11-4-2016 12:00 AM

Channel Form and Processes in a Formerly Glaciated Terrain

Nathaniel Bergman, *The University of Western Ontario*

Supervisor: Marco Van De Wiel, *The University of Western Ontario*
A thesis submitted in partial fulfillment of the requirements for the Doctor of Philosophy degree in Geography

Abstract

Despite that many places around the world in general, and North America in particular, were glaciated during the last ice age, relatively little is known about rivers that evolved over glaciated landscapes once they deglaciated. They are commonly categorized as alluvial with a glacial legacy, and often described as simple gravel-bed or sand-bed rivers. Alternatively, they are classified as bedrock rivers because the glacial deposits have been eroded and the underlying rock is exposed. However, the glacial history of these rivers is important and my research shows that classification for them is "semi-alluvial". This work shows that classification is important, not only for scientific accuracy but for river management that must be based on a sound understanding of river form and process. Poor understanding can be costly (i.e. restoration and management that do not achieve their goals; failed infrastructure), leading to wasted resources and inefficient functioning of the river.

Two major geomorphologic phases can be discerned in formerly glaciated terrains:

Deglaciation which exposed the landscape to erosion when ponded meltwater was abundant and led to outburst flooding. These catastrophic floods cannot occur under the modern climate of southern Ontario. Evidence for glacial lakes is found in lacustrine deposits upstream, and on top of moraines while spillways indicate where glacial lakes drained and carved deep valleys into the terrain. Spillways reveal outburst flooding with boulder lags, terraces, flow scars and possibly mounds and berms away from the modern misfit channel.

Postglacial incision and deposition during a single flood event or a single hydrological year. Human disturbance over the past two centuries, includes agriculture, channelization, millponds and weirs.

The dissertation addresses two parallel paradigms in fluvial geomorphology: Which processes are responsible for most of the geomorphic work that shapes the landscape - catastrophic flood(s) or small-scale erosion and deposition that bit by bit give the channel and valley its present morphology? My work will show that the outburst flooding of Glacial Lake London (paper 1) sets the stage for the postglacial morphology and small-scale processes we see today (paper 2). The third paper attempts to explain these small-scale processes using a 1D hydraulic model that can answer dynamic questions like bankfull discharge, water surface slopes of discharges, and velocity reversal hypothesis.

The findings show that the outlet of the Thames River near the neighborhood of Byron is a deeply-incised spillway channel that formed from the catastrophic drainage of Glacial Lake London during deglaciation. The dry lake bed serves today as the baselevel for the upper Thames River and its tributaries that incised and exposed the lacustrine and glacial sediments. The till exerts a strong control over channel form and resulting processes. Hydrological and sedimentary metrics often used to assess a river's condition produce contradictory results, doubting their validity without a-priory knowledge of what is the reference till river. With very little research done on semi-alluvial rivers in till, we propose that the scientific and engineering community focus on these rivers as they are quite abundant in N. America. Even so, evidence from till-bedded rivers and this study show they are quite different than alluvial and bedrock channels.

Co-Authorship Statement

This thesis contains three manuscripts. The first manuscript is entitled: "The drainage of the last Glacial Lake London, Ontario: Mapping and hydraulic modeling". This manuscript is co-authored with Marco Van De Wiel (supervisor), Steve Hicock (supervisor) and Yannick Rousseau (fellow PhD student). I conducted all fieldwork, data collection and analysis and wrote the paper under the supervision of Marco Van De Wiel and Steve Hicock who provided insight into the modeling and various glacial processes. Yannick Rousseau helped with the technical aspects of GIS work, assistance in Komoka Provincial Park and discussed aspects of the paper.

The second manuscript is entitled: “Semi-alluvial stream channel crossing the interlobate Arva Moraine, Medway Creek, southern Ontario, Canada. 1: Sedimentary characteristics and morphologic change”. This manuscript is co-authored with Marco Van De Wiel (supervisor) and Steve Hicock (supervisor). I conducted all fieldwork, data collection and analysis and wrote the paper under the supervision of Marco Van De Wiel and Steve Hicock who provided insight into the various glacial processes, and helped reviewing and editing the manuscript. Marco Van De Wiel provided funding. Steve Hicock also provided logistical support for sieving of river bed gravels and interpretation of the local tills and glacial deposits.

The third manuscript is entitled: "Semi-alluvial, till-bedded stream channel crossing an interlobate moraine in Southern Ontario, Canada. 2: Morphometry, hydrology and hydraulics of Medway Creek". This manuscript is co-authored with Marco Van De Wiel (supervisor) and Steve Hicock (supervisor). I conducted all fieldwork, data collection and analysis and wrote the paper under the supervision of Marco Van De Wiel and Steve Hicock who provided insight into the various modeling phases, and helped reviewing and editing of the manuscript.

I will be the first author on all three publications. All paper are planned to be published in the near future. The first in a special GSA Special Paper about paleofloods while the other two in Earth Surface Dynamics journal.

Dedication

For my parents, Hannah and Adam Bergman

My Dad Adam suddenly passed away during my 2nd year of the PhD in late 2011. He was the one that instilled the love of nature, outdoors and rivers in numerous hikes and road trips in Israel and North America during my childhood and teens.

Acknowledgments

First and foremost, I would like to express my sincere gratitude to Dr. Marco Van De Wiel and Dr. Steve Hicock who were a positive influence on this project when it seemed that it was not going to materialize after encountering significant academic difficulties. Marco always challenged me to find the best solutions to any field-encountered or modeling problems while Steve provided a perspective on glacial processes and always knew where to find the right source or person with an answer.

I'm indebted to Dr. Katrina Moser for her belief in me when progress was sluggish and pushing me forward, first as department graduate chair, later in many personal talks about lakes, rivers and climate in the department and as an integral part after joining my advisory committee. Dr. Adam Yates gave interesting perspectives about rivers I had not thought of and I also appreciate his time and effort as part of my advisory committee. Yannick Rousseau, my friend, office mate, field mate and expert technical advisor on GIS and modeling was instrumental in making large leaps in all stages of work. My external reviewers Dr. Joe Desloges and Dr. Guy Plint together with my internal reviewer Dr. James Voogt improved the work significantly after my defense. Dr. Bob Jarrett advised me on Manning's roughness and has been a huge support ever since I started my PhD. Dr. Jim O'Connor of the USGS Portland provided the data of moraine-dammed lakes. Dr. Dan Shrubsole, department chair, I thank you all for your support and advice over the years. Belinda Dodson is thanked for interesting conversations. To Dr. Peter Ashmore for initial ideas about Medway Creek.

Lori Johnson, the department's graduate administrator, was my rock in all matters, personal, academic and was my best advisor and support of how to tackle all matters in Canadiana. Caroline Majeau, Suzanne Dudzic, Barbara Thomas and Lelanya Milley helped me to handle endless bureaucracy that I'm terrible in dealing with. Erika Hill was helpful on many technical matters and lab analyses. Karen Van Kerkoerle drew and improved most figures. Joe Smrekar helped on various computing issues. Tim Marchand, Melody Cairns and Rhonda Card of Ontario Parks provided the permission to work in Komoka Provincial Park. Kimi Maruoka was a huge help on many technical issues and a great friend.

Finally, I am truly thankful for the love and support of my family along the years; Mum Hannah, Dad Adam, Eyal and Haim my brothers and their families.

Table of contents

Introduction

Brief glacial history of southern Ontario and its landforms

Southern Ontario is part of the gentler lowlands of the Great Lakes and St. Lawrence River, underlain by Paleozoic sedimentary rocks. The Laurentide Ice Sheet completely covered southern Ontario twenty thousand years ago at the last glacial maximum (LGM). This continental glacier complex covered most of Canada (80%) and extended into the northern states of the United States. The Wisconsinan glaciation began approximately 115,000 years BP and ended about 10,000 years BP. The Wisconsinan is divided into three major parts: Early, Middle and Late Wisconsinan. The Early Wisconsinan (115,000 - 60,000 years BP) marks the inception and growth of the Laurentide Ice Sheet coupled with a cooling climate. The Middle Wisconsinan (60,000 - 30,000 years BP) was warmer and southern Ontario was essentially ice-free. During the Late Wisconsinan (30,000 – 10,000 years BP) southern Ontario was ice-covered again with many advances and retreats of the ice lobes. Ice flow was controlled by the broad topographic depressions of the Great Lakes basins. Lobes of ice extending beyond the main body of the ice sheet developed in these basins and acted at times independently in response to local conditions at the base of the glacier rather than, or in addition to, climatic change. By 15,000 years BP the Laurentide Ice Sheet front started its final retreat from southern Ontario. While the Late Wisconsinan glaciation is the last time southern Ontario was completely glaciated, other earlier glaciations during the Quaternary Period as well as during the Precambrian covered the region but their geologic records are incomplete due to erosional processes as Ontario is located in the center of the North American landmass (Barnett, 1992).

The surficial geology of southern Ontario is composed predominantly of tills and unconsolidated rock debris deposited during the last glaciation when the Laurentide Ice Sheet occupied the region (Fig. 1). The tills are composed of unsorted mixture of clay, sand and gravel. The tills are spread over the landscape in a relatively flat terrain or sometimes molded into gentle hills called drumlins. Where the ice made temporary halts or re-advances, the till was pushed into long, rough ridges and sometimes covered with sand and gravel. These areas of rough topography are end moraines and are the most common high-relief landform of the region. The moraines are often breached by modern rivers and these breaches are either a result of a glacial spillway channel or post-glacial incision. The spillway channels carried drainage from the melting ice and today's modern rivers flowing through them are clearly smaller, making them underfit or misfit channels. Other glacial features of the region are long, narrow ridges of sand and gravel termed eskers and irregularly-shaped gravel hills called kames (Putnam and Kerr, 1966).

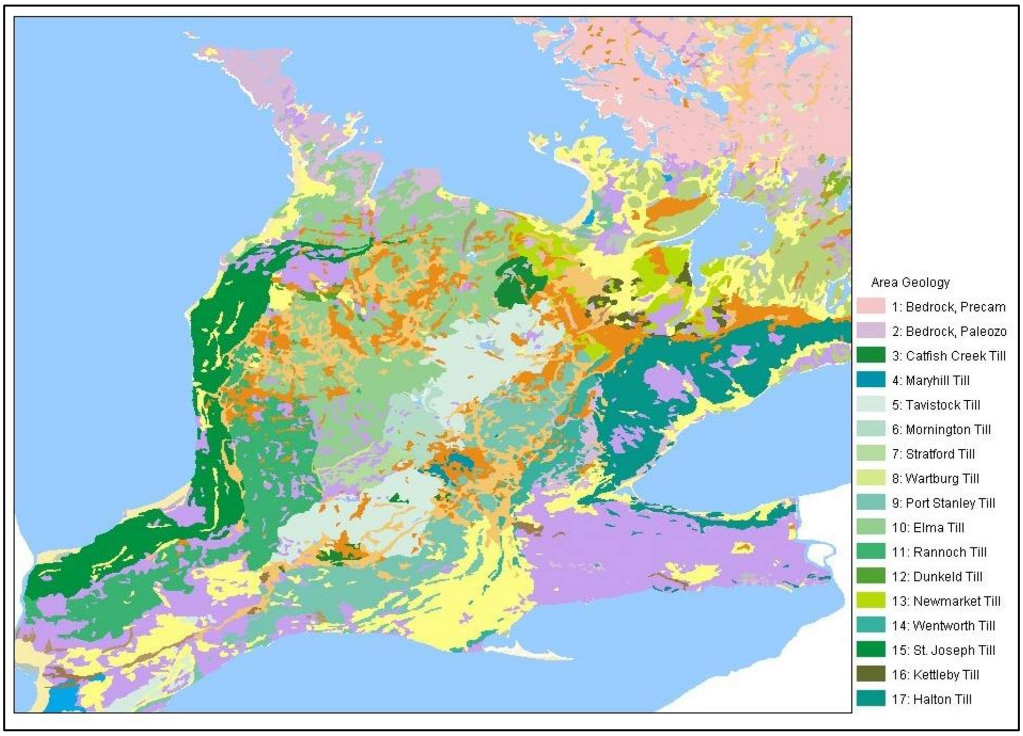


Figure 1. Surficial geology of southern Ontario showing the dominance of tills and some bedrock exposures (Ontario Geological Survey).

The Thames River, that is the study area of this dissertation together with its tributary Medway Creek, can be easily divided into two physiographic parts: 1. The upper watershed which is predominantly composed of till moraines, till plains and many dissecting spillway channels. 2. The lower watershed southwest of London all the way to Lake St. Clair outlet that is composed mainly of clay, sand and till plains (Fig. 2.). The transition point between the two watershed parts (around Delaware on the map) is a large spillway channel that drained the London Basin (through Glacial Lake London) and will be the focal point of the deglaciation process this work describes in the first manuscript. The marked differences in landforms are also expressed in the Thames River valleys' morphology as the upper watershed is generally deeply incised compared to the lower subdued basin (Fig. 3).

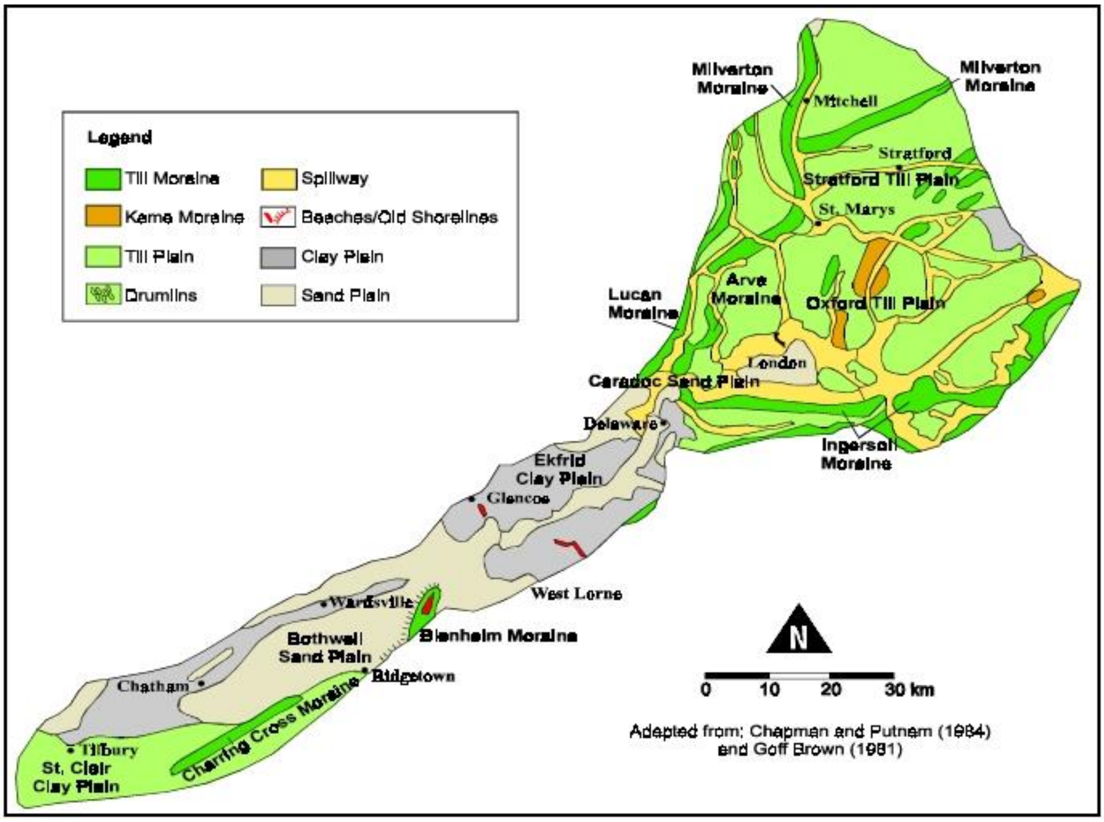


Figure 2. The Thames River watershed landforms (Upper Thames Conservation Authority, 1999).

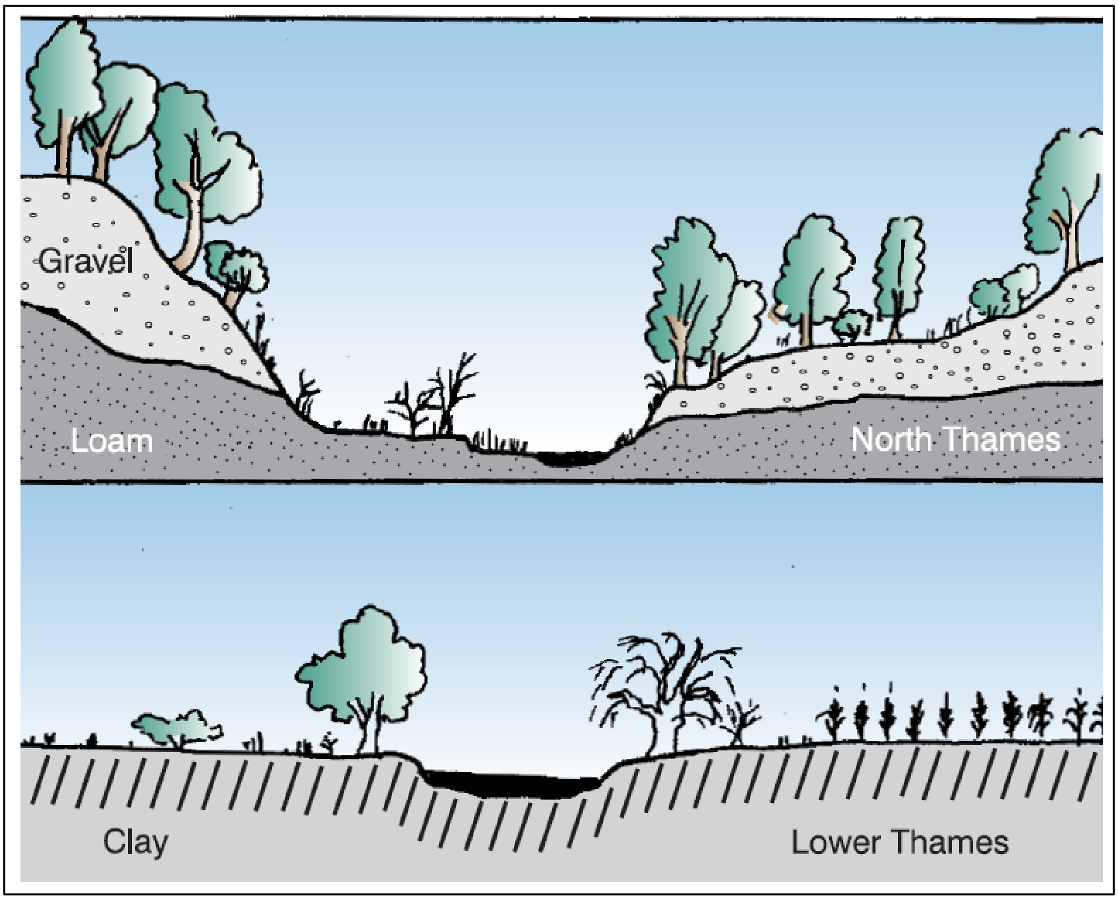


Figure 3. Typical cross-sections of the upper Thames River valley (top) and the lower Thames River valley (bottom) showing marked difference in topography (Upper Thames Conservation Authority, 1999).

While the tills were formed and deposited during glacial advance or retreat, the ice lobes generated considerable volumes of meltwater. There was a cyclical annual production of meltwater: summer release of water stored during the fall, winter and spring months but there was also a diurnal summer cycle release which was associated with rising day temperatures and cooler night temperatures. The diurnal cycles were less important in a large continental ice sheet like the Laurentide Ice Sheet as their water volumes were much smaller. The meltwater was either ponded in lakes close to the ice margins or flowed into streams and rivers draining away from the ice mass. The meltwater carried with them large quantities of sediments into the local rivers (glacio-fluvial outwash deposits) and lakes (glacio-lacustrine and lacustrine sediments) and these sediments overlie the tills.

History of catastrophic floods and their evidence

The field of paleoflood hydrology (i.e. ancient flood reconstruction) is strongly linked to an historic divergence between two opposing paradigms originating in 19th Century geology: uniformitarianism vs. catastrophism. Most scientists believed that processes in nature and associated landscape evolution are very slow (100s-1000s of years); erosional processes gradually remove sediments and denude the land into the present-day landscape and these processes are identical through time. Prominent geologists like Charles Lyell and James Hutton supported uniformitarianism and this also helped Charles Darwin articulate the revolutionary biological evolution of the origin of species in 1859. In contrast, scientists like Georges-Louis Leclerc, Comte de Buffon and later Georges Cuvier and Joseph Fourier argued that Earth underwent a series of short-term natural sudden catastrophic events (scale of a few minutes to days or months) such as violent volcanic eruptions, floods and other upheavals that shaped the landscape but these ideas were less common and often rejected (Baker, 1998).

These two paradigms clashed in fluvial geomorphology when J. Harlan Bretz suggested in the 1920s the Channeled Scabland of Eastern Washington was a result of a massive cataclysmic flood. The fact that Bretz had no water source to corroborate these findings and his discovery relied on peculiar landforms met intense criticism from the geologic community that was strongly entrenched in uniformitarianism ideas. Only when USGS scientist Joseph T. Pardee published his findings about Glacial Lake Missoula drainage in 1945, cataclysmic floods were accepted by the geologic community as an important geomorphic agent shaping the landscape almost instantaneously. Later, megaflood evidence was found in other breached lakes (i.e. Altai floods in Siberia, Bonneville Flood in Idaho, ancestral Great Lakes and Glacial Lake Agassiz) and other planetary bodies such as Mars (O'Connor et al., 2013).

What is the evidence of a catastrophic flood and what makes it so unusual compared to modern-day flood deposits or erosion? The first striking feature about an unusual flood is that it might leave behind a spillway morphology and misfit channel, (Kehew and Lord, 1986) that seem uncommon for the local river hydrology to create during ordinary floods, usually overbank flows that occur during the lifespan of humans (i.e. 10-100 years). Secondly, there are deposits or erosion scars high above the present-day river and it is quite evident that they originate from a fluvial process and not a hillslope or glacial process. Furthermore, if the river is gauged it is possible to associate modeled peak discharges with deposits/erosion remnants and check whether their elevations match. Paleoflood Geomorphologists divide these deposits/erosion remnants into two: Slack water deposits (SWDs; Patton et al., 1979) and Paleo stage indicators (PSIs; Jarrett, 1990). SWDs are sandy sediments that are in suspension during a large flood and deposit in dead zones or in localities which flow velocity decreases (i.e. caves, crevasses, tributary confluences and other irregular bank/river valley topography). PSIs are a variety of depositional/erosional features such as erosion scars, terraces, benches, boulder bars and boulder lags that mark a minimum level of the paleo-flow, usually far from the present-day channel and coupled with their large scale are ordinarily preserved for a long time after the flood occurrence (Costa, 1983). The hydraulic engineering community, which at first completely rejected paleoflood hydrology, adopted the methodology of the field and often uses these techniques to predict (hazard risk assessment) or reconstruct dam-break floods coupled with development of extensive parametric breaching equations that are based on physical characters of the reservoir stored behind the dam. Consequently, paleoflood hydrology is not only a mature scientific sub-discipline of geomorphology but also an applicable science (Baker, 2003).

Stream classifications

In order to define stream types, geomorphologists seek to characterize rivers usually according to their general morphology on a variety of scales. The classification allows assessing the channel morphology, organizing information about specific types as a framework, detecting change over time, comparing to other rivers and in many cases allowing determination of what a healthy or impaired river is (Kondolf et al., 2003). The earliest and simplest classification was suggested by Leopold et al. (1964) and that included straight, meandering, braided and anastomosing morphological patterns. Later classifications are more complex relating to the supply of sediment and bedforms. For example, the often-used Montgomery and Buffingtons' classification (1997) for mountain rivers relates to the substrate the channel is made of and the associated bedforms (Fig. 4).

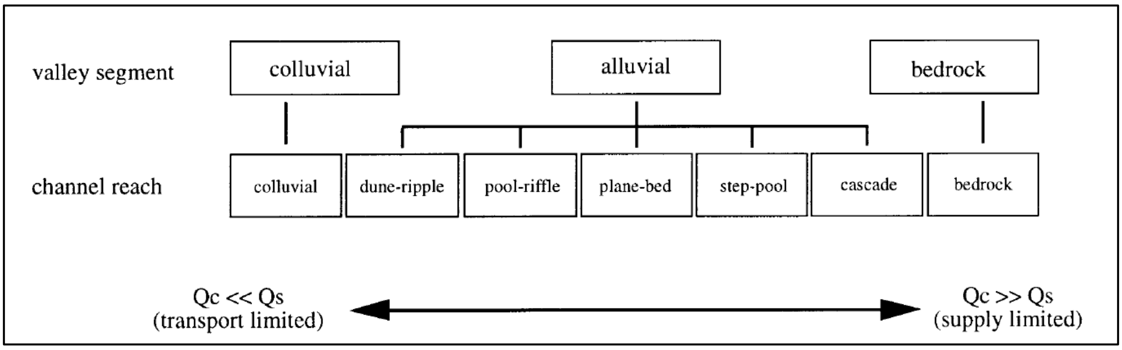


Figure 4. The Montgomery and Buffington (1997) channel classification for mountain rivers with relations to bedforms, transport capacities and sediment supply.

The main problems of most channel classifications are that they usually ignore processes and are thus very simplistic or that their developers worked in a specific region and consequently they cannot be generalized to other localities or environments. For example, the controversial Rosgen (1994, 1996) classification (Fig. 5; also termed Natural Channel Design - NCD), that is presenting very detailed channel morphologies of Western US rivers, receives heavy scrutiny from the academic community for its restoration failures (Simon et al., 2007) while the river practitioners community enthusiastically adopted this comprehensive classification scheme.

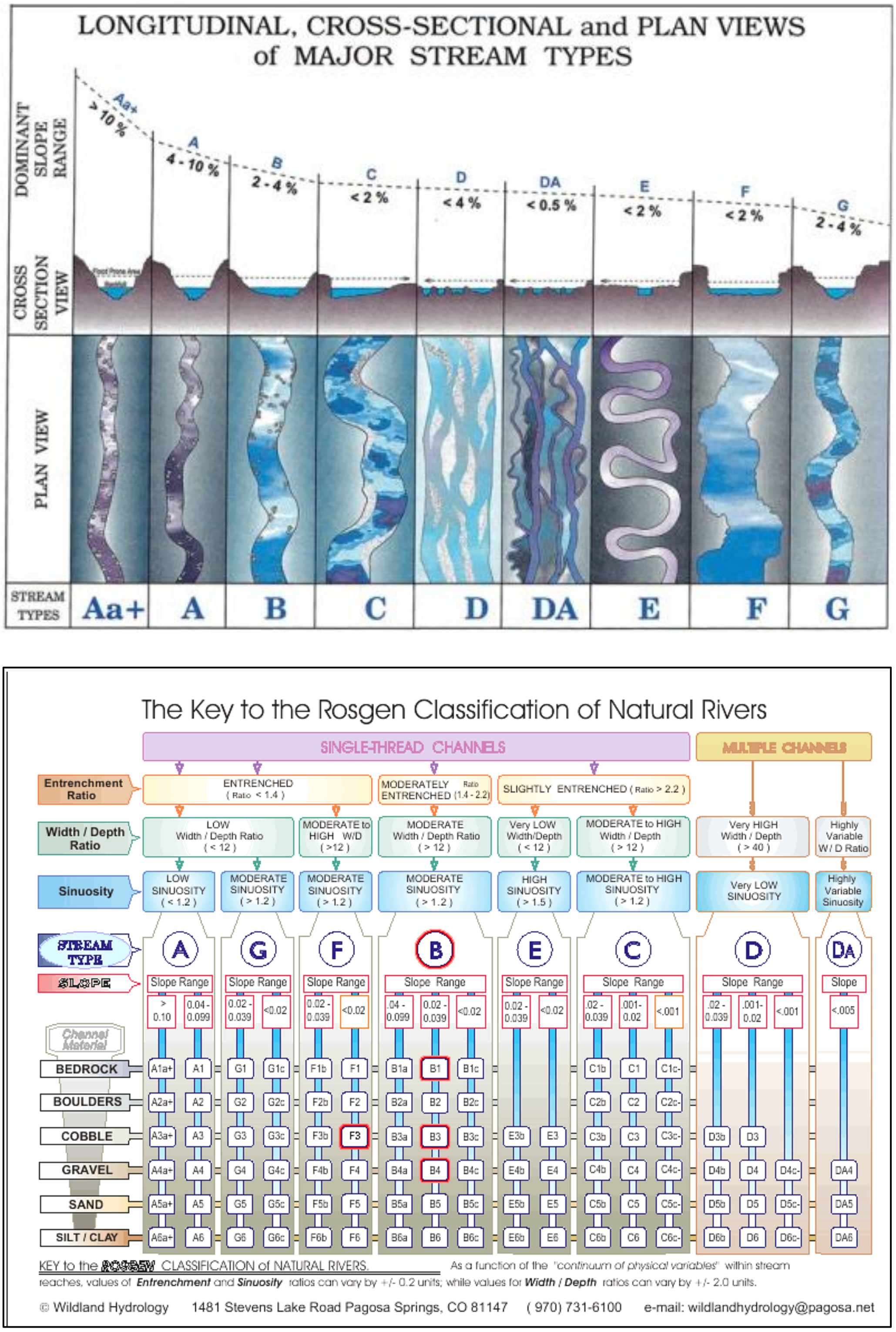


Figure 5. The complex Rosgen (1994, 1996) classification scheme of Natural Channel Design (NCD) often used in river restoration projects.

The failure to have an inclusive channel classification leads to an understanding that channels within a basin can be several types of rivers even at the same time according to changing topography, changes in geology and in the last couple of centuries land use changes, climate change and hydrology imposed by humans. Furthermore, all channel classifications published to date (2016) do not include semi-alluvial rivers. These rivers have some kind of geologic constraint (explained in manuscript two) and it would therefore be difficult to categorize them properly without a new category. For example, Martini (1977) and O'Connor et al. (1986) describe bedrock channels with well-developed alluvial bedforms. Looking back at Fig. 4 on the Montgomery and Buffington classification shows that bedrock rivers are not supposed to have bedforms in the first place and are expected to be devoid of alluvium - just plain bedrock. The second and third manuscript of this work show that a semi-alluvial river in a formerly glaciated terrain (Medway Creek) is a hybrid between alluvial rivers and bedrock rivers and together with the relatively little material published about this river type can be classified on its own. Phillips and Desloges (2014; 2015a) and Thayer et al. (2016) recently demonstrated the contemporary glacial conditioning of several southern Ontario till rivers and there were also recent attempts to classify them according to their floodplain (Phillips and Desloges, 2015b; Thayer and Ashmore, 2016). Our research adds to these studies by using a variety of hydrological and sedimentological indices often used in river studies both in academia and practicum of management and restoration.

Both the long-term instantaneous erosion (i.e. dam-break catastrophism during deglaciation) and contemporary short-term form and erosion processes (i.e. uniformitarianism) generate a continuum of a deglaciating landscape evolution and challenges common notions about how to view and manage these rivers.

References

Baker, V.R., 1998. Catastrophism and uniformitarianism: logical roots and current relevance in geology. In: Blundell, D.J. and Scott, A.C. (Eds.) Lyell: the Past is Key to the Present. Geological Society, London, Special Publications, 143, 171-182.

Baker, V.R., 2003. A bright future for old flows: origins, status and future of paleoflood hydrology. In: Thorndycraft, V.R. Benito, G. Barriendos, M. and Llasat, M.C. (Eds.) Palaeofloods, Historical Floods and Climatic Variability: Applications in Flood Risk Assessment, Proceedings of the PHEFRA Workshop, Barcelona, 16-19th October, 2002, Chapter 2, 13-18.

Barnett, P.J., 1992. Quaternary Geology of Ontario. In: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.) Geology of Ontario. Ontario Geological Survey, Special Volume 4 part 2, Chapter 21, 1011-1088.

Costa, J.E., 1983. Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range. Geological Society of America Bulletin 94, 986-1004.

Jarrett, R.D., 1990. Paleohydrologic techniques used to define the spatial occurrence of floods. Geomorphology 3, 181-195.

Kehew, A.E., Lord, M.L., 1986. Origin and large-scale erosional features of glacial-lake spillways in the northern Great Plains. Geological Society of America Bulletin 97, 162-177.

Kondolf, G.M., Montgomery, D.R., Piegay, H., Schmitt, L., 2003. Geomorphic classification of rivers and streams. In: Kondolf, G.M. Piegay, H. (Eds.) Tools in Fluvial Geomorphology, Wiley, pp. 171-204.

Leopold, L.B., Wolman, M.G., and Miller, P., 1964. Fluvial Processes in Geomorphology. W.H. Freeman and Company, 522 p.

Martini, I.P., 1977. Gravelly flood deposits of Irvine Creek, Ontario, Canada. Sedimentology 24, 603-622.

Montgomery, D.R., Buffington, J.M., 1997. Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin 109, 596-611.

O'Connor, J.E., Webb, R.H., Baker, V.R., 1986. Paleohydrology of pool-and-riffle pattern development: Boulder Creek, Utah. Geological Society of America Bulletin 97, 410-420.

O'Connor, J.E., Clague, J.J., Walder, J.S., Manville, V., Beebee, R.A., 2013. Outburst floods. In: Shroder, J. (Ed.) Treatise on Geomorphology. AP, San Diego, 475-510.

Patton, P.C., Baker, V.R., Kochel, R.C., 1979. Slack water deposits: a geomorphic technique for the interpretation of fluvial paleohydrology. In: Rhodes, D.D., Williams, G.P. (Eds.), Adjustments of the Fluvial System. Kendall/Hunt Publishing, Dubuque, IA, 225-253.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low-relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Phillips, R.T.J., Desloges, J.R., 2015a. Alluvial floodplain classification by multivariate clustering and discriminant analysis for low-relief glacially conditioned river catchments. Earth Surface Processes and Landforms 40, 756-770.

Phillips, R.T.J., Desloges, J.R., 2015b. Glacial legacy effects on river landforms of the southern Laurentian Great Lakes. Journal of Great Lakes Research 41, 951-964.

Putnam, D.F., Kerr, D.P., 1966. Physical Geography of Southern Ontario. In: A Regional Geography of Canada, J.M. Dent and Sons, Toronto, Chapter 8, pp. 185-200.

Rosgen, D.L., 1994. A classification of natural rivers. Catena 22, 169-199.

Rosgen, D.L., 1996. Applied River Morphology. Wildland Hydrology, Pagosa Springs, CO.

Simon, A., Doyle, M., Kondolf, M., Shields, F.D., Rhoads, B., McPhillips, M., 2007. Critical evaluation of how the Rosgen classification and associated "Natural Channel Design" methods fail to integrate and quantify fluvial processes and channel response. JAWRA Journal of the American Water Resources Association 43, 1117-1131.

Thayer, J.B., Ashmore, P., 2016. Floodplain morphology, sedimentology, and development processes of a partially alluvial channel, Geomorphology 269, 160-174.

Thayer, J.B., Phillips, R.T.J, Desloges, J.R., 2016. Downstream channel adjustment in a low-relief, glacially conditioned watershed. Geomorphology 262, 101-111.

Upper Thames Conservation Authority. 1999. Glaciers shape the land. Thames Topics, Booklet 3, Changing Environment, p. 1.

Background

Current erosion rates of 19 Ontario rivers (most of them in southern Ontario with drainage areas of 3-4000 km^2) are low: between 0.117 and 0.200 tons km^{-2} yr^{-1} (average of 0.15 tons km^{-2} yr^{-1}) (Church et al., 1999) suggesting contemporary fluvial stability. There are at least three mechanisms that could explain the formation of these valleys: 1. Gradual post-glacial incision where a new drainage network developed within the glacial deposits. 2. These valleys existed before glaciation with an existing drainage network, were completely filled, buried and covered by ice during the Laurentide Ice Sheet (LIS) and once they deglaciated and became exposed again, fluvial erosion resumed. 3. Some of the valleys are spillway channels either formed on the edge of the ice lobes as tunnel channels or are a result of catastrophic glacial lake drainage.

Phillips and Robert (2005) have associated river valley evolution in the Humber River (a tributary of Lake Ontario in the Toronto area) as post-glacial incision following the end of the last glaciation (15-13 Ka ^{14}C). Subsequently the Humber valley aggraded once its base level rose (~ 6.5 Ka BP) and vertical aggradation has been the dominant mode of floodplain formation in the Humber during the past 4000 years. Lateral channel migration was unlikely because rising base level would have reduced the amount of energy available for channel shifting (Weninger and McAndrews, 1989). For example, Arbogast et al. (2008), working in the Muskegon River in Michigan (a tributary of Lake Michigan that was deglaciated ~ 16Ka BP), found four post-glacial river terraces that they related to downcutting as a response to a reduction in flow magnitude (the flow geometry was more concentrated). In south-central New York, Scully and Arnold (1981) studied the Unadilla River and the Susquehanna River (deglaciated 14-12 Ka BP, Atlantic drainage) and found two distinct terraces. They linked the channel entrenchment to hydraulic effects or to an adjustment to an increase in channel slope caused by postglacial isostatic uplift of the region. Both examples are of till-bedded rivers.

A second mechanism of an incised valley's formation is inheritance of pre-glacial drainage. Several studies suggested that modern rivers of the Great Lakes are imposed on ancient bedrock rivers that are now buried under glacial sediments and are therefore 'tectonically predesigned' (Eyles et al., 1997; Lajeunesse, 2014). The Thames River (the river of this study and Medway Creek is one of its tributaries) does not exist in Grabau (1901), Spencer (1890; 1907), Karrow (1973), Flint and Lolcama (1986) and Gao (2011) studies that reconstructed the regional (bedrock) drainage of southern Ontario during glaciation, while today's modern Thames channel flows into Lake St. Clair. Spencer (1907) and later Flint and Lolcama (1986) located the London, Ontario, area between the pre-glacial Erigan River to the south and the Laurentian River to the north. According to Dreimanis et al. (1998), postglacial modern stream courses follow or are blocked by end moraines that determine their present paths. This supports the findings of Hack (1965) in Michigan that claimed the streams of Glacial Lake Duluth (now Lake Superior) followed existing glacial grooves that extended downslope towards the lake. In the much steeper bedrock terrains of the Swiss Alps, Montgomery and Korup (2010) suggested that despite repeated glaciations, the valleys are preserved and the topography does not start the incision again after each glaciation ends as it is persistent over time.

The third possible mechanism for the formation of incised river valleys is that they are catastrophic spillway channels (see Kehew et al., 2012 for a comprehensive review). Chapman and Putnam (1984) mapped many of the local rivers of southern Ontario as spillway channels, including both our study sites at Medway Creek and the

Thames River exiting London. As the LIS retreated it left numerous meltwater glacial lakes that then breached underneath the unconsolidated glacial material and ice sheets. The works of Shaw and Gilbert (1990) and Brennand and Shaw (1994) in southern Ontario support this mechanism of river valley creation. This interpretation of central southern Ontario tunnel channels is consistent with the view that the subglacial landsystem (drumlins, valleys and s-forms) was eroded by a regional meltwater underburst - the Algonquin event - that unsteadily evolved from sheet to a single vast channelized flow (e.g. Shaw and Gilbert, 1990). As fan deposits are not observed at the ends of channels it is likely that channel formation was contemporaneous with an underburst event that eroded drumlins in Lake Ontario and swept away most sediment derived from channel erosion. Ice sheet thinning and flattening associated with underbursts facilitated deglaciation by regional downwasting and stagnation (Brennand et al., 2007). This third mechanism of valley formation stands alone from the other two as it implies almost instant shaping of the topography and it brings up the argument that catastrophic flood events (cataclysms) dominate the landscape (Baker, 1977; Wolman and Gerson, 1978) over the smaller - medium events that do much less geomorphic work but are more frequent (Wolman and Miller, 1960).

According to Costello and Walker (1972), who worked on the sedimentology of the Credit River (Niagara Escarpment, Lake Ontario drainage), the post-glacial rivers of the area initially resembled braided outwash streams. As sediment amounts decreased and channels began to stabilize they achieved a single-thread channel form filling the outer valley and the braids that became their floodplain. This initial channel evolutionary stage fits the first two forming mechanisms described above but not the third that suggests almost instantaneous formation of the glacial valleys. Rivers incised into modern valley fill as a result of colonial milldam abandonment (Walter and Merritts, 1998; Merritts et al., 2011) are not comparable to southern Ontario and the American Midwest as these areas of the Mid-Atlantic (Pennsylvania) were not glaciated during the Wisconsinan. However, there is evidence of Holocene accretion of floodplains on the Grand River (Lake Erie drainage, see Walker et al., 1997) and Thames River (Stewart and Desloges, 2014) that were used by the native population for settlements and agriculture. From these previous studies it is clear that different perspectives exist on the evolution of the post-glacial fluvial landscape in southern Ontario. What is not clear, however, is how these different perspectives can be reconciled or how early post-glacial processes and contemporary processes combine in creating the contemporary fluvial geomorphology of southern Ontario.

This work will investigate the London Ontario area from various fluvial perspectives. The work is divided into three parts in a paper format. While each paper can stand alone, together they give a comprehensive picture of how rivers in the area evolved during two of the most dramatic geomorphic periods: 1. During the end of the ice age when the region first deglaciated and meltwater and remaining glacial sediments first interacted as fluvial processes (as opposed to earlier glacial processes); and 2. In recent times, under present-day fluvial processes following the British-European Colonial era that drastically modified the landscape by clearcutting of dense forests, draining wetlands, artificially channeling small tributaries and turning them into a vast agricultural and urban setting. The intervening period was less dramatic as braided outwash streams filled and turned into single-thread channels (Costello and Walker, 1972) and developed distinct buried soil horizons (Scully and Arnold, 1981).

Research questions

Specifically, this thesis seeks to answer the following research questions:

1. Do fluvial systems in the London area result from catastrophic floods or are today's frequent small to medium floods the main drivers of landscape evolution?
2. Do the geomorphic characteristics and dynamics of fluvial systems incised in tills differentiate them from alluvial rivers and bedrock rivers?
3. Are channel bed stability metrics originally developed for alluvial rivers suitable for use in till-bedded channels?

Thesis structure

The thesis is made up of three core papers. The first paper titled "The drainage of the last Glacial Lake London, Ontario: Mapping and hydraulic modeling" describes the London area during deglaciation, reconstructs Glacial Lake London minimum and maximum extents and its catastrophic drainage into what is today's Komoka Provincial Park and the Caradoc Delta using conventional paleoflood techniques. Paper 2 titled "Semi-alluvial till-bedded stream channel crossing the interlobate Arva Moraine, Medway Creek, Southern Ontario, Canada. 1: Sedimentary characteristics and morphologic change" shows that small rivers incised into till are different than ordinary alluvial rivers and resemble more soft bedrock-controlled channels with differing spatial and temporal bed erosion rates. A thin alluvial cover with varying sedimentology overlies the till within unorganized bedforms. The third paper titled "Semi-alluvial till-bedded stream channel crossing an interlobate moraine in Southern Ontario, Canada. 2: Morphometry, hydrology and hydraulics of Medway Creek" further explores the channel, floodplain and valley attributes coupled to the typical low, medium to high flow hydrology and simulated by 1D hydraulic modeling. Channel morphometry and hydrologic characteristics are compared to similar semi-alluvial rivers as well as to rivers with alluvial and bedrock channels. They illustrate that current classification, understanding, management and restoration of these rivers is lacking or flawed and that more research is needed to establish a reliable body of knowledge about them.

Research on semi-alluvial rivers in general, and a till-bedded channel in particular, goes beyond the scientific interest of river studies: it has social and economic implications as these types of terrains continue to be exposed under a rapidly warming climate in glaciated regions. These implications include active management and restoration of these rivers and placing housing development and infrastructure on their channel/floodplain/river valley.

While the work is area-specific, its inferences apply to many areas in North America that were affected by ice lobes (such as the US Midwest, US North Atlantic states, Great Lakes, Northern Great Plains) and other places around the world that deglaciated when the ice age ended, but also areas that are currently deglaciating due to anthropogenic climate change. The work is comparative and has examples from similar studies and other river types to show that these fluvial settings with glacial legacies need to be separated and distinguished from the classic alluvial-bedrock channel classifications that either ignore them or view them as alluvial or bedrock channels with geologic histories that can be dismissed or mentioned briefly.

References

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-central Lower Michigan, U.S.A. Geomorphology 102, 615-623.

Baker, V.R., 1977. Stream channel response to floods with examples from central Texas. Geological Society of America Bulletin 88, 1057-1071.

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

Brennand, T.A., Russell, H.A.J., Sharpe, D.R., 2007. Tunnel channel character and evolution in central southern Ontario. In: Knight, P.G., (Ed.) Glacier Science and Environmental Change. Blackwell Publishing, Malden, MA, USA, Chapter 6.

Campo, S.H., Desloges, J.R., 1994. Sediment yield conditioned by glaciation in a rural agricultural basin of southern Ontario, Canada. Physical Geography 15, 495-515.

Chapman, L.J., Putnam, D.F., 1984. Physiography of Southern Ontario. University of Toronto Press, Toronto, Ontario, Canada.

Church, M., Hamm, D., Hassan, M., Slaymaker, O., 1999. Fluvial clastic sediment yield in Canada: scaled analysis. Canadian Journal of Earth Sciences 36, 1267-1280.

Costello, W.R., Walker, R.G., 1972. Pleistocene sedimentology, Credit River, Southern Ontario: A new component of the braided river model. Journal of Sedimentary Petrology 42, 389-400.

Dreimanis, A., Winder, C.G., Aaltonen, R.A., 1998. London Ontario: Geology, geomorphology, geodata. In: Karrow, P.F. and White, O.L. (eds.) Urban Geology of Canadian Cities. Geological Society of Canada Special Paper 42, 237-260.

Eyles, N., Arnaude, E., Scheidegger, A.E., Eyles, C.H., 1997. Bedrock jointing and geomorphology in southwestern Ontario: an example of tectonic predesign. Geomorphology 19, 17-34.

Flint, J.J., Lolcama, J., 1986. Buried ancestral drainage between Lakes Erie and Ontario. Geological Society of America Bulletin 97, 75-84.

Foster, G.E., 1998. Morphology of three semi-alluvial stream channels in southern Ontario. Unpublished Masters Thesis. The University of Western Ontario, 215 pp.

Gao, C., 2011. Buried bedrock valleys and glacial and subglacial meltwater erosion in southern Ontario, Canada. Canadian Journal of Earth Sciences 48, 801-818.

Grabau, A.W., 1901. Guide to the geology and paleontology of Niagara Falls and vicinity. N.Y. State Museum Bulletin 45, 1-284.

Hack, J.T., 1965. Postglacial drainage evolution and stream geometry in the Ontonagon area, Michigan. USGS Professional Paper 504-B.

Karrow, P.F., 1973. Bedrock topography in southwestern Ontario: a progress report. Geological Society of Canada 25, 67-77.

Kehew, A.E., Piotrowski, J.A., Jørgensen, F., 2012. Tunnel valleys: Concepts and controversies - A review. Earth-Science Reviews 113, 33-58.

Lajeunesse, P., 2014. Buried preglacial fluvial gorges and valleys preserved through Quaternary glaciations beneath the eastern Laurentide Ice Sheet. Geological Society of America Bulletin 126, 447-458.

Merritts, D., and 27 authors., 2011. Anthropocene streams and base-level controls from historic dams in the unglaciated mid-Atlantic region, USA. Philosophical Transactions of the Royal Society A 369, 976-1009.

Montgomery, D.R., Korup, O., 2010. Preservation of inner gorges through Alpine glaciations. Nature Geoscience 4, 62-67.

Phillips, R.T.J., Robert, A., 2005. River incision in relation to post-glacial events in the Humber River basin, Ontario. Géographie physique et Quaternaire 59, 17-30.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low-relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Scully, R.W., Arnold, R.W., 1981. Holocene alluvial stratigraphy in the upper Susquehanna River basin, New York. Quaternary Research 15, 327-344.

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

Spencer, J.W., 1890. Origin of the basins of the Great Lakes of America. Quaternary Journal Geological Society of London 46, 523-533.

Spencer, J.W., 1907. Falls of the Niagara: Their evolution and varying relations to the Great Lakes; characteristics of the power and effects of its diversion: Canada. Geological Survey of Canada Publication 970, 490 p.

Stewart, A.M., Desloges, J.R., 2014. A 9000-year record of vertical and lateral accretion on the floodplain of the lower Thames River, southwestern Ontario, Canada, and implications for archaeological research. Quaternary Research 342, 161-172.

Walker, I.J., Desloges, J.R., Crawford, G.W., Smith, D.G., 1997. Floodplain formation processes and archaeological implications at the Grand Banks Site, lower Grand River, southern, Ontario. Geoarchaeology: An International Journal 12, 865-887.

Walter, R.C., Merritts, D.J., 2008. Natural streams and the legacy of water-powered mills. Science 319, 299-304.

Weninger, J.M., McAndrews, J.H., 1989. Late Holocene aggradation in the lower Humber River valley, Toronto, Ontario. Canadian Journal of Earth Sciences 26, 1842-1849.

Wolman, M.G., Gerson, R., 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surface Processes and Landforms 3, 189-208.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54-57.

Drainage of the last Glacial Lake London, Ontario: Mapping and hydraulic modeling

Nathaniel Bergman[1, 2, 3], Marco J. Van De Wiel[1, 4], Stephen R. Hicock[5], Yannick Y. Rousseau[6]

1. Department of Geography, Western University, Canada
2. Yigal Alon Kinneret Limnological Laboratory, Israel Oceanographic and Limnological Research, Israel
3. Department of Geography and Environmental Studies, University of Haifa, Israel
4. Centre for Agroecology, Water and Resilience, Coventry University, United Kingdom
5. Department of Earth Sciences, Western University, Canada
6. Department of Geography, Planning and Environment, Concordia University, Canada

Abstract

Glacial Lake London existed when southern Ontario was deglaciated during Glacial Lake Maumee III or Glacial Lake Whittlesey phases - the early ancestral phases of modern Lake Erie. Whether one or several lakes, it was an ice-marginal Laurentide Ice Sheet meltwater lake which was dammed behind convergent Arva and Ingersoll Moraines and was breached in its southwestern part to generate a catastrophic flood. Evidence for this high magnitude flood is a large v-shaped spillway channel downstream of the lake outlet, where the modern Thames River flows, and which is an underfit channel under the current hydro-climatic conditions. This study reconstructs paleolake surface area, depths, and volumes for Glacial Lake London based on contemporary topography, lacustrine stratigraphy found in Medway Creek's Arva Moraine bluffs, and the upper lake level topographic constraint prior to overtopping.

The resulting flood that drained the lake is also reconstructed using two different approaches: 1) applying a variety of published parametric breaching equations to find the peak discharge (13,000-241,000 m^3/s) based on breach geometric characteristics and reconstructed maximum lake volume (9.1 km^3); and 2) field mapping of a series of unusual geomorphic features found downstream of the Lake London outlet and attributed to the dam-break flood. These features are used for HEC-RAS step-backwater modeling of water surface slope profiles to calculate peak discharge. Reconstructed peak flood discharges (13,400-75,600 m^3/s) are in a similar range as other dam-break floods from relatively small ice-marginal lakes of the Laurentide Ice Sheet and 20 montane moraine-dammed lake outbursts. Our results confirm that contemporary hydrology could not have formed the Thames River valley downstream of the glacial lake outlet and that its formation required extreme discharges one or two orders of magnitude greater than historical maximum gauged floods (~1,500 m^3/s).

The study shows that outburst floods and resulting spillway channels originating from moraine-dammed lakes are not only typical of ice-marginal lakes in steep montane environments, or proglacial lakes in front of a glacier, but also occur in subdued deglaciating terrains of a receding ice sheet. Studies of these landforms can better determine whether local or regional landscape evolution has resulted from catastrophic events or gradual postglacial erosion and incision of the landscape. Furthermore, these dramatic events coupled with local deglaciating paleoclimate can inhibit temporary development of flora, fauna and human settlement of the region following the transition from a harsh cold ecosystem to a warmer habitat.

Key words: Southern Ontario; Glacial Lake London; outburst moraine dam-break flood; peak discharge reconstruction; parametric breach equations; step-backwater modeling

1. Introduction

The Thames River outside the city of London, southern Ontario, Canada (Fig. 1) flows through a deep river valley or canyon called "The Trench" by early French explorers and settlers ("La Tranchée"; Historica Dominion Institute Canada, 2015). Earlier studies concluded that this river valley is a spillway channel that drained Glacial Lake London through the Caradoc Delta near the townships of Komoka and Kilworth Heights into Glacial Lake Whittlesey or the earlier Glacial Lake Maumee III (the predecessors of Lake Erie; Chapman and Putnam, 1984; Dreimanis et al., 1998) once the Laurentide Ice Sheet had retreated northward during the end of the Wisconsinan Stage (Dyke and Prest, 1987). The advances and retreats of the Huron and Erie lobes (Chapman and Putnam, 1984) and other ice lobes (Teller and Kehew, 1994) left numerous meltwater lakes. It took at least 17 lacustrine phases until reaching modern Lake Erie, some of them lasting for only a few decades (Fullerton, 1980; Calkin and Feenstra, 1985; Totten, 1985; Barnett, 1985; Coakley and Lewis, 1985; Karrow et al., 2000). The exact timing of the formation of Glacial Lake London and the subsequent outburst flood(s) are currently unknown, although Glacial Lake Whittlesey (Hough, 1958; Forsyth, 1959) in the Erie and Huron basins existed during the Port Huron Stadial, about 13,000 radiocarbon years BP (Dreimanis, 1966; Barnett, 1979) or earlier during Glacial Lake Maumee III (Leverett and Taylor, 1915; Hough, 1958, Forsyth, 1959; Stewart, 1982) about 14,500 BP (Barnett, 1985; Calkin and Feenstra, 1985; Barnett, 1992). However, the last drainage of Glacial Lake London through the modern Thames River Valley has never been thoroughly investigated. Instead it was interpreted as a spillway channel, based on the modern underfit Thames River inner channel and surrounding valley dimensions (Chapman and Putnam, 1984; Dreimanis et al., 1998), but without corroborating evidence. A spillway channel is defined as an outlet of a lake (or dam) into a downstream river. The drainage mode can be catastrophic and short-lived (i.e., quick erosion of the dike and emptying of the lake within several hours or days) or result from gradual erosion of the dike and downstream channel until it achieves some kind of equilibrium in the long profile.

While misfit streams are a prominent feature of formerly glaciated terrains (Dury, 1960; 1964), the notion that the present-day topography is the result of one or more catastrophic high-magnitude low-frequency floods and not of postglacial incision (Phillips and Robert, 2005; Arbogast et al., 2008) possibly through channel migration (meandering, cutoffs and avulsion) needs corroboration from direct morphologic and stratigraphic evidence, hydrology, and hydraulic modeling.

Geologic evidence from the Quaternary shows that proglacial lakes were important landscape features during deglaciation (Blown and Church, 1985; Carrivick and Tweed, 2013) and their breaching and generation of glacial lake outburst floods (GLOFs, or jökulhlaups in Icelandic) are common even today under a rapidly warming climate

(Carrivick, 2010; 2011; Worni et al., 2014). GLOFs' high magnitude and destructiveness sometimes pose a natural hazard to downstream communities and infrastructure, so understanding these flood processes has both societal (risk preparedness) and academic benefits (Laenen et al., 1987; Mathews and Clague, 2000; Vilímek et al., 2005; Geertsema, 2013).

From a regional perspective, GLOFs are especially important because of the complex chain of hydrogeomorphic-hydroclimatic events (ice retreats and re-advances; Dreimanis, 1977; Dyke and Prest, 1987) that formed the modern Great Lakes (Karrow and Calkin, 1985; Barnett, 1992; Larson and Schaetzl, 2001). Understanding these process can add valuable information for the paleoenvironmental reconstruction of the region with further implications for local paleoclimate at the end of the ice age during deglaciation (Severinghaus and Brook, 1999; Derouin et al., 2007), appearance of flora (Goldthwait, 1958; Delcourt and Delcourt, 1984; Bartlein et al., 1986; Dreimanis et al., 1989; Jacobson et al., 1987; Yu, 2000), local fauna (Adams, 1905; Dreimanis, 1967; Dreimanis, 1977; Gibbard and Dreimanis, 1978; Mandrak and Crossman, 1992; Morris et al., 1993; Bajc et al., 1997; Yu, 2000; Karrow et al., 2007) and early human settlement of the area by the Palaeo-Indians (Jackson et al., 2000; Thieme, 2003; Ellis et al., 2011; Stewart and Desloges, 2014).

This work addresses this knowledge gap about the last Glacial Lake London and its drainage and has four main goals:

1. Establish dimensions (volume, area and depth) of the last Glacial Lake London;
2. Estimate peak discharge and other flow variables of the flood(s);

3. Assess field evidence to support a single catastrophic flood vs. several floods or gradual drainage;
4. Compare this flood with other ice-marginal lakes and outburst floods of the Laurentide Ice Sheet and other montane moraine-dammed lakes.

2. Regional setting and study site

The Thames River is the second largest basin in southern Ontario, draining an area of 5,825 km^2 into Lake St. Clair, the connecting lake between Lake Huron and Lake Erie (Fig. 1). Two main tributaries of the Thames, the North and South Thames, converge in the city of London. Downstream of London, the Thames River channel is entrenched into a deep river valley with relatively steep walls, 60-70 m high above the current channel bed for more than 15 km. The channel gradient in this reach is low - only 0.0015. The river exits this deep valley at the Caradoc Delta, the downstream limit of the study area.

The city of London is situated on a dry lake bed overlying tills, whereas the Thames River and its two major tributaries are incised into those lacustrine and glacial deposits (Dreimanis et al., 1998). Before the construction of Fanshawe Dam in 1952, the city suffered from large floods and heavy damages, especially during 1883 and 1937. In March 1977, an extreme spring melt reached 0.6 m below the crest of the dam (271.3 m asl) and the water released from the reservoir flooded London again (Helsten and Davidge, 2005; Upper Thames River Conservation Authority, 2015) demonstrating the poor soil drainage of the former lake and the narrow spillway at its outlet which presumably restricted drainage of the water across the city area. While it is difficult to recognize lacustrine sediments within the city outside of a few non-residential areas, an early painting of the city from 1893 clearly shows the flat landscape with the surrounding

Arva Moraine (Fig. 2), although this could potentially also reflect artistic license. Glacial Lake London was formed between the Arva Moraine and the Ingersoll Moraine on a sand outwash plain (Fig. 3). Where these two end moraines met, there was a structural change and more sand and gravel was deposited by meltwater than till by ice (seen at the large Byron gravel pit - Ontario Geological Survey, 1983) allowing the water to erode a spillway (Fig. 4). The two lake-constraining moraines (Arva and Ingersoll) are described below.

2.1 The Arva Moraine

The Arva Moraine is an end moraine that extends in NEE-SSW direction from SE of St. Marys in the north to north London in the south for a total distance of 17 km. Its maximum width is 3.2 km near the village of Arva. The moraine is presently breached in six different locations but only Medway Creek and the Thames River are possible spillway channels for Glacial Lake London. In order for Medway Creek to flow out of Glacial Lake London there would have to be drainage reversal of the present channel and an abandoned channel between Arva and Komoka Provincial Park but there is currently no field evidence to support this notion. Given the fact that the breached Arva Moraine on Medway Creek contains lacustrine deposits from Glacial Lake London on top of the moraine's bluffs (Fig. 5), it is ruled out that this river valley is the lake outlet. Furthermore, we conclude that the valley was already present when the lake existed and was an inlet basin of the lake highstand. The other breach opening occurs at the present Thames River course between the Arva Moraine and the Ingersoll Moraine (near the neighborhood of Byron) that were connected during the existence of Glacial Lake London (Sado and Vagners, 1975; Fig. 3; Fig. 4).

The materials composing this Late Wisconsinan moraine are: Huron Lobe sandy-silt loam till, sandy lacustrine deposits, sandy ice-contact stratified drift, gravelly ice-contact stratified drift, Huron Lobe sand-silt till, deltaic and some complex deposits, sandy outwash and gravelly outwash (Sado and Vagners, 1975). Drift is a generic term (that includes till) and refers to any glacial sediments whether they were transported, deposited or reworked by ice (glacial or glaciofluvial processes) while till was deposited by ice and was never reworked by subsequent processes. In general, the tills of the area can be associated with Catfish Creek Drift that was deposited during the Nissouri Stade, 23-16.5 ka BP (de Vries and Dreimanis, 1960; Terasmae et al., 1972). The Arva Moraine formed prior to the end of Lake Maumee II (14,200-14,000 BP), but it could have already existed during Lake Maumee I (Fullerton, 1980).

2.2 The Ingersoll Moraine

The Ingersoll Moraine is an end moraine that was formed during the retreat of the Erie Lobe (late Gary substage). The tills of this end moraine, associated ground moraines, and stagnant ice moraines range from clayey silt till to silty clay till, and they are classified as part of the Port Stanley drift (Dreimanis et al., 1963). The moraine is aligned in a west-east direction from Byron within London into Oxford County to a point just south of the city of Ingersoll, for a total length of 25 km (Fig. 3). Reynold Creek cuts through the moraine at Putnam and this valley was another flooded basin of Glacial Lake London. Fullerton (1980) noted that Ingersoll Moraine formation is not linked stratigraphically to a specific Lake Maumee phase and it may have been formed during Lake Maumee I (14500±150 BP, White, 1982).

3. Methodology

3.1 Reconstruction of Glacial Lake London

The only prior attempt to reconstruct Glacial Lake London was by Dreimanis et al. (1998) (Fig. 3), who used many unpublished theses at Western University and his own personal observations. The map produced by Dreimanis et al. (1998) was used here as a base layer map for initial verification of glacial landform features (moraines, till planes, lakes, deltas and spillways) after it was digitized and georeferenced into Arc-GIS and imposed on the Ontario 10 m resolution DEM (Ontario Ministry of Natural Resources, 2003). A similar lake reconstruction approach was suggested by Leverington et al. (2002) and Curry et al. (2014). Lacustrine silty-clay in Medway valley on top of Arva Moraine till (overlying a Huron Lobe till above Erie Lobe deposits - see Whittaker, 1986) at an elevation of 265.6 m asl is the highest known lacustrine deposit of Glacial Lake London, and therefore, represents the highest evidence of the lake bed (Fig. 5). The lacustrine sediments appear on top of all four bluffs along Medway Valley so we can assume they form a continuous crest line of the maximum elevation of water. The top DEM elevation of the two breached moraines (276.2 m) was the uppermost water surface prior to overtopping. Buried beach deposits of Glacial Lake London were found at elevations of 253-270 m asl, but there is a single anomalous higher shoreline deposit at an elevation of 280 m asl NW of Fanshawe Reservoir that is associated with drainage to Lake Leverett to the southwest (Dreimanis and Packer, 1959). Hence this topographic constraint of 276.2 m asl is the upper water level limit of the lake draining south (Fig. 4). The lake bed maximum (i.e. highest lake sediments) and the highest lake elevation possible constrain the elevation range for the lake (10.6 m difference). Evidence of lake elevations are currently limited and a further study focusing on the lake stands and the timing of its existence is needed to give a better understanding about the lake. We arbitrarily added two equal intervals of elevations of 3.5 m within the minimum and maximum water levels to see the gradual expansion of the water over the present topography. Dreimanis et al. (1998) suggested there were several Lake London phases, but our work will only relate to the last one, for which we have direct field evidence. We assumed the Ontario 10 m DEM of our study area does not need isostatic rebound correction to account for uplift (Lewis and Anderson, 1985; Calkin and Feenstra, 1985). However, it has to be acknowledged that no isostatic rebound contrasts an earlier value of 45 m suggested by the study of Jensen (1967). Since the lake was relatively small in comparison to the nearby vast ancestral Great Lakes we assumed that the London area uplift was insignificant or it was homogenous throughout the entire lake. Therefore, crustal deformation and rebound is not considered in this study.

