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Reconstruction of multi-century flood histories from oxbow lake sediments, Peace-Athabasca Delta, Canada

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Abstract:

Floods caused by ice-jams on the Peace River are considered to be important for maintaining hydro-ecological conditions of perched basins in the Peace-Athabasca Delta (PAD), Canada, a highly productive and internationally recognized northern boreal ecosystem. Concerns over the potential linkages between regulation of the Peace River in 1968 for hydroelectric production and low Peace River discharge between 1968 and 1971 during the filling of the hydroelectric reservoir, absence of a major ice-jam flood event between 1975 and 1995, and low water levels in perched basins during the 1980s and early 1990s have sparked numerous environmental studies largely aimed at restoring water levels in the PAD. Lack of sufficient long-term hydrological records, however, has limited the ability to objectively assess the importance of anthropogenic factors versus natural climatic forcing in regulating hydro-ecological conditions of the PAD. Here, we report results of a paleolimnological study on laminated sediments from two oxbow lakes in the PAD, which are located adjacent to major flood distributaries of the Peace River. Sediment core magnetic susceptibility measurements, supported by results from several other physical and geochemical analyses as well as stratigraphic correspondence with recorded high-water events on the Peace River, provide proxy records of flood history spanning the past ~180 and ~300 years in these two basins. Results indicate that inferred flood frequency has been highly variable over the past 300 years but in decline for many decades beginning as early as the late nineteenth century, well before Peace River regulation. Additionally, several multi-decadal intervals without a major flood have occurred during the past 300 years. While climate-related mechanisms responsible for this variability in flood frequency remain to be determined, as does quantifying the relative roles of river regulation and climate variability on hydro-ecological conditions in the PAD since 1968, these results suggest that ecosystem management strategies for the PAD need to explicitly account for natural variations in flood recurrence intervals. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS flood reconstruction; delta; Peace River; regulation; paleolimnology; magnetic susceptibility

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INTRODUCTION

Floods caused by ice-jams on the Peace and Athabasca rivers are important for recharging the Peace-Athabasca Delta (PAD), Canada, and are thought to play a significant role in regulating hydro-ecological conditions in this highly productive northern boreal ecosystem (Prowse and Lalonde, 1996; Prowse *et al.*, 2002a). The longer growing season, nutrient-enriched soils, numerous lakes and wetlands, and reduced permafrost in the PAD compared to surrounding regions provide an attractive oasis for wildlife such as buffalo, muskrat, and waterfowl (PADTS, 1996). The PAD is recognized by the Ramsar Convention on the Conservation of Wetlands

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of International Importance, and approximately 80% of it is located within Wood Buffalo National Park, a UNESCO (United Nations Educational, Scientific and Cultural Organization) World Heritage Site.

Over the past 35 years, controversy has developed regarding the cause of recently observed dry conditions in the PAD which have been linked to reduced flood frequency (Prowse and Conly, 1998; Leconte *et al.*, 2001). Concerns were first raised shortly after construction of the WAC Bennett Dam in 1968, a hydroelectric dam located near the headwaters of the Peace River in northern British Columbia (PADPG, 1973; Townsend, 1975). Initial filling of the Williston Lake reservoir over the subsequent 4 years (1968 – 1971) coincided with low discharge on the Peace River and no significant flooding of the PAD. Substantial changes in the landscape, vegetation, and wildlife of the PAD occurred during this 4-year interval (PADPG, 1973; Townsend, 1975). A second dry interval began after 1974, the year of a major ice-jam-induced flood, during which many perched basins in the higher elevation landscape of the PAD, especially near the Peace River, experienced low water levels (Prowse and Conly, 1998). This dry interval lasted until the next major ice-jam flood event of 1996 (Prowse *et al.*, 2002a). Concerns over the possible impact of flow regulation have led to several environmental studies, including the Peace-Athabasca Delta Project Group (PADPG, 1973), the Peace-Athabasca Delta Implementation Committee (PADIC, 1987), the Peace-Athabasca Delta Technical Studies (PADTS, 1996), and the Northern River Basins Study (Gummer *et al.*, 2000), aimed largely at identifying and implementing engineering solutions to restore water levels in the PAD. Recent studies, however, suggest that the absence of a high-magnitude flood between 1975 and 1995 may be a result of climatic variation that increased the temperatures during the ice-cover season, reduced the snow-pack depths, and altered the intensity and duration of the pre-melt period (Prowse and Conly, 1998; Prowse *et al.*, 2002b). Thus, reduced spring ice-jam flooding due to regulation of the Peace River and climate variability represent competing hypotheses to explain what are perceived to be unusually dry conditions and low lake levels in the PAD during much of the past 35 years. Rigorous scientific assessment of these hypotheses has been hampered, however, because only 9 years of pre-dam discharge data are available for the Peace River near the PAD at Peace Point. The absence of sufficient long-term records limits the ability to assess the importance of anthropogenic factors versus natural climatic forcing in regulating hydro-ecological conditions of the PAD.

To address this shortcoming, Traditional Knowledge and written documents were used to develop a history of Peace River spring ice-jam floods as a part of the latest series of environmental studies on the PAD (Peterson, 1995; PADTS, 1996). Results, subsequently reported and updated by Timoney *et al.* (1997) and Timoney (2002), indicate that 28 major ice-jam floods occurred on the Peace River between 1826 and 2000 which, on average, is 1 flood approximately every 6 years over this period. Interestingly, these historical records indicate the occurrence of a multi-decadal dry-wet-dry cycle of lower than average flood frequency during the late 1800s and late 1900s and intervening higher than average flood frequency during the early 1900s (Timoney *et al.*, 1997). The significance and magnitude of many of the flood events recorded by historical sources, however, have been questioned because of the high spatial variability of flooding and possible observer bias (Prowse and Conly, 2002). Hence, generating long-term records of flood frequency for the PAD from additional data sources is required to assess objectively the impact of river regulation on hydro-ecological conditions of the PAD. Improved knowledge of flood frequency prior to river regulation, in particular, is essential for designing effective and appropriate ecosystem management strategies in the PAD in light of ongoing climatic change and variability, as well as the likelihood of increased demand on water resources for hydroelectricity and other uses. These data are also required for providing benchmark targets to evaluate the success of any future efforts to increase water levels in the PAD (PADTS, 1996). As recommended by Peterson, following compilation of the Traditional Knowledge – based and historical-based flood records (1995, p. 14):

'A more extensive flood history database... may possibly be constructed from analysis of sediment cores taken from the Delta lakes. Sediment cores taken from perched basins may provide information on the frequency of ice-jam induced overland floods. The possibility of using this technique to enhance the body of flood history data should be investigated.'

Here, we report results from physical and geochemical analyses of laminated lake sediments from two oxbow lakes in the PAD, which are located adjacent to major flood distributaries of the Peace River. High sedimentation rates in these basins are the result of their close proximity to these distributaries, which flood into the oxbows when river water levels exceed the basin sill elevations. Preservation of the physical and geochemical signatures of these events in the sediment records allows reconstruction of flood histories over the past ~180 and ~300 years for these lakes. These results are then used to address two questions that have profound implications for ecosystem management of the PAD:

1. Has flood frequency declined since the upper Peace River was regulated in 1968? and
2. Did prolonged multi-decadal intervals (20+ years) between major floods (analogous to the 1975 – 1995 interval) occur prior to regulation of the Peace River?

HYDROLOGICAL SETTING OF THE PEACE-ATHABASCA DELTA AND SITE DESCRIPTIONS

The hydrology of the PAD is complex because of the low relief and the presence of distributary channels that can experience flow reversals. Under normal flow conditions, Lake Athabasca and the Athabasca River drain northwards via the Rivière des Rochers, Revillon Coupé, and Chenal des Quatre Fourches to the Peace River, where they join to form the Slave River (Figure 1). During high-water events on the Peace River, accentuated by ice-jam conditions near the confluence of the Peace and Slave rivers, water can flow southwards from the Peace River through these channels and cause overland flooding of the northern Peace sector of the PAD. Periodic ice-jam flooding is thought to be important for maintaining the water balance of many perched basins in the Peace sector of the PAD, which provide extensive wildlife habitat, as high Peace River discharge during the open-water season is generally insufficient to cause widespread flooding (Prowse and Lalonde, 1996). In the absence of flooding, many of the perched basins may desiccate and terrestrial vegetation may replace productive wetland habitat because long-term average open-water evaporation exceeds precipitation (Prowse and Lalonde, 1996; Prowse and Conly, 1998). Widespread flooding of the Peace sector of the PAD last occurred in 1996 and 1997 (Prowse *et al*., 2002a; Timoney, 2002).

Groundwater exchange is considered to be a minor component of lake water balances because most basins in the PAD are underlain by low permeability (10^{-8} – 10^{-10} m s^{-1}) clay and silt deposited by floodwaters and horizontal gradients are very low (PADPG, 1973; Prowse *et al*., 1996). The region lies within the zone of discontinuous permafrost, which may further restrict local groundwater movement.

In this study, we report the results of physical and geochemical analyses on lake sediment records obtained from 'PAD 54' (local name: Horseshoe Slough) and 'PAD 15' (local name: Pete's Creek). PAD 54 is a 6-m deep cut-off meander (surface area ~18 ha) of the Chenal des Quatre Fourches, located 4 km south of the Peace River (Figure 1). The basin is normally closed hydrologically, but is intermittently connected to the river via a small crevasse channel across the levee that defines the sill elevation (209·92 m above sea level; D. Peters, National Water Research Institute, personal communication, 2000). This lake was cored because it was likely to provide a highly sensitive record of flood events and overbank flow from the Chenal des Quatre Fourches, which connects with the Peace River immediately upstream of the sharp bend at Rocky Point (Figure 1). This bend is considered to be a site where ice-jams frequently develop (Prowse and Lalonde, 1996).

PAD 15 is a 4-m deep oxbow lake formed by a meander cut-off (~16 ha) from the Revillon Coupé, which is also a major distributary of the Peace River during flood events (Figure 1). This basin is also hydrologically closed except for intermittent connections during high-water events on the river. The precise sill elevation of this lake is not known, but it is likely slightly higher than that of PAD 54 (see below). This lake was cored because it was considered likely to provide a highly sensitive record of flood events and overbank flow from the Revillon Coupé.

