Influences of Confluences on Reach Scale Morphology of Southern Ontario Stream Channels

By

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Graduate Department of Geography

The University of Toronto

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2013

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**Abstract**

Downstream adjustment in stream channel morphology is examined in the context of stream channel confluences. Stream channel confluences represent areas of point specific increases in discharge, flow energy and potential erosion in a river system which will in turn affect the post-confluence downstream morphology. Analysis of 12 confluence junctions from southern Ontario streams, constituting 36 channel reaches in total, show an internally consistent hydraulic geometry relationship but with specific controls on channel morphology related to boundary conditions. Predictions of mainstem morphologies is possible using tributary attributes but reach specific channel confinement and material type add significant influence.
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Compiled Channel Reach Characteristics
1. Introduction

1.1. Defining the Problem

It is commonly understood for natural alluvial river systems the downstream adjustment in channel dimensions and morphological form (e.g. width and depth) will be controlled by sediment load, discharge, bed material grain size and gradient (Knighton, 1987). An increase in discharge in the downstream direction as drainage area increases should produce predictable changes in channel slope and median grain size (Charlton, 2010). However, these empirical relations are complicated when local variations in the ability of a channel to do erosive and transportational work are impeded. Further complicating the empirical relations is the potential for local changes in channel boundary composition between channel reaches, so that the maintenance of channel form for one boundary type may not apply to another (Knighton, 1987).

Downstream adjustment along a river channel is regarded as intermittent due to the inputs (i.e. discharge and sediment) from tributaries. Therefore, stream channel confluences represent areas of point specific increases in discharge, flow energy and potential erosion in a river system making confluences unique locations for studying the potential for changes in downstream river channel morphology. It is expected that the channel morphology and sedimentology downstream from confluences should respond abruptly to the rapid change in hydrology and hydraulic conditions as two tributaries join. For example, tributary inputs have a tendency to alter the bed material composition downstream from the confluence by introducing coarser bed material which in turn alters the bed forms and sediment transport along the downstream mainstem reach (Knighton, 1980).
So for southern Ontario, a region that is heavily influenced by glacially inherited landforms (Chapman and Putnam, 1984; Sharpe et al., 1997), the downstream adjustment in channel form is subject to multiple influences along the longitudinal profile.

1.2. Research Questions

The research presented within this thesis aims to address the following three questions in formerly glaciated landscape of southern Ontario:

1. How do tributaries influence post-confluence mainstem channel morphology?
2. To what extent do boundary conditions effects channel morphology?
3. To what extent can post-confluence mainstem channel morphology be predicted?

These questions are imbedded in the concept that ‘adjusted’ rivers should show consistent relationships amongst a number of morphologic parameters (e.g. width and depth) and the factors that influence them (e.g. material strength, slope, channel confinement, etc.).
2. Downstream Changes in Channel Morphology

2.1. Controls on Stream Channel Morphology

Variations in river channel morphology are inherently controlled by discharges that are high enough to induce sediment transport for a particular grain size along a river reach with a particular gradient. Over geologic time, the landscape in which channels form can be influenced by numerous processes including tectonic uplift, landscape erosion and climate change that can alter the discharge and sediment characteristics of a river channel. Over historical time scales, changes in discharge, sediment supply and land use alterations can influence the changing morphologies of stream channels (Montgomery and Buffington, 1998). Major changes can become apparent over decadal to century time scales and can influence both local reach characteristics as well as systematic downstream alterations.

A convenient reference point for assessing the degree of channel change is the equilibrium channel. A channel is in equilibrium, or is a graded system, when no aggradation or degradation of the channel bed is occurring. This will happen when the channel has adjusted its slope in accordance with the available discharge to transport the sediment load supplied to it from upstream (Makin, 1948). This equilibrium state may persist over a period of several years to decades and signifies a balance between the sediment load entering through the upstream channel reach and the load that is carried out of the reach. Equilibrium river form, in theory, could span a much longer time but because of isostatic adjustments of southern Ontario following deglaciation, at best a dynamic equilibrium might occur where sometimes stable steady-state channel forms evolve under progressive entrenchment as the landscape rises.
differentially. However, channel equilibrium may only exist briefly or not at all. This can occur when natural flows vary significantly over time so that, for example, higher flows that are actually capable of transporting the sediment occur infrequently within the system (Lane, 1954). In some instances the channel bed can be lowered in elevation during high flows or can be filled as flows lessen and the transport capacity of the channel is reduced, both have an effect on channel gradient.

If there is a change in conditions, such as discharge or sediment supply at a river channel cross-section that is in equilibrium, predictable changes in the downstream morphology of the channel can be made. These predictions originally derived from Lane’s (1954) empirical relationship:

\[ Q_s D \propto Q_w S \]  

where \( Q_s \) is the sediment discharge, \( D \) is the diameter of the characteristic sediment size along the intermediate particle axis, \( Q_w \) is the water discharge, and \( S \) is the channel slope. If this balance is observed, channel characteristics define the stability and equilibrium of the cross-section or channel reach. Adjustments in the four variables in the Lane relation can occur slowly as a result of gradual changes to the system (e.g. geologic time scale) or they can occur suddenly (human time scales) shifting the channel out of equilibrium into and aggrading or degrading system (Makin, 1948).

In conjunction with Lane’s (1954) relation, it is also imperative to consider the transporting power of the channel in terms of velocity or shear stress and stream power. Velocity will vary as a result of changing channel slope and discharge, and with any change in the way energy is dissipated within the system either as internal or external frictional losses. Internal friction losses result from energy dissipation within a turbulent current whereas external fiction losses
result from boundary friction along the wetted perimeter. The transporting competence or stream power of a channel will directly affect how much sediment can be routed through the system over a given period of time and is a function of both discharge and slope (Makin, 1948). On shorter time scales and on the more spatially focused river reach scales lateral stability, as opposed to vertical stability (equilibrium), is more dependent on local and regional sediment supply and the stream power available to transport the sediments in the channel reach. So the adjustment of channel form as defined by lateral stability might be most appropriately tied these more restricted time and space scales.

2.1.1. **Discharge**

For a stream channel to be in equilibrium it must be adjusted to the range of discharges that transport the majority of the annual sediment load (Andrews, 1980). Day to day fluctuations in the flow stage of river channels is dependent on a variety of conditions such as climatic setting, seasonal changes in precipitation and local weather conditions. There are several ways to define discharge thresholds in a way that is significant to characterize channel forming flows. Critical discharge, effective discharge, dominant discharge, and bankfull discharge are all common terms in the literature, and all relate to different aspects of channel forming flow.

**Critical Discharge**

Critical discharge refers to the discharge that initiates sediment transportation within the channel. This stage is important, especially in gravel-bed channels, since bed material is motionless and immobile during low flow conditions (Ferguson, 1994). Critical flow conditions are often predicted using the Shields equation:

\[
T_c^* = \frac{T_c}{(\rho_s - \rho) g D_i}
\]  

[2.2]
Were $T^*_{c}$ represents the critical dimensionless shear stress, $T_c$ is the dimensionless shear stress, $\rho_s$ and $\rho$ are the sediment and water densities respectively, $g$ is the acceleration due to gravity and $D_i$ is the particle diameter. In channels with uniform beds, the critical dimensionless shear stress is considered to be between 0.05 and 0.06 when the bed particles are greater than 0.1 mm (Ferguson 1994). Observed shear stress can be calculated as:

$$T = \rho g d S$$  \[2.3\]

where $d$ is the channel depth and $S$ is the slope. However, for channels with non-uniform bed materials, the relationship between the various particle sizes changes because of hiding and protrusion effects. Therefore the critical shear stress to entrain a particle of diameter $D_i$ on a bed with a median grain size of $D_{50}$ depends on its relative size $D_i/D_{50}$. A new critical flow equation was given by Andrews (1983) to incorporate the influence of relative particle size:

$$T^*_{cl} = T^*_{c50} \left(\frac{D_i}{D_{50}}\right)^{-x}$$  \[2.4\]

Critical shear stress as a measure of critical discharge is often used depending on available measured variables, but is not necessarily a superior method of indicating critical discharge. Bathurst (1987) proposed a critical discharge equation that uses the same relative size of $D_i/D_{50}$ for steep narrow channels, but relies only on unit discharge data where:

$$q_{cl} = q_{c50} \left(\frac{D_i}{D_{50}}\right)^b$$  \[2.5\]

where $q_{cl}$ is the critical unit discharge for the movement of particle $D_i$, $q_{c50}$ is the critical unit discharge for the reference particle size $D_{50}$ which in turn can be calculated by:

$$q_{c50} = 0.15 g^{0.5} D^{1.5} S^{-1.12}$$  \[2.6\]
Effective and Dominant Discharge

Effective discharge and dominant discharge can be used interchangeably as they both refer to the discharge that transports the largest fraction of the annual sediment load under equilibrium conditions (Carling, 1988; Andrews, 1980; Emmett and Wolman, 2001). The effectiveness of a given discharge over a period of years is determined by flow magnitude and frequency (Figure 2.1). Sediment transport rate is illustrated by Curve A in Figure 2.1, and is used to express the exponential increase of mass transported with increasing discharge. Sediment transport rate \((I_b)\) is proportionally related to discharge to the power \(n\), where:

\[
(I_b \propto Q^n)
\]  

[2.7]

As discharge increases there will be a doubling of the sediment transport rate. Curve B in Figure 2.1 represents the frequency to which discharge varies over days to years. Moderate flows carrying smaller amounts of sediment are the most frequent and the largest flows, which carry the very high sediment loads, are rare. The product of the sediment transport rate and frequency of flow occurrence represented by Curve C in Figure 2.1 shows the relative effectiveness of a particular discharge. As illustrated in Figure 2.1, there is a range of intermediate discharges that transport the largest portion of the annual sediment load. This range is called the effective discharge, and is represented by the top peak of Curve C (Andrews, 1980).
Figure 2.1: Relationship between discharge and sediment transport rate, frequency of occurrence and the product of frequency and transport rate (from Wolman and Miller, 1960; Andrews, 1980).

Bankfull Discharge

Bankfull discharge refers to channel flow that is held within the banks of a river cross-section at a level just lower than the floodplain and where the flow magnitude is considered to be connected and influential to the overall morphology of the channel (Navratil, et al., 2006). This discharge is used as a means of comparing the morphology of river reaches and is highly correlated to effective discharge (Wolman and Miller, 1960). There are many different methods for determining bankfull discharge ($Q_b$) as outlined in Williams (1978), and researchers must recognize these when taking field measurements. In fluvial geomorphic research, the height of
the active floodplain has been adopted as the preferred indicator of bankfull conditions which corresponds to a break in slope at the top of bank along the channel cross-section (Navratil, et al., 2006).

Although bankfull level may be relatively easy to identify along a cross-section using indicators such as top of bank and slope inflections (indicated as ToB and BI respectively in Figure 2.2), bankfull discharge is best determined from several cross-sections at a reach scale where approximately 15 bankfull channel widths define a reach. As numerous studies have shown there is a wide range of variability in bankfull level at cross-sections within a reach (Williams, 1987). As well, bankfull discharge generally increases downstream, so there is a relationship between bankfull discharge and drainage area (Andrews, 1980). The flow frequency that defines bankfull conditions is dependant on the river system in question. Wolman and Leopold (1975) indicate that a return period of 1 to 2 years is common for channel in a humid temperate climate. While Williams (1978) indicates that return periods for channel capacity conditions, not necessarily bankfull, range from 1 to 32 years.
2.2. Predicting Downstream Variation in Channel Pattern and Morphology

2.2.1. Changes in Downstream Channel Morphology

The hydraulic characteristics of a river channel (width, depth and velocity), have been related to the downstream increase in discharge (Leopold and Maddock, 1953). While regional climate and geology have some influence, the downstream hydraulic geometry relations have shown remarkable consistency. In this study the hydraulic relations of tributaries are compared to the downstream mainstem of the river to determine the influence of abrupt discharge changes at tributary junctions. Empirical relationships between a representative discharge and the hydraulic characteristics can be expressed by the power relationships:
\[ w = aQ^b \]
\[ d = cQ^f \]
\[ v = kQ^m \]

where \( Q \) is discharge, \( w, d, \) and \( v \) are water-surface width, average depth and average velocity respectively. For which:

\[ b + f + m = 1.0 \]  \[2.9\]

and

\[ a \times c \times k = 1.0 \]  \[2.10\]

Leopold and Maddock (1953) noted that in the downstream direction, the majority of channel adjustment to a representative discharge is taken up almost entirely by width and depth \((b = 0.4 \) and \( f = 0.5)\), and that for both cross-sectional and downstream relations width, depth and velocity will increase with discharge.

### 2.2.2. Predicting Channel Patterns

Early research on predicting channel plan-form morphology correlated channel slope to bankfull discharge (Lane, 1954; Leopold and Wolman 1957). Lane (1954) proposed a proportional relationship outlined previously (Eq. 2.1) that indicated channel morphology is controlled by the relationship between sediment discharge, grain size, water discharge and slope. Leopold and Wolman (1957), were the first to propose a discriminating relation based on slope and bankfull discharge that would indicate the transition between braided channel morphology and meandering alluvial channels. The slope of the discriminating line is;
\[ S = 0.06Q_b^{-0.44} \]  

where \( S \) is the channel slope and \( Q_b \) is bankfull discharge. This relationship was used as the standard for predicting channel patterns for decades until the importance of median grain size \( (D_{50}) \) to the transition into braiding became more evident. Ferguson (1987) showed that with increasing median grain size the slope of the discriminating line would also increase.

Further investigations also indicate that the transition between braiding and meandering channel patterns are less dependent on bankfull discharge alone and more related to stream power. Van den Berg (1995) expressed potential specific stream power \( (\omega_v) \) for gravel bed rivers as:

\[ \omega_v = 3.3 S \bar{Q}_b \]  

The potential specific stream power for bankfull conditions is plotted against the median grain size \( (D_{50}) \) in order to define a stable channel vs. unstable channel discriminating threshold line:

\[ \omega_{v,bf} = 900 D_{50}^{0.42} \]  

Potential specific stream power \( (\omega_v) \) is calculated by:

\[ \omega_v = \rho gQS/w \]  

Van den Berg (1995) discovered that the discriminating line, expressed in equation 2.13 and displayed in Figure 2.3, accurately predicted the threshold between braided and single thread rivers with a sinuosity of \( P > 1.5 \).
Further revision of van den Berg (1995), and the recognition that channel patterns form a continuum rather than a distinct types (Ferguson, 1987), was provided by Bledsoe and Watson (2001). They expressed the discriminating line from equation 2.13 as the 50% probability of braiding when the median grain size is between 0.1 and 100 mm. Kleinhans and van den Berg (2011) further expanded upon equation 2.13 by predicting four stability fields based on channel bar patterns based on natural rivers in unconfined alluvium that were experiencing slow changing equilibrium conditions. The new threshold lines distinguish between channels that are stable and single thread, meandering with scrolls, moderately braided-meandering with scrolls

**Figure 2.3:** Channel pattern in relation to median grain size ($D_{50}$) and potential specific stream power calculated using bankfull discharge (from van den Berg, 1995).
and chutes and highly braided (Figure 2.4). The discriminating line A in Figure 2.4 is the same as equation 2.13, and lines B and C are stream powers of $\omega = 285D_{50}^{0.42}$ and $\omega = 90D_{50}^{0.42}$, respectively.

