Lake Ontario basin to the WSW. These are ornamented and/or reworked by later ice flow from the east and southeast. Note that unlike the previous examples, no moraine is visible to mark the extent of the late-stage ice advance, but subtle recessional ridges of the Brighton Moraine are shown in orange. Nearshore erosion along the former Glacial Lake Iroquois (GLI) shoreline may be partially responsible for the erasure of a moraine, if it was deposited. C,D. 5 m cell size hillshade and interpreted elevation model for the area south of the ORM and PDF. Multiple apparent comma-forms (labelled with lowercase c's on map) are marked. Inset shows detail on 2 m DEM, showing tcomma-forms consist of a broad hill oriented NE-SW (interpreted as older drumlin forms), with a fine linear ridge emanating from their NE end indicating a clockwise shift in ice flow directions (arrows labelled 1-5) and re-moulding by ice flowing NW and NNW from within the Lake Ontario basin during the formation of the LODF, with a downflow limit along the flanks of the ORM in the north. E,F. 2 m cell size hillshade and interpreted elevation model for the Ottawa Valley area. The Precambrian-Paleozoic unconformity is marked with a dashed black line.

Fig. 5.8. Relative age of flow sets within in the central part of the PDF (modified from Marich 2016). Long axes of individual streamlined bedforms are colour coded by each flow set, from oldest (1; green) to youngest (5; pink). Note that flow set are typically separated by primarily till-cored ridges associated with the Lake Simcoe Moraines (black lines). Note the occurrence of older (green) bedforms within younger flow sets, suggesting a widespread drumlinization of the regional till sheet followed by local modification by small-scale readvances. See Figs 5.4-5.6 for details. The step-like change in grey tones on the shaded relief map arises from the intersection of the margins of the 2-m (eastern) and 5-m (western) elevation model boundaries.

Fig. 5.9. Generalized correlation of regional flow sets with major Late Wisconsinan events in south-central Ontario. Phase 1 shows regional sheet flow and drumlinization directions. The transition from phase 1 to phase 2 and higher is characterized by significant reorganization of the LIS and the onset of ice streaming and major flow switching and readvances in the region (see Ross *et al*. 2006; Eyles & Doughty 2016; Marich 2016; Mulligan *et al*. 2016; Sookhan *et al*. 2018). The distributions of landforms and resulting flow sets bear strong resemblance to the time-transgressive bedform pattern model of Stokes & Clark (2001: Fig. 5.8) showing multiple bedform populations within flow sets arising from several episodes of bedform generation. See Figs 5.4-5.8 for details.

Fig. 5.10. Conceptual model of landscape development in the PDF showing (A) transition from hard bed with local patches of thin till on Shield terrain, into a mixed bed mosaic of bare limestone bedrock and increasingly thick patches of till southward to the soft bed underlain by thick successions of unconsolidated sediments. An early phase of till sheet drumlinization occurred beneath a broad ice stream (Ontario-Erie Ice Stream: OEIS) prior to (B) lateglacial flow switching initiating the break-up of the OEIS into smaller ice streams, the Halton Ice stream (HIS) in the south and the Simcoe Ice Stream (SIS) in the north. Flow switching promoted the development of new generations of highly elongate bedforms, and modification/removal of pre-existing bedforms. The Oak Ridges Moraine was deposited between the two ice streams (e.g. Sookhan *et al*. 2018a). Figure modified from Eyles & Doughty (2016).

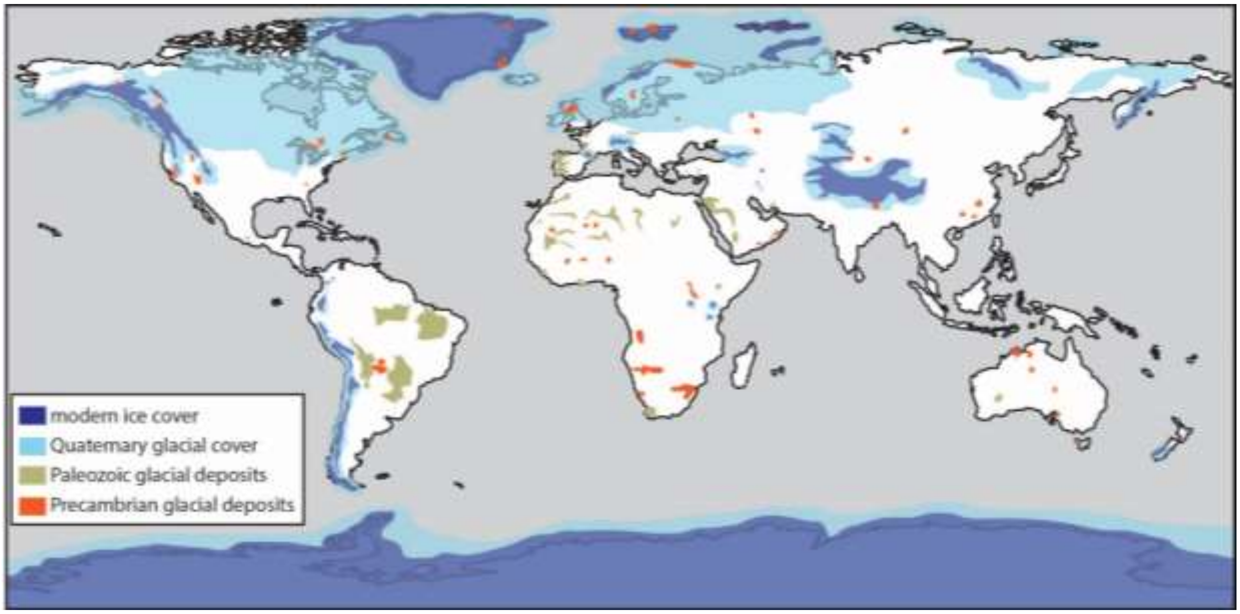
CHAPTER 1: INTRODUCTION

1.1. INTRODUCTION

Glacial sediments occur on all continents of the globe (Fig. 1.1) and locally form or host significant resources. Despite their significance, much remains to be learned about the nature of erosional and depositional processes and patterns associated with glacial deposits across a full range of spatial and temporal scales. In southern Ontario, Canada, a wealth of geologic data have been collected, as thick successions of Pleistocene glacial deposits provide aggregate material for growing cities and towns and host or control recharge pathways to aquifers that provide drinking water for more than three million people (Barnett, 1992; Sharpe et al., 2014).

Mapping of Ontario's glacial deposits began in the late 1800's (e.g. Spencer, 1890; Goldthwait, 1910; Taylor, 1913; Coleman, 1939), and government agencies began intensifying mapping efforts in the 1960's and 1970's (see Karrow, 1974). In the early 1990's, an increased focus on improving understanding of sediment architecture in southern Ontario led to the collection of high-quality surface and subsurface information (Boyce et al., 1995; Howard et al., 1995; Barnett et al., 1998; Pugin et al., 1999; Pullan et al., 2002; Sharpe et al., 2002). These efforts were drastically increased following the events at Walkerton in 2000, when seven residents died and thousands more became ill following contamination of the municipal drinking water supply with *E. Coli* bacteria (Worthington and Smart, 2017). One of the many lessons learned from this tragedy was that, in many areas of southern Ontario, there was insufficient knowledge of the subsurface setting of groundwater systems supplying public (and domestic) water sources (O'Connor, 2002). In response, the Ontario Geological Survey (OGS) developed a groundwater initiative in 2002 that began with data assembly and standardization of available

Fig. 1.1: Generalized map of ancient (Precambrian and Paleozoic) glacial deposits and recent (Quaternary and modern) ice cover worldwide. Data from multiple sources, (Hoffman and Schrag, 2002; Ehlers and Gibbard, 2007; Le Heron et al. 2018)


modern ice cover
Quaternary glacial cover
Paleozoic glacial deposits
Precambrian glacial deposits

geologic information, but quickly shifted to collection of new, high-quality subsurface data to better map groundwater geochemistry, and the three-dimensional (3D) properties and architecture of Paleozoic bedrock geology, as well as overlying glacial sediment successions (e.g. Bajc et al., 2018).

3D sediment mapping involves a basin analysis approach, integrating surficial sediment and landform mapping, regional geophysical data collection, and drilling of continuously-cored boreholes in order to better define bedrock topography, characterize the properties and architecture of subsurface sediments, and identify potential aquifers, groundwater flow zones, recharge areas, and surface water – groundwater interactions (Sharpe et al., 2002). To date, the OGS has drilled over 300 continuously-cored boreholes, most of which penetrate the thick (locally >195 m) Quaternary sediments down to underlying bedrock. These boreholes provide critical insights into the distribution and properties of subsurface sediments and aquifer systems (Bajc et al., 2014; 2018; Benoit et al., 2016; Burt, 2018; Mulligan and Bajc, 2018).

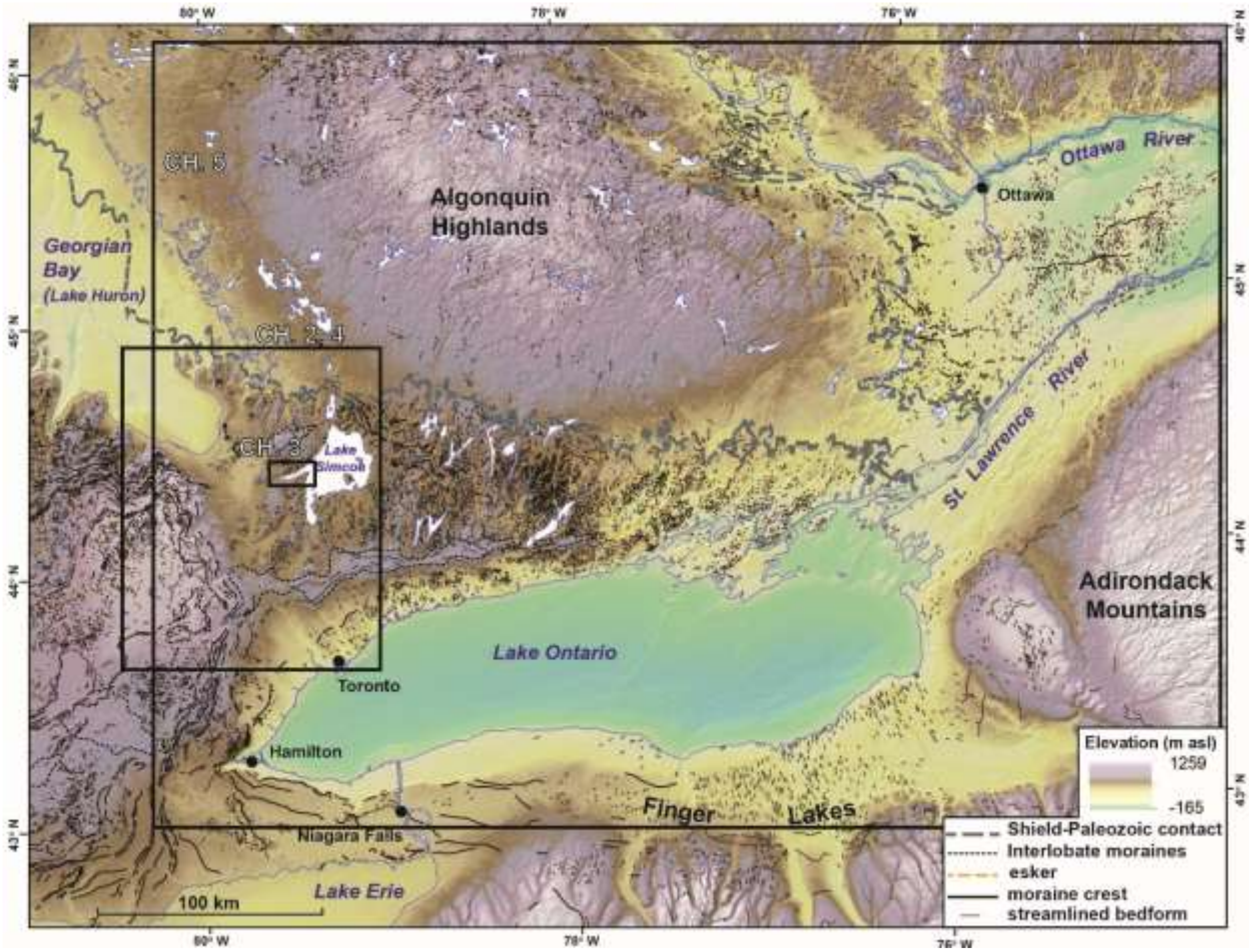
Data collected in support of 3D sediment mapping in Simcoe County, as well as from other recently-released high-resolution regional terrain data sets and geophysical surveys across southern Ontario, provide the high-quality support for the analysis of glacial event chronology and paleoenvironmental conditions (Bajc et al., 2015; Mulligan and Bajc, 2018; Mulligan et al., 2018a). The collection of these data permits analysis of regional trends in bedrock, sediment, and landform characteristics on similar spatial scales to past ice sheets, but also provides high-quality and high-resolution local constraints on the physical properties of significant geologic units. These data also support the development of conceptual models to describe evolving conditions beneath the former Laurentide Ice Sheet (LIS) during the last glacial cycle and will eventually permit development of computational and quantitative glaciological models in the future.

Within Simcoe County, these high-quality data sets form the basis for an integrated analysis of large (>30 km long, 150m deep, up to 8.5 km wide) tunnel valleys that form the western part of an even larger system of valleys that spans across south-central Ontario (Brennand and Shaw, 1994; Sharpe et al., 2002), as well as the internal sediment facies and properties of the regional Late Wisconsin (Marine Oxygen Isotope Stage (MIS) 2) till sheet. Tunnel valleys are a critical component of the subglacial hydrologic system, which is both a driver of, and is influenced by, glacier dynamics (Kehew et al., 2012; van der Vegt et al., 2012; Ravier et al., 2015). Understanding the regional architecture and local-scale facies variations of glacial sediments within this complex system will assist in understanding the transient and spatial variations processes related to subglacial meltwater production, storage, and discharge, and deciphering the interrelationships between subglacial drainage, ice dynamics and till deposition across an irregular and heterogeneous glacier bed. The insights gained from these analyses may shed light on investigations of subglacial tills, tunnel valleys, or subglacial bedform genesis along the peripheral regions of other modern or previously glaciated landscapes.

1.2. STUDY AREA

South-central Ontario is bordered to the south and west by the Great Lakes basins, and to the east and north by structural valleys occupied by the St. Lawrence and Ottawa rivers (Fig. 1.2). The northern part of south-central Ontario is dominated by the rugged topography of the Algonquin Highlands of the Canadian Shield (~300-600 m asl). To the south, the Shield rocks are unconformably overlain by flat-lying Paleozoic carbonate and clastic rocks that dip to the southwest (Armstrong and Carter, 2010). The bedrock surface is lowest within the Great Lakes

Fig. 1.2: Glacial landforms and major lithologic boundaries overlain on a combined topographic and bathymetric map of south-central Ontario and surrounding area. Note the prominent physiographic features including Shield highlands, overdeepened Great Lakes basins, and low-lying terrain occupied by the St. Lawrence and Ottawa rivers. Black boxes show the extent of study areas for individual chapters (CH 2-5) presented herein.



80° W
78° W
76° W
46° N
45° N
44° N
43° N
CH. 5
CH. 2, 4
CH. 3
Algonquin Highlands
Georgian Bay (Lake Huron)
Lake Simcoe
Ottawa River
Ottawa
St. Lawrence River
Adirondack Mountains
Lake Ontario
Toronto
Hamilton
Niagara Falls
Finger Lakes
Lake Erie
100 km
Elevation (m asl)
1259
-165
Shield-Paleozoic contact
Interlobate moraines
esker
moraine crest
streamlined bedform

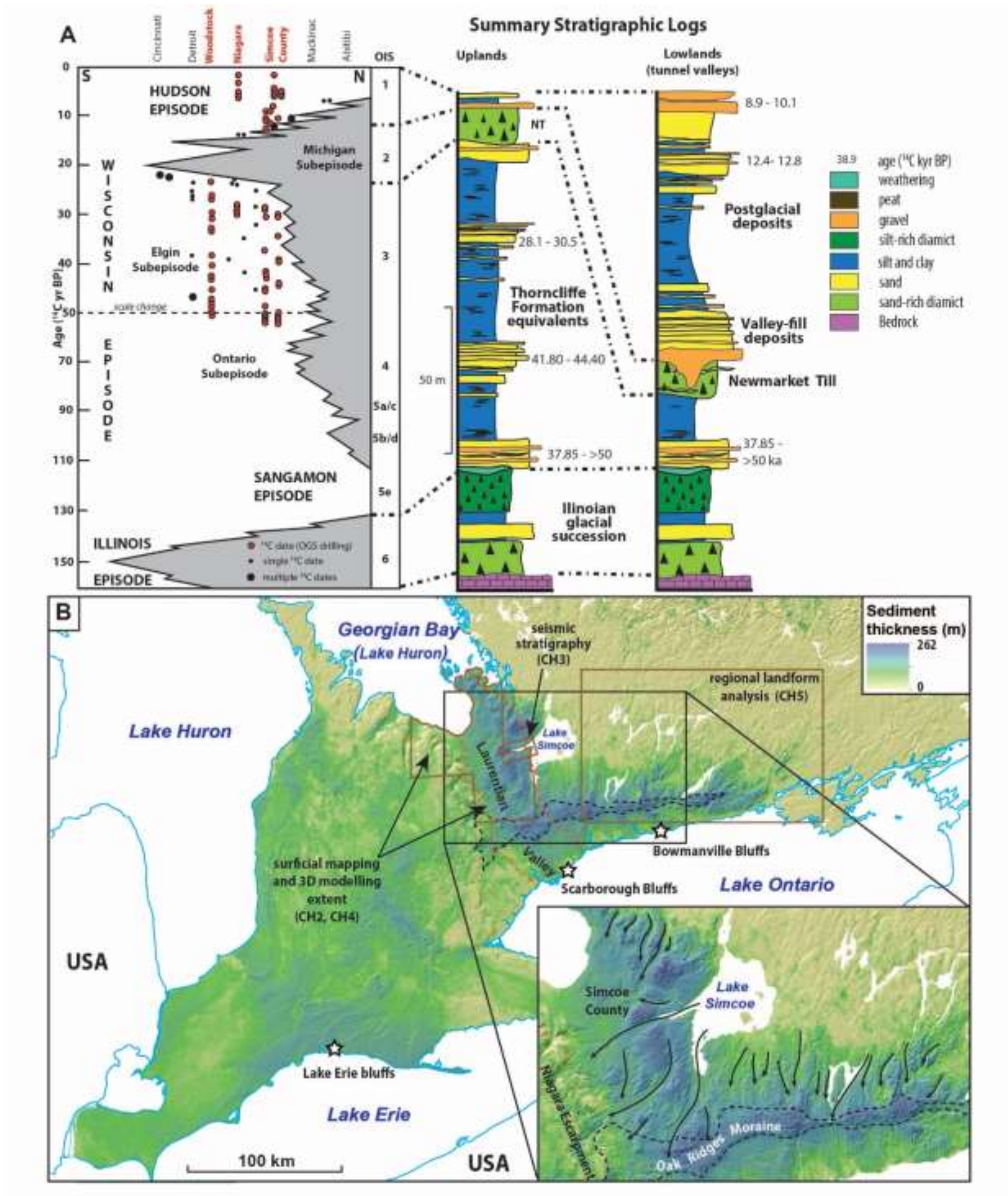
basins, which have been significantly overdeepened by erosion during multiple Quaternary glaciations. The Niagara Escarpment forms a regionally continuous scarp, punctuated by numerous parallel re-entrant valleys, along the western boundary of the study area (Fig. 1.2). The scarp has total relief of >350 m, but its topographic expression has been muted by the cover of glacial sediments (locally up to 262 m thick; Gao et al., 2006) that occur immediately to the east. Some of the thickest Quaternary sediment successions in southern Ontario overlie a broad, but poorly-defined, low on the bedrock surface that lies east of the Niagara Escarpment and connects the Georgian Bay and Lake Ontario basins (Laurentian Valley/Trough; Spencer, 1890; Eyles et al., 1985; Gao, 2011; Mulligan and Bajc, 2018; Sharpe et al., 2018; Fig. 1.3).

Quaternary sediments overlying bedrock in south-central Ontario span at least the last two glacial episodes (Eyles, 1987; Mulligan and Bajc, 2018; Fig. 1.3). Irregular deposition and localised erosion of sediments during the Late Wisconsin has produced the characteristic upland – lowland configuration that characterizes much of the landscape (Chapman and Putnam, 1984; Fig. 1.3). Recent surficial mapping (Mulligan and Bajc, 2012; Mulligan 2017a, b) and sediment drilling investigations (Bajc et al., 2015; Mulligan 2016; 2017c; 2018) have identified complex architectural relationships between upland and lowland regions. These relationships are complicated by thick accumulations of postglacial deposits in the southern part of the study area, where successions of sand, silt and gravel comprising the Oak Ridges Moraine (ORM) and thick (locally up to 80 m) successions of glaciolacustrine sediment bury the Late Wisconsin till sheet.

1.3. RATIONALE AND OBJECTIVES

This study investigates the stratigraphic and morphological relationships between Quaternary landforms and sediments observed within upland and channelized lowland areas in south-central Ontario. The existence of this characteristic upland-lowland landscape has been

Fig. 1.3: A) time-distance diagram (modified from Karrow et al., 2000) showing the ice limits and radiocarbon age determinations from organic material in Ontario (see Appendix for radiocarbon dates from Simcoe County). Generalized representative stratigraphic columns for uplands and lowlands (tunnel valleys) from deep sediment drilling in Simcoe County (Mulligan and Bajc, 2018), show typical sediment facies related to changes in ice front positions over the last two glacial cycles. B) sediment thickness map of southern Ontario (data from Gao et al., 2006) showing the location of the study areas (brown outlines) and the thick sediment infill of the Laurentian Valley/Trough (dashed red lines) and beneath the Oak Ridges Moraine (dashed black lines). Inset shows the Laurentian Valley/Trough area in more detail with black arrows denoting the axes (and inferred paleoflow directions) of tunnel valleys that produce the upland-lowland landscape across south-central Ontario.


A
Summary Stratigraphic Logs
Uplands
Lowlands (tunnel valleys)
HUDSON EPISODE
Michigan Subepisode
WISCONSIN EPISODE
Elgin Subepisode
Ontario Subepisode
SANGAMON EPISODE
ILLINOIS EPISODE
Thorncliffe Formation equivalents
Postglacial deposits
Valley-fill deposits
Newmarket Till
Illinoian glacial succession
NT
50 m
8.9 - 10.1
12.4- 12.8
28.1 - 30.5
41.80 - 44.40
37.85 - >50
37.85 - >50 ka
age (¹⁴C kyr BP)
weathering
peat
gravel
silt-rich diamict
silt and clay
sand
sand-rich diamict
Bedrock
B
Georgian Bay (Lake Huron)
seismic stratigraphy (CH3)
regional landform analysis (CH5)
Sediment thickness (m)
Lake Huron
Lake Simcoe
Laurentian Valley
surficial mapping and 3D modelling extent (CH2, CH4)
Bowmanville Bluffs
Scarborough Bluffs
Lake Ontario
USA
Lake Erie bluffs
Lake Erie
100 km
Simcoe County
Niagara Escarpment
Oak Ridges Moraine

known for decades, but the underlying processes responsible for its formation have remained elusive. Lowland regions were originally viewed as areas of focussed erosion by fast-flowing and/or high fluxes of glacier ice (Straw, 1968), but were later interpreted as tunnel channels (Barnett, 1990) that form part of a broader integrated, anabranching network stretching from the Niagara Escarpment to Kingston (Brennand and Shaw, 1994; Sharpe et al., 2004; Fig 1.3).

The two conflicting explanations for the upland-lowland landscape have led to drastically different predictions of subsurface sediment type and architecture. These differences significantly affect any resultant paleoglaciological reconstruction efforts or modern groundwater flow modelling. Given the large existing population in south-central Ontario, its forecasted growth, and the high demand on groundwater for municipal, domestic and agricultural uses, a thorough understanding of regional architecture, properties, and hydrogeologic characteristics of subsurface sediments in the region is essential to effectively and safely develop growth plans and to make informed land use decisions. Recent studies have highlighted the dominant sediment properties, architecture and groundwater characteristics of older strata coring regional uplands (Mulligan and Bajc, 2018) and within the uppermost strata infilling the lowland valleys (Mulligan et al., 2018a); however, neither of these investigations describe deeply-buried sediments within the valleys, or their relationship with upland sediments or the groundwater hosted within them

Integrated analyses of multiple subsurface data sets provide the most thorough understanding of the subsurface geometry, connectivity and hydrogeological function of late Quaternary sediments found in upland and lowland settings in south-central Ontario. This thesis summarizes and interprets high-quality surficial, subsurface, and geophysical data in order to provide a thorough analysis of the local-scale composition of sediments and landforms and evaluate their significance within regional-scale landform-sediment assemblages. This methodology permits detailed understanding of local-scale heterogeneity, which can be used to gain insight into processes operating at local scales and allows for integration with regional data that record the broader systems representative of former ice lobes/streams.

In particular, this thesis aims to provide a detailed understanding of Late Wisconsin subglacial processes and the evolution of tunnel valleys found in lowland regions in south-central Ontario, through analysis of extensive surficial outcrops, 58 continuously-cored boreholes, and high-resolution sub-bottom profile data from the submerged parts of one of the tunnel valleys. These data can be used to characterize the substrate sediments and bedrock, composition of the regional till sheet (the Newmarket Till), tunnel valley infill sediment architecture, and interregional landform relationships. Developing an improved understanding of the tunnel valleys, including their three-dimensional geometry and the processes responsible for their formation, such as glacial sediment erosion, transport, and deposition, assists in developing an improved understanding of subglacial dynamics and is essential to reliably discover, characterize, and efficiently and safely manage groundwater resources found in such settings.

1.4. BACKGROUND INFORMATION

This thesis describes sediments and landforms that developed subglacially during the Late Wisconsin and were either buried or modified by ice-marginal and proglacial processes during deglaciation of the region. Below is a brief outline of some key concepts relating to the geologic history of southern Ontario during the build-up to, and the retreat from, the last glacial maximum, the nature of subglacial processes and subglacial till production, and the distribution, characteristics, and evolution of tunnel valley systems across formerly glaciated landscapes.

1.4.1. Glacial history of southern Ontario

Sediments in southern Ontario are attributed to (at least) the last two glacial-interglacial episodes (Bajc et al., 2014; Mulligan, 2016; Mulligan and Bajc, 2018; Fig. 1.3). The sedimentary succession deposited during this time consists of a lower (Marine Oxygen Isotope Stage (MIS) 6 and/or older) glacial package comprising two tills and interbedded glaciolacustrine deposits, with a regional subaerial unconformity developed on a deeply weathered surface of the upper till (Fig. 1.3). The unconformity is overlain by non-glacial alluvial, lacustrine, and peat deposits that span the Sangamon (MIS 5) through Middle Wisconsin (MIS 3) Episodes (Mulligan and Bajc 2018). These older deposits are buried by thick successions of glaciolacustrine rhythmites with interbedded sand bodies bearing detrital organic remains that record progressive flooding of the landscape during advance of the Laurentide Ice Sheet (LIS) beginning in the Early Wisconsin (MIS 4; Eyles et al., 1985; Mulligan and Bajc, 2018). The pre-Late Wisconsin sediments observed in the study area correlate well with well-studied sediment exposures along the Scarborough (Karrow, 1967; Eyles and Eyles, 1983), Bowmanville (Brookfield et al., 1982), and Lake Erie bluffs (Dreimanis, 1992; Fig. 1.3).

Capping these pre-Late Wisconsin sediments is the Newmarket Till, that records Late Wisconsin (MIS 2) ice advance into south-central Ontario and southward into the northern United States (Fig. 1.3). The surface of the Newmarket Till is highly undulating and drumlinized and the ice advance responsible for its deposition also contributed to significant erosion of earlier sediments and Paleozoic bedrock across broad areas of southern Ontario. The large lowland valleys separating broad drumlinized uplands on the modern landscape were originally attributed to enhanced lateglacial erosion beneath fast-flowing zones in the LIS (Straw, 1968). However, analysis of borehole data and geophysical surveys carried out in the region led to the identification of truncations of the Newmarket Till in the valleys (Barnett, 1990), which were later attributed to erosion during regional subglacial sheet floods (Sharpe et al., 2004; 2018; Sharpe and Russell, 2016).

In the southernmost part of the study area, the Newmarket Till is deeply buried by sand, silt and gravel of the ORM, which was deposited during initial break-up of the LIS into semi-independent ice lobes flowing out of the major lake basins in southern Ontario (Barnett et al., 1998). Sediments in the ORM have been interpreted to record the distal and late phases of subglacial outburst floods that eroded the tunnel channel network across southern Ontario then filled the valleys and constructed the core of the ORM (Brennand and Shaw, 1994; Sharpe et al., 2002; 2004; 2018; Sharpe and Russell 2016). Other studies interpret the ORM as a composite set of stacked subaquatic fans and delta complexes deposited in a growing interlobate lake fed by meltwater from ice lobes/streams occupying the areas to the north and south (Chapman and Putnam, 1984; Chapman, 1985; Howard et al., 1995; Barnett et al., 1998; Maclachlan and Eyles, 2013; Mulligan et al., 2016; Sookhan et al., 2018; Eyles et al., 2018).

Relatively minor post-glacial modification of the landscape has taken place in the study area, although the flanks of uplands are commonly gullied and in some locations, particularly on northwest-facing slopes, wave erosion in deglacial lakes has locally eroded the Newmarket Till, redepositing it as sand and gravel comprising large spits (Mulligan, 2014; 2017a,b; Schaetzl et al., 2016; Mulligan et al., 2018a). In other areas, groundwater piping has created large erosional amphitheatres up to a few kilometers wide and >60 m deep along the flanks of uplands where groundwater is discharged through coarse-grained sediments that underlie the Newmarket Till and overlie thick fine-grained rhythmite successions (Bajc and Rainsford, 2010; Mulligan, 2013; Mulligan et al., 2018b).

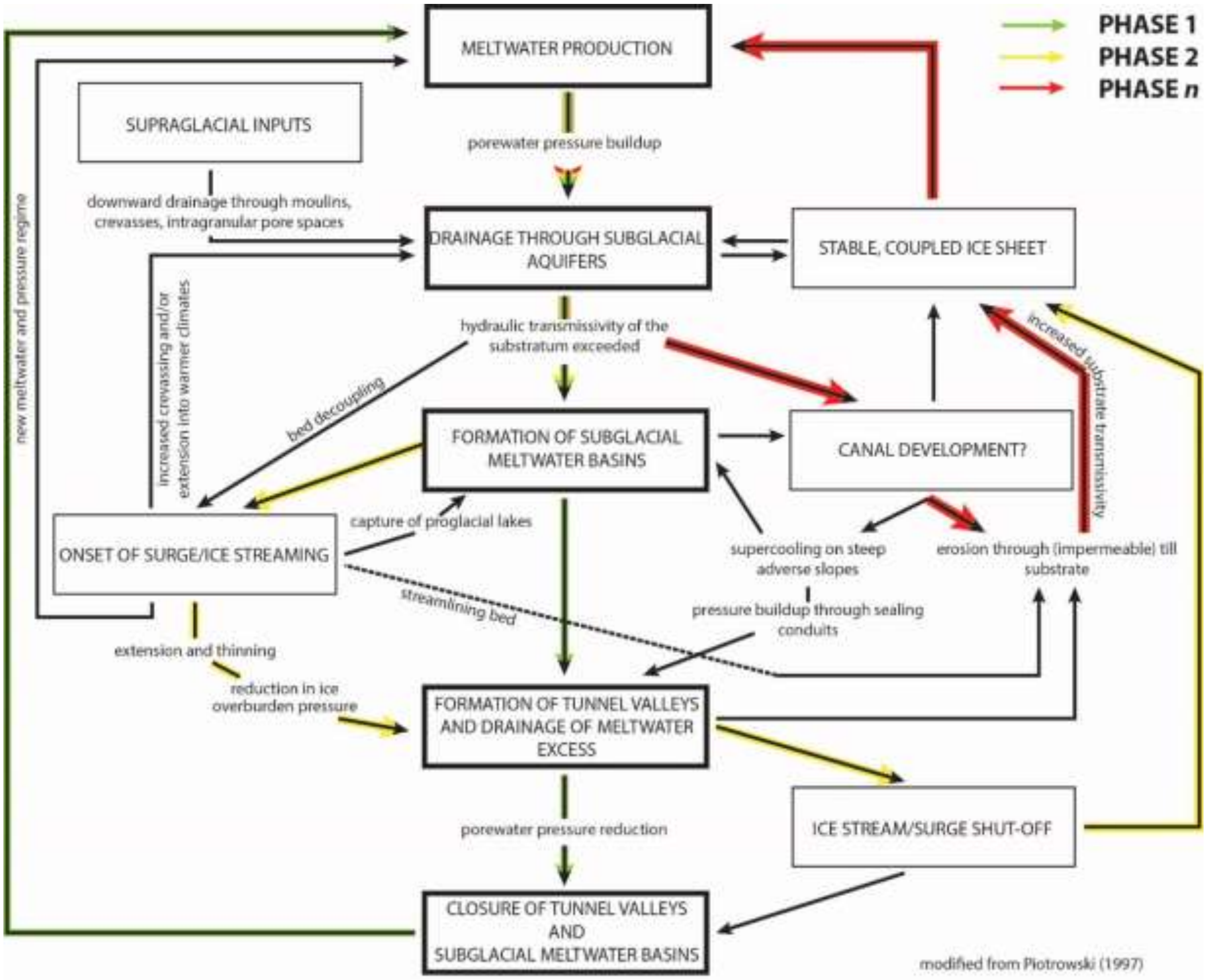
1.4.2. Subglacial processes and till deposition

For such seemingly simple features – large masses of ice that flows downhill when the force of gravity exceeds the internal strength of the ice or frictional resistance of the substrate – glaciers are remarkably complex. Actual observations of glaciers and glacial deposits identify the spatial and temporal complexity and variability of processes at work within the ice, at the interface between the ice and its bed, and within the substrate materials, regardless of whether it is rock, till, stratified sediments, or water. Englacial and subglacial processes are highly dynamic and governed largely by feedback-driven responses to internal and external forcings (Fig. 1.4).

Subglacial till accretes either top-down, through increasing deformation and incorporation of substrate material into a mobile deforming bed, or bottom-up, by melt-out of englacial debris and deposition onto a stiff substrate (e.g. Benn and Evans, 1996; Evans et al., 2006; Menzies et al., 2006). Multiple 'end-member classifications' of subglacial till exist: subglacial traction till (includes deformation and lodgement tills), glaciotectonite, and melt-out till, though in reality, almost all tills are hybrids (Evans et al., 2006). The formation of subglacial till is intimately linked to ice dynamics, hydrology, substrate character and rheology, and paleogeographic setting – all of which undergo significant changes during different phases of glaciation.

In south-central Ontario, for instance, the character and nature of deposition of till sheets beneath a regional ice cover was affected by: the presence of extensive proglacial lakes that occupied much of southern Ontario during advance and retreat phases of the LIS, the existence of overdeepened Great Lakes basins and prominent bedrock escarpments, sharp changes in bedrock lithology (crystalline Shield rocks, limestone, shale, dolostone, evaporate), the direction of ice flow relative to pre-existing bedrock structures and valleys, ice thickness, flow velocity,

Fig. 1.4: conceptual flow model of the evolution of processes and feedbacks that affect ice flow dynamics, meltwater production and drainage, and substrate rheology (bold black boxes are from Piotrowski (1997). Multiple phases of ice flow can occur through a single period of ice cover. In the hypothetical situation depicted above, basal meltwater production during an early phase of ice advance (phase 1) exceeded drainage capacity of the substrate, storing water in subglacial basins until fluid overpressure resulted in N-channel formation leading to tunnel valley genesis until the substrate is able to effectively drain meltwater again. During a subsequent period of meltwater excess (Phase 2), the development of subglacial meltwater basins triggers a surge or streaming phase of ice flow, resulting in extension and thinning of the ice and reduction of ice overburden pressure over the subglacial meltwater reservoir, until meltwater drainage via tunnel valleys reduces porewater pressures, increases bed coupling, and ends the life cycle of the ice stream. These first two phases of ice flow significantly alter the nature and distribution of substrate sediments during future phases of ice flow. If a shift towards higher bulk transmissivity of substrate occurs, development of meltwater basins and large-scale fluid overpressures may be inhibited, leading to a more stable and strongly coupled ice sheet.


PHASE 1
PHASE 2
PHASE n
MELTWATER PRODUCTION
SUPRAGLACIAL INPUTS
porewater pressure buildup
downward drainage through moulins, crevasses, intragranular pore spaces
DRAINAGE THROUGH SUBGLACIAL AQUIFERS
STABLE, COUPLED ICE SHEET
new meltwater and pressure regime
increased crevassing and/or extension into warmer climates
hydraulic transmissivity of the substratum exceeded
bed decoupling
increased substrate transmissivity
FORMATION OF SUBGLACIAL MELTWATER BASINS
CANAL DEVELOPMENT?
capture of proglacial lakes
ONSET OF SURGE/ICE STREAMING
supercooling on steep adverse slopes
erosion through (impermeable) till substrate
streamlining bed
pressure buildup through sealing conduits
extension and thinning
reduction in ice overburden pressure
FORMATION OF TUNNEL VALLEYS AND DRAINAGE OF MELTWATER EXCESS
porewater pressure reduction
ICE STREAM/SURGE SHUT-OFF
CLOSURE OF TUNNEL VALLEYS AND SUBGLACIAL MELTWATER BASINS
modified from Piotrowski (1997)

basal temperature, subglacial meltwater generation and drainage style, ice flow switching and the onset of ice streams, and the existence of extensive proglacial lakes fronting the ice sheet during deglaciation.

In general, fine-grained substrates (such as those that accumulated in the former lake basins fronting the advancing Laurentide Ice Sheet) do not permit adequate drainage of meltwater at the ice-bed-interface, resulting in increasing porewater pressures in subglacial sediments. This has the effect of reducing the effective pressure, which reduces grain-to-grain contacts and thus the shear strength of the sediment, leading to ductile failure and flow of the sediment beneath the ice (Benn and Evans, 1996; Roberts and Hart, 2005). This mobile sediment layer is commonly referred to as the soft deforming bed, and it plays an important role in contributing to enhanced ice flow velocities by creating a decollement within subglacial sediments, where a thin, saturated and dilatant upper mobile layer (A-Horizon; Boulton and Hindmarsh, 1987) is actively moving with the ice, over a more rigid bed below (B-Horizon; Boulton and Hindmarsh, 1987).

Coarser-grained substrates (such as fractured or karst carbonate bedrock, outwash systems or subaquatic fans) promote much different responses during ice advance, largely due to their higher bulk permeability, which enables enhanced drainage of meltwater at the bed, leading to reduced porewater pressures and increased effective pressure. Coarse beds transfer applied normal stress deeper into the bed and, combined with a higher shear strength compared to fine-grained substrates, promote a strongly coupled ice-bed-interface (Passchier et al., 2010).

As till sheets accrete, the influence of the underlying substrate deposits is diminished. The character of the resultant traction till/glaciotectonite succession and delivery of new material to the ice-bed interface begin to influence till deposition and meltwater processes operating at the bed. As even relatively coarse-grained tills thicken, the ability of water at the bed to penetrate into subglacial aquifers is diminished, and surpluses of water can build up beneath the ice. These require drainage through more efficient mechanisms than Darcian flow, via distributed canal systems (Walder and Fowler, 1994; Ng, 2000) or channels (Roethlisberger, 1972; Shreve, 1985; Piotrowski, 1997), to prevent decoupling of the ice and bed, which may trigger an outburst flood (Russell et al., 2007) and/or surge (Kamb et al., 1985; Fig. 1.4). The build-up and discharge of subglacial porewater pressures plays a critical role in the dynamics of overlying ice. As rising porewater pressures contribute to instability, evolving drainage mechanisms must keep pace; one of the most efficient means to drain large quantities of meltwater and reduce porewater pressures is the development of tunnel valleys and/or tunnel channels (Kehew et al., 2012).

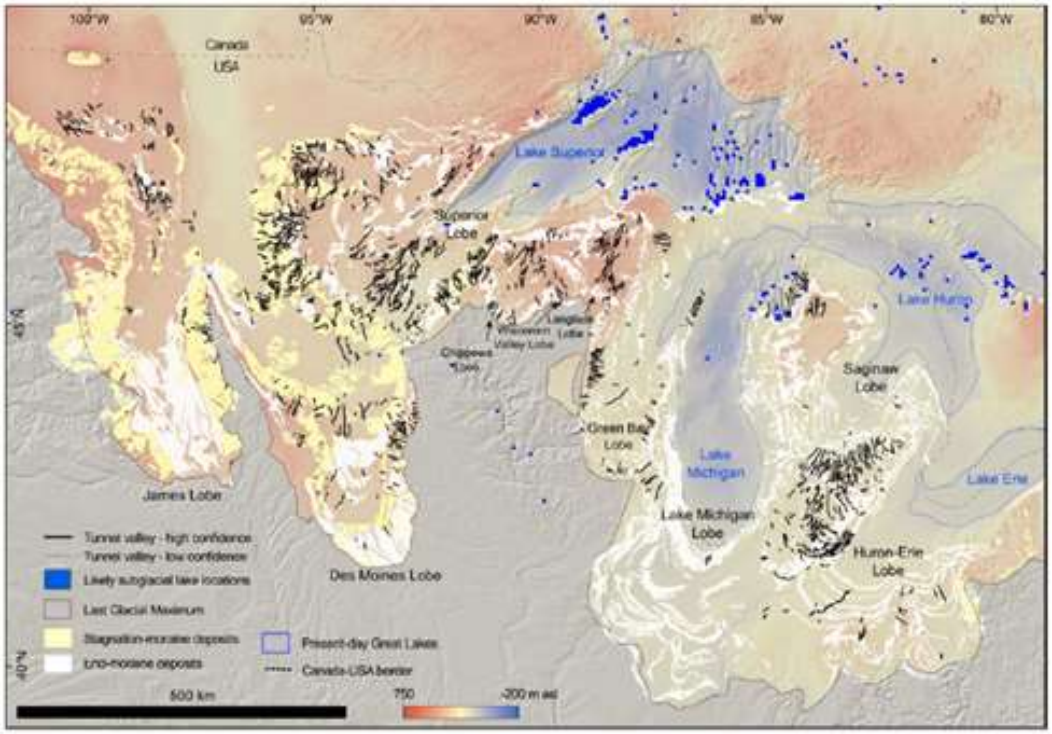
1.4.3. Tunnel valley formation and landform-sediment associations

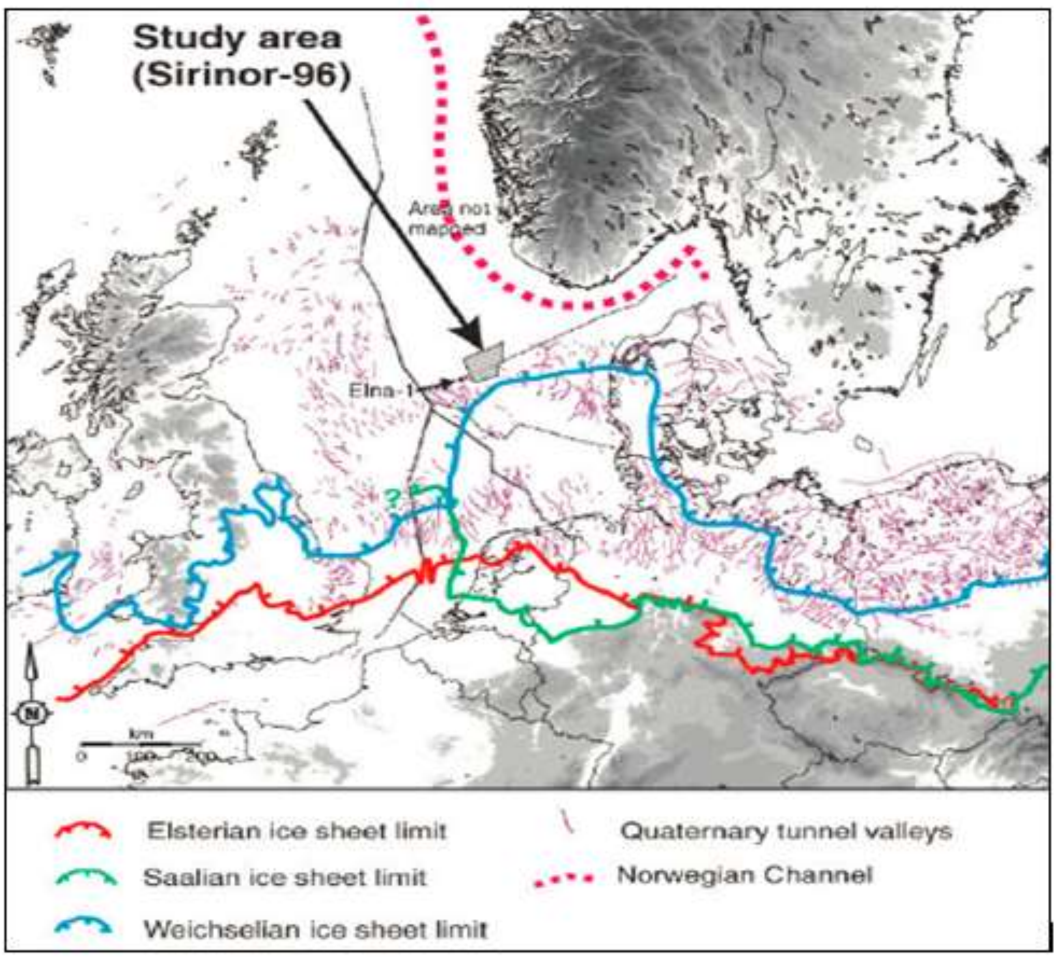
1.4.3.1. History and diagnostic criteria

Elongate deep incisions cut into glacial sediments have been documented since the 19th century (Jentzsch, 1884; Ussing, 1903). Early theories attributed them largely to subglacial meltwater erosion, but their wider occurrence in a variety of settings across broad areas of NW Europe in terrestrial and offshore settings has led to a wide variety of subsequent interpretations of valley genesis (see review by Huuse and Lykke-Anderesen, 2000).

Tunnel valleys are common features in glacial landscapes (Fig. 1.5) and a set of diagnostic criteria has been developed to identify and differentiate them from incisions or channels carved by other processes. Tunnel valleys are 10-100’s of km long, 250-5000 m wide, and 10-s to 100’s of m deep; they are aligned roughly parallel to former ice flow directions and commonly terminate at former ice margins with undulating, adverse, or convex-up long profiles; and they may contain eskers or other subglacial landforms. The term tunnel valley describes all

Figure 1.5: Distribution of tunnel valleys mapped in a) the Great Lakes Region of the USA (Livingstone and Clark, 2016); and b) northwest Europe (Kristensen et al., 2008). Note that the European map shows the distribution of valleys from (at least) the last three glaciations (MIS 2, 6, 12).


100°W
95°W
90°W
85°W
80°W
Canada
USA
Lake Superior
Superior Lobe
Lake Huron
Saginaw Lobe
Green Bay Lobe
Lake Michigan
Lake Michigan Lobe
Lake Erie
Huron-Erie Lobe
James Lobe
Des Moines Lobe
Tunnel valley - high confidence
Tunnel valley - low confidence
Last Glacial Maximum
Present-day Great Lakes
Canada-USA border
500 km
750
-200 m asl


Study area
(Sirinor-96)
Area not mapped
Elna-1
N
km
0
Elsterian ice sheet limit
Saalian ice sheet limit
Weichselian ice sheet limit
Quaternary tunnel valleys
Norwegian Channel

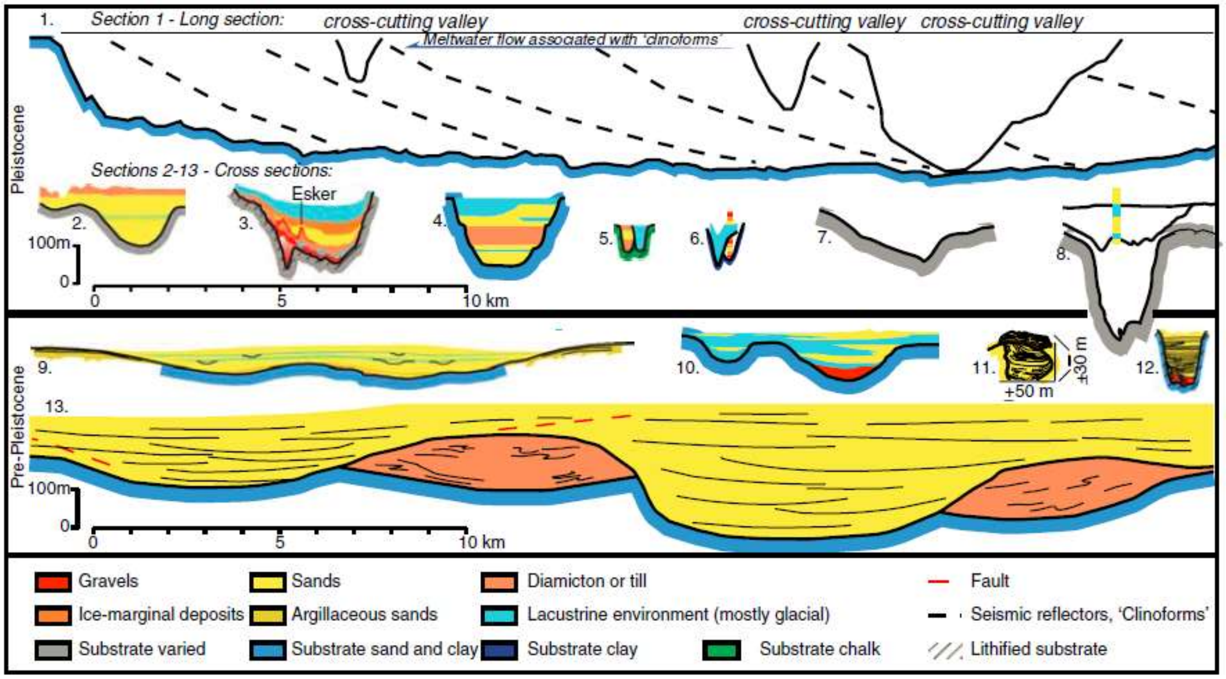
erosional features attributed to subglacial meltwater erosion into the bed, but where it can be identified that the valleys were formed through bankfull flow conditions (the cross-sectional area of the valley ≈ the cross-sectional area of the meltwater flow), they are termed tunnel channels (Clayton et al., 1999).

1.4.3.2. Morphology

Tunnel valleys observed in formerly glaciated terrains are much larger than modern analogues (Russell et al., 2007) and display a wide variety of forms, but there is a consistent trend that longer valleys tend also to be wider (Livingstone and Clark, 2016). Widths of valleys range up to 5 km, with very rare exceptions exceeding this (O'Cofaigh, 1996; Huuse and Lykke-Andersen, 2000; Kehew et al., 2012). Many have argued that valleys exceeding ≈ 5 km in width likely experienced enhanced direct glacial erosion which attributed to their enhanced widths (Van der Vegt, 2012; Mulligan et al., 2018b). Depth of erosion in tunnel valleys is highly variable and is likely influenced by substrate geology (resistance to erosion; Janszen et al., 2012), duration (and repetition) of ice cover and flow routing (Piotrowski, 1994), availability of meltwater and magnitude of discharges (Passchier et al., 2010), and the possible reoccupation of the valley during successive glaciation(s) (Jorgensen and Sandersen, 2006).

The cross-sectional form of tunnel valleys can take a variety of shapes, ranging from v-shaped, to U-shaped or parabolic, flat-floored, or complex (Huuse and Lykke-Andersen, 2000; Kristensen et al., 2008; Van der Vegt et al., 2012). Complex valley profiles generally have a broad, irregular cross-sectional form, with smaller-scale features superimposed, such as multiple v-shaped incisions along the floor, broad terraces along the flank/shoulder, or asymmetric geometry with one parabolic side and the other V-shaped or composite (Huuse and Lykke-Andersen, 2000; Kristensen et al., 2008; Ravier et al., 2015). Longitudinal sections commonly

Fig. 1.6: Summary of cross-sectional profile morphology of tunnel valleys from Pleistocene and Paleozoic examples worldwide (from Van der Vegt et al., 2012).



1. Section 1 - Long section:
cross-cutting valley
Meltwater flow associated with 'clinoforms'
cross-cutting valley
cross-cutting valley
Pleistocene
Sections 2-13 - Cross sections:
Esker
100m
0
0
5
10 km
Pre-Pleistocene
±30 m
±50 m
100m
0
0
5
10 km
Gravels
Ice-marginal deposits
Substrate varied
Sands
Argillaceous sands
Substrate sand and clay
Diamicton or till
Lacustrine environment (mostly glacial)
Substrate clay
Substrate chalk
Fault
Seismic reflectors, 'Clinoforms'
Lithified substrate

show complex gradients, which may be adverse, convex-upward, undulating, or marked by a series of significantly overdeepened (>100m) reaches (Rains et al., 2002; Fisher et al., 2005; Kristensen et al., 2008; Stewart et al., 2011; Kehew et al., 2013; Fig. 1.6).

Tunnel valleys may be observed in isolation, but are more commonly observed as part of a broader network of similar features (Fig. 1.5). They are generally oriented nearly parallel to former ice flow directions, controlled by net subglacial gradient (ice surface gradient – bed gradient) and substrate geology, and terminate at end/recessional moraines or outwash fans (Ussing, 1903; Sandersen et al., 2009; Livingstone and Clark, 2016; Fig.1.5). Tunnel valleys commonly form straight to slightly sinuous networks orthogonal to moraine crests and may diverge or converge, depending on the configuration of relatively thin (<1 km) ice lobes (Boulton et al., 2009; Sandersen et al., 2009; Kehew et al., 2013; Livingstone and Clark, 2016). In some areas, a relatively consistent spacing of valleys is observed (Kehew et al., 2013), although this spacing varies between geographic regions (Livingstone and Clark, 2016).

The best examples of tunnel valley network morphology come from 3D seismic surveys in the North Sea (e.g. Praeg, 2003). In this region, a complex network of hundreds of cross-cutting valleys is observed, with highly sinuous paths (Huuse and Lykke-Andersen, 2000; Kristensen et al., 2008; Stewart et al., 2011; Van der Vegt et al., 2012). Early work had interpreted an anabranching network of valleys, but improved resolution and further data acquisition has led to the identification of >7 generations of valleys in the region, relating to at least three distinct episodes of ice cover (MIS 2, MIS 6, MIS 12; Kristensen et al., 2008; Stewart et al., 2013). Improved visualization has greatly enhanced understanding of the tunnel valley networks throughout the region; this in turn has permitted more detailed reconstruction of past glacial history and hydrological processes. Though the 3D seismic investigations provide valuable insight into the morphology of entire tunnel valley networks, there is comparatively little data on the sediment infill within the valleys.

1.4.3.3. Infill sediments

Tunnel valleys may be completely buried or partially under-filled. Sediments infilling tunnel valleys are rarely exposed – under-filled valleys commonly host lakes and/or misfit streams, and buried valleys have no surface expression. Where available, surficial exposures generally reveal only the uppermost sediments within the valleys, which record the latest stages of their infilling (Eyles and McCabe, 1989; Piotrowski et al., 1999; Russell et al., 2003; Korus et al., 2016). In a few cases, large outwash fans consisting of boulder-rich gravels at the downflow ends of tunnel valleys have been attributed to deposition at the terminal end of flows associated with tunnel channel excavation in terrestrial settings (Cutler et al., 2002; Fisher et al., 2005).

Pre-Pleistocene tunnel valleys have also been identified in Neoproterozoic and Ordovician strata (Beuf et al., 1971; Eyles and de Broekert, 2001; Le Heron et al., 2005; Ravier et al., 2015). In these examples, uplift of the strata has locally exposed the entire valley form, permitting analysis of entire successions, from substrate to primary tunnel valley fill and postglacial sedimentation. Many studies on pre-Pleistocene tunnel valleys show fluted surfaces on the tunnel valley walls, a general absence of diamicts and/or tills, and the common observation that the majority of the fills consist of proglacial sediments (Van der Vegt et al., 2012 and references therein). Recent work has begun to relate the form and sedimentary infill (including depositional processes) to the position of the valley relative to ice stream activity, implying a significant control on valley development exerted by local glaciological factors (Ravier et al., 2015).

Though exposures through Quaternary tunnel valley sediments are rare, an increasing number of studies are investigating tunnel valley infill stratigraphy through continuously-cored boreholes, sometimes complemented by geophysical investigations (Russell et al., 2003; Jorgensen and Sandersen, 2006; Bajc et al., 2014; Mulligan et al., 2018b). Through detailed facies analysis, reconstructions of paleoenvironments and flow histories are possible (Russell et al., 2003), and geophysical studies (seismic reflection and electromagnetic surveys) can greatly assist in constraining sediment architecture within the valleys (Pugin et al., 1999; Korus et al., 2016). One problem with studies of borehole sediment successions is that, due largely to the prohibitive costs of sediment drilling, tunnel valleys are commonly only represented by a single borehole, rendering insufficient assessment of lateral facies variations and sediment architecture (e.g. Kehew et al., 2013).

Many studies identify coarse-grained sediments near the base of the valleys, and finer sediments toward the upper parts of the succession (Russell et al., 2003; Kehew et al., 2013). Repeated meltwater discharges can lead to complex infill successions through successive episodes of erosion and deposition (Jorgensen and Sandersen, 2006). Generally, till and diamicts comprise a small proportion of the infill sediments, except in rare cases and/or older valleys that were re-occupied during later glaciations (Piotrowski, 1994; Jorgensen and Sandersen, 2006; Cummings et al., 2012; Van der Vegt et al., 2012; Mulligan et al., 2018b). Sediments within tunnel valleys are generally deposited during ice retreat, recording subglacial and ice-proximal conditions at the base, passing upwards into proglacial (glaciofluvial, -lacustrine, -marine) conditions in the upper parts of the succession.

1.4.3.4. Seismic data

Through the collection of a wealth of high-resolution seismic data, the internal architecture of tunnel valleys has been subjected to increasingly more detailed study (Pugin et al., 1999; Huuse and Lykke-Andersen, 2000; Praeg, 2003; Kristensen et al., 2008; Ottesen et al., 2016). Generally, the seismic facies identified within tunnel valley infills are consistent with those interpreted from boreholes and sedimentological studies. Seismic profiles across tunnel valleys commonly reveal a lowermost unit of chaotic reflectors overlain by well-bedded sub-horizontal reflectors, capped by a drape of parallel, horizontal, low-amplitude or transparent facies (Huuse and Lykke-Andersen, 2000; Janszen et al., 2012). This common seismic stratigraphic succession is in general agreement with the facies geometry expected for transitions from subglacial to proglacial sedimentation. An exception to this generalization is the observation of large clinoforms (up to 12 km long) dipping up-ice, that form a large proportion of the valley infills (Praeg, 2003; Kristensen et al., 2008). Seismic investigations coupled with EM surveys and borehole control also reveal locally complex cut-and-fill geometries with highly variable internal sediment units (Jorgensen and Sandersen, 2006).

1.4.3.5. Terrestrial vs subaquatic settings

The majority of seismic data covering tunnel valleys are from subaquatic settings (glaciolacustrine or glaciomarine). Because sediments routed through the valleys debouched at the base of a waterbody, their architecture and infill successions were governed by different controls than those in subaerial settings. Tunnel valleys identified in terrestrial settings in the southern Great Lakes region of the USA contain thick successions of sand and gravel glaciofluvial sediments, and locally pass downflow from subglacial tunnel valleys into proglacial spillway features (Kehew et al., 2013). Tunnel valleys in the ORM area of southern Ontario bear similar infill characteristics (Russell et al., 2003; Sharpe et al., 2013), but their infill sediments have a much more complex architecture, related to multiple discharge events of varying magnitudes, and deposition into standing water in a growing interlobate lake with fluctuating regional base levels during their development (Barnett et al., 1998).

Deposition of tunnel valley/channel sediments into a standing water body limits the downflow sediment transport capacity for coarse-grained sediments, enhances the potential for deposition of muds, and may promote enhanced calving, and increased rates of ice retreat; all of these factors may lead to more localized distributions of coarse-grained sediments within the valleys (Janszen et al., 2013; Mulligan et al., 2018b).

1.4.4. Interpretations of tunnel valley genesis

The genesis of tunnel valleys has been debated since their first identification, more than one hundred years ago (Ussing, 1903), although their formation is generally ascribed to erosion primarily by subglacial meltwater, with subsequent modification or enhancement by direct glacial erosion in some cases (O'Cofaigh, 1996; Huuse and Lykke-Andersen, 2000; Kehew et al., 2012; Van der Vegt et al., 2012; Livingstone and Clark, 2016). Central to the remaining questions on tunnel valley development are considerations regarding the mode of meltwater erosion, particularly the magnitude and duration of meltwater discharges (*cf.* Brennand and Shaw, 1994 vs. Boulton et al. 2009, vs. Jorgensen and Sandersen, 2006), and the significance of direct glacial erosion to the final form of the valleys (Huuse and Lykke-Andersen, 2000). Conceptualizations of tunnel valley genesis generally fall into three categories: steady-state, time-transgressive, or catastrophic.

1.4.4.1 Steady-state

Theories of steady-state erosion invoke subglacial porewater pressure gradients as a primary control on valley formation. As meltwater generated at (or delivered to) the subglacial bed begins to exceed the capacity of the substrate to drain it through Darcian flow, distributed or channelized drainage networks develop (Nye, 1976; Kamb et al., 1985; Ng, 2000). If channels reach the glacier margin, they create low-pressure pathways for meltwater drainage, which begin to divert porewaters toward the channels as the path of least resistance for meltwater evacuation at the ice bed (Boulton et al., 2009). Channels enlarge and erode upglacier and laterally, sapping progressively more groundwater into them; as high-pressure groundwater seeps into the low-pressure channels, sediments are carried with them, then eroded downflow to the terminus (Fig. 1.7).

The steady state model for tunnel valley formation relies on soft, saturated substrates and high porewater pressures in subglacial sediments (Nye, 1976). Support for the steady-state model comes from field instrumentation of glacier forefields (Engelhart et al., 1978; Boulton et al., 2009), identification of deformation tills within small tunnel valleys in Germany (Piotrowski et al., 1999), interpretation of high porewater pressures based on deformation structures observed in substrate sediments beneath Ordovician tunnel valleys in Morocco (Ravier et al., 2015), and till-floored valleys overlying highly deformed sediments in southern Ontario (Chapter 2). Criticisms of this theory arise from the modelling of channel stabilization due to till and sediment creep into the developing low (e.g. Alley, 1992) and the observations of tunnel valleys incised into lithified substrates (which are resistant to piping) and boulder gravels at the downflow ends of valleys, interpreted to record high-magnitude discharges (Wright, 1973; O'Cofaigh, 1996; Cutler et al., 2002; Kehew et al., 2013). It should be noted that once a valley is established, any subsequent surpluses of meltwater at the IBI (through percolation of supraglacial meltwater in crevasses or moulins) would likely follow pre-existing pathways, and could increase discharge by several orders of magnitude, which could erode the substrate and produce boulder accumulations at the ice margin.

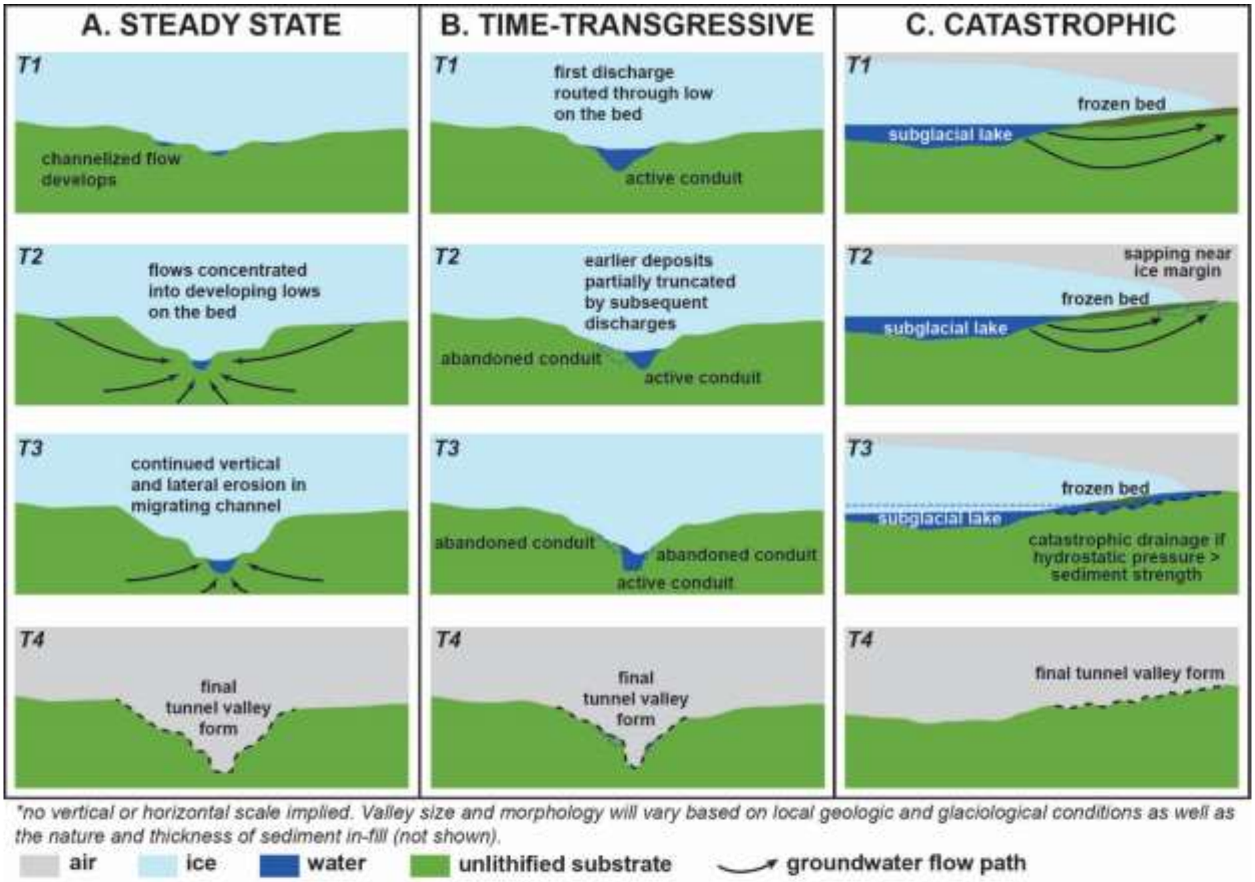
1.4.4.2. Time-transgressive

Seismic data and boreholes revealing complex architectures and highly variable sediment facies and valley characteristics point to a mechanism of valley formation involving repeated meltwater discharges. In this view, small channels are eroded by meltwater, then filled with ice, before a later flow erodes the substrate further (Fig. 1.7). The overall form of the valley is inferred to be much greater that the magnitude of individual flows that formed it (Huuse and Lykke-Andersen, 2000; Jorgensen and Sandersen, 2006; Sandersen et al., 2009). An emerging theory invokes simultaneous erosion and backfilling of valleys due to glaciohydraulic supercooling at the downflow termination of valleys to explain the presence of long (>10 km) ‘clinoforms’ dipping upglacier in some parts of the North Sea (Kristensen et al., 2008).

1.4.4.3. Catastrophic

Large, flat-floored valleys with consistent width and orientations and with boulder accumulations at their downflow ends are generally interpreted to have been eroded by catastrophic discharge(s) (Wright, 1973; Clayton et al., 1999; Cutler et al., 2002). A catastrophic origin of these features is more convincing if they can be related to an upflow reservoir (subglacial or proglacial), or if there is strong evidence of permafrost in the region (Hooke and Jennings, 2006; Van der Vegt et al., 2012; Fig. 1.7c). Some authors have extended the catastrophic hypothesis to suggest that entire networks of valleys were active simultaneously, as part of an integrated, anabranching tunnel channel system (Brennand and Shaw, 1994; Fisher et al., 2005; Sharpe et al., 2018).

Fig. 1.7. Simplified conceptual models of tunnel valley/channel development through time (*T1-T4*). A) Once the hydraulic transmissivity of the substrate is exceeded, meltwater is conveyed into channels/canals/conduits. These progressively sap sediment and meltwater from adjacent areas in the bed. Continued deformation of sediment into channels and flushing towards the margin promotes tunnel valley growth as the channel migrates. B) meltwater excesses are drained by channels eroded into the bed, which may be subsequently infilled with sediment during waning flow stages. Ice creeps into the channel during periods of low flow. Subsequent flows create new channels that enhance and/or truncate previous deposits. The final valley form is the sum of erosion and deposition from all discharge events. C) Catastrophic erosion of tunnel channels is most likely when there is potential connection to a significant meltwater reservoir, and when a physical barrier (such as bedrock or frozen sediment) is present, enabling growth of the reservoir and increase in hydrostatic pressures. Slow groundwater sapping initiated at the ice margin erode headwards beneath the margin and can progressively increase the pressure gradient between reservoir and proglacial systems, triggering catastrophic failure of the reservoir seal. Note that the panels for catastrophic drainage show a longitudinal section, whereas cross-sections are shown for steady state and time-transgressive.


A. STEADY STATE
T1
channelized flow develops
T2
flows concentrated into developing lows on the bed
T3
continued vertical and lateral erosion in migrating channel
T4
final tunnel valley form
B. TIME-TRANSGRESSIVE
T1
first discharge routed through low on the bed
active conduit
T2
earlier deposits partially truncated by subsequent discharges
abandoned conduit
active conduit
T3
abandoned conduit
abandoned conduit
active conduit
T4
final tunnel valley form
C. CATASTROPHIC
T1
frozen bed
subglacial lake
T2
sapping near ice margin
frozen bed
subglacial lake
T3
frozen bed
subglacial lake
catastrophic drainage if hydrostatic pressure > sediment strength
T4
final tunnel valley form
*no vertical or horizontal scale implied. Valley size and morphology will vary based on local geologic and glaciological conditions as well as the nature and thickness of sediment in-fill (not shown).
air
ice
water
unlithified substrate
groundwater flow path

1.4.5. Tunnel valleys within glacial landsystems

Subglacial hydrology is an important control on ice dynamics and the stability of ice sheets (Piotrowski, 1997; Greenwood et al., 2016; Ravier and Buoncristiani, 2018). If meltwater excess at the ice-bed interfaces are insufficiently drained, the ice may decouple from its bed, which makes it prone to surging and can lead to destabilization and or/shut-down (Kehew et al., 2012; 2013). Tunnel valleys provide an efficient means of meltwater evacuation and therefore represent a significant influence on ice dynamics (Lelandais et al., 2018).

Discharges of meltwater related to tunnel valley formation are commonly interpreted to occur during deglaciation, and the subsequent reduction in subglacial porewater pressures that ensues is thought to contribute to re-coupling of the ice to its bed and limit ice surging/streaming activity in the area. If this is the case, then there should be consistent relationships between tunnel valleys and landforms associated with ice streaming. Studies that invoke meltwater sheetfloods present descriptions of the relationship between tunnel channels and other bedforms (Brennand and Shaw, 1994; Sharpe et al., 2004; Fisher et al., 2005; Sharpe et al., 2013), but recent investigations in Simcoe County have shown sediment architectures and landsystem relationships that are at odds with existing catastrophic sheetflood erosion models (Bajc et al., 2014; Mulligan and Bajc, 2018; Mulligan et al., 2018b).

1.5. ORGANIZATION OF THE THESIS

To address the project objectives, four main investigations were undertaken to analyze the characteristics and interrelationships of distinct components of the landscape and underlying glacial sediments, and to provide insights into the evolution of processes that governed changes in sediment properties and the development of complex sediment architectures in the region. The results of these four investigations are presented in Chapters 2 through 5 of this thesis.

Chapter 2 presents an integrated analysis of the surficial geology, morphology, and regional subsurface architecture of the large tunnel valleys that create the upland-lowland terrain in Simcoe County, south-central Ontario. The investigation is based on surficial mapping of a nearly 4000 km^2 region lying between the Niagara Escarpment in the west, Oak Ridges Moraine in the south, Lake Simcoe in the east, and Georgian Bay in the north. Mapping is supplemented by geophysical surveys (particularly regional-scale high-resolution, multi-component seismic reflection profiles; Pugin et al., 2018) and extensive drilling of continuously-cored boreholes from surface to bedrock, through both valleys and intervening uplands (Bajc et al., 2015; Mulligan, 2016; 2017c; 2018). Results indicate protracted evolution of the valleys, through multiple phases of meltwater build-up and discharge enhanced by piping from subglacial aquifers, as well as through direct glacial erosion and deformation by active ice that occupied the valleys. Tunnel valley formation largely predates regional till sheet drumlinization. The infill of the valleys is complex, composed of variably textured ice-contact sediments (sand, gravel, silt and diamict), overlain by thick (up to 76 m) successions of proglacial lake sediments.

