Quantifying Dominant Heat Fluxes in an Arctic Alaskan River with Mechanistic River Temperature Modeling

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QUANTIFYING DOMINANT HEAT FLUXES IN AN ARCTIC ALASKAN RIVER WITH MECHANISTIC RIVER TEMPERATURE MODELING

by

Tyler V. King

A dissertation submitted in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

Civil and Environmental Engineering

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2018
ABSTRACT

Quantifying Dominant Heat Fluxes in an Arctic Alaskan River with Mechanistic River Temperature Modeling

by

Tyler V. King, Doctor of Philosophy
Utah State University, 2018

Major Professor: Dr. Bethany T. Neilson
Department: Civil and Environmental Engineering

River temperatures exert a primary control on physical, chemical, and biological processes, which impact water quality, habitat suitability, and nutrient cycling. While the fundamental processes that influence river temperatures are the same across geographic regions, their relative importance varies significantly. Little is known about the processes that control water temperature in Arctic rivers. This lack of knowledge limits our ability to quantify the impacts of climate change on river temperatures in a region where changes in climate are most strongly manifested. This dissertation addresses this knowledge gap by incorporating detailed field measurements with mechanistic river temperature modeling to estimate the relative importance of heat fluxes affecting temperatures in an Arctic river underlain by continuous permafrost. Results indicate that shortwave radiation and net longwave radiation are significant heat fluxes for the entire length of the river under a wide range of flow conditions. In the headwater portion of the
watershed, however, heat fluxes associated with hyporheic exchange and lateral inflows become significant under low-flow and high-flow conditions, respectively. Additional field observations and modeling of a lower order reach were used to quantify the heat fluxes associated with hyporheic exchange, which were found to reduce the diel amplitude in river temperature fluctuations. As lateral inflows were observed to vary spatially, high-resolution aerial imagery was used to develop a method for estimating river discharge, and, therefore, lateral inflows, remotely. This method has similar accuracy as in-situ gauging stations and alternative remote sensing approaches, presenting a potential solution to barriers in performing river temperature modeling in other parts of the Arctic. In total, the work presented in this dissertation is a foundation upon which the impacts of climate change on Arctic river temperatures can be assessed.

(205 pages)
PUBLIC ABSTRACT

Quantifying Dominant Heat Fluxes in an Arctic Alaskan River with Mechanistic River Temperature Modeling

Tyler V. King

Temperatures strongly affect physical, chemical, and biological processes in rivers and streams. The processes that influence river temperatures are known across most geographic regions, but the relative importance varies significantly. Little is known about what controls water temperature Arctic rivers, limiting our ability to understand the impacts of climate change. This dissertation addresses this knowledge gap by incorporating field measurements with river temperature modeling to estimate the relative importance of key factors that affect Arctic river temperatures. Results indicate that shortwave radiation (e.g., sunlight) and net longwave radiation are significant throughout an Arctic watershed in all flow conditions. In areas where the river is smaller, however, exchange of water with the riverbed and inputs of water from the landscape become significant under low-flow and high-flow conditions, respectively. Additional field observations and modeling were used to quantify the water and heat exchanges between the river and the riverbed. These heat exchanges were found to cool the river and reduce the daily range of temperatures. To better estimate the flow of water from the landscape to the river, a new method for estimating river flow was developed using high-resolution aerial imagery. This method allows us to estimate river flow without depending on field measurements, and presents a potential solution to barriers in performing river temperature modeling in other parts of the Arctic.
ACKNOWLEDGMENTS

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CHAPTER 1

INTRODUCTION

Water temperature is a primary water quality parameter for rivers and streams given its influence on instream physical [Arp et al., 2012a; Pohl et al., 2007; Rawlins et al., 2010; Syvitski, 2002], biological [Deegan et al., 1999; Mohseni et al., 1998], and chemical [Cory et al., 2013; McNamara et al., 2008] processes. The importance of water temperature as a water quality metric has led to extensive research on the mechanisms that control river temperature [Arscott et al., 2001; Brown, 1969; Hannah et al., 2008a; Mohseni et al., 1999; Sinokrot and Stefan, 1993; Webb et al., 2003; Webb et al., 2008]. Mechanistic, or process focused, studies are advantageous to statistical approaches [Arismendi et al., 2014; Kelleher et al., 2012] in a predictive capacity, making them preferable for determining effects of climate change.

River temperatures are controlled by heat fluxes across the air-water interface, the water-sediment interface, and heat and mass loading from groundwater and lateral inflows [Brown, 1969; Caissie, 2006; Webb and Zhang, 1997; Webb et al., 2008]. While the same general heat fluxes are present across a broad range of regions around the world, the importance of various heat fluxes vary significantly based on local conditions, creating the need for regional or local assessment. Components of river energy budgets have been quantified in forested catchments [Benyahya et al., 2012; Brown, 1969; Hebert et al., 2011; Johnson, 2004; Leach and Moore, 2010; Sinokrot and Stefan, 1993], proglacial rivers [Cardenas et al., 2014; Chikita et al., 2010; Khamis et al., 2015; Magnusson et al., 2012], hot and cold desert environments [Cozzetto et al., 2006; Neilson
et al., 2010a], moorland streams [Evans et al., 1998; Garner et al., 2014; Hannah et al., 2004; 2008b; Webb and Zhang, 2004], and coastal catchments [Moore et al., 2005]. However, analysis of dominant heat fluxes in Arctic regions is limited [Blaen et al., 2013; Merck et al., 2012], and a process based understanding of these controls has not been investigated thoroughly in part because of the remote nature of the Arctic. Understanding the processes controlling arctic river temperatures is imperative in order to predict the influence of extensive changes in Arctic climate [Hinzman et al., 2005; Houghton et al., 2001; Mann et al., 2010; Serreze et al., 2000].

While arctic river temperatures are influenced by the same basic heat and mass fluxes as their temperate counterparts, the relative importance and variability of these fluxes have not been well quantified for Arctic systems and may be unique. For example, subsurface flow within Arctic basins is limited to supra-permafrost flow through seasonally thawing soils known as the active layer [McNamara et al., 1997; McNamara et al., 2008]. The shallow nature of the active layer (less than 1 m) minimizes landscape storage. Runoff generation is therefore dominated by shallow hillslope drainage processes that have rapid initial response times, high runoff/precipitation ratios, and extended recessions [McNamara et al., 1997; 1998; Zhang et al., 2000]. This essentially removes deeper, regional groundwater influences on river temperature which have been shown to be significant in temperate rivers [Erickson and Stefan, 2000; Mohseni et al., 1999].

Additionally, heat fluxes between Arctic rivers and their surroundings are anticipated to be unique in some aspects given the thermal and meteorological
composition of the Arctic environment. The presence of permafrost below thawed streambed sediments imposes a zero degree boundary in close proximity to the main channel that may produce strong temperature gradients through the sediment. These thermal gradients could enhance conductive and convective bed heat fluxes from the river into the subsurface during the warm season. The potential for permafrost to act as a heat sink presents the potential for fundamentally unique river energy balances when compared with temperate rivers. Meteorological conditions are also known to be unique for high latitude regions. Chief among these are cool summer air temperatures, long periods of solar radiation, and high relative humidity. These atmospheric conditions may create Arctic specific significance among surface heat fluxes.

The objectives of this dissertation were to fill this knowledge gap by identifying the dominant heat fluxes in Arctic rivers. Through detailed data collection and modeling, the significance of many heat fluxes were quantified allowing for identification of dominant heat fluxes. This provides a foundation for scenario testing to determine the sensitivity of river temperatures to changes in climate and associated landscape and hydrologic responses. Understanding the controls on river temperatures helps to identify the minimum data required to accurately predict river temperature, which is of particular interest in remote Arctic basins where data collection is logistically challenging and expensive.

This work starts with the application of a basic river temperature model applied in two distinct portions of the Kuparuk River, Alaska (Chapter 2). Dominant heat fluxes were identified for the headwater and coastal portions of the watershed based on
processes known to be significant in temperate rivers. Longitudinal and temporal trends in these heat fluxes were identified. Model performance was used to identify locations and hydrologic conditions for which model assumptions were reasonable, and where additional model refinement was required.

Initial modeling results over estimated diel temperature changes in the headwater portion of the watershed under low-flow conditions. This led to a chapter focusing on refining river temperature modeling for this region through the inclusion of hyporheic exchange (Chapter 3). In this work, I built upon the understanding of hyporheic exchange processes as a mechanism for buffering river temperatures in temperate systems [Arrigoni et al., 2008] and knowledge of hyporheic exchange in Arctic rivers [Edwardson et al., 2003] to determine the associated thermal influences on river temperatures in the lower order portion of the Kuparuk River under low-flow conditions.

The sensitivity of river temperatures to net heat fluxes are directly related to river discharge [Gu et al., 1998; Webb et al., 2003]. As such, accurate estimates of discharge are critical for river temperature modeling. Access to remote field sites on Alaska’s North Slope has historically limited hydrologic studies to either headwater basins (e.g., Imnavorit Creek, Oksrukuyik Creek, Toolik Lake Inlet Stream, Upper Kuparuk) or outlets of entire watersheds (e.g., Kuparuk, Ob, Lena, Putuligayuk, Fish Creek watersheds) where hydrologic and thermal signals are integrated over large spatial and temporal scales. This sparse network of in-situ gauging stations poses a significant barrier to river temperature modeling in Arctic basins. To lower this barrier, a method was developed to estimates river discharge from aerial imagery (Chapter 4). This method is advantageous
in that it can be applied at many locations along a river to estimate lateral inflows. This method also complements satellite based remote sensing techniques by providing estimates of river discharge for smaller rivers that are undetectable in satellite imagery.
CHAPTER 2

WATER TEMPERATURE CONTROLS
IN LOW ARCTIC RIVERS

Abstract

Understanding the dynamics of heat transfer mechanisms is critical for forecasting the effects of climate change on arctic river temperatures. Climate influences on arctic river temperatures can be particularly important due to corresponding effects on nutrient dynamics and ecological responses. It was hypothesized that the same heat and mass fluxes affect arctic and temperate rivers, but that relative importance and variability over time and space differ. Through data collection and application of a river temperature model that accounts for the primary heat fluxes relevant in temperate climates, heat fluxes were estimated for a large arctic basin over wide ranges of hydrologic conditions. Heat flux influences similar to temperate systems included dominant shortwave radiation, shifts from positive to negative sensible heat flux with distance downstream, and greater influences of lateral inflows in the headwater region. Heat fluxes that differed from many temperate systems included consistently negative net longwave radiation and small average latent heat fluxes. Radiative heat fluxes comprised 88% of total absolute heat flux while all other heat fluxes contributed less than 5% on average. Periodic significance was seen for lateral inflows (up to 26%) and latent heat (up to 18%) in the lower and higher stream order portions of the watershed respectively. Evenly distributed lateral inflows from large scale flow differencing and temperatures from representative tributaries provided a data efficient method for estimating the associated heat loads. Poor

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1Coauthored by Tyler V. King, Bethany T. Neilson, Levi D. Overbeck, and Douglas L. Kane
model performance under low flows demonstrated need for further testing and data collection to support inclusion of additional heat fluxes.

Introduction

Given the influence of river temperature on instream physical [Arp et al., 2012a; Pohl et al., 2007; Rawlins et al., 2010; Syvitski, 2002], biological [McNamara et al., 2008; Mohseni et al., 1998], and chemical [Cory et al., 2013; McNamara et al., 2008] processes, and the severity of changes in arctic climate [Hinzman et al., 2005; Houghton et al., 2001; Mann et al., 2010; Serreze et al., 2000], it is imperative to identify and quantify the processes controlling arctic river temperatures. While mechanisms that control river temperature in temperate climates have been the focus of extensive research [e.g., Arismendi et al., 2014; Arscott et al., 2001; Brown, 1969; Hannah et al., 2008a; Mohseni et al., 1999; Sinokrot and Stefan, 1993; Webb et al., 2003; Webb et al., 2008], understanding of controls on river temperature in the Arctic is limited [Blaen et al., 2013; Lammers et al., 2007; Merck et al., 2012; Yang et al., 2014] and a process based understanding of these controls has not been investigated thoroughly.

River temperatures are controlled by heat fluxes across the air-water interface, the water-sediment interface, and heat and mass loading from groundwater and tributaries [Caissie, 2006; Webb and Zhang, 1997; Webb et al., 2008]. While arctic river temperatures are likely influenced by the same basic heat and mass fluxes as their temperate counterparts, the relative importance and variability of these fluxes are
expected to be unique for arctic systems. For example, subsurface flow within the basin is limited to supra-permafrost flow through seasonally thawing soils known as the active layer [McNamara et al., 1997; McNamara et al., 2008]. The shallow nature of the active layer (less than 1 m in the hillslope) minimizes hillslope storage. Runoff generation is therefore dominated by shallow hillslope drainage processes that have rapid initial response times, high runoff/precipitation ratios, and extended recession [McNamara et al., 1997; 1998; Zhang et al., 2000].

Hillslope drainage contributes significantly to arctic hydrology through a variety of stream morphologies including water tracks, alluvial/colluvial streams, and beaded streams [McNamara et al., 1999; Zhang et al., 2000]. These drainage networks have unique hydrologic and thermal responses over time and space [Arp et al., 2012b; McNamara et al., 1997; 1998; 1999; Merck et al., 2012; Zhang et al., 2000]. Determining the appropriate treatment of hillslope contributions or lateral inflow heat/mass loading (defined here as both surface and subsurface flow into a river) is complicated by the diversity of hillslope drainage or stream morphologies and the absence of steady hydrologic conditions where base flow characteristics could be assessed [Arp et al., 2012b; Kane et al., 2000; Kane et al., 2008; McNamara et al., 1998]. As a result, it is not possible to apply common assumptions about lateral inflows (e.g., dominated by groundwater, assumed steady or insignificant) or to measure spatial influences directly. Identifying the influence of lateral inflows on river temperature is a key component for understanding the implications of changes in arctic hydrology, the latter of which has been the focus of significant research [McClelland et al., 2004;
Only with an understanding of the combined hydrologic and climatological influences will it be possible to evaluate the impact of future climate change on arctic river temperatures. Following the work of Smith and Lavis [1975], Polehn and Kinsel [2000], Arscott et al. [2001], Caissie [2006] and others, we anticipated certain hydrologic conditions would produce lateral inflows that would rival surface heat flux influences on river temperature in the lower order drainages of the watershed, but that surface heat fluxes would dominate energy exchange in the higher order portion of the watershed. To test this, we collected data over two open water seasons for the Kuparuk River, Alaska, USA to support the development of a dynamic stream temperature model. The model included heat fluxes known to be significant for temperate systems and was used to determine the mechanisms controlling temperature dynamics throughout the basin. Identifying dominant heat fluxes, and therefore minimum data required to accurately predict river temperature, is of particular interest in these remote low arctic basins where data collection is logistically challenging and expensive.

Site Description

The Kuparuk River watershed, located approximately 250 km north of the Arctic Circle, drains 8140 km² of the Alaskan North Slope from the foothills of the Brooks Range to the Beaufort Sea (Figure 2-1A). A narrow transect across a physiographic gradient, the watershed ranges in elevation from 0 to approximately 1500 meters above
Figure 2-1. Kuparuk River watershed with meteorological data collection locations (red triangles) and significant channels (A). Main channel (red circles) and tributary (T) (blue circles) data collection sites are shown in lower (B) and upper (C) model domains. The location of main channel sites in river kilometers from site 1 (0 km) are shown in parenthesis.

sea level with an average elevation of 210 meters above sea level and an average slope along the Kuparuk River of 2.6%. Along the elevation gradient from the headwater to the coastal plain, river discharge increases, bed slope decreases, median sediment size decreases (7 cm [Oatley, 2002] to 2 cm [Best, 2002]), bed material sorting transitions from poorly sorted to well sorted, channel hydraulic geometry relationships shift due to change in ice dynamics [McNamara and Kane, 2009], dominant hillslope drainage
mechanisms shift from runoff [Zhang et al., 2000] to evapotranspiration [Rovansek et al., 1996], permafrost thickness increases from 250 m to 600 m thick [Osterkamp and Payne, 1981], and land-cover transitions from sedge dominated to moist acidic tussock tundra to wet tundra dominated by rich fens [Nelson et al., 1997; Walker et al., 1989].

The Kuparuk River watershed is classified as an arctic desert receiving approximately 320 mm of precipitation annually with approximately 38% falling as snow [McNamara et al., 1997]. Underlain by continuous permafrost, the watershed is isolated from deep groundwater flow paths [Kane et al., 2013], and subsurface flow is limited to the active layer where soils generally thaw 0.2 – 0.6 m below the ground surface during the summer depending on local conditions such as soil type, vegetation, and topography [Hinzman et al., 1998; Hinzman et al., 1991; McNamara et al., 1997; Nelson et al., 1997].

Drainage networks within the arctic generally shift from colluvial to beaded to alluvial morphologies with increasing drainage area [Arp et al., 2015]. The shallow active layer promotes the development of colluvial preferential flow paths called watertracks in regions with sufficient topographic relief [Arp et al., 2012b; Arp et al., 2015; Zhang et al., 2000]. Watertracks transport water along relatively straight plan forms from steep hillslopes to valley bottoms where flow often becomes spatially distributed and may become shallow subsurface flow before entering streams [McNamara et al., 1998]. Beaded headwater streams are comprised of a series of pools connected by shallow chutes [Merck and Neilson, 2012] and are found throughout the Kuparuk river watershed. Beaded streams and water tracks drain into alluvial rivers (e.g.,
the Kuparuk River) that have meandering plan forms, steep cut banks, gravel point bars, and increasingly frequent braids with distance downstream. The Toolik River follows this trend and is the largest tributary within the Kuparuk watershed, draining 35% of the watershed (Figure 2-1A and 2-1B).

Two long-term stream gaging stations exist on the main stem of the Kuparuk River. The U.S. Geological Survey (USGS) maintains a gage approximately 15 km from the Arctic Ocean (USGS station 15896000) at a location that drains 8140 km² (Site 1 in Figure 2-1B) while the University of Alaska Fairbanks maintains a gaging station 309 km farther upstream where the Dalton Highway crosses the Kuparuk River at a location that drains 142 km² (Site 9 in Figure 2-1C). Hydrographs from these locations are characterized by substantial discharge during snowmelt in late May, followed by highly variable discharge during the open water season driven by precipitation events. Snowmelt coincides with maximum annual discharge at the USGS gage, when an estimated 60% of annual runoff occurs [Dery et al., 2005; McNamara et al., 1998]. At the headwater gages, however, precipitation events have been observed to produce discharge levels that meet or exceed those during snow melt [Kane et al., 2008].

Methods

Due to the highly variable discharge regimes within these arctic basins during the open water season, a dynamic temperature modeling approach coupled with field data collection was adopted to: 1) estimate the dominant controls on temperature in the
Kuparuk River, and 2) determine if basic field data used in model setup adequately represent these controls. Identifying dominant heat fluxes provides fundamental knowledge necessary for predicting changes in river temperatures due to a changing climate over large regions of the arctic (e.g., North Slope of the Brooks Range in Alaska). Data collected during the open water seasons (June – August) of 2013 and 2014 were used as model inputs and for evaluation of the model. Due to extreme winter freezing of the river and typically destructive ice-flows during spring breakup, data collection outside of the open water season was beyond the scope of the work presented here.

Model Formulation

The dynamic temperature model applied to the study area uses a kinematic wave approach (Eqn. 1 and 2) for hydraulic routing [Chapra, 1997], and calculates heat fluxes across the air-water interface and water-sediment interface similar to the approach outlined in Cardenas et al. [2014] (see Figure A-1 for conceptual model diagram). Lateral inflows were treated as either distributed or point sources of mass and heat. Negative lateral inflows were treated as distributed losses of mass and heat. Governing heat and mass equations are solved with an explicit Euler method with upwind differencing scheme [Chapra, 1997].

\[
\frac{\partial Q}{\partial x} + (\alpha \beta Q^{\beta - 1}) \frac{\partial Q}{\partial t} = q
\]  

(1)

\[
\alpha = \left( \frac{n B^{2/3}}{\sqrt{S_0}} \right)^{3/5}
\]  

(2)

Where \( Q \) = volumetric discharge (m\(^3\) s\(^{-1}\)), \( x \) = distance downstream (m), \( \beta = \frac{3}{5} \), \( t \) = time (s), \( q \) = lateral inflows (treated as point or distributed) per unit length (m\(^3\) s\(^{-1}\) m\(^{-1}\)), \( n \) =
Manning’s roughness coefficient, $B = \text{channel width (m)}$, and $S_o = \text{channel slope (m m}^{-1}\text{)}$.

The energy balance equation to predict the change in temperature over time and space is

$$\frac{\partial (A_c T)}{\partial t} = \frac{\partial (Q T)}{\partial x} + q_d T_d + q_{pt} T_{pt} + \frac{B}{\rho_C P} (J_f + J_w + J_{sed}) \quad (3)$$

Where $A_c = \text{river cross sectional surface area (m}^2\text{)}$, $T = \text{water temperature (°C)}$, subscripts $d$ and $pt$ refer to distributed and point source, respectively, $\rho = \text{density of water (kg m}^{-3}\text{)}$, $C_P = \text{specific heat capacity of water (J kg}^{-1}\text{ °C}^{-1}\text{)}$, $J_f = \text{friction heat flux from Theurer et al. [1985] (W m}^{-2}\text{)}, J_w = \text{total heat flux across the air-water interface (W m}^2\text{)},$ and $J_{sed} = \text{conductive sediment heat flux (W m}^2\text{)}. \text{Total heat flux across the air-water interface (}J_w, \text{Eqn. 3) is expressed as}$

$$J_w = J_{sn} + J_{an} + J_{br} + J_c + J_e \quad (4)$$

Where $J_{sn} = \text{shortwave solar radiation (W m}^2\text{)}, J_{an} = \text{atmospheric longwave radiation (W m}^2\text{)}, J_{br} = \text{atmospheric longwave radiation from the water surface (W m}^2\text{)}, J_c = \text{sensible (conduction and convection) heat flux (W m}^2\text{)},$ and $J_e = \text{latent heat flux (evaporation and condensation) (W m}^2\text{)}. \text{The shortwave solar radiation flux was based on field observations as described below and riparian shading was negligible due to limited shrubs and no trees in the riparian areas. Longwave radiation estimates include atmospheric radiation only and assumes vegetative influences to be insignificant. Surface fluxes }J_{an}, J_{br}, J_c, \text{ and } J_e \text{ are estimated based on Chapra [1997] and detailed equations are provided in the Appendix A. As derived in Neilson et al. [2010b], the bed conduction heat flux } (J_{sed}, \text{Eqn. 3) is given by}$

$$J_{sed} = \frac{\rho_{SED} C_{P,SED} \alpha_{SED}}{\gamma_{SED}} (T_{SED} - T) \quad (5)$$

and sediment temperatures ($T_{SED}$) in this shallow layer are estimated by
\[
\frac{dT_{SED}}{dt} = \alpha_{SED} \frac{A_{SED}}{V_{SED}} \left[ \frac{(T - T_{SED})}{Y_{SED}} + \frac{(T_{gr} - T_{SED})}{Y_{gr}} \right]
\]  

(6)

In the equations above, \( \rho_{SED} \) = sediment density (kg m\(^{-3}\)), \( C_{P,SED} \) = specific heat capacity of sediment (J kg\(^{-1}\) °C\(^{-1}\)), \( \alpha_{SED} \) = coefficient of thermal diffusivity in sediment (m\(^2\) sec\(^{-1}\)), \( Y_{SED} \) = bed depth (e.g., depth of the shallow sediment layer below the river) (m), \( T_{SED} \) = predicted temperature of the bed sediment layer adjacent to the surface water (°C), \( A_{SED} \) = area of water-sediment interface (m\(^2\)), \( V_{SED} \) = volume of sediment (m\(^3\)), \( T_{gr} \) = boundary condition temperature of the deeper sediments below the shallow layer (assumed to be 0 °C), and \( Y_{gr} \) = the distance to the frozen boundary from \( Y_{SED} \). This approach treats heat exchange across the sediment-water interface as simple conduction, ignoring advective heat transfer due to hyporheic exchange.

Lateral inflows have the potential to significantly contribute energy without imparting a significant change in river temperature (e.g., if lateral inflows are significant but river and lateral inflow temperatures are similar at their confluence). If trying to quantify the influence of lateral inflows on instream temperature, the use of a standard thermal datum of 0 °C would produce an overestimate. Instead, we use a relative thermal datum and following Kurylyk et al. [2015a], and define the apparent sensible heat flux from lateral inflows (\( J_{lat,i} \), Eqn. 7) as the portion of the heat flux from lateral inflows that affects a change in river temperature resulting from a difference in temperatures between main channel and lateral inflow.

\[
J_{lat,i} = \frac{\rho_{C_p} q_i (T_{lat,i} - T_i)}{B_i \Delta x}
\]  

(7)

Where for every model cell \( (i) \), \( J_{lat} \) = apparent sensible heat flux from lateral inflows (W m\(^{-2}\)), and \( T_{lat} \) = distributed \( (T_d) \) or point source \( (T_{pt}) \) lateral inflow temperature.
(°C). The apparent sensible heat flux is formulated to be comparable with other heat fluxes (Eqn. 4 and Eqn. 5).