Incorporating a differentiation between meandering, multiple-thread anabranching and braided channels, Eaton et al. (2010) produced thresholds that relate the critical slope associated with a change in channel pattern to dimensionless discharge and relative bank strength. Bank strength is thought to be an important variable having a strong influence on channel geometry (Eaton and Giles, 2009).

**Figure 2.4**: Patterns of equilibrium alluvial rivers plotted with the potential specific stream power related to valley gradient and predicted width where the data is subdivided by bar pattern (after Kleinhans and van den Berg, 2011).
For a channel where relative bank strength ($\mu'$) does not change with channel size (i.e. in the downstream direction), $\mu'$ can be incorporated into the $S \propto Q$, hydraulic geometry power, relation (Millar, 2005). Eaton et al. (2010) proposed a threshold to define the onset of anabranching as the formation of mid channel bars:

$$S^* = 0.40\mu'1.41Q^{*-0.43}$$  \hspace{1cm} [2.15]

where $S^*$ is critical slope, $Q^*$ is dimensionless discharge, and $\mu'$ is the dimensionless relative bank strength given by the ratio of the critical shear stress for entrainment of the channel banks to the critical shear stress for the channel bed. When the bed and banks are comprised of similar material $\mu'$ can be set to a value of 1. After the formation of mid channel bars, channels can become stable by dividing into $N$ anabranches which effectively reduces $Q^*$ for each bifurcated channel such that any further mid-channel bar growth is blocked and the system becomes a series of interwoven but stable single channels. The subsequent threshold between stable anabranching channels, with fewer than four anabranches, and fully braided channels is developed by Eaton et al. (2010) is expressed as:

$$S^* = 0.72\mu'1.41Q^{*-0.43}$$  \hspace{1cm} [2.16]

### 2.3. Stream Channel Confluences

Stream channel confluences represent areas of point specific increases in discharge, flow energy and potential enhanced erosion in a river system to which the downstream main-stem channels must adjust accordingly. Confluences are marked by complex flow patterns and sediment transport that lead to the development of specific bed and bank morphologies along the
downstream channel that can have profound impacts not only on the geomorphology of the stream but on the stream ecology as well (Rice, et al., 2001).

2.3.1. Confluence Morphology

Previous research into stream channel morphology has noted several common characteristics that develop at these locations including; avalanche faces at the mouth of each tributary, a scour hole, a tributary-mouth bar, sediment accumulation along the upstream confluence corner, downstream mid-channel bars, and downstream lateral bars (Figure 2.5) (Rhoads, et. al., 2009; Best and Rhoads, 2008; Kenworthy and Rhoads, 1995; Best, 1986). These features are widely dependent on time and space and they will vary in magnitude depending on the discharge received from each contributing tributary channel.

Two key characteristics that largely control confluence morphology are the junction angle ($\alpha$), and the ratio of discharge between the two contributing tributaries (Figure 2.5). These features will have the greatest impact on the location of the mixing interface and shear layer of the combining flows and ultimately the location and size (width and depth) of the scour hole.
2.3.2. Flow Dynamics and Sediment Transport at Confluences

The complex flow structures observed at confluences depends on the junction angle, the planform symmetry of the confluence and the momentum flux ratio ($M_r$) of the incoming flows (Mosley, 1976). In particular the path of the shear layer, or the orientation of the two combining tributary discharges, is controlled by the ratio of discharges of the tributaries and their relative momentums. As seen in Figure 2.6a, the location of the shear layer will fluctuate between the downstream banks depending on which tributary has a greater discharge. As the two tributary flows converge within the confluence, the relative momentum ratio between them will alter the position of the shear layer and affect the degree of cross-channel sediment mixing. For instance, assuming a confluence where the relative widths of both tributaries are similar, but the discharge differs; the position of the shear layer will be deflected across the main channel (Figure 2.5). As
the momentum ratio increases in favor of one tributary, the deflection across the other tributary becomes greater and vice versa (Kenworthy and Rhoads, 1995).

Apart from the highly turbulent shear layer within the confluence, flow separation, flow acceleration and flow stagnation are observed. This zone has been termed the confluence hydrodynamic zone (CHZ). The spatial extent of the CHZ corresponds to the distance downstream over which the combined flow is influenced by hydraulic pressure gradients connected with the convergence and realignment of the combining tributary flows (Kenworthy and Rhoads, 1995). The downstream boundary of the CHZ is marked by the gradual divergence and deceleration of flow (Figure 2.5).

Erosion along the shear layer producing a scour hole within the confluence is associated with turbulent and helicoidal flow cells generated by the converging flows. The depth and cross-sectional area of the scour hole increases with increasing turbulence and maximum size is reached when both contributing flows are equal (Mosley, 1976). Confluence flow surveys conducted by Roy et al. (1988) concluded that flow through the confluence is concentrated through the scour, and is accelerated as discharge rises so that at bankfull conditions velocity within the confluence is 1.6 times higher than either tributary. Conversely, at low flow stages there is a distinct loss of flow momentum as tributary flows converge at the confluence and the scour hole acts as a storage location for both water and sediment (Best, 1986).
Figure 2.6: Conceptual model of relationships between hydrologic inputs and spatial patterns of suspended sediment concentration at stream confluences. (a) Variation in the path of the shear layer as momentum ratio ($M_r$) increases. Size of arrows is proportional to momentum fluxes (i.e. discharge) of incoming flows, $C_1$ and $C_2$ refer to mean suspended sediment concentrations of these flows. (b) Idealized cross-channel patterns of suspended sediment at a cross-section (A-A') immediately downstream of a confluence for various combinations of momentum ratio and sediment concentration ratio ($S_r$). Degree of shading indicates relative sediment concentration (from Kenworthy and Rhoads, 1995).

A detailed analysis of near-bed flow patterns conducted by Boyer et al. (2006) revealed that within the shear layer throughout the confluence, the turbulence generated at bankfull stage is associated with intense upward flow movements contributing to high sediment transport rates through the confluence scour thereby enhancing erosion. Turbulence is also intensified by discordant bed heights between the tributaries.

Sediment particles being transported through the confluence follow a distinct pattern that depends upon the separation of flow between each tributary as they enter the confluence.
Routing of the sediment plays a major influence on bed morphology downstream along the combined channel (Best, 1988). Roy and Bergeron (1990) noted that particle pathways depend on timing within the season and flow stage. During a particle seeding experiment, Roy and Bergeron (1990) monitored the progression of several particles, corresponding to $D_{16}$, $D_{50}$ and $D_{64}$, as they moved through the confluence over the course of one spring to fall season. They found that particle transport through the confluence shows a dual pattern. From spring to early summer the path of the particles was mostly controlled by the local bed gradient and the scour hole, where sediment mixing was common. From late summer to fall, the particles moved parallel to the banks and rarely mixed. These particle pathways correspond to the distribution of flow velocity vectors at various flow stages, where, as stage increases the vectors become aligned with the scour hole.

The flow dynamic and sediment pathways through the confluence are important for the development of the confluence bed morphology. Rhoads et al. (2009) showed that for high discharge ratios ($Q_r > 1$), regardless of momentum, where the main tributary is hydrologically dominant, the bed morphology within the confluence does not change dramatically downstream despite sediment transport and erosion and deposition occurring. However, when low discharge ratios occur ($Q_r < 1$) and the lesser tributary becomes more dominant, major changes to confluence morphology occur. The changes that occur include a realignment and enlargement of the scour hole through the center of the channel and significant erosion of the tributary mouth bar along the inner bank. These changes create a near symmetrical channel profile in the downstream combined channel that lasts for approximately one channel width. These changes are reverted once the discharge ratio becomes higher and the main tributary becomes hydrologically dominant.
In addition to complex flow dynamics throughout the confluence, there are distinct patterns in grain size that contribute to bed morphology. Within the zone of flow stagnation at the upstream confluence corner (see Figure 2.5) there is a marked decrease in particle size. The scour hole is marked by coarse grained particles. The upstream junction corner is marked by a rapid decrease, followed by a gradual increase downstream, in grain size along the downstream tributary mouth bar within the flow stagnation zone (see Figure 2.5) (Best, 1988). The distribution of grain size within the confluence will depend on the sediment supply from each tributary as well as their hydrologic characteristics (Rhoads et al., 2009; Biron et al., 1993; Roy and Bergeron, 1990).

2.3.3. Downstream Main-stem Adjustment

Changes in channel geometry downstream from a confluence have been investigated by numerous authors. Miller (1958), studied mountain streams to produce a method for determining the change in channel width downstream of the confluence:

\[ w_c = p(w_{MT} + w_T) \]  

Where \( w_c \), \( w_{MT} \) and \( w_T \) are the widths of the combined mainstem channel, the main tributary and the lesser tributary respectively. The \( p \) value is a function of branching symmetry and should fall between 0.5 and 1.0. This equation can also be applied to other geometry characteristics such as channel depth, cross-sectional area, slope and grain size.

Richards (1980) attempted to improve upon Miller’s equation by estimating downstream width changes as a ratio of channel discharge magnitudes \( n \):
where \( n_C \) and \( n_{MT} \) are the magnitudes of the combined downstream channel and main tributary respectively and \( k_2 \) is equal to 0.6. This model can be applied to confluences with both symmetrical and asymmetrical planform geometries, however it does not take into account the width or discharge of the minor tributary. By ignoring the influence of the minor tributary Roy and Woldenberg (1986) noted that a significant amount of variation in downstream characteristics was left unexplained.

Instead of finding an alternative to the classic hydraulic geometry relationships to predict post confluence changes, (cf. Richard, 1980), Roy and Woldenberg (1986) transformed the hydraulic geometry relationships into a continuity of flow equation so that a downstream variable such as width can be predicted by:

\[
 w = w_{MT}^{x} + w_{T}^{x} 
\]  
[2.19]

For this equation width \( (w) \) can be substituted for any other hydraulic geometry variable (depth, velocity, slope, cross-section area). The exponent \( x \) represents the hydraulic geometry exponent related to the variable in question for an individual confluence, and represents an average for the entire system. The exponent \( x \) can be calculated as the reciprocal of exponent \( x \).

The example outlined by Roy and Woldenberg (1986) uses data on confluences collected for one river system by Richards (1977 and 1980) where the hydraulic geometry exponent \( (x) \) for width averaged between riffles and pools is 0.34 and \( x \) is therefore equal to 2.94. These exponents were substituted into Equation 2.19 and the resulting predicted downstream widths
were compared with recorded widths for the River Fowey (Figure 2.7). A similar approach will be used in this thesis.

**Figure 2.7:** Comparison of expected and observed widths for the River Fowey (England) (from Roy and Woldenberg, 1986)
3. Introduction to Study Sites

Study sites were selected based on a variety of criteria including tributary size ratio. A number of co-dominant tributaries where bankfull stage is proportional were selected as well as a number of sites where one tributary is obviously more dominant. The ratio between the lesser tributary and the main tributary average between 1:1, and 1:3, with a max of 1:5 along the Ausable River. Other factors explored include differences in slope between the tributaries as well as variations in river channel boundary conditions such as glacial and alluvial bed and banks. Based on boundary type categorization, confluence sites in this study can be separated into three groups; alluvial (Type 1), semi-alluvial (Type 2), and non-alluvial (Type 3). These categories are assigned based on the extent of non-alluvial conditions, such as glacial sediments or bedrock, which is present within the beds and banks of the study reaches. Accessibility was also a factor in choosing each study site. Potential study sites were reviewed using Google Earth Imagery (2012) as well as preliminary reconnaissance and consultation with researchers who had previously visited the sites. Twelve study sites were chosen that best represent a range of conditions that are thought to be typical of a small watershed in southern Ontario.

Study sites are mostly in south central Ontario with the exception of one in the south west (Figure 3.1). Table 3.1 is a summary of the 12 confluences sites which includes: Fourteen Mile Creek, Sixteen Mile Creek, Ausable River, Catfish Creek, Credit River, Don River, Duffins Creek, Humber River and the Mad River. In the cases of the Ausable, Credit and Humber rivers, multiple confluence locations were selected throughout these watersheds. At each confluence location a reach was defined to extend at least ten channel bankfull widths upstream of the
confluence along each of the tributaries and ten bankfull channel widths downstream from the confluence. Each site consisted of three channel reaches and therefore 36 study reaches in total.
Figure 3.1 Map of southern Ontario, Canada displaying the general location of study confluences as well as other Ontario drainage networks and the underlying physiography. Numbers correspond with site descriptions on Table 3.1. Surficial geology is from Chapmand and Putnam (1984) and shows the dominant material/terrain types.
Table 3.1: Description of each study confluence.

<table>
<thead>
<tr>
<th>#</th>
<th>Study Site</th>
<th>Confluence Latitude, Longitude</th>
<th>River Kilometer</th>
<th>Channel</th>
<th>Average Width (m)</th>
<th>Average Depth (m)</th>
<th>Slope (m/m)</th>
<th>Bankfull Discharge (m³/s)</th>
<th>Surficial Geology</th>
<th>Channel Pattern (Reach Scale)</th>
<th>Alluvial Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Ausable River</td>
<td>43°05'40&quot;N, 81°48'48&quot;W</td>
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<td>Ausable River</td>
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<td>2</td>
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<td></td>
<td></td>
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<td></td>
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<td>0.0024</td>
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<tr>
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<td>Ausable River</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Ausable River</td>
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<td>164.21</td>
<td>Till Moraine</td>
<td>Mod. Meandering with lateral bars and chutes</td>
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<tr>
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1Table includes: the latitude and longitude of each confluence site, river kilometer upstream from the river watershed outlet with one of the Great Lakes, channel names, average width and depth for bankfull conditions, water surface slope, bankfull discharge, surficial geology of each channel based on Chapman and Putnam (1984) designations, observed channel pattern at reach scale and the alluvial type grouping which ranges from 1-alluvial, 2-semi-alluvial and 3-non-alluvial.
Table 3.1: continued.