Chapter 3 provides an analysis of sub-bottom profile data collected in Kempenfelt Bay, the eastern (submerged) portion of one of the valleys studied in Chapter 2. High frequency (2-16 kHz) data creates decimetre-scale resolution of the architecture of bed contacts comprising the valley fill beneath the bay and provides a critical test of concepts developed through subsurface mapping presented in Chapter 2. The data show a prominent undulating and ridged high-amplitude reflector, interpreted as the drumlinized regional Late Wisconsin till sheet, forming the floor of the valley-fill succession. Lows on this basal reflector are overlain by transparent or chaotic facies with local anticlinal structures interpreted as eskers and subaquatic fans. These sediments are draped by thick successions of well-laminated deposits (silt and clays with thin sandy(?) interbeds) with local deformed zones. Deformed zones occur commonly along the flanks of the valley (interpreted as slumps/ mass movement deposits), and record postglacial disturbances of the sedimentary successions, possibly related to neotectonic activity and/or major storm events. The data presented also provide insights into the hydrogeologic setting of the valleys and groundwater recharge to the modern lake body.

Chapter 4 describes the character and internal composition of the regional Late Wisconsin till sheet, the Newmarket Till (NT), that occurs throughout south-central Ontario as an undulating and drumlinized till plain and spans three physiographic regions in the study area (Chapman and Putnam, 1984). The NT caps regional sediment-cored uplands where internal facies display the highest degree of consistency. In lowland regions, it shows greater heterogeneity in terms of its internal composition and physical properties. Along the Niagara Escarpment, a wide variety of diamict facies, pebble lithologies, bedform orientations and morainic deposits has previously led to the identification of three regional tills: the Catfish Creek Till, the Elma Till, and the NT. However, detailed analysis of the morphology of the NT across the three physiographic regions indicate that it (and previous subdivisions of Catfish Creek, Elma and Allenwood tills) forms part of a regionally continuous, but locally variable and heterogeneous, till sheet that blankets southern Ontario. Internal facies and physical characteristics are strongly dependent on substrate material and local bed topography as well as the nature of subglacial drainage at different locations. The data presented in this chapter provide a regional-scale view of heterogeneity within the till, compared to detailed local-scale investigations of the NT undertaken to the south of the study area. The internal characteristics of the NT are used to develop a revised understanding of the depositional evolution of the unit in south-central Ontario, and the implications that interpretations of a heterogeneous internal composition have for regional groundwater studies.

Chapter 5 examines the geomorphological characteristics of south-central Ontario to reconstruct conditions that existed on the bed of the Laurentide Ice Sheet on the scale of former ice lobes and ice streams. Recently-released high-resolution (2- and 5-m cell size) terrain data were used to map subglacial bedforms (drumlins, mega-scale glacial lineations (MSGL), grooved/streamlined bedrock), ice-marginal landforms (end and recessional moraines, eskers, subaquatic fans), and proglacial shoreline features (deltas, bluffs, bars and spits) that occur throughout south-central Ontario, from the Ottawa Valley, to the Algonquin Highlands on the Canadian Shield, through to the flat-lying carbonate bedrock terrain and the thick package of glacial sediments underlying the Peterborough drumlin field (PDF). Mapping reveals complex landform associations across south-central Ontario that can be used to delineate multiple 'flow sets' – areas of consistent bedform orientations with a well-defined spatial extent, usually demarcated by terminal (and possibly lateral) moraines, and associated meltwater features. The data support recent suggestions of ice streaming in southern Ontario (Eyles, 2012; Maclachlan and Eyles, 2013; Eyles and Doughty, 2016; Sookhan et al., 2018a, b; Eyles et al., 2018) and highlight the morphological complexity of bedforms within the PDF. A series of till-cored ridges are the newly-identified southeastern extension of the Lake Simcoe Moraine. These ridges span the entire PDF from Lake Simcoe to the east of the Oak Ridges Moraine and mark an important divide within the region: south and west of the moraines, a single population of bedforms occurs; north and east of the moraines, multiple cross-cutting and superimposed bedforms are observed, as well as abundant complex drumlin morphologies (asymmetric, parabolic, comma-form, spindle, bisected/grooved, multi-tailed) interpreted to record multiple phases of ice flow.

Additional large-scale (tens of km wide) cross-cutting flow sets record significant reorganization of late glacial ice masses within the Lake Ontario basin and Ottawa-St. Lawrence valleys. Well-studied drumlins in the PDF are transitional to drumlins, MSGL, and streamlined Shield bedrock identified across the Algonquin Highlands. The observed relationships hold significant implications for paleoice sheet reconstructions and the data highlight the utility of the newly-released terrain data in future site-scale field investigations.

Chapter 6 provides a summary of the major findings of each of the four investigations comprising this thesis and presents suggestions for future work in southern Ontario and surrounding Great Lakes region. Together, the data provide a robust understanding of how the Laurentide Ice Sheet advanced and retreated from the region, preliminary inferences of evolution of subglacial processes operating during the last glacial maximum, and important implications for groundwater investigations in the region. The data also raise numerous significant questions on the nature of spatiotemporal variations in subglacial conditions governing the accretion of a regional till sheet over variable substrate materials, development of large-scale tunnel valley systems in unconsolidated sediment and bedrock, and the onset, evolution and flow switching of ice sheet and ice stream flows on variable beds. The application of past hypotheses of landscape evolution are discussed and a few speculative solutions to these questions are provided, along with suggestions for possible future investigations.

1.6. PREFACE

This thesis consists of 4 interlinked chapters containing research led by the author. Chapter 2 provides an overview of the sedimentary architecture and landform relationships within Late Wisconsin tunnel valleys and is published in *Quaternary Science Reviews 197*, 49-74. R. Mulligan is the primary author, completing the fieldwork, literature review, writing, figure drafting. Coauthor A.F. Bajc logged 18 of the boreholes presented in the study, conducted additional fieldwork with the lead author and provided invaluable discussions on sediment relationships in the region; Coauthor C.H. Eyles (supervisor) visited field sites, assisted in literature review, and provided essential discussions and editing. Chapter 3 will be formatted for submission to *Journal of Paleolimnology* but has not yet been submitted. Numerous co-authors were involved at various stages of the work. M. Pozza, J. Boyce, M. Doughty, M. Armour, and N. Eyles collected the data and helped with data conversion. J.I. Boyce processed the data and prepared it for analysis. R. Mulligan analyzed the data, conducted the literature review, drafted the figures and prepared the manuscript. C.H. Eyles (supervisor) assisted with the literature review and provided significant discussion and editing throughout the preparation of the manuscript. Chapter 4 is submitted to *Quaternary Research* but has not yet completed peer-review. Author contributions are the same as Chapter 2 (described above). Chapter 5 is published in *Boreas 48(3),* 635-657. A.S. Marich conducted detailed surficial mapping in the central part of the region in 2013-2015. R. Mulligan began manual landform mapping into a GIS in 2011. South-central Ontario was chosen as a site to highlight because of the coverage by numerous high-resolution terrain models, the complexity and interrelationships of landforms in the area, and their possible relation to evolving ice dynamics due to changes in substrate topography and lithology along former ice flow lines, and because the recently-completed surficial mapping provides the necessary ground-truthing for remote analysis. R. Mulligan wrote the manuscript and prepared the figures, assisted by C.H. Eyles (supervisor) who added significant discussion and edits. A.S. Marich provided essential information on landform composition and discussion of concepts presented in the paper.

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CHAPTER 2: DRUMLINIZED TUNNEL VALLEYS IN SOUTH-CENTRAL ONTARIO

*this is the accepted version of a paper with the same title, published in *Quaternary Science Reviews 197*, 49-74. https://doi.org/10.1016/j.quascirev.2018.07.026

Abstract

A network of Late Wisconsin valleys deeply incise thick successions (up to 200 m) of Quaternary sediment and Paleozoic bedrock across south-central Ontario. The valleys have previously been interpreted as an integrated network of tunnel channels recording catastrophic releases of subglacial meltwater across the bed of the Laurentide Ice Sheet (LIS). Recent geologic investigations in the western part of the valley network integrate information gained from surficial sediment and landform mapping, continuously-cored boreholes that penetrate the entire Quaternary sediment succession in valleys as well as intervening uplands, and both water-borne and land-based geophysical surveys. Together, these data provide new insights and a refined interpretation of the genesis, timing and paleoglaciological significance of these valleys.

Valleys within the study area are some of the largest observed within the regional network – up to 30 km long, 7 km wide and >175 m deep. They are oriented both north-south and in a radial pattern, varying from NE-SW in the southern part of the study area, to ESE-WNW in the north. An undulating Late Wisconsin till sheet (Newmarket Till) caps the regional uplands and is commonly observed along the flanks and floors of the valleys – all of which show evidence of a drumlinization phase that postdates excavation of most of the valleys. At least two generations of valleys are interpreted from cross-cutting relationships, with abrupt heads and locally perched downflow ends of some valley forms. Radiocarbon age determinations of organic material from both below and above the Newmarket Till constrain the timing of valley excavation to between 28.05 and 12.81 ^{14}C kyr BP. Borehole and outcrop data within the valleys reveal four stratigraphic units (SU1-SU4) that form the floor and sediment infill of the valleys. The Newmarket Till (SU1) is commonly underlain by deformed substrates. The till is absent in parts

of some valleys and the lowermost parts of the valley fill commonly consist of coarse-grained gravels (SU2) that pass upwards into sands (SU3). The bulk of the sediment infill consists of glaciolacustrine silt-clay rhythmites (SU4) that pass upwards into organic- and mollusk-bearing nearshore and fluviodeltaic deposits that record drainage of proglacial lakes from the study area during deglaciation.

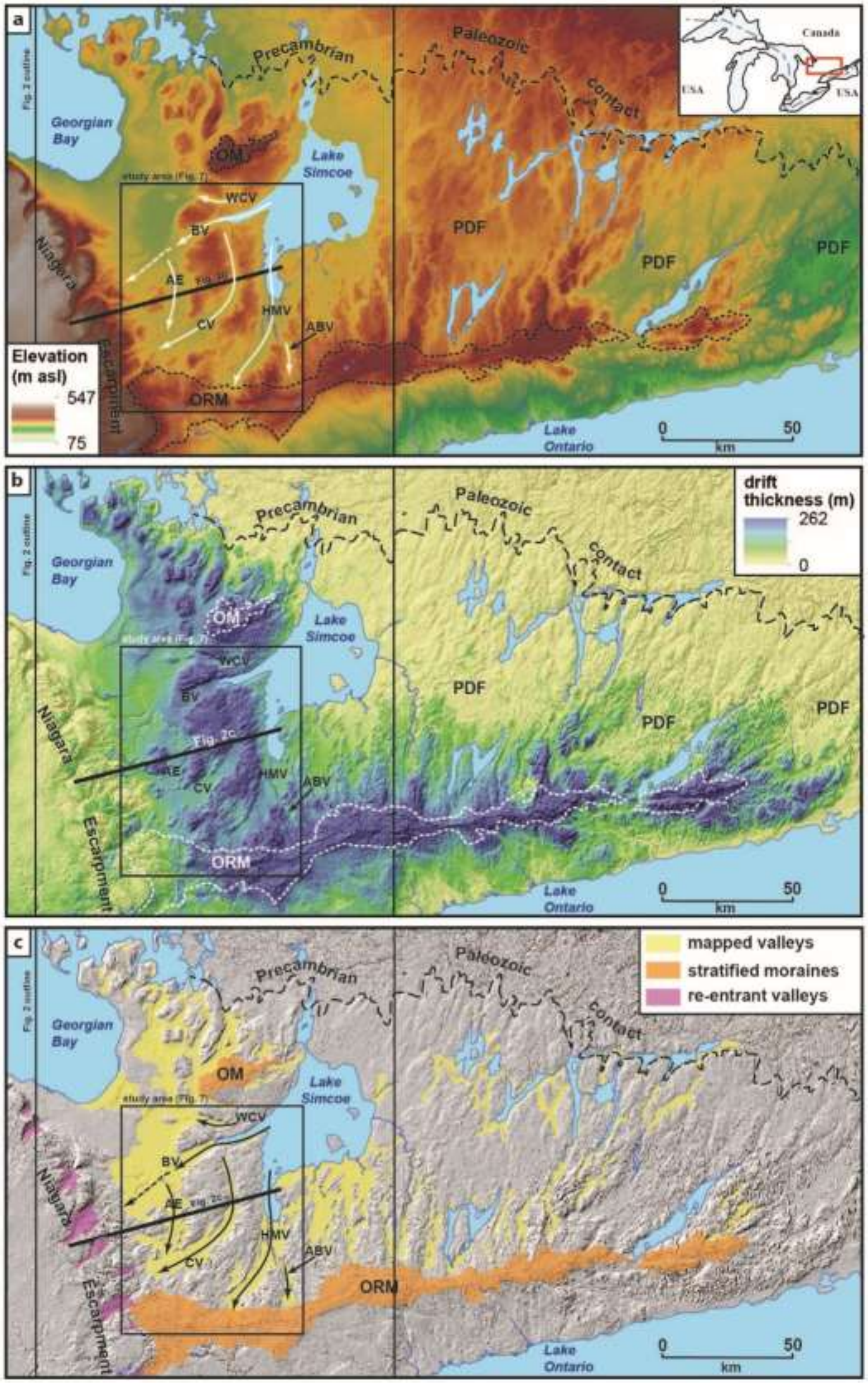
It is proposed that these features are a previously unrecognized type of tunnel valley that developed subglacially during Late Wisconsin ice cover. Meltwater surpluses would be generated at the ice-bed interface due to the low-transmissivity substrate in the study area with flow preferentially routed into lows on the bed. Downward incision into confined and pressurized aquifers likely facilitated rapid headward erosion of the valleys enhanced by piping. Waning meltwater flow velocities/discharges allowed ice creep to re-occupy newly-formed valleys permitting renewed till deposition and subsequent drumlinization of the valleys. The valleys also provided efficient routing for meltwater draining the retreating ice marginal zone during deglaciation and allowed localized sand and gravel deposition as eskers and ice-proximal subaquatic fans in the Oak Ridges Moraine and within proglacial lakes that developed during ice retreat. However, there is no evidence for regional-scale catastrophic release of subglacial meltwater in the valley infills. This integrated dataset represents one of the most complete characterizations of subglacial valleys in a terrestrial setting in North America.

2.1. INTRODUCTION

Glacial sediments are found across the globe and host or form a variety of important resources, ranging from groundwater to hydrocarbons, aggregates and industrial minerals (Huuse et al., 2012). Developing improved models of glacial erosion, transport, deposition and associated sediment facies architecture is an essential prerequisite for reliably discovering and characterizing, as well as efficiently and safely managing these resources. In southern Ontario, thick successions (>200 m) of glacial sediments host regional and local aquifers that supply drinking water to millions of residents, assist in maintaining ecologically sensitive habitats, and form aggregate resources essential for infrastructure development (Barnett, 1992a; Sharpe et al., 2013).

Demand for improved geologic information in support of groundwater studies across southern Ontario has led to the collection of a wealth of high-quality geologic data over the last 25 years (Barnett, 1993; Barnett et al., 1998; Sharpe et al., 2002; Burt and Dodge, 2011). These data have shed light on and highlighted the importance of large surficial and buried Quaternary-aged valleys in regional groundwater systems (Desbarats et al., 2001; Sharpe et al., 2013). Recent investigations in Simcoe County, Ontario (Fig. 2.1) have integrated detailed surficial mapping with extensive, high quality, continuous sediment coring and regional geophysical surveys (Bajc and Rainsford, 2010; 2011; Mulligan, 2011; 2013; 2014; 2015; 2016; 2017a,b,c; Bajc et al., 2012; 2014; 2015; Mulligan and Bajc, 2012) to provide new insights into the properties and sedimentary architecture of regionally identified stratigraphic units in the area (Mulligan and Bajc, 2018; Mulligan et al., 2018).

Fig. 2.1: a) topography and physical setting, b) drift thickness and c) annotated classified map of the study area (black rectangle) in south-central Ontario showing the network of valleys previously interpreted as tunnel channels (simplified from Brennand et al., 2006). The central axes of tunnel valleys and inferred paleoflow direction are highlighted with arrows. Elongate lakes and bays fill several of the larger valleys in the region. Inset map in a) shows location of figure within Great Lakes region of North America. PDF = Peterborough Drumlin Field; ORM = Oak Ridges Moraine; OM = Oro Moraine; WCV = Willow Creek valley; BV = Barrie valley; AE = Alliston Embayment; CV = Cookstown valley; HMV = Holland Marsh valley; ABV=Aurora buried valley. Location of Figs 2 and 7a shown in black lines, Fig. 2.2c labelled.

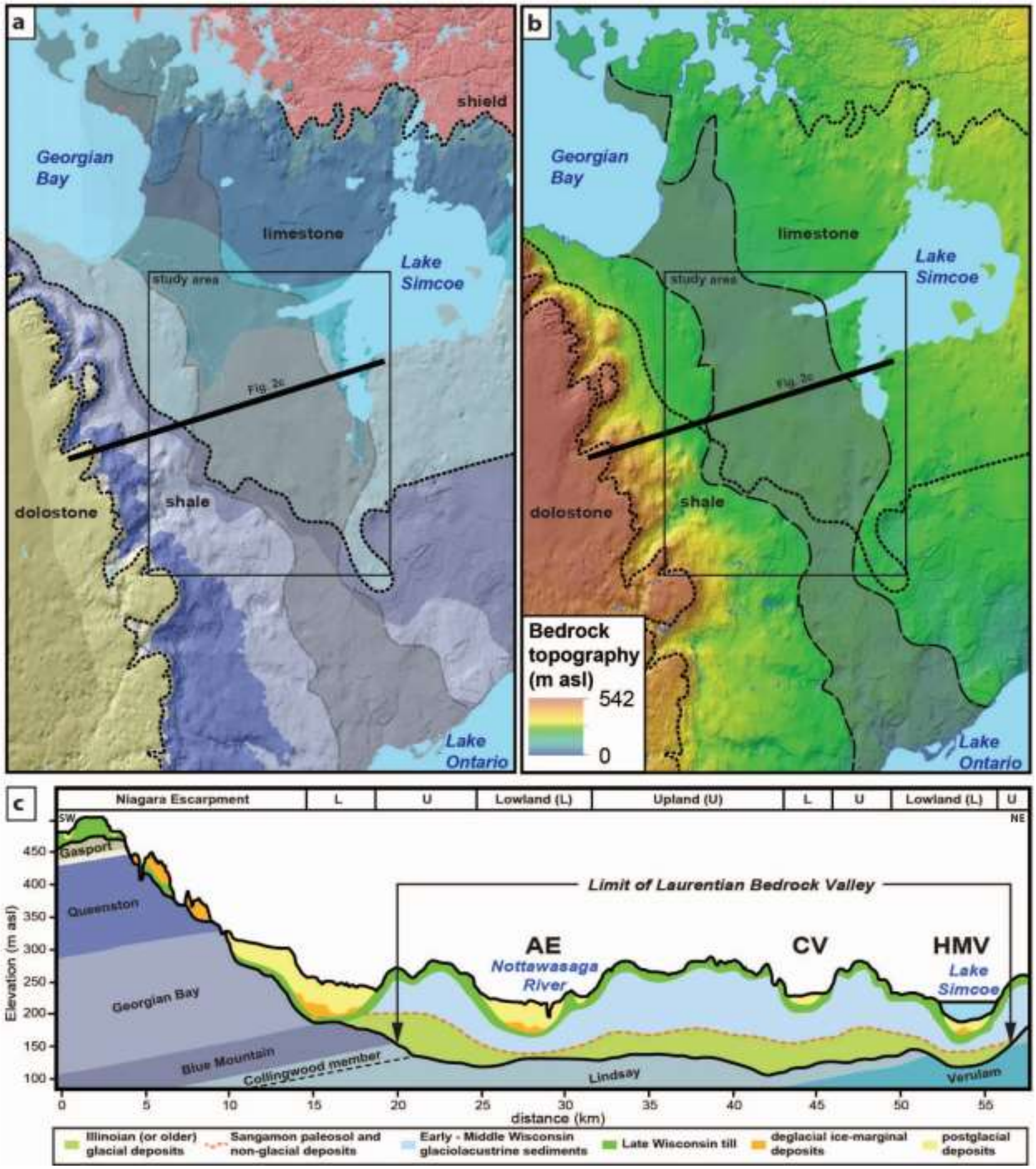

a
Precambrian
Paleozoic
contact
Canada
USA
Georgian Bay
Lake Simcoe
OM
WCV
BV
AE
CV
HMV
ABV
ORM
PDF
Niagara
Escarpment
Elevation (m asl)
547
75
Lake Ontario
0
50
km
b
drift thickness (m)
262
0
c
mapped valleys
stratified moraines
re-entrant valleys

The study area comprises parts of two 3D mapping areas in Simcoe County, where a number of deeply incised valleys truncate late Pleistocene glacial units, locally extending into underlying Paleozoic bedrock (Figs. 2.1, 2.2). The valleys are only partially in-filled, giving the present landscape a characteristic upland-lowland configuration (Chapman and Putnam, 1984). They have previously been interpreted to form part of a broad, integrated tunnel channel network that spans much of south-central Ontario (Brennand and Shaw, 1994; Sharpe et al., 2004; Fig. 2.1). Their presence has been known for decades (Straw, 1968; Barnett, 1990; Sharpe et al., 2004; Sookhan et al., 2018), but consensus on the underlying processes responsible for their formation has not been reached. This paper provides new insights on the geomorphic relationships between uplands and valleys and describes the character and architecture of major sediment units comprising the valley infill, as well as the underlying regional strata. An emphasis is also placed on the implications of these observations for understanding valley genesis and their relationship with changing subglacial conditions beneath the Laurentide Ice Sheet (LIS).

2.2. REGIONAL GEOLOGIC SETTING

The study area is bordered by Georgian Bay in the northwest, the Precambrian Shield peneplain to the northeast, Lake Simcoe and the till plain of the Peterborough drumlin field (PDF) to the east and southeast, high hummocky topography of the Oak Ridges Moraine (ORM) to the south, and the Niagara Escarpment to the west (Fig. 2.1). Within southern Simcoe County, predominantly drumlinized upland terrain (the western extension of the PDF) is underlain by thick successions (up to 200 m) of Quaternary sediment. The uplands are truncated by a series of

Fig. 2.2: a) bedrock geology and b) topography of the study area (black rectangle) and Laurentian Valley (shaded with long dashed outline). Short dashed lines denote major breaks in bedrock lithology from crystalline shield rocks in the northeast, to limestone, shale, and dolostone further southwest. See Fig. 2.1 for location, Fig 2.2c shown in thick black line; c) cross-sectional profile showing the bedrock topography and lithology as well as Quaternary sediment thickness across the Laurentian Valley in the central part of the study area. Modern hydrological features added for reference and valley names (see Fig. 2.1) shown to highlight broad-scale sediment architecture and upland-lowland physiography (see Fig. 2.4). Note the significant thinning of Early-Middle Wisconsin sediments beneath lowlands compared to intervening uplands. Vertical exaggeration ~30x; bedrock unit dips are not to scale, see legend in a).


a
shield
Georgian Bay
limestone
Lake Simcoe
study area
Fig. 2c
dolostone
shale
Lake Ontario
b
Georgian Bay
limestone
Lake Simcoe
study area
Fig. 2c
dolostone
shale
Bedrock topography (m asl)
542
0
Lake Ontario
c
Niagara Escarpment
L
U
Lowland (L)
Upland (U)
L
U
Lowland (L)
U
SW
NE
Gasport
Queenston
Georgian Bay
Blue Mountain
Collingwood member
Lindsay
Verulam
Limit of Laurentian Bedrock Valley
AE
Nottawasaga River
CV
HMV
Lake Simcoe
Elevation (m asl)
450
400
350
300
250
200
150
100
0
5
10
15
20
25
30
35
40
45
50
55
distance (km)
Illinoian (or older) glacial deposits
Sangamon paleosol and non-glacial deposits
Early - Middle Wisconsin glaciolacustrine sediments
Late Wisconsin till
deglacial ice-marginal deposits
postglacial deposits

valleys in-filled by up to 120 m of deglacial sediments and are presently occupied by underfit streams and rivers, extensive wetlands, and elongated bays in Lake Simcoe (Figs. 2.1, 2.2).

2.1.1. Bedrock geology

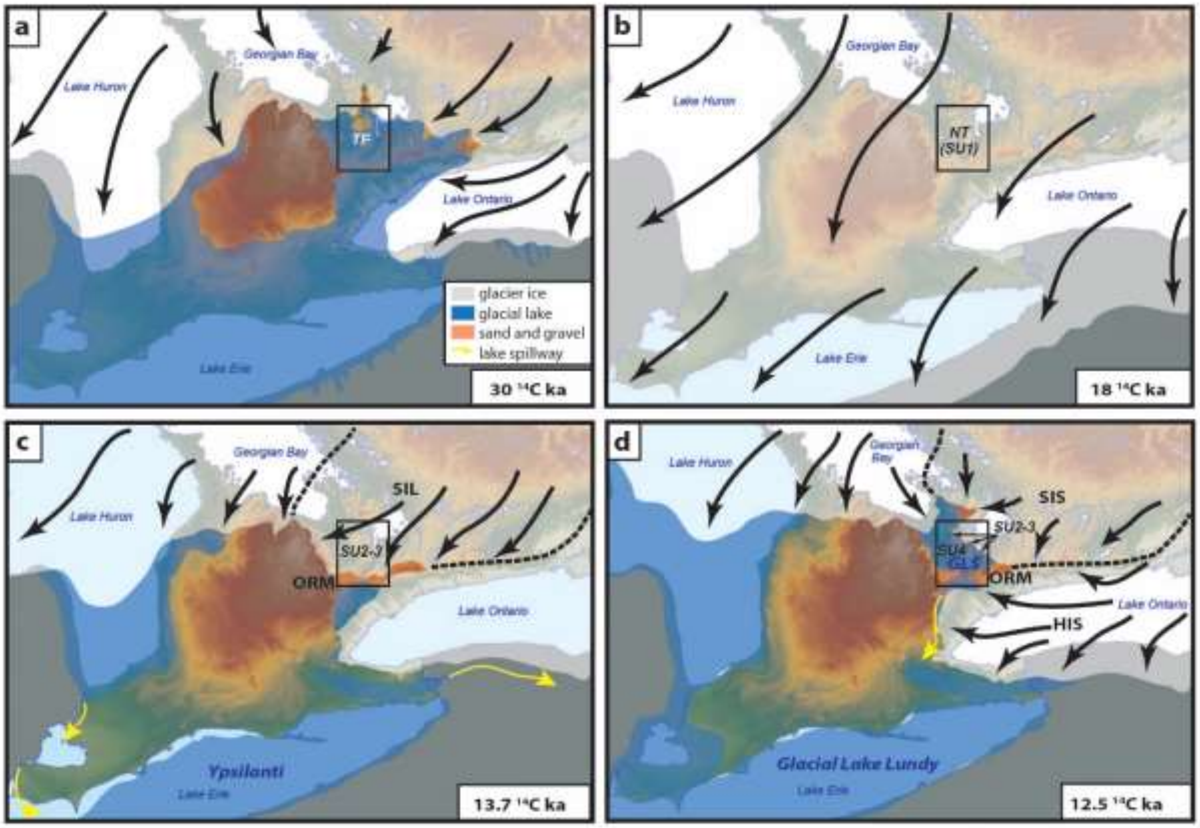
Beneath the study area, gently-dipping strata of Paleozoic limestone and shale onlap the buried Precambrian shield which outcrops approximately 45 km to the north (Fig. 2.2a), although an increasing number of possible inliers of Precambrian basement rocks have been intersected during recent drilling investigations immediately north of the study area (Mulligan, 2016; 2017c). Paleozoic strata dip gently to the southwest and are deeply eroded within a broad valley/trough extending from Georgian Bay to Lake Ontario (Fig. 2.2b; Gao et al., 2006). The trend and location of the valley are generally thought to be inherited from underlying structural features in Precambrian basement rocks (Eyles et al., 1997) and it may mark the possible location of a preglacial fluvial system (Laurentian Valley; Spencer, 1890) draining the upper Great Lakes. Recent papers have advocated a polygenetic origin for the trough invoking various degrees of glacial/glacial meltwater modification of the preglacial landscape throughout the Quaternary (Gao, 2011; Mulligan and Bajc, 2018; Sharpe et al., 2018). The Laurentian trough is bounded to the west by the Niagara Escarpment, a prominent (>300 m) bedrock scarp interrupted by a series of near-parallel re-entrant valleys (Figs. 2.1, 2.2; Straw, 1968; Eyles et al., 1997). The eastern and northern extents of the Laurentian Valley, as well as details on potential tributaries and discrete channel features within the main valley/trough, remain poorly-defined (Sharpe et al., 2018; Mulligan, 2017c), largely due to the lack of deep subsurface information constraining the geometry of the bedrock surface (Fig. 2.2c).

2.1.2. Quaternary geology

Sediments infilling the Laurentian valley/trough in the study area are attributed to at least the last two glacial-interglacial episodes (Karrow, 1967; Bajc et al., 2014; Mulligan, 2016; 2017c; Mulligan and Bajc, 2018). The succession consists of a lower (Illinoian and/or older) glacial package (tills and interbedded glaciolacustrine deposits) overlying Paleozoic bedrock (Fig. 2.2c), with a regionally identified subaerial unconformity on its upper surface. Thin, non-glacial, alluvial, lacustrine, and peat deposits that span Sangamon (last interglacial) through to Middle Wisconsin time overlie the unconformity (Mulligan and Bajc, 2018; Mulligan, 2017c; Fig. 2.2c). The regional unconformity and its associated deposits are buried by thick successions of glaciolacustrine rhythmites, the earliest deposits of which lack evidence of proximal ice. Glacially influenced rhythmites higher in the succession are interbedded with sand bodies often containing detrital organic accumulations. Together, these lake deposits record a flooding event resulting from ice advance into the St. Lawrence valley beginning in the Early Wisconsin that prevented eastward drainage of the Lower Great Lakes (Figs. 2.2c, 2.3a; Eyles et al., 1985). Within the study area, the sand interbeds within the glaciolacustrine succession both thin and fine towards the south (Mulligan and Bajc, 2018). However, units to the north and south of the study area, containing significant thicknesses of sand and locally, gravel, have been identified at the same stratigraphic position (Sharpe et al., 2002; Burt and Dodge, 2011; Mulligan, 2016; 2017c; Gerber et al., 2018).

Capping the glaciolacustrine sequence is the Newmarket Till, recording Late Wisconsin ice advance into the study area and southward into the northern United States (Figs. 2.2c, 2.3b; Gwyn and Dilabio, 1973; Sharpe et al., 1994; Bajc and Rainsford, 2010). The surface of the Newmarket Till is highly undulating, drumlinized, and well-exposed across the majority of

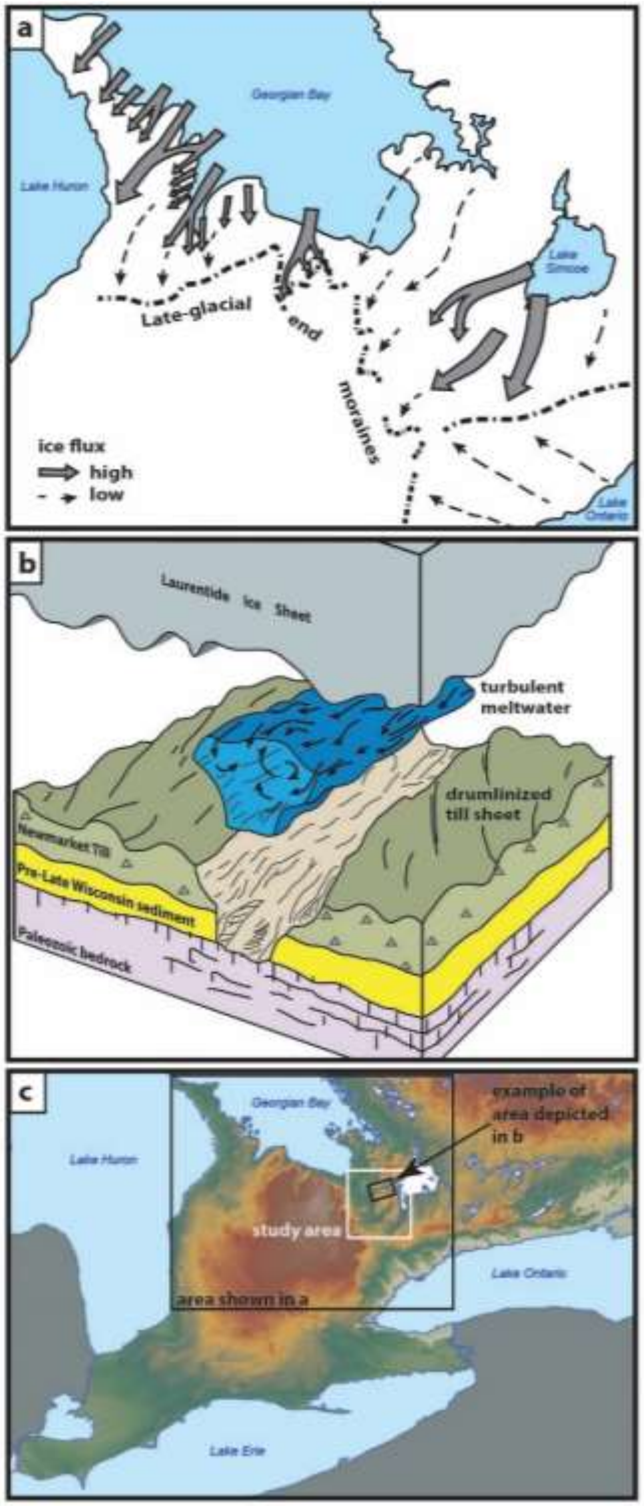
Fig. 2.3: a) advance of the LIS prior to the last glacial maximum (LGM). Middle Wisconsin Thorncliffe Formation (TF) sediments were deposited in a large proglacial lake; b) deposition of Newmarket Till (NT; SU1) beginning during LGM ice cover. Generalized ice flow lines shown in black arrows; c) Mackinaw phase ice retreat. Deposition of thick sediments associated with the ORM (SU2-3). Dashed black lines mark approximate boundaries of the Simcoe Ice Lobe (SIL); d) Post-Port Huron phase ice and glacial lake extents in southern Ontario, with coarse-grained sediments deposited near ice margins (SU2-3) and settling of fines in ice distal areas (SU4). Study area shown in black rectangle, proglacial lake spillways shown in yellow arrows. Ages are approximate, data from multiple sources, see references in text. GLS= glacial Lake Schomberg, ORM= Oak Ridges Moraine; SIS= Simcoe Ice stream, HIS= Halton Ice Stream; see Sookhan et al. (2018).


a
Georgian Bay
Lake Huron
TF
Lake Ontario
glacier ice
glacial lake
sand and gravel
lake spillway
Lake Erie
30 ¹⁴C ka
b
Georgian Bay
Lake Huron
NT (SU1)
Lake Ontario
Lake Erie
18 ¹⁴C ka
c
Georgian Bay
Lake Huron
SIL
SU2-3
ORM
Lake Ontario
Ypsilanti
Lake Erie
13.7 ¹⁴C ka
d
Lake Huron
Georgian Bay
SIS
SU2-3
SU4
GLS
ORM
Lake Ontario
HIS
Glacial Lake Lundy
Lake Erie
12.5 ¹⁴C ka

upland areas throughout the region, but its full topographic expression is less well understood beneath modern lakes and under extensive ice-marginal and postglacial glaciolacustrine deposits within the ORM and lowland settings (Figs. 2.3c, d; Deane, 1950; Todd et al., 2008; Mulligan et al., 2018).

The valleys that separate drumlinized uplands were originally attributed to focussed erosion along zones of fast-flowing ice beneath the LIS (Fig. 2.4a; Straw, 1968). Cored boreholes coupled with reflection seismic and downhole geophysical investigations carried out in south-central Ontario failed to record Newmarket Till within the valleys, suggesting truncation of the unit (Barnett, 1990). This was attributed to subglacial meltwater erosion event(s) (Barnett, 1990) that were later interpreted to record a widespread regional unconformity (Fig. 2.4b; Brennand and Shaw, 1994; Sharpe et al., 2002, 2004; Sharpe and Russell, 2016). In the southernmost part of the study area, the till surface is deeply buried by sand, silt and gravel of the ORM, deposited during initial break-up of the LIS into semi-independent ice lobes flowing out from the lake basins (Lake Ontario, Georgian Bay and Lake Simcoe) and off the Algonquin Highlands in southern Ontario (Barnett, 1992a; Eyles and Doughty, 2016; Mulligan et al. 2017; Sookhan et al., 2018; Fig. 2.3c). Sediments in the ORM have been interpreted to record the distal and late phases of subglacial outburst floods responsible for the erosion of the tunnel channel network across southern Ontario (Brennand and Shaw, 1994; Sharpe et al., 2004; 2013; Sharpe and Russell 2016). Alternative interpretations of ORM development invoke a more time-transgressive origin, resulting from deposition of sediments into a growing interlobate lake, where sediment facies and landform morphology are influenced by fluctuating ice margins that controlled sediment input points and base levels (Duckworth, 1979; Chapman, 1985; Howard et al., 1995; Barnett et al., 1998; Sookhan et al., 2018).

Fig. 2.4: a) subglacial valley erosion primarily through subglacial exhumation of substrate materials due to inferred high ice fluxes (from Straw, 1968); b) tunnel channel erosion by turbulent meltwater following collapse of a regionally extensive subglacial sheetflood (from Sharpe et al., 2002); c) extent of a,b, and study area in southern Ontario.


a
Georgian Bay
Lake Huron
Lake Simcoe
Lake Ontario
Late-glacial
end
moraines
ice flux
high
low
b
Laurentide Ice Sheet
turbulent meltwater
drumlinized till sheet
Newmarket Till
Pre-Late Wisconsin sediment
Paleozoic bedrock
c
Georgian Bay
Lake Huron
example of area depicted in b
study area
Lake Ontario
area shown in a
Lake Erie

The flanks of uplands are commonly gullied and in some locations, particularly on northwest-facing slopes, wave erosion during inundation by postglacial lakes has locally eroded and reworked the Newmarket Till along steep, high bluffs, redepositing it as sand and gravel in large spits and beach/bar complexes (Chapman and Putnam, 1984; Schaetzl et al., 2016; Mulligan, 2017a,b). In other areas, groundwater piping along the margins of the uplands has created large erosional amphitheatres up to a few kilometers wide and >60 m deep (Barnett, 1997; Bajc and Rainsford, 2010; Mulligan, 2013).

2.3. DATA COLLECTION AND METHODS

Recent field investigations in the study area were initiated in 2010 and 2011 with surficial sediment mapping (Bajc and Rainsford, 2010; Mulligan, 2011; 2013; 2014; 2015; 2017a, b; Mulligan and Bajc, 2012; Fig. 2.5). Surficial sediments were logged using standard sedimentological logging techniques and a landsystems approach was used for mapping surficial features to assist in evaluating sediment-landform assemblages and spatiotemporal trends in sediment properties (Evans and Twigg, 2002; Evans, 2003; Kehew et al., 2012a; Mulligan, 2014). Delineation of surficial sediments and geomorphic features was facilitated by the analysis of high-resolution (5 m and 2 m cell size) digital elevation and surface models (MNRF, 2010; 2013) as well as aerial photography and satellite imagery (Mulligan and Bajc, 2012; Mulligan 2017a, b).

Regional geophysical surveys (ground gravity, seismic reflection, and airborne EM) were undertaken to improve characterization of the subsurface and assist in the selection of borehole locations (Bajc and Rainsford, 2011; Bajc et al., 2012; 2014; Pugin et al., 2018). In total, 28

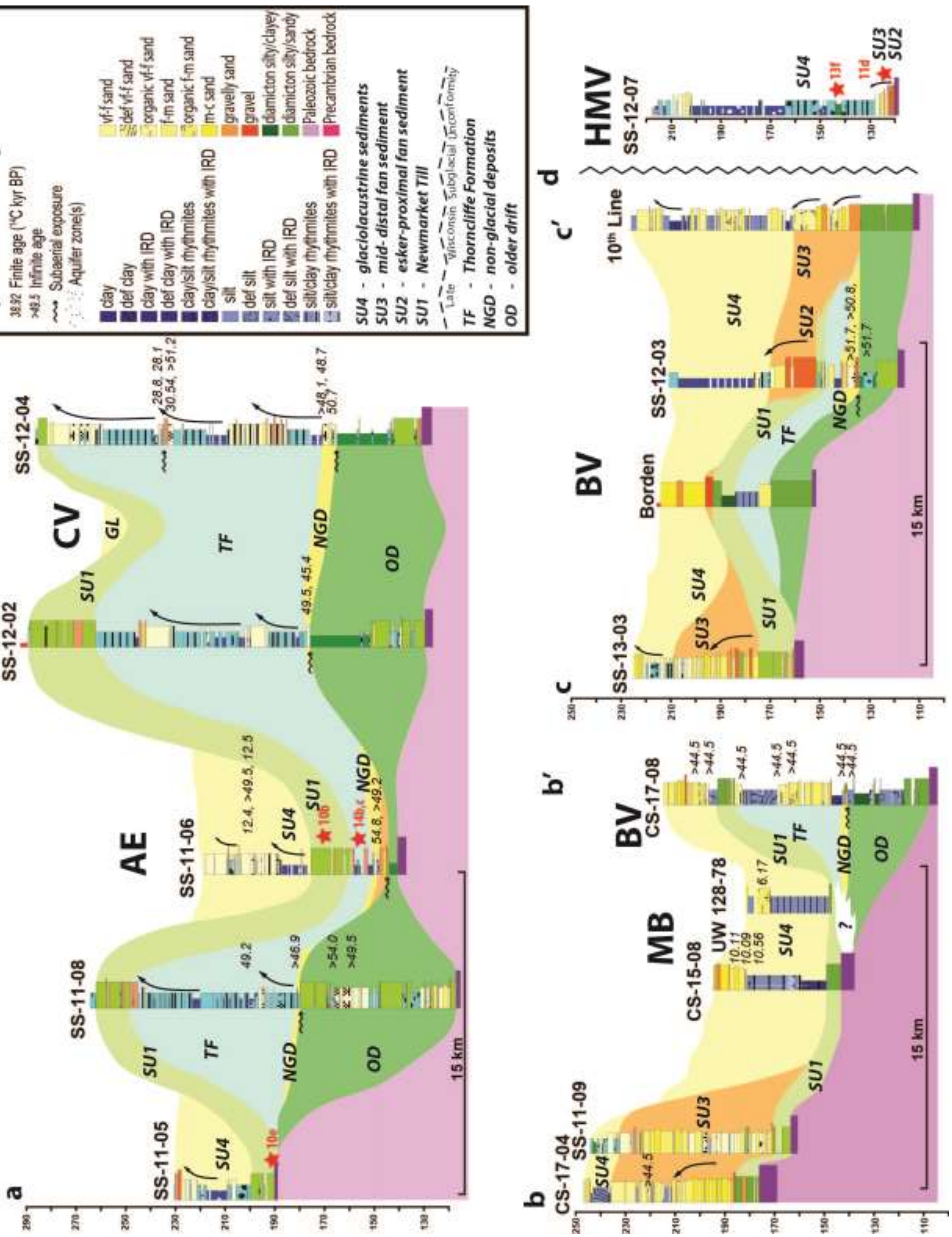
Fig. 2.5: Data distribution in the study area (thick black rectangle) overlain on hillshade showing regional physiography, simplified from Chapman and Putnam (1984). Recent OGS 3D sediment mapping areas in central and southern Simcoe County shown in black dashed line, recently completed surficial maps shown in thin black line. Other cored boreholes and seismic reflection surveys (pink lines), shown for reference (see references in text). Locations of cross-sections and boreholes in Fig. 2.6 shown in thick dashed white lines.


Precambrian Shield
Georgian Bay
Lake Simcoe
Lowland
Upland
Oro
Elmvale
Borden
Schomberg
10th line
WCV
BV
AE
CV
HMV
Dundalk Plain
Niagara Escarpment
Oak Ridges Moraine
Fig. 9a
Fig. 9b
Fig. 9c,d
0
15 km
Cored borehole to bedrock
Cored borehole (not reaching bedrock)
Borehole shown in Fig. 5
Seismic reflection Survey

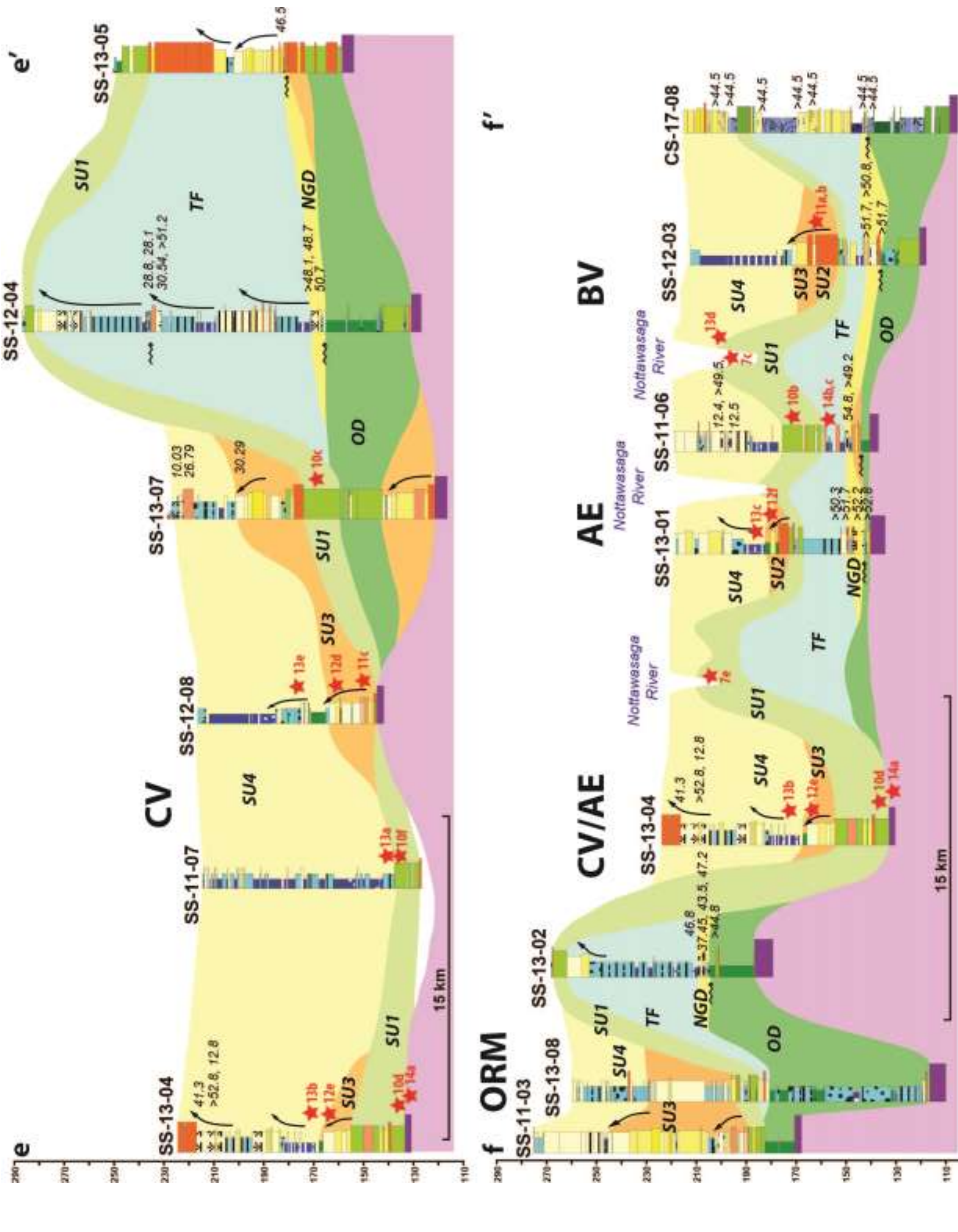
continuously-cored boreholes were drilled in the study area, with all but one reaching bedrock; an additional 71 boreholes were drilled to the north and northwest by the Ontario Geological Survey (Burt and Dodge, 2011; Mulligan 2016; 2017c; Figs. 2.5, 2.6) and three previously cored boreholes provide regional data constraint (Borden log provided by M. Tetreault, and 10th line log provided by R. Gerber; see Crow et al., 2015; data for UW-128-78 from Fitzgerald, 1982).

Boreholes drilled in this study were cored using a mud-rotary system equipped with a Christiansen core barrel retrievable by wireline providing excellent preservation of fine sedimentary structures. PQ diameter cores (approximately 8.5 cm) were logged on-site, recording colour, grain size, sorting, sedimentary structures, nature of bedding contacts, reactivity with 10% hydrochloric acid and degree of consolidation. Analysis of sediment successions exposed in boreholes within the valleys highlighted the existence of four stratigraphic units (SU1-SU4) that composed the floor and infill of the valleys in the region. Photos were taken at 25 cm intervals and additional photos were taken of small-scale structures and significant contacts. Representative samples were taken at regular intervals (1 m) and across all major changes in sediment type for particle size analysis. Diamict units were analyzed for a number of additional parameters including total matrix carbonate, calcite:dolomite ratios and heavy mineral content. Organic-bearing units were sealed in vinyl wrap in the field and saved for detailed sampling and processing in the lab to isolate material suitable for radiocarbon age dating, macrofossil analysis and pollen determinations (Fig. 2.5; Bajc et al., 2014; Mulligan and Bajc, 2017a; see Appendix). Logs of all boreholes described in this report are publically available online (Bajc, 1994; Burt and Dodge, 2011; Bajc et al., 2015; Mulligan, 2016; 2017c).

Fig. 2.6: Generalized cross-sectional profiles showing radiocarbon age determinations (Bajc et al., 2014) and unit geometry interpreted from continuously-cored boreholes in the study area. a) E-W profile showing upland-valley architecture. Note radiocarbon dates above and below Newmarket Till that help to constrain timing of valley erosion; b,c) longitudinal profiles upflow (left to right) in the northwestern and southwestern parts of the BV, respectively; d) sediment infill exposed in a single borehole in the HMV; e) longitudinal profile upflow (left to right) in the CV. Note the progressive deepening of the valley in the downflow direction and localized deposition of sandy valley-fill deposits (SU3); f) S-N profile showing relationship between ORM, CV, AE, and BV. The base of AE is much shallower than BV and CV; Red stars and numbers show locations of photos shown in Figs. 2.6-2.12. The upper parts of SU4 (shown in beige) record sediment facies changes associated water plane fluctuations during the later stages of ice retreat and are not discussed in this paper. See Mulligan et al. (2018) for details, radiocarbon dates for Simcoe County provided in Appendix.



Key to Sediment Logs
SU4 - glaciolacustrine sediments
SU3 - mid- distal fan sediment
SU2 - esker-proximal fan sediment
SU1 - Newmarket Till
TF - Thorncliffe Formation
NGD - non-glacial deposits
OD - older drift
AE
CV
MB
BV
HMV
15 km


e
e′
f
f′
CV
ORM
CV/AE
AE
BV
SS-13-04
SS-11-07
SS-12-08
SS-13-07
SS-12-04
SS-13-05
SS-11-03
SS-13-08
SS-13-02
SS-13-01
SS-11-06
SS-12-03
CS-17-08
Nottawasaga River
15 km

2.4. SURFICIAL GEOLOGY AND GEOMORPHOLOGY

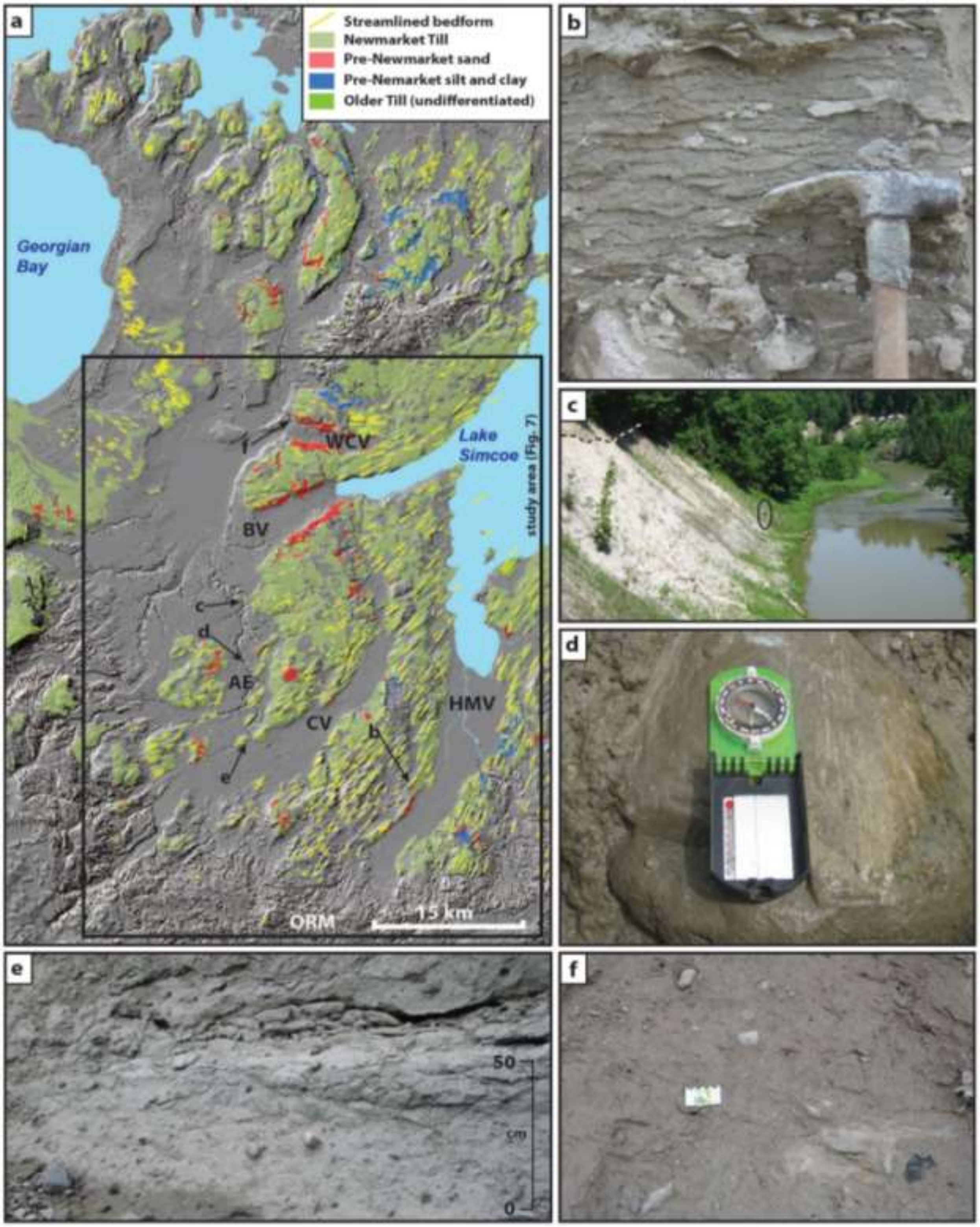
Surficial mapping investigations in the study area have shed light on important geomorphological relationships between the valleys and other glacial landforms. These help guide understanding of the distribution of major sediment packages between uplands and lowlands. Uplands are cored primarily with fine-grained glaciolacustrine rhythmites with localized sand and gravelly deposits and capped by a drumlinized till sheet whereas the lowlands represent significant removal of pre-Late Wisconsin sediments and contain the same drumlinized till sheet along their flanks and floors, as well as host thick successions of postglacial valley infill sediments.

2.4.1. Drumlinized surficial till sheet

Drumlins occur within a broad range of settings within the study area. They are most commonly observed in large groups ornamenting upland terrain, although they can be found flanking valley walls and as isolated ridges within or along the margins of valleys (Figs. 2.5, 2.7, 2.8). Drumlins are generally large and broad with younger flutings being relatively uncommon compared to the landscapes created beneath other late glacial ice lobes in southern Ontario (Mulligan et al., 2016; Chapter 5). An overall trend in drumlin long-axis orientation is observed; drumlins in the northern part of the study area have long axes oriented to the WNW or WSW and drumlins in the south have S or SSW long axis orientations (Figs. 2.7a, 2.8a). Drumlins in the southern part of the study area, particularly those lying at lower elevations, have been partly buried by glaciolacustrine sediments, reducing their overall topographic expression.

The uplands in the study area are capped by a moderately- to over-consolidated, brown to grey or grey-brown, predominantly massive diamict that forms the core of the majority of streamlined bedforms (Fig. 2.6). Matrix grain size is fairly consistent, varying from silty sand to sandy silt with a fining trend southward off the Precambrian shield. Exposures through drumlins

Fig. 2.7: a) surficial distribution of sediments that pre-date deglaciation in the region (NT in green, older sands in red, clays in blue, and tills in lime green; data from outside recent mapping areas from OGS, 2010). Drumlin long axes shown in yellow, locations of b-f marked; b) fissile dense till exposed in a drumlin on uplands; c) large outcrops exposing thick (>13 m) successions of Newmarket Till (SU1) overlain by fine-grained glaciolacustrine sediments (SU4); contact shown with dashed line, person circled in black for scale; d) striated bullet-shaped boulder with parallel striae and long axis indicating ice flow toward the southwest; e) overconsolidated massive fissile till at the base of a cut bank exposure along the southern flank of a large drumlin; f) massive moderately- to well-consolidated till with deformed sand lenses and pods. Scale bar on card is in cm.


a
Streamlined bedform
Newmarket Till
Pre-Newmarket sand
Pre-Nemarket silt and clay
Older Till (undifferentiated)
Georgian Bay
Lake Simcoe
study area (Fig. 7)
WCV
f
BV
c
d
AE
CV
HMV
b
e
ORM
15 km
b
c
d
e
50
cm
0
f

in upland areas reveal primarily massive, sandy diamict showing well-developed fissility (Fig. 7b) with occasional thin (generally <5 cm) sorted interbeds of sand and silt. Amalgamated diamict beds of variable texture are observed at some locations, separated in places by stratified deposits. Clast content commonly ranges between 5 and 15% and consists mainly (70-90%) of local Paleozoic carbonate and shale rocks, generally less than 20 cm in diameter. Boulders up to 3 m in diameter have been observed at some locations. Striated and faceted boulders within the matrix indicate a southwestward ice flow direction, consistent with the long-axis orientations of drumlins in the area (Fig. 2.7d).

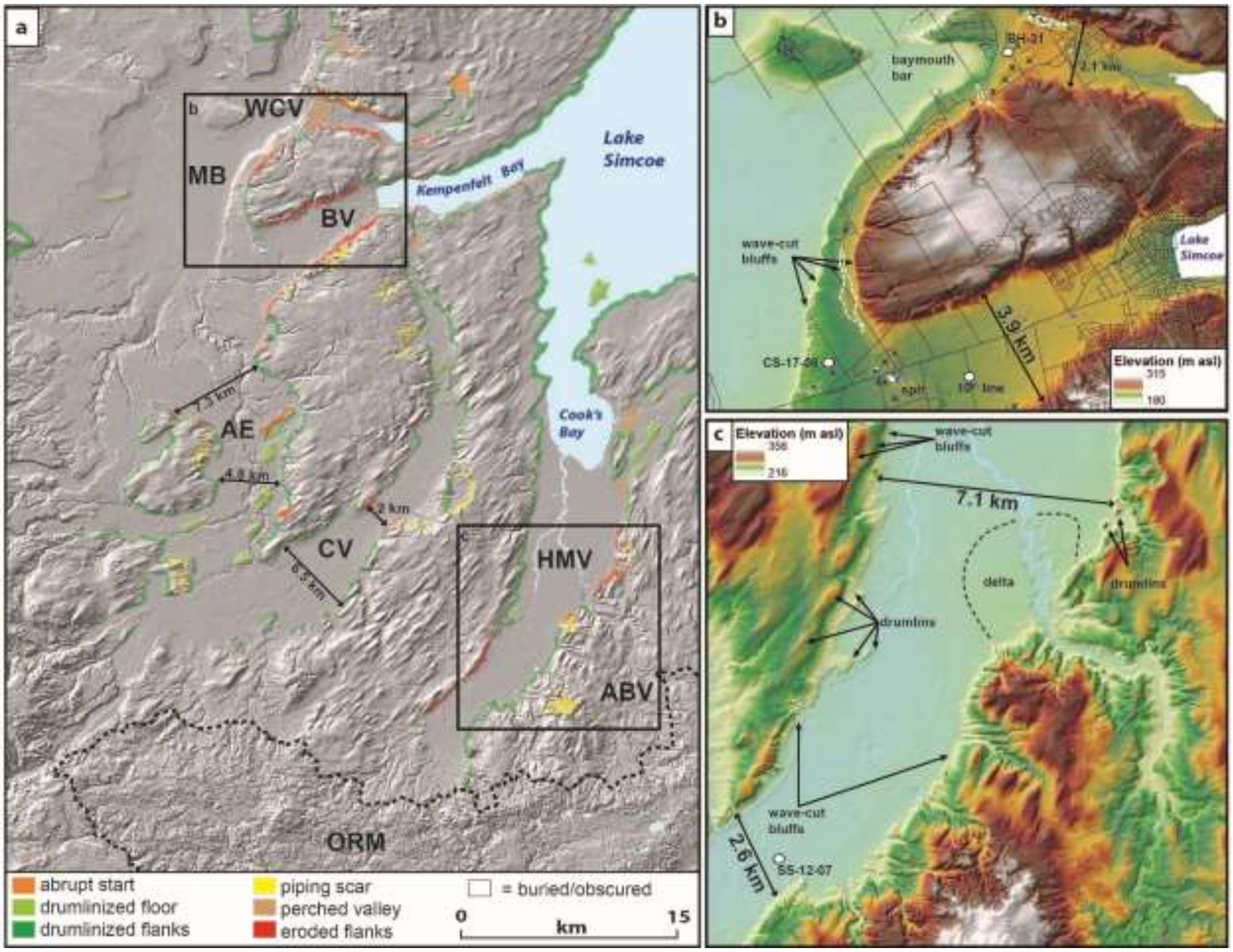
The diamict is commonly observed along the flanks of the uplands/valleys with little discernable variations in facies, striation orientations or sediment properties (Fig. 2.6). A higher degree of interbedding of diamict beds with varying textures is locally observed at surface along the flanks of the Cookstown valley and southern part of the Holland Marsh Valley (Figs. 2.1, 2.6d, 2.8). In some places, the flanks of the valleys are blanketed by interbedded silt and fine-grained diamicts with bed contacts that follow the general slope of the present valley. The western half of the Barrie Valley is a notable exception, where extensive exposures of sediments that underlie the diamict occur along the valley walls (Fig. 2.7a). Throughout the study area, an emerging relationship exists between the elevation of the base of the surficial diamict and the substrate type observed. At higher elevations (usually above 270 m asl), the base of the diamict is commonly underlain by thick successions (up to 30 m) of sand, whereas fine-grained silt and clay rhythmites are typically exposed beneath the diamict at lower elevations.

It is often possible to trace the upper surface of the diamict from the uplands down into the shallow subsurface within the valleys by walking out the contact in road and ditch cuts or basement excavations in new housing developments and augering shallow holes. Exposures of the diamict near these occurrences, and presumed to be correlative, are also seen in nearby stream and river cut banks (Fig. 2.7a,c). The diamict exposed in outcrops within the valleys exhibits similar physical properties and paleoflow indicators (striated and faceted boulders, shear planes, etc.) to those observed on uplands (Fig. 2.6c-f; Mulligan et al., 2018). However, fissility is generally not as well developed and increased interbedding with sorted sediments (ranging from clay to gravel) and greater variations in diamict matrix textures and clast content are observed. Striae and long-axis orientations of faceted, bullet-shaped boulders are consistent with those observed on the uplands (Fig. 2.7d) as well as regional drumlin trends. Drumlins that rise up above the flat-lying valley floors (Figs. 2.7, 2.8) are oriented parallel to and display similar morphologies to those observed on adjacent uplands.

2.4.2. Valley geometry and spatial relationships

Valleys separating drumlinized uplands in the study area commonly have north-northeast to south-southwest orientations, but the spatial relationships of individual valleys and valley-upland transitions are complex. CV and HMV (Figs. 2.1, 2.4-2.7) run nearly north-south in their upflow ends, then deflect toward the southwest, parallel to drumlins on the uplands, in their downflow ends to the south (Fig. 2.8). Parts of some valleys (e.g. AE; Fig. 2.8) run almost exclusively north-south, and merge with (or are cross-cut by) other valleys (e.g. BV, CV; Fig. 2.8). In the north, valleys trend east-northeast to west-southwest (e.g. BV; Fig. 2.8), or deflect toward the northwest (e.g. WCV; Fig. 2.7). Further complexities in the spatial arrangement of valleys in the north arise due to the presence of extensive wetlands that occupy the Minesing basin (Fig. 2.8), where a lack of subsurface information hinders evaluation of the relationship between valleys and broad lowland areas to the north and those observed in the study area.

Fig. 2.8: a) morphology of the flanks of tunnel valleys overlain on shaded relief map (5m cell size). Note the common drumlinization of valley flanks (green), particularly in upflow areas, the perched head of the CV and downflow end of the WCV (brown), and abrupt start of the WCV (orange) in Wisconsin age sediment. Many areas with eroded flanks (red) have been emphasized by nearshore erosion in deglacial lakes. Locations of Figs b and c outlined in black; b) close up view of the morphology of the flanks along the upland separating the BV and WCV. Note the large bar and spit features and abundance of aggregate pits, which indicate significant erosion of the flanks of the uplands following deglaciation, but white dashed lines show extent of surficial exposures of Newmarket Till below the former shorelines; c) close up view of the morphology of the central part of the HMV and ABV. The flat surface within the valleys is due to the thick infill of glaciolacustrine sediments. Steep slopes at the transition between the flat floors and the valley flanks are partially due to intense wave erosion in glacial Lake Algonquin; a subtle raised delta surface is visible where the E. Holland River meets the flat valley floor. Note the significant variation in valley width over a short distance downflow. The morphology of valley flanks alternates from drumlinized to eroded, and several large groundwater piping scars are visible, especially along the southeastern flank. The relationship between the HMV and ABV is incompletely understood (but see Gerber et al., 2018).


a
b
WCV
MB
BV
Kempenfelt Bay
Lake Simcoe
7.3 km
AE
4.8 km
2 km
CV
6.3 km
Cook's Bay
HMV
ABV
ORM
abrupt start
drumlinized floor
drumlinized flanks
piping scar
perched valley
eroded flanks
= buried/obscured
0
km
15
baymouth bar
BH-31
2.1 km
wave-cut bluffs
Lake Simcoe
3.9 km
CS-17-08
spit
Elevation (m asl)
315
180
c
Elevation (m asl)
356
216
wave-cut bluffs
7.1 km
delta
drumlins
drumlins
wave-cut bluffs
2.6 km
SS-12-07

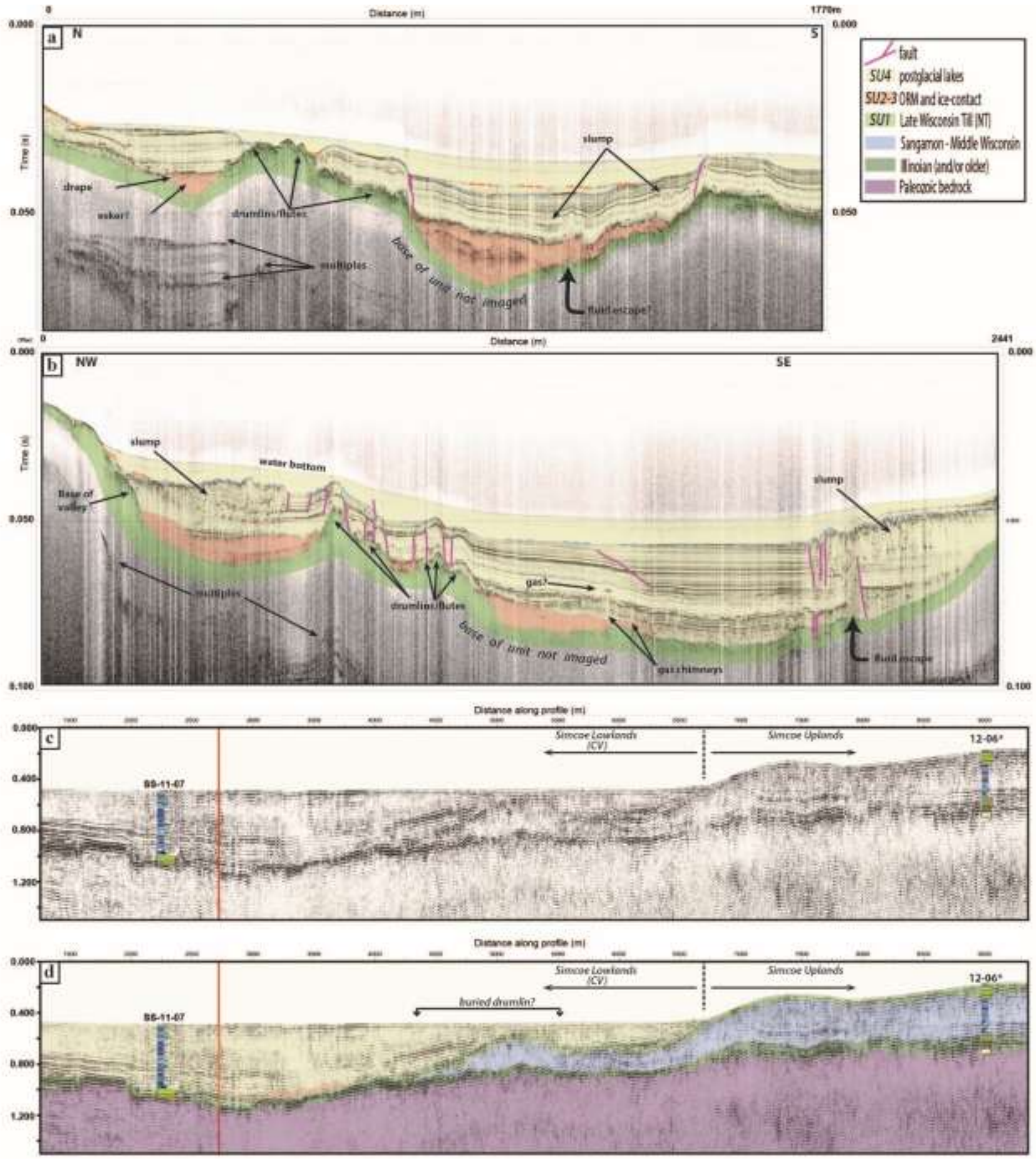
Many of the upflow parts of valleys running through Simcoe County begin beneath modern waterbodies (Fig. 2.8a), rendering analysis of their upflow ends (as well as full valley lengths) difficult, although sub-bottom profiles in Lake Simcoe have provided insights into the geometry of the submerged parts of valleys (Todd et al., 2008; Fig. 2.9). The upflow ends of the valleys appear to lie in areas of thin drift, commonly near the contact between Paleozoic and Precambrian rocks to the north of the study area (Fig. 2.1). Two exceptions include the WCV and the CV (Fig. 2.8a), which originate in areas of thick glacial sediment cover (>150m; Fig. 2.1b). Their bases are floored by Newmarket Till, which can be observed in surficial exposures (Figs. 2.7a, 2.14) and they lie as much as 80 m above adjacent valley segments, with paths that are nearly orthogonal to the axes of WCV and CV.