Channel geometry coupled with large ranges in discharges led to widely varying wetted widths ($B$) (Figure A-2), and therefore surface areas ($A_s = B \cdot \Delta x$). As surface areas significantly affect modeled river temperature, and are related to wetted width, wetted width to discharge relationships were developed for main channel sites 2 – 9 (Figure 2-1B and 2-1C). These relationships followed a power function (Eqn. 8) and were linearly interpolated spatially between sites to cover all model cells.

$$B = \gamma Q^\delta$$

(8)

Where $\gamma$ and $\delta$ are fitting parameters calibrated with a least squares regression for each station. This relationship provided wetted width estimates for each model cell that varied as a function of discharge and resulted in temporally and spatially varying estimates of $A_s$.

The study area within the Kuparuk basin was divided into a 252 km long lower model domain (site 4 – 1 in Figure 2-1B) and a 34 km long upper model domain (site 9 – 5 in Figure 2-1C) to exclude a 23 km long section of river that includes a significant surficial ice feature (the Kuparuk Aufeis described in Yoshikawa et al. [2007]) and associated sections of river that have been observed to become subsurface under low flow conditions during the open water season. The model was discretized longitudinally such that estimates of numerical [Chapra, 1997] and actual [Fischer, 1973] dispersion were within a factor of two of each other based on observed depth, width, velocity, and bed slope following Neilson et al. [2010b]. A calculation time step was selected that satisfied
dispersion and Courant condition requirements [Chapra, 1997]. As determined by data availability throughout each study area, the 2013 model runs covered 34 and 37 days for the upper and lower model domains, respectively, while the 2014 model runs covered 35 and 42 days for the upper and lower model domains, respectively (Table 2-1). Model discretization and parameters were held constant between model years, but in some cases varied between domains (Table 2-1). Channel top widths were allowed to vary with discharge based on power law relationships (Eqn. 8) developed from observations.

Table 2-1. Modeling information by model domain

<table>
<thead>
<tr>
<th>Variable</th>
<th>Upper</th>
<th>Lower</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model Segments</td>
<td>( \delta x )</td>
<td>100 m</td>
</tr>
<tr>
<td>Number of Segments</td>
<td>( N )</td>
<td>342</td>
</tr>
<tr>
<td>Channel top width</td>
<td>( B )</td>
<td>7 – 30 m</td>
</tr>
<tr>
<td>Channel Slope</td>
<td>( S_o )</td>
<td>0.0075</td>
</tr>
<tr>
<td>Mean Elevation</td>
<td>( Z )</td>
<td>774 m</td>
</tr>
<tr>
<td>Median Downstream Boundary Discharge*</td>
<td>( Q )</td>
<td>10 m(^3) s(^{-1})</td>
</tr>
<tr>
<td>Manning’s Roughness</td>
<td>( n )</td>
<td>0.1</td>
</tr>
<tr>
<td>Bed Depth</td>
<td>( Y_{SED} )</td>
<td>0.5 m</td>
</tr>
<tr>
<td>Depth to frozen ground</td>
<td>( Y_{fr} )</td>
<td>1 m</td>
</tr>
<tr>
<td>Sediment Thermal Diffusivity</td>
<td>( \alpha_{SED} )</td>
<td>1.26 x 10(^{-6}) m(^2) s(^{-1})</td>
</tr>
<tr>
<td>Bulk Sediment Density</td>
<td>( \rho_{SED} )</td>
<td>1400 kg m(^{-3})</td>
</tr>
<tr>
<td>Specific Heat Capacity of Sediment</td>
<td>( C_{P,SED} )</td>
<td>710 J kg(^{-1}) °C(^{-1})</td>
</tr>
<tr>
<td>Specific Heat Capacity of Water</td>
<td>( C_P )</td>
<td>4187 J kg(^{-1}) °C(^{-1})</td>
</tr>
<tr>
<td>Water Density</td>
<td>( \rho )</td>
<td>1000 kg m(^{-3})</td>
</tr>
<tr>
<td>Emissivity of water</td>
<td>( \varepsilon )</td>
<td>0.97</td>
</tr>
<tr>
<td>Model period 2013</td>
<td>July 2 – Aug 5</td>
<td>June 27 – Aug 3</td>
</tr>
<tr>
<td>Model period 2014</td>
<td>June 30 – Aug 4</td>
<td>June 24 – Aug 4</td>
</tr>
</tbody>
</table>

* = values are based on both 2013 and 2014 model periods.

Data Collection

In each of the 2013 and 2014 study years, seven main channel data collection locations were instrumented along the main stem of the Kuparuk with four (sites 1, 2, 3, and 4) in the lower model domain and three (sites 5, 8, and 9) in the upper model domain
Sub-reaches in this paper are defined as the sections of river between consecutive main channel data collection locations. In 2014, site 3 (Figure 2-1B) was moved 6.3 km downstream from its position in 2013 to address concerns that a secondary braid may divert some of the flow under high flows.

**Discharge**

Discharge records for site 1 were provided by the USGS while discharge records for sites 2, 3, 4, 5, 8, and 9 (Figure 2-1) were estimated from rating curves produced from a minimum of 4 discharge measurements per site per season and a continuous record of 15 minute stage measurements from Campbell Scientific® CS450 (Logan, UT) pressure transducers (accuracy = 0.005 m). Rating curves were graphically matched between years to incorporate both years of observations into a single rating curve, similar to the approach taken in Overbeck [2015]. Some stages at sites 5 and 8 exceeded the range of observed flows and therefore exceeded the developed rating curves (Table A-2). To address this issue, composite rating curves were produced for these sites based on the long term rating curve reported for site 9, similar to the approach taken in Kane et al. [2003]. Low flow discharge measurements were made using a SonTek Flowtracker™ (San Diego, CA) handheld Acoustic Doppler Velocimeter (accuracy = 1% of measured velocity), while a Teledyne RD Instruments® StreamPro™ (Poway, CA) Acoustic Doppler Current Profiler (velocity and depth accuracy = 1% of measured value) was used at high flows. Imnnavait Creek (T11) and the Toolik River were gaged in both years and rating curves were developed to produce time series of discharge.

**Water Temperature**

Water temperatures used for upstream boundary conditions and model evaluation
were collected at 15 minute intervals using Onset Corporation® HOBO ProV2™ (Borne, MA) thermistor temperature loggers (accuracy = ±0.2°C). Thermistor accuracy was verified by comparing thermistor readings against those of a Fluke® 5610™ thermistor (American Fork, UT) (accuracy = ±0.01 °C) in continuously stirred homogeneous baths at approximately 0 °C and 22 °C. All sensors were found to be within the manufacture’s reported accuracy, and no corrections were made to thermistor records. Temperature loggers were deployed at all of the main channel data collection locations (site 1 – 9 in Figure 2-1), and in four tributaries in 2013 (The Toolik River, T4, T7, and T11). Based on the diversity of temperatures observed in the three tributaries in the upper model domain during 2013, an additional 13 tributaries were instrumented for temperature in the upper model domain in 2014 for a total of 16 instrumented tributaries (T1 – T16 in Figure 2-1C). The Toolik River is the only tributary in the lower model domain whose catchment is similar in size to that of the Kuparuk River at their confluence, and as such was the only instrumented tributary in the lower model domain for both 2013 and 2014.

**Lateral Inflows**

Net lateral inflows were estimated by differencing hydrographs from consecutive gaging stations. The lack of steady flow conditions required that travel times be considered. This led to routing observed hydrographs from upstream to downstream using the kinematic wave approach described above (Eqn. 1 and Eqn 2) and subtracting the simulated hydrograph from the observed hydrograph at the downstream gaging station. This produced time varying estimates of net lateral inflows between subsequent gaging stations that represented a combination of surface water contribution and supra-
permafrost subsurface discharge for two sections in the upper model domain, and three sections in the lower model domain. Lateral inflows were assumed to primarily enter the model reaches as surface flow based on very low Darcy flux estimates in the Imnavait Creek watershed. These calculations incorporated time series of near stream hydraulic gradients, local estimates of hydraulic conductivity, and observations of depth to thaw [B.T. Neilson, unpublished data].

**Meteorological Data**

Meteorological forcing data necessary to estimate surface heat fluxes (Eqn. 4) were collected from a network of stations throughout the watershed (Figure 2-1A). Air temperature, relative humidity, windspeed, and precipitation for both years were collected at the Upper Kuparuk (UK) and the North White Hills (NWH) meteorological stations maintained by the Water and Environmental Research Center at the University of Alaska Fairbanks. A Campbell Scientific® Model 207 (Logan, UT) was used to measure air temperature (accuracy = 0.4 °C) and relative humidity (accuracy = 5%) at 2 m above the ground, and a Met One® Model 104A (Grants Pass, OR) anemometer (accuracy = 0.11 m s⁻¹ over 0–45 m s⁻¹) was used to measure windspeed 3 m above the ground. Precipitation was measured in 0.254 mm increments using a Texas Electronics tipping bucket with an Alter windshield. Shortwave radiation for the upper model domain was measured in 2013 at Imnavait Creek (IC) with a Hukseflux® LP02 (Manorville, NY) pyranometer (spectral range = 285-3000 nm) and in 2014 at site 9 with a Hukseflux® NR01 (Manorville, NY) 4-component net-radiometer (spectral range = 305-2000 nm). Shortwave radiation for the lower model domain was measured in 2013 at the Anaktuvuk
River Burn (ARB) site with a Campbell Scientific® CNR-1 (Logan, UT) 4-component net-radiometer (spectral range = 305-2800 nm) and in 2014 at sites 3 and 4 with Hukseflux® LP02 (Manorville, NY) pyranometers. Riparian and topographic shading were not considered in the temperature modeling given the lack of substantial riparian vegetation and prominent topographic features. The influence of differential cloud cover was approximated by averaging radiation values among pyranometers along each model domain where data were available.

Model Testing

Because of the minimal subsurface inflows, temperatures of lateral inflows were estimated from observations of tributary temperatures. The spatial distribution of surface lateral inflows, coupled with the temporally variable nature of their temperature regimes led to four basic scenarios to determine how to best represent heat fluxes from lateral inflows. For the upper model domain, lateral inflows were simulated as distributed sources for the first three scenarios and as point sources in the fourth scenario. The tributaries in the upper model domain (Figure 2-1C) were classified as beaded and non-beaded for scenario 4 based on the presence or absence of pools along the tributaries’ main stem as seen in a DigitalGlobe® satellite image from June 2010 (Table 2-2). Only the first two scenarios are applicable to the lower model domain.

Scenario 1 (S1): Lateral inflows are assigned temperatures that match their receiving reaches. This results in the lateral inflows adding volume and heat such that they do not change the temperature of the receiving reach (e.g., the apparent sensible heat flux from lateral inflows, $J_{lat}$, is zero). This ensures accurate representation of the volumes given no information about lateral inflow temperatures.

Scenario 2 (S2): Lateral inflow temperatures are set to that of a
“representative” surface water inflow. For the upper model domain this is T11 (Imnavait Creek) and for the lower model this is the Toolik River.

Scenario 3 (S3): Temperatures for lateral inflows are set to an average of tributary temperatures for each sub-reach. This only applies to the Upper Kuparuk where multiple tributaries were instrumented.

Scenario 4 (S4): Lateral inflows along the upper model domain are treated as point sources. Total lateral inflows are divided between known tributaries and scaled by tributary drainage area. Lateral inflow temperatures are determined by the morphology of surface water sources (beaded or non-beaded), which were estimated via averages for each morphology.

Table 2-2. Morphology classification and drainage area for identified surface water lateral inflow sources in the upper model domain

<table>
<thead>
<tr>
<th>Tributary</th>
<th>Morphology</th>
<th>Drainage Area [km²]</th>
<th>Percent of Section Total Tributary Drainage Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site 5 – 8</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Beaded</td>
<td>58</td>
<td>22%</td>
</tr>
<tr>
<td>2</td>
<td>Beaded</td>
<td>6</td>
<td>2%</td>
</tr>
<tr>
<td>3</td>
<td>Non-Beaded</td>
<td>9</td>
<td>3%</td>
</tr>
<tr>
<td>4</td>
<td>Non-Beaded</td>
<td>164</td>
<td>61%</td>
</tr>
<tr>
<td>5</td>
<td>Non-Beaded</td>
<td>1</td>
<td>&lt;1%</td>
</tr>
<tr>
<td>6</td>
<td>Beaded</td>
<td>1</td>
<td>&lt;1%</td>
</tr>
<tr>
<td>7</td>
<td>Non-Beaded</td>
<td>17</td>
<td>6%</td>
</tr>
<tr>
<td>8</td>
<td>Non-Beaded</td>
<td>5</td>
<td>2%</td>
</tr>
<tr>
<td>9</td>
<td>Beaded</td>
<td>3</td>
<td>1%</td>
</tr>
<tr>
<td>10</td>
<td>Beaded</td>
<td>5</td>
<td>2%</td>
</tr>
<tr>
<td>Site Total Tributary Drainage Area</td>
<td>286</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

| Site 8 – 9|            |                      |                                               |
| 11        | Beaded     | 17                   | 36%                                           |
| 12        | Beaded     | 3                    | 6%                                            |
| 13        | Non-Beaded | 3                    | 5%                                            |
| 14        | Beaded     | 5                    | 9%                                            |
| 15        | Non-Beaded | 9                    | 20%                                           |
| 16        | Non-Beaded | 12                   | 24%                                           |
| Site Total Tributary Drainage Area | 48 | -                     |

| Site 5 – 9|            |                      |                                               |
| Total    | Beaded     | 97                   |                                               |
|          | Non-Beaded | 220                  |                                               |
|          | All Tributaries | 317              |                                               |
|          | Entire Upper Model Domain | 420 |                                               |
Model performance for each scenario was assessed for each main channel data collection location with a Root Mean Square Error (RMSE) objective function calculated for each day and over each simulation period. Daily RMSE values were calculated and plotted against mean daily discharge to assess model performance under various flow conditions. Because of an observed relationship between discharge and temperature, a high/low flow threshold was set at the median discharge \((Q_{50})\) for each main channel location over the model periods for both years of data collection. This allowed for analysis of model performance under high and low flow conditions. The underdetermined nature of the system and uncertainty in key attributes (e.g., channel geometry) limited model calibration to manning’s roughness to ensure accurate travel times.

**Heat Flux Comparisons**

Heat fluxes in Eqn. 4, Eqn. 5, and Eqn. 7 were calculated for every model cell for every time step, and time series were produced by spatially averaging heat fluxes for each model domain. Heat fluxes were compared for the model scenario that produced the best overall model fit. Percent contribution from each flux was determined as the magnitude of each heat flux normalized by the total heat flux magnitude. For comparison with other heat fluxes, net longwave radiation is defined as the sum of incoming and outgoing radiation \( (J_{in} = J_{an} + J_{bn}) \).
Results

Observations

In the upper model domain there was a consistent warming trend with distance downstream, while in the lower model domain temperatures rose between site 4 and site 3, and showed inconsistent trends between site 3 and site 1 (Figure 2-2). The magnitude of diurnal temperature fluctuations averaged 3.1 °C and 2.1 °C in the upper and lower model domains, respectively. The aufeis feature acted as a thermal reset in river temperature with temperatures at site 4 on average 2.2 °C colder than site 5. Discharge increased with distance downstream, was higher in 2014 than 2013 and was highly episodic with virtually no stable flow conditions in 2014 (Figure 2-3 and Table A-2).

Lateral inflows, estimated from differences in routed hydrographs, were temporally dynamic in the upper domain, with limited steady flow conditions while the lower model domain had periods of more constant lateral inflows (Figure A-3). Lateral inflows were greater in 2014 than in 2013 in both model domains. Discharge records for the instrumented tributaries verify that the Toolik River accounts for more than 45% of the lateral inflows in the lower model domain, while T11 (Imnavait Creek) accounts for 5% of the lateral inflows in the upper model domain. Lateral inflows entered the upper model domain at an average rate of 0.20 m$^3$ s$^{-1}$ per kilometer. Lateral inflows entered the
Figure 2-2. Observed temperatures in the main channel of the upper (A, C) and lower (B, D) model domains for 2013 (A, B) and 2014 (C, D). Sites are listed from upstream to downstream in the legends. The upper model domain (A, C) experienced larger diurnal fluctuations in temperature than the lower model domain (B, D), a continuous increase in temperature with distance downstream, and more variable temperatures in 2013 than 2014. The lower model domain (B, D) experienced both increases and decreases in temperature with distance downstream.

Observed temperatures for the three surface water lateral inflows T4, T7, and T11 (Figure 2-1C) in the upper model domain from 2013 showed a large range in lower model domain at an average rate of 0.32 m$^3$ s$^{-1}$ per kilometer when including the effect of the Toolik River, and 0.17 m$^3$ s$^{-1}$ per kilometer when the contribution from the Toolik River was removed.
Figure 2-3. Observed precipitation (blue) and discharge for the upper (A, C) and lower model (B, D) domains in 2013 (A, B) and 2014 (C, D). Sites are listed from upstream to downstream in the legends. Precipitation records from UK and NWH meteorological stations are shown for the upper and lower model domains, respectively. Precipitation and discharge were both greater in 2014 than 2013.

temperatures (Figure A-4) which led to the instrumentation of 13 additional surface water sources in the upper model domain in 2014 (Figure 2-1C). When grouped, the average temperature records for beaded and non-beaded morphologies showed that beaded streams were generally warmer than non-beaded streams (6.4 °C warmer on average) with differences in temperature between streams varying between 2.1 and 13.3 °C (Figure A-5). This temperature range was greatest during periods without precipitation when lateral inflows were at their lowest (Figure A-5). Temperature data collected within the Toolik River were similar in range between years and to the lateral inflows in the upper
model domain (Figure A-6).

Meteorological conditions varied by year and by model domain (Figure A-7, Figure A-8, and Table A-3). July received the most precipitation (75.5 mm on average) and was the warmest month (mean air temperature of 10.6 °C) for both model domains for both years. On average, the upper model domain received 70% more precipitation, 24% less intense solar radiation, and 31% lower wind speeds than the lower model domain over both years. Average air temperature and relative humidity were within 5% of each other between model domains. For both model domains, 2014 had greater precipitation, colder air temperature, higher relative humidity, and less intense solar radiation than 2013, concurrent with observations of greater discharge in 2014 than 2013.

Modeling

Lateral Inflows

Assigned lateral inflow temperatures varied by up to 9 °C between model scenarios with temperatures of the non-beaded sources in S4 being consistently colder than all other lateral inflow temperatures (Figure 2-4). In the upper model domain, lateral inflow temperatures were more homogeneous and discharge values greater in 2014 than 2013. All lateral inflow temperatures in the lower model domain were set to the Toolik River temperature records (Figure A-6).

Hydraulics

While modeled hydrographs reproduced observed hydrographs well (Figure A-9), manually calibrated Manning’s roughness coefficients \( n \) (Table 2-1) were higher than published values [Kane et al., 2003]. This indicated that the simplified representation of
Figure 2-4. Temperatures assigned to lateral inflows in the upper model domain for 2013 (A) and 2014 (B) scenarios. The range in temperatures between scenarios was larger in 2013. Non-beaded streams in S4 were consistently the coldest between all scenarios.

channel geometry does not adequately capture actual hydraulic variability through the study reaches.

Water Temperature

*Upper model domain:* The range in simulated temperatures between scenarios averaged 0.8 °C (SD = 0.6 °C) at site 5 (Figure 2-5), and S2 provided the best model fit for the upper model domain averaged over both years (Table A-4). Modeled and observed river temperatures agreed better in 2014 than in 2013 (site 5 shown in Figure 2-5). RMSE values between scenarios and sites range from 1.0 - 2.4 °C in 2013 and these values were improved in 2014 to 0.6 - 1.5 °C. For both years modeled river temperatures underestimated mean river temperatures while they overestimated the magnitude of diurnal temperature fluctuations (Table A-4). Given the differences in flow regimes between 2013 and 2014 and the highly inaccurate predictions of temperature during low
flows (e.g., July 13th to 18th, 2013, Figure 2-5), it was necessary to compare model performance for various flows (Figure 2-6 and Table A-5). Mean daily RMSE values were 1.55 °C and 0.99 °C for days with mean discharge below and above the median discharge, respectively (Table A-5). The worst model performance occurred under extremely low flow conditions with residuals at times up to 8.2 °C (Figure 2-5) resulting in daily RMSE values of up to 5.5 °C (Figure 2-6). Model performance was better at site 8 than site 5 (Tables A-4 and A-5) likely due to the dominant influences of the upstream boundary condition at site 8 (13 km from boundary) compared to site 5 (34 km from boundary) [Heavilin and Neilson, 2012].

Figure 2-5. Observed shortwave solar radiation and precipitation for the upper model domain in 2013 (A) and 2014 (C). Site 5 modeled (grey shaded region) and observed (dashed black line) water temperatures and observed discharge (solid blue line) for 2013 (B) and 2014 (D). Shaded regions represent the range in modeled temperatures at site 5 for the four model scenarios (S1-S4). Periods of low flow coincide with periods of poor model performance (e.g., July 14th 2013).
Figure 2-6. Daily RMSE values vs. average daily discharge for the upper (left) and lower (right) model domains. Vertical and horizontal lines represent the median discharge and mean RMSE, respectively.

Lower model domain: The range in simulated temperatures between scenarios averaged 0.45 °C (SD = 0.3 °C) at site 1 (Figure 2-7, Table A-6), and S2 provided the best model performance for all sites in the lower model domain. Estimated river temperatures for sites 1, 2, and 3 in the lower model domain (site 1 shown in Figure 2-7) and model period statistics for observed and modeled water temperatures from these same locations illustrate a mixed model bias (Table A-6). Water temperature was underestimated at site 3 in 2014 and site 2 in both years, overestimated at site 1 in both years and overestimated at site 3 in 2013. Modeled daily temperature ranges were overestimated for sites 1 and 2 and underestimated for site 3 (Table A-6). Daily RMSE values over all sites and scenarios ranged from 0.8 – 1.9 °C in 2013 and 0.8 – 1.5 °C in 2014 (Table A-6). The highest RMSE values occurred under lower flow conditions (Figure 2-6). Mean RMSE values for flows below and above the median flow were 1.55
°C and 0.93 °C, respectively (Table A-5). Overall model performance was similar between years (Table A-6) despite the very different flow (Figure 2-3) and lateral inflow regimes (Figure A-3). There was a period of exceptionally poor model performance for site 1 from July 12th through July 17th 2013 where the daily mean and diurnal range in temperatures at the downstream boundary were both overestimated (Figure 2-7B).

Heat Fluxes

Heat flux values calculated for S2 were used for heat flux comparisons (Table 2-3). Solar radiation, $J_{sn}$, had the largest maximum magnitude, reaching peaks up to 823 W m$^{-2}$ on clear days while longwave incoming and outgoing radiation, $J_{an}$ and $J_{br}$, had the

![](image)

Figure 2-7. Observed shortwave solar radiation and precipitation for the lower model domain in 2013 (A) and 2014 (C). Site 1 modeled (grey shaded region) and observed (dashed black line) water temperatures, and observed discharge (solid blue line) for 2013 (B) and 2014 (D). Shaded regions represent the range in modeled temperatures at site 1 for the two model scenarios (S1 – S2).
largest mean values at 245 and 357 W m$^{-2}$, respectively. Net longwave radiation, ($J_{ln} = J_{an} + J_{br}$), was a loss of heat with an average value of 112 W m$^{-2}$. Sensible heat flux, $J_c$, latent heat flux, $J_e$, sediment heat flux, $J_{sed}$, and the apparent sensible heat from lateral inflows, $J_{lat}$, were both positive and negative (Table 2-3). Of the total heat fluxes, on average 88% is attributed to $J_{sn}$ (17%), $J_{an}$ (29%), and $J_{br}$ (42%) while the rest of the heat fluxes contributed less than 5% on average (Table 2-3). While the average contribution from $J_e$ was only 4%, it became significant periodically in the lower model domain contributing up to 18% of the total heat flux on clear warm days with low relative humidity and high windspeed (Table 2-3 and Figure 2-8). Similarly, $J_{lat}$ contributed 2% on average for the entire model domain, but was periodically significant in the upper domain contributing up to 26% to total heat flux shortly after precipitation events when apparent sensible heat flux from lateral inflows was elevated and solar radiation was depressed (Table 2-3 and Figure 2-8). Contributions from the other heat fluxes remained

**Table 2-3. Statistics of heat fluxes (W m$^{-2}$) and percent contribution to the total heat flux (%) as calculated under S2 for the upper and lower model domains and for the entire Kuparuk River over 2013 and 2014**

<table>
<thead>
<tr>
<th></th>
<th>$J_{sn}$</th>
<th>$J_{an}$</th>
<th>$J_{br}$</th>
<th>$J_c$</th>
<th>$J_e$</th>
<th>$J_f$</th>
<th>$J_{sed}$</th>
<th>$J_{lat}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper</td>
<td>Min</td>
<td>0 (0)</td>
<td>17 (16)</td>
<td>-427 (22)</td>
<td>-51 (0)</td>
<td>-149 (0)</td>
<td>1 (0)</td>
<td>-40 (0)</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td>178 (17)</td>
<td>244 (29)</td>
<td>-353 (42)</td>
<td>3 (2)</td>
<td>-22 (3)</td>
<td>28 (3)</td>
<td>-9 (1)</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>823 (53)</td>
<td>291 (40)</td>
<td>-23 (57)</td>
<td>62 (7)</td>
<td>30 (12)</td>
<td>84 (11)</td>
<td>4 (4)</td>
</tr>
<tr>
<td>Lower</td>
<td>Min</td>
<td>0 (0)</td>
<td>66 (16)</td>
<td>-396 (24)</td>
<td>-104 (0)</td>
<td>-241 (0)</td>
<td>2 (0)</td>
<td>-26 (0)</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td>185 (18)</td>
<td>246 (29)</td>
<td>-362 (43)</td>
<td>-5 (3)</td>
<td>-44 (5)</td>
<td>8 (1)</td>
<td>-11 (1)</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>705 (50)</td>
<td>307 (40)</td>
<td>-87 (37)</td>
<td>91 (12)</td>
<td>40 (18)</td>
<td>19 (5)</td>
<td>5 (3)</td>
</tr>
<tr>
<td>Kuparuk River</td>
<td>Min</td>
<td>0 (0)</td>
<td>42 (16)</td>
<td>-412 (23)</td>
<td>-78 (0)</td>
<td>-195 (0)</td>
<td>1 (0)</td>
<td>-33 (0)</td>
</tr>
<tr>
<td></td>
<td>Mean</td>
<td>181 (17)</td>
<td>245 (29)</td>
<td>-357 (42)</td>
<td>-1 (2)</td>
<td>-33 (4)</td>
<td>18 (2)</td>
<td>-10 (1)</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>764 (52)</td>
<td>299 (40)</td>
<td>-55 (57)</td>
<td>77 (9)</td>
<td>35 (15)</td>
<td>52 (7)</td>
<td>5 (3)</td>
</tr>
</tbody>
</table>
Figure 2-8. Time series of heat fluxes that were calculated to be periodically significant under model scenario 2 (S2) for 2013 (A, B) and 2014 (C, D) for the upper (A, C) and lower (B, D) model domains. $J_{sn}$, $J_{ln}$, and $J_{lat}$ were periodically significant (>15% of the total heat flux) in the upper model domain (A, C) while $J_{sn}$, $J_{ln}$, and $J_{e}$ were periodically significant in the lower model domain (B, D).
below 15% of the total heat flux at all times (Table 2-3, Figure A-10, and Figure A-11). $J_c$ and $J_e$ were larger in magnitude and relative contribution in the lower model domain while friction, $J_f$, $J_{sed}$, and $J_{lat}$ were larger in magnitude and relative contribution in the upper model domain (Table 2-3).