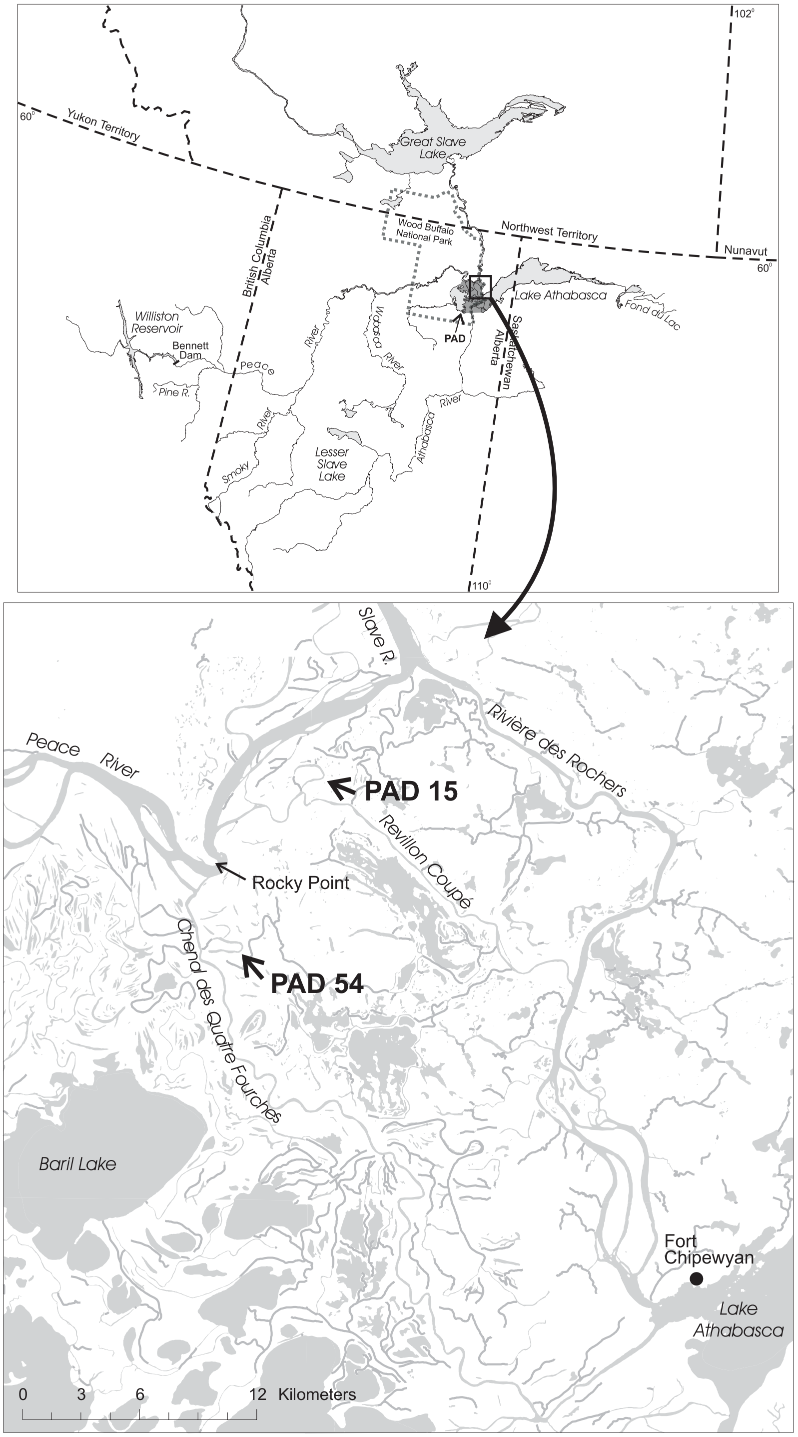

Figure 1. The Peace-Athabasca Delta (PAD) is located in northeastern Alberta, Canada, to the west of Lake Athabasca. Most of the PAD is located within Canada's largest national park, Wood Buffalo National Park. The WAC Bennett Dam was constructed on the Peace River 1180 km upstream of the PAD in 1968. PAD 54 and PAD 15 are oxbow lakes located in the northern Peace sector of the PAD

Observations we have made over the past few years, as well as ongoing hydrological monitoring using water isotope tracers, indicate that PAD 54 and PAD 15 are very susceptible to flooding, as evidenced by turbid, isotopically depleted river waters entering these lakes during high river stages in summer 2001 (PAD 54 only) and spring 2003 (both basins). River waters entering PAD 54 in the summer of 2001, but not PAD 15, indicate that the sill elevation may be slightly lower at PAD 54. Both events were minor as floodwaters were mainly contained along and immediately adjacent to the main river channels and did not cause widespread overland flooding. Since autumn 2000, we have not observed substantial seasonal draw down of water levels or strong evaporative isotopic enrichment of the lake waters that would indicate volumetric reduction (i.e. evaporation exceeding precipitation and inflow). This suggests that direct precipitation on the lakes and catchment runoff are sufficient to maintain water levels in these basins during years without flooding (also see Wolfe *et al.*, 2005), unlike some perched basins in the Peace sector of the PAD (Prowse and Lalonde, 1996).

METHODS

Field collection

Sediment cores were collected through lake-ice in March 2001 from the north arms of PAD 54 and PAD 15, within a few 100 m of the point of river water entry, and midway between north and south shorelines. Two cores of unconsolidated surface sediments (upper ~20 – 30 cm) were obtained from each lake using a Glew (1989) gravity corer (diameter 7·5 cm), which were then transported to the field station in a helicopter and sectioned into 0·5-cm intervals within 24 h of collection. Two cores were obtained from each coring site to provide sufficient material for laboratory analyses described below.

More compact, deeper sediments (i.e. below 10-cm sediment depth) were collected using a 10-cm diameter Russian peat corer with extension rods. The Russian corer collects a 1-m long sediment core section each time it is deployed. Consequently, to ensure recovery of a continuous sedimentary sequence and to correlate the 1-m core sections, we collected adjacent 1-m core sections such that they contained a 30-cm sequence of sediment that overlapped with the section immediately above it. The adjacent 1-m core sections were collected from two coring locations situated ~1 m apart, so as to avoid sediment disturbances that arise when adjacent sections are collected from the same coring site. From PAD 54, a total length of 363·5 cm of sediment core, comprising five 1-m overlapping sections, was recovered. From PAD 15, a total length of 406·0 cm of sediment core, consisting of six 1-m overlapping sections, was recovered. Immediately after recovery, core sections were wrapped in plastic sheets to prevent desiccation, and placed in supportive PVC trays. All gravity core samples and core sections were transported to the University of Waterloo, Canada, and placed in cold storage (4 °C) prior to laboratory analyses.

Laboratory analyses

The gravity cores and the upper 1-m section of the core sequences collected with the Russian corer from PAD 54 and PAD 15, which were sectioned at 0·5 cm intervals in the lab, were subject to several physical and geochemical analyses as described below. The deeper 1-m core sequences collected with the Russian corer were left intact and analyzed only for magnetic susceptibility at 0·2 cm resolution. Extensive physical and geochemical analyses were conducted on the upper 1 m of sediment from PAD 54 and PAD 15 to develop sediment core chronologies and to characterize the origin, sedimentology, and depositional conditions that produced the banded sediments. The complete 363·5 cm and 406·0 cm long sediment cores from PAD 54 and PAD 15, respectively, were used to generate a proxy record of flood history from magnetic susceptibility measurements, after correcting for reduced compaction and higher water content of surficial sediments and identifying that the peaks in magnetic susceptibility correspond to high-water levels on the Peace River recorded at Rocky Point (Figure 1) since 1972.

Radiometric dating. Radioactive isotopes ^{210}Pb, ^{226}Ra, and ^{137}Cs were measured at the Environmental Radiochemistry Laboratory, Freshwater Institute, Winnipeg, Canada. Dry sediment (1 – 5 g) was sealed in 60 × 15 mm plastic petri dishes, aged for 30 days, and counted on a gamma spectrometer (Ge(Li) or HPGe semiconductor detector) for the determination of ^{137}Cs and ^{226}Ra (Joshi, 1987). One- to 3-g subsamples were analyzed for ^{210}Pb by leaching in 6N HCl in the presence of a ^{209}Po tracer, autoplating Po onto a silver disc (Flynn, 1968), and counting the disc on an alpha spectrometer to determine ^{210}Pb via its ^{210}Po decay product. ^{226}Ra was determined by the radon de-emanation technique (Mathieu, 1977; Wilkinson, 1985).

For PAD 54, ^{210}Pb analyses were conducted at every 0·5-cm interval between 0- and 23·5-cm depth, on combined 2-cm slices at 5-cm intervals between 25- and 90-cm depth, and at 299 – 300-cm depth. ^{226}Ra analyses were measured on samples spanning depth intervals 0 – 0·5 cm, 4·5 – 5·0 cm, 9·5 – 10·0 cm, 14·5 – 15·0 cm, 22·0 – 22·5 cm, and 69 – 71 cm. ^{137}Cs analyses were initially performed on every second 0·5-cm interval up to 23·5-cm depth and then on combined 2-cm slices at 5-cm intervals between 25- and 90-cm depth. Several additional analyses were then conducted in the upper 23·5 cm to increase the resolution of the ^{137}Cs profile.

For PAD 15, ^{210}Pb analyses were conducted at every 0·5-cm interval between 0- and 13·5-cm and at 372 – 373-cm depth. ^{226}Ra analyses were measured on samples from 4·5 – 5·0 cm and 9·5 – 10·0 cm. ^{137}Cs analyses were performed at every second 0·5-cm interval between 0- and 23·5-cm depth and on combined 2-cm slices at 5-cm intervals between 25- and 90-cm depth.

Radiocarbon dating of bulk organic sediments sampled from dark-coloured beds (PAD 54: 155 – 160 cm and 323·5 – 328·5 cm; PAD 15: 396 – 401 cm) was performed by liquid scintillation counting at the University of Waterloo – Environmental Isotope Laboratory (UW-EIL). A twig sampled at 236·5-cm depth in the PAD 15 core sequence was analyzed by accelerator mass spectrometry (AMS) at the Rafter Radiocarbon Laboratory (Lower Hutt, New Zealand) after initial sample preparation and CO_2 gas collection at the UW-EIL.

Moisture and loss-on-ignition determinations. Samples for moisture content, organic matter content, and total carbonate content were evaluated by weight loss on heating to temperatures of 85 °C, 500 °C, and 1000 °C, respectively (Dean, 1974). Analyses were performed on 0·5-cm intervals of the upper 1 m for PAD 54 and PAD 15 at the University of Manitoba.

Grain size. Particle size spectra for each subsample were determined using an automated laser optical particle size analyser (Galai CIS-1; Last, 2001a; Aharonson *et al.*, 1986) after removal of organic matter by hydrogen peroxide treatment. Percentages of clay, silt, and sand, mean and median grain size, and population standard deviation (sorting) were calculated from these spectra. Replicate analyses indicate that the precision of the particle size data is approximately ±4%. Analyses were performed on continuous 0·5-cm intervals on the upper 1 m for PAD 54 and PAD 15 at the University of Manitoba.