3.2 Field evidence of possible flood features used as paleostage indicators (PSIs)

Since the London area receives today about 1000 mm of precipitation per year, of which about a third occurs as snow (2 m/yr; Thornthwaite and Associates, 1964; Peck et al., 2012), the preservation potential of depositional/erosional bedform features after thousands of years is low. The best preservation occurs in desert environments where erosion processes after the floods occurrences are slow and minimal (Greenbaum et al., 2006), but southern Ontario never had this climate since deglaciation and it was always a humid climate (Edwards and Fritz, 1986). However, the preservation potential is higher for bedform features or coarse lags away from the present active channel of the Thames River or along the valley walls of Komoka Provincial Park, which are relatively well preserved.

The distribution of geomorphic features in terms of their elevation and proximity to each other allowed us to assess whether the flood was a single flow event or if there were multiple floods. The former would generate erosional/depositional features on one top level replicating the peak water surface while the latter may generate a variety of forms along the valley at different elevations, such as terraces (Arbogast et al., 2008). The waning flow of a catastrophic flood may generate additional flood features at lower elevations but these are usually of smaller size as shear stress and stream power are reduced, and are often destroyed by a combination of subsequent (lower magnitude) floods and erosional slope processes. Furthermore, depositional features formed close to the modern channel and floodplain have a higher probability of being eroded by medium and smaller floods that would entrain the sediments, bury or rework them (Baker, 1984; Lewin and Macklin, 2003). However, we also recognize that, in the case of multiple floods, a high magnitude flood could have erased all evidence of its smaller lower-level predecessors. In selected locations, sediment samples were taken from the bedform features.

3.3 Determining peak discharge of the flood

We used two approaches to reconstruct the peak discharge of the flood(s): 1. Parametric breaching equations based on breach and lake volume dimensions (Wu et al., 2011). 2. Step-backwater modeling that is based on paleo-stage indicators (PSI; Baker, 1987) found within Komoka Provincial Park.

3.3.1 Parametric modeling breaching equations

The parametric modeling approach uses statistically derived regression equations for estimating the embankment breach characteristics (Costa, 1988; Walder and O'Connor, 1997; Wahl, 1998; Pierce et al., 2010; Thornton et al., 2011; Wu et al., 2011). Such equations for dam breaching have been developed based on data from historic dam failures (and occasionally flume experiments) that estimate from certain dam parameters the peak outflow discharge (Q_p) (MacDonald and Langridge-Monopolis, 1984; Soil Conservation Service, 1986; Froehlich, 1995; Walder and O'Connor, 1997; Tahershamsi et al., 2003; Thornton et al., 2011). Dam parameters typically include:

average width of the embankment (W),

breach shape,

breach side slope,

cross-sectional area of the embankment at the breach,

failure time (t_f in minutes or hours),

water volume released (V_{max}) or total water behind the dam (V),

mode of failure (overtopping, piping or seepage and foundation defects),

material erodibility (concrete or earthen embankment or a combination of the two),

the cross-sectional area of the embankment at the breach water surface elevation (A),

height of water above breach bottom (H_w), and

height of dam breach (d) (Soil Conservation Service, 1986; Froehlich, 1995; Thornton et al., 2011; De Lorenzo and Macchione, 2014). The main problem of the parametric approach is that it contains many uncertainties (Wahl, 2004). Hagen (1982) and later Walder and O'Connor (1997) claimed that V_{max}, d and H_w do not have any particular advantage as predictors of Q_p. However, the alternate computer modeling approach below also contains inherent uncertainties and assumptions that can change the flood routing outcome. Hence we take the view that these two approaches complement each other. A similar approach was taken by Clayton and Knox (2008) who obtained a range of values spanning two orders of magnitude for the peak discharge for Glacial Lake Wisconsin outburst flood. Clark et al. (2008) also used the same method for Glacial Lake Oshkosh volumes and its resulting outburst flood. We compare and contrast our findings with both these studies. Carling et al. (2010) found that dam failure does not have to occur at a maximum lake depth (i.e., 47.4 m deep) but flood routing models favor a scenario that the lake emptied by overtopping under conditions of maximum water level. Although an over-topping model is favored, collapse of the dam due to piping or high waves when the lake was below maximum capacity cannot be ruled out.

3.3.2 HEC-RAS s*tep backwater modeling, geomorphic features and modern hydrology*

We used the 1D HEC-RAS (Hydrologic Engineering Center - River Analysis System) 4.1 model, which has dam breaching capabilities. It is based on the Saint Venant equations for routing the flood downstream of the lake using channel cross-sections and assigned roughness (HEC-RAS, 2015). This engineering program is the most widely used 1D model for simulating the hydraulics of water flow through natural rivers and artificial watercourses and its main advantage is its simplicity, the relatively limited data it requires for running steady, unsteady and mixed flow simulations, generating water surface profiles and hydraulic parameters for each cross-section with stable runs. HEC-RAS' main limitation is that it has no direct modeling of the hydraulic effect of cross-section shape changes, bends, and other 2D and 3D aspects of the flow. Step-backwater modeling of the flood from Glacial Lake London was performed for a reach downstream of the breach at Byron (Fig. 4) all the way to the Caradoc Delta head, for a total length of

15.3 km of the Thames River (Fig. 6). The upstream part of the reach outside London is initially straight, but after 2.8 km there are two large meander bends within Komoka Provincial Park. The community of Kilworth Heights (part of Middlesex Township) is located on the right bank of the river valley creating some human disturbance but most of the development is outside the river valley flood course and has not influenced the postglacial cross-sectional shape. Valley cross-sections were extracted from the Ontario 10 m DEM that has ± 5 m vertical reliability (Ontario Ministry of Natural Resources, 2003). Farthest downstream spacing of cross-sections was 1.0 km, although much denser spacing was used near the paleoflood geomorphic features described in section 3.2 (Figs. 6 and 7). Larger spacing was used downstream of Komoka Provincial Park approaching the Caradoc Delta.

We are aware that the present DEM topography includes not only postglacial flood downcutting, but also subsequent postglacial Holocene incision by the modern Thames River. Choosing the correct bed datum of a large flood is a known problem in paleohydrology (Williams and Costa, 1988; Carling et al., 2003). Unless there are preserved terraces (Hack, 1965, Arbogast et al., 2008) or other clear geomorphic features that represent the past river bed from the flood of interest, the paleohydraulic modeling uses the contemporary river bed profile (Enzel et al., 1994; House and Baker, 2001; Kite et al., 2002; Greenbaum, 2007; Greenbaum et al., 2014). The logic behind this assumption is that contemporary discharges and the part of the cross-section they occupy are almost insignificant for the calculations of much larger floods (Baker, 1987).

Channel roughness values can have a large impact on modelled flow properties in step-backwater modelling (Fread, 1991a; Greenbaum. 2007), and their estimation requires careful consideration. Form resistance from vegetation is assumed to be negligible, as vegetation was not yet established in this newly deglaciated landscape which took a long time to shift from barren tundra (or periglacial desert) to tundra vegetation (Morris et al., 1993; Yu, 2003). Form resistance of stationary erratics (Ferro, 1999) were not found along the present-day flood course. However, major flow roughness can come from the eroding glacial sediments themselves (till and drift) in the form of hyperconcentrated flow (i.e. the flow has high ($\geq$ 60% volume or 80% by weight), mostly fine, suspended sediment concentration and the sediment plays and integral role in the flow behavior and mechanics; Pierson, 2005; Cao et al., 2006), large boulders once the material is transported (Elfstrom, 1987) or possible ice slabs (Prowse and Beltaos, 2002) within the lake. We used Manning's *n* roughness coefficient values of 0.07 for the channel and 0.10 close to the banks, as suggested by Matsch (1983) for River Warren, the southern outlet of Glacial Lake Agassiz. That flood event also reflects a deglaciating environment draining catastrophically. Flume experiments show that bed roughness decreases the velocity but not time-to-peak velocity. Also, suspended sediments reduce velocity variability (Carrivick et al., 2009) but it does not affect peak flow depth and peak flow velocity (Staines and Carrivick, 2015). We performed a sensitivity analysis, using a variety of *n* values to see how the choice of roughness parameter affected the simulated peak discharge (Wohl, 1998; Pappenberger et al., 2005; Carling et al., 2010). In sensitivity analysis simulations, we chose a Δn of 0.02 between the channel and the floodplain *n* roughness values in order to allow the floodplain to act as the main flow path to maintain downstream boundary conditions (Pappenberger et al., 2005). These selected Δn roughness values are small compared with the Δn values suggested by Horritt and Bates (2002) of 0.04-0.05, but their flood simulations were of a modern urban river with large floodplain roughness not a deglaciating one without vegetation.

Starting from a suitable paleo-DEM and using contemporary measured peak discharge, the step-backwater approach iteratively increases simulated discharges until features of interest are covered by the flow. This topographic approach requires identification of relevant bedform features (previous section), as well as a suitable paleo-DEM and a representative contemporary flood discharge (O'Connor and Webb, 1988). The ideal gauging station for our study is at Byron on the Thames River (station number 02GE001 draining 3,080 km^2) that is located exactly on the spillway of Glacial Lake London. However, the gauging station has been experiencing attenuated peak flow due to the existence of Fanshawe Dam and Reservoir since 1952 (drainage area of 1,450 km^2). Pittock Dam on the South Thames River (northeast of Woodstock) and Wildwood Dam on Trout Creek (northeast of St. Marys before the confluence with the North Thames River) are insignificant in terms of discharge attenuation. Therefore, in order to reconstruct modern maximum hydrology of Thames River without the attenuation of Fanshawe Dam, we summed up peak discharges of the two main tributaries: North Thames River near Thorndale (draining 1,320 km^2, station number 02GD015) upstream of Fanshawe Reservoir and South Thames River (draining 1,340 km^2 near Eeling, station number 02GD001) before the confluence (hydrological data from 1954-2013). When available, Medway Creek (draining an additional 203 km^2, below Fanshawe Dam, station number 02GD008) and Stoney Creek (draining an additional 37.3 km^2, below Fanshawe Dam, station number 02GD028) were also added to increase the peak discharge precision. We therefore covered about 86-94% of the total drainage area relative to Byron gauging station but with no attenuation. When maximum instantaneous flow was available, it was preferred over daily discharge. It is obvious that the timing of the peak flows rarely occurs simultaneously in all channels (Khan, 1993), but the combined peak flows give theoretical contemporary maximum discharge values that can happen under the current climate.

The HEC-RAS model uses this input discharge which then is iteratively raised in equal increments until it yields a water surface profile covering the field evidence (O'Connor and Webb, 1988; O'Connor and Baker, 1992; Enzel et al., 1994; D'Urso, 2000; Kite et al., 2002; Greenbaum, 2007). Besides peak flow, HEC-RAS also calculates flow depths, velocities and allows calculation of maximum shear stress and stream power. Finally, peak flood values reconstructed using hydraulic modeling are compared with four other cases of ice-marginal lakes of the Laurentide Ice Sheet and nineteen montane moraine-dammed lake breaching studies.

4. Results

4.1 Geomorphological evidence for a large flood

We identified and mapped 21 landform features along the Thames River valley left bank (facing downstream, southern bank) within Komoka Provincial Park (Fig. 7a, b). These features are possible paleostage indicators (Patton et al., 1979; Ely and Baker, 1985; Baker, 1987; Kochel and Baker, 1988; Jarrett, 1990). The most prominent feature is a 100 m long and 50 m wide surficial boulder lag (Fig. 7c) with clasts up to 1.2 m (b-axis) that is unusual compared with the present channel bed grain size distribution (Fig. 7d) (Costa, 1983; Waythomas and Jarrett, 1994; Fisher, 2004).The boulders are far (150-

200 m) from the modern channel, scattered, occasionally imbricated and many are partially buried within the soil with only their tops exposed. Other bedforms include several erosional landforms - three wash terraces (Fig. 7e) (Kehew and Lord, 1986; Kozlowski et al., 2005; Carrivick, 2007), six paleo-flow marks (a horizontal scar high on the valley wall that is a result of fluvial erosion rather than slope processes; Fig.7f) and four benches. The difference between a wash terrace and a bench is the scale: the first is a large-area flat surface 100s m long and wide while the second is only a few meters in width and length. In addition, two berms (an elongated low-elevation hump in the flow direction; Fig. 7g) and five mounds (2.5-3.0 m high and cone-shaped; Fig. 7h) were identified. These mounds may not necessarily result from a large flood and may be related to other ice-related processes (Clayton et al., 2008; Curry et al., 2010), although they could potentially be flood-deposited bergmounds (Bjornstad, 2014) or pendant bars (Malde, 1968). Today there is no apparent obstruction upstream of the berms and mounds such as bedrock or resistant glacial deposit. However, since this location was at the edge of the flow, the blockage could have been giant ice blocks. Following Bohorquez and Darby (2008), we subjectively classified each feature (low, medium and high certainty) as to whether or not it originated from a large flood. This uncertainty is important to recognize during hydraulic modeling and determination of peak flood and corresponding water surface slope. Minimum and maximum flood levels were reconstructed using only features identified with high certainty of being of fluvial origin (i.e., the boulder lag, and the fluvial terraces). However, these lower and upper profiles constrained all the other features (benches, mounds, berms, scars) between them besides one bench.

4.2 Reconstruction of Glacial Lake London

Reconstruction of the lake from minimum level (265.6 m) to maximum level prior to overtopping (276.2 m) is presented in Fig. 8 and Table 1. At maximum level, the lake was elongated: 45 km long east-west and 16 km wide north-south. The shoreline of Glacial Lake London had numerous irregularities, including spits and inlets typical of proglacial lakes (Clayton and Attig, 1989; Carrivick and Tweed, 2013). The lake was relatively shallow (19.9-26.7 m on average) with the deepest part next to the lake spillway at Byron - just below 50 m. At maximum level, Glacial Lake London had five major basins (Fig. 8d): London-Dorchester basin (the main body of water at all water levels), Medway Creek basin in the NW part of the lake, North Thames River - Fanshawe Lake basin in center NW, South Thames River basin in the NE next to the city of Woodstock, and Reynolds Creek basin in the SE. The latter three basins are outside of the reconstruction by Dreimanis et al. (1998) (Fig. 3), and are therefore newly identified parts of the lake. Dreimanis et al. (1998) excluded Medway Creek as being part of the lake, but the lacustrine stratigraphic evidence (Fig. 5) suggests the channel was already flooded by Glacial Lake London at that time and extended to the NE all the way to the village of Arva. Since there are currently no dated sediments and precise chronology of one or several lakes, our assumption is that each ice lobe's (Huron and Erie) readvance erased existing lacustrine sediments and thus only the last glacial lake sediments are preserved. After the lake drained, Medway Creek further incised into Arva Moraine while adjusting to the falling baselevel of the drained lake. There are two known paleodeltas within the paleolake: London Delta (seen in Fig. 3) and Medway Creek's Delta (Winder, 1980). Both are submerged and inactive today within Fanshawe Reservoir (Dreimanis et al.,

1998) and the North Thames River, respectively (Winder, 1980). The South Thames River basins and North Thames River basins probably bordered the retreating Laurentide Ice Sheet to the north as the Erie and Huron Lobes were separating from the interlobate zone (Dreimanis and Goldthwait, 1973; Barnett, 1985). The lake was relatively small compared to ancestral Great Lakes that existed at the time in the southern Laurentide Ice Sheet region (Eschman and Karrow, 1985; Hansel et al., 1985; Barnett, 1985; Teller and Kehew, 1994) and their respective spillways (Clayton and Attig, 1989; Clayton and Knox, 2008; Kehew et al., 2009). The total drainage area into the lake, based on present topography, was 2,700 km^2.

While the reason for dam failure is unknown, moraine dams commonly fail due to overtopping and erosion of the outlet (Costa and Schuster, 1988; Clague and Evans, 1994; 2000; Richardson and Reynolds, 2000; O'Connor and Beebe, 2009; O'Connor et al., 2013). However, if the dam breached due to piping or foundation failure (Blown and Church, 1985) of the area where Arva and Ingersoll Moraines converged, the highest lake elevation was 265.6 m asl. (10.6 m lower than full capacity and about a third of the maximum water volume). With current field evidence, it is not possible to define the exact lake dimensions and extent at the time of failure.

Lake volumes presented in Table 1 relate only to liquid water (i.e. the lake was proglacial not ice-marginal). If the lake was in contact with receding ice, especially during the calving phase (Schomacker, 2010), it is possible that ice was also a major component of lake volume (Carrivick, 2011). This condition implies a reduction in peak discharge due to obstruction by ice jams near the outlet (Smith and Pearce, 2002) and high roughness around stranded ice blocks (Russell, 1993). Consequently, peak discharges could be overestimated by the parametric breaching equations (Table 2) if these ice-processes affected discharge close to the breach. On top of the bluffs along Medway Creek, lacustrine sediment is massive and homogenous with no apparent stratigraphy (Fig. 5). Smith and Ashley (1985) and Ashley (1975; 1995) associated such a sedimentation pattern to glacial lakes with underflow circulation, no thermocline and with direct contact with the ice margin. If this sedimentologic assumption is correct for Glacial Lake London, rhythmites would only be found in the deeper parts of the lake and not in beach areas where only homogenous mud would settle (Carrivick and Tweed, 2013).

4.2. Peak discharge of the flood

4.2.1 Parametric breaching equations

Table 2 presents the results of peak discharge for Glacial Lake London using 21 selected equations proposed by thirteen authors and agencies. The results span two orders of magnitude, with a range of 13,000-241,400 m^3/s, an average of 64,900 m^3/s and a median of 44,100 m^3/s. Maximum calculated peak discharges for the Thames River (non-attenuated; Fig. 9), based on the gauged modern hydrology, range 137-1,493 m^3/s (average of 735 m^3/s, standard deviation 334 m^3/s, median 707 m^3/s). These average and median peak discharge values are about the same as modelled hydrographs proposed by Cunderlik and Simonovic (2005) for the Thames River. The attenuated bankfull discharge of the lower Thames River is 411 m^3/s (Stewart and Desloges, 2014). The highest calculated modern discharge (1,493 m^3/s) is one to two orders of magnitude smaller than any of the peak discharges calculated from parametric equations while the average and median flood discharges are three orders of magnitude smaller.

4.2.2 Hydraulic modeling

The maximum calculated modern discharge (1,493 m^3/s) was rounded to 1,500 m^3/s for incremental iterative increase of flow in the HEC-RAS modeling. When the simulated water elevation was close to the target elevation, this was refined to 150 m^3/s increments until the landforms of interest were precisely covered by the peak discharge (O'Connor and Webb, 1998). The HEC-RAS modeling results (Fig. 10) show a minimum discharge of 13,350 m^3/s covering the boulder lag, to 75,600 m^3/s covering the three terraces. The medium to low flood-origin features are contained within that range beside Bench 2, which is higher (85,000 m^3/s). The modern peak discharge (~1,500 m^3/s) is 11% of the modeled discharge needed for covering the boulder lag but only 2% of the discharge needed to cover the terraces. This suggests that peak paleodischarges of Glacial Lake London were 9 to 50 times greater than maximum floods derived from snowmelt runoff in the modern Thames River. Comparing the above parametric breaching equations in section 4.2.1, they are similar to the lowest peak discharge covering the boulder lag (13,350 m^3/s vs. 13,000 m^3/s), but differ from the upper elevation peak discharge covering the three terraces (75,600 m^3/s vs. 241,400 m^3/s).

It is not unusual to find disparity between peak discharge released from a lake based on its given water volume (i.e., parametric breaching equations) and peak discharge based on modeling that covers bedforms, paleostage indicators and slackwater deposits (e.g. Rathburn, 1993). The discrepancy is associated with the basic assumptions and focus of each method: The parametric breaching equations rely on the impounded water (lake or reservoir) and breach (or dam) characteristics before they drain downstream into the river or watercourse. These equations are often based on real case studies to calculate peak discharge (Wu et al., 2011). In many instances, the equations are clumped into different types of materials building the dikes to reduce the uncertainty and increase precision and ability to generalize (O'Connor et al., 2013). In contrast, 1D step-backwater modeling (such as the HEC-RAS model) relates to river hydraulics and conveying the water (energy) between points (cross-sections) while taking into account the friction (Manning's roughness) and expansion/contraction as the energy losses. The momentum equation is used when the channel is complex and has hydraulic jumps, structures that obstruct the flow and intervening water inputs from tributaries. For unsteady flow, the model fully solves 1D Saint Venant Equations using an implicit, finite difference method. Even when the dam is incorporated into the model and breached during the simulation, the river hydraulics remain the focus not the impounded lake.

The reliability of modeled peak discharges based on subjective choice and uncertainty of Manning’s *n* roughness coefficients deserve scrutiny (Burnham and Davis, 1990; Fread, 1991a; Aronica et al., 1998; Pappenberger et al., 2005). The results presented above were obtained using an estimated Manning’s *n* roughness of 0.07 and 0.1 for the channel and banks, respectively. Wohl (1998) found that variation of *n* by ±25% can lead to 20% change in discharge. A 20% error margin above or below the actual peak flow is considered a good engineering estimate (Trieste and Jarrett, 1987). The literature presents contradicting results regarding the effects of roughness on routing a dam-break flood. A comparison between a physical model (i.e. based on actual experiments on a reduced-size replica of the study area) and numerical dam-break models (SMPDBK and DAMBRK, respectively) found that the sensitivity of the latter to bottom friction of the channel is responsible for the large discrepancies in water surface slope (Bozkus and Kasap, 1998), similar to Greenbaum's (2007) results that also used the HEC-RAS model. Fread (1991a) claimed that the effect of altering *n* to a higher value is significant in raising the stage. O'Connor and Webb (1988) found opposite results, *i.e.* the peak discharge is not that sensitive to choosing Manning's *n*. Begnudelli and Sanders (2007) also found that the most accurate simulation of the St. Francis dam-break flood is produced when using a uniform Manning's *n* and that varying it only changes flood wave travel times, not the extent of flooding. Carling et al. (2010) found in the Altai Floods that varying Manning's *n* by 30% along the channel in a 1D model results in only an 8% variation in the upstream discharge. In 2D modeling the sensitivity was even less due to deep flow depths. Our use of a generic roughness for all cross-sections in the HEC-RAS model dam-break simulation simplifies this methodologic problem by assuming that till is homogeneous in terms of spatial erodibility and the resulting sediment transport. Fread (1991a), developer of the DAMBRK (Fread, 1991b) and FLDWAV (Fread and Lewis, 1998) models, suggested increasing Manning's *n* values from past (gauged) floods, especially from large ones. In an extensive study of HEC-RAS inundation extent, Pappenberger et al. (2005) found that a variety of roughness values may yield identical results in water surface profiles.

We therefore conducted a sensitivity analysis to determine whether our study is sensitive or not to Manning's *n* choice while running a variety of roughness combinations for the lower boulder lag and three higher terraces using the HEC-RAS model. Varying the *n* for the lower limit of the boulder lag results in differences in simulated discharges of up to 18% (Table 3), which falls within the 20% tolerance commonly applied in engineering design of dam infrastructure to withstand failure during extreme flood discharges (Trieste and Jarrett, 1987). However, for the upper limit of the flood (three terraces) there is a large discrepancy between the various simulations (Table 3).

We conclude that, considering the lake volume(s) in Table 1, the results from parametric breaching equations (Table 2), and our original Manning's *n* choices, it is improbable that peak discharge exceeded 100,000 m^3/s. Only four equations out of twenty one (19%) in Table 2 exceed that value, while the remaining discharges are all within a magnitude smaller. Therefore, the first two simulations in Table 3 for the terraces are likely erroneous and it is improbable that low Manning's roughness *n* values are realistic for a flood of this type and magnitude. Simulations 3-6 fall within the 20% permissible error, compared to our original Manning's *n* choice. The last simulation (7), with extreme high roughness values, also does not seem realistic and yields a discharge that is too low, although it somewhat resembles (11% difference) the median discharge from the parametric breach equations (49,100 m^3/s vs. 44,100 m^3/s). Considering that we cannot choose an alternate reach or find other paleostage indicators to verify various simulations (O'Connor and Webb, 1988), the upper limit of the flood is highly uncertain, unlike the lower boulder lag simulations that yield results with relatively small changes for a variety of Manning's roughness values. Despite the large uncertainty in the upper limit of peak discharge, our work provides an initial estimate, using two accepted techniques (parametric breach equations and modeling) for a flood downstream of the breach.

4.2.3 Peak flow variables

The modeled water surface slope of the outburst flood is 0.0027 for the lower boulder lag and 0.0031 for covering the higher three terraces (Fig. 10) over the 15.3 km reach.

These water surface slopes are steeper than present-day bed slope (0.0017) and maximum modern peak discharge water surface slope (0.0021). The differences in slopes can be explained by the valley's heterogeneous geometry with steeper valley slopes in the downstream direction next to the geomorphic features and energy losses in moderate valley slopes that are typical to terrain with high roughness. Furthermore, from the boulder lag to the farthest terrace, elevation loss is highest with a water surface slope of 0.0089, compared with bed gradient of only 0.0034. As a result of the relative steepness of this reach, this was the subreach with highest flood velocities at 10.0-11.4 m/s.

Calculated stream power for the dam-break flood was 620-1,270 W/m^2. These values are low compared to GLOFs in mountain environments (Cenderelli and Wohl, 2003; Staines and Carrivick, 2015) and can be explained by the low bed gradient of the Thames Valley. The calculated shear stress range was 760-1,050 N/m^2. Critical shear stress (τ_c) needed to entrain maximum clast size derived from many rivers was proposed by Komar (1988) using units of N/m^2 and mm:

$$\tau_c = (26.6\,(d/10)^{1.21})/10 \qquad (1)$$

The basic assumption with this method is that the formula is universal and is suitable for every river and flood type. Results suggest clasts 1.1-1.4 m could have been entrained which fits the largest clasts found within the boulder lag (up to 1.2 m). No other boulder erratics were found that could be associated with a large flood along the flood course. However, Arva and Ingersoll Moraines do contain larger boulders (b-axis > 2 m) than the boulder lag and these are occasionally scattered within modern river courses (Dreimanis and Packer, 1959; Bergman et al., 2016), gravel pits (Ontario Geological Survey, 1983; Krzyszkowski and Karrow, 2001) and fields upstream of the breach in what used to be the Glacial Lake London bed. These boulders remained stationary during the lake drainage and the local energy was not sufficient to transport them. While this sedimentary evidence of flow competence derived from boulder deposits is common in paleohydraulic reconstructions (Costa, 1983; Williams, 1983; Fisher, 2004), it is not without problems. Wilcock (1992) determined that the use of extreme sediment size values are prone to large errors as the sample size is relatively small, heavily biased towards the boulders, missing other grain fractions (as in our case) and does not represent the real sediment transport during a flood. Consequently, boulders are not reliable predictors of critical shear stress or flow magnitude (Wilcock, 1992). However, Carling (1987) associated boulder berms with flow separation downstream of an abrupt change in channel morphology that is consistent with the transition from the relatively wide lake outlet in Byron (Fig. 4) into the deeply incised river valley within Komoka Provincial Park. Considering we only use the boulder lag for our lower limit peak discharge, our paleohydraulic reconstruction is made more robust by using several geomorphic features and methods rather than only this single deposit.

The Froude number for the same reach is 0.5-0.8, subcritical but still much higher than most of the other subreaches with values of 0.1-0.3. The largest flow depth was above the boulder lag at 46.0 m. The shallowest flow depth was at Caradoc Delta in the end of the reach at 18.2 m.

5. Discussion

5.1 Why a catastrophic flood(s) and not gradual postglacial incision?

For the deep valley of Thames River (50-70 m high) downstream of London to gradually incise since the end of the ice age, a steady rate of 3.8-5.4 mm/yr for 13,000 BP time period is needed, or 3.4-4.8 mm/yr for 14,500 BP time period. This postglacial gradual incision approach does not fit the classic spillway morphology of the valley as described in detail by Kehew (1982) for underfit channels with lag deposits on the upper scoured surfaces of the valley. The question arises: was such an incision rate possible in a warming postglacial climate for southern Ontario?

Current local channels are relatively stable with negligible incision and sediment production (Ashmore and Church, 2001) and morphologic changes are usually associated with modern anthropogenic activities (Campo and Desloges, 1994; Tinkler and Parrish, 1998; Schottler et al., 2013). During the Holocene, the channels adjusted to Great Lakes water levels and isostatic rebound (Phillips and Robert, 2005). The maximum non-attenuated flood based on the gauged record (Fig. 9) fills only 2-5% of the river valley's area. We, therefore, looked for analogue rivers in the Great Lakes area that have an established chronology since the end of the ice age and compare their morphology with our study site. Arbogast et al. (2008) have shown that in the upper Muskegon River valley (Michigan, draining 6000 km^2 and Lake Michigan drainage) has four distinct terraces that were all dated to postglacial incision. They look different than the steep Thames River valley walls (Fig. 7a, 7b) with long flat surfaces and minimal elevation changes. The river valley is several km wide. Even the inner valley close to the channel has no morphologic resemblance to the Thames River. The Humber River (Ontario, draining 900 km^2 and Lake Ontario drainage) has incised about 30 m since the end of the ice age. It is quite similar to the Thames River valley as its headwater starts from the Oak Ridges Moraine and flows 21 km to its confluence with East Humber River (Phillips and Robert, 2005). However, the Humber River has many terraces and a wide valley reflecting gradual postglacial incision. Also, the valley reaches appear relatively straight, yet have abandoned meander loops (i.e. ancestral drainage) preserved on either side of the valley. This morphology suggests gradual downcutting after abandonment of the meander loops as the river hydrology changes from abundant glacial meltwater to predominantly rain and snowmelt hydrology. The dating of the valley features of the Humber River show that the rate of postglacial incision was highest during the first 4900 years (21 m or 4.3 mm/yr), then fell over the next 3900 years (8 m or 2.1 mm/yr) and finally in the last 6200 years changes were minimal (only 1 m or 0.2 mm/yr) (Phillips and Robert, 2005). While initial postglacial incision matches Thames River valley incision rates, the other two postglacial time periods of the Humber River incision rates are too small. Our conclusion is that the Thames River valley was cut by catastrophic lake drainage(s). The morphology of unusual features preserved along the valley's left wall (Table 4), lack of terraces in the Upper Thames River due to its inundation by a lake and preservation of large meander loops suggests rapid rather than gradual incision. Terraces exist in the Middle and Lower Thames River but that part of the basin evolved at a later stage when the river was diverted to the southwest and was no longer flowing to the nearby predecessors of Lake Erie to the south.

5.2 Single catastrophic flood vs. multiple floods

We did not find any deposit along the main Thames Valley that represents multiple drainages and backponding that produces repeated rhythmites as found elsewhere in outburst floods (Waitt, 1985). The lack of preservation of multiple floods is possibly associated with insufficient backwater areas like in some bedrock channels or could be due to the last catastrophic flood erasing the evidence of its predecessors. Grain size distributions of some of the geomorphic features indicate that the boulder lag is coarser than all other valley and lacustrine sediments (Fig. 11); it is completely devoid of any gravel < 256 mm, although this could be a result of later erosion and winnowing. None of the depositional geomorphic features presented in Table 4 resemble lacustrine silt-clay sediments that were found on Medway Creek's top bluff (Fig. 5). This sedimentary outcome could suggest several scenarios: 1) the lacustrine sediment was not evacuated from the breached lake, e.g. 'clear water' release (Lord and Kehew, 1987); 2) the dam-break released lake sediments and they mixed with river valley glacial sediments; 3) the lake did not have much fine lacustrine sediments since it was short-lived or the incoming meltwater was not turbid or 4) the sediment could have originated solely from the valley's glacial deposits (Fig. 7a) and bank till deposits (Fig. 7b) being scoured during the outburst torrent leading to numerous debris flows (Maizels, 1997).

Mound 1 (Fig. 7h) was sampled and contains three distinct layers. The 10-200 cm layer contains clasts up to cobble size and negates the idea that this depositional feature is a result of reworking of the fine lacustrine sediment as their grain size distributions do not match. It also casts doubt on a fluvial origin: how did such coarse particles reach so high in the water column as suspended sediment when the valley was flooded unless they are from a local source? The region's Wisconsinan history is complex and characterized by several advances and retreats of the local ice lobes (Dreimanis and Goldthwait, 1973) so possibly the mounds could be remnants of older glacial processes (Kleman, 1994). Yet, the few dam-break studies of similar subdued terrain (see the discussion) do not allow us to rule out these geomorphic features as non-fluvial. Consequently all mounds were classified as low degree of certainty in the hydraulic modeling (Table 4; Section 4.2.2).

Since the sedimentology cannot resolve single or multiple flood events in our study, the location, elevation and proximity of the features to each other and to the channel may provide another clue. Carling et al. (2009) stressed the importance of 'association' between geomorphic features where association refers to whether the sedimentary landform is contiguous with other landforms, including self-similar forms, which would have been formed during the flood. Our peak flood reconstruction relies on the four most prominent features: boulder lag and the three wash terraces that have high degree of flood-generated certainty. The HEC-RAS simulated upper and lower bound water profiles contain all the other features between them, except Bench 2. We used HEC-RAS again, specifically for each feature with the assumption that all features are flood-generated.

5.3 Comparison to other moraine dam-break floods

Evidence for the occurrence of outburst floods from glacial lakes in the Great Lakes region at the end of the Wisconsinan is quite extensive (Breckenridge and Johnson, 2009). However, it would be wrong to compare a small ice-marginal or proglacial lake like Glacial Lake London with the ancestral Great Lakes outlets (Bretz, 1951; Bleuer and Moore, 1979; Karrow and Calkin, 1985; Fraser and Bluer, 1988; Rea et al., 1994;

Rayburn et al., 2005; Kehew et al., 2009; Curry et al., 2014) or Glacial Lake Agassiz spillways (Fisher 1993; 2003; Colman et al., 1994; Leverington and Teller, 2003) that were associated with Laurentide Ice Sheet decline. These were megalakes on a regional scale, and their resulting outburst megafloods caused global climatic change and sea level rise (Broecker et al., 1989; Barber et al., 1999; Clarke et al., 2003; 2004). Megalakes are defined as exceptional for their extent and/or their volume (Baker, 2009), far greater than Glacial Lake London extent and volume. Outburst floods from megalakes are 'megafloods', with peak discharge exceeding 10^6 m^3/s (Baker, 2002; Clarke et al., 2003). We, therefore, compare Glacial Lake London with four proglacial lakes that were below the megaflood threshold: Glacial Lake Wisconsin (in Wisconsin; Clayton and Attig, 1989; Clayton and Knox, 2008), Glacial Lake Oshkosh (Wisconsin; Clark et al., 2008), Glacial Lake Watts and Glacial Lake Edmund (Pennsylvania; D'Urso, 2000). The criterion was a peak discharge < 1 million m^3/s, which is one order of magnitude larger than peak discharge generated by Glacial Lake London.

We also included twenty moraine-dammed montane lakes from data compiled by O'Connor and Beebee (2009) and two other moraine dam-break case studies of Lake Zurich (Strasser et al., 2008) and Lake Ventisquero Negro (Worni et al., 2012). The reason for inclusion of these montane lakes is that they are all dam-break floods that occurred within glacial sediments which reduces the geologic uncertainty (i.e., parent material and failure mechanism of the moraine-dammed lake) when looking for a correlation between the peak discharge released during the outburst flood vs. lake volume (O'Connor et al., 2013). However, glacier lake outburst may be caused by different mechanisms and might not be fully comparable to moraine dam-break outbursts. The results (Fig. 12) suggest that there is a medium correlation ($0.50 < R^2 < 0.80$; Doornkamp and King, 1971) between peak discharge and lake volume for montane lakes (R^2=0.68), but only a weak connection for proglacial lakes of the Laurentide Ice Sheet (R^2=0.36). Combined, a medium correlation is obtained (R^2=0.72). Most montane lakes are at least an order of magnitude smaller than the Laurentide Ice Sheet ice marginal lakes. Since most (77%) of the data points were obtained from the montane moraine-dammed lakes, which exhibit a large scatter around the best-fit line, it would be wiser to wait until more data are gathered from the Laurentide Ice Sheet ice-marginal lakes to better address the peak discharge vs. lake volume correlation of this flat glaciated terrain.

For Glacial Lake London, Fig. 12 shows that for both data points, there is notable variability (at least one order of magnitude) within the proglacial lake data. This gap can be explained by two possible reasons: 1) spatial erodibility of the glacial material that is highly variable among lodgement till (Kamphuis et al., 1990; Shugar et al., 2007; Khan and Kostachuk, 2011; Mier and Garcia, 2011) or softer deformation till (Hicock and Dreimanis, 1992) that was easier for the water to erode than other glacial deposits; 2) the flood route was ‘tectonically predesigned’ and followed an older lineament or glacial groove (Hack, 1965; Eyles et al., 1997), or even a preferential path of an older Glacial Lake London flood that was infilled by glacial readvance thus reducing the roughness the flood had to overcome. Despite the low bed gradient of the Thames River channel bed, the river valley spillway is deeply incised into a complex glacial landscape, a spatial phenomenon termed ‘glacial conditioning’ (Phillips and Desloges, 2014). Marren and Toomath (2014) claimed that topographic forcing of moraine-confined reaches of proglacial rivers is the major cause for their channel course/pattern, and that other factors such as hydrology and sediment supply play a minor role. This condition suggests that the flood peak released during a dam-break is strongly associated with the parent material(s) arrangement (topography and fluvial erodibility) and lake volume and dam height (such as used in the parametric breaching) are not the only factors in accurate prediction and the breaching process is more complex (Wu et al., 2011).

D'Urso (2000) addressed the large discrepancy between a HEC-RAS simulation fitting paleostage indicators (4,900 m^3/s) and the Slippery Rock Gorge constriction through which the outburst flood was routed and claims the actual discharge might have been less than an order of magnitude smaller if there was no overflow, similar to modern gauged peak discharges. This anomaly exemplifies the complexity of paleoflood reconstruction in a humid climate where preservation of paleostage indicators is uncertain, while the postglacial topography remains highly active (including further channel incision, bank erosion and hillslope mass wasting).

6. Conclusions

Glacial Lake London was reconstructed based on Dreimanis work, present-day topography, and lacustrine sediments found on top of the Arva Moraine bluffs in Medway Creek. For the first time, the lake is presented in its entirety at a variety of possible water levels. Catastrophic drainage was reconstructed based on breach dimensions, and geomorphic features found in the Komoka spillway area. Both of these approaches suggest that modern peak hydrology generated by snowmelt or rain-induced floods is 1-2 orders of magnitude less than the discharge generated by an outburst flood from a glacial lake. Furthermore, modern Thames River occupies a small area during large floods and is clearly a misfit river channel. However, with current field evidence we cannot determine whether drainage was a single event of repeated lake fill and drain cycles. Moraine-dammed lakes typically drain once and do not refill unless there is progressive lowering of the basin, which favors a single flood event. Following the many uncertainties, this work needs further expanding by conducting a thorough dating program of lacustrine sediments, mapping of beaches and dating the downstream flood deposits in order to have a better chronology of events.

Comparison of the Glacial Lake London breach to four other outburst floods from ice-marginal lakes of the Laurentide Ice Sheet and 20 montane moraine-dammed lakes show there is a link between lake water volume and discharge released from the montane lakes (R^2=0.68). However, unlike montane moraine-dammed lakes, there is not enough data for ice-marginal lakes of the Laurentide Ice Sheet (R^2=0.36) and a comparison to any of the megalakes spillways such as the ancestral Great Lakes or Glacial Lake Agassiz would be erroneous, due to the highly different magnitudes of water volumes (megalakes) and the resultant energetic release (megafloods).

The inclusion of this proglacial lake into the regional chronology is important from a variety of perspectives. 1) It offers an alternative explanation to gradual postglacial incision of Great Lakes rivers into glacial deposits and shows that proglacial lakes played a major role in landscape evolution of the Great Lakes and the American Midwest. 2) It proves that dramatic spillway topography can develop within low-relief moraine topography and complements the tunnel channel mechanism of rapid erosion and incision of the landscape that also occurs at receding ice margins, especially in mountainous terrains. 3) The deglaciation process was not only characterized by slow retreats of ice lobes but also by rapid removal of meltwater and ice by flowing water, thus allowing flora, fauna and humans to encroach on the exposed deglaciating landscape or disappear when the ice readvanced. 4) It is important to put rivers in formerly glaciated terrains in the right context as catastrophic landscape-shaping flood events still affect them and in most instances these rivers cannot be treated as ordinary self-formed alluvial channels.

Acknowledgments

Ontario Ministry of Natural Resources (MNR) Ontario Parks gave us permission to work and collect data in Komoka Provincial Park. Bob Jarrett advised on roughness values and HEC-RAS modeling. Jim O'Connor of the USGS Portland supplied the data of moraine dam-break floods. Chris Smart is thanked for interesting discussions regarding finds in Komoka Provincial Park. Erika Hill helped with laboratory analysis and field equipment. The Geological Association of Canada (AGC) is thanked for allowing us to use the Dreimanis et al. (1998) original figure (Fig. 3) of the glacial landscape of London. Joe Smrekar helped with various computing issues. Karen Van Kerkoerle drew and improved most maps, and some additional figures. Caren Raedts shared her ideas of the Thames River evolution following the Wisconsinan. Mike Hensen of the London Free Press took the photograph in Fig. 7c. This paper is part of the first author's PhD dissertation under the supervision of M. Van De Wiel and S. Hicock.

References

Adams, C.C., 1905. The postglacial dispersal of the North American biota. Biological Bulletin 9, 53-71.

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-central Lower Michigan, U.S.A. Geomorphology 102, 615-623.

Aronica, G., Hankin, B., Beven, K.J., 1998. Uncertainty and equifinality in calibrating distributed roughness coefficients in a flood propagation model with limited data. Advances in Water Resources 22, 349-365.

Ashley, G.M., 1975. Rhythmic sedimentation in Glacial Lake Hitchcock, Massachusetts-Connecticut. In: Jopling, A.V. and McDonald, B.C. (Eds.) Glaciofluvial and Glaciolacustrine Sedimentation 23, 304-320.

Ashley, G.M., 1995. Glaciolacustrine environments. In: Menzies, J. (ed.) Modern Glacial Environments, Vol. 1, Butter-worth-Heinemann, Oxford, 417-444.

Ashmore, P., and Church, M. 2001. The impact of climate change on rivers and processes in Canada. Available from the Geological Survey of Canada, Ottawa, Ont. Bulletin 555.

Bajc, A.F., Morgan, A.V., Warner, B.G., 1997. Age and paleoecological significance of an early Postglacial fossil assemblage near Marathon, Ontario, Canada. Canadian Journal of Earth Sciences 34, 687-698.

Baker, V.R., 1984. Flood sedimentation in bedrock fluvial systems. In: Koster, E.H., and Steel, R.J., (Eds.), Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists Memoir 10, 87-98.

Baker, V.R., 1987. Palaeoflood hydrology and extraordinary flood events. Journal of Hydrology 96, 79-99.

Baker, V.R., 2002. High-energy megafloods: planetary settings and sedimentary dynamics. In: Martini, I.P., Baker, V.R., Garzon, G., (Eds.), Flood and Megaflood Deposits: Recent and Ancient Examples. IAS Special Publication 32, 3-15.

Baker, V.R., 2009. Glacial Megalakes. In: Gornitz, V. (Ed.) Encyclopedia of Earth Sciences Series: Encyclopedia of Paleoclimatology and Ancient Environments, 380-382.

Barber, D.C., Dyke, A., Hillaire-Marcel, C., Jennings, A.E., Andrews, J.T., Kerwin, M.W., Bilodeau, G., McNeely, R., Southon, J., Morehead, M.D., Gagnon, J.-M., 1999. Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes. Nature 400, 344-348.

Barnett, P.J., 1979. Glacial Lake Whittlesey: the probable ice frontal position in the eastern end of the Erie basin. Canadian Journal of Earth Sciences 16, 568-574.

Barnett, P.J., 1985. Glacial retreat and lake levels, north central Lake Erie Basin, Ontario. In: Karrow, P.F., and Calkin, P.E., (Eds.) Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 185-194.

Barnett, P.J., 1992. Quaternary Geology of Ontario. In: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Scott, G.M. (Eds.). Geology of Ontario. Ontario Geological Survey, Special Volume 4, 1011-1088.

Bartlein, P.J., Prentice, I.C., Webb, T.III, 1986. Climatic response surfaces from pollen data for some eastern North American taxa. Journal of Biogeography 13, 35-57.

Begnudelli, L., Sanders, B.F., 2007. Simulation of the St. Francis dam-break flood. Journal of Engineering Mechanics 133, 1200-1212.

Bergman, N., Van De Wiel, M.J., Hicock, S.R., 2016. Semi-alluvial stream channel crossing the interlobate Arva Moraine, Medway Creek, Southern Ontario, Canada. 1: Sedimentary characteristics and morphologic change. This dissertation, chapter 2.

Bjornstad, B.N., 2014. Ice-rafted erratics and bergmounds from Pleistocene outburst floods, Rattlesnake Mountain, Washington, USA. Quaternary Science Journal 63, 44-59.

Bleuer, N.K. Moore, M.C., 1979. Glacial stratigraphy of the Fort Wayne area and the draining of Glacial Lake Maumee. Indiana Academy of Science 81, 195-209.

Blown, I., Church, M., 1985. Catastrophic lake drainage within the Womathko River basin, British Columbia. Canadian Geotechnical Journal 22, 551-563.

Bohorquez, P., Darby, S.E., 2008. The use of one- and two-dimensional hydraulic modelling to reconstruct a glacial outburst flood in a steep Alpine valley. Journal of Hydrology 361, 240-261.

Bozkus, Z., Kasap, A., 1998. Comparison of physical and numerical dam-break simulations. Turkish Journal of Engineering and Environmental Science 22, 429-443.

Breckenridge, A., Johnson, T.C., 2009. Paleohydrology of the upper Laurentian Great Lakes from the late glacial to early Holocene. Quaternary Research 71, 397-408.

Bretz, J H., 1951. The stages of Lake Chicago: their causes and correlations. American Journal of Science 249, 401-429.

Broecker, W.S., Kennett, J.P., Flower, B.P., Teller, J.T., Trumbore, S., Bonani, G., Wolfli, W., 1989. Routing of meltwater from the Laurentide Ice Sheet during the Younger Dryas cold episode. Nature 341, 318-321.

Burnham, M.W., Davis, D.W., 1990. Effects of data errors on computed steady-flow profiles. Journal of Hydraulic Engineering, ASCE, 116, 914-929.

Calkin, P.E., Feenstra, B.H., 1985. Evolution of the Erie-Basin Great Lakes. In: Karrow, P.F., and Calkin, P.E., (Eds.), Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 149-170.

Campo, S.H., Desloges, J.R., 1994. Sediment yield conditioned by glaciation in a rural agricultural basin of southern Ontario, Canada. Physical Geography 15, 495-515.

Cao, Z., Pender, G., Carling, P., 2006. Shallow water hydrodynamic models for hyperconcentrated sediment-laden floods over erodible bed. Advances in Water Resources 29, 546-557.

Carling, P.A., 1987. Hydrodynamic interpretation of a boulder berm and associated debris-torrent deposits. Geomorphology 1, 53-67.

Carling, P.A., Kidson, R., Cao, Z., Herget, J., 2003. Paleohydraulics of extreme flood events: Reality and myth. In: Gregory, K.J., Benito, G., (Eds.) Paleohydrology: Understanding Global Change. John Wiley & Sons, Chichester, UK, 325-336.

Carling, P.A., Burr, D.M., Johnsen, T.F., Brennand, T.A., 2009. A review of open-channel megaflood depositional landforms on Earth and Mars. In: Burr, D.M., Carling, P.A., Baker, V.R., (Eds.), Megaflooding on Earth and Mars. Cambridge University Press, 33-49.

Carling, P.A., Villanueva, I., Herget, J., Wright, N., Borodavko, P., Morvan, H., 2010. Unsteady 1D and 2D hydraulic models with ice dam break for Quaternary megaflood, Altai Mountains, southern Siberia. Global and Planetary Change 70, 24-34.

Carrivick, J.L., 2007. Modelling coupled hydraulics and sediment transport of a high magnitude flood and associated landscape change. Annals of Glaciology 45, 143-154.

Carrivick, J.L., 2010. Dam break - Outburst flood propagation and transient hydraulics: A geosciences perspective. Journal of Hydrology 380, 338-355.

Carrivick, J.L., 2011. Jökulhlaups: geological importance, deglacial association and hazard management. Geology Today 27, 133-140.

Carrivick, J.L., Tweed, F.S., 2013. Proglacial lakes: character, behaviour and geological importance. Quaternary Science Reviews 78, 34-52.

Carrivick, J.L., Xie, Z., Sleigh, A., Hubbard, M., 2009. Experimental modelling of flow – bed interactions in jökulhlaups. Poster presented in EGU 2015, Vienna, Austria.

Cenderelli, D.A., Wohl. E.E., 2003. Flow hydraulics and geomorphic effects of glacial-lake outburst floods in the Mount Everest region, Nepal. Earth Surface Processes and Landforms 28, 385-407.

Chapman, L.J. Putnam, D.F., 1984. The physiography of Southern Ontario**.** Ontario Ministry of Natural Resources, 3RD edition, Ontario Geological Survey special volume 2, 270 p.

Clague, J.J., Evans, S.G., 1994. Formation and failure of natural dams in the Canadian Cordillera. Geological Survey of Canada Bulletin 464, 35 p.

Clague, J.J., Evans, S.G., 2000. A review of catastrophic drainage of moraine-dammed lakes in British Columbia. Quaternary Science Reviews 19, 1763-1783.

Clark, J.A., Befus, K.M., Hooyer, T.S., Stewart, P.W., Shipman, T.D., Gregory, C.T., Zylstra, D.J., 2008. Numerical simulation of the paleohydrology of glacial Lake Oshkosh, eastern Wisconsin, USA. Quaternary Research 69, 117-129.

Clarke, G., Leverington, D., Teller, J., Dyke, A., 2003. Superlakes, megafloods, and abrupt climate change. Science 301, 922-923.

Clarke, G.K.C., Leverington, D.W., Teller, J.T., Dyke, A.S., 2004. Paleohydraulics of the last outburst flood from glacial Lake Agassiz and the 8200 BP cold event. Quaternary Science Reviews 23, 389-407.

Clayton, J.A., Knox, J.C., 2008. Catastrophic flooding from Glacial Lake Wisconsin. Geomorphology 93, 384-397.

Clayton, L., Attig, J.W., 1989. Glacial Lake Wisconsin. Geological Society of America Memoir, vol. 173. Boulder, Colorado, 80 p.

Clayton, L., Attig, J.W., Ham, N.R., Johnson, M.D., Jennings, C.E., Syverson, K.M., 2008. Ice-walled-lake plains: implications for the origin of hummocky glacial topography in middle North America. Geomorphology 97, 237-248.

Coakley, J.P., Lewis, C.F.M., 1985. Postglacial lake levels in the Erie Basin. In: Karrow, P.F., and Calkin, P.E., (Eds.) Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 195-212.

Colman, S.M., Keigwin, L.D., Forester, R.M., 1994. Two episodes of meltwater influx from glacial Lake Agassiz and their climatic contrast. Geology 22, 547-550.

Cooper, A.J., Baker, C., 1978. Quaternary Geology of the Bothwell-Ridgetown Area, Southern Ontario. Ontario Geological Survey Preliminary Map P.1973, Scale 1:50000.

Costa, J.E., 1983. Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range. Geological Society of America Bulletin 94, 986-1004.

Costa, J.E., 1985. Floods from dam failures. U.S. Geological Survey Open-File Report 85-560, Denver, Colorado.

Costa, J.E., Schuster, R.L., 1988. The formation and failure of natural dams. Geological Society of America Bulletin 100, 1054-1068.

Costa, J.E., 1988. Floods from dam failures. In: Baker, V.R. Kochel, R.C. and Patton, P.C. (Eds.), Flood Geomorphology. John Wiley and Sons Inc., 439-463.

Cunderlik, J.M., Simonovic, S.P., 2005. Hydrological extremes in a southwestern Ontario river basin under future climate conditions. Hydrological Sciences Journal 50, 631-654.

Curry, B.B., Konen, M.E., Larson, T.H., Yansa, C.H., Hackley, K.C., Alexanderson, H., Lowell, T.V., 2010. The DeKalb mounds of northeastern Illinois as archives of deglacial history and postglacial environments. Quaternary Research 74, 82-90.

Curry, B.B., Hajic, E., Befus, K., Clark, J. and Carrell, J., 2014. The Kankakee Torrent and other large meltwater flooding events during the last deglaciation, Illinois, USA. Quaternary Science Reviews 90, 22-36.

De Lorenzo, G., Macchione, F., 2014. Formulas for the peak discharge from breached earthfill dams. Journal of Hydraulic Engineering, ASCE, 140, 56-67.

Delcourt, P.A., Delcourt, H.R., 1984. Late Quaternary paleoclimate and biotic responses in eastern North America and the western North Atlantic Ocean. Palaeogeography, Palaeoclimatology, Palaeoecology 48, 263-284.

Derouin, S.A., Lowell, T.V., Hadjas, I., 2007. Evolution and deglaciation of the Upper Peninsula of Michigan: An examination of chronology and stratigraphy in kettle lake cores. Journal of Great Lakes Research 33, 875-886.

de Vries, H., Dreimanis, A. 1960. Finite radiocarbon dates of the Port Talbot Interstadial deposits in Southern Ontario. Science 131, 1738-1739.

Doornkamp, J.C., King, C.A.M., 1971. Numerical analysis in geomorphology: an introduction. Edward Arnold, London, 372 pp.

Dreimanis, A., 1966. Lake-Arkona-Whittlesey and post-Warren radiocarbon dates from “Bridgetown Island” southwestern Ontario. The Ohio Journal of Science 66, 582-586.

Dreimanis, A., 1967. Mastodons, their geologic age and extinction in Ontario, Canada. Canadian Journal of Earth Sciences 4, 663-675.

Dreimanis, A., 1977. Late Wisconsin glacial retreat in the Great Lakes region, North America. Annals of the New York Academy of Sciences 288, 70-89.

Dreimanis, A., 1995. Landforms and structures of the waterlain west end of St. Thomas moraine, SW Ontario, Canada. Geomorphology 14, 185-196.

Dreimanis, A., Vagners, U.G., McKenzie, G.D., Etherington, J.R., 1963. Pleistocene geology of the St. Thomas area (west half). Ontario Department of Mines, Preliminary Geological Map No. 238.

Dreimanis, A., Goldthwait, R.P., 1973. Wisconsin Glaciation in the Huron, Erie, and Ontario Lobes. Geological Society of America Memoirs 136, 71-106.

Dreimanis, A., Packer, R., 1959. Friends of Pleistocene Geology Eastern Section Field Trip Guide. Contribution of the Department of Geology, University of Western Ontario, No. 25, 22nd Reunion, May 16-17, 1959, London, Ontario, 30 p.

Dreimanis, A., Liivrand, E., Raukas, A., 1989. Glacially redeposited pollen in tills of southern Ontario, Canada. Canadian Journal of Earth Sciences 26, 1667-1676.

Dreimanis, A., Winder, C.G., Aaltonen, R.A., 1998. London Ontario: Geology, geomorphology, geodata. In: Karrow, P.F. and White, O.L. (eds.) Urban Geology of Canadian Cities. Geological Society of Canada Special Paper 42, 237-260.

D’Urso, G.J., 2000. Revised glacial margins and Wisconsin meltwater paleoflood hydrology in Slippery Rock Creek Basin, Central Western Pennsylvania. Unpublished PhD Dissertation, West Virginia University, 174 p.

Dury, G.H., 1960. Misfit streams: Problems in interpretation, discharge, and distribution. Geographical Review 50, 219-242.

Dury, G.H., 1964. Principles of underfit streams: General theory of meandering valleys. U.S. Geological Survey Professional Paper 452A, Washington, D.C.

Dyke, A.S., Prest, V.K., 1987. Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Géographie physique et Quaternaire 41, 237-263

Edwards, T.W.D., Fritz, P., 1986. Assessing meteoric water composition and relative humidity from ^{18}O and ^{2}H in wood cellulose: paleoclimatic implications for southern Ontario, Canada. Applied Geochemistry 1, 715-723.

Elfstrom, A., 1987. Large boulder deposits and catastrophic floods. A case study of the Båldakatj Area, Swedish Lapland. Geografiska Annaler 69, 101-121.

Ellis, C.J., Carr, D.H., Loebel, T.J., 2011. The Younger Dryas and Late Pleistocene peoples of the Great Lakes region. Quaternary International 242, 534-545.

Ely, L.L., Baker, V.R., 1985. Reconstructing palaeoflood hydrology with slackwater deposits: Verde River, Arizona. Physical Geography 6, 103-126.

Enzel, Y., Ely, L.L., Martinez-Goytre, J., Vivian, R.G., 1994. Paleofloods and dam-failure flood on the Virgin River, Utah and Arizona. Journal of Hydrology 153, 291-315.

Eschman, D.F., Karrow, P.F., 1985. Huron basin glacial lakes: a review. In: Karrow, P.F., Calkin, P.E. (Eds.), Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 79–93.

Evans, S.G., 1986. The maximum discharge of outbursts floods caused by the breaching of man-made and natural dams. Canadian Geotechnical Journal 23, 385-387.

Eyles, N., Arnaude, E., Scheidegger, A.E., Eyles, C.H., 1997. Bedrock jointing and geomorphology in southwestern Ontario: an example of tectonic predesign. Geomorphology 19, 17-34.

Ferro, V., 1999. Friction factor for gravel-bed channel with high boulder concentration. Journal of Hydraulic Engineering, ASCE, 125, 771-778.

Fisher, T.G., 1993. Glacial Lake Agassiz: The N.W. Outlet and Paleoflood Spillway, N.W. Saskatchewan and N.E. Alberta. Unpublished Ph.D. thesis, University of Calgary, Canada.

Fisher, T.G., 2003. Chronology of glacial Lake Agassiz meltwater routed to the Gulf of Mexico. Quaternary Research 59, 271-276.

Fisher, T.G., 2004. River Warren boulders, Minnesota, USA: catastrophic paleoflow indicators in the southern spillway of glacial Lake Agassiz. Boreas 33, 349-358.

Forsyth, J.L., 1959. The beach ridges of northern Ohio. Ohio Department of Natural Resources, Division Of Geological Survey Information Circular No. 25, 10 p.

Fraser, G.S., Bleuer, N.K., 1988. Sedimentological consequences of two floods of extreme magnitude in the late Wisconsin Wabash Valley. In: Clifton, H.E. (Ed.) Sedimentologic Consequences of Convulsive Geologic Events. Geological Society of America Special Paper 229, 111-126.

Fread, D.L., 1991a. Flood routing models and Manning n. In: Channel Flow Resistance: Centennial of Manning's Formula. Water Resources Publications, Littleton, Colorado, 421-435.

Fread, D.L., 1991b. The NWS DAMBRK model: theoretical background/user documentation. HRL-256, Hydrologic Research Laboratory, National Weather Service, Silver Spring, Maryland, 315 p.

Fread, D.L., Lewis, J.M., 1998. NWS FLDWAV model theoretical description and user documentation. National Weather Service (NWS), NOAA, Silver Spring, Maryland, 1.1-27.1.

Froehlich, D.C., 1995. Peak outflow from breached embankment dam. Journal of Water Resources Planning and Management 121, 90-97.

Fullerton, D.S., 1980. Preliminary Correlation of Post-Erie Interstadial Events (16,000-10,000 Radiocarbon Years Before Present), Central and Eastern Great Lakes

Region, and Hudson, Champlain, and St. Lawrence Lowlands, United States and Canada. US Geological Survey Professional Paper 1089, 52 p.

Geertsema, M., 2013. Jökulhlaups. Encyclopedia of Natural Hazards, Encyclopedia of Earth Science Series 2013, 567-569.

Gibbard, P.L., Dreimanis, A., 1978. Trace fossils from late Pleistocene glacial lake sediments in southwestern Ontario, Canada. Canadian Journal of Earth Sciences 15, 1967-1976.

Goff, K., Brown, D.R., 1981. Groundwater resources of the Thames River basin. Ontario Water Resources Report 14, Ontario Ministry of the Environment, 5 p.

Goldthwait, R.P., 1958. Wisconsin age forests in western Ohio. I. Age and glacial events. Ohio Journal of Science 58, 209-219.

Greenbaum, N., 2007. Assessment of dam-failure flood and a natural, high-magnitude flood in a hyperarid region using paleoflood hydrology, Nahal Ashalim catchment, Dead Sea, Israel. Water Resources Research 43, W02401, doi:10.1029/2006WR004956.

Greenbaum, N. Porat, N. Rhoads, E., Enzel, N., 2006. Large floods during late Oxygen Isotope Stage 3, southern Negev desert, Israel. Quaternary Science Reviews 25, 704-719.

Greenbaum, N., Harden, T.M., Baker, V.R., Weisheit, J., Cline, M.L., Porat, N., Halevi, R., Dohrenwend, J., 2014. A 2000 year natural record of magnitudes and frequencies for the largest Upper Colorado River floods near Moab, Utah. Water Resources Research 50, 10.1002/2013WR014835.

Hack, J.T., 1965. Postglacial drainage evolution and stream geometry in the Ontonagon area, Michigan. USGS Professional Paper 504-B.

Hagen, V.K., 1982. Re-evaluation of design floods and dam safety. Transactions of the 14th International Congress on Large Dams, vol. 1, International Commission on Large Dams, Paris, 475-491.

Hansel, A.K., Mickelson, D.M., Schneider, A.F., Larsen, C.E., 1985. Late Wisconsinan and Holocene history of the Lake Michigan basin. In: Karrow, P.F., Calkin, P.E. (Eds.), Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 39-53.

HEC-RAS, 2015. http://www.hec.usace.army.mil/software/hec-ras/ (Last accessed 9 March 2015).

Helsten, M., Davidge, D., 2005 Flood damage estimation in the Upper Thames River Watershed CFCAS project: Assessment of water resources risk and vulnerability to changing climatic conditions. Upper Thames River Conservation Authority, Project Report VII, 46 p.

Hicock, S.R., Dreimanis, A., 1992. Deformation till in the Great Lakes region: implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. Canadian Journal of Earth Sciences 29, 1565-1579.

Historica Dominion Institute Canada., 2015. Thames River. In: http://www.thecanadianencyclopedia.com/articles/thames-river. Last accessed: March 30, 2015.

Horritt, M.S., Bates, P.D., 2002. Evaluation of 1D and 2D numerical models for predicting river flood inundation. Journal of Hydrology 268, 87-99.

