Constraining accretion rates in a tide-dominated, freshwater river (Pitt River, Canada) and implications for lateral accretion of channels in the tidal-fluvial transition

by
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Abstract

A vibracore-based investigation of channel and floodplain deposits in the Pitt River Valley (PRV) was conducted to spatially and temporally confine the evolution of the PRV floodplain and the tide-dominated Pitt River. Sedimentological and ichnological assessment of cores is supplemented by geochronological and palynological analyses.

The lateral migration rate of a Pitt River meander is quantified using Carbon-14 age dates of organic detritus at the base of a channel-margin core and the position of the core relative to the present-day channel profile. The Pitt River meander bar is shown to laterally migrate between 0.16 and 0.28 metres per year. A comparison of this rate to previous studies reveals that channels modulated by tides are capable of migrating at a rate equivalent to slowly meandering purely fluvial systems, and tidally affected channels migrate at less than 1.5% of the channel’s width each year.

Keywords: vibracore; tide-dominated; mixed tidal-fluvial; lateral migration; bank accretion; bend curvature
Restlessness is discontent and discontent is the first necessity of progress. Show me a thoroughly satisfied man and I will show you a failure.

Thomas Edison
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Chapter 1.

Introduction

Our appreciation of deposition in mixed tidal-fluvial systems is restricted mainly to empirical observations of sediment distributions and flow conditions over geologically insignificant timespans. Consequently, our ability to reliably predict how these complex systems evolve with time is limited, and characterization of sediments deposited where tides and rivers interact is largely based on systems in which the tidal signature and brackish-water signature have not been separated. The Pitt River, British Columbia, Canada is a distinctive depositional environment that presents an opportunity to expand our understanding of tidal-fluvial systems. In the case of the Pitt River, flow is largely tide-dominated but waters are fresh. Characterization of sediments in this system enables us to investigate the tidal signature independent of brackish-water, and to evaluate the longer-term evolution of a tidal river.

The research presented herein is based upon a vibracoring program across the lower Pitt River Valley (PRV) in the municipality of Pitt Meadows. Evaluation of the vibracores comprises two related but distinct projects. In Project 1 (Chapter 2), the data are used to augment the current model of post-glacial evolution of the PRV (i.e., Clague et al. 1983). Nine sites were drilled and cored to assess the depositional history of the Pitt River and to define the character and distribution of Pitt River deposits (vs. Fraser River Delta and PRV floodplain deposits). In addition, palynological data were collected in an attempt to confirm a marine or non-marine signature. Acquired age dates are used to constrain the age of PRV sediments and Pitt River deposits. In Project 2 (Chapter 3), the stratigraphy and time-line established in Project 1 is used to consider the evolution of the tide-dominated Pitt River. The lateral migration rate of the channel is derived from vibracore data and air photo interpretation. These results are compared to published lateral migration rates of purely fluvial channels to show that channel migration in the
tidal-fluvial transition does occur (i.e., Van De Wiel 2003; Nicoll and Hickin 2010), though at a rate equivalent to slowly meandering purely fluvial systems.

1.1. Research objectives

The objectives of this research can be summarized into three key sets of questions:

1) Is the presently accepted model for the evolution of the Pitt River Valley accurate? Do our data confirm the previously interpreted position of stratigraphic and geographic boundaries, and can the chronology for the marine-to-non-marine transition be refined?

2) What are the sedimentological and ichnological characteristics of the Pitt River deposits? Is there evidence of tides, and if so, what are they? How do these criteria differ from those of brackish-water and tidal deposits?

3) Has the Pitt River migrated laterally with time? If so, what was the rate of lateral accretion and how does this compare to those of purely fluvial systems?

1.2. The Pitt River tributary: previous work and background

The Pitt River is a wholly freshwater, tide-dominated tributary that joins the lower Fraser River 30 km inland of the southern British Columbian coastline (Fig. 1.1). The Pitt tributary drains 1640 km² of the Coast Mountains. The tributary possesses a river-lake-river morphology (upper Pitt River, Pitt Lake, Pitt River) that extends over 70 km in a north-south direction. The focus of this work is on the Pitt River, which is 20.1 km long and has a sinuosity of 1.20 (Ashley 1977).

The Pitt River is relatively under-studied in comparison to the lower Fraser River. Ashley (1977, 1978) and Smith (1985) conducted field studies on the Pitt River, and several regional studies (e.g., Johnston 1922; Milliman 1980; Clague et al. 1983) have described the Pitt tributary as a component of the lower Fraser River. Cores obtained from Pitt Lake and Pitt River point bars reveal rhythmic alternations of sand and mud,
interpreted as a result of deposition in a tide-dominated setting (Johnston 1922; Ashley 1977, 1978; Smith 1985).

1.2.1. The Pitt River Valley (PRV)

The Pitt River Valley (PRV) is one of multiple north-south oriented valleys that connect to the Fraser River in the lower mainland of British Columbia. During the last deglaciation, the westerly valleys, including the PRV, were inundated with seawater and formed fjords. The westward progradation of the Fraser River during the Holocene infilled the mouth of the PRV and led to the gradual transition of the PRV fjord into a freshwater lake. The Pitt River connects Pitt Lake to the Fraser River, and is both tide-
dominated and filled with freshwater (Ashley 1977). The saltwater wedge extends up the Fraser River to a position 10 km seaward of the Pitt-Fraser confluence (Ashley 1977).

1.2.2. Flow and tides in the Pitt River

The Pitt River is deemed tide-dominated because the majority of water and bedload-transported sediment moving through the river is sourced from downstream (via the Fraser River); flow in the channel reverses with the tidal cycle, and sediment transport is driven by flood-oriented flows. Fluctuations in discharge in the Pitt River reflect seasonal changes and mixed semi-diurnal tidal cyclicity. While over 50% of discharge through the Pitt River during the freshet (annually from May to August) reflects catchment-basin drainage, only 5% can be attributed to catchment-basin drainage during base flow (September to April; Ashley 1977). The remaining 95% of water flowing through the Pitt River during base flow is sourced from the Fraser River, forcing water up the Pitt River during the flood tide. Flow reverses in the Pitt River as the water level rises in the tidal backwater zone of the lower Fraser River. Bidirectional flow persists in the Pitt River during base flow, with unidirectional flow occurring only during the freshet. Even during the freshet, however, tidal backwater is generated (Ashley 1977).

Tidal flow is unrestricted at both ends of the Pitt River due to the reservoir capacity of Pitt Lake. The tidal range in Pitt Lake is up to 1.2 m (Johnston 1922; Ashley 1977; Milliman 1980). Flood tidal currents commonly have greater peak velocities than ebb tidal currents during base flow (Fig. 1.3; Table 1.1; Ashley 1977, 1978). Conversely, ebb tides exceed flood tides in duration throughout the entire year and the net discharge direction is downstream (Fig. 1.2). As such, the Pitt River must still be considered a river, albeit a tide-dominated one. Flood tidal dominance is discernable from the northward-oriented bedforms, as over 65% of bedforms during flood and ebb tides in both winter and summer months are flood oriented (Ashley, 1977). In addition, sediments comprising the Pitt Delta, deposited at the southern end of Pitt Lake, are derived from the Fraser River and are transported as bedload (Ashley 1977).
Figure 1.2. Pitt River tidal curves. Ebb tides persist for longer than flood tides, but flood tides exceed ebb tides in peak flow velocity, and constitute the dominant sand transport mechanism. From Ashley (1977).

Figure 1.3. Ashley’s (1977) flow velocity recording points (A-C) in reference to this study’s vibracore drill site S3 (red halo). Google Earth image.
Table 1.1. Peak velocities recorded by Ashley (1977) at 0.4 channel depth. Note that locations A and B are located proximal to vibracore site S3 (Fig. 1.3).

<table>
<thead>
<tr>
<th>Location</th>
<th>Time of year</th>
<th>Pitt River max. mean flood velocity (m s(^{-1}))</th>
<th>Pitt River max. mean ebb velocity (m s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>NNE of S3 Base flow (March 11, 1975)</td>
<td>0.52</td>
<td>0.40</td>
</tr>
<tr>
<td>B</td>
<td>NWW of S3 Base flow (October 8, 1975)</td>
<td>0.64</td>
<td>0.38</td>
</tr>
<tr>
<td>C</td>
<td>Pitt-Fraser confluence End of freshet (August 11, 1975)</td>
<td>0.59</td>
<td>0.44</td>
</tr>
</tbody>
</table>

Ashley (1977) calculated maximum discharge at the northern limit of Pitt River during the end of the freshet as well as base flow at both peak flood and ebb flow periods. She determined that the greatest magnitude of discharge occurred during base flow at peak flood flow (2400 m\(^3\) s\(^{-1}\)). In contrast, base flow discharge at peak ebb flow was 2080 m\(^3\) s\(^{-1}\). At the waning end of the freshet, peak flood and ebb discharge rates were proposed to reach 1800 m\(^3\) s\(^{-1}\) and 950 m\(^3\) s\(^{-1}\), respectively.

1.2.3. Anthropogenic modification of the PRV

Dyking of the Pitt and Fraser rivers to facilitate reclamation of the PRV dates back to the 1890s (Collins 1975). However, the original dykes proved to be largely ineffective, and were supplanted by a new dyke system constructed in the 1950s, following the Fraser River flood of 1948. The dykes that train the Pitt River channel and the south end of Pitt Lake have halted any migration of the system for the past 60 years.

1.3. Methods

The results of this thesis are drawn from nine vibracores collected across the PRV floodplain from June to November 2014 (Fig. 1.4). A Wink Vibracore drill was used to collect all nine cores, and an in-house manual for vibracore drilling developed during the summer of 2014 can be viewed in Appendix A.

Spacing between neighbouring cores is approximately 2000–2500 m, with the exception of cores S3 and S9, which were drilled 270 m apart. Cores were drilled in the Pitt-Addington Marsh Wildlife Management Area and along the margins of private and commercial farming fields. General localities for drilling sites were selected to have a
transect orientation along strike of the PRV (NNE-SSW) and a short intersecting transect across the mouth of the PRV (roughly NW-SE). A variety of factors dictated the specific locations of drill sites, primarily: convenience for the landowners (i.e., no drilling on land occupied by crops); the likelihood of collecting an undisturbed sample (cores had to be drilled away from packed or dyked zones and where the ground wasn’t significantly reworked/tilled); and preferred proximity to the truck (to reduce the distance that the heavy equipment needed to be manually transported). These factors significantly limited the viable areas for data collection.

Depth of penetration of the vibracores ranged from 11.9 to 15.1 m. All cores were taken back to the laboratory and processed. Cores were slabbeted using a Dremel™, fixed with a diamond-coated cutting wheel to cut through the rigid PVC core tubes. Cores were X-rayed in ~30 cm increments while the samples were still saturated with water. Once they dried completely, cores were described to characterize sedimentological and ichnological variation. Selected sand-dominated intervals were resin peeled using a two-part epoxy resin.

Six samples from various cores were selected for palynological analysis. Samples were processed at the Le Geotop-UQAM facility in Montréal. Additionally, Carbon-14 radiocarbon ages from five samples of organic material were processed in the Keck-UCIAMS facility at UC Irvine after being prepared by Dr. Duane Froese at the University of Alberta. The results of these data helped to spatially and temporally constrain the paleodepositional environments inferred from the sedimentology and ichnology of the core samples.
Figure 1.4. Locations of the nine vibracores drilled in Pitt Meadows. The dashed white box outlines the areal extent photographed in Figure 3.2.
Chapter 2. Mid-Holocene evolution of the Pitt River Valley floodplain

The interpretation of the post-glacial evolution of the Fraser lowland following the Fraser Glaciation has been refined several times over the last 45 years, primarily based on lithological assessments of cores and radiocarbon dates (e.g., Mathews et al. 1970; Armstrong and Hicock 1980; Armstrong 1981; Clague et al. 1982, 1983; Williams and Roberts 1989; Clague and James 2002). Presently, it is agreed that a series of sea level changes occurred along the SW coast of British Columbia at the end of the Fraser Glaciation 13 k $^{14}$C BP. These sea level changes reflected the complex interaction between eustatic sea level rise, isostatic rebound, and tectonism (Mathews et al. 1970; Clague et al. 1982). Sedimentation on the Fraser River floodplain and in adjacent coastal regions is largely the product of these changes.