Mineralogy. All samples analyzed for mineralogy were air-dried at room temperature, disaggregated in a mortar and pestle, and passed through a 62·5-μm sieve after removal of organic matter by hydrogen peroxide treatment. Bulk mineralogy was determined using standard X-ray diffraction techniques as outlined by Last (2001b). Percentages of the various minerals were estimated from the bulk mineral diffractograms using the intensity of the strongest peak for each (Last, 2001b). Duplicate analyses indicate that the precision of the mineralogical data is approximately ±8%. Analyses were performed on continuous 0·5-cm intervals on the upper 1 m for PAD 54 and PAD 15 at the University of Manitoba.

Magnetic susceptibility. Analyses were performed on the upper 1 m for PAD 54 and PAD 15 on discrete 1-cm^3 subsamples of wet sediment at continuous 0·5-cm intervals using a balanced alternating current (AC) bridge sensor (Sapphire Instruments, Model SI2B). Whole core magnetic susceptibility of intact 1-m long core sections collected with the Russian corer from PAD 54 and PAD 15 was measured using a surface scanning sensor (Bartington Instruments, Model MSE2) and an automated measuring stage with a stratigraphic

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resolution of 0·2 cm. Magnetic susceptibility is expressed as initial volume susceptibility (κ). Measurements of magnetic susceptibility were performed at the Environmental Variability and Extremes Laboratory, Queen's University, Kingston, Canada.

Carbon and nitrogen geochemistry. All samples were pre-treated with 10% (by volume) HCl at 60 °C to remove carbonate minerals and shells. This was followed by rinsing with de-ionized water, freeze-drying, and sieving (500 μm) to remove coarse organic debris that may be of terrestrial origin. Carbon (C) and nitrogen (N) content, C/N weight ratios, and carbon isotope composition of the acid-washed <500-μm fraction was measured by an elemental analyzer interfaced with a continuous flow – isotope ratio mass spectrometer. Isotope results are expressed as δ values, representing deviations in per mil (‰) from the Vienna-PeeDee Belemnite (VPDB) standard for carbon, such that $\delta_{\text{sample}} = [(R_{\text{sample}}/R_{\text{standard}}) - 1]^{*}10^{3}$, where R is the $^{13}C/^{12}C$ ratio in the sample and the standard. Analyses were performed on continuous 0·5-cm intervals on the upper 1 m for PAD 54 and PAD 15 at the UW-EIL. Results of repeat $\delta^{13}C$ analyses of samples are generally within ±0·2‰.

RESULTS AND DISCUSSION

General stratigraphy and core description

Sediment cores recovered from PAD 54 and PAD 15 consist of light to dark grey, slightly calcareous silty clay and clayey silt, and exhibit striking colour banding of alternating dark- and light-coloured beds and laminations of variable thicknesses. Most bed thicknesses are <10 mm but show a range from <1 mm to approximately 10 cm. Contacts between beds are even and sharp to diffuse. There is no evidence of any erosional contacts or sedimentary structures produced by traction currents, although several distinctly fining-upward graded beds are present in PAD 15. There is no evidence of subaerial exposure or desiccation, such as pedogenic horizons, which are common in many other lakes in southwestern Canada (*cf* Last and Vance, 1997; Vance *et al.*, 1997; Teller and Last, 1982). Bioturbation is not present and there was virtually no macrofossil debris observed except for a twig at 236·5-cm depth in PAD 15, which was sampled for ^{14}C dating.

Overall, these general stratigraphic observations suggest that relatively deep water conditions have persisted in both basins during the time-span represented by the sediment records. Lacustrine processes in these oxbow lakes have been dominated by suspension-settling sedimentation associated with periodic influx of detrital, minerogenic sediments derived from flooding from the adjacent river channels.

Sediment core chronology

Laboratory analyses on the stratigraphic records from PAD 54 and PAD 15 included the use of radioactive isotopes ^{210}Pb, ^{137}Cs, ^{226}Ra, and ^{14}C to establish sediment core chronologies. Results indicated that measurements of ^{210}Pb and ^{14}C on bulk sediment samples could not be used to establish reliable chronologies, owing to rapid sedimentation rates that diluted unsupported ^{210}Pb activities and dilution of the ^{14}C signal by 'old carbon' material (Table I; Figures 2, 3). For example, total ^{210}Pb analyses throughout the upper 1 m of sediment from PAD 54 show no change in concentration, and the values are similar to concentrations of ^{226}Ra measured at selected intervals (Figure 2a). An additional sample analyzed at 299 – 300-cm depth resulted in a similar ^{210}Pb concentration as in the upper 1 m (0·0334 Bq g^{-1}), which is comparable to supported ^{210}Pb concentrations from a number of other lake sediment cores that we analyzed in the PAD (Hall *et al.*, 2004). Although fewer ^{210}Pb and ^{226}Ra analyses were conducted on sediments from PAD 15 than from PAD 54, the results are similar to those obtained from PAD 54 (Figure 3a). Specifically, total ^{210}Pb activity of a sample at 372 – 373-cm depth is equivalent to activities of sediments in the upper 15 cm (0·0333 Bq g^{-1}). These results indicate very rapid sedimentation rates at both lakes leading to dilution of unsupported ^{210}Pb.

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Concentrations of total ^{210}Pb are thus virtually equivalent to supported ^{210}Pb and prevent the use of ^{210}Pb dating models for developing a chronology for these sediments. While similar ^{210}Pb profiles can be obtained as a result of bioturbation or physical sediment mixing (Appleby, 2001), these explanations are unlikely, given the laminated structure of the sediment. Additionally, analyses of ^{14}C on bulk organic sediment from PAD 54 and PAD 15 show clear evidence of old carbon contamination (Table I), as the radiocarbon dates from bulk sediment samples are impossibly old (12 400 to 16 450 ^{14}C yr BP), given that this area was covered by the Laurentide Ice Sheet until at least 10 500 ^{14}C yr BP (Lemmen *et al.*, 1994). Consequently, chronologies were developed primarily on the basis of ^{137}Cs measurements of bulk sediment (both lakes), with additional support from AMS ^{14}C analysis of a plant macrofossil from PAD 15, as described below.

In contrast to the ^{210}Pb profiles, stratigraphic variation is evident in the ^{137}Cs profiles of both lakes (Figures 2b, 3b). Pronounced maxima in ^{137}Cs concentration occur at 72-cm depth in the sediment core from PAD 54 and at 50-cm depth in the core from PAD 15. These stratigraphic horizons are interpreted to represent 1964, the year of maximum atmospheric concentration of radionuclides produced from atmospheric testing of nuclear weapons (Appleby, 2001). Minor ^{137}Cs peaks above the 1964 horizon in both sediment records

Table I. Radiocarbon data for lakes PAD 54 and PAD 15

Lab number	Material	Core depth (cm)	Reported age (^{14}C yr BP)
PAD 54			
WAT-4226	Bulk sediment	155 – 160	12 650 ± 200
WAT-4227	Bulk sediment	323·5 – 328·5	12 400 ± 170
PAD 15			
R-26544/2	Twig	236·5	126 ± 60
WAT-4223	Bulk sediment	396 – 401	16 450 ± 420

WAT = University of Waterloo Environmental Isotope Laboratory; R = Rafter Radiocarbon Laboratory.

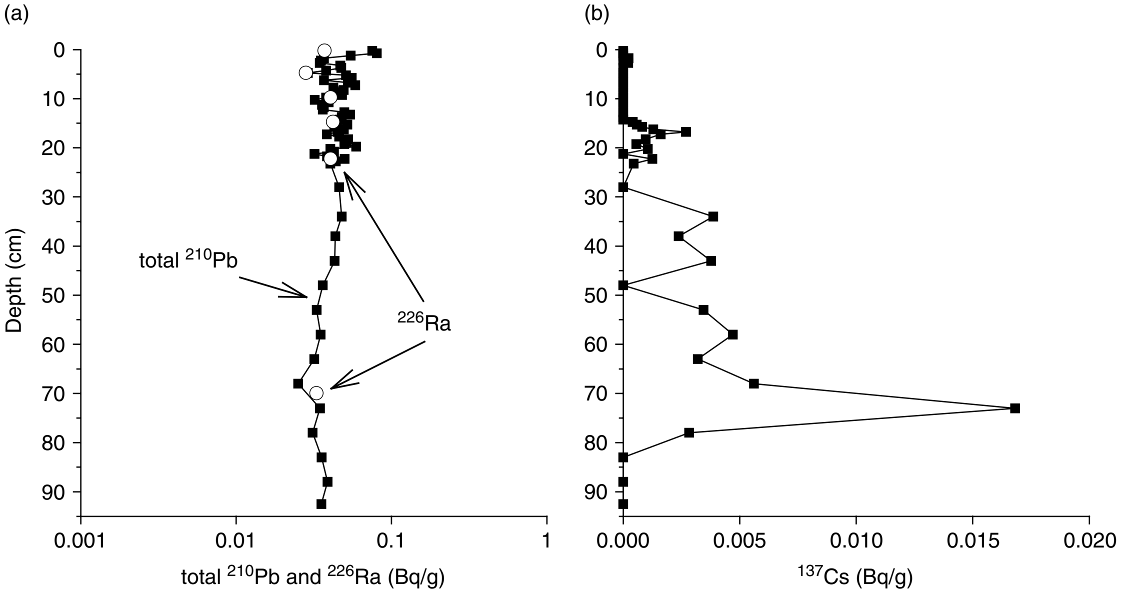

Figure 2. Profiles of activities of (a) total ^{210}Pb and ^{226}Ra, and (b) ^{137}Cs in sediments from PAD 54

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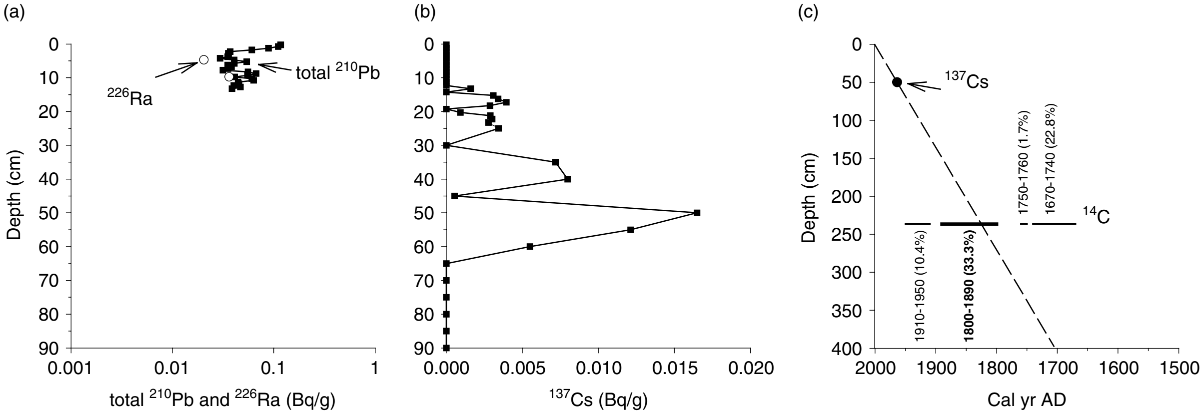