Hough, J. L., 1958. Geology of the Great Lakes. University of Illinois Press, Urbana, 313 p.

House, K.P., Baker, V.R., 2001. Paleohydrology of flash floods in small desert watersheds in western Arizona. Water Resources Research 1825-1839.

Jackson, J.L., Ellis, C., Morgan, A.V., McAndrews, J.H., 2000. Glacial lake levels and Eastern Great Lakes Palaeo-Indians. Geoarchaeology: An International Journal 15, 415-440.

Jacobson, G.L., Jr., Webb, T., III, Grimm, E.C., 1987. Patterns and rates of vegetation change during the deglaciation of eastern North America. In: Ruddiman, W.F. and Wright, H.E., Jr. (Eds.) North America and adjacent oceans during the last deglaciation. Boulder Colo., Geological Society of America, The Geology of North America, v. K-3, 277-288.

Jarrett, R.D., 1990. Paleohydrologic techniques used to define the spatial occurrence of floods. Geomorphology 3, 181-195.

Jensen, L.S., 1967. A study of the Pleistocene geology of the campus of the University of Western Ontario. Unpublished Senior Thesis, Department of Geology, University of Western Ontario, London, Ontario, 139 p.

Kamphuis, J.W., Gaskin, P.N., Hoogendoorn, E., 1990. Erosion tests on four intact Ontario clays. Canadian Geotechnical Journal 27, 692-696.

Karrow, P.F., Calkin, P.E., (Eds.) 1985. Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 258 p.

Karrow, P.F., Dreimanis, A., Barnett, P.J., 2000. A proposed diachronic revision of Late Quaternary time-stratigraphic classification in the Eastern and Northern Great Lakes area. Quaternary Research 54**,** 1-12.

Karrow, P.F. Morris, T.F. McAndrews, J.H. Morgan, A.V. Smith, A.J. Walker, I.R. 2007. A diverse late-glacial (Mackinaw Phase) biota from Leamington, Ontario. Canadian Journal of Earth Sciences 44, 287-296.

Kehew, A.E., 1982. Catastrophic flood hypothesis for the origin of the Souris Spillway, Saskatchewan and North Dakota. Geological Society of America Bulletin 93, 1051-1058.

Kehew, A.E., Lord, M.L., 1986. Origin and large-scale erosional features of glacial-lake spillways in the northern Great Plains. Geological Society of America Bulletin 97, 162-177.

Kehew, A.E., Lord, M.L., Kozlowski, A.L., Fisher, T.G., 2009. Proglacial megaflooding along the margins of the Laurentide Ice Sheet. In: Burr, D.M., Carling, P.A., Baker, V.R., (Eds.) Megaflooding on Earth and Mars. Cambridge University Press, Ch. 7, 104-127.

Khan, I., Kostachuk, R., 2011. Erodibility of cohesive glacial till bed sediments in urban stream channel systems. Canadian Journal of Civil Engineering 38, 1363-1372.

Khan, M.H., 1993. Muskingum flood routing model for multiple tributaries. Water Resources Research 29, 1057-1062.

Kirkpatrick, G.W., 1977. Evaluation guidelines for spillway adequacy. The Evaluation of Dam Safety, Engineering Foundation Conference, ASCE, Reston, Virginia, 395-414.

Kite, S.J., Gebhardt, T.W., Springer, G.S., 2002. Slackwater deposits as paleostage indicators in canyon reaches of the central Appalachians: Reevaluation after the

1996 Cheat River flood. In: House, P.K., Webb, R.H., Baker, V.R., Levish, D.R. (Eds.), Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology. Water Sciences and Application, vol. 5. American Geophysical Union, Washington DC, 257-266.

Kleman, J., 1994. Preservation of landforms under ice sheets and ice caps. Geomorphology 9, 19-32.

Kochel, R.C., Baker, V.R., 1988. Palaeoflood analysis using slackwater deposits. In: Baker, V.R., Kochel, R.C., Patton, P. C., (Eds.), Flood geomorphology, Wiley, New York, 357-376.

Komar, P.D., 1988. Sediment transport by floods. In: Baker, V.R., Kochel, R.C., and Patton, P.C., (Eds.), Flood Geomorphology. John Wiley & Sons, 97-111.

Kozlowski, A.L., Kehew, A.E., Bird, B.C., 2005. Outburst flood origin of the Central Kalamazoo River Valley, Michigan, USA. Quaternary Science Reviews 24, 2354-2374.

Krzyszkowski, D., Karrow, P.F., 2001. Wisconsinan inter-lobal stratigraphy in three quarries near Woodstock, Ontario. Géographie physique et Quaternaire 55, 3-22.

Laenen, A., Scott, K.M., Costa, J.E., Orzol, L.L., 1987. Hydrologic hazards along Squaw Creek from hypothetical failure of the glacial moraine impounding Carver Lake near Sisters, Oregon. U.S. Geological Survey Open-File Report 87-41, 48 p.

Larson, G., Schaetzl, R., 2001. Origin and evolution of the Great Lakes. Journal of Great Lakes Research 27, 518-546.

Leverett, F., Taylor, F.B., 1915. The Pleistocene of Indiana and Michigan and the history of the Great Lakes. U.S. Geological Survey Monograph 53.

Leverington, D.W., Teller, J.T., 2003. Paleotopographic reconstructions for the eastern outlets of glacial Lake Agassiz. Canadian Journal of Earth Sciences 40, 1259-1278.

Leverington, D.W., Teller, J.T., Mann, T., 2002. A GIS method for reconstruction of late Quaternary landscapes from isobase data and modern topography. Computers & Geosciences 28, 631–639.

Lewin, J., Macklin, M.G., 2003. Preservation potential for Late Quaternary river alluvium. Journal of Quaternary Science 18, 107-120.

Lewis, C.F.M., Anderson, T.W., 1985. Postglacial lake levels in the Huron Basin: Comparative uplift histories of the basins and sills in a rebounding glacial marginal depression. In: Karrow, P.F., and Calkin, P.E., (Eds.) Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 147-148.

Lord, M.L., Kehew, A.E., 1987. Sedimentology and paleohydrology of glacial-lake outburst deposits in southeastern Saskatchewan and northwestern North Dakota. Geological Society of America Bulletin 99, 663-673.

MacDonald, T.C., Langridge-Monopolis, J., 1984. Breaching characteristics of dam failures. Journal of Hydraulic Engineering, ASCE, 110, 567-586.

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

Malde, H.E., 1968. The catastrophic Late Pleistocene Bonneville Flood in the Snake River Plain, Idaho. United States Geological Survey Professional Paper 596, 52 p.

Mandrak, N.E., Crossman, E.J., 1992. Postglacial dispersal of freshwater fishes into Ontario. Candian Journal of Zoology 70, 2247-2259.

Marren, P.M., Toomath, S.C., 2014. Channel pattern of proglacial rivers: topographic forcing due to glacier retreat. Earth Surface Processes and Landforms 39, 943-951.

Mathews, W.H., Clague, J.J., 2000. The record of jokulhlaups from Summit Lake, northwestern British Columbia. Canadian Journal of Earth Sciences 30, 499-508.

Matsch, C., 1983. River Warren, the southern outlet of Glacial Lake Agassiz. In: Teller, J.T., Clayton, L. (Eds.), Glacial Lake Agassiz. Geological Association of Canada, Special Paper, vol. 26. St. Johns, NL, pp. 231-244.

Mier, J.M., Garcia, M.H., 2011. Erosion of glacial till from the St. Clair River (Great Lakes basin). Journal of Great Lakes Research 37, 399-410.

Morris, T.F., McAndrews, J.H., Seymour, K.L., 1993. Glacial Lake Arkona - Whittlesey transition near Leamington, Ontario: geology, plant, and muskox fossils. Canadian Journal of Earth Science 30, 2436-2447.

O'Connor, J.E., Webb, R.H., 1988. Hydraulic modeling for paleoflood analysis. In: Baker, V.R., Kochel, R.C., and Patton, P.C., (Eds.), Flood Geomorphology. John Wiley and Sons Inc., 393-402.

O'Connor, J.E., Baker, V.R., 1992. Magnitudes and implications of peak discharges from glacial Lake Missoula. Geological Society of America Bulletin 104, 267-279.

O’Connor, J.E., Beebee, R.A., 2009. Floods from natural rock-material dams. In: Burr, D.M., Carling, P.A., Baker, V.R. (Eds.), Megaflooding on Earth and Mars. Cambridge University Press, Cambridge, UK, pp. 128-171.

O’Connor, J.E., Clague, J.J., Walder, J.S., Manville, V., Beebee, R.A., 2013. Outburst floods. In: Shroder, J. (Ed.) Treatise on Geomorphology. AP, San Diego, 475-510.

Ontario Geological Survey., 1983. Aggregate resources inventory of London Township Middlesex County Southern Ontario. Aggregate Resources Inventory paper 55, Staff of the Engineering and Terrain Geology Section, Ontario Ministry of Natural Resources, 34 p.

Ontario Ministry of Natural Resources., 2003. Provincial Digital Elevation Model - Tiled Data Set, version 2.0.0, 10 m DEM, Geomatics Service Centre, Peterborough. Available in: http://www.lib.uwo.ca/madgic/geospatial/mnrdems_2005_data.htm

Pappenberger, F., Beven, K.J., Horritt, M.S., Blazkova, S., 2005. Uncertainty in the calibration of effective roughness parameters in HEC-RAS using inundation and downstream level observations. Journal of Hydrology 302, 46-69.

Patton, P.C., Baker, V.R., Kochel, R.C., 1979. Slack water deposits: a geomorphic technique for the interpretation of fluvial paleohydrology. In: Rhodes, D.D., Williams, G.P. (Eds.), Adjustments of the Fluvial System. Kendall/Hunt Publishing, Dubuque, IA, 225-253.

Peck , A., Prodanovic, P., Simonovic, P.P.S., 2012. Rainfall intensity duration frequency curves under climate change: City of London, Ontario, Canada. Canadian Water Resources Journal 37, 177-189.

Phillips, R.T.J., Robert, A., 2005. River incision in relation to post-glacial events in the Humber River basin, Ontario. Géographie physique et Quaternaire 59, 17-30.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low-relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Pierce, M.W., Thornton, C.I., Abt, S.R., 2010. Predicting peak outflow from breached embankment dams. Journal of Hydrologic Engineering, ASCE, 15, 338-349.

Pierson, T.C., 2005. Hyper-concentrated flow - transitional process between water flow and debris flow. In: Jakob, M., Hungr, O., (Eds.), Debris-Flow Hazards and Related Phenomena. Springer, Berlin, Chapter 8, 159-202.

Price, J.T., Lowe, G.W., Garrison, J.M., 1977. Unsteady flow modeling of dam-break waves. Proceedings of Dam Break Flood Routing Model Workshop, Bethesda, Maryland. U.S. Water Resources Council, Hydrology Committee, National Technical Information Service, Springfield, Virginia, PB-275-437.

Prowse, T.D. Beltaos, S., 2002. Climatic control of river-ice hydrology: a review. Hydrological Processes 16, 805-822.

Rathburn, S.L., 1993. Pleistocene cataclysmic flooding along the Big Lost River, east central Idaho. Geomorphology 8, 305-319.

Rayburn, J.A., Knuepfer, P.L.K., Franzi, D.A., 2005. A series of large, Late Wisconsinan meltwater floods through the Champlain and Hudson Valleys, New York State, USA. Quaternary Science Reviews 24, 2410-2419.

Rea, D.K., Moore, T.C., Jr, Anderson, T.W., Lewis, C.F.M., Dobson, D.M., Dettman, D.L., Smith, A.J., Mayer, L.A., 1994. Great Lakes paleohydrology: complex interplay of glacial meltwater, lake levels, and sill depths. Geology 22, 1059-1062.

Richardson, S.D., Reynolds, J.M., 2000. An overview of glacial hazards in the Himalayas. Quaternary International 65, 31-47.

Russell, A.J., 1993. Obstacle marks produced by flow around stranded ice blocks during a glacier outburst flood (jökulhlaup) in west Greenland. Sedimentology 40, 1091–1111.

Sado, E.V. Vagners, U.G. 1975. Quaternary Geology of the Lucan Area, Southern Ontario. Ontario Division of Mines, Preliminary Map p.1048. Geological Survey, scale 1:50000, Geology 1971, 1972.

Schomacker, A., 2010. Expansion of ice-marginal lakes at the Vatnajökull ice cap, Iceland, from 1999 to 2009. Geomorphology 119, 232-236.

Severinghaus, J.P. Brook, E.J. 1999. Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. Science 286, 930-934.

Shugar, D., Kostachuk, R., Ashmore, P., Desloges, J., Burges, L., 2007. In situ jet-testing of the erosional resistance of cohesive streambeds. Canadian Journal of Civil Engineering 34, 1192-1195.

Singh, K.P., Snorrason, A., 1984. Sensitivity of outflow peaks and flood stages to the selection of dam breach parameters and simulation models. Journal of Hydrology 68, 295-310.

Smith, D.G., Pearce, C.M., 2002. Ice jam-caused fluvial gullies and scour holes on northern river flood plains. Geomorphology 42, 85-95.

Smith, N.D., Ashley, G.M., 1985. Proglacial lacustrine environment. short course 16, Tulsa, OK. In: Ashley, G.M., Shaw, J., Smith, N.D. (Eds.), Glacial Sedimentary Environments, Society of Economic Paleontologists and Mineralogists, 135-216.

Soil Conservation Service., 1981. Simplified dam-breach routing procedure. U.S. Department of Agriculture, Technical Release 66, Washington, D.C. 25 p.

Soil Conservation Service., 1986. A study of predictions of peak discharge from a dam breach. Engineering Division, Natural Resources Conservation Service, U.S. Department of Agriculture, Washington, D.C.

Staines, K.E.H., Carrivick, J.L., 2015. Geomorphological impact and morphodynamic effects on flow conveyance of the 1999 jökulhlaup at sólheimajökull, Iceland. Earth Surface Processes and Landforms 40, 1401-1416.

Stewart, A.M., Desloges, J.R., 2014. A 9000-year record of vertical and lateral accretion on the floodplain of the lower Thames River, southwestern Ontario, Canada, and implications for archaeological research. Quaternary International 342, 161-172.

Stewart, R.A., 1982. Glacial and glaciolacustrine sedimentation in Lake Maumee III near Port Stanley, southwestern Ontario. PhD dissertation, Department of Geology, University of Western Ontario, 350 p.

Strasser, M., Schindler, C., Anselmetti, F.S., 2008. Late Pleistocene earthquake-triggered moraine dam failure and outburst of Lake Zurich, Switzerland. Journal of Geophysical Research 113, F02003.

Tahershamsi, A. Ponce, V.M., Shetty, A.V., 2003. Embankment dam breaching: Geometry and peak outflow characteristics. Journal of Dam Engineering 14, 73-82.

Teller, J.T., Kehew, A.E., 1994. Introduction to the late glacial history of large proglacial lakes and meltwater runoff along the Laurentide Ice Sheet. Quaternary Science Reviews 13, 795-799.

Terasmae, J., Karrow, P.F, Dreimanis, A. 1972. Quaternary stratigraphy and geomorphology of the eastern Great Lakes region of southern Ontario. XXIV International Geological Congress Guidebook, Excursion A42.

Thieme, D.M., 2003. Archaeological site formation in glaciated settings, New Jersey and southern New York. In: Cremeens, D.L., Hart, J.P. (Eds.), Geoarchaeology of Landscapes in the Glaciated Northeast, Bulletin 497. New York State Museum, New York State Education Dept., Albany, New York, pp. 163-179.

Thornthwaite and Associates., 1964. Publications in Climatology 17(2). Technical Report No. 6 NSF Contract C266 Centerton NJ.

Thornton, C.I., Pierce, M.W., Abt, S.R., 2011. Enhanced predictions for peak outflow from breached embankment dams. Journal of Hydrologic Engineering, ASCE, 16, 81-88.

Tinkler, K.J., Parrish, J., 1998. Recent adjustments to the long profile of Cook Creek, an urbanized bedrock channel in Mississauga, Ontario. In: Tinkler, K.J., and Wohl, E.E., (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union Geophysical Monograph 107, Washington, D.C., 167-188.

Totten, S.M., 1985. Chronology and nature of the Pleistocene beaches and wave-cut cliffs and terraces, northeastern Ohio. In: Karrow, P.F., and Calkin, P.E., (Eds.) Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, 171-184.

Trieste, D.J., Jarrett, R.D., 1987. Roughness coefficients of large floods. In: James, L., and English, M., (Eds.), Proceedings of the Irrigation and Drainage Division of the American Society of Civil Engineers and the Oregon Section. ASCE, Portland, Oregon, 32-40.

Upper Thames River Conservation Authority., 2015. Fanshawe Dam & Reservoir. Available in: http://thamesriver.on.ca/water-management/flood-control-structures/fanshawe-dam. Last accessed 13 April, 2015.

U.S. Bureau of Reclamation (USBR)., 1988. Downstream hazard classification guidelines. ACER Tech. Memorandum No. 11, U.S. Department of the Interior, Denver, 57.

Vilímek, V., Zapata, M.L., Klimeŝ, J., Patzelt, Z., Santillân, N., 2005. Influence of glacial retreat on natural hazards of the Palcacocha Lake area, Peru. Landslides 2, 107-115.

Wahl, T.L., 1998. Prediction of embankment dam breach parameters: a literature review and needs assessment. Dam Safety Research Rep. DSO-98-004, U.S. Bureau of Reclamation, 60 p.

Wahl, T.L., 2004. Uncertainty of predictions of embankment dam breach parameters. Journal of Hydraulic Engineering, ASCE, 130, 389-397.

Waitt, R.B., 1985. Case for periodic, colossal jökulhlaups from Pleistocene glacial Lake Missoula. Geological Society of America Bulletin 96, 1271-1286.

Walder, J.S., O'Connor, J.E., 1997. Methods for predicting peak discharge of floods caused by failure of natural and constructed earthen dams. Water Resources Research 33, 2337-2348.

Waythomas, C. F., Jarrett, R.D., 1994, Flood geomorphology of Arthurs Rock Gulch, Colorado: Paleoflood history. Geomorphology 11, 15-40.

White, G.W., 1982. Glacial geology of Northeastern Ohio. Ohio Geological Survey Bulletin 68, 75 p.

Whittaker, W.A., 1986. Till geochemistry and glacial geology of Lower Medway Valley, London, Ontario. Unpublished senior thesis, Department of Geology, University of Western Ontario, 166 p.

Wilcock, P.R., 1992. Flow competence: Criticism of a classic concept. Earth Surface Processes and Landforms 17, 289-298.

Williams, G.P., 1983. Paleohydrological methods and some examples from Swedish fluvial environments, I: Cobble and boulder deposits. Geografiska Annaler 65A, 227-243.

Williams, G.P., Costa, J.E., 1988. Geomorphic measurements after a flood. In: Baker, V.R., Kochel, R.C., and Patton, P.C., (Eds.), Flood Geomorphology. John Wiley and Sons Inc., 65-77.

Winder, G., 1980. Origin of the Medway Valley revealed. The Medway Valley News, Winter 1979-1980, 4-6.

Wohl, E.E., 1998. Uncertainty in flood estimates associated with roughness coefficients. Journal of Hydraulic Engineering, ASCE, 124, 219-223.

Wolman, M.G., 1954. A method of sampling coarse bed material. Eos Transactions, AGU 35, 951-956.

Worni, R., Stoffel, M., Huggel, C., Volz, C., Casteller, A., Luckman, B., 2012. Analysis and dynamic modeling of a moraine failure and glacier lake outburst flood at Ventisquero Negro, Patagonian Andes (Argentina). Journal of Hydrology 444-445, 134-145.

Worni, R., Huggel, C., Clague, J.J., Schaub, Y., Stoffel, M., 2014. Coupling glacial lake impact, dam breach, and flood processes: A modeling perspective. Geomorphology 224, 161-176.

Wu, W., Altinakar, M.S., Al-Riffai, M., Bergman, N., Bradford, S.F., Cao, Z., Chen, Q.J., Constantinescu, S.G., Duan, J.G., Gee, M.D., Greimann, B., Hanson, G., He, Z., Hegedus, P., Van Hoestenberghe, T., Huddleston, D., Hughes, S.A., Imran, J., Jia, Y., Jorgeson, J.D., Kahawita, R., Klumpp, C.C., Lai, Y., Langendoen, E.J., Liu, S., Moreda, F., Morris, M., Morvan, H., Orendorff, B., Pak, J., Peeters, P., Reed, S., Sanders, B.F., Scott, S.H., Soares-Frazao, S., Song, C.R., Sutherland, J., Teal, M.J., Tsubaki, R., Wahl, T.L., Weston, D.M., Williams, D.T., Zech, Y., Zhang, L., 2011. ASCE/EWRI Task Committee on Dam/Levee Breaching. Earthen embankment breaching. Journal of Hydraulic Engineering 137, 1549-1564.

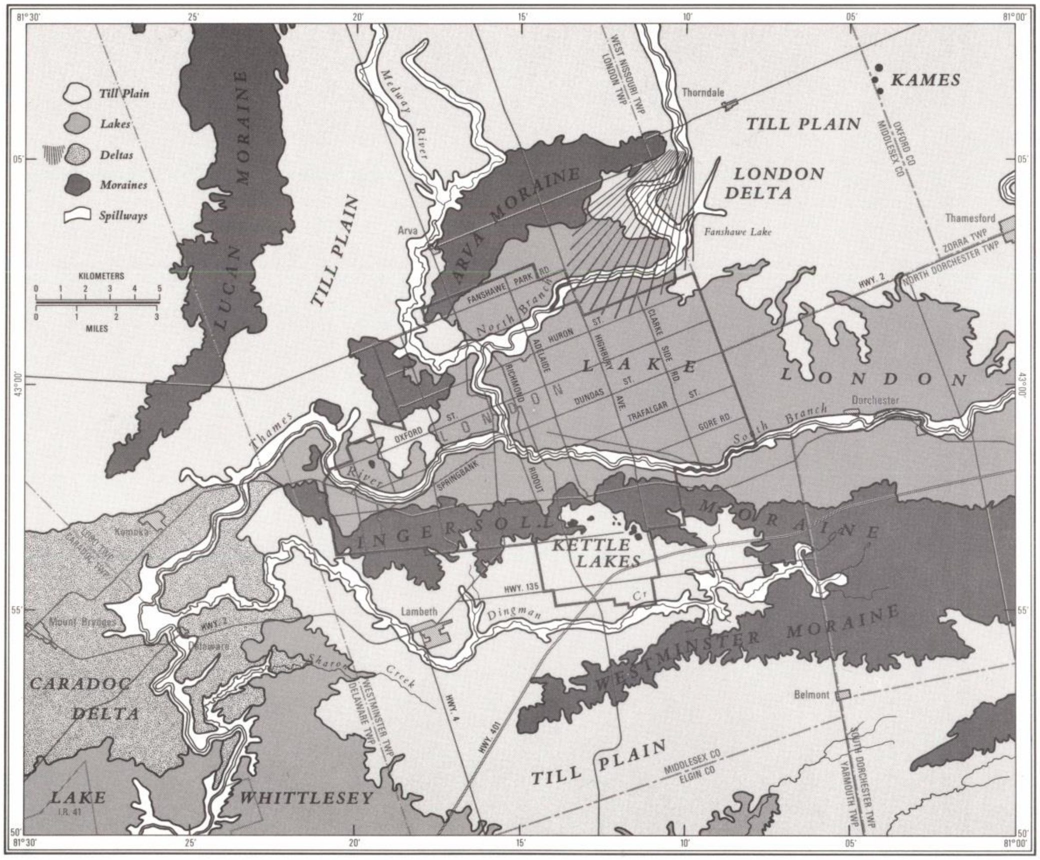
Yu, Z., 2000. Ecosystem response to Lateglacial and early Holocene climate oscillations in the Great Lakes region of North America. Quaternary Science Reviews 19, 1723-1747.

Yu, Z., 2003. Late Quaternary dynamics of tundra and forest vegetation in the southern Niagara Escarpment, Canada. New Phytologist 157**,** 365-390.

Figures and Tables

Figure 1. Location of Thames River watershed, major cities and locations mentioned in text. Major moraines within the watershed are indicated: 1. Arva Moraine, 2. Ingersoll Moraine and adjoining Westminister Moraine to the south, 3. St. Thomas Moraine, 4. Blenheim Moraine and 5. Charing Cross Moraine. Adapted from Chapman and Putnam (1984) and Goff and Brown (1981).

Figure 2. An early drawing of the city of London Ontario from 1893. View is from south to north. Thames River tributaries and main stem incise into lacustrine sediments of former Glacial Lake London. Arva Moraine is in the background. Artist was probably sitting on a high point of the Ingersoll Moraine. Source: Toronto Lithographing Company.



THE GLACIAL LANDSCAPE OF LONDON

Figure 3. The glacial landscape of London, Ontario by Dreimanis et al. (1998). Reproduced with permission of Geological Association of Canada.

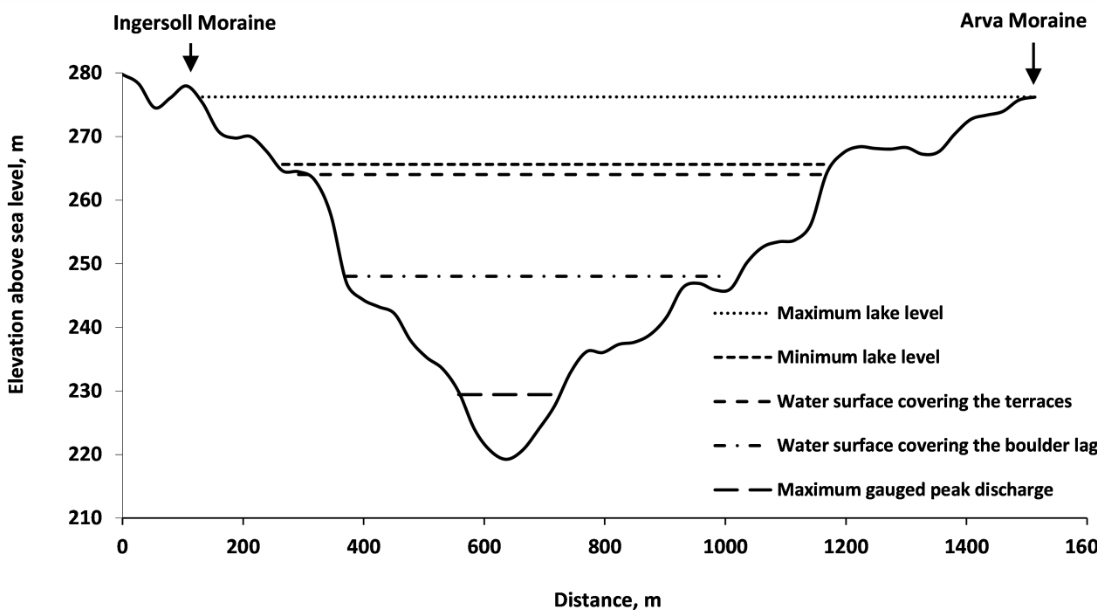


Figure 4. The cross-section of the breach of Glacial Lake London where the Arva and Ingersoll moraines were connected near the London neighborhood of Byron and associated water levels of the lake (see text), downstream geomorphic features (see text) and gauged hydrology. The difference between the top and bottom water surface elevations was used as the height of the moraine dam (H_w = 46.8 m) in the parametric breaching equations.

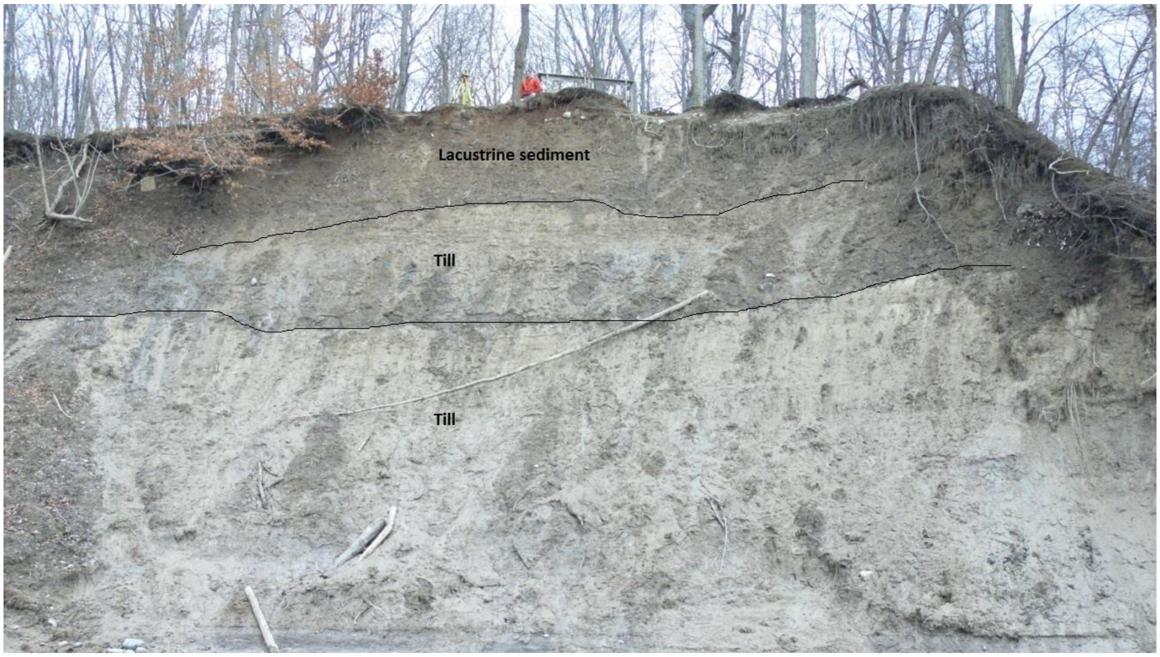


Figure 5. Arva Moraine tills and overlying lacustrine sediments of Glacial Lake London on a right bank bluff of Medway Creek, next to Brescia University College, London. This is the highest known deposit of the lake bed at 265.6 m. Note that lacustrine sediment does not show evident stratigraphic layering. View from north to south. Yannick Rousseau peeking on top of bluff for scale - bluff is 18 m high above river bed.

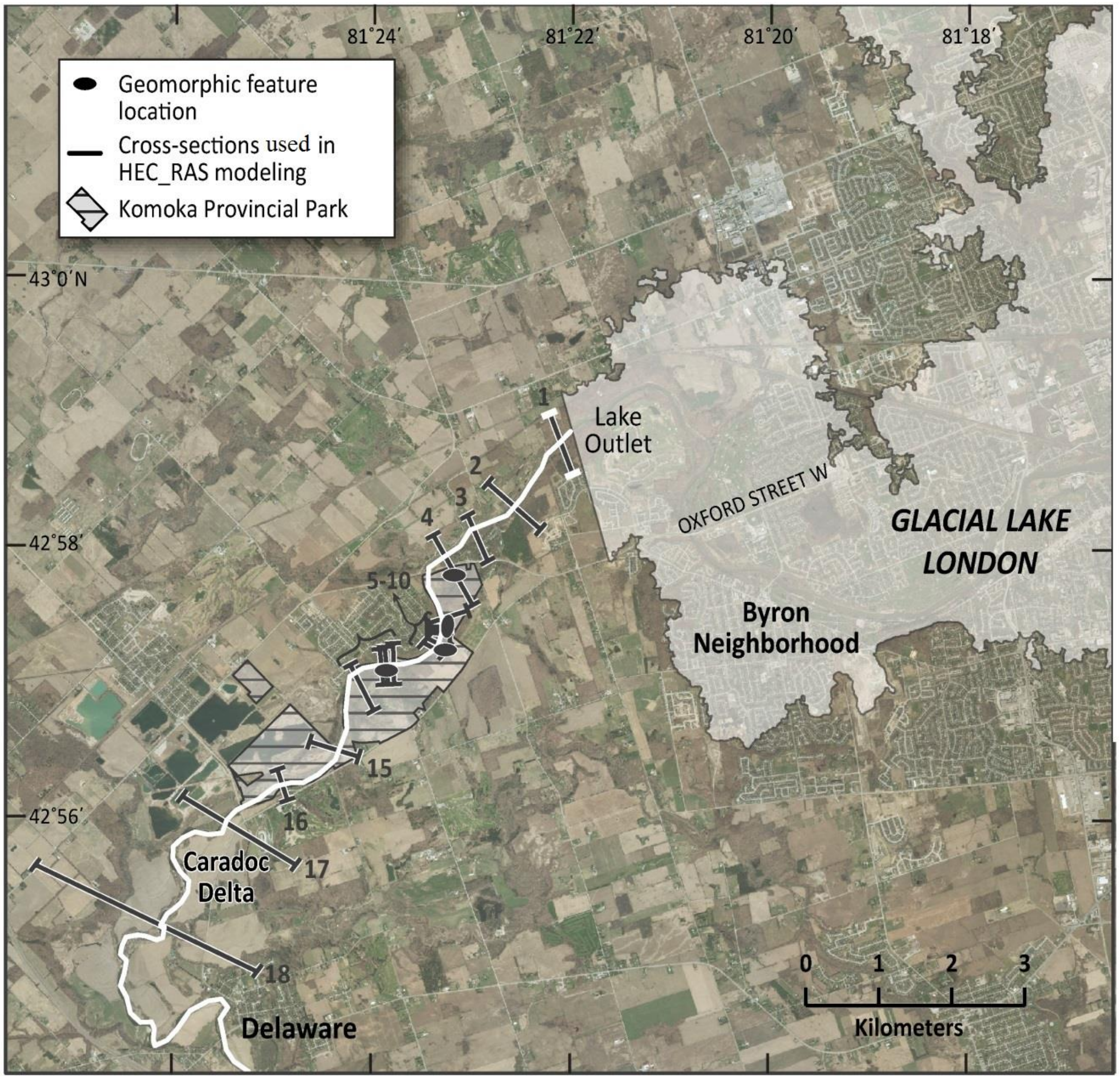


Figure 6. SW part of Glacial Lake London imposed on an aerial photograph, Thames River spillway channel into Komoka Provincial Park, Caradoc Delta at bottom left and cross-sections used in HEC-RAS modeling (black dots indicate location of landforms).

Figure 7. Some landforms in Komoka Provincial park (a) Top of valley bluff (view downstream), (b) Till bank of Thames River (view upstream), (c) Boulder lag beginning more than 150 m away from Thames River channel (d) Pointbar of Thames River with boulders up to 600 mm, (e) Wash Terrace 3 with featureless topography (view upstream), (f) Paleo erosional flow mark scar at the forefront of the photograph (delineated by a black line) exposed from vegetation well above modern channel floods on the valley wall, (g) Longitudinal Berm 1 with grassy top and (h) Mound 1 (view downstream).

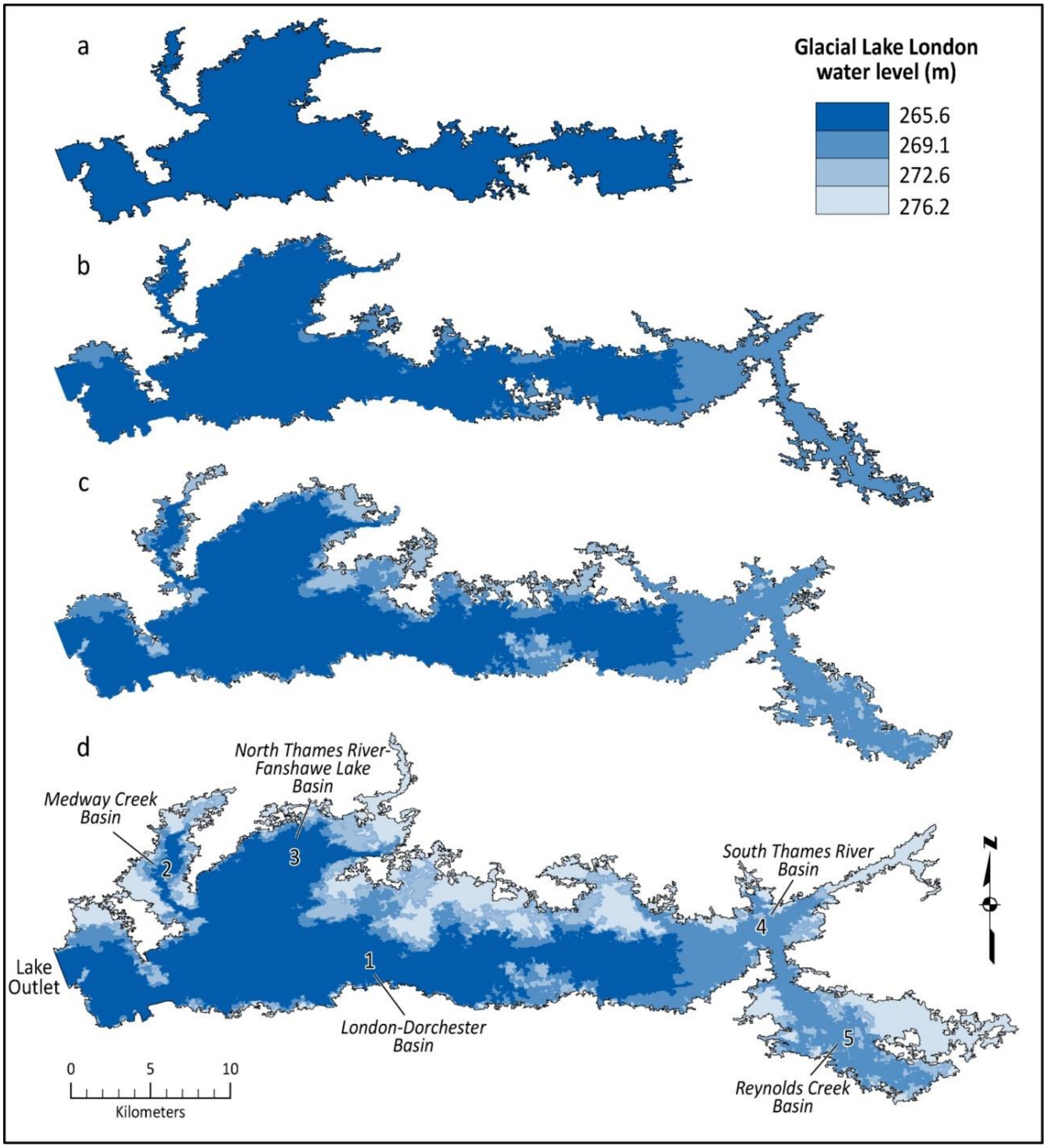


Figure 8. Reconstruction of Glacial Lake London. (a) smallest reconstruction, based on lacustrine stratigraphic evidence from Medway Creek; (d) largest reconstruction, based on highest elevation of Arva and Ingersoll Moraines' convergence where the breach occurred; (b, c) lake elevations at 33% and 67% between minimum and maximum levels.

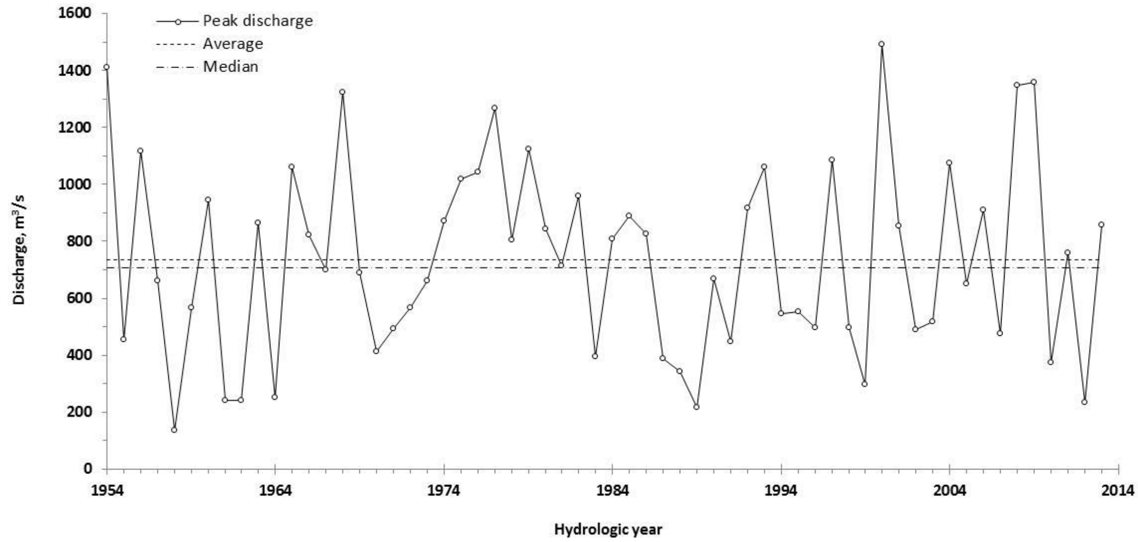


Figure 9. Reconstructed non-attenuated peak flows for Thames River in London during six decades (1954-2013). Average and median are 735 m^3/s and 707 m^3/s, respectively. Highest peak flow (1493 m^3/s) occurred in 2000.

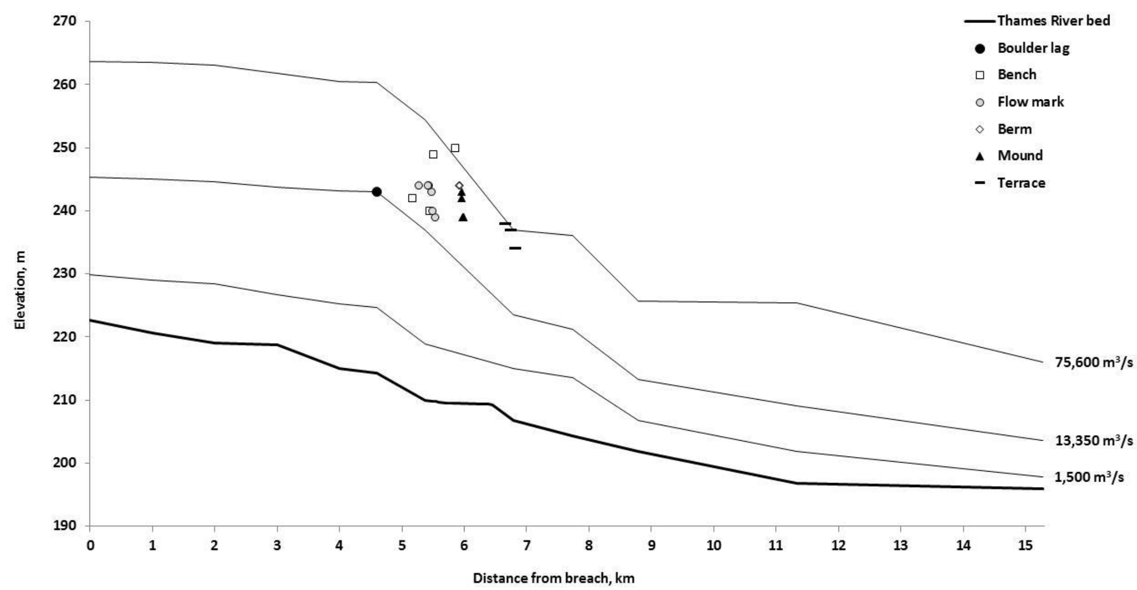


Figure 10. Calculated water surface profiles of Thames river flood course based on HEC-RAS modeling for modern peak discharge and minimum (boulder lag) and maximum (terraces) paleo discharges.

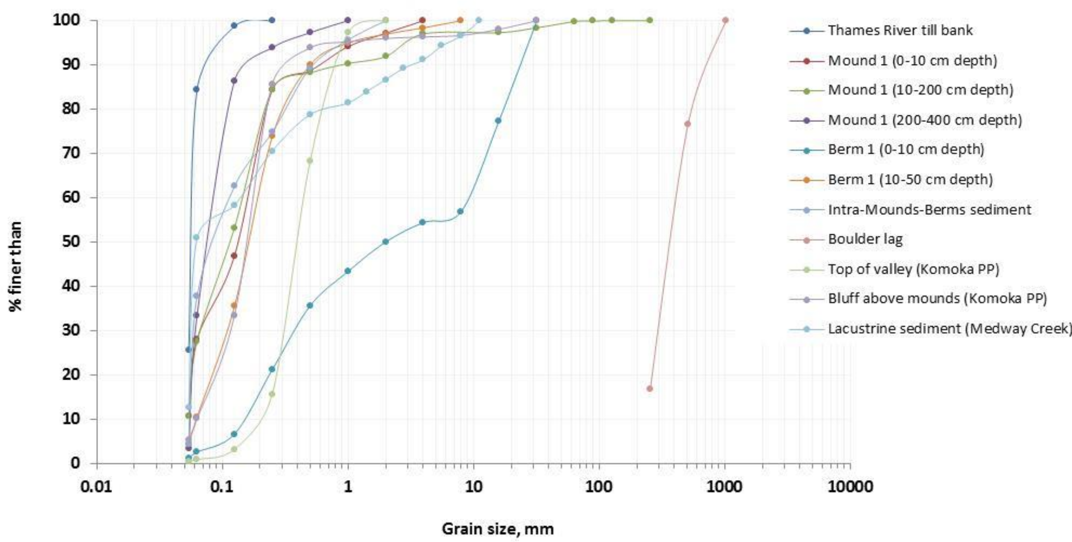


Figure 11. Grain size distributions of various landforms within Komoka Provincial Park, illustrating the diverse nature of sediments along the flood course. Lacustrine sediment from Medway Creek's highest bluff (265.6 m) was added. Note that coarse boulder lag is completely devoid of grain size fractions < 256 mm but this may be due to post-flood erosion and reworking of the deposit in this humid continental climate.

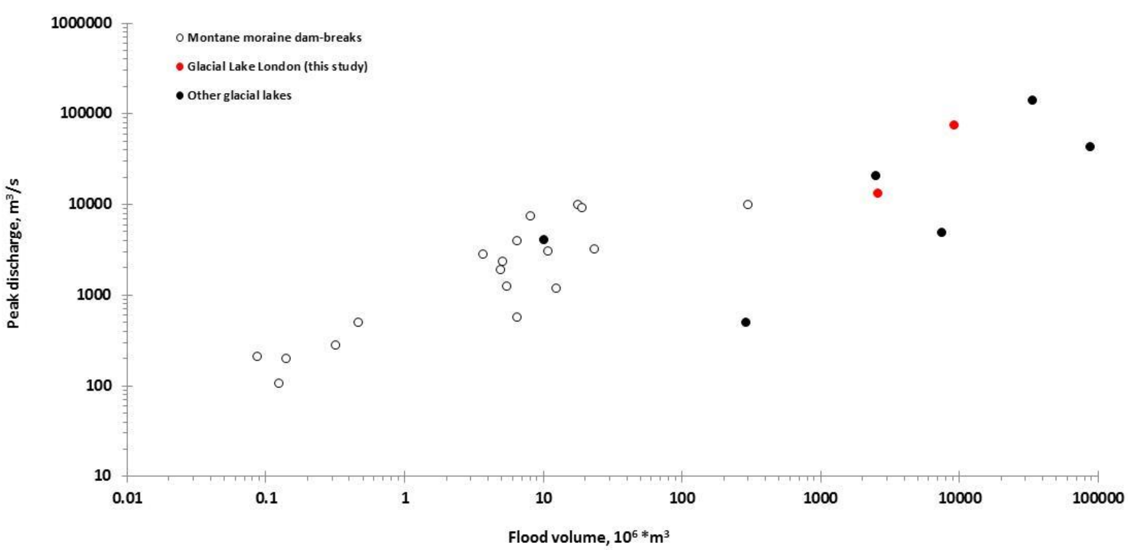


Figure 12. Peak discharges of dam-break floods of montane moraine-dammed lakes and ice-marginal lakes of the Laurentide Ice Sheet and their flood volumes. Glacial Lake London has lower and upper peak discharges from modeling vs. minimum and maximum lake volumes.

Table 1: Paleohydrographic characteristics of Glacial Lake London.

Elevation, m asl	Lake surface area, km^2	Lake perimeter, km	Maximum depth, m	Average depth, m	Lake volume, km^3
265.6[a]	130.0	340.6	36.9	19.9	2.6
269.1	191.1	613.0	40.4	23.1	4.4
272.6	242.5	690.3	43.9	23.8	5.8
276.2[b]	339.7	815.3	47.4	26.8	9.1

Notes: a. Lower bound: elevation of lacustrine sediments on top of Arva Moraine

b. Upper bound: elevation of overtopping of Ingersoll and Arva Moraines

Table 2: Summary of parametric breach equations for Glacial Lake London.

#	Reference/Source	Relation proposed for peak discharge, Q_p	Q_p for Glacial Lake London*, m^3/s	Comments
1	**Kirkpatrick (1977)**	$Q_p = 2.297(H_w+1)^{2.5}$	19,300	Transformed from cubic feet per second.
2	**Price et al. (1977)**	$Q_p = 8/27g^{0.5}Hw^{1.5}(0.4b+0.6T)$	44,100	g is the acceleration due to gravity, b is width of breach base, T is top width of breach at initial water level. Ignores frictional and turbulent resistance and therefore tends to be larger than slope-area and drawdown rate methods.
3	**Soil Conservation Service (1981)**	$Q_p = 65H_w^{1.85}$	20,400	Based on 13 dam failures. Transformed from cubic feet per second.
4	**Hagen (1982)**	$Q_p = 370(H_wV_{max})^{0.50}$	23,300	Based on seven dams, transformed from cubic feet per second.
5	**MacDonald and Langridge-Monopolis (1984)**	$Q_p = 1.175\ (V_{max}\ H_w)^{0.41}$	68,900	Includes concrete dam failures which are larger than embankment dams so probably overpredicts peak flows for earth dams.
6	**Singh and Snorrason (1984)**	$Q_p = 1.776V_{max}^{0.47}$	85,200	Based on 20 dam failures and 8 simulations. Q_p based on the

				simulations.
		$Q_p = 13.4H_w^{1.89}$	19,200	
7	**Costa (1985)**	$Q_p = 10.5H_w^{1.87}$	13,900	r^2=0.8 and standard deviation error is 82%.
		$Q_p = 325(H_wV_{max})^{0.42}$	75,200	Regression equation using 29 dam failures, r^2=0.75 and standard deviation error is 95%.
8	**Soil Conservation Service (1986)**	$Q_p = 0.00042\ B_r^{1.35}$ where $B_r = V_{max}H_w/A$	36,300	B_r is the breach factor and A is the cross-sectional area of the embankment at the breach water surface elevation. The discharge range is $1.77\ H_w^{2.5} < Qp < 16.6\ H_w^{1.85}$ or 26,521-248,727 m^3s^{-1}.
9	**Evans (1986)**	$Q_p = 0.72\ V_{max}^{0.53}$	136,700	Based on the relationship proposed by Clague and Mathews (1973) for jökulhlaups.
10	**U.S. Bureau of Reclamation (1988)**	$Q_p = 19.1\ H_w^{1.85}$	23,500	This is an envelope equation based on 21 case studies.
11	**Froehlich (1995)**	$Q_p = 0.607\ V_{max}^{0.295}\ H_w^{1.24}$	62,000	Provides good agreement between measured and computed peak outflows over entire range of values according to the author.
12	**Walder and O'Connor (1997)**	$Q_p = 1.16V_{max}^{0.46}$	44,200	

		$Q_p = 2.5d^{2.34}$	20,200	*d* is drop in lake level, assumed as dam height
		$Q_p = 0.61\ (H_w V_{max})^{0.43}$	61,100	
13	**Pierce et al. (2010)**	$Q_p = 0.784\ H_w^{\ 2.688}$	24,200	Linear
		$Q_p = 2.325\ \ln H_w^{\ 6.405}$	13,000	Curvelinear
		$Q_p = 0.00919\ V_{max}^{0.745}$	241,400	
		$Q_p = 0.0176\ (H_w V_{max})^{0.606}$	196,300	
		$Q_p = 0.038\ V_{max}^{0.475} H_w^{1.09}$	135,200	

[*] H_w is dam height, V_{max} is reservoir's maximum volume

[**] Values used for Glacial Lake London: H_w = 46.8 m (153.5 ft), V_{max} = 9100*10^6 m^3

Table 3: Roughness value combinations and resultant peak discharge variables for Glacial Lake London drainage using HEC-RAS model simulations.

Boulder Lag

HEC-RAS	**Manning's *n* roughness**		**Peak discharge**	**Max velocity**	**WSS**	**Max shear stress**	**Max stream power**	**Froude #**
Simulation	**Channel**	**Banks**	**m^3/s**	**m/s**		**N/m^2**	**W/m^2**	**range**
1	0.04	0.06	14,000	12.8	0.0026	1140	680	0.1-0.5
2	0.05	0.07	13,800	10.0	0.0027	790	660	0.1-0.7
3	0.06	0.08	13,600	10.0	0.0027	790	640	0.1-0.6
4	0.07	0.09	13,400	10.0	0.0025	780	580	0.1-0.5
5	0.08	0.10	13,100	8.7	0.0028	650	610	0.1-0.8
6	0.09	0.11	12,300	7.4	0.0028	510	570	0.1-0.7
7	0.10	0.12	11,500	6.5	0.0028	420	530	0.1-0.6

Terraces

HEC-RAS	**Manning's roughness**		**Peak discharge**	**Max velocity**	**WSS**	**Max shear stress**	**Max stream power**	**Froude #**
Simulation	**Channel**	**Banks**	**m^3/s**	**m/s**		**N/m^2**	**W/m^2**	**range**
1	0.04	0.06	145,000	19.5	0.0028	2160	1810	0.2-0.8

2	0.05	0.07	121,400	16.3	0.0028	1666	1600	0.2-0.8
3	0.06	0.08	94,500	12.7	0.0030	1160	1430	0.2-0.8
4	0.07	0.09	76,700	11.1	0.0031	956	1290	0.1-0.8
5	0.08	0.10	63,700	9.8	0.0031	796	1190	0.1-0.7
6	0.09	0.11	61,600	9.0	0.0032	703	1140	0.1-0.6
7	0.10	0.12	49,100	7.8	0.0032	568	1010	0.1-0.6

Table 4: 21 geomorphic features found on the left bank in Komoka Provincial Park.

#	Feature	UTM	Easting	Northing	Elevation m	Degree of certainty	Discharge m^3/s
1	Boulder lag	17	468533	4756848	243	High	13,350
2	Bench 1	17	468453	4756033	249	Medium	50,000
3	Bench 2	17	468437	4755888	250	Medium	85,000
4	Bench 3	17	468448	4756090	240	Medium	18,000
5	Bench 4	17	468431	4756350	242	Medium	18,000
6	Flow mark 1	17	468394	4756011	239	Medium	18,200
7	Flow mark 2	17	468422	4756050	240	Medium	19,000
8	Flow mark 3	17	468417	4756052	243	Medium	25,300
9	Flow mark 4	17	468440	4756105	244	Medium	25,800
10	Flow mark 5	17	468426	4756100	244	Medium	25,800
11	Flow mark 6	17	468433	4756248	244	Medium	21,500
12	Berm 1	17	468395	4755832	244	Low	50,000
13	Berm 2	17	468358	4755804	244	Low	50,000
14	Mound 1	17	468332	4755799	243	Low	46,700
15	Mound 2	17	468343	4755792	242	Low	43,000
16	Mound 3	17	468328	4755791	239	Low	30,000
17	Mound 4	17	468321	4755794	239	Low	30,000
18	Mound 5	17	468310	4755780	239	Low	30,000
19	Terrace 1	17	467525	4755528	234	High	55,000
20	Terrace 2	17	467594	4755537	237	High	75,600
21	Terrace 3	17	467689	4755534	238	High	75,600

Semi-alluvial stream channel crossing the interlobate Arva Moraine, Medway Creek, Southern Ontario, Canada. 1: Sedimentary characteristics and morphologic change

Nathaniel Bergman[1, 2, 3], Marco J. Van De Wiel[1, 4], Stephen R. Hicock[5]

1. Department of Geography, Western University, Canada
2. Yigal Alon Kinneret Limnological Laboratory, Israel Oceanographic and Limnological Research, Israel
3. Department of Geography and Environmental Studies, University of Haifa, Israel
4. Centre for Agroecology, Water and Resilience, Coventry University, United Kingdom
5. Department of Earth Sciences, Western University, Canada

Abstract

We describe a detailed study of an incisional till-cobble-bed river reach of a small channel crossing a moraine, including the boundary till characteristics, bare till patch features, and for the first time their annual erosion rates, bedform dimensions and spacing, and grain size distributions of bedforms. Results show that the boundary till is predominantly fines (silt and clay) with some sand, while gravel contribution to the alluvium is relatively small. There is evidence of some clast *in situ* and transport rounding. Till exposures constitute a relatively small portion of the bed and till erosion rates are relatively high compared to bedrock rivers although highly variable between till patches and within patches. Bedforms consist of flats, pools, riffles, bars and a couple of cobble-boulder steps. The bedforms are not well organized in terms of spacing and show high morphologic variability. The sediment forming the bed is poorly sorted, and grain sizes of the bedforms show high variability ranging from fines to large boulders beyond channel competence. As expected, riffles and steps are coarser while flats and pools are finer-grained.

Sedimentary stability metrics show that riffles are unstable, while pools and flats are more stable. Grain size distributions of three additional fine-rich rivers demonstrate bimodal or polymodal distributions, with poorly sorted beds, suggesting all these channels can be defined as unstable. We conclude that till-bedded channels differ from their alluvial and bedrock counterparts in a variety of ways. Consequently, we recommend that semi-alluvial rivers be differentiated from their alluvial and bedrock counterparts in future channel classifications. Such a practice will not only be holistic in nature, but will also be useful for the river research and practitioners' community to gain the appropriate research tools needed for assessment, management and restoration practices for these rivers.

1. Introduction

Channels incised into till are a characteristic of the landscape in southern Ontario (Campo and Desloges, 1994; Foster, 1998; Thayer, 2010; Hrytsak, 2012; Phillips and Desloges, 2014; 2015a, b: Thayer et al., 2016) and the Great Lakes region in general (Hack, 1965; Arbogast et al., 2008). Further, the large areal extent of the continental Laurentide Ice Sheet (LIS) during the last Wisconsinan glaciation (Clark et al., 1993) and its consequent melting (Teller, 1990) exposed glacial sediments in many Atlantic watersheds (Brakenridge et al., 1988; Marchand et al., 2014), northern-draining Hudson Bay rivers (Nielsen et al., 1986), and northern Mississippi-Missouri drainage watercourses (Bhowmik, 1979; Gran et al., 2009, Stout et al., 2014), which many contain in their underlying bed and bank boundaries. A semi-alluvial channel incised into till can be defined as “a channel that cannot substantially widen, lower or shift its bed without eroding till”, an analogue borrowed from the bedrock literature (e.g. Turowski et al., 2008; Meshkova et al., 2012). Once alluvium enters a till channel it becomes a mixed channel setting, i.e. a till-alluvial system, if the channel is either till constrained (the bed is till), till confined (the banks are till), or both. Again, the definition is adapted from the bedrock literature (Meshkova et al., 2012).

These till-bedded rivers initially evolved after the retreat of the LIS following the end of the last glaciation and are still responding to base level fall (and sometimes rise) of the Great Lakes (Thornbush and Desloges, 2011), the Mississippi River (Belmont, 2011; Shen et al., 2012; Gran et al., 2013; Wickert et al., 2013), Hudson Bay (Fraser et al., 2005) and the Atlantic Ocean (Clark and Fitzhugh, 1992), in addition to glacial-isostatic rebound of the landscape from the ice sheet loading (Lewis et al., 2005) and water deformation (i.e. the large water mass also deforms the crust; Clark et al., 2007). Furthermore, modern human-induced climate change (Ashmore and Church, 2001; Kling et al., 2003; Novotny and Stefan, 2007) and drastic land modification (Annable et al., 2011; 2012; Woltemade, 1994; Campo and Desloges, 1994; Miller and Nudds, 1996; Fitzpatrick et al., 1999; Belmont et al., 2011) have made these streams much more erosive than in the past because of increasingly flashy discharge from urban land surfaces and quick agricultural drainage with minimal water retention (Knox, 1989; Schottler et al., 2013), adding a major human stressor to the landscape. In many cases, these streams no longer resemble natural watercourses, and have altered geometries, low water quality, no floodplain, high sediment loads (Dickinson and Green, 1988), and tile drainage. Many do not function according to common geomorphic principles, such as forming discharge and bankfull discharge that link morphology to process (Powell, 2006). The expected fluvial outcome when the hydrology, climate and sediment delivery ratios are altered is geomorphic instability that is often expressed in extensive channel widening (i.e. bank collapse) and deep incision of the beds. However, the geomorphic instability is not always conclusive in southern Ontario (Campo and Desloges, 1994; Annable et al., 2012) and the American Midwest (Doyle et al., 2000; Stout et al., 2014), as sediment sources of these channels are highly variable and site-specific (Wilkin and Hebel, 1982; Wilcock et al., 2009). Nevertheless, under certain circumstances, when bed alluviation is sufficient, semi-alluvial channels may behave as regular alluvial rivers as was shown in Irvine Creek, a bedrock tributary of the Grand River, Ontario (Martini, 1977).

Compared to the vast literature on alluvial and bedrock river, the understanding of how semi-alluvial channels incised into till (or other glacial deposits) function and evolve is quite limited. Other than the limited literature directly investigating them, information has been derived from studies conducted for a different purpose, such as rivers with some kind of substrate forcing (MacVicar and Roy, 2011), channelization (Landwehr and Rhoads, 2003; Ward et al., 2008), habitat quality (Wang et al., 1997; Blann et al., 2009; Marchildon et al., 2011), bedforms (Hartley, 1999), archeology (Stewart and Desloges, 2014) or sediment contamination (Rhoads and Cahill, 1999; Wilcock et al., 2009). Direct work focusing on the subject of glacial legacy on river/valley form was recently done in mountainous terrains (Brardinoni and Hassan, 2006; 2007; Amerson et al., 2008; Addy et al., 2011; Gomez and Livingstone, 2012; Prasicek et al., 2014; Addy et al., 2014; Hassan et al., 2014), and occasionally in lowland alluvial settings (Collins and Montgomery, 2011; Phillips and Desloges, 2014, 2015a, 2015b).

Recent anthropogenic change has resulted in altered hydrology which leads to more frequent extreme flood events (Diffenbaugh et al., 2005), changes in sediment delivery rates (Syvitski and Milliman, 2007; Belmont et al. 2011) and modifications in land use (Taylor, 1977; Matson et al., 1997; Knox, 2006; Ellis, 2011). Hence, there is a need to establish a systematic, scientific understanding of semi-alluvial rivers, similar to our understanding of alluvial (Bridge, 2003) and bedrock channels (Tinkler and Wohl, 1998a, b) and within a framework similar to the channel classification proposed by Montgomery and Buffington (1997) for mountain streams, by Turowski et al. (2008) for bedrock channels and Sutfin et al. (2014) for ephemeral rivers.