The geomorphic complexity of the relationships between the valleys in the study area is well illustrated by the WCV (Figs. 2.1, 2.8). It is the only valley with a WNW long-axis orientation, parallel to a poorly-developed set of drumlins and flutings developed on uplands northwest of Lake Simcoe (Barnett, 1992b; Mulligan, 2017b). It has a curved course, with an abrupt upflow end, running first southerly, and then deflecting westward and eventually northwestward, where it ends perched high above the base of a much deeper incised south-southwest oriented valley (Fig. 2.The character of the valley flanks changes downflow – in the upflow part, a sloping, drumlinized till plain occurs on the uplands outside of the valley as well as on its flanks (Fig. 8a), before dipping down below glaciolacustrine sediments that partially infill the valley. Downflow, the valley margins are steep and highly dissected (Figs. 2.7, 2.8a) exposing a truncated till high in the valley walls and thick sandy deposits below (Fig. 2.7a). Despite its truncation high along the valley flanks, till (presumably correlative with Newmarket Till observed on the uplands, based on physical characteristics, clast striae and faceting orientations, and a lack of similar diamicts encountered beneath regional upland strata locally) is observed in stream cut banks and boreholes at elevations well below prominent former glacial lake shorelines within the WCV (Figs. 2.7a, 2.8b, 2.14) as well as in valleys immediately to the north (Barnett, 1992b). Similar downflow trends from drumlinized to eroded valley wall morphology are clear in the BV, CV and HMV (Figs. 2.7, 2.8), whereas the AE (Figs. 2.1, 2.8) shows almost entirely drumlinized flanks, except in small areas where till is eroded in groundwater piping scars and wave-cut bluffs (Fig. 2.8; Mulligan and Bajc, 2012). Seismic reflection suggest the drumlinized flanks of the CV extend into the subsurface beneath younger glaciolacustrine sediments (Fig. 2.9)

Valley width ranges between <1 and >7 km but no consistent trend within the study area or within individual valleys exists (Fig. 2.7). The WCV and BV (Figs. 2.1, 2.8) display variable widths along their lengths (2.6 to 7.1 km wide at their base at surface). In the AE, CV and HMV, both broad and severely constricted segments of the valleys are observed (Fig. 2.8a). These variations show no obvious relationship to length along the valley, orientation/curvature of the valleys, or ice-marginal landforms, which are generally absent within the study area, except for a few ice-contact slopes and minor moraines and/or crevasse fill ridges mapped near Barrie and west of Lake Simcoe (Barnett, 1992b; Mulligan and Bajc, 2012). The constrictions in the valleys tend to occur in areas where drumlinized till extends into the valley (Fig. 2.8a).

CV and BV align well with, and appear to extend into, prominent re-entrants in the Niagara Escarpment (Fig. 2.1). Above the escarpment, downflow of many of the re-entrant valleys, thick accumulations of sand and gravel occur locally, either as eskers (Straw, 1968; Kor and Cowell, 1998, Mulligan, 2015) or as stacked subaquatic fans/kames comprising large

Fig 2.9: a,b) High-resolution CHIRP sub-bottom profile across the eastern part of Kempenfelt Bay (submerged eastern extension of the BV; see Figs. 2.1, 2.4-2.8). A strong undulating reflector rises towards the northern and southern shorelines of the modern bay and is characterized by abundant smaller-scale (few m) ridges (drumlins/flutes?) on local highs beneath the valley. The drumlinized reflector is locally overlain by transparent or chaotic facies interpreted as coarse-grained sediments. An anticlinal reflector (esker?) is draped by horizontal strata (subaquatic fan?) in a). Well-laminated strata comprise the bulk of the sediment infill (modified from Mulligan, in prep); c) uninterpreted and d) interpreted S-wave seismic reflection profile (data from Pugin et al., 2018) with borehole logs and surficial mapping as geologic constraint (Figs. 2.6, 2.7). SS-11-07 is 200 m south of profile and SS-12-06 is projected from 2.7 km south for reference. A mound capped by strong amplitude reflectors shows similar 2D geometry to drumlins on uplands and is interpreted as a buried drumlin capped by NT (SU1). See Fig. 2.6 for additional information.



a
N
S
Distance (m)
1770m
Time (s)
fault
SU4 postglacial lakes
SU2-3 ORM and ice-contact
SU1 Late Wisconsin Till (NT)
Sangamon - Middle Wisconsin
Illinoian (and/or older)
Paleozoic bedrock
slump
drape
esker?
drumlins/flutes
multiples
base of unit not imaged
fluid escape?
b
NW
SE
2441
slump
water bottom
Base of valley
multiples
drumlins/flutes
gas?
gas chimneys
fluid escape
base of unit not imaged
c
Distance along profile (m)
Simcoe Lowlands (CV)
Simcoe Uplands
12-06*
BS-11-07
d
buried drumlin?

moraine complexes (Burt, 2018). HMV has a more southward course and is buried beneath or extends south of the ORM (Sharpe et al., 2018).

2.5. SUBSURFACE SEDIMENTS IN VALLEYS

Four stratigraphic units (SU1-SU4; Figs. 2.6, 2.9-2.12) overlie pre-Late Wisconsin deposits in the study area and comprise the floor and infill succession of the drumlinized tunnel valleys cut into both unconsolidated sediment and Paleozoic bedrock substrates. A moderately- to over-consolidated silty sand diamict with local sand, silt and gravelly interbeds (SU1) commonly forms the floor and flanks of the valleys and is correlated with till outcropping throughout drumlinized upland terrain (described above). Gravel (SU2) locally overlies, or fills incisions cut into SU1. Both lower units are locally overlain by medium- to coarse-grained sand passing upward into very fine-grained sand with interbedded silt-rich diamict beds (SU3). Fine-grained silt and clay rhythmites with clasts and thin silty diamict interbeds (SU4) drape the underlying units and form the bulk of the sediment infill within the valleys. SU4 grades upwards into organic-bearing fluviodeltaic, distal glaciolacustrine, and lacustrine sediments that cap the sediment sequence.

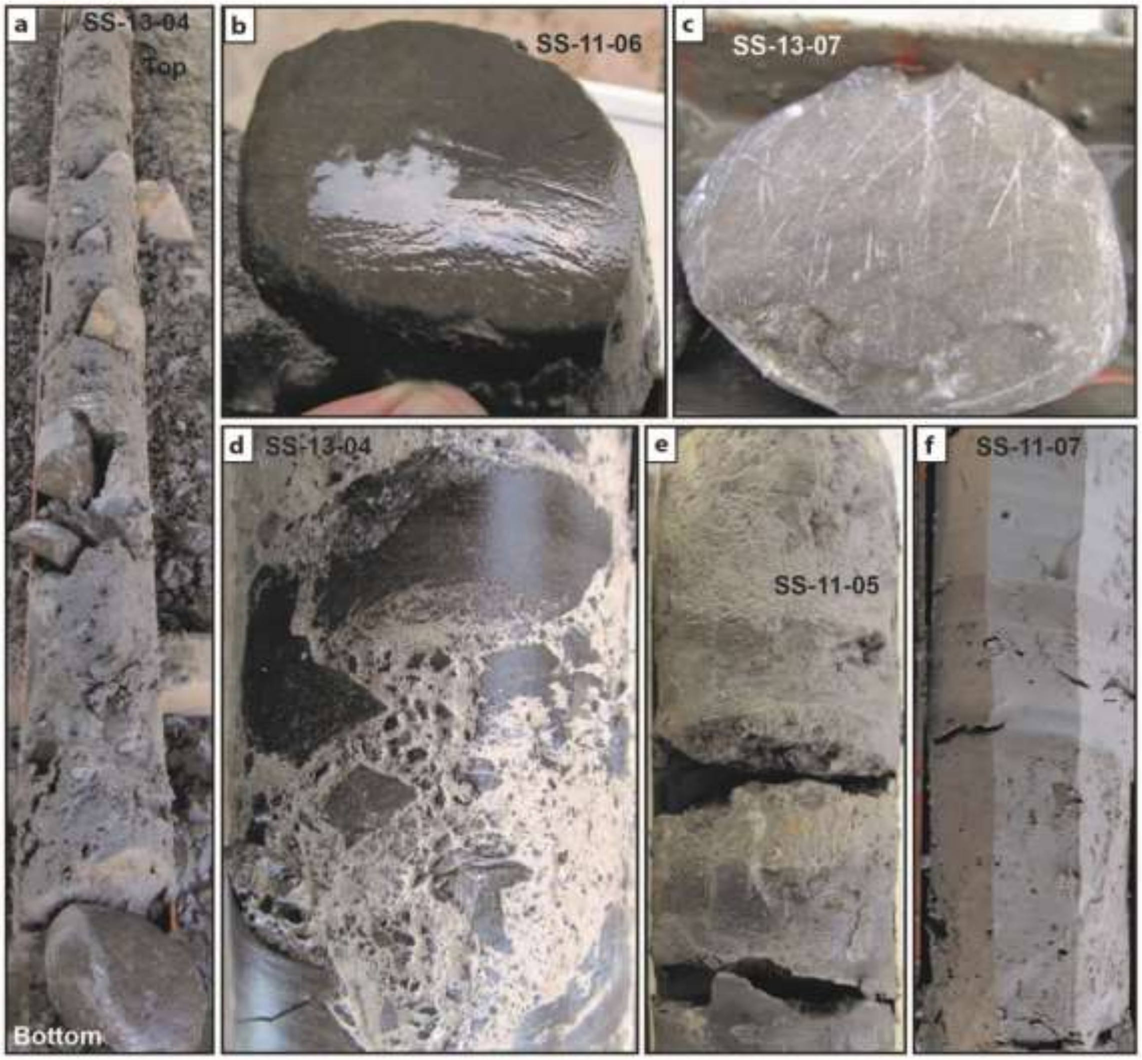
2.5.1. SU1 – The Newmarket Till

A grey to brown-grey silty sand to sandy silt diamict, up to 15 m thick, floors the sediment succession within many of the valleys in southern Simcoe County (Figs. 2.6, 2.7, 2.9, 2.10). Matrix consolidation is variable but generally ranges from moderately- to over-consolidated with rarer, loose or poorly consolidated facies becoming more common in the upper parts of the unit. Clast content ranges between 5 and 15% with locally-derived Paleozoic carbonate and shale clasts comprising 70-95% of the lithologies observed (Fig. 2.10a). Paleozoic clasts are subangular to subrounded and commonly have sub-horizontontal upper and lower surfaces with either parallel or chaotic striae orientations (Figs. 2.7, 2.10b,c). Clasts are commonly more angular at the base of the unit when SU1 is observed directly overlying bedrock (Figs. 2.6, 2.10d,e). In these zones, the local bedrock comprises a large proportion of the observed clasts in the lower 50-100 cm of the unit. Evidence of clast crushing and disaggregation is abundant (Fig. 2.10d), locally forming linear, sub-horizontal, or rotational clast trains of small broken material emanating from a larger parent clast. The basal contact with soft bedrock lithologies is highly irregular, appearing to be interbedded in some boreholes (Figs. 2.10e, 2.14a).

Although massive (homogenous) in places, SU1 is most commonly composed of a stacked sequence of massive diamict beds (0.5 – 4 m thick) either directly juxtaposed, or separated by sorted interbeds generally 5-20 cm thick but, in places, up to 1 m thick (Fig. 2.6). Variations in the characteristics of diamict beds is due to the juxtaposition of silt-enriched and –depleted facies and clast-rich and clast-poor facies that have generally the same overall colour and matrix consolidation (Figs. 2.7, 2.10). Bed contacts are generally sharp, undulating or planar, and locally deformed, although some sand or silt beds underlying or within SU1 become increasingly clast-rich upwards transitioning into massive diamict beds of SU1.

Sorted interbeds within SU1 consist primarily of fine- to coarse-grained sand and silt with lesser gravelly sand. Sands are generally planar bedded or ripple cross-laminated and silts are finely planar laminated with both facies typically being highly deformed with pervasive folding and faulting (high-angle reverse and low-angle normal step faults) of laminated beds. Gravelly sand facies are poorly recovered in core but exposures of likely correlative deposits in outcrop reveal well-preserved foreset beds indicating paleoflow toward the west. Organic-bearing deposits occurring above and below SU1 and intersected in drill core constrain deposition of SU1 to after 28 060 ± 230 and before 12 805 ± 40 ^{14}C yr BP (Fig. 2.6; Bajc et al.,

Fig. 2.10: SU1 facies. a) cobble-rich, dense, massive diamict; b) clast with sub-parallel striae and gouges; c) clast with chaotic striae, likely indicating transport in a mobile deforming bed; d,e) monolithic, angular clasts immediately above shale bedrock; f) silty upper facies sharply overlain by fine-grained rhythmites with dropstone layers (SU4). See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.


a
SS-13-04
Top
Bottom
b
SS-11-06
c
SS-13-07
d
SS-13-04
e
SS-11-05
f
SS-11-07

2014; Appendix). SU1 is commonly overlain by SU3 deposits, but is directly overlain by SU4 in several areas (Figs. 2.6, 2.10f). It is thin or absent where gravelly deposits of SU2 is encountered.

2.5.1.1. Interpretation

The poorly sorted, massive, overconsolidated matrix, combined with striated clasts and boulders suggests SU1 is a subglacial till (Boulton and Deynoux, 1981; Benn and Evans, 1996; Slomka et al., 2015). Clast disaggregation features and evidence of clast crushing and incorporation of local bedrock suggest abrasion and active transport in a mobile subglacial bed (Evans et al., 2006). Rotational clast structures and poorly-developed clast fabrics may indicate that bed deformation was important locally and/or transiently during till deposition (Piotrowski and Klaus, 1997; Roberts and Hart, 2005). Interbedding of various diamict facies and increased proportions of sorted interbeds within SU1 in the valleys are consistent with a complex depositional history (Hicock and Dreimanis, 1992). We suggest this likely involved repeated episodes of meltwater build-up and discharge, resulting in fluctuations in sediment strength and effective pressures at the ice-bed interface within the valleys compared to intervening upland areas.

SU1 is interpreted as the Newmarket Till, the regional Late Wisconsin Till sheet in south-central Ontario (Gwyn and Dilabio, 1973; Sharpe et al., 1994; Sharpe and Russell 2016; Mulligan et al., in press). This correlation is based on its physical and chemical properties (lithology, facies, consolidation, clast and striae orientations, heavy mineral signature; Bajc et al., 2015), combined with its mapped continuity from drumlinized uplands to river exposures within valleys. Additionally, stratigraphic analysis and chronostratigraphic markers in several boreholes indicate the Sangamon Episode subaerial unconformity is locally preserved beneath the base of the valleys and no tills have been encountered stratigraphically between Illinoian and Late

Wisconsin Newmarket Till in the area (Mulligan and Bajc, 2018; Fig. 2.6). Regionally, the base of the Newmarket Till is undulating, and is interbedded with older sediments and/or has a deformed basal contact (Boyce et al., 1995; Mahaney et al., 2014; Mulligan and Bajc, 2017), or may be planar and sharp, recording local bed decoupling (Piotrowski et al., 2006), erosion at the base of the ice sheet (Eyles et al., 2016) or passive sedimentation (e.g. Pugin et al., 1999; Sharpe et al., 2018).

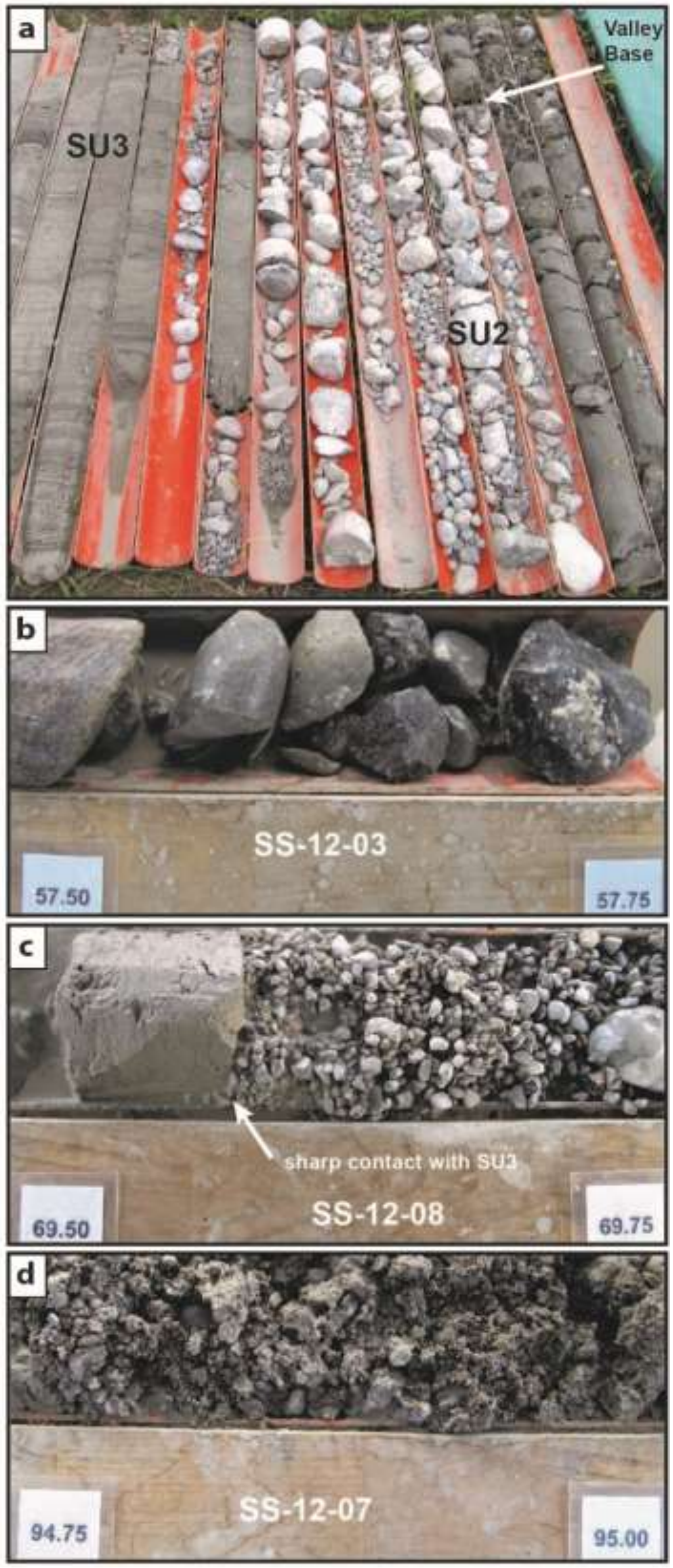
2.5.2. SU2 – valley-fill gravel

SU2 consists of gravelly sediments located at the base of the valley infill succession (Figs. 2.6, 2.9, 2.11). These sediments were not encountered in all valley borings and were poorly recovered in core. The maximum observed thickness of SU2 is 12 m and, in most cases, little matrix was recovered (a reflection of challenges associated with mud rotary drilling methods), complicating any attempt to characterize the bedding within SU2 (Fig. 2.11a). Stratification of pebble- and cobble-rich facies is suggested by their discrete zonation in core (Fig. 2.11a-c). In general, larger clasts were encountered at the base of the unit than at the top, suggesting an overall fining-upward succession in SU2. The upper part of SU2 is often more poorly sorted, with higher proportions of matrix fines recovered (Fig. 2.11d). Subangular to rounded clasts, up to 15 cm in diameter, were observed and consist of a mixture of carbonate, shale, and shield lithologies (Fig. 2.9b) with carbonate clasts being dominant. SU2 is always overlain by SU3 (Fig. 2.6); the upper contact may be sharp (e.g. Fig. 2.11c), gradational, or interbedded (Fig. 2.11a).

2.5.2.1. Interpretation

Poor overall recovery of SU2 and a lack of geometric control on its internal facies and variability prohibit detailed interpretations of its origin. Its restricted spatial extent within the central, deeper parts of valleys suggests a confined depositional environment (Ghienne and

Fig. 2.11. SU2 facies. a) cobble-rich SU2 facies showing an interbedded contact with overlying SU3 deposits. Matrix material was poorly recovered, but there is apparent stratification of pebble- and cobble-rich facies; b) close-up of clasts in SU2 showing large, sub-rounded clasts of diverse lithologies; c) openwork pea gravel at the top of SU2 directly overlain by silty very fine-grained sand (SU3); d) poorly sorted gravel with coarse-grained sand matrix near the base of SS-12-07. See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.

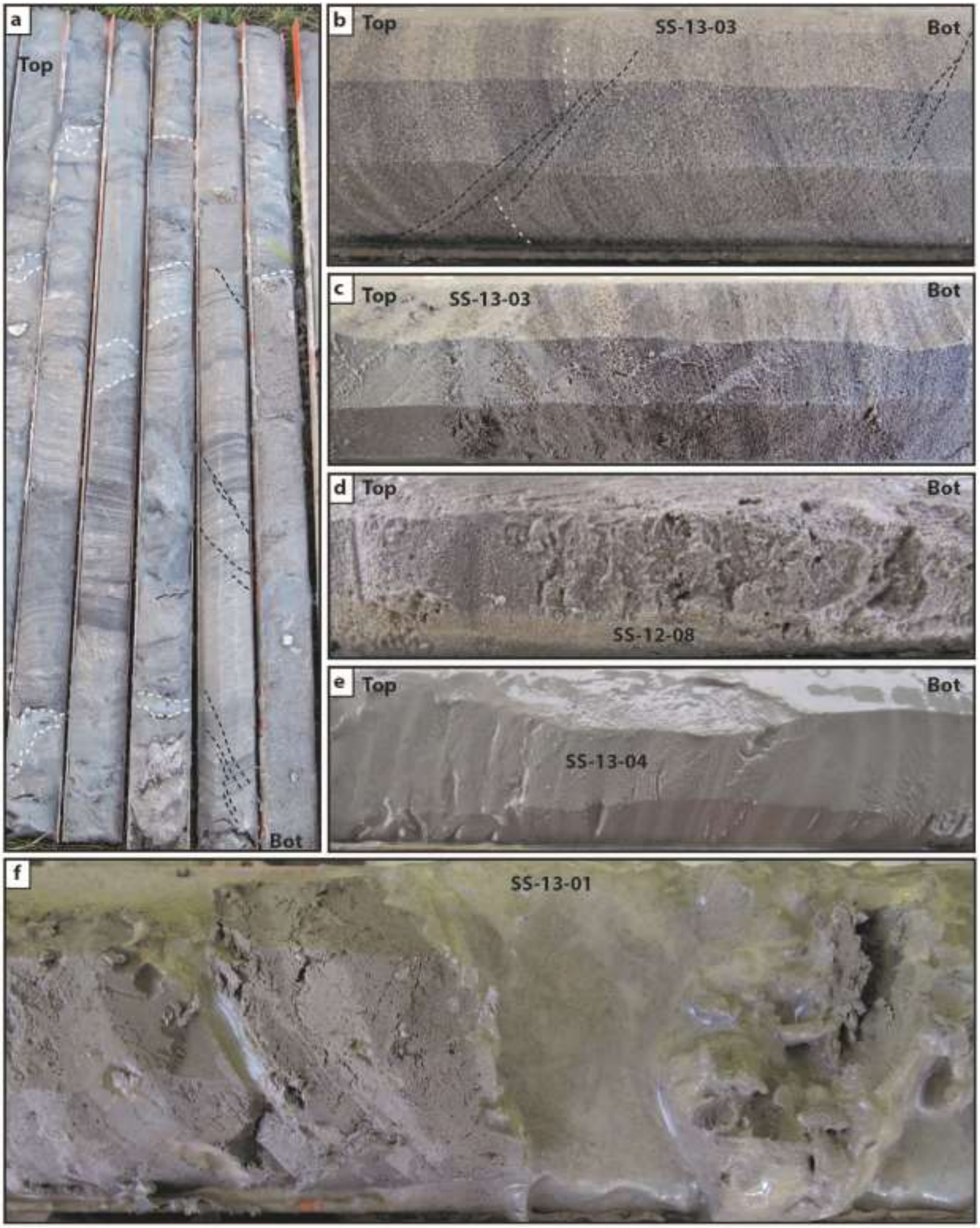

a
SU3
Valley Base
SU2
b
SS-12-03
57.50
57.75
c
sharp contact with SU3
SS-12-08
69.50
69.75
d
SS-12-07
94.75
95.00

Deynoux, 1998) and large, rounded cobbles are consistent with moderate to high magnitude flow velocities (Cutler et al., 2002). From the information available from boreholes, it is not possible to confidently discern subglacial vs. proglacial setting for deposition of SU2. Fining upward trends in average clast size and the gradational upper contact between SU2 and SU3 are consistent with waning flow velocities during deposition. These relationships are typical of many flood/outburst deposits (e.g. Russell et al., 2003; Russell et al., 2007), but we interpret them here to record the transition from esker to subaquatic fan setting (see description of SU3 below; Fig. 9), recording the transition to a more ice distal conditions during glacier retreat (Krzyszkowski and Zieliński, 2002; Winsemann et al., 2007). Interbedded facies record fluctuating flow rates, produced by meltwater pulses and/or local ice marginal fluctuations during deposition of SU2.

2.5.3. SU3 – valley-fill sand

SU3 occurs as either sharply overlying SU1 or transitionally overlying SU2 (Figs. 2.6, 11a). Thick fining upward sequences, locally >40 m thick (boreholes 11-09, 13-03, 17-04; Fig. 2.6), were intersected in boreholes and typically grade from massive to planar-bedded, pebbly medium- to coarse-grained sand with heavy mineral banding (Figs. 2.6, 2.12a-c), into ripple-scale cross-laminated (Type-A and B; Jopling and Walker, 1968), planar laminated and deformed medium-fine- to very fine-grained sand with interbedded silt at the top (Fig. 2.12c-e). Coarse-grained sands near the base of the unit are composed of stacked, fining-upward beds displaying low-angle truncations and faulting (reverse and normal) with vertical offsets of up to 5 cm (Fig. 2.12a,b). Sand-rich diamict beds up to 50 cm thick are locally encountered near the base of SU3. In the finer-grained upper facies, soft-sediment deformation features are abundant; ball and pillow and flame structures up to 30 cm in amplitude commonly mark the contacts between fine sand and silt beds. Large contorted beds (folded) are common within sand packages (Fig. 2.12a). SU3 is generally well or very well sorted except in the upper few meters

Fig. 2.12: SU3 facies. a) upward transition from planar laminated with well-developed heavy mineral laminations and abundant faulting (dashed black lines) into interbedded fine-to medium-grained sand and silty sand, with deformed contacts (dashed white lines). Rare pebbles are dispersed throughout; b) close up of medium-grained sand bed with normal fault offsets up to 3.5 cm; c) transitional facies within SU3 showing the introduction of interstitial fines upward in the succession; d) massive to faintly horizontally stratified fine-grained sand with gas bubbles; e) rhythmically-bedded very-fine grained sand and silt couplets in the uppermost part of SU3; f) stratified basal contact of diamict capping SU3. Note the dipping beds and small-scale deformation along sharp bed contacts defined by light and dark colour changes; See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.


a
Top
Bot
b
Top
SS-13-03
Bot
c
Top
SS-13-03
Bot
d
Top
Bot
SS-12-08
e
Top
Bot
SS-13-04
f
SS-13-01

where interstitial silt content increases (Fig. 2.12e) and within diamict beds up to 6 m thick that locally mark the transition from SU3 to overlying SU4 (Figs. 2.6, 2.12f). Diamict beds within SU3 are distinguished from the thin (10 cm) fine-grained diamict beds near the base of overlying SU4 deposits (described below), based on their coarser matrix texture, higher degree of matrix consolidation, low clast content (1%), massive to very crudely stratified and deformed matrix, and abundance of variably-textured interbeds, separated by sharp to gradational, commonly deformed contacts (Fig. 2.12f). Clasts are generally small (<5 cm) and show no obvious preferential alignment in cores. The basal contact with underlying sands may be gradational, interbedded, or sharp, and is commonly deformed. Sharp contacts with silt and clay interbeds up to 5 cm thick within the upper part of SU3 are marked by small-scale brittle and ductile deformation structures including normal and reverse microfaulting, ball and pillows, and minor shearing structures (Fig. 2.12f).

2.5.3.1. Interpretation

Sands in the lower part of SU3 were deposited in moderate to high energy subaqueous environments, likely in ice-marginal subaquatic fans (Rust and Romanelli, 1975; Fig. 2.9). Planar laminated, coarse-grained sand is interpreted to record laminar to turbulent flow from an upper flow regime (Clerc et al., 2013), possibly as hyperconcentrated density flows (Mulder and Alexander, 2001). Low angle truncations record non-uniform flow conditions, producing scours associated with channel migration or temporal changes in flow velocity/geometry (Russell et al., 2003). It is argued here that restricted penetration of the faults (<20 cm) points to repeated discrete events rather than a single or continuous event and may record combination(s) of gravity destabilization during fan spreading (Ravier et al., 2014b), deformation by grounded ice regaining contact with the bed (Denis et al., 2010), or disturbance from minor readvances/oscillations of the ice front (Weaver and Arnaud, 2011).

Upward fining in SU3 records a reduction in flow competence. Soft sediment deformation features in the upper finer part of SU3 are consistent with sustained rapid sedimentation rates and record sediment remobilization through water escape, slumping, or shock from minor earthquakes following glacier retreat (Weaver and Arnaud, 2011; Pisarska-Jamrožy and Weckwerth, 2013; Doughty et al., 2014). Thick packages of rippled sands (Type-A and B) record persistently high sedimentation rates with increasing proportions of sediment contribution from suspension (Jopling and Walker, 1968; Jobe et al., 2011). The up-section increase in silt abundance, both in discrete beds and interstitially, records at least local cessation of coarse-grained sediment delivery due to either shutting off of the flow system, channel avulsion, or temporary re-configuration of the drainage network (Winsemann et al., 2007). The overall upward-fining trend in SU3 records decreasing influence of meltwater flows into the lake basin. It is likely that the decrease in meltwater activity is what permitted deposition of the diamicts observed in the uppermost parts of SU3. They likely record subaqueous sediment gravity flows emanating from the retreating ice margin and/or the steep slopes along the valley walls (Barnett and Karrow, 2018; Mulligan et al., 2018).

2.5.4. SU4 – glaciolacustrine rhythmites

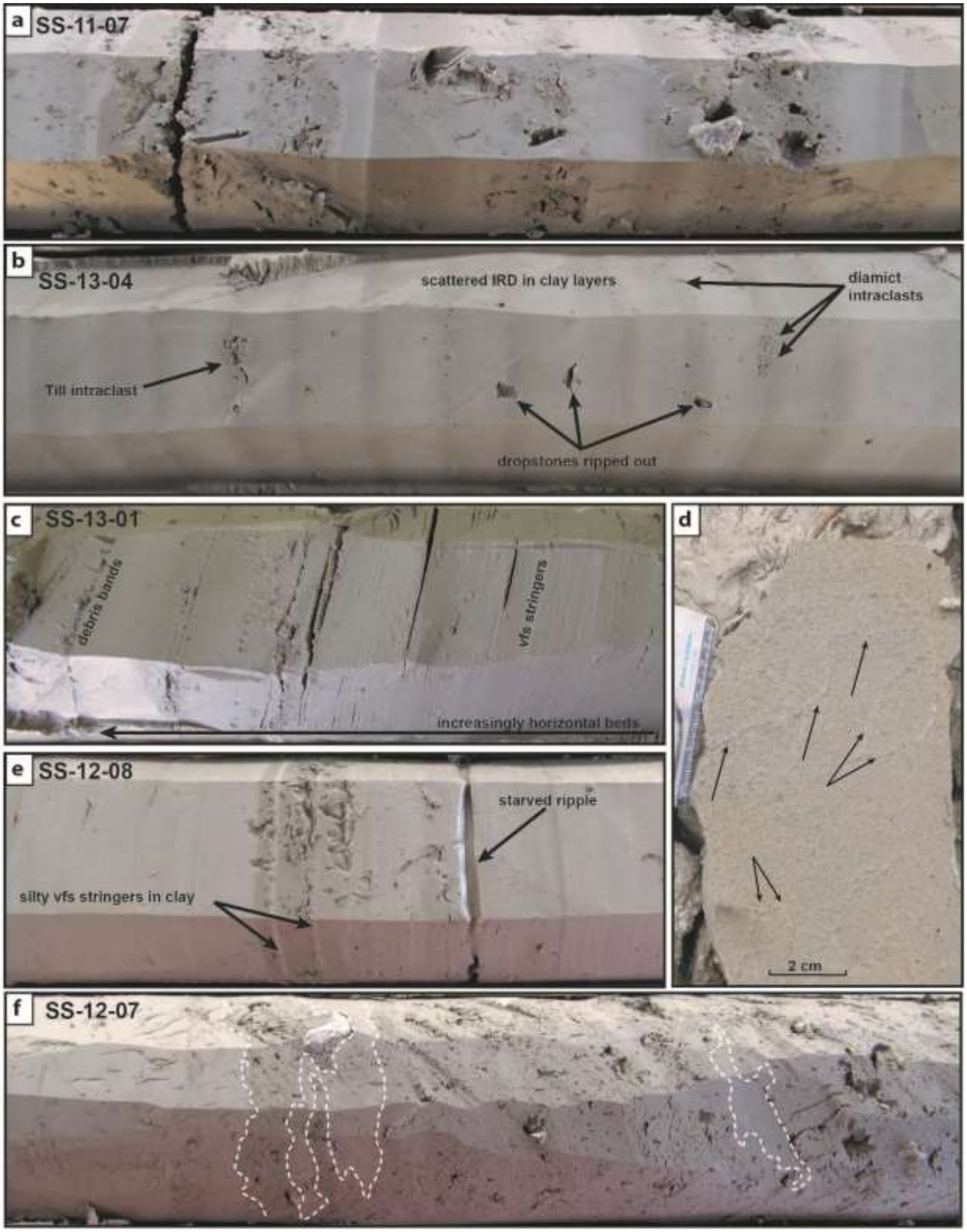
SU4 is characterised by fine-grained, planar-laminated rhythmites consisting of alternating beds of silt and clay (0.5-5 cm thick) that form couplets <1 - 27 cm thick. The rhythmites are very soft (resulting in poor recovery of some intervals, particularly if cobbles were encountered by the drill bit) and commonly comprise the bulk of the sedimentary infill within the valleys (Figs. 2.6, 2.9). The character of the lower parts of SU4 and its basal contact with underlying units (SU1 or SU3) is highly variable. Undisturbed, uniform couplets of silt and clay with abundant ice-rafted debris (IRD) directly overlie SU1 in many surficial exposures as well as in cores within the valleys (SS-11-05, -06, -07, CS-15-08 and CS-17-08; Figs. 2.6, 2.13a,b).

The rhythmites drape the undulating upper surface of SU1 but local, low-angle truncation horizons up to 10 m wide are observed in some outcrops where the transition between the two units is well exposed. Silt-clay couplets in the lower few meters of SU4 are interrupted by massive to stratified clayey silt diamict interbeds up to 10 cm thick (Fig. 2.13a,b). The diamict interbeds generally have sharp bases and grade upward into finely laminated clay (Fig. 2.13a). Clasts are common throughout both the fine and coarse layers of rhythmites in the lower part of SU4 (Fig. 2.13b). They account for a small proportion of the sediment (<1% by volume) and their abundance decreases rapidly up-section (Fig. 2.13c). At several locations in the study area, horizontal feeding traces are observed along the silty bedding planes within SU4, even within the lower 0.5 m of the unit where clasts and diamict interbeds are common (Fig. 2.13d).

In one borehole (SS-12-07; Figs. 2.6d, 2.13e), a 2.5-m thick, massive, diamict package was intersected 13 m above the base of SU4. It bears similarities to diamict beds that cap SU3 (see above) and is distinguished from the thin beds near the base of SU4 described above based on its thickness, degree of consolidation relative to adjacent deposits, higher clast content and larger average clast size, its massive character and sharp upper contact with overlying rhythmites.

Where SU4 overlies SU3, the contact may be abrupt or interbedded with multiple facies types observed over thicknesses of 1-3 m along the transition between the two units (Fig. 2.13c). Thin interbeds (0.5 - 5cm thick) of ripple cross-laminated to planar laminated very fine- to fine-grained sand are common near the base (Fig. 2.13c,d). The sands decrease in thickness and abundance upwards forming a gradational transition between the two units. Thin diamict

Fig. 2.13: SU4 facies. a) rhythmically bedded silt and clay with diamict interbeds up to 5 cm thick and abundant till intraclasts; b) rhythmically bedded silt and clay with rare debris (IRD and till intraclasts) in both the silt and clay beds; c) finely laminated silt and clay with clay intraclasts at the base, passing upwards into finely laminated silt with abundant very fine-grained sand (vfs) stringers and horizontally bedded silt and clay with thin debris bands rich in sand and granules; d) horizontal trace fossils (arrowed) along a silty, debris-rich bedding plane near the base of SU4; e)finely laminated silt and clay with silty vfs stringers and starved ripples. Thickness of laminated silt separated by clay-rich beds (dark colours) is approximately 15 cm; f) fine-grained moderately- to well-consolidated, crudely stratified and deformed diamict within (interrupting) SU4. Dashed white lines denote deformed contacts and incorporation of clayey silt in the uppermost part of the diamict; See Fig. 2.5 for borehole locations, cores are approximately 8.5 cm wide; see Fig. 2.6 for photo locations.


a SS-11-07
b SS-13-04
scattered IRD in clay layers
diamict intraclasts
Till intraclast
dropstones ripped out
c SS-13-01
debris bands
vfs stringers
increasingly horizontal beds
d
2 cm
e SS-12-08
starved ripple
silty vfs stringers in clay
f SS-12-07

interbeds that are common where SU4 overlies SU1, are absent or rare. SU4 grades upwards into organic-bearing fluviodeltaic, distal glaciolacustrine, and lacustrine sediments that provide a record of changing environments during later phases of deglaciation (Mulligan et al., 2018).

2.5.4.1. Interpretation

SU4 records suspension settling of fines in a low-energy subaqueous environment (Ridge et al., 2012; García et al., 2015). Silt and clay couplets record alternating periods of underflow and suspension settling of clays (Ashley, 1975). Sandy interbeds near the base of SU4 may record the final phases of flow from subaquatic fans originating from the retreating ice margin (SU2 and SU3; Gravenor and Coyle, 1985) or from prograding fluviodeltaic systems entering the former lakes and/or intense storm events reworking the unconsolidated and largely unvegetated uplands surrounding the basin (Schaetzl et al., 2016; Mulligan et al., 2018). The presence of trace fossils, gradational and/or interbedded contacts commonly observed between SU3 and SU4, and the gradational upward transition of SU4 into mollusk-bearing fluviodeltaic deposits (see Mulligan et al., 2018) implies continuous, uninterrupted sedimentation between the two units and suggests an ice-marginal rather than subglacial setting for SU3 and SU4 (McCabe and O'Cofaigh, 1994; Livingstone et al., 2015). No detailed work has been completed that could definitively determine whether the couplets are varves that represent annual sedimentation cycles (Antevs, 1929). Discrete, thin, debris-rich diamict interbeds represent dumps from ice berg flipping or debris flows originating from the valley walls (Major, 2000; Bennett et al., 2002). Scattered clasts near the base of the unit record rainout of debris transported into the basin from floating ice, either as icebergs calved off the retreating ice margin, or as seasonal ice cover or ice shelf (Eyles and Eyles, 1983; Martini et al., 1993; García et al., 2016).

A subaqueous ice-marginal environment for the lower part of SU4 is also supported by the presence and characteristics of silty diamict beds that separate SU3 and SU4, which are interpreted to record subaqueous debris flows derived from either glacial sources (Shomacker and Kjaer, 2008) or the flanks of the valleys (Janszen et al., 2013). During periods of higher meltwater inputs (i.e., during deposition of SU2-3), the preservation potential of debris flows along the base of the valley would be low, compared to later phases when the flows waned and/or ice retreated. Similar diamict facies are observed in surface exposures throughout the study area (Mulligan and Bajc, 2012; Mulligan et al., 2018).

2.5.5. Valley substrates

Study of the sediments and bedrock strata occurring immediately beneath the valleys provides additional constraint on formative processes that cannot be gained from analysis of their fills alone (e.g. Ravier et al., 2015a).

2.5.5.1. Borehole Data

The valleys in the study area are locally incised into Paleozoic bedrock that displays striated and grooved upper surfaces when cut into resistant limestone strata (boreholes 11-05, 11-09, 15-08, 17-04; Fig. 2.6). Where the valley bottom rests on softer shale bedrock, the upper units of rock are brecciated and contain a significant number of vertical fractures (Fig. 2.14a). Horizontal intervals of brecciated bedrock containing a sandy silt matrix separated by undisturbed rock are commonly observed in the upper 1.5 m of shale bedrock strata (Fig. 2.14a). Significant mud losses during drilling in the highly brecciated intervals indicate highly transmissive conditions resulting from brecciation.

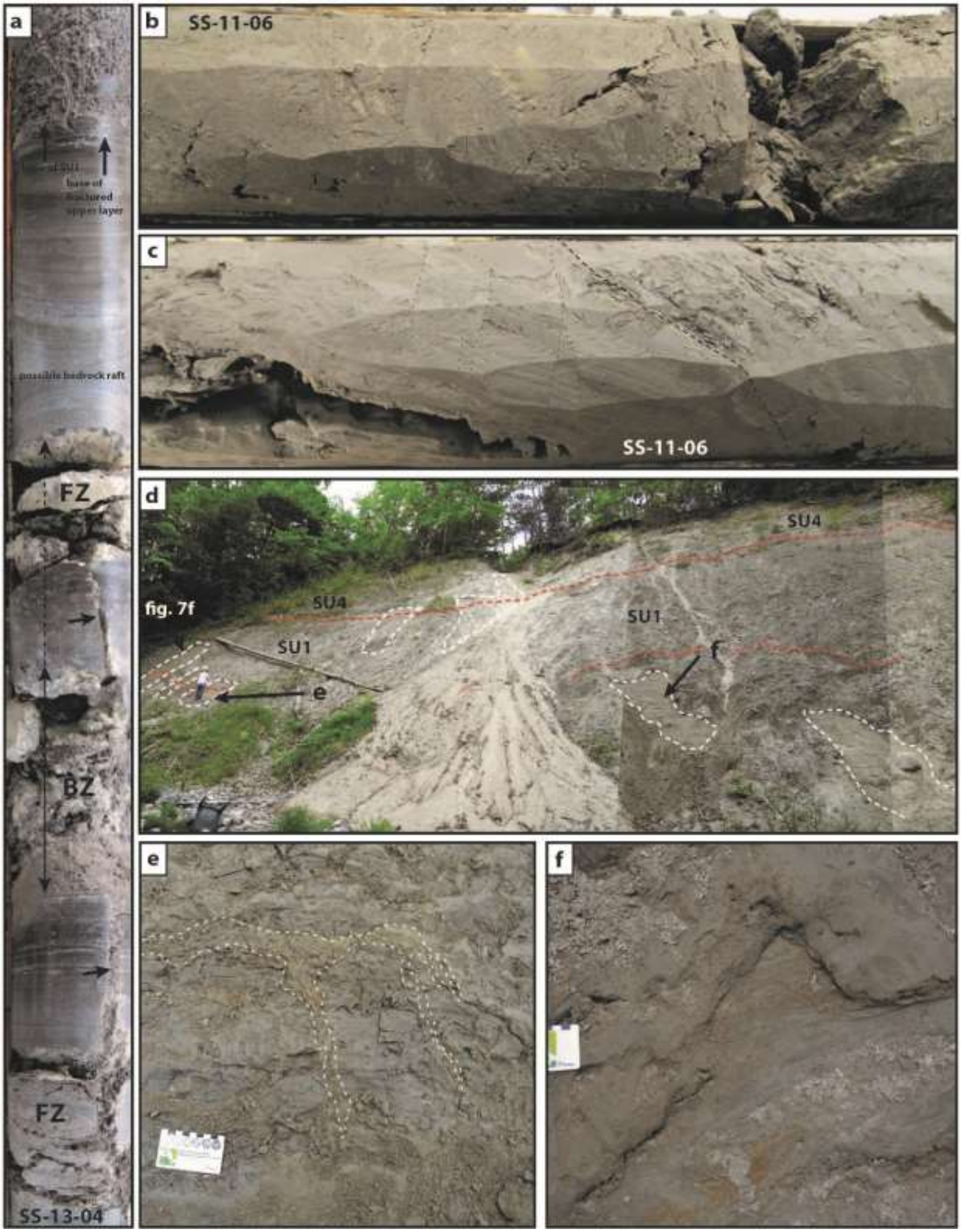
Borehole SS-12-03 provides a record of valley incision and infill in the confluence area of the AE and BV (Figs. 2.5, 2.6c). The base of the valley is eroded into compacted Early-

Middle Wisconsin silt and clay rhythmites which are abruptly overlain by a thin diamict that is likely correlative with the lowermost part of SU1. In SS-11-06, AE incises into compacted rhythmites that are highly contorted; shearing and microfaulting with diverse orientations are observed directly beneath SU1 (Fig. 2.14b), and near vertical bedding, cut by sub-horizontal faults characterize the highly deformed upper 2 m of the rhythmite sequence (Fig. 2.14c). The degree of deformation decreases downwards with shells and delicate plant macrofossils preserved within older non-glacial (Sangamon – Middle Wisconsin) deposits 7 m below the base of the valley.

2.5.5.2. Surface Exposure

Only one near surface exposure of substrate sediment below SU1 was discovered within a tunnel valley setting in Simcoe County. A 25 m high cut bank along Willow Creek, north of Barrie (Figs. 2.1, 2.5, 2.14d) exposes a full succession of SU1 and up to 4 m of underlying substrate sediments. The contact between the substrate sediments and the overlying sediments of SU1 (interpreted as Newmarket Till) is partially interbedded, undulating, and highly deformed. As in borehole SS-11-06, the substrate sediments are predominantly fine-grained rhythmites, with minor very fine-grained sand and clast-rich bands 2-15 cm thick, all of which are highly deformed. The intensity of deformation decreases downward within these deposits to the base of the exposure. Large-scale deformation features within the upper 1.5 m of the substrate material are truncated by sub-horizontal sheared beds and reverse faults. At one location, a discontinuous and deformed lens of silty very fine-grained sand crosscuts subhorizontal beds of silt and clay as downward-thinning injection dikes(?) (Fig. 2.14e). Elsewhere in the outcrop, bedding within the substrate sediments is more chaotic with thicker, deformed (faulted and folded) sands truncated by highly deformed (folded) silt beds (Fig. 2.14f).

Fig. 2.14: Valley substrate material. a) shale bedrock with brecciated beds (BZ), intensely fractured zones (FZ) and deeply-penetrating vertical fracture indicating possible glaciotectonization of the uppermost shale beds following valley erosion. Note the sharp upper contact with monolithic diamict composed entirely of angular shale clasts; b) highly sheared unconsolidated sediments 1.5 m beneath SU1 in sediment hosted valley (AE); c) polyphase deformation of substrate material indicated by near-vertical beds (ductile regime) truncated by sub-horizontal faults (brittle regime), 2 m below SU1 in sediment hosted valley; d) outcrop exposure of SU1 (dashed red lines) and underlying substrate material. Person for scale at left: locations of e,f and till facies shown in Fig. 2.6f marked, see Fig. 2.6a for outcrop location; e) possible injection structure (white dashed line; 0.75 m below base of SU1), indicating downward pressure gradients during or following valley erosion; f) folded clay layer truncating deformed fine-grained sand indicating mobile sediments, likely associated with high relative porewater pressures. See Fig. 2.6 for borehole locations, Fig. 2.6 for photo locations. Cores are approximately 8.5 cm wide, scale bar on cards in e and f is in cm.


a
base of fractured upper layer
possible bedrock raft
FZ
BZ
FZ
SS-13-04
b
SS-11-06
c
SS-11-06
d
SU4
fig. 7f
SU4
SU1
SU1
e
f
e
f

2.5.5.3. Interpretation

Striated limestone bedrock underlying SU1 at the base of the valleys suggests warm-based ice and supports active erosion and debris transport at the glacier sole and debris transport during/following erosion of the valley (Bennett and Glasser, 2009). Brecciated beds in shale strata (Fig. 14a) provide further support for active ice within the valleys and may indicate glaciotectonization of the bedrock strata (see Vaughan-Hirsch et al., 2013). A monolithic array of clasts in the lower parts of SU1 directly above the bedrock contact attests to intense erosion of the substrate by active ice with high strain rates and a strongly coupled IBI (Boulton and Hindmarsh, 1987; Evans et al., 2006; Maclachlan and Eyles, 2011).

Where the valleys are floored by unconsolidated, stratified sediments, a complex strain history is documented. Near-vertical beds truncated by sub-horizontal faults below the valley floor in SS-11-06 (Fig. 2.14c) record changing styles of deformation – vertical bedding was produced by large-scale ductile deformation whereas sub-horizontal faults were generated by a later phase of brittle deformation (e.g. Roberts and Hart, 2005). This change in strain signature is consistent with changing hydrologic conditions (and ice dynamics) at the IBI, with the early phase occurring while the sediments were saturated and under high hydrostatic pressures and the later phase occurring following drainage of the substrate sediments and a resultant increase in sediment shear strength (Evans et al., 2006; Ravier et al., 2015a). Downward propagating and thinning wedges of sediment observed in outcrop are interpreted as injection structures (Ravier et al., 2015b) and suggest that at times, pressure gradients favoured downward recharge of substrate sediments from the IBI (Boulton et al., 2009).

2.6. DISCUSSION

2.6.1. Synthesis

2.6.1.1. Occurrence and distribution of NT (Newmarket Till)

The Newmarket Till forms a regional stratigraphic marker bed that records Late Wisconsin ice advance across southern Ontario – its occurrence, characteristics, and distribution in the vicinity of tunnel valleys in the study area can be used to gain insight into the evolution of subglacial conditions that occurred during tunnel valley genesis in the region. The Newmarket Till (SU1) exhibits significant topographic variation across the study area (>175 m total relief; Figs. 2.6, 2.9), creating a complex regional stratigraphic architecture between uplands and tunnel valleys, where the till overlies Paleozoic Bedrock, older (Illinoian) glacial deposits, or Early-Middle Wisconsin glaciolacustrine sediments (Figs. 2.6, 2.9). Assignment of SU1 in the tunnel valleys to the Newmarket Till is achieved by: tracing the continuity of the till in continuous exposures from the flanks of valleys and/or drumlins in the valleys into the subsurface in cut banks and borehole successions; assessment of stratigraphic relationships, particularly where older tills occur beneath the valley base in the study area (boreholes SS-11-06; SS-12-03; SS-13-01; CS-17-08; Fig. 2.6). Additional support for the correlation of SU1 with the Newmarket Till in the valleys consists of distinct heavy mineral signatures compared to the older sediments (borehole SS-13-07; Fig. 2.6; Bajc et al., 2015); correlation of reflectors observed on offshore and land-based seismic reflection data in the area (Figs. 2.5, 2.9; Todd et al., 2008; Pugin et al., 2018; Mulligan et al., in prep) to borehole strata and surficial exposures; and chronostratigraphic markers from radiocarbon age determinations (boreholes SS-11-06; SS-13-07; SS-13-04; Fig. 2.6; Bajc et al., 2014). The regional architecture and internal properties of the Newmarket Till reported here (Figs. 2.6-2.10, 2.14) conflict with the descriptions of a conformable, planar basal contact by Sharpe et al. (2018) interpreted to record widespread passive sedimentation (as melt-out till) in a subglacial lake within the study area. Sharpe et al. (2018) interpret diamict units encountered at depth within the valleys as debris flows or part of the older (Illinoian) glacial succession, based on textural data and correlation of elevation ranges of the deposits beneath uplands. The wealth of subsurface data presented here indicate a more complex depositional setting for Newmarket Till, and all available information must be considered in an attempt to correlate between uplands and valley settings.

Striated bedrock at the base of the valleys in the study area records active ice erosion subsequent to valley excavation. Combined with fractured, brecciated, and glaciotectonizated shale strata and the multi-phase soft-sediment deformational characteristics of the substrate sediments (Fig. 2.14), a partial record of spatio-temporal variations in bed properties, meltwater production and/or thermal regime along the base and flanks of the tunnel valleys during valley excavation compared to adjacent upland areas is emerging (Kristensen et al., 2008; Janszen et al., 2012). The widespread occurrence of Newmarket Till within the valleys in Simcoe County suggests that the tunnel valleys formed primarily during phase(s) of active ice advance (e.g. Van der Vegt et al., 2012) and may indicate that valley excavation was partially contemporaneous with till deposition and pre-dates regional drumlinization, which is generally thought to have occurred towards the end of Late Wisconsin ice cover (Figs. 2.7-2.9; Crozier, 1975; Maclachlan and Eyles, 2013; Eyles et al., 2016; Sookhan et al., 2018).

In addition to the drumlinized Newmarket Till within the valleys (Figs. 2.6-2.9), support for glacial erosion contributing to tunnel valley excavation is provided by their excessive widths, which can locally exceed 10 km (Fig. 2.8a) – many workers have suggested that subglacial valleys greater than 5 km wide are considered to represent glacier troughs rather than tunnel valleys (Ghienne et al., 2007; Van der Vegt et al., 2012). Tunnel valleys of comparable widths (up to 5 km) are generally only reported in areas where the valleys have been subjected to repeated erosion beneath multiple ice sheets during successive glaciations (Piotrowski, 1994; Jorgensen and Sandersen, 2006). Lows on the topography of the underlying Illinoian glacial deposits may have locally influenced valley location in the study area (AE; Fig. 2.6a) by influencing the patterns of proglacial sediment deposition prior to Late Wisconsin ice cover.

Although detailed study of the character and facies variability within the Newmarket Till (SU1) has not yet been undertaken in the study area (Chapter 4), it has been noted that there is a common tendency for a higher degree of interstratification of diamicts and sorted beds within the valleys (Bajc and Rainsford, 2011; Bajc et al., 2012; 2014; Mulligan, 2013). The increased variability in sediment facies and till properties within the valleys in the study area is probably the result of a significantly more complex depositional history compared to adjacent upland regions (e.g. Hicock and Dreimanis, 1992) resulting from increased meltwater action at the IBI in low-lying areas of the LIS. The absence of SU1 along the central (deepest) parts of some valleys suggests that meltwater systems were operating throughout (or subsequent to) valley excavation, inhibiting deposition of, or eroding, the Newmarket Till (Barnett et al., 1998; Sharpe et al., 2004).

2.6.1.2. Substrate controls on tunnel valley genesis

Large amounts of sand and gravel underlying the Newmarket Till (SU1) to the north (Burt and Dodge, 2011; Bajc et al., 2014; Mulligan, 2016; 2017c), and deep channels infilled with sand and gravel incised into sediments underlying Newmarket Till to the south (Sharpe et al., 2011; 2013; 2018; Gerber et al., 2018), record meltwater systems that pre-date till deposition and likely promoted the development of tunnel valleys in the study area.

Meltwater systems north of the study area delivered significant volumes of sediment to the proglacial zone during the advance of the LIS into the study area (Mulligan, 2016). Thicker sand bodies occur at higher elevations than the underlying fine-grained lake bottom deposits and have higher transmissivity which allows them to more efficiently drain meltwater at the IBI, resulting in a greater sediment strength and resistance to subglacial erosion (e.g. Boulton and Hindmarsh, 1987). These differences provided the blueprint for the development of the subglacial bed topography during the Late Wisconsin – areas underlain by thick sandy deposits formed obstacles and remained as highs on the subglacial bed, enhancing erosion and ice flow to intervening low areas as ice streamlined around the resistant obstacles, promoting higher rates of subglacial erosion (Eyles et al., 2016) and preferential routing of subglacial meltwater at the IBI. Together, these create a positive feedback loop wherein ice and meltwater get progressively preferentially routed into the developing low, resulting in the large size of the final tunnel valley form.

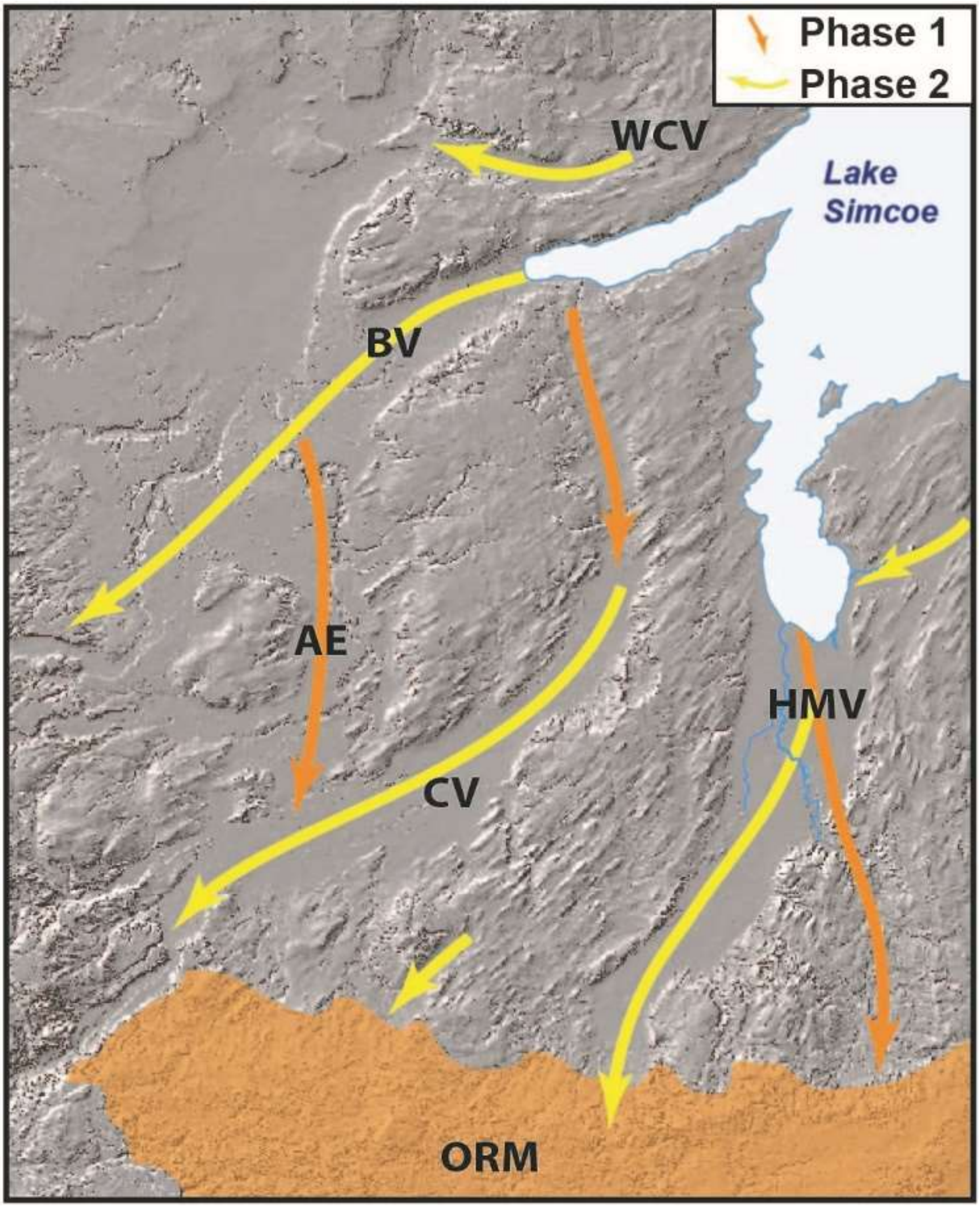
A direct relationship between the meltwater systems responsible for the deposition of thick, coarse-grained sediment accumulations beneath uplands and the erosion of the tunnel valleys within the study area remains difficult to determine, as the factors that are responsible for the shift from aggradational systems in pre-Late Wisconsin sediments to deep channel erosion are unknown. However, this shift may simply record the advance of ice into the study area and a transition from proglacial glaciolacustrine to subglacial environments. Paleocurrent measurements generally record south to southwestward flow directions (Sibul and Choo-Ying, 1971; Barnett, 1986; Bajc and Rainsford, 2010; Sharpe et al., 2013; 2018; Mulligan, 2014). The upland aggradational systems predate advance of the LIS into the study area and southerly paleocurrent directions observed in sands and gravels underlying Newmarket Till (SU1) in the flanks of the BV and WCV are parallel to the upper reaches of the HMV, CV, and AE, but are orthogonal to the E-W and SE-NW axial trends of the BV and WCV, respectively (Figs. 2.7, 2.8). These conflicting inferred flow directions and the cross-cutting relationships of the valleys in Simcoe County (Fig. 2.6), indicate multiple phases of tunnel valley excavation, likely spanning a shift in ice surface gradients and flow directions (Fig. 2.15; see below).

2.6.1.3. Geometry and architecture of the tunnel valley systems

We interpret that N-S oriented valleys formed first, then were either modified, enhanced, or truncated by a second generation of valleys, with a radial orientation in line with regional drumlin trends (Fig. 2.15). The N-S oriented head of Cookstown valley lies at right angles to, and is truncated by, the BV which is floored ≈80 m lower (Fig 2.9; Chapter 3) and NE-SW trending drumlins that cross the AE, CV, and HMV (Figs. 2.7, 2.8). This evidence suggests that there were at least two discrete phases of valley excavation likely under two different ice gradient/ice dynamic regimes.

Perched valleys (AE, CV, WCV) floored with Newmarket Till (SU1) lie immediately adjacent to significantly deeper valleys (where till is apparently truncated, such as the confluence of the BV and AE; Fig. 2.6c,f) and likely record re-routing of meltwater along different pathways following an initial phase of excavation (Jorgensen and Sandersen, 2006; Passchier et al., 2010). New pathways could have developed if the IBI along the valley floors became sealed by active ice and renewed till deposition following meltwater flow event(s) or the development of overdeepenings and steep adverse slopes. This would cause reorganization of the subglacial hydrological system (Cook and Swift, 2012) by plugging the valleys with sediment or by changing the regional ice dynamics/gradient. The deeper valleys likely remained active throughout the multiple phases of erosion and the southwestward to radial deflections in their

Fig. 2.15: simplified schematic illustrating the relative timing of different valleys or segments of valleys. Note the termination of southern valleys beneath the ORM; past drilling and geophysical surveys have identified their extension beneath the moraine (e.g. Barnett et al., 1998; Sharpe et al., 2018).


Phase 1
Phase 2
WCV
Lake
Simcoe
BV
AE
CV
HMV
ORM

downflow ends record changing gradients on the ice surface profile which are significantly (~10x) more important in governing meltwater flow than bed topography (Clarke, 2005).

Although the upflow ends of the valleys commonly occur beneath modern lakes, their locations do not coincide with the margins of potential large reservoirs of subglacial meltwater (Evatt et al., 2006; Livingstone et al., 2013) that would be required to generate a catastrophic drainage event (Kehew et al., 2012b). Additionally, the widespread evidence for proglacial lakes during build-up of the LIS (Mulligan and Bajc, 2018) and during deglaciation in the region (Schaetzl et al., 2016; Mulligan et al., 2018), combined with the sedimentological evidence described above, suggest that the LIS in south-central Ontario was warm-based. Basal temperatures regionally at or above pressure melting point reduce the likelihood of permafrost developing under the ice margin, therefore restricting the potential for the development of pressurized subglacial reservoirs in southern Ontario compared to the outer margins of late glacial ice lobes further south in the Great Lakes region (e.g. Cutler et al., 2002; Hooke and Jennings, 2006).

Although the valleys do not appear to be linked to major subglacial reservoirs, the coincidence of many of the Simcoe County valleys downflow of significant valleys and/or aeromagnetic anomalies observed in shield terrain to the north (Brennand and Shaw, 1994), suggests a structural control to the location and orientation of some of the valleys (Brennand et al., 2006). Within the study area, the WCV is unique, in that it has an abrupt origin on the drumlinized upland north of Barrie (Figs. 2.1, 2.7, 2.8). The abrupt start, disconnected from local overdeepenings or potential basins, suggests that meltwater was likely sourced from supraglacial conduits (moulins) reaching the ice bed (e.g. Livingstone and Clark, 2016) and/or headward growth of the valley through evacuation of groundwater stored beneath the ice.

2.6.1.4. Valley-fill sediments

Consistent gradational transitions between stratigraphic units within the valley infills (SU2-SU4) suggest a largely continuous, uninterrupted sediment record. Fining-upward successions passing from gravel to fine-grained sand and silt are interpreted here as waning flow deposits related to esker-subaquatic fan sedimentation along a retreating ice margin (Rust, 1977; Barnett et al., 1998; Ahokangas and Mäkinen, 2014). The localised distribution of thick successions of coarse-grained SU2 deposits restricted to the deepest, central parts of the valleys (Figs. 2.6, 2.9) and the lack of evidence of extensive scouring of SU1 elsewhere in the valleys suggest restricted deposition within subglacial channels and/or at the ice margin, due to grounded ice along the base of much of the valleys. Proglacial (ice marginal or sub-ice shelf transitioning to distal glaciolacustrine) rather than subglacial depositional settings are suggested for the transition between SU2 to SU3-4, based on vertical and spatial grain size trends, thickness of coarse-grained sediment accumulations, gradational contacts between units (and into overlying fluviodeltaic deposits), and the widespread occurrence of trace fossils throughout SU4 (*cf* Sharpe et al., 2018). The rate of ice retreat during deglaciation of the study area likely outpaced the delivery of coarse-grained sediments to the ice margin after the ORM was deposited, resulting in the apparent localized distribution of esker and/or fan deposits (e.g. Livingstone and Clark, 2016; Fig. 5).

The character of SU4 sediments is consistent with evolving depositional processes and sedimentation patterns in an ice-contact lake during ice retreat. Thin diamict beds within the silt and clay rhythmite succession record ice-rafting from a calving glacier margin, with localised debris dumps caused by overturning melting icebergs or grounding along the flanks of the valleys (Barnett and Karrow, 2018; Mulligan et al., 2018). The abrupt decrease in ice-rafted debris suggests a rapid retreat of the ice front from the study area. However, sedimentation rates within the valleys remained high, possibly due to numerous local sediment sources such as the ice margin, fluvial systems entering the basin from the southwest and west, and downslope contributions from resedimentation along steep valley slopes (Mulligan et al., 2018).

The thickness of fine-grained sediments overlying Late Wisconsin till in the study area is striking and likely arises from the unique paleogeography of deep valleys offering >150 m of accommodation space and creating preferential depocentres in a high-level proglacial lake (Glacial Lake Schomberg; Chapman and Putnam, 1984) that was dammed to the north, east and south by ice lobes (Mulligan, 2013). Sediments were delivered to the lake from fluvial inputs to the north, west, and south, subglacial conduits along the retreating ice margin to the northeast, and through wave erosion along unconsolidated and unvegetated islands in the lake (see Mulligan et al., 2018). Gradational upward transitions into fluvial and shallow-water glaciolacustrine sediments bearing articulated mollusk shells and delicate plant macrofossils record changing depositional environments consistent with water plane fluctuations during later phases of ice retreat (Deane, 1950; Karrow et al., 1975; Mulligan et al., 2018).

2.6.1.5. Relationships with the ORM

Strong relationships exist between the valley infills in the study area and thick accumulations of sand, silt and gravel that comprise the ORM, the largest glacial landform in southern Ontario, immediately south of the study area (Sharpe et al., 1994; Barnett et al., 1998; Sharpe and Russell, 2016; Fig 2.6f). The downflow ends of many of the tunnel valleys within the broader network (Figs. 2.1, 2.8, 2.15) extend to or beneath the ORM (Sharpe et al., 1994; Pugin et al., 1999; Sharpe et al., 2018). As such, ORM sediments are stratigraphic equivalents of SU2-4 reported here, though significant differences in sediment facies and architecture exist, largely due to differences in the stability of ice margin positions, sediment input points, and accommodation space available within water bodies between two melting ice lobes/streams (Barnett et al., 1998; Russell et al., 2003; Sharpe and Russell, 2016; Sookhan et al., 2018). It should be noted that only the valley fills are demonstrably equivalents - although the tunnel valleys in the study area largely predate the ORM, they undoubtedly created preferential pathways for subglacial meltwater flow that contributed sediments to subsequent ORM deposition (Barnett et al., 1998; Russell et al., 2003). Conceptualizations involving integration of tunnel valley erosion and fill into a single erosional-depositional sequence for the ORM (Sharpe et al., 2004; Sharpe and Russell, 2016; Sharpe et al., 2018) require re-assessment in light of the new understanding of the timing and duration of tunnel valley erosion presented herein.

2.6.2. Implications for ice sheet dynamics

It is well established that hydrologic conditions at the IBI exhibit strong controls on the behaviour and stability of an ice sheet/stream/lobe (Rothlisberger, 1972; Piotrowski and Kraus, 1997; Boulton et al., 2001; Greenwood et al., 2016). Deciphering the evolution of hydrological processes and drainage characteristics beneath former ice sheets/streams/lobes remains a daunting challenge. Many meltwater landforms developed during deglaciation remain exposed on the former ice sheet beds providing valuable clues to how the drainage system developed during ice retreat/downwasting (Brennand, 2000). The record of subglacial meltwater activity during the earlier phases of a glacial advance is far less well understood, due in large part to their poor preservation potential as a result of reworking by subglacial processes. Regardless, several field studies have identified deposits attributed to subglacial meltwater activity during decoupling events (Clark and Walder, 1994; Piotrowski and Klaus, 1997; Boyce and Eyles, 2000; Roberts and Hart, 2005; Ravier et al., 2014b; Slomka et al., 2015). We suggest the sand and gravel interbeds locally observed within the Newmarket Till (SU1) within the tunnel valleys record small-scale decoupling during repeated bed separation events, but their character, thickness, and limited distribution are inconsistent with regional synchronous decoupling event(s) reported elsewhere (Sharpe et al., 2004; 2018).

Many studies have demonstrated a relationship between tunnel valley location, formation and geometry compared to substrate transmissivity (Huuse and Lykke-Andersen, 2000; Jorgensen and Sandersen, 2006; Sandersen and Jorgensen, 2012; Janszen et al., 2012). The thickness and overall fine-grained nature of the substrate across southern Simcoe County likely promoted high relative porewater pressures. During periods of high meltwater production, it is unlikely that subglacial drainage could have been achieved by Darcian flow alone (Boulton et al., 2001; 2009; Piotrowski et al., 2006), requiring the development of a more effective channelized drainage system at the IBI in order to drain excess meltwater (Kehew et al., 2012b). Water contained within the confined sandy aquifers interbedded with the predominantly fine-grained substrate sediments (10-100 m below the ice-bed interface) would have been significantly overpressured from overriding ice more than several hundred meters thick (Boulton et al., 2001; Klemen et al., 2013). As progressive deepening of lows on the bed occurred (through combined glacial and meltwater erosion; Björnsson, 1996), groundwater pressure gradients between confined aquifers and the IBI would increase until pressure differences exceeded the sediment strength, leading to failure and drastically increased erosion rates. The configuration of substrate sediments in the study area, and their resistance to erosion, is similar to that reported in the southern North Sea by Janszen et al. (2012); their model of substrate control on the progressive deepening of tunnel valleys related to undercutting and exhumation of non-cohesive sediment beds seems largely applicable in this case, despite the unconsolidated nature of the substrate material in the study area.

The study area is unique within southern Ontario, not only because of the large size of the tunnel valleys, but it is also the only region that has not been significantly impacted by lateglacial fluting/drumlinization – there are no cross-cutting bedforms or large recessional/readvance moraines within the study area (Mulligan et al., 2016; 2017). It is tempting to infer that the development of the valleys during the Late Wisconsin may have been a contributor to significantly increasing the stability of the Simcoe Ice Lobe/stream (due to the development of a highly efficient drainage system delivering excess meltwater to the ice margin; Figs. 2.3, 2.16), but several other factors likely exerted important controls. The high terrain upflow of the study area on the Canadian Shield where the Algonquin Highlands are located and the presence of the deep basins of Lakes Huron and Ontario to the north and south, respectively, would have contributed to significant topographic steering of the ice sheet/streams (Ross et al., 2006; Eyles and Doughty, 2016; Mulligan et al., 2017; Sookhan et al., 2018), leading to more dynamic ice in the Great Lakes basins compared to the inter-lake setting of the study area. Additionally, the steep slope of the Niagara Escarpment immediately downflow of the study area impacted, and likely entrenched, ice flow directions and meltwater routing. Eskers commonly occur directly above modern re-entrant valleys, recording meltwater routing through and up valleys cut into the escarpment face (Straw, 1968; Sharpe, 1990; Kor and Cowell, 1998; Mulligan, 2015; Mulligan et al., 2016), locally terminating in extensive stacked subaquatic fans and/or ice-contact deltas comprising large sandy moraines (Burt, 2018). Streamlined bedforms commonly diverge from the heads of re-entrants, terminating at recessional moraines that roughly parallel the crest of the escarpment (Straw, 1968; Mulligan, 2015).

The large scale of the tunnel valleys in the study area suggests meltwater outbursts alone cannot explain their genesis (e.g. Sandersen et al., 2009). The large scale of the valleys may simply be a product of a relatively stable ice lobe/stream and significant thicknesses of unconsolidated Wisconsin Episode sediments within the study area (Mulligan and Bajc, 2018) compared to areas in the central and eastern parts of the valley network (Fig. 1). Björnsson (1996) describes a glacial trench of similar dimensions (20 km long, 4 km wide, >200 m deep) to the valleys in Simcoe County beneath Breidamerkerjökull in southern Iceland that formed in only a few hundred years. The development of lows on the bed creates positive feedback mechanisms which promote faster ice flow velocities, increased temperatures at the IBI (through frictional melting and thermal insulation due to thicker ice cover), increased meltwater abundances by increasing the subglacial catchment area, and increased melting and ice flow rates. This increase in subglacial meltwater production flushes sediment from the IBI toward former ice margins (e.g. Björnsson, 1996; Bennett, 2003) and permits the genesis of the large tunnel valleys within a single phase of ice cover, lasting ~15 000 years.