Several regional studies include data collected across and/or south of the Pitt River Valley (PRV) (e.g., Armstrong and Hicock 1980; Armstrong 1981; Clague et al. 1982, 1983; Williams and Roberts 1989). Comparatively little research has focused on the evolution of the PRV (Collins 1975; Ashley 1977; Smith 1985; Locher 2006, 2014; Katzie Development Corporation Archaeology 2010). This chapter incorporates findings from this research with those presented in technical reports (e.g., Smith 1985; Locher 2014), geological publications (e.g., Ashley 1977) and anthropological studies (e.g., Locher 2006), in order to build a more precise spatial and temporal model for the evolution of the PRV through the mid- to late-Holocene.

2.1. Holocene sea level history of the Fraser lowland

Following the retreat of the Cordilleran ice sheet at the end of the Fraser Glaciation 13 k $^{14}$C BP, rapid relative sea level fall occurred in response to isostatic rebound (Fig. 2.1; Mathews et al. 1970; Clague et al. 1982). Sea level fell from 200 m above present-day mean average sea level (masl) 13 k $^{14}$C BP, to 20-30 m above masl by 11 k $^{14}$C BP. By 10 k $^{14}$C BP, glaciers in the Fraser lowland had completely melted,
and sea level continued to fall from 10 m above masl to at least 12 m below masl by 9 k \(^{14}\text{C} \text{BP}\) (Clague et al. 1982; Williams and Roberts 1989).

Sea level began to rise towards present-day sea level approximately 8 k \(^{14}\text{C} \text{BP}\) (Williams and Roberts 1989), and much of this increase occurred between 8 and 6 k \(^{14}\text{C} \text{BP}\) (Fig. 2.1; Clague et al. 1982). The Fraser lowland rapidly aggraded during this time (~5 mm y\(^{-1}\); Clague et al. 1982). Following a stillstand between 6.2 and 5.8 k \(^{14}\text{C} \text{BP}\), relative sea level rise continued until it reached its present position approximately 2.25 k \(^{14}\text{C} \text{BP}\) (Williams and Roberts 1989).

The evolution of the PRV through the Holocene is summarized as the following series of events:

13-11 k \(^{14}\text{C} \text{BP}\): While regional sea level (RSL) fell rapidly, the PRV remained glaciated and ice from the PRV glacier calved directly into the sea (Armstrong 1981; Clague et al. 1983).

11-10.5 k \(^{14}\text{C} \text{BP}\): By 10.5 k \(^{14}\text{C} \text{BP}\), the Pitt valley glacier had melted. The “Pitt Fjord” was connected directly to the sea (Armstrong 1981; Clague et al. 1982, 1983).

10.5-10 k \(^{14}\text{C} \text{BP}\): RSL continued to fall and the Fraser River delta prograded westward across the mouth of the PRV (Fig. 2.2). The Pitt Fjord became disconnected from
the sea, forming a proto-Pitt Lake and the incipient Pitt River. A mudflat/estuarine environment spanned across the southern end of the PRV, extending from Maple Ridge to New Westminster (Mathews et al. 1970; Clague et al. 1982; Clague and James 2002).

10-8 k \(^{14}\)C BP: RSL reached and maintained a low of around 12 m below sea level 9 k \(^{14}\)C BP. Sedimentation across the PRV floodplain occurred while sea level remained low (Clague et al. 1982; Williams and Roberts 1989).

8-6 k \(^{14}\)C BP: Sea level began to rise between 8-6 k \(^{14}\)C BP, during which time rapid (~5mm y\(^{-1}\)) aggradation of PRV floodplain occurred (Clague et al. 1982; Williams and Roberts 1989).

6 k \(^{14}\)C BP-present: Sea level rise slowed and the PRV floodplain stabilized. This led to soil development and continued human occupation at the south end of the PRV floodplain (south of the PRV’s bedrock valley walls) dating back to 5.0 k \(^{14}\)C BP (Williams and Roberts 1989; Katzie Development Corporation Archaeology 2010).

Figure 2.2. Westward progradation of the mouth of the Fraser River between 10.5 and 10 k \(^{14}\)C BP. This is the time period in which the Pitt Fjord was cut-off from the Strait of Georgia. From Clague et al. (1983).

### 2.2. Previous studies

Regional mapping of surface sediments in the PRV by Armstrong and Hicock (1980) showed that the floodplain was dominated by deltaic sediments sourced from the Fraser River (Fig. 2.3). Secondary sediment sources include mountain streams and
Pleistocene-aged glacial deposits. Mapping of shallow subsurface sediments was partly achieved by acquiring sediment cores, many of which have been described by Armstrong (1981), Clague et al. (1982), and Williams and Roberts (1989). The cores listed in Table 2.1 were drilled in various locations along the Fraser River in close proximity to the Pitt-Fraser confluence (Figs. 2.3 and 2.4). These cores are dominated by silt and are devoid of significant (m-scale) peat horizons and coarse sand packages that are common in Fraser River channel deposits further west and south of the PRV (e.g., Clague et al. 1983; Williams and Roberts 1989). Radiocarbon dates from these cores yield ages of 8-7 k \(^{14}\)C BP at approximately 10 m below masl.

In addition to regional geological studies that include the PRV, archeological research inside the PRV has yielded geologically relevant data (Figs. 2.3 and 2.4). Cores collected for an archaeological impact assessment (AIA) at the northern end of the PRV floodplain range in length from 20 to 30 m and are dominated by silt and silty sand (Locher 2014). A high degree of lateral heterogeneity between cores inhibited the correlation of lithological packages across the valley. This lithological heterogeneity is considered to reflect frequently shifting depositional conditions in a mostly subaqueous environment. These deposits are interpreted as a combination of tidally derived sediment from the Fraser River and overbank deposits from freshet floods.

Radiocarbon dates obtained from Locher’s (2014) cores showed that, in accordance with the rest of the Fraser lowland, rapid vertical accretion occurred between ~8-6 k \(^{14}\)C BP in the PRV (from 7790 ± 20 \(^{14}\)C BP at 6.6 m below masl to 6030 ± 15 \(^{14}\)C BP at <2.8 m above masl; Table 2.1). Stabilization of the PRV floodplain over the past 2 ky is evidenced by the radiocarbon dates within the top 1 m of soil, which range from 2700 ± 25 to 2240 ± 25 \(^{14}\)C BP.

Diatom analysis from one of Locher’s (2014) cores, drilled < 600 m from the channel margin of the Pitt River, indicates that a brackish to freshwater shift occurred when sediment was deposited ~13 m below masl (roughly 20 m core depth), approximately 9080 ± 20 \(^{14}\)C BP. This transition is interpreted as a shift from a brackish lagoon to a freshwater embayment (Locher 2014).
Figure 2.3. The surficial geology of the lower PRV defined by Armstrong and Hicock (1980). Bedrock (pink; PT) bounds Fraser River sediment (silt and sand; light yellow; Fb and Fc) in the PRV floodplain. East of Pitt River, a postglacial alluvial fan deposit (sand and gravel; orange; SAj,1) and zones of Pleistocene Sumas Drift (gravel and sand; grey; SAb), Vashon Drift (till with sandy matrix; green; Va), and Fort Langley Formation (stoney silt and sand; blue; FLc) exist. Site locations from previous studies within the map area are noted with O symbols (see Fig. 2.4 and Table 2.1) and core locations from this project are noted with X symbols (see Fig. 2.5).
Figure 2.4. Sites (cores) described in previous studies, either proximal to or within the PRV study area. Summarized in Table 2.1, Armstrong (1981) described sites 1 and 2; Clague et al. (1982) site 3; Williams and Roberts (1989) site 4. Archeological site DhRp-52 (site 5) was documented by Katzie Development Corporation (2010). Site 6 encompasses the cores collected by Locher (2014). Selected data results from Locher (2014) are presented in Table 2.1.
Table 2.1. Previously collected radiocarbon ages derived from sites in proximity to the PRV.

<table>
<thead>
<tr>
<th>Source</th>
<th>Site # (Fig. 2.4)</th>
<th>Lab #</th>
<th>Dominant lithology</th>
<th>Depth of organic sample relative to masl (m)</th>
<th>(^{14})C age (BP)</th>
<th>±</th>
<th>Total depth drilled, relative to masl (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Armstrong (1981)</td>
<td>1</td>
<td>GSC-229</td>
<td>peaty silt</td>
<td>-10.4</td>
<td>8290</td>
<td>140</td>
<td></td>
</tr>
<tr>
<td>Armstrong (1981)</td>
<td>2</td>
<td>S-99</td>
<td>peaty silt</td>
<td>-10.3</td>
<td>7300</td>
<td>120</td>
<td></td>
</tr>
<tr>
<td>Clague et al. (1982)</td>
<td>3</td>
<td>GSC-3099</td>
<td>(clayey, sandy) silt</td>
<td>-10</td>
<td>7710</td>
<td>80</td>
<td>&gt;-10</td>
</tr>
<tr>
<td>Williams and Roberts (1989)</td>
<td>4</td>
<td>D24</td>
<td>organic rich silt</td>
<td></td>
<td></td>
<td></td>
<td>&lt;-12</td>
</tr>
<tr>
<td>Katzie Development Corporation Archaeology (2010)</td>
<td>5</td>
<td>UCIAMS-67988</td>
<td>soil</td>
<td>3.5</td>
<td>4985</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>Locher (2014)</td>
<td>6</td>
<td>UCIAMS-135516</td>
<td>clayey silt, silty sand</td>
<td>-6.6</td>
<td>7790</td>
<td>20</td>
<td>&gt;-13</td>
</tr>
<tr>
<td>Locher (2014)</td>
<td>6</td>
<td>UCIAMS-135522</td>
<td>clayey silt, silty sand</td>
<td>2.8</td>
<td>6030</td>
<td>15</td>
<td>&gt;-13</td>
</tr>
<tr>
<td>Locher (2014)</td>
<td>6</td>
<td>UCIAMS-140043</td>
<td>clayey silt, silty sand</td>
<td>5.1</td>
<td>2700</td>
<td>25</td>
<td>-24</td>
</tr>
<tr>
<td>Locher (2014)</td>
<td>6</td>
<td>UCIAMS-140040</td>
<td>clayey silt, silty sand</td>
<td>5.5</td>
<td>2240</td>
<td>25</td>
<td>-24</td>
</tr>
</tbody>
</table>

2.3. Results

Two coring transects were devised and vibracores were collected along each transect (Fig. 2.5). The depths of all cores and their exact position is given in Table 2.2. The north-south (N-S) transect (Figs. 2.5 and 2.6) runs along strike of the PRV, and the east-west (E-W) transect (Figs. 2.5 and 2.7) intersects the long transect at the southern end of the PRV. Samples of organic material were extracted from cores for radiocarbon dating, and mud samples were taken for palynological analysis (see Sections 2.2.1 and 2.2.2)
Table 2.2. Summary of core locations and depths.