**Discussion**

Between scenarios, lateral inflow temperatures were seen to vary significantly, at times varying up to 9 °C (Figure 2-4). Model results, however, varied far less (up to 0.8 °C) between scenarios indicating that the treatment of lateral inflow temperatures had little influence on resulting modeled river temperature. There were both similarities and differences between heat flux dynamics in the low arctic Kuparuk River and temperate systems. Similar to Hebert et al. [2011] and Caissie [2006], the magnitudes of surface heat fluxes ($J_{sn}$, $J_{an}$, $J_{br}$, $J_c$, and $J_e$) were on average greater in the higher stream order portion of the study area (Table 2-3). Surface fluxes were greater than bed fluxes in both model domains, similar to Hebert et al. [2011] and Moore et al. [2005] (Table 2-3). Friction heat flux, $J_f$, was a relatively small portion of the energy budget in both model domains, averaging 3% in the upper and 1% in the lower model domains. Friction contributed up to 11% in the upper model domain under high flows similar to Webb and Zhang [1997]. Shortwave solar radiation, $J_{sn}$, was the main daytime heat source (Figure 2-8) which is consistent with Neilson et al. [2009], Caissie [2006], and Cozzetto et al. [2006] who all found shortwave radiation to dominate instream heat budgets across a wide range of temperate and polar systems. Consistent with a forested basin in New
Brunswick, Canada [Maheu et al., 2014], sensible heat fluxes, $J_c$, were on average positive in the upper model domain and negative in the lower model domain (Table 2-3). Following Smith and Lavis [1975], Polehn and Kinsel [2000], Arscott et al. [2001], Caissie [2006], lateral inflows had greater significance in the upper model domain, at times contributing up to 26% of the total heat flux. Net longwave radiation was consistently negative throughout the Kuparuk River, indicating that radiative emission from water (Eqn. A-4) exceed that from air (Eqn. A-1) at all times. This is similar to the glacier-fed alpine stream reported in Khamis et al. [2015], but different from Sinokrot and Stefan [1993] and Benyahya et al. [2012] who report positive values under some conditions. While latent heat flux, $J_e$, at times contributed up to 15% of the total energy budget in the Kuparuk River, its average contribution was only 4% making it a relatively small heat flux. This differs from Webb and Zhang [1997], Cozzetto et al. [2006], Hannah et al. [2008b], and Khamis et al. [2015] who all found latent heat to be a significant heat loss in systems ranging from moorlands to Antarctic deserts to alpine streams. The low average contribution from latent heat flux in the Kuparuk River is likely due to relatively cool air temperatures and high relative humidity – two factors that distinguish the North Slope desert from its temperate and polar counterparts, respectively.

The basic modeling approach and data used here captured the temporal and spatial variability of instream temperature responses during a significant portion of the model duration. Flow dependent accuracy indicated that under low flow conditions: 1) some mechanisms that were not considered in this modeling effort were significant; 2) channel geometry was not adequately represented; and/or 3) river temperatures were most
susceptible to inaccuracies in local meteorological conditions. Some processes that were
not included in this initial approach that may prove to be important include: transient
storage, dynamic depth to thaw, and subsurface lateral inflows. Zarnetske et al. [2007]
found transient storage in this basin to be greatest under low flow conditions, and
hyporheic exchange has been shown to dampen diurnal temperature fluctuations
[Arrigoni et al., 2008]. In this model formulation, bed conduction was calculated over a
constant distance to the frozen layer, and advective hyporheic influences in the bed
sediments, which are likely significant [Edwardson et al., 2003], were not accounted for.
By limiting streambed heat exchange to conduction, vertical and lateral advective mass
and heat transport are ignored as are their influences on vertical thermal gradients within
the sediment. These impacts, in turn, affect estimates of conductive heat fluxes.
Zarnetske et al. [2007] showed that thaw depth increased throughout the entire open
water season for Oksrukuyik Creek in a neighboring basin, indicating that a constant
distance to frozen ground could result in an overestimate early in the model periods, and
an underestimate late in the model periods [Brosten et al., 2009]. Zarnetske et al. [2008]
also showed that hyporheic exchange can be constrained by the depth of thaw early in the
season, but that once a threshold depth is surpassed, hyporheic exchange is dependent
solely on hydraulic head gradients. Dynamic depth of thaw could be incorporated into
the temperature modeling by calibration of a bed thaw model (e.g., Kurylyk [2015]) and
hyporheic exchange influences could be incorporated following Neilson et al. [2010b].

Channel geometry is critical to river temperature modeling as surface heat fluxes
are scaled by surface area. Under low flows, increased surface area to volume ratios
influence residence times and therefore, river temperature [Schmadel et al., 2015]. This was seen via a sensitivity analysis where a +/- 20% change in river width produced +/- 0.1 °C and +/- 0.4 °C change in mean daily RMSE values for the bottom 50th and 10th percentile discharge values, respectively. Sensitivity of river temperature to channel geometry under low flows was exacerbated by a lack of stabilizing effects from the boundary condition propagation and significantly reduced lateral inflows under low flow conditions [Gu and Li, 2002; Heavilin and Neilson, 2012; Schmadel et al., 2015]. High resolution remote sensing techniques hold promise for improving estimates of channel geometry (e.g., cross sections at small enough spatial scales to adequately capture reach heterogeneity [Schmadel et al., 2015]) and other model parameters (e.g., delineation of surface transient storage zones from thermal imagery [Bingham et al., 2012]). More accurately constraining channel geometry will reduce overall uncertainty in model predictions and allow for focused calibration of parameters associated with additional heat transfer processes (e.g., hyporheic exchange and lateral inflow partitioning).

Sparse data necessitated the approximation of meteorological conditions across the entire basin from stations up to 150 hundred kilometers away from some model cells. Hebert et al. [2011] showed that meteorological conditions, especially solar radiation and windspeed, do not interpolate well over long distances. As an example, high daily average river temperatures were simulated at site 1 from July 12th through July 17th 2013 (Figure 2-7), and this period of poor model performance concluded with a sudden rise in observed temperatures on July 17th. Given that estimates of width and lateral inflows did not change significantly on the 17th, it is likely that inaccurate estimates of
meteorological conditions for this period are the source of model error. Solar radiation for this period was measured at the Anaktuvuk River Burn (ARB) site, 150 kilometers away from downstream portions of the lower model domain (Figure 2-1A), and likely was not representative of conditions along the entire study reach. Differences in elevation between meteorological stations and the river could have influenced accuracy of estimating air temperature through adiabatic cooling and windspeed through spatial heterogeneity in weather patterns. These errors could be minimized with denser meteorological station networks and/or estimation of spatial distribution of metrological conditions using a local meteorological model like that suggested by Liston and Elder [2006].

While large scale differencing of routed hydrographs provided a data efficient method to accurately reproduce system hydraulics under most conditions, peak discharge was slightly overestimated for some storm events in the lower model domain, and there were periods of rapid change in lateral inflow estimates in both model domains. These artifacts are due to errors in rating curve development (e.g., observations of discharge and stage observations) and hydraulic routing (e.g., estimates of channel geometry, roughness, and channel slope and simplifying assumptions of distributed lateral inflow contributions over long distances). Further, while simple treatment of lateral inflows was adequate for capturing the heat loads from the landscape to the main channel at large spatial scales, evaluating model performance at smaller spatial scales will likely reveal that these simplifications neglect spatial heterogeneity in river temperature vital to suitable species habitat [Brierley and Fryirs, 2005; Kurylyk et al., 2015b]. A more
nuanced approach (e.g., separating lateral inflows into surface and subsurface components, accounting for simultaneous gains and losses within sub-reaches, and increasing the spatial resolution of lateral inflow estimates) may better capture thermally significant dynamics of lateral inflows.

Parsing lateral inflows into surface and subsurface components requires increasing understanding of near-stream active layer characteristics and hydraulic properties. The importance of separating lateral inflows into surface and subsurface components may increase in the future with changes in discharge [Overeem and Syvitski, 2010; Peterson et al., 2002] coupled with significant changes in landscape architecture [Liljedahl et al., 2016] that could alter subsurface flow. Incorporation of subsurface lateral inflows and hyporheic exchange into the current temperature modeling effort could buffer river temperature and improve modeled diurnal temperature fluctuations (Tables A-4 and A-5). Regression analyses and hydrologic models may prove valuable for estimating lateral inflow temperature and discharge from meteorological conditions [Flint and Flint, 2008; Pohl et al., 2007; Rwetabula et al., 2007; Webb et al., 2003].

Data requirements for river temperature modeling varied over space and time. Predicting river temperatures for the lower model domain may be feasible with, at a minimum, temperature and flow data to represent upstream and downstream boundary conditions, representative cross section geometries, and meteorological data. This modeling effort illustrates the need for accurate records of solar radiation to produce accurate river temperature predictions in the lower model domain. More sophisticated modeling approaches and data collection methods are required to adequately describe the
thermal behavior of lower order headwater portions of basins where data requirements vary by flow regime. Similar to the lower modeling domain, temperature and flow data are required to represent upstream and downstream boundary conditions. In addition, estimates of heat fluxes from lateral inflows are needed for simulating river temperature under high flow conditions and accurate estimates of channel geometry also required under low flow conditions. The spatial scale at which channel geometry is required is, however, currently unclear and should be the focus of future work in lower order headwater sections of arctic rivers. Additionally, more data may be needed to include additional heat fluxes not accounted for in this effort, such as hyporheic exchange.

Conclusion

A relatively basic river temperature model that accounts for heat fluxes known to be relevant for temperate systems was applied to the Kuparuk River, Alaska to determine controls on water temperature in this low arctic basin. Model performance was used to identify regions and flow conditions for which further model refinement is required. There were both similarities and differences in relative importance of heat fluxes between low arctic and temperate rivers. The similarities provided evidence that the extensive research on temperate rivers provides a reasonable foundation upon which to build, while the differences demonstrated the need for further testing and data collection to support the inclusion of additional heat transfer mechanisms.

As anticipated, surface heat fluxes dominated in higher order, lower gradient
portions of the river basin, while competition between surface fluxes and lateral inflows increased with distance upstream and varied with hydrologic condition. The spatial and temporal variability in observed lateral inflow temperatures necessitated the testing of various treatments of lateral inflow heat fluxes. Differences in simulated river temperatures between these scenarios were small (average range in simulated temperatures were 0.8 °C at site 5 and 0.45 °C at site 1) and the largest differences in simulated temperatures occurred under low lateral inflow conditions.

Model performance varied directly with discharge, indicating that simplifications assumed in this approach are more valid under high flow conditions. However, it is not clear if poor model performance under low flow conditions can be attributable to missing heat transfer mechanisms, inaccurate representation of river geometry, or inaccurate representations of meteorological conditions. Future work should focus on: 1) improved estimates of channel geometry with a focus on low-flow conditions, and 2) including additional heat transfer mechanisms that could include transient storage processes, thaw bulb progression, and partitioning lateral inflows into surface and subsurface components. Remote sensing techniques hold promise for improving channel geometry estimates and could additionally be used to estimate parameters associated with the incorporation of additional heat transfer mechanisms.
CHAPTER 3

QUANTIFYING THE ROLE OF HYPORHEIC EXCHANGE ON
ARCTIC RIVER TEMPERATURE IN AREAS OF
CONTINUOUS PERMAFROST

Abstract

Hyporheic exchange has the potential to significantly influence river temperatures in regions of continuous permafrost under low-flow conditions given the strong thermal gradients that exist in river bed sediments. However, there is limited understanding of the impacts of hyporheic exchange on Arctic river temperatures. To address this knowledge gap, heat fluxes associated with hyporheic exchange were estimated in a fourth order Arctic river using field observations coupled with a river temperature model that accounts for hyporheic and surface transient storage influences. Temperature time series and tracer study solute breakthrough curves were measured in the main channel and river bed at multiple locations and depths to characterize hyporheic exchange and provide parameter bounds for model calibration. Model results for low-flow periods from three years indicated that hyporheic exchange contributed up to 20% of the total river energy balance, reduced the main channel diel temperature range by up to 2.3 °C, and reduced mean daily temperatures by 0.23 °C over a 13.1 km long study reach. These influences are due to main channel heat loss during the day and gain at night via hyporheic exchange, and heat loss from the hyporheic zone to the ground below via conduction. Main channel temperatures were found to be sensitive to simulated changes.

Coauthored by Tyler V. King and Bethany T. Neilson
in ground temperatures due to changes in hyporheic exchange heat flux and deeper ground conduction. These results suggest that the heat sink influence of hyporheic exchange could be reduced if ground temperatures warm in response to projected increases in permafrost thaw below rivers.

Introduction

Water temperatures exert a primary control on river ecosystem function with influences on habitat suitability [e.g., Boisneau et al., 2008] and growth rates [e.g., Deegan et al., 1999; Nicieza and Metcalfe, 1997] for aquatic species, as well as physical [Arp et al., 2012b; Merck and Neilson, 2012; Pohl et al., 2007; Rawlins et al., 2010; Syvitski, 2002] and chemical [Cory et al., 2013; McNamara et al., 2008] river processes. Given the ecological importance of river temperature, understanding the controlling heat fluxes is key to effective river management [Dugdale et al., 2017; Hannah and Garner, 2015; Poole and Berman, 2001]. This can be even more important in a changing climate where energy balances can shift [Caldwell et al., 2015; Luce et al., 2014; Muñoz-Mas et al., 2016; van Vliet et al., 2013] and has already been shown to impact some rivers [e.g., Isaak et al., 2012].

The application of process based river temperature models provides insight regarding the influences of specific heat fluxes and typically account for heat transfers across the air-water and water-sediment interfaces. Decades of research have identified heat fluxes across the water-sediment interface, referred to here as bed heat fluxes, as a
substantial component of some river heat budgets, especially in shallow rivers and streams [e.g., *Hondzo and Stefan*, 1994]. Early research that included bed heat fluxes focused on conductive heat fluxes only [*Brown*, 1969; *Evans et al.*, 1998; *Sinokrot and Stefan*, 1993], while more recent work has shown advective heat fluxes to be an important component of heat exchange across the water-sediment interface [*Burkholder et al.*, 2008; *Caissie and Luce*, 2017; *Cozzetto et al.*, 2006; *Neilson et al.*, 2010b; *Story et al.*, 2003]. Together, these bed heat fluxes typically act to reduce the amplitude of diel river temperature fluctuations [*Hondzo and Stefan*, 1994; *Johnson*, 2004; *Neilson et al.*, 2009; *Norman and Cardenas*, 2014; *Storey et al.*, 2003], which has been attributed to temporary storage of heat in streambed sediments that is subsequently returned to the main channel via conductive and/or advective heat fluxes [*Arrigoni et al.*, 2008; *Burkholder et al.*, 2008]. Bed heat flux magnitudes are sensitive to sediment temperature gradients [*Evans et al.*, 1998], rates and extent of advective exchange [*Hester et al.*, 2009], and sediment thermal [*Hondzo and Stefan*, 1994] and hydraulic [*Sawyer and Cardenas*, 2009; *Tonina and Buffington*, 2009] properties. In rivers where bed heat fluxes are significant, river temperature modeling accuracy necessitates reasonable estimates of these river bed characteristics and processes.

Despite extensive study of bed heat fluxes in temperate rivers, little work has been done to understand their influence on rivers in regions of continuous permafrost [Chapter 2; *Wankiewicz*, 1984]. Work that has been done indicates that bed conduction alone does not explain sediment temperature profiles, nor main channel temperature time series under low-flow conditions [Chapter 2; *Wankiewicz*, 1984]. *Wankiewicz* [1984] suggested
that advective exchange is required to explain the temperature profiles in streambed sediments, and the results from Chapter 2 suggest the same mechanism may buffer instream temperatures from the influences of surface heat fluxes under low-flow conditions. Studies of hyporheic exchange in regions of continuous permafrost have documented sufficient hydraulic gradients to produce hyporheic exchange, that the hyporheic zone is important for biochemical regeneration, and that the depth of thaw does not generally limit the depth of hyporheic exchange [Edwardson et al., 2003; Greenwald et al., 2008; Zarnetske et al., 2008]. However, these studies do not estimate the influences of hyporheic exchange on instream river temperatures.

In the Arctic, changes in climate are projected to result in warmer air temperatures, increased precipitation, and increased thaw depth [AMAP, 2012; Chylek et al., 2009; Hinzman et al., 2005; Rawlins et al., 2010]. Estimates of current controls on river temperatures have shown that surface heat fluxes dominate the river energy balance under high flows, however, less is known about the significant heat fluxes under low-flow conditions as presented in Chapter 2. Changes in Arctic climate, coupled with limited understanding of river energy balances to our objective of quantifying the dominant heat fluxes in areas of continuous permafrost under low-flow conditions. Based on the thermal influences of hyporheic exchange on instream temperatures in temperate systems [e.g., Arrigoni et al., 2008], and the presence of a cold (e.g., zero degree Celsius) boundary below permafrost rivers, we hypothesize that hyporheic exchange is an important heat flux that reduces the amplitude of diel temperature fluctuations and reduces mean river temperatures under low-flow conditions. We test
these hypotheses by analyzing heat flux estimates produced with a process based river temperature model that accounts surface and bed heat fluxes. We further provide an estimate of how river temperatures may respond to projections of deeper thaw depths by modeling the sensitivity of river temperatures to changes in ground temperatures below the river bed.

Methods

At reach scales, hyporheic exchange is commonly estimated by using a one zone transient storage model [Bencala, 1983; Haggerty and Reeves, 2002; Hall et al., 2002; Runkel, 1998] that is generally informed by tracer study breakthrough curves. Others have developed models that account for both surface and hyporheic transient storage separately, called two zone models [Briggs et al., 2009; Choi et al., 2000; Harvey and Wagner, 2000]. Neilson et al. [2010b] developed a two zone solute model to also account for heat fluxes and temperature responses. This allows for the combined use of conservative solute breakthrough curves and observations of river temperatures to estimate the influences of surface and subsurface transient storage on river temperature. Accounting for surface and subsurface transient storage zone independently can be important in some systems given that the surface transient storage zone is influenced by atmospheric conditions, river temperatures, and shallow sediment temperatures, while the hyporheic transient storage zone is primarily influenced by the river and deeper sediment temperatures.
While acknowledging the limitations of tracer studies in transient storage model calibration due to the bias towards shorter flow paths [Harvey and Wagner, 2000; Payn et al., 2009], here we coupled tracer study data with temperature observations in the main channel and the subsurface to provide multiple lines of evidence for estimating the reach average vertical extent of the hyporheic zone and exchange rates with the main channel via two zone temperature and solute model calibration at a 1.5 km scale during stable, low-flow conditions. As scale dependence of transient storage model calibration is a known issue when using tracer studies to inform model calibration [e.g., Gooseff et al., 2013], the resulting parameters are then applied to a larger 13.1 km reach to determine if the role of all heat fluxes, including hyporheic exchange, are consistent at larger scales.

For the purpose of this work, we use the term “hyporheic transient storage zone” to refer to the portion of the streambed sediments directly below the main channel with flow paths short enough to influence solute breakthrough curves and main channel river temperatures via advection. By extension, “hyporheic exchange” refers to advective exchanges of mass between the main channel and the hyporheic transient storage zone and “hyporheic exchange heat flux” ($J_{HTS}$) refers to the heat flux associated with hyporheic exchange, which is distinct from the conductive heat flux across the water-sediment interface ($J_{bed}$).

Model Formulation

To support our objective of determining the role of hyporheic exchange on main channel temperatures, the Two-Zone Temperature and Solute (TZTS) transient storage model described in Neilson et al. [2010a]; Neilson et al. [2010b] was applied. This
model is a temporally dynamic, spatially discrete, solute and heat transport model that accounts for individual heat fluxes. Heat and solute are accounted for in the main channel (MC), surface transient storage (STS), hyporheic transient storage (HTS), and surface transient storage sediment (STS$_{sed}$) zones (Figure 3-1). Atmospheric (atm) influences and ground (gr) temperatures are treated as boundary conditions based on observations. In addition to downstream, main channel transport of heat and solute, heat fluxes are estimated across the air-MC, MC-HTS, HTS-gr, air-STS, STS-STS$_{sed}$, STS$_{sed}$-gr, and MC-STS interfaces, while solute fluxes are estimated across the MC-HTS, and MC-STS interfaces only (Figure 3-1).

Figure 3-1. Cross-sectional schematic of a model cell showing zones, boundaries, heat fluxes (solid arrows) and solute fluxes (dashed arrows) between zones. Downstream transport is simulated, but not shown in this figure. Adapted from Neilson et al. [2010a].
Following Neilson et al. [2010a], the governing equations for temperature in the main channel (\(T_{MC}\)), surface transient storage zone (\(T_{STS}\)), hyporheic transient storage zone (\(T_{HTS}\)), and the sediment below the surface transient storage zone (\(T_{STS,sed}\)) are:

\[
\frac{\partial T_{MC}}{\partial t} = -U_{MC} \frac{\partial T_{MC}}{\partial x} + D \frac{\partial^2 T_{MC}}{\partial x^2} + \frac{J_{atm}}{\rho_C Y_{MC}} + \frac{\alpha_{STS} Y_{STS}}{A_{cs,MC} \beta B_{tot}} (T_{HTS} - T_{MC}) + \frac{Q_{HTS}}{V_{MC}} (T_{HTS} - T_{MC}) + \frac{\rho_{sed} c_{p, sed} \alpha_{sed}}{\rho_C Y_{MC} Y_{HTS}} (T_{HTS} - T_{MC})
\]

\(1\)

\[
\frac{\partial T_{STS}}{\partial t} = \frac{J_{atm,STS}}{\rho_C Y_{STS}} + \frac{\alpha_{STS}}{\beta B_{tot}} (T_{MC} - T_{STS}) + \frac{\rho_{sed} c_{p, sed} \alpha_{sed}}{\rho_C Y_{HTS} Y_{STS}} (T_{STS,sed} - T_{STS})
\]

\(2\)

\[
\frac{\partial T_{HTS}}{\partial t} = \frac{\rho_C Q_{HTS}}{\rho_{sed} c_{p, sed} Y_{HTS}} (T_{MC} - T_{HTS}) + \frac{\alpha_{sed}}{Y_{HTS} Y_{gr}} (T_{MC} - T_{HTS}) + \frac{\alpha_{sed}}{Y_{HTS} Y_{gr}} (T_{gr} - T_{HTS})
\]

\(3\)

\[
\frac{\partial T_{STS,sed}}{\partial t} = \frac{\alpha_{sed}}{Y_{HTS} Y_{gr}} (T_{STS} - T_{STS,sed}) + \frac{\alpha_{sed}}{Y_{HTS} Y_{gr}} (T_{gr} - T_{STS,sed})
\]

\(4\)

where \(T\) = temperature (°C), \(t\) = time (s), \(U\) = velocity (m s\(^{-1}\)), \(x\) = distance downstream (m), \(D\) = longitudinal dispersion (m\(^2\) s\(^{-1}\)), \(\rho\) = density of water (kg m\(^{-3}\)), \(C_p\) = specific heat capacity of water (J kg\(^{-1}\) °C\(^{-1}\)), \(Y\) = zone depth (m), \(\alpha_{STS}\) = surface transient storage exchange coefficient (m\(^2\) s\(^{-1}\)), \(A_{cs}\) = cross sectional surface area (m\(^2\)), \(\beta\) = STS fraction of total width, \(B_{tot}\) = total channel width (m), \(Q_{HTS}\) = hyporheic exchange coefficient (m\(^3\) s\(^{-1}\)), \(V\) = zone volume (m\(^3\)), \(\rho_{sed}\) = density of sediment (kg m\(^{-3}\)), \(C_{p,sed}\) = specific heat capacity of sediment (J kg\(^{-1}\) °C), \(\alpha_{sed}\) = coefficient of thermal diffusivity of sediment (m\(^2\) s\(^{-1}\)), \(T_{gr}\) = temperature of the ground (°C) at depth \(Y_{gr}\) (m) from water-sediment interface, and \(J_{atm}\) = atmospheric heat flux (W m\(^{-2}\)), defined as

\[
J_{atm} = J_{sn} + J_{NetLW} - J_c - J_e
\]

\(5\)

where \(J_{sn}\) = net shortwave radiation (310 – 2800 nm) (W m\(^{-2}\)), \(J_{NetLW}\) = net longwave
radiation calculated as the sum of atmospheric and water longwave radiation (W m\(^{-2}\)), \(J_c\) = sensible heat flux (W m\(^{-2}\)), and \(J_e\) = latent heat flux (W m\(^{-2}\)). Equations for \(J_{sn}\), \(J_{NetLW}\), \(J_c\), and \(J_e\) are given in Chapra [1997].