Figure 3. Profiles of activities of (a) total ^{210}Pb and ^{226}Ra, and (b) ^{137}Cs in sediments from PAD 15. (c) ^{14}C results obtained from a twig at 236·5-cm depth and calibrated to calendar years indicate that the interval of highest probability is consistent with an average sedimentation rate of 1·39 cm yr^{-1}, estimated from ^{137}Cs data. The ^{14}C date (Table I) was calibrated to calendar years by the probability distribution method using the OxCal Program v3·3 (Bronk Ramsey, 1999) and the Intcal98 calibration curve (Stuiver *et al*., 1998)

likely reflect re-worked sediments introduced to the lakes by post-1964 flood events. Similar overall patterns in the ^{137}Cs profiles at PAD 54 and PAD 15 suggest that the use of ^{137}Cs to establish the 1964 stratigraphic horizon is robust in these lacustrine sediments and unlikely to be affected by pore-water diffusion as observed in other studies (e.g. Crusius and Anderson, 1995). On the basis of the ^{137}Cs data and extrapolation to the surface, average linear sedimentation rates are estimated at 2·0 cm yr^{-1} for PAD 54 and 1·39 cm yr^{-1} for PAD 15. Using these sedimentation rates, the onset of measurable ^{137}Cs occurs between 1958 and 1961 at PAD 54 and between 1953 and 1957 at PAD 15, which are consistent with atmospheric ^{137}Cs fallout in the Northern Hemisphere reaching significant levels after 1953 (Appleby, 2001). Further down-core, ^{14}C dating on a twig at 236·5-cm depth from PAD 15 indicates that the interval of highest probability (1800 – 1890 cal yr AD) is in agreement with the ^{137}Cs-inferred average sedimentation rate of 1·39 cm yr^{-1} for the sediment core (Figure 3c). Other calibrated dates associated with this ^{14}C analysis are possible, but they have lower statistical probability. Since additional chronological information would be required to establish different sedimentation rates with equal or greater confidence than the estimates based on ^{137}Cs profiles, we applied sedimentation rates of 2·0 cm yr^{-1} for PAD 54 and 1·39 cm yr^{-1} for PAD 15 to the full stratigraphic records in order to estimate the sediment core chronologies. On this basis, the 363·5-cm sediment core from PAD 54 spans approximately 180 years, whereas the 406-cm sediment core from PAD 15 spans about 300 years. Clearly, assumptions of linear sedimentation rates may be problematic in these deposits, especially over short time intervals (e.g. annual to decadal), which are characterized by widely varying energy conditions. As we illustrate below, however, reconstructed variability in flood frequency spanning the length of the sediment records is strongly compatible with independent historical evidence. Moreover, in the absence of more highly resolved chronological control, use of a linear sedimentation rate represents a conservative approach with respect to assessing the changes in flood frequency because flood frequency is likely to be underestimated for periods when floods were common and overestimated when floods were relatively rare.

Physical and geochemical stratigraphy

We used a combination of physical and geochemical approaches to characterize the origin, sedimentology, and processes leading to deposition of the dark and light-coloured laminations and beds in PAD 54 and PAD 15. Analyses of grain size, mineralogy, magnetic susceptibility, organic C and N content, C/N weight ratios, and bulk organic δ^{13}C, reveal a consistent record of oscillatory energy conditions at PAD 54 and PAD 15. The physical parameters, namely, grain size, mineralogy, and magnetic susceptibility, were used because they can generally provide information about the source of the deposit, mechanisms responsible for the transport of the material, past physical conditions at the depositional site within the basin, and paleoclimatic and paleohydrological conditions within the surrounding watershed (Last, 2001a,b; Dearing, 1999; Sandgren and Snowball, 2001). In detrital-dominated lacustrine sediments, these measurements frequently have been used to assess energy conditions related to the impact of past fluvial activity or catchment erosion (e.g. Campbell *et al.*, 2000).

Texturally, the inorganic fraction of the upper 1 m of sediment recovered from the two basins is mainly clayey silt. Mean grain size for PAD 15 averages 5·8 μm (very fine silt), with a range from 3·3 μm to 12·7 μm. Sediments from PAD 54 are slightly finer-grained; mean grain size averages 4·8 μm and ranges from 2·0 to 8·4 μm. The mineral suite in both cores is nearly identical: mainly clay minerals (average = 65%) and quartz (20%), with smaller but equal proportions of feldspar minerals (6%) and carbonate minerals (5%). Magnetic susceptibility in the upper 1 m from PAD 54 and PAD 15 varies substantially, with values ranging from approximately 50 to >400 κ.

Results from analyses of mean grain size, siliciclastic mineralogy, and magnetic susceptibility on the upper 1 m of sediment from PAD 54 and PAD 15 show that coherent sub-annual oscillations define these profiles, which are consistent with variations in detrital input and energy conditions (Figure 4a, b). Statistically significant positive correlations between magnetic susceptibility and mean and median grain size, quartz and total feldspar content, and inverse correlation with carbon and nitrogen content (Appendix Table Ia, b) support a detrital interpretation of the magnetic susceptibility profile (*cf* Bennett *et al.*, 1990; Zolitschka, 1998). Stratigraphic increases in mean grain size coincide with increases in quartz and total feldspar content, decreases in clay mineral content, and increases in magnetic susceptibility (see arrows in Figure 4a, b; see Appendix Table I for respective linear correlation coefficients). We interpret these trends to be associated with high-energy conditions characterized by the influx of relatively coarse-grained, quartz-, feldspar-, and magnetic-rich detrital material. On the other hand, intervals defined by relatively low mean grain size, low quartz and total feldspar content, high clay mineral content, and low magnetic susceptibility reflect lower-energy intervals, reduced detrital influx and suspension-settling of relatively fine-grained, clay mineral-rich sediment.

Measurements of bulk organic carbon and nitrogen content and carbon isotope composition were also employed because these indicators can provide information on various processes that influence carbon and nitrogen cycling regimes in lakes and their watershed (Dean, 1999; Meyers and Teranes, 2001). In a similar study, Brown *et al.* (2000) identified flood events in lake sediment cores in New England by decreases in carbon and nitrogen percent, due to both dilution by increased inorganic (fluvial) sediment supply and reduced lake productivity, and increases in C/N ratios due to influx of terrestrial organic debris. Measured bulk organic δ^{13}C values also indicated that the organic fraction in the flood deposits represented a mixture of aquatic and terrestrial sources (Bierman *et al.*, 1997). As discussed in the subsequent text, results from the measurement of organic carbon and nitrogen content, C/N weight ratios, and carbon stable isotope composition on the <500 μm fraction from the upper 1 m of sediment from PAD 54 and PAD 15 are similar to that reported by Bierman *et al.* (1997) and Brown *et al.* (2000), and provide further support for our interpretation of the physical, mineralogical, and magnetic susceptibility data.

Organic geochemical signatures from both PAD 54 and PAD 15 sediment cores are similar, characterized by high-frequency fluctuations (Figure 4a, b). Overall, sediments are low in organic content with carbon and nitrogen values $<3\%$ and $<0{\cdot}3\%$, respectively. C/N ratios range from 8 to 15. Bulk organic δ^{13}C values range between -28 and -25‰. Stratigraphic variability in organic geochemistry is comparable to that described by Brown *et al.* (2000), with decreases in carbon and nitrogen content associated with the high-energy events as defined above (e.g. high magnetic susceptibility; see Figure 4a, b, and Appendix Table I for respective linear correlation coefficients), likely indicating minerogenic dilution and reduced algal productivity due to elevated turbidity of the lake water. Increases in C/N ratios also characterize the high-energy events (Figure 4a, b; Appendix Table I), although we speculate that this mainly reflects influx of C-rich and N-poor fine-grained hydrocarbon compounds such as bitumen derived from oil sands upstream (see subsequent text) rather than