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Total depth (m)</th>
<th>Total depth penetrated relative to masl (m)</th>
<th>Mazama ash present? (✓/✗)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>49° 20.00' N</td>
<td>122° 38.24' W</td>
<td>12.0</td>
<td>− 6.8</td>
<td>✓</td>
</tr>
<tr>
<td>S2</td>
<td>49° 18.92' N</td>
<td>122° 37.25' W</td>
<td>9.5</td>
<td>− 4.8</td>
<td>✓</td>
</tr>
<tr>
<td>S3</td>
<td>49° 16.87' N</td>
<td>122° 40.79' W</td>
<td>11.9</td>
<td>− 6.0</td>
<td>✓</td>
</tr>
<tr>
<td>S4</td>
<td>49° 16.80' N</td>
<td>122° 38.47' W</td>
<td>11.4</td>
<td>− 7.5</td>
<td>✓</td>
</tr>
<tr>
<td>S5</td>
<td>49° 14.99' N</td>
<td>122° 37.99' W</td>
<td>11.9</td>
<td>− 11.3</td>
<td>✗</td>
</tr>
<tr>
<td>S6</td>
<td>49° 14.99' N</td>
<td>122° 40.63' W</td>
<td>12.1</td>
<td>− 7.0</td>
<td>✓</td>
</tr>
<tr>
<td>S7</td>
<td>49° 17.83' N</td>
<td>122° 38.35' W</td>
<td>13.0</td>
<td>− 9.7</td>
<td>✗</td>
</tr>
<tr>
<td>S8</td>
<td>49° 15.84' N</td>
<td>122° 39.35' W</td>
<td>14.9</td>
<td>− 10.3</td>
<td>✓</td>
</tr>
<tr>
<td>S9</td>
<td>49° 16.73' N</td>
<td>122° 40.70' W</td>
<td>15.1</td>
<td>− 9.2</td>
<td>✓</td>
</tr>
</tbody>
</table>
Figure 2.5. Core sites (S1-S9) in the PRV. The N-S transect runs parallel to the PRV and the E-W transect is normal to the N-S transect near the southern end of the PRV.
Figure 2.6

Depositional units:
- D1: Soil
- D2: Cohesive mud
- D3: Disturbed silt
- D4: Alternating vf sand & mud
- D5: Coarse sand

Legend:
- Distance between cores
- Mean average sea level (scale relative to masl)
- Mazama ash in D3
- Radiocarbon sample (¹⁴C age)
- Palynological sample (% + marine indication)
Figure 2.7

S3

S9

0%

0.27 km

2.32 km

2.28 km

Pitt River Valley floodplain

W

Mazama ash

Pitt River point-bar

6355 ± 15

6355 ± 15

7835 ± 20

7805 ± 20

0%

2980 ± 15

0%

0% 6020 ± 20

Alluvial fan deposit

S5

E

0 m (masl)

-2 m

-4 m

-6 m

-8 m

-10 m
Figure 2.6.  *Figure 2.5* The N-S transect. For core locations, see Fig. 2.5. The datum for this cross section is mean average sea level (masl). Note that the grain size scale is labeled under the S6 core (furthest right). In the N-S transect, all palynological samples lack a positive marine affinity (0% marine indication).

Figure 2.7.  *Figure 2.5* The E-W transect. For core locations, see Fig. 2.5. For the legend for this transect, refer to Fig. 2.6. The datum for this cross section is mean average sea level (masl). Note that the grain size scale is labeled under the S5 core (furthest right). In the E-W transect, all palynological samples lack a positive marine affinity (0% marine indication), but one sample (S9) has an indeterminable source (?%; see Section 2.2.2).

### 2.3.1. Radiocarbon analysis

A total of five samples of unidentified macerated plant material were selected for radiocarbon dating analysis in order to construct a temporal framework for deposition in the study area (Table 2.3; see Figs. 2.6 and 2.7). Samples were processed in the Keck Carbon Cycle AMS Facility at UC Irvine via sample preparation at the University of Alberta by Dr. Duane Froese. Samples were obtained from carbonaceous detritus rich horizons in cores S3, S5 and S8.

Table 2.3.  Radiocarbon data results. Calendar ages were calibrated to two standard deviations (2σ) using IntCal 13 calibration software (Reimer et al. 2013).

<table>
<thead>
<tr>
<th>UCIAMS #</th>
<th>Core</th>
<th>Depth (m)</th>
<th>Depth relative to sea level (m)</th>
<th>δ¹³C (‰ ± 0.1)</th>
<th>¹⁴C age (BP)</th>
<th>±</th>
<th>Calendar age range (cal BP) (2σ probability)</th>
</tr>
</thead>
<tbody>
<tr>
<td>152358</td>
<td>S3</td>
<td>11.6</td>
<td>- 5.7</td>
<td>-28.7</td>
<td>2980</td>
<td>15</td>
<td>3075 – 3210</td>
</tr>
<tr>
<td>152354</td>
<td>S5</td>
<td>7.0</td>
<td>- 6.4</td>
<td>-28.3</td>
<td>6355</td>
<td>15</td>
<td>7260 – 7315</td>
</tr>
<tr>
<td>152355</td>
<td>S8</td>
<td>5.4</td>
<td>- 0.8</td>
<td>-25.9</td>
<td>6020</td>
<td>20</td>
<td>6795 – 6930</td>
</tr>
<tr>
<td>152357</td>
<td>S8</td>
<td>14.2</td>
<td>- 9.6</td>
<td>-28.6</td>
<td>7835</td>
<td>20</td>
<td>8560 – 8640</td>
</tr>
<tr>
<td>152356</td>
<td>S8</td>
<td>14.8</td>
<td>- 10.2</td>
<td>-28.6</td>
<td>7805</td>
<td>20</td>
<td>8545 – 8625</td>
</tr>
</tbody>
</table>

**S3 (UCIAMS-152358)**

Results: The radiocarbon sample obtained from the base of S3 (2980 ± 15 ¹⁴C BP 5.7 m below masl; see Fig. 2.7) is > 3 k ¹⁴C BP younger than the four samples obtained from the S5 and S8 cores. In S9, < 300 m from S3, the Mazama ash deposit
(6730 ± 40 $^{14}$C BP; Hallet et al. 1997) is located 2.7 m below masl. This proves that the sediments in S3 were deposited subsequent to sediment in S9 and S8 (both of which contain Mazama ash deposits).

Interpretation: S3 was drilled through a point-bar deposit formed during the westward lateral migration (accretion) of a Pitt River meander. The river cut across the floodplain (preserved in S9) as the channel migrated across the PRV through S3 ~3 k $^{14}$C BP.

**S5 (UCIAMS-152354)**

Results: Organic matter sampled between sand lenses in S5, ~30 cm above the thick coarse sand zone was dated at 6355 ± 15 $^{14}$C BP (see Fig. 2.7).

Interpretation: The thick, poorly sorted coarse-grained sand deposit at the base of S5 is inferred to be a topographic high that impeded deposition from occurring until it was buried below aggrading floodplain deposits. The paleosol on top of the thick sand (below the radiocarbon sample) suggests that the top of the sand was subaerially exposed prior to 6355 ± 15 $^{14}$C BP.

**S8 (UCIAMS-152355, 152356, and 152357)**

Results: Samples obtained above and below a coarse-grained sand lens at the base of S8 provide inverted radiocarbon dates (relative to burial depth), but have overlapping age ranges (UCIAMS-152357, 7835 ± 20 $^{14}$C BP 9.6 m below masl and UCIAMS-152356, 7805 ± 20 $^{14}$C BP 10.2 m below masl; see Fig. 2.6 or 2.7). The shallowest radiocarbon sample in S8 (UCIAMS-152355, 6020 ± 20 $^{14}$C BP at 0.8 m below masl) was taken roughly 4.3 m above the preserved Mazama ash deposit in S8.

Interpretations: The small (albeit inverted) age gap between samples at the base of S8 (UCIAMS-152357 and 152356) is interpreted as a valid indication that a period of non-deposition and/or significant erosion did not occur after deposition of the sand lens, and hence the sand lens is not depositionally equivalent to the thick sand unit at the base of S5. In fact, the sand in S8 may be a rapidly deposited bed of sediments sourced from the sand bed in S5.
In line with radiocarbon results collected by Locher (2014), the shallowest sample in S8 (UCIAMS-152355), at < 1 m below masl, suggests that sea level had essentially reached masl ~6 k \(^{14}\text{C}\) BP.

The age dates from the radiocarbon samples in S8 bounding the tephra deposit are used to confirm that the ash layer is the Mazama ash deposit (6730 ± 40 \(^{14}\text{C}\) BP; Hallet et al. 1997). Further, the depths of the preserved ash in this study’s cores (S4, S6, S8 and S9) are consistent with the depths of Mazama ash preserved in other cores (between 2-6 m below masl; Clague et al. 1982; Williams and Roberts 1989; Locher 2014). The depth of the Mazama ash relative to the bounding radiocarbon samples in S8 are used to infer that aggradation on the PRV floodplain was more rapid between ~6.7 to 6.0 \(^{14}\text{C}\) BP than ~7.8 to 6.7 \(^{14}\text{C}\) BP (~6 mm y\(^{-1}\) and ~4 mm y\(^{-1}\), respectively).

### 2.3.2. Palynology

Six clay-rich samples were selected for palynological assessment to gauge the salinity of the paleoenvironment (Table 2.4; see Figs. 2.6 and 2.7). Samples were analyzed at Geotop-UQAM’s micropaleontology and marine palynology laboratory. Raw semi-quantitative results recorded by Geotop-UQAM are provided in Appendix B.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (m)</th>
<th>Depth relative to masl (m)</th>
<th>Palynological count</th>
<th>Possible marine or brackish affinity? (√/x)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pollen (%)</td>
<td>Spores (%)</td>
</tr>
<tr>
<td>S3</td>
<td>2.2</td>
<td>+ 3.7</td>
<td>98.2</td>
<td>1.8</td>
</tr>
<tr>
<td>S3</td>
<td>11.6</td>
<td>− 5.7</td>
<td>90.9</td>
<td>7.4</td>
</tr>
<tr>
<td>S4</td>
<td>9.8</td>
<td>− 5.9</td>
<td>94.7</td>
<td>5.3</td>
</tr>
<tr>
<td>S8</td>
<td>5.2</td>
<td>− 0.6</td>
<td>97.7</td>
<td>2.3</td>
</tr>
<tr>
<td>S8</td>
<td>13.6</td>
<td>− 9.0</td>
<td>86.7</td>
<td>13.3</td>
</tr>
<tr>
<td>S9</td>
<td>14.9</td>
<td>− 9.0</td>
<td>82.3</td>
<td>16.9</td>
</tr>
</tbody>
</table>

**Palynological results and interpretations**

The palynological results did not provide a positive indication of marine influence in any of the samples. Analysis of the fine fraction (< 106 µm) revealed that all six samples are extremely rich in terrestrial palynomorphs (Larix, Tsuga, angiosperm and
bisaccate pollen grains, and monolete and trilete spores). Rare freshwater algal remains were recorded at the base of S3 (Botryococcus and Pediastrum). In the S9 sample, one Halodinium specimen was identified in the slide count.

The acritarch genus *Halodinium* was first identified by Bujak (1984) in Pleistocene sediments of the Bering Sea and northern North Pacific. These specimens have an unknown affinity, but have been associated with glacial meltwater plumes and transitional freshwater-to-marine environments, primarily in Arctic settings (de Vernal et al. 1989; Mudie 1992; Head 1993).

Qualitative assessment of the coarse fraction (> 106 μm) of the samples led to the identification of terrestrial vegetal remains and mineral grains, neither of which could provide a positive indication of marine influence at the time of deposition.

*Halodinium* in S9

Palynological samples at the base of S8 and S9 were obtained at roughly the same depth below sea level. It is inferred that these sediments were deposited coeval with one another, so that the horizon used in S9 for palynological analysis was deposited close to 7.8 k ¹⁴C BP (see Fig. 2.7). The presence of *Halodinium* in S9 at this time could be linked to an inundation of brackish water from the Strait of Georgia up the Fraser River during the onset of transgression. Brackish and marine palynomorphs may have been swept up the Pitt River due to tidal draw and deposited on the PRV floodplain next to the Pitt River in an overbank deposit.