Simplified forms of the temperature governing equations were used for solute concentration (\(C\)) given fewer exchange interfaces and processes:

\[
\frac{\partial C_{MC}}{\partial t} = -U_{MC} \frac{\partial C_{MC}}{\partial x} + D \frac{\partial^2 C_{MC}}{\partial x^2} + \frac{\alpha_{STS}}{\alpha_{MC} \beta_{tot}} (C_{STS} - C_{MC}) + \frac{q_{HTS}}{v_{MC}} (C_{HTS} - C_{MC})
\]  

(6)

\[
\frac{\partial C_{STS}}{\partial t} = \frac{\alpha_{STS}}{(\beta_{tot})^2} (C_{MC} - C_{STS})
\]  

(7)

\[
\frac{\partial C_{HTS}}{\partial t} = \frac{q_{HTS}}{v_{HTS}} (C_{MC} - C_{HTS})
\]  

(8)

Hydraulic routing is approximated using a kinematic wave approach similar to Cardenas et al. [2014]

\[
\frac{\partial Q}{\partial x} + \left( \left( \frac{n \beta_{tot}^{2/3}}{\sqrt{S_o}} \right)^{3/5} Q^{-2/5} \right) \frac{\partial Q}{\partial t} = q
\]  

(9)

where \(Q\) = volumetric discharge (m\(^3\) s\(^{-1}\)), \(n\) = Manning’s roughness coefficient, \(S_o\) = channel slope (m m\(^{-1}\)), and \(q\) = lateral inflows per unit length (m\(^3\) s\(^{-1}\) m\(^{-1}\)). The only modification to the model from earlier applications [e.g., Neilson et al., 2010a] was to define depth to ground (\(Y_{gr}\)) relative to the river bed, rather than to the bottom of the hyporheic transient storage zone.

Model cell length (\(\Delta x\)) and calculation time step were set to 10 m and 10 s respectively to approximate estimated longitudinal dispersion with numerical dispersion, while also meeting the Courant condition [Chapra, 1997; Fischer, 1973; McQuivey and Keefer, 1974]. All governing heat and mass equations were solved with an explicit Euler
method using an upwind differencing scheme [Chapra, 1997]. Consistent with Gooseff et al. [2013], mean residence time in the hyporheic zone \( R_{HTS} \) is estimated as:

\[
R_{HTS} = \frac{(B_{tot}(1-\beta) + \Delta x + Y_{HTS})}{Q_{HTS}}
\]  

(12)

where the numerator on the right-hand side of the equation represents the volume of the hyporheic transient storage zone.

**Study Site**

A fourth-order segment of the Kuparuk River, Alaska (Figure 3-2) was selected to represent a common Arctic alluvial river underlain by continuous permafrost. The study area is composed of a 1.5 km long study reach used for model calibration (Figure 3-2a) located within a 13.1 km long test reach (Figure 3-2b) used for evaluation of reach scale influences of hyporheic exchange on river temperatures. Site names are consistent with Chapter 2.

The mean annual air temperature for this region is approximately \(-10\, ^\circ C\), with mean summer (June-Aug) air temperatures of \sim 10\, ^\circ C\. The site is underlain by continuous permafrost \( 300 - 600 \) m thick [Osterkamp and Payne, 1981], with a perennial surface thaw reaching a maximum depth of less than one meter in terrestrial environments [Nelson et al., 1997].

River discharge in this basin has a strong seasonal component with high flows coinciding with the spring freshet, usually in May. After the perennial snow pack has melted, summer (June-Aug) discharge recedes rapidly to predominantly low-flow conditions interrupted by short duration, high flow events driven by convective storms and a lack of groundwater storage due to shallow terrestrial thaw.
Figure 3-2. Map of 1.5 km long calibration reach (A) within the 13.1 km model test reach (B) from Site 9 to Site 8. Arrows indicate the direction of flow. Distances are from tracer injection site at the upstream boundary of the calibration reach. Main channel temperature, specific conductance, and discharge were measured at calibration and test reach boundary sites. Main channel and subsurface temperature and specific conductance were measured at the piezometer sites.

[McNamara et al., 1998]. Discharge typically increases in autumn (Sept) in response to large-scale frontal precipitation events before returning to a frozen state through the winter. During the three months of ice free and low-flow conditions (June-Aug), river
temperatures in the Kuparuk River are generally observed to increase in the downstream direction with diurnal temperature ranges of ~6 °C +/− 3 °C [Chapter 2].

Bed substrate in the study reach is characterized as cobble deposited following the Wisconsinian glaciation with mean particle size of 70 mm [Oatley, 2002]. Thaw depths under alluvial streams in the vicinity of the Kuparuk River are highly variable between geomorphic features, and reach maximum depths between 0.5 and 3 meters in August before refreezing in the fall and winter [Brosten et al., 2006; Brosten et al., 2009].

**Data Collection**

Data were collected to provide: 1) field observations of hyporheic exchange, 2) external forcing and boundary condition data for temperature modeling, 3) calibration data for a 1.5 km long calibration reach (Figure 3-2a), and 4) test data for the 13.1 km long test reach (Figure 3-2b). Data collection included measurements of temperature in the main channel and sediments, solute breakthrough curves in the main channel and sediments, discharge, meteorological conditions, and river top widths every meter within both reaches (Table B-1, Figure 3-2).

**Field Observations of Hyporheic Exchange**

*Solute Data:* Pulse injections of sodium chloride at the upstream boundary of the tracer study reach on 4, 7, and 10 July 2017 produced solute breakthrough curves recorded in the main channel and river bed of the calibration reach. 38.4, 39.0, and 30.0 kg of sodium chloride (NaCl) were used in these tracer studies, respectively, and were injected nearly instantaneously. Specific conductivity was measured at one minute intervals with AquaTroll200 conductivity-temperature-depth loggers deployed in the
main channel at 500 m, 600 m, and 1200 m, and 1500 m downstream from the injection site (Figure 3-2). Specific conductance breakthrough curves were calculated as the deviation from background specific conductivity and converted to solute breakthrough curves using standard curves developed using stream water collected on site [Goosoff and McGlynn, 2005; Payn et al., 2009]. Background specific conductance was collected at the upstream boundary with an AquaTroll200 deployed 10 m upstream of the tracer study injection location.

Solute breakthrough curves were also measured in the sediments with AquaTroll200’s in partially screened, three cm outside diameter piezometers driven into the river bed to depths between 14 and 100 cm at 500 m (n = 2-5), 600 m (n = 2-3), and 1200 m (n = 3) downstream of the injection site. Some piezometers were relocated between tracer studies in response to preliminary data processing. Piezometers were installed in the river bed by driving a small diameter iron bar into the sediments, driving an iron casing around the bar, replacing the bar with the piezometer, removing the casing, and allowing the river bed to collapse around the piezometer. Reported piezometer depths are to the top of the 11 cm long screens. AquaTroll200’s deployed in these piezometers were set to record at 1 minute intervals for the duration of each tracer study. Absence of a specific conductance breakthrough curve within any given piezometer was assumed to be caused by that piezometer being located below the hyporheic transient storage zone.

Temperature Data: The AquaTroll200 loggers also recorded temperature time series in the main channel and piezometers, which were used as a second line of evidence
of surface-subsurface advective exchange. Locations with similar diel amplitude to the main channel provided evidence of hyporheic exchange, while temperature time series that differed significantly from the main channel suggested that piezometer screens were located below the hyporheic transient storage zone. In the latter case, daily variability in temperature were taken to primarily represent the influences of bed conduction only.

Similar to Fanelli and Lautz [2008], temperature amplitude ratios were used to evaluate the similarity streambed temperatures to the main channel. Amplitude ratios for the subsurface temperature records are expressed as a percentage of the main channel temperature amplitudes:

\[
A_{r,d,x} = \frac{(T_{\text{max}} - T_{\text{min}})_{\text{PZ,d,x}}}{(T_{\text{max}} - T_{\text{min}})_{\text{MC,x}}} \times 100
\]

(13)

where \(A_{r,d,x}\) is the temperature amplitude ratio for depth \(d\) (cm) at piezometer location \(x\) (m), \(T_{\text{max}}\) and \(T_{\text{min}}\) are the maximum and minimum temperatures, respectively, and subscripts PZ and MC are the piezometer and main channel, respectively.

Data collection for river temperature modeling

Meteorological information needed to estimate the surface heat fluxes \(J_{\text{sn}}, J_{\text{NetLW}}, J_{c}\), and \(J_{e}\) were recorded at the University of Alaska, Fairbank’s Upper Kuparuk Meteorological Station (UK Met, Figure 3-3-2a). These parameters include hourly measurements of air temperature, wind speed, relative humidity, and incoming shortwave radiation. These observations were used for both model calibration and testing.

Sediment temperature measurements in addition to those made in the piezometers
were recorded every four hours at 10, 20, 30, 40, 50 and 60 cm below the water-sediment interface at Site 9 using a T-Rod (Alpha-Mach, Montreal, Canada). Temperature recordings showing less than 10% of diel ranges observed in the main channel were used to represent the temperature of the ground below the hyporheic transient storage zone \( (T_{gr}) \) [Silliman and Booth, 1993; Storey et al., 2003]. These \( T_{gr} \) values were assumed to hold for the calibration and test reaches while the reach-average depth to the location of this temperature \( (Y_{gr}) \) was subject to calibration.

Upstream boundary conditions of flow and temperature were collected for the test and calibration reaches (Figure 3-2). Observations of discharge were made using a FlowTracker ADV (Onset Corporation, Borne, Massachusetts). For Site 9, discharge observations made over the summers of 2013-2017 were related to stage measured with a CS-451 pressure transducer (Campbell Scientific, Logan, Utah) and logged in a CR200x data logger (Campbell Scientific, Logan, Utah) to produce stage-discharge rating curves. Hydrographs for Site 9 were produced from half-hourly measurements of stage. For the calibration reach, point observations of discharge using the FlowTracker were made for each study period. Discharge estimates at the downstream locations of the two study reaches, 1500 m and Site 8, were made similar to those for the injection site and Site 9, respectively. Upstream and downstream boundary condition flows were monitored for stability and differenced to verify negligible lateral inflows (e.g., shallow groundwater contributions from the landscape) for the duration of the simulations. Temperatures at the upstream boundaries were recorded with a Hobo ProV2 temperature logger (Onset Corporation, Borne, Massachusetts) at Site 9 recording every half hour for the test reach,
and an AquaTroll200 sensor at US TS logging every one minute for the calibration reach.

Due to the potential sensitivity of heat transfer across the air-water interface, it was necessary to determine accurate river surface areas. High resolution (14 cm) near infrared (NIR) aerial imagery collected in the summers of 2013 – 2015 with a custom payload mounted to a Robinson 44 helicopter was used to estimate river wetted width every one meter along the test reach (see Chapter 4 for details). River widths were extracted from mosaicked, orthorectified, and georeferenced NIR imagery using a binary threshold of brightness to differentiate between water and land and validated with ground observations.

**Model Calibration**

In this model formulation, there are seven model parameters that require calibration: surface transient storage fraction ($\beta$), channel roughness ($n$), surface transient storage cross sectional area ($A_{C,STS}$), surface transient storage exchange coefficient ($\alpha_{STS}$), depth of hyporheic exchange ($Y_{HTS}$), hyporheic exchange coefficient ($Q_{HTS}$), and depth to ground temperature ($Y_{gr}$). Potential parameter ranges permitted within calibration for $Y_{HTS}$ and $Y_{gr}$ were set based on field observations. $Y_{HTS}$ values were set to range from zero to the maximum depth of observed solute breakthrough curves and near main channel temperature patterns in piezometers. Note that including a zero depth in the calibration range for $Y_{HTS}$ allows for parameter selection that excludes hyporheic exchange thus allowing us to test if hyporheic exchange is necessary to reproduce observed main channel temperature and solute records. The calibration range for $Y_{gr}$ was set based on the range of piezometers depths at which sediment temperatures measured in
piezometers were similar to temperatures observed at the 60 cm depth at Site 9 (Figure B-1). The remaining model parameters were set from literature (see the supporting information for a full description; Text B-1, Table B-2).

The Multi Objective Shuffled Complex Evolutionary Metropolis (MOSCEM) calibration algorithm [Vrugt et al., 2003] was used to calibrate the 7 free model parameters using observed main channel temperature time series and solute breakthrough curves at 1500 m during the 7 July 2017 simulation. The calibration routine iteratively samples the seven-dimensional parameter space given the bounds for each parameter (Table B-2) and evaluates the predicted temperature and solute time series against observations using the Nash Sutcliff Efficiency (NSE) [Nash and Sutcliffe, 1970] objective function. This process was repeated 10,000 times in an attempt to determine a parameter set that minimizes objective functions for both temperature and solute. Calibrating against two objective functions (e.g., temperature and solute) results in a tradeoff in objective function space with equally ideal solutions existing along a Pareto Front with improvement in one objective function resulting in a reduction in the complementary objective function (e.g., Figure B-2). The Pareto Optimal solution has the shortest Euclidian distance from the origin of objective function space, provides a compromise between the two objective functions, and was selected as the best model parameter set. The optimal parameter sets for individual objective functions are located at opposite ends of the Pareto Front. This multiple objective function approach allowed us to identify the degree of tradeoff between information provided by temperature and solute observations.
Model Testing

To determine if the parameter set calibrated for the 1.5 km long study reach captured temperature responses over longer length scales, the optimal parameter set was applied to the 13.1 km long test reach from Site 9 to Site 8. This was first done for a three day period in 2017 that includes the 7 July 2017 tracer study to ensure that the only difference between the calibration and test simulations was a change in spatial scale. To determine if the role of hyporheic transient storage was consistent between different low-flow periods, the model was then tested for a seven day low-flow period in June 2015, and a four day low-flow period in July 2013 identified in Chapter 2. NSE and root mean square error objective functions (RMSE) were used to evaluate model performance for these simulations.

Heat Flux Analysis

Individual heat flux magnitudes were analyzed to determine the dominant processes controlling temperatures and provide information on the diurnal patterns and competing processes. Dominant heat fluxes were identified as those contributing at least 20% to the total main channel energy budget at any time over the simulation period. The total main channel energy budget is calculated as the sum of magnitudes for all heat fluxes across the air-water, water-sediment, and MC-STS interfaces shown in Figure 3-1, and does not include downstream advective heat transport. Temporal patterns were analyzed by grouping heat fluxes by hour of day and determining the mean, minimum, and maximum heat flux magnitudes and percent contributions as a function of time of day.
Model Scenarios

Two model scenarios were run, with the first scenario providing information on the influence of hyporheic exchange on river temperatures. For this scenario, the 13.1 km long model test runs for 2013, 2015, and 2017 were re-run setting the hyporheic exchange coefficient ($Q_{HTS}$) to zero to simulate the absence of hyporheic exchange. Model results with and without hyporheic exchange were compared to determine the influence of hyporheic exchange on river temperatures. The second model scenario was run with the objective of understanding potential impacts of deeper thaw depths on river temperature. We assume that thaw depth and ground temperature are highly correlated, similar to what is seen for shallow sediments in terrestrial environments [Frauenfeld et al., 2004]. As such, we varied observed ground temperatures by +/- 1, 2 and 4 °C to determine the sensitivity of river temperatures to these changes that act as an analog to sensitivity to thaw depth. Further sensitivity analyses were performed to determine which heat fluxes contribute to the main channel temperature response to changes in ground temperatures.

Results

Field Observations of Hyporheic Exchange

Solute Breakthrough Curves: Solute breakthrough curves were observed in the main channel at all piezometer locations (Figures 3-3, B-3, and B-4). No subsurface solute breakthrough curves were observed in the piezometers at 500 m (Table 3-1, Figures 3-3A, B-3A, and B-4A). Subsurface solute breakthrough curves were observed
Table 3-1. Presence of solute breakthrough curves (BTC) and temperature amplitude ratio ($A_r$) for piezometers records

<table>
<thead>
<tr>
<th>Site</th>
<th>Depth [cm]</th>
<th>4 July 2017</th>
<th>7 July 2017</th>
<th>10 July 2017</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$A_r$ BTC</td>
<td>$A_r$ BTC</td>
<td>$A_r$ BTC</td>
<td></td>
</tr>
<tr>
<td>500 m</td>
<td>14</td>
<td>44% No</td>
<td>37% No</td>
<td>64% No</td>
</tr>
<tr>
<td></td>
<td>18</td>
<td>43% No</td>
<td>- -</td>
<td>- -</td>
</tr>
<tr>
<td>600 m</td>
<td>30</td>
<td>30% No</td>
<td>14% No</td>
<td>- -</td>
</tr>
<tr>
<td></td>
<td>57</td>
<td>17% No</td>
<td>- -</td>
<td>- -</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>7% No</td>
<td>- -</td>
<td>- -</td>
</tr>
<tr>
<td>1200 m</td>
<td>18</td>
<td>- -</td>
<td>- -</td>
<td>96% Yes</td>
</tr>
<tr>
<td></td>
<td>31</td>
<td>- - 48% No</td>
<td>No</td>
<td>52% No</td>
</tr>
<tr>
<td></td>
<td>69</td>
<td>- - 10% No</td>
<td>- -</td>
<td></td>
</tr>
<tr>
<td></td>
<td>73</td>
<td>- - 10% No</td>
<td>- -</td>
<td></td>
</tr>
<tr>
<td></td>
<td>57</td>
<td>99% Yes</td>
<td>97% Yes</td>
<td>- -</td>
</tr>
<tr>
<td></td>
<td>64</td>
<td>75% Yes</td>
<td>63% Yes</td>
<td>- -</td>
</tr>
<tr>
<td></td>
<td>77</td>
<td>38% No</td>
<td>16% No</td>
<td>- -</td>
</tr>
</tbody>
</table>

at a depth of 18 cm, but not at 31 cm at 600 m (Table 3-1, Figures 3-3C, B-4C). At 1200 m, solute breakthrough curves were observed at 64 cm, but not 77 cm depths (Table 3-1, Figures 3-3E and B-3C). These observations indicate that the depth of hyporheic exchange ($Y_{HTS}$) varied from less than 14 cm at 500 m to up to 77 cm at 1200 m.

*Temperature Time Series*: Diel temperature ranges in piezometers where solute breakthrough curves were observed were similar to the main channel (Figures 3-3, B-3, and B-4) having an average amplitude ratio of 86% (StDev = 16%) (Table 3-1). In comparison, the average amplitude ratio in piezometers where solute breakthrough curves were not observed was only 31% (StDev = 18%) (Table 3-1). These temperature observations provided a second line of evidence, in addition to the solute observations, for setting the calibration bounds on the depth of hyporheic exchange ($Y_{HTS}$), which was set to 0 – 70 cm (Table B-2).
Figure 3-3. 7 July 2017 tracer study solute breakthrough curves (left) and temperature time series (right) in the main channel (black lines) and at various depths in the sediment (colored lines) at 500 (top), 600 (middle row), and 1200 (bottom) meters downstream from the injection location.

Sediment Temperature Profiles: Similar to the temperature records collected in the piezometers, the sediment temperature profile at Site 9 showed notable decreases in mean and amplitude of temperatures with depth during the period of tracer studies (4 July – 12 July 2017) (Figure 3-4). Temperatures at 60 cm had a daily amplitude less than 10% of the main channel and were used to prescribe $T_{gr}$. Sediment temperatures at 57 cm at
Figure 3-4. Distribution (left) and time series (right) of main channel (MC) and sediment temperatures measured at Site 9 in 10 cm increments from the water-sediment interface over the period of tracer studies in 2017. Mean and range in temperature values decreased with depth. Temperatures from the 60 cm deep sensor were used as the ground temperature boundary ($T_{gr}$) in transient storage modeling.

500 m and 69 cm at 600 m were similar to $T_{gr}$ observed at the 60 cm depth at Site 9 while the temperatures at site 1200 m were much warmer than at Site 9 to depths of 77 cm (Figure B-1). From these observations, the calibration for $Y_{gr}$ was set to 50 – 100 cm (Table B-2).

**Model Calibration Results**

The Pareto optimal parameter set reproduced observed solute BTC and temperature at 1500 m reasonably well (solute NSE = 0.91, solute RMSE = 0.79 mg Cl\textsuperscript{-} L\textsuperscript{-1}, temperature NSE = 0.87, temperature RMSE = 0.61 °C) (Figures 3-5A and 3-5B). When compared with the calibration parameter ranges, the Pareto optimal model calibration selected relatively high values for channel roughness ($n$) and surface transient storage cross sectional area ($A_{CS,STS}$), moderate values for hyporheic transient storage exchange rate ($Q_{HTS}$), surface transient storage exchange rate ($a_{STS}$) and storage width fraction ($\beta$), and low values for depth of hyporheic exchange ($Y_{HTS}$) and ground depth...
Figure 3-5. 7 July 2017 measured (blue points) and modeled (red lines) solute (left) and temperature (right) records at the downstream end of the model calibration reach using the Pareto optimal (A, B), solute end-member (C, D), and temperature end-member (E, F) parameter sets. Grey regions in A and B represent the range of Pareto front model calibrations for solute and temperature, respectively. Solute is plotted in hours from injection time, and temperature records as time. Period of record in the solute panels is indicated by vertical dashed lines in temperature panels. Solute and temperature data were collected and analyzed at one minute intervals, and are shown at five and 30 minute intervals, respectively, for graphical clarity.

\((Y_{yr})\) (Table B-2, Figure B-5). The Pareto optimal parameter set is more similar to the solute end-member than the temperature end-member, showing that the solute predictions are more sensitive to parameter changes (Table B-2, Figure B-5). This is further seen by all Pareto front calibrations reproducing temperatures reasonably well (as illustrated by narrow grey bands in Figure 3-5B), while the temperature end-member calibration did
not reproduce the observed main channel solute breakthrough curves (Figure 3-5E) and the Pareto front resulted in large solute bounds (Figure 3-5A). Calculated residence times in the hyporheic zone ($R_{HTS}$) were 2.7, 1.9, and 13.1 minutes for the Pareto optimal, solute, and temperature end-members, respectively (Table B-2). Understanding that the model produced a reach-average representation of hyporheic exchange influences, predicted hyporheic transient storage temperature and solute responses lag the main channel signals by 11 minutes and are dampened compared with the main channel (Figure B-6), similar to observations at 57 cm depth at 1200 m (Figure 3-3).

**Model Testing Results**

Using the optimal calibration parameter set, the model produced reasonable temperature results for the 13.1 km long test reach from Site 9 to Site 8 for the three day test period in 2017 (NSE = 0.75, RMSE = 0.81 °C), indicating that the 1.5 km long reach average parameter calibration is reasonable over longer spatial scales (Figure 3-6A). Flow conditions were low, less than one m$^3$ s$^{-1}$, and stable for the model test periods in 2013, 2015, and 2017 allowing for comparison of hyporheic heat fluxes under similar conditions across multiple years (Figure B-7). Model results from 2013 (NSE = 0.91, RMSE = 0.64 °C) and 2015 (NSE = 0.92, RMSE = 0.49 °C) further indicate that the 2017 calibration held for low-flow conditions between years (Figures 3-6C, 3-6E). This suggests that the influences of hyporheic exchange have a similar impact on instream temperatures during these warm-season, low-flow conditions.