<table>
<thead>
<tr>
<th>#</th>
<th>Study Site</th>
<th>Confluence Latitude, Longitude</th>
<th>River Kilometer</th>
<th>Channel</th>
<th>Average Width (m)</th>
<th>Average Depth (m)</th>
<th>Slope (m/m)</th>
<th>Bankfull Discharge (m³/s)</th>
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<td>Till Plain, bedrock</td>
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<td>2</td>
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Table includes: the latitude and longitude of each confluence site, river kilometer upstream from the river watershed outlet with one of the Great Lakes, channel names, average width and depth for bankfull conditions, water surface slope, bankfull discharge, surficial geology of each channel based on Chapman and Putnam (1984) designations, observed channel pattern at reach scale and the alluvial type grouping which ranges from 1-alluvial, 2-semi-alluvial and 3-non-alluvial.
Table 3.1: continued.

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<th>#</th>
<th>Study Site</th>
<th>Confluence Latitude, Longitude</th>
<th>River Kilometer</th>
<th>Channel</th>
<th>Average Width (m)</th>
<th>Average Depth (m)</th>
<th>Slope (m/m)</th>
<th>Bankfull Discharge (m$^3$/s)</th>
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<th>Channel Pattern (Reach Scale)</th>
<th>Alluvial Type</th>
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<td>Mod. Meandering with lateral bars</td>
<td>2</td>
</tr>
<tr>
<td>12</td>
<td>Sixteen Mile Creek</td>
<td>43°28'24&quot;N, 79°47'26&quot;W</td>
<td>15.5</td>
<td>16 Mile East</td>
<td>15.7</td>
<td>0.7</td>
<td>0.0062</td>
<td>24.72</td>
<td>Till Moraine, bedrock</td>
<td>Meandering with chutes</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>16 Mile West</td>
<td>27.8</td>
<td>0.68</td>
<td>0.0074</td>
<td>19.12</td>
<td>Till Moraine, bedrock</td>
<td>Meandering with chutes, stable islands</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>16 Mile Creek</td>
<td>27</td>
<td>0.95</td>
<td>0.0046</td>
<td>42.7</td>
<td>Till Moraine, bedrock</td>
<td>Meandering with chutes, stable islands</td>
<td>3</td>
</tr>
</tbody>
</table>

Table includes: the latitude and longitude of each confluence site, river kilometer upstream from the river watershed outlet with one of the Great Lakes, channel names, average width and depth for bankfull conditions, water surface slope, bankfull discharge, surficial geology of each channel based on Chapman and Putnam (1984) designations, observed channel pattern at reach scale and the alluvial type grouping which ranges from 1-alluvial, 2-semi-alluvial and 3-non-alluvial.
3.1. Geography and Geomorphic Context

The southern Ontario landscape, and in particular southern Ontario fluvial landscapes, have been influenced by repeated glaciations, post-glacial isostatic uplift, as well as Holocene climate change and geomorphic erosion. Holocene adjustments to inherited glacial sediments and landforms continues today, however the tempo and magnitude of these adjustments are controlled by Paleozoic bedrock and a sequence of Quaternary overburden deposits which are dominated by tills, outwash and glaciallacustrine sediments that underlay all of the watersheds in this study (see Figure 3.1). Adjustments within the study site watersheds are also controlled by the base levels of the lakes to which they eventually drain. The watersheds drain into Georgian Bay, Lake Ontario, Lake Erie and Lake Huron (Chapman and Putnam, 1984) (Figure 3.1).

The last major glacial advance ended about 18 ka with the Late Wisconsin glaciation. Followed by the Laurentide Ice Sheet withdrawal into local sites of origin: Lake Huron, Georgian Bay, Lake Erie and Lake Ontario basins by 14 – 13 ka. The ice sheet deposited significant amounts of till (Anderson, et al., 2007). These lodgment tills, particularly Halton Till, dominate most of the larger watershed areas within this study and are either overlain by, or adjacent to, thick outwash deposits of sand and gravel or thin glaciallacustrine clay plains. Occasionally, shale or limestone bedrock outcrops within the river channel bottoms or along the banks.

River confluence reaches within this study that are controlled by bedrock are believed to be Holocene (10 ka – present) landscape features that are laterally stable with minimal migration due to erosion over century time scales. River confluence reaches that have formed within till and outwash deposits, forming floodplains and a self-forming alluvial planform, are expected to have migrated due to erosion and deposition over the same time scales. However, the alluvial
reaches are often entrenched within the floodplain having down-cut through the glacial sediments. Entrenched reaches are considered to be laterally stable in a similar manner to bedrock reaches so that there is very little or no migration due to non-erodible bed and banks.

3.2. Climate and Hydrology

The climate in southern Ontario is generally temperate with mild winters and warm summers. However there are distinct variations in climate depending on local topography and proximity to the Great Lakes as well as the frequency and strength of weather systems that cross the area (Brown, et al., 1980). Precipitation is evenly distributed throughout the year with snowfall in the winter and rain in the spring, summer and fall. Location and topography come into play with the amount of precipitation a region will receive, with the greatest amount occurring on the lee of Lake Huron and Georgian Bay where elevations reach 450 meters above sea level and lowest levels of precipitation occurring on the eastern shores of Lake St. Clair and northern shores of Lake Erie as well as below the Niagara Escarpment from Hamilton to Toronto (Singer, et al., 2003).

Meso-scale climates around southern Ontario can vary between watersheds as well as within a watershed as elevation and topography take effect. For instance, depending on the season the higher elevation of the Oak Ridges Moraine affects the precipitation patterns of the Credit, Humber and Don River watersheds’ creating greater amounts of precipitation in the upper reaches of the respective watersheds (Barrett, 2008). The Oak Ridges Moraine causes approximately 100 mm more of mean annual precipitation to in the upper reaches of the Humber River, Credit River and Don River watersheds (Barrett, 2008; TRCA, 2009; CVC, 2007).
Temperature also varies slightly around southern Ontario and within each watershed. Mean annual temperature is 7° Celsius, but there are slightly higher temperatures in south-western Ontario and cooler temperatures on the south shore of Hudson Bay (Brown, et al., 1980). Proximity to the Great Lakes can also influence the temperature within a watershed. Since the Great Lakes are so large, they act as a moderating influence on the temperature of the air masses that move across them. This allows air temperature to be cooler by the shore in summer and slightly warmer in the winter. This effect can be seen for up to ten kilometers away from the shore (TRCA, 2002).

Annual runoff for Ontario follows the precipitation patterns with high runoff occurring in the south-west and low runoff occurring in the north-east (Environment Canada, 2013). Peak flows for rivers in this study are displayed in Table 3.2, where peak flows correspond to spring snow melt.
Table 3.2: Discharge data for study site rivers for 2010 (Water Survey of Canada, 2013).

<table>
<thead>
<tr>
<th>River and Gauge ID</th>
<th>Mean Daily Discharge Aug-Feb (m³/s)</th>
<th>Mean Daily Discharge Mar-Jul (m³/s)</th>
<th>Max Instantaneous Discharge (m³/s)</th>
<th>Min Instantaneous Discharge (m³/s)</th>
<th>Average Daily Discharge (m³/s)</th>
<th>Total Yearly Discharge (m³/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ausable River (02FF010)</td>
<td>1.42</td>
<td>1.52</td>
<td>4.42 - Mar</td>
<td>1.53 - Feb</td>
<td>no data</td>
<td>no data</td>
</tr>
<tr>
<td>Catfish Creek (02GC018)</td>
<td>1.43</td>
<td>2.58</td>
<td>51.3 - Apr</td>
<td>0.107 - Sept</td>
<td>2.17</td>
<td>792.50</td>
</tr>
<tr>
<td>Credit River (02HB025)</td>
<td>4.81</td>
<td>9.26</td>
<td>72.3 - Mar</td>
<td>4.5 - Feb</td>
<td>6.68</td>
<td>2,439.37</td>
</tr>
<tr>
<td>Don River (02HC024)</td>
<td>3.64</td>
<td>5.00</td>
<td>40.70 - Aug</td>
<td>1.30 - May</td>
<td>4.22</td>
<td>1,540.43</td>
</tr>
<tr>
<td>Duffins Creek (02HC049)</td>
<td>2.36</td>
<td>4.75</td>
<td>52.0 - Mar</td>
<td>2.07 - Jul</td>
<td>3.37</td>
<td>1,230.13</td>
</tr>
<tr>
<td>Fourteen Mile Creek (02HB027)</td>
<td>0.20</td>
<td>0.47</td>
<td>10.10 - Mar</td>
<td>0.02 - Sept</td>
<td>0.31</td>
<td>113.90</td>
</tr>
<tr>
<td>Humber River (02HC003)</td>
<td>4.65</td>
<td>10.10</td>
<td>88.0 - Mar</td>
<td>1.71 - Sept</td>
<td>6.94</td>
<td>2,533.86</td>
</tr>
<tr>
<td>Mad River (02ED015)</td>
<td>1.70</td>
<td>4.63</td>
<td>24.9 - Mar</td>
<td>0.676 - Sept</td>
<td>2.94</td>
<td>1,071.64</td>
</tr>
<tr>
<td>Sixteen Mile Creek (02HB004)</td>
<td>0.71</td>
<td>1.90</td>
<td>43.5 - Mar</td>
<td>0.58 - Aug</td>
<td>1.22</td>
<td>445.30</td>
</tr>
</tbody>
</table>
3.3. Study Sites

The twelve confluence sites can be grouped into three geographical groupings; Greater Toronto Area, Southwestern and Central Ontario. The groupings are advantageous since the underlying surficial geology as well as climatic conditions are more similar for each group.

3.3.1. Greater Toronto Area Confluence Sites

Study confluences in the Greater Toronto Area, hereby referred to as the GTA, are Fourteen Mile Creek, Sixteen Mile Creek, Credit River (two sites), Don River, Duffins Creek and Humber River (two sites). The former three watersheds have headwaters above the Niagara Escarpment and the latter two have headwaters on the Oak Ridges Moraine. The Humber River has headwaters originating on both the Niagara Escarpment and the Oak Ridges Moraine. Each river flows through the geomorphic regions of the South Slope (characterized by glaciofluvial sediments), the Peel Plain and Iroquois Plain (characterized by glaciolacustrine sediments) before draining into Lake Ontario (Chapman and Putnam, 1984). Dominant bedrock formations underlying these watersheds are the Queenston shales and the Georgian Bay shales. The bedrock can be found in isolated outcroppings where a river flows over the Niagara Escarpment or where a channel becomes heavily incised. Halton Till is the main surficial component that makes up the South Slope in the western portion of the GTA region. It is comprised of 1-2% stone content and can occur in till or lake plain deposits ranging from 1 to 15 meters thick, often interbedded with fine sand, silt and clay (Sharpe et al., 1997). In the eastern GTA, Newmarket Till becomes more dominant. Newmarket Till has a larger stone content and can range in thickness from 1-50 meters (Sharpe et al., 1997). In areas of former glacial lake basins, glacial lake deposits of
poorly sorted silt, sand and clay dominate. These deposits are often closer to present day Lake Ontario shoreline as well as along insized reaches of rivers, and can range from 1-50 meters in thickness (Sharpe et al., 1997).

3.3.1.1 Credit River

The Credit River drains approximately 1000 km$^2$ with headwaters originating in the Orangville area on the Niagara Escarpment. The channel flows south for approximately 90 km before draining into Lake Ontario at Port Credit (CVC, 2007). Numerous tributaries enter the river forming at least twenty sub-watersheds. There are two study confluences along the Credit River; one at 21 km and one at 71 km upstream from the Lake Ontario outlet.

The study confluence along Credit River at 21 km is located at the confluence of the Credit and Fletchers Creek and is a 720 m long study reach (Figure 3.2). At this point, drainage area of the main Credit River is 715 km$^2$ making the contrast in discharge with Fletchers Creek, only 40 km$^2$, more substantial. This site is an asymmetrical confluence with Fletchers Creek entering the main Credit at a near 80° angle. Both channels are moderately incised into a low valley with undercut banks and heavily vegetated channel slopes above the identified bankfull stage. At the confluence there is a significant scour pool and deposition zone along the downstream tributary side. Along the tributary there is significant large woody debris influencing flow direction throughout the reach, and along the main Credit River upstream of the confluence minor stabilization structures have been put into place along the left bank, looking upstream. This site is categorized as a Type 2 (semi-alluvial) channel for this study.

Credit River at 71 km is the confluence between the Credit and West Credit rivers that join at the Forks in the Credit south of Caledon. At this confluence the Credit River drains only 219 km$^2$ and the West Credit drains 108 km$^2$, the study reach is 630 m long (Figure 3.3). This site is
a symmetrical planform confluence categorized as a Type 2 (semi-alluvial) channel due to the large boulder sized ($D_{\text{max}} > 100$ mm) sediment clasts, that appear to be glacially derived materials, present within the reach as well as large pebble and cobble sized materials being carried into the reach from upstream during bankfull flows. The study reach is also incised with undercut banks.

**Figure 3.2**: Credit River at 21 km. **A**: Plan view of the confluence (Google Earth, 2013), **B**: Looking downstream along the main Credit River past the confluence (field image), **C**: Looking upstream along Fletchers Creek (field image). (Arrows show the direction of flow).
Figure 3.3: Credit River at 71 km field images. A: Looking upstream along the confluence with the West Credit on the left and the main Credit River on the right, B: Looking downstream along Credit River. (Arrows show the direction of flow).

3.3.1.2 Don River

The Don River site is 24 km upstream from the Lake Ontario outlet and is located at the confluence of the East Don River and German Mills Creek. The entire study reach is 410 m long. At this point the East Don River drains approximately 61 km$^2$ and German Mills Creek drains 42 km$^2$ (Figure 3.4). Although the confluence is situated within a park setting, the effects of urbanization and increased runoff into the channels has made it necessary to install several flood control and bank stabilization structures along the study reach. Upstream from the confluence along the East Don River, an old weir structure is in place marking the start of the surveyed channel. Along German Mills Creek the channel is lined with concrete blocks to stabilize the banks. This confluence is symmetrical with near same bankfull discharges of 9.6 m$^3$/s for the East Don River and 8.1 m$^3$/s for German Mills Creek making this a co-dominant junction.
Upstream from the confluence, the East Don River forms a sharp meander bend where continuous flow hitting the bank forms an undercut which has subsequently collapsed releasing sediment ranging from sand to gravel into the channel (Figure 3.4B). Lateral bar deposits form along the banks of the channel with a median grain size of 26 mm. The morphology of German Mills Creek is almost entirely controlled by engineering structures throughout the study reach.