New till-bedded rivers are currently evolving around the world, as new landmass is deglaciated and exposed to fluvial erosion processes as the climate warms. However, in urban and suburban areas the relative lack of scientific understanding of how these streams operate negatively affects management, modern conservation practices and river restoration techniques (Hassan et al., 2014). Using the controversial Natural Channel Design (NCD; Rosgen, 1996), often applied in alluvial river restoration practices in North America (Wilcock, 2012), to semi-alluvial till channels may yield disastrous results that do not reflect natural form and process of these streams (Ness and Joy, 2002; Geomorphic Solutions, 2009). These results include unnatural stability as a result of using very large boulders to stabilize the channel bed and banks (i.e., boulder armoring), erroneous use of bankfull discharge values and a mismatch between design (form) and hydrological and sedimentary processes (Ness and Joy, 2002). Champoux et al. (2003) investigated long-term restoration of a small stream in Wisconsin and found that the morainic section of the channel is highly unstable (i.e. rapid bank collapses and bed incision), further highlighting that restoration schemes should be based on local geomorphic context rather than on an unsuitable restoration design.

This work is intended to add new information by exploring contemporary channel form and processes in a formerly glaciated terrain and comparing it to other semi-alluvial, alluvial and bedrock channels. Ordinarily such studies were done on a large basin-scale (Hack, 1965; Campo and Desloges, 1994; Phillips and Robert, 2005; Arbogast et al., 2008; Wilcock et al., 2009; Gran et al., 2009; 2011; 2013; Belmont, 2011; Belmont et al., 2011; Stout et al., 2014; Phillips and Desloges, 2014; Hassan et al., 2014; Phillips and Desloges, 2015a, b), but our work focuses on a smaller scale, i.e. a 1.5 km reach of a channel crossing a till-cored moraine and affected by intense upstream agricultural practices, local urbanization and an historic millpond upstream of the study site. The advantage of our study is its ability to detect short-term (single hydrological years) spatial details and temporal changes that are overlooked or cannot be identified when working on a larger basin-scale.

One of the main issues when looking at till-bedded rivers is the boundary control which till exerts on channel form and processes. The till boundary erodibility or tensile strength was investigated under controlled conditions using a variety of techniques such as laboratory flume or deploying geotechnical instruments in the field (Kamphuis, 1983; Kamphuis et al., 1990; Shugar et al., 2007; Mier and Garcia, 2011; Khan and Kostaschuk, 2011; Pike, 2014) and many studies investigated the till properties itself such as its cracks and fractures (Grisak and Cherry, 1975). However, very few studies looked at till erodibility under natural conditions. For example, Hill (1973) looked at till banks and found high annual erosion rates (30-60 mm/yr) associated with high shear stress and frost action, although the erosion distribution was highly variable. The question that arises is: Can small-scale reach data aid in categorizing semi-alluvial rivers in formerly glaciated terrain in broader scientific context and consequently help in deploying management and restoration protocols?

2. Study area

2.1. Regional setting

Medway Creek drains 205 km^2 and is a typical 3rd order tributary flowing into the North Thames River, the larger confluent of the Thames River (5830 km^2) that flows southwest across southern Ontario into Lake St. Clair. The channel has two main headwater branches (Fig. 1). The West Fork begins east of the town of Lucan and flows in south-south east direction. The East Fork starts south of Elgenfield Rd. and flows first in a southwest and then a southeast direction. At their headwaters, the two channel forks resemble shallow ditches, with extensive tile drainage from the surrounding agricultural land. The watershed is confined to the west and northwest by the Lucan Moraine and the channels flow through a vast till plain. The East and West Forks converge just north of the village of Arva, after which the main stem flows south along the western toe of the Arva Moraine and enters a millpond at Arva Dam. Below Arva Dam the channel starts to incise and meander in a distinct glacial valley. About 3 km south of Arva, Medway Creek crosses the Arva Moraine into the basin in which the city of London is located. According to Chapman and Putnam (1984), this glacial valley is a spillway but there is insufficient geomorphic evidence to support this (Bergman et al., 2016a). Such evidence would include a dry, ephemeral, abandoned channel that is detached from the Medway watershed, a misfit stream or a tributary connecting to the Thames River near Arva. It needs to be acknowledged such drainage could be buried under postglacial sediments and farmed land. Assuming there was no flow reversal due to isostatic rebound, the Arva Moraine breach of Medway Creek was spilling into Glacial Lake London. If there was drainage reversal, Medway Creek might have been Glacial Lake London's secondary outlet together with the Thames River. Presently, Medway Creek meanders 12 km from the Arva Dam to its confluence with the North Thames (Dreimanis et al., 1998).

Medway Creek below Arva Dam has incised into several stratigraphic sedimentary units described by Dreimanis et al. (1998). The oldest unit is Devonian bedrock of limestone and shale (Dreimanis et al., 1998). The second unit from the bottom is 10 m of glacial drift and the third unit from the bottom is the Dorchester Till from the Erie Lobe where the river bed is currently situated. An additional small till unit exists below on the eastern part of the river valley, but is completely missing on the western side. The fourth unit from the bottom is Tavistock Till from the Huron Lobe (Dreimanis et al., 1998) while the fifth unit from the bottom is silt and clay from an unspecified source. The top unit is the Arva Moraine till of the Huron Lobe. This unit is thicker to the east and is almost missing in the western part of Medway Valley. On the four bluffs of Medway Valley there are lacustrine sediments from Glacial Lake London (Bergman et al., 2016a). These sedimentary units were derived from generalized cross-sections (Dreimanis et al., 1998); they thin, thicken or disappear upstream or downstream through the valley due to the competing Huron and Erie Lobes of the LIS (Hicock and Dreimanis, 1992). It is also possible to treat the entire study area as Catfish Creek Till that is prevalent on the north shore of Lake Erie and was partly formed beneath a floating, overriding ice shelf by basal melting (Gibbard, 1980). The Catfish Creek Drift was deposited during the Nissouri Stade (23-16.5 ka BP, Terasmae et al., 1972).

Soils in the London area are moderately to poorly-drained with a tendency to gley when wet. Boulders, cobbles and pebbles are common. Low-lying land along stream courses is subject to flooding and is quite variable in texture (i.e. grain sizes). Hillsides along stream courses are frequently eroded with numerous cobbles and boulders. Soil orders that have been noted in the London area are Luvisolic, Brunisolic, Gleysolic, Regosolic and organic (O.A.C., 1931; Hagerty and Kingston, 1992) with Gleysolic (poorly drained and prolonged saturation) and Luvisolic (under forested canopy) dominating the river and stream corridors.

The climate of the Medway Creek basin is affected by the Great Lakes which surround it on three sides (Fig. 1a). Climate is temperate, with warm summers, mild winters and reliable precipitation. Local variations in climatic conditions are created by topography, the proximity to the Great Lakes and prevailing wind. Annual variations on the other hand, depend on the nature and frequency of the weather systems that cross the local area. Monthly precipitation ranges from 800-1000 mm (Brown et al., 1968). Mean annual runoff ranges from less than 200 mm to greater than 450 mm. Streams carry about a third of the total precipitation (Chapman and Putnam, 1984). Medway Creek is located in the Great Lakes snow belt, and snowfall occurs from October until April with strong lake-effect snowstorms (Niziol et al., 1995). During winter, occasional warm air masses from the Gulf of Mexico bring rains and warmer temperatures that create moderate temperature and melt events.

During the European-American settlement, the natural forests were removed and replaced by agricultural land and urban development. Current land use of Medway Creek watershed is 83% agricultural (all types of crops; seasonal, permanent and livestock), 11% natural and 6% urban (Upper Thames Conservation Authority, 2013). Total vegetation cover on Medway Creek is 11.3%, of which 9.6% is forest (Upper Thames Conservation Authority, 2013). The main vegetative cover is deciduous forest, followed by meadow. 34% of the riparian zone (a 30 m buffer on both sides of a watercourse) is in permanent vegetation (forest + meadow). Forest cover surrounds the river valley from Arva all the way to the confluence with the North Thames River. Sugar maple and American beech are the main tree species with silver maple and white elm hardwood in swamps and several species of hickory on well drained sites (City of London Corporate Services Department, 2000).

2.2. Hydrology

The 5-year mean annual flow of Medway Creek is 2.9 m^3/s (Upper Thames Conservation Authority, 2013), and the largest peak flow recorded was 147 m^3/s in 1977. The natural hydrologic regime follows strong seasonal trends associated with local climate and inherent geologic controls. Richards (1990) classified small streams in the London area as hydrologically 'variable', referring to their rapid response times to rainstorms or snowmelt events. Land use and soil type appear to be more important than watershed size in determining the runoff characteristics of a given stream or river in southwestern Ontario (Richards, 1990).

As precipitation is fairly evenly spread over the year and periodic thaws occur throughout winter at many locations, classic snowmelt hydrographs are rare (Allan and Hinz, 2004). The rain-on-snow events generated from warm air masses arriving from the Gulf of Mexico during late winter decrease the magnitude of the following spring flood. As temperatures warm during April-May, the (remaining) snow begins to melt and the spring flood (freshet) ensues. Spring flood duration and magnitude depend on the amount of snow on the ground and the rate of increase in temperature. A gradual increase yields a moderate peak discharge and long duration rising limb, while a sharp increase in temperatures may yield a high peak flow and short duration rising limb. If temperatures fluctuate, the flow may contain many sub-peaks, piggy-back bores and the major peak discharge is relatively subdued (Fig. 2). An additional hydrologic situation that needs to be acknowledged is that the river conveys large ice slabs, either in winter or early spring when the ground is still frozen. While these float on the water surface, an obstruction (such as a large fallen log or bridge foundations) or just the intensity of ice flow collisions can create a massive ice jam. Once the snow completely melts during late spring, the freshet recession limb wanes, the rain amounts generally increase during summer but river discharge decreases due to the high evaporative losses and the drying of soils in the watershed. During this time the entire river discharge is based on groundwater outflow (baseflow) with occasional urban floods, direct rain above the channel and no significant overland flow is generated from the catchment. Consequently, the upper parts of the basin become disconnected after the spring flood until late September or October when the basin's soils become saturated again by rain. Flow is only sustained year-round in the lower incised reach below Arva Dam. The Arva Mill occasionally uses the millpond water during the summer low flows and pulses the downstream channel for a few hours but it is almost insignificant to the concurrent discharge. In rare cases, antecedent high moisture ground conditions coupled with high rainfall may generate a large summer flood (Fitzpatrick et al., 2008). As temperatures drop during late summer into fall, autumn rains wet the catchment's bare soils and reconnect the upper basin to the river - generating flash floods after almost every significant rainfall.

2.3. Medway Creek study site

The study site is a 1.5 km long, full meander of Medway Creek (~7th meander of the main stem, with a drainage area of 200 km^2) (Figs. 1b and 1c), which runs through the Medway Heritage Forest and is located just upstream of the confluence with the North Thames River. Medway Creek has been gauged since the 1940s by Environment Canada. The gauging station (# 02GD008) is currently operated by the Upper Thames Conservation Authority (UTRCA). There are two steep (almost vertical) bluffs, one in the center of the meander (right bank) and one next to the upstream part of the reach (left bank), that contribute large quantities of sediment to the channel during precipitation events and after the snowmelt. Elevations range between 242 m (river bed) to 265.4 m (top of the right bank Arva Moraine bluff). The top of the valley is 272.8 m. Medway Creek's bed is about 5 m above Devonian limestone and shale bedrock and separated from it by Catfish Creek Till (Dreimanis et al., 1998).

Bedforms along the reach include pools, riffles and flats (Fig. 3), point bars, one braid bar and two steps that are surprising for a reach with such a low slope (0.003%). Bed till exposures vary in length from a few cm to several channel widths. The main channel does not preserve any bed knickpoints or waterfalls within the erodible till although till benches and ledges usually exist next to bank toes. However, small tributaries and gullies without sufficient stream power to scour the till do have exposed till waterfalls that imply channel evolution. There are scattered coarse boulders (> 1024 mm) along the reach that are beyond the channel's competence. Large woody debris (LWD) plays a key role in channel morphology as it protects the banks from erosion when logs are parallel to the flow below the bank's toe, but also causes complex flow patterns when they are perpendicular or diagonal to the flow direction, causing localized scour or aggradation of sediments.

3. Methods

A comprehensive survey of channel morphologies (mapping and bedforms) and sedimentology (of alluvium and till) was performed on the full 1.5 km meander of the Lower Medway Valley (Figs. 1b and 1c). Analyses included: 1. Description of till sediment characteristics based on 44 clasts collected from the river bed; 2. Topographic and morphologic mapping and bedform analysis of the reach; 3. Mapping of till patches;

4. , Grain size distribution analysis of all bedforms and alluvium character, and 5. Annual monitoring of till erosion. Details of the methods are provided in the Appendix. The results were compared to other semi-alluvial rivers of the region, alluvial and bedrock channels in order to give a broader context to similarities and differences. In the discussion we relate our findings to geomorphic stability using a variety of sedimentary metrics. Geomorphic stability is defined as a channel being adjusted to its hydrology and sedimentology meaning that bank and bed erosion rates are within natural variability for that area and the river has the capability to transport incoming sediments from hillslopes and upstream without going through major deposition or scour. Finally, we argue for classifying semi-alluvial channels as a separate river type between alluvial and bedrock channels.

4. Results

4.1 Till sediment characteristics

There is a size limit to the till clasts we collected - ~ 200 mm (a-axis). Clasts larger than 200 mm picked up from the river bed and exposed to air were found to be very unstable: first cracking, then breaking into many smaller fractions and consequently could not be analyzed properly in the lab. Even when kept in river water or tap water, the > 200 mm till clasts crumbled and broke down into a slurry. Therefore, the clasts analyzed range from 60-210 mm (a-axis), b-axis from 50-140 mm, and c-axis from 20-110 mm. This size limit biases our sample collection as it cannot contain large stones that occasionally occur in the till banks and in the bed of Medway Creek. Sphericity of the clasts ranges between 0.12-0.32 (average 0.17, mean 0.16), which falls into the bladed shape category. The roundness values of the till clasts (range 0.1 to 0.8, average and median 0.3) imply some rounding in place or actual transport of the entire till clasts as bedload during floods. The color of the 44 dry till clasts collected from the river bed alluvium are light grey (10YR7/1 Munsell color) and grey (10YR5/1) when the clasts are submerged and saturated. The amounts of sand (8-32%, average of 18%, median of 17%) and gravel (2-17%, average and median of 6%) are highly variable, with fines (silt and clay; < 0.063 mm) the dominant size fraction (57-89%, average of 76%, median of 77%) (Fig. 4a; Table 1).

Bulk density of the till ranges from 1.32 to 2.54 g/cm^3 (average 1.62 g/cm^3 and median 1.52 g/cm^3). These average values are slightly higher than Ohio tills (range of 1.86-2.02 g/cm^3) reported by Fausey et al. (2000) for the London Moraine Darby Till and higher than the Kalamazoo River bed cores (Lake Michigan drainage) - 1.05-1.97 g/cm^3 (average 1.39 g/cm^3) (McNeil and Lick, 2004), but similar to Sunnybrook Till (from Highland Creek) - 1.7-2.2 g/cm^3 and Halton Till (from Fletcher's Creek) with 1.5-2.2 g/cm^3 in southern Ontario reported by Khan and Kostaschuk (2011). Porosity values range from 0.5-12% (average 2.6%, median 1.5%).

Twenty-three of the till clasts (52%) were bioturbated (Fig. 4b) and 21 (48%) showed no macro-biological activity (Fig. 4c). The holes' diameters range from 2 to 8 mm in the bioturbated clasts and up 10 mm deep. The bioturbation is the first step in the breakdown of the clast to smaller fractions and removal of the outer layer although the till is massive and does not show apparent bedding. Till that was not bioturbated had a wet to dry mass ratio close to 1.0 (1.01-1.05, average 1.02) suggesting the till was heavily consolidated with no significant voids or cracks that allow water into the material. However, bioturbated clasts had a lower wet to dry mass ratio (0.88-0.99, average 0.97), indicating that some till clasts endured breakdown and erosion. In some cases burrowing chironomids (blood worms) were seen within till clasts when removed from the water (Fig. 4d).

Four borehole (BH) drillings done during July 2013 for a proposed development on top of the downstream left bank bluff included SPT (Standard Penetration Test) in the field and Atterberg limits in the lab. Seventeen tests were conducted in the Medway Valley till with sediment taken from the boreholes (Fig. 5). We excluded the topsoil and overlying lacustrine silt-clay layers of Glacial Lake London from analysis as these are not relevant to our study. The N-value (standard penetration resistance) for the SPT range is 8-24 blows, average and median of 16 blows to penetrate 15 mm using 63.5 kg hammer and 50 mm cone. There was no difference with elevation (Fig. 5). The till contained intermittent wet sand layers. The Atterberg limits (Plasticity Index – PI) of the till range is from 9-29 PI (slightly plastic to medium plastic), average of 18 and median of 17 PI (both medium plasticity) suggesting high clay content (Tridon Properties Limited, 2014). Using the SPT N-values, it is possible to calculate the undrained shear strength of the tested till:

$$c_u = f_1 \mathrm{N} \qquad (1)$$

where c_u is the undrained shear strength in kN/m^2, N is the SPT blow count (N-value), and f_1 is a factor depending on till plasticity (actual PI can be used as f_1 for each sample instead of average PI). The undrained shear strength ranges from 99 to 483 kN/m^2 (average and median are 280 kN/m^2) which are very high values. Trenter (1999) suggested that the best use of this approach is site-specific when only SPT N-values are available in some boreholes and other complementary *in situ* geotechnical tests (such as

CPT; Cone Penetration Test, a piezocone penetrometer probe, helical probe test; HPT and vane shear test) were not performed. Furthermore, this SPT method can only be used in tills that contain gravels while many geotechnical tests require that the sample will be devoid of gravel. For comparison, Shugar et al. (2007) and Khan and Kostaschuk (2011) tested the Sunnybrook till using jet testing and found a critical shear stress range of 0.001-63.3 kN/m^2 (average of 12.9 kN/m^2 and median of 6.01 kN/m^2); much lower than Medway Creek's till. Flume work on the St. Joseph Till from the St. Clair River bed (Mier and Garcia, 2011) is a better representative of shear strength of the till as it replicates horizontal water shearing rather than vertical shearing that the SPT and jet-testing induce on the till. Their study found that turbid water, once critical shear stress is attained, is much more erosive on the till than clear water. This result corroborates earlier work which showed that sand acts as a grinding agent on the till when it saltates or moves over the bed (Kamphuis, 1983; Kamphuis et al., 1990). Mier and Garcia (2011) also found that the erosion of the till has uncovered, to some extent, the embedded materials inside the clay mixture, which appear to comprise a wide range of particle sizes. Under flow conditions achieved at the critical point, those particles would be entrained if the material was cohesionless but till particle cohesion elevates the shear stress needed to erode them. Mier and Garcia (2011) shear stress values in the flume are 4.14-4.32 kN/m^2 with an average of 4.23 kN/m^2, which are low compared to most studies but similar to bed values of shearing of the Grand River (in Michigan) - 0.4 to 3.2 kN/m reported by Jepsen et al. (2000) using bed cores. The comparison between different sites and different types of till is difficult because of the different geotechnical techniques utilized by various researchers.

Pike (2014) recently examined till erosion in a unidirectional flume of samples collected from the area downstream of the right bank bluff in our study. Her results showed that mass erosion of till dominated and occurred around natural planes of weakness and irregularities, such as gravel particles, within the material and average shear stress of about of 8 kN/m^2. However, when the samples were air-dried and reintroduced into water the shear stress needed was only 1 kN/m^2 with complete disintegration of the till. Addition of an alluvium into the flume increased the erodibility of the till through direct particle impacts. Pike's (2014) results demonstrate that the different conditions the till is exposed to can lead to heterogeneous spatial erodibility. Furthermore, the tools and cover effect (Turowski and Rickenmann, 2009) depend on sediment supply from upstream, and till patch exposures tend to erode whether the water is clear or turbid. The complete cover effect of the alluvium protecting the till can only happen when the sediment is thick enough that even if some of the alluvium is in motion the till does not come into contact with any moving particles or the shearing action of the flow. It is worth noting that the above flume experiments/cores were done under controlled laboratory conditions with no alluvial cover/partial alluvial cover to protect or erode the till (i.e. enough to transport grains but not enough to cover); the till/clay samples are all disturbed to some extent when removed from their original location (the importance of shear stress). Temperature changes were relatively minor during the tests but the erosion/preservation processes are complex in a natural river setting. An experimental low strength boundary material channel mimicking bedrock by Finnegan et al. (2007) showed the complexity of interplay of sediment supply, river incision, and channel morphology that all the above mentioned geotechnical tests were not set up to test.

4.2 Till exposures, boulders and erosion of the boundary till

Similar to bedrock, till (of subglacial or supraglacial origin) is not spatially homogenous and contains different architectural elements and associated lithofacies (Boyce and Eyles, 2000). For example, these properties include different ratios of sand-silt-clay and rock fragments, grain size distributions, grain sorting, rock shapes, carbonate content and cementation, lamination, bulk density and void ratios (Dreimanis and Reavely, 1953; Dreimanis, 1959; Arneman and Wright, 1959; Easterbrook, 1964; Evenson et al., 1977; Hicock and Dreimanis, 1992). Tills are also known to contain cracks and fractures that are potential weak spots as water moves within them (Grisak and Cherry, 1975). An example of cracked till from Medway Creek's bed can be seen on Fig. 6b and Fig. 7a.

We detected and measured 24 patches of till which have at least ≥ 1 m of exposure as one of their dimensions (width or length). Such exposures are visible while walking along the channel at almost all flow conditions as their color is quite distinct from neighboring gravels (Fig. 6c, 6d and Fig. 7a-c). Till patches of this size constitute 598 m^2 of the bed which is 0.03% of the study reach. However, we estimate that 5-10% of the bed is devoid of alluvium (i.e. till exposures) and the smaller estimate is due to the mapping technique that "misses" very small exposures and is highly subjective. These values of alluvium coverage (90-95% by area) are higher than Hrytsak's (2012) estimate of 70% on Medway Creek but these can be explained by different sampling procedures, different sampling locations, and different time of sampling.

In places where alluvium is thin it is enough to remove a single clast to expose the till substrate. Till patches appear in any of the bedforms (pools, riffles and flats) except steps and bars. In many cases, the till patch is a continuation of the bank till and forms a ledge and complex cross-section with an inner terrace-like shape. In cases where there is a steep gradient between the ledge and bed below, the till patch cannot alluviate from bedload transport as it is disconnected from the nearby deeper bed (Fig. 6b and 6c). Often the till is partially alluviated (i.e. it has gravel on top) from *in situ* gravel that is embedded within the till matrix (Fig. 7c), as was found by Mier and Garcia (2011) when they eroded a till slab in a flume. Occasionally, loose coarse material covering the till surface can originate from collapsed bank material.

In bedrock channels the assumption is that coarse stationary material, especially boulders that occupy a large portion of the cross-section and create high roughness and turbulence, is responsible for alluviation and protection of the boundary from incision (Chatanantavet and Parker, 2008). We identified 275 boulders ($\geq$ 500 mm) along the study reach and correlated their distance to the 24 till patch exposures, expecting a negative correlation (e.g. an exposed till patch will not be close to a boulder and the areal size of the patches will not be correlated to the nearest boulder upstream). The results presented in Fig. 8a and 8b show that boulders are not the direct cause for the boundary exposures, neither in terms of distance, nor in size.

Our observations over four years have shown that in many instances the top 1-2 cm of the till bed becomes soft with time, while the underlying till remains highly consolidated. Broadly speaking, this observation corresponds to the proposal by Chatanantavet and Parker (2009) to a bedrock “battering layer” overlying an aging layer. We did not observe any plucking processes (Whipple et al., 2000a). Nor did we observe the variety of bedforms that appear in bedrock channels (Richardson and Carling, 2005), besides ledges next to banks (Fig. 6b and 6c). Bioturbation of the till by chironomid larvae (blood worms) occurred only when they appeared in till ‘clasts’ (Figs. 2b and 2d) but not when the till was part of the bed or banks (Figs. 6b, 6c, 7a and 7c).

Recovery rates of the erosion pins during four years were 57-100% with an average of 65%. During two of the hydrological years (HY 2011-2012 and HY 2012-2013), the largest erosion pin patch (53 pins) next to the right bank was buried by a massive collapse of the till bank (estimated initially at 15 m^3) that merged into the bed so recovery of pins by the end of the hydrologic year was calculated as zero. The patch was re-exposed during HY 2013-2014 after the removal of the reworked till, about half each year. There are two to six different patches per year and a total of eleven patches (Table 2). Bed lowering erosion rates range from 6-260 mm, average 61 mm and median of 39 mm. The 260 mm lowering value of one of the pins seems unusually high but this can be explained by its location in a pool that was possibly heavily scoured and a chunk of till around the erosion pin was removed. Another possibility is that once the erosion pin was protruding into the flow it served as an obstacle to moving grains and it was impacted by them time and again further increasing the bed till erosion. Spatial autocorrelations using Moran *i* reveal that there is no dependence of neighboring erosion pins within each patch, nor between the patches. Fig. 9a presents HY 2013-2014 results, when we had the most monitored patches, and exemplifies the variability of observed erosion rates. Unfortunately, there are currently no published erosion rates for bed lowering in till channels to which we can compare our results. Erosion of till banks in two Irish rivers found erosion rates of 30-60 mm also with high variability within a site and between sites (Hill, 1973). The spatial heterogeneity in erosion within patches and between patches rates is probably associated with the till tensile strength. The more consolidated the till is the harder it is for the flow and particle abrasion to erode it and vice versa. An alternative explanation is that at this patch-scale the bed relates to the till patch as a local baselevel within the bedform it is in. Once the thin alluvium is removed or part of it is in motion the particles accelerate over the till due to the sudden change in boundary roughness and continue downstream. As the flow wanes the thin alluvium settles again but the till patch remains bare and lower than before. Thus, there is a positive feedback mechanism in which the till patch "attracts" bedload particles, those particles erode the till through rolling and sliding over the relatively smooth surface, they lower it further and continue downstream. This mechanism of till patch differential lowering due to serving as local baselevel can be seen in Figs. 6d, 7a and 7b where the till patch is a few centimeters lower than the nearby surrounding alluvium.

Comparison of Medway Creek to erosion studies of bedrock channels indicates that Medway Creek’s maximum erosion rates are higher than all of the reported bedrock channels, but not by much (Fig. 9b). Relative to the till geologic erosion rate (2.6 mm/yr), our short term values are very high. However, other studies in bedrock channels also found similar results (Stock et al., 2005). This could be explained by periods of rapid valley incision and periods of relaxation when the channel reaches a hard substrate or is climate-induced (i.e. less precipitation and flows). It may also indicate that the incision is not homogenous in space and time over the river valley and it occurs in a patchy manner depending on local hydraulic/sedimentary conditions.

4.3 Bedform dimensions and spacing

We identified a total of 38 continuous bedforms, excluding the 5 bars (4 pointbars and one braid bar). Bars were excluded because, besides the braid bar, they are forced (meander) pointbars and lack downstream connectivity to the upstream and downstream bedforms, except during high flows when they are activated. Therefore, they will only be discussed in section 4.4 regarding their grain size distributions. Of the 38 continuous bedforms, the most common were riffles (39.5%), then pools (34.2%), flats (21.1%) and 2 cobble-boulder steps (5.3%). Table 3 and Fig. 10 present the bedform characteristics.

Although we followed the detailed bedform classification of Montgomery and Buffington (1997), the distinction between pools and flats, which are both scour features and have some similar dimensional characteristics (Figs. 3 and 10c), is problematic and prone to observer bias and considerable error in forested channels (Wood-Smith and Buffington, 1996), especially when the determination criteria is visual. The bed boundary exposures of till can be considered separate from bedforms, so they appear below independently.

The spatial bedform arrangement is quite inconsistent. The intervening bedforms between identical bedforms can be of different kinds or can be repeated more than once. Therefore, the typical alluvial rivers bedform coupling of pools-riffles, flats-riffles and steps-pools does not apply to the study reach as in many instances there are intervening bedforms that disrupt this organization. Spacing between the 15 riffles ranges from 12 to 321.5 m (average of 84.1 m, median of 47.3 m). Nine of the intervening bedforms are pools, three are flats and two are a variety of bedforms. Spacing between the 13 pools ranges from 5.1 to 195 m (average of 33.5 m, median of 14.5 m), of which the intervening bedforms are eight riffles, one flat and three are a variety of bedforms. Spacing between the 8 flats range from 0.5 to 369.3 m (average of 121.3 m, median of 67 m). The intervening bedforms are two riffles, one step, one pool and the other three are a variety of bedforms.

The two boulder steps (the downstream one is seen in Fig. 6a) were termed by Martini (1977) as “small scale riffle bars” or “large scale transverse ribs”. While they are very distinct and unusual for such a low gradient channel (0.003 on average), the distance between them in the study reach is more than 1 km, so we can neither relate to spacing of other bedforms nor to any relations between them. Both boulder steps span the whole channel width: the upstream one is 10.4 m wide and the downstream one 13.3 m wide. However, these unusual bedforms do appear in other nearby rivers and were observed on the North Thames River, Maitland River and the Bayfield River to the north that are also semi-alluvial channels incised into till (authors’ personal observations). Martini (1977) described these coarse steps on Irvine Creek as perpendicular to the flow imbrications either isolated or in sets of two or three. In order for them to form, the channel needs to have strong convergent flow during floods and the steps separate steep reaches into a series of riffles and pools. The upstream step separates two flats and the downstream one is between a pool and flat. Despite their coarse size, the steps are not a stagnant sedimentary body and they break down and re-form during high flows (see grain size distributions below).

4.4 Grain size distributions, alluvium structure and thickness

Downstream fining is a main sedimentary feature in alluvial rivers that is often investigated and infers channel competence reduction (Ashworth and Ferguson, 1989) or sediment attrition (Adams, 1979). While downstream fining usually occurs over long distances when the channel changes its gradient and stream power (Knighton, 1980), other factors such as bimodality (Wolcott, 1988), tributary disruption (Rice, 1998) and patchiness (Paola and Seal, 1995) can also cause the opposite effect of downstream coarsening (Brummer and Montgomery, 2003). Using this rationale, we hypothesize that the channel expansion and contractions along Medway Creek (discussed in Bergman et al., 2016b) and the till boundary forcing (i.e. till control) will also exhibit some kind of sedimentary pattern (Sklar et al., 2006), despite the relatively short distance of the study reach. Ferguson et al. (1996) also used this approach in the Allt Dubhaig (Scotland) for a 2.5 km reach. Gran et al. (2013) attributed downstream coarsening on the Le Sueur River (Minnesota) to break-up of the till which leaves behind coarse lags and strongly affects the longitudinal profile. A study by Renwick (1977) in a small channel in New York has shown that very coarse material originating from glacial deposits is introduced into the main channel during extreme storms.

On Medway Creek, benches and ledges (Figs. 4b, 4c and 5c) are the only sculpted forms that are preserved within the till geometry and the abundant forms described for bedrock channels (Richardson and Carling, 2005) are absent. We did observe a boulder lag at the toe of the right bank bluff, almost bar-like in shape (Fig. 6d). Since the bluff is an outer meander cut bank, it leads us to believe this is a result of the channel's inability to transport these boulders during most flows, as there are till exposures nearby that suggest high shear stress during floods (Fig. 6d).

The grain size distributions of 43 bedforms (pools, riffles, flats, bars and steps) are presented in Fig. 11a. Most of the grain size distributions (GSDs) are bimodal (65%). Of the fifteen riffles, eleven (73%) distributions are bimodal and four (27%) are unimodal. Three of the bars are bimodal (60%) and two are unimodal. Ten of the pools (77%) are bimodal and three (23%) are unimodal distributions. The eight flats have the highest unimodal distributions with five (63%) and the other three are bimodal (37%). By bedform, most of the bed is poorly sorted (84%) while the remaining bedforms are moderately sorted. By bed area, 77% of the bed is poorly sorted while the rest is moderately sorted. It is quite difficult to observe any trends as the grain sizes of the bedforms overlap both in the fine and coarse tails, except for pointbar 3 which is much finer than the rest. It is noteworthy that both the fine and coarse tails of GSDs are prone to large errors associated with sampling procedures. Furthermore, the connection of the tails to the main curve using sometimes completely different sampling techniques (i.e. Wolman pebble count and sieving) is also problematic. To minimize the errors all GSDs in this study were done by one individual and with identical sampling program for all bedforms.

To overcome this difficulty, we average the GSDs of all bedforms by their type (Fig. 11b). This shows that pools are finest, followed by bars, flats are intermediate while the riffles and steps are coarsest. The pools and flats are almost identical in their coarse tail (d_{70} and above). The steps show a much coarser distribution than all other bedforms but, considering this GSD is based only on two bedforms and their limited total area of the bed, it would be wrong to draw any major conclusions about the whole bed from them. However, the steps' GSDs changes with time are a good indicator of the behavior of the coarse bedload fractions and whether boulders of certain sizes are mobile or not (see below).

There are two types of bars along the reach: four pointbars and one braid bar just below the downstream step. The bars' sedimentary importance lies in the fact that they are slightly elevated above the immediate channel and dry most of the time, i.e. they represent the sediment composition that is transported as bedload during high flows (Leopold, 1994) although the sand fraction can be entrained in suspension (Nunally, 1967). They constitute < 1.0% of the total area of the reach. Looking at their coarse tails shows that they contain few large cobbles and small boulders (256-512 mm). Pointbar 3 is significantly finer than its counterparts and is entirely composed of sand.

Selected percentiles of gravels along the study reach are presented in Fig. 11c. Very slight downstream fining trend is observed within d_5 to d_{75} (trend line slope range is from -0.0175 to -0.0028). However, there is a slight coarsening trend for the d_{95} percentile and the $d_{max,}$ (trend line slope of 0.012 and 0.041, respectively) possibly suggesting that the very coarse boulder tail is non-fluvially controlled (Wolcott, 1988; Brummer and Montgomery, 2003). It is possible that our 1.5 km study reach is not long enough to capture such subtle sedimentary changes and lies within natural variability, but the fact that only two coarse percentiles are different than their finer counterparts is indicative of supply external control (exposure of boulders from the till bed and banks) rather than competence and those boulders being transported from upstream. Thayer et al. (2016) found downstream fining on the Little Rouge River (115 km^2, Lake Ontario drainage) correlated to slope and disturbed by coarse cobble-boulder lags generated by the till but their study was done over a 45 km long reach where such differences are pronounced, not on a reach-scale. The sorting index (SI) of all bedforms ranges from 1.7 to 4.4 (average of 2.9, median of 2.8) conforming to poorly sorted to very poorly sorted gravels. If partitioned into bedform types the average and median sorting of the riffles (1.9-3.7 range) and flats (1.7-3.6 range) are identical at 2.7, lower and better than the pools' average and median sorting of 3.5 (2.6-4.4 range). The four pointbars have an average sorting of 2.4 (range 1.7-4.0, median 2.0). The two steps containing mainly cobbles and boulders are best sorted at 2.2 average (range 2.1-2.3) together with the braid bar at 1.9.

The sand fraction of the distributions (≤ 2 mm) that is known to be a destabilizing component of the bed when appearing in high values (Wilcock et al., 2001) shows that despite the till's predominantly fine-grained composition it is not directly reflected in the bedforms. As expected, pools have the highest sand representation of 4.2-39.4% (average 15.5%, median 11.1%) followed by flats with 1.7-13.8% (average 6.9%, median 5.3%). Riffles have 0.7-11.4% (average 5.5%, median 4.5%) while the steps have the lowest sand fractions of 2.6-2.7% (average and median 2.7%). Besides one pointbar that was entirely made of sand (Pointbar 3), the other three pointbars together with the braid bar were devoid of any sand during the sampling survey. These results require further analysis to define bedform and reach stability of the study site. Such a stability analysis using a variety of accepted metrics appears in the discussion section below.

As stated above, steps are the coarsest bedform (Fig. 11b) and they were stationary during the study period, i.e. did not change location or migrate, somewhat like a steady local baselevel (Bowman, 1977). However, comparing their annual GSDs reflects the behavior of the mobile coarse fraction of the entire channel, similar to higher elevation pointbars in alluvial channels (Nunally, 1967). Peak discharges during the four year study period ranged 26-64 m^3/s, discharges that are sub-bankfull to bankfull depending on which cross-section in the reach is chosen (see Bergman et al., 2016b). Figure 11d demonstrates that there were sedimentary changes in the steps' GSDs over four years. The d_{50} of the upstream and downstream steps fined (from 159 down to 118 mm and from 182 down to 178 mm, respectively). On the other hand, the coarser tails of the GSDs, from d_{84} and above, show that the upstream step coarsened while the downstream one fined (d_{max} of 770 went up to 920 mm and 840 declined to 700 mm, respectively). Hence, boulders of certain sizes were mobile. Blair and McPherson (1999) classified this sediment fraction as "medium boulders" (512-1024 mm). There are "coarse boulders" (1024-2048 mm) in the study reach (the largest one with a b-axis of 2000 mm can be seen in the background of Fig. 6a upstream of the step), but we have no GSD indication or visual observation that any of these boulders had moved during our study. Out of 275 boulders sampled, only 10 (3.6%) fall into this category and we conclude this is an immobile fraction of the bed, originated from the local bed and banks till rather than fluvially transported from upstream. The fine tails of gravel and sands in both steps were non-existent in the early sampling and are not a result of sampling truncation or error. This suggests the steps can trap the finer bed fractions in large voids between the cobbles and boulders. Mobility of this coarse bedform relies on sampling accuracy being done by the same individual, is reproducible, and incorporates all material present in the steps as its transverse and narrow alignment is easily characterized as one grain population, as suggested by Kondolf (1997).

A key sedimentary feature that implies sediment availability from banks and hillslopes (or river valley) and extent of the mobility of the alluvium during floods is the bed vertical structure, i.e. stratigraphic armoring of the bed (Parker and Klingeman 1982; Dietrich et al., 1989) or its absence (Laronne et al., 1994). Whereas armoring of the bed suggests low sediment supply from upstream or lateral sediment inputs leaving a coarse lag at the surface when the fines are transported and diminished and consequently minimal bed mobility occurs (Parker, 1990), a thick non-graded alluvium (often with upward grain coarsening) implies a vast supply of sediment and high sediment mobility (Powell, 1998). However, moving in the upstream-downstream direction can also yield variations in armoring even if sediment supply is high (Lisle and Madej, 1992) as the river bed is patchy and contains a very wide grain size distribution (Paola and Seal, 1995). In order for any bed structure to form, a certain alluvium thickness is needed unless the boundary is exposed and there is not enough sediment to cover the entire channel bed. Analysis of 26 random locations along the study reach thalweg, alongside 213 m downstream of the downstream right-bank bluff, revealed that alluvium thickness ranges from 0 (i.e. no alluvium at all and exposure of the boundary till) to 62 cm at riffles' tails (average depth 20 cm, median 10 cm). In the same river, Hrytsak (2012) found in two reaches an average alluvium depth of 24 cm, which is quite similar to our results. In many pools, flats and riffles' heads the thickness is only one grain thick suggesting that the boundary is highly erodible if the protecting alluvium is removed during sediment transport (cover becoming tools effect; Turowski and Rickenmann, 2009), almost like a thin veneer. In contrast, in the riffles' centers and tails the alluvium thickens and the cobbles are highly interlocked with sandy fines filling the matrices. These sedimentary structures are, therefore, much more difficult to break. However, in these alluvium deposits there is no classic two-layered armor (Dietrich et al., 1989), but rather a more homogenous (non-censored) vertical structure from the surface all the way down to the till boundary's subsurface. This finding of no bed armor corresponds to the extensive GSDs survey of Annable (1996) that found that none of the local rivers were armored using the d_{50} armor ratio ($d_{50Surface}/d_{50Subsurface} < 2$). Even when using d_{84} for the armor ratio, only six streams of 47 (13%) were found to be armored (ratio > 2). Specifically relating to six streams classified like Medway Creek (E4 type), armor ratios were very low (≤ 1) using d_{50} ratio but even when using d_{84} ratio, none passed the armor threshold of 2. Two rivers (33%) were very close to the armor threshold but the other four (67%) were with a ratio equal to or about 1 (Annable, 1996). If the armor ratio metric is valid for rivers like Medway Creek as it is for alluvial rivers, there is no shortage of sediment supply from upstream, till banks and bed. This notion is highly important as it implies that till exposures are not a direct result of sediment supply shortage (i.e. there is enough within-reach sediment to blanket the entire bed) but rather a result of local hydraulics.

5. Discussion - Channel (in)stability

The fact that local river practitioners use the Natural Channel Design (NCD) method (Ashmore, 1999; Annable, 1999) borrowed from the alluvial literature and its consequent failure (Ness and Joy, 2002; Annable, 2003) exemplifies the need for a more robust, quantitative approach based on actual field measurements transformed into stability indices. We therefore look for indices that are solely based on quantitative measured geomorphic attributes. The use of bankfull discharge as a stability metric (Olsen et al., 1998) seems inappropriate for our study reach (Annable et al., 2011; Floresheim et al., 2013; Hassan et al., 2014). This metric of bankfull discharge is addressed in Bergman et al. (2016b).

5.1 Channel stability based on sedimentary attributes

We apply three different indicators of channel stability: the Log Relative Bed Stability ($LRBS^*$) index of Kaufmann et al. (2008; 2009), Kappesser's (2002) Riffle Stability Index (RSI), and fine sediment abundance (Lisle and Hilton, 1992, 1999). The $LRBS^*$ method describes the tendency of the bed particles to resist erosion relative to the reference bankfull discharge. The essential reach-wide field data are a systematic particle count, a longitudinal profile of thalweg depth, bankfull width and height above present water level, and reach-wide bed slope:

$$RBS^* = 1.66\theta d_{gm} /(RS) \qquad (2)$$

where θ is the critical dimensionless Shields number (see below), d_{gm} is geometric mean bed surface particle diameter (m), R is hydraulic radius (m) and S is bed slope approximated to water surface slope. d_{gm} is calculated by:

$$d_{gm} = (d_{84}d_{16})^{0.5} \qquad (3)$$

θ range is 0.04-0.5 and is calculated from the Reynolds particle number:

$$Re_p = [(gRS)\ 0.5d_{gm}] /v \qquad (4)$$

g is gravitational acceleration (m/s^2), v is the kinematic viscosity of water ($1.02\ 10^{-6}\ m^2/s$ at 20˚C). When $Re_p \leq 26$ θ is 0.04 and when $Re_p > 26$ θ is 0.5. The difference in θ values is associated with the gradient of the streams, where for high gradient streams Shields numbers are large while small θ values represent mobility under conditions with form roughness absent (i.e. assigning θ values according to GSDs). The reference bankfull discharge values (width and height above present water level) we used were the actual measured values for each bedform. In order to simplify the results, Kaufmann et al. (2009) suggested to add a log to RBS^* so the results will have a narrow range for comparison purposes and be termed LRBS*. The higher the LRBS* value the higher the stability and the closer it is to zero or even negative the bed is unstable and prone to scour and induce high rates of sediment transport.

The results of this analysis are highly unexpected in the sense that the pools, which would be expected to scour during high flows, are the most stable bedforms (range = 0.16-1.96; average = 1.12; median = 0.96) followed closely by flats (range = 0.43-1.85; average = 1.09; median = 1.13) (Fig. 12a). The depositional riffles are the least stable bedform in the reach (range = -0.03-1.28; average = 0.38; median = 0.24) (Fig. 12a). The LRBS* of the whole reach equals 0.8. While there is no absolute LRBS* value which distinguishes a functioning and impaired river system, LRBS* does give a robust measure of the bed stability for comparison. For example, Pahl (2006) found much lower stability values than we report for 13 alluvial reaches of the Carson River in Nevada (range = -1.89-0.26; average = -0.96; median = -0.85). Although Medway Creek is an incised channel, and its stability results are higher than the Carson River values, it would be wrong to make assumptions about instream habitat quality based on sedimentary/hydraulic criteria without actual measurement of biological inventory (Duncan et al., 2011).

Kappesser's (2002) Riffle Stability Index (RSI) is defined by calculating the geometric mean (d_{gm}) of the riffle of interest and then checking on its cumulative GSD what the corresponding percentile for that value is. The resulting RSI value is unit-less. The higher the RSI value, the higher the loading of fines. The main assumption behind this index is that the amounts of fines reflect the degree of landscape disturbance. An abnormally eroding landscape would choke the channel with large amounts of fine sediments that it cannot evacuate during sediment-carrying flows. Medway Creek's riffles fall within a relatively narrow RSI range of 31-48 (average and median 38) (Fig. 12b). These low values indicate that riffles possess similar attributes, with relatively low textural variability. For comparison, streams of Idaho have higher median RSI values of 58 and 80 for natural and managed rivers respectively (Kappesser, 2002). Virginia streams have a mean and median of 70, with low values of 40 and high values approaching 100 (Kappesser, 2002). The riffle stability results of Medway Creek also corroborate the GSDs minor changes in the upstream – downstream direction, seen in Fig. 11c, without any significant textural changes. However, due to lack of research on till channels, it is hard to determine whether these low RSI values are normal, indicate another control exerted by the local till textural composition, or reflect disturbance from land uses within the basin upstream of the study site.

Since $LRBS^*$ and RSI are not enough to indicate conclusive results, we applied a third stability metric. Fine-sediment abundance can indicate a reduction in transport capacity without a compensating decrease in sediment supply. This assumption comes from the graded river concept of Mackin (1948) - if the river maintains equilibrium conditions over time, no bed deposition or scour occur – although the GSDs can change as a function of sediment supply. However, it is expected that, if the bed is starved of sediment, a coarse surface layer of armor will develop once the fines are winnowed out (Dietrich et al., 1989). Conversely, if sediment supply exceeds transport capacity, the river will supply no armor and the fines will dominate the surface layer (Laronne et al., 1994; Bergman et al., 2010). Ecologically, excess fines in a river bed are perceived as a negative component of the GSD, as they block gravel pores, smooth the surface, reduce roughness and consequently affect habitat quality and spawning of fish (Everest et al., 1987). Fines also increase bed instability by reducing the shear stress needed to entrain coarser fractions (Wilcock et al., 2001). The question is: is fine sediment abundance a reliable stability metric in a river where the boundary material produces large amounts of fines? Pools are usually the best indicator for fine sediment abundance because: 1) this fraction is deposited within them during the waning stage of floods; 2) it is the first fraction to be winnowed out of the pool; 3) the residence time is therefore relatively short (Lisle, 1995); and 4) it reaches full fractional mobility during a typical flood season (Church et al., 1991). To test an abundance of fines metric on Medway Creek we implement the technique proposed by Lisle and Hilton (1992, 1999) of relative volume of fine bed material on the 13 pools of the study reach. Dimensionless relative volume of bed material V^* is represented by:

$$V^* = V_f/(V_f + V_r) \qquad (5)$$

where V_f and V_r are respectively the fine sediment pool volume and residual pool volume for each pool. The scoured pool volume ($V_f + V_r$) is the residual volume of a pool if the fine sediment were removed (Lisle and Hilton, 1992). On a channel like Medway Creek that has a wide GSD from silt-clay till patches up to medium boulders, choosing a coarser truncation point for fines than the traditional sand-gravel transition (2 mm) is fitting to reduce the sampling bias of the fine tail of the distribution. Since we conducted the GSD sampling with 32 mm truncation between Wolman pebble counts and sieved fractions (coarse gravel to very coarse gravel transition), we used this clear boundary to define the finer material (V_f). Results show that the relative volume of fine sediment in pools is not high (ranging = 0.01-0.15; average = 0.04; median = 0.03; Fig. 12c). The fifth pool (V^* =

0.15) is unusually high compared to other pools. This pool is just downstream of the right bank bluff, which might be the cause for fines loading, although the first pool downstream of the left bank bluff does not show the same fines loading. If compared to the results from Bear Creek, California, a river rich in fines with granites and metasediments as parent material (Lisle and Hilton, 1992), our results are quite similar to undisturbed pools of Bear Creek, but significantly lower than Bear Creek pools that were disturbed by illegal mining activity (values up to 0.55; Fig. 12c). Lisle and Hilton (1999) differentiated between river channels with fines from poor parent material and fines from rich parent materials. The latter, of interest to our study, show values ranging from 0.01-0.50, average 0.2 and median 0.19 (n =22). The pools' relative volumes of fine sediment on Medway Creek support the previous stability analysis of Kaufmann et al. (2008; 2009) that pools are the most stable bedform of the study reach (Fig. 12a).

If the three methods are a robust viable measure of stability in till channels, as they are for alluvial ones (Rathburn and Wohl, 2003; Pahl, 2006), they show that the bed is stable as the riffles, which present instability, only constitute 15% of the total bed area of the study reach. However, in a recent paper Lisle et al. (2015) cast doubt on the ability of stability metrics to adequately describe channel impairment, especially with human interference (anthropogenic sources). Geomorphic channel impairment is generally defined as a river that has a constant imbalance between its hydrology and sedimentology (process and form) and thus it is not functioning properly. This negative state contrasts natural variability in which the river's channel form and process maintain quasi-equilibrium. Lisle et al. (2015) suggest having a two tiered evaluation where tier one uses rapid assessment to provide a coarse analysis (such as the stability metrics above) while for tier two an historical contextual assessment is needed. Presently it is not possible to conduct the tier two phase as there is no “template” or “reference” till river that sufficiently describes form and process. Considering that the till parent material produces large amounts of fines (Fig. 4a; Tables 1 and 2) and that there has been human disturbance in the upstream catchment for more than two centuries, the use of the phrase “sediment impairment” needs scrutiny to distinguish normal levels of fines ("background" or ambient sediment) from unusual amounts of fines loading (Belmont et al., 2011). Wilkin and Hebel (1982) worked on the Middlefork River (Illinois, Mississippi River drainage) that, like Medway Creek, is deeply incised into glacial sediments. They found that sediments are predominantly eroding from in-channel sources (bed, banks and farmed floodplains) rather than farmed uplands.

From an ecological point of view, Medway Creek (and specifically our study site that is annually sampled) has the highest fish biodiversity in any southern Ontario stream with 44 species (John Schwindt, Upper Thames River Conservation Authority, personal communication). The high fish biodiversity would imply a healthy and stable river system making the geomorphic-sedimentary stability metrics inadequate at this stage to make decisions on assessing channel state, especially for habitat quality use for which these sedimentary metrics were designed. The sedimentary metrics are also disconnected from water quality parameters (i.e. it does not matter to the results if the river is polluted or not) so connecting channel sedimentology and biological attributes without proper context is flawed.

5.2 Channel stability based on GSDs of fine-rich rivers

It is accepted that in gravel-bed rivers the presence or absence of abundant bed fines is a reliable indicator of channel stability determining its bed mobility (Lisle and Hilton, 1992, 1999; Wilcock et al., 2001; see previous section). The increase in fines destabilizes the coarser bed particles and reduces the shear stress needed to entrain them while the fine particles themselves (usually sand up to pebbles) are easily mobilized at relatively low shear stress flow events. Thus, an armored bed (i.e. the upper layer of the bed is coarse-grained and overlies finer particles) is much harder to entrain than a mixed bed dominated by fines. By comparing Medway Creek's generalized GSD to other river beds where the local geology produces rapid sedimentary breakdown and high supply of fines, it is possible to put these bed characteristics in a wider context. By giving each GSD of the bedforms a weighted percentage of the channel area it occupies, we are able to transform all the GSDs of Medway Creek (Fig. 11a) into a single GSD representing the entire study reach (Fig. 11b).

We compare Medway Creek's bed to three other rivers with very high fine sediment supply: Nahal Eshtemoa of the northern Negev Desert (Israel) (Powell et al., 1999), a semi-alluvial semiarid gravel-bed channel incised into late Pleistocene loess deposits; Nahal Me’arot in NW Israel (Greenbaum and Bergman, 2006), a watershed with Mediterranean climate draining an ancient volcanic valley with eroded tuffs and limestone clasts; and the West Walker River flowing out of formerly glaciated Sierra Nevada in California into a rain-shadow desert with eroded granites (Bergman et al., 2010). We caution that Medway Creek and the West Walker River are perennial rivers dominated by spring snowmelt hydrographs, while Nahal Me'arot and Nahal Eshtemoa are flashflood-prone ephemeral wadis. While the common notion is that bed structure and composition are mainly controlled by sediment supply (Parker and Klingeman, 1982; Dietrich et al., 1989; Laronne et al., 1994), hydrograph shape and duration effects on bed stability producing similar sedimentary stability outcome cannot be ruled out (Hassan et al., 2006).

Our analysis shows that Medway Creek falls between ephemeral rivers and the perennial desert river, with Medway Creek's GSD finer than the West Walker River but coarser than the Mediterranean volcanic and semiarid GSDs (Fig. 13). There does not appear to be an unusual amount of fine sediment loading (< 2 mm), although if the truncation of fines is increased to 32 mm (medium gravels) it constitutes about a third of the bed. This value possibly suggests that shear stress needed to entrain a large portion of the bed is not high indicating bed instability.

We look at four different sedimentary attributes of these four rivers: modality, sorting, skewness and kurtosis of their GSDs. Sediment modality is important for bed stability as bimodal sediment is less stable than unimodal sediment and consequently results in higher transport rates and sediment yields (Wang et al., 2015). Furthermore, bimodal beds may represent a distinct threshold between gravel-bed and sand-bed states (Sambrook Smith, 1996). This could reflect the parent material composition as each peak represents a distinct population of different material supplied to the stream, or material of the sizes between the peaks could be structurally unstable and, therefore quickly breaks down to sand-size particles (Wolcott, 1988). All four rivers have bimodal or polymodal distributions, with very poorly sorted beds, the fine tails dominate with negative skew (as expected) and besides the West Walker River that is very leptokurtic, the other three rivers are quite similar and mesokurtic (Table 4; Fig. 13). The reason that West Walker River is leptokurtic may be the proximity of sampling sites to the Sierra Nevada (i.e. relatively short distance downstream from the narrow mountainous canyon reach), explaining the presence of very coarse bed material in the channel, while for other rivers the distributions are possibly linked to some downstream fining and better sorting processes. In order to quantify each GSD's ability to resist flow shear, some measure of modality is needed. If the sediment mixture is weakly bimodal, the critical shear stress of individual fractions will show little variation in grain size and will depend only on the mean grain size of the mixture (i.e. d_{50}). For strongly bimodal sediments, fractional critical shear stress increases with grain size, an apparent result of lateral segregation of the finer and coarser fractions on the bed surface that causes fractional critical shear stress to deviate from size independence in the direction of unimodal (Shields) values. Wilcock (1993) suggested a bimodality parameter *B* to quantify the effect of bimodality on critical shear stress of the bed:

$$B = (d_c/d_f)^{0.5} \sum P_m \qquad (6)$$

where d_c is the coarse mode, d_f is the fine mode and P_m is proportional to the bimodal sediments. A mode is assumed to have a width of one Φ unit (factor of two). Each mode is defined as the four contiguous 1/4Φ units containing the largest proportion. Therefore, $\sum P_m$ can take a maximum value of unity for a purely bimodal mixture. If the *B* ratio presents low values ($B < 1.0$) modality is weak, while high values ($B > 1.0$) indicate strong modality. When the modes (d_c and d_f) are close, the GSD is narrow and tends to behave like a unimodal sediment mixture (Wilcock, 1993) with higher bed stability (Wang et al., 2015). We use d_{16} and d_{84} for d_c and d_f, respectively. Results show that all four river beds are high and consequently unstable: the two ephemerals (loess and volcanic sediments) have an identical *B* (4.2), the West Walker River (granites) produces the lowest *B* (2.3), while till from Medway Creek presents the highest value of *B* (5.9). Interestingly, the sum proportion in mode ($\sum P_m$) divides the ephemeral channels and the perennial channels into two distinct groups with values of 0.6 and 0.8 respectively. We conclude that loading of fine sediments in Medway Creek is not abnormal. However, when dealing with rivers that have unique geology, then sedimentary metrics should be used in context with background bed fine levels or as part of a long-term monitoring program in which changes can be detected and cause and effect distinguished. More sedimentary datasets are needed from fine-rich channels to isolate the non-alluvial/geologic control or signature from ordinary fluvial process variability typical of alluvial channels free of non-fluvial constraints (Wolcott, 1988; Sklar et al., 2006).

6. A case for a new channel classification category

In all historic channel classifications in humid-climate environments (e.g. Leopold and Wolman, 1957; Kellerhals et al., 1976; Rust, 1977; Rosgen; 1994; Kondolf, 1995; Montgomery and Buffington, 1997; Newson and Newson, 2000; Church, 2002), semi-alluvial channels in general and till-bedded channels in particular do not receive specific treatment. The river researcher or practitioner needs to assume they are a hybrid between bedrock channels and alluvial channels, often just treated as low gradient, sinuous gravel-bed rivers (Hartley, 1999; MacVicar and Roy, 2011; Marchildon et al., 2011). This lack of attention to till-bedded rivers is surprising as vast areas of North America were glaciated by the Laurentide and Cordilleran Ice Sheets (Ehlers and Gibbard, 2004a), and it is also relevant to Europe (Trenter, 1999; Ehlers and Gibbard, 2004b; Prasicek, 2015), and other continents (Ehlers and Gibbard, 2004c, Prasicek, 2015). Furthermore, as climate warms, more and more areas will be deglaciated and glacial deposits will be exposed (Marren and Toomath, 2014).

Our extensive dataset and anecdotal research from other till-bedded rivers show that semi-alluvial till-bedded channels are different from both alluvial channels and bedrock channels in many aspects: 1) till erodibility within the same till and between tills; 2) bedform arrangement; 3) lack of one dominant discharge such as bankfull or effective discharge; 4) anomalous grain sorting processes that are a mix of fluvial and non-fluvial controls; 5) large spatial variation in alluvium thickness; and 6) alluvium structure that does not resemble the 2-layered armor structure typical of perennial alluvial rivers. We therefore argue that semi-alluvial till-bedded channels deserve a category of their own between bedrock channels and alluvial channels in all future channel classifications. The differentiation of a new category has three main implications:

1. The geologic constraint, conditioning or legacy of semi-alluvial rivers should be explicitly considered rather than merely an overlooked background detail. Specifically for till-bedded channels, to date (2016), the entire literature consists of less than twenty papers that focus on glacial conditioning. The historical context may better explain form and processes and reveal the non-fluvial component of the landscape. Such a legacy exerting some control on river evolution is not necessarily tills or other glacial deposits as discussed in this study but could also be incised loess channels (Seginer, 1966; Schumm et al., 1984; Simon and Hupp, 1986; Simon, 1994; Rozin and Schick, 1996; Hanson and Simon, 2001; Craddock et al., 2010; Bergman et al., 2014) or channels

loaded with eroded volcanic materials such as tuff, pumice, tephra and ash (Collins et al., 1983; Major et al., 2000; Hayes et al., 2002; Greenbaum and Bergman, 2006).

2. The degree of control that non-alluvial material exerts might be associated with time scale and changing climate - if the material was completely exhumed the channels may behave as ordinary alluvial or bedrock channels. For example, in the Great Lakes region, both Hack (1965) and Martini (1977) describe channels that incise into bedrock, i.e. the glacial deposits were completely removed since the end of the last ice age. In contrast, in this study and the studies of Phillips and Robert (2005), Arbogast et al. (2008), Wilcock et al. (2009), Gran et al. (2009; 2011 and 2013), Belmont (2011), Belmont et al. (2011), Khan and Kostaschuk (2011) and Phillips and Desloges (2014; 2015a, b); Thayer et al., (2016), rivers encounter tills or other glacial deposits. The early work of Ruhe (1952) identified age as a crucial factor in drainage density and drift exhumation - the younger moraine tills of the De Moines Lobe are less eroded than those that were uncovered earlier thus creating distinct topographic surfaces with various degrees of incision, even though they are all from the last Wisconsinan glaciation. On a shorter time scale in a tropical climate, Hayes et al. (2002) described extreme fine sediment loading following the eruption of Mount Pinatubo in 1991 on the Pasig-Potrero River (Philippines) which was covered by 33% of pyroclastic flows. In the same river, Gran and Montgomery (2005) showed the rapid evacuation of those volcanic sediments and the change from braided pattern with massive sediment transport to single-thread armored

channel with clear water indicating the recovery and return to geomorphic stability.

3. Once semi-alluvial channels become their own category, it will be possible to implement appropriate management and restoration schemes instead of using practices borrowed from alluvial rivers, such as Rosgen's NCD that proved to be inadequate (Ness and Joy, 2002), or from river ecology such as fish habitat enhancement using installation of in-stream structures that also failed to improve habitat (Champoux et al., 2003). Specifically, it will be possible to reconstruct an official archetypal river model or to design a handbook that managers and practitioners will be able to use and focus their actions based on sound river science.

The establishment of a theoretical framework coupled with real case studies that identify the uniqueness of semi-alluvial till-bedded rivers needs to be supported by appropriate metrics as used in this work. However, rapid assessment protocols are not enough to judge the immediate state of a river (Lisle et al., 2015) and deeper understanding based on long-term monitoring of a variety of effective variables is needed to supplement such a program done every few years. Presently, the sources of till within channels are poorly understood and it would be beneficial to understand these sources by fingerprinting (Belmont et al., 2011), as was also recently done for mountainous bedrock channels (Riebe et al., 2015). Understanding where the sediments originate from, whether from the watershed (ultimate source) or simply the till lining the channel bed and banks (local source), is a crucial step in any management program that allows river personnel to understand what are normal levels of sedimentation (within natural variability) and what rates indicate degradation and impairment that might need artificial intervention to restore the channel to a healthy state.

7. Conclusions and recommendations

A comprehensive data set of a glacially-conditioned 1.5 km long meander on Medway Creek, London, southern Ontario, Canada is presented. The study reach does not have any significant external water and sediment inputs from tributaries which makes it an ideal site for characterizing a semi-alluvial cobble-bed channel incised into till. The data set includes till characteristics, till erosion rates, bedform features and bed stability analysis using a variety of sedimentary metrics and a comparison to other rivers with abundant fines. Bed till exposures constitute a relatively small portion of the total bed area. Nearby upstream boulders of various sizes do not seem to be directly related to till patch area regardless of their position. Erosion rates are comparable to channels incising into soft bedrock and are very high compared to the local incision rate. Bedforms are highly disorganized and show no regular spacing that is typically found in stable autogenic bedrock and alluvial rivers. The channel does not show typical alluvial sorting processes such as downstream fining; there is slight downstream coarsening, implying some non-fluvial control. Alluvium thickness is also highly variable from non-existent (exposed till substrate) or one grain thick up to 60 cm deep at riffle tails where they plunge into a subsequent pool or flat. The typical two-layered armor structure ordinarily seen in perennial alluvial rivers is completely missing. The sand fraction that is a known destabilizing fraction of the bed in alluvial channels is not as dominant along the various bedforms as one would expect. Channel (in)stability analysis, using three different sedimentary attributes, does not yield conclusive results.

We recommend that research of till rivers receive more attention from the fluvial scientific and practitioners' communities as knowledge about them is currently lacking. While till rivers cover vast extents of formerly glaciated areas around the world, they are completely ignored in present river classifications. Based on our study and other investigations of similar channels within glacial deposits, semi-alluvial channels deserve to be a category on their own. This is not only relevant for river classification, but this knowledge gap also has implications for everyday management and restoration practices and whether current indices such as channel geometry and stability metrics are suitable and adequate to determine impairment and river health.