Tunnel valleys in the study area developed as a largely self-organizing system in response to changes in meltwater availability and substrate transmissivity and strength (e.g. Boulton et al., 2009). Topography on the underlying glacial succession, the form and location of buried bedrock valleys, escarpment re-entrants, and other structural weaknesses all exerted some control the locations of the valleys across south-central Ontario. The valleys in the study area are in many ways unique: their large size exceeds the typically reported widths of tunnel valleys (Van der Vegt et al., 2012) and they are (to our knowledge) the only examples of tunnel valleys floored by drumlinized till deposited during the same glacial phase as valley excavation. Unlike many tunnel valleys, which form late-stage incisions through the subglacial bed and typically mark the end of dynamic ice action (e.g. Wellner et al., 2006; Kehew et al., 2013), the valleys in the study are a much more integral part of the subglacial bed, and contained active ice until final deglaciation of the region. Further exploration of the causative links of tunnel valley genesis and ice dynamics is warranted: both are strongly controlled by favourable bed topography, substrate strength and transmissivity, meltwater availability and porewater pressure at the IBI as well as ice surface gradients (see Bennett, 2003 and Kehew et al. 2012b for reviews). Recent and ongoing work within southern Ontario has highlighted the importance of major reconfiguration of the LIS into semi-independent ice streams during the lateglacial period (Ross et al., 2006; Kleman and Glasser, 2007; Eyles, 2012; Eyles and Doughty, 2016; Mulligan et al., 2016; Sookhan et al., 2018; Chapter 5). Future developments in understanding ice stream onset and evolution in the region may assist in better constraining the causative relationships between regional ice dynamics and subglacial meltwater systems.

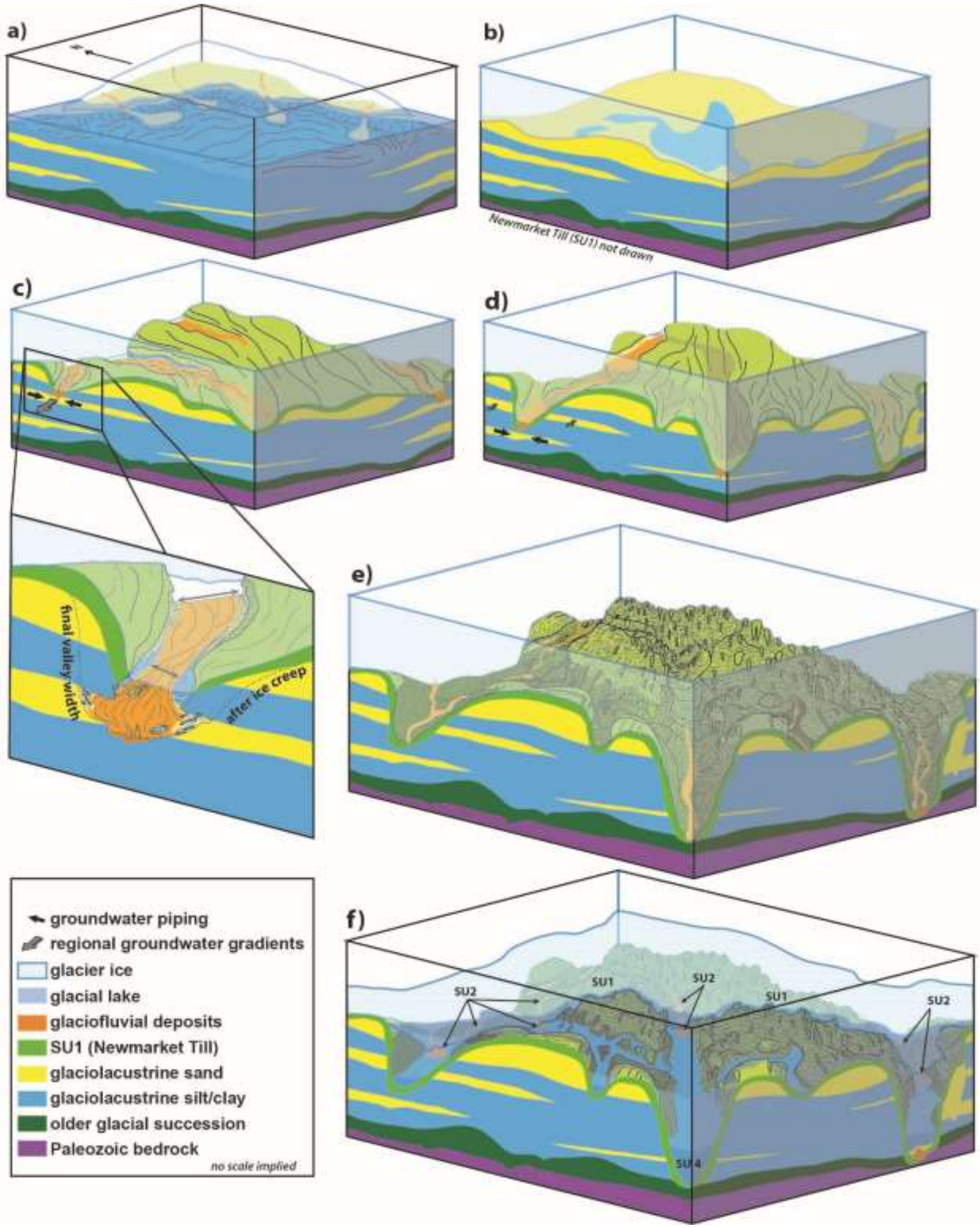
2.6.3. Proposed mechanisms for valley genesis in Simcoe County

A protracted sequence of events is proposed here to explain the landform associations, unit geometries and sediment characteristics documented in the study area and their relationship to valley erosion and infill sedimentation processes (Fig. 2.16).

i) Thick successions of largely fine-grained sediments (tills and glaciolacustrine sediments) were deposited in low-lying bedrock regions in south-central Ontario (Fig 2.16a). Although this cover serves to mute the topographic variability of the bedrock surface, significant valleys and cuestas are clearly visible immediately west of the study area (Fig. 2.1).

ii) Coarse-grained sediments were deposited locally in front of the advancing LIS (Fig. 2.16a) producing a mosaic of glaciofluvial and fine-grained glaciolacustrine facies beneath Newmarket Till (Fig. 2.16b). The thicker, coarse-grained bodies acted as resistant cores that the ice deformed and streamlined around, resulting in enhanced

Fig. 2.16: Conceptual model of valley development and evolution. a) thick successions of fine-grained glaciolacustrine sediments, with coarser-grained sand interbeds are deposited on the bedrock and older glacial sediments within proglacial lakes fronting the advancing ice margin during the Wisconsin Episode. Subaquatic fans immediately in front of the ice sheet deposit variable thicknesses of sand bodies capping the glaciolacustrine deposits; b) localised sediment deposition produces a mosaic of substrate materials at the IBI during Late Wisconsin ice cover. Thicker sand bodies act as well-drained, resistant cores, resulting in ice deforming around them contribution to higher ice fluxes and erosional capabilities in intervening areas. Newmarket Till (SU1) not drawn; c) the predominantly fine-grained substrate is unable to drain meltwater excesses and a more efficient drainage system develops (canals or discrete conduits), enhancing vertical erosion in swales. Predominantly N-S drainage developed due to regional ice gradients, but meltwater systems could have been captured by re-entrant valleys in the Niagara Escarpment to the west. Breaching of cohesive, fine-grained units and till provides connection between pressurized water in confined subglacial aquifers and drainage systems at the IBI. This increases meltwater discharges and sediment erosion by piping and/or turbulent flow. Inset shows the influence of gradients within the confined aquifers contributing to flow and sediment evacuation during breaches, which could have undercut valley flanks, leading to sediment collapse and significant widening of the valley during reoccupation by ice (dashed black line); d) during periods of reduced meltwater production, ice fills the valleys and till deposition (SU1) is resumed. Changing ice sheet configuration during late phases of glaciation changes ice surface gradients towards the southwest, and a second phase of valley systems oriented parallel to ice flow cross-cut older N-S systems. Following shut/slowdown of meltwater production, till deposition resumes; e) changes in ice dynamics and/or basal thermal regime shifts basal ice conditions from till accretion to till sheet drumlinization across both uplands and valleys; patches of till may have also been removed within valleys by solely glacial erosion; f) localised deposition of SU2 and SU3 as ice-contact subaquatic fans, primarily within the valleys, in ice-dammed lakes during deglaciation. SU4 is deposited as fine-grained glaciolacustrine sediments in more ice distal locations during progressive ice retreat. Diagrams are for illustrative purposes - no horizontal or vertical scale is implied.


a)
b)
Newmarket Till (SU1) not drawn
c)
d)
final valley width
after ice creep
e)
f)
SU2
SU1
SU2
SU1
SU2
SU4
groundwater piping
regional groundwater gradients
glacier ice
glacial lake
glaciofluvial deposits
SU1 (Newmarket Till)
glaciolacustrine sand
glaciolacustrine silt/clay
older glacial succession
Paleozoic bedrock
no scale implied

iii) ice fluxes and erosion within intervening areas where sandy substrates were either lacking or very thin.

iv) The low transmissivity substrate was likely unable to efficiently drain the meltwater during period of increased melting. Local lows on the bed would have been the preferential pathways for routing of basal meltwater during periods of enhanced meltwater production/delivery to the IBI (Fig. 2.16c). The exact nature of this drainage network remains unclear, but meltwater flowing in discrete canals/channels or conduits is suggested. These would have created linear zones of lower basal porewater pressure, allowing sediment to creep in from adjacent areas then be transported downflow by continued meltwater activity.

v) Continued downward incision into the substrate increased gradients between pressurized confined aquifers hosted within coarse-grained interbeds in the glaciolacustrine succession and meltwater at the IBI (Fig. 2.16c). Once the gradients across the boundary exceed the strength of the bed material, the units are susceptible to failure – pressurized water within these confined aquifers would have contributed significant volumes of meltwater and sediment into the active subglacial conduit(s), significantly enhancing rates of sediment excavation (Russell et al., 2003). Headward and lateral erosion would be accelerated in response to failure of the conduit walls within the loosely-packed sandy aquifer sediments and vertical erosion would have been enhanced by the addition of significant volumes of groundwater that had been stored within the confined aquifer (Fig. 2.16c).

vi) Channel growth decreases the ability of the walls to contribute sediments and water, hence erosion rates diminish, eventually allowing ice to regain contact with the bed, sealing the aquifer from the IBI through renewed till deposition (Fig. 2.16d). New valleys may have been created following the closure of valley segments and after a change in regional ice sheet configuration while others may have continued to be occupied by meltwater during the entire glacial episode (Fig. 2.16d).

vii) The Newmarket Till (SU1) within tunnel valleys and uplands is drumlinized through erosion of the substrate by active ice. Due to the erosional nature of drumlins in the region (Sharpe et al., 2004; Englert et al., 2015; Eyles et al., 2016), parts of the floors and/or flanks of the valley may have had material stripped away by glacial erosion during till sheet drumlinization (Fig. 2.16e).

viii) During ice retreat, eskers developed primarily within the valleys and supplied sediment to the ice margin. Localized coarse-grained sediments (SU2-3) were deposited on or within incisions into Newmarket Till (SU1) at the mouths of conduits entering proglacial lakes (Fig. 2.16f). Settling of finer-grained sediments (SU4) in distal settings and in quiet waters adjacent to conduits persisted once the ice margin (primary sediment source) retreated northward (Fig. 2.16f).

2.7. CONCLUSIONS

Recent geologic investigations in southern Simcoe County have highlighted the complex relationships and topography exhibited by the Late Wisconsin Newmarket Till sheet. The drumlinized till sheet caps broad, regional uplands and can be consistently traced from the flanks of the uplands into the subsurface, flooring large broad tunnel valleys in places, where it is drumlinized as well. This relationship presents a challenge to existing theories for landscape evolution in southern Ontario that invoke catastrophic releases of subglacial meltwater and the development of deep and extensive subglacial lakes (e.g. Sharpe et al., 2018). A new model for tunnel valley genesis in Simcoe County is presented here based on the demonstrated continuity of the Newmarket Till within the valleys, evidence of active ice within the valleys, late-stage drumlinization of the till sheet, cross-cutting of valley segments, and the localised distribution of coarse-grained valley-fill sediments. In this new model, subglacial topography was generated as a result of the distribution and erodibility of pre-Late Wisconsin sediments. Lows on the subglacial bed formed preferential pathway for meltwater evacuation (through conduits/canals/channels) at the ice-bed interface when meltwater availability exceeded the capacity of subglacial aquifers. Once vertical erosion breached confined aquifers hosted in pre-Late Wisconsin strata, sediment supply and meltwater discharges increased rapidly and lateral and headward valley exhumation was accelerated. Slowing of valley excavation rates due to excessive widening and intersection of more cohesive materials allowed downward creep of ice to regain contact with the bed, resuming till deposition and sealing the aquifers from the drainage system at the IBI. Some channels may have remained active for longer periods, possibly until the final phases of deglaciation, and contributed to routing of large volumes of meltwater and sediment to the ORM. Coarse-grained valley-fill sediments in the study area were deposited in local subaquatic fans along the retreating ice front, fed by esker systems that are locally preserved within the valleys. Drumlinized tunnel valleys in Simcoe County form a more integrated part of the subglacial landscape compared to examples of tunnel valleys studied elsewhere in North America and Europe. We suggest they represent a previously unidentified class of tunnel valley and the landform relationships, sediment geometry and properties observed in the study area hold significant implications for understanding tunnel valley and landform genesis as well as providing a new framework to study evolving dynamics beneath the former Laurentide Ice Sheet.

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CHAPTER 3: SEISMIC STRATIGRAPHY AND ARCHITECTURE OF GLACIAL AND POST-GLACIAL SEDIMENTS INFILLING A LATE WISCONSIN TUNNEL VALLEY, KEMPENFELT BAY, LAKE SIMCOE, CANADA

Abstract

21 Line-km of high-frequency (2-16 kHz) sub-bottom profile data permit the delineation of the sediment architecture in Kempenfelt Bay in Lake Simcoe, which occupies a large (>30 km long, up to 4 km wide) Late Wisconsin tunnel valley in south-central Ontario. Five seismic stratigraphic successions (SSS I – SSS V) comprise the floor and infill of the tunnel valley. SSS I is a high-amplitude undulating and ridged reflector that forms the acoustic basement in the survey data. The unit is lowest in the central parts of the valleys and rises towards both shorelines. SSS II is characterized by chaotic to transparent facies, with local horizontal or anticlinal reflectors, which preferentially occur within lows on SSS I. The bulk of the valley infill consists of well bedded parallel reflectors, which generally increase in reflectivity upward (SSS III). Two sub-facies are observed: parallel and horizontal (SSS IIIa) and deformed (SSS IIIb), the latter occurring predominantly along the steep margins of the valley. Local thin, transparent wedges (SSS IV) occur along the margins of the valley overlying SSS III deposits, and the entire infill is blanketed by transparent to faintly bedded deposits (SSS V). The five units record the latest stages of ice occupying the tunnel valley, depositing the drumlinized Newmarket Till along the valley's floor (SSS I), followed by submarginal meltwater depositing esker and subaquatic fan sediments (SSS II) during the earliest phase of ice retreat, which later resulted in the development of a deep lake basin and proglacial mud deposition (SSS III) within the valley, which was affected by significant mass movement events disrupting deposition. Lake level fall to a lowstand during the early-mid Holocene exposed the valley walls to erosion and resedimentation as grainflows (SSS IV) before water level rise to modern levels resulted in deposition of mud and gyttja (SSS V). Deformation features observed within the data (extensive

faulting, lake floor pitting, and slumps may create preferential pathways for fluid migration, including modern groundwater discharge into Kempenfelt Bay.

3.1. INTRODUCTION

Quaternary glacial deposits host significant resources across a large area of mid-latitude North America and Europe (Huuse et al. 2012; Sharpe et al. 2014). Integrated geophysical and subsurface investigations, involving intensive drilling and sampling programs, are being employed increasingly to map Quaternary deposits, provide hydrostratigraphic frameworks for groundwater resource evaluation (e.g. Légaré-Couture et al. 2018), and improve understanding of landscape evolution and glacial erosional and depositional processes.

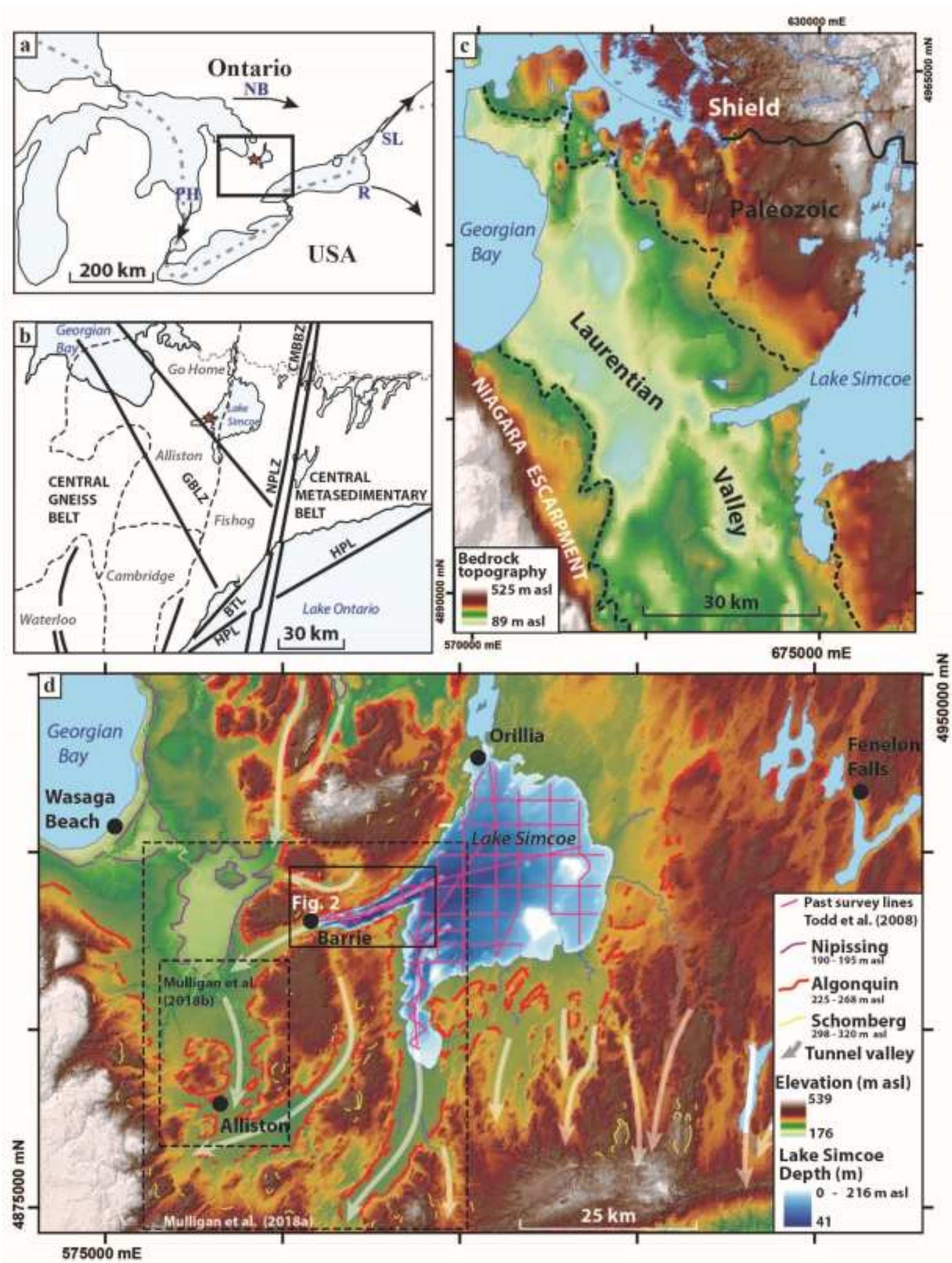
Tunnel valleys are important features of glacial landscapes as they form in response to spatial and temporal changes in subglacial meltwater conditions and ice dynamics during glacier overriding. The development of tunnel valleys is one of the most efficient ways to drain excess volumes of subglacial meltwater (van der Vegt et al. 2012), and they therefore exert a critical control on the evolution of ice sheets (Lelandais et al. 2018). Developing an improved understanding of the controls on tunnel valley genesis across a heterogeneous substrate provides key insights into understanding the interrelationships between meltwater storage and discharge events and ice dynamics (Ravier et al. 2015) and will help to better predict their sediment infill architecture and relationships with regional stratigraphic units for modern resource evaluations and risk assessment (Huuse et al., 2012).

Kempenfelt Bay in Lake Simcoe (Fig. 1), occupies a Late Wisconsin (MIS 2) tunnel valley that extends from beneath the Lake to the Niagara Escarpment, 50 km to the west (Fig 1c). This is one of five large (>30 km long, 2-8.5 km wide) and deep (>150m) tunnel valleys that incise through thick (up to 200 m) Quaternary sediment successions, locally down to underlying Paleozoic bedrock in Simcoe County, which are part of an even broader network of tunnel valleys that spans much of south-central Ontario (Fig. 1d). This paper presents high-resolution chirp sub-bottom profile data throughout Kempenfelt Bay in Lake Simcoe, and integrates it with past sub-bottom investigations in Kempenfelt Bay (Todd et al. 2008; Doughty et al. 2014), as well as detailed surficial mapping (Barnett 1998; Mulligan and Bajc 2012; Mulligan 2017a,b) and sedimentological logging of outcrops (Mulligan et al., 2018a, b) and continuously-cored boreholes (Bajc et al. 2015; Mulligan 2017; Mulligan and Bajc 2018) conducted as part of regional three-dimensional sediment mapping investigations throughout Simcoe County (Bajc et al., 2019). These data provide an enhanced framework for constraining conceptualizations of tunnel valley genesis and infill in the region, identifying potential hydraulic connections between subsurface aquifers and surface water in Lake Simcoe, and refining understanding of the environmental conditions of deglacial lakes in Simcoe County, previously interpreted from onshore geomorphological (Deane 1950; Mulligan et al. 2015) and sedimentological (Karrow et al., 1975; Mulligan et al. 2018b), as well as previous sub-bottom seismic investigations (Todd et al. 2008; Doughty et al. 2014).

3.2. STUDY AREA AND GEOLOGIC SETTING

Lake Simcoe (744 km^2) is located 60 km northwest of Toronto, Ontario (Fig. 3.1). The lake basin lies to the south of the Precambrian Shield and is underlain by Paleozoic sedimentary strata that onlap more deeply buried Late Proterozoic metasedimentary rocks of the Grenville Province (Fig. 3.1b; Armstrong and Carter 2010). Kempenfelt Bay lies across a northwest-southeast trending shear zone that forms a major terrane boundary between the Go Home and Fishog domains (Pozza 2002). The terrane boundary is offset by several west-east trending basement fault zones that have been mapped in regional and local-scale magnetic data (Fig. 3.1b; Boyce and Morris 2002; Boyce et al. 2002).

Fig. 3.1: a) location of the Kempenfelt Bay study area (red star) and major outlets for postglacial lakes within the Great Lakes region of Ontario and USA. NB = North Bay; PH = Port Huron; R = Rome; SL = St. Lawrence. Black rectangle shows extent of Fig. 3.1d; b) structural features and Precambrian lithotectonic domain boundaries (dashed black lines, labelled in grey italics; from Easton and Carter 1995) in south-central Ontario. Bold black lines show major magnetic lineaments: GBLZ = Georgian Bay Linear Zone, CMBBZ = Central Metasedimentary Belt Boundary Zone, NPLZ = Niagara Pickering Linear Zone, HPL = Hamilton-Presqu'ile Lineament, BTL = Burlington-Toronto Lineament. Redrawn from Boyce and Morris (2002); c) bedrock topography of the study area, showing the location of the Precambrian-Paleozoic contact (solid line) and approximate extent of the Laurentian valley (dashed line; data from Gao et al. 2006); d) Topography of the study area (MNRF 2010), showing prominent abandoned postglacial shorelines (glacial Lakes Schomberg and Algonquin and the Nipissing Phase of the upper Great lakes) and bathymetry of Lake Simcoe (MNRF 2002). Location of the Kempenfelt Bay survey area (Fig. 3.2) shown in black box, axes of regional tunnel valleys shown in transparent white arrows other studies mentioned in text are highlighted.

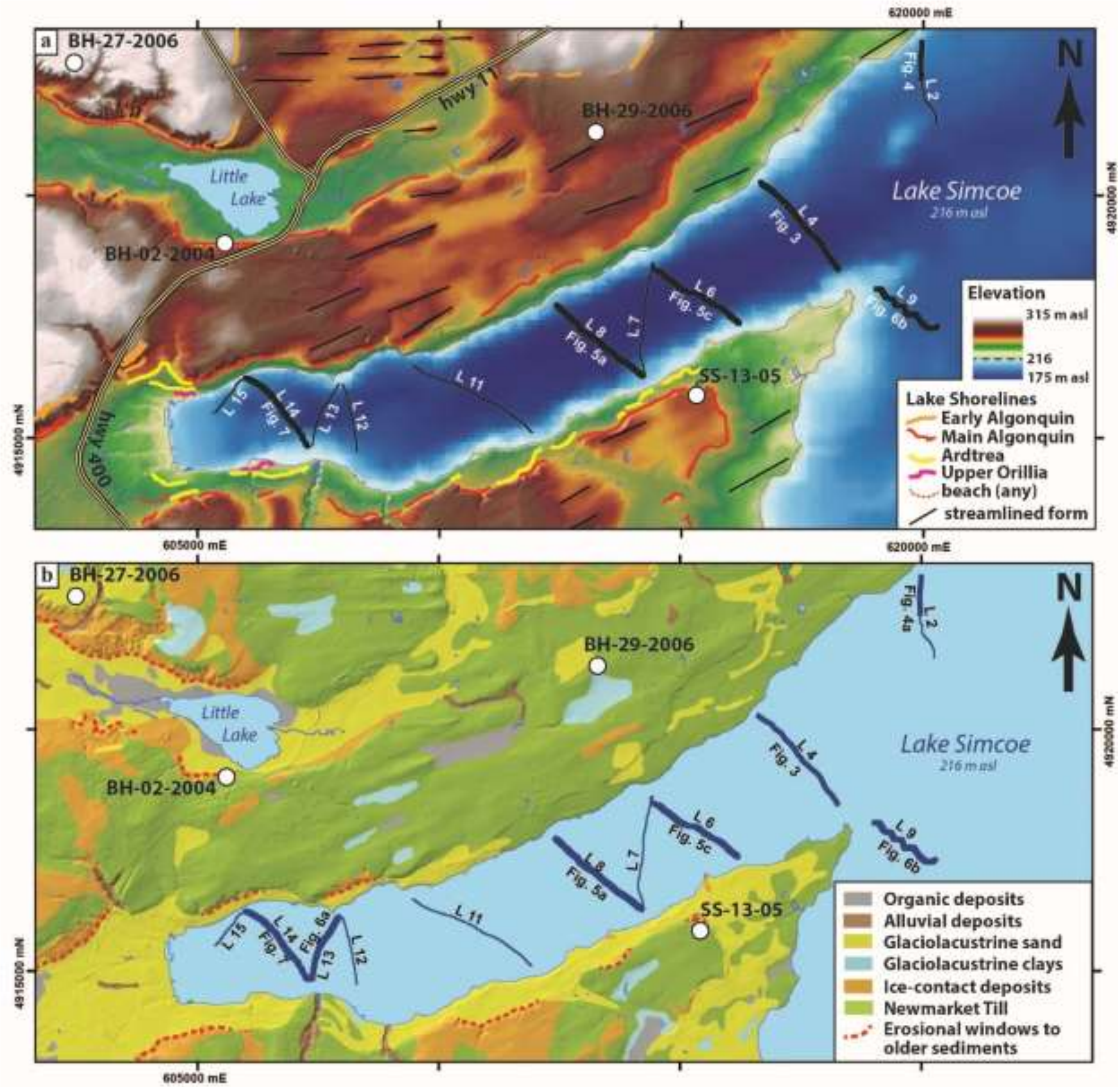

a
Ontario
NB
SL
PH
R
USA
200 km
b
Georgian Bay
Go Home
CMBBZ
Lake Simcoe
Alliston
GBLZ
NPLZ
CENTRAL GNEISS BELT
CENTRAL METASEDIMENTARY BELT
Fishog
HPL
Cambridge
BTL
Lake Ontario
Waterloo
HPL
30 km
c
630000 mE
4965000 mN
Shield
Paleozoic
Georgian Bay
Laurentian
Lake Simcoe
NIAGARA ESCARPMENT
Valley
Bedrock topography
525 m asl
89 m asl
30 km
4890000 mN
570000 mE
675000 mE
d
4950000 mN
Georgian Bay
Orillia
Fenelon Falls
Wasaga Beach
Lake Simcoe
Fig. 2
Barrie
Past survey lines Todd et al. (2008)
Nipissing
190 - 195 m asl
Algonquin
225 - 268 m asl
Schomberg
298 - 320 m asl
Tunnel valley
Elevation (m asl)
539
176
Lake Simcoe Depth (m)
0 - 216 m asl
41
Mulligan et al. (2018b)
Alliston
4875000 mN
Mulligan et al. (2018a)
25 km
575000 mE

Kempenfelt Bay and Cook's Bay (Fig. 3.1) overlie the eastern margin of the Laurentian Valley, a poorly-defined, broad low on the Paleozoic bedrock surface that marks the position of a former river system connecting Georgian Bay to Lake Ontario (Fig. 3.1c; Spencer 1890; Eyles et al. 1993; data from Gao et al. 2006) and was enhanced by glacial activity during Quaternary ice advances (Gao 2011; Mulligan and Bajc 2018; Sharpe et al. 2018). Thick Quaternary deposits (in places over 175 m thick) deposited during the last two glacial episodes overlie the Paleozoic bedrock low and create extensive upland areas in the region (Fig. 3.1; Burt and Dodge 2011; Mulligan and Bajc 2018). Near-complete sediment successions are preserved beneath these upland areas, but extensive erosion during the Late Wisconsin carved large, complex tunnel valleys that truncate upland strata, in places down to the underlying bedrock (Sharpe et al. 2004; 2018; Mulligan et al. 2018a). This has created a characteristic upland-lowland topography in Simcoe County (Fig. 1d; Chapman and Putnam 1984). Lake Simcoe occupies an extensive lowland area comprising multiple overdeepened basins relative to the surrounding terrain. Its two prominent arms, Cook's Bay and Kempenfelt Bay, each occupy large tunnel valleys separating adjacent uplands (Figs. 3.1c, 3.2).

The tunnel valleys of Simcoe County were initially attributed to enhanced late glacial erosion by ice (Dreimanis 1954; Straw 1968) and were later interpreted as tunnel channels (*sensu stricto*; Clayton et al. 1999), recording catastrophic releases of stored subglacial meltwater (Barnett 1990; Sharpe et al. 2004; Todd et al. 2008; Barros 2011). More recently, an origin involving multiple cycles of meltwater incision and direct glacial enhancement over a protracted time period was proposed (Mulligan et al. 2018a). Regardless of the mechanism(s) behind their formation, these valleys represent significant postglacial depositional basins, which now host up to 90 m of sediment deposited in extensive lakes that developed during ice margin retreat (Lewis et al. 2008; Bajc et al. 2015; Mulligan et al. 2018a).

In Simcoe County, glacial Lake Schomberg (elevation 300-320 m asl; Fig 3.1d), was the first to develop while ice surrounded the study area to the north, east, and south, and abutted the Niagara Escarpment to the west, preventing drainage to the Great Lakes basins (Chapman at Putnam 1984; Mulligan 2013). Incremental drainage to form early Lake Algonquin (240-260 m asl; Figs. 3.1d, 3.2a) was achieved as ice margins retreated; water levels in early Lake Algonquin (ELA) then fell rapidly (as much as 30 m) as the isostatically-depressed Fenelon Falls outlet was exposed and the Kirkfield low phase was reached around 12.5 ^{14}C kyr (15 000 cal yr) BP (Finamore 1985; Mulligan et al. 2018b; see Appendix). Subsequent uplift of the outlet resulted in significant water level rise (15 – 20 m) south and west of the outlet to form main Lake Algonquin (MLA; Figs. 3.1d, 3.2; Larson and Schaetzl 2001; Lewis et al. 2005; Schaetzl et al. 2016). Final lake drainage was initiated after approximately 10.5 ^{14}C kyr BP when ice receded north of the Algonquin Highlands, permitting drainage to the east through the French River lowlands and North Bay (Fig. 3.1a; Harrison 1972; Karrow et al. 1975). This drainage phase, combined with significant climatic shifts during the Holocene, resulted in the enclosure of numerous isolated basins, including Lake Simcoe, as regional base levels fell more than 150 m below current Georgian Bay levels, and produced a widespread unconformity in lake sediment profiles across Ontario (Todd et al. 2008; Doughty et al. 2014). This lake level fall occurred prior to a mid-Holocene transgression that formed the Nipissing phase of the upper Great Lakes (Fig. 3.1d).

Fig. 3.2: a) topography and bathymetry of the survey area, showing locations of survey lines (L1 – L15) and figures in text, as well as geologic data from boreholes drilled for recent groundwater resource studies in the area. Note the multiple elevations of shoreline features (author's unpublished data); b) surficial geology of the study area (simplified from Barnett, 1997) overlain on 5-m hillshade relief model. Areas where sediments pre-dating the last ice advance are exposed (erosional windows) are highlighted by dashed red lines. Note the continuity of the Newmarket Till surface from high elevations surrounding Kempenfelt Bay down to the modern shoreline.


a
BH-27-2006
BH-29-2006
BH-02-2004
SS-13-05
hwy 11
hwy 400
Little Lake
Lake Simcoe
216 m asl
620000 mE
605000 mE
4920000 mN
4915000 mN
N
Elevation
315 m asl
216
175 m asl
Lake Shorelines
Early Algonquin
Main Algonquin
Ardtrea
Upper Orillia
beach (any)
streamlined form
b
Organic deposits
Alluvial deposits
Glaciolacustrine sand
Glaciolacustrine clays
Ice-contact deposits
Newmarket Till
Erosional windows to older sediments

3.3. METHODS

The Kempenfelt Bay survey data consist of 11 single-channel sub-bottom seismic profiles (21 line km) acquired across the strike of the tunnel valley in July, 2000 (Fig. 3.2a,b; Boyce et al., 2001). The seismic data were collected using an Edgetech X-STAR sub-bottom profiling system with a swept frequency CHIRP source (Eyles et al., 2001). The high frequency (2-16 kHz) source provided decimeter-scale resolution of sub-bottom layers and a maximum penetration in silt and muds of about 60 m. The CHIRP source has limited penetration in coarse-grained sediments and bedrock resulting in incomplete imaging of the valley fill where compact sediments (e.g. till) and bedrock underlie the lake floor.

Seismic data processing was performed in Vista 2D software and included conversion of data to standard SEG-Y format, automatic gain control (20 ms), bandpass filtering (5-15 kHz) and application of a down-trace running average (3 traces) and trace mix filters to suppress low frequency noise. Processed SEG-Y data were exported to IHS Kingdom Suite for display, reflector picking and horizon mapping. The seismic sections were depth-migrated using average

Fig. 3.3: Line 4 profile showing the overall architecture of seismic stratigraphic successions in the survey (see Fig. 3.2 for location). SSS I has an undulating topography that forms the base of the valley and deepens towards the center of the basin, marked by local ridges (interpreted here as drumlins/flutes). SSS II has a wedge or lenticular geometry infilling local lows and a transparent of chaotic internal fill. SSS III forms the bulk of the valley infill, consisting of well-layered parallel draped reflectors affected by near-vertical normal faulting, but is locally highly deformed (compressional folding and reverse faulting) high up within the succession and along the flanks of the valley (SSS IIIb). SSS IV locally occurs as wedges along valley flanks or local lows on the upper surface of SSS I or III. SSS V is nearly transparent and forms a drape of fairly consistent thickness across the base of the valley. Note numerous small gas chimneys and possible gas pockets trapped within the sediment fill, as well as some larger areas of fluid (gas and porewater?) escape along fault planes. In general, the upper surfaces of SSS I and SSS III form stronger multiples than the lake bottom (SSS V). X axis in meters, y axis in seconds.

NW
SE
fault
top of SSS V
top of SSS IV
top of SSS III
top of SSS II
top of SSS I
channel?
Base of valley
IV
water bottom
IIIb
multiples
I
drumlins/flutes
V
extent of Fig. 4b
IIIa
gas?
II
gas chimneys
IIIb
downlap
onlap
fluid escape

velocities for freshwater column (1450 ms^{-1}) and post-glacial infill sediments (1650 ms^{-1}; Todd et al., 2008).

3.4. RESULTS

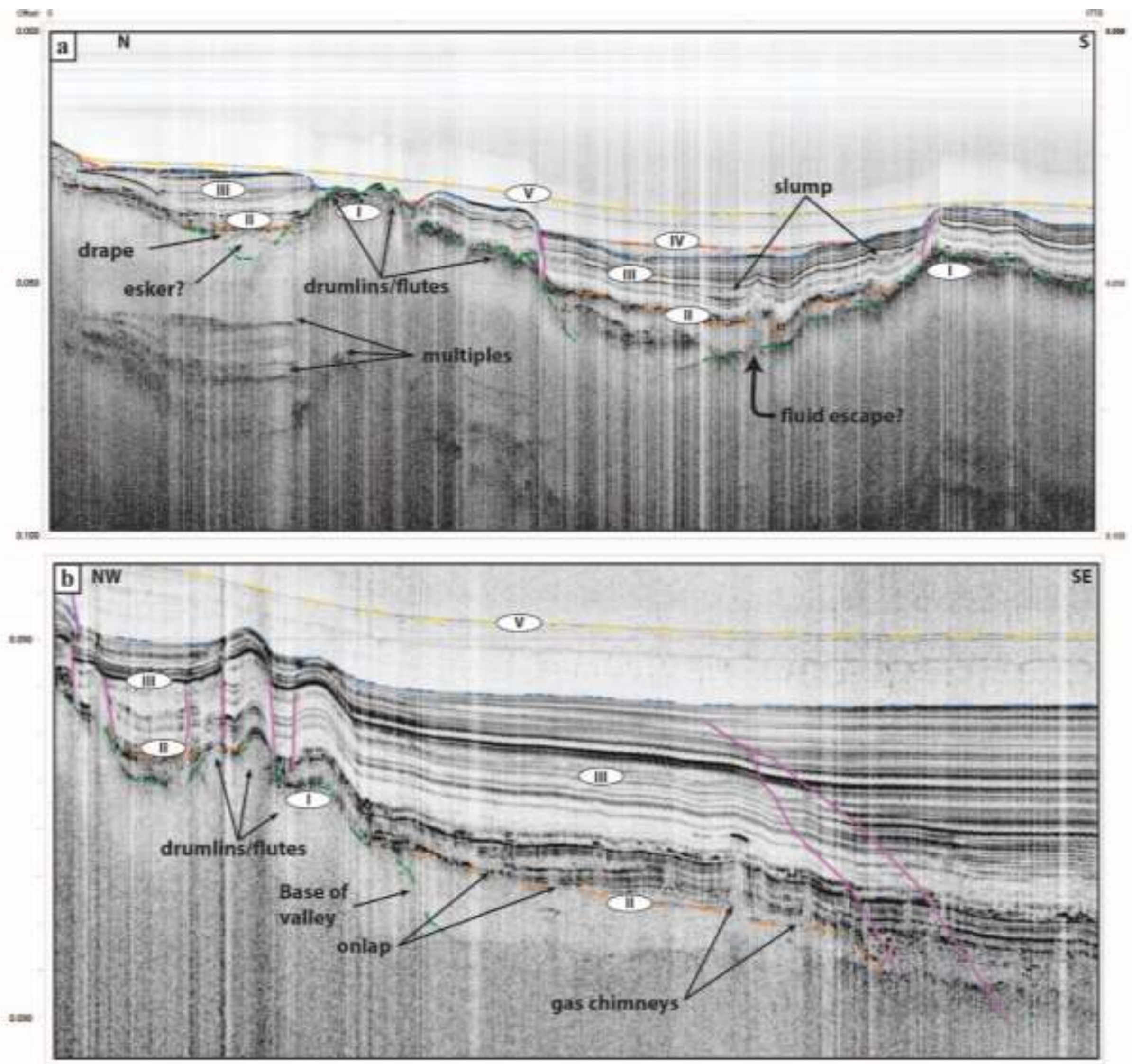
Five seismic stratigraphic successions (SSS I – V; Fig. 3.3) were identified in seismic profiles based on reflection amplitudes, continuity, geometry, internal architecture and stacking patterns (Fernández et al., 2017. These units represent infill of the tunnel valley presently occupied by Kempenfelt Bay. Together, the five successions locally exceed 40 ms (approx. 66 m) thickness, which represents the maximum penetration of the signal through the sediments. No reflectors attributed to underlying Paleozoic bedrock were identified in this study.

3.4.1. SSS I

SSS I is the lowermost unit and forms acoustic basement in the seismic data. The upper reflector of SSS I is characterized by a strong positive amplitude reflection. It is a highly undulating marker bed, forming a generally smooth base for the overlying valley infill succession. No internal bedding or structures are visible within SSS I (Figs. 3.3, 3.4) and its upper surface creates numerous multiples visible throughout many profiles, where it is encountered at shallow depths and where it is more deeply buried by overlying SSSs, producing stronger multiples than all other interfaces observed in this study.

SSS I forms the flanks, and parts of the floor of the valley currently occupied by Kempenfelt Bay and its upper surface in all profiles increases in elevation toward the modern shorelines (Figs. 3.3-3.5). The surface slope is steepest within the confines of Kempenfelt Bay and is much gentler on slopes outside the eastern edge of the bay, where it is encountered at relatively

Fig. 3.4: a) Line 2 profile showing sediment architecture in the eastern part of Kempenfelt Bay. Note numerous high points on surface of SSS I (interpreted as streamlined bedforms(drumlins/flutes); arrowed), anticlinal reflector within SSS II at 0.04s in the low on SSS I in the north (left side of profile) draped by flat-lying reflectors. Numerous disturbed zones (small slumps) occur within SSS III and upward fluid escape structures are visible through the deepest (poorly imaged) part of the basin; b) central part of Line 4 (see Figs. 3.2, 3.3 for location). Individual highs (bedforms) on SSS I are clear on the overall undulating surface, which forms the oldest acoustic basement of the survey. SSS I is not imaged in the central part of the profile where SSS II is thickest. Note the transparent seismic facies of SSSII in the deepest part of the profile, compared to the chaotic facies in the local low in the northwest (0.0625s).


a
N
S
III
II
I
V
IV
III
II
I
slump
drape
esker?
drumlins/flutes
multiples
fluid escape?
b
NW
SE
V
III
II
I
III
II
drumlins/flutes
Base of valley
onlap
gas chimneys

shallow water depths (Figs. 3.3, 3.4). Irregularities are observed in the generally smooth to undulating character of the upper surface of SSS I. These include small channels (concave-up geometries), isolated hills/ridges, and notches or small benches along the flanks where SSS I occurs at relatively shallow depths (Figs. 3.3, 3.4). The upper SSS I reflector is not visible in the deepest (central) parts of the surveyed area, where overlying sediments are thickest.

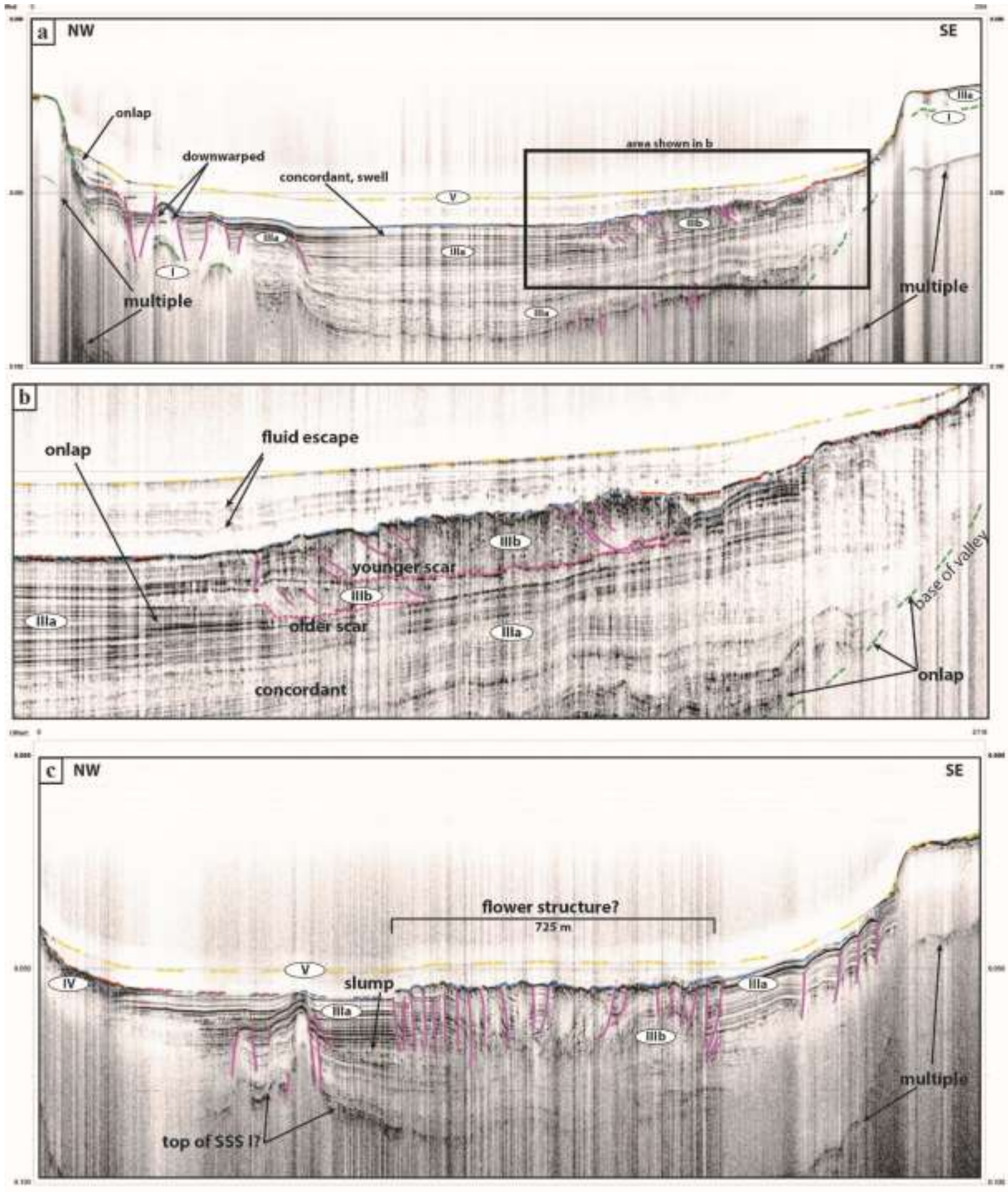
3.4.2. SSS II

SSS II is only observed on a few profiles within the survey where overlying sediments are relatively thin (Lines 2, 4, 9, 15; Fig. 3.2) and locally in-fill depressions on, or channels incised into, SSS I (Figs. 3.3, 3.4). SSS II is thickest overlying channels or other localized lows on SSS I, and has a generally lenticular unit geometry. It is generally massive or has faint, chaotic internal reflectors. In one profile (Line 2; Fig. 3.3, 3.4), an anticlinal reflector 85 m wide and approximately 9.9 m (6 ms) high is observed internally within SSS II, infilling a local low on the upper surface of SSS I. Other faint to moderate amplitude reflectors in SSS II at this location drape and onlap the flanks of the anticlinal feature and overlie it, forming continuous parallel reflectors within the low on SSS I.

3.4.3. SSS III

SSS III drapes SSS I-II along the base of the valley, and onlaps SSS I along the margins. It forms the bulk of the infill stratigraphy of Kempenfelt Bay, exceeding 35 ms (> 30 m) in thickness. Internally, SSS III is generally well-bedded, with laterally continuous concordant reflectors in the central parts of the valley, and onlapping and downlapping reflectors along the flanks (Figs. 3-6). Well-bedded facies are given the designation SSS IIIa on all figures herein. More transparent facies are common near the base of the unit, but faint reflectors show clear drape and onlapping relationships with underlying units. Reflectors show much greater local

Fig. 3.5: a) Line 8 profile showing facies and architecture of SSS III. Note near-vertical normal faulting along the base and northwestern flank of the valley, with downwarping of adjacent reflectors in SSS III. Reflectors are largely parallel and conformable, high amplitude events occur along the flanks of highs and taper out towards central parts of lows. Aggradation in the deepest parts of the valley, likely due to traction current activity, creates a swell consisting of a broad anticlinal feature with concordant internal reflectors. A large zone of deformed sediments occurs along the southeastern flank of the valley and is shown in greater detail in b; b) close-up view of deformed zone (SSS IIIb) showing it is composed of (minimum) two distinct disturbed zones (mass movement events; shown by dashed pink lines). Truncation of underlying reflectors points to localized erosion at the base of the slumps. Note stacked reverse faults and rotated strata indicating movement toward the basin center. Local downwarping of reflectors within SSS V may indicate fluid escape through the upper sediment packages; c) Line 6 profile showing a wide zone of deformed SSS IIIb sediments in the central part of the valley. Radial changes in fault planes (pink lines) are consistent with a flower structure indicating transverse movement beneath the visualized succession, but additional data is required in order to confirm relation of the deformation to underlying features. Strong reflectors in the lower part of the profile may be consistent with SSS I, but data quality is insufficient for confident identification.


a
NW
SE
onlap
downwarped
concordant, swell
area shown in b
V
IIIa
IIIb
I
multiple
b
onlap
fluid escape
IIIb
younger scar
older scar
IIIa
concordant
base of valley
c
NW
SE
flower structure?
725 m
IV
V
slump
IIIa
IIIb
multiple
top of SSS I?

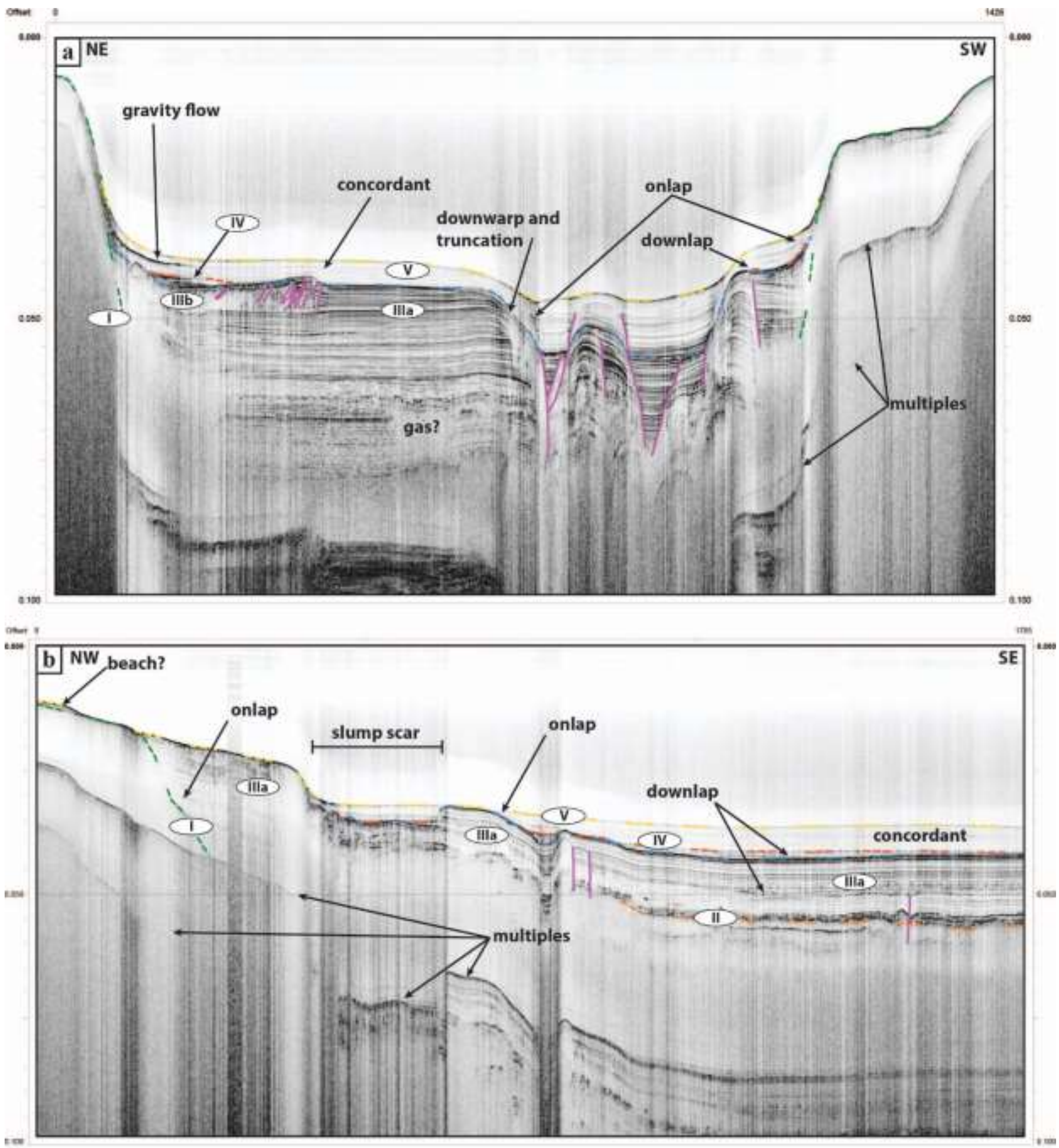
topographic variations near the base of SSS III, but become more regular, continuous, flat-lying, and increase in amplitude upward in the succession (Figs. 3.3-3.7). The high-amplitude internal reflectors appear to have parallel or lenticular geometries and are generally conformable, with no channelized, cross-cutting, or cut-and-fill relationships observed.

The upper surface of SSS III is marked with a widespread unconformity where it is observed at shallow depths or where it is overlain by SSS IV; the upper contact is conformable, but is characterized by a high amplitude reflector where the interface is encountered at greater depths (below ~ 18 m / 25 ms; Fig. 3.6b). Bed thickness appears to be fairly uniform throughout SSS IIIa, except in thick wedge-like bodies observed along the flanks of highs, thinning toward central parts of the basin, which are given the designation SSS IIIb (Figs. 3.3, 3.5, 3.6). Internally, these wedge-like bodies are transparent, chaotic, or contain intensely deformed reflector horizons with abundant near-vertical discontinuities,

Discontinuities in reflectors, interpreted as the result of faulting, are commonly observed in both SSS IIIa and IIIb (Figs. 3.3-3.6). Near-vertical breaks in the beds commonly form graben-like structures in SSS IIIa. The footwalls on graben margins are commonly down-warped, whereas the graben is characterized by horizontal internal reflectors (Fig. 3.6). In some areas, discrete zones of signal attenuation are observed along the footwall of the faults and diffractions are observed on the top of the fault planes. At one location, the modern lake floor is depressed where a fault penetrates the entire observed valley infill succession (SSS III-V; Fig. 3.7). The depression on the lake floor is approximately 50 m wide and 1.7 m deep.

Along the flanks of Kempenfelt Bay, particularly in the deep area in the central part of the bay, discrete packages within SSS III have highly irregular bounding surfaces, with contorted, faulted, and/ or chaotic internal facies (Figs. 3.3-3.6). These are given the designation SSS IIIb.

Fig. 3.6: a) Line 13 profile showing significant topographic variation in the central part of the valley. The low on the lake bed (SSS V) coincides with numerous graben features penetrating SSS III into SSS V. The lower part of SSS V contains higher amplitude reflector that onlap the flanks of grabens, while the upper part is more transparent, with concordant reflectors draping the entire valley fill succession; b) Line 9 profile showing SSS IV infilling local lows (base of slump scar and small low on SSS III in the center of the profile) and forming thin wedge-like bodies that taper out towards the basin center. A large slump scar within SSS III is entirely infilled with SSS V muds. A small ridge of SSS V in the northwesternmost part of the profile (.011-0.12s, approx. 8 m depth) may correspond to a beach deposit formed during lowered lake levels.

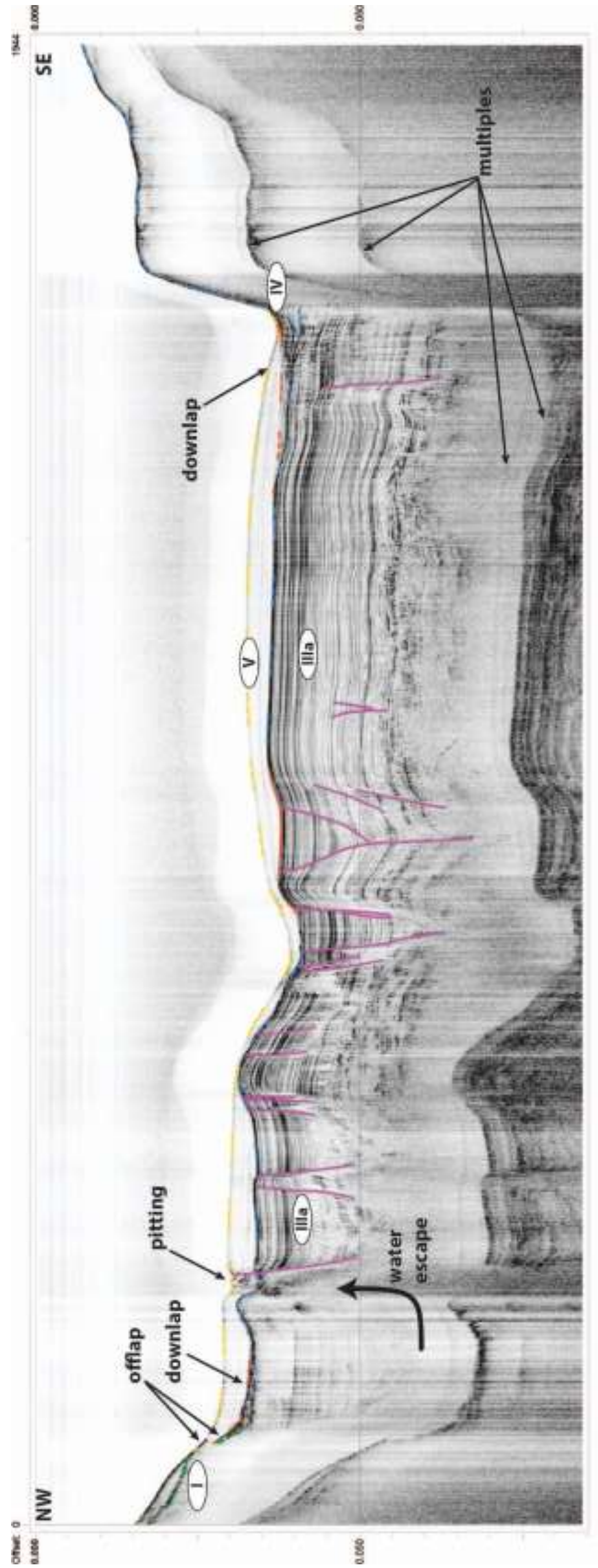

a
NE
SW
gravity flow
concordant
onlap
downwarp and
truncation
downlap
IV
V
IIIb
IIIa
I
multiples
gas?
b
NW
beach?
SE
onlap
slump scar
onlap
downlap
IIIa
V
IIIa
IV
concordant
I
IIIa
II
multiples

Low-angle discontinuities interpreted as thrust faults are common within these bodies where beds are still discernable. Along the flank of the valley, the features thin and onlap the valley margin. Their bases may be (apparently) conformable, truncate, or show interbedding with underlying SSS IIIa facies. Laterally, toward the basin center, these bodies are sharply juxtaposed beside SSS IIIa facies. One occurrence of SSS IIIb, observed in the central parts of the basin floor (Fig. 3.5c) is heavily faulted, with various fault plane orientations (all near-vertical), and undisturbed SSS IIIa facies flanking both sides. At another location, east of Kempenfelt Bay, vertical truncation of SSS III facies is observed, forming a box-shaped erosional feature (Fig. 3.6b) that is partially infilled with SSS IV-V sediments. Numerous local ‘bright spots’ are observed within SSS III, particularly within the lower, more transparent facies of the unit (SSS IIIa; Figs. 3.3, 3.4b, 3.5a). These typically overlie vertical zones of chaotic facies within SSS II and III. Multiple zones of chaotic facies are observed within thicker parts of SSS III, commonly concentrated along faults, or emanating upwards from highly irregular surfaces on underlying SSS I and II facies (Figs. 3.3, 3.4a, 3.7).

3.4.4. SSS IV

SSS IV is characterized by wedge-like geometries that are common along steep slopes on SSS I, directly overlying SSS III (Fig. 3.6). SSS IV typically infills lows/channels and generally mutes the slope breaks on underlying units, where SSS III onlaps the steep flanks of SSS I. The basal surface commonly truncates SSS III reflectors (Fig. 3.6). Where SSS IV wedges begin along the valley flank (SSS I), they thin and pinch out toward the basin center, but if they originate along highs on the surface of SSS III, they may form thin, laterally continuous units (Fig. 3.6). Internal facies range from transparent and homogenous to faintly chaotic. In line 9 (Fig. 3.6), the depth of the base of SSS IV increases with distance away from the lake shoreline.

Fig. 3.7: Line 14 profile showing thick successions of SSS III that are largely undisturbed except for minor normal faulting. SSS V has a positive relief in the central part of the valley, downlapping onto SSS III-IV near the valley margin in the southeast. Fluid escape (likely including porewaters) is evident along the base of a fault plane. Pitting of the lake bed indicates recent removal of sediment by fluid flow, either due to recent expulsion of water caused by activation of the fault, or through sustained fluid flow due to strong upward hydrologic gradients within the valleys.


SE
multiples
IV
downlap
V
IIIa
pitting
IIIa
water escape
offlap
downlap
I
NW

Basin-dipping reflectors are locally observed downlapping onto underlying SSS III as the unit pinches out towards the basin center (Fig. 3.6).

3.4.5. SSS V

SSS V drapes the underlying units, forming a planar tabular geometry that blankets SSS I-IV, draping the topography of underlying units except along the steepest slopes of SSS I-III. It is commonly 4-8 m (5-10 ms) thick. Internally, faint concordant reflectors characterize SSS V, although in some lines, the central part of the unit may display an internal package of higher-amplitude reflectors (Figs. 3.4, 3.5b, 3.7). SSS V onlaps highs on the underlying units and may display local downlapping patterns toward the basin center along breaks in slopes (Figs. 3.4, 3.6). In one area in the western part of Kempenfelt Bay (line 14; Fig. 3.7), the onlapping pattern along valley flanks is not observed, and the unit has a lenticular geometry, with downlapping reflectors terminating along the upper surface of SSS III near the southern margin of the valley (Fig. 3.7).

3.5. INTERPRETATION

3.5.1. SSS I

SSS I is defined by an undulating high-amplitude reflection that represents ‘acoustic basement’ below Kempenfelt Bay (Figs 3.3-3.7). The high surface relief on this unit and its reflectivity suggest that it is a compact deposit (Pirrotta et al. 2013), likely a subglacial till, that predates the post-glacial lacustrine infill. The stratigraphic position of SSS I and its common tendency to rise toward the margins of the lake indicate that is likely correlative with the Newmarket Till (Gwyn and Dilabio, 1973; Todd et al., 2008; Bajc and Rainsford, 2010) which is exposed in drumlinized uplands on the flanks of the Kempenfelt valley (Fig. 3.2; OGS 2010; Mulligan and Bajc 2012; Mulligan 2017a,b; Mulligan et al. 2018a). In addition, SSS I shows similar characteristics to the Newmarket Till reported from existing lake-based seismic data sets (Todd et al. 2008), and has a stratigraphic position at the base of the tunnel valley infill succession (Todd et al. 2008; Bajc et al. 2012; Mulligan et al. 2018a, b*)*. The strong reflector at the upper surface of SSS I is consistent with high P-wave velocities (2000 to >3000 ms^{-1}) measured for the Newmarket Till in the region (Pullan et al. 2002; Crow et al. 2015; 2018). The local topography displayed on the upper surface of SSS I is consistent with drumlinization of the till sheet as seen both on land (Fig. 3.2; Sharpe et al. 2004; Sookhan et al. 2018; Mulligan et al. 2018a; Eyles et al. 2018) and in the subsurface/ beneath Lake Simcoe (MNRF 2002; Todd et al. 2008). The undulating upper surface of SSS I is also consistent with the form of subglacial erosional features comprising local overdeepenings and/or internal sills (Cook and Swift 2012).

Recent investigations have revealed that the bases of some tunnel valleys in the area (and the surface elevation of Newmarket Till within them) tend to deepen progressively downflow to the south and west (Mulligan et al. 2018a; Bajc et al. 2019). Unfortunately, the lack of acoustic penetration through the thick overlying SSS III deposits precludes full assessment of the geometry of SSS I along the deepest parts of the valleys in this investigation. The lack of definitive identification of the upper surface of Newmarket Till in the central parts of the survey area may also be due to its erosion/non-deposition at the base of the valleys (Sharpe et al. 2004; 2018; Todd et al. 2008).

Interpolation of bedrock elevations along the flanks of Kempenfelt Bay (Gao et al. 2006), and comparison with the depth intervals at which SSS I is encountered, suggest that in many areas, SSS I overlies significant thicknesses of older sediments. The lack of identification of a basal reflector of SSS I is likely due to strong impedance contrasts reported for Newmarket Till (P-wave velocities 2000 - 3000 ms^{-1}compared to 1500-1800 ms^{-1} for overlying postglacial deposits; Pullan et al. 2000; Sharpe et al. 2002; Crow et al. 2018), which reflect energy at the interface.

3.5.2. SSS II

The massive and chaotic seismic facies and the occurrence of SSS II within lows and/or incised into SSS I, suggest that it is correlative with coarse-grained valley-fill deposits identified in previous water-borne seismic investigations throughout Lake Simcoe (Todd et al. 2008) and in recent drilling and geophysical investigations in the region (Sharpe et al. 2002; 2018; Burt and Dodge 2011; Bajc et al. 2014b; 2015; Mulligan et al. 2018a). SSS II represents the initial deglacial sediments, deposited as eskers and subaquatic fans at or near the ice margin, either infilling lows on the Newmarket Till in the tunnel valley, or in meltwater conduits (R-channels) at the base of the ice sheet. In this study, only one seismic line permitted visualization of internal bedding architecture in SSS II (Line 2; Fig. 3.4a); the anticlinal surface with chaotic internal bedding is interpreted as a small esker (Pugin et al. 1999) and overlying horizontal and parallel reflectors are interpreted to record the transition from subglacial to subaquatic fan sedimentation (Ross et al. 2006). Significant diffraction events that could be interpreted to record the presence of large boulders (recording high magnitude meltwater flows) within SSS II were not observed. In the study area, SSS II has a limited thickness relative to the entire valley fill succession, which suggests that ice retreat was fairly rapid, inhibiting the deposition of thick and continuous coarse-grained deposits along the ice margin (Mulligan et al. 2018a).

3.5.3. SSS III

SSS III comprises the bulk of the sediment infill of Kempenfelt Bay, and its regular, continuous internal reflectors draping underlying units suggest distal suspension settling of fine-grained deposits (Eyles et al. 2000; Nzekwe et al. 2018). Together, SSS III sediments are attributed to deposition in glacial lakes Schomberg and Algonquin (Deane 1950; Chapman and Putnam 1984; Lewis et al. 2005; Todd et al. 2008; Mulligan et al. 2018b).

The graben features observed within SSS III in Kempenfelt Bay record faulting of the sediments, likely due to a combination of melting of buried ice blocks, isostatic rebound of the crust, loading by younger valley infill sediments, and possibly neotectonic activity (Doughty et al. 2014). Higher-amplitude reflectors within the succession are interpreted to record thicker and coarser-grained interbeds deposited by traction currents originating from marginal streams carrying spring melt, and/or debris flows derived from reworking of shoreline and shallow water deposits during storm events (Eyles et al. 2000; Mulligan et al. 2018b).

The consistent occurrence of high amplitude reflectors within the upper part of SSS III suggests an allocyclic control may be responsible for their deposition. These beds are interpreted to record enhanced erosion of valley walls and sediment progradation (Ashley 1975), consistent with a base level fall during the Kirkfield low phase (Karrow et al. 1975; Finamore 1985; Mulligan et al. 2018b). This interpretation is supported by the many wedge-shaped bodies observed along the flanks of the valleys within SSS III, interpreted as slumps and subaqueous density flows that grade laterally into typical SSS IIIa facies towards the deeper valley center, and recording decreasing competency of the flows with increasing distance away from the slope margin (Mulder and Alexander 2001; Vardy et al. 2010). Bright spots within SSS IIIa commonly overlie vertical chaotic zones or pipes, and likely indicate the presence of gas hosted within the sediments (Okay and Aydemir 2016)

The large lobate bodies containing deformed and faulted internal reflectors are interpreted as subaqueous slump blocks or cohesive debris flows, similar to those observed in local surficial investigations (Barnett 1997; Mulligan 2013) and in water-borne CHIRP profiles elsewhere (Dingler et al. 2009). Slumping could be caused by a variety of factors including storms, downfaulting of the central parts of the basin (due to unloading of glacier mass, melting of buried ice blocks, or neotectonic activity), changing base levels, or syn-depositional slope oversteepening (Kaszycki 1987; Shilts et al. 1989; Shilts and Clague 1992). Small slumps are observed at various intervals within SSS III, but the largest features consistently occur in the uppermost beds, and are conformably draped by SSS V. It is possible that a single seismic event could have triggered the near-simultaneous development of the larger slumps (Doughty et al. 2014). Detailed subsequent study, with the collection of additional lines, including lower-frequency signal to enhance depth penetration, should be undertaken to validate this suggestion.

Support for a seismic trigger for the slumps might also be provided by the occurrence of SSS IIIb in the central part of the valley in Line 6 (Fig. 3.5c), where the deformed unit is flanked on both sides by undisturbed strata (SSS IIIa). Here, the faulting and pervasive deformation are consistent with transverse (strike-slip) displacement, since there is no observed head of a slump up-gradient (Fig. 3.5c). The configuration of highly faulted beds amid undisturbed strata is consistent with the geometry of a positive flower structure in a wedge thrust-fold (Civile et al. 2014). Significant offsets observed at the base of some of the fault traces, rotated fault blocks with radial fault plane orientations, and penetration of the faults through much of the valley infill succession, influencing the modern lake bottom, suggest that neotectonic stresses, rather than less intense or local factors, were the dominant controls in developing the deformation features (Doughty et al. 2014). However, even though the faults observed in the sediments overlie deep-seated structural features (Easton and Carter 1995; Boyce et al. 2002; Fig. 3.1b), without directly observing the faults penetrating and offsetting the underlying Paleozoic bedrock, other mechanisms cannot be definitively ruled out (Shilts and Clague 1992).

3.5.4. SSS IV

The truncation of deposits beneath SSS IV (and SSS V) marks a significant unconformity, one that is better developed in shallower parts of the basin, particularly east of Kempenfelt Bay (Figs. 3.4a, 3.6b). This truncation surface is correlated to the early Holocene unconformity identified in water borne seismic investigations in Lake Simcoe and across central Ontario (Todd et al. 2008; Doughty et al. 2014) and in sediment investigations (Fitzgerald 1985; Bajc and Rainsford 2010; Mulligan et al. 2018). A significant regional base level fall is responsible for the formation of this unconformity as falling water levels exposed the steep, unconsolidated and largely unvegetated valley walls to littoral erosion and downslope resedimentation processes. The wedge-shaped geometry of SSS IV (Figs. 3.3, 3.4a, 3.6b) is consistent with downslope resedimentation into the basin. The transparent to chaotic internal facies suggest the wedges are composed primarily of sand; these grade laterally into adjacent SSS V, recording a fining of grain size and loss of flow competency, and support a density flow origin for SSS IV (Vardy et al. 2010). Channelized features incised into SSS I and III are attributed to initial erosion of the substrate during base level fall, followed by infilling during the subsequent transgression (Mulligan et al. 2018b).