2.3.3. Sedimentological descriptions and interpretations of Depositional units

Depositional unit 1 (D1): Poorly sorted, organic-rich clay to medium-grained sandy soil (Soil; Fig. 2.8A)

D1 is present at the top of all cores except S6 and ranges in thickness between 0.2 and 3 m (Figs. 2.6 and 2.7). In S5, an ~25 cm thick interval of D1 is present between 6.6 and 6.9 m below masl between two coarse-grained packages of sand (Fig. 2.7).
Figure 2.8. Selected core intervals representative of depositional units found in PRV floodplain deposits. Samples A-C and E are photographs of preserved core; samples D and D* are resin peels. A: D1, poorly sorted material (soil). B: D2, cohesive mud. C: D3, disturbed silt. D: D4, alternating parallel laminated sand and silty mud. D*: D4*, alternating current ripple laminated sand and silty mud. E: D5, coarse sand.
Vertical roots are common throughout D1, and commonly extend over 10 cm into the underlying sediments. Organic content (i.e., macerated plant material) is typically present and variable in concentration. D1 is mottled and ranges in grain size from clay to medium-grained sand. D1 is rich in organic debris and lacks sedimentary structures, but commonly contains 10-30 cm thick bands of soils that vary in colour, concentration of organic matter, packing, and grain size. Boundaries between distinct soil horizons are sharp.

D1 represents late-Holocene soil and/or anthropogenic fill. Banded soil horizons match cores recovered by Locher (2014), who dated such horizons back to 2.7 k $^{14}$C BP. These soils indicate that the PRV floodplain was primarily subaerial and has not been subjected to significant erosion for ~3 k $^{14}$C BP.

**Depositional unit 2 (D2):** Massive, clay-dominated cohesive mud with common organic-rich horizons (Cohesive mud; Fig. 2.8B)

Grey to beige clay-rich cohesive mud zones are dm-scale in thickness (commonly 20-60 cm), with the exception of an ~2 m thick zone in S4. Sedimentary laminations and animal burrows are not observed in this unit. Randomly oriented organic flecks and organic-rich horizons (1-3 cm thick) hosting horizontal roots (> 5 mm diameter) are common. Thin (< 1 mm, 10-25 cm long) vertical roots may be present. D2 commonly grades over 2-5 cm into D3 and otherwise shares sharp contacts with D3 and D4. Mazama ash deposits are preserved within D2 in cores S4 and S6.

D2 is interpreted as fine-grained sediment that accumulated in a submerged tidal freshwater marsh environment. The preservation of the Mazama ash in S4 and S6 and presence of roots (vertical and horizontal) indicate that this unit was deposited in a low-energy environment in which the Mazama ash was not eroded away and plants would have been able to sustain growth. However, the marsh water would have experienced tidal flow, inhibiting formation of a peatland.

**Depositional unit 3 (D3):** Silty mud (or silty vfL sand) with disturbed clay or sand lenses (Disturbed silt; Fig. 2.8C)
Soft-sediment deformation (SSD) varies from mm- to cm-scale (< 2mm to > 80 mm) and extends vertically and/or horizontally. SSD lenses and beds are (internally) lithologically homogeneous and very rarely have an observable lining. Identifiable ichnological structures include Planolites (≤ 12 mm diameter), and rare Palaeophycus, Teichichnus, and Skolithos. Rare wavy parallel lamination is generally disturbed or obstructed by cm-scale SSD lenses. These laminated bedsets range in thickness from 5 cm to < 20 cm. Sideritized nodules (2-5 cm thick) are sporadically located throughout disturbed silt zones. Mazama ash deposits are preserved in D3 in cores S8 and S9.

D3 commonly grades into D4. Contacts with D2 may be gradational (< 10 cm) or sharp. Within D3, the dominant grain size may shift gradually over ≤ 20 cm or abruptly across a sharp contact.

D3 is interpreted as stacked backswamp deposits that accumulated episodically during freshet flooding of the Fraser lowland. During floods, suspended sediment sourced from the Pitt River and Fraser River would have rapidly spilled out onto the PRV floodplain. The disturbance in D3 is primarily interpreted as a result of soft-sediment deformation that occurred while drilling. Deformation caused by drilling likely overprints deformation by dewatering and bioturbation that occurred following the rapid deposition of silty overbank sediment. During base flow, D3 sediments may have remained submerged under tidal marshland water before being subaerially exposed as flood water evaporated or infiltrated the sediment; this may have enabled the formation of sideritized horizons. The Mazama ash hosted in D3 in two cores would have been preserved in subaqueous conditions between flooding events.

**Depositional unit 4 (D4):** Dm- to m-scale zones of alternating vf sand and silty mud bedsets (Alternating vf sand + mud; Fig. 2.8D)

D4 deposits are typically > 50 cm to > 2 m thick. Silty mud bedsets are commonly ≤ 5 cm thick and locally clean upward over < 3 cm into sandy bedsets. Sandy bedsets are dominated by vfL sand and may extend over 10 cm without mud laminae. Successive sandy bedsets may increase or decrease in relative thickness and mud content. Sand-dominated bedsets are up to 30 cm thick.
Organic-rich laminae are uncommon and are limited to 5-15 cm thick zones, mainly within sandy bedsets. Twigs (< 4 mm diameter) may be present in these organic-rich zones. Lamination in these bedsets is dominated by (wavy) parallel lamination. Rare current ripple lamination is observed in the sandy bedsets. Traces are rare (Bl: 0-1), and include Planolites (≤ 2 mm diameter) and mud-lined, sand-filled Palaeophycus (< 8 mm diameter). This depositional unit typically has sharp, undulatory contacts with D3.

D4 packages are interpreted as seasonal packages of floodplain fines. Upon freshet flooding of the Fraser lowland, suspended silt and vFL sand would have been deposited onto the submerged PRV marshland, similar to D3, except D4 deposits may reflect deposition closer to point sources and in relative depositional lows.

**Depositional unit 4* (D4*):** Current ripple laminated, alternating vF sand and silty mud bedsets (Alternating vF sand + mud; Fig. 2.8D*)

A subset of the alternating parallel laminated sand and mud unit, D4* deviates from D4 in that its sandy bedsets are predominantly current rippled (with local bidirectional current rippled bedsets). D4 and D4* otherwise share similar sedimentological (e.g., bedset thicknesses, organic content) and ichnological characteristics.

D4* is observed throughout S3 and at the base of S7, and is interpreted to represent lateral accretion of a fluvial channel. In S3, D4* represents point-bar deposition and reflects the lateral (westward) migration of the Pitt River (Chapter 3). A series of crevasse splay deposits or a migrating tributary running across the PRV floodplain from the Coast Mountain foothills could be the source of this current ripple-dominated zone at the base of S7.

**Depositional unit 5 (D5):** Massive, poorly sorted vc- to fU-dominated sand (Coarse sand; Fig. 2.8E)

D5 beds of fU to mU sand are observed in S4, S5, and S8. These beds are 5-20 cm thick and have sharp contacts. D5 is also observed as a zone of coarse-grained sand extending vertically over 4 m at the base of S5. D5 lacks mud or organic content. At the base of S5, D5 fines upwards from a very coarse-grained sand with clasts.
(granules to medium pebbles) to medium-grained sand devoid of clasts. Grains are sub-angular and clasts are sub-rounded. The contact between the top of the coarse zone in S5 and the paleosol (D1) between 6 and 7 m below masl is not preserved, but a cobble was collected at the top of D5.

The S5 core lies within Armstrong and Hicock's (1980) mountain stream deltaic and channel-fill deposit, with sediments sourced from Jacobs Lake in foothills of the Coast Mountains (Fig 2.3). The D5 zone at the base of S5 is interpreted as a post-glacial fan deposit that formed upon the release of glacial meltwater from the foothills. This deposit would have formed a topographically high mound, sheltered from erosion by the foothills of the Coast Mountains. The paleosol on top of this coarse-grained sand validates this interpretation – the soil had to have formed in a subaerial environment above the floodplain. D5 lenses in S4, S5 and S8 likely represent pulses of coarse sediment shed onto the PRV from the foothills during glacial meltwater flood events.

**Summary of depositional units**

D1-D5 units are interpreted to have been deposited in a tidal freshwater marsh that was seasonally inundated by freshet floodwater. During the freshet, suspended sediment carried in the Fraser River would have settled out of suspension and deposited over the PRV floodplain as the Fraser (and Pitt) rivers spilled over their banks and onto the greater Fraser lowland, forming stacked seasonal floodplain fines. During base flow, most of the PRV floodplain would have remained submerged until around 6 k $^{14}$C BP (minus highlands, such as at S5, see Section 2.2.1).

D3 and D4 volumetrically dominate the cores (except S3 and S5) and represent regional floodplain (alluvial) deposition consistent with the cores described by Armstrong (1981), Clague et al. (1982), Williams and Roberts (1989) and Locher (2014). It is anticipated that without anthropogenic interference, a layer of D1 formed and is preserved across the PRV floodplain. Intermittent zones of D2 are present in these regional deposits, but are not correlatable because they likely accumulated in isolated topographic lows. D4* in S3 records the lateral migration of Pitt River. D5, at the base of S5, represents an alluvial fan sourced from the Coast Mountain foothills.
2.4. Discussion

The lake-river morphology developed in the PRV was established before the initiation of mid-Holocene sea level rise 8 k $^{14}$C BP. The PRV’s marine to freshwater transition was not captured in our cores, as the vibracores in this study were not drilled deep enough into the subsurface. The sediments preserved in all of the vibracores reflect deposition that occurred in a completely freshwater environment.

2.4.1. Paleogeographic reconstructions of the PRV floodplain

The results from this study have been integrated with the findings of Armstrong and Hicock (1980), Clague et al. (1982, 1983), Katzie Development Corporation Archaeology (2010) and Locher (2014) to produce paleogeographic reconstructions of the PRV floodplain from the initiation of mid-Holocene sea level rise to present. Time slices for 8, 6 and 3 k $^{14}$C BP are presented.

Onset of sea level rise in the mid-Holocene (8-6 k $^{14}$C BP)

A tidally influenced, submerged (m-scale water depth) tidal freshwater marshland likely existed across the PRV floodplain approximately 8 k $^{14}$C BP (Fig. 2.9). Upon transgression (initiated ~8 k $^{14}$C BP), sediments carried in the Fraser River were regularly deposited on the Fraser lowland during annual floods (Clague et al. 1982; Locher 2014; this study). With the onset of this sea level rise, the PRV floodplain and the rest of the Fraser lowland began to aggrade rapidly (~5 mm y$^{-1}$; Clague et al. 1982; this study).

Radiocarbon ages from Locher (2014) and this study indicate that the southern limit of Pitt Lake was north of S2 (and Locher’s transect) prior to 7.8 k $^{14}$C BP (~8 k $^{14}$C BP; Tables 2.1 and 2.3). Based on shared similarities between S1 and S2 as well as Locher’s (2014) cores, it is inferred that Pitt Lake’s southern limit was north of S1 around 8 k $^{14}$C BP. Correspondingly, the maximum rate of northward progradation of the PRV floodplain/southern limit of Pitt Lake is restricted to < 2 km over 8 ky, representing < 0.25 m y$^{-1}$.
Figure 2.9. Paleogeographic reconstruction of the PRV floodplain around 8 k $^{14}$C BP. The present PRV floodplain was submerged under m-scale depths of freshwater at the onset of the sea level rise 8 k $^{14}$C BP. Cores collected from Locher (2014) indicate that the southern margin of Pitt Lake was already north of S2 by 8 k $^{14}$C BP, and probably north of S1. The projected limit of constant subaerial exposure is delineated.
Rapid aggradation of the PRV floodplain (6-3 k $^{14}$C BP)

Radiocarbon ages from this study and Locher (2014) demonstrate that deposition of sediment ~6 k $^{14}$C BP occurred around masl (< 1.5 m below masl to > 2 m above masl), indicating that the marshy floodplain had variable m-scale topography and that sea level had essentially reached the present-day position 6 k $^{14}$C BP (Clague et al. 1983; Williams and Roberts 1989). Areas of higher topography on the PRV floodplain, specifically in the south (around S5) would have remained subaerial, permitting the initiation of soil development (Fig. 2.10). The rest of the PRV floodplain would have remained (intermittently) shallowly submerged, and possibly modulated by tides.