Acknowledgements

This work is part of the PhD dissertation of the first author under the supervision of the second and third coauthors. Yannick Rousseau assisted in fieldwork and GIS analysis and is greatly acknowledged. Francisco Flores-Santiago was the photographer on many field days. Erika Hill provided field equipment, helped in fieldwork and also helped in lab analysis. We thank Peter Wilcock for allowing us to use the West Walker River grain size distribution data. John Laronne and Ian Reid provided the Nahal Eshtemoa grain size distribution data. Karen Van Kerkoerle drew and improved most maps, and some additional figures. Joe Smrekar gave tech support. Upper Thames Conservation Authority and the City of London gave us permission to work in this environmentally protected area. Peter Ashmore helped in earlier stages of the work. Katrina Moser is thanked for fruitful discussions that improved the manuscript.

References

Aberle, J., Nikora, V., Walters, R., 2004. Effects of bed material properties on cohesive sediment erosion. Marine Geology 207, 83-93.

Adams, J., 1979. Wear of unsound pebbles in river headwaters. Science 203, 171-172.

Addy, S., Soulsby, C., Hartley, A.J., Tetzlaff, D., 2011. Characterisation of channel reach morphology and associated controls in deglaciated montane catchments in the Cairngorms, Scotland. Geomorphology 132, 176-186.

Addy, S., Soulsby, C., Hartley, A.J., 2014. Controls on the distribution of channel reach morphology in selectively glaciated catchments. Geomorphology 211, 121-133.

Allan, J., Hinz, L., 2004. An Assessment of Flows for Rivers of the Great Lakes Basin. School of Natural Resources & Environment, The University of Michigan, Ann Arbor, Michigan, Final Report to the Great Lakes Protection Fund, October, 2004, Available from: http://www-personal.umich.edu/~dallan/pdfs/GLPF_Report.pdf. Last accessed on June 5th, 2013.

Amerson, B.E., Montgomery, D.R., Meyer, G., 2008. Relative size of fluvial and glaciated valleys in central Idaho. Geomorphology 93, 537-547.

Annable, W.K., 1996. Morphologic relationship of rural watercourses in southern Ontario and selected field methods in fluvial geomorphology. Credit Valley Conservation Authority, Ontario Ministry of Natural Resources, Peterborough, ON, 92 pp.

Annable, W.K., 1999. On the design of natural channels: decisions, directions and design. In: Stream Corridors, Adaptive Management and Design: Proceedings of the Second International Conference on Natural Channel Systems. Niagara Falls, Ontario.

Annable, W.K., 2003. Short comings in applying regional hydraulic and morphologic databases in Natural Channel Design projects. Hydrology Days.

Annable, W.K., Lounder, V.G., Watson, C.C., 2011. Estimating channel-forming discharge in urban watercourses. River Research and Applications 27, 738-753.

Annable, W.K., Watson, C.C. Thompson, P.J., 2012. Quasi-equilibrium conditions of urban gravel-bed stream channels in southern Ontario, Canada. River Research and Application 28, 302-325.

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-central Lower Michigan, U.S.A. Geomorphology 102, 615-623.

Arneman H.F., Wright, H.E., 1959. Petrography of some Minnesota tills. Journal of Sedimentary Petrology 29, 540-554.

Ashmore, P., 1999. What would we do without Rosgen?: rational regime relations and natural channels. In: Stream Corridors, Adaptive Management and Design: Proceedings of the Second International Conference on Natural Channel Systems. Niagara Falls, Ontario.

Ashmore, P., Church, M., 2001. The impact of climate change on rivers and processes in Canada. Available from the Geological Survey of Canada, Ottawa, Ont. Bulletin 555.

Ashworth, P.J., Ferguson, R.I., 1989. Size-selective entrainment of bed load in gravel bed streams. Water Resources Research 25, 627-634.

Belmont, P. 2011. Floodplain width adjustments in response to rapid baselevel fall and knickpoint migration. Geomorphology 128, 92-102.

Belmont, P., Gran, K.B., Schottler, S.P., Wilcock, P.R., Day, S.S., Jennings, C., Lauer, J.W., Viparelli, E., Willenbring, J.K., Engstrom, D.R., Parker, G., 2011. Large shift in source of fine sediment in the Upper Mississippi River. Environmental Science and Technology 45, 8804–8810.

Bergman, N., Schmidt, J. Wilcock, P.R., 2010. The sedimentary characteristics of a coarse gravel-bed river choked with fine sediment, West Walker River, California. Poster presented in the Spring Runoff Conference, Utah State University, Logan, Utah. Available at: http://works.bepress.com/peter_wilcock/25

Bergman, N., Van De Wiel**,** M.J., Hicock, S.R., 2016a. The drainage of the last Glacial Lake London, Ontario: Mapping and hydraulic modeling. This dissertation, chapter 1.

Bergman, N., Van De Wiel, M.J., Hicock, S.R., 2016b. Semi-alluvial stream channel crossing the interlobate Arva Moraine, Medway Creek, Southern Ontario, Canada. 2: Channel morphometry, hydrology and 1D hydraulic modeling. This dissertation, chapter 3.

Bergman, N., Sholker, O., Roskin, Y., Greenbaum, N., 2014. The Nahal Oz Reservoir dam-break flood: Geomorphic impact on a small ephemeral loess-channel, Negev Desert, Israel. Geomorphology, 210, 83-97.

Bhowmik, N.G., 1979. Hydraulics of Flow in the Kaskaskia River, Illinois. Illinois State Water Survey, Report of Investigations 91, Urbana, 116 pp.

Blair, T.C., McPherson, J.G., 1999. Grain-size and textural classification of coarse sedimentary particles**.** Journal of Sedimentary Research 69, 6-19.

Blann, K.L., Anderson, J.L. Sands, G.R., Vondracek, B., 2009. Effects of agricultural drainage on aquatic ecosystems: A Review. Critical Reviews in Environmental Science and Technology 39, 909-1001.

Bowman, D., 1977. Stepped-bed morphology in arid gravelly channels. Geological Society of America Bulletin 88, 291-298.

Boyce, J.I., Eyles, N., 2000. Architectural element analysis applied to glacial deposits: Internal geometry of a Late Pleistocene till sheet, Ontario, Canada. Geological Society of America Bulletin 112, 98-118.

Brakenridge, G.R., Thomas, P.A., Conkey. L.E., Schiferle, J.C., 1988. Fluvial sedimentation in response to postglacial uplift and environmental change, Missisquoi River, Vermont. Quaternary Research 30, 190-203.

Brardinoni, F., Hassan, M. A., 2006. Glacial erosion, evolution of river long profiles, and the organization of process domains in mountain drainage basins of coastal British Columbia. Journal of Geophysical Research 111, F01013.

Brardinoni, F., Hassan, M.A., 2007. Glacially-induced organization of channel-reach morphology in mountain streams. Journal of Geophysical Research 112, F03013.

Bridge, J.S., 2003. Rivers and Floodplains: Forms, Processes, and Sedimentary Record. Wiley-Blackwell, Malden, MA, 504 pp.

Brown, D.M., Mckay, G.A., Chapman, L.J., 1968. The climate of southern Ontario. Climato-Logical Studies No. 5, Department of Transport, Meteorological Branch, Toronto, 50 pp.

Brummer, C.J., Montgomery, D.R., 2003. Downstream coarsening in headwater channels. Water Resources Research 39, 1294.

Campo, S.H., Desloges, J.R., 1994. Sediment yield conditioned by glaciation in a rural agricultural basin of southern Ontario, Canada. Physical Geography 15, 495-515.

Champoux, O., Biron, P.M., Roy, A.G., 2003. The long-term effectiveness of fish habitat restoration practices: Lawrence Creek, Wisconsin. Annals of the Association of American Geographers 93, 42-54.

Chapman, L.J., Putnam, D.F., 1984. Physiography of Southern Ontario. University of Toronto Press, Toronto, Ontario, Canada.

Chatanantavet, P., Parker, G., 2008. Experimental study of bedrock channel alluviation under varied sediment supply and hydraulic conditions. Water Resources Research 44, W12446.

Chatanantavet, P., Parker, G., 2009. Physically based modeling of bedrock incision by abrasion, plucking, and macroabrasion. Journal of Geophysical Research 114, F04018.

Chin, A. 1989. Step-pools in stream channels. Progress in Physical Geography 13, 391-408.

Chin, A. 1998. On the stability of step-pool mountain streams. The Journal of Geology 106, 59-70.

Church, M., 2002. Geomorphic thresholds in riverine landscapes. Freshwater Biology 47, 541-557.

Church, M. Jones, D. 1982. Channel bars in gravel bed rivers. In: Hey, R.D., Bathurst, J.C., Thorne, C.R. (Eds.), Gravel-Bed Rivers: Fluvial Processes, Engineering and Management. Wiley, Chichester, 291-338.

Church, M.A., McLean, D.G., Wolcott, J.F., 1987. River bed gravels: sampling and analysis. In: Thorne, C. R. Bathurst, J. C. Hey, R.D. (Eds.) Sediment transport in gravel-bed rivers. Chichester, Wiley, 43-88.

Church, M.A., Wolcott, J.F., Fletcher, W.K., 1991. A test of equal mobility in fluvial sediment transport: Behavior of the sand fraction. Water Resources Research 27, 2941-2951.

City of London Corporate Services Department. 2000. Guidelines document for the evaluation of ecologically significant woodlands, 1-7.

Clark, J.A., Zylstra, D.J., Befus, K.M., 2007. Effects of Great Lakes water loading upon glacial isostatic adjustment and lake history. Journal of Great Lakes Research 33, 627-641.

Clark, P.U., Clague, J.J., Curry, B.B., Dreimanis, A., Hicock, S.R., Miller, G.H., Berger, G.W., Eyles, N., Lamothe, M., Miller, B.B., Mott, R.J., Oldales, R.N., Stea, R.R., Szabo, J.P., Thorleifson, L.H., Voncent, J.-S., 1993. Initiation and development of the Laurentide and Cordilleran ice sheets following the last interglaciation. Quaternary Science Reviews 12, 79-114.

Clark, P.U., Fitzhugh, W.W., 1992. Postglacial relative sea level history of the Labrador Coast and interpretation of the archaeological record. In: Johnson, L.L. (Ed.) Paleoshorelines and Prehistory: an Investigation of Method. CRC Press, London, 189-213.

Collins, B.D., Montgomery, D.R., 2011. The legacy of Pleistocene glaciation and the organization of lowland alluvial process domains in the Puget Sound region. Geomorphology 126, 174-185.

Collins, B.D., Dunne, T., Lehre, A.K., 1983. Erosion of tephra-covered hillslopes north of Mount St. Helens, Washington: May 1980-May 1981, Zeitschrift für Geomorphologie 46, 103-121.

Coulthard, T.J. Van De Wiel, M.J., 2007. Quantifying fluvial non-linearity and finding self-organized criticality? Insights from simulations of river basin evolution. Geomorphology 91, 216-235.

Craddock, W.H., Kirby, E., Harkins, N.H., Zhang, H., Shi, X., Liu, J., 2010. Rapid fluvial incision along the Yellow River during headward basin integration. Nature Geoscience 3, 209-213.

Culley, R.W., 1971. Effect of freeze-thaw cycling on stress-strain characteristics and volume change of till subjected to repetitive loading. Canadian Geotechnical Journal 8, 359-371.

Dasenbrock, D.D., Schneider, J.A., Mergen, E.M., 2010. Cone penetration testing experience in glacial geological conditions of Minnesota, USA. Minnesota Department of Transportation, 8 pp.

Dickinson, W.T., Green, D.R., 1988. Characteristics of sediment loads in Ontario streams. Canadian Journal of Civil Engineering 15, 1067-1079.

Dietrich, W.E., Kirchner, J.W., Ikeda, H., Iseya, F., 1989. Sediment supply and the development of the coarse surface layer in gravel-bedded rivers. Nature 340, 215-217.

Diffenbaugh, N.S., Pal, J.S., Trapp, R.J., Giorgi, F., 2005. Fine-scale processes regulate the response of extreme events to global climate change. PNAS 102, 15774-15778.

Doyle, M.W., Harbor, J.M., Rich, C.R., Spacie, A., 2000. Examining the effects of urbanization on streams using indicators of geomorphic stability. Physical Geography 21, 155-181.

Dreimanis, A., 1959. Rapid macroscopic fabric studies in drill-cores and hand specimen of till and tillite. Journal of Sedimentary Petrology 29, 459-463.

Dreimanis, A., Reavely, G.H., 1953. Differentiation of the lower and the upper till along the north shore of Lake Erie. Journal of Sedimentary Petrology 23, 238-259.

Dreimanis, A., Winder, C.G., Aaltonen, R.A., 1998. London Ontario: Geology, geomorphology, geodata. In: Karrow, P.F. and White, O.L. (eds.) Urban Geology of Canadian Cities. Geological Society of Canada Special Paper 42, 237-260.

Duncan, W.W., Goodloe, R.B., Meyer, J.L. Prowell, E.S., 2011. Does channel incision affect in-stream habitat? Examining the effects of multiple geomorphic variables on fish habitat. Restoration Ecology 19, 64-73.

Elliot, K.A. 1998. The forests of southern Ontario. The Forestry Chronicle 74, 850-854.

Ellis, E.C., 2011. Anthropogenic transformation of the terrestrial biosphere. Philosophical Transactions of the Royal Society A 369, 1010-1035.

Easterbrook, D.J., 1964. Void ratios and bulk densities as means of identifying Pleistocene tills. Geological Society of America Bulletin 75, 745-750.

Ehlers, J., Gibbard, P.L., (Eds.) 2004a. Quaternary Glaciations Extent and Chronology: Part II: North America. Developments in Quaternary Science 2, Elsevier, Amsterdam.

Ehlers, J., Gibbard, P.L., (Eds.) 2004b. Quaternary Glaciations Extent and Chronology: Part I: Europe. Developments in Quaternary Science 2, Elsevier, Amsterdam.

Ehlers, J., Gibbard, P.L., (Eds.) 2004c. Quaternary Glaciations Extent and Chronology: Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science 2, Elsevier, Amsterdam.

ESRI., 2012. ArcGIS for Desktop. Release 10.1. Redlands, CA.

Evenson, E.B., Dreimanis, A., Newsome, J.W., 1977. Subaquatic flow tills: a new interpretation for the genesis of some laminated till deposits. Boreas 6, 115-133.

Everest, F.H., Beschta, R.L., Scrivener, J.C., Koski, K.V., Sedell, J.R., Cederholm, C.J., 1987. Fine sediment and salmonid production: A paradox. In: Salo, E., Cundy, T., (Eds.) Streamside Management: Forestry and Fishery Interactions, University of Washington, Seattle, 98-142.

Fausey, N.R., Hall, G.F., Bigham, J.M., Allred, B.J., Christy, A.D., 2000. Properties of the fractured glacial till at the Madison County, Ohio, Field Workshop Pit Site. The Ohio Journal of Science 100, 107-112.

Ferguson, R.I., Hoey, T.B., Wathen, S., Werritty, A., 1996. Field evidence for rapid downstream fining of river gravels through selective transport. Geology 24, 179-182.

Finnegan, N.J., Sklar, L.S., Fuller, T.K., 2007. Interplay of sediment supply, river incision, and channel morphology revealed by the transient evolution of an experimental bedrock channel. Journal of Geophysical Research Earth Surface 112, F03S11.

Fitzpatrick, F.A., Peterson, E.M., Stewart, J.S., 1996. Habitat Characteristics of Benchmark Streams in Agricultural Areas of Eastern Wisconsin. U.S. Geological Survey Water-Resources Investigations Report 96-4038-B, 35 p.

Fitzpatrick, F.A., Knox, J.C., and Whitman, H.E., 1999. Effects of historical land-cover changes on flooding and sedimentation, North Fish Creek, Wisconsin. U.S. Geological Survey Water-Resources Investigations Report 99-4083, 12 p.

Fitzpatrick, F.A., Peppler, M.C., Walker, J.F., Rose, W.J., Waschbusch, R.J., Kennedy, J.L., 2008. Flood of June 2008, southern Wisconsin. U.S. Geological Survey Scienific Investigations Report 2008-5235, 24 p.

Floresheim, J.L., Chin, A., Gaffney, K., Slota, D., 2013. Threshold of stability in incised “Anthropocene” landscapes. Anthropocene 2, 27-41.

Foster, G.E., 1998. Morphology of three semi-alluvial stream channels in southern Ontario. Unpublished Masters Thesis. The University of Western Ontario, 215 pp.

Fraser, C., Hill, P.R., Allard, M., 2005. Morphology and facies architecture of a falling sea level strandplain, Umiujaq, Hudson Bay, Canada. Sedimentology, 52, 141-160.

Fripp, J.B., Diplas, P., 1993. Surface sampling in gravel streams. Journal of Hydraulic Engineering 119, 473-490.

Geomorphic Solutions., 2009. Evaluating the effectiveness of ‘Natural’ Channel Design projects: An introduction and preliminary assessment of sites in TRCA’s jurisdiction. Prepared by: Geomorphic Solutions, Sernas Group Inc., The Toronto and Region Conservation Authority, LGL Limited, February 2009, 116 p.

Gibbard, P., 1980. The origin of stratified Catfish Creek Till by basal melting. Boreas 9, 71-85.

Gomez, B., Livingstone, D.M., 2012. The river it goes right on: Post-glacial landscape evolution in the upper Waipaoa River basin, eastern North Island, New Zealand. Geomorphology 159-160, 73-83.

Grabowski, R.C., Droppo, I.G., Wharton, G., 2011. Erodibility of cohesive sediment: The importance of sediment properties. Earth Science Reviews 105, 101-120.

Gran, K.B., Montgomery, D.R., 2005. Spatial and temporal patterns in fluvial recovery following volcanic eruptions: Channel response to basin-wide sediment loading at Mount Pinatubo, Philippines. GSA Bulletin 117, 195-211.

Gran, K.B., Belmont, P., Day, S.S., Jennings, C., Johnson, A., Perg, L., Wilcock, P.R., 2009. Geomorphic evolution of the Le Sueur River, Minnesota, USA and implications for current sediment loading. In: James, L.A., Rathburn, S.L., Whittecar, G.R., (Eds.) Restoration of fluvial systems with broad historical changes and human impacts. GSA Special paper 451, 119-130.

Gran, K.B., Belmont, P., Day, S.S., Finnegan, N., Jennings, C., Lauer, W.J., Wilcock, P.R., 2011. Landscape evolution in south-central Minnesota and the role of geomorphic history on modern erosional processes. GSA Today 21, 7-9.

Gran, K.B., Finnegan, N., Johnson, A.L., Belmont, P., Wittkop, C., Rittenour, T., 2013. Landscape evolution, valley excavation, and terrace development following abrupt postglacial base-level fall. Geological Society of America Bulletin 125, 1851-1864.

Greenbaum, N., Bergman, N., 2006. Formation and evacuation of a large gravel-bar deposited during a major flood in a Mediterranean ephemeral stream, Nahal Me'arot, NW Israel. Geomorphology 77, 169-186.

Grisak, G.E., Cherry, J.A., 1975. Hydrologic characteristics and response of fractured till and clay confining a shallow aquifer. Canadian Geotechnical Journal 12, 23-43.

Hack, J.T., 1965. Postglacial drainage evolution and stream geometry in the Ontonagon area, Michigan. USGS Professional Paper 504-B.

Hagerty, T.P., Kingston, M.S., 1992. The soils of Middlesex County. Report no. 56 of the Ontario Centre for Soil Resource Evaluation, Resources Management Branch, Ontario Ministry of Agriculture and Food, Guelph, Ontario, volume 1, 213 pp.

Hancock, G.S., Anderson, R.S., Whipple, K.X., 1998. Beyond power: Bedrock river incision process and form. In: Tinkler, K., and Wohl, E.E., (Eds.), Rivers over rock: Fluvial processes in bedrock channels. American Geophysical Union Monograph 107, 35-60.

Hanson, G.J. Simon, A., 2001. Erodibility of cohesive streambeds in the loess area of the midwestern USA. Hydrological Processes 15, 23-38.

Hartley, A.M., 1999. The pool-riffle sequence as the principal design component of low-gradient, meandering, gravel-bed channels. M.Sc. Thesis, School of Engineering, University of Guelph, Guelph, Ontario, Canada, 158 pp.

Haschenburger, J.K., Rice, S.P., Voyde, E., 2007. Evaluation of bulk sediment sampling criteria for gravel-bed rivers. Journal of Sedimentary Research 77, 415-423.

Hassan, M.A., Egozi, R., Parker, G., 2006. Experiments on the effect of hydrograph characteristics on vertical grain sorting in gravel bed rivers. Water Resources Research 42, W09408.

Hassan, M.A., Brayshaw, D., Alila, Y., Andrews, E., 2014. Effective discharge in small formerly glaciated mountain streams of British Columbia: Limitations and implications. Water Resources Research 50, 4440-4458.

Hayes, S.K., Montgomery, D.R., Newhall, C.G., 2002. Fluvial sediment transport and deposition following the 1991 eruption of Mount Pinatubo. Geomorphology 45, 211-224.

Hicock, S.R. Dreimanis, A. 1992. Deformation till in the Great Lakes region: implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. Canadian Journal of Earth Sciences 29, 1565-1579.

Hill, A.R. 1973. Erosion of river banks composed of glacial till near Belfast, Northern Ireland. Zeitschrift für Geomorphologie 17, 428-442.

Hrytsak, T.N., 2012. Alluvial cover dynamics in a semi-alluvial channel. MSc Thesis, Department of Geography, Western University, 156 pp.

Jepsen, R., McNeil, J., Lick, W., 2000. Effects of gas generation on the density and erosion of sediments from the Grand River. Journal of Great Lakes Research 26, 209-219.

Jordan, B.A., Annable, W.K., Watson, C.C., Sen, D., 2010. Contrasting stream stability characteristics in adjacent urban watersheds: Santa Clara Valley, California. River Research and Applications 26, 1281-1297.

Kamphuis, J.W., 1983. On the erosion of consolidated clay material by a fluid containing sand. Canadian Journal of Civil Engineering 10, 223-231.

Kamphuis, J.W., Gaskin, P.N., Hoogendoorn, E., 1990. Erosion tests on four intact Ontario clays. Canadian Geotechnical Journal 27, 692-696.

Kappesser, G., 2002. A riffle stability index to evaluate sediment loading to streams. Journal of the American Water Resources Association 38, 1069-1081.

Kaufmann, P.R., Faustini, J.M. Larsen, D.P. Shirazi, M.A., 2008. A roughness-corrected index of relative bed stability for regional stream surveys. Geomorphology 99, 150-170.

Kaufmann, P.R., Larsen, D.P., Faustini, J.M., 2009. Bed stability and sedimentation associated with human disturbances in Pacific Northwest streams. Journal of American Water Resources Association 45, 434-459.

Kellerhals, R. Bray, D.I. 1971. Sampling procedures for coarse fluvial sediments. Journal of the Hydraulics Division, ASCE, 97, 1165-1179.

Kellerhals, R., Church., M. Bray, D.I., 1976. Classification and analysis of river processes. Journal of the Hydraulics Division, 102, 813-829.

Khan, I., Kostaschuk, R., 2011. Erodibility of cohesive glacial till bed sediments in urban stream channel systems. Canadian Journal of Civil Engineering 38, 1363-1372.

Kling, G.W., Hayhoe, K., Johnson, L.B., Magnuson, J.J., Polasky, S., Robinson, S.K., Shuter, B.J., Wander, M.M., Wuebbles, D.J., Zak, D.R., Lindroth, R.L., Moser, S.C., Wilson, M.L., 2003. Confronting Climate Change in the Great Lakes Region: Impacts on our Communities and Ecosystems. Union of Concerned Scientists, Cambridge, Massachusetts, and Ecological Society of America, Washington, D.C.

Knighton A.D., 1980. Longitudinal changes in size and sorting of stream bed material in four English rivers. Geological Society of America Bulletin 91, 55-62.

Knox, J.C., 1989. Long and short-term episodic storage and removal of sediment in watersheds of southwestern Wisconsin and northwestern Illinois. In: Hadley, R.F., Ongley, E.D., (Eds.), Sediment and the Environment. International Association of Hydrological Sciences, Wallingford, UK, 157-164.

Knox, J.C., 2006. Floodplain sedimentation in the Upper Mississippi Valley: Natural versus human accelerated. Geomorphology 79, 286-310.

Kondolf, M.G., 1995. Geomorphological stream channel classification in aquatic habitat restoration: uses and limitations. Aquatic Conservation: Marine and Freshwater Ecosystems 5, 127-141.

Kondolf, M.G., 1997. Applications of the pebble count notes on purpose, method and variants. Journal of the American Water Resources Association 33, 79-87.

Landwehr, K., Rhoads, B.L., 2003. Depositional response of headwater stream to channelization, east central Illinois, USA. River Research and Application 19, 77-100.

Laronne, J.B., Reid, I., Yitshack, Y., Frostick, L.E., 1994. The non-layering of gravel streambeds under ephemeral flood regimes. Journal of Hydrology 159, 353-363.

Leopold, L.B. 1994. A view of the river. Harvard University Press, Cambridge MA, 298 pp.

Leopold, L.B., Wolman, M.G., 1957. River channel patterns: braided, meandering, and straight. Geological Survey Professional Paper 282-B, 85 pp.

Lewis, C.F.M., Blasco, S.M., Gareau, P.L., 2005. Glacial isostatic adjustment of the Laurentian Great Lakes basin: using the empirical record of strandline deformation for reconstruction of early Holocene paleo-lakes and discovery of a hydrologically closed phase. Géographie physique et Quaternaire 59, 187-210.

Lisle, T.E., 1995. Particle size variations between bedload and bed material in natural gravel bed channels. Water Resources Research 31, 1107-1118.

Lisle, T.E. Hilton, S., 1992. The volume of fine sediment in pools: An index of the supply of mobile sediment in stream channels. Water Resources Bulletin 28, 371-383.

Lisle, T.E., Hilton, S., 1999. Fine bed material in pools of natural gravel-bed channels. Water Resources Research 35, 1291-1304.

Lisle, T.E., Madej, M.A., 1992. Spatial variation in armouring in a channel with high sediment supply. In: Billi, P., Hey, R.D. Thorne, C.R., Tacconi, P. (Eds.) Dynamics of Gravel-Bed Rivers. John Wiley & Sons, New York, 277-293.

Lisle, T.E., Buffington, J.M., Wilcock, P.R., Bunte, K., 2015. Can rapid assessment protocols be used to judge sediment impairment in gravel-bed streams? A commentary. Journal of the American Water Resources Association 51, 373-387.

Mackin, J.H., 1948. Concept of the graded river. Geological Society of America Bulletin 48, 463-512.

MacVicar, B.J., Roy, A.G., 2011. Sediment mobility in a forced riffle-pool. Geomorphology 125, 445-456.

Major, J.J., Pierson, T.C., Dinehart, R.L., Costa, J.E., 2000. Sediment yield following severe volcanic disturbance - a two-decade perspective from Mount St. Helens. Geology 28, 819-822.

Marchand, J-P., Buffin-Bélanger, T., Hétu, B., St-Onge, G., 2014. Stratigraphy and infill history of the glacially eroded Matane River Valley, eastern Quebec, Canada. Canadian Journal of Earth Sciences 51**,** 105-124.

Marchildon, M.A., Annable, W.K., Imhof, J.G., Power, M., 2011. A high-resolution hydrodynamic investigation of brown trout (*Salmo trutta*) and rainbow trout (*Oncorhynchus mykiss*) redds. River Research and Application 27, 345-359.

Marren, P.M., Toomath, S.C., 2014. Channel pattern of proglacial rivers: topographic forcing due to glacier retreat. Earth Surface Processes and Landforms 39, 943-951.

Martini, I.P., 1977. Gravelly flood deposits of Irvine Creek, Ontario, Canada. Sedimentology 24, 603-622.

Matson, P.A., Parton, W.J., Power, A.G., Swift, M.J., 1997. Agricultural intensification and ecosystem properties. Science 277, 504-509.

McNeil, J., Lick, W., 2001. Effects of sediment bulk properties on erosion rates. The Science of the Total Environment 266, 41-48.

McNeil, J., Lick, W., 2004. Erosion rates and bulk properties of sediments from the Kalamazoo River. Journal of Great Lakes Research 30, 407-418.

Meshkova, L.V., Carling, P.A., Buffin-Be´langer, T., 2012. Nomenclature, complexity, semi-alluvial channels and sediment-flux-driven bedrock erosion. In: Church, M., Biron, P.M., Roy, A.G., (Eds.) Gravel-bed Rivers: Processes, Tools, Environments. John Wiley & Sons, Ltd., Chapter 31, 424-431. Michigan Department of Natural Resources., 1991, Qualitative biological and habitat survey protocols for wadable streams and rivers. Michigan Department of Natural Resources, Surface Water Quality Division, Great Lakes and Environmental Assessment Section, GLEAS Procedure 51, 40 p.

Mier, J.M., Garcia, M.H., 2011. Erosion of glacial till from the St. Clair River (Great Lakes basin). Journal of Great Lakes Research 37, 399-410.

Miller, M.W., Nudds, T.D., 1996. Prairie landscape change and flooding in the Mississippi River Valley. Conservation Biology 10, 847-853.

Montgomery, D.R., Buffington, J.M., 1997. Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin 109, 596-611.

Myers, T.J., Swanson, S., 1996. Temporal and geomorphic variations of stream stability and morphology: Mahogany Creek, Nevada. Water Resources Bulletin 32, 253-265.

Ness, R., Joy, D.M., 2002. Performance of Natural Channel Designs in southwestern Ontario. Canadian Water Resources Journal 27, 293-315.

Newson, M.D., Newson, C.L., 2000. Geomorphology, ecology and river channel habitat: mesoscale approaches to basin-scale challenges. Progress in Physical Geography 24, 195-217.

Nielsen, E., Morgan, A.V., Morgan, A., Mott, R.J., Rutter, N.W., Causse, C., 1986. Stratigraphy, paleoecology, and glacial history of the Gillam area, Manitoba. Canadian Journal of Earth Sciences 23, 1641-1661.

Niziol, T.A., Snyder, W.R., Waldstreicher, J.S., 1995. Winter weather forecasting throughout the eastern United States. Part IV: Lake effect snow. Weather Forecasting 10, 61-77.

Novotny, E.V., Stefan, H.G., 2007. Stream flow in Minnesota: Indicator of climate change. Journal of Hydrology 334, 319-333.

Nunally, N.R. 1967. Definition and identification of channel and over-bank deposits and their respective role in floodplain formation. Professional Geographer 19, 1-4.

O.A.C., 1931. Soil survey of the county of Middlesex, province of Ontario. Soil Survey Report, Department of Chemistry, O.A.C., Guelph.

Olsen, D.S., Whittaker, A.C., Potts, D.F., 1998. Assessing stream channel stability thresholds using flow competence estimates at bank-full stage. Journal of the American Water Resources Association 33, 1197-1207.

O'Neill, M.P., Abrahams, A.D., 1984., Objective identification of pools and riffles. Water Resources Research 20, 921-926.

Pahl, R., 2006. Carson River relative bed stability investigation. Nevada Division of Environmental Protection, Bureau of Water Quality Planning, 15 pp.

Paola, C., Seal, R., 1995. Grain size patchiness as a cause of selective deposition and downstream fining. Water Resources Research 31, 1395-1407.

Parker, G., 1990. Surface-based bedload transport relation for gravel rivers. Journal of Hydraulic Research 28, 417-436.

Parker, G., Klingeman, P.C., 1982. On why gravel bed streams are paved. Water Resources Research 18, 1409-1423.

Peterson, T.J., Western, A.W., Argent, R.M., 2012. Analytical methods for ecosystem resilience: A hydrological investigation, Water Resources Research 48, W10531.

Pfankuch, D.J., 1975. Stream reach inventory and channel stability evaluation. U.S. Department of Agriculture Forest Service. Region 1. Missoula, Montana, 25 pp.

Phillips, J.D., 2003. Sources of nonlinear complexity in geomorphic systems. Progress in Physical Geography 26, 339-361.

Phillips, J.D., 2014. Global tipping points: Some lessons from studies of thresholds in Geomorphology. Available in: https://geography.as.uky.edu/sites/default/files/Tipping%20Points.pdf. Last accessed October 5, 2014.

Phillips, R.T.J., Robert, A., 2005. River incision in relation to post-glacial events in the Humber River basin, Ontario. Géographie physique et Quaternaire 59, 17-30.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low-relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Phillips, R.T.J., Desloges, J.R., 2015a. Alluvial floodplain classification by multivariate clustering and discriminant analysis for low-relief glacially conditioned river catchments. Earth Surface Processes and Landforms 40, 756-770.

Phillips, R.T.J., Desloges, J.R., 2015b. Glacial legacy effects on river landforms of the southern Laurentian Great Lakes. Journal of Great Lakes Research 41, 951-964.

Pike, L., 2014. The dynamics of glacial till erosion: hydraulic flume tests on samples from Medway Creek, London, ON. Unpublished Master of Civil Engineering thesis, McGill University, Montreal, 99 pp.

Powell, D.M., 1998. Patterns and processes of sediment sorting in gravel-bed rivers. Progress in Physical Geography 22, 1-32.

Powell, D.M., Reid, I., Laronne, J.B., 1999. Hydraulic interpretation of cross-stream variations in bed-load transport. Journal of Hydraulic Engineering 125, 1243-1252.

Powell, G., 2006. Examination, application, and evaluation of geomorphic principles and resulting water quality in Midwest agricultural streams and rivers. PhD dissertation, Food, Agricultural, and Biological Engineering Ohio State University. Available from: https://etd.ohiolink.edu/

Prasicek, G., Otto, J-C., Montgomery, D.R., Schrott, L., 2014. Multi-scale curvature for automated identification of glaciated mountain landscapes. Geomorphology 209, 53-65.

Prasicek, G., Otto, J-C., Montgomery, D.R., Schrott, L., 2015. Glaciated valleys in Europe and western Asia. Journal of Maps 11, 361-370.

Rathburn, S., Wohl, E., 2003. Predicting fine sediment dynamics along a pool-riffle mountain channel. Geomorphology 55, 111-124.

Renwick, W.H., 1977. Erosion caused by intense rainfall in a small catchment in New York State. Geology 5, 361-364.

Rhoads, B.L., Cahill, R.A., 1999. Geomorphological assessment of sediment contamination in an urban stream system. Applied Geochemistry 14, 459-483.

Rice, S., 1998. Which tributaries disrupt downstream fining along gravel-bed rivers? Geomorphology 22, 39-56.

Rice, S.P., Church, M., 1996. Sampling surficial fluvial gravels: The precision of size distribution percentile estimates. Journal of Sedimentary Research 66, 654-665.

Rice, S.P., Hascheburger, J.K., 2004. A hybrid method for size characterization of coarse subsurface fluvial sediments. Earth Surface Processes and Landforms 29, 373-389.

Richards, R.P., 1990. Measures of flow variability and a new flow based classification of Great Lakes tributaries. Journal of Great Lakes Research 16, 53-70.

Richardson, K., Carling, P.A., 2005. A typology of sculpted forms in open bedrock channels. Geological Society of America Special Papers 392, 108 pp.

Riebe, C.S., Sklar, L.S., Lukens, C.E., Shuster, D.L., 2015. Climate and topography control the size and flux of sediment produced on steep mountain slopes. Proceedings, National Academy of Sciences (PNAS) 112, 15574-15579.

Roberts, J., Jepsen, R., Gotthard, D., Lick, W., 1998. Effects of particle size and bulk density on erosion of quartz particles. Journal of Hydraulic Engineering 124, 261-1267.

Rosgen, D.L., 1994. A classification of natural rivers. Catena 22, 169-199.

Rosgen, D.L., 1996. Applied River Morphology. Wildland Hydrology, Pagosa Springs, Colorado.

Rozin, U., Schick, A.P., 1996. Land use change, conservation measures and streamchannel response in theMediterranean/semiarid transition zone: Nahal Hoga, southern coastal plain, Israel. Erosion and Sediment Yield: Global and Regional Perspectives. IAHS Publ. 236, 427-444.

Ruhe, R.V., 1952. Topographic discontinuities of the Des Moines lobe. American Journal of Science 250, 46-56.

Rust, B.R., 1977. A classification of alluvial channel systems. Fluvial Sedimentology Memoir 5, 187-198.

Sambrook Smith, G.H., 1996. Bimodal fluvial bed sediments: origin, spatial extent and processes. Progress in Physical Geography 20, 402-417.

Schottler, S.P., Ulrich, J., Belmont, P., Moore, R., Lauer, W.J., Engstrom, D.R., Almendinger, J.E., 2013. Twentieth century agricultural drainage creates more erosive rivers. Hydrological Processes 28, 1951-1961.

Schumm, S.A., 1973. Geomorphic thresholds and complex response of drainage systems. In Morisawa, M. (Ed.) Fluvial geomorphology. Binghamton, NY, Publications in Geomorphology, 299-310.

Schumm, S.A., Harvey, M.D., Watson, C.C., 1984. Incised Channels: Morphology, Dynamics and Control. Water Resources Publication, Littleton, Co. 200 pp.

Seginer, I., 1966. Gully development and sediment yield. Journal of Hydrology 4, 236-253.

Shen, Z., Törnqvist, T.E., Autin, W.J., Mateo, Z.R.P., Straub, K.M., Mauz, B., 2012 Rapid and widespread response of the Lower Mississippi River to eustatic forcing during the last glacial-interglacial cycle. Geological Society of America Bulletin 124, 690-704.

Shugar, D., Kostaschuk, R., Ashmore, P., Desloges, J.R., Burge, L., 2007. In situ jet testing of the erosional resistance of cohesive streambeds. Canadian Journal of Civil Engineering, 34, 1192-1195.

Simon, A., 1994. Gradation processes and channel evolution in modified West Tennessee streams: Process, response and form. U.S. Geological Survey Professional Paper 1470, 63 pp.

Simon, A., Hupp, C.R., 1986. Chanel Evolution in Modified Tennessee channels. Proceedings, IV Federal Interagency Sedimentation Conference, Las Vegas, Nevada, March 24-27, volume 2, 5-71-5-82.

Sklar, L.S., Dietrich, W.E., Foufoula-Georgiou, E. Lashermes, B., Bellugi, D., 2006. Do gravel bed river size distributions record channel network structure? Water Resources Research 42, W06D18.

Stewart, A.M., Desloges, J.R., 2014. A 9000-year record of vertical and lateral accretion on the floodplain of the lower Thames River, southwestern Ontario, Canada, and implications for archaeological research. Quaternary Research 342, 161-172.

Stock, J.D., Montgomery, D.R., Collins, B.D., Dietrich, W.E., Sklar, L., 2005. Field measurements of incision rates following bedrock exposure: Implications for process controls on the long profiles of valleys cut by rivers and debris flows. Geological Society of America Bulletin 117, 174-194.

Stout, J.C., Belmont, P., Schottler, S.P., Willenbring, J.K., 2014. Identifying sediment sources and sinks in the Root River, southeastern Minnesota. Annals of the Association of American Geographers 104, 20-39.

Sutfin, N.A., Shaw, J., Wohl, E.E., Cooper, D.A., 2014. A geomorphic classification of ephemeral channels in a mountainous, arid region, southwestern Arizona, USA. Geomorphology 221, 164-175.

Syvitski, J.P.M., Milliman, J.D., 2007. Geology, geography, and humans battle for dominance over the delivery of fluvial sediment to the coastal ocean. The Journal of Geology 115, 1-19.

Taylor, C.H., 1977. Seasonal variations in the impact of suburban development on runoff response: Peterborough, Ontario. Water Resources Research 13, 464-467.

Taylor, C.H. Roth, D.M. 1979. Effects of suburban construction on runoff contributing zones in a small southern Ontario drainage basin. Hydrological Sciences-Bulletin 24, 279-301.

Teller, J.T., 1990. Volume and routing of late-glacial runoff from the southern Laurentide Ice Sheet. Quaternary Research 34, 12-23.

Terasmae, J., Karrow, P.F, Dreimanis, A. 1972. Quaternary stratigraphy and geomorphology of the eastern Great Lakes region of southern Ontario. XXIV International Geological Congress Guidebook, Excursion A42.

Thayer, B.J., 2010. Floodplain development for a semi-alluvial river. Unpublished Honor Thesis, Department of Geography, Western University, 87 p.

Thayer, B.J., Phillips, R.T.J., Desloges, J.R., 2016. Downstream channel adjustment in a low-relief, glacially conditioned watershed. Geomorphology, in press.

Thornbush, M.J., Desloges, J.R., 2011. Palaeoenvironmental reconstruction at an oxbow lake situated at the lower Nottawasaga River, southern Ontario, Canada. Environmental Archeology 16, 1-15.

Tinkler, J., Wohl, E.E., 1998a. (Eds.) Rivers Over Rock: Fluvial Processes in Bedrock Channels. Tinkler, K.J., and Wohl, E.E., (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union Geophysical Monograph 107, Washington, D.C., 323 p.

Tinkler, K., Wohl, E.E., 1998b. A primer on bedrock channels. In: Tinkler, K.J., and Wohl, E.E., (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union Geophysical Monograph 107, Washington, D.C., 1-18.

Tinkler, K.J., Parrish, J., 1998. Recent adjustments to the long profile of Cook Creek, an urbanized bedrock channel in Mississauga, Ontario. In: Tinkler, K.J., and Wohl, E.E., (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union Geophysical Monograph 107, Washington, D.C., 167-188.

Trenter, N.A., 1999. Engineering in Glacial Tills: CIRIA (Construction Industry Research and Information Association) Report C504, London, 259 pp.

Tridon Properties Ltd., 2014. 161 Windermere Road Vacant Land Condominium Development Environmental Impact Study. AECOM, Kitchener, ON, Appendix B – Borehole Logs.

Turowski, J.M., Hovius, N., Wilson, A., Horng, M., 2008. Hydraulic geometry, river sediment and the definition of bedrock channels. Geomorphology 99, 26-38.

Turowski, J.M., Rickenmann, D., 2009. Tools and cover effects in bedload transport observations in the Pitzbach, Austria. Earth Surface Processes and Landforms 34, 26-37.

Upper Thames Conservation Authority., 2013. Medway Creek 2012 Watershed Report Card. Available from: http://thamesriver.on.ca/wp-content/uploads//WatershedReportCards/RC_Medway.pdf. Last accessed: June 1, 2014.

Wang, L., Lyons, J., Kanehl, P., Gatti, R., 1997. Influences of watershed land use on habitat quality and biotic integrity in Wisconsin streams. Fisheries 22, 6-12.

Wang, L., Cuthbertson, A., Pender, G., Cao, Z., 2015. Experimental investigations of graded sediment transport under unsteady flow hydrographs. International Journal of Sediment Research 30, 306-320.

Ward, A.D., Mecklenburg, D., Powell, G.E., Jayakaran, A.D., D'Ambrosio, J., Draper, J., 2008. Creating ditches with floodplains. 16th National Nonpoint Source Monitoring Workshop September 14-18, 2008 Columbus, Ohio, 12 p.

Whipple, K.X., Hancock, G.S., Anderson, R.S., 2000a. River incision into bedrock: Mechanics and relative efficacy of plucking, abrasion, and cavitation. Geological Society of America Bulletin 112, 490-503.

Whipple, K.X, Snyder, N.P., and Dollenmayer, K., 2000b. Rates and processes of bedrock incision by the Upper Ukak River since the 1912 Novarupta ash flow in the Valley of Ten Thousand Smokes, Alaska. Geology 28, 835-838.

Wickert, A.D., Mitrovica, J.X., Williams, C., Anderson, R.S., 2013. Gradual demise of a thin southern Laurentide ice sheet recorded by Mississippi drainage. Nature 502, 668-674.

Wilcock, P.R., 1993. Critical shear stress of natural sediments. Journal of Hydraulic Engineering 119, 491-505.

Wilcock, P.R., Kenworthy, S.T., Crowe, J.C., 2001. Experimental study of the transport of mixed sand and gravel. Water Resources Research 37, 3349-3358.

Wilcock, P.R., and 12 others., 2009. Identifying sediment sources in the Minnesota River Basin. Minnesota River Sediment Colloquium June 30, 2009, Minnesota Pollution Control Agency, 16 p.

Wilcock, P.R., 2012. Stream restoration in gravel-bed rivers. In: Church, M., Biron, P.M., Roy, A.G., (Eds.) Gravel-bed Rivers: Processes, Tools, Environments. John Wiley & Sons, Ltd., Chapter 12, 137-146.

Wilkin, D.C., Hebel, S.J., 1982. Erosion, redeposition, and delivery of sediment to Midwestern streams. Water Resources Research 18, 1278-1282.

Wolcott, J., 1988. Nonfluvial control of bimodal grain-size distributions in river-bed gravels. Journal of Sedimentary Research 58, 979-984.

Wolman, M.G., 1954. A method of sampling coarse bed material. Eos Transactions, AGU 35, 951-956.

Woltemade, C.J., 1994. Form and process: Fluvial geomorphology and flood-flow interaction, Grant River, Wisconsin. Annals of the Association of American Geographers 84**,** 462-479.

Wood-Smith, R.D., Buffington, J.M., 1996. Multivariate geomorphic analysis of forest streams: implications for assessment of land use impacts on channel condition. Earth Surface Processes and Landforms 21, 377-393.

Appendix

Methods

a. Till sediment characteristics: Sediment bulk properties (Roberts et al., 1998; McNeil and Lick, 2001), particle size properties (Aberle et al., 2004), sediment cohesion properties and biological activity (Grabowski et al., 2011) determine material erodibility, in addition to other factors that interact with them in traction or impact (e.g. sand and gravel as bedload or particle saltation hits, air/water temperature that causes freeze-thaw cycles, its desiccation and cracking; Culley, 1971; Hill, 1973; Pike, 2014). Till was characterized by randomly collecting 44 till clasts of various sizes from the river bed. Each clast was air dried, color determined according to Munsell system, measured for size along its three axes (short, intermediate and long), sphericity calculated from the axes, precisely weighed with an analytical scale, volume determined (the sample within a nylon bag was inserted into a water container with known volume and the water displacement was recorded), as well as its submerged weight (clast weighed under water), bulk density (mass/volume) and porosity. Subsequently, each clast was disintegrated in water and when fully dispersed, passed through 2 mm and 63 micron sieves to separate the gravel, sand and fines and get their percentages after drying in an oven at 105°C. Each sample was photographed and described morphologically (cracks, borings of aquatic insects leading to bioturbation and roundness). Material strength was obtained from four borehole drillings, which were done at the top of the valley (above the left bank downstream for a proposed development) which included SPT (Standard Penetration Test) in the field and Atterberg limits for plasticity index in the lab (Tridon Properties

Limited, 2014). This technique is similar to the CPT (Cone Penetration Test) technique used by Dasenbrock et al. (2010) to characterise tills and other glacial deposits in Minnesota.

b. Topographic and morphological mapping: The 1.5 km reach of Medway Creek was mapped for bed, bank and floodplain elevations using a differential GPS (5500 points) during low flow (~ $1 m^3/sec$). The resulting topographic DEM (Digital Elevation Model) was used to extract a channel longitudinal profile. The profile was corrected for noise (Phillips and Desloges, 2014), after which precise subunits (~ 1 m resolution) were identified based on dominant bedforms (pools, riffles, flats, bars and steps) and exposed patches of till. Bedforms were identified from the DEM based on local slope and curvature of the long profile (Church and Jones, 1982 for bars; O'Neill and Abrahams, 1984 for pools and riffles; Chin, 1989 for step-pools and general bedform definition from Montgomery and Buffington, 1997). This bedform classification was visually verified in the field at low water level. Foster (1998) has shown this method is an effective procedure of bedform identification in three other semi-alluvial till-bedded streams of the area (Dingman Creek, Nissouri Creek and Oxbow Creek).

Both bank heights were measured in the field at 5 meter intervals using a stadia rod (307 points for the left bank, 309 for the right bank). These bank elevations were later corrected relative to the thalweg datum (i.e. the height between bank toe and thalweg elevation were added) and imposed on it to create two additional long profiles of the reach. This approach proved more precise than extraction of bank elevations from the DEM (or LiDAR) as it allowed finding a precise break in the cross-sectional slope into the floodplain according to high flow marks, top of a bar or lack of vegetation when the exposed till/soil topography was highly complex (and bankfull discharge was hard to determine). Channel widths were determined with a measuring tape in the field at 25 m intervals (62 points) to account for the cross-sectional complexity that cannot be accurately extracted from a DEM for reasons mentioned above. These data aided in developing the channel morphometry parameters, especially downstream hydraulic geometry, width-depth ratios and determining bankfull discharge.

c. Till patch mapping: Unlike channel bedforms, for which identification protocols exist (see **b**), there is currently no established procedure to identify and extract till exposures from a topographic map. Consequently, a separate GPS survey of the channel was conducted in the field to map all till patch exposures. This survey gave information about: how many patches exist, their size, their areal percentage of the total bed, spacing between them and relationship to other bedforms and large boulders. In addition, since till patches have varying degrees of exposures, they were visually classified to account for how much alluvium covers them (in percent). This classification helped in determining the annual erosion rates of till patches and their sampling strategy to represent the entire study site (see **h**). While **b** and **c** complement each other, the reason for separating them during the mapping phase is the bedforms and till patches do not always coincide (in terms of their borders and sizes). This allows better distinction and scale delineation of the till patches.

d. Morphologic analysis: Once the bed-feature map was created, it was analyzed in ArcGIS 10.1 (ESRI, 2012) and verified in the field to determine the spatial and morphologic characteristics (dimensions, spacing and distribution) of the dominant channel bedforms. A similar analysis of bedform morphology from a DEM (or long profile) was developed for step-pool channels by comparing it to visual measurements of the bedforms (Zimmermann et al., 2008). This determines the accuracy of the DEM.

e. Surface alluvium grain size distribution surveying: Each bedform (see **b**) underwent a grain size distribution (GSD) survey in order to characterize the alluvial cover. Each GSD survey of a bedform included at least three cross-sections of Wolman (1954) pebble counts (upstream, center and downstream) with at least 300 clast measurements of the b-axis per bedform as a minimum (Rice and Church, 1996). Due to the tendency to sample coarse material rather than finer sediment (Fripp and Diplas, 1993), and to avoid the misrepresentation of the fine tail, every time a clast smaller than 32 mm was encountered it was recorded as 'fines'. Once the GSD cross-section was complete, 2-5 kg of fines (< 32 mm) bulk sample from several locations in that bedform were bagged, taken to the lab, the silt-clay fractions were washed in a 63 micron sieve to avoid particle cohesion, air-dried and subjected to sieve analysis. The Wolman pebble counts (grid by number) and the bulk sieve analysis (volume by weight) can be combined into a single GSD for each bedform, without requiring conversion (Kellerhals and Bray, 1971). Such hybrid procedure was utilized successfully by Rice and Haschenburger (2004), as the larger samples of < 0.1% d_{max} criteria (Church et al., 1987) are impractical to sample due to their large size (~5 tons for a d_{max} > 128 mm; Haschenburger et al., 2007) and the disturbance such an invasive procedure would have on the bed structure. Relating the grain size distributions to the bedform map (see **b**) using statistical sedimentary attributes such as selected grain percentiles (d_5, d_{25}, d_{50}, d_{75}, d_{84}, d_{95}) and d_{max} also enabled detecting spatial trends, such as downstream fining/coarsening and deploying bed stability indices (Kaufmann et al., 2008; 2009; Kappesser, 2002; Lisle and Hilton, 1992; 1999).

f. Statistical grain size analysis: Statistical attributes (percentiles, sorting and types of distributions) were calculated for each bedform's GSD (see **e**). These statistics allowed comparison among similar bedforms and between different bedforms in the same channel.

g. Alluvial cover thickness and stratigraphy: In 26 random locations along the thalweg the alluvium (or its lack thereof) was characterized by measuring its depth to the boundary till by picking up clasts. In locations where alluvium was thick enough, pits were hand-dug. These pits revealed the vertical sedimentary structure of the alluvium (e.g. armor existence, framework or censored gravel stratigraphy).

h. Erosion measurement: To identify annual erosion rates of the till on a local scale, 94 erosion pins 25-30 cm long and 0.5 cm in diameter were inserted into three exposed till patches (between 25 and 35 pins per site): at the right bank of the channel (in 2010), at the center of the channel (in 2010), and in a riffle (in 2012). Using a 5 kg sledgehammer, erosion pins were carefully inserted, *i.e.* without visibly cracking the surface till (although subsurface cracks and micro-cracks may have developed due to the erosion pin's insertion), until flush with the bed. At the end of each hydrologic year, the exposure of each pin was precisely measured with a ruler, after which the pin was reinserted until it was again flush with the bed. Preliminary results indicated high erosion rates and substantial within-patch erosion variability. Consequently, the sampling strategy for the much larger scale of the entire study site was adapted. During hydrologic year 2013-2014, three additional sites were added on Medway Creek. The results were compared to longer term (geologic) incision rate from a topographic map (valley-scale) with the top of the valley serving as reference level divided by the time since the end of the last ice age. Results from the erosion analysis were compared to short-term channel incision rates for bedrock channels (Tinkler and Wohl, 1998a and references within; Stock et al., 2005). This comparison should account for the fact that Medway Creek is by not a natural river and accelerated incision is often associated with land clearance (Elliot, 1998), agriculture (Knox, 1989; Campo and Desloges, 1994; Woltenmade, 1994; Fitzpatrick et al., 1996) and particularly the construction of impervious surfaces in urban areas (Taylor and Roth, 1979; Fitzpatrick et al., 1999; Annable et al., 2012).

i. Stability analysis: Hydrologic systems, including rivers, bedforms (Chin, 1998) and reaches (Kaufmann et al., 2009), are often investigated for their current stability or over a longer time scale if historic data is available. Such resilience can be hydrologic (Peterson et al., 2012) or geomorphic (Doyle et al., 2000), and can be defined as: 1) steady state (equilibrium), 2) quasi-equilibrium, or 3) instability due to some kind of internal or external perturbation (Schumm, 1973). Since natural river systems are non-linear (Phillips, 2003) and could exhibit self-organized criticality (Coulthard and Van De Wiel, 2007), it is possible for a system to have several steady and unsteady states which require crossing of thresholds to move from one state to the next (Church, 2002). In geomorphology, at least three distinct notions of stability are used. Most commonly stability/instability is used as general shorthand to distinguish landscapes and landforms that are static (exhumed and no longer going through erosion or deposition), slowly changing, or in steady state (quasi-equilibrium - changes are within natural variability), vs. those that are undergoing rapid change. A second notion is that of mechanical stability, which concerns the conditions under which change occurs. Finally, there is dynamic stability which determines resilience of geomorphic systems and whether they are sensitive to small perturbations or minor variations in initial conditions (Phillips, 2014). We relate to the first only, as our data set is too short to detect longer term changes.

On the reach-scale, stability is generally defined as the ability of the stream to transport water and sediment of its watershed while maintaining its dimension, pattern and profile over time, i.e. without either aggrading or degrading (Pfankuch, 1975). Although stability changes in space and time (Myers and Swanson, 1996), the fact that each bedform repeats itself several times within the study reach allows stability determination at a specific point in time of that bedform, of all bedforms of that type and the generalization of the entire reach stability by combining all of them. Since different stability methods use different metrics, results might not be conclusive, and could even be contradictory to common assumptions (Doyle et al., 2000; Jordan et al., 2010; Annable et al., 2012).

There are two approaches to asses channel stability: qualitative and quantitative. A qualitative approach is based on professional judgment in the field of observed mass wasting of channel banks, riparian vegetation condition and bank profiles (Simon, 1994). The main advantage of this technique is its simplicity. Its major flaws are user bias and for our case also a lack of established protocol for a “healthy” stable reference reach to compare to, as these till channels have been disturbed for about two centuries (Campo and Desloges, 1994). For example, the six stage model of incised channel evolution by Schumm et al. (1984), based entirely on channel morphology, would classify our study reach as Stage I. This stage conforms to disequilibrium, sediment transport capacity exceeds sediment supply, bank height that is less than critical bank height, a U-shaped cross-section and W/D ratios at bankfull that are highly variable (discussed in Bergman et al 2016b). Phillips and Desloges (2014) recently described parts of southern Ontario large rivers as "glacially conditioned" but their work is hard to apply on a smaller, short reach-scale like our case study as they looked into entire watersheds long profiles and concurrent stream powers. Fitzpatrick et al. (1996) ranked twenty sites in agricultural areas of eastern Wisconsin as part of stream habitat characteristics in order to create benchmark streams. However, this complex classification (80 parameters) aimed at determining habitat quality according to the Michigan Department of Environmental Quality, Great Lakes Environmental Assessment Section (GLEAS) Procedure 51 (Michigan Department of Natural Resources, 1991), found no relation between GLEAS scores and relatively homogeneous units (RHU's) or the percentage of agricultural land in the drainage basins above the benchmark-stream sites. One of the RHU's they used included clayey surficial deposits like till on carbonate bedrock.

Figures

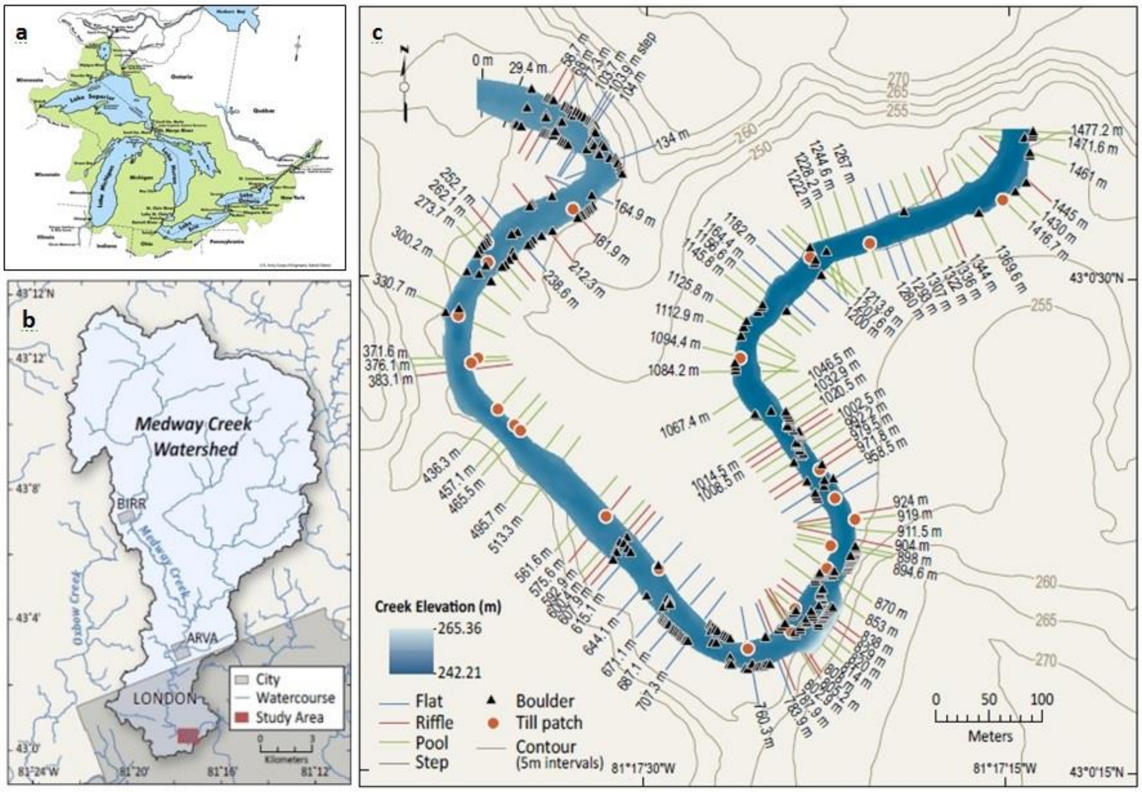


Figure 1. (a) Great Lakes, (b) The watershed of Medway Creek and (c) View of the study area with length along the thalweg profile and bed coloration according to elevation.

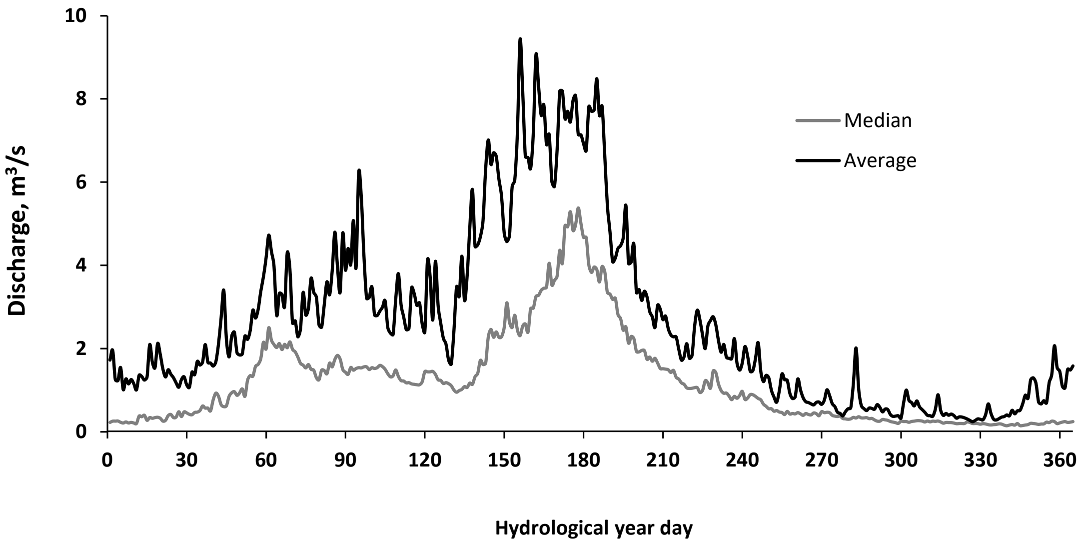


Figure 2. Average and median hydrographs for Medway Creek showing the flashy nature of the flows with a spring freshet peak discharge. Data is based on gauged data from 1945-2014.

Figure 3. The three most common bedforms on Medway Creek: pool (top), riffle (center) and flat (bottom).