3.5.5. SSS V

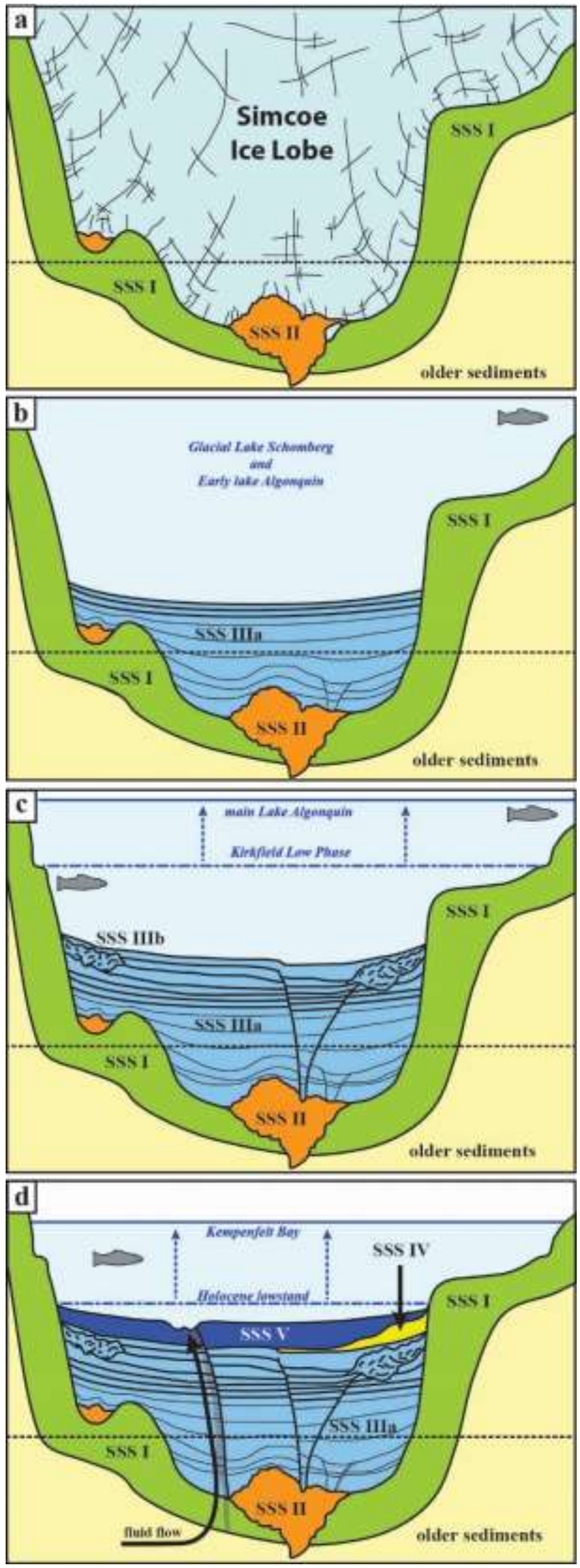
SSS V passively drapes SSS IV and the underlying unconformity and records deposition during the time interval from the development of the underlying unconformity to modern times. SSS V represents basinal equivalents of the SSS IV sediment wedges along the valley flanks, and channel fills further toward the basin center. The angular unconformity along the basal contact of SSS V with underlying units has been used to constrain the maximum base level fall during the early Holocene (Figs. 3.3, 3.6b, 3.8d; Lewis et al. 2008).

3.6. DISCUSSION

3.6.1. Depositional setting and evolution of Kempenfelt Bay

The presence of Newmarket Till (SSS I) flooring the eastern parts of the tunnel valley underlying Kempenfelt Bay, and ridges on its undulating surface that are consistent with drumlins, provide support for a subglacial origin for the valley (Barnett et al., 1998; Todd et al., 2008), at least during the final stages of its excavation (Fig. 3.8a; Mulligan et al., 2018a). Recession and melting of the Simcoe Ice Lobe promoted enhanced meltwater activity during deglaciation leading to deposition of localized, coarse-grained, ice-proximal and submarginal esker and fan sediments (SSS II; Fig. 3.8a). Continued ice retreat exposed Kempenfelt Bay, allowing it to fill with meltwater and permitting deposition of thick accumulations of glaciolacustrine sediments; the lacustrine deposits are characterized by fine-grained homogenous sediments at the base (Fig. 3.8b), passing upward into variably-textured (and thicker?) rhythmites near the top of the succession as regional base levels fluctuated (Fig. 3.8b). Melting of ice blocks, differential sediment loading, and/or isostatic effects resulted in faulting of the lower strata, whereas a significant event (or events) probably triggered large slump slides and scars that deform the uppermost part of the glaciolacustrine sediments (SSS III; Fig. 3.8c). Faults

Fig. 8: Conceptual model for infilling of the Kempenfelt Bay valley; a) drumlinization of the Newmarket Till (SSS I) followed by deposition of coarse-grained sediments (SSS II) along lows within the valley. Early sedimentation was likely in conduits confined by the glacier (eskers) which were later draped by more laterally extensive sediments (subaquatic fans). Ice blocks are locally buried by sediment; b) ice retreat terminates the coarse-grained sediment supply, and fine-grained rhythmite deposition dominates in post-glacial lakes Schomberg and Early Algonquin. Melting of buried ice blocks leads to the development of non-penetrative faulting along the base of SSS III. Reflector amplitude typically increases upward due to falling water levels and re-working of unconsolidated and unvegetated valley walls by wave erosion; c) falling water levels to the Kirkfield low phase triggers instability along the valley walls, leading to the extensive mass movement deposits in the upper part of SSS III. Sedimentation continues as water level rises to main Lake Algonquin stage; d) following incremental water level fall in post-Algonquin lake levels, a climatically-induced lowstand during the early-mid Holocene (e.g. Lewis et al., 2008) triggered mass movement events and gravity flow deposition (SSS IV). During transgression to modern levels, mud and gyttja accumulated across the basin. Recent groundwater flow reaches the lake floor along fault planes.


a
Simcoe
Ice Lobe
SSS I
SSS I
SSS II
older sediments
b
Glacial Lake Schomberg
and
Early lake Algonquin
SSS I
SSS IIIa
SSS I
SSS II
older sediments
c
main Lake Algonquin
Kirkfield Low Phase
SSS I
SSS IIIb
SSS IIIa
SSS I
SSS II
older sediments
d
Kempenfelt Bay
SSS IV
Holocene lowstand
SSS I
SSS V
SSS IIIa
SSS I
SSS II
fluid flow
older sediments

that penetrate the younger (<3.4 ^{14}C kyr BP; Todd et al. 2008) uppermost Holocene sediments (SSS IV-V) in Kempenfelt Bay may provide a record of neotectonic events within Lake Simcoe (Figs. 3.7, 3.8d).

The sandy sediment wedges (SSS IV) that unconformably overlie the glaciolacustrine sediments of SSS III, are interpreted to record base level fall during the early Holocene. The unconformity surface between SSS III and SSS IV is consistent with the development of erosional conditions associated with closure of the Lake Simcoe basin during the Holocene lowstand(s) after 8.3-8.5 cal ka BP (Todd et al. 2008). It is not possible to discern whether the sandy sediment bodies were deposited during lake level fall after isolation from the glacial Lake Algonquin basin, beginning during the Penetang phase of post-Algonquin lakes (Deane 1950; Lewis et al. 2005), or whether they represent the final lowstand system within the basin during the Early Holocene.

The low-lying elevation of SSS IV and its thickness, lateral continuity, and lateral gradation into transparent facies associated with widespread Holocene mud (SSS V) supports a density flow origin for these sediments. Small notches cut into the Newmarket Till along valley flanks (SSS I) might record shoreline erosion during time(s) when water levels were below modern water plane. Transparent to faintly laminated facies in SSS V record mud and gyttja deposition beginning during Holocene water level rise toward modern day Kempenfelt Bay.

The high-frequency sub-bottom seismic data presented here provide high-resolution visualization of sediment to about 60 m, but do not penetrate into deeper sediments identified previously by other studies (Todd et al., 2008; Sharpe et al., 2018).

Table 3.1: correlation of units identified in sub-bottom studies and sedimentological investigations in the Lake Simcoe Area (Fig. 3.1d). MLA=Main Lake Algonquin; KLA=Kirkfield Low phase of Lake Algonquin; ELA=early Lake Algonquin; GLS=glacial Lake Schomberg.

Study	Todd et al. 2008	Mulligan et al. 2018b	**This study**	*Formation*
Unit name	Blue Sequence	SU 8	**SSS IV - V**	*Holocene sediment*
	Green Sequence	SU 6-7	**SSS III**	*MLA and post-Algonquin*
		SU 4-5		*KLA*
		SU 3		*ELA/GLS*
	Purple Sequence	SU 2	**SSS II**	*Valley fill/early deglacial*
	Red Sequence	SU 1	**SSS I**	*Newmarket Till*
	Brown Sequence	n/a	**n/a**	*Thorncliffe Fm.*

The interpretation of SSS I as the Newmarket Till is based primarily on the correlation of the geometry and topography of the upper reflector with mapped till distribution and thickness along the modern shoreline of Kempenfelt Bay (Barnett 1997). Additional support comes from interpretation of Newmarket Till beneath the floor of Lake Simcoe in other sub-bottom investigations (Todd et al. 2008) and recent drilling investigations in Simcoe County, that intersect the Newmarket Till at various depths within tunnel valleys that formed bays in glacial Lake Algonquin (Fig. 3.1; Bajc et al. 2015; Mulligan 2016; 2017c; Mulligan et al. 2018a, b). Recent conclusions presented by Sharpe et al. (2018), who interpret the tunnel valley occupied by Kempenfelt Bay as a tunnel channel formed by collapse of late-stage subglacial meltwater sheet floods that eroded drumlins on the Newmarket Till, are inconsistent with the data presented here.

The limited thickness of SSS II contrasts with the findings of previous studies from Lake Simcoe and the surrounding region (Fig. 3.1d), that interpret thick coarse-grained valley fill deposits (Pugin et al., 1999; Todd et al. 2008; Barros 2011; Sharpe et al. 2013; 2018); however, this may be explained by the time-transgressive nature of ice retreat. Thicker valley fill deposits are observed in the tunnel valley beneath Cook's Bay and further to the south, particularly in closer proximity to the ORM (Russell et al. 2003a, 2003b; Barros 2011; Sharpe et al. 2018), where ice marginal positions may have been more stable for a longer period of time, allowing for thicker coarse-grained sediment accumulation. In contrast, the tunnel valley beneath Kempenfelt Bay was mostly ice-filled during ORM deposition and relatively sediment starved compared to the western (downflow) part of the valley (Mulligan et al. 2018a). Tunnel valleys occurring throughout Simcoe County were almost entirely occupied by glacier ice during the lateglacial, except where localized subglacial conduits existed along valley thalwegs, inhibiting deposition or eroding the Newmarket Till (SSS I) and promoting sediment transport and deposition in ice marginal areas. These factors caused the localized and highly variable thickness of coarse-grained deposits along the length of the valleys by severely restricting accommodation space compared to ice-marginal areas (Barnett et al. 1998; Mulligan et al. 2018a). Thicker successions of coarse-grained material have been intersected by recent drilling investigations in the western extension of Kempenfelt Bay, in the Barrie and base Borden areas (Fig. 3.1; Bajc et al. 2015; Mulligan 2016; 2017c; Mulligan et al. 2018a).

Previously-identified channelized features identified in outcrops of correlative glaciolacustrine successions to SSS III (Mulligan et al. 2018b) were not observed in the present study, likely due to the low-lying valley floor and resulting quiet water environment that persisted in Kempenfelt Bay, even during lowered regional base levels (Lewis et al. 2008). Small (5x4 m) channelized features infilled with coarse-grained deposits were observed in sediment exposures and interpreted in boreholes onshore at elevations between 218 and 207 m asl (Bajc et al. 2015; Mulligan et al. 2018b). The lake bottom in Kempenfelt Bay lies at elevations below this interval (Figs. 3.2-3.7) and would therefore have permitted continuous fine-grained sediment accumulation by suspension settling and minor density flows throughout multiple periods of lake level fluctuation.

3.6.2. Origin of sediment deformation

Brittle faulting, minor folding and slumping of sediments identified in SSS III and IV record post-depositional deformation of the valley fill sediments in Kempenfelt Bay. Evidence for sediment deformation is common throughout Lake Simcoe (Todd and Lewis 1993; Todd et al. 2008; Doughty et al. 2014). Large lobes of deformed and contorted sediment (SSS IIIb) on the flanks of the basin (Figs. 3.3, 3.5, 3.6a) likely record mass wasting event(s), leading to downslope mobilization and slumping of cohesive sediment masses and debris flow lobes following deglaciation (Fig. 3.8c).

Near-vertical brittle faults observed in otherwise undisturbed sediment could result from a variety of processes, including collapse of sediments above melting ice blocks buried during ice retreat (Shilts and Clague 1992), differential sediment compaction and neotectonic faulting induced by either glacial unloading, or by reactivation of deep-seated basement structures underlying Kempenfelt Bay (Boyce et al. 2002; Doughty et al. 2014). The penetration of some faults to various intervals through SSS III and SSS V suggests that these disturbances persisted throughout deglaciation and into modern time, likely precluding an ice-melt origin as cause for all of the observed faulting. The interpreted positive flower structure (Fig. 3.5c; Civile et al. 2014) provides evidence of deformation, either as a result of post-depositional melt-out of buried ice or potential tectonic disturbance by reactivated basement structures. A neotectonic origin for the faults is consistent with the observation of pop-up structures deforming the glacial Lake Algonquin shoreline to the northeast of Lake Simcoe (Rutty and Cruden 1993). Sedimentological investigations in adjacent parts of the paleo Lake Algonquin plain have not encountered any consistent, widespread deformation features in outcrops at corresponding intervals (Mulligan et al. 2018b), but deformed zones are common at various intervals within rhythmite successions deposited in the tunnel valleys elsewhere in Simcoe County (Bajc et al. 2019).

3.6.3. Groundwater resources

Groundwater has previously been suggested to seep into Kempenfelt Bay, from both deeply-buried and near-surface aquifers (Longstaffe et al. 2008; Roy and Malenica 2013; Ji 2016). The high-resolution data presented in this study provide an excellent tool for visualizing the potential connections between pressurized buried aquifers and surficial water reservoirs (Todd et al.

2008). These hydraulic connections may be enhanced in areas where the Newmarket Till is penetrated by faults, contains coarse-grained interbeds (Gerber et al. 2001; Mulligan et al. 2018b) or is absent (eroded?) at the base of the valley fill succession (Desbarats et al. 2001; Sharpe et al. 2002; Mulligan et al. 2018a). Gas/fluid escape features observed along fault planes, indicate that faulting of sediments infilling the valley beneath Kempenfelt Bay has provided enhanced connection for vertical fluid flow between older sediments and SSS II (Todd et al. 2008). With the current data set, it is not possible to identify whether the fluid escape features would have developed during discrete events, synchronous with the development of the faults (Kuscu et al. 2005), or whether these represent persistent migration pathways exploited by pressurized aquifers underlying lowlands throughout Simcoe County (Golder 2004; Bajc et al. 2014; Mulligan 2017c). The pitting observed where faults intersect the lake floor (Fig. 3.7) may provide support for the latter, as depressions caused by discrete fluid migration events in the past would likely have been infilled with recent sediments, and geochemical sampling supports modern groundwater exchange in deep parts of Kempenfelt Bay (Longstaffe et al. 2008).

3.7. SUMMARY

The findings of this study of the tunnel valley infill sediments underlying Kempenfelt Bay support many previous studies of Quaternary processes and stratigraphy based on lake-based seismic (Todd et al. 2008) and onshore data in the region (Barnett 1997; Mulligan and Bajc 2012; Mulligan et al. 2018a 2018b). High-resolution chirp data provide an excellent foundation for the identification of sediment characteristics including fine-scale bedding architecture within the upper parts of the valley infill, as well as fault networks and extensively deformed horizons. The identification of drumlinized Newmarket Till flooring the eastern part of the tunnel valley adds support to recent suggestions of a complex, protracted genesis for the tunnel valleys in the region and demonstrates that the valley form predates till sheet drumlinization (Mulligan et al. 2018a). Eskers on the floor of the valley record the activity of confined subglacial meltwater systems during deglaciation. The sub-bottom stratigraphy beneath Kempenfelt Bay also shows evidence of large-scale deformation events, including possible strike-slip features that may indicate reactivation of Late Proterozoic basement faults during deglaciation. The data here provide a visual picture of how sediment architecture and deformation features form (or enhance) potential hydraulic connections between buried aquifers and modern surface water bodies.

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CHAPTER 4: LOCAL VARIATIONS IN A REGIONAL TILL SHEET IN SOUTH-CENTRAL ONTARIO

Abstract

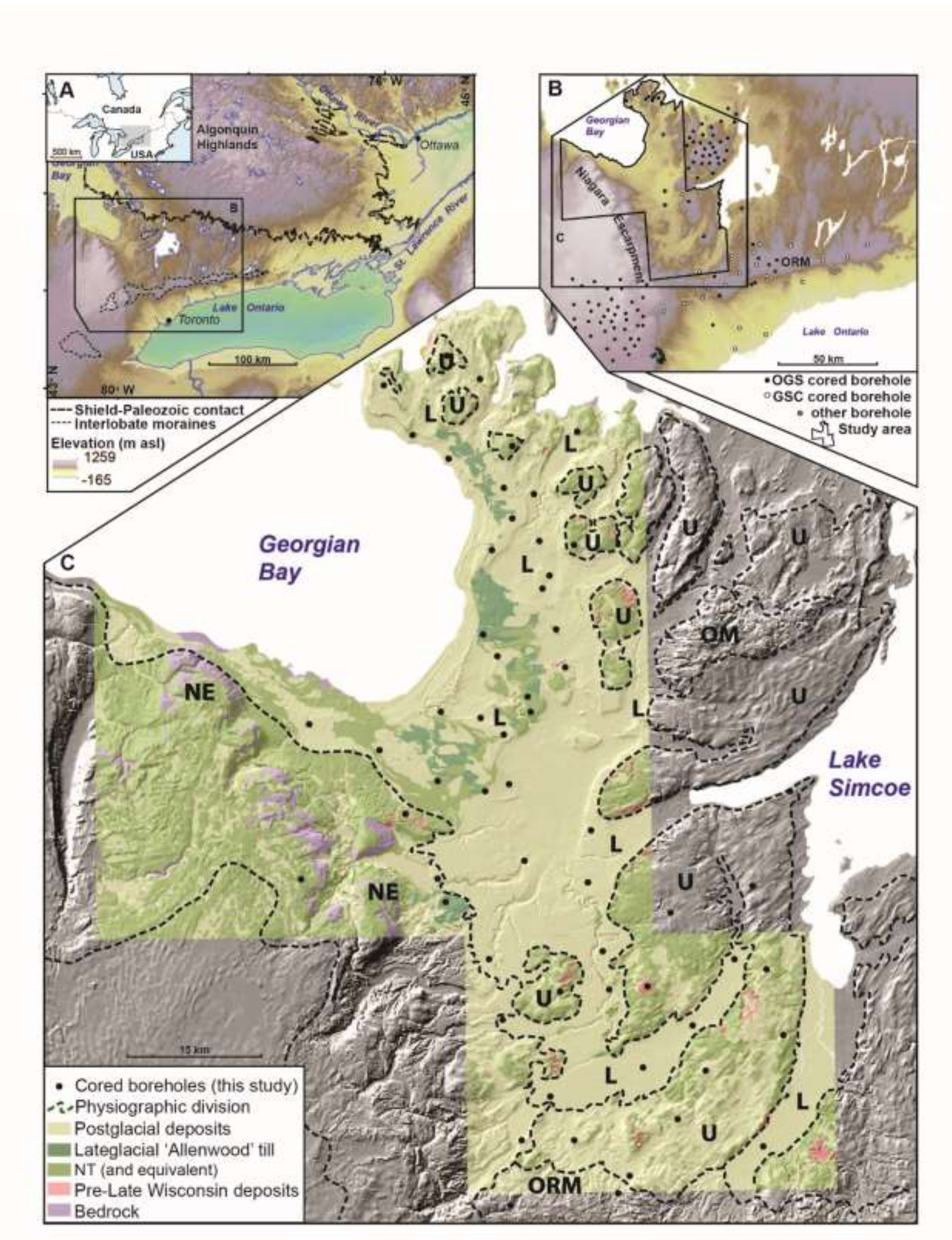
The Newmarket Till (NT) records the advance of the Laurentide Ice Sheet over south-central Ontario during the last glacial maximum. It occurs across three prominent physiographic regions in the study area: the Niagara Escarpment, a regional scale bedrock scarp 200-300 m high; drumlinized uplands underlain by >150 m of Pleistocene sediments; and beneath lowland s. Detailed surficial mapping of over 3750 km^2 has identified consistent internal facies characteristics within the NT capping uplands, but significant spatial and vertical changes in till matrix texture, colour, consolidation, and clast content, fabric and lithology within intervening lowlands and along the Niagara Escarpment. These local variations are governed by changes in substrate lithology and topography that promoted fluctuations in former subglacial erosional and depositional conditions including stress regimes, porewater pressures and subglacial drainage. Combined sedimentological analysis of the till, with stratigraphic information gained from 58 continuously-cored boreholes, and morphological data from high-resolution terrain models, guides the correlation of the NT across the three regions. Documenting local variations within the NT provides a better framework for understanding the controls that subglacial topography and substrate lithology have on till deposition, evolving conditions at the ice-bed interface during glaciation, and the local hydrogeological significance of the NT.

4.1. INTRODUCTION

Despite over a century of study, understanding of processes operating at the base of ice sheets is still incomplete. Detailed investigations of modern glacier forefields have provided a wealth of information as sediment facies observed in modern glacial deposits can, in places, be related to specific events (such as surges/streaming phases/meltwater discharges) documented by satellite imagery and/or aerial photographs (Evans and Twigg, 2002), field instrumentation (Boulton and Zatsepin, 2007), detailed sedimentological logging (Kjaer et al., 2006), or historical records (Bjornsson, 1996). These sources of information, combined with data on the physical conditions at the base of modern glaciers provided by geophysical and drilling investigations (e.g. Smith et al., 2007), are providing critical information for glacial sedimentologists reconstructing processes that created the glaciated landscapes we observe today.

In south-central Ontario, the main phase of the Late Wisconsin (Marine Oxygen Isotope Stage; MIS 2) advance of the Laurentide Ice Sheet (LIS) is recorded by a regionally extensive diamict unit, the Newmarket Till (NT), also referred to as the Northern Till and/or Bowmanville Till (Martini and Brookfield, 1999; Boyce and Eyles, 2000; Sharpe et al., 2004; Sharpe and Russell, 2016; Sookhan et al., 2018). The NT has a highly variable surface expression across the region but is reported to have relatively consistent physical properties that allow it to be interpreted as a subglacial till and make it a reliable regional marker bed in stratigraphic, geophysical, and hydrogeological studies (Boyce et al., 1995; Gerber et al., 2001; Sharpe et al., 2002; Crow et al., 2018). However, despite its regional extent and significance to a wide variety of studies, detailed investigations of its internal characteristics, facies, and physical properties are relatively scarce (see Boyce and Eyles, 2000; Gerber et al., 2001; Sharpe et al., 2002; Meriano and Eyles, 2009; Kjaarsgard et al., 2018). Investigations of the NT to date have focussed

Fig. 4.1: A. Geography and topography of southern Ontario. Overdeepened lake basins are separated by intervening highs, where interlobate moraine complexes (short dashed lines) were locally deposited. Inset shows location of map (shaded area) within Great Lakes region of North America. Location of B outlined in black rectangle. B. Location of study area within south-central Ontario and the location of previous cored boreholes drilled by the Geological Survey of Canada (GSC), Ontario Geological Survey (OGS), and others. C. Simplified surficial geology of the study area overlain on hillshaded elevation model. Locations of cored boreholes supporting this study are shown in black and the divisions between physiographic regions described herein are shown with dashed black lines. NE = Niagara Escarpment; L = Simcoe Lowlands; U = Simcoe Uplands; ORM = Oak Ridges Moraine; OM = Oro Moraine; NT = Newmarket Till. Uncoloured areas lie outside the detailed mapping presented here, but physiographic divisions are drawn for context based on previous mapping (Chapman and Putnam, 1984) and recent DEM analysis.


A
Canada
USA
500 km
Algonquin
Highlands
Ottawa
Georgian
Bay
St. Lawrence River
Lake Ontario
Toronto
100 km
74° W
45° N
80° W
43° N
Shield-Paleozoic contact
Interlobate moraines
Elevation (m asl)
1259
-165
B
Georgian Bay
Niagara Escarpment
C
ORM
Lake Ontario
50 km
OGS cored borehole
GSC cored borehole
other borehole
Study area
C
Georgian Bay
Lake Simcoe
NE
L
U
OM
ORM
15 km
Cored boreholes (this study)
Physiographic division
Postglacial deposits
Lateglacial 'Allenwood' till
NT (and equivalent)
Pre-Late Wisconsin deposits
Bedrock

primarily on documentation of the properties of the till from drumlinized upland regions, whereas NT occurrences in other settings are largely unstudied (*cf*. Sharpe et al., 2011; 2018; Mulligan et al., 2018a, b).

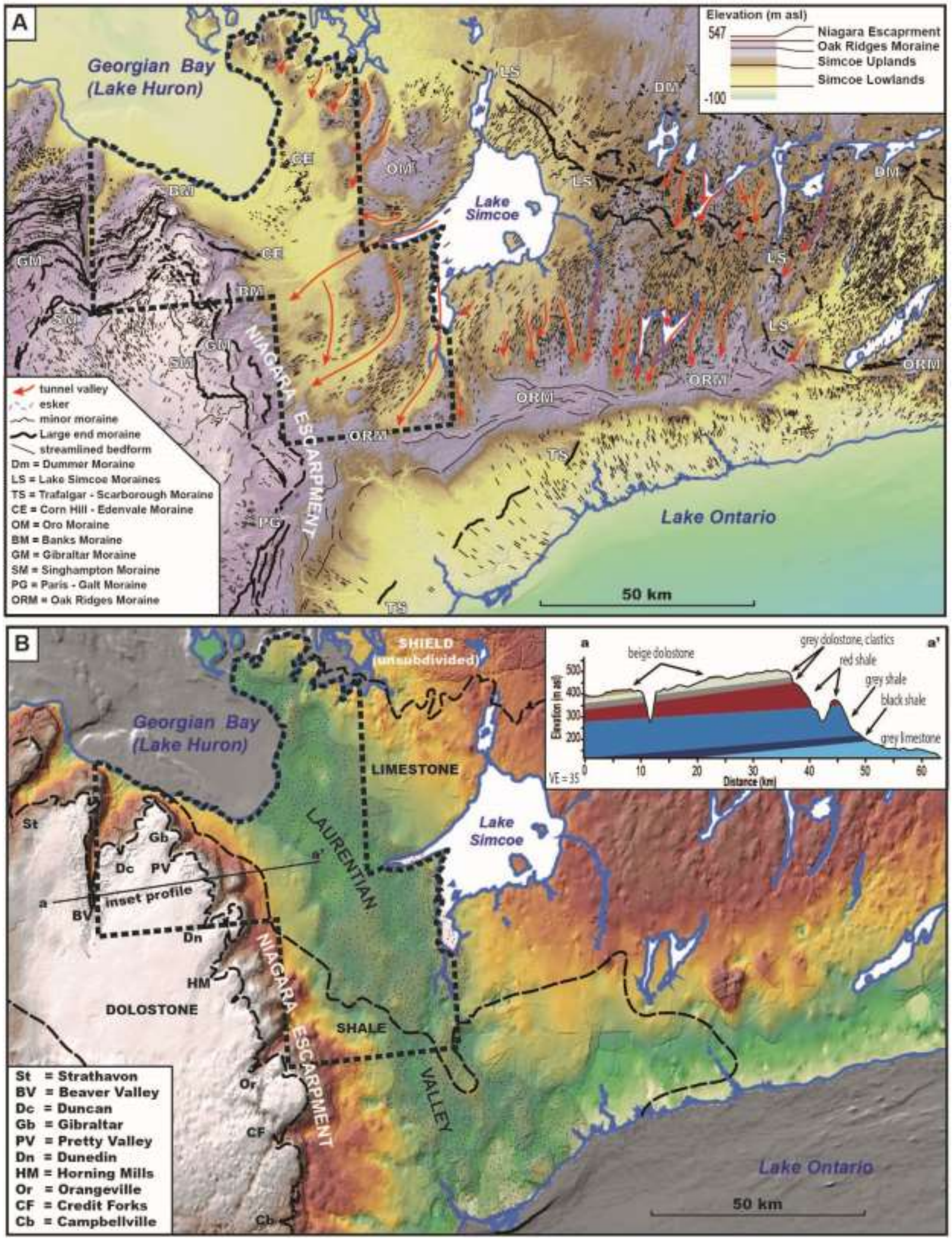
This study presents descriptions of the morphological properties, internal characteristics, and stratigraphic relationships of the NT in multiple physiographic settings within south-central Ontario (Fig. 4.1). These properties record deposition of the NT under a range of subglacial conditions resulting from the spatial and temporal variability in processes operating below the LIS as it advanced over a heterogeneous substrate and irregular bed topography. Descriptions are presented from three distinct settings: the Niagara Escarpment, Simcoe uplands, and Simcoe lowlands, areas that remain as prominent physiographic regions on the modern landscape (Figs. 4.1 and 4.2; Chapman and Putnam, 1984). The purpose of this paper is to highlight local variations in the sedimentology and physical properties of the NT across a complex glaciated landscape. Understanding the nature of these variations within a single till sheet, commonly viewed as a regional stratigraphic marker bed, can provide further insight into the depositional evolution of regional tills with respect to local geological controls and the nature of transient bed conditions beneath former ice sheets.

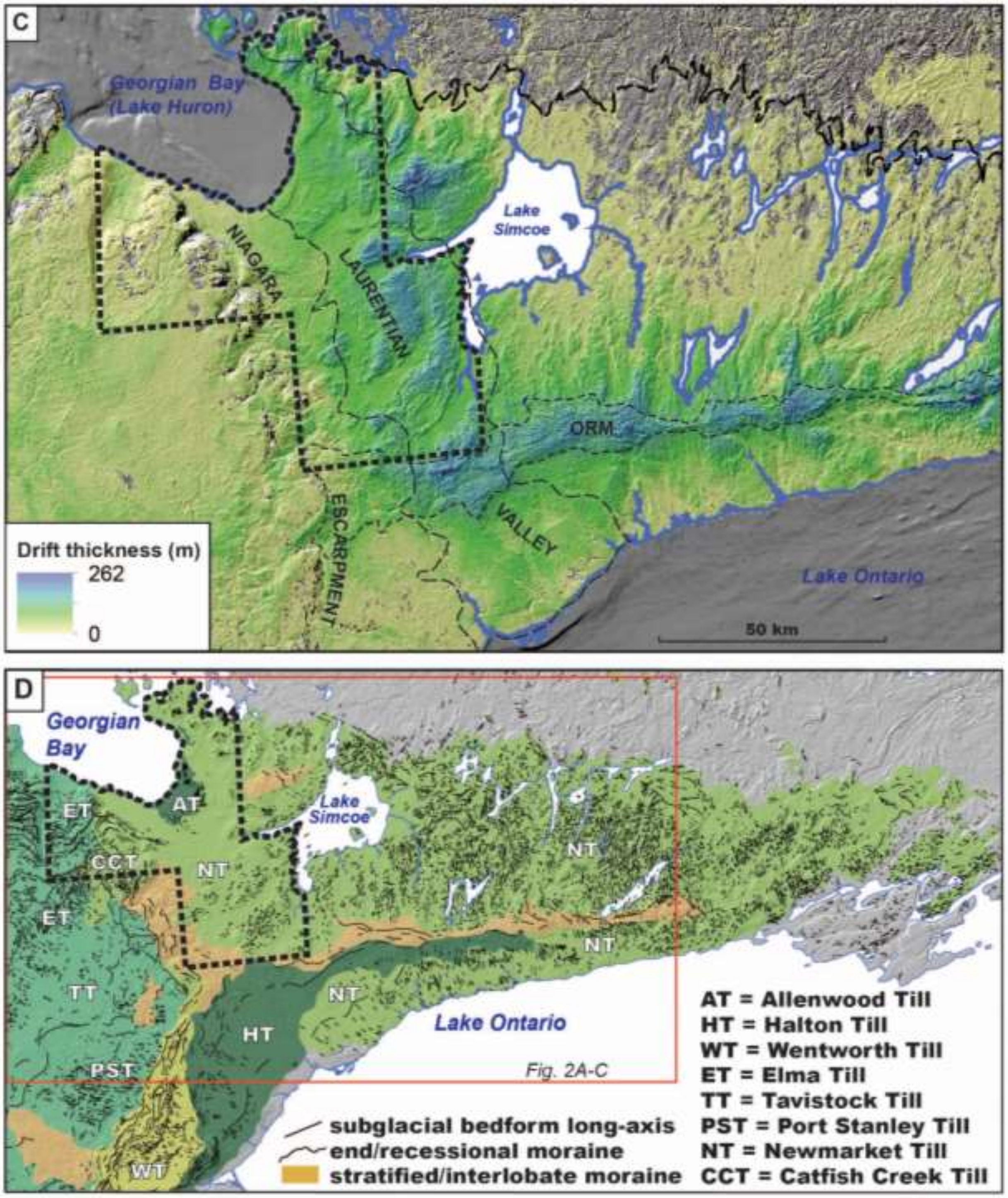
4.2. BACKGROUND

4.2.1. Regional geologic setting

The study area lies to the southwest of the Algonquin Highlands of the Canadian Shield and to the southeast of the overdeepened Georgian Bay basin (Fig. 4.1). The Niagara Escarpment lies along the western margin of the study area, forming the most prominent landform in southern Ontario (Figs. 4.1-4.4). The Oak Ridges Moraine, a sandy interlobate landform

Fig. 4.2: A) topography and glacial landforms in south-central Ontario. Prominent moraine systems are labelled in white acronyms (see key in lower left); B) major bedrock lithological divisions (long dashed lines; Armstrong and Carter, 2010) and locations of re-entrant valleys (Straw, 1968; see key in lower right) overlain on bedrock topography (Gao et al., 2006). Inset profile a-a' in upper right (see thin black line for location) shows generalized stratigraphy of the Niagara Escarpment cuesta. The Laurentian Valley/trough is shaded and stippled; C) drift thickness map of south-central Ontario (Gao et al., 2006). Little to no sediment occurs on Shield terrain and along the face of the Niagara Escarpment, whereas thick sediment successions occur within the Laurentian Valley/trough and the Oak Ridges Moraine (ORM); D) previously-mapped distribution of Late Wisconsin (MIS 2) till sheets in south-central Ontario (Ontario Geological Survey, 2010). In all frames, the study area is represented by bold black dashed polygon.


A
Georgian Bay
(Lake Huron)
Lake Simcoe
Lake Ontario
Elevation (m asl)
547
-100
Niagara Escaprment
Oak Ridges Moraine
Simcoe Uplands
Simcoe Lowlands
NIAGARA ESCARPMENT
50 km
tunnel valley
esker
minor moraine
Large end moraine
streamlined bedform
Dm = Dummer Moraine
LS = Lake Simcoe Moraines
TS = Trafalgar - Scarborough Moraine
CE = Corn Hill - Edenvale Moraine
OM = Oro Moraine
BM = Banks Moraine
GM = Gibraltar Moraine
SM = Singhampton Moraine
PG = Paris - Galt Moraine
ORM = Oak Ridges Moraine
B
SHIELD
(unsubdivided)
Georgian Bay
(Lake Huron)
LIMESTONE
LAURENTIAN
Lake Simcoe
inset profile
DOLOSTONE
SHALE
VALLEY
NIAGARA ESCARPMENT
Lake Ontario
50 km
beige dolostone
grey dolostone, clastics
red shale
grey shale
black shale
grey limestone
Elevation (m asl)
Distance (km)
VE = 35
St = Strathavon
BV = Beaver Valley
Dc = Duncan
Gb = Gibraltar
PV = Pretty Valley
Dn = Dunedin
HM = Horning Mills
Or = Orangeville
CF = Credit Forks
Cb = Campbellville


C
Georgian Bay
(Lake Huron)
Lake
Simcoe
NIAGARA
LAURENTIAN
ORM
ESCARPMENT
VALLEY
Lake Ontario
Drift thickness (m)
262
0
50 km
D
Georgian
Bay
Lake
Simcoe
AT
ET
CCT
NT
ET
TT
PST
WT
HT
NT
NT
NT
Lake Ontario
Fig. 2A-C
subglacial bedform long-axis
end/recessional moraine
stratified/interlobate moraine
AT = Allenwood Till
HT = Halton Till
WT = Wentworth Till
ET = Elma Till
TT = Tavistock Till
PST = Port Stanley Till
NT = Newmarket Till
CCT = Catfish Creek Till

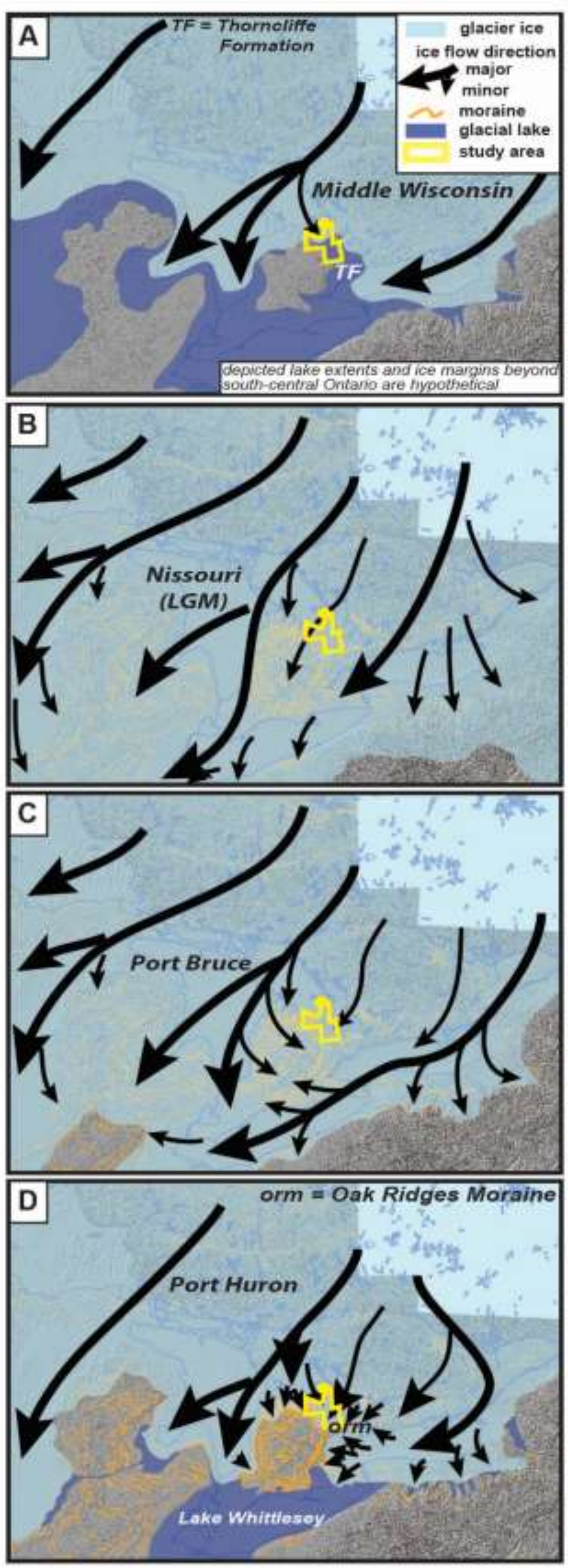
complex that stretches over 160 km from the Niagara Escarpment in the west toward the eastern end of Lake Ontario, forms the southern margin of the study area (Figs. 4.1, 4.2A, 4.4). Uplands capped by drumlinized NT and intervening lowlands (interpreted as tunnel channels/valleys; *cf.* Brennand and Shaw, 1994; Mulligan et al., 2018a) stretch from the study area to the eastern end of Lake Ontario (Fig. 4.2A).

South-central Ontario is underlain by gently dipping Paleozoic strata of Upper Ordovician and Silurian age that onlap the Canadian Shield and the Algonquin Highlands to the north (Fig. 4.2B). Bedrock topography of both the Shield and overlying carbonate and shale strata is lowest near the overdeepened Great Lakes basins (Fig. 4.2B; Gao et al., 2006). The Niagara Escarpment forms a 200-300 m high cuesta comprising Silurian caprock dolostone overlying more easily eroded siliciclastic rocks of Early Silurian and Late Ordovician age (Fig. 4.2B; Brunton et al., 2009). The face of the cuesta is marked by numerous sub-parallel re-entrant valleys oriented NE-SW in the southern part of the study area and N-S in the northern parts (Fig. 2B; Straw, 1968; Eyles et al., 1997). The re-entrants are fairly regularly spaced, but display highly variable morphologies, ranging from 1.5 - 8 km wide, 2.5 – 16.5 km long, and can exceed 300 m of relief (Fig. 4.2B; Straw, 1968). East of the Niagara Escarpment, a broad low on the Paleozoic bedrock surface spans the area between Georgian Bay and Lake Ontario (the Laurentian Valley, Fig. 2B), but remains poorly-defined, despite recent collection of geological and geophysical data, due to a thick cover (>200 m locally) of Pleistocene sediment (Fig. 4.2C; Spencer, 1890; Gao et al., 2006; Mulligan, 2017a; Sharpe et al., 2018; Bajc et al., 2019).

4.2.2. Glacial history

Glacial sediments in southern Ontario were deposited primarily during the Wisconsin Episode (MIS 2-4). Extensive lakes developed within the Great Lakes region during build-up and

Fig. 4.3: Glacial history of the study area. A) extensive proglacial lakes fronting the advancing LIS towards the end of the Middle Wisconsin Episode (MIS 3); B) regional-scale southwestward flow during the last glacial maximum (LGM; MIS 2); C) thinning of ice sheets during deglaciation promotes the development of topographically-controlled ice lobes/streams flowing out of the Great Lakes basins; D) late-glacial ice margins and ice flow directions in southern Ontario. Ice flowed radially from within lake basins where the thickest residual ice masses remained.


A
TF = Thorncliffe Formation
glacier ice
ice flow direction
major
minor
moraine
glacial lake
study area
Middle Wisconsin
TF
depicted lake extents and ice margins beyond south-central Ontario are hypothetical
B
Nissouri (LGM)
C
Port Bruce
D
orm = Oak Ridges Moraine
Port Huron
orm
Lake Whittlesey

advance of the LIS (Fig. 4.3A) beginning in the Early Wisconsin (MIS 4). During the peak of Late Wisconsin (MIS 2) glaciation, regional ice flow was to the south and southwest, and largely unaffected by local bed topography (Fig. 4.3B). However, during deglaciation, thinning of the ice sheet initiated the development of topographically constrained and independent ice lobes (Barnett, 1992) or ice streams (Eyles, 2012; Eyles and Doughty, 2016; Sookhan et al., 2018a; Eyles et al., 2018; Mulligan et al., 2019) controlled by the deep Great Lakes basins and intervening structural highs (Fig. 4.3B,C). Late Wisconsin (MIS 2) till plains are commonly buried by glaciolacustrine sediments deposited in extensive proglacial lakes (Fig. 4.3D), which also facilitated rapid ice flow and ice lobe/stream re-advances (Eyles et al., 2018).

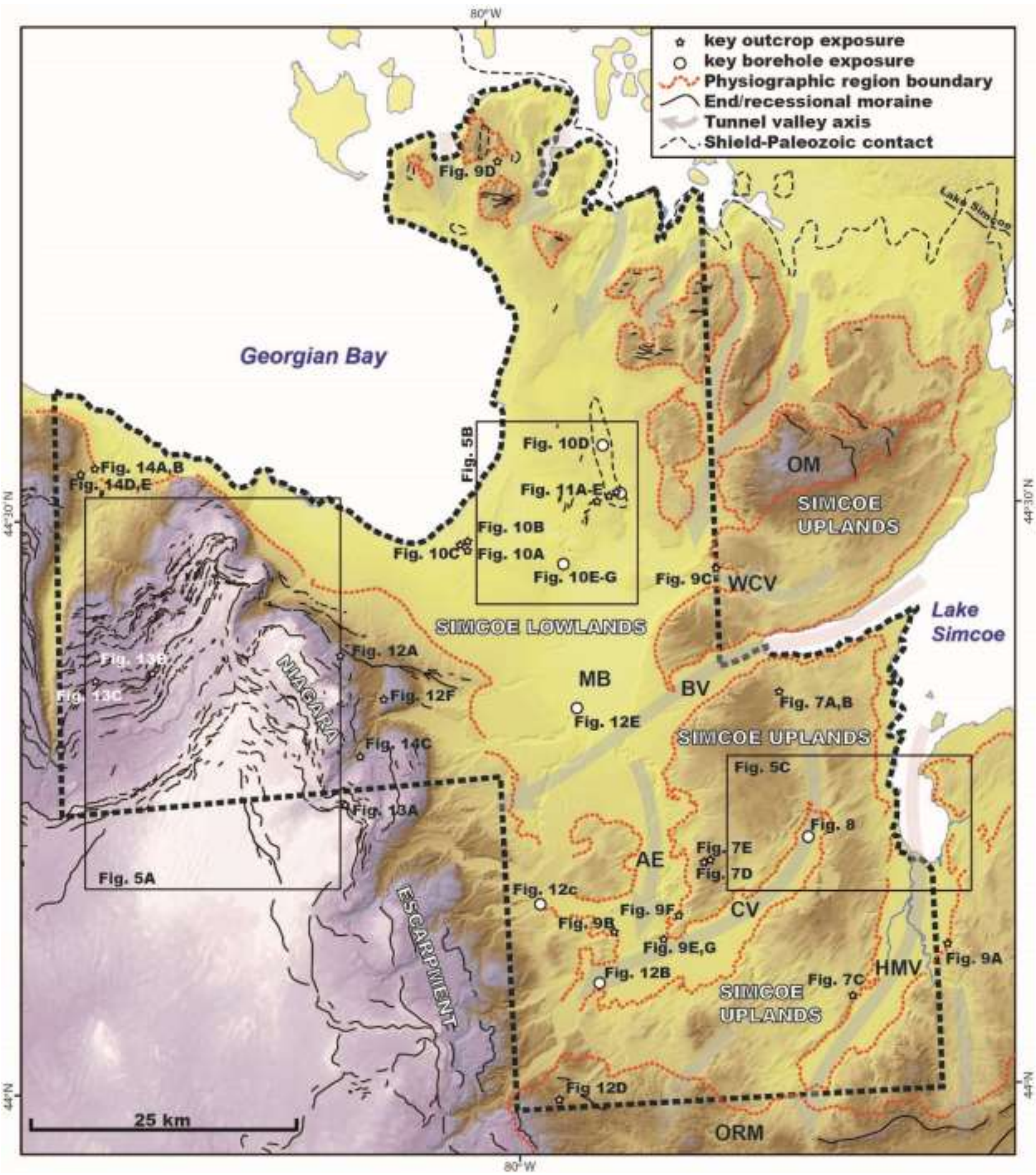
Well studied outcrops of the NT occur mostly on broad drumlinized uplands (Boyce et al., 1995; Boyce and Eyles, 2000; Sharpe et al., 2004; 2018; Mulligan et al., 2018a); however, recent investigations indicate that the NT also occurs within intervening tunnel valleys buried beneath glaciolacustrine sediments (Fig. 4.2A; Bajc et al., 2015; Mulligan et al., 2018a, b). The NT has also been mapped above the Niagara Escarpment, but only as far west as the Singhampton Moraine, which marks the previously-interpreted downflow limit of ice responsible for deposition of the NT (Figs. 4.2D and 4.4; Gwyn and Cowan, 1976). West of the Singhampton Moraine, Late Wisconsin (MIS 2) ice flow is recorded by the Catfish Creek Till and a series of younger tills including the Port Stanley, Tavistock, Elma, Wentworth and Halton tills (Figs. 4.2D and 4.4; see Karrow, 1974; Barnett, 1992 for discussion). A distinct younger till, the Allenwood Till was previously interpreted to overlie the NT in the study area immediately south and east of Georgian Bay and was assumed to have been deposited during a local re-advance during deglaciation of the area (Burwasser, 1974; Figs. 4.2D and 4.4).

4.3. METHODS

Data presented here were collected during 3D sediment mapping investigations conducted by the Ontario Geological Survey in support of regional groundwater investigations beginning in 2010 (Bajc and Rainsford, 2010; Mulligan, 2014). Detailed sediment and landform mapping was undertaken across the study area and immediate surroundings by analyzing high-resolution terrain data sets (Ministry of Natural Resources and Forestry, 2010; 2014; 2016) and examining natural exposures along stream and river cut banks, wave-cut bluffs, groundwater piping scars, and in manmade exposures in gravel pits, quarries, excavations, road and ditch cuts, auger and sediment probes, and test pits. Sections were cleaned, photographed and logged using standard sedimentological logging techniques recording texture, sorting, sedimentary and/or deformation structures, clast content and lithology, long axes of boulders and striated/faceted cobbles.

Subsurface data presented here were collected from continuously-cored boreholes drilled using a mud rotary drilling system equipped with a Christiansen core barrel retrievable by wireline. Core recovery for the project was very high (>90%), except in gravelly sediments and poorly-consolidated diamicts, due to limitations of the drilling method. Cores were photographed at 25-cm intervals, logged using standard logging techniques, and sampled at least every meter or where changes in sediment lithology were noted (Bajc et al., 2015). Sediment samples were analyzed for particle size, calcite:dolomite ratios, and heavy mineral abundances.

Fig. 4.4: Topography of the study area (thick dashed line) and location of key exposures of the NT in outcrops (stars) and boreholes (circles). Tunnel valleys (transparent grey arrows) are labelled WCV = Willow Creek Valley; BV = Barrie Valley; MB = Minesing Basin; AE = Alliston Embayment; CV = Cookstown Valley; HMV = Holland Marsh Valley. The Oak Ridges (ORM) and Oro (OM) interlobate moraines and physiographic regions are outlined in red.



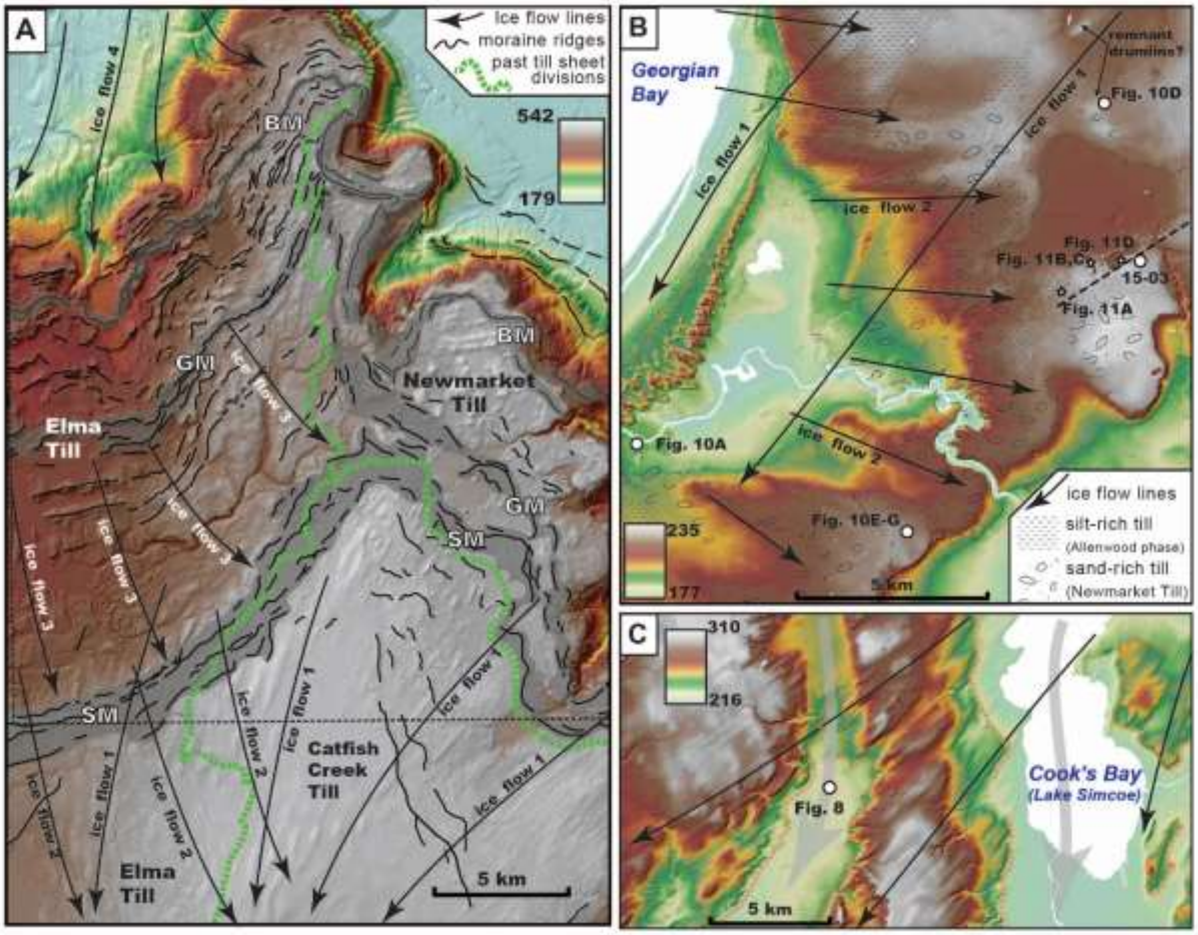
key outcrop exposure
key borehole exposure
Physiographic region boundary
End/recessional moraine
Tunnel valley axis
Shield-Paleozoic contact
Georgian Bay
Lake Simcoe
SIMCOE LOWLANDS
SIMCOE UPLANDS
NIAGARA ESCARPMENT
OM
WCV
MB
BV
AE
CV
HMV
ORM
Fig. 5A
Fig. 5B
Fig. 5C
Fig. 9D
Fig. 10D
Fig. 11A-E
Fig. 10B
Fig. 10A
Fig. 10C
Fig. 10E-G
Fig. 9C
Fig. 14A,B
Fig. 14D,E
Fig. 12A
Fig. 12F
Fig. 12E
Fig. 7A,B
Fig. 14C
Fig. 13A
Fig. 8
Fig. 7E
Fig. 7D
Fig. 12c
Fig. 9F
Fig. 9B
Fig. 9E,G
Fig. 12B
Fig. 7C
Fig. 9A
Fig 12D
25 km
80°W
44°30'N
44°N

4.4. RESULTS

The NT forms a broad belt of drumlinized terrain stretching from the Niagara Escarpment to the east end of Lake Ontario, and southward into the New York Drumlin Field (Figs. 4.1, 4.2A,D and 4.5). Multiple cross-cutting bedform populations occur in the vicinity of the Niagara Escarpment (Fig. 4.5A) and in the broad lowlands southeast of Georgian Bay (Fig. 4.5B); multiple till sheets have previously been mapped in these areas (OGS, 2010; Fig 4.2D). A single set of bedforms ornaments the NT across the surface and flanks of the Simcoe uplands and into the subsurface beneath intervening lowlands (Fig. 4.5C).

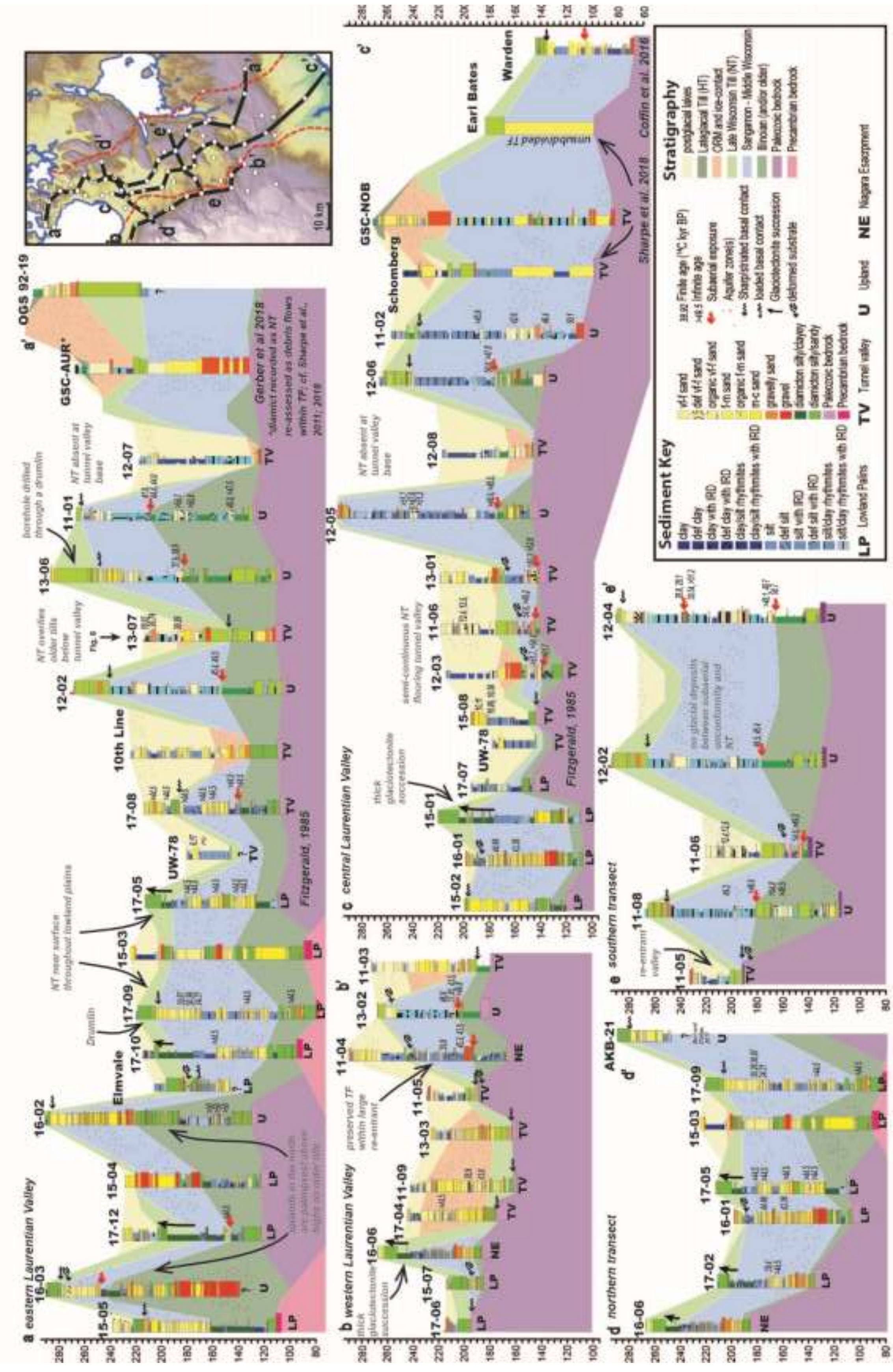
The surface of the NT shows a considerable amount of morphological variation across the Simcoe uplands, lowlands and Niagara Escarpment in south-central Ontario, and the development of Late Wisconsin (MIS 2) tunnel valleys (Mulligan et al., 2018a) produced a complex regional stratigraphic architecture (Fig. 4.6). The NT occurs at a range of elevations, overlying thick (>150 m) pre-Late Wisconsin sediment successions beneath uplands and the lowland plains south and east of Georgian Bay (Fig. 4.6), underlying >100 m of sediments within tunnel valleys in the central parts of the study area, and directly overlying Paleozoic bedrock in parts of some tunnel valleys as well as along the Niagara Escarpment (except in parts of some large re-entrant valleys, where thick, pre-Late Wisconsin sediment successions are preserved; Fig. 4.6). NT facies are described below from these three distinct, but related physiographic regions (simplified from Chapman and Putnam, 1984); uplands (primarily drumlinized areas of Simcoe County); lowlands (tunnel valleys and lowland plains); and the Niagara Escarpment (escarpment flanks, crest, and re-entrant valleys, Figs. 4.1 and 4.4).

Fig. 4.5: High-resolution digital elevation models (MNRF 2010; 2016) displaying the surface morphology of the NT within selected parts of the study area (see Fig. 4.4 for locations). A) Up to four generations of streamlined bedforms exist on the NT-Elma-Catfish Creek Till surface near the Beaver Valley along the Niagara Escarpment, several extending beyond previously-mapped till sheet boundaries (dashed green lines) and beneath major moraine systems (shaded grey areas; SM = Singhampton Moraine; GM = Gibraltar Moraine; BM = Banks Moraine); B) southeastward streamlining of coarse- (stipple pattern, ice flow 1) and overlying fine-grained till (formerly the Allenwood Till; dashed pattern, ice flow 2) southeast of Georgian Bay indicating late-stage ice flow switching. C) drumlinized Newmarket Till surface along uplands and the flanks of intervening tunnel valleys (grey arrows). Note the strongly oblique angle of drumlins to tunnel valley long axes in the north. Dashed red line marks the main Lake Algonquin shore bluff and boundary between upland and lowland physiographic regions.



A
ice flow lines
moraine ridges
past till sheet divisions
542
179
BM
GM
SM
Elma Till
Newmarket Till
Catfish Creek Till
ice flow 1
ice flow 2
ice flow 3
ice flow 4
5 km
B
Georgian Bay
remnant drumlins?
Fig. 10D
Fig. 11D
Fig. 11B,C
15-03
Fig. 11A
Fig. 10A
Fig. 10E-G
235
177
silt-rich till
(Allenwood phase)
sand-rich till
(Newmarket Till)
C
310
216
Fig. 8
Cook's Bay
(Lake Simcoe)

Fig. 4.6: Generalized stratigraphic architecture throughout the Laurentian Valley (dashed red lines on inset map; Fig 2B). Key landform-sediment successions are highlighted in arrows and italics. Radiocarbon dates are provided in ^{14}C kyr BP (Bajc et al., 2014; Mulligan et al., 2018b; see Appendix). Acronyms below logs denote the physiographic region within which the borehole was drilled. Boreholes beginning with 11-, 12-, 13- are from Bajc et al. (2015). Boreholes beginning with 15-, 16-, 17- are from Mulligan (2016; 2017c). Sources for other cored boreholes provided below the logs. Def = deformed; IRD = Ice-rafted debris; vf-f = very fine to fine; m-c = medium to coarse.

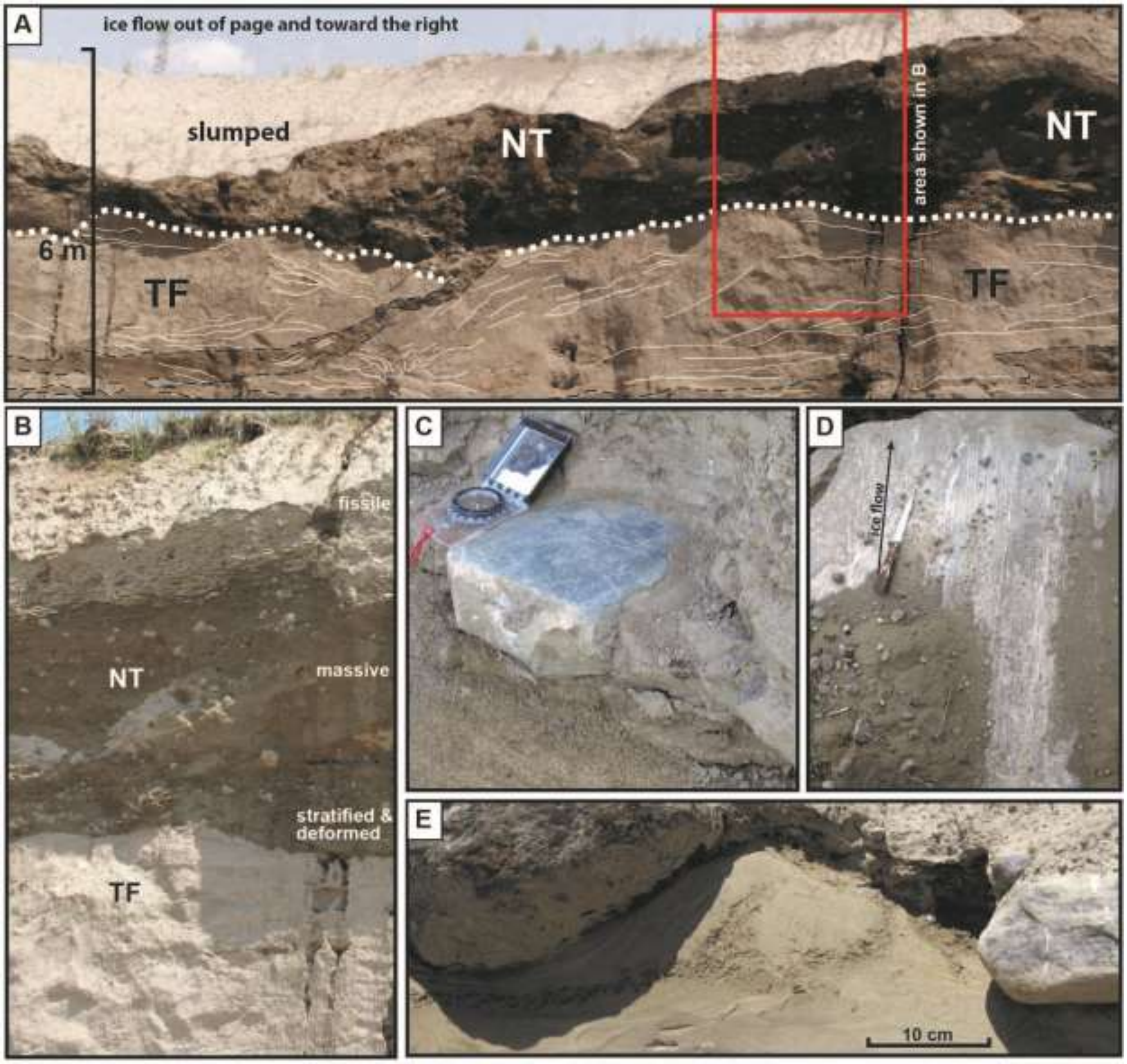


4.4.1. Uplands

The NT shows the most consistent facies characteristics where it caps broad uplands in Simcoe County at elevations of between 225 and 330 m asl. It generally consists of a well to highly-consolidated, grey to buff brown diamict that forms successions 8-12 m thick, locally reaching thicknesses in excess of 30 m (Fig. 4.6). Across the majority of the uplands region, the NT is massive, although crude stratification of stacked, massive diamict beds 0.25 – 5 m thick is locally observed (Fig. 4.7A,B). In the extensively drumlinized terrain southwest of Lake Simcoe (Figs. 4.4 and 4.5C), the cores of drumlins are composed of dense fissile till (Fig. 4.7B), whereas the flanks and lee sides of the drumlins are locally looser and more sand-rich (Deane, 1950). Carbonate clasts are commonly faceted and ornamented with long, continuous striae and gouges (Fig. 4.7C,D) and are aligned with sub-horizontal shear planes which parallel the long axes of local drumlins. Stratified interbeds within the NT are rare, generally thin (<20 cm), and discontinuous (<10 m long). Massive, planar laminated or deformed sands predominate within stratified interbeds, with rarer gravel or silt.

The basal contact of the NT is variable across upland regions. Locally, clasts at the base of the till are faceted and striated and form part of a planar and sub-horizontal basal contact with underlying sand or silt and clay successions (Figs. 4.6 and 4.7A-D). Planar contacts occur preferentially where massive and/or fissile facies form the entire NT succession. Sands below the planar basal contact may be intact, truncated, deformed (folded, sheared, faulted), and in places sand is incorporated into the base of the till (Fig. 4.7A). In other areas, the basal contact of the NT is transitional and commonly loaded, passing from interbedded and deformed silt, sand and diamict, upwards into massive and/or fissile NT forming a highly irregular base to the till (Fig. 4.7E). Sand below loaded NT basal contacts, are commonly undisturbed, except for a thin (2-15

Fig. 4.7: A) surficial exposure of the NT and the underlying upper Thorncliffe Formation (TF) which is deformed in the uppermost 1.5 m, above a 20-40 cm thick silt bed (dashed black lines). Red box denotes location of close-up image in B), modified from Mulligan and Bajc (2018); B) vertical facies gradation within the NT from stratified and deformed diamict and gravelly sediments upwards into highly consolidated massive and fissile silty sand till; C) striated and faceted boulder at the base of the NT where it overlies coarse sand of the TF. Ice flow was from the upper right (northeast); D) striated and faceted boulder at the base of the NT where it overlies thick silt and clay deposits. Ice flow from the bottom of the photo (towards the WSW), note termination of striae at plucked surface in the upper part of the photo, knife is 25 cm long; E) loaded contact (see flame developed in dark brown silt between cobbles) between very fine sand and silt of the TF and interstratified diamict and sands forming the base of the overlying NT succession, approximately 700 m east of photo in D). See Fig. 4.4 for photo locations.


A
ice flow out of page and toward the right
slumped
NT
NT
area shown in B
6 m
TF
TF
B
fissile
NT
massive
stratified & deformed
TF
C
D
ice flow
E
10 cm

cm) horizon of water escape and load structures in the uppermost part of the underlying sediment (Fig. 4.7E).

4.4.1.1. Interpretation

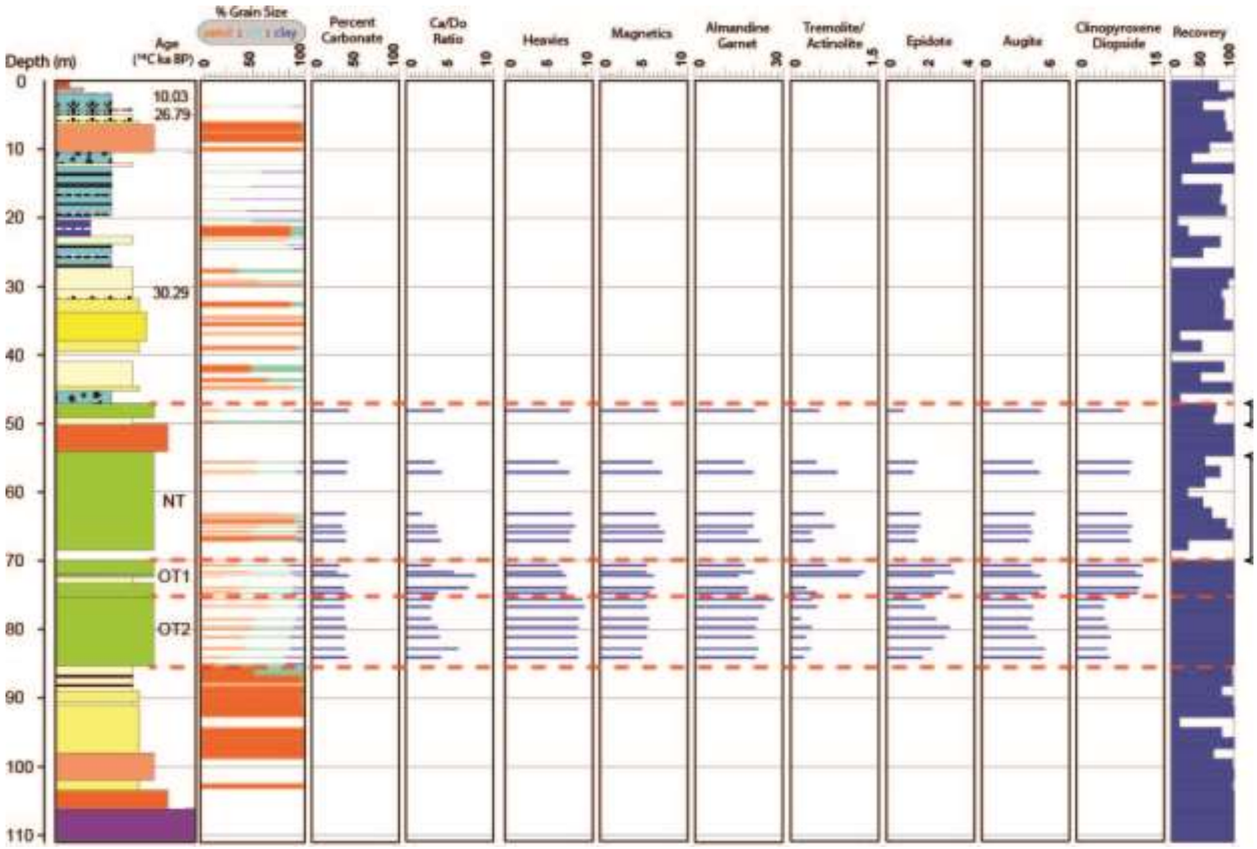
The NT underlying upland regions shows considerable consistency in physical characteristics but displays local variability that indicates complex depositional controls beneath the LIS during its deposition. The high degree of matrix consolidation has typically been attributed to high strain within the deforming zone beneath the LIS (Boyce and Eyles, 2000), but secondary cement may be a contributor to this (Kjarsgaard et al., 2018). Extensive zones of well-developed fissility suggest vertically-isolated shearing processes operated within the accreting till (Phillips et al., 2018). This, together with the presence of faceted and striated boulders that parallel shear planes and regional drumlin orientations, indicate a strongly coupled ice-bed interface and top-down lodgement deposition and incremental thickening of the NT across extensive tracts of uplands in the study area. Massive NT facies record homogenization of subglacial debris within the traction zone at the ice base (e.g. Evans et al., 2006). Rare stratified interbeds, deposited in local conduits or cavities, suggest brief periods of local ice-bed separation (Boyce and Eyles, 2000); the sheared contacts separating stacked massive diamict beds comprising the NT may be the result of periods of basal sliding interrupting till accretion and bed deformation (e.g. Piotrowski et al., 2006).