By 5 k $^{14}$C BP, evidence of human occupation south of this project’s cored locations indicate that the southern end of the floodplain was subaerial and sufficiently stable for continued occupation (Katzie Development Corporation Archaeology 2010; Figs. 2.3 and 2.4).

Stabilization of the PRV floodplain (< 3 k $^{14}$C BP)

Locher (2014) recorded up to 0.9 m of preserved organic soil at the top of two cores, which varied in age from 2.7-2.1 k $^{14}$C BP (Table 2.1). Preservation of late-Holocene soil indicates that the PRV floodplain has been subaerial and has avoided significant erosion for thousands of years (Fig. 2.11). With comparison to Locher’s (2014) cores, the cores drilled across the PRV in this study suggest the floodplain’s elevation and morphology has not changed significantly since 3 k $^{14}$C BP, which has barred lateral migration of the Pitt River (Chapter 3).
Figure 2.10. Paleogeographic reconstruction of the PRV marshy floodplain around 6 k \(^{14}\)C BP. At 6 k \(^{14}\)C BP, parts of the PRV floodplain reached masl (Locher 2014; this study). The floodplain was mostly waterlogged and submerged in shallow water. The areal extent of freshet flooding across the PRV floodplain at 6 k \(^{14}\)C BP would have been more restricted than 8 k \(^{14}\)C BP. The projected limit of constant subaerial exposure is delineated.
The PRV floodplain is much the way it is today, excluding lateral migration of the Pitt River (Chapter 3). The PRV floodplain would have been primarily subaerial, except during extreme floods.

Figure 2.11. Paleogeographic reconstruction of the PRV floodplain around 3 k $^{14}$C BP. The PRV floodplain is much the way it is today, excluding lateral migration of the Pitt River (Chapter 3). The PRV floodplain would have been primarily subaerial, except during extreme floods.
Chapter 3.

Quantification of lateral migration in the tide-dominated Pitt River, Canada: Implications for recognizing tide-influenced meandering river deposits in the rock record

The mechanisms that drive meandering in rivers have not been isolated, but it is understood that meander migration rates reflect a multitude of interdependent factors (Knighton 1998). To understand and predict meander behaviour, data have been collected in the laboratory (cf. Bagnold 1960) and from modern meandering systems (e.g., Hickin and Nanson 1975; Hooke 1980; Hudson and Kesel 2000). These studies provide lateral migration rates (via bank erosion) over time periods spanning months to hundreds of years, generally through recurring field observations or assessment of historical records (i.e., maps and aerial photographs; see Lawler 1993). Multiple review articles (Hooke 1980; Lawler 1993; Van De Wiel 2003) provide a substantial database of lateral migration rates of meanders and corresponding river parameters (e.g., drainage area, mean discharge). Sedimentological and geochronological evidence from a laterally accreting point bar along the tide-dominated Pitt River, British Columbia, Canada has been used to calculate a long-term meander migration rate via bank accretion. Few studies present measurements of lateral migration by way of bank accretion, providing an inadequate basis for comparison (Nanson and Beach 1977; Hupp 1992). Moreover, only one paper provides meander rates in a tide-influenced river reach (the lower Mississippi River; Hudson and Kesel 2000).

3.1. Introduction

The calculated meander migration rate from the Pitt River bar evaluated in this study is compared to published rates of lateral migration via bank erosion from purely fluvial meandering rivers, accepting that short term (e.g., decade-scale) observations of bank erosion can outpace the rate of bank accretion by an order of magnitude (Hupp
Lateral migration in the Pitt River is compared to lateral migration of meandering bars in the lower Mississippi (Hudson and Kesel 2000), and in rivers across western Canada (Hickin and Nanson 1984; Nicoll and Hickin 2010). The intent of this comparison is to assess the rate of meandering in tide-influenced channels relative to purely fluvial channels as a means of enhancing the accuracy of predicting the accretion rate of tide-influenced channel bars in the rock record.

3.1.1. Meander migration

Meandering behaviour reflects the interaction between erosive and resistive forces acting on a floodplain (Nicoll and Hickin 2010). Erosive force is controlled by stream power or flow energy, which is the product of discharge and gradient (Hickin and Nanson 1984; Nanson and Hickin 1986; Knighton 1998; Richard et al. 2005). Using statistical analyses of meandering systems in western Canada, Nanson and Hickin (1986) and Nicoll and Hickin (2010) found that stream power had the greatest influence on meander migration rates. Resistive forces, which act against stream power and meander migration, are linked to riparian vegetation (Smith 1976; Burckhardt and Todd 1998), sediment calibre at the channel bed (Nanson and Hickin 1986), and bank and channel morphology (Hooke 1980). Channel morphology includes bank height (Hickin and Nanson 1984), gradient, and bend geometry (e.g., Mathes 1941; Bagnold 1960). Cohesive sediment (i.e., clay plugs) and sedimentological heterogeneities in the floodplain also impede the rate at which meanders are able to migrate (Hudson and Kesel 2000) or may force downstream accretion and counter point-bar formation (Smith et al. 2009; Fustic et al. 2012).

The influence of tides on meander migration in fluvial channels has not been directly addressed in the literature. Tides may contribute as a resistive force against meander migration, as tides reverse and/or intermittently dampen flow, inherently reducing stream power by temporarily reversing or retarding downstream discharge. Moreover, the relatively high proportion of fine-grained (cohesive) sediment linked to: 1) deposition in the tide-influenced turbidity maximum zone (La Croix and Dashtgard 2014); 2) changes in bend geometry (Dalrymple and Choi 2007) in response to high-frequency fluctuations in discharge and increased mud content; and 3) the shift in vegetation type on the coastal floodplain (Hupp 1986), all could serve to further resist meander migration.
in the tidal reach of a meandering river system. Herein, the term ‘tidal reach’ is used to describe the downstream length of a fluvial channel where tides affect flow and sedimentation, and includes the backwater, tidal backwater and brackish zones of the channel at the seaward terminus of a river.

### 3.1.2. Migration and bend curvature

While relations between planform geometry and meander migration rate are apparent, the ways in which bend parameters influence meander migration rates are not fully understood (e.g., Bagnold 1960; Leopold and Wolman 1960; Hickin 1974, 1978; Knighton 1998). Emphasis has been placed on the relationship between meander migration (\(M; \text{m y}^{-1}\)) and bend curvature (\(r/w\)). Bend curvature (\(r/w\)) is the ratio of a meander bend’s radius (\(r; \text{in metres (m)}\)) and mean channel width (\(w; \text{in metres (m)}\)), and has been identified as a factor that influences meander migration (Fig. 3.1; Leopold and Wolman 1960). Early parameterization of meandering systems showed that meander bend curvature most commonly falls between 1.4 and 4.3 (Leopold and Wolman 1960). Hickin (1974) suggested that a range of \(r/w\) between 2 and 3 reflects equilibrium in natural channels, with meanders gradually tightening towards an \(r/w\) of 3 and then rapidly shifting towards an \(r/w\) of 2. This acceleration in migration (bank erosion) in tighter meanders is attributed to the initiation of flow separation at the outer (concave) bank once bend curvature falls below \(r/w = 3\) (Hickin 1978; Hickin and Nanson 1984; Hooke 1997, 2003).

Preliminary field observations of meander migration rates relative to bend curvature show that migration rates indeed peak from \(2 < r/w < 3\), with a normal distribution of migration rates on either side of this optimal bend curvature window (Fig. 3.2; Hickin and Nanson 1975). It should be noted, however, that the scope of Hickin and Nanson’s (1975) study was on a freely meandering river with a silty sand to gravel bed in a relatively homogeneous, unconsolidated floodplain lacking in mud content and vegetation. Such homogeneous river systems are rare, and subsequent comprehensive datasets exhibit greater complexity in the relationship between migration rate and bend curvature (e.g., Hickin and Nanson 1984; Hooke 1997; Hudson and Kesel 2000; Nicoll and Hickin 2010).
3.1.3. Tidal indicators in sedimentary deposits

Flow conditions downstream of a meandering river’s tidal limit are constantly fluctuating (Dalrymple and Choi 2007), and there is temporal and spatial variability in suspended sediment content (concomitant with migration of the turbidity maximum), salinity, and flow velocity (from unidirectional flow modulated by tidal forces to flow...
reversals). Sediments deposited in the tidal reach of rivers exhibit these dynamic conditions. Evidence of tidal influence on deposition from (1) the Fraser River, Canada, and (2) ancient fluvial-tidal systems, are summarized below.

(1) In the tidal reach of the Fraser River, the thickness, uniformity, frequency, and lateral continuity of mud beds increase downstream from the freshwater, fluvially dominated tidal backwater zone towards the brackish, admixed tidal-fluvial transition (Dashtgard et al. 2012; Sisulak and Dashtgard 2012; Johnson and Dashtgard 2014; La Croix and Dashtgard 2014). Heterolithic bedding (i.e., flaser, wavy, and lenticular) is absent in the tidal backwater zone, but lenticular bedding becomes common in the time-averaged turbidity maximum zone, where the channel shifts from being fluvially dominated to fluvially influenced (La Croix and Dashtgard 2015). Discrete, alternating sand and mud bedsets are persistent across the mixed tidal-fluvial transition and tide-dominated stretch of the Fraser River, but are absent in the tide-influenced reach of the channel. These sand and mud couplets are principally attributed to seasonal fluctuations in discharge (Sisulak and Dashtgard 2012). However, increasing rhythmicity and uniformity of these couplets have been linked to increasing tidal influence in the seaward direction (Johnson and Dashtgard 2014). Trace fossils in the tidal backwater zone are diminutive and patchy in distribution (BI 0-1), and show very low diversity (1-2 forms; La Croix and Dashtgard 2015). Bioturbation intensity, diversity and uniformity increase into the time-averaged turbidity maximum zone.

The Fraser River’s tidal backwater limit is marked by a decrease in the grain size of bedload sediments in the seaward direction (Venditti et al. 2010; Dashtgard et al. 2012), and the presence of marine dinoflagellate cysts (dinocysts) throughout the tidal backwater zone (Czarnecki et al. 2014). The gravel to sand transition occurs over a few kilometres up to 102 river km upstream from the mouth of the Fraser River, and coincides with the landward maximum of effective tidal influence on fluvial discharge during base flow (Venditti et al. 2010; Dashtgard et al. 2012). The shift in sediment calibre at the bed, in part, reflects the decrease in the carrying capacity of the river due to tidal dampening of fluvial flow in the tidal backwater zone (Dashtgard et al. 2012; Nittouer et al. 2012). Czarnecki et al.’s (2014) palynological analysis of sediments across the tidal reach of the Fraser River indicates that marine dinocysts comprise < 1% to 0% of the palynological assemblage in sediments across the (freshwater) tidal
backwater (with decreasing abundance in the landward direction). The upstream transport of these marine micro-organisms through the tidal backwater has been primarily attributed to either: 1) flow separation on channel margins as tides raise the water levels and/or 2) eolian transport by onshore-directed winds (Czarnecki et al. 2014).

(2) Analyses of ancient fluvial systems suggest multiple non-unique sedimentological indicators observed in conjunction with one another can be used to interpret deposition that occurred across the tidal reach of a river, primarily in the brackish, admixed tidal-fluvial transition (e.g., Ghosh et al. 2005; Plink-Björklund 2005; van den Berg et al. 2007). These indicators include bidirectional cross-strata (herringbone cross stratification), cyclical alternation of relatively fine- and coarse-grained bedsets (tidal bundles; e.g., Visser 1980; De Boer 1989), relatively high proportions of mud drapes and reactivation surfaces (de Mowbray and Visser 1984; van den Berg et al. 2007), fluctuations in organic content (Martinius and Gowland 2011), irregular foresets and foreset climbing ripples (van den Berg et al. 2007; Choi 2010; Martinius and Gowland 2011), shovel-toe foresets to bottomsets (van den Berg et al. 2007), and fluctuations in topset (brinkpoint) and bottomset thicknesses (van den Berg et al. 2007; Martinius and Gowland 2011). These features have been interpreted as lines of evidence for high frequency fluctuations in flow conditions across the tide-influenced and tide-dominated extents of meandering rivers.