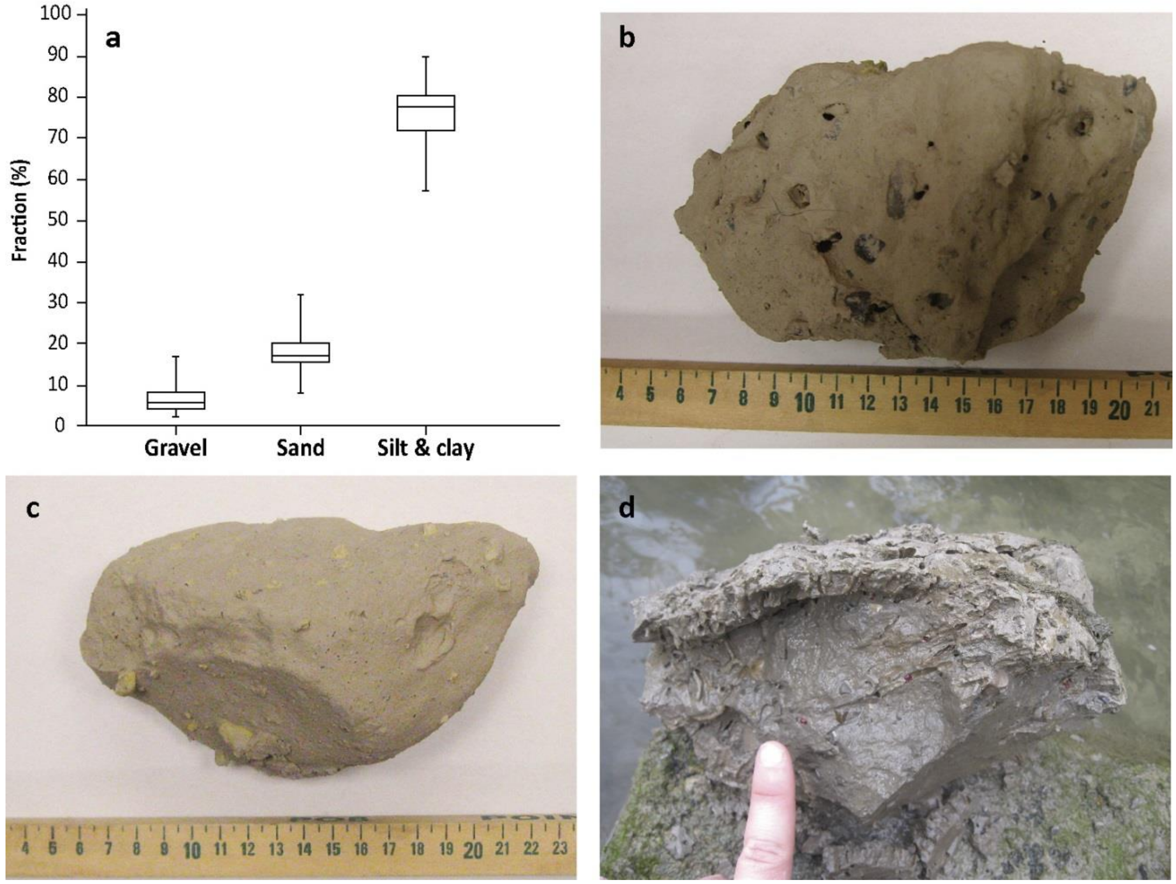


Figure 4. **Till clasts from Medway Creek bed. (a) Fractional breakdown of sediment sizes among all collected clasts (n = 44). The boxes indicate 25th and 75th percentiles as well as the median; the whiskers indicate 5th and 95th percentiles. (b) Sample dry clast showing bioturbation. (c) Sample dry clast showing protruding stones of the gravel fraction, covered by algae. (d) Sample wet clast, immediately following recovery from channel bed, showing burrowing chironomids (blood worms).**

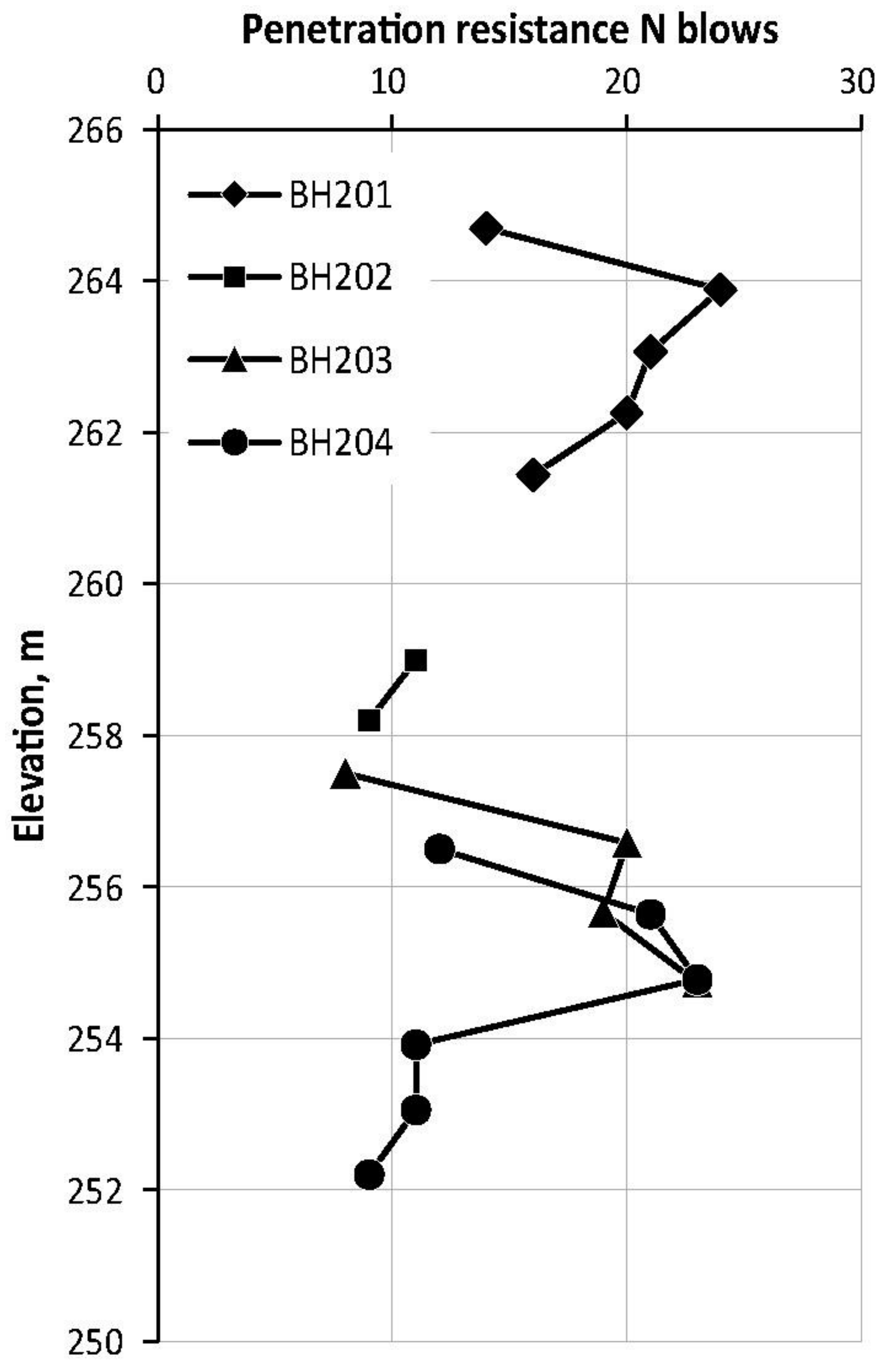


Figure 5. Penetration resistance (N blows) using Standard Penetration Test (SPT) into the local Medway Creek till conducted within four boreholes for a proposed development above the river valley. The topsoil and lacustrine layers are not included (Source: Tridon Properties Ltd., 2014).

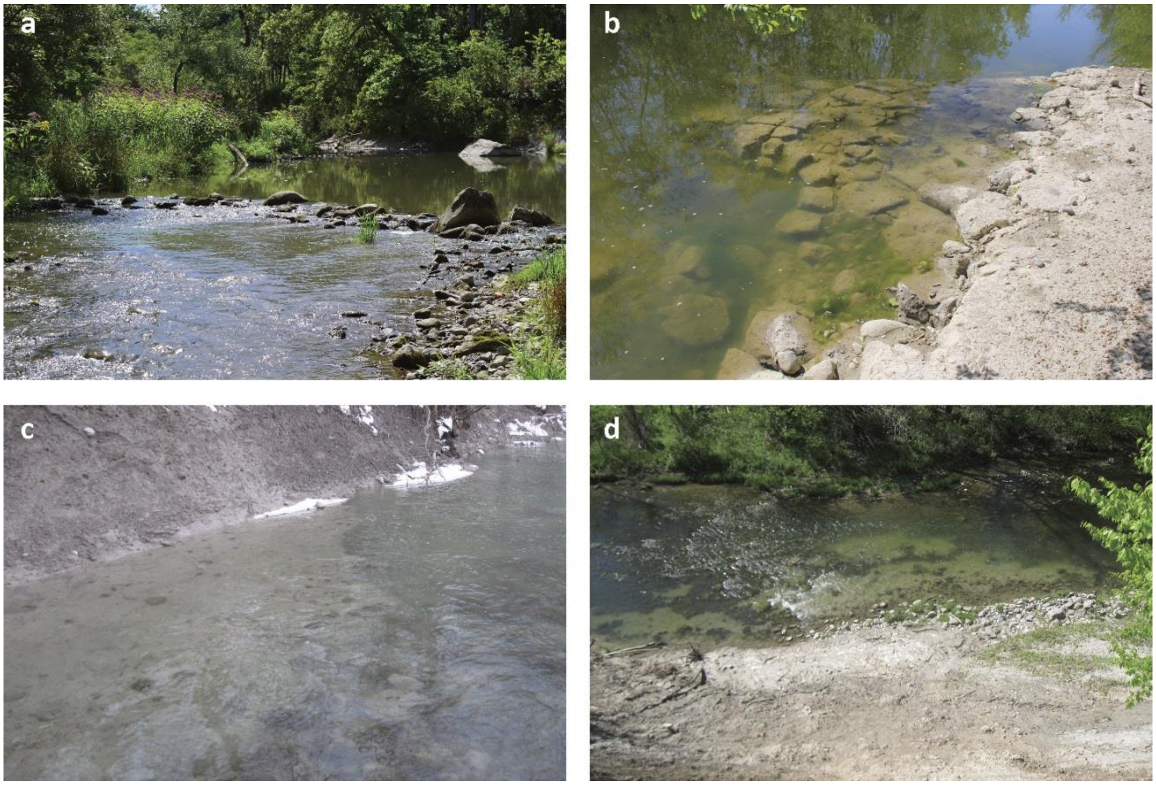


Figure 6. **Morphologic features along Medway Creek. (a) The downstream cobble-boulder step (view is upstream). Similar features were observed in other till rivers of the area. Note a giant 2 m boulder in the background. (b) Bed till ledge connected to the bank. During summer low flows the till is partially dry, cracks and breaks down into block-like clasts of various sizes that are incorporated into the channel bed alluvium (view is downstream). (c) Submerged bed till ledge during winter high flow. The ledge is smoothed and a new layer of till is exposed (view is upstream). (d) A planar view from the top of the right bank bluff into the channel (flow direction is left to right.). The most apparent features are the bed till exposures and the coarse boulder lag in the center-right of the photo.**

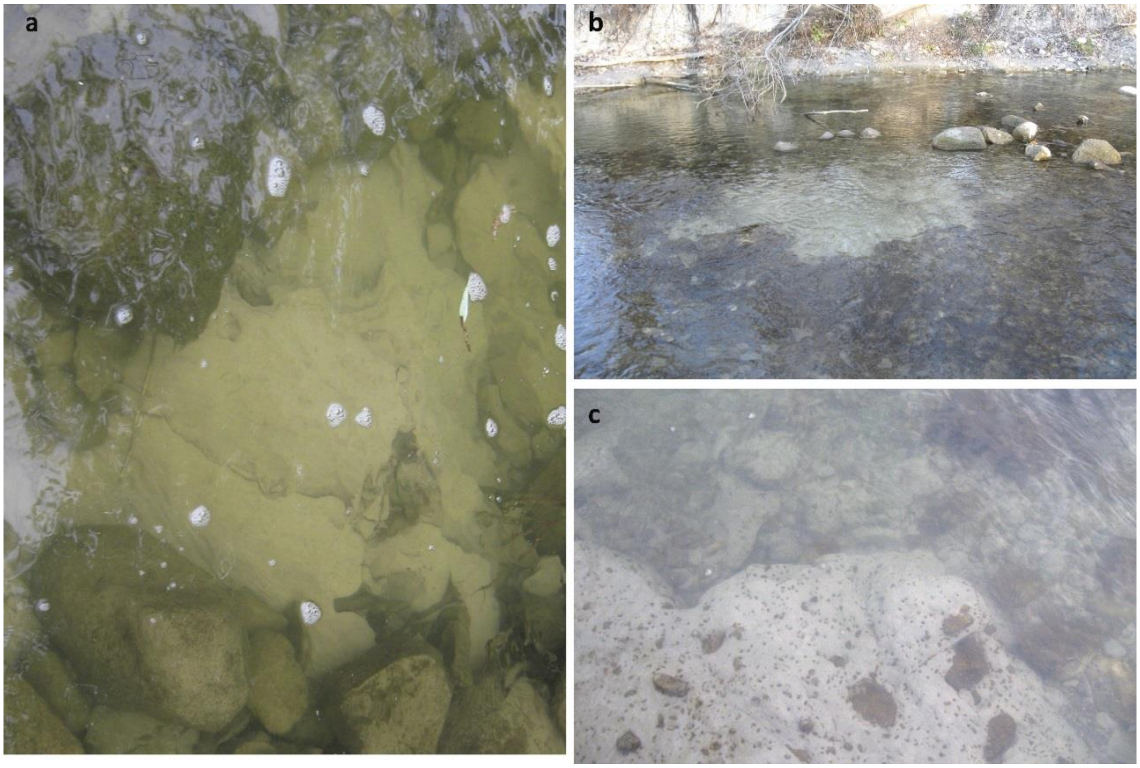


Figure 7. **Till exposures along Medway Creek. (a) Submerged bed till exposure with cracks and irregularities that are potential weak spots for erosion by the flow and surrounding gravels (flow direction is from top to bottom). (b) Center-channel till patch surrounded by much coarser stones with distinct coloration differences during low flow conditions (flow direction is from right to left). The boulders on the center right might be responsible for the boundary exposure which is located directly in their wake. (c) Till exposure with *in situ* gravel-sized stones (flow direction is from left to right). Note the large differences in grain sizes from few cobbles and boulders down to numerous granules that dot the till.**

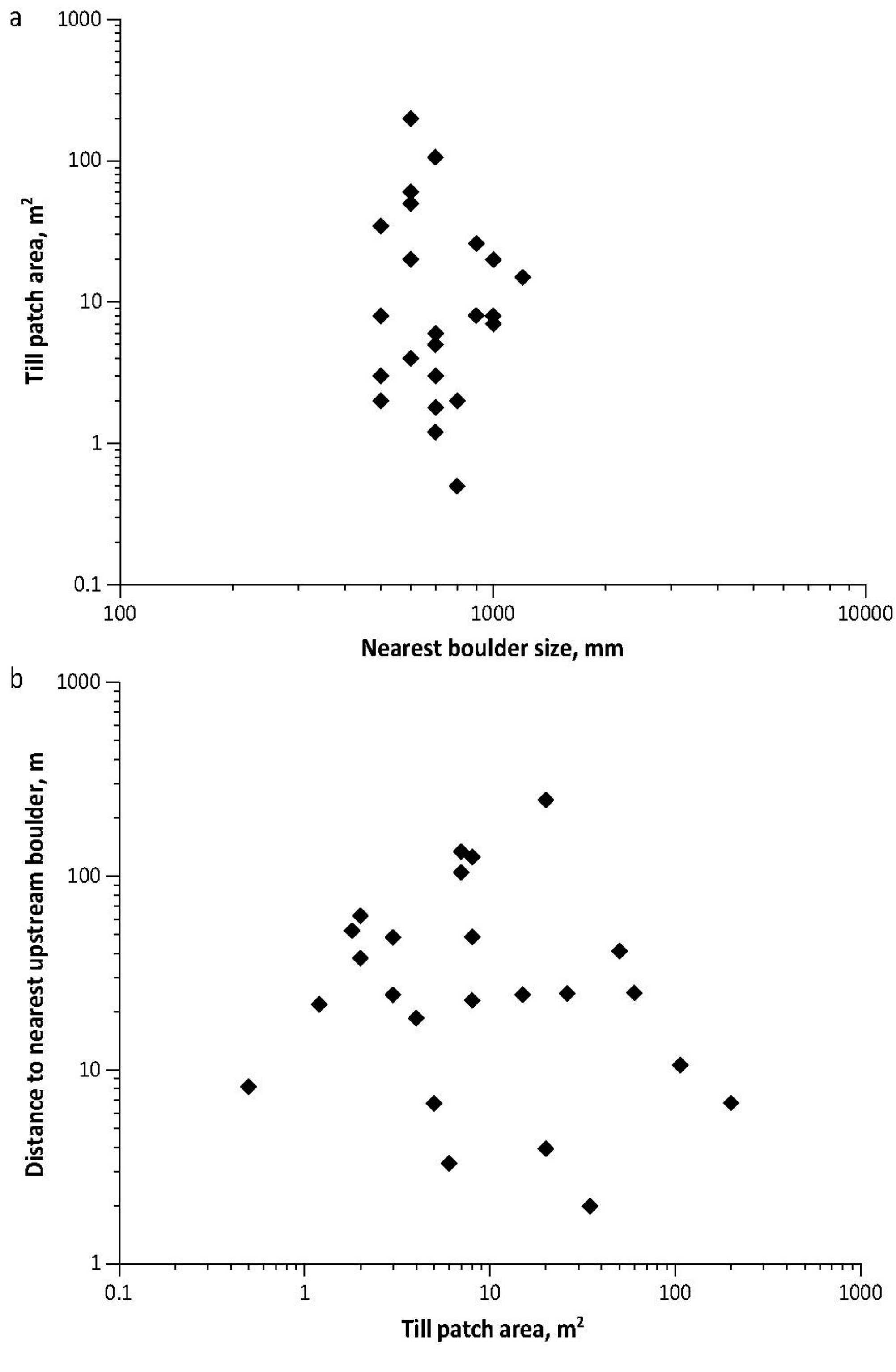


Figure 8. Relation between boulder size and till patch area (a) and between till patch area and distance to upstream boulder (b). These results show that the till patches are not controlled by the size and location of boulders.

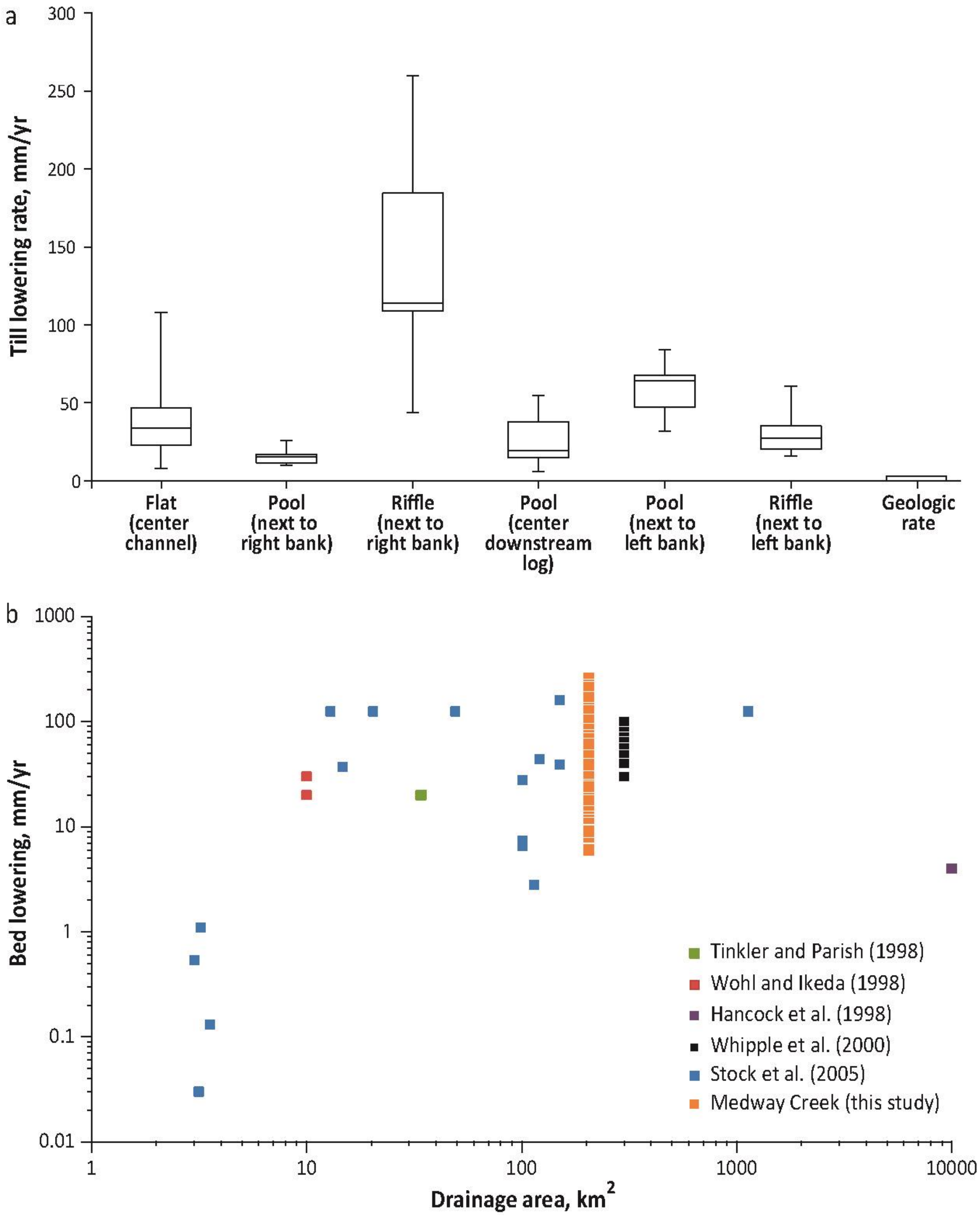


Figure 9. (a) Annual values for till erosion rates for HY 2013-2014 for all six patches using erosion pin measurements. The boxes indicate 25th and 75th percentiles as well as the median; the whiskers indicate 5th and 95th percentiles. Note the low average geologic rate in comparison. (b) A comparison of all Medway Creek's erosion pin data with other short-term rates of bedrock channel incision based on direct measurements.

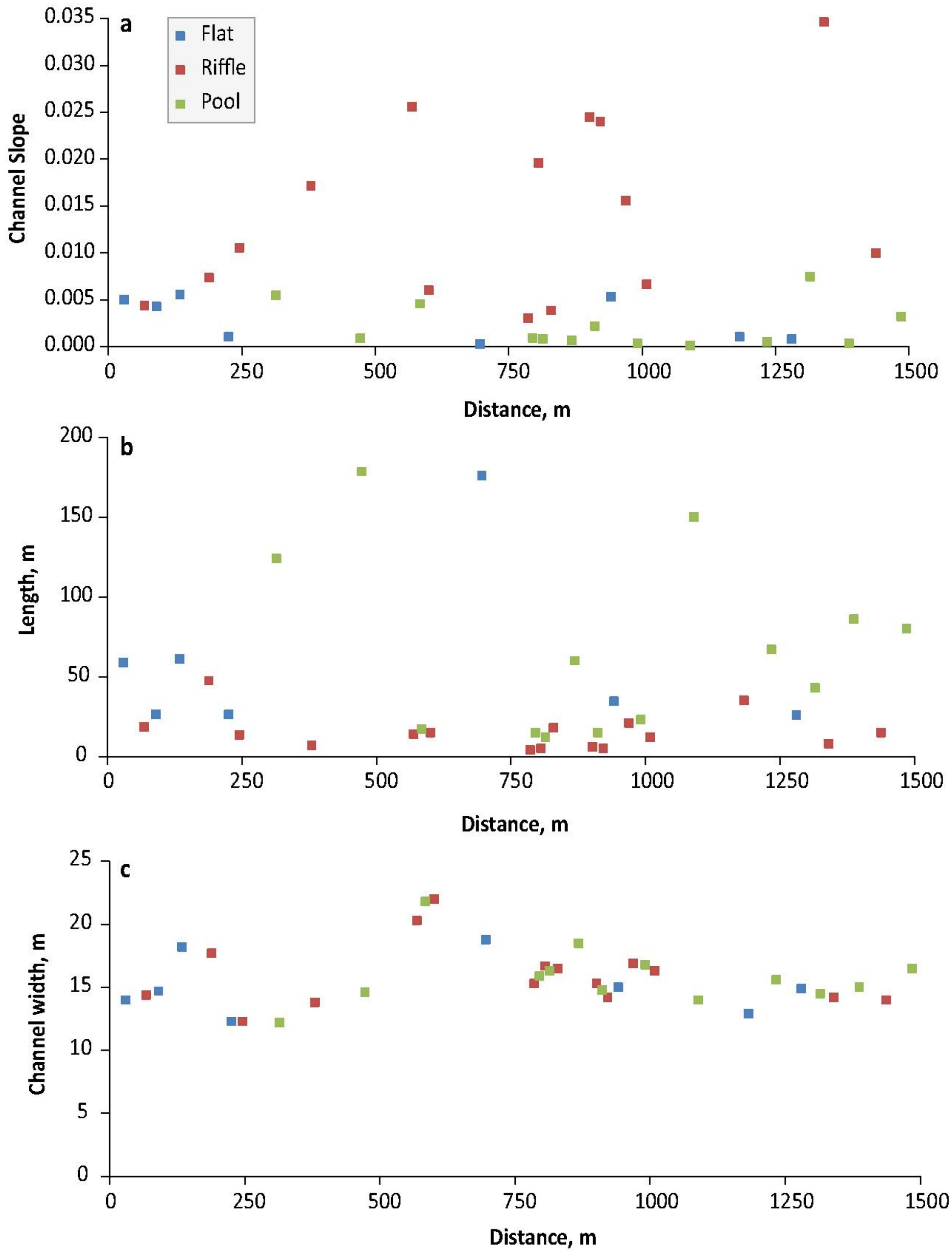


Figure 10. Bedform characteristics: (a) Slope. (b) Length. (c) Channel widths along the study reach.

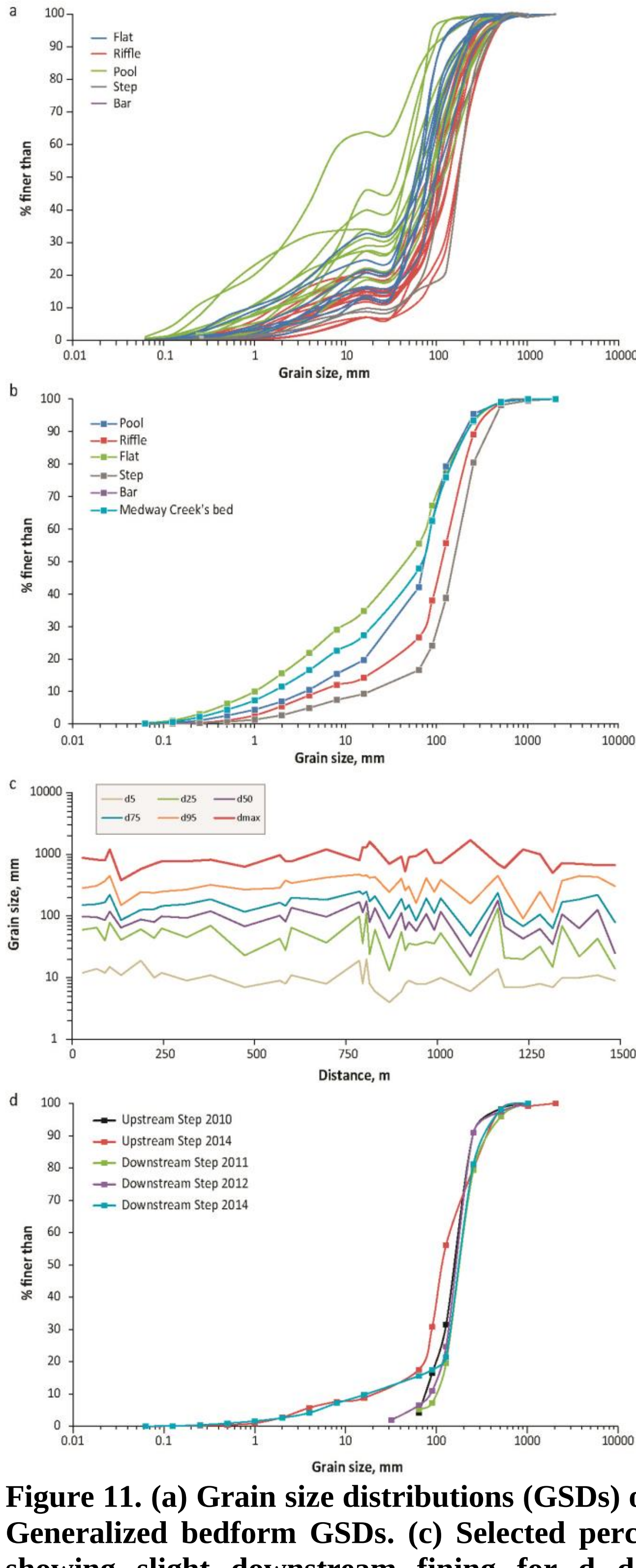


Figure 11. (a) Grain size distributions (GSDs) of all bedforms in the study reach. (b) Generalized bedform GSDs. (c) Selected percentiles of the GSDs along the reach showing slight downstream fining for d_5-d_{75} while d_{95} and d_{max} show slight coarsening. (d) Changes in the steps' GSDs.

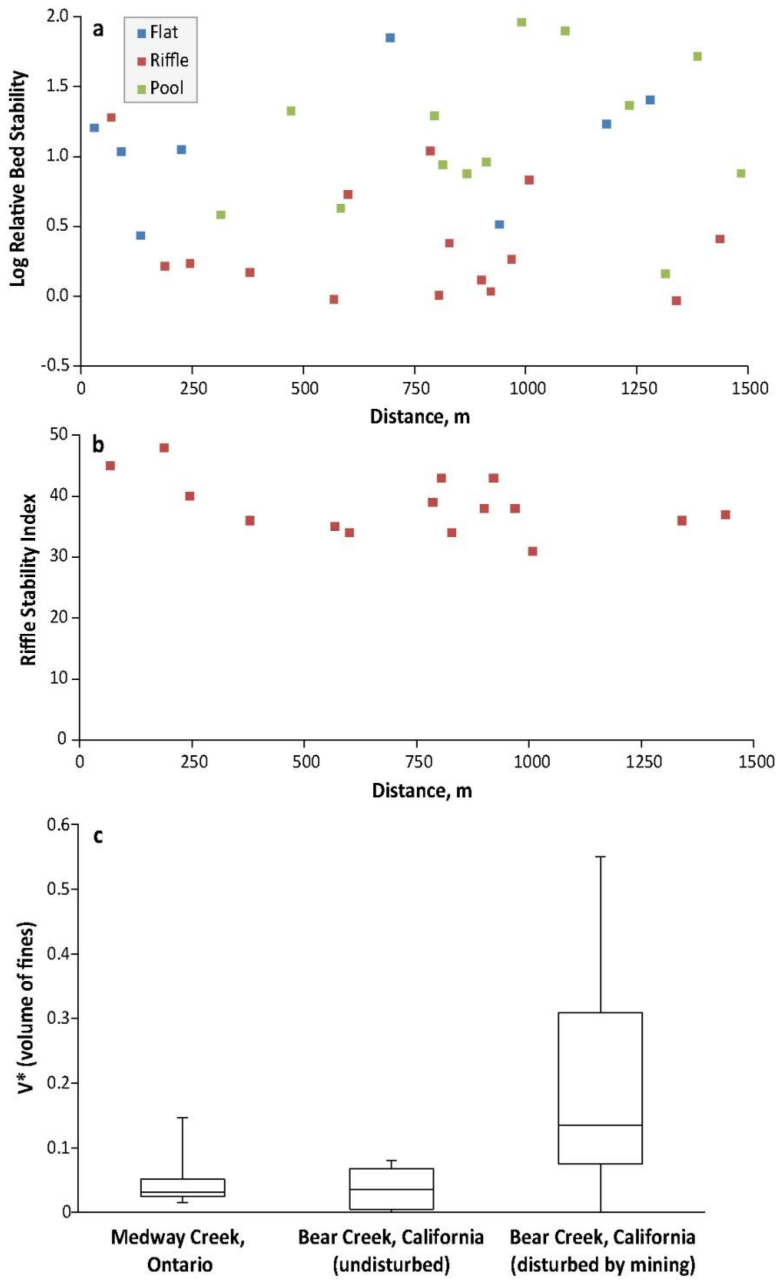


Figure 12. Three indices used to describe bed stability of the study reach. (a) The Kaufmann et al. (2008; 2009) LRBS* method. (b) Riffle Stability Index (RSI) according to Kappesser (2002). (c) relative volume of fine sediment in pools (Lisle and Hilton, 1992; 1999). Data from Lisle and Hilton (1992) of disturbed and undisturbed pools on Bear Creek California for comparison. The boxes indicate 25th and 75th percentiles as well as the median; the whiskers indicate 5th and 95th percentiles.

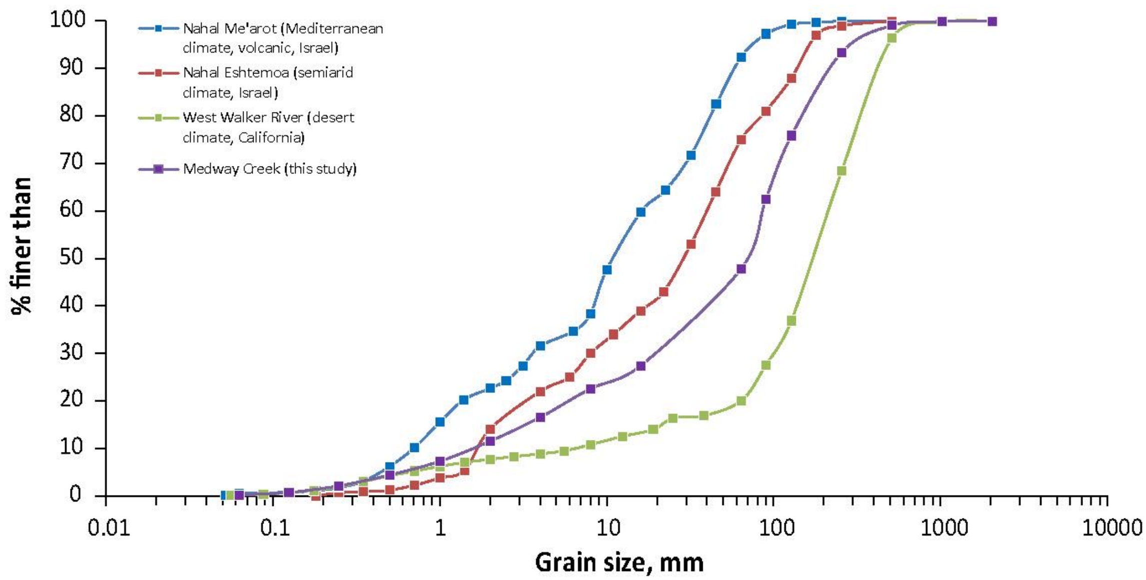


Figure 13. Comparison of Medway Creek GSD to three other streams that have an abundance of fines originating from their parent material.

Tables

Table 1. Till characteristics of selected clasts collected from the river bed.

#	Dry mass gr	Submerged mass gr	a-axis mm	b-axis mm	c-axis mm	Sphericity	Roundness	Volume equivalent in water cm^3	Bulk density p, gr/cm^3	Porosity %	Gravel gr	% Gravel	Sand gr	% Sand	Silt and clay gr	% Silt and clay	Bioturbated
1	190	199	90	80	20	0.23	0.3	139	1.37	4.52	6	3.16	15	7.89	169	88.95	No
2	238	254	80	60	70	0.12	0.2	137	1.74	6.30	5	2.10	24	10.08	209	87.82	Yes
3	323	336	105	80	60	0.15	0.5	217	1.49	3.87	12	3.72	38	11.76	273	84.52	Yes
4	1340	1375	160	120	50	0.21	0.4	666	2.01	2.55	33	2.46	109	8.13	1198	89.40	Yes
5	734	814	120	90	110	0.12	0.2	289	2.54	9.83	34	4.63	131	17.85	569	77.52	Yes
6	384	397	90	80	50	0.14	0.2	181	2.12	3.27	15	3.91	61	15.89	308	80.21	No
7	753	827	140	100	50	0.20	0.4	469	1.61	8.95	51	6.77	115	15.27	587	77.95	Yes
8	869	884	130	120	50	0.17	0.5	598	1.45	1.70	99	11.39	276	31.76	494	56.85	No
9	1385	1409	210	140	30	0.32	0.4	1020	1.36	1.70	102	7.36	212	15.31	1071	77.33	Yes
10	1209	1370	160	120	80	0.16	0.2	785	1.54	11.75	114	9.43	336	27.79	759	62.78	Yes
11	445	452	110	90	80	0.13	0.6	336	1.32	1.55	32	7.19	97	21.80	316	71.01	Yes
12	287	303	100	70	70	0.14	0.4	216	1.33	5.28	16	5.57	60	20.91	211	73.52	Yes
13	238	252	80	60	60	0.13	0.3	169	1.41	5.56	10	4.20	59	24.79	169	71.01	Yes
14	922	953	130	90	60	0.18	0.4	528	1.75	3.25	37	4.01	226	24.51	659	71.48	Yes
15	598	620	130	70	50	0.22	0.2	251	2.38	3.55	61	10.20	141	23.58	396	66.22	No
16	207	209	100	60	40	0.20	0.5	140	1.48	0.96	18	8.70	26	12.56	163	78.74	No
17	984	989	170	100	80	0.19	0.5	642	1.53	0.51	52	5.28	152	15.45	780	79.27	No
18	405	408	100	70	60	0.15	0.2	262	1.55	0.74	20	4.94	68	16.79	317	78.27	No
19	303	306	110	100	30	0.20	0.5	216	1.40	0.98	8	2.64	50	16.50	245	80.86	No
20	1087	1103	140	90	80	0.16	0.1	566	1.92	1.45	62	5.70	164	15.09	861	79.21	Yes
21	400	404	110	90	60	0.15	0.4	180	2.22	0.99	22	5.50	48	12.00	330	82.50	No
22	202	207	70	60	50	0.13	0.2	126	1.60	2.42	12	5.94	32	15.84	158	78.22	Yes
23	454	459	100	100	70	0.12	0.5	218	2.08	1.09	28	6.17	74	16.30	352	77.53	Yes
24	1575	1576	170	120	90	0.16	0.4	648	2.43	0.06	78	4.95	184	11.68	1313	83.37	No
25	385	387	100	70	60	0.15	0.4	250	1.54	0.52	30	7.79	76	19.74	279	72.47	No
26	1488	1490	170	110	100	0.16	0.6	906	1.64	0.13	128	8.60	208	13.98	1152	77.42	No
27	426	430	100	80	60	0.14	0.3	256	1.66	0.93	32	7.51	72	16.90	322	75.59	No
28	390	393	110	80	60	0.16	0.1	246	1.59	0.76	30	7.69	62	15.90	298	76.41	Yes
29	1318	1334	170	120	90	0.16	0.4	792	1.66	1.20	104	7.89	216	16.39	998	75.72	No
30	847	852	140	130	70	0.15	0.5	432	1.96	0.59	70	8.26	192	22.67	585	69.07	No
31	173	177	70	60	50	0.13	0.3	124	1.40	2.26	10	5.78	32	18.50	131	75.72	Yes
32	244	247	90	80	60	0.13	0.3	182	1.34	1.21	13	5.33	44	18.03	187	76.64	Yes
33	309	311	100	80	20	0.25	0.4	218	1.42	0.64	35	11.33	88	28.48	186	60.19	No
34	219	222	100	60	40	0.20	0.1	120	1.83	1.35	10	4.57	41	18.72	168	76.71	Yes
35	117	121	70	50	30	0.18	0.2	83	1.41	3.31	3	2.56	20	17.09	94	80.34	No
36	210	212	100	60	40	0.20	0.5	155	1.35	0.94	27	12.86	40	19.05	143	68.10	No
37	138	139	80	70	30	0.17	0.1	96	1.44	0.72	23	16.67	34	24.64	81	58.70	No
38	341	344	90	90	40	0.15	0.8	257	1.33	0.87	12	3.52	66	19.35	263	77.13	Yes
39	295	298	100	90	30	0.19	0.3	198	1.49	1.01	11	3.73	40	13.56	244	82.71	Yes
40	90	92	60	50	50	0.12	0.1	66	1.36	2.17	4	4.44	16	17.78	70	77.78	Yes
41	100	101	70	60	40	0.14	0.2	66	1.52	0.99	4	4.00	16	16.00	80	80.00	No
42	198	199	80	70	40	0.15	0.2	134	1.48	0.50	18	9.09	46	23.23	134	67.68	No
43	118	119	80	60	30	0.19	0.1	80	1.48	0.84	4	3.39	18	15.25	96	81.36	Yes
44	431	440	110	90	60	0.15	0.3	269	1.60	2.05	43	9.98	79	18.33	309	71.69	Yes
Minimum	90	92	60	50	20	0.12	0.1	66	1.32	0.06	3	2.10	15	7.89	70	56.85	
Maximum	1575	1576	210	140	110	0.32	0.8	1020	2.54	11.75	128	16.67	336	31.76	1313	89.40	
Average	531	546	111	85	56	0.17	0.3	317	1.64	2.40	35	6.38	93	17.80	403	75.82	
Median	385	390	100	80	55	0.16	0.3	218	1.54	1.28	25	5.64	64	16.85	289	77.37	

Table 2. Erosion rates of till patches during the study period.

Hydrological year	Peak discharge m^3s^{-1}	Site	Location in channel	Erosion pins #	Configuration W*L	Recovered #	Recovery %	Average erosion rate mm	Median erosion rate mm	Minimum erosion rate mm	Maximum erosion rate mm	Inter-patch erosion spatial autocorrelation Moran I	Moran I P-value
2010-2011	84	1	Flat (center channel)	20	4*5	19	95	143	134	106	225	0.019737	0.621129
		2	Pool (next to right bank)	53	5*10+3	48	91	80	78	44	143	0.399527	0.000116
2011-2012	23	1	Flat (center channel)	20	4*5	19	95	70	66	19	125	-0.01319	0.80719
		2	Pool (next to right bank)	53	5*10+3	Buried	0	-	-	-	-		
2012-2013	68	1	Flat (center channel)	20	4*5	20	100	32	32	19	53	-0.07054	0.914024
		2	Pool (next to right bank)	53	5*10+3	Buried	0	-	-	-	-		
		3	Riffle (next to right bank)	24	6*4	24	100	48	39	13	111	0.203058	0.102673
2013-2014	33	1	Flat (center channel)	20	4*5	20	100	16	16	10	26	-0.04184	0.937087
		2	Pool (next to right bank)	53	5*10+3	30	57	125	114	44	260	0.408976	0.000731
		3	Riffle (next to right bank)	24	6*4	24	100	24	20	6	55	0.203058	0.102673
		4	Pool (center downstrear	16	4*4	16	100	40	34	8	108	0.137664	0.238612
		5	Pool (next to left bank)	12	3*4	11	92	59	64	32	84	-0.0569	0.845384
		6	Riffle (next to left bank)	18	4*4+2	18	100	30	28	16	61	0.314021	0.028285
			Total	386		249	65	61	39	6	260		

* In Moran's I Values range from −1 (indicating perfect dispersion) to +1 (perfect correlation). A zero value indicates a random spatial pattern.
For statistical hypothesis testing, Moran's I values can be transformed to Z-scores in which values greater than 1.96 or smaller than −1.96 indicate spatial autocorrelation that is significant at the 5% level.

Table 3. Bedform characteristics.

		Width, m			Length, m			Channel widths*			Channel slope			
	#	Min	Max	Average	Min	Max	Average	Min	Max	Average	Min	Max	Average	Median
Pools	13	12	20	16	13	279	70	0.9	15.3	4.7	0.0001	0.0075	0.0021	0.0009
Flats	8	12	18	15	19	186	62	1.3	12.4	4.1	0.0003	0.0056	0.0029	0.0027
Riffles	15	12	21	15	5	31	12	0.3	2.1	0.8	0.0031	0.0346	0.0142	0.0106

* Using an average channel width of 15 m for the whole reach

Table 4. Bed characteristics of streams with abundance of fines originating from the parent material.

	Medway Creek	Nahal Me'arot	Nahal Eshtemoa	West Walker River
Bed type (d_{50})	Cobbles (67 mm)	Medium gravel (11 mm)	Coarse gravel (28 mm)	Cobbles (171 mm)
Modality	Bimodal	Trimodal	Polymodal	Bimodal
Sorting	Very poorly sorted (2.8)	Very poorly sorted (2.5)	Very poorly sorted (2.2)	Very poorly sorted (2.8)
Skewness	Very fine skewed (0.51)	Very fine skewed (0.33)	Fine skewed (0.16)	Very fine skewed (0.61)
Kurtosis	Mesokurtic (1.07)	Mesokurtic (1.07)	Mesokurtic (1.02)	Very leptokurtic (1.69)

Semi-alluvial till-bedded stream channel crossing an interlobate moraine in Southern Ontario, Canada. 2: Morphometry, hydrology and hydraulics of Medway Creek

Nathaniel Bergman[1, 2, 3], Marco J. Van De Wiel[1, 4], Stephen R. Hicock[5]

1. Department of Geography, Western University, Canada
2. Yigal Alon Kinneret Limnological Laboratory, Israel Oceanographic and Limnological Research, Israel
3. Department of Geography and Environmental Studies, University of Haifa, Israel
4. Centre for Agroecology, Water and Resilience, Coventry University, United Kingdom
5. Department of Earth Sciences, Western University, Canada

Abstract

In the first paper we focused on semi-alluvial till channel characteristics from a static perspective, including: lithology of the local till, till exposures, bedform morphology and sedimentology, and the alluvium they contain. The only dynamic aspect was the annual erosion rate of the till. We argued for classifying such semi-alluvial channels as a stand-alone category, separate from other river types, by distinguishing their differences induced by past glacial activity. Here we add the dynamic perspective of channel morphometry coupled to hydrologic and hydraulic processes. We then compare our results and analyses to other river types.

The study reach contains rapid width contractions and expansions, which are associated with till control and erodibility as there are no significant water and sediment inputs from tributaries. The incised channel is confined with poor connection to the overlying floodplain. Width/depth (W/D) ratios oscillate along the reach and are higher than another till-bedded river when measured on a reach-scale. However, when compared to other alluvial, semi-alluvial and bedrock rivers on a basin scale, it is evident that Medway Creek's average W/D value is in the middle of the range and till control is negatively related to drainage area – it decreases as drainage size increases. Bankfull discharge, whether determined by hydrological statistical equations or numerical modeling for each cross-section, yields many different values due to hydrological flashiness or varying downstream hydraulic geometry, respectively. Even the highest flood that filled the floodplain and nearby river valley shows rapid expansions and contractions due to incision into the Arva Moraine and constraining by steep bluffs. Bedform morphologies maintain identical patterns at a range of discharges and no modeled channel velocity reversals are detected in pools and riffles as in some alluvial rivers. Channel stability (based on bedform downstream hydraulic geometry; DHG) suggests that most riffles are well adjusted to DHG, while the few that are not are anomalous. The majority of pools have poorly developed DHG and flats are generally in between riffles and pools. These results contrast with sedimentologic stability analysis of the DHG, indicating the complexity of this type of channel in a formerly glaciated terrain. Thus there is a need to improve hydrologic metrics in order to better manage these rivers and implement proper restoration procedures for them.

1. Introduction

Alluvial rivers have been investigated for more than a century for a variety of practical reasons such as engineering and dam works (Alexander et al., 2012), river restoration practices (Palmer et al., 2010) and general scientific understanding (Bridge, 2003). During that time, research on bedrock rivers was quite rare (Baker, 1984; Ashley et al., 1988), although these rivers have received a lot of attention from the earth science-engineering research community in the last two decades (Tinkler and Wohl, 1998; Turowski, 2012 and references within). In contrast to these two river types, semi-alluvial channels in general, and till-bedded channels in particular, are only now receiving research interest (Arbogast et al., 2008; Gran et al., 2009; Merten et al., 2010; Mier and Garcia, 2011; Gran et al., 2011; Khan and Kostachuk, 2011; Phillips and Desloges, 2014, 2015a, 2015b; Thayer et al., 2016). This is despite their abundance in formerly glaciated North America and other regions around the world that are currently deglaciating due to modern anthropogenic climate change. Earlier work identified the importance of former glaciation on modern (post-glacial) geomorphic fluvial processes and form (Holmes, 1937; Ruhe, 1952; Hack, 1965; Hill, 1973; Brakenridge et al., 1988; Campo and Desloges, 1994; Ashmore and Church, 2001; Phillips and Robert, 2005; Brardinoni and Hassan, 2007) and the sedimentary difference between semi-alluvial and alluvial channels (Davis, 1958).

Many conceptual models and empirical equations have been developed to investigate or explain the relationship between flow and form in river channels; most relate to different environments with different climate and lithology, ordinarily with the watershed-scale serving as the analysis unit (Leopold and Maddock, 1953; Annable, 1996; Wohl, 2004; Eaton and Church, 2007) while others seek to explain transitions between different bedforms on a reach-scale (Keller and Melhorn, 1978). Flow-form relations (regime theory) and hydraulic modeling have been extensively applied to alluvial channels (Leopold and Maddock, 1953; Nicholas, 2013, respectively). However, there have been few attempts to look into the channel morphometry of semi-alluvial channels incised into till and their downstream hydraulic geometry (DHG). The limited modeling attempts have focused mainly on a large river (St. Clair River between Lake Huron and Lake St. Clair) subjected to intense human shipping disturbance (Czuba et al., 2011; Liu et al., 2012) and do not represent the abundant, smaller till channels with an agricultural-urban setting typical to the Great Lakes region in general and SW Ontario in particular (Campo and Desloges, 1994; Annable, 1996). Thayer et al. (2016) recently looked at the Little Rouge River, a low relief till channel, and found basin-scale relations for the DHG despite the fact that glaciation legacy disrupts it on a reach-scale. Phillips and Desloges (2014) investigated 22 large southern Ontario rivers (also basin-scale analysis) and found that glaciation effects are expressed in the bed slopes where reaches are over-steepened or under-steepened. The over-steepened reaches with high stream powers are moderated by very coarse sediment caliber originating from glacial sediments or high bank strength that does not allow the channel to widen.

The goals of this paper are to investigate the hydrogeomorphic metrics of a semi-alluvial till channel that crosses a moraine and to compare it with other till rivers, as well as alluvial and bedrock rivers. This analysis includes width/depth ratios, bankfull and maximal discharge using a variety of techniques, peak discharge, modeled channel flow velocities for each bedform and finally a stability analysis of bedforms based on a downstream hydraulic geometry discriminator by bedform types.

2. Study area

The complete description of the regional setting of southern Ontario, Medway Creek's watershed and study area appear in Bergman et al. (2016). Briefly, Medway Creek is a 3rd order tributary of the Thames River (5830 km^2) in southern Ontario and drains 205 km^2 of a predominantly agricultural catchment (83%) while the lower basin part is a deep glacial valley flowing mainly within the city of London and incised through the Arva Moraine, where the study site is located. The channel's alluvium is a cobble-bed; bedforms consist of pools, riffles, flats, and occasional steps and till exposures while the floodplain has thin soils overlying the till. The local climate is humid temperate typical of the Great Lakes with warm summers, mild winters, 800-1000 mm precipitation, and about a third of the precipitation is translated to surface flow. During the European-American settlement, the natural forests were removed and replaced by vast agricultural land and urban development. However, the main vegetative cover in the study area is deciduous forest and meadow.

The mean annual flow of Medway Creek is 2.9 m^3/s (Upper Thames Conservation Authority, 2013), and the largest peak flow recorded was 147 m^3/s in 1977. Flow is only sustained year-round in the lower incised reach below Arva Dam. The study site is a 1.5 km long, full meander of Medway Creek (Fig. 1). Medway Creek has been gauged since the 1940s by Environment Canada and by the Upper Thames Conservation Authority. There are two steep bluffs, one in the center of the meander (right bank) and one next to the upstream part of the reach (left bank), that contribute large quantities of sediment to the channel during precipitation events and after snowmelt.

3. Methods

Detailed data-collection procedures of this study are presented in the appendix of Bergman et al. (2016). Data presented here are based on morphometric mapping of the channel and floodplain/river valley DEM as well as field measurements, assigned roughness based on the grain size distributions (GSDs) and till exposures coupled to the gauged hydrology to produce a variety of metrics such as bankfull discharge and the largest flow on the measured record. Equations from the literature and data from other studies from a variety of fluvial environments (semi-alluvial, alluvial and bedrock channels) were used and compared to our results.

Modeling was done using the HEC-RAS 4.1 software model (U.S. Army Corps of Engineers, 2016), using 94 cross-sections that represent the reach's bedform division (Fig. 1). In places where the bedform was very long (especially pools) additional cross-sections were automatically interpolated by the program. Manning's roughness values were calculated according to the d_{84} of the GSD of bedforms (Cheng, 2016; see the GSDs in Bergman et al., 2016). In places where a till patch occurred in part of a cross-section, a roughness value of 0.013 of smooth concrete was assigned to it (Wong and Lim, 2006). Flows were calculated using iterations for a specific cross-section in a similar repetitive technique often used in paleoflood studies for covering flood deposits (O'Connor and Webb, 1988) or for the entire reach using all the cross-sections with a single value discharge.

4. Results and discussion

4.1 Channel geometry and inferred hydrologic and hydraulic characteristics

4.1.1 Width/depth ratios

The right bank heights of Medway Creek in the study reach range from 0.4 to 20.4 m (downstream Arva Moraine bluff; average 2.8, median 2.0; n = 62 field measurement points). If the bluff is excluded from the analysis, maximum bank height is 6.5 m (average 2.2, median 1.9). The left bank heights range from 0.3 to 18.0 m (upstream Arva Moraine bluff; average 2.5, median 1.8; n = 62). If the left bank bluff is excluded from the analysis, maximum bank height is 5.6 m (average 1.9, median 1.7), similar to the right bank values (Fig. 2a, 2b).

Channel width was defined as a measurement between two opposite points, one on each bank, where the gradient of the bank changes and above it the river would burst into the floodplain (i.e. the channel is at maximum capacity - similar to bankfull discharge determination in alluvial rivers; Williams, 1978). While this is trickier in an incised channel, in cases there was a high vertical bank or steep bluff, only the opposite bank point was used for measuring width as an imaginary horizontal line. Channel widths (W) vary from a minimum of 10.0 to a maximum of 22.9 m (average 15.4, median 15.0; Fig. 2a, b; n = 62). Considering that there are only two small tributaries entering the study reach (from the right bank) with negligible flow and sediment contributions to the main stream, local channel width oscillations do not result from downstream changes in water discharge or tributaries' sediment flux, but can be attributed to till control, the two bluffs' mass wasting processes, and perhaps within-reach scour and deposition. Human disturbance could affect channel width (Faustini et al., 2009) but is not directly relevant for the 1.5 km Medway Creek study reach as the millpond dam is 12 km upstream. There are three tributaries entering the right side of Medway Creek (Snake Creek + two smaller creeks downstream), but they are all upstream of the study reach. There are three more tributaries entering the left side (two above the study reach and one downstream of it). All these small tributaries drain heavily urbanized areas, and which, after heavy rainstorms, contribute a substantial discharge on a timescale of hours. Two gully tributaries within the meander study site, one upstream and one downstream of the right bluff, do not contribute any unusual water inputs or sedimentary input that may influence channel geometry, and there are no direct excavations or reinforcements of the channel banks or valley walls. However, channel width might be associated with larger basin-scale changes induced by human activity.

Although the study reach's sinuosity of 3.14 is a tortuously meandering value for incised E type streams, its width/depth (W/D) ratios range from 4.4 to 45.0 (average 16.2, median 13.9; Fig. 3a), which covers the entire range from low (< 12) to high (> 12) W/D ratios of the Rosgen classification (Rosgen, 1994). The question that arises is: are the W/D ratios of Medway Creek highly unusual for formerly glaciated terrain (FGT) rivers? The main problem is to find suitable comparison data sets that look at channel geometry on a similar short reach-scale rather than a whole basin-scale, as most DHG work is done on the latter (Leopold and Maddock, 1953; Schumm, 1963). We compare Medway Creek to Kaskaskia River (Illinois, Mississippi drainage) that also has a data set with two short reach-scale sections (Bhowmik, 1979; Fig. 3a). We also compare the basin-scale of the W/D ratios of 66 alluvial rivers and 126 FGT rivers from different physiographic regions (Fig. 3b; Appendix Table 1). The regularity in the behavior of the width and depth both at-a-station (reach-scale) and in the downstream direction (basin-scale) implies quasi-equilibrium conditions in the channel (Wolman, 1955).

Medway Creek does have relatively high W/D ratios compared to the two reaches of the Kaskaskia River (Fig. 3a) although the latter is a much larger river (draining 3445 km^2 and 7045 km^2). In general, small FGT rivers have on average relatively low basin-scale W/D ratios (e.g. Bhowmik, 1979; McCandless and Everett, 2002, McCandless, 2003; Mulvihill et al., 2009; Medway Creek - this study) although there's considerable scatter (Fig. 3b). Large FGT rivers, such as the Scioto River Ohio and the St. Lawrence River (Leopold and Maddock, 1953), and probably the Kaskaskia River from the reach-scale analysis, are capable of scouring their valleys and consequently have higher W/D ratios. The most extreme W/D ratios appear on desert alluvial rivers (Pelletier and DeLong, 2004), while sandy prairie rivers present the lowest W/D ratio values (Schumm, 1963). Schumm (1960) attributed the W/D ratio of sand-bed streams outcome to the fine content (silt and clay) content of the cohesive banks' material - the higher the fines content the harder it is for the channel to adjust to the forming flows (i.e. negative relationship) so the channels deeply incise instead of widen.

Variable W/D ratios are usually associated with unstable river systems (see Chapter 2). Phillips (1990) suggested that variability in channel geometry is entirely consistent with the concept of an unstable system, as an unstable system would not be expected to maintain a constant relationship between hydraulic and geometric variables. Repetto et al. (2002), using 2D and 3D models and flume experiments, found that width oscillations lead to planimetric instability, although they related this mainly to bifurcations of braided rivers with an assumption that the channel is laterally stable. Pelletier and DeLong (2004) described and modelled an oscillating pattern of narrow, deeply incised reaches and wide, shallow reaches with a characteristic wavelength in arid desert rivers and attributed them to width, depth and slope instabilities. Dodov and Foufoula-Georgiou (2004) postulated that variation in the channel cross-sectional shape under preservation of the momentum, water and sediment is a result of systematic increase of channel asymmetry downstream induced by scale-dependent fluvial instability. Finnegan et al. (2005) asserted that, in bedrock channels, the W/D ratios and roughness are a function of the boundary material composition and that W/D ratios (and roughness) can be assumed constant, a claim not valid for till of our study. However, they also demonstrated that using conventional scaling relationships for channel width can result in underestimation of stream-power variability in channels incising bedrock and their model improves estimates of spatial patterns of bedrock incision rates.

In a channel like Medway Creek, these oscillations would most likely relate to the local strength of the till boundary material as the elevated forest floodplain vegetation is somewhat disconnected from the channel. However, vegetation effects on W/D ratios cannot be completely ruled out (Trimble, 1997). Shugar et al. (2007) and Khan and Kostaschuk (2011) investigated two local tills (Halton Till and Sunnybrook Till) of two southern Ontario rivers using a variety of geotechnical methods and found that, similar to glacially-derived loess deposits in the US Midwest (Hanson and Simon, 2001), the material composition (silt-clay and organic ratios with negligible sand and gravel content) determines the cohesive material erodibility (Grabowski et al., 2011). Khan and Kostaschuk (2011) found that the upper layer of tills (Sunnybrook till and the Halton till) is weaker than its underlying parts because it is exposed to water action and bioturbation.

Their results corroborate the assertion by Chatanantavet and Parker (2009) for bedrock channels that the exposed top layer of the boundary is much weaker (has less compressive strength) than the underlying submerged bedrock creating a vertical strength variation in addition to the material spatial variability. This creates a certain period of boundary exposure, during which the water and sediments gradually weather the till until it becomes more prone to erosion and the flow can incise into it and expose the harder substrate. The boundary till has spatial and temporal cohesive strength properties and internal architecture (Menzies et al., 2006) that determine the fluvial erodibility (Grabowski et al., 2011). These lithologic controls are expressed in Medway Creek's cross-sectional morphology and consequent W/D ratios and bedform long profile (i.e. enlargement, incision or both), similar to eroding soft bedrock channels (Allen and Narramore, 1985; Stock et al., 2005; Annandale, 2006).

4.1.2 Bankfull discharge and floodplain characteristics

Bankfull discharge (Q_{Bf}) that is reflected in a given cross-section morphology (at-a-station) is dependent on local W/D ratio and basin-scale longitudinal downstream hydraulic geometry (DHG) as the discharge varies (usually increases as tributaries join the main stem). Therefore, it is often assumed that the bankfull discharge concept that is applicable to alluvial channels (Leopold and Maddock, 1953; Leopold et al., 1964; Williams, 1978) does not apply to deeply incised channels that have lost their connection to the floodplain due to differential lowering of the bed relative to the banks (Schumm et al., 1984; Darby and Simon, 1999). If the bankfull discharge is not reflecting the river's hydrology, climate and sediment erodibility, one would expect that W/D ratios and the DHG would also not follow power-law relationships between discharge, width, depth and slope. However, not all researchers agree that channel downstream hydraulic geometry is irrelevant to entrenched streams and width-depth ratios are ordinarily reported for incised bedrock channels (Baker, 1984; Wohl and Merritts, 2001; Wohl, 2004). Furthermore, Montgomery and Gran (2001) suggested that the classic hydraulic geometry parameters used for alluvial rivers are the same for bedrock channels despite the rock boundary control, as was shown by Wohl (2004) for a variety of river material types unless the boundary control exceeds stream power. Thayer et al. (2016) recently showed that for another till-bedded river (Little Rouge River) in southern Ontario, despite the glaciation legacy, there are very good hydraulic geometry relations on a basin-scale.

Hassan et al. (2014) proposed that in formerly glaciated terrains, in order to determine the effective discharge, sediment mobility allows discrimination between three types of streams: 1) streams in which gravel (defined as sediment > 8 mm diameter; Hassan et al., 2014) moves frequently and effective discharge occurs during gravel transport (Frequently Mobile Gravel); 2) streams in which gravel moves infrequently but effective discharge nonetheless occurs during gravel transport (Infrequently Mobile Gravel); and 3) streams in which sand (defined as sediment < 8 mm diameter; Hassan et al., 2014) moves over largely immobile gravel and effective discharge occurs frequently during sand-phase transport (Sand over Immobile Gravel). Only the Infrequently Mobile Gravel streams have large, rare effective discharges that approximate the bankfull discharge; in Frequently Mobile Gravel and Sand over Immobile Gravel streams the effective discharge is much more frequent and smaller than bankfull. Only in the Infrequently Mobile Gravel streams does the effective discharge approximate a channel-forming discharge. In Frequently Mobile Gravel and Sand over Immobile Gravel streams, the effective discharge bears little relation to the size or dimensions of the channel and is at best a channel-maintaining flow (Hassan et al., 2014).