This site is categorized as a Type 1 (alluvial) channel due to the alluvial sediments exposed by channel incision. The channel incision is mainly attributed to rapid urbanization within the watershed increasing runoff into the channels.

*Figure 3.4:* Don River field images. **A:** Looking downstream past the confluence along the East Don River, **B:** Bank erosion upstream from the confluence on the East Don River. (Arrows show direction of flow).
3.3.1.3 Duffins Creek

Along Duffins Creek, the study confluence is located 7.5 km upstream from the Lake Ontario outlet at the confluence of the West Duffins and East Duffins creek branches (Figure 3.5). The entire length of all study reaches is 880 m long. At this location the West Duffins drains 134 km² and the East Duffins drains 121 km². The headwaters of both branches originate on the Oak Ridges Moraine. Channel substrate along the two tributary reaches varies significantly.

Along West Duffins Creek, greater bankfull discharge as well as higher channel slope has led to a coarser substrate, where D₅₀ is approximately 30 mm. Along this channel there are several erosional banks on the meander bends contributing finer sand and silt into the channel. In the channel there are areas of course grained deposits forming lateral channel bars. There is also evidence of chute channels present along multiple lateral bars. At the confluence, large woody debris blocks flow but not enough to create a pool (Figure 3.5C). Along the East Duffins Creek the gradient is significantly lower so that a backwater pool is able to form at the mouth of the tributary as it enters the confluence. Large woody debris is also found throughout this section of the study reach, impacting flow and creating areas of enhanced deposition and scour within the channel. Median grain size for the tributary is 25 mm. Downstream from the confluence there are substantial bar deposits that bifurcates the channel. There are also lateral bar formations with chute channels. The median grain size downstream of the confluence is 36 mm. While the floodplain is moderately to heavily vegetated above bankfull level with grasses an trees, there is still evidence of flooding overtopping the banks during high flows. This study site is categorized as a Type 1 (alluvial) channel.
Figure 3.5: Duffins Creek. A: Mid-channel bar along East Duffins Creek (field image), B: plan view of the confluence (Google Earth, 2013), C: Large woody debris impacting flow into the confluence from West Duffins Creek (field image), D: Looking upstream towards the confluence (eroded bank in background is East Duffins Creek), (field image). (Arrows show direction of flow).

3.3.1.4 Fourteen Mile Creek

Fourteen Mile Creek is a small watershed of only 30.2 km² with headwaters originating on the South Slope (Chapman and Putnam, 1984) below the Niagara Escarpment. The creek flows east draining into Lake Ontario at the town of Oakville. The study confluence in located 6 km upstream from the Lake Ontario outlet, where an unnamed tributary joins Fourteen Mile
Creek from the north. The total combined study reach is 470 m long (Figure 3.6). Flowing into the confluence, the main branch of Fourteen Mile Creek drains and area of 12.6 km$^2$ and the tributary drains 4 km$^2$. The tributary is characterized as non-alluvial Type 3 because of the extensive presence of Queenston shale outcroppings along the channel banks indicating that bedrock plays a large role in limiting vertical and lateral migration (Figure 3.6A). Despite the limiting factor of bedrock along the channel, the tributary exhibits the tendency for lateral instability due to the extensive presence of undercut banks (often undermining the shale) as well as the presence of chute channels forming on lateral bars. The main channel, both upstream and downstream from the confluence, exhibits a more active floodplain with alluvial deposits and is categorized as a Type 2 (semi-alluvial) channel. However, this floodplain is a bench step below the valley flat which in turn corresponds to the inactive floodplain along the tributary. Although there are lateral bars present along the inner corner of meander bends and outcroppings of Queenston shale, the main channel does not exhibit an overall tendency towards lateral instability.
Figure 3.6: Fourteen Mile Creek field images. A: Undercut Queenston shale outcropping along the banks of the tributary, B: Looking downstream along the Fourteen Mile Creek just upstream from the confluence, C: Tributary flows entering the confluence at a near 90° angle, D: Downstream from the confluence along the main channel. (Arrows show direction of flow).

3.3.1.5 **Humber River**

The Humber River watershed is located within the central Greater Toronto Area with headwaters originating on the Niagara Escarpment and the Oak Ridges Moraine. The watershed is comprised of five major sub-watersheds draining 903 km² into Lake Ontario at Humber Bay. There are two study confluences on the Humber River; the first site is located at the confluence
of the Humber River and West Humber which is 18 km upstream from the outlet. At this point the Humber River drains 580 km$^2$ and the West Humber 195 km$^2$. The confluence is asymmetrical with the two channels converging at an angle just above 90° (Figure 3.7). Along the entire reach the channel is incised into the floodplain with erosion scars and undercut banks present along the exposed vertical banks.

Flowing into the confluence, the main Humber River is associated with a steeper gradient and a larger bankfull discharge. Immediately upstream from the confluence there is a large lateral bar deposit of cobble to boulder size material that is distinct in the fact that the remainder of the upstream channel substrate is comprised of fine sand.

High energy and larger discharge from the main Humber River effectively truncates the tributary discharge flowing into the confluence. This creates a backwater pool just upstream on the West Humber River that over-widens the channel, and can be especially seen during low flows. Along the West Humber River, lateral channel bars are constant throughout the study reach where $D_{50}$ is 46 mm. A significant lateral bar has formed at the upstream junction corner extending for several meters downstream along the main Humber River. This entire confluence site is categorized as a Type 1 (alluvial) channel.

The second site is located at the confluence of the Humber River and Cold Creek which is 56 km upstream from the outlet. At this location the Humber River drains approximately 200 km$^2$ and Cold Creek drains 65 km$^2$ (Figure 3.8). This confluence is asymmetrical, however both channels have similar low gradients and median grain size of approximately 10 mm upstream from the confluence. Downstream from the confluence, the reach gradient steepens and median grain size increases to approximately 25 mm. The reach is incised into the floodplain with frequent evidence of erosion scars along the exposed banks. However, there is also evidence of
floodplain connection with recent alluvial deposits at the upper and lower extents of the reach. This confluence is categorized as a Type 1 (alluvial) channel.

**Figure 3.7:** Humber River at 18 km. **A:** Planar view of the confluence (Google Earth, 2013), **B:** West Humber flows draining into the confluence from the left, main Humber River flows from the top of the picture (field image). (Arrows show direction of flow).

**Figure 3.8:** Humber River at 56 km. **A:** Planar view of the confluence (Google Earth, 2013), **B:** Humber River flows draining from the bottom left, Cold Creek flows from the top middle (field image). (Arrows show direction of flow).


### 3.3.1.6 Sixteen Mile Creek

Sixteen Mile Creek watershed is comprised of three main tributary branches, West, Middle and East, that flow southward through Milton and Mississauga before joining to form the main channel near the town of Oakville and eventually draining into Lake Ontario. The resulting watershed drains approximately 372 km². Similar to Fourteen Mile Creek, Sixteen Mile Creek headwaters originate along the Niagara Escarpment (Chapman and Putnam, 1984). The study site is located at the confluence of the East and West branches 15.5 kilometers upstream from Lake Ontario. The total study reach is 1000 m long. The confluence has a symmetrical ‘Y’-shaped, planform where both joining channels have near identical bankfull discharges making this a co-dominant confluence. However, the East branch has a larger drainage area of 202 km² whereas the West branch has a drainage area of 133 km² (Figure 3.9). The Sixteen Mile Creek study reach is categorized as a Type 3 (non-alluvial) channel because of the presence of a bedrock channel bottom as well as a Queenston Shale bluff along the West branch immediately upstream from the confluence. The tributary channels are also moderately incised, where undercut banks are present alongside a floodplain which is also heavily forested.
Figure 3.9: Sixteen Mile Creek. A: Queenston shale bluffs along West Sixteen Mile Creek (field image). B: Bedrock incision along East Sixteen Mile Creek (field image). C: Planar view of the Sixteen Mile Creek confluence (Google Earth, 2013). (Arrows show direction of flow).
3.3.2. Southwestern Ontario Confluence Sites

Study confluences in South-Western Ontario are along the Ausable River at km 70 and 71.5, and Catfish Creek at 24 km.

3.3.2.1 Ausable River

The Ausable River is the largest watershed in this study as it drains an area of 1189 km$^2$ into Lake Huron at the town of Port Franks. The underlying physiography of this watershed is dominated by the till based Wyoming Moraine. The channel flows parallel to the moraine for several kilometers until it makes an abrupt change in flow direction to the north cutting through the moraine and producing a steep sided gorge. As the channel crosses the moraine, significant down cutting has occurred exposing bedrock at numerous locations. Downstream from the gorge, the channel is influenced by Lake Huron shoreline sand plain deposits from the most recent glaciation (Chapman and Putnam, 1984).

There are two confluence study sites along the Ausable River, both are within the Gorge that cuts across the Wyoming Moraine and they are approximately one and a half kilometers apart (Figure 3.10). These sites were chosen for the steep gradients of the tributaries compared to the rest of southern Ontario sites. However, the tributaries are significantly smaller than the main channel with drainage areas of less than 2 km$^2$, with the source water coming from agricultural runoff at the top of the Gorge. Ausable River at 70 km has a tributary with a gradient of 0.06 m/m and a median ($D_{50}$) surficial grain size of 69 mm. Ausable River at 71.5 km has a tributary with a gradient of 0.18 m/m and a median grain size also of 74 mm. Both tributaries are deeply incised with bedrock underlying small sections of the channel. Undercut banks are also present along much of the channels as well as large woody debris blockades. At the confluence with the main Ausable River, both tributaries form sediment bars at the upstream
and downstream junction corners, however a significant lateral bar is present at the downstream junction corner of Ausable River at 70 km. This bar significantly alters the flow path of the main Ausable, creating a deep stagnant pool (> 2 m at bankfull flow) on the upstream side and an area of rapid turbulent flow downstream and parallel to the bar. Apart from this location, there is very little change in the main Ausable River channel at either site except a gradual decrease in slope downstream. The channel is located at the bottom of a steep walled valley where there is evidence of relic terraces formed from channel incision. Median grain size of the mainstem Ausable River is 67 mm at both confluences. Banks are heavily vegetated with grass and shrubs and there is no identifiable floodplain. Since bedrock directly influences the tributaries they are categorized as Type 3 (non-alluvial) channels, were as the main Ausable River both upstream and downstream of the confluences are categorized as Type 2 (semi-alluvial) channels since there is no direct association with bedrock.
Figure 3.10: Ausable River at 70 km and 71.5 km. **A:** Looking downstream along the tributary at 71.5 km (field image), **B:** Looking upstream along the tributary at 70 km (field image), **C:** Planar view of both Ausable River confluence sites (Google Earth, 2013). (Arrows show direction of flow).
3.3.2.2  *Catfish Creek*

Catfish Creek is a watershed that drains approximately 490 km$^2$ and flows into Lake Erie with the outlet at Port Bruce Ontario. Catfish Creek headwaters are located along the St. Thomas, Norwich and Tillsonburg moraines which form the northern and eastern watershed boundaries (Martyn and Ashbaugh, 2010). The confluence is approximately 24 km upstream from the Lake Erie outlet at the junction of the West Catfish and East Catfish Creek branches. East Catfish Creek is the dominant tributary and drains approximately 150 km$^2$ of land where as the West Catfish Creek drains 100 km$^2$ (Figure 3.11).

The confluence forms a symmetrical “Y”-shaped junction where backwater effects have over widened the mouth of West Catfish Creek where a significant amount of water pools even at low summer flow observed during surveying. Immediately downstream from the confluence the banks have been engineered to support a road overpass over the channel. Approximately 100 m downstream from the confluence, the channel makes an abrupt 90° turn to flow east along an exposed till bluff. It is along this section of the reach that the channel gradient increases significantly and bed substrate coarsens to a median grain size of 66 mm.

East Catfish Creek, upstream from the confluence, is incised into the floodplain, with nearly vertical banks that are often vegetated and displaying areas of erosion. The bed substrate ranges from coarse sand to gravel (median grain size approximately 10-20 mm) with bar formations along the water’s edge as well as in the center of the channel bifurcating the flow.

Upstream from the confluence, West Catfish Creek remains over widened for approximately 100 m. It is through this section of the reach that the banks are lined with armour stones for bank stabilization. A lower bench develops further upstream with evidence of recent flooding. Bed
substrate is again sand to coarse gravel with median grain size on lateral bars approximately 10-20 mm. All channels at this study confluence are Type 1 (alluvial) channels.

Figure 3.11: Catfish Creek. A: Looking upstream along the confluence, West Catfish Creek flows from in from the left and East Catfish Creek from the right (field image), B: Looking downstream along the main channel (field image), C: Planar view of the confluence (Google Earth, 2013). (Arrows show direction of flow).
3.3.3. Mad River Confluence Site

The Mad River is a tributary to the larger Nottawasaga River watershed which drains into Georgian Bay at Wasaga Beach. The Mad River watershed drains approximately 354 km$^2$. The study confluence is located at the junction of the Mad River and Noisy River approximately 40 km upstream from the Nottawasaga River junction and 74 km upstream from Georgian Bay (Figure 3.12). Both the Mad and Noisy Rivers have headwaters originating above the Niagara Escarpment where they both carved out incised valleys into the Amabel dolostone and Queenston shale bedrock formations as they flow off of the Escarpment and continue across the Simcoe Lowlands sand plains (Berger Group, 2006; Post and Rodie, 2010). At the confluence site the Mad River drains 130 km$^2$ and the Noisy River 96 km$^2$. The tributaries merge at a symmetrical confluence, complicated by the fact that the Mad River bifurcates into two channels separated by a heavily vegetated small island forming a three channel confluence (Figure 3.12A). Upstream, both the Mad and Noisy Rivers have steep gradients and median grain size of 91 mm. Substrate is dominated by fractured shale material, of which some is contributed from further upstream and some is eroded from the banks. The channels are situated within a valley bottom, but there is an active floodplain present. The Mad River exhibits a greater tendency for lateral instability with chute channels and slack water pools remaining from previous high flows. Downstream from the confluence, the channel gradient decreases and the substrate coarsens with boulders within the channel. However, median grain size along lateral bar deposits decreases to 61 mm. The downstream section of the reach also flows alongside a steep embankment that releases eroded sediments during high flows and also limits the lateral mobility of the channel. The tributary channels are categorized as Type 3 (non-alluvial) channels and the main-stem downstream of the confluence is categorized as a Type 2 (semi-alluvial) channel.
Figure 3.12: Mad River field images. **A:** Looking upstream at the confluence where the Noisy River joins from the far left, **B:** Looking downstream towards the confluence (background) along the Mad River, **C:** Looking downstream towards the confluence (background) along the Noisy River. (Arrows show direction of flow).
4. Methods

In order to properly investigate the changing morphologies downstream from a confluence junction, several techniques were employed in both the field and in the lab setting.