**Heat Flux Analysis Results**

Shortwave radiation ($J_{sw}$) was consistently positive (gain in heat by the main
Figure 3-6. Left: Modeled (red line) and observed (blue dots) temperatures for the test reach at Site 8 for low-flow periods in 2017 (A), 2015 (C), and 2013 (E) using parameters estimated from the calibration reach in 2017. Right: Mean (lines) and range (whiskers) of diurnal heat flux patterns for the dominant heat fluxes $J_{sn}$ (black), $J_{HTS}$ (red), $J_{NetLW}$ (blue), and $J_{e}$ (green) for the simulation periods in 2017 (B), 2015 (D), and 2013 (F).

channel), net longwave radiation ($J_{NetLW}$) was consistently negative (loss of heat from the main channel), and the influence of the remaining main channel heat fluxes ($J_{c}, J_{e}, J_{HTS}, J_{bed}, J_{sts}$) varied between positive and negative within each 24 hour period. On average, $J_{HTS}, J_{sts}, J_{bed},$ and $J_{e}$ were negative while $J_{c}$ was positive. $J_{sn}, J_{NetLW}, J_{HTS},$ and $J_{e}$ were the
dominant heat fluxes, each contributing at least 20% to the total heat flux at any given
time in the simulation periods (Table 3-2). $J_{sn}$ had by far the largest in magnitude,
reaching values of 706 W m$^{-2}$. The mean (~30%) and maximum (~60%) percent
contribution for $J_{sn}$ and $J_{NetLW}$ were similar (Table 3-2). Maximum percent contributions
from $J_{sn}$ occurred during the day under peak shortwave radiation, while the maximum
percent contribution from $J_{NetLW}$ occurred at night (Figure 3-6). Maximum percent
contributions from $J_{HTS}$ (31%) occurred during the day while maximum percent
contributions from $J_e$ (30%) occurred at night when $J_{HTS}$ was transitioning from a
negative to a positive heat flux (Figure 3-6). $J_{HTS}$ ranged from -267 to 120 W m$^{-2}$ and
represented the largest range in heat fluxes that act as both a heat source and a heat sink.
Bed conduction ($J_{bed}$) was essentially negligible, representing ≤1% of the total heat flux
for all simulations. The diurnal patterns in dominant heat fluxes show that the hyporheic
exchange heat flux directly opposed shortwave radiation for all simulation periods
(Figures 3-6B, 3-6D, 3-6F).

**Scenario Results**

The first model scenario, where river temperatures were modeled over the 13.1
km long test reach without hyporheic exchange (e.g., $Q_{HTS} = 0$), resulted in an increase in
average river temperatures of 0.23 °C (0.02 °C km$^{-1}$) and an increase in diurnal amplitude
of instream temperatures of up to 2.3 °C (0.2 °C km$^{-1}$) (Figure B-8). The degree of
buffering varied by year with 2013 and 2017 showing greater buffering than 2015, when
flows were highest. Results from the second model scenario, where sensitivity of main
channel temperatures to ground temperatures were tested, revealed that main channel
Table 3-2. Statistics for main channel and ground heat flux magnitudes (W m\(^{-2}\)) and percent of total main channel energy balance accounting for heat fluxes shown in Figure 3-1 (%) for the 13.1 km long test reach from Site 9 to Site 8.

<table>
<thead>
<tr>
<th>Year</th>
<th>J_{sn} W m^2</th>
<th>J_{NetLW} W m^2</th>
<th>J_{HTS} W m^2</th>
<th>J_{STS} W m^2</th>
<th>J_{bed} W m^2</th>
<th>J_{e} W m^2</th>
<th>J_{e} W m^2</th>
<th>J_{gr} W m^2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Min</td>
<td>8</td>
<td>-186</td>
<td>10</td>
<td>-267</td>
<td>&lt; 1%</td>
<td>-17</td>
<td>&lt; 1%</td>
<td>-13</td>
</tr>
<tr>
<td>2017</td>
<td>Mean</td>
<td>300</td>
<td>37%</td>
<td>-153</td>
<td>30%</td>
<td>-84</td>
<td>15%</td>
<td>-7</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>702</td>
<td>56%</td>
<td>-121</td>
<td>60%</td>
<td>87</td>
<td>23%</td>
<td>3</td>
</tr>
<tr>
<td>Min</td>
<td>3</td>
<td>&lt; 1%</td>
<td>-178</td>
<td>10%</td>
<td>-193</td>
<td>&lt; 1%</td>
<td>-72</td>
<td>&lt; 1%</td>
</tr>
<tr>
<td>2015</td>
<td>Mean</td>
<td>232</td>
<td>33%</td>
<td>-143</td>
<td>29%</td>
<td>-10</td>
<td>13%</td>
<td>-27</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>648</td>
<td>65%</td>
<td>-115</td>
<td>60%</td>
<td>120</td>
<td>31%</td>
<td>7</td>
</tr>
<tr>
<td>Min</td>
<td>4</td>
<td>1%</td>
<td>-186</td>
<td>10%</td>
<td>-291</td>
<td>&lt; 1%</td>
<td>-54</td>
<td>&lt; 1%</td>
</tr>
<tr>
<td>2013</td>
<td>Mean</td>
<td>294</td>
<td>35%</td>
<td>-151</td>
<td>30%</td>
<td>-93</td>
<td>14%</td>
<td>-3</td>
</tr>
<tr>
<td></td>
<td>Max</td>
<td>706</td>
<td>60%</td>
<td>-121</td>
<td>61%</td>
<td>86</td>
<td>23%</td>
<td>7</td>
</tr>
</tbody>
</table>

*Note: Percentages will not sum to 100% as the timing of percent contribution statistics vary between heat fluxes.*
temperatures varied by 0.2 °C for every 1 °C alteration to ground temperature during 7
day period in 2015 (Figure B-9). $J_{gr}$ and $J_{HTS}$ varied by 8 and 9 W m$^{-2}$ per °C change in
ground temperature, respectively, while all other heat fluxes, including $J_{bed}$, varied by
less than 1 W m$^{-2}$ per °C change in ground temperature.

Discussion

From observations of solute breakthrough curves in the subsurface (Figures 3-3, B-3, and B-4) it is clear that hyporheic exchange is present to depths of up to 64 cm in the
calibration reach. Additionally, the absence of solute breakthrough curves in some
piezometers indicates that depth of thaw does not limit the depth of hyporheic exchange,
confirming model and tracer study results reported of a smaller, second order alluvial
stream within the Kuparuk River watershed [Greenwald et al., 2008; Zarnetske et al.,
2008]. From temperature records at locations where solute breakthrough curves were
observed, it is apparent that hyporheic exchange impacts sediment temperatures (Figure
3-3). This exchange, however, is spatially heterogeneous as seen in solute breakthrough
curve depths ranging from less than 14 cm to at least 64 cm within a 700 m reach of river.
Such spatial heterogeneity in hyporheic exchange attributed to local hydraulics [Norman
and Cardenas, 2014; Zarnetske et al., 2008] and heterogeneity in stream bed sediments
[Cozzetto et al., 2013; Pryshlak et al., 2015; Sawyer and Cardenas, 2009; Tonina et al.,
2016] is well documented, including in Arctic alluvial rivers [Edwardson et al., 2003;
Zarnetske et al., 2008]. While spatial heterogeneity makes it difficult to capture
subsurface conditions at small spatial scales, these modeling results presented here
capture the bulk of the main channel response of the solute breakthrough curves and
temperature signals (Figure 3-5).

Pareto front tradeoffs in model calibration, resulting from calibration to
temperature and solute, again highlights that the model does not perfectly capture the heat
and solute transport processes influencing main channel responses. One likely reason for
this tradeoff is the difference in windows of detection between temperature and solute
[Harvey et al., 1996]. The solute end-member on the Pareto front favors shallow depths
of hyporheic exchange ($Y_{HTS} = 9$ cm) and transient storage with short residence times
($R_{HTS} = 1.9$ minutes). This is different from the temperature end-member that selected
deeper hyporheic depths ($Y_{HTS} = 33$ cm) resulting in longer average travel times ($R_{HTS} =
13.1$ minutes) (Table B-2). This suggests that the flow paths influencing main channel
temperatures are likely longer than those affecting main channel solute breakthrough
curves. The Pareto optimal solution, representing a compromise between these two
windows of detection that is notably influenced more by solute than temperature,
incorporates the information from both signals ($Y_{HTS} = 11$ cm, $R_{HTS} = 2.7$ minutes).
While the Pareto front end-members perform better for their respective objective
function, the Pareto optimal solution still captures the majority of the temperature and
solute signals, indicating that the model reasonably captures the dominant processes.

For both the Pareto optimal and temperature end-member model calibrations, $J_{HTS}$
is seen to be a significant heat flux in this system (Table 3-2), consistent with our
hypotheses. The magnitude and percent of the overall heat flux contributed from $J_{HTS}$
observed here are higher than for other systems where these metrics have been reported [Arrigoni et al., 2008; Caissie, 2006; Hester et al., 2009; Loheide and Gorelick, 2006; Neilson et al., 2009]. For example, the 31% percent contribution from hyporheic exchange greatly exceeded the 3-5% reported in Neilson et al. [2009] for the Virgin River, Utah. Further, we estimate that the magnitude of $J_{HTS}$ ranged from -291 to 120 W m$^{-2}$, which is wider than the range of -145 to 120 W m$^{-2}$ reported by Caissie et al. [2014] for Catamaran Brook, Canada. A majority of this difference in heat flux magnitude is in the negative term, indicating that these systems differ in heat loss rather than heat gain from hyporheic exchange. A likely cause for this is the presence of permafrost below the Kuparuk River, which enhances the heat loss from hyporheic exchange through strong sediment thermal gradients during the day (Figure 3-4).

The large diurnal swings in $J_{HTS}$ (Figure 3-6) resulted in a buffering of main channel temperatures from surface heat fluxes. This is consistent with temperate systems [Brown, 1969; Burkholder et al., 2008; Caissie and Luce, 2017; Evans et al., 1998; Sinokrot and Stefan, 1993]. Model testing results, suggesting that the diurnal swing in $J_{HTS}$ buffers main channel temperatures up to 0.2 °C per km, indicate that $J_{HTS}$ is likely the missing mechanism identified in Chapter 2 and should be accounted for in Arctic river temperature modeling under low-flow conditions.

As mentioned before, the magnitude of heat loss from $J_{HTS}$ is larger than its heat gain. The net effect is that $J_{HTS}$ acts to cool the main channel with an average heat flux of -62 W m$^{-2}$ (Table 3-2). This is consistent with our hypotheses, and is counter to the findings of Neilson et al. [2009] where hyporheic exchange was identified as net heat
source. Conduction of heat from the hyporheic transient storage zone to the ground below \( J_{gr} \) likely causes \( J_{HTS} \) to act as a net heat sink. This is highlighted by the fact that mean heat fluxes for \( J_{gr} \) and \( J_{HTS} \) are within 3 W m\(^{-2} \) of each other in all three test periods, showing that the same amount of heat that is passed to the sediment via \( J_{HTS} \) is removed by \( J_{gr} \) (Table 3-2).

While \( J_{gr} \) currently influences \( J_{HTS} \) to act as a heat sink, increases in ground temperatures may reduce the heat sink capacity of \( J_{HTS} \), as seen by the sensitivity of main channel temperatures to ground temperatures (Figure B-9). This is consistent with our hypotheses, and is driven by reduction in heat losses via \( J_{HTS} \) and \( J_{gr} \) under scenarios of warmer ground temperatures (Figure B-9). The sensitivity of main channel temperatures to ground temperatures suggests that analyses of climate change influences on Arctic river temperatures should include attempts to estimate the impact of projected increases in thaw depths [e.g., AMAP, 2012] on ground temperatures. Additionally, analyses of the impacts of thaw depth on temperature dependent biogeochemical processes in the hyporheic zone should include changes in temperature in addition to changes (or lack thereof) in residence times and vertical extent as suggested by Zarnetske et al. [2008] and Greenwald et al. [2008].

While this work captures the dominant thermal influences of hyporheic exchange, multiple questions remain. First, this work assumes that hyporheic exchange does not vary with sediment temperature. Recent work by Cozzetto et al. [2013] shows that diurnal variation in hyporheic transient storage zone temperatures impact the depth of hyporheic exchange in an alluvial stream in Antarctica. This builds upon the work of
that demonstrated temperature dependency of hydraulic conductivity due to changes in viscosity. Future work on hyporheic exchange in Arctic rivers should focus on these temperature dependent effects on flow through the hyporheic zone. Second, while model calibration held for low-flow conditions across multiple years, it is likely limited to periods with similar hydraulic (e.g., the Darcy-Weisbach friction factor and unit stream power [Zarnetske et al., 2007]) and antecedent conditions [Ward et al., 2013]. As such, additional testing is required to determine the thermal influences of hyporheic exchange under intermediate flow conditions. We predict that the influence of hyporheic exchange on main channel temperatures is highest under low-flow conditions, and will decrease with discharge as surface heat fluxes and lateral inflows become dominant [Chapter 2]. Lastly, future work should focus on extending these analyses to other Arctic rivers to determine the transferability and limitations of the findings presented here.

Conclusions

This paper identifies hyporheic exchange as a controlling process on main channel temperatures in an Arctic alluvial river under low-flow conditions. Additionally, this work highlights the utility of using both temperature and solute records to provide multiple lines of evidence for the influence of hyporheic exchange on main channel temperatures through interpretation of field observations and model calibration.

Results from field observations of temperature and solute presented clear
evidence of hyporheic exchange, while highlighting spatial heterogeneity in this process. Solute breakthrough curves and temperature records observed in the river bed that are very similar to those seen in the main channel provided evidence of rapid hyporheic exchange in some locations. Despite spatial heterogeneity in hyporheic exchange, reach average approximations of transient storage processes within the heat and solute transport model captured the majority of main channel solute and temperature records. These approximations held over multiple spatial scales and over multiple years under similar, low-flow conditions.

When calibrating to main channel temperature and solute records simultaneously, we found tradeoffs between objective functions that may indicate differences in windows of detection for solute and temperature. This is seen in parameter selection for calibration end-members that indicate the flow paths influencing main channel temperatures are likely longer than those affecting main channel solute breakthrough curves. Despite these differences, a compromise between these calibration end-members captured the majority of the main channel temperature and solute breakthrough curves.

Model results showed that hyporheic exchange is a dominant heat flux under low-flow conditions that acts to reduce the amplitude of diel temperature fluctuations by temporarily storing heat in the hyporheic zone, and reduces mean river temperatures by delivering heat to the sediments that is subsequently conducted from the hyporheic zone to the colder, underlying ground. These findings suggest that hyporheic exchange plays a significant role in reducing mean and maximum daily temperature in Arctic rivers under low-flow conditions, when temperatures can be limiting to aquatic species. We further
estimate that the magnitude of heat loss through the hyporheic zone will decrease in the presence of warmer sub-river ground temperatures, and suggest that understanding how these temperatures will respond to projected increase in riverbed thaw are necessary for determine the impacts of climate change on Arctic river temperatures.
CHAPTER 4

ESTIMATING DISCHARGE IN LOW-ORDER RIVERS WITH HIGH-RESOLUTION AERIAL IMAGERY

Abstract

Remote sensing of river discharge promises to augment in situ gauging stations, but the majority of research in this field focuses on large rivers (>50 m wide). We present a method for estimating volumetric river discharge in low-order (<50 m wide) rivers from remotely-sensed data by coupling high-resolution imagery with 1-dimensional hydraulic modeling at so-called virtual gauging stations. These locations were identified as locations where the river contracted under low flows, exposing a substantial portion of the river bed. Topography of the exposed river bed was photogrammetrically extracted from high-resolution aerial imagery while the geometry of the remaining inundated portion of the channel was approximated based on adjacent bank topography and maximum depth assumptions. Merged full channel bathymetry was used to create hydraulic models that encompassed virtual gauging stations. Discharge for each aerial survey was estimated with the hydraulic model by matching modeled and remotely-sensed wetted widths. Based on these results, synthetic width-discharge rating curves were produced for each virtual gauging station. In situ observations were used to determine the accuracy of wetted widths extracted from imagery (mean error 0.36 m), extracted bathymetry (mean vertical RMSE 0.23 m), and discharge (mean percent error 7% with a standard deviation of 6%). Sensitivity analyses were conducted to determine the influence of inundated channel bathymetry and roughness parameters on estimated discharge.

Coauthored by Tyler V. King, Bethany T. Neilson, and Mitchell T. Rasmussen
discharge. Comparison of synthetic rating curves produced through sensitivity analyses show that reasonable ranges of parameter values result in mean percent errors in predicted discharges of 12% to 27%.

Introduction

River discharge records are central to fundamental understanding of hydrologic processes. While in situ gauging stations are considered the most accurate approach for estimating river discharge, they are expensive to maintain and in many cases, the sites and resulting data can be technically, logistically, and politically difficult to access [Fekete and Vörösmarty, 2007]. The limitations of and global decline in in situ gauging stations has given rise to estimation of river discharge from remotely-sensed hydraulic variables as a means to augment and extend in situ gauging-station networks [Alsdorf and Lettenmaier, 2003; Dingman and Bjerklie, 2006; Smith et al., 1996].

A wide range of techniques has been developed to estimate river discharge from remotely-sensed hydraulic variables, including substituting remotely-sensed stage or width observations for in situ observations at established gauging stations [e.g., Bjerklie et al., 2003; Kouraev et al., 2004; Smith, 1997], quantitative imagery analysis [e.g., Johnson and Cowen, 2016; Legleiter et al., 2017; Stumpf et al., 2016], and coupling remotely-sensed observations of wetted width, water-surface elevation, and/or free surface slope with well-established open-channel flow equations [e.g., Durand et al., 2014; Garambois and Monnier, 2015; Liu et al., 2015; Wilson et al., 2015]. This last
approach has received the most attention and is seen as the most applicable to remote, ungauged basins given the potential to use only remotely-sensed data products to estimate discharge.

One of the largest impediments to remote sensing of discharge is the need for estimates of channel bathymetry. A wide range of techniques have been proposed to develop empirical correlations between spectral properties and observed water depths [Legleiter et al., 2009; Legleiter et al., 2004; Lyzenga, 1981; Su et al., 2008; Westaway et al., 2003]. These approaches, however, require known depths for calibration and relatively low turbidity in order for light to penetrate the water column and reflect off the river bed. Green LiDAR and radar can be used to map inundated bathymetry in shallow (< 2m deep) waters, but require highly specialized and expensive payloads and/or platforms that have prevented widespread adoption [Bailly et al., 2010; Hilldale and Raff, 2008; Melcher et al., 2002]. A number of approaches have been presented where channel bathymetry and slope are estimated from the assimilation of river width, slope, and/or water surface elevation with hydrologic modelling [Biancamaria et al., 2011; Durand et al., 2010; Durand et al., 2008; Durand et al., 2014; Mersel et al., 2013; Yoon et al., 2012; Yoon et al., 2016]; however, these methods often use hydrologic models to provide initial estimates of discharge from ancillary hydrometeorological observations, which are not always available in remote locations. As a result, channel bathymetry is often treated as: 1) available from an outside source, which may be true for large, socially and economically important water ways [e.g., Liu et al., 2015] but is unlikely for smaller, remote rivers, 2) a set of calibration parameters that are estimated in conjunction with
discharge using data assimilation techniques and assumptions about mass conservation over large (10 km) spatial scales [e.g., Durand et al., 2014; Durand et al., 2016; Garambois and Monnier, 2015], or 3) terms to be integrated into a simplified longitudinal trend in hydraulic geometry within the at-many-stations hydraulic geometry framework of Gleason and Smith [2014]. For each approach there are tradeoffs between accuracy, a priori data requirements, and the scale at which the method can be applied.

The vast majority of reported techniques for remote sensing of river discharge have utilized satellite observations with the goal of obtaining global coverage. While these satellite-based approaches have the potential to greatly expand global coverage of river discharge estimates [e.g., Pavelsky et al., 2014], their applicability is limited to larger rivers where widths are greater than 100 m. Lower-order river reaches that fall below the width threshold imposed by the resolution of satellite observations are not only ubiquitous [Allen and Pavelsky, 2015; Downing et al., 2012], but are also biogeochemically important [e.g., Ågren et al., 2007], provide critical aquatic habitat [e.g., Rosenfeld et al., 2002], and have the lowest coverage of in situ gauging stations [Pavelsky et al., 2014]. As such, alternative methods to in situ gauging stations and satellite-based remote sensing approaches are necessary to quantify river discharge and answer basic hydrologic questions within these smaller basins.

In contrast to the coarse resolution of satellite observations used for the extraction of hydraulic properties at global scales, high-resolution imagery has been used for decades to produce digital terrain models of non-inundated river channels using photogrammetric techniques [Collin and Chisholm, 1991; Lane et al., 1994]. Recent
advances in hardware (cameras, platforms, global positioning systems, inertial momentum units) and post processing software have greatly advanced the use of aerial photography for high-resolution three-dimensional reconstruction of non-inundated geomorphic units [e.g., Flener et al., 2013; Javernick et al., 2014; Watanabe and Kawahara, 2016]. These studies demonstrate the accuracy of extracting digital surface models (DSMs) of river banks from aerial imagery, but stop short of integration with hydraulic modelling to estimate river discharge.

To address the need for and lack of gauging stations in lower-order rivers, we present a technique to estimate river discharge in lower-order rivers that uniquely combines the techniques of extracting channel morphology from high-resolution aerial imagery and estimating river discharge using well established open-channel flow hydraulics. Our technique builds upon previous approaches of remote sensing of river discharge while taking advantage of the high-resolution information that is available with aerial imagery to estimate river discharge in portions of the watershed that are unobservable with satellite-based observations. With a case study from the Kuparuk River in Arctic Alaska, we take advantage of the properties of Arctic hydrology and geomorphology to demonstrate the feasibility of this method.

Methods

With the objective of estimating river discharge remotely, the following general approach was adopted. Aerial surveys were conducted from which wetted widths and channel bathymetry were extracted. Open-channel hydraulic routing models were
created for locations identified as virtual gauging stations (VGS) using the extracted bathymetry. The hydraulic models were used to estimate discharge required to reproduce the wetted widths observed in the aerial imagery. These modeled discharges and extracted widths were used to develop synthetic width-discharge rating curves. In situ observations were used to determine the accuracy of this approach. These individual steps are detailed in the following sections.

**Study Site**

This method was developed and tested for a 35 km long study reach near the headwaters of the Kuparuk River basin, Alaska (Figure 4-1). The hydrologic regime is typical of basins over continuous permafrost with flows that are dominated by snowmelt in the spring and rapid responses to precipitation events throughout the summer (Figure 4-2) [McNamara et al., 1998]. Top-down thawing of soils in the basin through the summer months only thaws the top tens of centimeters, producing a very thin hydrologically active subsurface layer, or “active layer” which is bounded underneath by effectively impermeable, frozen soil. Limited storage in the active layer produces high runoff/rainfall ratios, rapid and significant rises in streamflow in response to precipitation events throughout the summer months, and recession to very low flows between precipitation events. These discharge regimes, coupled with reaches in the cobble lined alluvial channel with low angle, transverse bed slopes, allow for extensive bed exposure under low flows and substantial expansion and contraction in wetted width in response to changes in discharge. A lack of overhanging vegetation allows these changes in width to be observable from above. More details on the Kuparuk River basin are provided in Chapter 2.
Figure 4-1. A 22 km portion of a 35 km study reach (A) where hydraulic models were developed for regions (red boxes) that encompassed the three virtual gauging stations (black boxes in B and C). Wetted widths were extracted from NIR imagery for each aerial survey and were seen to vary significantly between low (B) and high flows (C). Exposed channel bathymetry was photogrammetrically extracted under low flows (grey lines in D and E) and bathymetry of the inundated portion of the channel (black lines in D and E) was approximated with trapezoidal channel bathymetry using bank slopes extracted from the adjacent exposed channel topography. Note that the exposed channel under low flows (down to the red line in D) was inundated during subsequent aerial surveys (blue lines in D). Channel bathymetry was determined for all transects in the hydraulic model domains (E) and discharge required to produce extracted widths within the virtual gauging station for each aerial survey was determined using a one-dimensional hydraulic routing model.
Figure 4-2: Observed discharge and dates of aerial surveys (vertical colored lines) for 2014 (A) and 2015 (B). Note different y-axis scales between years. Aerial surveys were distributed over a wide range of flows and antecedent conditions, with some flights occurring under similar flows at some locations. Imagery from 9 July 2015 was used to extract channel bathymetry.

Data Collection

Virtual gauging stations (VGSs) are locations within river reaches where volumetric river discharge is related to a remotely-sensed attribute (e.g., wetted width) of the inundated river channel. With a goal to establish VGSs in river reaches by combining remotely-sensed bathymetry, wetted widths, and hydraulic modeling, we conducted repeat field campaigns and aerial surveys along a 35 km long reach of the Kuparuk River in Arctic Alaska (Figure 4-1A). Data collection included aerial surveys over a range of discharges to provide high-resolution visible (RGB) and near infrared (NIR) imagery, and in situ observations to evaluate method accuracy.

Aerial imagery was collected from a custom imagery acquisition payload constructed by AggieAir™ at Utah State University. The battery powered payload was
mounted to the outside of a Robinson R-44 helicopter (Torrance, CA) and controlled wirelessly by an operator within the aircraft. Position, trajectory, and imagery data were collected simultaneously with a fully integrated VectorNav GPS (Dallas, TX), inertial measurement unit, and image acquisition system. RGB imagery was collected with a Canon S-95, which has a 10 megapixel CCD sensor with 8-bit radiometric resolution and ISO range of 80-3200. NIR imagery was collected with a camera identical to that used for RGB imagery, with the notable exception that the manufacturer’s optical filter was replaced with a Wratten 87 NIR filter, which selects for 750 nm radiation.