The sedimentary characteristics of the bedforms allow us to find the corresponding geometric characteristics along the study reach where the increase in drainage area is minor, lateral sediment and water inputs from tributaries are negligible and therefore the channel should maintain its form, unless the till control (e.g. erodibility) plays a major role. According to the commonly used Rosgen (1994) channel classification, Medway Creek is an entrenched, single thread channel type E4 (channel slope < 2%, with d_{50} of cobbles). This type of channel is not considered a high ratio bedload stream (sediment supply in relation to stream discharge) suggesting that most sediment transport is washload and suspended load, not bedload. This type of classification received heavy scrutiny as it is entirely based on morphology and sedimentology but does not integrate and quantify the channel process and their responses to these changes (Simon et al., 2007). Annable (1996) analyzed 47 rural rivers in southern Ontario and classified six (13%) of them as type E4 so Medway Creek general morphology is not unusual for this area, although it is not the most common channel type. He suggested the following relationship:

$$Q_{Bf} = 0.52A_d^{0.75} \qquad (1)$$

where A_d is the basin's area (in km^2), which translates to 28 m^3/s. Using the channel width and assuming (cohesive) bank material it is also possible to extract bankfull discharge from a width-bankfull discharge relationship (Anderson et al., 2004) from rivers with high silt-clay bank content (> 10%):

$$Q_{Bf} = 3.87W^{0.5} \qquad (2)$$

giving Medway Creek a bankfull discharge of 32 m^3/s (based on a maximum width of 23 m in our study reach). If we use Thayer et al. (2016) bankfull widths discharge relations from their case study of the Little Rouge River, the bankfull discharge on Medway Creek yields 7 m^3/s (for 10 m width) to 40 m^3/s (for 23 m width). By measuring bedload transport using a large Helley-Smith bedload sampler (76 mm x 76 mm opening) on Medway Creek's spring flood, we found that at a discharge of ~32 m^3/s there is incipient motion of the bed, mainly of the sand fraction but some clasts up to 60 mm were also in motion (authors' unpublished data).

An alternative to using channel dimensions for finding bankfull discharge in such a problematic non-alluvial river setting is using the 2-year flood (Q_{2yr}) recurrence interval (RI; Wolman and Leopold, 1957) or 1.5-year flood event ($Q_{1.5yr}$; Leopold et al., 1964) and correlating these discharges to actual flow elevations. The 2-year flood on Medway Creek is 76 m^3/s and the 1.5-year flood is 60 m^3/s based on the maximum instantaneous discharge recorded by the Environment Canada gauging station downstream of the study site. The use of maximum instantaneous discharge is advantageous to the use of average daily maximum discharge values, which is likely to result in underestimation of the flows associated with various return periods. This is because the daily data used to perform the flood frequency analysis are averaged over a 24-hour period and as such have smoothed out the extreme values unlike the maximum daily discharge. A continental-scale relationship (for the US) between bankfull width and drainage area for single-thread channels by Wilkerson et al. (2014) gives Medway Creek a bankfull width of 10 m which translates to a discharge of ~10 m^3/s but it is worth mentioning that semi-alluvial channels affected by the Laurentide Ice Sheet are treated as regular alluvial channels in their analysis. This analysis underestimates bankfull discharge as maximum width is 23 m, more than double this estimate but it does fit well with the smaller widths cross-sections. The continental scaling is averaging a variety of geologies, topographies, river histories and human disturbance, hydrology and climate and thus is less accurate, especially for a river that has gauged data.

The entrenchment ratio and degree of incision cannot be easily calculated for it without *a priori* knowledge of the bankfull discharge as in many localities the flow is contained within the channel cross-section and cannot overspill into the high disconnected floodplain. An example of the problematic metric of bankfull discharge in this river setting is taken from the spring flood of 2011 with a peak discharge of 37.5 m^3/s (Fig. 4a). While one section of the channel experiences bankfull discharge covering the inner point bar (Nunally, 1967), the nearby upstream section of the channel contains the entire flow and is well below the top elevation of both banks. This observation of the ability of the channel to contain most floods with very little spillage into the floodplain caused Knox (1989) to describe such till channels as "flume-like" with efficient flood routing through considerable velocity and erosive force. The floodplain would therefore be considered a low energy environment with very little accretion of fine suspended sediment during high magnitude floods that exceed the high banks. Coarser sediments (> sand) cannot reach the floodplain unless they originate from the underlying till and lacustrine sediments of the floodplain or the topsoil or deposit during rare floods of high magnitude and low frequency.

The genetic channel floodplain classification of Nanson and Croke (1992) would term the floodplain of Medway Creek as Class C Order C1 - laterally stable single-channel floodplain. The dense deciduous Carolinian forest on the banks ensures that even if a high flow reaches the floodplain the topsoil will remain intact and not erode, making this a poorly-developed floodplain as there is minimal river-floodplain sediment exchange. This finding corroborates earlier observations by Ritter (1975) and Patton (1988) in lowland rivers of formerly glaciated terrains of New England and Illinois, respectively. During large floods the floodplain experiences very few localities of erosion or deposition but natural weathering processes (such as valley hillslope processes during snowmelt) erase these evidences. If deposition of coarse material remains in place it is expressed as gravel lenses underneath the topsoil so is not visible on the surface (Ritter, 1975; Patton, 1988). The Nanson and Croke (1992) classification for Medway Creek contradicts the recent findings from the nearby Rouge River that found that soil profiles in the floodplain often suffer from overbank sedimentation producing pedostratigraphic successions. On the Rouge River, active erosion across these surfaces insures that soils lack sufficient weathering time to produce B horizons even on distal reaches of the floodplain (Mahaney et al., 2016). Another recent classification with four types of floodplain developed specifically for low-relief formerly glaciated channels (Phillips and Desloges, 2015a), does not fit the study reach of Medway Creek: it would be a C type according to till grain sizes (predominantly silt and clay, some sand with little gravel and cobbles; see Bergman et al., 2016), according to the high W/D ratios of the channel (section 4.1) it would be a B type, and strictly by channel morphology visualization it would be a combination of S and M types. Thayer (2010; Thayer and Ashmore, 2016) investigated the floodplain of Medway Creek, including our study site, and proposed two competing mechanisms for its formation: 1) Low relief, vertically accreted floodplain through slow meander extension and avulsion, and 2) Floodplain is formed through meander translation, extension, and chute cut-offs. The difference between the floodplain formation mechanisms are not only the processes but the energy required to create them with 2) more energetic. We suggest the difficulty of assigning a classification type or its anomalous categorization results from our reach incising the interlobate Arva Moraine.

Since bankfull discharge changes from one cross-section to the next and is highly variable between different methods, for our study and other similar incised till-bedded channels, the solution might be to use a fuzzy number rather than a single deterministic value (Johnson and Heil, 1996). While this approach is useful for engineers basing designs on a range of possible values, it is less useful when using metrics that require a single discharge value for scientific purposes (Wilcock, 1997). Sholtes and Bledsoe (2016) recently compared bankfull discharge from a morphologic perspective (field-based evidence) to process-based predictors (sediment transport data) from gauged sites across the United States including coarse, bed load-dominated channels and fine, suspended load-dominated channels with varying drainage areas. They found the best bankfull discharge estimate was associated with 50% of cumulative sediment yield based on the flow record (Q_h). This Q_h has the lowest relative error in predicting bankfull discharge for coarse and fine bed sites when compared with the statistical approach (Q_{eff}, $Q_{1.5yr}$ and Q_{2yr}). Specifically for coarse-bed sites like our study, Q_{eff} and Q_h both perform well in predicting bankfull discharge, followed by $Q_{1.5yr}$ (Sholtes and Bledsoe, 2016). This process-based approach might be the solution for predicting bankfull discharge in a problematic setting like ours, but its main disadvantage is that it requires prior knowledge about sediment yield resulting from the collection of data for several hydrologic years.

Alternatively, it is possible to use a computational fluid dynamic (CFD) model (Van De Wiel et al., 2007) or a simpler program with sediment transport capabilities (Pitlick et al., 2009) or a sediment transport equation (Barry et al., 2004) to predict the sediment transport rates. Nevertheless, results would be more realistic if some of them could be calibrated against measured field data to reduce the modeling uncertainty and explain variability based on qualitative observations and assumptions.

4.1.3 Hydraulic modeling of the bankfull discharge

The bankfull discharge (Q_{bf}) is a major geometric and design metric (Williams, 1978) that needs to be predicted in the most accurate way possible, especially when managing or restoring a river (Shields et al., 2003; Doyle et al., 2007). Here we use the HEC-RAS model (U.S. Army Corps of Engineers., 2016) to determine for each cross-section the exact discharge that just fills the channel before spilling onto the floodplain. The model was calibrated against high flow marks of the spring flood of 2013 (peak discharge of 52 m^3/s) that were flagged on the floodplain. For the channel, we used bedform grain roughness (d_{84}) and lower values for smooth till patches (Bergman et al., 2016). The roughness value used for till was 0.013 - similar to smooth concrete (Wong and Lim, 2006; Chow, 1959). For the floodplain we used a generic Manning's *n* roughness of 0.08 to account for trees and the thick vegetation of the river valley. The results presented in Fig. 4b show that there is not a single bankfull discharge but rather 44 different values (for 94 cross-sections, not including extrapolated ones) starting as low as 7 m^3/s up to 99 m^3/s (median 33, average 36 m^3/s). The significance of this finding is that one would expect to have a narrow discharge range (few values) reflecting small changes in proximal cross-sections (i.e., width and depth) of the reach but instead the rapid fluctuations in bankfull discharge imply downstream hydraulic geometry is strongly controlled by till tensile strength and erodibility. Due to this large variability in bankfull discharge values, all the scaling correlations between depths and widths to bankfull discharges are weak, whether looking at the whole reach or by bedform type (Table 1).

Unlike a basin-scale analysis that can mask glacial conditioning on the DHG by increasing discharge downstream (Thayer et al., 2016), it is apparent that reach-scale analysis with no sediment or water inputs exposes the control exerted by the moraine and cohesive till. Eaton and Church (2007) claimed that the widespread similarity of downstream scaling form of equations suggests that they express some important underlying regularities in the morphology of stream channels through the drainage network. But in environments in which bank strength varies greatly, scaling relationships fail to capture the rapid changes in widths and depths. Hydrologic response often exhibits considerable scatter that is difficult to interpret when working with a small number of observations, although this scatter disappears when increasing the number of observations (Zehe et al., 2007). The scatter thus is not measurement error but is related to the scale of analysis.

The modeled average and median bankfull discharges are almost identical to extraction of bankfull discharge from maximum channel width (23 m) proposed by Anderson et al. (2004) that gives 32 m^3/s, and slightly higher than the bankfull discharge equation proposed by Annable (1996) that gives 28 m^3/s for southern Ontario rural streams. The localities that produce a very high bankfull discharge are areas where a former meander loop enters the channel creating a compound cross-section with a large area, while localities with very low bankfull discharge values are in places where the

Arva Moraine is not constraining the channel; incision into the till banks is small and consequently both banks are relatively low, allowing the flow to easily expand into the floodplain.

Large and variable values of bankfull discharge in our study reach, unlike graded and stable alluvial rivers (Mackin, 1948; Lane, 1955), show the channel is not in quasi-equilibrium (Stevens et al., 1975) but rather reflects a history of many flows with the combination of local till properties and its erodibility. Vertical incision and bank collapse widening are not compensated by deposition and opposite bank advancement and the channel records numerous local geometry adjustments that even the perennial shrub and forest of the floodplain cannot stabilize. The DHG measurement uncertainty (Harman et al., 2008) in our cross-sections was reduced by the mapping methodology. The mapping was not only based on the GPS-produced DEM but supported by bank heights and channel widths measured in the field and an assigned bed roughness for each cross-section that actually reflects local bed sediments and till exposures (Fig. 2a). The accuracy of measuring the exact points of bankfull discharge inflection (i.e. where the slope breaks from the channel into the floodplain) is especially important as flat areas are less accurate in the DEM and that could expand the floodplain at the expense of the channel. It is estimated that the DEM threshold vertical inaccuracy is 0.25 m while for the width it is 2.0 m (Borgniet et al., 2003).

The bankfull modeling results support Thayer's (2010) and Thayer and Ashmore (2016) first formation mechanism: vertical accretion of suspended sediment as a result of annual overbank flooding. Meanders are extended during overbank flows, they migrate and neck-cutoffs leave behind abandoned channels. These newly inactive channels fill and are buried with time and therefore the alluvium above them is thicker than the nearby floodplain that is a result of incision and is overlain by a thin soil layer. This mechanism negates the second mechanism of lateral accretion of bars and vertical accretion on the channel margins as there are only 5 bars (Bergman et al., 2016); only one is a medial bar while the rest are forced (pointbars), but many bankfull discharge values surpass them in elevation besides the small discharge that remain in-channel. Since overbank flow can occur at a range of discharges (Fig. 4b), local floodplain sedimentation rates can account for up to 1.5 m of sediment in buried channels but only a few cm of sandy fines in the flatter floodplain that is topographically higher (Thayer, 2010; Thayer and Ashmore, 2016). Thornbush and Desloges (2011) and Oliva et al. (2016) showed that oxbow lakes hold key hydrologic and stratigraphic evidence for overbank flows (using archeological artifacts and paleofloods, respectively) in a temperate humid climate. However, the relatively confined morphology of Medway valley incised into the Arva Moraine does not allow these oxbow lakes to develop as the abandoned channels are filled and their only signature is seen either in the floodplain stratigraphy (Thayer, 2010; Thayer and Ashmore, 2016) or when bankfull discharge is modeled and the cross-sectional area is unusually large (the site of avulsion).

In larger rivers like the Lower Grand and Lower Thames, the floodplain's vertical and lateral accretion are better recorded (Walker et al., 1997; Stewart and Desloges, 2014) and it is possible to build a long term chronology of overbank flow events and associate them with sedimentation. Bankfull discharge models confirm the W/D ratios' variability (section 4.1.1, Fig. 3a, b) and that boundary till and the Arva Moraine (and the associated width constriction) control local channel morphometry on a reach-scale. These results will probably be opposite in a much larger till river (Leopold and Maddock, 1953) or when conducting basin-scale analysis on a small till channel (Thayer et al., 2016), implying bankfull flow (or other effective discharge) is scale-dependent.

4.1.4 Maximum discharge and 100 year flood for Medway Creek

Southern Ontario does not have an established envelope curve for peak discharges (Q_p) as unlike regional Q_{bf} there are many missing values. We therefore look for similar channels in the American Midwest and bordering US states that were also previously glaciated during the late Wisconsinan. Patton and Baker (1976) investigated morphometric relationships for small drainage basins across the US, including ten in Indiana (4-329 km^2 in size) that were all formerly glaciated by the LIS (Mickelson and Colgan, 2003) and have a similar flat topography like southern Ontario. They suggested a maximum discharge of 222 m^3/s for a channel with a drainage area like Medway Creek, i.e. about 7 times average bankfull discharge.

The modern flow record for Medway Creek is too short (71 years, but the early years' record is seasonal and incomplete) and the construction of probable maximum flood (PMF) and probable maximum precipitation (PMP) will need large extrapolations that will yield very large errors. Alternatively, the 100 year flood (Q_{100}) can be calculated using a proposed USGS method related to climatic and watershed properties:

$$Q_{100} = 0.471\, A_d^{0.715} * E^{0.827} * S_h^{0.472} \qquad (3)$$

where E is elevation (in m) and S_h is the basin shape factor, defined as the drainage area divided by the square of the main channel length. This gives a Q_{100} of 195 m^3/s for Medway Creek. A flood of this magnitude has not occurred on record (largest measured flood was 147 m^3/s in 1977). When nonparametric frequency analysis was performed on

183 gauging stations from Ontario and Quebec, unimodal and multimodal maximum annual flood density functions were discovered (Gingras et al., 1994). The stations with a unimodal density were subject to the spring snowmelt, while the multimodal densities were subject to snowmelt or to rainfall-only events. Gingras et al. (1994) suggested the following equation for southern Ontario:

$$\log Q_{100} = 0.873 \log A_d - 0.032 \qquad (4)$$

This yields $Q_{100} = 99$ m^3/s, which was exceeded twice since records started (1977 and 2007).

The wide range of discharge metrics (Q_{Bf}, Q_{2yr}, $Q_{1.5yr}$, Q_p and Q_{100}) demonstrates the inherent problem of applying alluvial river channel theory (DHG of Leopold and Maddock, 1953) to an incised channel that has a cohesive boundary. While DHG theory works quite well on a basin-scale (Thayer et al., 2016), on a much smaller reach-scale it is inadequate. Its geometry and dimensional analysis (Strahler, 1958) do not necessarily reflect the contemporary hydrology and climate (Rodriguez-Iturbe and Escobar, 1982), but a longer deglaciation legacy (Wilcock et al., 2009; Gran et al., 2009; Phillips and Desloges, 2014; 2015a; 2015b) intertwined with modern short-term anthropogenic processes (Campo and Desloges, 1994; Novotny and Stefan, 2007; Schottler et al., 2013). There are similarities between till channels (Kamphuis 1983; Kamphuis et al., 1990; Pike, 2014) and soft bedrock channels (Stock et al., 2005) in terms of tensile strength of the boundary material while active tectonic uplifting in bedrock rivers is replaced by isostatic rebound in young till (Lewis et al., 2005) but channel dimensions on a reach-scale are not enough to infer and solely explain DHG. In addition, the reach-scale and basin-scale comparisons are sometimes incompatible as the amount of intervening factors and internal complexity are quite different. Hence, a variety of methods and metrics might be needed to understand the complex channel morphometry of till channels, especially when working on a reach-scale.

This discharge variability is important during continuous active river management and especially when implementing restoration practices that are often based on hydrologic metrics that assume the various flow variables (often width, depth, their resulting cross-sectional area and bankfull discharge) are reflecting channel adjustment to discharge and sediment supply (Mackin, 1948; Lane, 1955), but do not account for excessive control by the boundary material (Wohl, 2004). Foster (1998) worked on 3 semi-alluvial streams crossing or in between moraines in the London area (Dingman Creek, Oxbow Creek and Nissouri Creek) and concluded that Natural Channel Design (NCD) and habitat restoration need to incorporate differences in the bedform features in order to succeed. For example, when prescribing flow releases from an upstream dam and the volume of water to be released downstream, determining the hydrograph shape (i.e. peak discharge, duration of the rising limb and the recession) is crucial in achieving geomorphic work rather than releasing a monotonic flow which has limited flushing effects (Kondolf and Wilcock, 1996). Such complex design eco-hydraulic models with different scenarios are already available in order to maintain pool-riffle morphology of gravel-bed rivers (Schwartz et al., 2015; Brown et al., 2016), but are lacking for till-bedded rivers.

4.1.5 Modeling largest flood on the measured record

Once the classic bankfull discharge concept fails for the study reach, a new question arises: is there another single discharge that shaped the floodplain and river valley? In many channels, there is an effective discharge (Q_{eff}; Andrews, 1980) or dominant discharge (Carling, 1988) which is not necessarily the bankfull discharge; this is especially true for formerly glaciated terrains (Hassan et al., 2014). The idea to use the largest flood on record stems from desert rivers' geometry (Graf, 1988), where often there are two separate morphologies: 1) The outer morphology formed by a very large high-magnitude low-frequency rare flood, and 2) The inner channel where most of the ordinary flows occur and that morphology is distant from the external one. The internal morphology occupies a small area of the external valley morphology due to its cohesive bed and banks and thus the external morphology can be preserved for a long time (i.e. control of rare floods of overall valley geometry). We use the largest flood on the measured record (147 m^3/s) to model if there is an external morphology that maintains any geometric relations. We use two different parameters to test this: widths (ASCE Task Committee on Hydraulics, Bank Mechanics, and Modeling of River Width Adjustment, 1998) and cross-sectional area (Bagnold, 1960). The cross-sectional area's main advantage is that it ignores each cross-section's simple (i.e. rectangular or trapezoidal) or compound shape but inherently includes widths and depths. If any of these do not show significant change along the Medway reach since the largest flood took place almost four decades ago (1977), it is possible that it is the only recorded geometry (Baker, 1977; Wolman and Gerson, 1978) while the classic bankfull discharge(s) only reflects relatively small-scale erosional processes (Wolman and Miller, 1960), not only of the flows but till erodibility (Kamphius, 1983; Khan and Kostachuk, 2011). Therefore it is not bankfull discharge but rather a rare flood event that we are considering here.

Results of maximum channel widths for the flood of record are presented in Fig. 5a. Widths were normalized relative to channel center in order to better visualize their changes without the effect of meander morphology (Fig. 1). As with non-constant channel bankfull discharges (Fig. 4b), outer widths of floodplain and river valley also show variability. Bank width irregularities are exaggerated within the river floodplain and valley floor, with a minimum channel width of 59 m, and a maximum flow width of 175 m. The widest channel areas are opposite the two bluffs while the narrowest are downstream of them, where flow converges back towards the channel. This results from channel width expansion in places where the channel is cutting through the Arva Moraine. The moraine's steep constraining bluff forces the floodplain flow on that side to converge back into the river, expand to the opposite side of the channel deep into the floodplain and river valley and possibly have major flow losses and backponding on the forest floor.

The river valley and floodplain are affected by the long-term glacial legacy (Gran et al., 2011; Phillips and Desloges, 2014; Thayer et al., 2016) and shorter term non-fluvial processes such as bluff mass wasting (Day et al., 2013) and sediment erosion due to freeze-thaw processes (Isard and Schaetzl, 1998). The cross-sectional areas are therefore expected to reflect these spatial changes for this low gradient channel even though forest and shrub vegetation (e.g. roughness and holding the soil) are homogenous. Fig. 5b presents the cross-sectional area of the largest flood on record. The largest flow area is before the right bank bluff (283 m^2), while the smallest area is exactly between the left bank bluff and the right bank bluff (62 m^2; average 140, median 137). The bluffs, representing river incision through the Arva Moraine, thus play a major role in determining flow structure during high flows as they serve as obstructions to flow and are the highest relief of the floodplain such that no natural flood can reach or exceed their tops. Furthermore, in places the cross-sectional area is relatively small and the flow converges into a confined section and cannot expand; energy, slope and velocity have to be compensated and rise dramatically (Froehlich, 1994; Phillips and Desloges, 2014). We did not observe terraces or outer floodplain/valley incision that supports a larger external morphology formed by a large flood although it is possible it was not preserved and removed with time, unlike other cases of till channels that preserve their erosional history well (Phillips and Robert, 2005; Arbogast et al., 2008). Thayer et al. (2016) showed that moraines play a major role in determining channel slope and stream power in southern Ontario, thus strengthening our modeling results from Medway Creek. Furthermore, moraine reaches are different not only in coarser sediments and geomorphic instability that characterizes them, compared to low-gradient channel reaches, but need a different approach when conducting channel rehabilitation (Champoux et al., 2003; Merten et al., 2010).

It seems that, for Medway Creek, both widths and cross-sectional areas of the largest flood are not reflected in external valley morphology, and that the Arva Moraine bluffs are the major control for constraining larger floods. Furthermore, the bluffs' steep high-relief topography means that there are raised floodplain/river valley patches that cannot be inundated by high discharge under the current climate and they are unrelated to modern fluvial processes. Internal morphology of the channel is controlled not only by Arva Moraine bluffs but also by cohesiveness/erodibility of the local till. There is, therefore, no single formative discharge that can be assigned to Medway Creek as with alluvial gravel-bed rivers that all conform to a regular range of formative discharges (Parker, 1978; Parker, 1979).

4.1.6 Modeling channel velocities

As seen in section 4.1.2, bankfull discharge is a key metric to describe channel form and function, although we found it challenging to determine a single value in our study reach. Here we simulate and compare the reach's average bankfull flow velocities (Q_{bf} = 36 m^3/s) to minimum bankfull velocities (Q_{bf} = 7 m^3/s) for each of the bedforms (Fig. 6a). Results indicate that velocities increase in an identical pattern for pools and riffles (slope *a* equals 1.18 and 1.15, respectively) and faster for flats (*a* = 1.61). Thus it is possible that flats might be the only bedform to experience velocity reversal according to Keller's (1971) concept at higher discharges. Using the classic Keller (1971) graph from minimum bank-full discharge (7 m^3/s) to the largest flood on record (147 m^3/s), using 5 m^3/s discharge modeling increments, shows this is not the case (Fig. 6b). Riffles maintain highest velocities at all discharges (a =0.010), pools have almost an identical trend increase (a = 0.011) while the flats are exactly identical to riffles (a = 0.010) and remain the lowest. However, at 67 m^3/s riffle and pool velocities almost intersect and are identical until 97 m^3/s, when riffle velocities rise quickly again. The rise in velocities in all three bedforms starts rapidly then flattens.

Despite Arva Moraine's topographic constrictions, pools and flats do not develop the expected velocity reversals (Keller, 1971; Thompson et al., 1999), as in most cross-sections the channel has one lower bank to spill into the floodplain and river valley. In a few cross-sections constrictions are on both banks and the flow is completely contained (Fig. 2a) and modeling allows us to see the hydraulic response. Examining the velocity profiles for three selected discharges (7, 36 and 147 m^3/s) shows that the two bluffs are associated with sharp rises in velocity while a third, just downstream outside the study reach, is responsible for another spike in flow acceleration (Fig. 7). Velocity patterns are identical at all discharges, with sharp fluctuations. Even though the largest flow on record (147 m^3/s) is 21 times the minimum bankfull discharge (7 m^3/s), and more than 4 times the average bankfull discharge (36 m^3/s), average velocities maintain a consistent pattern (R^2 of 0.6 and 0.8, respectively). This can be explained by rigid morphology of the channel exerted by till control and in general the inability of the flow to modify that cohesive moraine morphology aside from a few localities where weathered till is weaker and more erodible. The representative grain size roughness of d_{84} (Cheng, 2016) associated with each bedform (Bergman et al., 2016) is greatest just upstream of each bluff, while downstream there is a sharp drop in bed roughness values after the flow widened.

To summarize, simulated reach channel velocities are strongly associated with bluff constrictions and cross-sectional variability in widths and depths. Velocity oscillations maintain identical flow patterns at a range of flow magnitudes including overbank flows. When analyzing the bedforms, pools and riffles have an identical increase in velocity as the flow magnitude rises while flats have much faster rate of flow accelerations. However, this does not imply velocity reversal for any of the three bedforms whether the flow is contained in-channel or during overbank flows. At a flood range of 67 to 97 m^3/s pools' and riffles' velocities almost converge suggesting the bottom friction of the channel is less important as depth increases. This is not the case for flat velocities that remain lower than pools and riffles at all discharges. Bed roughness and velocities are high just before the bluffs constrain the flow on one bank and force it to expand towards the other lower bank and floodplain but that changes immediately downstream of the bluffs as the flow is no longer limited (except by local till banks). Therefore, the moraine exerts a strong and permanent lithologic reach-scale control of the velocity.

4.2 Channel stability based on bedform downstream hydraulic geometry

Wohl (2004) suggested a simple equation to discriminate whether a channel has poor DHG or developed DHG regardless of type of channel (including formerly glaciated rivers and bedrock channels). Wohl's (2004) method can also be tested on the bedform scale, which makes this method highly suitable for our study. The equation is based on a ratio of stream power to coarse percentile sediment size, Ω/d_{84}, which proved to be an effective discriminator between rivers with poorly developed DHG and well developed DHG (Wohl, 2004). The stream power, Ω (in kg m/s^3), is defined as:

$$\Omega = \gamma QS \quad (5)$$

where γ is specific weight of water (9800 N/m^3), Q is annual high flow (in m^3/s) (we used 32 m^3/s, see reasoning in section 4.1.2) and S is reach slope (unitless or m/m). When Ω/d_{84} is larger than 10,000 kg/s^3, DHG is well-developed; when below, it is poorly developed. The strength of the Ω/d_{84} ratio is that it takes into account the discharge and the grain size just like the RBS* (Relative Bed Stability, see adjoining paper Bergman et al., 2016; Kaufmann et al., 2008; 2009). The annual high discharge reflects the contemporary climate. The grain size reflects the control of parent material in terms of climate and erosion regime (e.g. what grain sizes it breaks down to) and hillslope coupling and bank stability. If the channel has a very coarse tail GSD beyond the competence of the channel, it will be difficult for the channel to adjust the DHG. In order to interpret the results, we defined the following criteria: DHG is considered well developed when Ω/d_{84} ratio > 10,000 kg/s^3, poorly developed when Ω/d_{84} ratio < 9000 kg/s^3 and marginal when 9000 kg/s^3 $\leq \Omega/d_{84} \leq$ 10000 kg/s^3. Most riffles are well adjusted to the DHG, while the few that are not are somewhat marginal (Fig. 8; Table 2). The majority of pools are with poorly developed DHG and flats are generally in between (Fig. 8; Table 2).

A possible explanation to bedforms adjustment is time-scale. Since pools are a scour bedform, we associate their poor DHG adjustment with strong till lithologic control. Conversely, for the depositional riffles, till control has less impact as sediment overlays the till and flow is not incising into it (i.e. alluvium is thickest; Bergman et al., 2016) thus they can better adjust to bed-moving discharges. This notion of longer (millennial) time-scale bears similarity to the observation by Carling et al. (2009) that "…while in alluvial channels the amplitude of both pools and riffles may adjust at the same time scale, in a bedrock river over a relatively short time scale there is a better opportunity to construct an alluvial riffle but excavation of a pool requires a longer time period.". Although Medway Creek's till is softer than bedrock, its erodibility (Pike, 2014; Bergman et al., 2016) might control the location of bedforms and their lack of cyclic organization and regular spacing (i.e. they are forced bedforms as a result of till control; Bergman et al., 2016). This notion fits with Ruhe's (1952) observation that on millennial time scales the differential incision and drainage densities of morainic channels are a function of time since deglaciation (i.e. corresponds to the Davisian theory of young and mature landform development). Very few cases of bedrock pool-riffle sequences are reported in the literature (Patton and Baker, 1978; Keller and Melhorn, 1978; Baker, 1984; O'Connor et al., 1986; Baker and Pickup, 1987; Carling et al., 2009) and thus formation/controlling mechanisms that we could use as analogues for semi-alluvial till channels are poorly understood relative to the vast alluvial channel architecture literature (Chartrand and Whiting, 2000). An interesting study by O'Connor et al. (1986) explained bedform formation in a small bedrock channel as a result of high discharges; bouldery riffles form where stream power drops, i.e. downstream of canyon expansions and upstream of canyon bends and constrictions. On Medway Creek we cannot use channel widths as a forming mechanism, as the average widths of the different bedforms is essentially identical and only bedform lengths and slopes vary between bedforms (see Bergman et al., 2016; Fig. 10 and Table 3). The largest boulders in the study reach are randomly positioned in places where till produced them (either from banks or the bed) and they appear in all three bedform types. We have no direct evidence that Medway Creek has ever had the competence to move them, although Carling et al. (2009) stated they might be associated with a longer time-scale of extreme discharges.

Since pools constitute most of the channel bed's area (55%) we conclude that channel form is in most instances poorly adjusted to discharge and coarse sediments in the bed. The results of the Wohl (2004) method (presented above) shows pools have a poorly developed DHG, which contradicts the Lisle and Hilton (1992, 1999) stability method. It also contradicts the $LRBS^*$ and RSI analyses which suggest that riffles are not stable (Bergman et al., 2016). The Wohl (2004) method does align well with the ecological fish-biodiversity assessment. Consequently, one would expect Medway bedforms to be rich in fine sediment to induce that instability (Wilcock et al., 2001), but this is not the case (Bergman et al., 2016).

To summarize, the downstream hydraulic geometry discriminator formula proposed by Wohl (2004) was implemented in order to analyze the ability of the 3 major bedform types to adjust to the control the moraine and till exert over the reach. Pools that are a scour bedform feature and occupy most of the channel's area (55%) are poorly adjusted. In contrast, most riffles are hydraulically well adjusted, perhaps because their sediment overlies till rather than cutting into it as with pools. The flats are a transitional bedform between pools and riffles. These results suggest that different bedform types operate on different time scales depending on the local erodibility/resistance of the till, implying that the channel is operating within strong glacial conditioning and has not yet reached a state of quasi-equilibrium that is seen in alluvial rivers and expressed as rhythmic spacing of bedforms. Our results are supported by a study by Livers and Wohl (2015) that compared alluvial valley rivers to rivers in formerly glacial valleys in low-order mountain streams of the Colorado Front Range. They found that formerly glacial valleys display much more variability in channel geometry, the DHG is not distinct and when looking on a smaller local reach bedform-scale (10^1-10^3 m) the local control (or process domains) overrides the larger drainage basin downstream relationships.

5. Conclusions

Bank incision into till and Arva Moraine bluffs is highly variable, creating a variety of bank elevations. Similarly, channel widths vary considerably, resulting in a wider range of width/depth (W/D) ratios relative to larger till-bedded channels on a reach-scale. When conducting a similar analysis on a basin-scale, the average W/D ratio of Medway Creek falls in between other river types in a variety of environments, and specifically for formerly glaciated terrains it seems that till control is associated with the scale of analysis. Bankfull discharge, a key metric in any hydrologic analysis, is found to produce a range of values when using equations or gauged statistics. Furthermore, precise modeling of bankfull discharge for each cross-section shows that there is not a single value discharge on a reach-scale or smaller bedform-scale. Maximum discharge also does not show clear geometric relations nor preserves an external morphology. The steep bluffs of the Arva Moraine are the dominant geomorphic feature of the reach. The velocity reversal hypothesis was not detected at a range of simulated in-bank and overbank flows.

Because of these hydrogeomorphic irregularities resulting from a strong glacial legacy, till channels cannot be classified as ordinary self-forming alluvial gravel-bed rivers. The lack of compatibility between sedimentologic stability (adjoining paper) and bedform DHG stability (the Wohl method) demonstrates that current metrics often used by fluvial scientists and river managers and practitioners are inadequate for a semi-alluvial channel with strong lithologic control. Similar to rapid assessment protocols and sedimentary metrics, hydrologic metrics used to detect channel stability or change must be used in fluvial context of a vast body of knowledge about similar rivers or as part of a long-term monitoring program which gives a reference for comparison. Alternatively, hydrologic and sedimentary metrics can be coupled to a biologic survey such as the Index of Biotic Integrity (IBI) when investigating semi-alluvial till channels. This is especially important when actively managing or restoring morainic semi-alluvial rivers that differ from their alluvial river counterparts. The distinction between temporal 'natural' processes and morphologic variability such as long-term glacial legacy and short-term hydrologic and morphologic changes linked to human activities over the landscape might prove difficult to separate. There are few rivers that can be considered completely natural and pristine. It is the responsibility of river scientists and practitioners to manage and restore these semi-alluvial rivers in adequate ways that take into account the complexity of intertwined long-term glacial legacy and modern short-term anthropogenic modifications.

References

Alexander, J.S., Wilson, R.C., Green, W.R., 2012. A brief history and summary of the effects of river engineering and dams on the Mississippi River system and delta: U.S. Geological Survey Circular 1375, 43 p.

Allen, P.M., Narramore, R., 1985. Bedrock controls on stream channel enlargement with urbanization, North Central Texas. Water Resources Bulletin 21, 1037-1048.

Anderson, R.J., Bledsoe, B.R., Hession, W.C., 2004. Width of streams and rivers in response to vegetation, bank material, and other factors. Journal of the American Water Resources Association 40, 1159-1172.

Andrews, E.D., 1980. Effective and bankfull discharges of streams in the Yampa River basin, Colorado and Wyoming. Journal of Hydrology 46, 311-330.

Annable, W.K., 1996. Morphologic relationship of rural watercourses in southern Ontario and selected field methods in fluvial geomorphology. Credit Valley Conservation Authority, Ontario Ministry of Natural Resources, Peterborough, ON, 92 p.

Annandale, G.W., 2006. Scour Technology: Mechanics and Engineering Practice. McGraw Hill Civil Engineering, 420 p.

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-central Lower Michigan, U.S.A. Geomorphology 102, 615-623.

ASCE Task Committee on Hydraulics, Bank Mechanics, and Modeling of River Width Adjustment., 1998. River width adjustment II: Modeling. Journal of Hydraulic Engineering 124, 903-917.

Ashley, G.M., Renwick, W.H., Haag, G.H., 1988. Channel form and processes in bedrock and alluvial reaches of the Raritan River, New Jersey. Geology 16, 436-439.

Ashmore, P., Church, M., 2001. The impact of climate change on rivers and river processes in Canada. Geological Survey of Canada Bulletin 555, 58 p.

Bagnold, R.A., 1960. Some aspects of the shape of river meanders. Geological Survey Professional Paper 282-E, United States Government Printing Office, Washington, 135-144.

Baker, V.R., 1977. Stream channel response to floods with examples from central Texas. Geological Society of America Bulletin 88, 1057-1071.

Baker, V.R., 1984. Flood sedimentation in bedrock fluvial systems. In: Koster, E.H., Steel, R.J., (Eds.) Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, 87-98.

Baker, V.R., Pickup, G., 1987. Flood geomorphology of the Katherine Gorge, Northern Territory, Australia. Geological Society of America Bulletin 98, 635-646.

Barry, J. J., Buffington, J. M., King, J. G., 2004. A general power equation for predicting bed load transport rates in gravel bed rivers. Water Resources Research 40, W10401.

Bergman, N., Van De Wiel, M.J., Hicock, S.R., 2016. Semi-alluvial stream channel crossing the interlobate Arva Moraine, Medway Creek, Southern Ontario, Canada. 1: Sedimentary characteristics and morphologic change. This dissertation, chapter 2.

Bhowmik, N.G., 1979. Hydraulics of Flow in the Kaskaskia River, Illinois. Illinois State Water Survey, Report of Investigations 91, Urbana, 116 p.

Borgniet, L., Estupina, V., Puech, C., Dartus, D., 2003. Methodologies for analyzing intrinsic and required DEM accuracy for Hydrological applications of flash floods. Canadian Journal of Remote Sensing 29, 734-740.

Brakenridge, G.R., Thomas, P.A., Conkey. L.E., Schiferle, J.C., 1988. Fluvial sedimentation in response to postglacial uplift and environmental change, Missisquoi River, Vermont. Quaternary Research 30, 190-203.

Brardinoni, F., Hassan, M.A., 2007. Glacially induced organization of channel-reach morphology in mountain streams. Journal of Geophysical Research 112, F03013.

Bridge, J.S., 2003. Rivers and Floodplains: Forms, Processes, and Sedimentary Record. Wiley-Blackwell, Malden, MA, 504 p.

Brown, R.A., Pasternack, G.B., Lin, T., 2016. The topographic design of river channels for form-process linkages. Environmental Management 57, 929-942.

Buffington, J.M., 2012. Changes in channel morphology over human time scales. In: Church, M., Biron, P.M., Roy, R.G., (Eds.) Gravel-bed Rivers: Processes, Tools, Environments. Chapter 32, John Wiley and Sons, Chichester, 435-462.

Campo, S.H., Desloges, J.R., 1994. Sediment yield conditioned by glaciation in a rural agricultural basin of southern Ontario, Canada. Physical Geography 15, 495-515.

Carling, P., 1988. The concept of dominant discharge applied to two gravel-bed streams in relation to channel stability thresholds. Earth Surface Processes and Landforms 13, 355-367.

Carling, P.A., Herget, J., Lanz, J.K., Richardson, K., Pacifici, A., 2009. Channel scale erosional bedforms in bedrock and in loose granular material: character, processes and implications. In: Burr, D.M., Carling, P.A., Baker, V.R., (Eds.), Megaflooding on Earth and Mars. Chapter 2, Cambridge University Press, 13-32.

Champoux, O., Biron, P.M., Roy, A.G., 2003. The long-term effectiveness of fish habitat restoration practices: Lawrence Creek, Wisconsin. Annals of the Association of American Geographers 93, 42-54.

Chartrand, S.M., Whiting, P.J., 2000. Alluvial architecture in headwater streams with special emphasis on step-pool topography. Earth Surface Processes and Landforms 25, 583-600.

Chatanantavet, P., Parker, G., 2009. Physically based modeling of bedrock incision by abrasion, plucking, and macroabrasion. Journal of Geophysical Research 114, F04018.

Cheng, N., 2016. Representative grain size and equivalent roughness height of a sediment bed. Journal of Hydraulic Engineering 142, 0.1061/(ASCE)HY.1943-7900.0001069, 06015016.

Chow, V.T., 1959. Open-channel hydraulics. New York, McGraw-Hill, 680 pp.

Czuba, J.A., Oberg, K.A., Best, J.L., Parsons, D.R., 2011. Coherent channel-scale secondary flow and sediment transport pathways in the Upper Saint Clair River. Journal of Great Lakes Research 37, 480-493.

Darby, S.E., Simon, A., (Eds.) 1999. Incised River Channels: Processes, Forms, Engineering and Management. John Wiley & Sons, Chichester, 442 p.

Davis, S.N., 1958. Size distribution of rock types in stream gravel and glacial till. Journal of Sedimentary Petrology 28, 87-94.

Day, S.S., Gran, K.B., Belmont, P., Wawrzyniec, T., 2013. Measuring bluff erosion part 1: terrestrial laser scanning methods for change detection. Earth Surface Processes and Landforms 38, 1055-1067.

Dodov, B., Foufoula-Georgiou, E., 2004. Generalized hydraulic geometry: Insights based on fluvial instability analysis and a physical model. Water Resources Research 40, W12201.

Doyle, M.W., Shields, D., Boyd, K.F., Skidmore, P.B., DeWitt, D., 2007. Channel-forming discharge selection in river restoration design. Journal of Hydraulic Engineering, ASCE, 133, 831-837.

Eaton, B.C., Church, M., 2007. Predicting downstream hydraulic geometry: A test of rational regime theory. Journal of Geophysical Research 112, F03025.

Faustini, J.M., Kaufmann, P.R., Herlihi, A.T., 2009. Downstream variation in bankfull width of wadeable streams across the conterminous United States. Geomorphology 108, 292-311.

Finnegan, N.J., Roe, G., Montgomery, D.R., Hallet, B., 2005. Controls on the channel width of rivers: Implications for modeling fluvial incision of bedrock. Geology 33, 229-232.

Foster, G.E., 1998. Morphology of three semi-alluvial stream channels in southern Ontario. Unpublished Masters Thesis. The University of Western Ontario, 215 p.

Froehlich, D.C., 1994. Width and depth-constrained best trapezoidal section. Journal of Irrigation and Drainage Engineering, ASCE, 120, 828-835.

Gingras, D., Adamowski, K., Pilon, P.J., 1994. Regional flood equations for the provinces of Ontario and Quebec. Water Resources Bulletin, American Water Resources Association, 30, 55-67.

Grabowski, R.C., Droppo, I.G., Wharton, G., 2011. Erodibility of cohesive sediment: The importance of sediment properties. Earth Science Reviews 105, 101-120.

Graf, W.L., 1988. Definition of flood plains along arid-region rivers. In: Baker, V.R., Kochel, R.C., and Patton, P.C., (Eds.), Flood Geomorphology. John Wiley, New York, 231-242.

Gran, K.B., Belmont, P., Day, S.S., Jennings, C., Johnson, A., Perg, L., Wilcock, P.R., 2009. Geomorphic evolution of the Le Sueur River, Minnesota, USA and implications for current sediment loading. In: James, L.A., Rathburn, S.L., Whittecar, G.R., (Eds.) Restoration of fluvial systems with broad historical changes and human impacts. GSA Special paper 451, 119-130.

Gran, K.B., Belmont, P., Day, S.S., Finnegan, N., Jennings, C., Lauer, W.J., Wilcock, P.R., 2011. Landscape evolution in south-central Minnesota and the role of geomorphic history on modern erosional processes. GSA Today 21, 7-9.

Hack, J.T., 1965. Postglacial drainage evolution and stream geometry in the Ontonagon area, Michigan. USGS Professional Paper 504-B.

Hanson, G.J. Simon, A., 2001. Erodibility of cohesive streambeds in the loess area of the midwestern USA. Hydrological Processes 15, 23-38.

Harman, C., Stewardson, M., DeRose, R., 2008. Variability and uncertainty in reach bankfull hydraulic geometry. Journal of Hydrology 351, 13-25.

Hassan, M.A., Brayshaw, D., Alila, Y., Andrews, E., 2014. Effective discharge in small formerly glaciated mountain streams of British Columbia: Limitations and implications. Water Resources Research 50, 4440-4458.

Hill, A.R., 1973. Erosion of river banks composed of glacial till near Belfast, Northern Ireland. Zeitschrift für Geomorphologie 17, 428-442.

Holmes, C.D., 1937. Glacial erosion in a dissected plateau. American Journal of Science 33, 217-232.

Isard, S.A., Schaetzl, R.J., 1998. Effects of winter weather conditions on soil freezing in southern Michigan. Physical Geography 19, 71-94.

Johnson, P.A., Heil, T.M., 1996. Uncertainty in estimating bankfull conditions. Water Resources Bulletin 32, 1283-1291.

Khan, I., Kostachuk, R., 2011. Erodibility of cohesive glacial till bed sediments in urban stream channel systems. Canadian Journal of Civil Engineering 38, 1363-1372.

Kamphuis, J.W., 1983. On the erosion of consolidated clay material by a fluid containing sand. Canadian Journal of Civil Engineering 10, 223-231.

Kamphuis, J.W., Gaskin, P.N., Hoogendoorn, E., 1990. Erosion tests on four intact Ontario clays. Canadian Geotechnical Journal 27, 692-696.

Kaufmann, P.R., Faustini, J.M. Larsen, D.P. Shirazi, M.A., 2008. A roughness-corrected index of relative bed stability for regional stream surveys. Geomorphology 99, 150-170.

Kaufmann, P.R., Larsen, D.P., Faustini, J.M., 2009. Bed stability and sedimentation associated with human disturbances in Pacific Northwest streams. Journal of American Water Resources Association 45, 434-459.

Keller, E.A., 1971. Areal sorting of bed-load material: the hypothesis of velocity reversal. Geological Society of America Bulletin 82, 753–756.

Keller, E.A., Melhorn, W.N., 1978. Rhythmic spacing and origins of pools and riffles. Geological Society of America Bulletin 89, 723-730.

Khan, I., Kostaschuk, R., 2011. Erodibility of cohesive glacial till bed sediments in urban stream channel systems. Canadian Journal of Civil Engineering 38, 1363-1372.

Knox, J.C., 1989. Long and short-term episodic storage and removal of sediment in watersheds of southwestern Wisconsin and northwestern Illinois. In: Hadley, R.F., Ongley, E.D., (Eds.), Sediment and the Environment. International Association of Hydrological Sciences, Wallingford, UK, 157-164.

Kondolf, M.G., Wilcock, P.R., 1996. The flushing flow problem: Defining and evaluating objectives. Water Resources Research 32, 2589-2599.

Lane, E.W., 1955. Design of stable channels. Transaction American Society of Civil Engineers 120, 1234-1260.

Leopold, L.B. Maddock, T. 1953. The hydraulic geometry of stream channels and some physiographic implications. U.S. Geological Survey Professional Paper 252, 1-57.

Leopold, L.B., Wolman, M.G., Miller, J.P., 1964, Fluvial processes in geomorphology. San Francisco, Calif., W.H. Freeman and Co., 522 p.

Lisle, T.E., Buffington, J.M., Wilcock, P.R., Bunte, K., 2015. Can rapid assessment protocols be used to judge sediment impairment in gravel-bed streams? A commentary. Journal of the American Water Resources Association 51, 373-387.

Liu, X., Parker, G., Czuba, J.A., Oberg, K.A., Mier, J.M., Best, J.L., Parsons, D.R., Ashmore, P., Garcia, M.H., 2012. Hydrodynamic modeling, sediment mobility, and bed armoring of the St. Clair River. Earth Surface Processes and Landforms 37, 957-970.

Livers, B., Wohl, E., 2015. An evaluation of stream characteristics in glacial versus fluvial process domains in the Colorado Front Range. Geomorphology 231, 71-82.

Mackin, J.H., 1948. Concept of the graded river. Geological Society of America Bulletin 59, 463-512.

Mahaney, W.C., Hancock, R.G.V., Somelar, P., Milan, A., 2016. Iron and aluminum soil/paleosol extractions as age/environment indicators: Some examples from a catchment in southern Ontario, Canada. Geomorphology 270, 159-171.

McCandless, T.L., Everett, R.A., 2002. Maryland stream survey: Bankfull discharge and channel characteristics of the streams in the Piedmont hydrologic region. U.S. Fish & Wildlife Service, Chesapeake Bay Field Office CBFO-S02-01, Annapolis Maryland, 40 p.

McCandless, T.L., 2003. Bankfull discharge and channel characteristics of streams in the Allegheny Plateau and the Valley and Ridge hydrologic regions. U.S. Fish and Wildlife Service, Chesapeake Bay Field Office CBFO-S03-01, Annapolis Maryland, 33 p.

Menzies, J., van der Meer, J.J.M., Rose, J., 2006. Till--as a glacial "tectomict", its internal architecture, and the development of a "typing" method for till differentiation. Geomorphology 75, 172-200.

Merten, E.C., Hemstad, N.A., Kolka, R.K., Newman, R.M., Verry, E.S., Vondracek, B., 2010. Recovery of sediment characteristics in moraine, headwater streams of Northern Minnesota after forest harvest. Journal of the American Water Resources Association 46, 733-743.

Mickelson, D.M., Colgan, P.M., 2003. The southern Laurentide Ice Sheet. Development in Quaternary Science 1, 1-16.

Mier, J.M., Garcia, M.H., 2011. Erosion of glacial till from the St. Clair River (Great Lakes Basin). Journal of Great Lakes Research 37, 399-410.

Montgomery, D.R., Gran, K.B., 2001. Downstream variations in the width of bedrock channels. Water Resources Research 37, 1841-1846.

Mulvihill, C.I., Baldigo, B.P., Miller, S.J., DeKoskie, D., DuBois, J., 2009, Bankfull discharge and channel characteristics of streams in New York State. U.S. Geological Survey Scientific Investigations Report 2009-5144, 51 p.

Nanson, G.C., Croke, J.C., 1992. A genetic classification of floodplains. Geomorphology 4, 459-486.

Nicholas, A., 2013. Morphodynamic Modeling of Rivers and Floodplains. In: Volume 2: Quantitative Modeling of Geomorphology, Treatise on Geomorphology, 160-179.

Nunally, N.R. 1967. Definition and identification of channel and over-bank deposits and their respective role in floodplain formation. Professional Geographer 19, 1-4.

O'Connor, J.E., Webb, R.H., Baker, V.R., 1986. Paleohydrology of pool-and-riffle pattern development: Boulder Creek, Utah. Geological Society of America Bulletin 97, 410-420.

O'Connor, J.E., Webb, R.H., 1988. Hydraulic modeling for paleoflood analysis. In: Baker, V.R., Kochel, R.C., and Patton, P.C., (Eds.), Flood Geomorphology. John Wiley and Sons Inc., 393-402.

Oliva, F., Viau, A.E., Bjornson, J., Desrochers, M.A. 2016. A 1300 year reconstruction of paleofloods using oxbow lake sediments in temperate southwestern Quebec, Canada. Canadian Journal of Earth Sciences 53, 378-386.

Palmer M.A., Menninger H.L., Bernhardt, E., 2010. River restoration, habitat heterogeneity and biodiversity: a failure of theory or practice? Freshwater Biology 55, 205-222.

Parker, G., 1978. Self-formed straight rivers with equilibrium banks and mobile bed. Part 2. The gravel river. Journal of Fluid Mechanics 89,127-146.

Parker, G., 1979. Hydraulic geometry of active gravel rivers, Journal of Hydraulics Division, ASCE, 105, 1185-1201.

Patton, P.C., Baker, V.R., 1976. Morphometry and floods in small drainage basins subject to diverse hydrogeomorphic controls. Water Resources Research 12, 941-952.

Patton, P.C., Baker, V.R., 1978. New evidence for pre-Wisconsin flooding in the Channeled Scabland of Eastern Washington. Geology 6, 567-571.

Patton, P.C., 1988. Geomorphic response of streams to floods in the glaciated terrain of southern New England. In: Baker, V.R., Kochel, R.C., Patton, P.C., (Eds.) Flood Geomorphology. John Wiley and Sons, New York, 261-277.

Pelletier, J.D., DeLong, S., 2004. Oscillations in arid alluvial-channel geometry. Geology 32, 713-716.

Phillips, J.D., 1990. The instability of hydraulic geometry. Water Resources Research 26, 739-744.

Phillips, R.T.J., Robert, A., 2005. River incision in relation to post-glacial events in the Humber River basin, Ontario. Géographie physique et Quaternaire 59, 17-30.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Phillips, R.T.J., Desloges, J.R., 2015a. Alluvial floodplain classification by multivariate clustering and discriminant analysis for low-relief glacially conditioned river catchments. Earth Surface Processes and Landforms 40, 756-770.

Phillips, R.T.J., Desloges, J.R., 2015b. Glacial legacy effects on river landforms of the southern Laurentian Great Lakes. Journal of Great Lakes Research 41, 951-964.

Pike, L., 2014. The dynamics of glacial till erosion: hydraulic flume tests on samples from Medway Creek, London, ON. Unpublished Master of Civil Engineering thesis, McGill University, Montreal, 99 pp.

Pitlick, J., Yantao, C., Wilcock, P., 2009. Manual for computing bed load transport using BAGS (Bedload Assessment for Gravel-bed Streams) Software. Gen. Tech. Rep. RMRS-GTR-223. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station, 45 p.

Ritter, D.F., 1975. Stratigraphic implications of coarse-grained gravel deposited as overbank sediment, southern Illinois. Journal of Geology 83, 645-650.

Rodriguez-Iturbe, I., Escobar, L.A., 1982. The dependence of drainage density on climate and geomorphology. Hydrological Sciences Journal 27, 129-137.

Rosgen, D.L., 1994. A classification of natural rivers. Catena 22, 169-199.

Ruhe, R.V., 1952. Topographic discontinuities of the Des Moines lobe. American Journal of Science 250, 46-56.

Schumm, S.A., 1960. The shape of alluvial channels in relation to sediment type. U.S. Geological Survey Professional Paper 352-B. U. S. Government Printing Office. Washington, D.C., 17-30.

Schumm, S.A., 1963. Sinuosity of alluvial rivers on the Great Plains. Geological Society of America Bulletin 74, 1089-1100.

Schumm, S.A., Harvey, M.D., Watson, C.C., 1984. Incised Channels: Morphology, Dynamics and Control. Water Resources Publication, Littleton, Co. 200 p.

Schwartz, J.S., Neff, K.J., Dworak, F.E., Woockman, R.R., 2015. Restoring riffle-pool structure in an incised, straightened urban stream channel using an ecohydraulic modeling approach. Ecological Engineering 78, 112-126.

Shields, D.F., Copeland, R.R., Klingeman, P.C., Doyle, M.W., Simon, A., 2003. Design for stream restoration. Journal of Hydraulic Engineering, ASCE, 129, 575-584.

Sholtes, J.S., Bledsoe, B.P., 2016. Half-yield discharge: Process-based predictor of bankfull discharge. Journal of Hydraulic Engineering, 142, doi: 10.1061/(ASCE)HY.1943-7900.0001137.

Shugar, D., Kostaschuk, R., Ashmore, P., Desloges, J.R., Burge, L., 2007. In situ jet testing of the erosional resistance of cohesive streambeds. Canadian Journal of Civil Engineering, 34, 1192-1195.

Simon, A., Doyle, M., Kondolf, M., Shields, F., Rhoads, B., McPhillips, M., 2007. Critical evaluation of how the Rosgen classification and associated ‘natural channel design’ methods fail to integrate and quantify fluvial processes and channel response. Journal of the American Water Resources Association 43, 1117–1131.

Stevens, M.A., Richardson, E.V., Simons, D.B., 1975. Nonequilibrium river form. Journal of Hydraulic Division, ASCE, 101, 557-566.

Stewart. A., Desloges. J.R., 2014. A 9000-year record of vertical and lateral accretion on the floodplain of the Lower Thames River, Southwestern Ontario, Canada. Quaternary International 342, 161-172.

Stock, J.D., Montgomery, D.R., Collins, B.D., Dietrich, W.E., Sklar, L., 2005. Field measurements of incision rates following bedrock exposure: Implications for process controls on the long profiles of valleys cut by rivers and debris flows. Geological Society of America Bulletin 117, 174-194.

Strahler, A.N., 1958. Dimensional analysis applied to fluvially eroded landforms. Geological Society of America Bulletin 69, 279-300.

Thayer, B.J., 2010. Floodplain development for a semi-alluvial river. Unpublished Honor Thesis, Department of Geography, Western University, 87 p.

Thayer, B.J., Ashmore, P., 2016. Floodplain morphology, sedimentology, and development processes of a partially alluvial channel. Geomorphology 269, 160-174.

Thayer, B.J., Phillips, R.T.J., Desloges, J.R., 2016. Downstream channel adjustment in a low-relief, glacially conditioned watershed. Geomorphology 262, 101-111.

Tinkler, K., Wohl, E.E., 1998. A primer on bedrock channels. In: Tinkler, J. Wohl, E.E. (Eds.) Rivers Over Rock: Fluvial Processes in Bedrock Channels. Geophysical Monograph Series vol. 107, AGU, Washington, D.C., 1-18.

Thompson, D.M., Wohl, E.E., Jarrett, R.D., 1999. Velocity reversals and sediment sorting in pools and riffles controlled by channel constrictions. Geomorphology 27, 229-241.

Thornbush, M.J., Desloges, J.R., 2011. Site reconstruction at an oxbow lake situated at the lower Nottawasaga River in southern Ontario, Canada. Environmental Archaeology 16, 1-15.

Trimble, S.W., 1997. Stream channel erosion and change resulting from riparian forests. Geology 25, 467-469.

Turowski, J.M., 2012. Semi-alluvial channels and sediment-flux-driven bedrock erosion. In: Church, M. Biron, P. Roy, A. (Eds.) Gravel Bed Rivers: Processes, Tools, Environments. John Wiley & Sons, Chichester, chapter 29, 401-416.

Upper Thames Conservation Authority., 2013. Medway Creek 2012 Watershed Report Card. Available from: http://thamesriver.on.ca/wp-content/uploads//WatershedReportCards/RC_Medway.pdf. Last accessed: June 1, 2014.

U.S. Army Corps of Engineers., 2016. HEC-RAS (Hydrologic Engineering Center – River Analysis System) Version 4.1. Available from: http://www.hec.usace.army.mil/software/hec-ras/downloads.aspx

Van De Wiel, M.J., Coulthard, T.J., Macklin, M.G., Lewin, J., 2007. Embedding reach-scale fluvial dynamics within the CAESAR cellular automaton landscape evolution model. Geomorphology 90, 283-301.

Walker, I., Desloges, J.R., Crawford, G., Smith, D., 1997. Floodplain formation processes and archaeological site integrity, Grand Banks, Southern Ontario. Journal of Geoarcheology 12, 865-887.

Wilcock, P., 1997. Friction between science and practice: The case of river restoration. Eos, Transactions American Geophysical Union 78, 454-454.

Wilcock, P.R., Kenworthy, S.T., Crowe, J.C., 2001. Experimental study of the transport of mixed sand and gravel. Water Resources Research 37, 3349-3358.

Wilkerson, G.V., Kandel, D.R., Perg, L.A., Dietrich, W.E., Wilcock, P.R., Whiles, M.R., 2014. Continental-scale relationship between bankfull width and drainage area for single-thread alluvial channels, Water Resources Research 50, 919-936.

Williams, G.P., 1978. Bank-full discharge of rivers. Water Resources Research 14, 1141-1154.

Wohl, E. 2004. Limits of downstream hydraulic geometry. Geology 32, 897-900.

Wohl, E.E., Merritts, D.M., 2001. Bedrock channel morphology. Geological Society of America Bulletin 113, 1205-1212.

Wolman, M.G., Gerson, R., 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surface Processes and Landforms 3, 189-208.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54-74.

Wong, T.S.W., Lim, C.K., 2006. Effect of loss model on evaluation of Manning roughness coefficient of experimental concrete catchment. Journal of Hydrology 331, 205-218.

Wood-Smith, R.D., Buffington, J.M., 1996. Multivariate geomorphic analysis of forest streams: implications for assessment of land use impacts on channel condition. Earth Surface Processes and Landforms 21, 377-393.

Zehe, E., Elsenbeer, H., Lindenmaier, F., Schulz, K., Blöschl, G., 2007. Patterns of predictability in hydrological threshold systems. Water Resources Research 43, W07434.

Figures

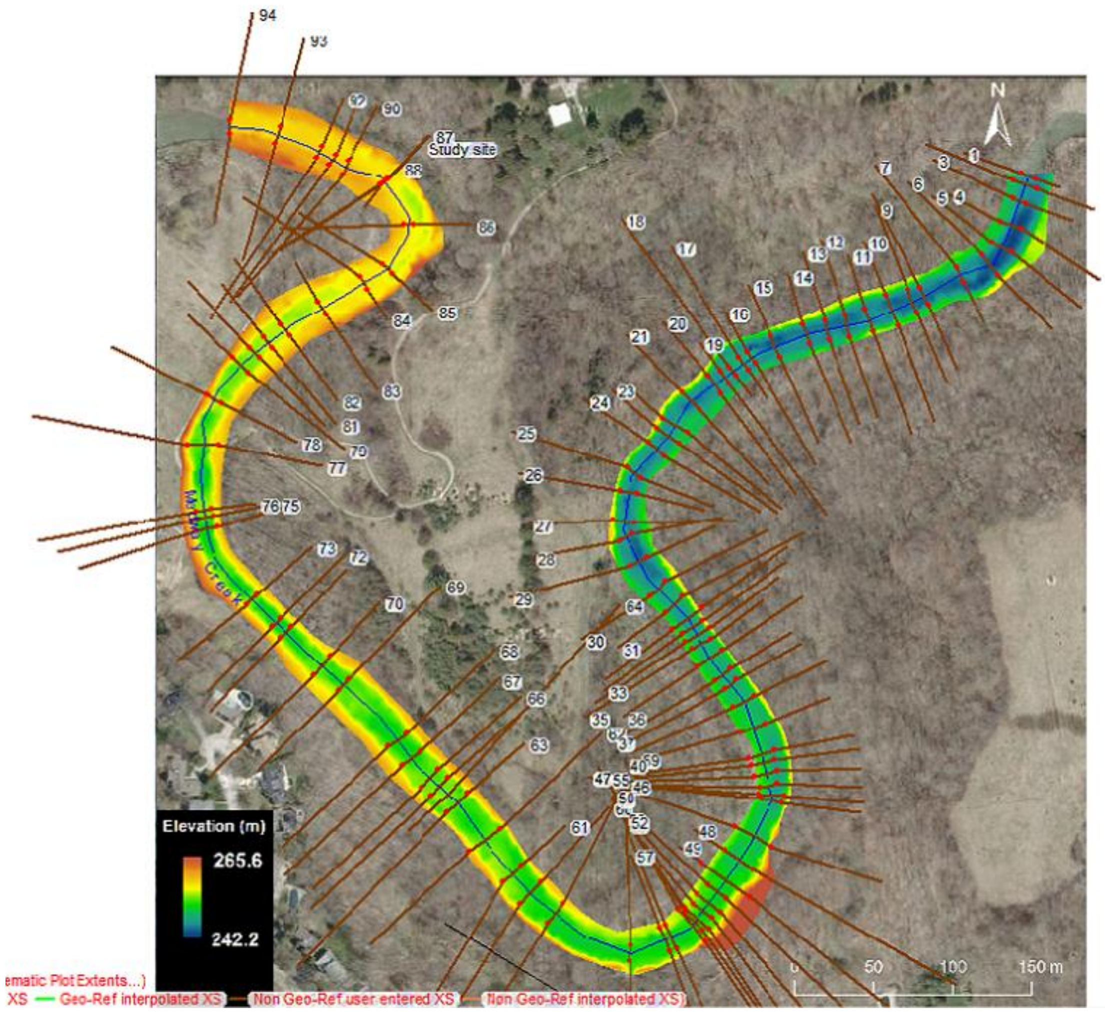


Figure 1. The 1D modeling setup of 94 cross-sections overlying Medway Creek's channel DEM and surrounding floodplain-valley aerial photograph.