4.1. Field Work

Field work was completed between May and August of 2012. Surveys conducted at each confluence site included cross-sections and longitudinal-profiles of the tributaries and main-stems using a Sokkia™ engineering level and stadia rod. Grain size measurements were made on exposed channel bars that were present.

4.1.1. Surveying

At each confluence site, at least two channel cross-sectional profiles were surveyed along each tributary channel and the main channel downstream from the confluence. The locations of the cross-sections were determined by visually identifying overall characteristic channel morphologies. For instance, cross-sections were not surveyed in areas that were identified as back-water pools. Cross-section spacing was distributed throughout the reach in order to capture a “characteristic” channel. Longitudinal profiles were surveyed to obtain low water surface slope. The profiles were surveyed throughout each reach at current water-levels for at least one reach length, approximately 10 times the bankfull channel width.
4.1.2. Grain Size Analysis

In addition to the channel surveys, the Wolman (1954) pebble count method was conducted along each tributary and main channel reach where the median surficial grain size was greater than 10 mm along the intermediate b-axis. When the dominant surficial grain size was smaller than 10 mm, the characteristic grain size was simply noted with a general description such as coarse sand. The surveys were conducted along lateral channel bars where longitudinal transects were set up and sediments were samples every 25 to 50 cm until at least 100 particles had been measured.

Figure 4.1: Wolman style pebble count along the Mad River (field image).
4.2. Determining Channel Forming (Bankfull) Discharge

Reach specific discharges were inferred from the channel cross-section surveys and velocity estimates based on the width and depth at bankfull discharge as well as channel roughness. In the field, numerous features along the bank were recorded in an attempt to accurately determine bank-full discharge and roughness at each cross-section location. Recorded in the field were: current water-level, bar tops, undercut banks, erosion scars and scouring, vegetation and tree lines, inflection points where the bank slope starts to decrease, and floodplain level. These features correspond with Williams (1978) indicators for determining bank-full flow stage. When the cross-sections were viewed graphically, average widths and depths could be calculated to determine the minimum width-depth ratio which was also subsequently included in determining cross-sectional areas at bank-full discharge.

When an appropriate bankfull flow stage was determined, bankfull discharge was estimated using two different velocity-area calculations. The first is the Manning’s relation defined as:

\[ V = \frac{d^{2/3} S^{1/2}}{n} \]  \[4.1\]

Where \( d \) is the average depth, \( S \) is the channel slope and \( n \) is the estimated Manning’s roughness coefficient determined using Cowan’s (1956) compound equation method:

\[ n = n_0 + n_1 + n_2 + n_3 + n_4 m_5 \]  \[4.2\]

The variables in the compound relation are; sediment size \( (n_0) \), degree of surface irregularity \( (n_1) \), variation in channel cross-section \( (n_2) \), effects of obstructions \( (n_3) \), vegetation \( (n_4) \), and
degree of channel meandering ($m_3$). A second approach for determining $n$ is the Manning-Strickler roughness determination method (cf. Richards, 1982) was also used where:

$$ n = 0.039 \, D_{50}^{0.167} \quad [4.3] $$

Where $D_{50}$ is the median or 50th percentile grain size sampled along the reach.

A second velocity calculation is based on the Darcy-Weisbach relation defined as:

$$ V = \frac{8 \, g \, R \, S}{f \, f} \quad [4.4] $$

Where $g$ is the acceleration due to gravity, $R$ is the hydraulic radius of the channel x-section at bankfull level, and $f \, f$ is the friction factor first developed by Limerinos (1969) to employ a relative smoothness measure (cf. Richards, 1982):

$$ f \, f = [1/(2 \, \log (d/D_{84}) + 1.16)]^2 \quad [4.5] $$

Where $D_{84}$ is the 84th percentile grain size measured for each channel reach, of which 84% of particles are smaller.

In order to improve the accuracy of these bank-full discharge estimates the results were compared with a regional flood regime and a local flood regime relation based on the correlation between the two year ($Q_2$) discharge and contributing drainage area where:

$$ Q_2 = a \, A_d^b \quad [4.6] $$

The coefficient $a$ and the exponent $b$ are calculated by a linear regression test for Log10 transformed discharge-drainage area data. The regional regime model is based on a standard flood frequency analysis of gauging records for southern Ontario obtained from the Water
Survey of Canada, where \( a \) is equal to 0.25 and \( b \) is equal to 0.91. The local regime model only uses gauge records for a particular geographical area divided by catchment authority (see Table 4.1) (Phillips, 2012). For these estimates drainage area was delineated using the Spatial Analyst Hydrology Tool in ArcGIS from the Ontario Provincial DEM of 10 m resolution (Merwade, 2009).

Discharges were also directly compared to mean daily discharges from local gauging stations whenever there was a gauge present within a reasonable distance from the study reach.

**Table 4.1:** Coefficients and exponents needed for the local regime and regional regime models.

<table>
<thead>
<tr>
<th>Catchment Authority</th>
<th>( a )</th>
<th>( b )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Toronto and Region/Credit Valley</td>
<td>0.28</td>
<td>0.83</td>
</tr>
<tr>
<td>South Lake Huron</td>
<td>0.4</td>
<td>0.9</td>
</tr>
<tr>
<td>Nottawasaga/Georgian Bay</td>
<td>0.2</td>
<td>0.86</td>
</tr>
<tr>
<td>Long Point</td>
<td>0.1</td>
<td>1.05</td>
</tr>
<tr>
<td><strong>Regional Model</strong></td>
<td><strong>0.25</strong></td>
<td><strong>0.91</strong></td>
</tr>
</tbody>
</table>
Chapter Five: Influences of Confluences on the Downstream Morphology of Stream Channels

5. Results

The twelve study sites constitute 36 study reaches (two tributaries and one mainstem channel for each site). The results in this chapter are set in the context of the most influential factors that are thought to control channel morphology including river gradient, bankfull discharge and grain size. Several studies looking at discriminating form and function of river channels have been attempted over many years, so the results from this study are compared to both classical channel planform discriminations of Leopold and Wolman (1957) as well as recently revised approaches of Kleinhans and van den Berg (2011). Both of these discriminating methods seek to predict the lateral stability of a channel reach based on planform morphology such as if a channel is meandering or braided, or to what extent there is sediment bar formation. A laterally stable channel is more likely to be a single-thread meandering or straight with minimal bar formations. A laterally unstable channel is more likely to have two or more channel threads with significant depositional bars with chute cutoffs and unstable banks. The lateral stability of a channel can relate to the equilibrium status of the reach. A channel that has significant sediment deposition in the form of bars can signify an imbalance in the ability of the channel to adequately transport the sediment load under the given discharge and channel slope (Lane, 1954), therefore, the channel is aggrading and in a state of non-equilibrium.
5.1. Channel Patterns at Confluences

Leopold and Wolman (1957), used the slope-discharge relationship is an attempt to define a threshold line that discriminates between laterally stable (meandering) and laterally unstable (braided) rivers. Figure 5.1 shows all 36 reaches in this study plotted with the Leopold and Wolman (1967) discriminating line.

Leopold and Wolman (1957) define a braided channel as one that has two or more anastomosing channels flowing around alluvial islands. They also indicate that braided channels as a whole are steeper, wider, and shallower then single thread meandering channel. For a given discharge, meandering channels will have a gentler slope, and for the same slope a braided channel will have a higher discharge. In the context of southern Ontario however, the notion of a fully braided channel developing is highly unlikely due to the relatively gentler slopes, finer grain size, and bank confinement present along many water courses. For this reason, a reinterpretation of Leopold and Wolman’s’ “braided” and “meandering” channels into “laterally unstable” and “laterally stable” respectively is necessary for a clearer interpretation of channel form in this environment. A laterally unstable channel is more likely to develop chutes, cut-off channels or to avulse and create a new channel through the floodplain, whereas laterally stable channel are likely to remain single-thread and in a more constant position.

Figure 5.1, shows that the study reaches generally plot both above and below the Leopold and Wolman (1957) threshold line. This suggests that approximately 50% of the reaches surveyed are energetic enough to become laterally unstable. The Ausable River tributaries plot anomalously high because of their steep gradients (S > 0.05 m/m) and low bankfull discharge (Qₘ < 1 m/s) in comparison to the rest of the southern Ontario sites.
Figure 5.1: Slope and bankfull discharge relationship for each study reach plotted with the Leopold and Woman (1957) threshold line (solid, grey). Sites that plot above the Leopold and Wolman threshold line are potentially laterally unstable and sites that plot below the line are potentially laterally stable.

When the 36 reaches are plotted based on boundary condition type as outlined in chapter 3, some clustering appears (Figure 5.2).
Figure 5.2: Study reaches grouped based on boundary condition classification.

All of the semi-alluvial and non-alluvial reaches consistently plotting either along or above the threshold line. Alluvial (Type 1) reaches, which comprise the majority of the study reaches, show a wider scatter both above and below the threshold line. However, within a single tributary-mainstem study context, the plotting positions can indicate a dependence on local river channel characteristics. For instance, Figure 5.3 shows that both of the Catfish Creek tributaries plot distinctly below the threshold line and the combined main channel plots above of the line. The Catfish Creek tributaries are both narrow, deep channels that are incised into the floodplain and have little chance for migration (see Table 3.1 and Figure 3.11). Downstream along the combined mainstem channel, however, the floodplain becomes more accessible and the channel widens. Even though the mainstem does not exhibit significant evidence for lateral instability, the higher gradient and wider channel present along the reach, indicates that this channel is still more likely to be unstable (i.e. more sensitive to change) in comparison to the two tributaries.
upstream. Therefore, plotting position of these three reaches accurately describes the conditions observed in the field.

**Figure 5.3:** Catfish Creek tributaries and mainstem plotted with the Leopold and Wolman (1957) discrimination line. ■ indicates the combined downstream channel, △ indicates the main tributary flowing into the confluence, and ⊱ indicates the lesser tributary flowing into the confluence.

**Figure 5.4:** All alluvial reaches plotted on the Leopold and Wolman (1957) style slope-discharge plot. ■ indicates the combined downstream channel, △ indicates the main tributary flowing into the confluence, and ⊱ indicates the lesser tributary flowing into the confluence.
Another example of where channel reaches downstream of a confluence change in potential stability is the Humber River at 56 km (Figure 5.4). Both of the tributaries flowing into the confluence plot below the threshold line while the downstream combined channel plots on the line indicating the contribution of the two tributaries leads to a higher potential for instability downstream. Both tributaries are again heavily confined and entrenched within the moderately vegetated (i.e. grasses and few shrubs) channel banks, with very limited migration and minor meander bend progression through bank erosion. Downstream from the confluence the channel widens and becomes shallower (see Table 3.1, and Figure 3.8). These channel conditions reflect the plotting position towards and beyond the Leopold and Wolman (1957) threshold line.

The Don River, Duffins Creek, and Humber River at 18 km sites all have the dominant tributary as well as the combined downstream channel plotting above the threshold line (Figure 5.4). The less dominant tributaries of the Don River and Humber River at 18 km plot along the threshold line, where as the Duffins Creek lesser tributary plots distinctively below the line. For these sites, the steeper slopes along the mainstem and dominant tributary results in the greater potential for a more laterally unstable pattern. Based on evidence present in the field (see site photos in Chapter 3), the increased lateral instability downstream from a confluence of these alluvial boundary sites is warranted. However, the degree of lateral instability predicted when reaches are plotted along the Leopold and Wolman (1957) threshold line can be difficult to interpret when compared to evidence observed in the field, especially when the dominant tributary and mainstem channels are concerned. For example, the dominant tributary of Humber River at 18 km plots as significantly unstable when there is no evidence that it is any more unstable then the downstream mainstem channel. The lesser tributaries on the other hand, have
single-thread channel morphologies (see site photos in Chapter 3) in accordance with their plotting position in Figure 5.4.

Channel reaches that have been classified as semi-alluvial and non-alluvial show a clearer pattern of being potentially more laterally unstable than their full alluvial counterparts. In particular, all non-alluvial reaches plot above the Leopold and Wolman (1957) threshold line. Commonly, these reaches have steeper slopes for a comparable discharge in alluvial reaches. For the semi-alluvial and non-alluvial reaches, the tributary to mainstem relationship shows a progression that is more predictable in a trend parallel to the threshold line when compared with the fully alluvial reaches (Figure 5.5 and Figure 5.6). Generally, the lesser tributaries have the highest slopes and lowest bankfull discharges and the mainstems have the lowest slopes and highest discharges, with the dominant tributary as transitional between the two. For semi-alluvial and non-alluvial sites, lesser tributaries plot as the most unstable in Figure 5.5. For example, the lesser tributary of Sixteen Mile Creek has significant bank erosion as well as multiple channel threads developing around stable mid-channel bars. Evidence of instability is less in the main tributary as well as the mainstem (see site photos in Chapter 3).
Figure 5.5: Semi-alluvial reaches plotted on the Leopold and Wolman (1957) style slope-discharge plot. ● indicates the combined downstream channel, + indicates the main tributary flowing into the confluence, and ○ indicates the lesser tributary flowing into the confluence. Not all reaches at one confluence site are characterised as the same alluvial type, see Table 3.1.

Figure 5.6: Non-alluvial reaches plotted on the Leopold and Wolman (1957) style slope-discharge plot. ▲ indicates the combined downstream channel, △ indicates the main tributary flowing into the confluence, and □ indicates the lesser tributary flowing into the confluence. Not all reaches at one confluence site are characterised as the same alluvial type, see Table 3.1.
Semi-alluvial and non-alluvial confluences that do not clearly follow the pattern of downstream mainstem channels forming a channel pattern that is a combination of the tributaries, are Fourteen Mile Creek and both of the Ausable River sites. The predicted bankfull discharge for the mainstem channel at Fourteen Mile Creek is lower than that predicted for the main tributary, so plotting positions in Figure 5.5 have a much higher degree of uncertainty for these two reaches. It is likely that either the main tributary bankfull discharge is overpredicted or the combined channel discharge is underpredicted from their cross-section areas. Since bankfull discharge was calculated from a top-of-bank geomorphic inflection in the cross section, either the main tributary is more entrenched into the floodplain than geomorphic indicators reflect, or the main downstream channel regularly overflows its banks at bankfull flows.