The ichnological signature of sediments across a river’s tidal reach should vary, reflecting the shift from wholly freshwater, fluvial channels to a brackish-water, tide-dominated setting at the seaward terminus of the channel. Terrestrial assemblages, principally associated with the Scyenia Ichnofacies, are expected to extend from the purely fluvial reach of the river into the tidal backwater zone (Buatois and Mángano 2004; Buatois et al. 2005; MacEachern et al. 2005), although this has not been reported from the rock record and is not supported from modern observations (La Croix et al. in press). Through the turbidity maximum zone, trace diversity and density remain low even though saline water is present at the bed. This is because of high turbidity levels and high sedimentation rates (e.g., Howard et al. 1975; Buatois et al. 2005). Seaward of the turbidity maximum zone, an increase in the persistence of saline water at the bed and the influx of food and larvae from the marine environment are reflected in a trace
assemblage dominated by simple, mainly vertical traces that increase in size in the seaward direction (Pemberton et al. 1982; Pemberton and Wightman 1992; Buatois et al. 1997; Gingras et al. 1999). Trace diversity, bioturbation intensity (and uniformity), and trace size all increase with increased persistence of brackish-water conditions, rising salinity towards normal marine (~35 psu) of resident water, decreased turbidity in the water column (Pemberton and Wightman 1992), and reduced sedimentation rate.

3.2. Study area

The Pitt River, British Columbia is a tide-dominated freshwater tributary that debouches into the tidal backwater of the Fraser River, approximately 10 km upstream of the landward limit of the Fraser River’s saltwater wedge (Ashley 1977). Due to the reservoir capacity of Pitt Lake, bidirectional tidal flow is induced in the Pitt River as water levels fluctuate in the tidal backwater of the Fraser River.

The lower Pitt River is 20.1 km long and is situated in the 19.7 km long Pitt River Valley (PRV); consequently, the sinuosity of the lower Pitt River is 1.2 (Table 3.1; Ashley 1977). Tidal range in the Pitt River tributary varies from 2 m at the Pitt-Fraser confluence to 1.2 m in Pitt Lake (Johnston 1922; Ashley 1977; Milliman 1980).

Table 3.1. Pitt River parameters. Data obtained from Ashley (1977).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Watershed/drainage area (km²)</td>
<td>1640</td>
</tr>
<tr>
<td>Floodplain width (km)</td>
<td>4.5-5</td>
</tr>
<tr>
<td>Floodplain length (km)</td>
<td>19.7</td>
</tr>
<tr>
<td>River length (km)</td>
<td>20.1</td>
</tr>
<tr>
<td>Sinuosity</td>
<td>1.2</td>
</tr>
<tr>
<td>Avg. grain size at bed</td>
<td>fine upper (fU) to medium upper (mU) sand</td>
</tr>
<tr>
<td>Avg. grain size at banks</td>
<td>silt to very fine (vf) sand</td>
</tr>
<tr>
<td>Max. tidal range at Pitt-Fraser confluence (m)</td>
<td>2</td>
</tr>
<tr>
<td>Max. tidal range in Pitt Lake (m)</td>
<td>1.2</td>
</tr>
<tr>
<td>Avg. channel width (m)</td>
<td>610</td>
</tr>
<tr>
<td>Avg. thalweg depth (m)</td>
<td>12.1</td>
</tr>
<tr>
<td>Width to depth ratio (w/d)</td>
<td>50</td>
</tr>
<tr>
<td>Anthropogenic influence</td>
<td>Dyking from 1890s onward (Collins 1975); current dyking network constructed in the 1950s</td>
</tr>
</tbody>
</table>
Flow in the Pitt River is bidirectional throughout most of the year. Year-round, ebb tides are longer in duration than flood tides, but throughout most of the year (during base flow), peak flood velocities exceed peak ebb velocities. Ebb-current velocities exceed flood velocities during the snowmelt induced freshet from April to August, and high fluvial discharge prevents or severely hampers flood-oriented flow. Peak flood and ebb flow discharge at the end of the freshet reaches 1800 m$^3$ s$^{-1}$ and 950 m$^3$ s$^{-1}$, respectively (Ashley 1977). From September to April (base flow), peak flood flow reaches approximately 2400 m$^3$ s$^{-1}$ and peak ebb flow reaches 2080 m$^3$ s$^{-1}$.

During base flow, only 5% of discharge flowing through the Pitt River is attributed to basin drainage – the remaining 95% of flow is sourced from the Fraser River (Ashley 1977). Flood tidal dominance in the Pitt River is reflected by flood-oriented bedforms and the net northward transport of sediment from the Fraser River into Pitt Lake, where it forms the Pitt River Delta (Ashley 1977).

### 3.3. Methods

Two vibracores were drilled proximal to the Pitt River channel: S3 and S9 (Figs. 2.5 and 3.3). These cores were drilled 270 m apart, approximately 450 m and 715 m from the Pitt River’s low water line, respectively. Regional air photos, taken in 1940 and 1949, reveal a meander scar, from which point the Pitt River has since migrated to the northwest (Fig. 3.3B). The meander scar is still observable, though muted, in recent satellite images of the area (see Fig. 3.3A).

Vibracore S3 was drilled between the meander scar and the modern channel into an anticipated point-bar deposit. Vibracore S9 was drilled southwest of the observed meander scar to confirm that S3 was not representative of regional floodplain deposition (refer to Section 2.3.3 for sedimentological interpretations of the regional cores). Geochronological and palynological samples were taken from cores across the PRV floodplain, including intervals from S3 and S9 (see Sections 2.3.1 and 2.3.2).
Figure 3.3. A) Vibracore S3 was drilled approximately 450 m from the low water line of the Pitt River. S9 was drilled 715 m from the low water line on the other side of the meander scar (Google Earth image, 2014). (B) The meander scar is obvious in historical photos taken prior to the construction of the current dyke network, which enabled cultivation of farmland on the floodplain (GeoBC aerial photo 1949; BC 779:84). See Fig. 1.4 for the location of these cores relative to the other vibracores taken on the PRV floodplain.
3.4. Results

3.4.1. Pitt River sedimentology, ichnology, geochronology & palynology

Sedimentology & ichnology

The S3 core is dominated by dm- to m-scale, rhythmically alternating current ripple laminated, very fine-grained sand bedsets and silty mud bedsets (see Section 2.3.3). Sandy bedsets are dominated by current rippled vfl sand, with common mud drapes and bidirectional current ripples. Sandy bedsets are up to 30 cm thick and can extend vertically over 10 cm without mud laminae. Silty mud bedsets are parallel laminated and typically ≤ 5 cm thick. Sand and mud bedsets have discrete contacts, but in some instances the silty mud bedsets grade upwards into sandy bedsets over a 1–3 cm interval. Uncommon organic-rich beds (5-15 cm in thickness) are present, mainly within sandy bedsets. Twigs (< 4 mm diameter) may be present in organic-rich zones. Trace fossils are rare (BI: 0-1), and include Planolites (≤ 2 mm diameter) and mud-lined sand-filled Palaeophycus (< 8 mm diameter). The S3 succession is interpreted as a point-bar succession.

The S9 core and all other cores drilled on the PRV floodplain contain dm-scale zones of cohesive mud, silt highly disturbed by soft-sediment deformation, and dm- to m-scale alternating parallel laminated very fine-grained sand and silty mud bedsets (see Section 2.3.3). These depositional units represent overbank deposition on the PRV floodplain (see Chapter 2 for detailed depositional unit descriptions).

Geochronology

Samples were taken from cores across the PRV floodplain to spatially and temporally confine the evolution of the floodplain through the mid- to late-Holocene. Samples were processed at the Keck Carbon Cycle AMS Facility at UC Irvine and sample preparation was done by Dr. Duane Froese at the University of Alberta. One sample of organic material was taken at the base of S3 from organics approximately 5.7
Palynology

Palynological analyses of clay-rich zones in S3 were completed to delineate the brackish to freshwater transition in the Pitt River system. Samples were analyzed at L'Université du Québec à Montréal’s (UQAM) Geotop micropaleontology and marine palynology laboratory. Palynological assessment of the fine fraction (< 106 µm) taken near the top and base of S3 indicates that the Pitt River has remained entirely freshwater for over 3 ky (Table 3.2).

Table 3.2. Palynological data. Rare freshwater algae are recorded at the base of S3 (Botryococcus and Pediastrum).

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (m)</th>
<th>Depth relative to sea level (m)</th>
<th>Palynological count</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pollen (%)</td>
</tr>
<tr>
<td>S3</td>
<td>2.2</td>
<td>+3.7</td>
<td>98.2</td>
</tr>
<tr>
<td>S3</td>
<td>11.6</td>
<td>-5.7</td>
<td>90.9</td>
</tr>
</tbody>
</table>

3.4.2. Quantified lateral migration rates in the Pitt River

The rate of lateral migration in the Pitt River bar was calculated using the $^{14}$C date in S3 collected 5.7 m below masl (Fig. 3.4). The lower limit of lateral migration (via accretion) is estimated by measuring the shortest distance from S3 to the nearest 5 m depth contour from Ashley’s (1977) channel-bed profile (distance: ~510 m; Fig. 3.5). This measurement is conservative, as the 5 m depth contour in the Pitt River only extends to 3 m below masl, which is 1.7 m higher than the depth of the $^{14}$C-dated sample in S3 (Fig. 3.4). Accretion of five hundred and ten metres over 3210 years (the longest probable time period) provides a mean lower limit for the lateral migration rate of 0.16 m y$^{-1}$.

The upper limit of calculated distance for lateral migration is approximately 870 m (Fig. 3.5). The channel has steeper banks on the distal side of the mid-channel bar, so the equivalent depth of the $^{14}$C sample (5.7 m below masl) should be proximate to the 5 m depth contour (3 m below masl). The upper limit of the rate of lateral migration for the Pitt River is therefore 0.28 m y$^{-1}$, calculated by dividing 860 m by 3075 years (the shortest probable time period).
Figure 3.4. The $^{14}$C sample (oldest probable age: 3210 cal BP) at the base of S3 is ~1.7 m below Ashley’s (1977) 5 m depth line, and represents a conservative estimate for the rate at which the Pitt River channel has laterally accreted.

Figure 3.5. The lower limit for lateral migration of the Pitt River bar is measured from S3 to the nearest 5 m depth line (510 m) delineated by Ashley (1977) and the upper limit for lateral migration is measured to the 5 m depth line on the distal side of the mid-channel bar (860 m). The thalweg in this stretch of the Pitt River is roughly 12 m in depth.
Uncertainties in the estimated rates of lateral migration include: (1) shifting of the 5 m depth contours and Pitt River thalweg since the 1970s, (2) development of the mid channel bar and interim evolution of the Pitt River channel since \( \sim 3 \text{ k}^{14} \text{C BP} \), and (3) possible reservoir effect and lack of precision related to radiometric dating of one unidentified plant sample at the base of S3.

(1) By using channel profile lines collected in the 1970s (Ashley 1977), the accuracy of the calculated lateral migration rates is reduced. Greater accuracy could have been achieved by collecting a higher-resolution channel profile so that an accurate distance could be measured from the same relative depth below mean sea level. Also, the channel profile has changed over the past 40 years and it is expected that the channel has continued to migrate during this time. However, since the dyke network along the Pitt River was constructed prior to the channel profiling, it is assumed that lateral accretion in the Pitt River since 1970s is negligible.