Undisturbed sand facies below the base of the NT indicate a lack of stress transfer into the bed, possibly during basal sliding (Piotrowski et al., 2001; Menzies and van der Meer, 2017), whereas deformed substrates indicate coupling of ice, till, and substrate sediments (e.g. Kjaer et al., 2006; Bird et al., 2018). Faceted and striated clasts observed at the basal contact of the NT indicate early lodging of clasts into an immobile substrate and subsequent overriding by a mobile traction layer. The preferential occurrence of loaded basal contacts where the lower part of the NT is composed of interbedded and deformed diamict and sorted sediments (interpreted here as subaqueous debris flows; e.g. Barnett and Karrow, 2018) suggests that debris flows were deposited prior to ice advance, probably in an ice marginal lake, protecting underlying substrate sediments from deformation. These stratified debris flow facies become increasingly homogenized upward, transitioning into massive and fissile NT and record overriding of the region by the LIS during the Late Wisconsin (MIS 2).

4.4.2. Lowlands

Analysis of digital elevation models suggests the NT extends from the drumlinized uplands, down their flanks and into the subsurface beneath thick successions of postglacial fine grained glaciolacustrine sediment and ice-contact sand and gravel infilling lowland tunnel valleys (Fig. 4.5C); this has been confirmed through surficial mapping and sediment drilling (Mulligan et al., 2018a, b) as well as geophysical surveys (Todd et al., 2008; Mulligan et al., 2018a). Within the broad lowland plains southeast of Georgian Bay, the NT occurs at or near the surface across a wide area (Figs. 4.2D and 4.5B) in NE-SW oriented remnant drumlins and as a wave-washed till plain ornamented by several mega-scale glacial lineations (MSGL) oriented perpendicular to the modern shoreline (Fig. 4.5B). In addition to its morphological continuity from the NT on uplands, stratigraphic information reveals the till encountered in lowlands to be Late Wisconsin (MIS 2) in age, as it overlies Middle Wisconsin glaciolacustrine sediments (>45 (beyond the limits of radiocarbon dating) – 28,060 ± 230 ^{14}C yr BP (A2449)) and underlies postglacial sediments (<12,805 ± 40 ^{14}C yr BP (A3041); Fig 4.6; dates from Bajc et al., 2014; see Appendix) and only the NT has been encountered at this interval in south-central Ontario.

Fig. 4.8: Stratigraphic log, particle size analysis from sampled intervals, carbonate and heavy mineral analysis of diamict units, and recovery statistics from borehole 13-07 drilled within a Late Wisconsin (MIS 2) tunnel valley (Figs. 4.4 and 4.5; Fig. 4.6 for legend). Thick diamict successions between 48 and 86 m can be correlated as three distinct till units (separated by dashed red lines) based on matrix texture, carbonate/dolomite ratios, heavy mineral abundances, and core recovery rates (strongly related to matrix consolidation). NT = Newmarket Till; OT1, OT2 = Illinoian (MIS 6) and/or older tills.

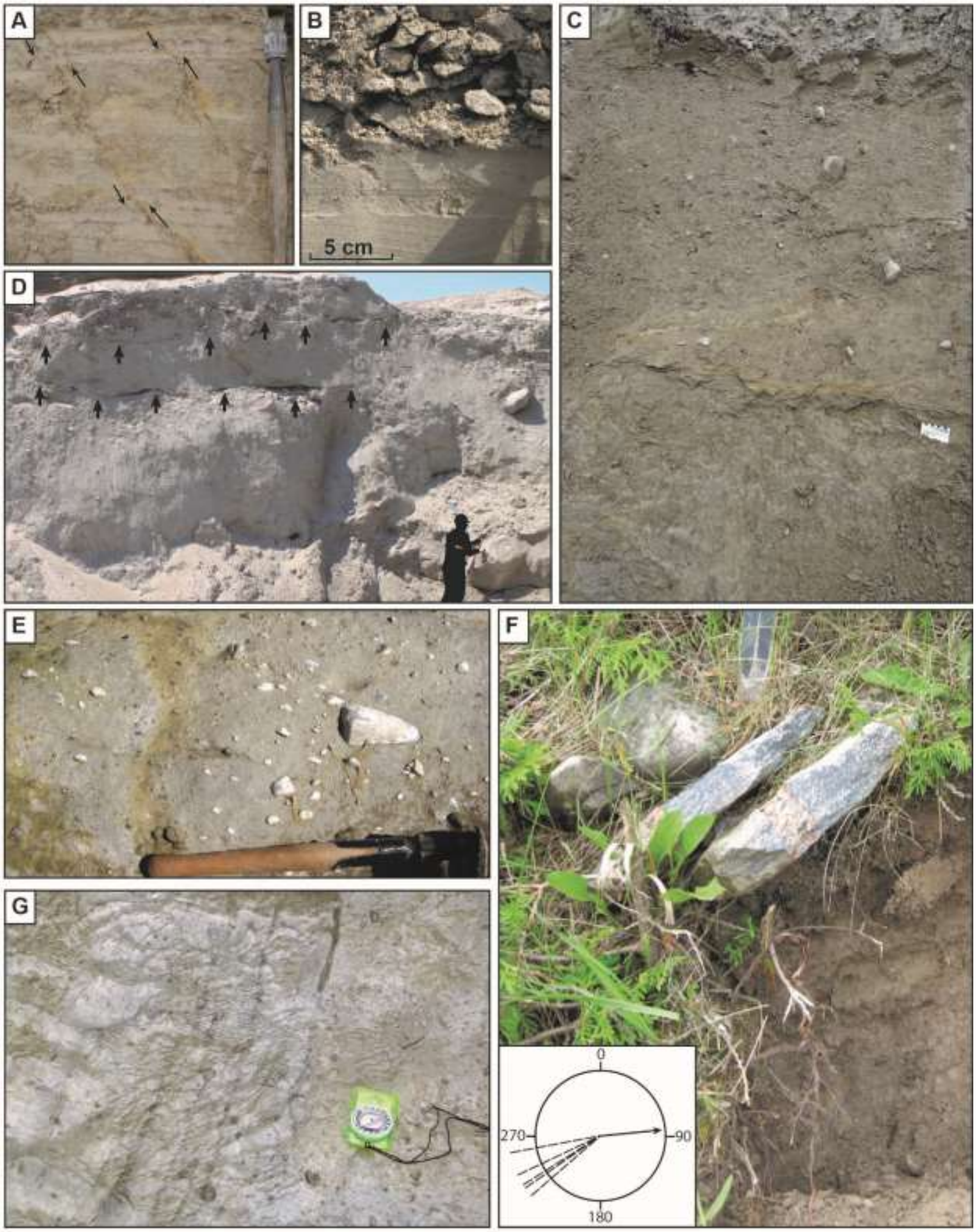


Depth (m)
Age (14C ka BP)
% Grain Size
Percent Carbonate
Ca/Do Ratio
Heavies
Magnetics
Almandine Garnet
Tremolite/ Actinolite
Epidote
Augite
Clinopyroxene Diopside
Recovery
10.03
26.79
30.29
NT
OT1
OT2

4.4.2.1. Tunnel Valleys

The NT directly overlies striated or deformed/fractured (glaciotectonized?) bedrock in the deepest and westernmost parts of some tunnel valleys (Mulligan et al., 2018a); in most tunnel valleys it overlies laminated silt, clay or sand of the Thorncliffe Formation (Fig. 4.6), and in one borehole it directly overlies texturally similar Illinoian tills (Figs. 4.6 and 4.8). Along the flanks of tunnel valleys, outcrop exposures of the base of the NT typically reveal a sharp contact with deformed silt-clay rhythmites (Fig. 4.9B) of the Thorncliffe Formation. The NT found within tunnel valleys displays a higher degree of heterogeneity and interbedding of sorted beds and lenses than in adjacent upland regions. Here, it is most commonly composed of stacked diamict beds, locally showing slight colour variations and rapid vertical changes in matrix texture and consolidation (Fig. 4.9C,D). Contacts between internal diamict beds comprising the NT are typically sharp, and successions may incorporate stratified interbeds of sand, silt or gravel up to 1 m thick, but more commonly 2-10 cm thick. Stratified interbeds are laterally discontinuous (<15 m wide) and are commonly highly deformed or massive; however, fine planar laminae are locally preserved, particularly within fine-grained silt and clay beds. Clasts generally comprise 5-15% of the volume of the till and display weakly to moderately well-developed fabrics (Fig. 4.9E), or rare stacking patterns with long axes dipping up-ice parallel to drumlin orientations (Fig. 4.9F). Numerous aligned striated clasts are present and indicate ice flow directions parallel to nearby drumlin axes on uplands and within the valleys (Mulligan et al., 2018b). Down-ice dipping fault planes are locally observed within the NT, striking 318° (orthogonal to ice flow direction; Fig. 4.9G). In parts of some valleys, the NT is absent and thick successions of sand and gravel form the base of the valley infill (Fig. 4.6; Mulligan et al., 2018a; Sharpe et al., 2018).

Fig. 4.9: A) finely laminated, thick-bedded (20-40 cm) silt-clay rhythmites comprising the Thorncliffe Formation exposed in the core of a small drumlin along the eastern margin of the Holland Marsh Valley (Fig. 4.4). The rhythmites are deformed by reverse faulting (highlighted by orange oxidation from modern groundwater flow); photo by Stephanie Kimmerle. Shovel handle is 50 cm long; B) highly consolidated, fissile, silty sand NT directly overlying laminated fine to medium sand of the upper Thorncliffe Formation. Photo is approximately 20 cm across; C) multiple diamict beds and lenses occurring in direct contact or separated by thin, deformed sand stringers comprising the NT. Scale card is 9 cm wide; D) over-consolidated NT occurring as stacked massive beds 2-4 m thick separated by thin, eroded sand stringers (arrowed) exposed in a sand pit; E) winnowing of matrix by modern fluvial processes highlights moderately developed fabric (long axes oriented to the SW) along the flank of a large drumlin on the floor of a tunnel valley; F) clast stacking along the margin of a tunnel valley. Clasts plunge 80° E (solid arrow in inset), nearly parallel to drumlin long axes in the vicinity (dashed lines in inset). Knife blade at top is 3 cm wide, ice flow was toward the right; G) parallel fault planes or joints accentuated by modern fluvial erosion strike NW-SE, orthogonal to drumlin long-axes. See Fig. 4.4 for photo locations.


A
B
5 cm
C
D
E
F
G
0
90
180
270

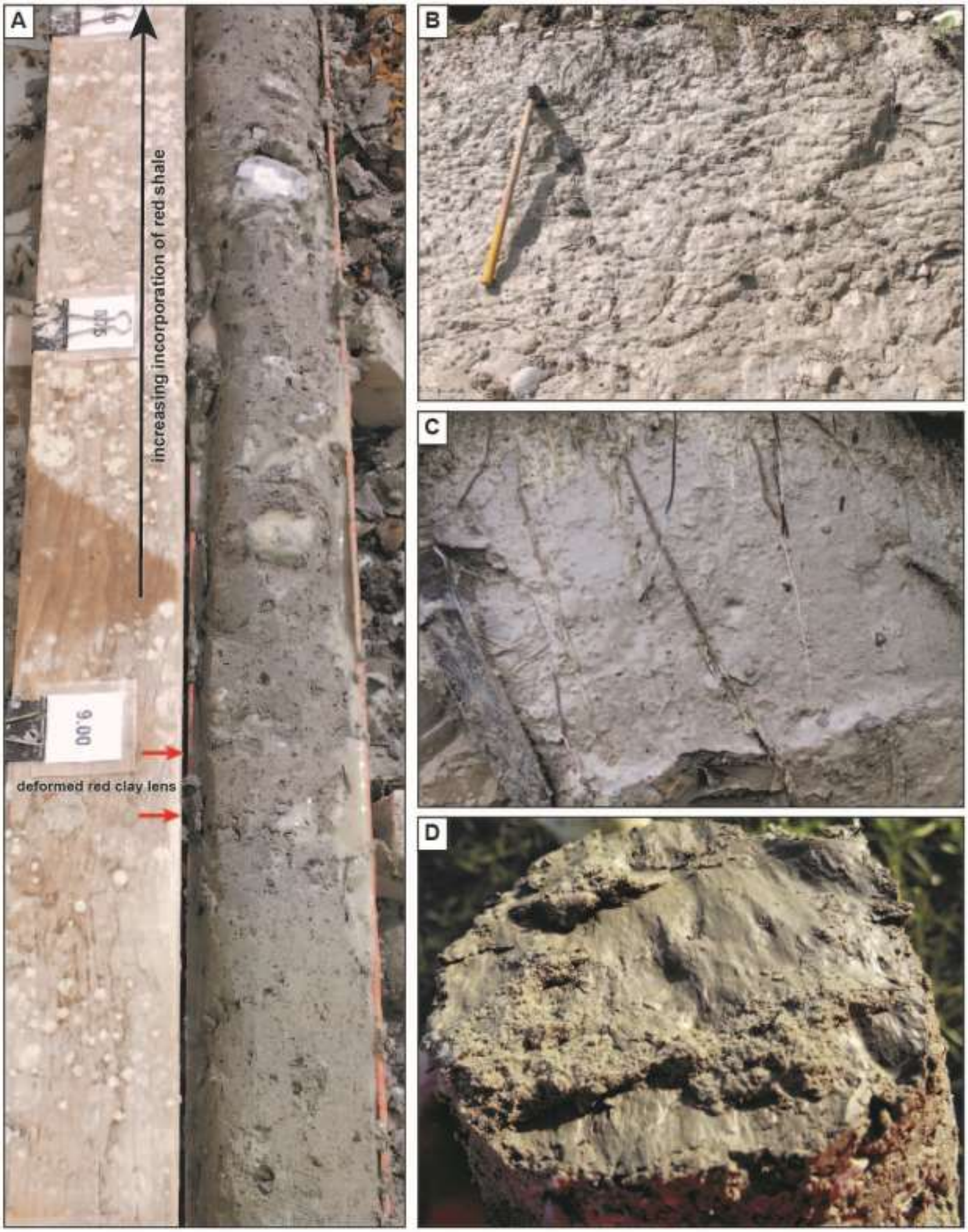
Interpretation

The overall appearance of the NT observed within tunnel valleys bears many similarities to the till on adjacent upland regions, but its greater heterogeneity reflects a more complex depositional history (e.g. Hicock and Dreimanis, 1992) due to the development of the tunnel valleys during the Late Wisconsin (MIS 2; Mulligan et al., 2018a). The presence of diamict beds with well-consolidated, fissile matrix and common faceted and striated clasts suggests lodgement processes controlled till accretion within parts of some of the valleys (e.g. Eyles et al., 1982). However, the common presence of stratified interbeds, weakly-developed clast fabrics, and the poor consolidation of some beds suggests a variety of other depositional processes promoted till development within the tunnel valleys (e.g. Evans and Hiemstra, 2005). Debris flows from collapse of valley walls (Barnett, 1990), melt-out of debris from the base of the ice overlying the valleys, and the significant variations in porewater pressures created by subglacial drainage events likely promoted the differences in diamict characteristics within the tunnel valleys, and individual diamict beds were variably incorporated into a basal traction layer during active ice flow within the tunnel valleys during and following their excavation (Fig. 4.5C; Mulligan et al., 2018a). A higher proportion of sorted beds within the NT records increased meltwater activity and bed separation events within the tunnel valleys, but their restricted thickness and spatial extent (<15 m wide) suggests limited subglacial accommodation space for meltwater deposits to accumulate and/or significant reworking during subsequent periods of stronger ice-bed coupling (Lee et al., 2015).

4.4.2.2. Lowland plains

The NT and an overlying fine-grained diamict facies, previously interpreted as the Allenwood Till (Burwasser, 1974), occur extensively in the lowland plains south and east of Georgian Bay (Figs. 4.2D and 4.4). An interbedded or gradational contact between the NT and overlying fine-

Fig. 4.10: A) gradational vertical transition from silty sand till (grey-brown colour below 9.1 m) to slightly sandy silt till (reddish gray above 8.6 m), formerly the Allenwood Till. Reddish colour derived from incorporation of red, Queenston Formation shale outcropping to the west; B,C) variation in till facies observed in successive outcrops along the Nottawasaga River. Note clast size and abundance in (B) compared to (C). Grub hoe in (B) is 1.1 m long, knife blade in (C) is 3 cm wide; D) slickensided clay seam within sand-rich NT exposed in a possible remnant drumlin indicating sub-horizontal shearing concentrated along the fine-grained seam within the NT. Core is approximately 8.5 cm wide; E-G) selected intervals of a thick glaciotectonite succession grading from deformed silt and clay with rare grit and IRD (E), to deformed silt-rich diamict and interstratified silt (F), into clast-rich and highly consolidated silty sand till (G). Cores are approximately 8.5 cm wide, see Fig. 4.4 for photo locations.

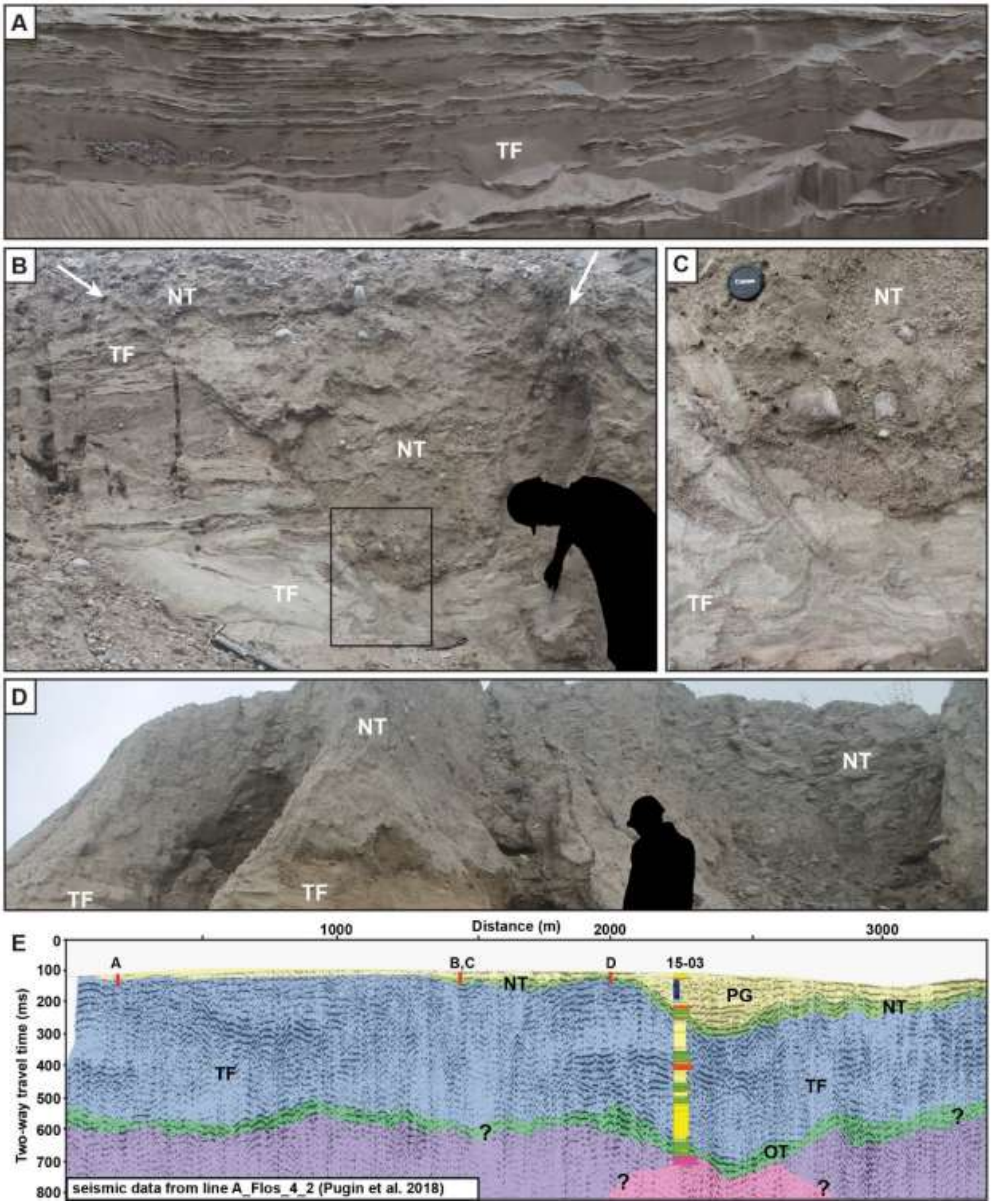

A
increasing incorporation of red shale
9.00
deformed red clay lens
B
C
D


E
22.25 m
22.45 m
F
19.50 m
19.70 m
G
12.30 m
12.50 m

grained diamict facies was observed during surficial mapping (Mulligan, 2014) and later confirmed by sediment drilling (Figs. 6 and 10A); local interbeds of deformed silt and clay with IRD may be present within the transitional zone between the two units, but the two diamict facies more commonly form a spatial and stratigraphic continuum. The lower, coarse-grained facies of the NT is typically enriched in clasts, many of which are of Shield derivation with long axes and striae oriented towards the southwest, parallel to upland paleoflow indicators. Upwards, there is locally correlation between the orientation of striae and clast long-axes orientations, both of which parallel the long axes of MSGL that trend ESE (Fig. 4.5C). The overlying fine-grained diamict facies is clast-poor (<5%) and contains >90% carbonate and shale clasts, including red shales derived from outcrops that lie to the northwest (Figs. 4.2B and 4.10A). Limestone clasts in the upper fine-grained diamict facies are typically ornamented with chaotic striae, or multiple cross-cutting sets of parallel striae and display a weakly-developed fabric indicating paleoflow towards the southeast.

Stratified interbeds are locally observed within both diamict facies and are commonly deformed; slickensided clay seams were observed in coarse-grained diamict facies in borehole 17-09 (Fig. 4.10D). The thickest successions of the NT within the lowland plains occur where it overlies gradational successions of undisturbed rhythmites with trace fossils of the Thorncliffe Formation, passing upwards into deformed silt and clay with ice-rafted debris and diamict lenses (Fig. 10F), and massive, well consolidated silty sand diamict with striated clasts (Fig. 4.10G), similar to NT facies observed on uplands. Rapid lateral changes in the physical character of the NT within the lowland plains are observed in exposures in adjacent aggregate pit operations and a nearby cored borehole (Figs. 4.5B and 4.11). The NT is absent in the westernmost pit; postglacial littoral sands directly overlie Middle Wisconsin (MIS 3) sand and gravel of the

Fig. 4.11: A) stratified sand and local gravelly beds deposited in subaquatic fans comprising the upper Thorncliffe Formation (TF). The NT was either not deposited or eroded by active littoral processes in postglacial lakes at this location. Section is approximately 7 m high; B) stratified TF sediments incised by a steep-walled, asymmetrical channel infilled with interstratified and deformed diamict and gravel, inset box showing location of (C) is approximately 0.5 x 0.75 m; C) sands in the TF within 20-30 cm of the channel are deformed by fluidization, injection structures, and minor faulting (see black box in (B) for location); D) Overconsolidated, blocky, silty sand diamict facies of the NT overlying deformed, fine to medium sand of the TF; E) part of a time-section of a S-wave seismic reflection profile (data from Pugin et al., 2018; Figs. 4.4 and 4.5). Borehole 15-03 was drilled to intersect a channelized feature in the subsurface (from approx. 2100-2700 m on the profile). Approximately 10 m of poorly- to moderately consolidated, interstratified diamict and fine-grained sand comprise the NT in the borehole. See Fig. 4.6 for legend. Locations marked with ‘?’ denote areas where reflector continuity limits confidence in unit correlations.


A
TF
B
NT
TF
NT
TF
C
NT
TF
D
NT
NT
TF
TF
E
Distance (m)
1000
2000
3000
Two-way travel time (ms)
0
100
200
300
400
500
600
700
800
A
B,C
D
15-03
NT
PG
NT
TF
TF
?
OT
?
?
?
seismic data from line A_Flos_4_2 (Pugin et al. 2018)

Thorncliffe Formation (Fig. 4.11A). 1 km to the east, the NT is poorly- to moderately-consolidated, stratified and deformed, silt-rich diamict with interbedded sands where it overlies, and infills small channels (2 m deep and 3.5 m wide) incised into, the Thorncliffe Formation. Further east, the NT consists of massive, highly-consolidated, blocky silty sand diamict with abundant Shield clasts and overlies deformed sands of the Thorncliffe Formation (Fig. 4.11D). A seismic reflection profile, constrained by a continuously-cored borehole less than 0.5 km to the southeast of the latter pit, identified larger-scale channel features floored by 10 m of interstratified poorly- to moderately consolidated diamict and interstratified silty fine sand comprising the NT in the borehole (Figs. 4.5B and 4.11E).

Interpretation

Within the lowland plains, complex facies associations observed within the NT are interpreted here to record the spatial and temporal evolution of depositional processes during multiple phases of Late Wisconsin ice cover. Thick sedimentary successions grading vertically from undisturbed rhythmites into fine-grained diamict and upwards into silty sand NT are interpreted as glaciotectonite successions (Evans et al., 2006; Maclachlan and Eyles, 2011; Fig. 4.6). These successions record the advance of the LIS into an ice-contact glaciolacustrine basin, followed by overriding by ice and progressive deformation and incorporation of sediment into a mobile deforming bed. The uppermost, sandy till facies capping the finer-grained successions record the main phase of regional ice advance from the northeast, contemporaneous with the deposition of the NT on uplands.

Locally, the complex facies associations observed within the NT in parts of the lowland plains bear many similarities to broader, regional trends observed within tunnel valleys within the study area. Meltwater activity preferentially routed through local lows may have contributed to increasing local heterogeneity, and interstratification with sorted beds, likely through repeated meltwater storage and discharge events routed through local channels in what was a broad low on the subglacial bed (Piotrowski et al., 1999; Kehew et al., 2012; Buechi et al., 2017). The small channels observed in pits and on seismic reflection profiles (Fig. 4.11) provide an analogue for the facies variations in the NT within the larger tunnel valleys in south-central Ontario. Intervening highs between channels (and tunnel valleys) experienced less intense fluctuations in meltwater drainage styles, which permitted the deposition of more uniform till facies.

The gradational to interbedded contact between the coarse-grained facies of NT and overlying fine-grained till facies (formerly the Allenwood Till) suggests that they form a depositional continuum, forming a 'hybrid till' (Stea and Brown, 1989; Stea and Finck, 2001) recording a change in ice flow directions, source area, and dominant substrate materials, without an intervening period of ice retreat (e.g. Eyles et al., 1982). Local beds of deformed silt and clay separating the two diamict units suggest either brief, local ice-marginal retreat-readvance cycles, or local ponding of subglacial water occurred (Livingstone et al., 2015).

4.4.3. Niagara escarpment

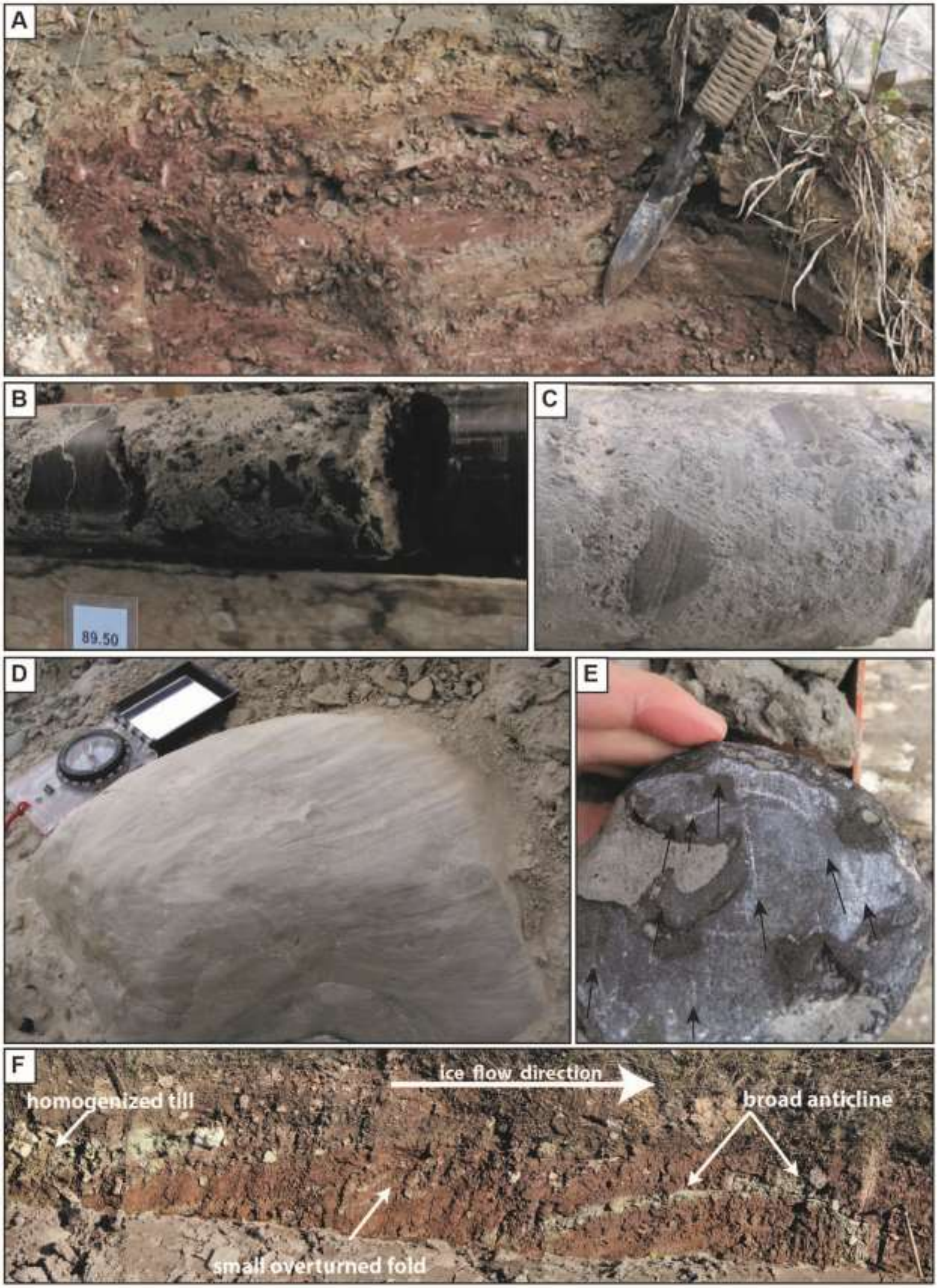
In the study area, the Niagara Escarpment is subdivided into three geomorphic regions comprising its flank, re-entrant valleys, and crest. Drumlinized NT (formerly mapped as the Elma Till) is observed within the Beaver Valley and above the crest of the Niagara Escarpment, where multiple phases of drumlinization are recorded on the Catfish Creek-Elma-Newmarket till plain (Figs. 4.2B,D and 4.5A). High-resolution DEMs show that bedforms on the surface of the NT extend uninterrupted beneath moraine ridges that previously defined mapped till extents (Fig. 4.5A). These data, combined with textural and mineralogical information obtained as part of surficial mapping, suggest the Catfish Creek, Elma and Newmarket Tills form a continuum in this region (e.g. Sharpe, 1990), with the moraines representing minor re-advances, still stands or pauses in the retreat of the ice margin. For simplicity, this composite till unit will be referred to as the NT in this paper, although the reader should note that Catfish Creek Till is the preferred term for the main Late Wisconsin (MIS 2) till occurring above the Niagara Escarpment in southwestern Ontario (Karrow, 1974; Bajc et al., 2018; Burt, 2018).

4.4.3.1. Escarpment Flank

The NT occurs in localized areas along the eastern flank of the Niagara Escarpment. Clast-rich, silty sand diamict (till) is predominant, but significant changes in colour, consolidation, sorting, clast content and lithology, and internal structure are observed at local scales (<1 km). Variations in till character are most pronounced where major changes in underlying bedrock lithology occur. Blue-grey till directly overlies blue-grey Georgian Bay Formation shales, red to reddish brown till occurs in direct contact with the red to terra cotta Queenston shale. Areas immediately downflow of a change in underlying bedrock lithology commonly show colour banded, crudely stratified and deformed till assemblages with interstratified and deformed sorted beds (Fig. 4.12A). The NT rapidly changes colour over short (100-500 m) distances at numerous sites along the escarpment.

Angular or flaggy clasts of limestone or shale derived from underlying bedrock are abundant in the till. In some low-lying areas adjacent to the escarpment, the basal (lower 20-60 cm) part of the NT is composed of monolithic angular clasts derived from underlying bedrock (Fig. 4.12B,C). However, depending on bedrock character, clast size and content can vary drastically; where NT directly overlies soft Queenston Shale, it is almost entirely red and may contain few to no visible clasts (Burwasser, 1974). Matrix grain sizes are also variable; above

Fig. 4.12: A) stratified and strongly colour banded NT immediately downflow from a lithologic boundary between grey shale and sandstone beds and red shales; B,C) monolithic clast-rich and angular breccia facies comprising the basal part of the NT directly above black shale bedrock; D) striated and faceted limestone boulder forming part of a crudely-developed pavement; E) parallel striae (arrowed) ornamenting a limestone cobble near the base of the NT; F) glaciotectonized red shales overlain by thin fine-grained NT on a bedrock outlier along the Niagara Escarpment. Multiple scales of folding are present. Shovel at bottom right is 0.5 m long. Cores are approximately 8.5 cm wide, see Fig. 4.4 for photo locations.


A
B
89.50
C
D
E
F
ice flow direction
homogenized till
broad anticline
small overturned fold

shale bedrock, clayey silt to silt matrix is common but in most areas the till has a grey-brown silty sand matrix.

Along the escarpment flanks the NT typically consists of stacked massive beds of diamict between 5 cm and 1 m thick (Fig. 4.6). Individual diamict beds are typically separated by sharp, undulating and deformed (locally faulted) contacts, or by thin (1-10 cm) beds of stratified silt or sand (Fig. 4.12A). Clast roundness commonly increases rapidly up-section, and colour banding is less pronounced upwards. In some areas, well-developed matrix fissility and clast fabric and faceting are observed parallel to local drumlins oriented NE-SW (Fig. 4.12D,E), locally forming crudely-developed, undulating pavements. At rare sites, deformation of shale strata is observed beneath the NT (Fig. 4.12F).

Interpretation

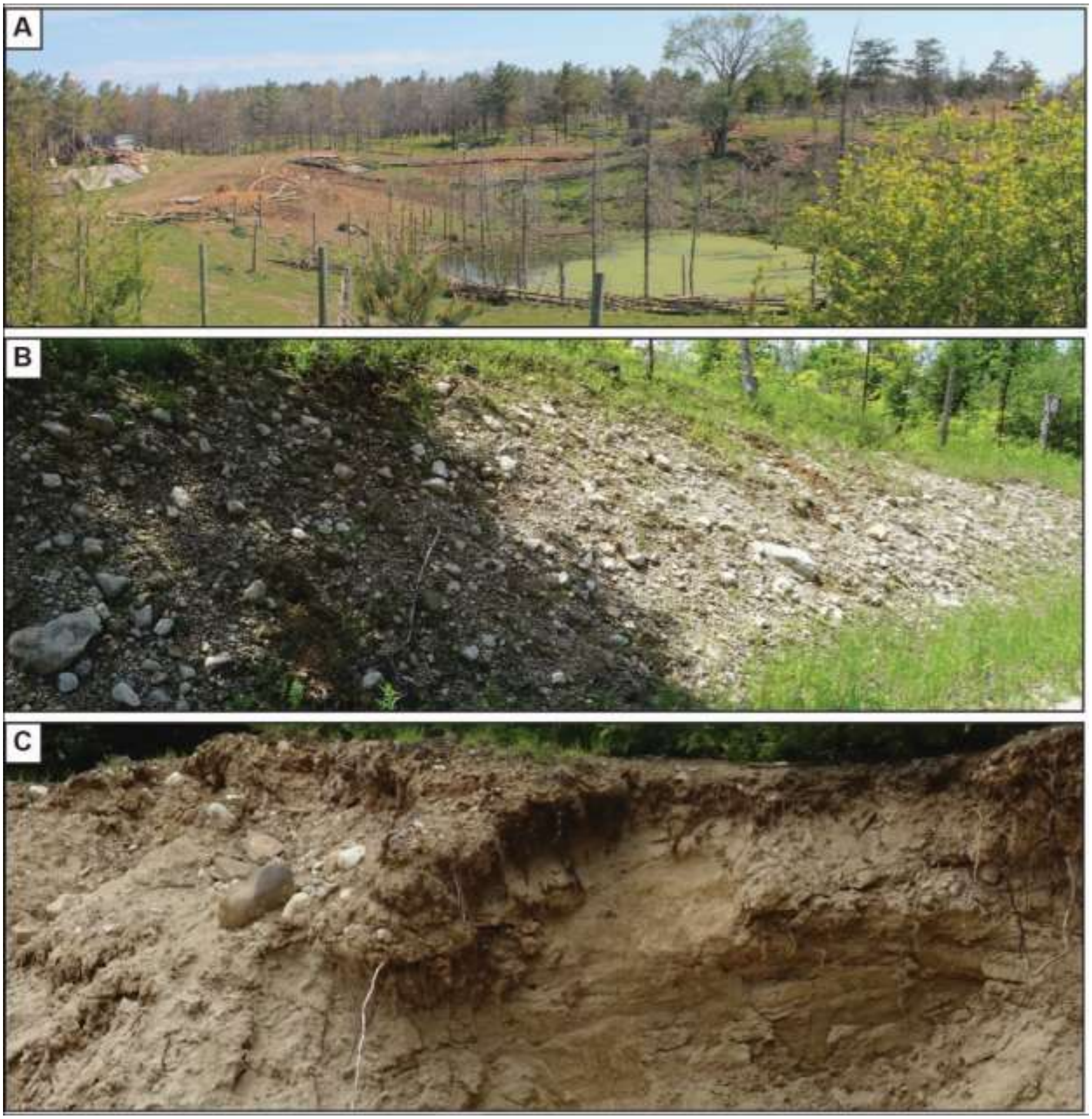
Rapid colour changes observed within the NT, coinciding with lithological changes in underlying Paleozoic bedrock, record incorporation of local substrate materials into the ice base along the flank of the Niagara Escarpment. Locally-derived angular clasts at the base of the till and their progressive increase in roundness upwards indicate rapid comminution of the soft bedrock, suggesting high strain rates and grain crushing within a mobile matrix (Evans et al., 2018). Stratification of matrix colours is interpreted to record migrating or alternating zones of local subglacial erosion, transport, and deposition at the ice base (e.g. Boulton et al., 2001) facilitated by highly compressive flow along the flanks of the escarpment and variations in subglacial meltwater storage and drainage. Poorly-developed clast fabrics, particularly in shale-rich NT facies, as well as vertical successions showing increasing homogenization of materials upwards, indicate these discrete zones do not persist long within the traction zone (Jennings, 2006), and likely indicate the presence of a mobile deforming bed at the base of the ice along the flank of the escarpment. Conversely, in areas where highly consolidated, fissile to massive silty sand diamict is observed overlying shale, lodgement and strong ductile shearing processes are likely to have operated. This suggests a mosaic of subglacial bed conditions (Piotrowski et al., 2001; 2006; Phillips et al., 2018) existed beneath the LIS along the flanks of the Niagara Escarpment.

4.4.3.2. Escarpment crest

The NT above and to the west of the Niagara Escarpment has previously been mapped as the Catfish Creek Till or the Elma Till (Figs. 4.2D and 4.5A), but these tills together with the NT form a depositional continuum in the study area (e.g. Sharpe, 1990). The till plain displays well-developed streamlined bedforms in between moraine ridges (Fig. 4.5A). South of the Singhampton Moraine (Fig. 4.2A) two generations of bedforms occur: large broad southwest-trending drumlins are superposed with low-relief southeast-trending lineations. North of the Singhampton Moraine, where sediment successions are thinner, bedforms fan out from the Beaver Valley (Fig. 4.5A). The NT is commonly grey to buff brown, highly consolidated, with a massive silty sand matrix showing local fissility. East of the Beaver Valley, the NT rapidly changes to a beige a sand-rich diamict with a less consolidated matrix and abundant angular clasts derived from the underlying dolostone. Previous analyses of clast fabric (Burwasser, 1974) are consistent with the orientation of cross-cutting bedforms and striae orientations recorded from bedrock exposures, indicating former ice flow from the north and northeast (Mulligan, 2014; 2015).

Numerous arcuate end and/or recessional moraines (Figs. 4.2A, 4.4, 4.5A) overlie the streamlined till plain, dominated by hummock and kettle topography (Fig. 4.13A). The moraines are variably composed of poorly to moderately-consolidated, sand-rich diamict, with high clast

Fig. 4.13: A) kame and kettle topography within the Gibraltar Moraine (fence posts in foreground are 1.5 m tall); B) cobble-rich sandy diamict comprising a small moraine ridge upflow of the Gibraltar Moraine east of the Beaver Valley. Section is approximately 3.5 m high.; C) sand, gravel and lenses of diamicton comprising a moraine ridge east of the Beaver Valley. Section is approximately 2.5 m high. See Fig. 4.4 for photo locations.


A
B
C

content (>25%; Fig 13B). Large blocks of angular to subangular beige Guelph Formation dolostone are particularly common in moraine ridges and glaciofluvial deposits overlying bedrock near the Beaver Valley (Figs. 4.5A, 4.13B). Other moraine ridges may be composed of stacked, heterogeneous and variably coloured diamicts and/or stacked successions of sand, gravel, or silt with interstratified diamict beds (Fig. 4.13C).

Interpretation

Till deposits observed above the crest of the Niagara Escarpment are typically enriched in dolomitic material and record the plucking and incorporation of bedrock units comprising the cuesta (Karrow, 1974; Burwasser, 1974). Highly consolidated fissile till underlying the streamlined till plain above the crest of the Niagara Escarpment records till emplacement by ductile shearing along vertically migrating shear zones below the ice (Phillips et al., 2018; Rice et al., 2018). Bedrock striations and multiple bedform orientations indicate subglacial debris transport during multiple phases of ice flow during the Late Wisconsin (MIS 2; Fig. 4.5A). The location of the most prominent moraines at the downflow end of a set of streamlined bedforms suggests they result from the downflow transport of material eroded during streamlining (Eyles et al., 2016; Sookhan et al., 2016), facilitated by subglacial meltwater delivery of sediment to marginal areas, which was particularly significant near re-entrant valleys (e.g. Kor and Cowell, 1998). The heterogeneity of sediment facies within the moraine ridges compared to the adjacent streamlined till plain records a transition to sub-marginal, supraglacial and proglacial depositional settings from the subglacial setting in which till was deposited (e.g. Johnson, 1990).

4.4.3.3. Re-entrant valleys

The NT observed within re-entrant valleys cutting into the face of the Niagara Escarpment displays local spatial variability in its characteristics as well as noticeable vertical trends. In the largest re-entrant, the Beaver Valley (Fig. 4.2), thick (>10 m) successions of till are exposed along river and stream cut banks (Fig. 4.14A). The NT is generally grey to grey-brown, massive and highly consolidated, with few thin (<10 cm) interbeds of sand, silt or gravel. The matrix is poorly sorted and varies from silty sand to sandy silt, with the finer matrix textures typically occurring higher in the exposed successions. The NT has a relatively high concentration of large clasts, consisting of a mix of Shield and local carbonate lithologies (approximately 35 and 65%, respectively) in the north, but the proportion of Shield material decreases rapidly southward from the shores of Georgian Bay. Outcrops of the NT commonly expose faceted and striated clasts near to the base of the till indicating ice flow to the SSE (156°). Up-section within the NT, clasts show less consistent long axis orientations and are marked by sets of cross-cutting and chaotic striae overprinted by a faint set of parallel striae oriented SE (144°; Fig. 4.14B).

Within smaller re-entrant valleys the NT displays more complex facies associations. Along the southern flank of the Dunedin re-entrant (Fig. 4.2B), exposures along road cuts reveal highly variable successions of interbedded till and stratified sediment. Near the base of the valley, highly consolidated, fissile, massive to very crudely stratified silty sand diamict is overlain by stratified sand and silt with paleocurrent indicators suggesting flow to the southwest (into the re-entrant). Further up the valley flank to the southwest, poorly sorted silt-rich diamict beds <10 cm thick are interstratified with fine-grained planar laminated sand beds 1-3 cm thick (Fig. 4.14C). Along the upper flanks of the re-entrant, moderately consolidated, massive, sand-rich diamict directly overlies striated bedrock of the escarpment crest.

Where thick pre-Late Wisconsin (MIS 2) sediments locally infill parts of the larger re-entrant valleys, particularly the Beaver Valley (Fig. 2B), the NT displays more consistent facies characteristics. Outcrop exposures along a stream cut bank in a tributary to the western part of

Fig. 4.14: A) 8-10 m of the NT (formerly Elma Till) exposed along an abandoned meander of the Beaver River. Field assistant is 2 m tall; B) small boulder exposed in the outcrop shown in A), showing an older set of gouges and grooves (double-ended arrows) cross-cut by striae (long arrows) parallel to clast long axis. Ice flow was toward the bottom of the photo, note streamlined upper edge of the clast. Scale on compass is in cm; C) interbedded sand-rich diamict and fine- to medium-grained sand lenses; D) stratified, interbedded and deformed grey diamict and beige-brown very fine sand (colour changes enhanced by oxidation from enhanced groundwater flow) near the base of the NT; E) basal contact of the NT (grey) with the underlying faulted (arrows show sense of displacement) and deformed TF sediments (red and blue-grey). Photo is approximately 1.5 m below (D). See Fig. 4.4 for photo locations.


A
postglacial lake sediment
NT
B
C
4 cm
D
E
dug
foothold
5 cm

the Beaver Valley (Clarksburg site; Fig. 4.5A) reveal fossiliferous alluvial, lacustrine and glaciolacustrine sediments deposited after 31,500 ± 1000 ^{14}C yr BP (BGS-182A; Warner et al., 1988) to 42,050 ± 509 ^{14}C yr BP (UOC-3286) (see Appendix). These deposits are increasingly deformed upwards towards the contact with overlying beds of grey to buff brown massive diamict interbedded with fine sand and silt with ice-rafted debris, comprising the lowermost bed of the NT (Fig. 4.14D,E). The exposed diamict succession is approximately 6 m thick consisting of stacked beds with varying matrix textures and interbedded deformed stratified sediments. Clasts generally display chaotic striae and poorly-developed fabrics, but where parallel striae are observed on larger cobbles and boulders, ice flow direction towards the south-southeast (140-170°) is indicated, consistent with the long axes of local drumlins trending south-southeast (Fig. 4.5A).

Interpretation

NT facies within re-entrant valleys appear to be strongly controlled by the lithology and local topography of the bedrock or substrate sediments. In the central parts of the Beaver Valley, where the NT overlies older sediment substrates, the internal facies and properties are highly consistent with NT facies observed on uplands in the study area because the effects of local bedrock topography on glacier dynamics in this region were diminished, allowing glacier ice to maintain contact with the bed across the entire valley floor. Substrate sediments record infilling of the valley during ice approach in the Early-Middle Wisconsin (Mulligan and Bajc, 2018). Deformation structures record a shift from proglacial to subglacial conditions (e.g. McCarroll and Rijsdijk, 2003), or changes in porewater pressures at the IBI (e.g. Phillips et al., 2008). The interbedded diamicts at the base of the NT in the Beaver Valley (Fig. 4.14D) suggest subaqueous debris flow deposition (Benn, 1996), and either scenario suggested above could account for the observed sediment and deformation characteristics. Within smaller re-entrants, the higher proportion of sorted sediment interbeds within the NT likely record deposition in subglacial cavities during repeated periods of de-coupling (Lessemann et al., 2010), facilitated by the significant local variations in bedrock topography (Buechi et al., 2017).

4.4.4. Regional trends

The high topographic variability between regions and the variable depths at which the NT is encountered in lowlands has historically led to difficulties in correlating the till across the study area. Recent geologic (Bajc et al., 2015; Mulligan, 2016; 2017a, b, c; 2018; Mulligan et al., 2018a, b) and geophysical (Todd et al., 2008; Mulligan et al., 2018a) data support the correlation of the NT from the surface of uplands into the subsurface beneath tunnel valleys. Documentation of the consistent stratigraphic position, paleoflow directional indicators and morphological characteristics of the NT permit its identification within the three physiographic regions in the study area.

A southward fining of grain size observed within the NT occurs throughout the area, spanning the uplands, lowland plains and Niagara Escarpment (Fig. 4.15). Particle size analysis demonstrates textural similarities between the NT and buried Illinoian/older tills (OT1, OT2; Figs. 4.6, 4.8 and 4.16; Sharpe et al., 2018). Despite this, there are differences in average silt and clay content (Fig. 4.16) and the three till units can be further distinguished by their lithological composition, degree of consolidation, chronostratigraphic data, and, where outcrops are available, by their paleoflow directional indicators (Mulligan and Bajc, 2018; Figs 4.6 and 4.8).

The NT is, on average, thinnest within the escarpment region and thickest within the uplands region (Fig. 4.17). If areas where the NT has been eroded by post-depositional

Fig. 4.15: Location map of surficial grain size samples collected from the NT in central Simcoe County (black line) overlain on physiographic region map. Matrix composition is displayed in pie charts, and a clear trend of decreasing sand content southward is evident across all three regions, except above the Niagara Escarpment in the southwesternmost corner of the study area, where high sand content is prevalent.

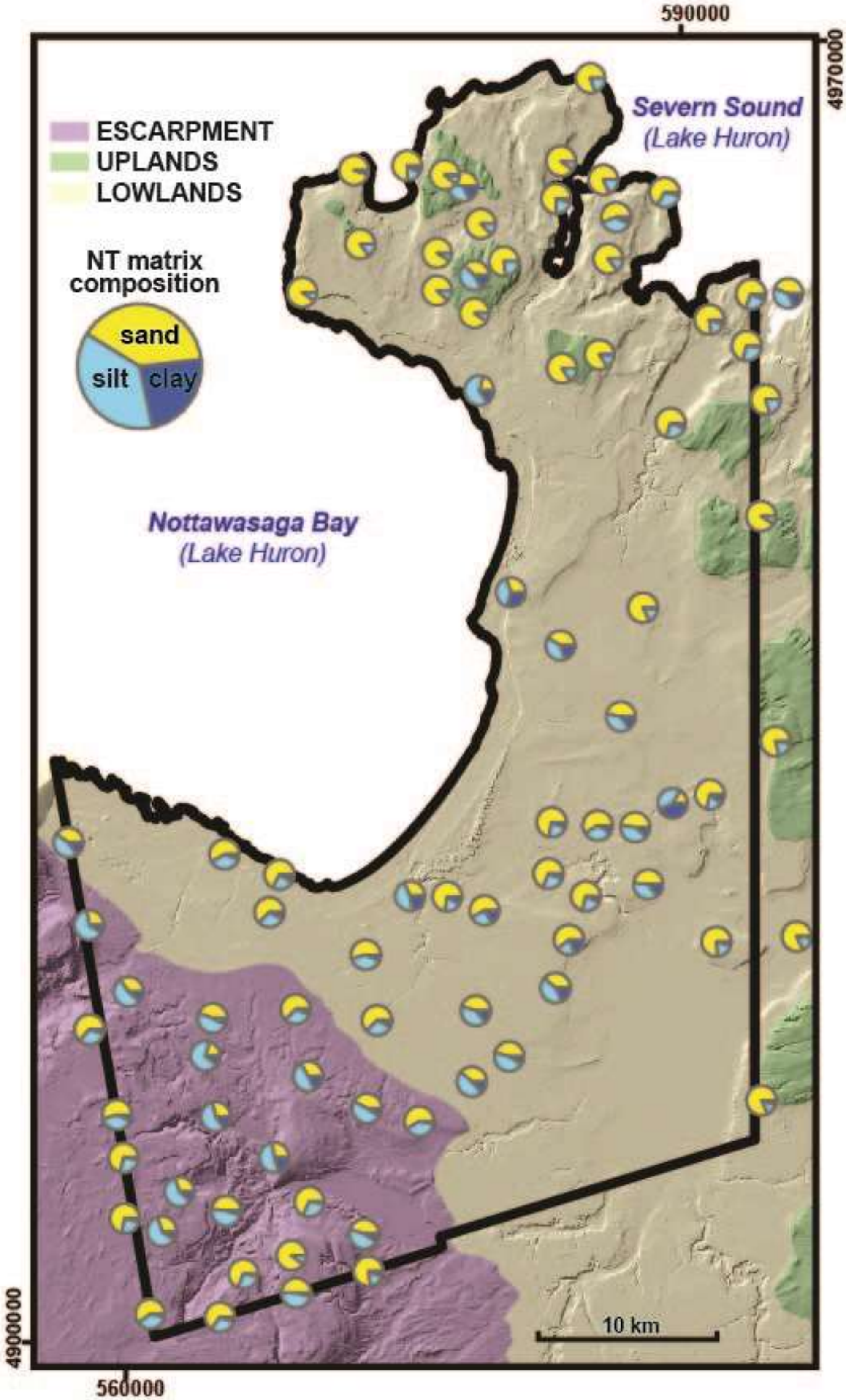

590000
4970000
Severn Sound
(Lake Huron)
ESCARPMENT
UPLANDS
LOWLANDS
NT matrix
composition
sand
silt
clay
Nottawasaga Bay
(Lake Huron)
10 km
4900000
560000

Fig. 4.16: Ternary diagrams showing variations in matrix grain size data from borehole samples collected from southern Simcoe County. The NT and buried Illinoian (MIS 6)/older tills (OT1 and OT2; see Fig. 4.6) share many textural characteristics, illustrating the need for numerous additional analytical parameters and stratigraphic information to differentiate between the tills.

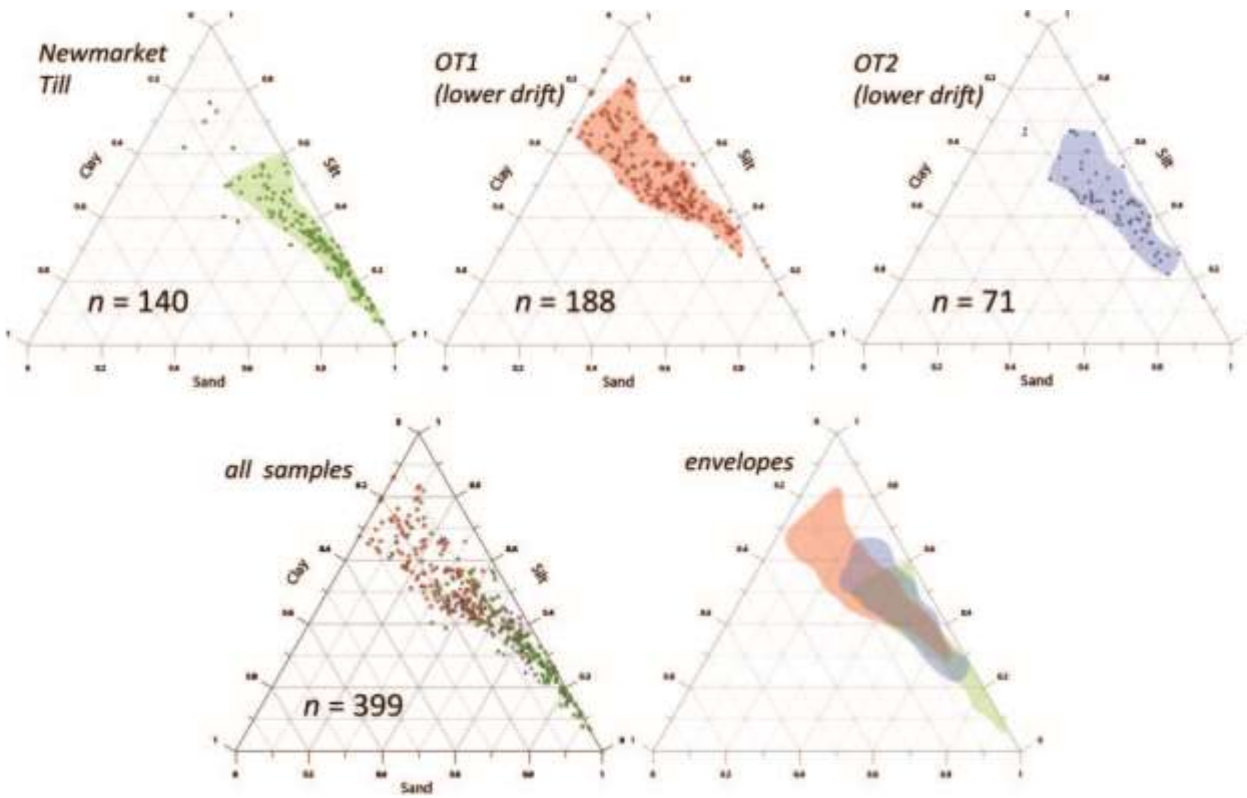
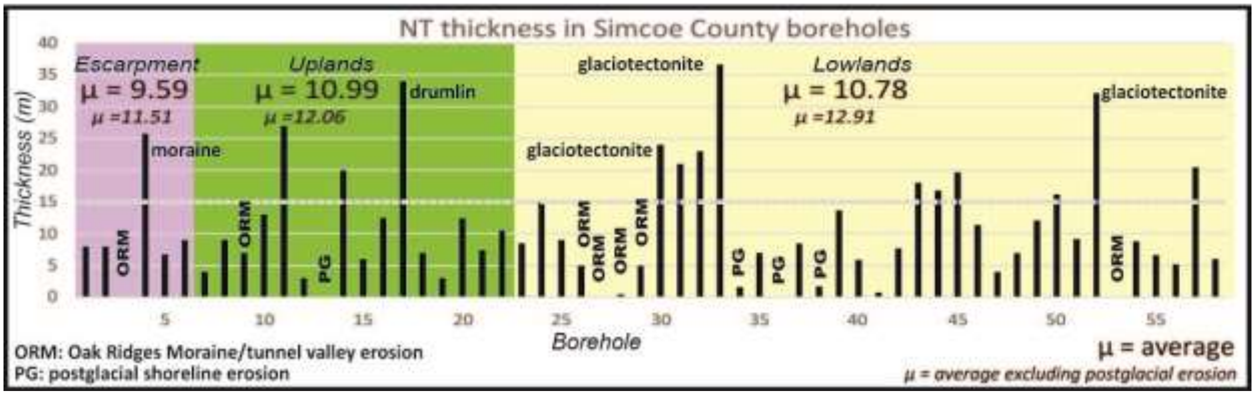

Newmarket Till
n = 140
OT1 (lower drift)
n = 188
OT2 (lower drift)
n = 71
all samples
n = 399
envelopes
Clay
Silt
Sand

Fig. 4.17: Thickness of the NT in all Simcoe County boreholes, grouped by physiographic region. Average thicknesses for all borehole intersections, and adjusted averages removing effects of post-depositional erosion within tunnel valleys/ ORM formation or along postglacial shorelines are displayed. The NT is generally <12 m thick, except in glaciotectonite complexes, drumlins, and ice-marginal complexes. See Fig. 4.6 for locations and stratigraphy of boreholes.


NT thickness in Simcoe County boreholes
Escarpment
μ = 9.59
μ =11.51
Uplands
μ = 10.99
μ =12.06
Lowlands
μ = 10.78
μ =12.91
moraine
drumlin
glaciotectonite
glaciotectonite
glaciotectonite
ORM
PG
Thickness (m)
Borehole
ORM: Oak Ridges Moraine/tunnel valley erosion
PG: postglacial shoreline erosion
μ = average
μ = average excluding postglacial erosion

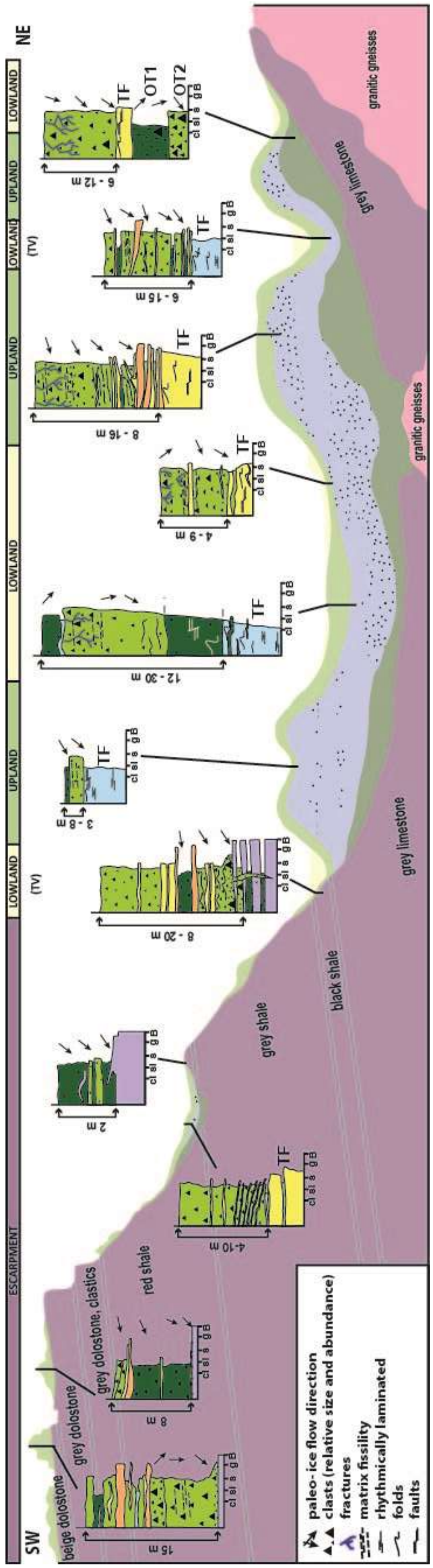
glaciofluvial, littoral, or groundwater piping processes are ignored, average thickness of the NT is greatest within the lowlands region (Fig. 4.17).

4.5. DISCUSSION

Identification of diamict beneath tunnel valley infills as the NT is achieved by walking out the contact of the NT from the flanks of uplands into the shallow subsurface in valleys, along road and ditch cuts, excavations for housing developments, and stream cut banks, and through correlation of marker beds identified in geophysical surveys. Regional stratigraphic drilling shows the presence of two older tills beneath the NT (Fig. 4.6) which are considered to be of Illinoian (MIS 6) and/or older age (Mulligan and Bajc, 2018); radiocarbon age determinations from organic material above and below the NT provides rough chronologic constraint for deposition of the till to the Late Wisconsin (MIS 2; Fig. 4.6) and only the NT has been encountered at this stratigraphic position in south-central Ontario. Land-based multi-component seismic reflection surveys (Fig. 4.11E; Pugin et al., 2018) and water-borne sub-bottom profiles (Todd et al., 2008) reveal undulating (drumlinized?) high-amplitude reflectors that rise towards the flanks of the valleys, where the NT is mapped at surface (Mulligan et al., 2018a; Mulligan, 2019).

These complementary data sets and detailed analyses of the internal composition of the diamict units permit re-evaluation of the genesis of the NT. Whereas previous workers invoked the existence of multiple till sheets deposited during significant ice-marginal fluctuations upon deglaciation of the region (Gwyn and White, 1973; Burwasser unpublished report; Barnett, 1992), or near-complete erosion of the NT from within lowland regions previously interpreted as tunnel channels (Sharpe et al., 2002; 2004; 2018), the NT can now be confidently correlated as a

Fig. 4.18: Summary diagram illustrating the stratigraphic architecture and facies variability within the NT across the physiographic regions in the study area. No scale implied but typical thickness ranges are provided at the left of each log (See Fig. 17). Arrows denoting paleo-ice flow directions are interpreted from clast provenance, till sheet morphology, clast faceting/striae orientations and/or substrate deformation features. TV=tunnel valley; See Fig. 4.6 for legend. TF = Thorncliffe Formation; OT1, OT2 = Older (Illinoian; MIS 6) tills.


SW
NE
ESCARPMENT
LOWLAND
UPLAND
(TV)
beige dolostone
grey dolostone
grey dolostone, clastics
red shale
grey shale
black shale
grey limestone
granitic gneisses
TF
OT1
OT2
15 m
8 m
4-10 m
2 m
8 - 20 m
3 - 8 m
12 - 30 m
4 - 9 m
8 - 16 m
6 - 15 m
6 - 12 m
cl si s g B
paleo- ice flow direction
clasts (relative size and abundance)
fractures
matrix fissility
rhythmically laminated
folds
faults

spatially heterogeneous unit capping sediment-cored uplands, flooring lowland plains and tunnel valleys, and blanketing the Niagara Escarpment in the study area (Figs. 4.6 and 4.18).

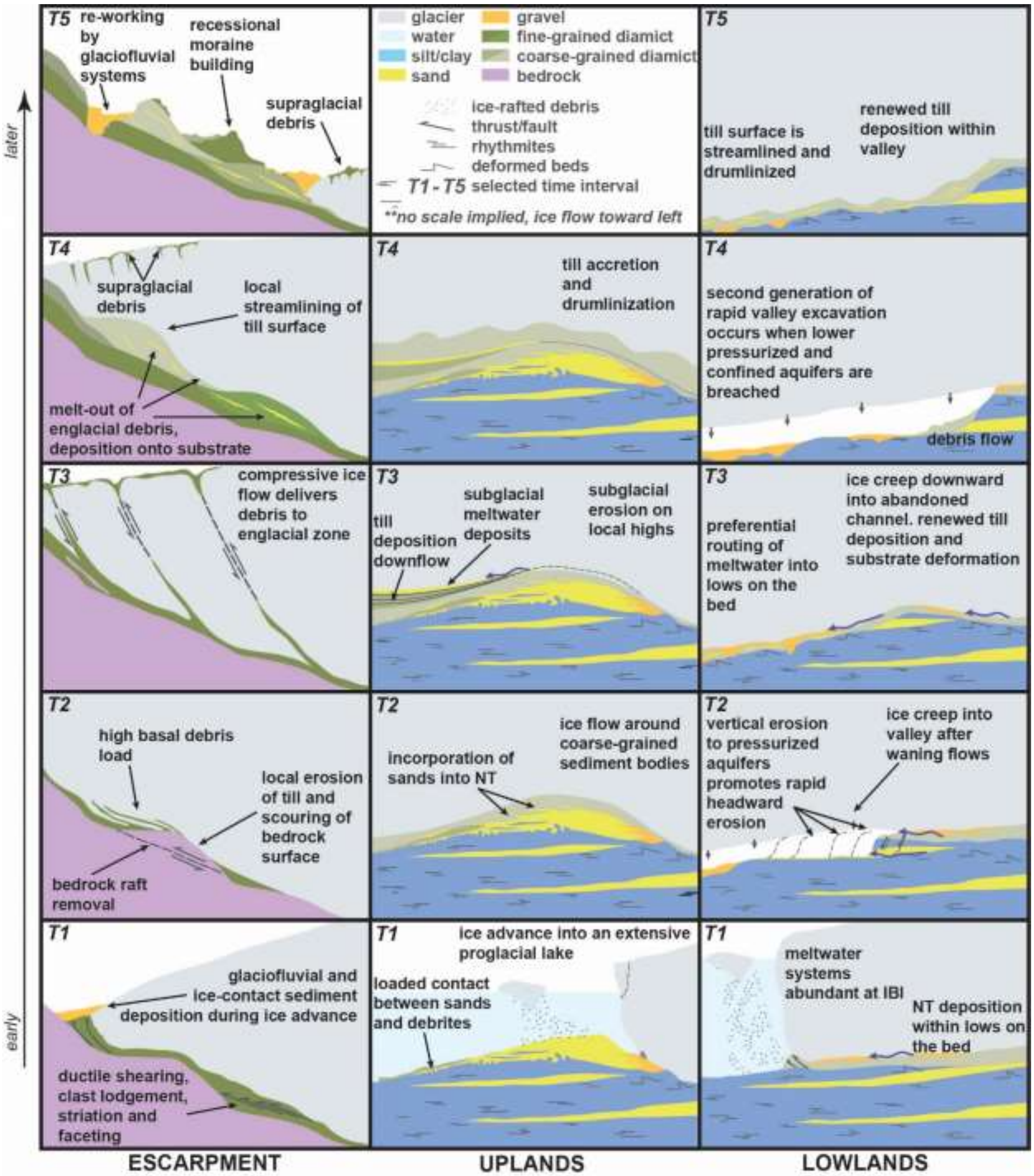
This understanding of internal heterogeneity and architectural complexity provides a new lens through which to analyze the NT characteristics and evaluate past conceptual models for regional till sheet deposition. It also underscores the importance of using multiple lines of evidence and multiple stratigraphical methods (see review in Lee, 2017) in efforts to correlate a till sheet0 across large and structurally complex geographic regions. Rather than a simple deformation or lodgement till, the NT is a complex hybrid deposit. It consists of multiple internal heterogeneous components that, when viewed as a whole, provide a record of a complex strain history resulting from a variety of subglacial erosional and depositional conditions governed by interactions with local bedrock and sediment substrates, spatial and temporal fluctuations in meltwater discharges and subglacial drainage styles, and changing ice flow velocity and direction (e.g. Eyles et al., 1982; Norris et al., 2018; Fig. 4.19).

4.5.1. Palaeoglaciological reconstructions

4.5.1.1. Uplands

Within the uplands region of the study area, the NT records progressive mixing of coarse-grained debris derived from north-facing escarpments along the Shield margin, and glaciofluvial sediment deposited along the advancing ice margin, with extensive fine-grained proglacial lake deposits that fronted the advancing LIS (Fig. 4.19; Eyles et al., 2018). The local nature of proglacial ice-contact deposition (largely as subaquatic fans and debris flows) produced a complex mosaic of substrate sediment types at the IBI, which affected the nature of till production (e.g. van den Berg and Beets, 1987). Debris flows predating ice overriding were loaded onto the uppermost sand packages within the Thorncliffe Formation (Fig. 4.19) and

Fig. 4.19. Summary diagrams depicting snapshots (T1-T5) of the evolution of the NT in different subglacial settings/physiographic regions. Bedrock topography and lithology create strong local contrasts in till properties and affect subglacial and proglacial drainage, increasing heterogeneity in the NT. Debris flows in glacial lakes fronting the advancing LIS shield parts of some upland areas from deformation during overriding. Strong ice-bed coupling and abundant sediment supply permit the development of uniform NT facies across large parts of uplands. Lowlands display a higher degree of NT facies variability and interbedding with stratified sediments due to the increased influence of subglacial meltwater and repeated periods of till erosion, deposition, and deformation throughout the Late Wisconsin (MIS 2). No scale is implied and snapshots T1-T5 did not necessarily occur sequentially or synchronously throughout each region.


T5 re-working by glaciofluvial systems
recessional moraine building
supraglacial debris
glacier
water
silt/clay
sand
gravel
fine-grained diamict
coarse-grained diamict
bedrock
ice-rafted debris
thrust/fault
rhythmites
deformed beds
T1 - T5 selected time interval
**no scale implied, ice flow toward left
T5
till surface is streamlined and drumlinized
renewed till deposition within valley
T4
supraglacial debris
local streamlining of till surface
melt-out of englacial debris, deposition onto substrate
T4
till accretion and drumlinization
T4
second generation of rapid valley excavation occurs when lower pressurized and confined aquifers are breached
debris flow
T3
compressive ice flow delivers debris to englacial zone
T3
till deposition downflow
subglacial meltwater deposits
subglacial erosion on local highs
T3
preferential routing of meltwater into lows on the bed
ice creep downward into abandoned channel. renewed till deposition and substrate deformation
T2
high basal debris load
local erosion of till and scouring of bedrock surface
bedrock raft removal
T2
incorporation of sands into NT
ice flow around coarse-grained sediment bodies
T2
vertical erosion to pressurized aquifers promotes rapid headward erosion
ice creep into valley after waning flows
T1
glaciofluvial and ice-contact sediment deposition during ice advance
ductile shearing, clast lodgement, striation and faceting
T1
ice advance into an extensive proglacial lake
loaded contact between sands and debrites
T1
meltwater systems abundant at IBI
NT deposition within lows on the bed
later
early
ESCARPMENT
UPLANDS
LOWLANDS

provided protection of the sands from subsequent subglacial deformation (e.g. Kruger and Kjaer, 1999). Progressively more massive diamict facies upwards in NT successions are interpreted to record homogenization of the actively deforming subglacial material in the traction zone (Evenson et al., 1977; Fig. 4.19). Stratified interbeds record periods of ice-bed separation during excessive meltwater production (Boyce and Eyles, 2000; Fig. 4.19). Their restricted lateral continuity is consistent with deposition in migrating shallow conduits operating at the ice bed during accretion of the NT. The decreasing abundance of stratified interbeds upwards and increasing matrix fissility suggests a transition from deforming bed to lodgement processes (Phillips et al., 2018). This transition could record a variety of factors: decreased strain rates, reducing till dilatancy and strengthening of the bed (Smalley and Unwin, 1968); decreasing porewater pressures, due to major drainage events associated with regional tunnel valley evolution contemporaneous with deposition of the NT (Mulligan et al., 2018a); development of more tightly-spaced channels/canals, leading to dewatering and overconsolidation of the till (Alley, 1992); or increased effective pressure resulting from thickening of overriding ice (Boulton and Dobbie, 1993). It is worth noting that the NT was accumulating while significant tunnel valleys were developing in the region; this suggests that changing subglacial drainage patterns and porewater/meltwater fluctuations were a primary driving factor in governing the local depositional style of the NT in upland regions.