The abrupt changes in channel slope and bankfull discharge downstream from confluences significantly influence the pattern of alluvial channel reaches as seen in the tributary to mainstem pattern in Figure 5.4. For most alluvial reaches, downstream changes include widening of the channel and steepening of the gradient. However, for reaches that are only partially alluvial or non-alluvial in nature the downstream steepening of the channel gradient is not experienced, producing a pattern where the transition downstream of the confluence is a linear pattern on the slope vs. discharge plots between the tributaries and mainstems reaches at each site.

Overall, based on channel features observed in the field, alluvial reaches experience a more abrupt change in channel pattern downstream from confluences. This observation is generally confirmed when data are displayed on the slope vs. discharge plots with the Leopold and Wolman (1957) threshold line as reference. However, a weakness of the Leopold and
Wolman (1957) approach is the degree of lateral instability is not adequately captured by a single threshold line.

The more robust approach of Kleinhans and van den Berg (2011) uses potential specific stream power ($\omega_{pv}$) of each reach versus median grain size ($D_{50}$). Stream power is regarded as a useful parameter for predicting channel patterns because it signifies the energy to move a characteristic size of sediment, in this case the median grain size (Kleinhans and van den Berg, 2011). Results from this study are plotted along with empirical thresholds developed by Kleinhans and van den Berg (2011) in Figure 5.7. For this study reference to channel “braiding” will be reinterpreted as “laterally unstable”, so that the greater $\omega_{pv}$, the greater the potential instability.

**Figure 5.7:** Study reaches grouped by boundary characteristics plotted with the Kleinhans and van den Berg (2011) discrimination thresholds.
Alluvial reaches again show the widest range in stream power and median grain size compared to semi-alluvial and non-alluvial reaches (Figure 5.7). The high variability in predicted channel pattern among the alluvial reaches again indicates that site specific characteristics, such as bank strength, channel entrenchment, etc. may be important determinants as to how these channels adjust to abrupt discharge increases downstream from the confluence.

Grain size and stream power for alluvial reaches continues to show that the combined downstream channels are increasingly unstable with a higher potential stream power compared to the contributing tributaries (Figure 5.8). For the very limited number of alluvial reaches studied here, there is a weak positive realation between $\omega_{pv}$ and $D_{50}$.

**Figure 5.8:** All Alluvial reaches plotted with the Kleinhans and van den Berg (2011) potential specific stream power – median grain size threshold lines. ■ indicates the combined downstream channel, + indicates the dominant tributary flowing into the confluence, and + indicates the lesser tributary flowing into the confluence.
When potential stream power and median grain size are plotted for the semi-alluvial and non-alluvial reaches (Figure 5.9 and Figure 5.10), the lesser tributaries flowing into each confluence generally have larger stream power due to higher slopes indicating that tributaries are the least potentially stable. Where there was a general trend to increasing grain size downstream of confluences in the alluvial reaches, this is not apparent in the majority of semi-alluvial and non-alluvial reaches. Except for Credit River at 72 km, Figure 5.9 shows the grain size of the combined downstream channel generally in an aggregate average of the grain sizes from the two contributing tributaries.

**Figure 5.9:** Semi-alluvial reaches plotted with the Kleinhans and van den Berg (2011) potential specific stream power – median grain size threshold lines. ● indicates the combined downstream mainstem channel, ▲ indicates the dominant tributary flowing into the confluence, and ○ indicates the lesser tributary flowing into the confluence. Not all reaches at one confluence site are characterised as the same alluvial type, see Table 3.1.
Figure 5.10: Non-alluvial reaches plotted with the Kleinhans and van den Berg (2011) potential specific stream power – median grain size threshold lines. ▲ indicates the combined downstream mainstem channel, ▲ indicates the dominant tributary flowing into the confluence, and ▲ indicates the lesser tributary flowing into the confluence. Not all reaches at one confluence site are characterized as the same alluvial type, see Table 3.1.

Field observations of actual channel morphology versus predicted channel stability in Figure 5.8, Figure 5.9 and Figure 5.10 provides some interesting results. The majority of the alluvial sites are predicted to be meandering channel with scrolls and these correspond with observed characteristics. For example, the channel patterns observed along the Catfish Creek reaches are single thread meandering channels with minor bank erosion present. Thus, the $\omega_{pv}$ vs. $D_{50}$ of these channels accurately predicts the observed stability. The Don River tributaries have undergone some channel stabilization, whereas the downstream mainstem is more fully alluvial. The current stream power and grain size characteristics of these tributary reaches appear to have adjusted to the more stable regime as confirmed in Figure 5.8. The downstream mainstem reach of the Don River is observed to have minor bank erosion and bar formation. The greater
instability of this reach is confirmed in Figure 5.8 where it plots as moderately unstable. The only channel reach that plots anomalously high towards instability is the dominant tributary at Humber River (18 km) site. This channel is deeply entrenched into the floodplain with a steep gradient but maintains a single thread channel with some evidence of bank erosion. Without entrenchment it is expected this channel would be much less stable in form.

Stream power and median grain size of the semi-alluvial and non-alluvial reaches predicts that the majority of channels should be meandering with scrolls. Most all of these channels are single thread meandering channels with some degree of bank erosion and mid-channel bar formation so median grain size and stream power thresholds appear to have reasonable discriminating power. However, the non-alluvial reaches have highly developed chute cutoffs and stable island development that are not as accurately represented by the plotting positions in Figure 5.10. Based on detailed measurements and results, and site visits to each of these confluences, some mainstems downstream from a confluence are at or near equilibrium conditions and some are poorly adjusted (non-equilibrium). Table 5.1 shows this division.
Table 5.1: Equilibrium status of mainstem reaches.

<table>
<thead>
<tr>
<th>Equilibrium State</th>
<th>Boundary Type</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mainstem Channel</strong></td>
<td><strong>Boundary Type</strong></td>
</tr>
<tr>
<td>Ausable River (x2)</td>
<td>Semi-Alluvial (Type 2)</td>
</tr>
<tr>
<td>Catfish Creek</td>
<td>Alluvial (Type 1)</td>
</tr>
<tr>
<td>Credit River (x2)</td>
<td>Semi-Alluvial (Type 2)</td>
</tr>
<tr>
<td>Humber River (56 km)</td>
<td>Alluvial (Type 1)</td>
</tr>
<tr>
<td>Mad River</td>
<td>Semi-Alluvial (Type 2)</td>
</tr>
<tr>
<td>Sixteen Mile Creek</td>
<td>Non-Alluvial (bedrock) (Type 3)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Non-Equilibrium State</th>
<th>Boundary Type</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mainstem Channel</strong></td>
<td><strong>Boundary Type</strong></td>
</tr>
<tr>
<td>Don River</td>
<td>Alluvial (Type 1)</td>
</tr>
<tr>
<td>Duffins Creek</td>
<td>Alluvial (Type 1)</td>
</tr>
<tr>
<td>Fourteen Mile Creek</td>
<td>Semi-Alluvial (Type 2)</td>
</tr>
<tr>
<td>Humber River (18 km)</td>
<td>Alluvial (Type 1)</td>
</tr>
</tbody>
</table>

5.2. Possible Causal Factors Controlling Channel Characteristics

Channel morphology characteristics of the tributaries and mainstems are analyzed for significant relationships using several sets of independent and controlling variables. Field measured channel characteristics of average width, average depth, slope and median grain size as well as the calculated bankfull discharge, form a set of variables that characterize channel reach attributes and the energy available to alter channel form. The hydraulic geometry of the aggregate data set can be explored by looking for associations between discharge and width and depth.

As expected, the variation in bankfull discharge for all study reaches is significantly related to channel width \(r^2 = 0.91\) (Table 5.1 and Figure 5.11A). Best subset regression
indicates that a total of 95% of bankfull discharge variation can be explained through the combination of channel width and depth. Figure 5.11A confirms the expected hydraulic geometry relation that width is proportional to discharge (i.e., w \( \propto Q^{0.5} \)) (cf. Leopold and Wolman, 1957). Depth also has a strong association with bankfull discharge confirming the expected hydraulic geometry relation of hydraulic geometry suggest d \( \propto Q^{0.4} \) (Figure 5.11B). Figure 5.11C shows that the aggregate data set of 12 rivers (36 sites) also confirms the well-established relation that as bankfull discharge increases channel slope decreases. The exponent b = -0.46 is close to that identified by Annabel (1996) for a larger group of other southern Ontario rivers. Median grain size and bankfull discharge are not well related (\( r^2 = 0.013 \)) as seen in Figure 5.11D.
Figure 5.11: Regression plots for bankfull discharge vs. bankfull width (A), bankfull depth (B), slope (C) and median grain size (D).
Table 5.2: Linear Regression Results for Log$_{10}$ Transformed Bankfull Discharge ($Q_b$) Data.

<table>
<thead>
<tr>
<th>Independent Variable</th>
<th>a</th>
<th>b</th>
<th>$r^2$</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>Constant Variance Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width</td>
<td>5.0</td>
<td>0.44</td>
<td>0.91</td>
<td>0.20</td>
<td>&lt;0.001</td>
<td>Passed ($P=0.519$)</td>
<td>Passed ($P=0.114$)</td>
</tr>
<tr>
<td>Depth</td>
<td>0.24</td>
<td>0.40</td>
<td>0.82</td>
<td>0.28</td>
<td>&lt;0.001</td>
<td>Passed ($P=0.753$)</td>
<td>Passed ($P=0.296$)</td>
</tr>
<tr>
<td>Slope</td>
<td>0.02</td>
<td>-0.46</td>
<td>0.41</td>
<td>0.51</td>
<td>&lt;0.001</td>
<td>Passed ($P=0.338$)</td>
<td>Passed ($P=0.625$)</td>
</tr>
<tr>
<td>$D_{50}$</td>
<td>0.036</td>
<td>0.04</td>
<td>0.013</td>
<td>0.65</td>
<td>0.504</td>
<td>Failed ($P=0.003$)</td>
<td>Passed ($P=0.059$)</td>
</tr>
</tbody>
</table>

Sample size (N) is 36 for all regression tests.

Channel slope for all 36 study reaches is moderately related to width and depth as expected from the hydraulic geometry relations (slope and bankfull discharge was previously assessed). The functional relationships for slope and depth (Figure 5.12B) and for slope and width (Figure 5.12A) show the strongest relationships. Slope and median grain size (Figure 5.12C) have a moderate positive relationship indicating that increased gradient does lead to a coarsening of the channel bed.

A multiple linear regression analysis of log$_{10}$ transformed data (Table 5.3) indicates that width, depth and bankfull discharge are highly inter-correlated therefore in a best subset regression model depth and median grain size together produce the highest explanation of slope variations in the 36 point data set. When both depth and median grain size are modeled together they explain 83% of slope variation (Table 5.3).
Figure 5.12: Regression plots for slope vs. width (A), depth (B) and median grain size (C).
Table 5.3: Linear Regression Results for Log\(_{10}\) Transformed Slope (S) Data.

<table>
<thead>
<tr>
<th>Independent Variable</th>
<th>a</th>
<th>b</th>
<th>(r^2)</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>Constant Variance Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width</td>
<td>2.1</td>
<td>-0.41</td>
<td>0.41</td>
<td>0.36</td>
<td>&lt;0.001</td>
<td>Passed (P=0.0266)</td>
<td>Failed (P=0.006)</td>
</tr>
<tr>
<td>Depth</td>
<td>0.06</td>
<td>-0.48</td>
<td>0.64</td>
<td>0.29</td>
<td>&lt;0.001</td>
<td>Passed (P=0.306)</td>
<td>Passed (P=0.547)</td>
</tr>
<tr>
<td>(D_{50})</td>
<td>0.17</td>
<td>0.27</td>
<td>0.25</td>
<td>0.41</td>
<td>0.002</td>
<td>Failed (P=0.013)</td>
<td>Passed (P=0.088)</td>
</tr>
</tbody>
</table>

Sample size (N) is 36 for all regression tests.

Table 5.4: Best Subset Regression Results for Log\(_{10}\) Transformed Slope (S) Data.

<table>
<thead>
<tr>
<th>Model # 1 (r^2 = 0.637)</th>
<th>Variable</th>
<th>Coef.</th>
<th>Std. Error</th>
<th>t</th>
<th>P</th>
<th>VIF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Constant</td>
<td>-2.447</td>
<td>0.051</td>
<td>-47.50</td>
<td>&lt;0.001</td>
<td>0.000</td>
<td>1.00</td>
</tr>
<tr>
<td>Depth</td>
<td>-1.323</td>
<td>0.171</td>
<td>-7.72</td>
<td>&lt;0.001</td>
<td>1.000</td>
<td>1.00</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Model # 2 (r^2 = 0.828)</th>
<th>Variable</th>
<th>Coef.</th>
<th>Std. Error</th>
<th>t</th>
<th>P</th>
<th>VIF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Constant</td>
<td>-1.306</td>
<td>0.191</td>
<td>-6.83</td>
<td>&lt;0.001</td>
<td>0.000</td>
<td>1.007</td>
</tr>
<tr>
<td>Depth</td>
<td>-1.264</td>
<td>0.120</td>
<td>-10.53</td>
<td>&lt;0.001</td>
<td>1.007</td>
<td>1.007</td>
</tr>
<tr>
<td>(D_{50})</td>
<td>0.819</td>
<td>0.135</td>
<td>6.07</td>
<td>&lt;0.001</td>
<td>1.007</td>
<td>1.007</td>
</tr>
</tbody>
</table>

Width and depth (Figure 5.13) and bankfull discharge (Figure 5.11D) do not have any significant relation with median grain size. It was expected that in normal alluvial settings, wider channel reaches may be related to channel splitting over a coarser bed substrate, but this is not reflected in the data. However, there is a weakly significant relationship with slope indicating some mutual adjustment across a range of channel boundary conditions (Table 5.2 and Figure 5.13).
As expected from normal hydraulic geometry relationships there are significant relations between channel width, depth and bankfull discharge indicating that the tributaries and mainstems together form a coherent hydraulic geometry relationship. For these regressions, the r-squared values are high and the data passes the parametric assumptions indicating that it is log-log normally distributed. Out of all the channel characteristics, slope is best related to depth but the relationship is only moderate (Table 5.2 and Figure 5.12B). Median grain size stands out as the least significant channel characteristic which is partially related to the non-normal distribution that grain size typically exhibits.