(2) Uncertainty exists in the evolution of the Pitt River since the deposition of channel-margin sediment at the base of S3. It is assumed that the Pitt River meander has continuously migrated towards the northwest since deposition of sediment preserved in S3. However, it is possible the Pitt River meander has episodically shifted its direction of migration or began migrating back towards the southeast after migrating to the northwest, in part developing the mid channel bar. Not accounting for these kinds of events could result in largely underestimating the rate at which the channel has migrated.

(3) Greater accuracy and precision in the rate of lateral accretion could be garnered by radiometrically dating more organic material at the base of S3 and ensuring that the organics came from terrestrial source. By selecting verified terrestrial detritus, as opposed to unidentified organic material, we could eliminate the possibility that the material dated at the base of S3 was aquatic in origin, which could be subject to a marine reservoir effect. If aquatic material was tested in S3, the apparent age of the \( ^{14} \text{C} \) date could be hundreds of years (older) than its true age. In this case, the rate of lateral migration in the Pitt River could be overestimated by around a decimeter per year.

Although uncertainties do exist, the values calculated are considered reasonable for the Pitt River bar. This is because the meander rates calculated bracket maximum
and minimum meander rates and the organic material appears to be terrestrial in origin. The greatest uncertainty relates to changes in the direction of lateral accretion, and if the direction of accretion changed, then the accretion rates calculated herein will underestimate the true rate of accretion.

3.4.3. Published rates of lateral migration in meandering rivers

Van De Wiel (2003) compiled meander migration rates (via bank erosion) from previous studies (including Hooke 1980 and Lawler 1993), and compared these rates to upstream drainage area, a surrogate for river size (Fig. 3.6). This collated dataset shows that meander migration increases with increasing drainage area, and by extension, river size. A similar trend is shared between lateral migration and mean discharge, an alternate proxy for river size (Fig. 3.7).

Drainage area and mean discharge may not be the most suitable means for comparing meander migration in tide-influenced channels, as water flow is not singly sourced from upstream and discharge is constantly fluctuating – both the direction and volume of discharge fluctuate depend on tidal influence. Even so, lateral migration in the tide-dominated Pitt River resides at the lower end of expected lateral migration for purely fluvial systems of similar size (Figs. 3.6 and 3.7; Van De Wiel, 2003).

Work by Hudson and Kesel (2000) on the lower Mississippi (i.e., seaward of Cairo, Illinois, > 1700 river km upstream from the mouth of the Mississippi) provides the only published record of lateral migration rates in a tide-influenced channel. Hudson and Kesel’s (2000) measurements demonstrate that the rate of meander migration decreases substantially once tidal forces begin to act on flow. The mean meander migration rate shifts from 46 m y\(^{-1}\) in the purely fluvial zone of the lower Mississippi to 32 m y\(^{-1}\) in the backwater zone, where fluvial flow is affected by tides only during periods of high discharge (roughly 600 river kilometres up from Head of Passes; Fig. 3.8; Nittrouer et al. 2012). In the tidal backwater zone, where fluvial flow is dampened by tides year-round, meander migration rates are reduced to an average of 6 m y\(^{-1}\). Downstream of the tidal backwater zone where saltwater incursion and bidirectional flow occurs at the channel bed, meander migration rates diminish to 3-4 m y\(^{-1}\).
Figure 3.6. Van De Wiel’s (2003) dataset of compiled lateral migration rates (via *bank erosion*) shows that meander migration is scaled to river size, wherein drainage area is used as a proxy for river size ($n = 220$). The upper and lower limits of bank accretion measured in the Pitt River are displayed in red triangles (0.28 and 0.16 m y$^{-1}$, respectively). The effective drainage area for the Pitt River is measured upstream from the meander (1407 km$^2$).
Figure 3.7. Meander migration rates via bank erosion (n = 105) compiled by Van De Wiel (pers. comm. 2014) and Nicoll (pers. comm. 2014). In general, meander migration and mean discharge scale to one another. The upper and lower limits of bank accretion measured in the Pitt River are displayed in red triangles (0.28 and 0.16 m y\(^{-1}\), respectively). Mean discharge in the Pitt River (288 m\(^3\) s\(^{-1}\)) was derived from Ashley (1977).
In order to compare Pitt River meander migration to the range of migration rates in the lower Mississippi, bend curvature is utilized (Fig. 3.9). The bend curvature of the Pitt River bar described herein is approximately 6.6 (Table 3.3). In Figure 3.9, migration rate (M; m y⁻¹) has been normalized using channel width (accepting channel width as a proxy for river size), so that the Pitt River can be plotted with values from the lower Mississippi River.

Table 3.3. Pitt River meander bend parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Meander bend radius (r; m)</td>
<td>4000</td>
</tr>
<tr>
<td>Average channel width (w; m)</td>
<td>610 (Ashley 1977)</td>
</tr>
<tr>
<td>Bend curvature (r/w)</td>
<td>6.6</td>
</tr>
<tr>
<td>Lateral migration rate (M; m y⁻¹)</td>
<td>0.16-0.28</td>
</tr>
<tr>
<td>Migration to width (M/w; y⁻¹)</td>
<td>0.00026-0.00046</td>
</tr>
</tbody>
</table>
Figure 3.9. Bend radii ($r; m$), meander migration ($M; m \, y^{-1}$) and channel width ($w; m$) values were collected in the purely fluvial zone (green; $n = 74$), and backwater and tidal backwater zones of the lower Mississippi (blue; $n = 34$; Hudson, pers. comm. 2015). Data was not collected from the reach of the Mississippi where saltwater incurs (red boxes in Fig. 3.8). All meanders within the tidal backwater zone of the Mississippi have migration to width values $< 0.04$. Data from Pitt River are plotted in the red triangle. The bend curvature of the Pitt River meander is approximately 6.6.

Normalized migration to width ($M/w$) and bend curvature ($r/w$) ratios have been used in previous studies to compare meandering fluvial systems (see Hooke 1997). Data from purely fluvial meandering rivers across western Canada were collected and analyzed by Hickin and Nanson (1984) and Nicoll and Hickin (2010), and exhibit a positively skewed distribution of data similar to that of meanders in the lower Mississippi (Fig. 3.10).
Figure 3.10. Migration to width (M/w) and bend curvature (r/w) ratios collected by (A) Hickin and Nanson (1984; n = 226) and (B) Nicoll and Hickin (2010; n = 247). Data from Pitt River are plotted in the red triangles. At approximately M/w = 0.04, a transition occurs between slowly migrating meanders (M/w < 0.04) of all bend curvatures, and tighter meanders (1 < r/w < 4) that can migrate relatively rapidly (M/w > 0.04).
3.5. Discussion

While lateral accretion in tide-influenced meandering rivers may be an order of magnitude slower than the rates of rapidly migrating meanders in purely fluvial settings, significant accumulations of laterally accreted sediment can form in tide-influenced river channels.

3.5.1. Migration and bend curvature

Bend curvature data are commonly displayed using envelop curves fit over the maximum migration (or migration to width) values present within given datasets (e.g., Hickin and Nanson 1984; Hudson and Kesel 2000; Nicoll and Hickin 2010). In this study, however, data points under the envelope curves are considered the curves are disregarded. Migration to width and bend curvature ratios are divided into relatively slow and rapid meander migration. The migration to width and bend curvature data collected by Hickin and Nanson (1984), Hudson and Kesel (2000), and Nicoll and Hickin (2010) possess similar positively skewed distributions with optimal bend curvatures between 1.5 < r/w < 4 (Figs. 3.9 and 3.10). Meanders can migrate relatively rapidly within this optimal bend curvature window, but meanders of all bend curvatures can have low migration to width ratios (and migrate relatively slowly). In the case studies presented by Hickin and Nanson (1984) and Nicoll and Hickin (2010) (Fig. 3.10), the transition between slowly migrating meanders and rapidly migrating meanders occurs at a migration to width (M/w) ratio of approximately 0.04, meaning that lateral migration occurs at a rate equivalent to 4% of the channel’s width every year. In the lower Mississippi, the transition between relatively slow and relatively rapid meander migration occurs at M/w = 0.015. The majority of slow meanders are those within the tide-influenced reach of the channel (i.e., in the backwater and tidal backwater zones; Fig. 3.9). Moreover, all meanders in the tidal backwater zone of the lower Mississippi migrate at less than 4% of the channel’s width per year (M/w < 0.04).

In the Pitt River, lateral migration (0.16-0.28 m y\(^{-1}\)) has occurred at an insignificant rate relative to channel width (0.03% or M/w = 0.0003). In purely fluvial
channels of comparable size to the Pitt River and with meanders that have bend curvatures similar to the Pitt River bar (r/w = 6), lateral migration rates can exceed 7% of the channel’s width annually (Hooke 1997).

The calculated migration to width ratios from the Mississippi and Pitt rivers provide a proxy for estimating lateral accretion rates in other tide-influenced channels (i.e., delta distributaries and estuarine channels). Equations derived from purely fluvial systems (i.e., Hickin and Nanson 1975; Van De Wiel 2003) cannot be applied to these systems, as they can grossly overestimate the rate of channel migration. Instead, it can be predicted that the rate of lateral accretion of ancient tide-influenced river channels occurred at a rate less than M = 0.04 w, and likely less than M = 0.015 w, with the rate of migration decreasing as tidal dominance over flow increases in the river channel.

3.5.2. **Tidal indicators in sedimentary deposits**

The sedimentological characteristics of the S3 core demonstrate that a completely freshwater tide-dominated system can still develop rhythmically alternating sand and mud beds. This suggests that the mechanism that controls mud deposition in a fluvial-tidal environment is not necessarily flocculation resulting from the mixing of fresh- and saltwater, but can also result from seasonal variations in flow velocity and intermittent flow dampening by tides. This indicates that rhythmic sand-mud couplets in fluvial deposits do not have to form in a brackish environment. Sedimentological characterization of ancient channel-margin deposits inferred to have a tidal component must be paired with ichnological and palynological analyses so that a definitive brackish signature can be deciphered. This kind of comprehensive assessment of channel-margin deposits (sedimentological, ichnological, and palynological) would enable a more accurate interpretation of the deposit’s relative position downstream of the tidal limit (i.e., from the freshwater backwater zone downstream into the brackish, admixed tidal-fluvial transition).

Relative to deposits more heavily influenced by tides downstream, fluvially dominated tidal backwater deposits should possess fewer mud drapes, relatively low mud contents, and exclusively unidirectional, seaward-directed indicators of flow (van den Berg et al. 2007; La Croix and Dashtgard 2015). Freshwater palynomorphs (with
possible rare marine palynomorphs transported upstream; Czarnecki et al. 2014), and very rare diminutive infaunal trace fossils may also be present. Mixed tidal-fluvial deposits are typically dominated by freshwater palynomorphs but with increasing numbers of marine palynomorphs (1-2 %), and cyclic colonization of beds by marine infauna that produce a trace assemblage dominated mainly by simple vertical burrows (e.g., Wightman et al. 1987; Pemberton and Wightman 1992; Czarnecki et al. 2014; Johnson and Dashtgard 2014). Moreover, mixed tidal-fluvial deposits generally contain a higher abundance of reactivation surfaces, possible bidirectional cross stratification and climbing ripples, and more mud (e.g., de Mowbray and Visser 1984; de Boer et al. 1989; van den Berg et al. 2007; Choi 2010; Martinius and Gowland 2011). Increased mud contents are manifest as common mud drapes, with greater thicknesses, uniformity, lateral continuity and rhythmicity of mud bedsets (e.g., Choi et al. 2004; Dalrymple and Choi 2007; Dashtgard et al. 2012; Sisulak and Dashtgard 2012; Johnson and Dashtgard 2014; La Croix and Dashtgard 2014, 2015).

Tidal influence in the Pitt River bar is expressed in the sedimentological signature of the deposit intersected in the S3 core, but is not reflected in the core’s ichnological or palynological signatures, as the system is completely freshwater in character. Tidal influence on deposition is detectable in the rhythmicity and uniformity of sand-mud interbeds, as opposed to irregular sand-mud interbedding observed in more fluvially dominated river reaches (e.g., Sisulak and Dashtgard 2012; Johnson and Dashtgard 2014). The presence of bidirectional current ripples, common mud drapes, and bed-to-bed variations in organic content also reflect the tide-dominated nature of flow in the Pitt River.