Twelve pairs of aerial targets were placed along the river corridor to provide validation of the orthomosaic production. The targets, produced with a quarter-square triangle pattern, were constructed from black and white acrylic sheeting to provide strong brightness contrast. Targets were at least 50 cm by 50 cm, or approximately 3.5 pixels square. For verification purposes, target locations were determined with Trimble R7 (Sunnyvale, CA) survey grade GPS units operated in Fast Static mode and post processed with a network adjustment from two continuously operating reference station sites. It should be noted that while aerial targets are useful for verifying accurate orthorectification, they were not used in the production of the orthorectified mosaics and are therefore not required for method application.

In situ gauging stations located 13, 20, and 35 km from terrestrial vehicle access were developed. These stations were installed for the summer months of June, July, and August, as the remote nature of these sites precluded construction of year-round infrastructure capable of withstanding the winter ice cover and spring breakup. At each
location steel posts were driven at least 50 cm into the river bed to which Campbell Scientific CS-450 (Logan UT) pressure transducers were mounted vertically within perforated PVC cages to diffuse any velocity head. The pressure transducers were mounted near the river bed to allow for measurement of stage under low flow conditions. Periodic discharge measurements were made throughout the periods of pressure transducer deployment using a SonTek FlowtrackerTM (San Diego, CA) handheld Acoustic Doppler Velocimeter for wadable conditions and a Teledyne RD Instruments® StreamProTM (Poway, CA) Acoustic Doppler Current Profiler for higher flows. A Trimble M3 (Sunnyvale, CA) total station was used to survey water surface elevation and local benchmarks to relate stage readings between deployments. Rating curves were derived using power law relationships from discharge and corresponding stage observations. Near continuous (15 min) hydrographs were produced from the observations of stage throughout the period of pressure transducer deployment (Figure 4-2).

Transects between aerial targets were surveyed with a total station to produce ground truthing observations of channel shape. Surveys included targets, vegetated riparian zone, dry river bank, and inundated bathymetry with observations made at locations with significant breaks in slope. Ground surface elevations were recorded which in some cases are up 1.5 m lower than the crown of the vegetation canopy. In these cases, elevations extracted from the photogrammetric DSM are higher than elevations from the total station surveys. This only becomes important for overbank flooding conditions, which this method is not intended to address.
Wetted widths were measured manually between pairs of aerial targets by either pulling a fiberglass tape measure across the river from bank to bank, or by using a Laser Technology TruPulse® 360r (Centennial, CO) laser range finder with a flat, broad plastic target held at the edge of water. Measurements were taken in triplicate and mean values used to evaluate the wetted widths extracted from aerial imagery.

**Image Processing**

Agisoft PhotoScan was used to produce orthomosaics from NIR imagery for all flights and photogrammetric DSMs from RGB imagery from the flight that corresponded to the lowest flows. Payload positioning information was ingested from on board GPS and IMU units, and point clouds were generated using feature-matching and bundle adjustment. Orthomosaics were produced by spatially averaging of the brightness values within the point cloud while pair-wise depth map computation algorithms were applied to the point cloud to produce photogrammetric DSMs [Agisoft, 2017].

Wetted widths extracted from the NIR orthomosaics were used to provide the spatial and flow-dependent metric that was then used to select VGS locations and evaluate volumetric flow rates using the hydraulic model. Note that while water surface elevation could have been used as the observed property in our VGSs, photogrammetric extraction of turbulent water surface elevations are highly inaccurate [Han and Endreny, 2014]. Wetted widths were determined by 1) producing orthomosaics from NIR aerial imagery, 2) extracting water masks from the orthomosaics with a binary NIR digital number threshold set manually to coincide with a local minima within the brightness histogram interpreted as the brightness of the narrow margin of wet soil along the water’s edges to compensate for incident light conditions, 3) manually extracting a river
centerline and producing transects perpendicular to the river centerline at 1 m intervals, and 4) determining the length of each transect intersecting the water mask. Wetted widths extracted from NIR orthomosaics were evaluated against ground-based observations of wetted widths as described below.

Channel bathymetry for the exposed portion of the channel was extracted from the photogrammetrically derived DSM created from RGB imagery collected under the lowest observed flows. Bathymetry of the shallow inundated portion of the river channel was approximated with trapezoidal cross-sections using side slopes extracted from the DSM. Bank slopes perpendicular to and within a few meters of the water’s edges were extended into the inundated regions until they either intersected the opposite bank or reached a maximum depth below the lowest observed water surface ($d$, Figure 4-1D). An initial guess for $d$ was set at 20 cm for all VGSs based on knowledge of the system and qualitative interpretation of the RGB aerial imagery. The sensitivity of discharge estimates to $d$ was determined as explained below. Limiting the bank slope used in bathymetry extrapolation to within a few meters of the water’s edge allowed us to avoid the issues of breaks in bank slope presented in Mersel et al. [2013]. The approximated inundated channel bathymetry was merged with the DSM to provide full channel geometry across the floodplain and river channel (Figure 4-1E). Vertical accuracy of the photogrammetrically derived DSM was evaluated against total station surveys as described below. Some alternative channel geometry processing methods include assuming rectangular cross sections below the minimum observed water surface or continuing bank slopes until they intersect without a maximum depth. The former does
not provide unique width estimates under low flows which would provide no lower estimate limit to flows, while the latter produces unstable hydraulic results where localized bank slopes vary significantly between subsequent cross-sections.

**Hydraulic Modeling**

As described above, VGS locations were selected where a wide range of wetted widths were observed. A subset of three potential locations near aerial targets and in situ gauging stations were selected to test and validate our method for estimating discharge. Hydraulic models of river reaches that encompassed the selected VGSs were produced in HEC-RAS 5.03 [Brunner, 2016] and run with steady state, subcritical flow routines. The downstream water surface elevations were extracted from the digital surface model for each flight as the elevation at the water mask’s edge, similar to the method used in Durand et al. [2014]. Hydraulic model domains had transects spaced every meter and the VGSs were located away from the downstream boundary of the model domains to minimize influences from the downstream boundary condition.

Volumetric flow rates within each hydraulic model domain were estimated for each aerial survey by varying flow in the 1-D hydraulic model until the wetted widths extracted from the NIR orthomosaics were reproduced by the model at the VGSs. Hydraulic simulations were run to ensure that the actual discharge were within the tested flows. For each tested flow, root mean square errors were calculated to compare simulated and extracted wetted widths for transects within the VGSs (wetted width RMSE). The simulated discharge that produced the minimum wetted width RMSE for a given flight was selected as the optimum simulated discharge. To determine the accuracy of this approach, the optimum simulated discharges were evaluated against in situ
gauging station discharges. The precision of this method is determined by comparing results from repeat aerial surveys under similar flow conditions.

In order to take advantage of well-established relationships between width and discharge, known as hydraulic geometries [Leopold and Maddock, 1953], power law width-discharge rating curves were produced for each VGS using widths extracted from imagery and flows that were either estimated from the hydraulic model (synthetic rating curves) or observed at in situ gauging stations (observed rating curves). The production of rating curves allows for discharge to be estimated at each location given only observations of wetted width, and for the identification of ranges in width or discharge for which the simulated and observed regressions are significantly different. To determine if the simulated and observed rating curves were significantly different, 95% confidence intervals were produced for each rating curve. Regions of the curves with overlapping confidence intervals were considered to not be significantly different.

**Accuracy Assessment**

Ground-based observations provide the data necessary to evaluate the accuracy of wetted widths extracted from orthomosaics, extracted channel bathymetry, and estimated river discharge. Accuracy of extracted wetted widths were evaluated as absolute error in meters:

\[
EB_{i,j} = \sqrt{(B_{obs,i,j} - B_{ext,i,j})^2}
\]  

(1)

where \(EB_{i,j}\) is the wetted width absolute error for VGS \(i\) for flight \(j\); \(B_{obs,i,j}\) is the wetted width for VGS \(i\) for flight \(j\) measured on the ground; \(B_{ext,i,j}\) is the wetted width for VGS \(i\)
for flight \( j \) extracted from orthomosaics. Percent width error is calculated from absolute error as:

\[
PEB_{i,j} = \frac{\sqrt{(B_{\text{obs},i,j} - B_{\text{ext},i,j})^2}}{B_{\text{obs},i,j}} \times 100
\]  

(2)

where \( PEB_{i,j} \) is the percent wetted width error for VGS \( i \) for flight \( j \). The accuracy of simulated wetted widths was evaluated against the extracted wetted widths with a root mean square error (RMSE) objective function as:

\[
RMSE_{i,j,k} = \sqrt{\frac{\sum_{m=1}^{n_i} (B_{\text{ext},i,j,m} - B_{\text{mod},i,j,k,m})^2}{n_i}}
\]  

(3)

where \( RMSE_{i,j,k} \) is the wetted width root mean square error for VGS \( i \) for flight \( j \) for tested flow \( k \); \( B_{\text{ext},i,j,m} \) is the wetted width for VGS \( i \) for flight \( j \) for transect \( m \) extracted from orthomosaics; \( B_{\text{mod},i,j,k,m} \) is the wetted width for VGS \( i \) for flight \( j \) for tested flow \( k \) for transect \( m \) produced from the 1D hydraulic model; \( n_i \) is the number of transects within VGS \( i \).

The photogrammetrically derived DSM (e.g., the portion of the river bed that was exposed under low flows) was compared against total station surveys of transects located between targets. The vertical accuracy was established as the RMSE of elevation between the surveyed and extracted elevations as:

\[
RMSE_o = \sqrt{\frac{\sum_{p=1}^{q_o} (Z_{\text{obs},o,p} - Z_{\text{ext},o,p})^2}{q_o}}
\]  

(4)

where \( RMSE_o \) is the vertical root mean square error for transect \( o \); \( Z_{\text{obs},o,p} \) is the elevation of position \( p \) on transect \( o \) from total station survey; \( Z_{\text{ext},o,p} \) is the elevation of position \( p \)
Estimated river discharge from the hydraulic modeling is evaluated against observed river discharge extracted from hydrographs for the given time of flights using a percent error:

\[
PEQ_{i,j} = \frac{Q_{\text{obs},i,j} - Q_{\text{mod},i,j}}{Q_{\text{obs},i,j}} \times 100
\]

where \( PEQ_{i,j} \) is the discharge percent error for VGS \( i \) for flight \( j \); \( Q_{\text{obs},i,j} \) is the discharge observed at the in situ gauging station \( i \) for flight \( j \); and \( Q_{\text{mod},i,j} \) is the discharge estimated from the hydraulic model to best reproduce the extracted widths for VGS \( i \) for flight \( j \).

Note that in situ (GS) and virtual (VGS) gauging stations with the same index \( i \) are located in close proximity to each other (Figure 4-1A).

**Sensitivity Analysis**

Sensitivity to model parameters of \( d \) used in bathymetry estimation and \( n \) used in hydraulic modeling were evaluated as percent differences between extracted rating curves and simulated rating curves produced from using plausible ranges of parameter values. Percent errors were calculated by: 1) prescribing a reasonable range of \( n \) values from published tables [Arcement and Schneider, 1989] and a reasonable range of possible reach averaged \( d \) determined from visual inspection of the high-resolution aerial imagery, 2) routing ranges of discharge through the 1-D hydraulic models for all combinations of reasonable depth and roughness values, 3) determining simulated discharge values that best reproduced the widths extracted for each simulation, 4) producing power law
discharge-width rating curves from the simulation results, and 5) calculating the percent errors between simulated and extracted rating curves for the region where there is significant difference in the curves for each simulation run (Eq. 6):

$$PEQ_{n,d,i} = \left( \frac{\sum_{w=w_{o,i}}^{r_i} \frac{Q_{sim,w,n,d,i} - Q_{obs,w,i}}{Q_{obs,w,i}}}{N_{widths}} \right) \times 100 \quad (6)$$

where $PEQ_{n,d,i}$ is the mean percent difference magnitude between observed and simulated discharges for VGS $i$ for the range in width where simulated and extracted rating curves were significantly different for simulations using roughness value of $n$, and depth of $d$; $w$ is the wetted width, $w_{o,i}$ is the statistical significance threshold width for VGS $i$ described above; $r_i$ is the maximum observed width for VGS $i$; $Q_{sim,w,n,d,i}$ is the simulated discharge for width $w$ at VGS $i$ using roughness $n$ and depth $d$ above the significance threshold; $Q_{obs,w,i}$ is the observed discharge for width $w$ at VGS $i$; and $N_{widths}$ is the number of simulated widths above the threshold $w_i$ and below the maximum observed width $r_i$.

Ranges for $d$ and $n$ values were determined to span the range of reasonable values. $d$ was initially set to 20 cm with a range of 10 - 50 cm based on knowledge of the system and visual inspection of the RGB imagery. Channel roughness ($n$) was initially set to 0.04 with a range of 0.03 to 0.05 based on the cobble substrate that is evident in the high-resolution aerial imagery and published literature on channel roughness values [Arcement and Schneider, 1989; Barnes, 1967]. Irregularities, variations in channel cross section, obstructions, vegetation, and meandering were negligible for these sites, negating the need to modify the base roughness value. Our selection of 0.04 is commensurate with
the work of Kane et al. [2003] for the same river.

From these percent error values, sensitivity of discharge estimates were calculated for each VGS in two ways: 1) the mean and standard deviation of percent error magnitudes from varying both depth and roughness (Eq. 7) and 2) the mean and standard deviation of the range of percent differences from varying one parameter at a time (Eq. 8 and 9). The former provides an estimate of the magnitude of the error associated with the range of both parameters, while the latter provides a measure of the range of the error associated with each parameter:

\[ E_i = \frac{\left( \sum_{n=0.03}^{0.05} (\sum_{d=10}^{50} (PEQ_{n,d,i}))) \right)}{N_{\text{depths}} \times N_{\text{roughnesses}}} \]  

(7) \[ E_{\text{depth},i} = \frac{\left( \sum_{n=0.03}^{0.05} (PEQ_{n,d=50,i} - PEQ_{n,d=10,i}) \right)}{N_{\text{roughnesses}}} \]  

(8) \[ E_{\text{rough},i} = \frac{\left( \sum_{d=10}^{50} (PEQ_{n=0.03,d,i} - PEQ_{n=0.05,d,i}) \right)}{N_{\text{depths}}} \]  

(9)

where \( E_i \) is the mean percent difference magnitude for VGS \( i \); \( N_{\text{depths}} \) and \( N_{\text{roughnesses}} \) are the number of depth and roughness values used in the sensitivity analysis respectively; \( E_{\text{depth},i} \) and \( E_{\text{rough},i} \) are the sensitivity of a VGS \( i \) to variations in \( d \) and \( n \), respectively.

Results

Ten aerial surveys conducted in 2014 and 2015 had average durations of 40 minutes, were conducted at approximately 300 m above ground at an average ground speed of 17 m s\(^{-1}\) and produced 0.14 m resolution imagery with imagery overlap of 30 to
40%. Wetted widths extracted from the resulting NIR orthomosaics had a mean accuracy of 0.36 m with a standard deviation of 0.28 m (or 0.36 m ±0.28 m) when compared against 11 field observations (Table 4-1). This level of accuracy corresponds with approximately two and a half pixels in the orthomosaics, and 2% ±1% of the observed wetted widths. Discharge ranged over an order of magnitude at each of the in situ gauging stations across all aerial surveys (Table 4-2, Figure 4-2), producing wetted widths ranging from 8 – 42 m.

<table>
<thead>
<tr>
<th>Date</th>
<th>8/7/2015</th>
<th>7/6/2015</th>
<th>7/22/2015</th>
<th>8/12/2015</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>VGS1</td>
<td>N/A</td>
<td>0.02</td>
<td>0.3</td>
<td>0.1</td>
<td>0.14</td>
</tr>
<tr>
<td>VGS2</td>
<td>0.3</td>
<td>0%</td>
<td>NA</td>
<td>NA</td>
<td>0.40</td>
</tr>
<tr>
<td>VGS3</td>
<td>0.4</td>
<td>1%</td>
<td>NA</td>
<td>0.4</td>
<td>0.53</td>
</tr>
<tr>
<td>Mean</td>
<td>0.35</td>
<td>2%</td>
<td>0.02</td>
<td>0.35</td>
<td>0.36</td>
</tr>
</tbody>
</table>

Table 4-1. Difference between observed and extracted wetted widths (EB) in absolute distance (m) and percentage (PEB) of observed widths (%). NA = observed data not available.

<table>
<thead>
<tr>
<th>Flight #</th>
<th>Date</th>
<th>Resolution m</th>
<th>Observed Discharge m³ s⁻¹</th>
<th>Estimated Discharge m³ s⁻¹</th>
<th>Percent Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2015-07-09</td>
<td>0.18</td>
<td>0.58, 0.70, 0.29, 0.61</td>
<td>0.86, 0.85, 0.52</td>
<td>5%</td>
</tr>
<tr>
<td>2</td>
<td>2015-07-09</td>
<td>0.12</td>
<td>0.72, 0.90, 0.51, 0.81</td>
<td>NA, NA</td>
<td>1%</td>
</tr>
<tr>
<td>3</td>
<td>2015-07-09</td>
<td>0.15</td>
<td>0.84, 0.99, 0.51, 0.81</td>
<td>1.15, 1.15</td>
<td>5%</td>
</tr>
<tr>
<td>4</td>
<td>2015-07-09</td>
<td>0.14</td>
<td>1.40, 1.80, 1.65, 1.37</td>
<td>1.67, 1.44</td>
<td>-2%</td>
</tr>
<tr>
<td>5</td>
<td>2015-07-09</td>
<td>0.10</td>
<td>1.10, 1.27, 1.73, 0.97</td>
<td>NA</td>
<td>-12%</td>
</tr>
<tr>
<td>6</td>
<td>2015-07-09</td>
<td>0.15</td>
<td>1.15, 1.58, 2.78, 1.16</td>
<td>1.69, 2.78</td>
<td>3%</td>
</tr>
<tr>
<td>7</td>
<td>2014-07-11</td>
<td>0.17</td>
<td>1.91, 2.59, 5.69, 2.12</td>
<td>2.69, NA</td>
<td>1%</td>
</tr>
<tr>
<td>8</td>
<td>2014-07-11</td>
<td>0.14</td>
<td>6.91, 8.84, 18.55, 5.74</td>
<td>8.49, 17.81</td>
<td>&lt;1%</td>
</tr>
<tr>
<td>9</td>
<td>2014-07-11</td>
<td>0.13</td>
<td>13.95, 21.39, 38.22, 13.11</td>
<td>22.67, 38.22</td>
<td>-6%</td>
</tr>
<tr>
<td>10</td>
<td>2014-07-11</td>
<td>0.13</td>
<td>13.95, 21.39, 38.22, 13.11</td>
<td>22.67, 38.22</td>
<td>-6%</td>
</tr>
</tbody>
</table>

Table 4-2. Aerial Survey Flight Specifics

aNA = imagery not available
Hydraulic models ranging from 53 to 120 m long encompassed the three selected VGS locations and were populated using DSMs produced from RGB imagery collected on 9 July 2015, the flight that corresponds with the narrowest extracted top widths. In these images, the inundated portion of the channel within the VGSs occupied as little as 20% of the wetted widths observed under high flows (Figures 4-1B and 4-1C). The photogrammetrically derived DSM captured the shape of the non-inundated portion of the channel well and was vertically accurate with an average vertical RMSE of 0.23 m ±0.05 m across 37 observations at three transects (Figure 4-3).

![Figure 4-3. Transect elevation profiles in meters above sea-level (mas) from VGS1 (A), VGS2 (B), and VGS3 (C) extracted from the photogrammetric DSM (red squares) and from total station surveys (black circles). Mean vertical RMSE is 0.23 m. Note that displayed DSM values are limited to stations where survey observations are available.](image-url)
Using our assumed initial parameter values of 20 cm for $d$ and 0.04 for $n$, remotely-sensed estimates of discharge fall within the 95% confidence intervals of the in situ rating curve, showing no statistically significant differences between the simulated and observed discharges (Figure 4-4). The mean percent error across all sites and all observed flows was 7% ±6% (Table 4-2). Repeat flights conducted under similar flow conditions (e.g., 7 Aug 2015 and 6 July 2015 for VGS1, and 7 Aug 2015 and 22 July 2015 for VGS3) produced discharge estimates within 5% of each other, indicating that between-flight variation is minimal and that there is a high level of method precision (Table 4-2).

Statistical significance thresholds from the sensitivity analysis were 15, 16, and 18 meters in width for VGS1, VGS2, and VGS3 respectively (Figure 4-5). Below these widths the extracted and simulated rating curves were not significantly different from

![Figure 4-4. Observed (black lines) and synthetic (red lines) rating curves for VGS1 (A), VSG2 (B), and VGS3 (C). All simulated discharges (red dots) fall within the 95% confidence intervals (grey regions) of the observed rating curves showing that there are no statistically significant differences between the observed and simulated discharges.](image-url)
Figure 4-5. Percent error between simulated and observed rating curves for VGS1 (top row), VGS2 (middle row), and VGS3 (bottom row) grouped by simulated depth (columns). Shaded region shows the range of percent errors for channel roughness values of 0.03 (top of shaded regions), 0.04 (black line), and 0.05 (bottom of shaded regions). Percent error statistics ($E_i$ on right of each set of figures, $E_{\text{depth}}$ and $E_{\text{rough}}$ in Table 4-3) are calculated for the statistically significant regions (blue shading right of the vertical lines). The percent errors for roughness values of 0.02 (top dashed grey line) and 0.06 (bottom dashed grey line) are shown for illustrative purposes, but are not included in the statistical analysis as these values are outside the reasonable range of representative channel characteristics.

Each other across all values of $d$ and $n$. Above these thresholds the rating curves become significantly different for at least some combinations of assumed parameter values. The mean percent error magnitudes increases to from 8% to 27% $\pm 16\%$ at VGS1, 9% to 25% $\pm 16\%$ at VGS2, and 4% to 12% $\pm 7\%$ at VGS3 (Figure 4-5). These represent scenarios where only general values for depth and roughness are available. If either depth or roughness values are well constrained, mean percent error magnitudes decrease by one-third. This analysis also illustrated that there is a transition in dominant sensitivity with
distance downstream (Table 4-3, Figure 4-6). Sensitivity to depth ($E_{depth}$) decreases and sensitivity to roughness ($E_{rough}$) remains constant with distance downstream. VGS1 is two times more sensitive to depth (72% ± 4%) than roughness (33% ± 6%), VGS2 is equally sensitive to depth (53% ± 8%) and roughness (42% ± 8%), and VGS3 is three times more sensitive to roughness (31% ±2%) than depth (9% ±3%) (Table 4-3).

Discussion

These results demonstrate that our approach of coupling high-resolution aerial imagery with open-channel hydraulic modeling produces remotely-sensed estimates of river discharge with accuracy levels that are on par with in situ gauging stations and other remote sensing approaches. Evaluating method performance across different discharges and between sites allows us to suggest conditions and VGS characteristics where this method works best. Percent error in model performance is highest for the bottom 20% of observed flows (Table 4-2). This is expected as under these low flow conditions the estimated channel bathymetry makes up the majority of the channel cross sectional area. Discharge estimates are also most accurate for VGS3 which had the greatest flow, the straightest planform, the most consistent river width with distance downstream, and widest ranges in wetted widths. These characteristics likely indicate beneficial traits in potential VGS sites and should be included in the selection of VGS locations. Other factors that could influence successful VGS selection include uniform bank slopes leading to and extending into the wetted portion of the channel and uniform bed forms.
Table 4-3. Percent error values (italicized text) for the statistically significant portions of the simulated rating curves for all combinations of depth and channel roughness.

<table>
<thead>
<tr>
<th>VGS1</th>
<th>Depth</th>
<th>Roughness</th>
<th>Range from varying roughness</th>
<th>E_{rough} d ± StDev</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n=0.03</td>
<td>n=0.04</td>
<td>n=0.05</td>
<td></td>
</tr>
<tr>
<td>d=10 cm</td>
<td>-17%</td>
<td>-29%</td>
<td>-42%</td>
<td>25%</td>
</tr>
<tr>
<td>d=20 cm</td>
<td>10%</td>
<td>-6%</td>
<td>-20%</td>
<td>29%</td>
</tr>
<tr>
<td>d=30 cm</td>
<td>36%</td>
<td>13%</td>
<td>-6%</td>
<td>42%</td>
</tr>
<tr>
<td>d=40 cm</td>
<td>49%</td>
<td>29%</td>
<td>14%</td>
<td>35%</td>
</tr>
<tr>
<td>d=50 cm</td>
<td>59%</td>
<td>42%</td>
<td>27%</td>
<td>32%</td>
</tr>
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<td>7%</td>
<td>47%</td>
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<th>E_{rough} d ± StDev</th>
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<td>11%</td>
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a Maximum ranges of percent errors resulting from varying depth for each simulated roughness value.
b Sensitivity to variations in depth defined in Eq. 8.
c Maximum ranges of percent errors resulting from varying channel roughness values for each simulated depth value.
d Sensitivity to variations in roughness defined in Eq. 9.
Figure 4-6: Discharge-width power law rating curves for VGS1 (A), VSG2 (B), and VGS3 (C). Extracted rating curves (red lines) reflect observed discharges vs extracted widths, and optimum rating curves (black lines) reflect simulated discharges vs simulated widths using a maximum depth of 20 cm and roughness of 0.04. The error bounds on the optimum rating curves reflect the range of rating curves that result from varying depth (blue regions) while holding roughness at 0.04, varying roughness while holding depth at 20 cm (red regions), and varying depth and roughness simultaneously (grey regions).