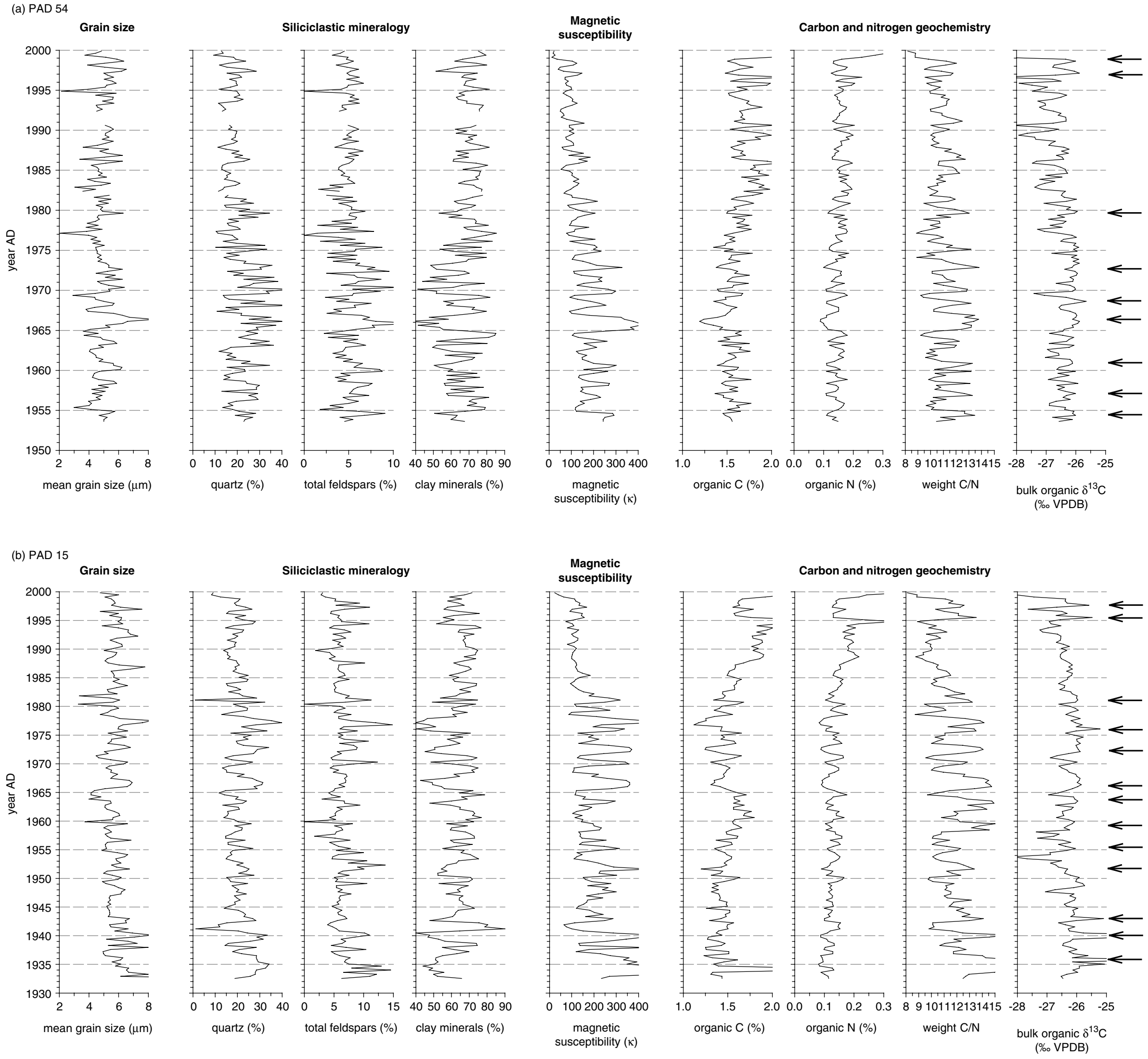

Figure 4. Mean grain size, siliciclastic mineralogy, magnetic susceptibility, and carbon and nitrogen geochemistry profiles for the upper 1 m of sediments from (a) PAD 54 and (b) PAD 15. Peaks in mean grain size, quartz content, total feldspar content, magnetic susceptibility, C/N ratios, and bulk organic $\delta^{13}C$ coincide with troughs in organic and clay mineral content and are interpreted as high-energy events characterized mainly by increased influx of detrital sediment (selected examples are identified by arrows). Intervening intervals are interpreted as low-energy events, typified by reduced detrital influx, increased lake productivity, and internal organic matter recycling within the lake water column. Breaks in mineralogy and grain size plots are samples that are not analyzed

terrestrial organic plant debris, because materials, such as wood and leaves, were virtually absent in the sediment cores (other than the dated twig). Results of $\delta^{13}C$ analyses indicate that this parameter is also sensitive to the energy conditions, with ^{13}C-enriched spikes associated with the inferred high-energy events (Figure 4a, b; Appendix Table I). These may reflect the input of dissolved inorganic carbon (DIC) with a ^{13}C-enriched signature, or perhaps it is more likely that the bitumen washed into the lake is ^{13}C-enriched relative to the aquatic organic matter that dominates the low-energy intervals, as suggested by the low C/N ratios (*cf* Meyers and Lallier-Vergès, 1999; Meyers and Teranes, 2001). Low $\delta^{13}C$ values during low-energy intervals may be caused by internal organic matter recycling within the lake water column, which contributes ^{13}C-depleted CO_2 to the DIC pool (*cf* Tyson, 1995; Hollander and Smith, 2001). High levels of recycling of organic matter, in addition to the influx of siliciclastic-dominated detrital sediment, likely account for the low organic content of these sediments.

In summary, detailed physical and geochemical analyses on the upper 1 m of sediment cores from PAD 54 and PAD 15 indicate that peaks in mean grain size, quartz content, total feldspar content, magnetic susceptibility, C/N weight ratios, and bulk organic δ^{13}C values are concurrent with troughs in clay mineral and organic (carbon and nitrogen) content. These stratigraphic intervals are interpreted as high-energy events. Conversely, troughs in mean grain size, quartz content, total feldspar content, magnetic susceptibility, C/N weight ratios, and bulk organic δ^{13}C values coincide with increases in clay mineral and organic content, interpreted as low-energy deposits formed by reduced detrital influx and increased lake productivity. These stratigraphic relationships correspond with the visible banded appearance of the sediment cores; dark-coloured strata are associated with high-energy events, whereas light-coloured strata are linked to low-energy conditions (Figure 5). We attribute this relationship to the presence of dark-coloured bitumen, as well as dark-coloured magnetic-rich minerals, in the high-energy deposits. Deposition of bitumen during high-energy events may account for the bulk organic matter ^{14}C dates measured on these beds that are contaminated with old carbon (Table I). A potential upstream source of bitumen is from oil sands eroded by the Wabasca River, a tributary of the Peace River (see Figure 1), which has also been cited as a possible source of contaminants found in fish collected from the river (Northern River Basins Study, 1996). Bitumen is contained in lower Cretaceous sandstones that subcrop in the watershed and has a very low specific gravity (5 – 7° API gravity; Kramers, 1974; Rottenfusser, 1980), which would favour suspension-settling sedimentation with river-derived clastic input to PAD 54 and PAD 15. The Wabasca River is thought to be a major supplier of suspended sediment to the Peace River during years of high runoff (Prowse and Conly, 1996).

On the basis of the ^{137}Cs-derived 50- and 70-year long records shown in Figure 4a and b, representing the upper 1 m of sedimentation at PAD 54 and PAD 15, respectively, oscillations in the various parameters do not appear to be obviously annual, and therefore it is unlikely that combinations of the dark- and light-coloured deposits represent varves (i.e. annual deposits). Rather, the most likely origin of the dark-coloured beds,

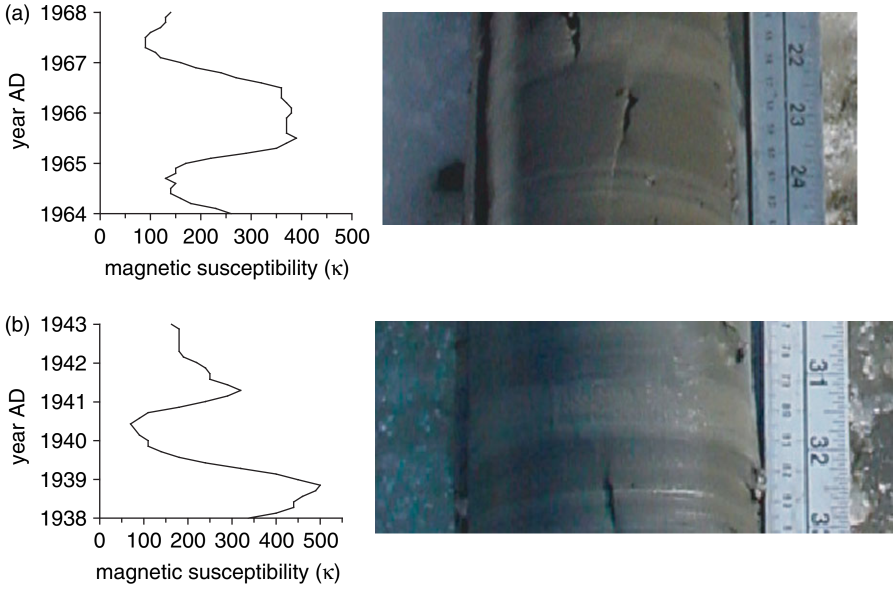

Figure 5. (a) Detail from PAD 54 sediments illustrating that peaks in magnetic susceptibility are associated with dark-coloured laminations. (b) Detail from PAD 15 sediments showing that troughs in magnetic susceptibility are associated with light-coloured laminations

characterized by parameters we interpret to reflect relatively high-energy conditions, are periodic flood events that provide an influx of mainly siliciclastic, magnetic-rich material from sediment-laden river waters that spill over the sills of these basins. Intervening light-coloured beds, which we interpret to reflect lower energy conditions, on the other hand, most likely originate from reduced input of magnetic-poor, siliciclastic material transported by local catchment runoff. An opposite scenario could be developed simply by considering that the rivers are a source of fine-grained material with low magnetic content, which would imply that the light-coloured beds represent flood events. If water levels remained deep enough to cause anoxia during non-flood periods, reduced influx of fluvial-derived fine-grained sediment could be associated with the generation of bacterially induced sulfide authigenesis resulting in elevated magnetic susceptibility signatures in the dark-coloured non-flood beds (e.g. Snowball *et al.*, 2002). However, we consider this latter scenario unlikely, given that results from the organic geochemistry analyses indicate that the organic fraction in the dark-coloured beds are also derived, at least partly, from allochthonous sources. Conversely, reduced turbidity and corresponding increased water clarity during lower-energy periods are probably key to supporting the increased autochthonous production and subsequent recycling of organic material, as inferred from slight increases in carbon and nitrogen content, and decreases in C/N ratios and bulk organic δ^{13}C values in the light-coloured beds.

Flood reconstructions

Magnetic susceptibility - a proxy for flood history. Variations in magnetic susceptibility in sediment cores from PAD 54 and PAD 15 capture the visibly laminated stratigraphy, which we interpret to reflect oscillating energy conditions between periodic flood events (dark-coloured strata) and more quiescent conditions (light-coloured strata). As described below, we use the complete magnetic susceptibility profiles from PAD 54 and PAD 15 to develop records of flood frequency after (1) applying conservative corrections to the magnetic susceptibility in the near-surface sediments to account for changes in compaction and water content of uppermost sediments (Dearing, 1999), and (2) comparing the magnetic susceptibility records to the post-1972 water level record that is available at Rocky Point on the Peace River, the nearest hydrometric station. The latter establishes that known high-water events are recorded in the magnetic susceptibility profiles, which are then used to estimate magnetic susceptibility thresholds for identifying paleoflood waters entering the oxbow lakes from down-core (pre-1972) measurements.

Increases in moisture content can dilute magnetic susceptibility measurements (Dearing, 1999; Sandgren and Snowball, 2001). In the upper part of sediment records, the moisture content naturally increases as a function of decreasing density. In the upper 1 m of sediments from PAD 54 and PAD 15, a 5-year running average of moisture content reveals a steady increase beginning at 50-cm depth (~1975) for PAD 54 and 28-cm depth (~1980) for PAD 15 as a result of lesser compaction (Figure 6). These smoothed trends are superimposed on highly fluctuating moisture content as captured in the raw data, in which lower moisture content is associated with higher magnetic susceptibility, indicating that more densely packed sediments characterize the higher-energy flood deposits (Figure 6). To correct for under-estimation of magnetic susceptibility as a result of decreasing compaction in the upper part of the sediment records (*cf* Dearing, 1999), linearly estimated increases in moisture content of 1·70% per year and 3·05% per year were applied to de-trend the magnetic susceptibility profiles post 1975 and post 1980 for PAD 54 and PAD 15, respectively (Figure 6). Notably, these magnetic susceptibility corrections are conservative in the upper few cm where the increases in moisture content are non-linear with sediment depth and, consequently, exceed the post-1975 (for PAD 54) and post-1980 (for PAD 15) linearly estimated changes.