The accumulation of fissile massive NT directly over sand, silt, or gravel with minimal to no deformation is also consistent with a depositional model involving deforming beds or a bed mosaic; the former suggests sliding over a saturated and overpressured water film/sediment layer (Piotrowski et al., 2006; Denis et al., 2010), while the latter indicates preferential zonation of deformation along a decollement at the base of the NT, with minimal vertical stress transfer into underlying sediments (e.g. Evans et al., 2006).

4.5.1.2. Lowlands

The occurrence of repeated meltwater drainage events, sediment evacuation, and subglacial deformation within lowlands produced more complex till successions compared to those found on adjacent uplands (Figs. 4.18 and 4.19). Small channels infilled with the NT overlying thick sandy successions observed in gravel pits in the lowland plains (Fig. 11B,C) provide an analog for the development of the larger tunnel valleys in the region: meltwater surpluses at the ice base are routed to low-pressure channels, which erode into the soft sediment substrate and are infilled with glacier ice and/or diamict following return to Darcian groundwater flow once the surpluses are evacuated (Piotrowski, 1994; Piotrowski et al., 1999; Kehew et al., 2012; see Mulligan et al., 2018a for discussion). This creates a positive feedback wherein incipient lows on the bed receive increasing amounts of meltwater, enhancing the heterogeneity of the accumulating sediment while intervening areas become increasingly well-drained and more uniform facies are deposited.

Within the broad lowland plains southeast of Georgian Bay, variable NT facies characteristics can be related to till production during multiple phases of ice flow, and transient subglacial drainage mechanisms with abundant meltwater surpluses (Fig. 4.19). It is likely that the broad lowland plains facilitated late-stage re-organization of the southwestward flowing LIS in the study area and promoted a late-stage flow event toward the southeast, out of Georgian Bay. This shift in ice flow locally deposited a fine-grained till facies, (formerly named the Allenwood Till), and re-oriented clasts within the upper part of the NT. These internal, morphological and stratigraphic relationships bear striking similarities to the transition from NT to Halton Till in the Lake Ontario basin to the south (Sharpe and Russell, 2016; Eyles et al., 2018; Sookhan et al., 2018; Mulligan et al., 2018c).

4.5.1.3. Niagara Escarpment

Along the flanks of the Niagara Escarpment, complex till assemblages record multiple phases of substrate erosion and deposition. Pre-Late Wisconsin subaerial erosion of the Niagara Escarpment crest produced abundant colluvial deposits (including large blocks of dolostone cap rock) along the flank of the escarpment (Barlow, 2002) and ice flowing up-gradient from the northeast would pond meltwater and accumulate sediments along the flank of the escarpment (e.g. Slomka and Utting, 2018). Compressive ice flow against the escarpment brought basal sediment into englacial (Tylmann et al., 2013) and supraglacial environments (Fig. 4.19) and promoted the development of complex till assemblages with multiple bands of distinct matrix colours (Fig. 4.19). Erodible shale bedrock promoted rapid comminution of mobilized material (rock flour and/or rafts) and allowed an abundant sediment supply to contribute to till deposition during periods of ice coupling (Krabbendam and Glasser, 2011). Areas of glaciotectonized Queenston shale strata indicate localized deep (1-2 m?) stress transfer into the bed (Hiemstra et al., 2007), likely during periods of reduced porewater pressures.

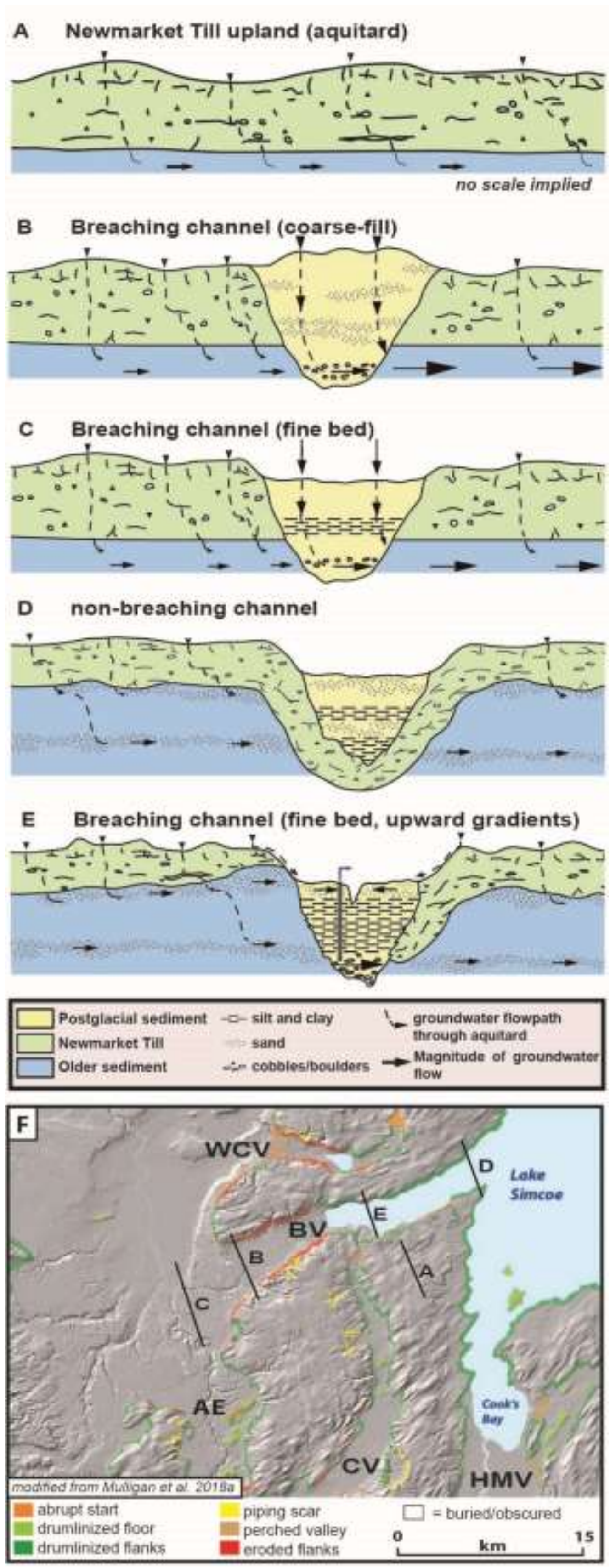
Within re-entrant valleys, the nature of NT deposition was strongly influenced by the size and orientation of the re-entrant, and its location and orientation with respect to meltwater features in the area, such as the large tunnel valleys that lay to the east (Mulligan et al 2018a). Large re-entrant valleys accumulated thick successions of predominantly fine-grained proglacial sediments that pre-date Late Wisconsin (MIS 2) ice advance and deposition of the NT (Pinch, 1977; Warner et al., 1988; Bajc et al., 2015). These sediments reduced the overall bed topography and promoted ice grounding, particularly in the Beaver Valley, where abundant drumlins and flutings are observed. NT facies in these regions indicate depositional processes dominated by high strain rates. Within smaller re-entrants, more pronounced variations in local topography and subglacial meltwater routing resulted in highly heterogeneous till successions due to the repeated opening and closure of subglacial cavities (e.g. Buechi et al., 2017).

Above the crest of the escarpment, rapid lateral changes in till composition and facies record drastic changes in subglacial conditions. Two critical geographic transitions influence these changes. The first is the transition from fine-grained substrates consisting of shales and fine-grained unconsolidated sediments below the escarpment, to coarse-grained, and karstic dolostone strata above the escarpment (Fig. 4.2B; Brunton and Dodge, 2008). The second is the transition from confined to unconfined drainage setting; below the escarpment, meltwater was ponded and the high topography of the escarpment created a strong gradient that promoted development of high hydrostatic pressures, while above the escarpment, free drainage toward the southwest was possible. Combined, these two factors promote a well-drained and strongly coupled IBI above the crest of the Niagara Escarpment where multiple bedform orientations indicate both regional-scale and topographically-constrained ice flow directions.

4.5.2. Implications for hydrogeology

The NT forms a regional stratigraphic marker bed and comprises a significant ‘leaky aquitard’ for regional groundwater systems (Sharpe et al., 2002). Local and field-scale evaluations of heterogeneity and groundwater flow through the NT have been undertaken to the south of the present study area (Gerber and Howard, 2000; Gerber et al., 2001; Desbarats et al., 2001). The variability of internal till facies observed within the NT documented herein support these studies and demonstrate a need for further testing to be undertaken at local and regional scales in order to support groundwater resource management in south-central Ontario.

Fig. 4.20. Hydraulic function of the NT and tunnel valleys. A. Limited recharge occurs through fractures and coarse-grained interbeds within the NT. B,C. Tunnel channel/valley erosion through the NT creates breaches through the NT. Depending on the permeability of infill sediments, these areas may form significant recharge pathways (A-C redrawn from Sharpe et al., 2002). D. Continuous NT forming the floor of some tunnel valleys limits vertical recharge. E. erosion through the NT and deposition of coarse-grained infill sediments creates hydraulic connection between valley fill and older sediments, but high topographic variability along the flanks of tunnel valleys and thick (>40 m) successions of fine-grained glaciolacustrine sediments promote upward gradients and locally strong artesian conditions within the valleys, severely restricting recharge through tunnel valleys. F. local distributions and heterogeneous composition of the NT and valley infill sediments creates rapid changes in hydraulic function within even a single tunnel valley (modified from Mulligan et al., 2018; see Chapters 2 and 3 for discussion).


A Newmarket Till upland (aquitard)
no scale implied
B Breaching channel (coarse-fill)
C Breaching channel (fine bed)
D non-breaching channel
E Breaching channel (fine bed, upward gradients)
Postglacial sediment
Newmarket Till
Older sediment
silt and clay
sand
cobbles/boulders
groundwater flowpath through aquitard
Magnitude of groundwater flow
F
WCV
D
Lake Simcoe
E
BV
B
A
C
AE
Cook's Bay
CV
HMV
modified from Mulligan et al. 2018a
abrupt start
drumlinized floor
drumlinized flanks
piping scar
perched valley
eroded flanks
= buried/obscured
0
km
15

Regional matrix textural variation may have a profound impact on the hydraulic properties of NT. Northward towards the Shield, silt and clay are increasingly depleted (Fig. 4.15). This has the effect of both increasing the bulk conductivity of the unit by reducing the amount of low-transmissivity fines, as well as reducing consolidation of the till, as the matrix particles cannot be packed together as tightly as within more silt-enriched facies.

Past regional-scale studies of the hydrogeology of the NT involved geostatistical mapping of leakance; the ratio of a unit's vertical hydraulic conductivity to its thickness (Desbarats et al., 2001; Sharpe et al., 2002). These studies relied upon contemporary changes to regional geologic models, from simple layer-cake stratigraphic successions, to ones that emphasized the presence and role of regional valleys interpreted as tunnel channels, where little to no NT was believed to occur (Sharpe et al., 2002; lowland region described herein). This early work helped to demonstrate the importance of breaches in the NT aquitard in creating hydraulic windows that enhance recharge to confined aquifers (Fig. 20A-C).

In light of new data presented here documenting the regional variability of NT facies and its stratigraphic architecture, conceptual models of recharge in the vicinity of tunnel valleys in south-central Ontario can be enhanced (Fig. 4.20D,E). The presence of thick and consolidated successions of the NT flooring the valleys creates a significant hydraulic barrier across large parts of some valley floors (Mulligan et al., 2018b). Coarse-grained valley-fill sediments have more localized distributions and limited thicknesses within the valley network than previously thought, as thick successions of fine-grained glaciolacustrine sediments comprise the bulk of the tunnel valley infills (Mulligan et al., 2018a; Fig. 4.6). Together, these data suggest that within the study area, tunnel valleys do not always form significant hydraulic windows that enhance recharge to buried aquifers. Based on the common occurrence of artesian conditions within sediments beneath tunnel valleys (Post and McPhie, 2018), these aquifers are likely sustained by recharge and infiltration in high ground along the adjacent Niagara Escarpment and Simcoe uplands. Significant variations in hydraulic function and preferential groundwater flow paths may exist, even within a single tunnel valley (Fig. 4.20F), indicating that a thorough understanding of the local-scale geological and hydrogeological conditions are essential for planning groundwater resource evaluation and protection.

4.6. CONCLUSIONS

The NT forms a nearly continuous unit that records the Late Wisconsin (MIS 2) ice cover across much of southern Ontario. Improved understanding of the vertical and spatial heterogeneity within this widespread unit is achieved through detailed analysis of the internal composition, facies, and physical characteristics of the till combined with integration of morphological, (chrono-)stratigraphical and geophysical data. Together, these data demonstrate that multiple discrete elements and sediment facies comprise the NT across diverse physiographic regions in south-central Ontario.

Across broad upland areas, strong ice coupling resulting from effective subglacial drainage and consistent substrate materials generates relatively uniform till characteristics including its matrix texture and high degree of consolidation, massive facies, as well as consistent striae, clast long-axis orientations and local bedform orientations. Heterogeneous till facies observed within tunnel valleys in lowland areas result from alternating episodes of glaciofluvial erosion and deposition during subglacial meltwater drainage events, and till deposition by subglacial deformation when ice occupies the tunnel valley floors. Complexities in the characteristics of till facies identified along the Niagara Escarpment developed as a result of alternating periods of bedrock and substrate erosion and till deposition. Compressive ice flow against the high (>300 m) relief of the bedrock escarpment produced significant amounts of englacial and supraglacial debris, which resulted in the deposition of texturally and lithologically distinct beds within the till.

The data presented here provide a glimpse of the variability in the internal composition and stratigraphic architecture of a significant regional marker bed. Through a multi-faceted analysis of the sediments and regional landforms, the multiple elements that comprise the NT have been identified. The diversity of these elements reflects significant spatial and temporal changes in depositional and erosional processes operating at the ice bed during a single glacial cycle and highlight the importance of understanding the range of possible properties of a single 'marker bed'. The types of variations described here for the NT provide a useful key for understanding glaciated landscapes along the peripheries of former ice sheets, where similar regional-scale changes in substrate lithology and topography are observed.

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CHAPTER 5: SUBGLACIAL AND ICE-MARGINAL LANDFORMS IN SOUTH-CENTRAL ONTARIO: IMPLICATIONS FOR ICE SHEET RECONFIGURATION DURING DEGLACIATION

*this is the accepted version of a paper with the same title published in *Boreas 48(3)*, 635-657.

Abstract

Regional-scale, high-resolution terrain data permit the study of landforms across south-central Ontario, where the bed of the former Laurentide Ice Sheet is well-exposed and passes downflow from irregular topography on Precambrian Shield highlands to flat-lying Palaeozoic carbonate bedrock, and thick (50 - >200 m) unconsolidated sediment substrates. Rock drumlins and megagrooves are eroded into bedrock and mega-scale glacial lineations (MSGL) occur on patchy streamlined till residuals in the Algonquin Highlands. Downflow, MSGL pass into juxtaposed rock and drift drumlins on Paleozoic bedrock and predominantly till-cored drumlins in areas of thick drift. The Lake Simcoe Moraines, now traceable for more than 80 km across the Peterborough drumlin field (PDF) form a distinct morphological boundary: downflow of the moraine system, drumlins are larger, broader and show no indication of subsequent reworking by the ice, whereas upflow of the moraines, a higher degree of complexity in bedform pattern and morphology is distinguished. Discrete radial and/or cross-cutting flow sets terminate at subtle till-cored moraine ridges downflow of local topographic lows, indicating multiple phases of late-stage ice flow with strong local topographic steering. More regional-scale flow switching is evident as NW oriented bedforms modify drumlins south of the Oak Ridges Moraine, and radial flow sets emanate from areas within the St. Lawrence and Ottawa River valleys. Most of the drumlins in the PDF formed during an early, regional drumlinization phase of NE-SW flow that followed the deposition of a thick regional till sheet. These were subsequently modified by local-scale, topographically-controlled flows that terminate at till-cored moraines, providing evidence that the superimposed bedforms record dynamic ice (re)advances throughout deglaciation of south-central Ontario. The patterns and relationships of glacial landform distribution and characteristics in south-central Ontario hold significance for many modern and palaeo-ice sheets, where similar downflow changes in bed topography and substrate lithology are observed.

5.1. INTRODUCTION

The Peterborough drumlin field (PDF; Fig. 5.1) is one of many well-studied clusters of subglacial streamlined bedforms. However, complex bedform patterns and landform relationships within the PDF have hindered the development of a full reconstruction of events that shaped the landscape. Early theories invoked a combination of subglacial deformation and erosion by the Laurentide Ice Sheet (LIS; Gravenor 1957; Crozier 1975; Boyce & Eyles 1991). This 'glacial' theory was later challenged by workers who linked broad tracts of erosional terrain to drumlin and tunnel channel formation in the region, and related them to catastrophic releases of subglacial meltwater (Shaw & Gilbert 1990; Brennand & Shaw 1994; Sharpe *et al.* 2004; 2018; Sharpe & Russell 2016). More recent studies have highlighted the importance of ice streams as geomorphic agents within southern Ontario and their ability to create extensive tracts of erosive and streamlined terrain (Ross *et al.* 2006; Eyles 2012; Margold *et al.* 2015; Eyles & Doughty 2016; Sookhan *et al.* 2018a, b; Eyles *et al.* 2018).

Ice streams are discrete regions of fast-flowing ice compared to adjacent areas of the ice sheet that account for as much as 90% of ice and sediment discharge along the margins of modern ice sheets (Bamber *et al.* 2000; Bennett 2003). The existence of former ice streams in southern Ontario was proposed over a century ago (Taylor 1913), but their importance as agents of landscape development has only recently been recognized by comparing the observed landform assemblages in the region to recently deglaciated landscapes and modern subglacial environments (Wellner *et al.* 2006; Smith & Murray 2009; Eyles & Doughty 2016; Sookhan *et al.* 2018a, b). Conceptual models involving the onset and evolution of palaeo-ice streams are proving to be increasingly effective in accounting for observed landform distributions, landsystems tracts, and glacial erosional systems in formerly glaciated parts of Canada and the United States (Dyke & Prest 1987; Veillette *et al.* 2017; Eyles *et al.* 2018; Grimley *et al.* 2018;

Sookhan *et al.* 2018b). Because the onset and evolution of ice streams are controlled by a variety of factors (including subglacial topography, geology, hydrology, substrate interactions, and internal glaciological controls; Bennett 2003; Winsborrow *et al.* 2010; Margold *et al.* 2015) the analysis of former ice stream bed morphology and sedimentology across large, geologically and topographically complex regions can provide valuable insights into the relative importance of local factors in governing ice flow velocities, landform genesis, and ice sheet/stream reorganization.

Despite over a century of investigation, a consensus on the origin of drumlins and other subglacial bedforms has yet to be reached (Schomacker *et al.* 2018). Recent theories on their formation fall into two general categories; selective erosion and streamlining of antecedent sediment by flowing media over an irregular and heterogeneous substrate (Eyles *et al.* 2016), or self-organizing localized deformation and remobilization of substrate materials due to instabilities at the ice bed (Stokes *et al.* 2013). Developing an improved understanding of streamlined bedform genesis is essential to enhancing knowledge of past ice dynamics, and the sensitivity and responses of ice sheets to external forcing agents.

One way to examine competing theories behind subglacial bedform generation involves detailed analysis of entire fields of bedforms and associated deposits. Areas where bedforms display strong coherence (parallel, curvilinear, or radial orientations), are consistent with general trends of esker orientations, and all landforms terminate (or show strong morphological changes)

Fig. 5.1: A. Regional topography, streamlined bedform orientations, and moraine distribution across south-central Ontario. Inset map in upper right shows location (red outline) within the Great Lakes Region of North America. Location of select figures shown in black outlines, thick grey dashed line marks the contact between Precambrian Shield and flat-lying Paleozoic rocks. Note the Lake Simcoe Moraines stretching from the north of Lake Simcoe to the east end of the ORM and the Brighton Moraine (shaded in grey shadow). Shield-Paleozoic contact shown in grey dashed lines. B. Distribution and orientation of streamlined bedforms in the Peterborough (PDF) and Lake Ontario (LODF) drumlin fields. The thick hatched black line in the west marks the crest of the Niagara Escarpment, the dashed line in the north marks the Precambrian-Paleozoic unconformity. Note the change in bedform orientation within the PDF and LODF to the north and south of the Oak Ridges Moraine (ORM; dashed red line). Stratified interlobate moraines shown in red ORM=Oak Ridges Moraine; OM=Oro Moraine)

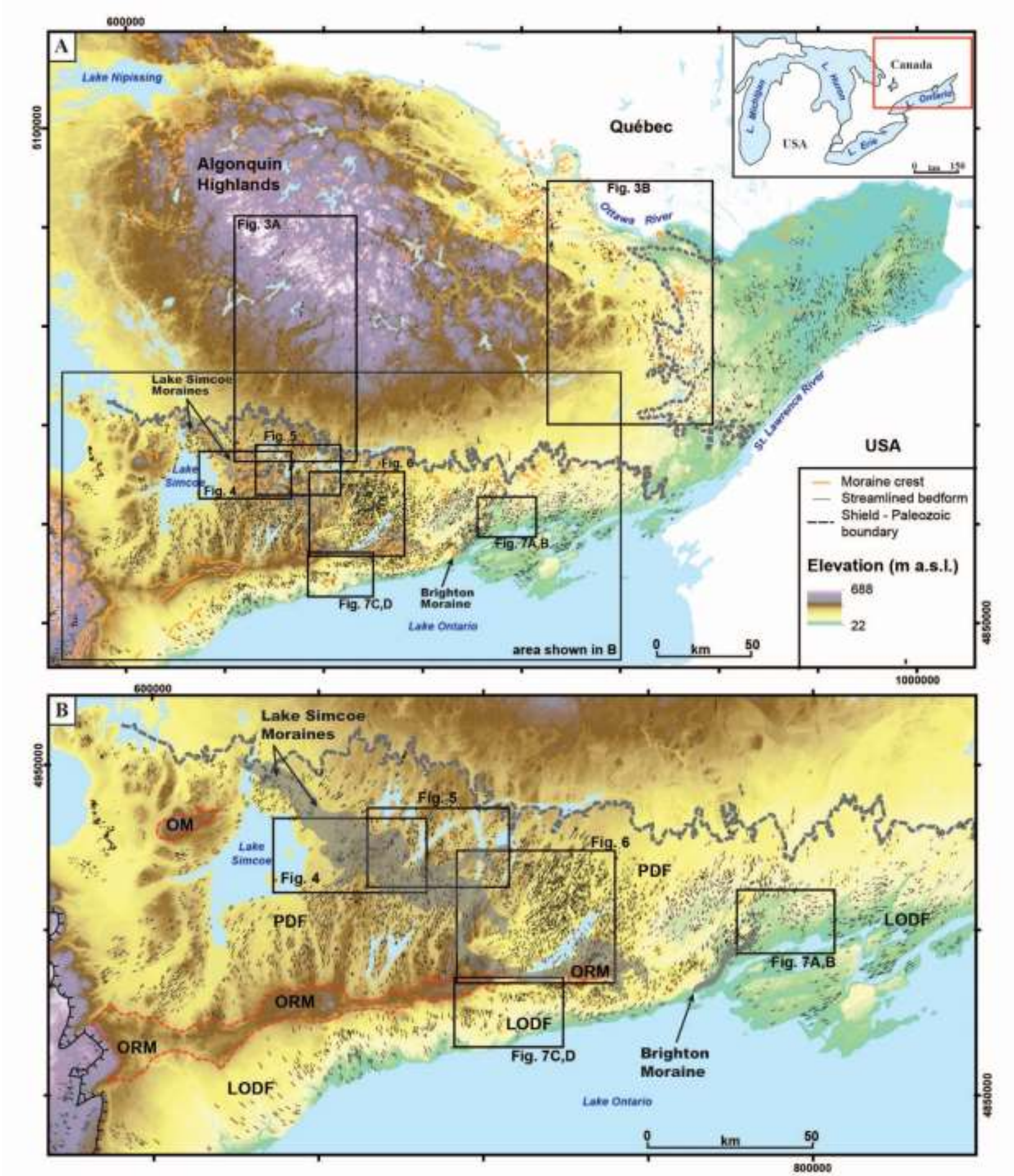

A
Lake Nipissing
Algonquin Highlands
Québec
Canada
USA
Fig. 3A
Fig. 3B
Ottawa River
St. Lawrence River
Lake Simcoe Moraines
Fig. 4
Fig. 5
Fig. 6
Fig. 7A,B
Fig. 7C,D
Brighton Moraine
Lake Ontario
area shown in B
Moraine crest
Streamlined bedform
Shield - Paleozoic boundary
Elevation (m a.s.l.)
688
22
km
600000
1000000
4850000
B
OM
PDF
ORM
LODF
800000

Table 5.1: Summary of landform characteristics recently mapped in south-central Ontario (area shown in Fig. 5.1A) and in the PDF and LODF (area shown in Fig. 1B). For additional information on bedform morphometry within the area the reader is referred to Maclachlan & Eyles (2013) and Eyles & Doughty (2016).

area	***feature***	***number***	***max length***	***avg. length***
1A	bedforms	18137	4.05	0.64
	moraines	5567	29.2	1.06
	eskers	711	71.9	2.18
1B	bedforms	8583	4.05	0.78
	moraines	1060	23.6	1.7
	eskers	230	71.9	3.85

at well-defined moraine ridges, define individual flow sets – areas of distinct landform populations traceable to a unique event (Clark 1993; Margold *et al.* 2015; Ely *et al.* 2018). Analysis of the characteristics and interrelationships of flow sets can provide valuable information on the origin and evolution of formerly glaciated landscapes.

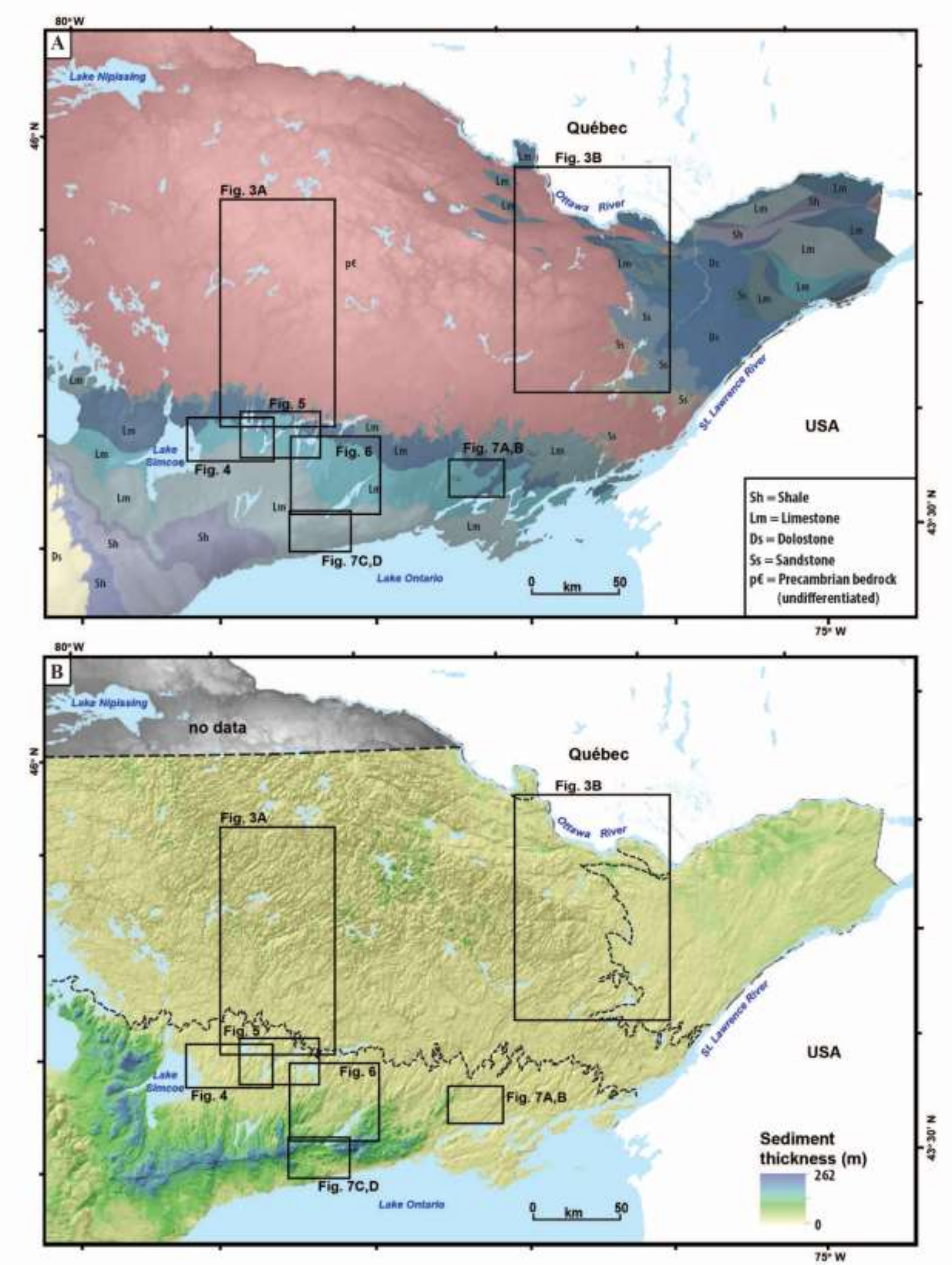
This paper presents preliminary results from regional digital landform mapping across south-central Ontario (Fig. 5.1A; Table 5.1). This work highlights the spatial arrangement and relationships of subglacial and ice-marginal landforms within the region in order to demonstrate the quality of recently-released terrain data and their utility in identifying and correlating regional landscape elements to guide and augment future field-based sedimentological investigations. The landform distribution and relationships presented here are used to test existing theories on regional landscape evolution in southern Ontario and highlight some significant questions regarding the evolution of this sector of the Laurentide ice sheet (LIS).

5.2. BACKGROUND

5.2.1. Regional geologic setting

The topography of south-central Ontario is dominated by the Algonquin Highlands of the Canadian Shield (Fig. 5.1A). This area of high ground is bordered to the west by the Georgian Bay basin, to the north by the French and Mattawa River lowlands, to the east by the Ottawa River valley and to the south by the St. Lawrence River valley and Lake Ontario basin. Thick successions of Cambrian and upper Ordovician strata onlap the flanks of the Algonquin Highlands in the Ottawa Valley and lowlands in the St. Lawrence River and Lake Ontario areas; a structural high in the basement rocks stretching from the Algonquin Highlands southeastward across the St. Lawrence valley into the Adirondacks of New York State disrupts this pattern (Fig.

Fig. 5.2: A. Generalized bedrock lithology of the study area, see key in lower right (Armstrong & Carter 2010). B. Sediment thickness within the study area (Gao *et al*. 2006). Note the rare and patchy distribution of thicker sediment accumulations within the Algonquin Highlands compared to areas underlain by Paleozoic rocks.


A
80° W
Lake Nipissing
Québec
Fig. 3B
Fig. 3A
Ottawa River
pЄ
Fig. 5
Fig. 4
Fig. 6
Fig. 7A,B
Fig. 7C,D
Lake Simcoe
Lake Ontario
St. Lawrence River
USA
Sh = Shale
Lm = Limestone
Ds = Dolostone
Ss = Sandstone
pЄ = Precambrian bedrock (undifferentiated)
0 km 50
46° N
43° 30' N
75° W
B
80° W
no data
Sediment thickness (m)
262
0

5.2; Armstrong & Carter 2010). The contact between Shield and Paleozoic terrain is marked by steep, north (up-ice) facing escarpments (up to 20 m high) separated by large re-entrant valleys in the south and is highly irregular in the east (Fig. 5.2) due to a more complex structural history characterized by significant syn- and post-depositional faulting of Paleozoic rocks.

The PDF lies to the south of the margin of the Algonquin Highlands (Fig. 5.1) and overlies upper Ordovician limestone plains (Fig. 5.2A). Within this area, thickness of glacial deposits increases rapidly toward the south, away from the Shield terrain; however, bare limestone terrain is common immediately east of the PDF and south near the margins of Lake Ontario and the St. Lawrence River (Fig. 5.2B; Gao *et al.* 2006). In these areas, drumlins, rock drumlins, whalebacks, megagrooves, and streamlined bedrock escarpment noses are directly juxtaposed and parallel, comprising parts of discrete flow sets (Sharpe *et al.* 2004; Eyles & Doughty 2016; Sookhan *et al.* 2018b). Previous mapping indicates that most of the drumlins in the PDF are cored by well- to highly-consolidated till (Newmarket/Northern Till; see discussion below), although stratified glaciofluvial and/or fine-grained glaciolacustrine sediment is observed in some drumlins (Gravenor 1957; Shaw & Sharpe 1987; Boyce & Eyles 1991; Eyles & Doughty 2016; Marich 2016). The drumlins have a predominantly northeast-southwest long-axis orientation north of the Oak Ridges Moraine (ORM; Fig. 5.1B; Maclachlan & Eyles 2013; Sharpe *et al.* 2013), whereas south of the ORM they show a northwest-southeast long axis orientation. These bedform orientations have previously been ascribed to shifting ice flow directions during the late glacial (Maclachlan & Eyles 2013; Eyles & Doughty 2016; Mulligan *et al.* 2016; Sookhan *et al.* 2018a, b) or deflection of flow lines during discrete, successive flood events from different directions (Shaw & Gilbert 1990; Sharpe *et al.* 2018).

5.2.2. Historical studies

The overall sequence of events that characterize the build-up and retreat of the LIS in the region is widely accepted, although many questions remain on the exact timing of events, and the nature of landform genesis in the region. Ice sheet growth during the Early Wisconsinan blocked eastern drainage of the Great Lakes, resulting in elevated lake levels in southern Ontario (Karrow 1967; Brookfield *et al.* 1982; Eyles & Eyles 1983). Ice subsequently advanced over the region from the north and northeast, depositing a regional till sheet (NT) during the Late Wisconsinan. Towards the end of glaciation, increased meltwater production led to the deposition of thick successions of stratified sand, gravel and silt comprising the ORM (Fig. 5.1) between two ice masses lying to the north and south (Chapman and Putnam 1984; Barnett *et al.* 1998). As ice retreated, high-level lakes formed, dammed against high topography and the receding ice margin until successively lower outlets were exposed permitting drainage of the lakes.

Early studies commented on glacial landforms (including drumlins in the PDF) as part of regional geomorphological research conducted within the Great Lakes region (Leverett and Taylor 1915) or southern Ontario (Putnam & Chapman 1936). Increased focus on systematic Quaternary geological mapping by government agencies during the mid to late 20th century produced more detailed maps of the region (Deane 1950; Gravenor 1957; Gadd 1980), and improved descriptions of the composition, distribution and relationships of local bedforms with other glacial features (Finamore & Courtney 1982; Barnett 1983). Gravenor (1957) proposed that drumlins within the PDF were erosional features, based on the occurrence of three types of drumlins; rock drumlins, drumlins composed of antecedent stratified sediments, and drumlins composed of the regional Late Wisconsinan till, the NT. Later studies related drumlin morphology to ice dynamics and evolving flow conditions (Boyce & Eyles 1991).

Following the proposal of the meltwater hypothesis for subglacial bedforms (Shaw 1983), renewed investigations began in the PDF (Shaw & Sharpe 1987; Shaw & Gilbert 1990; Shaw & Gorrell 1991; Kor *et al.* 1991; Brennand & Shaw 1994; Sharpe *et al.* 2004). These provided a unifying theory for landforms (bedrock escarpments, s-forms, drumlins, tunnel channels, eskers), and the ORM within the PDF area and areas to the south of Lake Ontario (Sharpe *et al.* 2004). The meltwater hypothesis has been widely criticized based on a number of issues including the source and fate of meltwaters, the fate of sediments eroded during the flood(s), and diachronous regional landform development within areas interpreted as former flood paths (see Clarke *et al.* 2005; Ó Cofaigh *et al.* 2010; Sookhan *et al.* 2018a; Eyles *et al.* 2018).

The most recent studies integrate a digital terrain model for the region (Kenny 1997) that was developed to support regional three-dimensional (3D) geological mapping for groundwater investigations in the ORM area (Barnett *et al.* 1998; Sharpe *et al.* 2002; Logan *et al.*, 2005). This, and later models (Ministry of Natural Resources and Forestry 2010, 2014a, b), greatly improved visualization of terrain elements and glacial landforms across southern Ontario, permitting more detailed, statistical analyses of drumlin morphology, orientations, distribution, and clustering within the PDF (Maclachlan & Eyles, 2013). Surficial mapping integrated with terrain analysis has identified new complex sediment-landform assemblages and erosional features within different parts of the PDF, recording different intensities of bedform genesis and postglacial erosion throughout the region (Marich 2014; 2016; Eyles & Doughty 2016; Mulligan *et al.* 2016; 2018a; Sookhan *et al.* 2018a). Recently-released high-resolution terrain models and aerial photographs (Ministry of Natural Resources and Forestry 2010; 2014a, b) provide the detailed morphological information required to shed new light on previous work within the PDF and surrounding areas, and to place observations of bedform relationships and composition into a broader regional context at the scale of former ice lobes/streams.

5.2.3. Nomenclature and terminology

A complex and confusing set of names for stratigraphic units and till sheets exists in southern Ontario (Brookfield *et al.* 1982); some of the issues related to nomenclature were highlighted in a recent review by Sharpe & Russell (2016). In this paper, the name Newmarket/Northern Till (NT) is given to the regional Late Wisconsin till sheet that stretches from the Niagara Escarpment in the west to the shores of Lake Ontario to the south and east. Undifferentiated Late Wisconsin tills found in patches in Shield terrain (e.g. Barnett, 1992a) are assumed here to be equivalent to the NT. The name Halton Till (HT) is given to a fine-grained surficial till found primarily in the western parts of the Lake Ontario basin and in close proximity to the southern flanks of the ORM (Sharpe & Russell 2016). Temporally equivalent deposits of till and stratified diamict are also observed locally on the northern flanks of the ORM (White 1975; Mulligan 2013). The PDF is defined in this study as predominantly NE-SW oriented streamlined terrain stretching from the Niagara Escarpment to the Brighton Moraine (Fig. 5.1, with a southern limit beneath the ORM (Fig. 5.1). The term 'Lake Ontario drumlin field' (LODF) is applied here for all streamlined bedforms oriented SE-NW observed south of the ORM, as well as for a broad flow set that cross-cuts PDF drumlins along its easternmost edge (Fig. 1B). MSGL are highly attenuated (l:w ratio >10), strongly parallel, and usually low-relief (<10 m high) bedforms composed of a variety of materials, typically subglacial till, or stratified sediments, that are viewed as diagnostic features of fast ice flow velocities associated with ice streams (Clark 1994; Stokes & Clark 1999; 2001; Spagnolo *et al.* 2014; Margold *et al.* 2015). Drumlins in the region are generally taller (commonly 20-30 m), shorter (800 – 1500 m) and wider (l:w ratios <5) than

MSGL, and may be composed of sediments or bedrock. Erosional streamlining of bedrock produces a wide variety of forms, depending on the lithology, orientation of structural features relative to ice flow direction, and glaciological factors (Krabbendam *et al*. 2016). The most common features that occur within the study area are streamlined escarpments, whalebacks/rock drumlins, roche moutonnées, mega grooves and subglacially sculpted bedrock (s-) forms (Gravenor 1957; Sharpe *et al*. 2004; Eyles & Doughty 2016).

5.3. METHODS

This study of landforms in central and eastern Ontario involved analysis of 2 m (Ministry of Natural Resources and Forestry 2104a, b) and 5 m (Ministry of Natural Resources and Forestry 2010) digital elevation and terrain models with hillshade relief maps. Landforms were manually mapped directly into a GIS using ArcMap 10.3 (Table 5.1). Where only subtle features were visible, or if the morphology from the terrain models was ambiguous, features were cross-checked using existing surficial geology maps (Ontario Geological Survey 2010), recently collected field data (>2500 stations) from surficial mapping in the central 2200 km^2 of the PDF (Marich 2016), air photos (Ministry of Natural Resources and Forestry 2014a, b), satellite imagery (Google Earth), or previous reports (e.g. Gravenor 1957; Chapman & Putnam 1984; Shaw & Gorrell 1991; Brennand & Shaw 1994; Eyles & Doughty 2016). The process of cross-checking features is particularly important for the 2 m terrain data sets, as they are hybrid surface/terrain models, where areas of dense coniferous vegetation have not been completely removed (Ministry of Natural Resources and Forestry 2014).

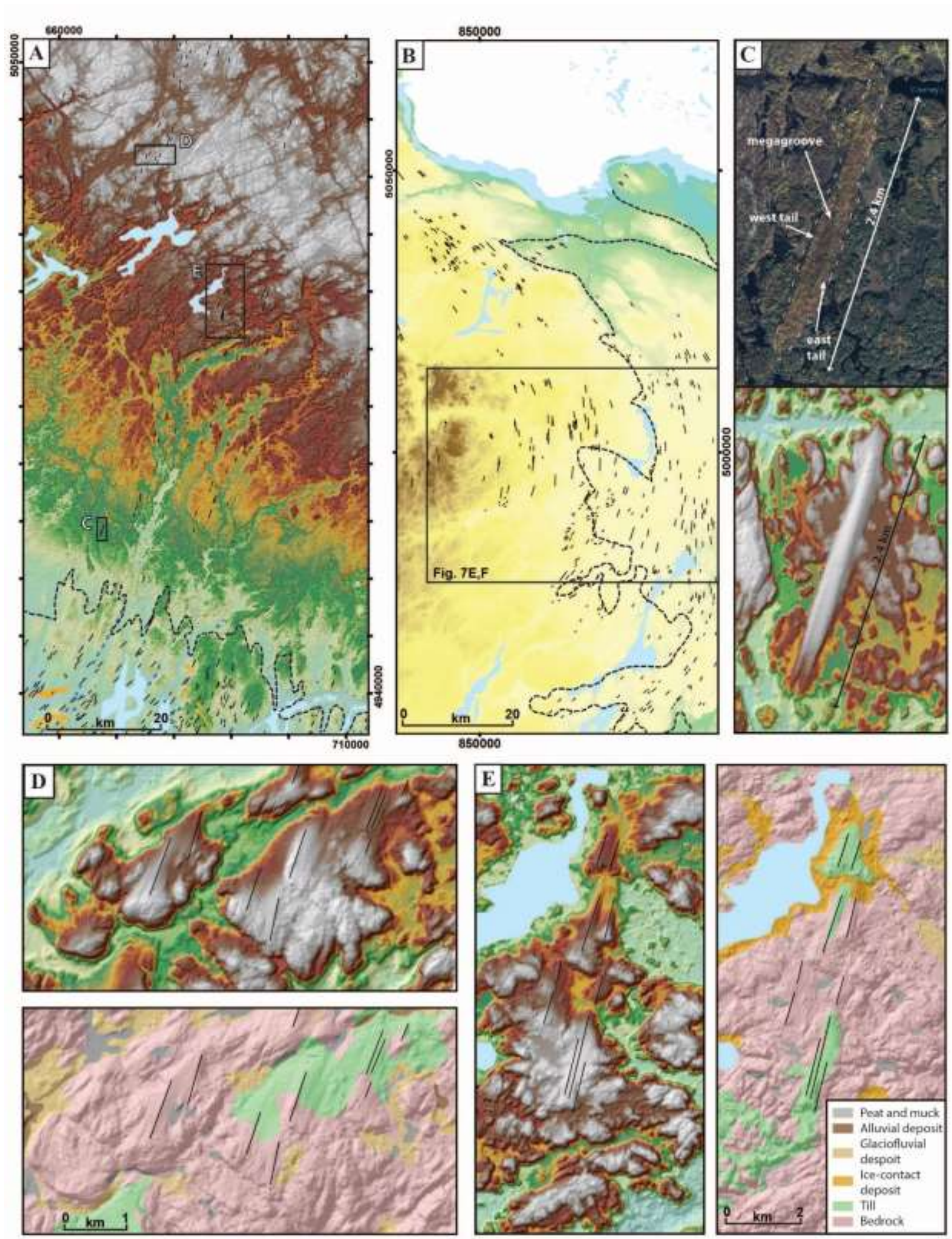
5.4. RESULTS

Given the large geographic area covered in this study, results are presented for smaller geographic sub-areas and common landsystem assemblages in order to present new landform inventories for areas that have previously received little attention, and to highlight the changing morphology/landform relationships observed across physiographic boundaries such as major lithologic or structural divisions. Presented below are new data on MSGL occurrences and widespread bedrock streamlining in Shield terrain, new moraine identification and correlation within the PDF, streamlined bedform characteristics within the PDF, and varying scales of cross-cutting flow sets (a few km to tens of km wide), both within local areas of the PDF and large areas within the Ottawa River, St. Lawrence River, and Lake Ontario lowlands (Fig. 5.1).

5.4.1. MSGL and streamlined bedrock in Shield terrain

Numerous MSGL and widespread areas of bedrock streamlining are observed across Shield terrain comprising the Algonquin Highlands of central Ontario (Figs 5.1, 5.3). Generally, they are more common in the central and eastern parts of the Shield terrain (Fig. 5.1), where drift is locally thicker (Fig. 5.2B) and where the surficial materials are commonly mapped as till (Barnett 1983; Ontario Geological Survey 2010). Local clusters of streamlined bedforms are also observed in the north and south; megagrooves and rock drumlins/whalebacks occur on bare bedrock and MSGL occur on adjacent patches of streamlined till that is reported as more consolidated and has lower stone and sand content than thin till in intervening zones between MSGL (Finamore & Courtney 1982). In the eastern part of the Algonquin Highlands, a curvilinear path is demarcated by the long axis trends of the MSGL and drumlins, following the flanks of the high topography in the interior of the Algonquin Highlands (Gadd 1980; Fig. 5.3A,B). In the central parts of the Algonquin Highlands, the trend of MSGL long axes is parallel

Fig. 5.3: A. 2-m terrain model showing topography of the south-central part of the Algonquin Highlands and Paleozoic terrain downflow (south of black dashed line). Streamlined bedforms shown in black, eskers in dashed blue, and moraine crests in solid orange lines. Note the generally parallel orientation of streamlined features on the Shield, and their relationship with features to the south. See Fig. 5.1 for location. B. 25-m elevation model showing orientation and distribution of streamlined bedforms (black lines) in the Ottawa- St. Lawrence lowlands. Note the curvilinear track of bedforms, parallel to the overall contours of the eastern part of the Algonquin Highlands. Location of Fig. 5.7E, F shown in black, see Fig. 5.1 for location. C. Satellite image (top; Google Earth) and 2-m elevation model of MSGL on a streamlined till residual in the southern Algonquin Highlands (see Fig. 5.3A). The feature is bisected along the central axis, creating two ‘tails’. D, E. 2-m elevation model and surficial geology overlain on hillshade (2-m cell size) showing morphology of streamlined features and their relation to mapped sediment cover (Ontario Geological Survey 2010).


A
B
C
D
E
660000
710000
5050000
4950000
850000
5000000
0 km 20
Fig. 7E,F
megagroove
west tail
east tail
2.4 km
0 km 1
0 km 2
Peat and muck
Alluvial deposit
Glaciofluvial despoit
Ice-contact deposit
Till
Bedrock

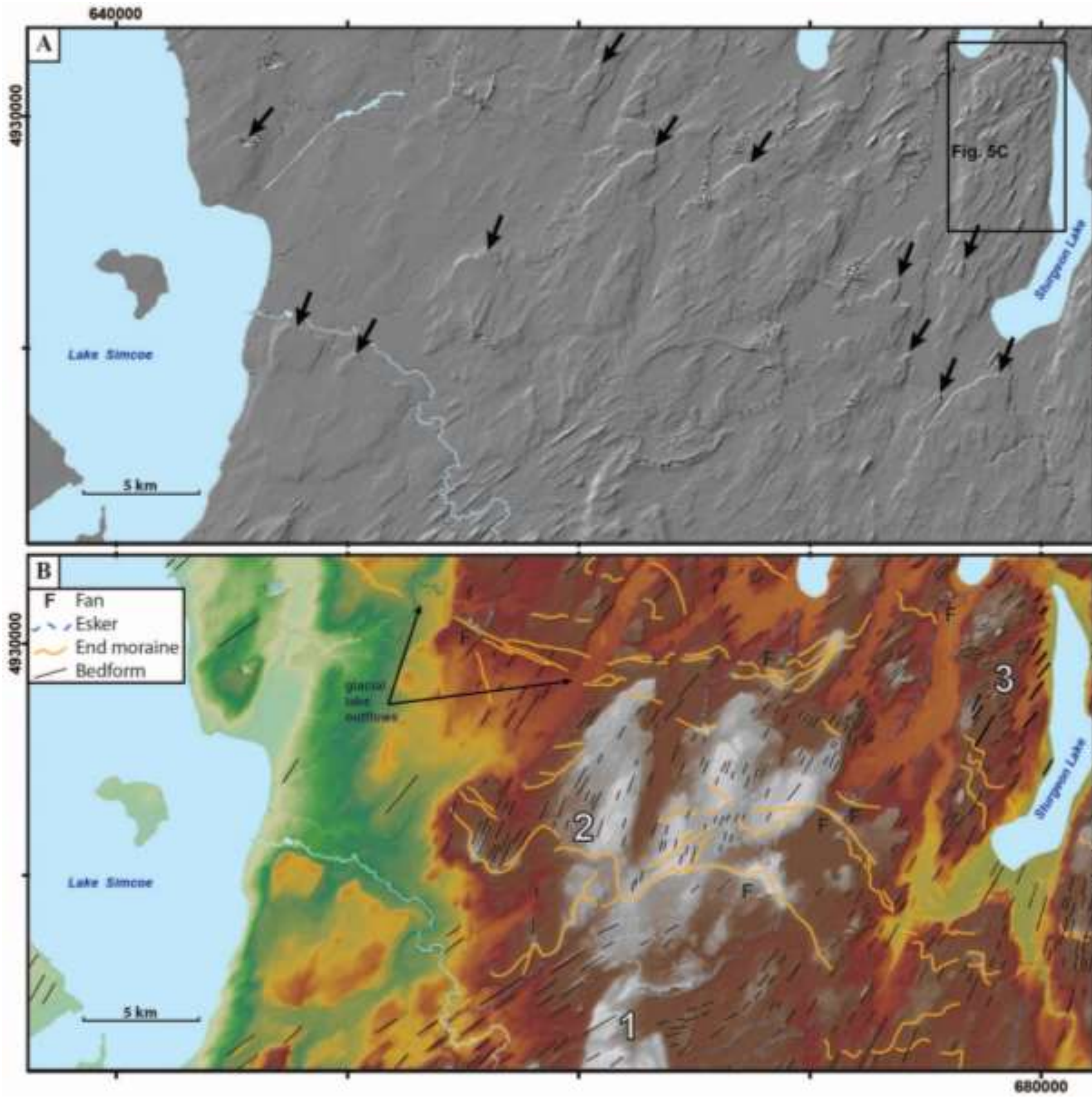
to regional striae orientations (generally NNE-SSW) reported from adjacent areas (Barnett 1983; Bajc 1994) and shows little to no variation (Fig. 5.3B,C; see below). Streamlined bedforms are much less common within the Algonquin Highlands compared to the PDF and LODF (Figs 5.1A, 5.3A,B).

Till-cored MSGL observed on Shield terrain are commonly 1-2.5 km long (Fig. 5.3C), but range up to 4 km long and 300 m wide. In many cases, clusters of MSGL occur in narrow bands oriented parallel to ice flow (Fig. 5.3E), or along transverse bands nearly orthogonal to former ice flow directions. Streamlined features that occur on areas where bedrock is mapped at surface are generally shorter (<700 m long) and narrower (<100 m wide), and are interpreted here as part of a spectrum of streamlined bedrock forms comprising grooves, megagrooves, whalebacks and rock drumlins. In the southern part of the Algonquin Highlands, MSGL and streamlined bedrock forms are parallel to, and grade southwestward into, drumlins and other bedforms within the PDF (see below).

5.4.2. Moraine identification and correlation in the PDF

Analysis of the new terrain data has led to the identification of numerous sinuous ridges that trend roughly east-southeast from near the northern shore of Lake Simcoe, past the southern end of Rice Lake, to (and wrapping around) the eastern edge of the ORM (Fig. 5.1C). Recent detailed surficial mapping identified segments of the ridges separated by broad patches of low-relief, hummocky till plain in central parts of the PDF (Marich 2016); the latest release of high-resolution terrain data in this region has facilitated the correlation of subtle ridges crests over larger areas which is presented here. The northwesternmost ridges are coincident with the previously-mapped 'Lake Simcoe Moraine' (Taylor 1913; Deane 1950; Sookhan *et al.* 2018b)

Fig. 5.4: A. 5 m hillshade of PDF east of Lake Simcoe (see Fig. 5.1 for location). Streamlined bedrock escarpment noses eroded into Ordovician limestones shown by black arrows (pointing in direction of former ice flow). B. Digital elevation model and glacial landform distributions for area shown in (A). Note the local curvilinear paths of bedforms in the northeast, and their correspondence to the location and orientations of valleys, lake basins, and gaps between streamlined escarpment noses. At least three generations of bedforms are visible, labelled 1-3 (in order of relative age), and are identified by strong obliquity of bedform orientations locally (center of image) and/or the overprinting of large (older) drumlins (shown with thicker black lines) with more subtle flutings (northeast corner, see Fig. 5.5C). The downflow extent of each flow set is marked by terminal moraines.


640000
4930000
A
Fig. 5C
Lake Simcoe
Sturgeon Lake
4920000
5 km
B
F Fan
Esker
End moraine
Bedform
glacial lake outflows
2
3
1
F
680000

and we suggest that the mapped ridges presented herein represent the newly-identified southeastern extension of this moraine system (see Discussion).

The sinuous ridges are grouped into two quasi-parallel sets extending from Lake Simcoe to the eastern part of the ORM (Fig. 5.1C), except in two areas; east of Lake Simcoe (Fig. 5.4), where a complex set of cross-cutting ridges is observed, and east of Sturgeon Lake, where the ridges show significant local deflections in the vicinity of lake basins (Fig. 5.5). In general, the ridges are subtle (up to 6 m high and 500 m wide), but thicker sediment accumulations (>15 m) are observed in some valleys that occur throughout the drumlinized NT plain (Figs 5.1, 5.2B, 5.4, 5.5). The ridges are composed primarily of silty sand till with lesser sand and gravelly interbeds (Marich 2016) and commonly have relatively smooth morphologies and maintain relatively uniform widths throughout much of their extent. However, in some locations in the west and northwestern part of the PDF, the ridges abruptly widen into bulbous lobes up to 1.5 km wide (Figs. 5.4B, 5.5B), which are cored with variable accumulations of ice-marginal sand and gravel deposits (subaquatic fans (F); Figs 5.4-5.6; Marich 2016). These lobes commonly have narrow, sinuous, gravelly (esker) ridges emanating from their northeast (upflow) and/or southwest (downflow) ends (Figs. 5.4B, 5.5B), and preferentially occur southwest of topographic lows, such as lake basins, or gaps between streamlined bedrock escarpment noses.

5.4.3. Streamlined bedforms in the PDF

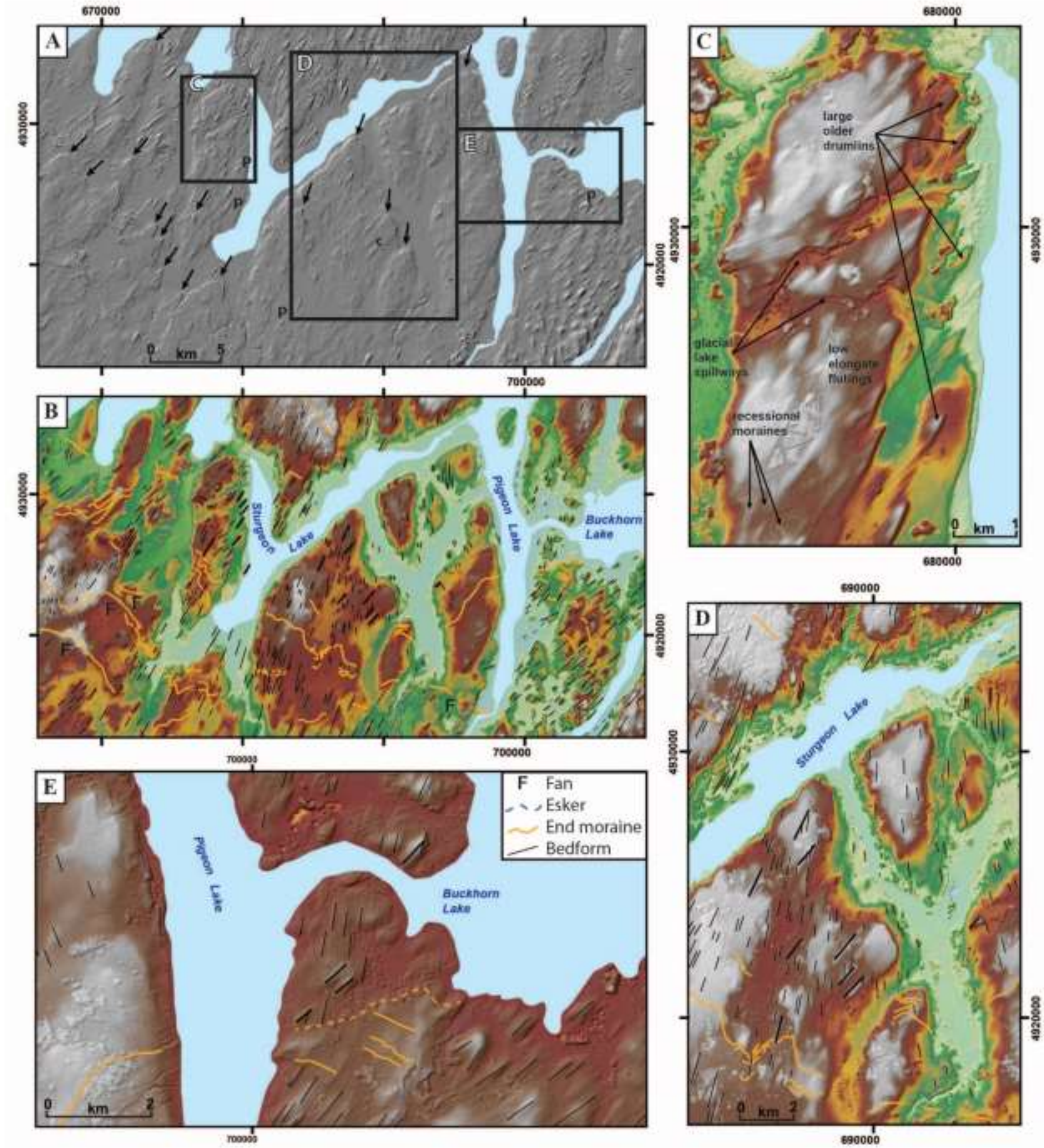
Complex bedform patterns and glacial landform relationships can be identified within the PDF. Much of the observed complexity occurs in areas in close proximity to (usually within ~15 km to the northeast of) newly-identified till-cored ridges of the Lake Simcoe Moraines that span the central part of the PDF (Figs 5.1C, 5.4, 5.5; see above).

In the southern and western parts of the PDF, drumlins show a broad, uniform pattern, changing from NNE-SSW in the southern parts, to NE-SW in the southwest near the Niagara Escarpment, and ENE-WSW to the west of Lake Simcoe (Fig. 5.1B; see also Sookhan *et al.* 2018a). In this part of the PDF, the drumlins are large and generally broad (up to1500 m long x 500 m wide), with more attenuated forms rarely observed. Drumlins throughout this part of the PDF ornament the highly undulating surface of the NT and are generally similar in size, except directly north of Lake Simcoe where they are commonly thinner and more elongate, juxtaposed with MSGL resting on Paleozoic limestones. Few small (< few m high), discontinuous transverse ridges are observed overlying drumlins along the western shores of Lake Simcoe and have previously been interpreted as 'ice-block ridges', crevasse fills, or minor moraines (Deane 1950; Barnett 1992b; Mulligan & Bajc 2012).

North and east (upflow) of the newly-mapped Lake Simcoe Moraines (Fig. 5.1C; see above), more complex bedform patterns are observed (Figs 5.4-5.7). In general, large, broad drumlins showing similar characteristics and orientations to those described above occur on high ground in these areas, but they are cross-cut and/or re-moulded by superimposed bedforms at oblique angles, or truncated entirely by a younger set of highly elongate streamlined bedforms (Figs 5.5-5.7).

The position and trend of superimposed bedforms within the PDF show strong correlation with the presence and orientation of gaps in prominent (up to 20 m high) bedrock escarpments or significant valleys cut into older Pleistocene sediments, and locally eroding into underlying Paleozoic bedrock (previously interpreted as tunnel channels; Brennand & Shaw 1994; Sharpe *et al.* 2004). Commonly, the superimposed bedforms occur within distinct flow sets displaying fanning-outward long-axis orientations that diverge from the south and west ends of topographic

Fig. 5.5. A. 5 m hillshade of PDF in the Sturgeon Lake area (see Fig. 5.1 for location). Streamlined bedrock escarpment noses eroded into Ordovician limestones shown by black arrows (pointing in direction of former ice flow). Multiple classes of streamlined bedforms are visible including comma-form (c), and parabolic (p). B. Digital elevation model and glacial landform distributions for area shown in (A). Note the general parallelism of large, broad (older) drumlins (bold black lines) compared to the radial fanning of younger low-relief flutings (thin black lines) that commonly terminate at end moraines. C. Detailed view of morphology west of Sturgeon Lake. Apparently parabolic forms shown in (A) can be explained through multiple discrete streamlining phases: an early drumlin-forming phase from ice flowing SSW, followed by a later phase by SW-flowing ice. Two sinuous channels cross-cut the upland, and are marked with flat-topped lobate features at their eastern ends. These are interpreted as glaciofluvial or glacial lake spillways that would have developed following deglaciation, while ice still blocked lower outlets to the north. D, E. close-up views (see (A) for location) of bedform orientations and relationships to end moraines. Note the radial fanning and orthogonal orientation of younger flutings to end moraines in (D) and the clear termination of a flow set identified by flutings at a set of small moraine ridges.

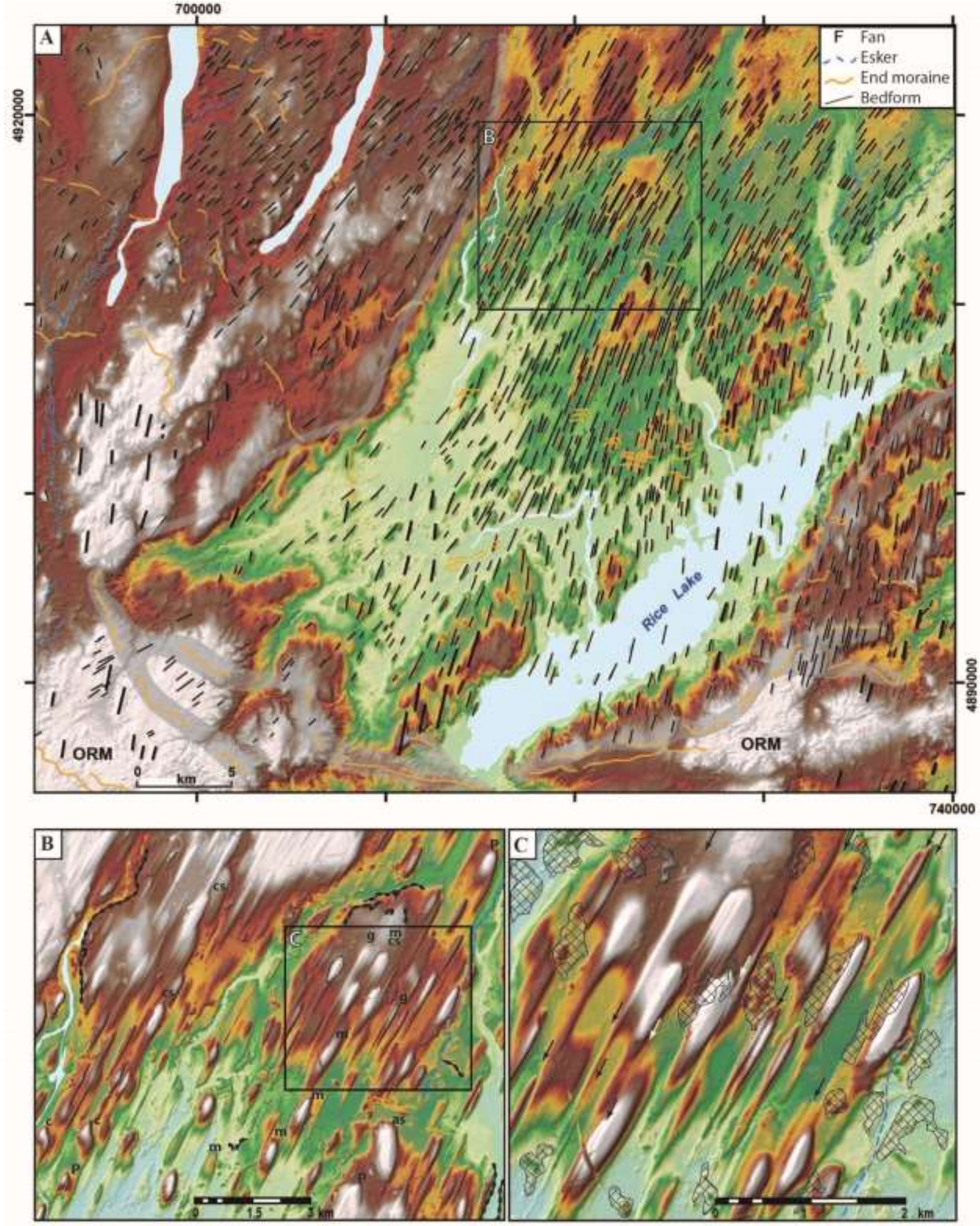

A
B
C
D
E
large older drumlins
glacial lake spillways
low elongate flutings
recessional moraines
Sturgeon Lake
Pigeon Lake
Buckhorn Lake
F Fan
Esker
End moraine
Bedform
670000
680000
690000
700000
4930000
4920000
0 km 5
0 km 1
0 km 2

lows and terminate at small, till-cored ridges associated with the Lake Simcoe Moraines (Figs 5.1, 5.4, 5.5, 5.6).

In the eastern part of the PDF, in the Peterborough and Rice Lake areas, a broad tract of low-lying terrain (herein referred to as the Rice Lake Corridor; Fig. 5.6) is characterized by densely distributed, highly elongate drumlins, grooves, and MSGL (Maclachlan & Eyles 2013; Marich 2014; Eyles & Doughty, 2016). The NT is mapped at surface throughout the floor and flanks of the Rice Lake Corridor as well as in upland areas surrounding it to the east and west, and on localised areas of high ground in the northeast and southwest of the corridor. These local areas of higher ground within the Rice Lake Corridor clearly affected the flow lines of the youngest flow set, which are recorded by divergent MSGL and complex drumlin forms around the highs within the corridor (Fig. 5.6A). Large, broad drumlins (likely equivalent to those observed in the southwestern part of the PDF) occur within and beyond the southern limit of the Rice Lake Corridor and are oriented oblique to the superimposed highly attenuated bedforms in the vicinity (Fig. 5.6). The high ground within the southwesternmost part of the Rice Lake Corridor are ornamented with arcuate, sub-parallel hummocky ridges of sand, gravel, and till (Ontario Geological Survey 2010) that form part of and/or partially overlie the northern flanks of the ORM (Fig. 5.6A).

The superimposed bedforms within the Rice Lake Corridor ornament and modify older broad drumlins and are highly attenuated and parallel, displaying a slightly curved path in the northeast (upflow) areas (Fig. 5.6A). Close inspection of the bedforms in the central part of the Rice Lake Corridor reveals a complex assemblage of highly attenuated MSGL (>4km long), drumlins, (mega)grooves, eskers, and later postglacial features including glacial lake shorelines, spillways, deltas (Fig. 5.6B; Marich 2016). As noted by previous workers, numerous drumlin

Fig. 5.6*:* A. Bedform distribution and orientation within the Rice Lake Corridor in the eastern part of the PDF (lateral boundaries shown by transparent grey lines, see Fig. 5.1 for location). This area was previously interpreted to mark the location of a lateglacial ice stream (Maclachlan & Eyles 2013) and has the densest distribution of streamlined forms within the PDF. Large, broad (older) drumlins are consistently oriented NNE-SSW and are best preserved outside the margins of the Rice Lake Corridor. A few older drumlins are preserved within the Rice Lake Corridor, particularly in the southwestern part, where it appears ice was thinner, given the deflection of bedforms around the upland near the southwest end of Rice Lake. A few large drumlins (bold lines) are visible among the otherwise low-relief MSGL in the central part of the Rice Lake Corridor. B. Close-up of the central part of the Rice Lake Corridor, where a wide range of streamlined bedform morphologies are visible. Spindle forms predominate, but comma-form (c), asymmetric (as), parabolic (p), multi-tailed (m), grooved/bisected (g), and features with crescentic scours around their stoss ends (cs). Many of these more complex forms originate downflow of large, high streamlined hills, which likely represent re-moulded older drumlins. Black hashed lines denote areas where Palaeozoic bedrock crops out at surface. C. Enlarged view of subsection of the area shown in (B) demonstrating the common occurrence of linear to slightly curved grooves (arrowed) that appear to entirely truncate numerous drumlin forms and possibly create the multi-tailed drumlins. Black hashed polygons show areas of dense coniferous vegetation has led to ground surface interpolation issues.



A
700000
4920000
4890000
740000
F Fan
Esker
End moraine
Bedform
Rice Lake
ORM
ORM
0 km 5
B
C
0 1.5 3 km
0 1 2 km

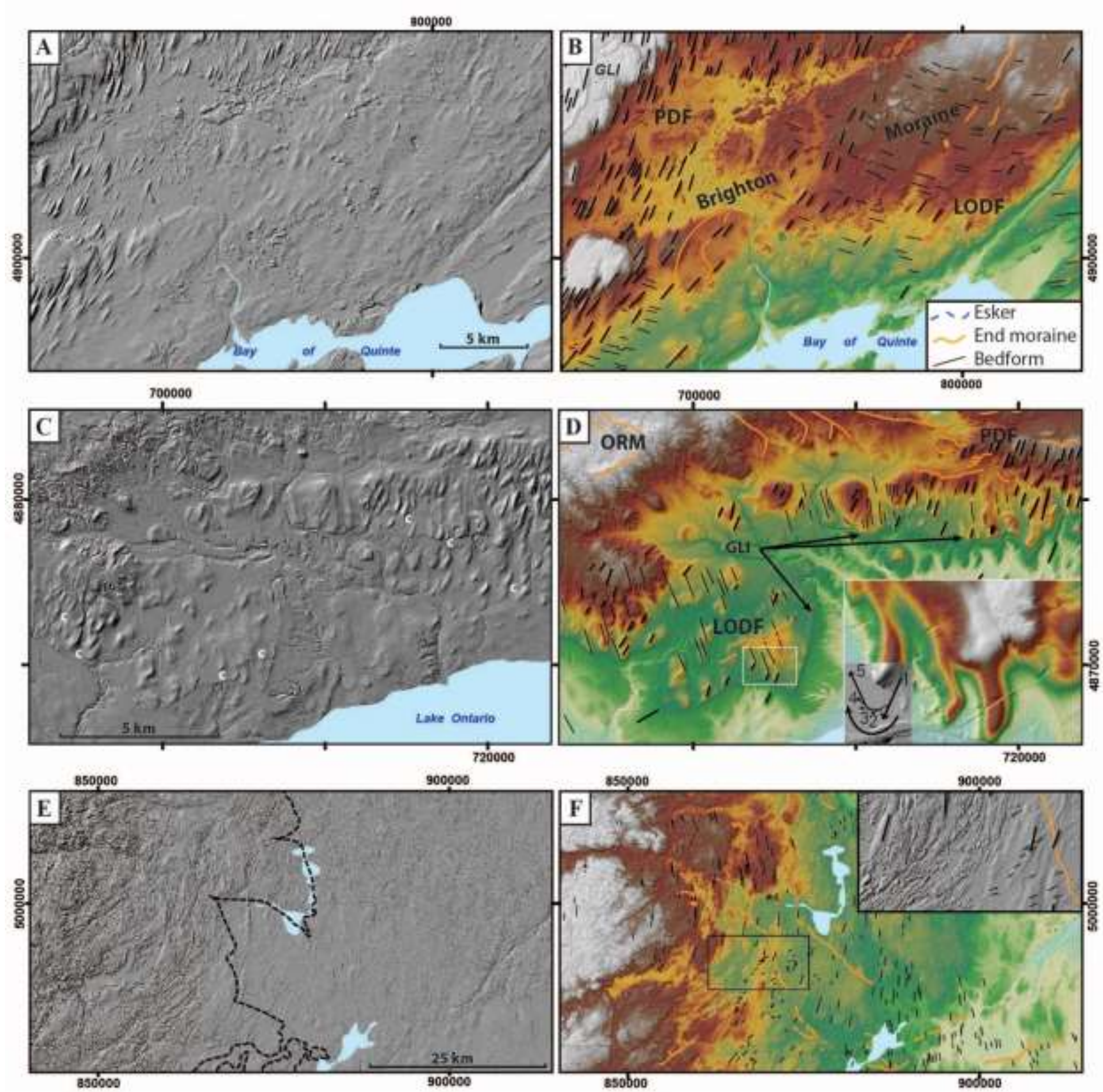
morphologies are visible, including spindle, comma, asymmetric, bisected, parabolic, 'multi-tailed', and forms with crescentic scours around their stoss ends (e.g. Crozier 1975; Shaw & Sharpe 1987; Boyce & Eyles 1991; Marich 2014; 2016; Eyles & Doughty 2016; Eyles *et al.* 2018; Fig. 5.6B,C). Drumlins occur *en echelon*, in nested groups, and in clusters. The long axis of the large (older) drumlins may be parallel or obliquely oriented to the longer, younger flutings comprising the flow set within the Rice Lake Corridor. In some cases, long, parallel grooves appear to completely bisect older drumlin/fluting forms (Fig. 5.6B); many of these grooves extend for considerable distances (>5 km), and may truncate drumlins/flutings downflow, or terminate at the stoss end of the next bedform to the southeast (Fig. 5.6B,C).