The fundamental approach of estimating river discharge from partial observations of the river channel and estimates of unobservable parameters (e.g., $d$ and $n$) has received a great deal of attention in recent years [e.g., Durand et al., 2014; Durand et al., 2016; Yoon et al., 2016]. Our proposed application of estimating discharge from remotely-sensed river widths using VGS rating curves is similar to the early attempts to remotely sense river discharge as outlined by Smith [1997] and Bjerklie et al. [2003] among others, who used empirical relationships between remotely-sensed hydraulic characteristics and ground-based observations of discharge. The notable difference is that our rating curves are produced from hydraulic models derived from remotely-sensed data products allowing our approach to be applied to ungauged basins where discharge observations are not available. Estimating inundated channel bathymetry from exposed channel geometry was demonstrated by Mersel et al. [2013] and informed our approach of extending river bank slopes to a maximum depth below the minimum observed water surface. However, in our approach we extract the channel geometry from one low flow flight while the
WSE-width pairwise observations method presented in Mersel et al. [2013] requires many repeat observations over the full range of discharges. We further take advantage of the very low flow conditions observed in our lower-order river reach to minimize the area for which bathymetry needs be estimated.

Integrating observations of hydraulic features with hydraulic modeling has also received attention in recent years. For example, Giustarini et al. [2011] and Neal et al. [2009] showed that assimilation of remotely-sensed hydrologic features with hydrodynamic models can improve estimates of river discharge given observations of channel bathymetry and a coupled hydrologic-hydraulic modeling framework. However, these approaches often require ancillary data to drive hydrologic models and/or provide bathymetry. Gleason and Smith [2014] made significant advancements in the field of estimating river discharge using only remote sensed observations by utilizing longitudinal trends in hydraulic geometry laws that scale with the range of observed wetted widths. Central to their approach is the assumption of mass conservation over long (10 km) river reaches allowing for parameter calibration using a genetic algorithm to preserve discharge continuity within the modeled reaches. Similarly, the GaMo and MetroMan algorithms presented in Garambois and Monnier [2015] and Durand et al. [2014], respectively, preserve continuity of discharge within study domains in order to estimate cross-sectional area of the channel at zero flow and channel roughness using optimization algorithms. Durand et al. [2016] compared these and other approaches to estimate river discharge using some or all of the hydraulic parameters that the forthcoming Surface Water and Ocean Topography (SWOT, https://swot.jpl.nasa.gov/) satellite mission is
expected to provide. This highly anticipated satellite mission promises to provide wetted width, water surface elevation, and slope for river reaches over 10 km long and 100 m wide with the possibility of reducing the threshold of observable widths down to 50 m [Fu et al., 2012]. Pavelsky et al. [2014] estimated the potential impact of estimating discharge globally for all rivers down to these two potential observability thresholds and determined that both levels of observation will greatly increase coverage compared with the fraction of basins observed by gauges in the Global Runoff Data Centre (GRDC) network. In their analysis, Pavelsky et al. [2014] illustrated a case where smaller observations thresholds were required to disaggregate upstream signals into the contributing basins. Our work is a direct continuation of this line of reasoning by further reducing the threshold of detection through the use of high-resolution data products.

Using high-resolution observations allows us to estimate discharge for reaches on the order of 100 m in length as opposed to the 10 km length required in the satellite based algorithms. This level of granularity is required to identify hydrologically significant processes in lower-order rivers. For example, discharge records from the in situ gauges in the Kuparuk River show that the GS1-GS2 reach is continuously a gaining reach, but that the GS2-GS3 reach is a losing reach for the five lowest flow flights and a gaining reach for the five highest flow flights (Figures 4-1A, 4-2, and Table 4-2). These longitudinal trends were identified in the remote sensing discharge estimates (Table 4-2), illustrating the utility of distributed estimates of river discharge to detect spatial trends in river discharge which impacts longitudinal chemical and thermal responses. Further, the establishment of additional VGSs within the study domain could be used to better
constrain lateral-flow trends including identifying transition zones between gaining and losing reaches, estimating discharge from surface water sources by differencing VGSs that bracket sources, and estimating gross groundwater interactions by differencing VGSs within a reach with no visible surface water sources.

Arctic tundra landscapes provide a compelling need for the presented method given the lack of in situ gauging stations at high latitudes [Lammers et al., 2001] and the need for better understanding of runoff generation processes contributing to the observed increases in discharge to the Arctic Ocean [Fu et al., 2012; McClelland et al., 2006; Peterson et al., 2002]. However, we suggest this method may be applicable to other systems where three basic requirements are met: 1) low flow conditions where a significant portion of the river bed is exposed, 2) a clear nadir view of the channel from above, and 3) a wide range of widths in response to discharge. A large fraction of Arctic, Antarctic, alpine, ephemeral, and desert rivers meet these criteria due to extended low (or no) flow periods, minimal overhanging vegetation, and large fluctuations in flow. Similar conditions also exist in streams and rivers where significant agricultural withdrawals dewater channels (e.g., Western United States), providing a clear management application for this method. Additional testing across multiple systems is required, and we acknowledge that the proposed method is not applicable in all river systems. However, no single solution exists for the challenge of remotely sensing river discharge and the approach presented here is intended to act as a member in the ensemble of methods for remotely sensing river discharge, similar to that presented by Durand et al. [2016]. Additional testing should also be done to determine if these results can be
achieved with high-resolution satellite observations. Doing so may provide a path towards larger spatial coverage and integration with additional sensors, such as altimetry observations, which could be used to develop multi-parameter rating curves which could improve method accuracy.

While the sensitivity analysis suggests that reasonable ranges for \( n \) and \( d \) produce acceptable ranges of discharge estimates, direct observation of these parameters would provide a significant advancement in the field and could make the method applicable to more systems. We suggest two possible paths forward to estimate these parameter values remotely, and one possible direction that could obviate the need to estimate channel roughness altogether. First, channel bathymetry of the inundated portion may be determined by spectral analysis following the techniques first presented by Lyzenga [1981] and demonstrated extensively since then [e.g., Carbonneau et al., 2006; Flener, 2013; Legleiter, 2015; Legleiter et al., 2009; Legleiter et al., 2004; Lyon et al., 1992; Marcus et al., 2003; Su et al., 2008].

We propose that by using the regions of the riverbed that are exposed under low flow conditions and inundated under high flow conditions to calibrate the empirical relationship between depth and spectral properties, the need for in situ observations of channel depth may be eliminated. Producing full channel geometry by merging dryland topography extracted photogrammetrically with inundated bathymetry approximated with spectral analysis is not a novel idea [e.g., Flener et al., 2013; Javernick et al., 2014; Westaway et al., 2003]. However, the approaches to date have required in situ observations of depth for model calibration while our suggestion is based solely on
remotely-sensed data products.

Second, analysis of the three-dimensional point-cloud that is produced as an intermediate product in the photogrammetric processing may provide estimates of channel roughness as suggested by Butler et al. [1998]. While this has not been done from aerial imagery, Vetter et al. [2011] have demonstrated this concept with a point cloud collected with airborne laser scanning. It is unclear what resolution and accuracy is needed for analysis of the point cloud to accurately represent the channel roughness, and this should be the focus of additional research. Lastly, known difficulty in assigning channel roughness a priori has been addressed through development of conveyance routines that do not include channel roughness [e.g., Dingman and Sharma, 1997; Jarrett, 1984; López et al., 2007; Riggs, 1976]. While these methods do obviate the need to estimate channel roughness per se, they also require estimation of fitting parameters which are often derived from regression analysis of large datasets. Future work could focus on integrating these alternative conveyance methods with the method presented here to reduce the dependency on accurate estimates of channel roughness, especially if fitting parameters can be related to remotely-sensed physical characteristics.

Conclusion

In a time of gauging station decline and increased hydrologic variability, there is need for methods that provide accurate estimates of river discharge via remote sensing with minimal a priori information. Satellite-based discharge algorithms are being
developed to provide estimates of discharge for rivers above approximately 50 to 100 meters wide. However, this technologically derived threshold holds little hydrologic significance and will exclude large and functionally important portions of drainage networks. We propose a method for remotely sensing river discharge for lower-order rivers by coupling high-resolution overlapping aerial imagery with hydraulic modeling. Channel geometry exposed under low flow conditions is extracted from high-resolution aerial imagery, leaving a small portion of the river bathymetry to be estimated from the surrounding topography. Wetted widths extracted from repeat aerial imagery under a range of flows are used to identify locations where width varies in response to discharge. Virtual gauging stations were selected as locations that had a strong variation in wetted width in response to discharge. One-dimensional hydraulic models were created for three locations with which synthetic rating curves were created. Discharge may be estimated from these synthetic rating curves given only an observation of channel wetted width.

With a mean accuracy of 7% ±6%, this method is comparable with in situ measurements of river discharge and other methods of remotely sensing volumetric river discharge. Limitations of this method include the need to estimate channel roughness (n) and maximum channel depth under the lowest observed flow conditions (d). The former can be estimated from the analysis of the substrate visible in the high-resolution aerial imagery, while the latter depends on knowledge of the system and subjective inspection of the imagery. However, sensitivity analyses investigating the influence of these parameters produced mean percent errors in predicted discharges ranging from 12% to 27% over ranges of reasonable assumed d and n values indicating that even a general
estimate of these parameters produces reasonably accurate estimates of river discharge.

We have tested and developed this method across multiple locations within an Arctic tundra watershed, and found that there are three basic requirements for method application: 1) low flow conditions that expose substantial portions of the river bed, 2) a clear nadir view of the entire river including the water’s edge (e.g., lack of overhanging riparian vegetation), and 3) geomorphic features that provide a wide range of widths as a function of discharge. Testing of this method across more and different channel types should be undertaken to determine the range of conditions for which this method remains applicable, however, most Arctic, Antarctic, alpine, desert, and ephemeral streams meet the method requirements identified.

Through the development of accurate spatially average discharge-width rating curves (virtual gauging stations), there is a clear opportunity to couple this approach with high-resolution satellite imagery to provide regular estimates of river discharge, even if these data products are not capable of extracting channel bathymetry with the accuracy of aerial imagery. This method provides an opportunity to extend gauging-station networks to ungauged rivers that are often inaccessible for technical, logistical, and political reasons and to densify existing gauging-station networks to quantify lateral inflows (e.g., groundwater exchanges and surface water contributions/abstractions) between in situ gauging stations. While the proof of concept used aerial imagery, this method could be used with high-resolution satellite imagery to provide global coverage.
CHAPTER 5

CONCLUSIONS

River temperatures exert a primary control on river physical, chemical, and biological processes, making this water quality parameter of social, environmental, and engineering significance. Global climate change is most strongly manifested at high latitudes, and understanding the implications of these changes on Arctic river temperatures necessitates research on fundamental controlling processes. Through mechanistic river temperature modeling informed by detailed field data collection, this dissertation identifies dominant heat fluxes and associated data requirements necessary for predicting temperatures in Arctic rivers. Specifically, this dissertation describes the application of a relatively basic river temperature model developed for temperate rivers to an Arctic river to estimate dominant heat fluxes and to identify regions and flow conditions for which model refinement is required (Chapter 2), reports on refinement of this river temperature model to include the influence of hyporheic exchange on river temperatures under low flow conditions (Chapter 3), and presents the development and application of a method for estimating river discharge via remote sensing by which river temperature modeling may be extended to larger portions of the Arctic (Chapter 4).

In Chapter 2, the dominant heat fluxes for the upper (headwater) and lower (coastal) portions of the Kuparuk River watershed are identified. Surface heat fluxes dominated the energy budget in the lower portion of the watershed, while competition between surface and lateral inflow heat fluxes increased with distance upstream and varied with hydrologic condition. Through scenario testing, it was shown that river
temperatures are generally insensitive to lateral inflow temperatures, while lateral inflow
discharge rates are necessary for accurate river temperature modeling. Model
performance was poor under low flow conditions in the upper portion of the watershed,
providing insight regarding possible model refinement. Specifically, predicted river
temperatures had far greater diurnal ranges under low-flow conditions in the upper
portion of the Kuparuk River than were observed. While the exact cause of these errors
was not clear, model error structure suggested that a buffering mechanism, likely related
to sediment heat exchanges, becomes significant under low-flow conditions and may
need to be added to the basic river temperature model. Other potential sources of error in
the model included using a constant depth to thaw, making simplifying assumptions of
channel geometry, and assuming that subsurface lateral inflows were negligible.

In Chapter 3, data focused on identifying the role of sediment heat exchanges on
river temperatures are presented that justify the inclusion of hyporheic exchange in the
temperature model. Observations of solute breakthrough curves and temperatures in the
subsurface that closely resemble those seen in the main channel during point injection
tracer studies provide multiple lines of evidence of extensive hyporheic exchange. A
multiple objective calibration of a two zone transient storage temperature and solute
model to main channel temperature time series and solute break through curves provided
further evidence of the importance of hyporheic exchange in buffering river temperatures
under low flow conditions. Analysis of heat fluxes from the calibrated model show that
heat fluxes associated with hyporheic exchanges are among the dominant heat fluxes
under low flow conditions and offset warming influences of shortwave radiation during
the day and cooling influences of net longwave radiation at night. Application of the calibrated model to the low-flow periods that showed poor model performance reported in Chapter 2 greatly improved model performance and demonstrated that inclusion of hyporheic exchange is necessary for modeling river temperatures under these conditions.

Given the significance of lateral inflows on river temperature modeling performance during higher flow conditions (Chapter 2) and the lack of in-situ gauging stations in the Arctic, new methods are needed for estimating discharge for river temperature modeling to be feasible at large scales in the Arctic. In Chapter 4, a new method for remote sensing of small, or lower order, rivers to assist in the estimation of river discharge is presented. Using aerial imagery collected under low-flow conditions, digital elevation models of the exposed river channel were created photogrammetrically. With simple approximations of channel shape for the remaining inundated portion of the river channel, hydraulic routing models were constructed and used to estimate width-discharge rating curves. Method accuracy over a broad range of flows was determined to be 7% +/- 6% by comparing model estimated discharge from wetted widths extracted from aerial imagery against discharge observations at in-situ gauging stations. This level of accuracy is comparable with in-situ gauging methods and other methods for remotely sensing river discharge. Sensitivity analysis of assumed model parameters of channel roughness and maximum channel depth demonstrated that reasonable ranges of these parameters produce mean model error ranging from 12 to 27%. From these results, it is clear that even general estimates for the assumed model parameters produce reasonable estimates of discharge.
Combining the findings from these three chapters, a comprehensive understanding of dominant heat fluxes in this Arctic river is apparent. Radiative heat fluxes are a significant portion of the Arctic river energy balances across spatial scales and a wide range of discharges. In lower order river reaches, however, additional heat fluxes of hyporheic exchange and lateral inflows become significant under low-flow and high-flow conditions, respectively. Hyporheic exchange significantly reduces the influences of radiative heat fluxes by providing a strong thermal connection between rivers and the underlying ground while lateral inflows connect the terrestrial and river environments by delivering heat and mass from the watershed to the main channel. Through identification of these dominant heat fluxes, it is clear that only with an understanding of the combined hydrologic and climatological influences will it be possible to evaluate the impact of future climate change on Arctic river temperatures.
CHAPTER 6

BROADER IMPACTS

Engineering Significance

Understanding the dynamics of thermal and hydrologic processes over both time and space fills a gap present in Arctic hydrology and engineering. River temperature is known to be a key attribute for biological, chemical, and physical processes, and this research provides the foundation for understanding how river temperatures may vary with changes in climate [e.g., Arismendi et al., 2014; van Vliet et al., 2013]. In an engineering context, changes in river temperatures have the potential to impact water quality, road construction practices in response to river freezing events, and infrastructure construction considerations in response to thermal erosion of bank sediments.

Further, access to remote field sites has historically limited hydrologic studies to either headwater basins (e.g., Innnavait Creek, Oksrukuyik Creek, Toolik Lake Inlet Stream, Upper Kuparuk) and or to outlets of entire watersheds (e.g., Kuparuk, Ob, Lena, Putuligayuk, Fish Creek watersheds) where hydrologic and thermal signals are integrated over large spatial and temporal scales. As a result, there has been limited understanding of longitudinal trends in thermal and hydrologic processes and information regarding controlling mechanisms. The findings presented in this dissertation help to fill the gap in both hydrologic and thermal regimes by focusing on the entire length of the Kuparuk watershed throughout the open water season to provide a quantitative understanding of spatial and temporal dynamics of the mechanisms that control arctic river temperature. This work provides a framework through which the impact of climate change on river productivity, carbon cycling, and nutrient export can be evaluated, as they pertain to river temperatures. This could be accomplished by estimating the sensitivity of river temperatures to changes in meteorological and hydrologic conditions projected under
climate change scenarios.

This research also explores techniques to estimate river discharge from remote sensing products. River discharge is crucial in numerous aspects of river management including flood warning, resource allocation, contaminant transport modeling, and habitat identification [Chapra, 1997]. Discharge estimates have historically been limited to a small subset of transects where significant investment is made to establish and maintain a calibrated empirical relationship between discharge and an observable hydraulic parameter, like stage. Remote sensing coupled with hydraulic modeling provides an opportunity to increase the scale and spatial resolution at which discharge is estimated while reducing the need for ground based measurements. While this method was developed and tested for an Arctic river, it has potential utility in other locations that meet basic requirements. Namely, significant portions of the river bed must be exposed under low flow conditions to allow for photogrammetric extraction of channel bathymetry. This work has significance by providing methods that could be used to expand our understanding of the global hydrologic cycle which has historically been limited to regions with enough infrastructure to support hydrologic monitoring programs. Ground based programs are vulnerable to infrastructure atrophy, political instability, and natural disasters. Remote sensing provides opportunities to expand our understanding of global hydrology to include regions previously inaccessible. It also provides an approach to bridge lapses in operation of ground based observation networks. Integrating these previously unstudied regions into our understanding of the global hydrologic cycle will afford opportunities for enhanced water resource management.

Social Significance

The significant gap in fundamental understanding of the key mechanisms that control river temperature in the Arctic limits our ability to anticipate impacts of climate change on the
livelihood of arctic. Some social implications of river temperature include local impacts on indigenous communities and impacts at the national scale through commerce and national security.

River temperatures affect habitat suitability for Arctic fish that indigenous communities depend on for sustenance. Alterations in aquatic habitat suitability could drive changes in abundance and distribution of native fish, leading to food stress on local communities. Local communities are also impacted by thermal erosion of river banks. River bank stability could be compromised with additional thawing of frozen soils under and around bridges, docks, pipelines, and other critical infrastructure.

At the national scale, river temperatures directly affect transportation to and from the Kuparuk Oilfields in Prudhoe Bay, AK, a petroleum producing region of importance to national security and American commerce. As the only ground transportation route to and from Kuparuk Oilfields, the Dalton Highway is a critical transportation corridor. Transportation of supplies and personnel along this 400 mile long road can be affected by river temperature through sediment transport, bank stability, and ice jam derived flooding which can bring transportation to a halt for weeks at a time.
CHAPTER 7

RECOMMENDATIONS FOR FUTURE WORK

This dissertation provides an initial assessment of dominant heat fluxes in one Arctic river, filling a gap in understanding of the global distribution of river energy budgets. In Chapter 2, dominant heat fluxes were shown to vary spatially and temporally within a single Arctic watershed. A clear extension of this work is to repeat similar analyses in other Arctic watersheds similar to the Kuparuk River watershed to determine the transferability of the findings presented here. Further, these findings should be tested in Arctic watersheds that are different from the Kuparuk River watershed to identify limitations. For example, many Arctic watersheds include glacial inputs, this type of basin should be included in future studies as they may require the inclusion of additional heat transfer mechanisms. Also in Chapter 2, lateral inflows are identified as a significant source of heat and mass for Arctic rivers. In this work, lateral inflow temperatures were estimated from measurements of surface water lateral inflow sources. This effectively ignores the difference in temperatures between surface and subsurface lateral inflows. Future work should focus on partitioning lateral inflows into surface and subsurface components as the relative contribution from these sources may shift in response to climate change.

Building on Chapter 3, where the thermal influence of hyporheic exchange was examined, next steps require better estimates of depth of thaw below Arctic rivers and incorporation of longer subsurface flow paths in subsurface modeling. The work presented here depended on intensive data collection of ground temperatures. This presents a barrier to wide spread river temperature modeling in Arctic basins. One way to lower this barrier is to incorporate detailed subsurface models of thaw (including both conduction and convective influences) to better represent the thermal regime within and below the hyporheic zone. In this way, a zero degree boundary may be prescribed in the absence of ground temperature
observations, and ground conduction and convective influences can be calculated with a better understanding of the thermal gradients from the main channel to the frozen surface. Further work is also needed to estimate the thermal significance of subsurface exchange with longer flow paths than were identified through tracer study calibration. Solute breakthrough curves can be used to estimate the transient storage zones through which the solute passes in the time of the tracer study; however, water flowing through longer flow paths is not captured in the solute breakthrough curve. Including temperature and solute in model calibration helps to alleviate this bias, but uncertainty in the hyporheic exchange flow paths still exist. Detailed groundwater/surface water models that include heat transport may be used to further assess the level of description needed to account for the influences of hyporheic exchange on instream temperatures.

Based on the method developed in Chapter 4, further testing of hydraulic models coupled with remote sensing should be explored. Remote sensing from aerial imagery complements other remote sensing techniques by focusing on rivers that are small enough to be unobservable in readily available satellite imagery, and the results warrant further investigation. Specifically, the key assumptions in the remote sensing method presented in Chapter 4 are related to setting maximum channel depth and effective channel roughness. Further work is needed to refine this method to provide remote sensing estimates of these parameters. A model of depth as a function of spectral signature could be calibrated in portions of the transects where water depths are known similar to the work of Legleiter [2015]. Areas of known channel depth could be estimated by differencing the water surface elevation from imagery collected under high flows and the digital elevation model produced under low flow conditions. The depth-spectral signature model could potentially be calibrated to this area and extended to the inundated portion of the channel under low flows to produce estimates of full channel bathymetry. Channel roughness may be estimated by analysis of photogrammetrically derived point clouds that can be used to estimate
surface roughness. With remote sensing methods for estimating channel roughness and maximum channel depth, river discharge may be truly remotely sensed without reliance on in-situ observations.
REFERENCES


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APPENDICES
Appendix A
Supporting Information for Chapter 2
Introduction

This appendix provides fundamental equations used to estimate surface heat fluxes (Text A-1), figures of the conceptual model, analyzed time-series, and model results (Figures A-1 through A-11), and tables that describe data collection locations, observed meteorological and hydraulic conditions, and modeling results (Tables A-1 through A-6). The equations in Text A-1 are adapted from Chapra [1997].