We then compared the 'compaction-corrected' magnetic susceptibility records from PAD 54 and PAD 15 to the daily water level record measured at Rocky Point, located near the junction of the Chenal des Quatre Fourches and the Peace River (Figures 1 and 7), to determine if known high-water events are recorded in the magnetic susceptibility profiles. The Rocky Point water-level record begins at 1972, although there are several notable gaps, including those during the spring of 1972, 1974, 1996, and 1997 when major ice-jam

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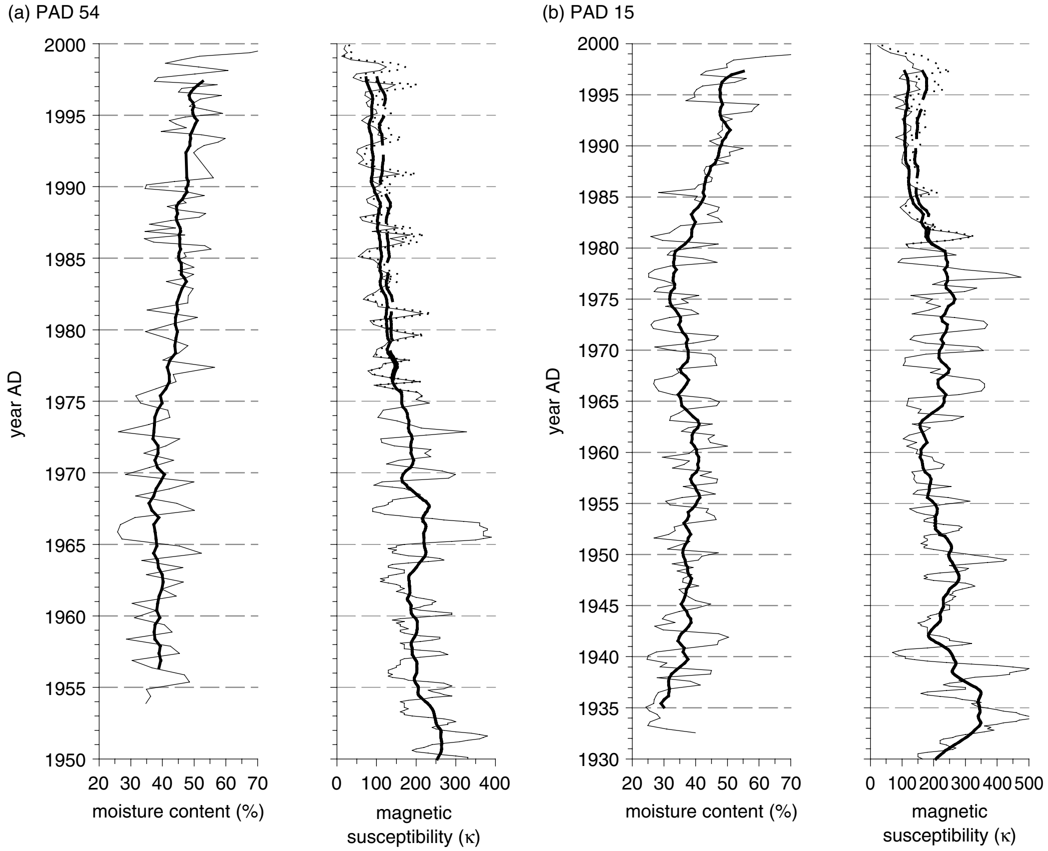

Figure 6. Moisture content and magnetic susceptibility profiles for the upper 1 m of (a) PAD 54 and (b) PAD 15. Both raw data and five-year running averages are shown. As described in the text, conservative corrections to the magnetic susceptibility profiles for dilution-effects caused by decreasing compaction in the upper part of the sediment records are shown as dotted (raw) and dashed lines (five-year running average)

events damaged the gauging station (D. Smith, University of Calgary, personal communication, 2002). Thus, it is likely that peak river levels during these years were not recorded. As shown in Figure 7a, lines are drawn to correlate water levels that approach or exceed the PAD 54 sill elevation (209·92 m) with the most closely associated peaks in the magnetic susceptibility record.

By visual inspection, it is evident that several high-water events closely coincide with peaks in the magnetic susceptibility profile performed at 0·5-cm intervals on the sediment core from PAD 54 (Figure 7a). These include the major spring ice-jam flood events of 1972, 1974, 1996 (high-water levels also occurred during the summer, which also led to flooding; Leconte *et al.*, 2001), and 1997, other reported high-water events including those of 1979 (spring ice-jam on Chenal de Quatre Fourches; Peters, 2003), 1990 (historic high open-water level measured at Peace Point; Prowse and Lalonde, 1996), and 1994 (spring ice-jam at Rocky Point; Peters, 2003), and several other lower magnitude events (1973, 1976, 1977, 1981, and 1987). Compaction-corrected magnetic susceptibility peaks associated with the flood events of 1996 and 1997 are underestimated because of very high water content in these uppermost sediments. As illustrated by the dashed line in Figure 7a, the two peaks in the magnetic susceptibility profile may be due to the 1974 (calendar year) ice-jam flood event because no other significant high-water levels can be correlated with the peak in C/N ratio and magnetic susceptibility at or about 1975 (estimated from ^{137}Cs dating). The largest peak in magnetic susceptibility record since 1970 coincides with the maximum C/N ratio, and is considered to be associated with the primary

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influx of detrital material during the 1974 ice-jam flood that caused widespread flooding of the Peace sector of the PAD (Pietroniro *et al.*, 1999). A similar comparison was made between the PAD 15 magnetic susceptibility record and the Rocky Point water level record (Figure 7b). The sill elevation for PAD 15 is unknown and therefore the PAD 54 sill elevation was tentatively used to identify the correlations between water levels that approach or exceed this elevation and peaks in the magnetic susceptibility record (Figure 7b). Allowing for minor stratigraphic offsets that may be associated with fluctuating sedimentation rates, the major spring ice-jam flood events of 1972, 1974, 1996, and 1997 also appear to be recorded in the PAD 15 magnetic susceptibility profile, as well as the other events described above, including those of 1973, 1976, 1977, 1979, and 1994. Overall, however, fewer numbers of water level peaks are recorded in the PAD 15 magnetic susceptibility profile compared to PAD 54, especially between 1980 and 1996. This does not appear to be an artifact of insufficient compaction-correction to the magnetic susceptibility profile (exceptions include flood events in 1996 and 1997) because, generally, low C/N ratios also occur during this interval and shifts in several other parameters (e.g. increase in clay mineral and carbon and nitrogen content) also indicate that this change is due to a period of reduced river influx into the basin (Figures 4b, 7b). Consequently, our results suggest that PAD 54 may be more prone to flooding than PAD 15, in line with our recent observation of river water entering PAD 54 but not PAD 15 during the summer of 2001, and may be further evidence that the sill elevation is higher at PAD 15 than at PAD 54. Alternatively, or additionally, the Chenal des Quatre Fourches (the river source for PAD 54) may more frequently experience high water compared to the Revillon Coupé (the river source for PAD 15) because the former connects with the Peace River immediately upstream of a 180° meander bend where ice-jams tend to develop (Prowse and Lalonde, 1996). Evidence for reduced river flooding at PAD 15 between 1980 and 1996 is nonetheless consistent with local observations during this period of substantial drying of many elevated, perched basins beyond the direct influence of the river channel network in the PAD (Prowse and Conly, 1998). As is the case for PAD 54, the largest peak in the post-1970 magnetic susceptibility record is likely associated with the primary influx of detrital material during the 1974 ice-jam flood and which is also correlated with a C/N ratio maximum (Figure 7b).

Close correspondence between peaks in magnetic susceptibility in the upper strata and high-water events recorded since 1972 at Rocky Point on the Peace River provide the basis for using down-core magnetic susceptibility measurements to identify previously unrecorded high-water events (Figure 8). Sill threshold values were estimated from Figure 7 for both PAD 54 and PAD 15, determined from the lowest magnetic susceptibility peaks that are associated with minor flood events. Because the 1974 ice-jam flood caused widespread inundation of the northern Peace sector of the PAD (Pietroniro *et al.*, 1999), the peak magnetic susceptibility value associated with this flood (Figure 7) was used to gauge the frequency of similar and larger events in the past.

Extrapolation of sill elevation and 1974 flood magnetic susceptibility threshold values prior to 1972 for PAD 54 and PAD 15 must be considered as preliminary because we have no means of estimating historical changes in sill elevations resulting from levee deposition and erosion. Continuous levee build-up (or erosion) could presumably lead to long-term apparent trends towards lower (or higher) flood frequency due to progressively higher (or lower) river water levels required for flooding, but no such trends are observed in the PAD 54 and PAD 15 magnetic susceptibility records (Figure 8). Rather, these records suggest that floods of varying magnitudes have persistently influenced PAD 54 and PAD 15, as indicated by nearly continuous, but highly variable, records of magnetic susceptibility values that exceed the estimated sill threshold values (Figure 8). Exceptions include very low magnetic susceptibility intervals from ∼1850 – 1865 at PAD 54 and from ∼1710 – 1730, ∼1745 – 1755, and ∼1760 – 1770 at PAD 15, which we interpret to reflect pre-1972 periods of particularly low flood frequency. Notably, the length of these intervals are likely underestimated because sedimentation rates during non-flood periods are undoubtedly lower than the long-term average. The lack of observable stratigraphic evidence for subaerial exposure or desiccation implies that input from local catchment runoff and precipitation were evidently sufficient to maintain water in these basins during these intervals of low flood frequency.