A rapid shift in drumlin/fluting morphology and abundance is observed across the western margin of the Rice Lake Corridor (Fig. 5.6A), coincident with a sharp rise in elevation of the terrain. Bedforms flanking the western margin of the Rice Lake Corridor have highly variable, complex, low-relief morphologies, and little to no consistency in their orientations (Fig. 5.6A).

5.4.4. Cross-cutting flow sets

In addition to the many large drumlins and MSGL that are locally ornamented with low-relief flutings or grooves within the PDF, extensive (tens of kms wide) tracts of cross-cutting flow sets are observed, consisting primarily of flutings, MSGL, and superimposed streamlined bedforms (Fig. 5.7). In some areas, the original drumlin forms are difficult to recognize. In general, drumlins north of the ORM and in the Ottawa valley are merely ornamented with younger, obliquely-angled flutings (except in the area southeast of Sturgeon Lake and Rice Lake Corridor; Figs. 5.5, 5.6), whereas original drumlin forms are extensively altered in the eastern part of the LODF and to the south of the ORM (Fig. 7B-E).

Fig. 5.7: A,B. 2 m cell size hillshade and interpreted elevation model for the area east of the ORM and PDF. Older PDF drumlins trace a broad, curvilinear path form the NNE toward the Lake Ontario basin to the WSW. These are ornamented and/or reworked by later ice flow from the east and southeast. Note that unlike the previous examples, no moraine is visible to mark the extent of the late-stage ice advance, but subtle recessional ridges of the Brighton Moraine are shown in orange. Nearshore erosion along the former Glacial Lake Iroquois (GLI) shoreline may be partially responsible for the erasure of a moraine, if it was deposited. C, D. 5 m cell size hillshade and interpreted elevation model for the area south of the ORM and PDF. Multiple apparent comma-forms (labelled with lowercase c's on map) are marked. Inset shows detail on 2 m DEM, showing tcomma-forms consist of a broad hill oriented NE-SW (interpreted as older drumlin forms), with a fine linear ridge emanating from their NE end indicating a clockwise shift in ice flow directions (arrows labelled 1-5) and re-moulding by ice flowing NW and NNW from within the Lake Ontario basin during the formation of the LODF, with a downflow limit along the flanks of the ORM in the north. E, F. 2 m cell size hillshade and interpreted elevation model for the Ottawa Valley area. The Precambrian-Paleozoic unconformity is marked with a dashed black line.


A
B
C
D
E
F
GLI
PDF
Brighton
Moraine
LODF
ORM
Bay of Quinte
Lake Ontario
Esker
End moraine
Bedform
5 km
25 km
800000
4900000
700000
720000
4880000
4870000
850000
900000
5000000

South of the ORM, a set of drumlins, MSGL, and irregular patches of streamlined terrain form the LODF (see also Maclachlan & Eyles 2013; Eyles & Doughty 2016; Eyles *et al.* 2018; Sookhan *et al.* 2018a). Bedforms comprising the LODF can be identified across all areas bordering Lake Ontario, including areas above the Niagara Escarpment to the west, fanning outward from the lake and terminating near the southern margins of the ORM (Fig 7D,E; Chapman & Putnam 1984; Mulligan *et al.* 2016; Burt & Mulligan 2017; Sookhan *et al.* 2018b). The most common bedforms within the LODF are low-relief drumlins and MSGL oriented nearly orthogonal to modern Lake Ontario shorelines. Oblate, asymmetric, or comma-form drumlins are also common (Fig. 5.7); rare large, broad drumlins oriented NE-SW (consistent with a continuous, older flow set observed in the southwestern parts of the PDF) are often ornamented with secondary 'tails', or are entirely reshaped, by a secondary flow set oriented towards the northwest (Fig. 5.7D,E). Bedforms with consistent orientations to those described here are also observed on the surface of the HT (and regional equivalents) and are associated with a series of subtle till-cored moraine ridges to the west (Karrow 2005; Eyles *et al.* 2011; Burt & Mulligan 2017) and south of Lake Ontario (Kozlowski *et al.* 2018).

To the east of the ORM, a broad set of low-relief bedforms cross-cuts older PDF drumlins in an area of thin drift overlying Paleozoic bedrock north of Lake Ontario (Fig. 5.7B,C). The bedforms record a curvilinear flowpath that parallels the St. Lawrence River and long-axis of Lake Ontario in the east and diverges toward the northwest in the western part of the flow set. The bedforms are generally shorter and more poorly-developed than those in many other areas of southern and central Ontario and their downflow extent remains uncertain as no continuous moraine ridge is observed. They are observed to the south of the former glacial Lake Iroquois shoreline (Fig. 5.7A,B) and in the southwesternmost part of the flow set, the bedforms end at a subtle sandy ridge, corresponding to the previously interpreted Brighton Moraine (Figs 5.1, 5.7A,B; Chapman & Putnam, 1984).

In the Ottawa Valley, multiple cross-cutting flowsets can be identified, particularly near the margins of the Shield. Low-relief bedforms are superimposed onto older MSGL and are ornamented orthogonal to, and terminate at, arcuate till-cored moraine ridges (Ontario Geological Survey 2010; Fig. 5.7E,F). Cross-cutting of streamlined forms is restricted to the western parts of flow sets, along the flanks of the Algonquin Highlands, while areas at greater distance from the Shield, lying at lower elevations, show only parallel features tracing a curvilinear path around the flanks of the Algonquin Highlands (Figs 5.1, 5.7E,F).

5.5. INTERPRETATION

Inter-regional analyses of landform types, patterns, and distributions provide the necessary context to compare previous local-scale studies and to view landform trends on scales compatible with the scale of former ice sheets/streams (Stokes & Clark 1999; Margold *et al.* 2015). Some of the newly-observed landform relationships described here have implications for the reconstruction of ice dynamics around the periphery of former ice sheets that overrode similar lithological and topographic variations, introducing questions regarding the spatiotemporal variability of subglacial processes on multiple scales.

5.5.1 Bedform patterns

5.5.1.1. Shield terrain

Curvilinear tracts of parallel drumlins and MSGL along the flanks of the Algonquin Highlands (Figs 5.1, 5.3, 5.7) show clear evidence of topographic funnelling of the LIS (Stokes & Clark 2001; Bennett 2003) down the axes of the Ottawa and St Lawrence River valleys. The concept of fast ice flow/ice stream(s) funneled through the St. Lawrence River lowlands and into southern

Ontario has been proposed by others (Gadd 1980; Ross *et al.* 2006; Margold *et al.* 2015; Eyles & Doughty 2016; Sookhan *et al.* 2018a, b; Eyles *et al.* 2018) and the data presented here provide strong support for this hypothesis.

An unexpected observation in this study was the abundance and parallelism of MSGL across significant (>300 m) topographic relief on Shield terrain in the central and southern parts of the Algonquin Highlands (Figs 5.1, 5.3A). The mapped distribution of MSGL is consistent with descriptions made by previous workers in the region (e.g. Barnett 1983) and existing surficial geology maps (Ontario Geological Survey 2010). The Algonquin Highlands area appears as a prime candidate for an ice stream 'sticky spot' (Stokes *et al.* 2007), given the high bedrock topography (Figs 5.1, 5.2), thin sediment (common lack of till; Figs 5.2, 5.3), and consolidation and texture of the sparse till deposits reported from the area. Despite these factors that would commonly contribute to sluggish ice flow, the presence of MSGL suggests fast flowing ice across the highlands and a possible ice stream that was able to pass rapidly over the bedrock high and hard bed; this was likely facilitated by lubrication from basal melting over the largely impermeable substrate (Krabbendam *et al.* 2016). A significant thickness of ice would be required to pass over the 400 m relief of the Algonquin Highlands without deflection to allow the development of such parallel landforms (Figs 5.1, 5.3).

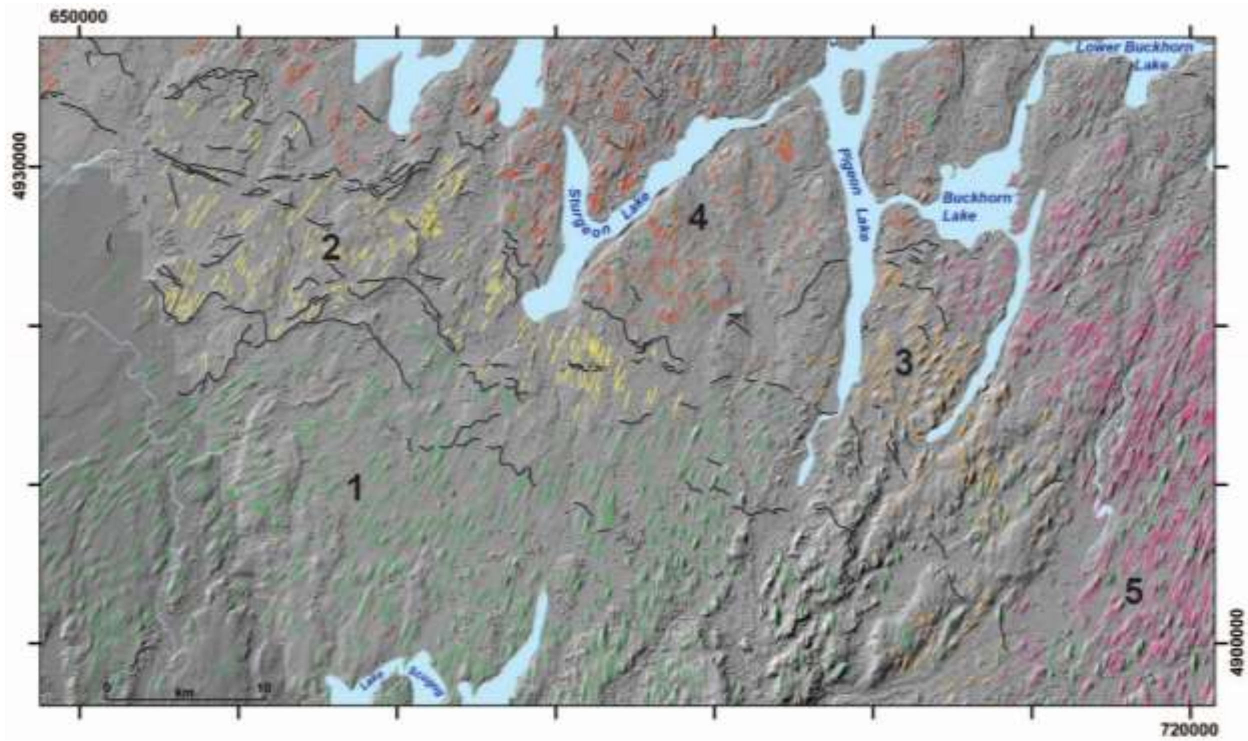
Eyles & Doughty (2016) described a transitional 'mixed bed' zone, where drumlins, flutes and rock drumlins are observed in thin drift terrain separating the 'hard bed' of the Shield and the 'soft bed' of thick sediments in the central parts of the PDF. The presence of drumlins and MSGL cored with till (and other stratified antecedent sediments), and their juxtaposition with local rock drumlins and streamlined bedrock escarpment noses, have previously been cited as evidence for an erosional origin for the bedforms across much of central Ontario (Figs 5.4, 5.5; Gravenor 1957; Shaw & Sharpe 1987; Clark et al. 2003; Englert *et al.* 2015; Eyles *et al.* 2016). The orientations of MSGL on the Algonquin Highlands and PDF drumlins are strongly coincident (Figs 5.1, 5.3A, 5.4-5.6), suggesting a landform continuum (and possibly synchronous development?) across the two areas. The occurrence of streamlined MSGL and megagrooves on bedrock and till within the Algonquin Highlands (Finamore & Courtney 1982; Krabbendam *et al.* 2016), suggests that streaming ice may have extended well upflow into shield terrain at various times during the Late Wisconsinan, prior to the onset of lateglacial ice streams occupying the Great Lakes basins (see Sookhan *et al.* 2018a, b for discussion).

5.5.1.2. PDF

The quasi-parallel sets of ridges spanning the PDF from north of Lake Simcoe to east of Rice Lake are interpreted as end moraines based on their orientation perpendicular to ice flow, their association with distinct local-scale bedform patterns, their relationship with ice-contact meltwater features such as eskers and subaquatic fans throughout the PDF (see below), and their mapped correlation with the location of the previously-mapped Lake Simcoe Moraine (Fig. 5.1; Taylor 1913; Deane 1950). These moraines are superposed on southwest-trending drumlins in the PDF, and mark the southwestern (downflow) limits of complex landform patterns interpreted to record multiple generations of superposed streamlined bedforms (Figs 5.4-5.7; see below).

From these relationships it is interpreted that an early, regional drumlinization phase occurred during the main phases of the Late Wisconsinan, which formed the large, broad drumlins north of Lake Ontario (and likely those in central New York State; Briner 2007; Kerr and Eyles 2007; Menzies *et al.* 2015). Subsequent phases of ice flow produced younger generations of streamlined bedforms that are superimposed onto, or entirely eroded/remoulded

Fig. 5.8: Relative age of flow sets within the central part of the PDF (modified from Marich 2016). Long axes of individual streamlined bedforms are colour coded by each flow set, from oldest (1; green) to youngest (5; pink). Note that flow set are typically separated by primarily till-cored ridges associated with the Lake Simcoe Moraines (black lines). Note the occurrence of older (green) bedforms within younger flow sets, suggesting a widespread drumlinization of the regional till sheet followed by local modification by small-scale readvances. See Figs 5.4-5.6 for details. The step-like change in grey tones on the shaded relief map arises from the intersection of the margins of the 2-m (eastern) and 5-m (western) elevation model boundaries.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Lower Buckhorn Lake
Sturgeon Lake
Pigeon Lake
Buckhorn Lake
2
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Lake Scugog
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720000

the older broad drumlins (Clark, 1993; Maclachlan and Eyles, 2013; Eyles and Doughty, 2016; Eyles *et al.*, 2018; Figs 5.4-5.10).

The large, broad shape of the drumlins across much of the PDF (Maclachlan & Eyles, 2013; Eyles and Doughty 2016) suggests formation under relatively slow, stable ice flow conditions (Sookhan *et al.* 2018a). Another possibility is that the large, broad drumlins represent immature forms and that, given additional time under stable ice flow conditions, would have developed into more elongate forms (Benediktsson *et al.* 2016). Support for this latter hypothesis arises from the observation that the majority of elongate forms occur either at significant distances upflow (e.g. Shield MSGL; Fig. 5.3), in topographic lows, where lateglacial

ice was preferentially routed (e.g. Ottawa Valley and Rice Lake Corridor bedforms; Fig. 5.7), or in areas that experienced multiple flow events (e.g. LODF and Ottawa-St. Lawrence River lowlands; Figs 5.7-5.10).

Close inspection of the dense group of bedforms that lies upflow of the Lake Simcoe Moraines in the Rice Lake Corridor (Figs 5.1B, 5.6) shows that older broad drumlins (similar to 'pristine' drumlins south and west of the Lake Simcoe Moraines; Figs 5.1B, 5.4, 5.5) are composed of NT or stratified sediments and preserved within the field of highly attenuated flutings and grooves. The drumlins are oriented slightly oblique to the main axis of the corridor, whereas the younger flutings show strong parallelism in line with the axis of the corridor. Together, these relationships suggest that the low-lying Rice Lake Corridor predates the regional drumlinization of NT, and that younger superimposed bedforms developed after ice thinned and the local topography exerted a stronger control on ice flow directions (Maclachlan & Eyles 2013). In the central part of the Rice Lake Corridor, the variety of streamlined bedform morphologies visible in close proximity (Fig. 5.6B), add further support to the idea of evolving streamlined bedform morphology from successive flow events (Clark 1993; Eyles *et al.* 2018). The older large drumlins acted as resistant obstacles to subsequent phases of ice flow. These obstacles then become seeding points for a new generation of bedforms, oriented parallel to the axis of the corridor, whose morphology is controlled by a combination of ice dynamics, subglacial hydrology, and bed/obstacle strength. The wide variety of features identified in the Rice Lake Corridor suggests a spectrum of bedforms is possible when ice flows over similar obstacles (equifinality). If the ice and/or deforming layer strength is greater than antecedent till, grooves may be ploughed/eroded into the substrate, truncating older drumlin forms (Clark *et al.* 2003; Eyles *et al.* 2016; Fig. 5.6B) but it is equally possible that if the older drumlin is less strong than the overriding ice, and oriented oblique to the later flow, the sides may be dragged/advected downflow, forming comma forms and parabolic drumlins (Boulton, 1987; Figs. 5.5-5.7), or creating multi-tailed drumlins (Clark 2010; Eyles *et al.* 2016; Fig. 5.6B). The morphological data presented here do not unequivocally support an erosional or depositional origin for streamlined bedforms, but suggest that local variations in bed characteristics likely play a role in bedform genesis (Evans *et al*. 2015). Regardless of their formative origin, the occurrence and arrangement of bedforms within the Rice Lake Corridor suggests that the complex drumlin forms (parabolic, asymmetric, comma-form, bisected) are not single bedforms, but palimpsest, polyphase features, whose genesis requires changing ice flow dynamics and/or directions (Clark 1993; Stokes & Clark 2001; Figs 5.8-5.10).

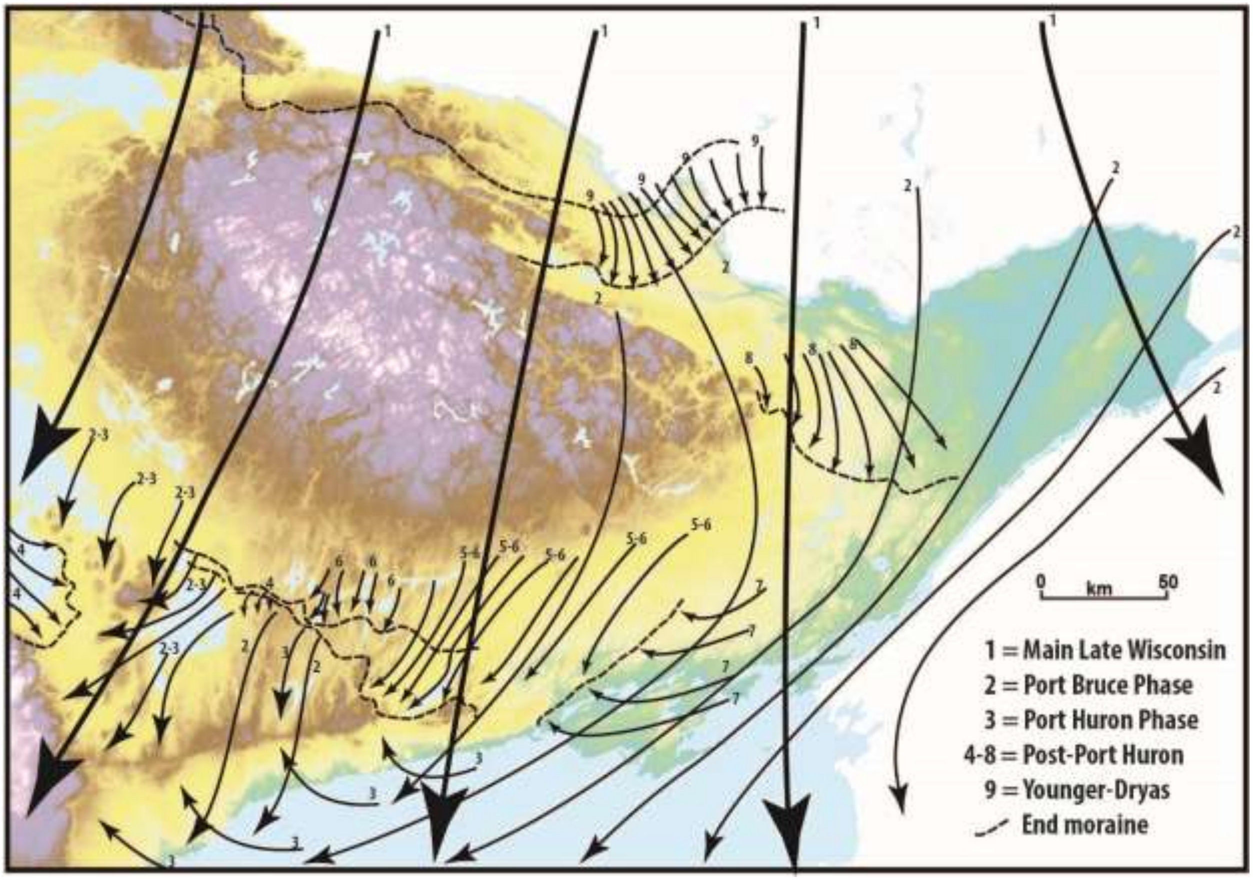
Crescentic scours around the upflow ends of some drumlins (Fig. 5.6B) may record meltwater scouring at the ice-bed interface after development of the drumlin form (Wellner *et al.* 2006), or erosion by a mobile layer of deforming till that sculpts the parent form (Eyles *et al.* 2016).

5.5.1.3. Cross-cutting flow sets

Multiple sub-regional, cross-cutting flow sets observed in the PDF and LODF attest to significant reorganization of the ice sheet into independent ice streams in central Ontario during deglaciation (Clark 1993; Maclachlan & Eyles 2013; Eyles & Doughty 2016; Sookhan *et al.* 2018a). The most extensive set of cross-cutting forms occurs within the LODF, along the entire length of the northern shore of Lake Ontario. Older PDF drumlins are ornamented with flutings, or are completely reshaped by ice that flowed NW out of the Lake Ontario basin (Fig. 5.7), likely during the development of the lateglacial Halton Ice Stream (Sookhan *et al.* 2018a) following shut down of a larger, SW-flowing ice stream (Ross *et al.* 2006; Eyles & Doughty, 2016) named the Ontario-Erie ice stream (Eyles *et al.* 2018). Comma-form and asymmetric drumlins can effectively be explained as pre-existing bedforms modified by a secondary ice flow phase consistent with fluting orientations within the LODF. The strongest evidence for this suggestion arises from the NW-oriented tails of apparent 'comma forms' that consistently emanate from the northeast end of stoss sides of former PDF drumlins (Fig. 5.7C, D). This records a shift from northeasterly to southeasterly ice flow during ice sheet reconfiguration/ ice stream flow switching and extends the previously described area affected by late glacial ice stream activity and flow switching further east within the Lake Ontario basin (Eyles & Doughty, 2016; Sookhan *et al.* 2018a).

West and northwest of Lake Ontario, younger HT and glaciolacustrine deposits blanket the drumlinized NT sheet and any PDF drumlins, complicating the analysis of ice flow from terrain data alone. Sedimentological studies from around the LODF support a change in ice flow directions, from the northeast to the northwest, either within the NT (Brookfield *et al.* 1982; Boyce *et al.* 1995; Boyce & Eyles 2000; Mahaney *et al.* 2014; T. Brennand, personal communication cited in Sharpe & Russell 2016), or between deposition of the NT and the HT,

Fig. 5.9: Generalized correlation of regional flow sets with major Late Wisconsinan events in south-central Ontario. Phase 1 shows regional sheet flow and drumlinization directions. The transition from phase 1 to phase 2 and higher is characterized by significant reorganization of the LIS and the onset of ice streaming and major flow switching and readvances in the region (see Ross *et al*. 2006; Eyles & Doughty 2016; Marich 2016; Mulligan *et al*. 2016; Sookhan *et al*. 2018). The distributions of landforms and resulting flow sets bear strong resemblance to the time-transgressive bedform pattern model of Stokes & Clark (2001: Fig. 8) showing multiple bedform populations within flow sets arising from several episodes of bedform generation. See Figs 5.4-5.8 for details.

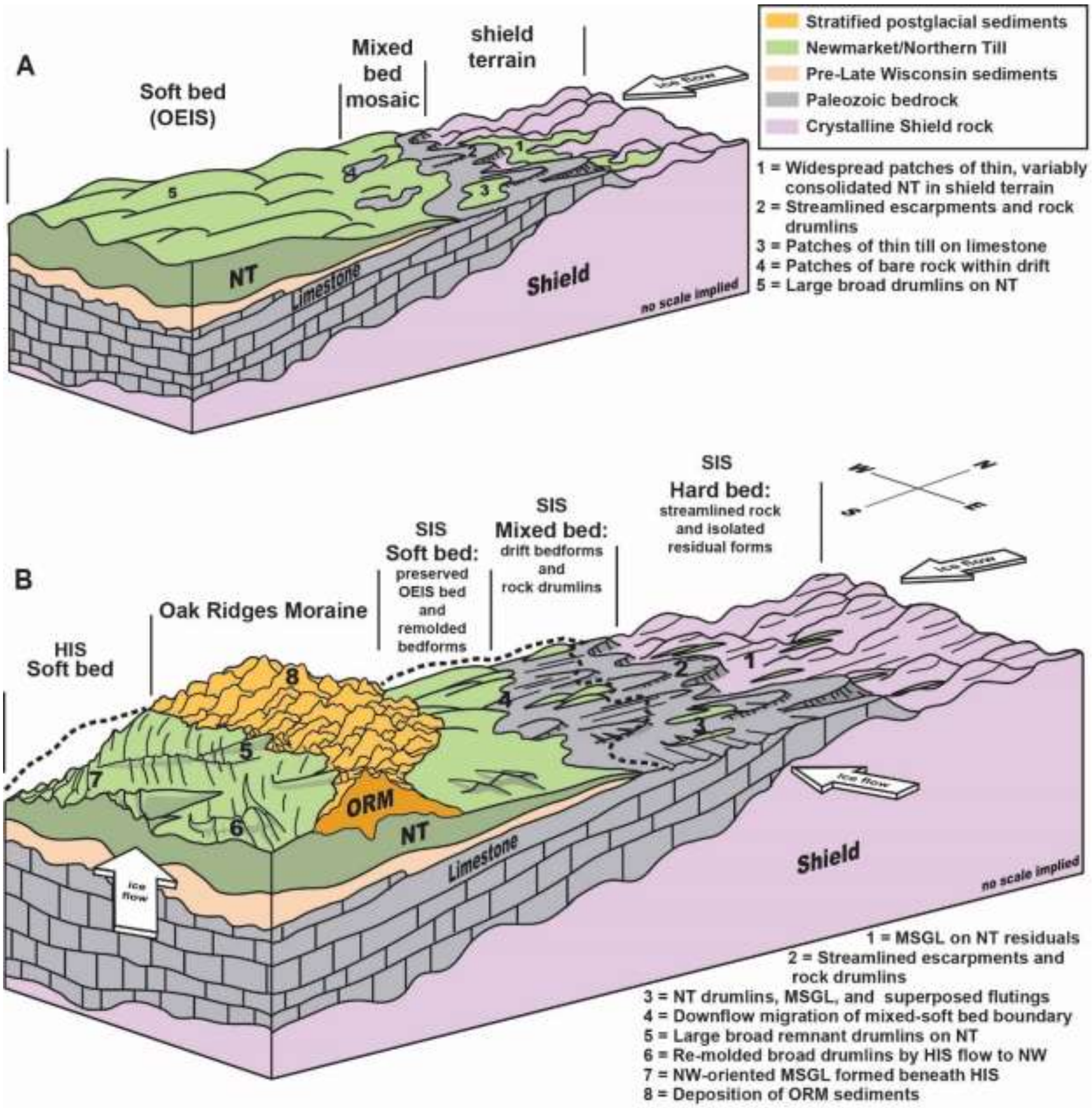

0 km 50
1 = Main Late Wisconsin
2 = Port Bruce Phase
3 = Port Huron Phase
4-8 = Post-Port Huron
9 = Younger-Dryas
End moraine

during a period of localized ice withdrawal (White 1975; Chapman and Putnam 1984; Sharpe & Barnett 1985; Karrow 2005; Slomka 2014). Areas where the ice re-advanced over lake sediments deposited during ice withdrawal are recorded by deposition (and streamlining) of fine-grained HT; areas that remained ice-covered during the flow switching display multiple bedform populations recorded on the surface of NT, as no new material was available for incorporation into a younger till sheet.

The change of ice flow directions between deposition of the NT and the HT is likely due to break-up of regionally southwest-flowing ice into smaller, independent ice lobes/streams around the axis of the ORM (Barnett 1992a; Sookhan *et al.* 2018a, b), which is underlain by a high on the subglacial bed (Howard *et al.* 1995; Sharpe *et al.* 2005; Kassenaar & Wexler 2006). Radial expansion of thicker ice along the axis of the Lake Ontario basin could be expected, producing the bedforms in the LODF consistently oriented perpendicular to the modern Lake Ontario shoreline (Eyles *et al.* 2011; Slomka 2014; Eyles & Doughty 2016; Burt & Mulligan 2017; Kozlowski *et al.* 2018; Sookhan *et al.* 2018a, b). Ice flowing along the deep Lake Ontario basin, characterized by fine-grained substrate sediments and a large ice-marginal lake, would have generated high porewater pressures and low effective stress at the IBI, facilitating fast (streaming?) ice flow velocities (Halton Ice Stream; Sookhan *et al.* 2018a). Studies at modern glaciers and ice streams have shown that flutings and drumlins can form under relatively thin (<200 m) ice lobes, in a period of months to years, respectively under surging/streaming (Boulton & Zatsepin 2006; Smith *et al.* 2007; Benediktsson *et al.* 2016; Dowling *et al*. 2016) or normal to sluggish ice flow velocities (Jónsson *et al.* 2016).

East of the ORM, the broad flow set emanating from the axis of the St. Lawrence River valley (Fig. 5.7A,B) records a later-stage radial expansion of the ice that remained within the

Fig. 5.10: Conceptual model of landscape development in the PDF showing (A) transition from hard bed with local patches of thin till on Shield terrain, into a mixed bed mosaic of bare limestone bedrock and increasingly thick patches of till southward to the soft bed underlain by thick successions of unconsolidated sediments. An early phase of till sheet drumlinization occurred beneath a broad ice stream (Ontario-Erie Ice Stream: OEIS) prior to (B) lateglacial flow switching initiating the break-up of the OEIS into smaller ice streams, the Halton Ice stream (HIS) in the south and the Simcoe Ice Stream (SIS) in the north. Flow switching promoted the development of new generations of highly elongate bedforms, and modification/removal of pre-existing bedforms. The Oak Ridges Moraine was deposited between the two ice streams (e.g. Sookhan *et al*. 2018a). Figure modified from Eyles & Doughty (2016).



A
Soft bed (OEIS)
Mixed bed mosaic
shield terrain
Stratified postglacial sediments
Newmarket/Northern Till
Pre-Late Wisconsin sediments
Paleozoic bedrock
Crystalline Shield rock
Ice flow
NT
Limestone
Shield
no scale implied
1 = Widespread patches of thin, variably consolidated NT in shield terrain
2 = Streamlined escarpments and rock drumlins
3 = Patches of thin till on limestone
4 = Patches of bare rock within drift
5 = Large broad drumlins on NT
B
HIS Soft bed
Oak Ridges Moraine
SIS Soft bed: preserved OEIS bed and remolded bedforms
SIS Mixed bed: drift bedforms and rock drumlins
SIS Hard bed: streamlined rock and isolated residual forms
W
N
S
E
Ice flow
ORM
NT
Limestone
Shield
no scale implied
1 = MSGL on NT residuals
2 = Streamlined escarpments and rock drumlins
3 = NT drumlins, MSGL, and superposed flutings
4 = Downflow migration of mixed-soft bed boundary
5 = Large broad remnant drumlins on NT
6 = Re-molded broad drumlins by HIS flow to NW
7 = NW-oriented MSGL formed beneath HIS
8 = Deposition of ORM sediments

lowland, in a similar manner to the advance/expansion of its predecessor, the Halton Ice Stream described above. Only a few small new and previously-identified moraine segments (Figs 5.7A,B, 5.9) mark the extent of the advance, which appears to coincide well with the feature previously identified as the Brighton Moraine by Chapman & Putnam (1984) and 'small moraines' described by Løken & Leahy (1964). The topographic step associated with the Brighton Moraine (Figs 5.1C, 5.7A,B) has alternatively been interpreted as a trim line marking the edge of former subglacial meltwater flood(s) flowing southwest through Lake Ontario and Erie basins and feeding into the Mississippi River valley (Lewis & Todd 2017). However, this view is inconsistent with the new terrain data, which shows the feature lying perpendicular to northwest-oriented drumlins and flutes cross-cutting older drumlins in the area and supports a morainal origin (Figs 5.1, 5.7, 5.9).

Similar examples of young flow sets exhibiting cross-cutting relationships with older bedforms exist in the Ottawa valley, and at local scales, within the PDF upflow of the Lake Simcoe Moraine(s) (Fig. 5.1C). Along the outermost lateral margins of the younger flow sets, older bedforms are ornamented with younger cross-cutting features, in intermediate positions strong modification of older bedforms is observed, and parallel or single bedform populations occur in the central and upflow areas (Figs 5.4, 5.5, 5.7-5.10). Similar bedform relationships occur locally across south-central Ontario (Mulligan *et al.* 2016) and bear striking resemblance to the conceptual model of time-transgressive landform development and preservation for terrestrially terminating ice streams (Stokes & Clark 2001: Fig. 5.8).

5.5.2 Landform relationships

The majority of individual late-stage flow sets of drumlins and superimposed bedforms terminate at, and are oriented orthogonal to, till-cored moraines (Figs 5.4-5.8). This consistent landform relationship assists in discriminating individual ice flow sets that record numerous discrete drumlinization/streamlining events and developing a clear landform-process model (Trenhaile 1975; Sookhan *et al.* 2016). Convergent and divergent patterns of bedform long axes strongly coincide with topographic lows, including gaps in bedrock escarpments and valleys cut into Pleistocene sediments (Figs 5.4, 5.5) that were previously interpreted as lateglacial tunnel channels (Brennand & Shaw 1994; Sharpe *et al.* 2004). However, sediment-hosted valleys in the western part of the valley network have recently been interpreted as tunnel valleys (*sensu stricto*; Clayton *et al.* 1999) that predate regional drumlinization of NT (Mulligan 2013; Mulligan *et al.* 2018a). The valleys within the study area, despite being smaller in scale, bear many morphological similarities to those observed to the west: they are commonly flanked by NT along their lateral margins and contain drumlins along their flanks and within central parts of valley axes. Further, the valleys show a strong association with late-stage bedform orientations, which commonly converge into the valleys in their northeast ends and diverge out from the valleys in their southwest ends (Figs 5.5, 5.6); associated flutings commonly terminate at subtle till-cored moraines. The valleys within the PDF are therefore interpreted here to have been excavated prior to (or contemporaneously with) regional drumlinization, and were occupied by active ice which experienced multiple cycles of retreat and minor re-advances/surges(?) throughout deglaciation (Fig. 5.5B; Mulligan et al., 2018a).

Many of the till-cored moraine ridges and associated bulbous subaqueous fans (composed primarily of sand and gravel deposits; Marich 2016) have a relatively smooth morphology and lack large hummocks or kettles (Figs 5.4B, 5.5B). This indicates the limited presence of buried ice blocks and there are no associated subaerial stream channels or distributary plains to suggest fluvial deposition (e.g. Lønne 1997). These features also lie at lower elevations than delta tops and shoreline features that have been mapped in the vicinity (Barnett *et al.* 1998; Mulligan *et al.* 2015). The bulbous bodies of sand and gravel are interpreted here as subaqueous fan deposits (Figs 5.4, 5.5) which lie down flow from continuous, well-preserved eskers and probably record R-channel development (Brennand 1994), and delivery of sediments to a subaqueous ice margin, prior to ice retreat.

5.6. DISCUSSION

As described above, the characteristics of regional streamlined bedforms, their occurrence within and along the flanks of valleys incised into sediments and bedrock, the abundant cross-cutting and divergent flow sets, and the termination of distinct flow sets at till-cored end moraines, point to a predominantly glacial influence on landscape development, involving the formation of valleys early during ice cover and active ice throughout deglaciation. The low-lying Rice Lake Corridor best exemplifies this, as multiple generations of drumlins are observed on the floor of the corridor, providing clear evidence of the existence of the low on the subglacial bed prior to early drumlinization (Figs. 5.6, 5.8, 5.9). Regionally, bedforms show strong evidence of the increasing influence of local bed topography as the ice sheet/stream thinned, leading to locally cross-cutting flow sets in upflow areas (Figs 5.4-5.9). The change in morphology and complexity of streamlined bedforms upflow of the previously-unidentified Lake Simcoe Moraine(s) is clear evidence of glacier ice as the landform shaping agent during repetitive re-advances/surges/streaming events throughout deglaciation.

5.6.1. Subglacial drainage and ice dynamics

The available data provide a means to investigate the evolution of subglacial dynamics beneath lateglacial ice lobes/streams to the north and south of the ORM. Drumlins and other sediment-cored bedforms in the region are part of coherent flow sets (Figs 5.8, 5.9) that are also characterized by striated and (mega-) grooved bedrock, streamlined bedrock escarpments and rock drumlins interpreted to record the erosion of the regional till plain and bedrock surface by glacier ice (Fig. 5.10). The switch from accretion of the NT to regional till sheet drumlinization may have been coincident with a shift from a distributed (low effective pressure) to a channelized (high effective pressure) drainage system (Walder & Fowler, 1994), which would produce more efficient and higher energy sediment delivery to the ice margins. This switch may have occurred in conjunction with thinning of the LIS, which would have facilitated meltwater delivery to the IBI.

Unlike the predominantly till-cored Lake Simcoe Moraines, the ORM is composed of predominantly stratified sand, silt and gravel, including thick successions of coarse-grained sediment deposited primarily by glaciofluvial systems at/near the ice margins (Barnett *et al.* 1998). The tunnel valleys that occur throughout the PDF and beneath the ORM (previously interpreted as tunnel channels; Brennand & Shaw, 1994; Sharpe *et al.* 2013, 2018) provided preferential pathways for meltwater and sediment transport from subglacial to sub-marginal and proglacial areas in the ORM (Brennand & Shaw 1994; Russell *et al.* 2003; Mulligan *et al.* 2018a). The valleys record the development of channelized subglacial drainage systems that began prior to regional bedform development and persisted throughout deglaciation, as indicated by the preferential occurrence of eskers within and along the flanks of valleys (Brennand, 1994; Brennand & Shaw 1994; Figs 5.4, 5.5), the incision of till within parts of some valleys, and partial infill of valleys with coarse-grained sediments, particularly in close proximity to the ORM (Sharpe *et al.* 2004, 2013, 2018; Marich 2016). The presence of large eskers within an integrated system across much of the PDF southward toward, but not always reaching the ORM records continued efficient channelized drainage systems operating at the ice bed throughout deglaciation.

Numerous studies demonstrate a link between meltwater abundance and the nature of subglacial drainage networks with ice dynamics (Parizek & Alley, 2004; Bartholomew *et al.* 2010). Increased meltwater availability due to climatic warming and fast ice flow rates in the region would favour a shift towards efficient channelized drainage towards the end of the last glacial cycle, providing an efficient way of evacuating sediments advected during bedform development to the ice margin (and ORM) through subglacial channels within tunnel valleys (N channels; Mulligan *et al.* 2018a) and eskers (R channels; e.g. Brennand 2000; Storrar *et al.* 2014). The importance of the tunnel valley network in this region is suggested by the presence of large eskers within an integrated system, occurring preferentially within valleys across much of the PDF southward to the ORM (Sharpe *et al.* 2004). Eskers record the persistence of efficient channelized drainage systems operating at the ice bed after development of the valleys. Although some eskers reach (and are subsequently buried by) the ORM at the southern limit of the PDF (Gadd 1980; Brennand and Shaw 1994), most do not, suggesting that much of the sediment comprising the bulk of the ORM predates deposition of the eskers.

5.6.2. Implications for ice sheet/stream dynamics

The study area encompasses a large region of the former bed of the southern part of the LIS, in an area previously interpreted to have been affected by ice streaming (Margold *et al.* 2015; Eyles & Doughty 2016; Eyles *et al.* 2018). The new terrain data provide a platform from which future sedimentological studies and quantitative geomorphic analyses can be launched.

Despite drastic, sharp changes in bed characteristics along former flow paths, the apparently seamless transition from MSGL on crystalline rocks on a large, isolated high in

Shield terrain (Algonquin Highlands; Fig. 5.1), to drumlins, flutings and rock drumlins on flat-lying bare Paleozoic or thin-drift terrain, into regionally well-organized drumlins resting on thick (>200 m locally), unconsolidated substrates (Fig. 5.2) points to the ability of streaming ice flow velocities to occur across a variety of topographic and lithologic boundaries. It remains unclear whether the parallel MSGL in the central Algonquin Highlands (Figs 5.1B, 5.3) formed contemporaneously with curvilinear paths running parallel to the structural lows along the Ottawa and St. Lawrence River.

Given that MSGL on the highlands tend to occur on isolated patches of till, with parallel streamlined bedrock forms observed in the immediate vicinity (Finamore & Courtney 1982; Fig. 5.3), an erosional origin is proposed for the streamlined bedforms (e.g. Sharpe *et al.* 2004; Eyles *et al.* 2016) and it is likely that the till sheet in the Algonquin Highlands region was once thicker and more extensive than it is at present. Similar observations have been made for the 'mixed bed' zone at the southern margin of the Shield (Eyles & Doughty 2016). The observations presented here suggest that the transition from hard bed conditions on the Canadian Shield, to mixed bed on Paleozoic sedimentary terrain is more of a patchwork or mosaic than a downflow gradation or a geographic division (Fig. 5.10).

The effects of changing subglacial drainage conditions on ice dynamics and landform genesis can also be investigated through this work. The initiation of regional tunnel valley development represents a shift from a distributed to a channelized drainage system (Kehew *et al.* 2012; Van der Vegt *et al.* 2012). Distributed and channelized drainage networks generally do not co-exist, as the pressure gradients created by the channels tends to sap meltwater from the higher pressure, but more sluggish, distributed systems (Walder & Fowler 1994). The increase in sediment strength due to dropping porewater pressure following entrenchment of the regional tunnel valley network could be a possible trigger for the switch from accretion to erosion and streamlining of NT. Conversely, the development of channelized drainage may have increased basal drag to slow basal sliding rates and inhibit erosion and streamlining.

The morphological characteristics and patterns of bedforms within the PDF point to a time-transgressive development of bedforms within the field, consistent with many areas along the periphery of the LIS (Colgan & Mickelson 1997; Sookhan *et al*. 2016; Kozlowski *et al*. 2018). An early, regional drumlinization phase occurred under thick ice flowing towards the SW (Figs 5.8-5.10). Later, as ice began to thin during deglaciation, local topography on the ice bed exerted stronger controls on ice flow and ice dynamics, documented by radial and cross-cutting flow sets within the PDF and remoulding of older drumlins and the common occurrence of northwest-oriented flutings in the LODF south of the ORM. The Lake Simcoe Moraines, now traceable for >80 km, form a significant morphological division within the PDF, separating 'pristine' drumlins to the south from variably eroded, re-sculpted and re-moulded forms to the north and east (Figs 5.8-5.10). Later phases of flow switching is identified by large-scale cross-cutting flow sets in the easternmost part of the PDF, across the LODF, and further up-ice in the Ottawa Valley (Fig. 5.9; Ross *et al.* 2006; Sookhan *et al.* 2018a, b). Tentative correlations of ice marginal positions are made to regionally dated events and shoreline morphological features from the Lake Huron, Lake Ontario and former Champlain Sea basins (Rayburn *et al.* 2011; Normandeau *et al.* 2017; Mulligan *et al.* 2018b; Fig. 5.9); future chronostratigraphic constraint will greatly assist in ice sheet reconstructions.

5.7. CONCLUSIONS

Through critical analysis of regional-scale terrain models in central and eastern Ontario, a large inventory of glacial landforms has been mapped. The distribution and pattern of subglacial bedforms, combined with their relationships with other glacial features across southern Ontario, supports recent interpretations of the former activity of ice streams in southern Ontario (Dyke & Prest 1987; Boyce & Eyles 1991; Clark & Stokes 2001; Eyles 2012; Maclachlan & Eyles 2013; Margold *et al.* 2015; Eyles & Doughty 2016; Eyles *et al.* 2018; Sookhan *et al.* 2018) and allows for assessment of the relative timing of major reorganizations of the LIS in the region. Juxtaposition of streamlined bedrock forms, drumlins, and MSGL provide support for past interpretations of an erosional origin for the majority of the bedforms in the region (Shaw & Sharpe 1987; Clark *et al*. 2003; Eyles *et al*. 2018). However, the variety of complex bedform morphologies observed within individual flow sets suggests local variations in substrate character may affect bedform genesis and morphology, and provide a means to guide future sedimentological investigations to test past theories of bedform generation. Parallel streamlined bedrock and MSGL observed in streamlined till residuals attest to the presence of active (fast-flowing?) ice overtopping the Algonquin Highlands, but tracts of drumlins and MSGL along the eastern flanks of the Algonquin Highlands point to strong topographic funnelling of flows during the lateglacial. It remains unclear whether these represent a contemporaneous drumlinized surface or multiple phases of bedform generation in the region. The ORM and numerous more subtle till-cored moraines, including newly-identified extensions of the Lake Simcoe Moraines mark the terminal extent of readvances or ice streaming events across the region that locally reshaped pre-existing landforms. Multiple local- and regional-scale cross-cutting flow sets occur in south-central Ontario, providing strong support to previous models of time-transgressive landform development during discrete episodes of landform genesis. Complex drumlin forms occur only in areas affected my multiple phases of ice flow, suggesting they represent a continuum of re-sculpted forms – detailed local analysis may provide insight into fundamental mechanisms of drumlin-formation. This work highlights the potential utility of newly-released (and publicly available) high-resolution terrain datasets that cover a key area where rapid changes in substrate material, sediment thickness, and subglacial bed topography played significant roles in governing the behaviour of the LIS as well as the development and evolution of palaeo-ice streams throughout deglaciation. South-central Ontario is an ideal area for future detailed sedimentological, and/or micromorphological studies to investigate the internal characteristics of these complex bedform sets, using multiple lines of evidence to better constrain their genesis and provide further insight into the evolution of local subglacial processes and timing of regionally significant events during the Late Wisconsinan.

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CHAPTER 6: CONCLUSIONS

6.1. INTRODUCTION

The objectives of this thesis were to provide new insights on the evolution of subglacial conditions and the processes of sediment erosion and deposition operating at the base of the former LIS in south-central Ontario. These were achieved through the interpretation of recently-collected high-quality geologic and geophysical data on the Quaternary sediment architecture, characteristics, and landform relationships across a large area of south-central Ontario where the bed of the Laurentide Ice Sheet is well-exposed and is characterized by complex topographic and lithological variations. The data constrain the nature of the genesis and infill of a regional-scale network of Late Wisconsin tunnel valleys (Chapters 2 and 3), the variability of sediment facies and subglacial conditions associated with the deposition of the regional Late Wisconsin till sheet (the Newmarket Till; NT; Chapter 4), and the time-transgressive nature of bedform genesis and landform relationships during significant reconfiguration of the LIS into ice streams following the last glacial maximum (Chapter 5). Together these data provide a means to test past theories on landscape evolution in south-central Ontario and to refine understanding of the links between the morphology and sedimentology of regional glacial landsystems with groundwater recharge and flow systems.

6.2. SUMMARY OF FINDINGS

Within the studied portion of the Late Wisconsin tunnel valley network that extends across south-central Ontario, a drumlinized surface extends from the flanks of upland regions into the subsurface beneath the valleys. This indicates that the valley forms largely pre-date the regional drumlinization of the till sheet and, based on the increasingly common observation of the NT within valleys, that valley development occurred prior to, and synchronous with, deposition of the NT. Cross-cutting morphological relationships indicate multiple phases of tunnel valley excavation in the region, some of which represent meltwater erosion within the valleys that post-dates deposition of the NT (e.g. Barnett, 1990; Barnett et al., 1998). The localized distribution and restricted thicknesses of coarse-grained valley in-fill sediments suggest limited accommodation space within the tunnel valleys and/or that the rate of ice retreat outpaced the rate of sediment delivery to the ice margin via eskers. The former notion is supported by the occurrence of eskers on the floors of the tunnel valleys, indicating they were filled with grounded ice during deglaciation. Recent suggestions that all the valleys within the network spanning south-central Ontario were simultaneously occupied by bankfull meltwater systems (Brennand and Shaw, 1994; Sharpe et al., 2002; 2004; 2013; 2018; Russell et al., 2003; Sharpe and Russell, 2016) is not supported by recent surficial mapping, sediment drilling, and geophysical investigations reported here. The tunnel valleys found across south-central Ontario are best explained as a product of repeated episodes of subglacial meltwater erosion, locally enhanced by groundwater piping, and direct subglacial erosion by active ice. This new understanding of regional sediment architecture and landform relationships presents an enhanced framework for the prediction of groundwater flow paths and the interactions between shallow and deep flow systems.

Through integration of detailed surficial mapping, sediment drilling, and geophysical investigations, tills observed flooring tunnel valleys that were previously interpreted as older glacial deposits (Gwyn, 1972; Sharpe et al., 2018) are now confidently correlated with the NT. Most of the previous detailed work on the NT comes from upland areas where sediment facies display consistency over large areas (e.g. Sharpe et al., 2002) and overlie thick successions (>100 m) of unconsolidated sediments consisting primarily of silt, clay and sand (Boyce et al., 1995; Barnett et al., 1998; Boyce and Eyles, 2000; Gerber et al., 2001; Meriano and Eyles, 2009; Crow et al., 2018; Sharpe et al., 2018). Comparative analysis of the NT on sediment-cored uplands, intervening lowlands and tunnel valleys, and along the Niagara Escarpment, provides detailed documentation of the wide range of possible facies and physical properties that can characterize the NT. These differences record spatial and temporal variations in subglacial depositional processes, which are inherited from local lithologic and topographic variations in the substrate. Documentation of differing styles of subglacial sediment deposition reveals the complex nature of the NT and indicates that it is not a simple, homogenized, stratigraphic marker bed, but a hybrid deposit composed of multiple texturally and lithologically diverse elements. The wide range of possible physical properties within the NT underscores the need for critical assessment of all available data (lithology, surface morphology, stratigraphic setting, paleoflow indicators, chronostratigraphic data, geochemistry, geophysical profiles) in developing reliable stratigraphic assessments of all diamict units in the region. The work presented here demonstrates the need for thorough understanding of the local setting of field sites in order to assess the significance of site-specific observations to regional-scale glacial processes and dynamics. It also highlights the wide range of sediment characteristics that exist within a single heterogeneous unit (the NT) that is of critical importance to regional and local aquifer recharge and vulnerability.

Improved characterization of the facies variability with the NT facilitates, and is augmented by, enhanced understanding of regional subglacial landscape evolution through morphological analyses of large parts of the till plain. The presence of active ice within the tunnel valleys is suggested by the characteristics of the sediment infills within the Simcoe County area (Chapters 2 and 3), but it is best exemplified by landform relationships observed in regional high-resolution digital elevation models (DEMs) that span the Peterborough drumlin field (PDF), where the bed of the former Laurentide Ice Sheet (LIS; recorded by the NT) is well exposed (Chapter 5). In this area, the tunnel valleys are generally narrower, shorter, and shallower than those that occur in Simcoe County, but share many geomorphic relationships, the most significant of which are the drumlinized flanks and the widespread occurrence of drumlins within the central parts of valleys. Recent surficial mapping by Marich (2016) identified segments of moraine ridges within the central parts of the PDF, which can be correlated from the north shore of Lake Simcoe to the east end of the ORM and their significance to the morphological properties of streamlined bedforms in the PDF can now be evaluated. South and west of the moraines, drumlins with simple planform morphologies predominate and no cross-cutting of landforms is observed. North and east of the moraines, several phases of younger ice flows are recorded by multiple cross-cutting lineations on bedforms. These younger bedforms typically show orientations that conform to local topographic lows (including tunnel valleys), and individual fields of younger bedforms terminate at predominantly till-cored end moraines. In a wide low-lying area (termed the 'Rice Lake Corridor'), significantly attenuated bedforms (mega-scale glacial lineations; MSGLs) are the most common feature, and numerous complex bedform morphologies are also observed (grooved, bisected, comma-form, asymmetric, parabolic, and multi-tailed/comet drumlins; Maclachlan and Eyles, 2013; Eyles et al., 2018). Together, these morphologic data shed light on the evolution and reconfiguration of the LIS during deglaciation. Significant shifts in ice flow directions occurred towards the end of the Late Wisconsin, increasingly controlled by subglacial topography as the ice sheet melted and thinned. Bedform characteristics within the PDF support previous suggestions that the onset and evolution of ice streams were important drivers in shaping the landscape of south-central Ontario (Maclachlan and Eyles, 2013; Margold et al., 2015; Eyles and Doughty, 2016; Eyles et al., 2016; Sookhan et al., 2018a, b; Eyles et al., 2018). Further, subglacial bedforms within the PDF developed time-transgressively, over a series of up to five distinct phases of active ice flow that persisted throughout deglaciation. The termination of bedform flow sets at till-cored end moraines is at odds with suggestions that these features are formed during regional-scale subglacial sheet floods (Shaw and Sharpe, 1987; Brennand and Shaw, 1994; Sharpe et al., 2004).

This new understanding of landform relationships, regional sediment architecture and properties in the vicinity of tunnel valleys has significant implications for predicting and modelling groundwater flow in these regions. The work completed in the 1990s in the ORM area was instrumental in documenting the truncation of the NT within the valleys (Barnett et al., 1998; Pugin et al., 1999; Sharpe et al., 2002; 2004), and their function as ‘hydraulic windows’ that significantly enhance leakance and recharge to buried aquifers (Desbarats et al., 2001; Sharpe et al., 2002). Analysis of the tunnel valleys within the primary study area (Chapters 2, 3 and 5) indicates that an even higher degree of architectural variability exists within tunnel valleys. Due to the potential for the NT to floor the entire valley, the limited distribution and thickness of coarse-grained valley infill sediments, and the thick overlying successions of glaciolacustrine silt-clay rhythmites, not all valleys may play a significant role in groundwater recharge. Perhaps more importantly, because of the wealth of new subsurface data available, it is now clear that the hydrogeologic setting and groundwater recharge potential within a single valley can change dramatically over just a few kilometres, based on localized erosion within the valleys, heterogeneous infill sediments, and local morphologic conditions that affect the rates and pathways of vertical groundwater flow from nearby uplands.

6.3. FUTURE WORK

Despite the wealth of recently collected high-quality data, additional local and regional-scale work is still required to adequately characterize the variability of, and fully understand the evolution of causative processes, that generated the regional stratigraphic framework of southern Ontario.

Data describing the setting and sedimentology of Late Wisconsin tunnel valleys in Simcoe County (Chapter 2) sheds a critical light on the relationships of major sediment groups in that particular area, but their correlation with complex systems to the north (Burt and Dodge, 2011; Mulligan 2016; 2017a; 2018) and south (Gerber et al., 2018; Sharpe et al., 2018) remains to be fully constrained, despite the local availability of high-quality data in those areas. Demonstrating the varying hydrogeologic properties of different parts of individual tunnel valleys illustrates a need for enhanced local-scale analyses of valley infills and their hydraulic properties for groundwater resource evaluations, work that is currently in its preliminary stages (Priebe et al., 2018).

The high-resolution sub-bottom profiles from Kempenfelt Bay (Chapter 3) enhance understanding of the architecture of the sediment infill of the tunnel valleys in south-central Ontario, and their relationship to modern(?) hydrogeology. The abundance of large mass-movement deposits, and the identification of widespread faulting within postglacial strata, suggest that seismic activity has significantly impacted valley fill sediments in this area (Todd et al. 2008; Doughty, 2014). Work on surficial exposures in the vicinity has yet to identify potentially coeval seismite beds; detailed logging of nearby stream cut banks, sediment cores, and additional collection of high-resolution sub-bottom profile data on a regular grid should be a focus of future study to determine the significance and/or frequency of recent seismic events.

The documentation of the heterogeneity of sediments comprising the NT should be viewed as the beginning of a new phase of research, rather than the end. Understanding the spatial complexity of a regional till sheet over various substrates can help to develop a better understanding of till rheological properties and till-forming processes, ultimately leading toward improved glaciological models in the future.

In addition to analysis of the internal sedimentological composition of the NT, regional-scale analysis of the morphology of the till sheet provides another lens with which to examine past ice dynamics (Chapter 5). Although a great deal of information on the relative timing of land-shaping events can be gained from analysis of DEMs which display the morphology of the NT in high-resolution, adding detailed, site-specific information from field sites across the former bed of the LIS will better elucidate the nature of bedform development in the region. Of particular interest for future analysis are the multitude of 'complex' drumlin morphologies that occur within the PDF. DEM analysis suggests these are not true variations of a drumlin form, but rather a hybrid landform developed under multiple phases of ice flow. Detailed (micro?)sedimentological and structural analysis of the NT and/or antecedent sediments may assist in providing new data to fuel discussion on the origin of drumlins. Though many of these bedforms have a predominantly erosional appearance (rock drumlins, mega grooves, MSGL, grooved/bisected drumlins), certain forms may record remobilization/remoulding of the original bedform. South-central Ontario presents a world-class field laboratory in which to investigate bed morphology and till sedimentology across a lithologically and topographically complex region around the periphery of a former ice sheet that underwent significant flow switching during deglaciation.

6.4. CONCLUSIONS

Late Wisconsin tunnel valleys form an integral part of the subglacial bed during evolving flow dynamics throughout the last glacial period. Local factors, including the lithology and topography of pre-Late Wisconsin sediments, provided a 'pre-design' for determining ice flow and subglacial meltwater routes which were drastically enhanced to form large (5 km wide, >30 km long, >150 m deep) tunnel valleys during successive periods of meltwater and subglacial erosion and deformation. Local substrate and topographic factors also promoted the development of a complex internal architecture within the NT that was deposited within tunnel valleys and across adjacent areas. The NT displays consistent facies and physical characteristics on sediment-cored uplands, but shows a high degree of local heterogeneity and vertical stacking of diverse sediments in tunnel valleys, below lowland plains, and along the Niagara Escarpment. Regional landform mapping also reveals details regarding the relative timing of events and provides preliminary constraints on reconfiguration of the LIS during deglaciation of southern Ontario.

Together, data from the chapters presented herein provide an integrated view of spatial and temporal changes in subglacial erosional and depositional processes that occurred across south-central Ontario during the Late Wisconsin. Evolving basal water pressures, meltwater drainage styles, and ice dynamics were critical factors controlling the development of the landscape in the region, but the nature of feedbacks that are initiated by fluctuations in these factors remains incompletely understood.

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APPENDIX

Introduction

Most of the data presented in this thesis was collected by government staff conducting fieldwork funded by provincial budgets and, as such, is available free of charge to the public. Not all the data has been quality-checked, field-checked, edited and/or formally released, but a list of published datasets and maps are provided below. Note that access to the files is subject to agreement with the terms of use for electronic information products from various government bodies (see terms of use on download pages).

Subsurface data:

South Simcoe Borehole logs, photos, PSA, geochemistry, radiocarbon ages, gamma and conductivity downhole logs
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=MRD324

Seismic Reflection data (land-based)
https://doi.org/10.4095/308388

Downhole geophysics
https://doi.org/10.4095/302783 - six central Simcoe boreholes
https://doi.org/10.4095/296884 - two south Simcoe boreholes

Central Simcoe borehole data
Contact author; data will be made available online at geologyontario.mndm.gov.on.ca upon completion of studies

Hydraulic conductivity of central Simcoe sediments
https://doi.org/10.4095/299048

pXRF chemostratigraphy south Simcoe
https://doi.org/10.4095/308450

Bedrock topography and sediment thickness maps
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=MRD207

Bedrock geology
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=MRD219

Surficial Geology (.pdf and ESRI .shp):

Alliston Area
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=P3768

Lindsay-Peterborough area
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=OFR6321

Collingwood area
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=P3815

Western half Barrie-Elmvale area
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=P3816

Penetanguishene-Christian Island
(release in late 2019)

Seamless surficial geology of southern Ontario
http://www.geologyontario.mndm.gov.on.ca/mndmaccess/mndm_dir.asp?type=pub&id=EDS014-REV

Terrain data

Download terrain data and/or high-resolution orthophotography using the search function in the Ontario Ministry of Natural Resources and Forestry Land Information Ontario metadata tool.
https://www.javacoeapp.lrc.gov.on.ca/geonetwork/srv/en/main.home

Dataset	***Search Keyword***
5-m DEM Greater Toronto Area	GTA DEM 2002, 2010
2-m DSM southwestern Ontario	SWOOP DEM
2-m DSM south-central Ontario	SCOOP DEM
2-m DSM Eastern Ontario	DRAPE DEM
2-m DSM central Ontario	COOP DEM
2-m DSM Algonquin Park	Algonquin DEM 2015
50-cm LiDAR Lake Erie	LiDAR Erie
50-cm LiDAR St. Lawrence Lowlands	LEAP 2009
50-cm LiDAR Peterborough County	LiDAR Peterborough

Radiocarbon ages

List of accelerator mass spectrometer (AMS) radiocarbon ages obtained from borehole and surficial samples collected as part of three-dimensional mapping investigations in Simcoe County. (Data from Bajc et al., 2014; Mulligan et al., 2018; Mulligan et al., in prep)

Lab ID	Borehole Number	Easting[1] (m)	Northing[1] (m)	Elevation[2] (m asl)	Depth[3] (m bgs)	Sample Material	Age[4] (^{14}C years BP)
A2053	SS-11-01	614442	4889524	267	55.9–56.1	*Dryas* leaves	41 800±600
A2443	SS-11-01	614442	4889524	267	56.1–56.2	*Dryas* leaves	44 400±1700
A2132	SS-11-01	614442	4889524	267	57.0–57.2	Wood	44 000±1800
A2054	SS-11-01	614442	4889524	267	80.8–81.1	Wood	>54 700
A2131	SS-11-01	614442	4889524	267	80.8–81.1	Wood	>50 800
A2055	SS-11-01	614442	4889524	267	120.3–120.5	Wood	48 800±1400
A2130	SS-11-01	614442	4889524	267	120.3–120.5	Wood	>47 500
A2133	SS-11-02	602975	4880338	256	70.25–71.10	Wood	>45 900
A2056	SS-11-02	602975	4880338	256	98.35–98.75	Wood	42 600±700
A2134	SS-11-02	602975	4880338	256	123.25–123.35	Wood	46 400±2500
A2057	SS-11-02	602975	4880338	256	142.5–142.65	Wood	50 100±1700
A2059	SS-11-04	586055	4878237	291	76.45	Wood	39 800±470
A2058	SS-11-04	586055	4878237	291	90.05–90.40	*Dryas* leaves	45 200±900
A2135	SS-11-04	586055	4878237	291	90.05–90.40	Wood	48 000±3000
A2445	SS-11-04	586055	4878237	291	90.05–90.40	*Dryas* leaves	43 900±1500
A2060	SS-11-06	594747	4898721	219	19.25–19.80	*Dryas* leaves	12 410±25
A2137	SS-11-06	594747	4898721	219	19.25–19.80	Wood	>49 500
A2446	SS-11-06	594747	4898721	219	19.25–19.80	*Dryas* leaves	12 455±40
A2061	SS-11-06	594747	4898721	219	69.2–69.3	Wood	54 800±3000
A2136	SS-11-06	594747	4898721	219	69.2–69.3	Wood	>49 200
A2062	SS-11-08	590082	4894188	264	63.65–64.35	Wood	49 200±1500
A2138	SS-11-08	590082	4894188	264	83.2–83.25	Succineidae sp.	>46 900
A2064	SS-11-08	590082	4894188	264	98.3–98.9	Wood	>54 000
A2139	SS-11-08	590082	4894188	264	105.8–106.1	Wood	>49 500
A2065	SS-11-09	577509	4903959	245	45.8–46.6	Wood	38 900±400
A2066	SS-11-09	577509	4903959	245	59.5–59.6	Wood	43 900±780
A2392	SS-12-02	602163	4902748	292	114.20–114.45	Wood	49 500±3100
A2447	SS-12-02	602163	4902748	292	114.45–114.75	*Dryas* leaves	45 400±1800
A2448	SS-12-03	593299	4906003	213	71.65–73.15	Wood	>51 700
A2393	SS-12-03	593299	4906003	213	72.00–72.85	Wood	>50 800
A2394	SS-12-03	593299	4906003	213	73.75–74.35	Wood	>51 700
A2395	SS-12-04	610758	4905514	286	50.40–50.95	*Dryas* leaves	28 840±240
A2449	SS-12-04	610758	4905514	286	50.40–50.95	*Dryas* leaves	28 060±230
A2450	SS-12-04	610758	4905514	286	53.3–53.8	*Dryas* leaves	30 540±300
A2451	SS-12-04	610758	4905514	286	53.3–53.8	*Dryas* leaves	>51 200
A2396	SS-12-04	610758	4905514	286	115.80–117.35	Wood	>48 100
A2452	SS-12-04	610758	4905514	286	115.80–117.35	*Dryas* leaves	48 700±2800
A2397	SS-12-04	610758	4905514	286	117.35–118.65	Wood	50 700±3600
A3329	N/A	576087	4925792	197	19	log	6895±25
A3328	N/A	576087	4925792	197	19	log	6370±25
A3330	N/A	576087	4925792	197	17	wood	6570±25
A3325	N/A	576196	4924966	193	10.5	log	6195±25
A3423	N/A	575820	4924476	191	6	wood	5915±15
A3422	N/A	575820	4924476	191	6	wood	5685±20
A3327	N/A	575697	4924585	201	9	wood	5000±20
UOC-3285	N/A	539788	4931199	261	10	wood	41349±429
UOC-3286	N/A	539788	4931199	261	10	wood	42054±509
UOC-3287	N/A	539788	4931199	261	10	wood	42016±521
UOC-3294	N/A	576244	4953638	189	1.5	wood	217±27
UOC-3295	N/A	576244	4953638	189	1.5	wood	172±27
UOC-3284	N/A	541454	4930183	229	4	wood	11776±40
UOC-3299	N/A	541454	4930183	229	4	wood	11863±34
UOC-3300	N/A	541454	4930183	229	4	wood	11828±34
UOC-3301	N/A	557610	4929842	180	1	wood	1450±27
UOC-3302	N/A	557610	4929842	180	1	wood	1197±27
UOC-3303	N/A	557348	4925397	228	2	wood	9531±30
UOC-3304	N/A	557348	4925397	228	2	wood	8912±38
UOC-3305	N/A	576087	4925792	197	3	charcoal	784±27
UOC-3279	CS-15-06	582692	4941869	226	102.95-104.05	wood	modern
UOC-3280	CS-15-06	582692	4941869	226	102.95-104.05	wood	modern

UOC-3281	CS-15-06	582692	4941869	226	102.95-104.05	wood	modern
UOC-3296	CS-15-08	586386	4908564	195	12.25-13.50	wood	10109±59
UOC-3297	CS-15-08	586386	4908564	195	12.25-13.50	wood	10093±48
UOC-3298	CS-15-08	586386	4908564	195	12.25-13.50	wood	10557±37
UOC-3288	CS-16-01	581428	4924041	196	21.75-22.70	wood	44655±1364
UOC-3289	CS-16-01	581428	4924041	196	37.70-37.75	wood	43355±1131
UOC-3290	CS-16-02	591694	4942461	291	130.45-130.5	wood	>50000
UOC-3291	CS-16-02	591694	4942461	291	133.15-133.9	wood	>50000
UOC-3292	CS-16-02	591694	4942461	291	134.3-134.4	wood	>50000
UOC-3293	CS-16-02	591694	4942461	291	138.45-138.6	wood	>50000
UOC-6104	CS-17-01	581934	4915757	209	2.25-3.0	leaf fragment	10204±59
UOC-6105	CS-17-01	581934	4915757	209	48.7-48.8	wood fragments	>44500
UOC-6106	CS-17-01	581934	4915757	209	51.4-51.5	leaf fragment	>44500
UOC-6107	CS-17-02	576813	4917116	214	47.1-47.2	leaf fragment	39424±333
UOC-6108	CS-17-02	576813	4917116	214	48.15-48.3	wood fragments	>44500
UOC-6109	CS-17-04	576804	4906589	242	24.2-24.3	leaf fragment	>44500
UOC-6110	CS-17-05	586945	4924564	212	32.75-33.0	leaf	>44500
UOC-6111	CS-17-05	586945	4924564	212	35.6-36.3	leaf fragments	>44500
UOC-6112	CS-17-05	586945	4924564	212	55.45-55.6	leaf fragments	>44500
UOC-6113	CS-17-05	586945	4924564	212	67.75-67.95	leaf fragments	>44500
UOC-6114	CS-17-05	586945	4924564	212	70.9-71.5	leaf fragments	>44500
UOC-6115	CS-17-05	586945	4924564	212	72.2-72.8	wood	>44500
UOC-6116	CS-17-06	562769	4922859	217	4.5-4.6	wood fragments	20262±1191
UOC-6117	CS-17-08	593510	4911632	215	14.5-14.7	leaf	>44500
UOC-6118	CS-17-08	593510	4911632	215	14.5-14.7	leaf	>44500
UOC-6119	CS-17-08	593510	4911632	215	16.3-16.5	leaf	>44500
UOC-6120	CS-17-08	593510	4911632	215	16.3-16.5	leaf	>44500
UOC-6121	CS-17-08	593510	4911632	215	16.3-16.5	wood fragments	>44500
UOC-6122	CS-17-08	593510	4911632	215	29.5-31.25	wood fragment	>44500
UOC-6123	CS-17-08	593510	4911632	215	46.75-47.25	leaf stem	>44500
UOC-6124	CS-17-08	593510	4911632	215	48.25-48.95	leaf stem	>44500
UOC-6125	CS-17-08	593510	4911632	215	48.25-48.95	twig	>44500
UOC-6126	CS-17-08	593510	4911632	215	51.75-52.0	wood fragment	>44500
UOC-6127	CS-17-08	593510	4911632	215	51.75-52.0	leaf	>44500
UOC-6128	CS-17-08	593510	4911632	215	52.25-53.3	*Salix* leaves	>44500
UOC-6129	CS-17-08	593510	4911632	215	63.4-64.05	leaf fragments	>44500
UOC-6130	CS-17-08	593510	4911632	215	71.75-72.35	leaf fragments	>44500
UOC-6131	CS-17-08	593510	4911632	215	72.5-72.75	*Dryas* leaf	>44500
UOC-6132	CS-17-08	593510	4911632	215	72.5-72.75	*Dryas* leaf	>44500
UOC-6133	CS-17-08	593510	4911632	215	72.5-72.75	twig	>44500
UOC-6134	CS-17-08	593510	4911632	215	72.9-73.1	twig	>44500
UOC-6135	CS-17-09	589677	4933424	220	39.25-39.5	leaf	30868±161
UOC-6136	CS-17-09	589677	4933424	220	40.25-41.1	twig	30263±155
UOC-6137	CS-17-09	589677	4933424	220	41.4-42.6	leaf	34308±218
UOC-6138	CS-17-09	589677	4933424	220	98.9-91.15	wood	>44500
UOC-6139	CS-17-09	589677	4933424	220	128.4-129.25	wood	>44500
UOC-6140	CS-17-10	588174	4937863	215	57.3-57.65	wood	>44500
UOC-6141	CS-17-12	583999	4948378	231	81.0-82.0	wood	>44500

[1]*Universal Transverse Mercator co-ordinates provided using North American Datum 1983 (NAD83), Zone 17.*
[2]*Elevation: metres above sea level (m asl).*
[3]*Depth: metres below ground surface (m bgs).*
[4]*Radiocarbon years before present (BP).*