Text A-1

Atmospheric Longwave Radiation:

\[ J_{an} = \sigma (T_{air} + 273)^4 \left( A + 0.0027\sqrt{e_{air}} \right) (1 - R_L) \]  

(Eq. A-1)

Where:
- \( J_{an} \) = atmospheric longwave radiation heat flux [W m\(^{-2}\)]
- \( \sigma \) = Stefan-Boltzmann constant \( 5.67 \times 10^{-8} \) [W m\(^{-2}\) K\(^{-4}\)]
- \( T_{air} \) = air temperature [°C]
- \( A \) = a coefficient (0.5 to 0.7)
- \( e_{air} \) = air vapor pressure [Pa]
- \( R_L \) = reflection coefficient (generally 0.03)

Air Vapor Pressure:

\[ e_{air} = \left( \frac{RH}{100} \right) * e_{sat} \]  

(Eq. A-2)

Where:
- \( e_{air} \) = air vapor pressure [Pa]
- \( RH \) = relative humidity [%]
- \( e_{sat} \) = vapor pressure at water surface [Pa]

Saturation Vapor Pressure:

\[ e_{sat} = 612.75e^{\frac{17.27T_{air}}{T_{air} + 237.3}} \]  

(Eq. A-3)

Where:
- \( e_{sat} \) = saturation vapor pressure [Pa]

Water Longwave Radiation:

\[ J_{br} = \varepsilon \sigma (T + 273)^4 \]  

(Eq. A-4)
Where:

\( J_{br} \) = water longwave radiation heat flux [W m\(^{-2}\)]
\( \epsilon \) = emissivity of water (approximately 0.97)
\( T \) = water temperature [\(^\circ\)C]

Conduction and Convection:

\[
J_c = c_1 f(U_w) (T - T_{air})
\] (Eq. A-5)

Where:

\( J_c \) = conduction and convection heat flux [W m\(^{-2}\)]
\( c_1 \) = Bowen’s coefficient \(~0.00353 [\text{Pa} \text{°C}^{-1}]\)
\( f(U_w) \) = wind transfer coefficient [W m\(^{-2}\) Pa\(^{-1}\)]

**Wind transfer coefficient:**

The wind transfer coefficient is calculated using the wind transfer coefficient equation as derived in *Brady et al.* [1969]:

\[
f(U_w) = 0.069 + 0.00345 U_w^2
\] (Eq. A-6)

Where:

\( f(U_w) \) = wind transfer coefficient [W m\(^{-2}\) Pa\(^{-1}\)]
\( U_w \) = wind speed measured seven meters above the water surface [m s\(^{-1}\)]

Evaporation/Condensation:

\[
J_e = f(U_w) (e_s - e_{air})
\] (Eq. A-7)

Where:

\( J_e \) = evaporation/condensation heat flux [W m\(^{-2}\)]
\( e_s \) = vapor pressure at the water surface [Pa]

**Vapor Pressure at Water Surface:**

\[
e_s = 612.75 e^{\frac{17.27 T}{237.3 + T}}
\] (Eq. A-8)

Where:

\( e_s \) = vapor pressure at water surface [Pa]

Fluid friction with the bed from *Theurer et al.* [1985].
\[ J_f = 9805 \frac{Q}{B} S_o \]  

(Eq. A-9)

Where:

\( J_f \) = heat from fluid friction with the bed \([\text{W m}^{-2}]\)

9805 = empirical value \([\text{s J m}^{-4}]\)

\( Q \) = volumetric discharge \([\text{m}^3 \text{s}^{-1}]\)

\( B \) = channel width \([\text{m}]\)

\( S_o \) = bed slope \([\text{m m}^{-1}]\)

Figure A-1. Schematic of model cell showing surface heat fluxes \((J_{sn}, J_{an}, J_{br}, J_e, J_c)\), bed heat fluxes \((J_f, J_{SED})\), lateral inflow discharge \((q_{(d \text{ or } pt)})\) and temperature \((T_{(d \text{ or } pt)})\), distance from bed to simulated sediment temperature \((Y_{SED}, T_{SED})\), depth from simulated sediment temperature to permafrost \(Y_{gr}\), sediment properties \((\rho_{SED}, \alpha_{SED}, C_{p,SED})\), water properties \((T, \rho, \epsilon, C_p)\), river width \((B)\), model cell length \((\partial x)\), and observed meteorological parameters \((RH, U_w, T_{air}, J_{sn})\).
Figure A-2. Width to discharge relationships for main channel gaging stations 2 – 9 (Figure 2-1) along the Kuparuk River.
Figure A-3. Precipitation and volumetric lateral inflows between subsequent gaging stations for the upper (A, C) and lower (B, D) model domains for 2013 (A, B) and 2014 (C, D).
Figure A-4. Temperatures of three instrumented surface water lateral inflows in the upper model domain from 2013. Temperatures in T4 and T11 were very similar, with T11 seeing larger diurnal temperature ranges. Mean temperatures in T7 (an alluvial stream) were on average 5.9 °C colder than T11 (a beaded stream) while their diurnal temperature ranges were similar.

Figure A-5. Average temperature records for beaded (solid black) and non-beaded (solid grey) surface water lateral inflows and precipitation (solid blue) in the upper model domain for 2014. Minimum (dashed black) and maximum (dashed grey) observed surface water lateral inflow temperatures are also presented to show the range in lateral inflow temperatures. Temperatures were seen to vary more than 13 °C between sites.
Figure A-6. Temperature records from the Toolik River in the lower model domain from 2013 (black) and 2014 (grey).
Figure A-7. Precipitation (A), air temperature (B), windspeed (C), relative humidity (D), and solar radiation (E) from 2013 for the upper (grey) and lower (black) model domains.
Figure A-8. Precipitation (A), air temperature (B), windspeed (C), relative humidity (D), and solar radiation (E) from 2014 for the upper (grey) and lower (black) model domains.
Figure A-9. Observed (grey solid) and modeled (black dashed) discharge at the downstream boundaries of the upper (A, C) and lower (B, D) model domains for 2013 (A, B) and 2014 (C, D).
Figure A-10. Time series of sensible ($J_c$), latent ($J_e$), friction ($J_f$), and sediment ($J_{sed}$) heat fluxes calculated for S2 for the upper model domain from 2013 (A-D) and 2014 (E-H). These heat fluxes had average and instantaneous percentage contributions below 15% for the upper model domain.
Figure A-11. Time series of sensible ($J_c$), apparent sensible lateral inflow ($J_{lat}$), friction ($J_f$), and sediment ($J_{sed}$) heat fluxes calculated for S2 for the lower model domain from 2013 (A-D) and 2014 (E-H). These heat fluxes had average and instantaneous percentage contributions below 15% for the lower model domain.
Table A-1. Location of and variables collected at each data collection site. Type: MC = main channel, T = tributary, MET = meteorological site. Parameter: Q = discharge, WT = Water temperature, AT = air temperature, SR = shortwave radiation, P = precipitation, RH = relative humidity, WS = windspeed. ** = only from 2013, * = only from 2014.

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Table A-2. Discharge statistics by site for discharges calculated with rating curves and gaged in the field. * = value is more than 20% beyond the limits of the respective rating curve. Site 3\textsuperscript{2013} was relocated between years 2013 and 2014. Site 3\textsuperscript{2014} was located 6.3 km upstream from site 3\textsuperscript{2014}.

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Table A-3. Meteorological conditions for upper and lower model domains for 2013 and 2014. Unless otherwise noted, upper model domain values were measured at the Upper Kuparuk (UK) meteorological station and lower model domain values were measured at the North White Hills (NWH) meteorological station. a = Imnavait Creek (IC), b = Site 3, c = Anaktuvuk River Burn (ARB). * = mean value reported, ** = cumulative value reported.

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a = Imnavait Creek (IC), b = Site 3, c = Anaktuvuk River Burn (ARB). * = mean value reported, ** = cumulative value reported.
Table A-4. Statistics of observed and modeled river temperatures and model performance (°C) for the upper model domain for the 2013 and 2014 model periods. Daily temperature range statistics were calculated as the difference between minimum and maximum temperatures observed or modeled for each calendar day.

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Table A-5. Model performance as average daily RMSE (°C) for low and high flow conditions over 2013 and 2014. Flow regimes were differentiated by the 50th percentile in discharge for each site as given in Table A-2.

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Table A-6. Statistics of observed and modeled river temperatures and model performance (°C) for the lower model domain for the 2013 and 2014 model periods. Daily temperature range statistics were calculated as the difference between minimum and maximum temperatures observed or modeled for each calendar day. *Model performance at site 1 evaluated for a reduced period of July 10th through August 5th due to delayed sensor deployment.

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Appendix B
Supporting Information for Chapter 3
Introduction

This document includes a description of how calibration parameter ranges were set in Chapter 3 (Text B-1), tables that describe data collection locations and model parameter calibration ranges (Tables B-1 and B-2), figures of the observed sediment temperature, an example Pareto front, tracer study results, distribution of calibration parameter values, result from model calibration, discharge and temperature records for Site 8, and results from model scenarios (Figure B-1 through B-8).

Text B-1: Parameter Calibration Ranges

In this model formulation, there are seven model parameters that require calibration: surface transient storage fraction \((\beta)\), channel roughness \((n)\), surface transient storage cross sectional area \((A_{C,STS})\), surface transient storage exchange coefficient \((\alpha_{STS})\), depth of hyporheic exchange \((Y_{HTS})\), hyporheic exchange coefficient \((Q_{HTS})\), and depth to ground temperature \((Y_{gr})\). Potential parameter ranges permitted within calibration were set based on field observations for \(Y_{HTS}\) and \(Y_{gr}\) and literature values for the remaining model parameters (Table B-2). The \(\beta\) calibration range (15 % to 30 %) was set based on values reported in Neilson et al. [2010a], and velocity transects collected during discharge measurements agree with these estimates (Neilson, unpublished data). \(n\) was set to range from 0.06 to 0.1 based on values from Chapters 2, 4, and Kane et al. [2003]. Ranges for \(n\) are higher than typically reported for cobble beds to account for emergent boulders that become significant under the low-flow conditions, channel sinuosity, and channel constrictions observed in this study [Arcement and Schneider, 1989; Barnes,
1967]. The range for $A_{C,STS}$ (0.52 to 1 m²) was set based on values reported by Edwardson et al. [2003], Neilson et al. [2010b], and Bingham et al. [2012]. Calibrated $a_{OTIS}$ values from Edwardson et al. [2003] were used to set the ranges for $a_{STS}$ (10 to 2500 cm² s⁻¹) and $Q_{HTS}$ (1000 to 33000 cm³ s⁻¹) using equations S1 and S2 from Neilson et al. [2010a]:

\[
\alpha_{OTIS} = \frac{a_{STS}Y_{STS}}{\beta B_{tot}A_{cs,MC}} \quad \text{(Eq. B-1)}
\]

\[
\alpha_{OTIS} = \frac{Q_{HTS}}{A_{cs,MC}\Delta x} \quad \text{(Eq. B-2)}
\]

where $B_{tot} =$ channel top width (m), $A_{cs,MC} =$ cross-sectional area of the main channel (m²), and $\Delta x =$ model cell length (m). The $a_{OTIS}$ values reported in Edwardson et al. [2003] are similar to those reported by Zarnetske et al. [2007], Runkel [1998], and Westhoff et al. [2010]. The resulting $a_{STS}$ and $Q_{HTS}$ values are similar to those used in Neilson et al. [2010b] and Bingham et al. [2012]. $Y_{HTS}$ values were set to range from zero to the maximum depth of observed solute breakthrough curves and near main channel temperature patterns in piezometers from field data collection. Note that including a zero depth in the calibration range for $Y_{HTS}$ allows for parameter selection that excludes hyporheic exchange thus allowing us to test if hyporheic exchange is necessary to reproduce observed main channel temperature and solute records. The calibration range for $Y_{gr}$ was set based on the range of piezometers depths at which sediment temperatures measured in piezometers were similar to temperatures observed at the 60 cm depth at Site 9. Model parameters of sediment specific heat capacity ($C_{p,SED} = 3.18$ J g⁻¹ °C⁻¹), thermal diffusivity ($a_{SED} = 0.009$ cm² s⁻¹), and density ($\rho_{SED} = 1.51$ g cm⁻³), were set to literature
values and were not varied in the calibration process (Table S2) [Chapra et al., 2008].
Table B-1. Description of data collection and application

<table>
<thead>
<tr>
<th>Observation</th>
<th>Site</th>
<th>Purpose</th>
<th>Stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discharge</td>
<td>Injection Site</td>
<td>Upstream flow boundary condition</td>
<td>Model calibration</td>
</tr>
<tr>
<td></td>
<td>1500 m</td>
<td>Downstream boundary condition for estimating lateral inflows</td>
<td>Model calibration</td>
</tr>
<tr>
<td></td>
<td>Site 9</td>
<td>Upstream flow boundary condition</td>
<td>Model testing</td>
</tr>
<tr>
<td></td>
<td>Site 8</td>
<td>Downstream boundary condition for estimating lateral inflows</td>
<td>Model testing</td>
</tr>
<tr>
<td>Meteorological Conditions</td>
<td>UK Met</td>
<td>External forcing for temperature modeling</td>
<td>Model calibration and testing</td>
</tr>
<tr>
<td>Sediment Solute BTCs</td>
<td>500 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>600 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>1200 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td>MC Solute BTCs</td>
<td>500 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>600 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>1200 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>1500 m</td>
<td>Model calibration</td>
<td>Model calibration</td>
</tr>
<tr>
<td>Sediment Temperatures</td>
<td>Site 9</td>
<td>Estimating $Y_{gr}$, forcing for ground conduction in temperature modeling</td>
<td>Parameter estimation, model calibration, and model testing</td>
</tr>
<tr>
<td></td>
<td>500 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>600 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td></td>
<td>1200 m</td>
<td>Estimating $Y_{HTS}$</td>
<td>Parameter estimation</td>
</tr>
<tr>
<td>River Top Widths</td>
<td>Site 9 - Site 8</td>
<td>Estimation of surface area of air-water and water-sediment interfaces</td>
<td>Model calibration and model testing</td>
</tr>
<tr>
<td>MC River Temperatures</td>
<td>Injection Site</td>
<td>Upstream temperature boundary condition</td>
<td>Model calibration</td>
</tr>
<tr>
<td></td>
<td>1500 m</td>
<td>Downstream temperature boundary condition for model calibration</td>
<td>Model calibration</td>
</tr>
<tr>
<td></td>
<td>Site 9</td>
<td>Upstream temperature boundary condition</td>
<td>Model testing</td>
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<tr>
<td></td>
<td>Site 8</td>
<td>Downstream temperature boundary condition for model evaluation</td>
<td>Model testing</td>
</tr>
</tbody>
</table>
Table B-2. Model parameter calibration ranges, sources, and results

<table>
<thead>
<tr>
<th>Parameter Description</th>
<th>Parameter Name</th>
<th>Calibration Range</th>
<th>Estimate Source</th>
<th>Pareto Optimal</th>
<th>Solute End-member</th>
<th>Temperature End-member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width of surface transient storage zone (%)</td>
<td>$\beta$</td>
<td>15 - 30</td>
<td>Neilson et al. 2010b, Field Observations</td>
<td>20.8</td>
<td>16.9</td>
<td>22.8</td>
</tr>
<tr>
<td>Manning roughness coefficient</td>
<td>$n$</td>
<td>0.05 - 0.1</td>
<td>Kane et al. 2003, Chapters 2, 4</td>
<td>0.1</td>
<td>0.1</td>
<td>0.08</td>
</tr>
<tr>
<td>Cross sectional area of STS (m2)</td>
<td>$A_{CS,STS}$</td>
<td>0.52 - 1.0</td>
<td>Neilson et al. 2010b, Edwardson et al. 2003, Bingham et al. 2012</td>
<td>0.86</td>
<td>0.92</td>
<td>0.72</td>
</tr>
<tr>
<td>STS exchange coefficient (cm2/s)</td>
<td>$\alpha_{STS}$</td>
<td>13 - 2439</td>
<td>Neilson et al. 2010b, Edwardson et al. 2003</td>
<td>1656</td>
<td>1769</td>
<td>69</td>
</tr>
<tr>
<td>HTS exchange coefficient (cm3/s)</td>
<td>$Q_{HTS}$</td>
<td>1000 - 32526</td>
<td>Neilson et al. 2010b, Edwardson et al. 2003</td>
<td>20464</td>
<td>19299</td>
<td>13525</td>
</tr>
<tr>
<td>Depth of hyporheic exchange (m)</td>
<td>$Y_{HTS}$</td>
<td>0 - 0.7</td>
<td>Field Observations</td>
<td>0.11</td>
<td>0.09</td>
<td>0.33</td>
</tr>
<tr>
<td>Depth to ground temperature (m)</td>
<td>$Y_{gr}$</td>
<td>0.5 - 1.0</td>
<td>Field Observations</td>
<td>0.51</td>
<td>0.50</td>
<td>0.65</td>
</tr>
<tr>
<td>Calculated HTS residence time (min)</td>
<td>$R_{HTS}$</td>
<td>-</td>
<td>-</td>
<td>2.7</td>
<td>1.9</td>
<td>13.1</td>
</tr>
</tbody>
</table>
Figure B-1. Main channel (thick black line) and sediment temperatures at each pieometer (blue lines) at 500 m (top), 600 m (middle), and 1200 m (bottom) downstream from the tracer study injection site. Site 9 sediment temperatures at 60 cm (thin black line) are similar 57 cm depth at 500 m, 69 cm depth at 600 m, and 73 cm depth at 600 m. These ranges of depth were used to bound the calibration range for $Y_{gr}$. 

Figure B-2. Objective space from dual objective function model calibration. Each point is a model run, with x and y coordinates representing 1-Nash Sutcliffe Efficiency for solute and temperature, respectively. Red points are the “Pareto front” where model runs are indistinguishable in terms of multi-objective function optimization. The “Pareto optimal” model run (black dot) is the optimum tradeoff between the dual objective functions and has the shortest Euclidian distance from the origin.

Figure B-3. 4 July 2017 tracer study solute break through curves (left) and temperature time series (right) in the main channel (black) and at various depths in the sediment (colored lines) at 500 (top) and 1200 (bottom) meters downstream from the injection location.
Figure B-4. 10 July 2017 tracer study solute break through curves (left) and temperature time series (right) in the main channel (black) and at various depths in the sediment (colored lines) at 500 (top) and 600 (bottom) meters downstream from the injection location.

Figure B-5. Parallel coordinate plot of model calibration parameter sets that produce Pareto front model results (grey lines). Bold lines indicate the Pareto optimal (black), temperature end-member (red), and solute end member (blue) model calibrations.
Figure B-6. Modeled main channel and hyporheic transient storage zone solute breakthrough curves and temperatures 1500 m downstream from the injection location using the Pareto optimal parameter set.

Figure B-7. Site 8 hydrographs (blue lines) and temperature time series (red lines) for June through late August for 2013 - 2017. Grey boxes in indicate the periods of poor model performance in 2013 reported in Chapter 2, the 7 day model test in 2015, and the three tracer studies in 2017.
Figure B-8. Effects of hyporheic transient storage on main channel temperatures. Observed (blue), modeled with hyporheic transient storage (solid red line), and modeled without hyporheic transient storage (dashed red line) temperatures are shown for the downstream end of the 13.1 km long study reach for 2017 (top), 2015 (middle), and 2013 (bottom). Averaged over the three simulation periods, hyporheic transiet storage buffers instream temperatures by 2.2 °C and lowers main channel temperatures by 0.23 °C.
Figure B-9. Sensitivity of main channel temperatures (A), ground conduction (B), hyporheic exchange (C), and bed conduction (D) to ground temperatures. River temperatures vary by 0.2 °C per 1 °C change in ground temperature. Ground conduction and hyporheic exchange heat fluxes vary by 8 and 9 W m$^{-2}$ per °C change in ground temperatures, respectively, while bed conduction is not sensitive to ground temperatures.
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VITA

Tyler V. King
Graduate Research Assistant, PhD Candidate
Department of Civil and Environmental Engineering
Utah Water Research Laboratory, Utah State University
Email: hydorty@gmail.com

Education

Doctor of Philosophy, Civil and Environmental Engineering, Anticipated August 2018
Utah State university, Logan, Utah

Master of Science, Hydrology December 2013
University of New Hampshire, Durham, New Hampshire

Bachelor of Science, Hydrology May 2010
University of New Hampshire, Durham, New Hampshire

Research Experience

2013-Current  Presidential Doctoral Research Fellow, Utah State University, Logan, UT.

2011-2012  Masters Student, University of New Hampshire, Dept. Earth Science, Durham, NH

2010-2011  Fulbright Student, Norwegian University of Science and Technology, Trondheim, Norway

2009  Student Researcher, Alaska Pacific River Forecasting Center, Anchorage, AK

Research Interests

Mechanistic river temperature modeling -
- Numerical modeling of river temperatures in high latitude regions
- Determination of dominant heat fluxes for evaluation of climate change projections
- Evaluation of spatial and temporal impacts of hydropower production on river temperatures

Hydroinformatics -
- Management of large hydrologic datasets
- Relational database management
Remote sensing in hydrology -
- Developing techniques for extracting channel geometry from aerial imagery
- Estimation of volumetric river discharge by coupling remotely sensed channel geometry with 1-D open channel flow models

Peer Reviewed Publications

Refereed Journal Articles:

Published


4. In Review


In Prep
1. King, T.V., B.T. Neilson “Influences of hyporheic exchange on Arctic river temperatures”. In preparation.


Professional Presentations (2014-Current)

First Author


Co-Author


where *=mentored student


Professional Activities

Invited Lectures/Presentations:

2017 Utah State University, CEE 5003/6003 Remote Sensing of Land Surfaces “The view from here: Remote Sensing in Climate Change” Logan, Utah, USA

2017 University of New Hampshire Earth Science Department. “Heat flux dynamics in low Arctic Rivers” Durham, New Hampshire, USA

2016 Max Planck Institute of Biogeochemistry. “Water Temperature Controls in Low Arctic Rivers” Jena, Germany.

Reviewer:

2017 Journal Reviewer, Geophysical Research Letters
Current Professional Societies

2017  American Water Resources Association
2017  Geological Society of America
2016  Permafrost Young Researchers Network
2016  United States Permafrost Association
2006  American Geophysical Union

Scientific Outreach

2017  Collaborating partner in Research Experience of Teachers, a National Science Foundation funded projects to incorporate current scientific research into public education curriculum by involving teachers in field work and involving researchers in curriculum development. Will host two high school teachers at Toolik Field Station, North Slope, Alaska for two weeks in July 2017, working to help develop their understanding of Arctic hydrology. Will continue to work with these teachers to develop curriculum for them to use in their classes. These lessons will be publically available.

2016  Authored the “Utah” entry for the U.S. State Department as part of the U.S. Chairmanship of the Arctic Council. This text is published online, and has been published as a printed book.


Awards

2017  Geological Society of America Hydrogeology Division Outstanding Graduate Research Proposal
2017  Outstanding Graduate Scholar Award, 2017 USU College of Engineering
2017  Outstanding Poster Presentation Award, 2017 USU Spring Runoff Conference
2016  USU Civil and Environmental Engineering Graduate Student Researcher of the Year 2016
2016  USU Graduate Student Travel Grant
2015  Outstanding Student Paper Award – 2015 AGU Fall Conference
2013  USU Presidential Doctoral Research Fellow
2012  Teaching Assistantship in Earth Science at the University of New Hampshire
2010  Student Fulbright Appointment
2009  NOAA Hollings Scholar
2009  John and Rose Mendelson Kurtz Scholarship
2008  Sidney and Kathleen Samuels Scholarship Fund
2008  New Hampshire Incentive Program Scholarship
2008  Penney Family Scholarship
2007  Glenice Dearborn Scholarship
2006  University of New Hampshire Dean’s Scholarship
2006  Message of Hope Award: Ulman Cancer Fund for Young Adults
2006  Kate Harvey Burns Scholarship
2006  General Henry H. Arnold Education Grant Program

**Funding**

2017  Geological Society of America Travel Grant ($250)
2017  Geological Society of America Graduate Student Research Grant ($2,500)
2016  USU Graduate School Travel Grant ($400)
2013  Presidential Doctoral Research Fellowship ($101,630)
2010  Student Fulbright Appointment ($18,500)
2009  NOAA Hollings Scholar ($26,000)
Teaching Experience

Instructor:

2017  **Introduction to Catchment Hydrology**, Utah State University, Intensive English Language Institute. Taught a two day, six hour, short course on the fundamentals of catchment hydrology to 12 students of Civil and Hydropower Engineering from the Nanchang Institute of Technology as a capstone of a three week Intensive English Language course. Lectured, lead discussions, and facilitated small in-class exercises on watershed delineation, water balance calculations, discharge calculation, rating curve development, and remote sensing of river discharge.

Teaching Assistantships:

2016  CEE 6740 – Surface Water Quality Modeling, Utah State University

2012  ESCI 810 – Groundwater Hydrology, University of New Hampshire

2011  ESCI 354 – Techniques in Environmental Science, University of New Hampshire

2010  CHEM 404 – General Chemistry II, University of New Hampshire

Student Mentoring

2017  Nikki Quinney, B.S. Research Assistant, Utah State University. I supervise and mentor Nikki in data management and processing. She is showing significant interest and investment in the research process.

2014  Mitchell T. Rasmussen, B.S. Environmental Engineering, Utah State University. I mentored and supervised Mitchell in lab work, filed work, data processing, and numerical modelling. Over the course of two years Mitchell participated in and took leadership roles in all aspects of the research process and demonstrated exceptional growth and personal development. Mitchell is now enrolled in a master’s program at Oregon State.

2013  Levi Overbeck, MS Civil and Environmental Engineering, University of Alaska Fairbanks. I worked with and mentored Levi in field work, data processing, and analysis for his thesis “Determination of lateral inflows in the Kuparuk river watershed, a study in the Alaskan Arctic.”
Professional Development

2018  Reviewing Undergraduate Research and Creative Opportunity Proposals Workshop, Logan, UT

2017  Strong Mentorship for Undergraduates Workshop, Logan, UT

2017  Finding/Applying for Teaching-Focused Higher Ed Positions Workshop, Logan, UT

2017  Tips for Teaching Undergraduate Courses, Graduate Training Series, Logan UT

2015  Write Winning Grants Workshop, Logan, UT

2016  National Science Foundation Arctic Field Training, Logan, UT

2011  National Science Foundation Responsible Conduct of Research, Durham, NH

Volunteer and Outreach Experience

Community Outreach

2016  Community night at USU Engineering Week. Developed and hosted a “thermal camera photo booth” where the public were encourage to learn about long wave radiation in the context of personal portraits of themselves.

2015  Utah Water Research Laboratory 50th anniversary open house. Presented approaches to and results from Arctic hydrology in a context that is meaningful for residents of Cache Valley, Utah through an open house at the Utah Water Research Laboratory.

2014  Public presentation, Kantishna Backcountry Lodge, Denali National Park, Alaska. Presented Arctic hydrology research to guests and staff of a backcountry lodge in Kantishna, Alaska as a guest presenter in their scientific enrichment program at the lodge.

Service

2015  AmeriCorps through the Utah Conservation Corps, Logan UT. I coordinated and
oversaw operations of grooming cross country ski trails in Bear River Mountains, Cache County Utah for Nordic United, a non-profit organization based in Logan, Utah.

**Non-Profit Memberships**

Winter Wildlands Alliance – Boise, ID  
Save Our Canyons – Salt Lake City, UT  
Wasatch Back Country Alliance – Salt Lake City, UT  
Nordic United – Logan, UT  
Union of Concerned Scientists – Washington, DC