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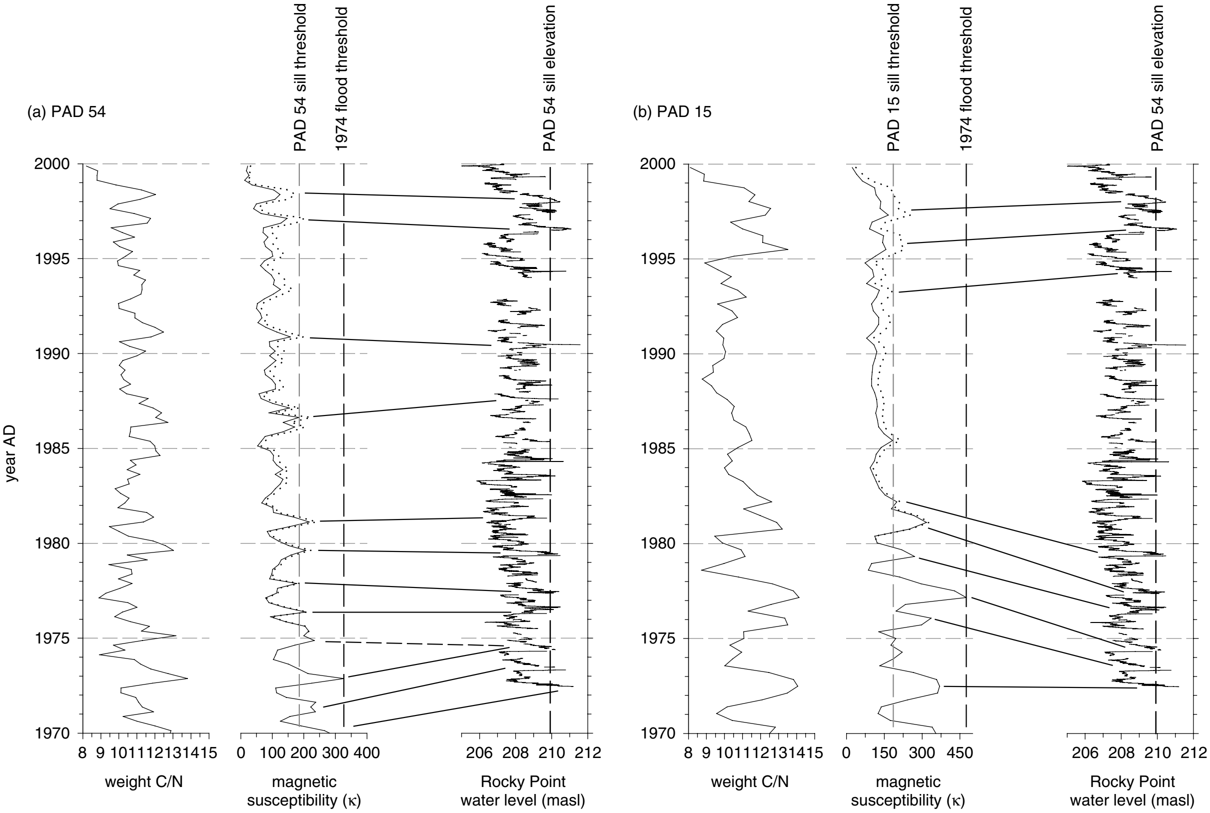

Figure 7. Comparison of the post-1970 magnetic susceptibility (raw data and compaction-corrected - see Figure 6) profiles for (a) PAD 54 and (b) PAD 15 and the Rocky Point daily water level record available since 1972. Lines are drawn between water levels that approach or exceed the PAD 54 sill elevation (209·92 m) and the most closely associated peaks in the magnetic susceptibility records. The sill elevation for PAD 15 is unknown and therefore the PAD 54 sill elevation was tentatively used to identify correlations between water levels that approach or exceed this elevation and peaks in the PAD 15 magnetic susceptibility record. Note that corrected magnetic susceptibility peaks for the 1996 and 1997 flood events in both records are underestimated because of rapid increase in water content in the uppermost sediments. Weight C/N ratio records (see Figure 4) are also shown for comparison

According to the PAD 54 record and utilizing the threshold magnetic susceptibility value associated with the 1974 flood, 25 major flood events of equal or greater magnitude than the 1974 flood have occurred at this site over the past ~180 years (Figure 8a). This corresponds to an average return time of 7·2 years, which is in remarkably close agreement with the Traditional Knowledge and historical record of ice-jam flood frequency (28 major floods between 1826 – 2000 with a return time of 6·2 years; Timoney, 2002). However, results indicate marked fluctuations in flood frequency over the past 180 years at this site. In general, flood frequency is low during the early to mid-1800s, rises in the late 1800s, attains a maximum during the early part of the twentieth century, and declines after ~1930, ~1950, and ~1975. The 21-year absence of a major ice-jam flood event between 1975 and 1995 and the 1968 – 1971 interval, which corresponds to the filling of the Williston Lake reservoir and which appears to be captured by a trough in the 5-year running average profile, are not unprecedented during the past ~180 years. Notably, extended multi-decadal intervals without major flooding of PAD 54 are recorded during the early to mid-1800s. In particular, the sediments record an absence of a 1974 – equivalent flood during a ~35-year period (~1836 – 1871), which is similarly recorded at PAD 15 (see below).

Although the PAD 15 record is substantially longer than PAD 54, far fewer (14 total) major flood events (i.e. equivalent to or greater than the 1974 flood) appear to have occurred at this site (Figure 8b). As discussed

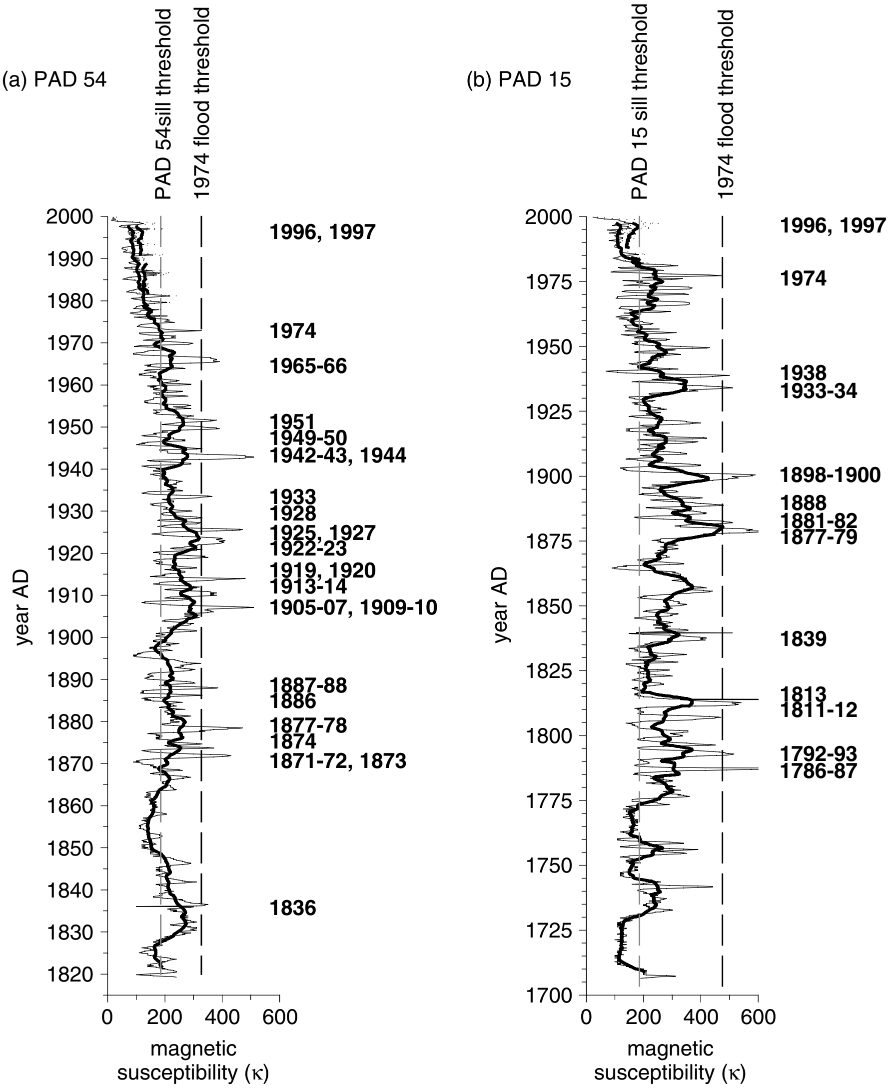

Figure 8. Inferred flood record for (a) past ~180 years at PAD 54 and (b) past ~300 years at PAD 15 from magnetic susceptibility. Chronology is developed from an average linear sedimentation rate of 2·0 cm year^{-1} for PAD 54 and 1·39 cm year^{-1} for PAD 15. Raw and compaction-corrected data are shown (black solid and dotted lines), which are analyses at sub-annual resolution. Five-year running averages based on the raw and compaction-corrected data are also shown (solid and dashed lines). Estimates of sill and 1974 flood thresholds are from Figure 7. Years along the right-hand side of the graphs are estimated dates of flood events that are equivalent to or exceed the 1974 flood threshold (note that 1996 and 1997 are included in these lists)

above with respect to the most recent part of the record, this reduced susceptibility to flooding may reflect that the sill elevation is higher at PAD 15 than at PAD 54 or the location of PAD 54 is more prone to flooding. Overall, results from PAD 15 also indicate marked fluctuations in flood frequency characterized by oscillating decadal-scale intervals of high and low flood frequency. Intervals of particularly high flood frequency include ~1785 – 1815 and ~1875 – 1900. The latter period is comparable to the ~1871 – 1890 period of high inferred flood frequency at PAD 54. As mentioned above, exceptionally low flood frequency occurred during several intervals in the 1700s (i.e. ~1710 – 1730, ~1745 – 1755, ~1760 – 1770). Twentieth century declines in flood frequency occurred after ~1900, ~1935, ~1950, and ~1975. Notably, several 20+-year intervals between major flood events occur throughout the record at PAD 15, which are comparable to the 1975 – 1995 period, including ~1839 – 1877 as observed in the PAD 54 magnetic susceptibility record.

Comparison with the Traditional Knowledge- and historical-based flood record. We compared the flood records inferred from magnetic susceptibility profiles from PAD 54 and PAD 15 to the Traditional Knowledge- and historical-based Peace River spring ice-jam flood record published by Timoney *et al.* (1997). The latter was derived from a number of different written sources including Hudson Bay Company archives and Catholic church records from Fort Chipewyan, Wood Buffalo National Park warden records, hydrometric records of Lake Athabasca, PAD water level records, as well as Fort Chipewyan oral history (Peterson, 1995). A database was developed in which the years with high-magnitude floods were assigned a value of 1, whereas years that were absent of floods or that experienced only minor flooding were assigned a value of 0. From these data, Timoney *et al.* (1997) produced a continuous record of flood frequency as shown in Figure 9. Allowing for some minor stratigraphic offset due to possible variability in sedimentation rates, a strong correspondence exists between sediment-inferred flood events of ≥1974 magnitude and intervals exceeding the long-term average flood frequency based on the Traditional Knowledge- and historical-based record. In particular, corresponding periods of frequent flooding (e.g. late 1800s to mid-1900s) and less frequent flooding (e.g. mid-1800s) are evident in all records. This lends considerable support to both the estimated sediment core chronologies for PAD 54 and PAD 15 as well as the integrity of the Traditional Knowledge- and historical-based flood record. While the precise dates of individual flood events may differ, the number of 1974-equivalent or greater floods in the more flood-prone PAD 54 magnetic susceptibility record (25 between 1819 and 2000) and the number of high-magnitude floods in the Traditional Knowledge- and historical-based flood record (28 between 1826 and 2000; Timoney, 2002) over nearly the same time interval are also in close agreement, as mentioned above.

CONCLUSIONS

Reconstructed flood histories from magnetic susceptibility measurements on laminated sediments in oxbow lakes PAD 54 and PAD 15 in the northern Peace sector of the PAD extend back in time by ~180 and ~300 years, respectively. Although our chronological analyses cannot account for the likelihood of variable sedimentation rates in these deposits, results based on long-term average sedimentation rates are in close agreement with the Peace River ice-jam flood record based on Traditional Knowledge and historical sources over the period of overlap (see Figure 9). The use of magnetic susceptibility, in select combination with other more laboratory-intensive physical and geochemical analyses, has shown to be a particularly sensitive and efficient tool for recognizing known events and identifying previously unrecorded flood events in these highly minerogenic sediments. These data provide the necessary framework for addressing the two research questions posed in the Introduction.