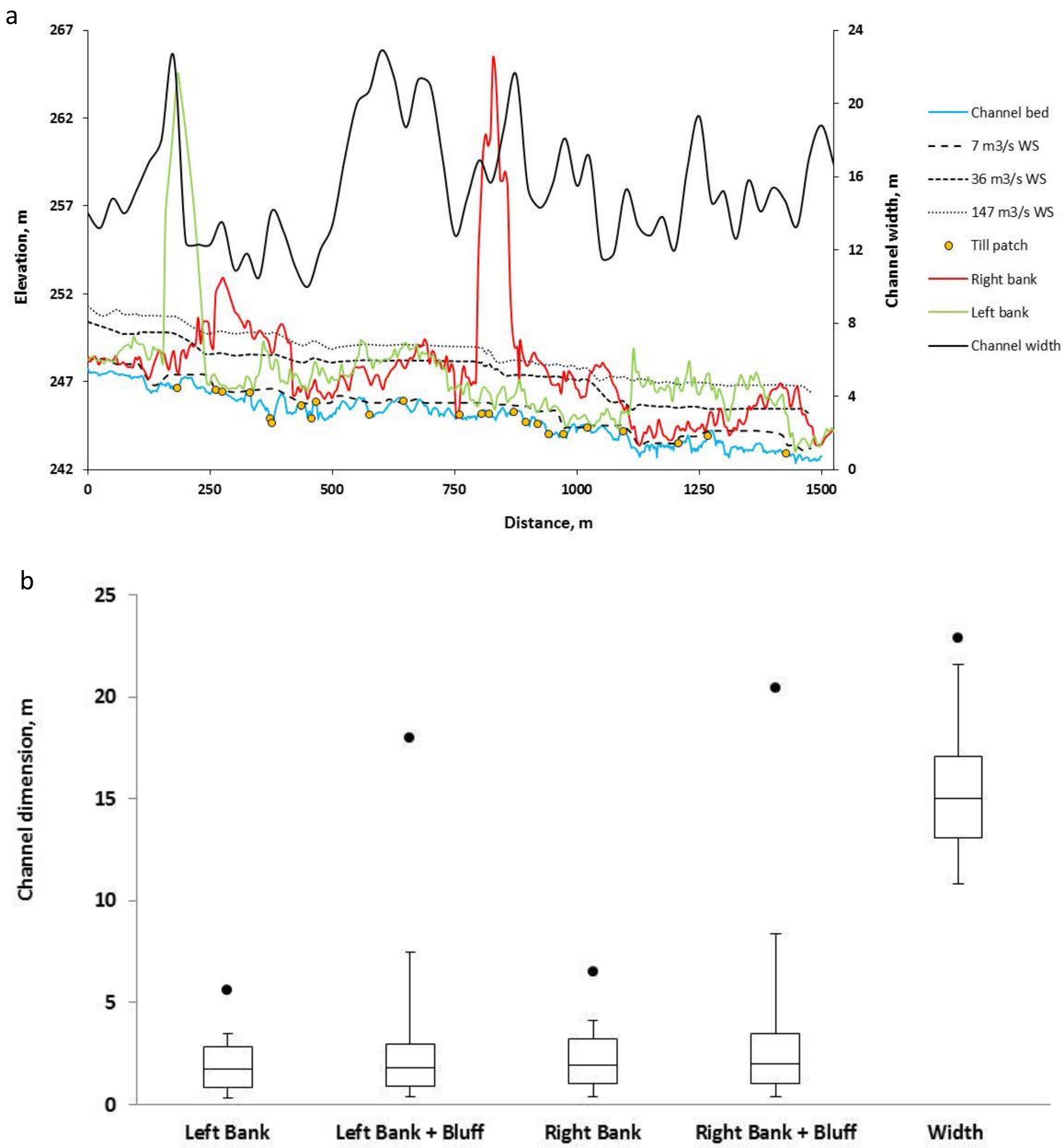


Figure 2. (a) The bed long profile, the right and left bank long profiles and channel widths along the study reach. Three selected long profiles of the minimum bankfull discharge, the average bankfull discharge and the largest flood on the measured record show that flow can be contained in many localities. The spike in elevation on each bank indicates a bluff from incision into the Arva moraine. (b) Channel width and bank heights comparison (n = 62). Boxes indicate 25th and 75th percentiles as well as the median; error bars indicate 5th and 95th percentiles. Dots indicate maximum values. Left and right bank heights (with and without bluffs heights) are similar.

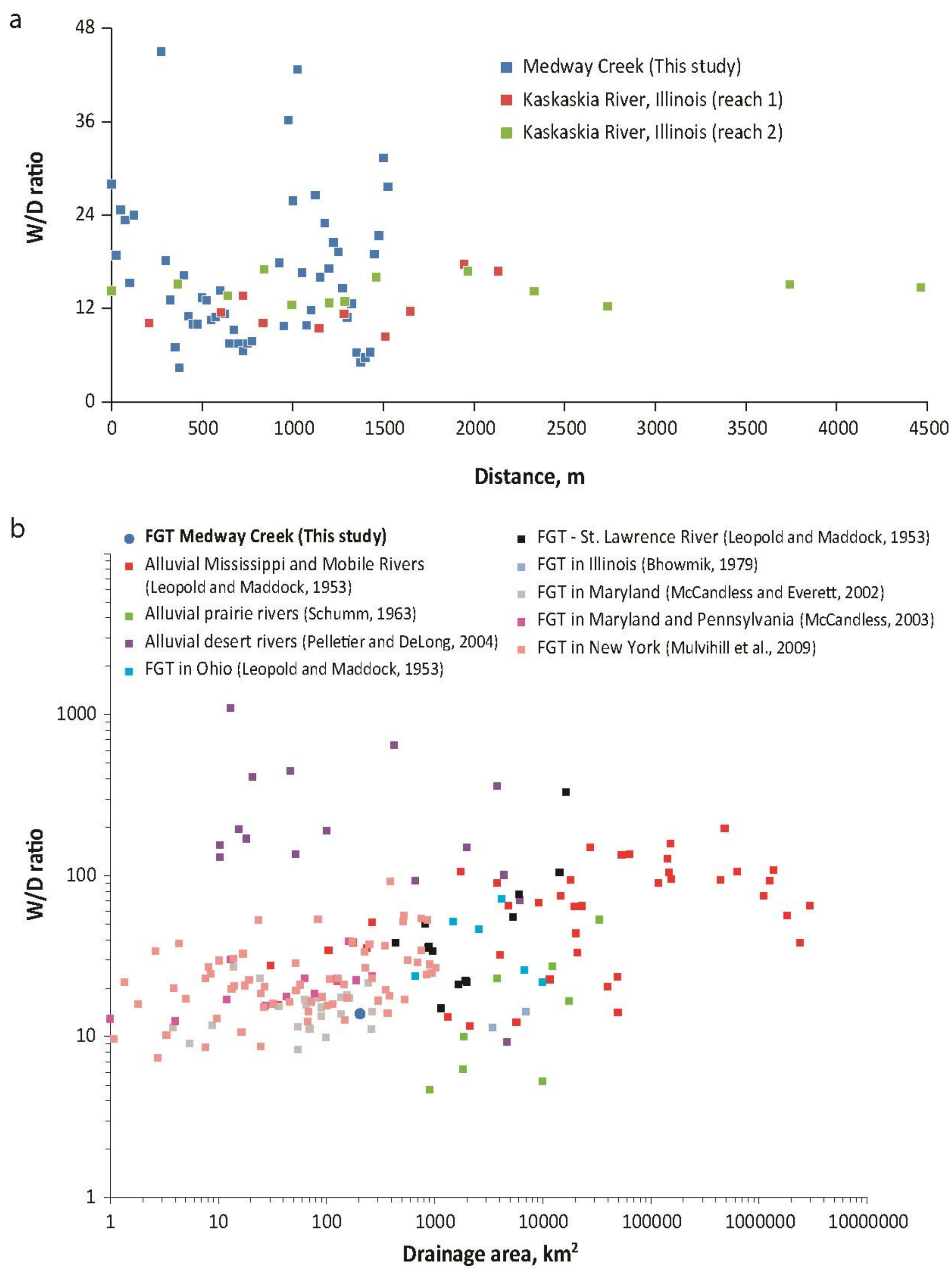


Figure 3. (a) Reach-scale W/D ratios for Medway Creek and a comparison with a channel with similar till boundary material – the Kaskaskia River in Illinois (Bhowmik, 1979). (b) Basin-scale W/D ratios of alluvial channels and formerly glaciated terrain (FGT) channels. Alluvial prairie rivers and alluvial desert rivers represent the two extremes of the W/D spectrum.

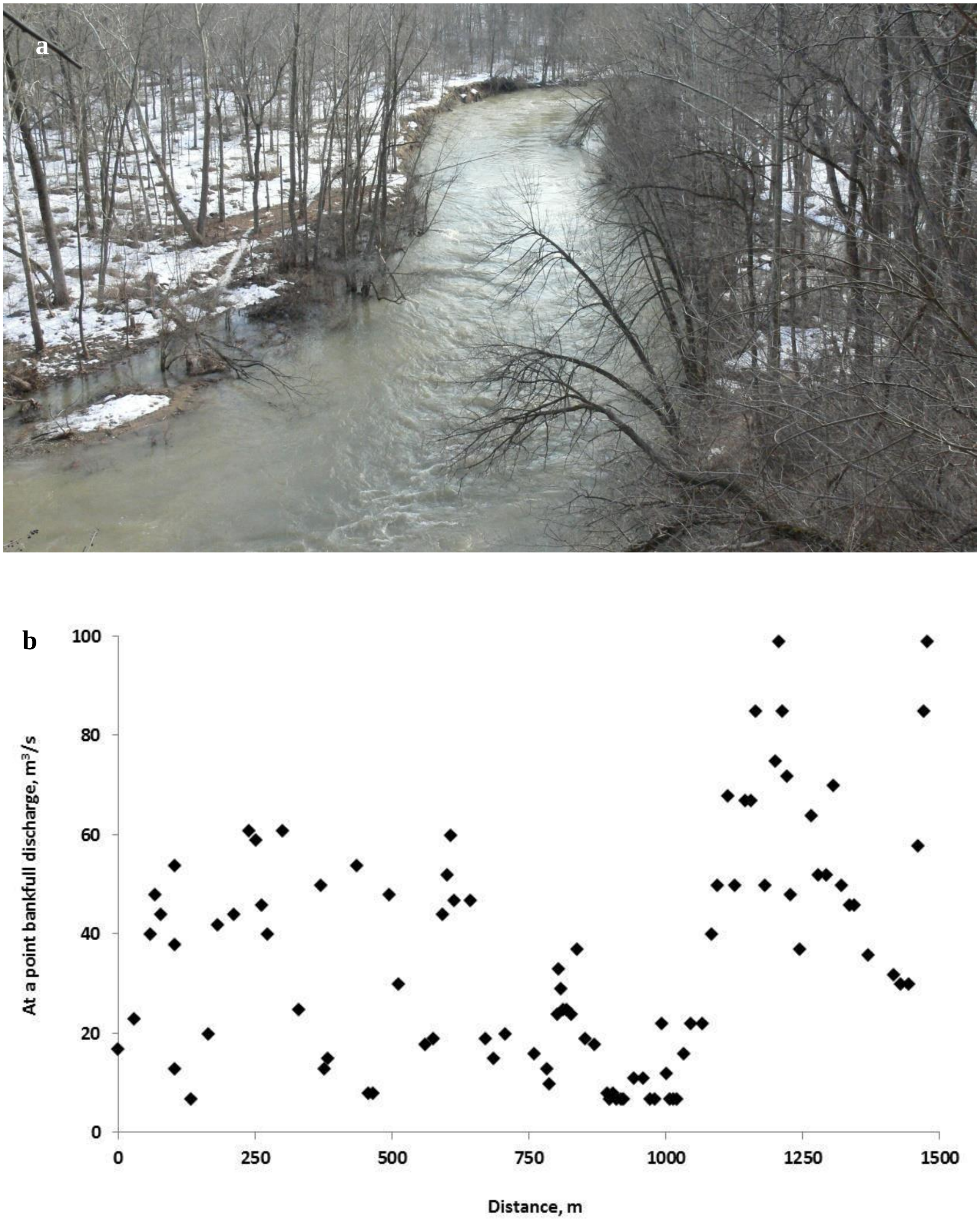
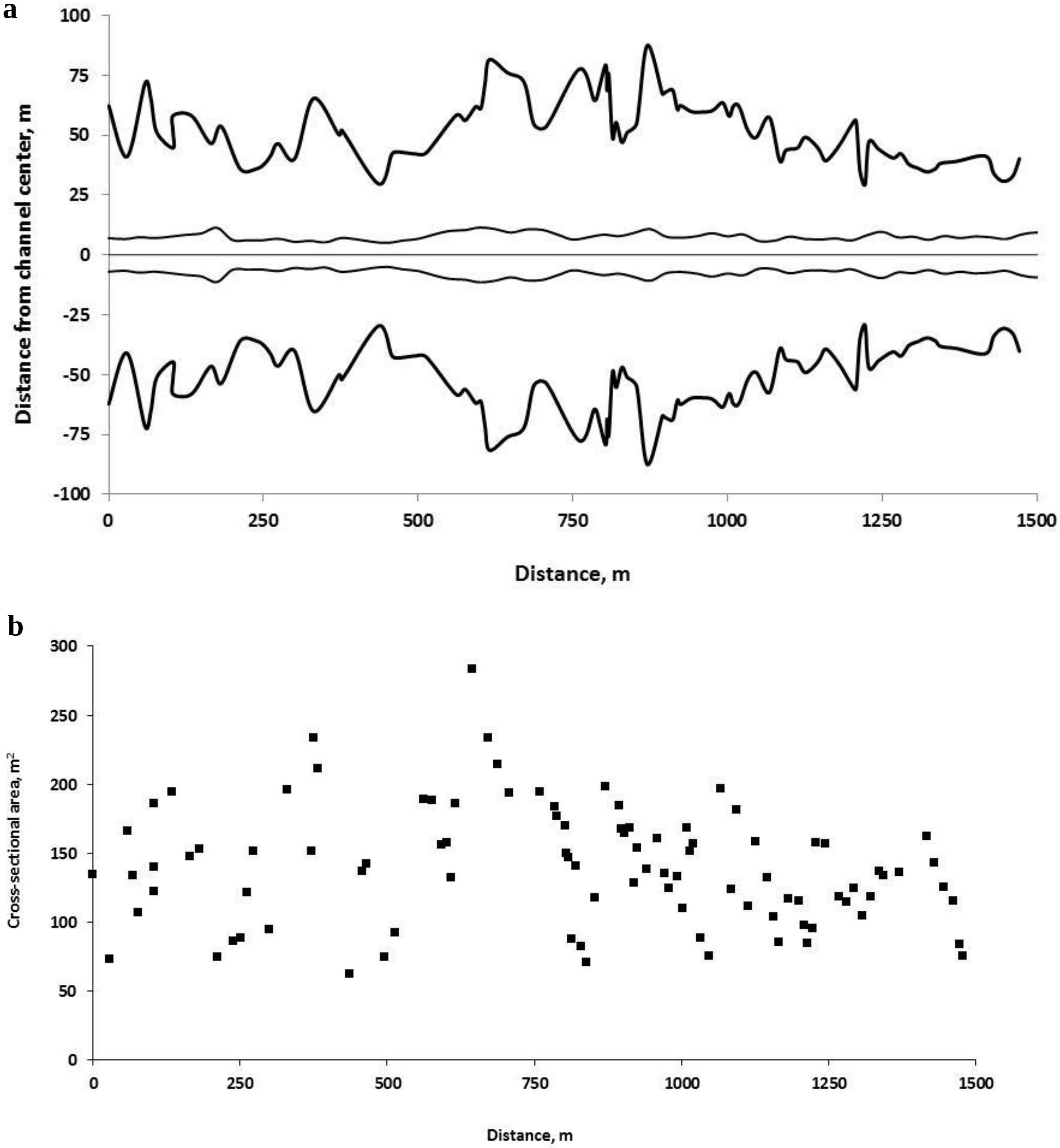


Figure 4. (a) An example of the problematic use of bankfull discharge during the March 18, 2011 spring melt flood (37.5 m^3/s) on a Medway Creek pool. Flow is overbank and covering the pointbar on the bottom left, while simultaneously upstream (top-center of the photo) the flow is entirely contained within the deep channel and is about half the right bank elevation. (b) Bankfull discharge for each cross-section derived using HEC-RAS simulations.

a



b

Figure 5. (a) Modeled channel widths during the largest flood on the measured record (147 m^3/s) normalized relative to the channel center and banks. (b) Cross-sectional area along the study reach. The minimum and maximum flow areas are associated with Arva Moraine bluffs.

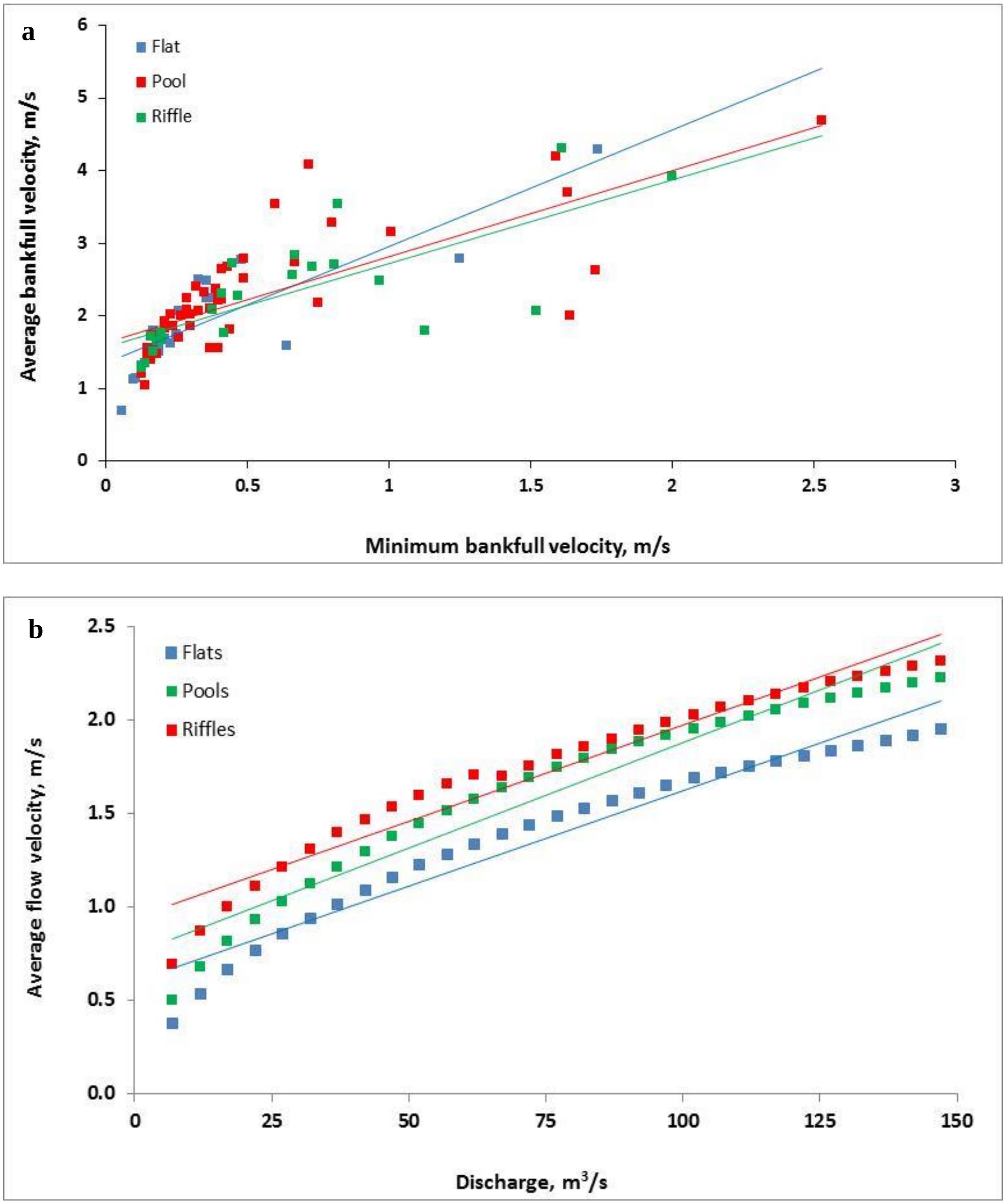


Figure 6. Comparison of modeled flow velocities of bedforms. (a) The average bankfull discharge velocities for all cross-sections (36 m^3/s) of the entire reach vs. the minimum bankfull flow velocities (7 m^3/s) for each bedform. (b) Average flow velocity of all three bedform types vs. incremental discharge increases of 5 m^3/s.

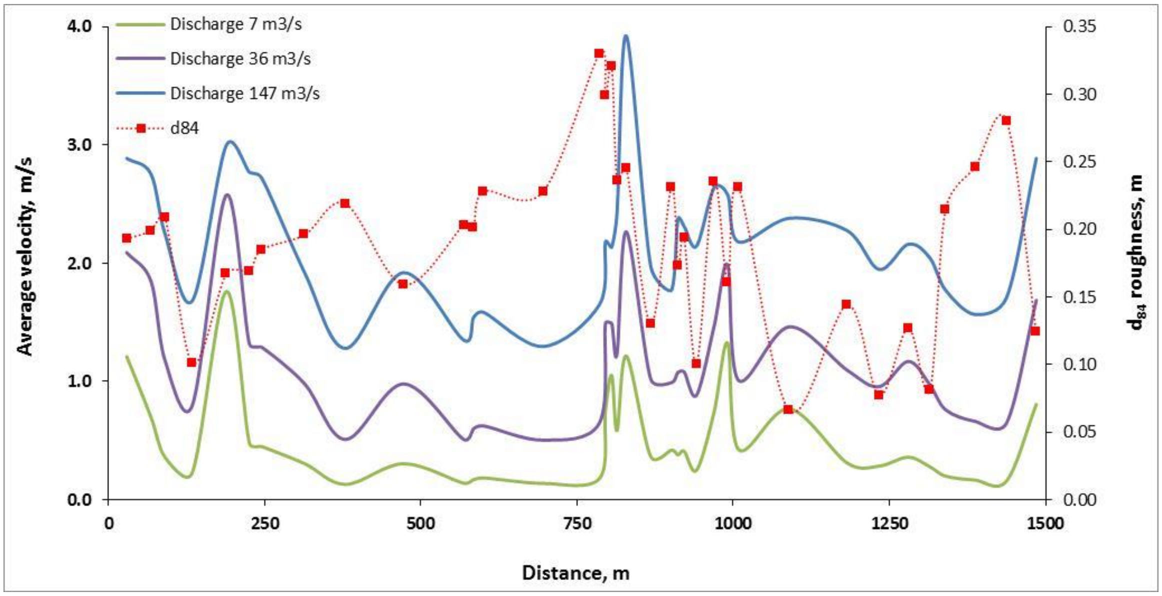


Figure 7. Average modeled velocities along the study reach coupled with bed roughness represented by d_{84}. Steep bluffs in the beginning, center and downstream (outside the study reach) are responsible for sharp velocity spikes at all discharges.

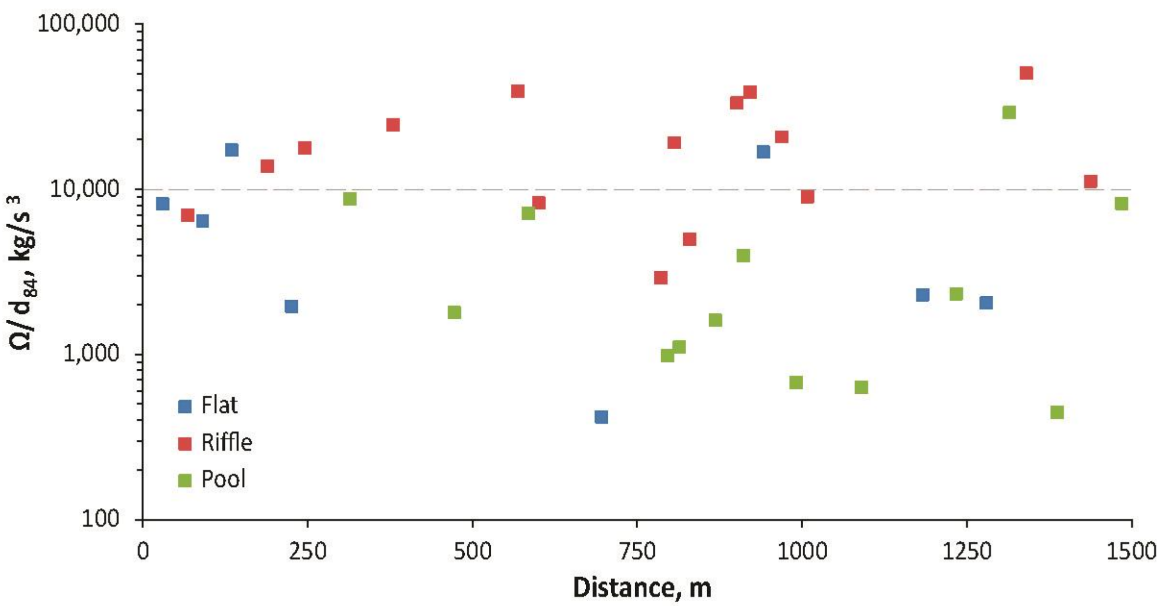


Figure 8. Bedform downstream hydraulic geometry discriminator according to the method proposed by Wohl (2004) for stream power vs. the 84th grain percentile. All values above 10,000 kg/s^3 (dashed grey line) are well developed while below it is undeveloped.

Tables

Table 1. Hydraulic geometry relations for width and depth, for the whole reach and per bedform.

	Width relations		Depth relations	
	Equation	R^2	Equation	R^2
Reach	$W_{reach} = 16.907Q_{bf}^{-0.03}$	0.0173	$D_{reach} = 2.2058Q_{bf}^{0.0217}$	0.0011
Pool	$W_{pools} = 15.164Q_{bf}^{-0.012}$	0.0032	$D_{pools} = 2.7683Q_{bf}^{-0.047}$	0.0085
Riffle	$W_{riffles} = 15.761Q_{bf}^{0.0101}$	0.0022	$D_{riffles} = 1.5642Q_{bf}^{0.1955}$	0.0637
Flat	$W_{flats} = 21.091Q_{bf}^{-1.071}$	0.0899	$D_{flats} = 1.7401Q_{bf}^{0.0381}$	0.0023

Table 2. DHG adjustment for all the bedforms in the study reach based on the Wohl (2004) method.

Hydraulic adjustment degree	Pools	Flats	Riffles	All* By count	All* By area
Well developed	1 (8%)	2 (25%)	10 (67%)	13 (36.1%)	19%
Marginally developed	3 (23%)	2 (25%)	2 (13%)	7 (19.4%)	11%
Poorly developed	9 (69%)	4 (50%)	3 (20%)	16 (44.4%)	70%

* Excluding the steps

Appendix
Table 1. W/D ratios for alluvial and formerly glaciated terrain (FGT) rivers.

Source	River and location	Scale	Drainage area, km^2	W/D ratio	Type of river
This study	Medway Creek (Ontario)	Reach	205	13.9	Formerly glaciated terrain
Leopold and Maddock (1953)	Mobile River Basin (Mississippi and Alabama)	Basin	5723.9	12.3	Alluvial
			2131.6	11.7	Alluvial
			11629.1	22.7	Alluvial
			1333.8	13.3	Alluvial
			40144.8	20.5	Alluvial
			49468.8	23.6	Alluvial
	Scioto River Basin (Ohio)		660.4	23.7	Formerly glaciated terrain
			1478.9	52.1	Formerly glaciated terrain
			2558.9	46.5	Formerly glaciated terrain
			4206.1	71.8	Formerly glaciated terrain
			6824.6	25.9	Formerly glaciated terrain
			9963.7	21.7	Formerly glaciated terrain
	Tennessee River Basin		175.9	38.5	Alluvial
			266.8	51.1	Alluvial
			30.3	27.8	Alluvial
			104.6	34.4	Alluvial
			1750.8	105.7	Alluvial
			4058.5	32.2	Alluvial
			4812.2	65.3	Alluvial
			235.2	35.6	Alluvial
			35.7	15.8	Alluvial
			23139.0	64.8	Alluvial
	St. Lawrence River Basin		955.7	33.9	Formerly glaciated terrain
			5306.9	55.7	Formerly glaciated terrain
			1150.0	15.0	Formerly glaciated terrain
			1983.9	21.8	Formerly glaciated terrain
			875.4	35.7	Formerly glaciated terrain
			435.1	38.1	Formerly glaciated terrain
			815.8	50.0	Formerly glaciated terrain
			888.4	36.1	Formerly glaciated terrain
			1665.4	21.1	Formerly glaciated terrain
			1939.9	22.2	Formerly glaciated terrain
			6032.1	77.0	Formerly glaciated terrain
			14322.6	104.8	Formerly glaciated terrain
			16353.2	331.8	Formerly glaciated terrain
	Kansas River Basin		3781.4	90.0	Alluvial
			27453.9	150.0	Alluvial
			53871.8	135.3	Alluvial
			63636.1	136.0	Alluvial
			9207.4	68.3	Alluvial
			14581.6	75.0	Alluvial
			18039.3	94.5	Alluvial
			19632.1	64.3	Alluvial
			20349.6	43.8	Alluvial
			21004.8	33.3	Alluvial
			49727.8	14.2	Alluvial
			117171.1	90.0	Alluvial
			143071.0	128.0	Alluvial
			146878.3	104.6	Alluvial
			151307.2	158.3	Alluvial
			155114.5	95.4	Alluvial
	Missouri River Basin		482774.1	197.0	Alluvial
			630662.6	105.7	Alluvial
			1098932.8	75.0	Alluvial
			1267023.1	93.4	Alluvial
			1368032.7	108.8	Alluvial
	Mississippi River Basin		444183.3	94.1	Alluvial
			1815583.0	56.6	Alluvial
			2415942.7	38.2	Alluvial
			2964243.6	65.1	Alluvial
Schumm (1963)	Arikaree River (Nebraska)	Basin	3,781	23.0	Alluvial
	White River (Nebraska)	Basin	906	4.7	Alluvial
	Powder River (Montana)	Basin	33,411	53.3	Alluvial
	Solomon River (Kansas)	Basin	17,534	16.6	Alluvial
	North Fork Republican (Nebraska)	Basin	12,354	27.4	Alluvial
	Sappa Creek (Nebraska)	Basin	9,946	5.3	Alluvial
	Prairie Dog Creek (Kansas)	Basin	1,867	10.0	Alluvial
	Red Willow Creek (Nebraska)	Basin	1,839	6.3	Alluvial
Bhowmik (1979)	Kaskaskia River (Illinois)	Reach	3,445	11.4	Formerly glaciated terrain, downstream of a dam
		Reach	7,045	14.3	Formerly glaciated terrain, downstream of a dam
Trimble (1997)	Coon Creek (Wisconsin, Driftless Area)	Reach	360	18.0	Grassy alluvial
		Reach	360	12.0	Grassy alluvial
		Reach	360	12.0	Grassy alluvial
		Reach	360	11.0	Grassy alluvial
		Reach	360	20.0	Forested alluvial
		Reach	360	18.0	Forested alluvial
		Reach	360	17.0	Forested alluvial
		Reach	360	16.0	Forested alluvial
McCandless and Everett (2002)	Baisman Run (Maryland)	Reach	4	11.39	Alluvial
	Basin Run (Maryland)	Reach	14	27.04	Alluvial
	Beaver Run (Maryland)	Reach	36	15.49	Alluvial
	Beaverdam Run (Maryland)	Reach	54	11.52	Alluvial
	Bennett Creek (Maryland)	Reach	163	17.41	Alluvial
	Big Elk Creek (Maryland)	Reach	136	17.57	Alluvial
	Big Pipe Creek (Maryland)	Reach	264	14.32	Alluvial
	Cranberry Branch (Maryland)	Reach	9	11.72	Alluvial
	Deer Creek (Maryland)	Reach	244	21.54	Alluvial
	Hawlings River (Maryland)	Reach	70	11.2	Alluvial
	Jones Falls (Maryland)	Reach	65	15.74	Alluvial
	Little Falls (Maryland)	Reach	137	13.79	Alluvial
	Little Patuxent River (Maryland)	Reach	98	9.88	Alluvial
	Long Green Creek (Maryland)	Reach	24	22.95	Alluvial
	Morgan Run (Maryland)	Reach	73	16.35	Alluvial

	Jones Falls (Maryland)	Reach	65	15.74	Alluvial
	Little Falls (Maryland)	Reach	137	13.79	Alluvial
	Little Patuxent River (Maryland)	Reach	98	9.88	Alluvial
	Long Green Creek (Maryland)	Reach	24	22.95	Alluvial
	Morgan Run (Maryland)	Reach	73	16.35	Alluvial
	Northeast Creek (Maryland)	Reach	63	17.01	Alluvial
	NW Br Anacostia River (Maryland)	Reach	55	8.32	Alluvial
	Patuxent River (Maryland)	Reach	90	13.37	Alluvial
	Piney Creek (Maryland)	Reach	81	17.41	Alluvial
	Seneca Creek (Maryland)	Reach	262	11.11	Alluvial
	Slade Run (Maryland)	Reach	5	9.11	Alluvial
	Western Run (Maryland)	Reach	155	18.13	Alluvial
	Winters Run (Maryland)	Reach	90	15.19	Alluvial
McCandless (2003)	Bear Creek (Maryland)	Reach	27	15.6	Alluvial
	Bear Creek (Maryland)	Reach	127	23.0	Alluvial
	Big Piney Run (Pennsylvania)	Reach	63	23.0	Alluvial
	Casselman River (Maryland)	Reach	162	39.3	Alluvial
	Crabtree Creek (Maryland)	Reach	43	17.8	Alluvial
	Ditch Run (Maryland)	Reach	12	17.1	Alluvial
	Evitts Creek (Pennsylvania)	Reach	78	18.6	Alluvial
	North Branch Potomac River (Maryland)	Reach	189	22.4	Alluvial
	Savage River (Maryland)	Reach	127	22.2	Alluvial
	Savage River (Maryland)	Reach	4	12.5	Alluvial
	Sawpit Run (Maryland)	Reach	13	30.1	Bedrock
	Sideling Hill Creek (Maryland)	Reach	264	23.9	Alluvial
	Toliver Run trib. (Maryland)	Reach	1	12.8	Alluvial
	Youghiogheny River trib. (Maryland)	Reach	1	13.0	Alluvial
Pelletier and DeLong (2004)	Bouse Wash (Arizona)	Basin	4364	101.0	Desert alluvial entrenched and braided
	Wild Burro Wash (Arizona)	Basin	21	410.0	Desert alluvial entrenched and braided
	Cottonwood Wash (Arizona)	Basin	47	450.0	Desert alluvial entrenched and braided
	Hassayampa River (Arizona)	Basin	3807	362.5	Desert alluvial entrenched and braided
	Vamori Wash (Arizona)	Basin	6185	70.0	Desert alluvial entrenched and braided
	Dead Mesquite Wash (Arizona)	Basin	10	130.0	Desert alluvial discontinuous
	North Airport Wash (Arizona)	Basin	18	170.0	Desert alluvial discontinuous
	Vail Wash (Arizona)	Basin	10	155.0	Desert alluvial discontinuous
	Vail-DM Wash (Arizona)	Basin	16	195.0	Desert alluvial entrenched and braided
	E La Quituni Valley A (Arizona)	Basin	13	1100.0	Desert alluvial distributary with channel fan
	E La Quituni Valley B (Arizona)	Basin	13	1100.0	Desert alluvial distributary with channel fan
	Cananda del Oro (Arizona)	Basin	660	92.9	Desert alluvial entrenched and braided
	Sycamore Creek (Arizona)	Basin	425	650.0	Desert alluvial entrenched and braided
	Peñitas Wash (Arizona)	Basin	101	190.0	Desert alluvial entrenched and braided
	Centennial Wash (Arizona)	Basin	4706	9.3	Desert alluvial entrenched and braided
	Bobaquivari Wash (Arizona)	Basin	52	136.7	Desert alluvial entrenched and braided
	Vamori Wash (Arizona)	Basin	1997	150.0	Desert alluvial entrenched and braided
Mulvihill et al. (2009)	Archer Creek (New York)	Reach	1.3	21.9	Formerly glaciated terrain
	Buck Creek	Reach	3.3	10.2	Formerly glaciated terrain
	Tributary to Mill Creek Tributary	Reach	4.3	38.1	Formerly glaciated terrain
	Vly Brook	Reach	8.5	24.5	Formerly glaciated terrain
	North Creek	Reach	16.9	32.7	Formerly glaciated terrain
	Hopkinton Brook	Reach	51.8	28.6	Formerly glaciated terrain
	Glowegee Creek	Reach	67.3	12.4	Formerly glaciated terrain
	Plum Brook	Reach	113.7	16.0	Formerly glaciated terrain
	Little Hoosic River	Reach	145.3	21.1	Formerly glaciated terrain
	Independence River	Reach	229.7	26.9	Formerly glaciated terrain
	West Branch Au Sable River	Reach	300.4	16.7	Formerly glaciated terrain
	Sandy Creek	Reach	354.8	19.6	Formerly glaciated terrain
	East Branch Au Sable River	Reach	512.8	51.7	Formerly glaciated terrain
	Bouquet River	Reach	699.3	28.9	Formerly glaciated terrain
	Moose River	Reach	940.2	24.7	Formerly glaciated terrain
	Batten Kill River (New York)	Reach	1025.6	26.8	Formerly glaciated terrain
	North Branch of Foulertons Brook (New Jersey)	Reach	1.1	9.7	Formerly glaciated terrain
	Stony Brook	Reach	5.0	17.2	Formerly glaciated terrain
	Horse Pound Brook	Reach	10.2	29.7	Formerly glaciated terrain
	Hunter Brook	Reach	19.2	22.6	Formerly glaciated terrain
	Valatie Kill	Reach	24.6	18.6	Formerly glaciated terrain
	Mahwah River	Reach	31.9	16.1	Formerly glaciated terrain
	Kisco River	Reach	45.6	16.5	Formerly glaciated terrain
	Roeliff Jansen Kill	Reach	71.2	16.6	Formerly glaciated terrain
	Fishkill Creek	Reach	148.4	12.6	Formerly glaciated terrain
	Ramapo River	Reach	225.1	33.6	Formerly glaciated terrain
	Tenmile River (Connecticut)	Reach	525.8	17.0	Formerly glaciated terrain
	Kinderhook Creek	Reach	852.1	24.4	Formerly glaciated terrain
	Sage Brook	Reach	1.8	16.0	Formerly glaciated terrain
	Cold Spring Brook	Reach	3.9	19.9	Formerly glaciated terrain
	Shackham Brook	Reach	7.6	23.0	Formerly glaciated terrain
	Little Elk Creek	Reach	9.7	12.9	Formerly glaciated terrain
	Merrill Creek tributary	Reach	13.8	30.6	Formerly glaciated terrain
	Albright Creek	Reach	17.6	20.8	Formerly glaciated terrain
	Mink Creek	Reach	26.9	20.4	Formerly glaciated terrain
	Trout Creek	Reach	52.3	19.4	Formerly glaciated terrain
	Steele Creek	Reach	67.9	14.3	Formerly glaciated terrain
	Little Delaware River	Reach	129.0	23.1	Formerly glaciated terrain
	Butternut Creek	Reach	154.6	17.4	Formerly glaciated terrain
	Otselic River	Reach	380.7	18.0	Formerly glaciated terrain
	Otselic River	Reach	562.0	29.8	Formerly glaciated terrain
	Tioughnioga River	Reach	756.3	34.4	Formerly glaciated terrain
	West Branch Delaware River	Reach	859.9	53.0	Formerly glaciated terrain
	Little Tonawanda Creek Tributary	Reach	2.6	34.0	Formerly glaciated terrain
	Stony Brook Tributary	Reach	8.2	27.0	Formerly glaciated terrain
	Ischua Creek Tributary	Reach	13.3	19.8	Formerly glaciated terrain
	Cuthrie Run	Reach	14.0	20.7	Formerly glaciated terrain
	Big Creek	Reach	16.4	10.7	Formerly glaciated terrain
	Ball Creek	Reach	23.5	53.0	Formerly glaciated terrain
	Little Tonawanda Creek	Reach	57.2	21.0	Formerly glaciated terrain
	Cayuga Inlet	Reach	91.2	17.7	Formerly glaciated terrain
	Catherine Creek	Reach	106.4	22.8	Formerly glaciated terrain
	Fivemile Creek	Reach	173.0	38.8	Formerly glaciated terrain
	Cayuga Creek	Reach	249.7	37.7	Formerly glaciated terrain
	Cazenovia Creek	Reach	349.6	36.7	Formerly glaciated terrain
	Catatonk Creek	Reach	391.1	92.3	Formerly glaciated terrain
	Conewango Creek	Reach	751.1	54.0	Formerly glaciated terrain
	Second Creek Tributary	Reach	2.8	7.4	Formerly glaciated terrain
	Canandaigua Outlet Tributary	Reach	7.6	8.6	Formerly glaciated terrain
	East Branch of Allen Creek	Reach	24.6	8.7	Formerly glaciated terrain
	Northup Creek	Reach	26.2	15.2	Formerly glaciated terrain
	Butternut Creek	Reach	83.4	53.7	Formerly glaciated terrain
	Irondequoit Creek	Reach	101.5	15.6	Formerly glaciated terrain
	Flint Creek	Reach	264.2	23.1	Formerly glaciated terrain
	Irondequoit Creek	Reach	367.8	14.0	Formerly glaciated terrain
	Oatka Creek	Reach	518.0	56.6	Formerly glaciated terrain
	Tonawanda Creek	Reach	903.9	28.2	Formerly glaciated terrain

Conclusions

Main findings

This work describes landscape evolution in a formerly glaciated terrain of low relief in southern Ontario. While the three papers concentrate on case studies of two streams (Thames River and Medway Creek), they are relevant to the entire Great Lakes region, the American Midwest and other North American areas affected by the Laurentide Ice Sheet and resulting glacial deposits. The work also addresses two contrasting themes in fluvial geomorphology: Do fluvial systems result from catastrophic floods (Baker, 1977; Wolman and Gerson, 1978) or are today's frequent small to medium floods (Wolman and Miller, 1960) the main drivers of landscape evolution? Instead of proving which theme dominates, this thesis shows that they can not only operate together but complement each other and represent a continuum of two different scales of erosion. Catastrophic floods leave a lasting legacy for smaller scale present-day erosion processes (i.e., the larger external morphology and glacial conditioning (Phillips and Desloges, 2014) set the stage for the modern landscape. Furthermore, in today's fluvial landscape it is difficult to distinguish between 'natural' erosion processes linked to glacial legacy and modern anthropogenic erosion associated with a variety of human activities.

Q1. Do fluvial systems in the London area result from catastrophic floods or are today's frequent small to medium floods the main drivers of landscape evolution?

The first paper sets the stage for a deglaciating environment. Glacial Lake London was a meltwater lake formed behind the Arva and Ingersoll Moraines. While it was a small lake compared to the ancestral Great Lakes, it dominated the area of what are now the City of London and the London Basin both of which are located on its dry bed. The former lake outlet begins near the neighborhood of Byron and flowed into a deep river valley within Komoka Provincial Park making the modern Thames River a misfit channel in this unusual spillway. Using local DEM, stratigraphic evidence, and topographic constraints, Glacial Lake London was fully reconstructed for the first time - expanding on the work of Dreimanis et al. (1998). This reconstruction determined peak discharge of lake drainage using parametric breach equations (Wu et al., 2011) as well as step-backwater HEC-RAS dam-break modeling based on unusual landforms found in Komoka Provincial Park. Various parameters of lake drainage were compared to other moraine-dammed lakes. Reconstruction of the lake and its catastrophic drainage reveal that geomorphic events and evolution of the landscape outside the ice margins were rapid and did not involve gradual erosion typical of postglacial fluvial incision (Hack, 1965; Phillips and Robert, 2005; Arbogast et al., 2008). Besides reconstructing local and regional chronology of deglaciation, the draining of Glacial Lake London has implications for allowing flora, fauna and humans to encroach on the exposed deglaciating landscape (Delcourt and Delcourt, 1984; Mandrak and Crossman, 1992; Yu, 2000; Elias, 2013) or even disappear when the ice readvanced (Dreimanis, 1967; Ellis et al., 2011).

But the lake and its drainage do not only affect the modern Thames River but have much wider implication for present-day Thames River tributaries too. Local tributaries of the Thames River incise into the lacustrine and underlying glacial sediments. The dry lake bed of the London Basin and the incising Thames River serve as their baselevel. Medway Creek discussed in papers 2 and 3 was a flooded inlet during Glacial Lake London existence; it was gradually incised into the lacustrine and till sediments after deglaciation and has a stagnant paleodelta at its outlet with the North Thames River suggesting it was probably a more powerful river after deglaciation. However, unlike the Thames River, there is no field evidence to support a spillway or flow reversal as Dreimanis et al. (1998) suggested. The London landscape was therefore shaped by a combination of a catastrophic flood(s) first and continued erosion of numerous small – medium floods as the climate warmed. The Arva Moraine that served as Glacial Lake London's western rim is still a prominent control on Medway Creek's channel evolution but it is a remnant of glacial processes not a catastrophic fluvial one.

Q2. Do the geomorphic characteristics and dynamics of fluvial systems incised in tills differentiate them from alluvial rivers and bedrock rivers?

The second paper documents a 1.5 km reach of Medway Creek - a small agricultural channel that cuts through the interlobate Arva Moraine within London. The channel incised into lacustrine sediments (seen on top of bluffs) and into till. The study reach lacks significant external water and sediment inputs from tributaries which makes it an ideal site for characterizing a semi-alluvial cobble-bed channel incised into till. Analyses of till samples collected from the channel bed showed that the till is not homogenous suggesting it has spatial and temporal erodibility, a result corroborated by geotechnical tests (SPT and plasticity) and till patch erosion rates using erosion pins placed in the bed over several years. Short-term erosion rates are comparable to channels incising into soft bedrock (Tinkler and Parrish, 1998; Stock et al., 2005) and are very high compared to the local (geologic) incision rate. However, till exposures constitute a small portion of the total bed area. Nearby upstream boulders of various sizes do not seem to be directly related to till patch area regardless of their position. Alluvium thickness varies from 0 (full till exposure) to 60 cm in riffle tails but in many flats and pools it is only one grain thick. Bedforms are highly disorganized and show no regular spacing typically found in stable autogenic bedrock and alluvial rivers. The channel lacks typical alluvial sorting such as downstream fining; there is slight downstream coarsening, implying some non-fluvial control. The typical two-layered armor structure normally seen in perennial alluvial rivers is completely missing. The sand fraction that is a known destabilizing fraction of the bed in alluvial channels is not as dominant along the bedforms as one would expect from a channel incised into fine sediments.

The third paper expands the second paper and investigates the morphometry, hydrology and hydraulics of Medway Creek. Bank incision into till and Arva Moraine is highly variable, creating a variety of bank elevations. Similarly, channel widths vary considerably, resulting in a wider range of width/depth (W/D) ratios relative to larger till-bedded channels on a reach-scale. When conducting a similar analysis on a basin-scale, the average W/D ratio of Medway Creek falls between other river types in a variety of environments, and specifically for formerly glaciated terrains it seems that till control is associated with drainage area and scale of analysis. Bankfull discharge has a range of values based on equations or gauged statistics. In fact, modeling of bankfull discharge for each cross-section shows that there is not a single value discharge on a reach-scale or smaller bedform-scale. Similarly, maximum discharge also does not show clear geometric relations nor does it preserve external morphology. The steep bluffs of Arva Moraine are the dominant geomorphic feature of the reach. The velocity reversal hypothesis was not detected at a range of simulated in-bank and overbank flows.

Together papers two and three reveal that because of hydrogeomorphic irregularities resulting from a strong glacial legacy (Phillips and Desloges, 2014; Phillips and Desloges, 2015a), till-bedded channels cannot be classified as ordinary self-forming alluvial gravel-bed rivers. Instead, they are semi-alluvial channels falling between the alluvial and bedrock channel continuum, in most instances are ignored

(Montgomery and Buffington, 1997), and are only now receiving attention (Phillips and Desloges, 2015b).

Q3. Are channel bed sedimentary, hydrologic and hydraulic stability metrics originally developed for alluvial rivers suitable for use in till-bedded channels?

In papers 2 and 3 we used a variety of quantitative sedimentary and hydraulic metrics to test whether Medway Creek is a stable channel. The stability determination is important because it answers whether the channel is in (quasi-) equilibrium state or adjusting to short or long-term perturbation. The short-term perturbation is on the scale of annual-decadal-century such as land use changes and forest clearing since the British-European settlement (Schottler et al., 2013) while the longer term perturbation is on millennial scale since the area deglaciated and encountered base level fall (Gran et al., 2011). Furthermore, the metrics results are inferring river health and degree of impairment therefore it is the first step in any channel assessment.

We first applied three different **sedimentary attribute** indicators of channel stability: the Log Relative Bed Stability (LRBS*) index of Kaufmann et al. (2008; 2009), Kappesser's (2002) Riffle Stability Index (RSI), and fine sediment abundance (Lisle and Hilton, 1992, 1999). Results were inconclusive. The LRBS* gave unexpected results that pools are the most stable bedforms while riffles are the least stable bedform. In Kappesser's (2002) Riffle Stability Index Medway Creek's riffles fall within a relatively narrow RSI range indicating that riffles possess similar sedimentary attributes, with relatively low textural variability. The technique proposed by Lisle and Hilton (1992, 1999) of relative volume of fine sediment in pools does not show high loading of these grain fractions. The pools' relative volumes of fine sediment on Medway Creek support the previous stability analysis of Kaufmann et al. (2008; 2009) that pools are the most stable bedform of the study reach. The conclusion from the three sedimentary stability metrics is that without a series of textural surveys that allows comparison to some kind of historic reference values it is difficult to determine whether the channel is in a healthy state or not. Additionally, the sedimentary metrics are relying on the fine fractions to indicate impairment while the till parent material is not taken into account although it is predominantly fine-natured in the first place.

The **hydrologic stability** of a river is best expressed in its cross-sectional morphology (i.e. graded river concept; Mackin, 1948) either reflecting bankfull discharge or another effective or dominant discharge. I used literature equations, hydrologic statistics and regional curves of similar till rivers but none gave satisfactory results with high degree of certainty that this is the correct value. When the reach cross-sections were modeled using a 1D step-backwater HEC-RAS model, similar to a method used in paleoflood hydrology, it was also found that there are numerous values of bankfull discharge proving that this important metric is problematic in this non-self-forming channel. While width contractions and expansions are typical of alluvial rivers when transitioning from one bedform to the next, drastic changes in cross-sectional area with minor change in slope suggests strong channel instability and that local till erodibility sometimes overcomes the ability of the flow to scour it. In order to determine this metric in a channel with strong till control, it might be necessary to determine a range of flows rather than a single value like is ordinarily determined for alluvial rivers.

In order to investigate the bedforms ability to adjust to till erodibility, climate and flow regime, we used a **downstream hydraulic geometry (DHG) metric** proposed by Wohl (2004). This equation is used as a discriminator: where the ratio of stream power to sediment size (Ω/d_{84}) exceeds 10,000 kg/s^3, downstream hydraulic geometry is well developed; where the ratio is 9000 - 10,000 kg/s^3 it is marginal and when the result falls below 9,000 kg/s^3, downstream hydraulic geometry relationships are poorly developed. Most riffles are well adjusted to the DHG, while the few that are not are somewhat marginal. The majority of pools are with poorly developed DHG and flats are generally in between. Since pools constitute most of the channel bed's area (55%) it's implied that channel form is in most instances poorly adjusted. The results of the Wohl (2004) method shows pools have a poorly developed DHG, which contradicts the Lisle and Hilton (1992, 1999) stability method. It also contradicts the RBS^* and RSI analyses which suggest that riffles are not stable.

The inconsistency between different alluvial channels' stability metrics (sedimentary, hydrologic and hydraulic) casts doubt on the simplicity and suitability to use them without a concurrent long-term monitoring program that can differentiate between inherent instability of till channels (i.e. glacial conditioning) and impairment or river health associated with human activities. Considering that in southern Ontario the monitoring is in most instances concentrating on stream hydrology, limited data on water quality and some biological inventory (for example, the 5-year Report Cards of UTRCA), the current use of stability metrics could lead to erroneous management and river restoration decisions that could be more damaging than beneficial. It is therefore recommended not to use stability metrics as a reliable methodology until the knowledge about till-bedded channels reaches a more advanced mature state of understanding of form and process. Furthermore, river practitioners in formerly glaciated regions should be aware that they cannot automatically use alluvial channel practices although these ordinarily seem cost-effective in terms of time and finances.

Recommendations

This work recommends that research of till rivers receive more attention from the fluvial community as knowledge about them is currently lacking. One major implication is that management and restoration of these rivers cannot be simply borrowed from practices used on alluvial rivers but require careful consideration of the strong till control based on monitoring and sound science. This is crucial when these rivers experience human intervention coupled with climate warming. While till rivers cover vast extents of formerly glaciated areas in North America and around the world, they deserve a classification of their own and should not be assumed to be some kind of hybrid river between alluvial and bedrock channels with a history of glacial conditioning. Based on our study and other investigations of similar channels in glacial deposits, semi-alluvial channels deserve to be classified separately, as we propose in Fig. 1 based on the Montgomery and Buffington (1997) classification. This is not only relevant for river classification, but this knowledge gap also has implications for everyday management and restoration practices and whether current indices such as channel geometry and stability metrics are suitable and adequate to determine river impairment and health.

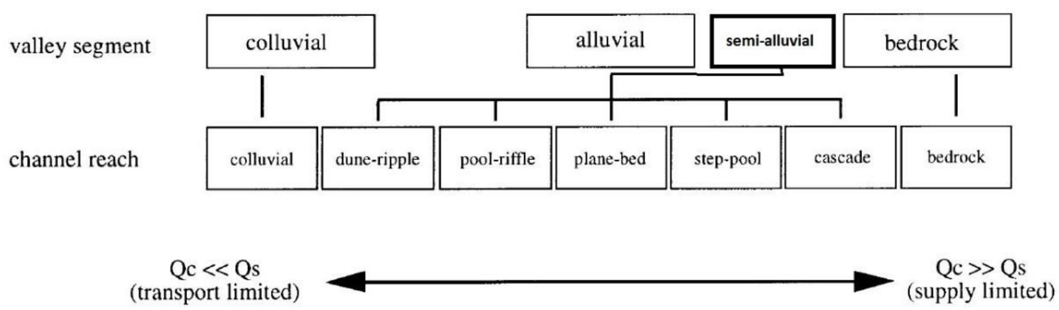


Figure 1. A revised Montgomery and Buffington (1997) channel classification with a new semi-alluvial category.

While till exerts a permanent geologic control on rivers in formerly glaciated terrains, its heterogeneity both in terms of different types of tills (Fig. 1 in the introduction) and even within the same type of till (as seen in paper 2 from the erosion pins), is expressed in channel form and processes. The spatial and temporal variability in till erodibility is what differentiates till rivers from alluvial rivers that are ordinarily easily shaped by fluvial erosion of their sediments. In that respect, tills resemble more soft bedrock that produces similar bedforms but their adjustment to channel hydrology and hydraulics operate on different time scales.

References

Arbogast, A.F., Bookout, J.R., Schrotenboer, B.R., Lansdale, A., Rust, G.L., Bato, V.A., 2008. Post-glacial fluvial response and landform development in the upper Muskegon River valley in North-central Lower Michigan, U.S.A. Geomorphology 102, 615-623.

Baker, V.R., 1977. Stream channel response to floods with examples from central Texas. Geological Society of America Bulletin 88, 1057-1071.

Delcourt, P.A., Delcourt, H.R., 1984. Late Quaternary paleoclimate and biotic responses in eastern North America and the western North Atlantic Ocean. Palaeogeography, Palaeoclimatology, Palaeoecology 48, 263-284.

Dreimanis, A., 1967. Mastodons, their geologic age and extinction in Ontario, Canada. Canadian Journal of Earth Sciences 4, 663-675.

Dreimanis, A., Winder, C.G., Aaltonen, R.A., 1998. London Ontario: Geology, geomorphology, geodata. In: Karrow, P.F. and White, O.L. (eds.) Urban Geology of Canadian Cities. Geological Society of Canada Special Paper 42, 237-260.

Elias, S.A., 2013. Beetle records: Postglacial North America. In: Elias, S., (Ed.), Encyclopedia of Quaternary Science. 2nd edition, Amsterdam, Elsevier, 282-290.

Ellis, C.J., Carr, D.H., Loebel, T.J., 2011. The Younger Dryas and Late Pleistocene peoples of the Great Lakes region. Quaternary International 242, 534-545.

Gran, K.B., Belmont, P., Day, S.S., Finnegan, N., Jennings, C., Lauer, W.J., Wilcock, P.R., 2011. Landscape evolution in south-central Minnesota and the role of geomorphic history on modern erosional processes. GSA Today 21, 7-9.

Hack, J.T. 1965. Postglacial drainage evolution and stream geometry in the Ontonagon area, Michigan. USGS Professional Paper 504-B.

Kappesser, G., 2002. A riffle stability index to evaluate sediment loading to streams. Journal of the American Water Resources Association 38, 1069-1081.

Kaufmann, P.R., Faustini, J.M. Larsen, D.P. Shirazi, M.A., 2008. A roughness-corrected index of relative bed stability for regional stream surveys. Geomorphology 99, 150-170.

Kaufmann, P.R., Larsen, D.P., Faustini, J.M., 2009. Bed stability and sedimentation associated with human disturbances in Pacific Northwest streams. Journal of American Water Resources Association 45, 434-459.

Lisle, T.E. Hilton, S., 1992. The volume of fine sediment in pools: An index of the supply of mobile sediment in stream channels. Water Resources Bulletin 28, 371-383.

Lisle, T.E., Hilton, S., 1999. Fine bed material in pools of natural gravel-bed channels. Water Resources Research 35, 1291-1304.

Mackin, J.H., 1948. Concept of the graded river. Geological Society of America Bulletin 48, 463-512.

Mandrak, N.E., Crossman, E.J., 1992. Postglacial dispersal of freshwater fishes into Ontario. Candian Journal of Zoology 70, 2247-2259.

Montgomery, D.R., Buffington, J.M., 1997. Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin 109, 596-611.

Phillips, R.T.J., Robert, A., 2005. River incision in relation to post-glacial events in the Humber River basin, Ontario. Géographie physique et Quaternaire 59, 17-30.

Phillips, R.T.J., Desloges, J.R., 2014. Glacially conditioned specific stream powers in low-relief river catchments of the southern Laurentian Great Lakes. Geomorphology 206, 271-287.

Phillips, R.T.J., Desloges, J.R., 2015a. Glacial legacy effects on river landforms of the southern Laurentian Great Lakes. Journal of Great Lakes Research 41, 951-964.

Phillips, R.T.J., Desloges, J.R., 2015b. Alluvial floodplain classification by multivariate clustering and discriminant analysis for low-relief glacially conditioned river catchments. Earth Surface Processes and Landforms 40, 756-770.

Schottler, S.P., Ulrich, J., Belmont, P., Moore, R., Lauer, W.J., Engstrom, D.R., Almendinger, J.E., 2013. Twentieth century agricultural drainage creates more erosive rivers. Hydrological Processes 28, 1951-1961.

Stock, J.D., Montgomery, D.R., Collins, B.D., Dietrich, W.E., Sklar, L., 2005. Field measurements of incision rates following bedrock exposure: Implications for process controls on the long profiles of valleys cut by rivers and debris flows. Geological Society of America Bulletin 117, 174-194.

Tinkler, K.J., Parrish, J., 1998. Recent adjustments to the long profile of Cook Creek, an urbanized bedrock channel in Mississauga, Ontario. In: Tinkler, K.J., and Wohl, E.E., (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union Geophysical Monograph 107, Washington, D.C., 167-188.

Wohl, E. 2004. Limits of downstream hydraulic geometry. Geology 32, 897-900.

Wolman, M.G., Gerson, R., 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surface Processes and Landforms 3, 189-208.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54-74.

Wu, W., Altinakar, M.S., Al-Riffai, M., Bergman, N., Bradford, S.F., Cao, Z., Chen, Q.J., Constantinescu, S.G., Duan, J.G., Gee, M.D., Greimann, B., Hanson, G., He, Z., Hegedus, P., Van Hoestenberghe, T., Huddleston, D., Hughes, S.A., Imran, J., Jia, Y., Jorgeson, J.D., Kahawita, R., Klumpp, C.C., Lai, Y., Langendoen, E.J., Liu, S., Moreda, F., Morris, M., Morvan, H., Orendorff, B., Pak, J., Peeters, P., Reed, S., Sanders, B.F., Scott, S.H., Soares-Frazao, S., Song, C.R., Sutherland, J., Teal, M.J., Tsubaki, R., Wahl, T.L., Weston, D.M., Williams, D.T., Zech, Y., Zhang, L., 2011. ASCE/EWRI Task Committee on Dam/Levee Breaching. Earthen embankment breaching. Journal of Hydraulic Engineering 137, 1549-1564.

Yu, Z., 2000. Ecosystem response to Lateglacial and early Holocene climate oscillations in the Great Lakes region of North America. Quaternary Science Reviews 19, 1723-1747.

Curriculum Vitae

Nathaniel Bergman

Postsecondary Education

2016 – PhD., Department of Geography, Western University, London, Ontario Canada.

2010 – Research Assistant, Department of Watershed Sciences, Utah State University, Logan, Utah, U.S.A.

2007 - M.A., Department of Geography and Environmental Development, Ben Gurion University of the Negev, Beer-Sheva, Israel.

1998 - B.A., Department of Geography and the Department of Political Science, University of Haifa, Haifa, Israel.

Grants and Awards

January 2016 – Israel Ministry of Science, Technology and Space – Natural Sciences Research grant ($345,000).

May 2014 – PSAC 610 (University of Western Ontario TAs and postdocs union) - Academic Achievement Scholarship ($500).

April 2015 – Faculty of Social Sciences Dean Graduate Research Award ($750).

April 2014 – Faculty of Social Sciences Dean Graduate Research Award ($750).

March 2013 – Graduate Teaching Assistants (GTA) Union - Scholarship for Outstanding Research Contributions ($500).

September 2012 – Best poster award Association of State Dam Safety Officials (ASDSO) ($250).

Fields of Research Interests

- Quaternary geology and geomorphology
- Sediment transport in gravel-bed streams
- Hydro-ecology
- Using water resources in conjunction with natural requirements
- Stream restoration
- Wetland hydrology and sedimentology
- Lakes and reservoirs as sediment sinks
- Catastrophic flooding and paleofloods as tools for predicting future flood hazards and climate change
- Anthropogenic effect on the riverine environment