Figure 5.13: Regression plots for median grain size vs. width (A) and depth (B).
Table 5.5: Linear Regression Results for \( \log_{10} \) Transformed Median Grain Size (\( D_{50} \)) Data.

<table>
<thead>
<tr>
<th>Independent Variable</th>
<th>a</th>
<th>b</th>
<th>( r^2 )</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>Variance Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width</td>
<td>30.6</td>
<td>0.16</td>
<td>0.02</td>
<td>0.251</td>
<td>0.445</td>
<td>Failed (P=0.033)</td>
<td>Failed (P=0.032)</td>
</tr>
<tr>
<td>Depth</td>
<td>0.58</td>
<td>-0.09</td>
<td>0.007</td>
<td>0.252</td>
<td>0.635</td>
<td>Failed (P=0.018)</td>
<td>Passed (P=0.213)</td>
</tr>
</tbody>
</table>

Sample size (N) is 36 for all regression tests.

5.2.1. Alluvial (Type 1)

It is useful to sub-divide the 36 study reaches into the three boundary conditions types to explore the influence of bank strength and valley conditions on channel form. Type 1 alluvial reaches comprise the majority of the total sample size (n=15), however, overall the log-linear regressions performed on this sub-data set indicate the weakest predicted co-variance between channel characteristics. The relationships between bankfull discharge and width, bankfull discharge and median grain size, slope and median grain size and width and median grain size are the four strongest of the alluvial reaches and are presented in Table 5.5 and Figure 5.14.

The relationship between width and discharge again confirms that the normal hydraulic geometry relationship holds true for this smaller sub-group (Figure 5.14A). However, other expected hydraulic geometry relations, such as depth and discharge, are not reflected in this smaller data set of alluvial reaches. The relationship between grain size and discharge is improved upon, compared to the full 36 reach data set, with an \( r^2 \) value of 0.44 indicating that for these fully alluvial sites width increases with grain size.

The weaker then expected relationships of the alluvial (self-forming) reaches indicates that there are other factors that should be considered when it comes to controls on channel characteristics. For alluvial channels, factors such as channel confinement and variability in
bank strength might affect the ability of the channel to widen and deepen under the current hydrologic regime.

Table 5.6: Linear Regression Results for Log$_{10}$ Transformed Type 1 Alluvial sub-group strongest relationships.

<table>
<thead>
<tr>
<th>Dependant Variable</th>
<th>Independant Variable</th>
<th>a</th>
<th>b</th>
<th>$r^2$</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>Constant Variance Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_b$</td>
<td>Width</td>
<td>4.82</td>
<td>0.45</td>
<td>0.60</td>
<td>0.115</td>
<td>&lt;0.001</td>
<td>Passed (0.340)</td>
<td>Passed (0.146)</td>
</tr>
<tr>
<td>$Q_b$</td>
<td>$D_{50}$</td>
<td>0.007</td>
<td>0.43</td>
<td>0.32</td>
<td>0.195</td>
<td>0.027</td>
<td>Passed (0.447)</td>
<td>Passed (0.209)</td>
</tr>
<tr>
<td>Slope</td>
<td>$D_{50}$</td>
<td>0.29</td>
<td>0.42</td>
<td>0.37</td>
<td>0.187</td>
<td>0.017</td>
<td>Passed (0.495)</td>
<td>Passed (0.257)</td>
</tr>
<tr>
<td>$D_{50}$</td>
<td>Width</td>
<td>126.5</td>
<td>0.51</td>
<td>0.44</td>
<td>0.137</td>
<td>0.007</td>
<td>Passed (0.386)</td>
<td>Passed (0.481)</td>
</tr>
</tbody>
</table>

Sample size (N) is 15 for all regression tests.
Figure 5.14: Regression relationships for Type 1 Alluvial sub-group. (A) discharge vs. width, (B) discharge vs. $D_{50}$, (C) slope vs. $D_{50}$, (D) $D_{50}$ vs. width.
5.2.2. Semi-Alluvial (Type 2)

Semi-alluvial reaches comprise the second largest subset with a sample size of 13. The relationships derived from log-linear regression analysis for semi-alluvial reaches are statistically stronger than what was found for the alluvial reaches. Higher predictability is seen between channel width, channel depth and bankfull discharge with some $r^2$ values higher than 0.8 and with exponents that confirm the normal hydraulic geometry relationships (Table 5.6). A moderate but significant relationship is seen between slope and median grain size with an $r^2$ value below the significance.

**Table 5.7:** Linear Regression Results for $\log_{10}$ Transformed Type 2 Semi-Alluvial sub-group strongest relationships.

<table>
<thead>
<tr>
<th>Dependant Variable</th>
<th>Independant Variable</th>
<th>a</th>
<th>b</th>
<th>$r^2$</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>Constant Variance Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width</td>
<td>Depth</td>
<td>24</td>
<td>0.81</td>
<td>0.68</td>
<td>0.164</td>
<td>&lt;0.001</td>
<td>Passed (0.658)</td>
<td>Passed (0.304)</td>
</tr>
<tr>
<td>Width</td>
<td>$Q_b$</td>
<td>5.7</td>
<td>0.40</td>
<td>0.91</td>
<td>0.084</td>
<td>&lt;0.001</td>
<td>Passed (0.252)</td>
<td>Passed (0.192)</td>
</tr>
<tr>
<td>Width</td>
<td>Slope</td>
<td>1.6</td>
<td>-0.50</td>
<td>0.36</td>
<td>0.23</td>
<td>0.031</td>
<td>Failed (0.002)</td>
<td>Passed (0.332)</td>
</tr>
<tr>
<td>Width</td>
<td>$D_{50}$</td>
<td>951.05</td>
<td>1.23</td>
<td>0.38</td>
<td>0.225</td>
<td>0.024</td>
<td>Passed (0.169)</td>
<td>Passed (0.121)</td>
</tr>
<tr>
<td>$Q_b$</td>
<td>Depth</td>
<td>0.25</td>
<td>0.39</td>
<td>0.83</td>
<td>0.121</td>
<td>&lt;0.001</td>
<td>Passed (0.200)</td>
<td>Failed (0.037)</td>
</tr>
<tr>
<td>$Q_b$</td>
<td>$D_{50}$</td>
<td>0.03</td>
<td>0.16</td>
<td>0.57</td>
<td>0.095</td>
<td>0.003</td>
<td>Passed (0.272)</td>
<td>Passed (0.269)</td>
</tr>
</tbody>
</table>

Sample size (N) is 13 for all regression tests.
Figure 5.15: Regression relationships for Type 2 Semi-Alluvial sub-group. (A) width vs. depth, (B) width vs. discharge, (C) width vs. slope, (D) width vs. grain size, (E) depth vs. discharge, (F) grain size vs. discharge.
5.2.3. Non-Alluvial (Type 3)

The final alluvial type grouping, non-alluvial, has the smallest sample size where n = 8. Despite the small sample size, non-alluvial channels have the highest overall predictable variance between the channel characteristics (Table 5.7 and Figure 5.16). For all variables except for median grain size, the $r^2$ value indicates very high predicted variance. The bankfull discharge vs. median grain size and width vs. slope relationships produce the expected hydraulic geometry exponents (Figure 5.16C, D and E). Median grain size (not shown) does not appear to vary systematically in non-alluvial reaches. This was confirmed by also looking at the coarse end of the distribution ($85^{th}$ percentile, $D_{85}$).

Since sample size is small, the non-alluvial reaches are strongly influenced by a few points (i.e. both the Ausable River tributaries as well as the Fourteen Mile Tributary are noted on Figure 5.16C as having the three steepest slopes). This indicates the importance of sampling broadly to capture southern Ontario variations in channel types.

Table 5.8: Linear Regression Results for Log$_{10}$ Transformed Type 3 Non-Alluvial sub-group strongest relationships.

<table>
<thead>
<tr>
<th>Dependant Variable</th>
<th>Indepenant Variabele</th>
<th>a</th>
<th>b</th>
<th>$r^2$</th>
<th>Standard Error</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
<th>P-Value</th>
<th>Normality Test (Shapiro-Wilk)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope Width</td>
<td></td>
<td>0.5</td>
<td>-0.74</td>
<td>0.97</td>
<td>0.077</td>
<td>&lt;0.001</td>
<td>Passed (0.556)</td>
<td>Passed (0.749)</td>
<td></td>
</tr>
<tr>
<td>Slope Depth</td>
<td></td>
<td>0.03</td>
<td>-0.60</td>
<td>0.95</td>
<td>0.081</td>
<td>&lt;0.001</td>
<td>Passed (0.726)</td>
<td>Passed (0.160)</td>
<td></td>
</tr>
<tr>
<td>$Q_b$ Slope</td>
<td></td>
<td>0.05</td>
<td>-0.64</td>
<td>0.96</td>
<td>0.123</td>
<td>&lt;0.001</td>
<td>Passed (0.165)</td>
<td>Passed (0.165)</td>
<td></td>
</tr>
<tr>
<td>$Q_b$ Width</td>
<td></td>
<td>4.4</td>
<td>0.50</td>
<td>0.97</td>
<td>0.072</td>
<td>&lt;0.001</td>
<td>Passed (0.487)</td>
<td>Passed (0.619)</td>
<td></td>
</tr>
<tr>
<td>$Q_b$ Depth</td>
<td></td>
<td>0.21</td>
<td>0.40</td>
<td>0.99</td>
<td>0.019</td>
<td>&lt;0.001</td>
<td>Passed (0.341)</td>
<td>Passed (0.072)</td>
<td></td>
</tr>
<tr>
<td>Depth Width</td>
<td></td>
<td>31.9</td>
<td>1.21</td>
<td>0.96</td>
<td>0.09</td>
<td>&lt;0.001</td>
<td>Passed (0.680)</td>
<td>Passed (0.160)</td>
<td></td>
</tr>
</tbody>
</table>

Sample size (N) is 8 for all regression tests.
Figure 5.16: Strongest regression relationships for Type 3 Non-Alluvial sub-group. (A) width vs. slope, (B) depth vs. slope, (C) slope vs. discharge, (D) width vs. discharge, (E) depth vs. discharge, (F) width vs. depth.
5.3. Predicting Changes in Channel Morphology Downstream from a Confluence

Roy and Woldenberg (1986), used equation 2.14 to predicted mainstem channel morphologies, such as width, depth and slope, downstream from a confluence. In order to use eq. 2.14, the ‘x’ and ‘x’ exponents must first be determined. The ‘x’ exponent in the equation is determined as the exponent in a power relation between the predicting variable (e.g. width) and the bankfull discharge to which it occurs for a set of reaches along a river system. In this instance, since the compiled data set of 36 study reaches complies well with normal hydraulic geometry relations (Figure 5.11), the exponent for width (b = 0.44) from this relation can be substituted into equation 2.14. The reciprocal exponent ‘x’ for width is therefore 2.27. The channel attribute downstream of the confluence can then be calculated using the upstream tributary pairs. Figure 5.17 shows a sample calculation for width using the East and West Duffins Creeks tributaries. Note that the Roy and Woldenberg (1986) relation modestly under predicts width at bankfull.

**Figure 5.17:** Example of mainstem width calculation using Roy and Woldenberg (2011).

<table>
<thead>
<tr>
<th>Bankfull Width (m)</th>
<th>Bankfull Discharge (m$^3$/s)</th>
<th>$x$</th>
<th>$I/x$</th>
<th>Calculated Mainstem Width (m)</th>
<th>Observed Mainstem Width (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>West Duffins</td>
<td>16.2</td>
<td></td>
<td></td>
<td>2.27</td>
<td>21.3</td>
</tr>
<tr>
<td>East Duffins</td>
<td>15.3</td>
<td>0.44</td>
<td></td>
<td>27.5</td>
<td>27.5</td>
</tr>
</tbody>
</table>
The predicted downstream characteristics of average width, depth and slope was made for all 12 study confluences using the Roy and Woldenberg (1986) method employing the hydraulic geometry exponents for the aggregate dataset. Figure 5.18 shows results compared to the actual observed measurements.

Figure 5.18: Observed versus predicted values (from Roy and Woldenberg, 1986) for average bankfull width, depth, and slope in all mainstem study reaches downstream of confluences.
As shown in Figure 5.18A and B, the predicted mainstem widths and depths using Roy and Woldenberg (1986) produce reasonably good predictions, whereas predicated slope (Figure 5.18C) tends to be lower in comparison to observed values, especially when considering gentler slopes. The absence of a perfect 1:1 prediction indicates the role of other factors such as channel confinement and bank strength as potentially important. In the case of slope, it may relate to the presence of coarser channel substrates (e.g. till beds) that reduce down-cutting and slope reduction and therefore the under prediction.

5.3.1. Alluvial Type Category Comparisons

When the study reaches are sub-divided into the three alluvial types based on boundary conditions of the observed downstream mainstem, the predicted channel characteristics of width, depth and slope show some interesting patterns (Figure 5.19). The distribution of all alluvial type channel widths is fairly evenly distributed along the line of perfect prediction (Figure 5.19A). However, Figure 5.19A also hints at consistent under prediction of the larger fully alluvial rivers which may be related to lack of available valley bottom room (i.e. confinement) in the mainstem of alluvial rivers downstream of the confluence. Figure 5.19B shows that there are some differences in the predicted depths of alluvial and semi-alluvial channels, with alluvial channels displaying a moderate over prediction and semi-alluvial channels showing an even distribution across the 1:1 line. The over prediction of depth in alluvial channels may indicate the presence of resistant (non-alluvial) channel beds in some alluvial reaches limiting down-cutting. The predicted channel slope (Figure 5.19C) does not show a difference between the boundary condition types, however downstream channel slopes for alluvial reaches are generally under predicted by the Roy and Woldenberg (1986) method.
Figure 5.19: Observed versus predicted values for average bankfull width, depth, and slope in all mainstem study reaches downstream of confluences.
Chapter Six: Discussion and Conclusions

6. Discussion and Conclusions

6.1. Discussion

6.1.1. Equilibrium Downstream from a Confluence?

Channels that are experiencing a dynamic equilibrium, one where there is no abrupt aggradation or degradation of the bed, may acquire this state due to a balance where the channel has adjusted its slope in accordance with the available discharge to transport a particular sediment size and sediment load supplied to it from upstream (Makin, 1948). In other words, for a river reach in equilibrium, the transport capacity of the discharge is in balance with the rate of supply. By this definition, rivers that experience migration through meander bend erosion and point bar formation can be considered to be in equilibrium (Chang, 1986). Equilibrium in channels can be disrupted over various time periods due to changes in climate, hydrology or tectonics. Over a decadal time scale, changes in the overall hydrology of river systems can dramatically influence the equilibrium of a river channel by forcing the observed balance in river channel characteristics (those of the Lane 1955 relation: sediment discharge, water discharge, grain size and slope) to cross a “threshold”. Therefore if one of the variables in question becomes altered, one or more of the other variables must adjust in order to restore equilibrium to the channel (Lane, 1955). Schumm (1969) introduced a process-response model of channel adjustments for different combinations of changing discharge and sediment rate. For alluvial rivers an increase in discharge is expected to lead to an increase in channel width and depth as well as a decrease in channel slope (Table 6.1) (Schumm, 1969).
Table 6.1: Qualitative Model of Channel Metamorphosis by Schumm (1969) (From: Chang, 1986). Where $B =$ channel width; $D =$ channel depth; $V =$ mean velocity; $Q =$ bankfull discharge; $L =$ meander wavelength; $S =$ channel slope; $F =$ width-depth ratio; $P =$ sinuosity.