Has flood frequency declined since the Peace River was regulated in 1968?

According to the ~180-year flood record generated from PAD 54, maximum flood frequency occurred during the first third of the twentieth century (see Figure 8). Between ~1900 and ~1935, 11 major floods were recorded at this site, and only eight major floods occurred during the remainder of the twentieth century (including 1996 and 1997) with half of these occurring during a 10-year period from about the early 1940s to early 1950s. Notably, this record shows an estimated 14-year hiatus of major floods between ~1951 and ~1965, immediately prior to the construction of the WAC Bennett Dam in 1968. At the apparently less flood-prone PAD 15 site, maximum flood frequency is inferred to have occurred earlier between ~1875 and ~1900 when four major floods can be identified (see Figure 8). Only 5 major floods can be distinguished (including 1996 and 1997) during the twentieth century. Although differences in the timing of maximum

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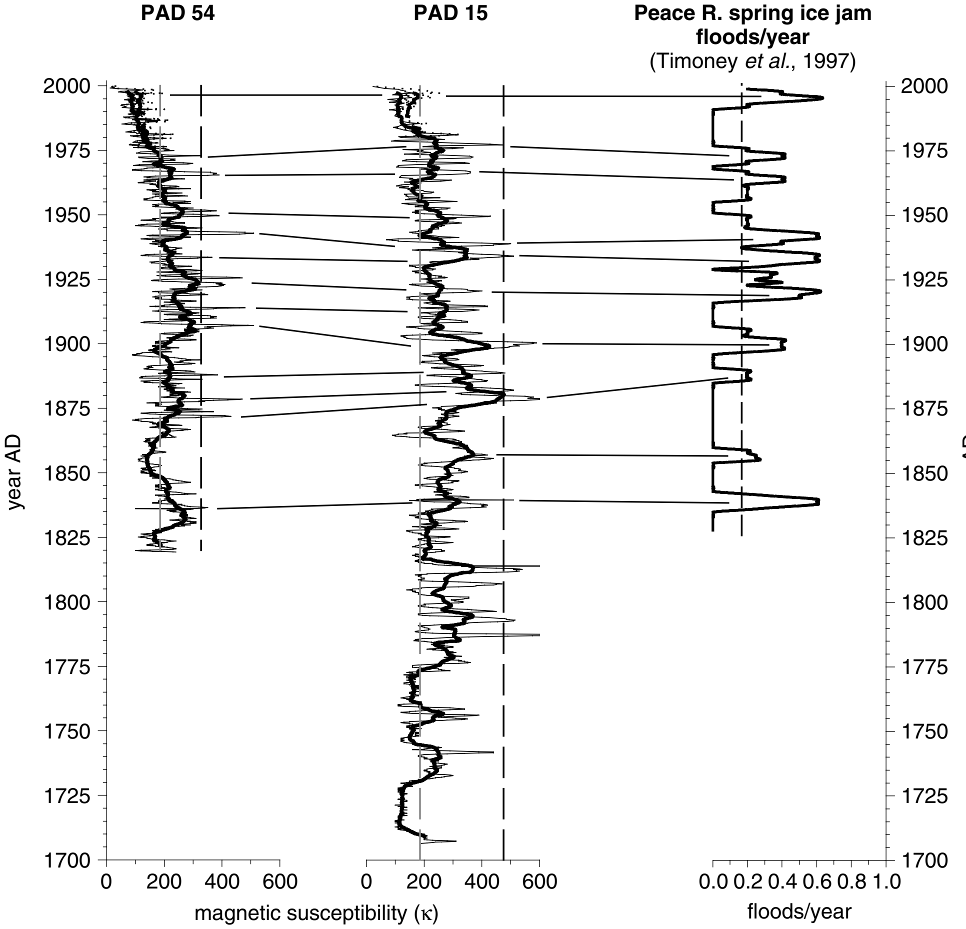

Figure 9. Comparison of magnetic susceptibility inferred flood records at PAD 54 and PAD 15 (raw data and five-year running average, as presented in Figure 8), and the five-year running mean of Peace River spring ice-jam flood frequency derived from Traditional Knowledge and historical sources (from Timoney *et al.*, 1997). Vertical dashed lines in PAD 54 and PAD 15 sediment records represent estimates of sill and 1974 flood thresholds, as in Figure 8. Vertical dashed line in the historical record is the long-term average flood frequency. Lines are tentatively drawn to correlate the sediment-based and historical flood records

flood frequency at these two sites are currently difficult to resolve, these data suggest that flood frequency had been in decline for several decades preceding Peace River regulation, beginning in the early to mid-twentieth century (at PAD 54) or perhaps as early as the late nineteenth century (at PAD 15). Allowing for the fact that magnetic susceptibility peaks associated with the 1996 and 1997 floods are underestimated because of the non-linear increase in water content in the uppermost sediments, no clear post-1968 inflection in flood frequency is evident in these records. While the 1980s through to the early 1990s stands out as an interval absent of flooding, on the basis of magnetic susceptibility measurements, additional detailed multi-proxy analyses spanning the past 50 and 70 years at PAD 54 and PAD 15, respectively, also fail to show substantial directional changes post-1968 (see Figure 4).

Have prolonged multi-decadal intervals (20+ years) between major floods (similar to the 1975 – 1995 interval) previously occurred under the influence of climate variability?

Several pre-1974 multi-decadal intervals without a major flood are evident in the sediment records from PAD 54 and PAD 15. For example, both PAD 54 and PAD 15 record a 35-year interval in the mid-1800s

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during which no major floods occurred. The longer record available at PAD 15 indicates several additional 20+ year periods without a major flood, including ∼1813 – 1839 and ∼1705 – 1786.

Implications for ecosystem management

These results indicate that flood frequency in the northern Peace sector of the PAD has been highly variable over the past ∼180 – 300 years and suggest that changes in hydrology are a natural feature of this ecosystem, independent of human influence or intervention. References to flood return times can be misleading and are not a particularly useful characterization of the variability, as the evidence at flood-prone PAD 54 suggests that major floods have occurred every 1 – 6 years over brief time intervals while during other periods these events have been separated by several decades. Variability in flood frequency is likely even more accentuated because we rely on the use of linear sedimentation rates to establish sediment core chronologies. Nonetheless, recognition of this inherent variability in flood frequency is critical for ongoing stewardship of water and ecological resources of the PAD by community-, government-, and industry-based stakeholders, as management decisions based on a single average flood return time may not be appropriate.

While our analysis provides the needed context for placing the past 35 years of flood history into a longer-term perspective, adding effectively to the Traditional Knowledge- and historical-based Peace River spring ice-jam flood reconstruction (Timoney *et al*., 1997), resolving whether significant river regulation impacts are superimposed upon natural climate-driven hydrological variability over this time interval remains uncertain. Further research on developing and integrating independent quantitative paleoclimate records are required to partition the relative roles of these stressors (Hall *et al*., 1999), as well as for generating future projections of flood frequency under climate change and variability scenarios and continued river regulation. Notably, available upstream paleo-precipitation data from tree-ring studies for the Banff-Jasper-Foothills area indicate that the ‘Little Ice Age’, an interval spanning the mid-1500s to late 1800s, was characterized by persistent dryness bracketed by precipitation maxima around 1530 – 1550 and 1890 – 1910 (Luckman and Watson, 1999; Watson and Luckman, 2001). The latter interval is in good agreement with our flood history reconstructions. Additionally, tree-ring reconstructions of streamflow for the North Saskatchewan, South Saskatchewan, and Saskatchewan rivers indicate prominent drought periods during the 1700s and mid-1800s (Case and MacDonald, 2003), which correspond with low flood frequency intervals in our sediment records, as does evidence for periodic desiccation in an elevated perched basin during the 1700s in the northern Peace sector of the PAD (Wolfe *et al*., 2005). While precipitation and river discharge are just two of the many factors that influence spring break-up conditions, these correlations are suggestive of a mechanistic linkage between regional climate variability and flood frequency in the Peace sector of the PAD. However, additional paleoclimate records are needed to determine if local and regional climatic conditions during the low flood frequency conditions, which characterized portions of the 1700s and mid-1800s, are appropriate analogs for low flood frequency during the late twentieth century.

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APPENDIX

Table A1.

	Clay	Silt	Sand	Mean	Med	Quartz	∑Feld	∑Clay	∑Silici	Mag	C	N	C/N	δ^{13}C
(a) Linear correlation coefficients for PAD 54 data (values significant at $\alpha = 0{\cdot}001$)														
OM			−0·3	−0·3		−0·3			−0·32					
	Clay	−0·95	−0·82	−0·3	−0·63	−0·32								
		Silt	0·77	0·65										
			Sand	0·35	−0·32	0·32	0·52	−0·38						
				Mean	0·63	0·38	0·65	−0·4		0·3			0·42	
					Med	0·36				0·34				
						Quart	0·48	−0·77	0·58	0·65	−0·38	−0·44	0·5	0·32
							∑Feld	−0·5	0·31	0·48		−0·32	0·52	
								∑Clay	0·58	−0·66	0·35	0·45	−0·66	−0·36
									∑Silici					
										Mag	−0·51	−0·55	0·65	0·45
											C	0·95	−0·5	−0·81
												N	−0·72	−0·84
													C/N	0·59
(b) Linear correlation coefficients for PAD 15 data (values significant at $\alpha = 0{\cdot}001$)														
OM						−0·65	−0·46	0·59		−0·71	0·64	0·74	−0·55	−0·53
	Clay	−0·98	−0·8	−0·66	−0·61									
		Silt	0·8	0·62	0·57									
			Sand	0·77	0·75									
				Mean	0·97	0·31	0·37			0·37				
					Median	0·33	0·36			0·42				
						Quart	0·51	−0·86		0·78		−0·52	0·66	0·45
							∑Feld	−0·63		0·58		−0·41	0·48	
								∑Clay	0·63	−0·78		0·5	−0·66	−0·42
									∑Silici	−0·3				
										Mag	−0·43	−0·67	0·76	0·51
											C	0·85	−0·3	−0·39
												N	−0·73	−0·55
													C/N	0·57

Parameters shown include: OM = organic matter; Clay = clay grain size; Silt = silt grain size; Sand = sand grain size; Mean = mean grain size; Med = median grain size; Quart = quartz mineral; ∑Feld = total feldspar minerals; ∑Clay = total clay minerals; ∑Silici = total siliciclastic minerals; Mag = magnetic susceptibility; C = organic carbon; N = total nitrogen; C/N = weight C/N ratio; δ^{13}C = organic carbon isotope composition