3.5.3. Implications for the rock record

Recognition of tidal influence on deposition in an ancient channel system should be based upon key sedimentological, ichnological, and palynological characteristics (see Sections 3.1.3 and 3.5.2). Once tidal influence is verified, a maximum probable rate of lateral accretion can be estimated. Data collected in the Pitt River and in the tide-influenced reach of the lower Mississippi (Hudson and Kesel 2000) demonstrate that geologically significant lateral migration occurs in tide-influenced meandering river
systems, at a rate equivalent to slowly migrating meanders in purely fluvial systems (Hickin and Nanson 1984; Hooke 1997; Van De Wiel 2003; Nicoll and Hickin 2010). In the tide-dominated Pitt River, more than 500 m of lateral accretion has occurred over ~3 ky. In the tide-influenced reach and mixed tidal-fluvial transition of the lower Mississippi, migration occurs on the scale of metres per year (Hudson and Kesel 2000). Using all data collected from tide-influenced river channels (Hudson and Kesel 2000; this study), it is estimated that lateral migration in most tide-influenced river meanders occurs at a rate less than 1.5% of the channel’s width each year.

In subsurface cores where the horizontal distance of a meander scroll and paleochannel parameters cannot be measured directly, an alternate method can be used to approximate the rate of lateral accretion using sedimentological and ichnological characteristics. First, seasonal cyclicity must be identified in the sediments being parameterized. In the tidal reach (i.e., from the mixed tidal-fluvial zone upstream to the tidal backwater zone), seasonal cycles are recorded in common unbioturbated sandstone dominated by current-generated bedforms sharply overlain by bioturbated to unbioturbated mud-dominated bedsets (e.g., Sisulak and Dashtgard, 2012; Jablonski, 2012; Nardin et al., 2012). Tidal influence on sedimentation in these sand-mud couplets is reflected in their rhythmicity and uniform thickness (cf. Choi et al. 2004; Dalrymple and Choi 2007; Johnson and Dashtgard 2014; La Croix and Dashtgard 2015). Using a basic trigonometric ratio (Eq. 1), the mean thickness (vertical accretion) of these annual sand-mud couplets and the mean dip angle of point-bar foresets observed in outcrops and/or wellbore imaging can be used to quantify contemporaneous lateral accretion of the point bar (Fig. 3.11):

\[ x = y (\tan \theta)^{-1} \]

(Equation 1)
Figure 3.11. The lateral accretion rate (m y\(^{-1}\)) of a point-bar can be estimated using the thickness (vertical accretion) of an annual sand-mud couplet (m y\(^{-1}\)) and the mean dip angle of the point-bar foresets (°).

**Rock record example: Middle McMurray Formation**

The middle unit of the McMurray Formation in northern Alberta, Canada is interpreted as amalgamated channel deposits in valleys with an upwards increase in marine influence (e.g., Jeletzky 1971; Pemberton et al. 1982; Wightman and Pemberton 1997). Evidence for tidal influence in the middle McMurray’s large-scale (kms long) point-bar deposits is primarily based on observed brackish to marine trace-fossil suites, the presence of marine and brackish-water dinocysts, and rhythmically alternating decimetre-scale sand and mud bedsets (e.g., Pemberton et al. 1982; Smith 1985).

Meander rates of McMurray paleochannels have only recently been discussed (e.g., Fustic et al. 2012; Jablonski 2012; Musial et al. 2012). For example, Musial et al. (2012) estimated the rate of McMurray point-bar accretion using the meander migration equation of Hickin and Nanson (1975) derived from a purely fluvial meandering system. Bend curvature values and floodplain ridge spacing were determined from seismic data, and used to infer that middle McMurray point-bars laterally accreted at a rate of 30-65 m y\(^{-1}\) (Musial et al. 2012).

An alternative lateral accretion rate for the tide-influenced McMurray channels in the example described by Musial et al. (2012) can be calculated by using the typical migration rate for tide-influenced channel meanders (< 1.5% of the channel’s width each year). Taking the McMurray channel parameters from Musial et al. (2012) (average paleochannel width = 750 m), it is estimated that lateral accretion probably occurred at a rate less than 11 m y\(^{-1}\) (M = 0.015 \times 750 m). Further, by using the typical mean lateral accretion dip of point-bars (10° ± 5°; Musial et al. 2012) and assuming that the
decimetre-scale sand-mud couplets in the middle McMurray reflect season deposition, a second estimate for lateral accretion can be calculated to be on the order of 2-6 m y\(^{-1}\) based on an average sand-mud bedset thickness of 0.50 m measured from subsurface cores.

### 3.6. Conclusions

Calculation of meander migration in the Pitt River demonstrates that tide-dominated (and correspondingly tide-influenced) river channels can migrate at a rate equivalent to slowly meandering purely fluvial systems. Sedimentological analyses and geochronological results from vibracores were used to calculate a long-term mean lateral accretion rate of 0.16-0.28 m y\(^{-1}\) for the Pitt River meander bar. This range in lateral meander migration is equivalent to previously recorded rates of lateral migration via bank erosion in slowly migrating and purely fluvial meandering systems of similar drainage area and mean discharge (cf. Van De Wiel 2003).

Data from the Pitt River and the tidal reach of the lower Mississippi (Hudson and Kesel 2000) provide the only empirically derived values of lateral migration in river channels affected by tides. The comprehensive study of Hudson and Kesel (2000) across the lower Mississippi demonstrated that meander migration rates decrease with increased tidal influence. While the results from the Pitt River and lower Mississippi indicate that tide-influenced channels migrate, it is not suitable to predict accretion rates of ancient tide-influenced channel bar deposits using equations derived from purely fluvial systems (e.g., Hickin and Nanson 1975; Van De Wiel 2003). Instead, parameterized meander behaviour in the Pitt and lower Mississippi rivers provides an estimated maximum annual rate of migration in tide-influenced channels equivalent to less than 1.5% of the channel's mean width, and this value decreases incrementally with increasing inferred tidal influence.
Chapter 4.

Conclusion

4.1. Research objectives

A vibracore-based project was conducted on the Pitt River Valley (PRV) floodplain to spatially and temporally constrain the mid-Holocene evolution of the floodplain and to quantify the lateral migration rates of the meandering, tide-dominated Pitt River. Three sets of key questions were posed as research objectives, and the results and conclusions drawn from this study are used to answer these questions below.

1) Is the presently accepted model for the evolution of the Pitt River Valley accurate? Do our data confirm the previously interpreted position of stratigraphic and geographic boundaries, and can the chronology for the marine-to-nonmarine transition be refined?

Vibracore data collected on the PRV floodplain confirm that rapid aggradation of the floodplain occurred between 8-6 k 14C BP (~5 mm y\(^{-1}\); Clague et al. 1982; Williams and Roberts 1989), and that by 6 k 14C BP, sea level had essentially reached its present level, resulting in the subsequent dissipation of vertical aggradation of the PRV floodplain.

This study and work by Locher (2014) serve to refine the evolutionary model for the PRV floodplain, as evident from cores collected in both projects, which indicate that the south end of Pitt Lake was likely within 2 km of its present position by 8 k 14C BP. Much of the northward progradation of the PRV floodplain must have occurred prior to 8 k 14C BP. The maximum mean rate of northward progradation of Pitt Lake over the last ~8 ky is 0.25 m y\(^{-1}\).
A marine-to-nonmarine transition could not be defined by the data collected in this study, as cores did not penetrate far enough into the subsurface. However, Locher (2014) was able to define a brackish-to-freshwater transition (via diatom analysis) at approximately 13 m below masl, 9080 ± 20 $^{14}$C BP.

2) What are the sedimentological and ichnological characteristics of the Pitt River deposits? Is there evidence of tides, and if so, what are they? How do these criteria differ from those of brackish-water and tidal deposits?

One vibracore (S3) was drilled through Pitt River channel-margin sediments (see Section 2.3.3). This core possessed rhythmically alternating bedsets of current rippled vFL sand and parallel laminated silty mud. Sandy bedsets have variable organic contents and common bidirectional current ripples and mud drapes. These sedimentological characteristics (bedsets of alternating sediment calibres, bidirectional current ripples, mud drapes, variance in organic content, etc.) indicate tidal influence on deposition. However, the ichnological (rare Planolites and Palaeophycus) and palynological signatures (strictly terrestrial palynomorphs) indicate that the Pitt River remained fresh when the sediments preserved in the S3 core were deposited (2980 ± 15 $^{14}$C BP). The S3 core demonstrates that rhythmicity and uniformity in sand-mud couplets reflects changes in flow conditions due to tidal and seasonal processes, and does not necessarily require freshwater and saltwater mixing.

3) Has the Pitt River migrated laterally with time? If so, what was the rate of lateral accretion and how does this compare to those of purely fluvial systems?

The Pitt River meander proximal to S3 has migrated laterally at approximately 0.16-0.28 m y$^{-1}$ over the past ~3 ky (Figs. 3.2 and 3.3). This rate is equivalent to slowly migrating meandering river channels that are purely fluvial (Figs. 3.4 and 3.5; Van De Wiel 2003; Nicoll and Hickin 2010).

At present, quantification of meander migration in tide-affected river channels is restricted to the Pitt River and lower Mississippi (Hudson and Kesel 2000). Data from these two systems are used to infer that meander migration in tidally influenced
channels occurs at a average rate < 1.5% of the channel’s width each year, and is even lower in river channels dominated by tides, such as in the Pitt River.

4.2. Future research

The Pitt River is anomalous in that it is a completely freshwater system. However, studying this system has enabled the evaluation of a tide-dominated system independent of saltwater influences and provided a foundation for future work on meander migration in the tidal reach of river systems.

One avenue for future research is the collection of field data (i.e., meander accretion rates) in other modern river systems, especially across and downstream of their tidal limits. The Mitchell River of Queensland, Australia is a low-gradient meandering channel that migrates across a tidal floodplain at its seaward terminus. Parameterization of meanders and lateral meander migration across the Mitchell River’s tidal floodplain (in the tide-dominated length of the channel) would add much-needed data to the presently limited data of lateral migration rates collected downstream of the tidal limit in fluvial channels (Hudson and Kesel 2000; this study). Moreover, by collecting more meander migration and bend curvature values, the estimated rate of maximum migration in channels influenced by tides can be augmented.

Additionally, the use of basic trigonometry to estimate lateral accretion based on the thickness of inferred annual deposits and average point-bar dip angles must be tested. First, more work must be done to definitively state that sand-mud couplets in fluvial-tidal channel-margin deposits are in fact seasonal (deposited annually). Once this is confirmed, case studies must be conducted on modern channels where lateral migration can be quantified by looking at long-term accretion in the same fashion that the Pitt River was parameterized. Next, the same channel-margin deposits should be parameterized with respect to average couplet thickness and bed dip angle so that the estimated rate of lateral accretion via geochronology and distance accreted can be compared to the estimated rate of lateral accretion via trigonometric ratios of sedimentological characteristics.
References


Appendix A.

The Wink Vibracore Drill: Overview and Instruction manual

Description:
The in house Wink Vibracore drilling manual developed for the ARISE Group at SFU by myself and Kristen Maduik. This manual was drafted in the summer of 2014 during our first field season with the Wink Vibracore drill.

File Name:
Wink_Vibracore_drilling_manual.pdf
Appendix B.

Palynological slide counts

Table B.1. Raw semi-quantitative palynological slide count results provided by Geotop-UQAM’s micropaleontology and marine palynology laboratory.

<table>
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<th>Larix</th>
<th>bisaccate</th>
<th>Tsuga</th>
<th>total pollen</th>
<th>monolete spore</th>
<th>trilete spores</th>
<th>Halodinium</th>
<th>Pediastrum</th>
<th>Botryococcus</th>
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