<table>
<thead>
<tr>
<th>Increase in discharge alone</th>
<th>Decrease in discharge alone</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q^+ \sim B^+ D^+ F^+ L^+ S^-$</td>
<td>$Q^- \sim B^- D^- F^- L^- S^-$</td>
</tr>
<tr>
<td>Increase in bed material discharge</td>
<td>Decrease in bed material discharge</td>
</tr>
<tr>
<td>$Q^+_i \sim B^+ D^- F^+ L^+ S^+ P^-$</td>
<td>$Q^-_i \sim B^- D^+ F^- L^- S^- P^+$</td>
</tr>
</tbody>
</table>

Discharge and bed material discharge increase together, e.g., during urban construction, or early stages of afforestation

$Q^+_i Q^+_i \sim B^+ D^+ F^+ L^+ S^+ P^-$

Discharge and bed material discharge decrease together, e.g., downstream from a reservoir

$Q^- Q^-_i \sim B^- D^+ F^- L^- S^- P^+$

Discharge increases as bed material discharge decreases, e.g., increasing humidity in an initially sub-humid zone

$Q^+ Q^-_i \sim B^+ D^- F^- L^- S^- P^+$

Discharge decreases as bed material discharge increases, e.g., increased water use combined with land-use pressure

$Q^- Q^+_i \sim B^+ D^- F^+ L^- S^+ P^-$

The abrupt increase in discharge and sediment input at river channel confluences, the focus of this thesis, has the potential to impact equilibrium of the channel and alter the observed channel morphology from what is viewed along the upstream contributing tributaries. Using the qualitative model for increased discharge as a guide, and disregarding channel boundary conditions, very few of the downstream mainstem channels studied in this research project strictly follow the adjustment that might be expected by Schumm (1969) (see Table 3.1). However, the channel change that occurs consistently in the downstream study reaches (except for along the Ausable River and Don River) is a predictable scaled increase in channel width with the proportional increase in discharge below the confluence. For the Ausable River, the
relatively small tributaries flowing into the mainstem appear to have no impact on the channel morphology. Although only the two Ausable River confluences of this nature were selected for study, it appears from this study that lesser tributaries must contribute at least one third or more of the discharge at the confluence to have a measureable effect along the downstream mainstem channel. For the Don River, the majority of downstream adjustment to increasing discharge appears to be taken up by increasing depth. A total of three downstream mainstem channels experience an increase in width and depth and a decrease in slope: they are Credit River at 21 km, Mad River and Sixteen Mile Creek. It is interesting to note that these three confluence systems are not all in the alluvial category and are influenced by bedrock and a larger median, bed sediment, grain size.

Adjustments to slope in the downstream mainstems are less obvious compared to adjustments in width and depth. A total of 4 mainstems see an increase in slope downstream from the confluence, 4 mainstems produce a slope that is steeper than the dominant tributary but gentler than the lesser tributary, and 4 mainstems produce a slope that is gentler then both or equal to the dominant tributary. For the four mainstems with a steeper slope, there is also an associated increase in median grain size (see Appendix A).

Based on field observations (Table 3.1, Appendix A), the downstream mainstems of the Ausable River (x2), Catfish Creek, Credit River (x2), Humber River at 56 km, Mad River and Sixteen Mile Creek appear to have adjusted to a state of equilibrium. These reaches represent a mixture of the three boundary types. The majority of these mainstems (six of eight), plot below the Kleinhans and van den Berg (2011) threshold for moderately braided ($w_{pv} = 285 \ D_{50}^{0.42}$), indicating a more laterally stable channel. However only three of the eight mainstems plot along or below the Leopold and Wolman (1957) threshold line, therefore Kleinhans and van den Berg
(2011) offers a slightly better prediction for these channels. Mainstems that appear to not be in a state of equilibrium are the Don River, Duffins Creek, Fourteen Mile Creek and Humber River at 18 km. These mainstems plot above the Leopold and Wolman (1957) threshold line, but not necessarily plot significantly higher than the other channels on the Kleinhans and van den Berg (2011) plots.

When all study reaches are viewed together, there is a strong indication that the mainstem channels tend to plot in the direction of lateral instability downstream of confluences. Therefore it can be assumed that even though the mainstems may be in a state of equilibrium currently, they are sufficiently close to the thresholds as defined by Leoplod and Wolman (1957) and Kleinhans and van den Berg (2011), that any change along the upstream tributary reaches could disrupt the observed balance.

6.1.2. Predicting Post-Confluence Morphology Changes

For analysis in this thesis, the bankfull discharge was considered to be close to or equivalent to the two-year flow \( (Q_2) \) of the channels studied (cf. Phillips, 2012). Bankfull discharge was adopted as the discharge that has the most influence on channel morphology (Navratil, et al., 2006). Site-specific bankfull discharge was inferred from channel area dimensions at the top-of-bank inflection along multiple channel cross-sections of a reach. Bankfull discharge is the preferred flow level to use in fluvial geomorphology research because it allows for the comparison of river reaches within the same fluvial system as well as other watersheds (Wolman and Miller, 1960). Using hydraulic geometry relations developed by Leopold and Maddock, (1953), bankfull discharge can be correlated to channel width, depth,
velocity and slope. Therefore, the increasing bankfull discharge downstream from a confluence can be used to make predictions of expected downstream channel dimensions.

In terms of matching observed versus predicted channel pattern and stability in tributaries and mainstems, Kleinhans and van den Berg (2011) method is generally more superior. This is due to the fact that stream power better represents the energy to move sediment masses (rather than energy needed to entrain individual particles) that control erosion and thus channel form. Stream power takes into account the combined influence of both discharge and slope on bed sediment size and thus better reflects the fundamental controlling influences on river pattern first described by Lane (1955) (Equation 2.1). There is still some utility in using the Leopold and Wolman (1957) method as it more closely reflects the fundamental hydraulic geometry control of discharge. While bank strength was not directly measured for this research, it was considered indirectly in the prediction of downstream changes in channel morphology through the subdivision of channel boundary types. Eaton et al. (2010) and Eaton and Giles, (2009) indicate that channel pattern will vary depending on the association between critical slope, discharge and relative bank strength. The qualitative subdivision of study reaches into alluvial, semi-alluvial and non-alluvial based on field observations was done in an attempt to identify significant variations in how downstream mainstem morphology changes. Bedrock confined reaches were originally considered to be laterally stable due to the assumption that channel bed and banks would be less erodible then the fully alluvial channels that are susceptible to major change as alluvium is eroded, transported and deposited (Schumm, 1985).
6.1.3. Confluence Morphology

All study confluences appear to show some variation of morphological characteristics as described by Kenworth and Rhoads (1995) and Best (1986). With the exception of the Ausable River where tributary inflows are minor, each confluence displayed a deep scour hole oriented to the centre of the downstream channel. Credit River at 21 km and Humber River at 18 km have the best examples of tributary mouth bars. For these two confluences, tributary angle is observed to be greater than 90° with the lesser tributary inputting a significantly larger median grain size. Channels where post-confluence mid-channel bars have formed are Duffins Creek and Humber River at 18 km. For the twelve confluence sites studied, downstream bar formation is observed when there is a distinct difference in tributary size ($Q_r > 1$). As $Q_r$ approaches 1, bar formation is less pronounced and may result from the fact that the near equal transporting capabilities of each tributary produce less backwater effects during flooding, more equal grain size inputs and therefore more persistent movement of sediment through the downstream confluence.

6.1.4. Interpretation of Results

In an attempt to adequately predict channel reach morphology two methods were used. The Leopold and Wolman (1957) relationship between slope and discharge was used in order to predict channel morphology as either laterally stable or laterally unstable. When all 36 study reaches are viewed as if they are independent of each other, there is a tendency for the reaches to plot either above or along the threshold line. In total 23 of the 36 reaches plot in accordance with observed morphologies. The Kleinhans and van den Berg (2011) stream power and grain size discrimination thresholds gives insight into channel stability based on bar patterns as a prediction
for meandering or braided (laterally stable, laterally unstable). Using this method, there is a slightly higher predictability with 28 of the 36 reaches plotting in accordance with observed morphologies.

When boundary conditions are incorporated, it is expected that fully alluvial channels should be more predictable in terms of channel pattern because they are “self-forming”. Based on both the Leopold and Wolman (1957) and Kleinhans and van den Berg (2011) alluvial (Type 1) grouping plots 14 of the 15 alluvial channels plot in accordance with observed morphologies. Mainstem alluvial channels that plot higher (Duffins Creek, Humber River at 18 km, Don River), are interpreted to be in a state of non-equilibrium and adjustment. When semi-alluvial and non-alluvial groupings are concerned the predictability of channel form decreases. The Leopold and Wolman (1957) approach predicts all non-alluvial (Type 3) channels to be laterally unstable, whereas Kleinhans and van den Berg (2011) approach predicts them to be meandering with scrolls. It was expected that the non-erodible bedrock banks would produce more single thread channel pattern as described by Schumm, (1985) but all exhibit reach morphologies with bars and chute channels indicating greater instability. The observed lateral stability/instability of semi-alluvial channels as predicted by the two methods has the overall weakest predictability with only two of thirteen accurately predicted by Leopold and Wolman (1957) and eight of thirteen accurately predicted by Kleinhans and van den Berg (2011). This is very significant in the southern Ontario context where many river reaches, and in some cases entire river networks, flow through semi-alluvial boundary conditions.

The hydraulic geometry of the aggregate data set of 36 channel reaches is not only in accordance with generally expected hydraulic geometry relationships (cf. Leopold and Maddock, 1953), but the results are also directly comparable to previous hydraulic geometry relations.
developed for Ontario streams (Annable, 1996). This indicates that the study reaches form a coherent hydraulic geometry data set in which predicted width and depth are proportional to changes in discharge. When assessed across three distinct categories of boundary materials, the hydraulic geometry results are somewhat consistent. For alluvial reaches (Type 1), only the relationship between width and discharge is observed to conform to expected results. Factors such as variability in channel confinement and bank strength might affect the ability of the alluvial channels to widen and deepen predictably under the current hydrologic regime. Semi-alluvial and non-alluvial channels, however, generally have non-erodible beds and thus may widen and deepen with increasing discharge in a somewhat more predictable pattern.

The Roy and Woldenberg (1986) method to predict changes in post-confluence width, depth and slope appear to produce adequate results. Channel width and depth of the downstream mainstem channel are predicted to a high degree, however slope is under predicted. Predicting downstream mainstem characteristics based on boundary type indicates that alluvial channels (Type 1) are more difficult to predict in terms of changes to width, depth and slope. However, both semi-alluvial and non-alluvial reach predictions generally fall closer to the 1:1 line indicating somewhat greater predictability. Again, this may reflect higher variability in the bed and bank materials.

Results from this study indicate that river channel confluences can produce significant but predictable changes in downstream channel morphologies. Confluences represent significant points of disruption in the transport of sediment and sometimes major adjustments in slope, width and depth of the channel. An accurate prediction of post-confluence channel morphology is needed in order to identify locations of potential non-equilibrium and assess if mainstem channels adjust quickly, and as predicted, to changing contributions from upstream tributaries.
This is particularly important in areas that are undergoing rapid land use change (e.g. GTA watersheds) where one tributary watershed might be undergoing dramatic change compared to the other joining tributary.

6.2. Conclusions

This thesis was an attempt to reduce some of the complexity associated with different boundary types and look for definable and predictable channel morphologies in selected confluence sites. Many of the existing confluence studies focus on flow dynamics (e.g. turbulence) and channel scour/deposition processes over a small area. What has been missing is more insight to the channel morphology, bank strength and hydraulic geometry changes at confluences. As well, confluences represent points of abrupt changes in discharge and thus adjustments in channel form downstream of the confluence could be considered a surrogate for climate change impacts on river systems. Based on the information and results presented in this thesis, conclusions can be drawn about the changing morphologies downstream from a confluence:

1. Rivers in southern Ontario are not homogenous self-forming alluvial rivers with highly predictable channel boundaries even at reach scale.

2. Glaciation and post-glacial entrenchment has introduced a heterogeneous mixture of alluvial, semi-alluvial, and non-alluvial channel reaches, which can sometimes occur within a single river reach.

3. When comparing select boundary condition types there are no overall differences in how mainstem channels adjust their morphologies. However certain characteristics are common between the boundary types; for channels that are fully alluvial,
mainstems tend to be wider with an intermediate or steeper slope. For channels that have a disruption in alluvial nature (i.e. bedrock influence) mainstems tend to be wider and deeper with an intermediate slope.

4. This limited sample of rivers generally shows that southern Ontario is dominated by reaches at or above the threshold for lateral instability. Other factors come into play that may limit the ability of a channel to anabranch or become extensively braided despite the stream power and median grain size conditions that indicate they should be. Channel/valley confinement and long term (Holocene) vertical entrenchment are believed to be limiting the ability of channels to braid.
References


Appendix A: Compiled Channel Reach Characteristics
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<th>Average Velocity (m/s)</th>
<th>w/d ratio</th>
<th>X-Section Area (m$^2$)</th>
<th>Wetted Perimeter (m)</th>
<th>Hyrdraulic Radius</th>
<th>Bankfull Discharge (m$^3$/s)</th>
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