Geoscientific contributions to understanding flood hazards in the Red River Valley, Manitoba

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ABSTRACT

Geoscience research is enhancing the flood history of the Red River and assessing the relevance of geological processes to the flood problem. Tree-ring and historic records of flooding indicate that major Red River floods in Manitoba over the last three and a half centuries have been clustered in the mid 18th century, the early to mid 19th century and the latter half of the 20th century. Tree-ring evidence also indicates that the Red River flood of 1826 was the most severe flood during the last 352 years. Dated organic materials contained in floodplain cores from near St. Jean Baptiste, Manitoba, reveal that the river channel has been migrating laterally at an average of about 0.04 m/yr over the past 1000 yr. This is slowly widening the river valley, but not significantly affecting the flood hazard in the short term. Rates of differential crustal rebound derived from lake gauge and paleo-shoreline data indicate that about half of the river gradient has been lost over the past 8000 years in the Emerson-Morris reach.

INTRODUCTION

The mitigation of Red River floods at Winnipeg has been addressed from a structural approach that includes the Red River Floodway, Portage Diversion, Shellmouth Dam and many kilometres of primary dyking (Mudry et al., 1981).

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This flood protection infrastructure has been very effective since the opening of the Floodway in 1968, and billions of dollars in flood damages have been avoided, particularly during the 1997 flood (Mudry et al., 1981; International Red River Basin Task Force, 2000). During this flood, the Floodway was utilized to slightly above design capacity (Rannie, 1998), but a minor change in weather conditions could have resulted in a catastrophic dyke failure causing large-scale flooding in Winnipeg (International Red River Basin Task Force, 2000). To reduce the risk of a multi-billion dollar flood disaster, recent studies have proposed enhancing the flood protection infrastructure at Winnipeg, specifically by expanding the Floodway or constructing a detention dam near St. Agathe (International Red River Basin Task Force, 2000; KGS Group, 2001). A key aspect of enhancing the flood protection is defining an adequate design discharge to mitigate against future floods.

For design purposes, the probability of flooding is generally estimated from the record of measured annual peak flows. However, the extrapolated return periods of extreme floods can be highly uncertain when the instrumental record: 1) is short in length; 2) contains flows from a mixed population of flood mechanisms; and/or 3) inadequately represents extreme floods generated by unusual events. The flood-frequency approach also assumes that flood-generating processes do not change over time and that floods are random in time and space (Baker et al., 2002), even though it is well known that flood frequencies are affected by, for example, climate variations (e.g., Ashmore and Church, 2001). Furthermore, landscape changes from geomorphic and/or geological processes can gradually alter channel or valley discharge capacities, thereby increasing or decreasing the stage associated with a given magnitude of flow. The understanding of flood recurrence and changes in flood hazard, however, can be enhanced through research that provides a long-term perspective on flood history and processes influencing flooding.

This paper reviews ongoing geoscientific research into the Red River flood hazard conducted by the Geological Survey of Canada and Manitoba Geological Survey. Significant outputs from this research include an enhanced flood history for the Red River, an assessment of changes to the valley discharge capacity caused by lateral channel migration and incision, and an extended hydroclimatic record for south-central Manitoba.

Paleoflood records for the Red River

Paleofloods are high flows that occurred in the past and were not recorded by direct or indirect human observation (Baker et al., 2002). These events can be investigated using natural archives, such as tree-ring records and lake deposits, in a wide variety of environmental settings. Paleoflood records can reduce the uncertainty in estimates of long return-period floods and add a historical component to flood hazard assessment (Baker et al., 2002).

Dendrochronology and Red River flooding

Flood rings and the Red River tree-ring network: Prolonged inundation from high magnitude Red River floods, such as the 1950 flood or larger, cause bur oak trees (*Quercus macrocarpa* Michx.) growing along the river to develop anatomically distinctive tree-rings, or 'flood rings', that can be used to identify older and previously unknown Red River floods (Fig. 1; St. George and Nielsen, 2000; St. George et al., in press). Flood rings contain anomalous shrunken vessels and other anatomical abnormalities within its earlywood (or spring wood) layer (Fig. 1). Key factors controlling the formation of flood rings include 1) a threshold stage of flooding that inundates the tree trunk, 2) the duration of submergence, and 3) the timing of the flood relative to the growing season (St. George et al., in press). This approach is particularly well suited for the study of Red River paleofloods because of the long duration (many weeks) of major floods, and the fact that oak trees growing adjacent to the river are located only on or immediately below the level of the prairie, which prevents the formation of flood rings from lesser flows.

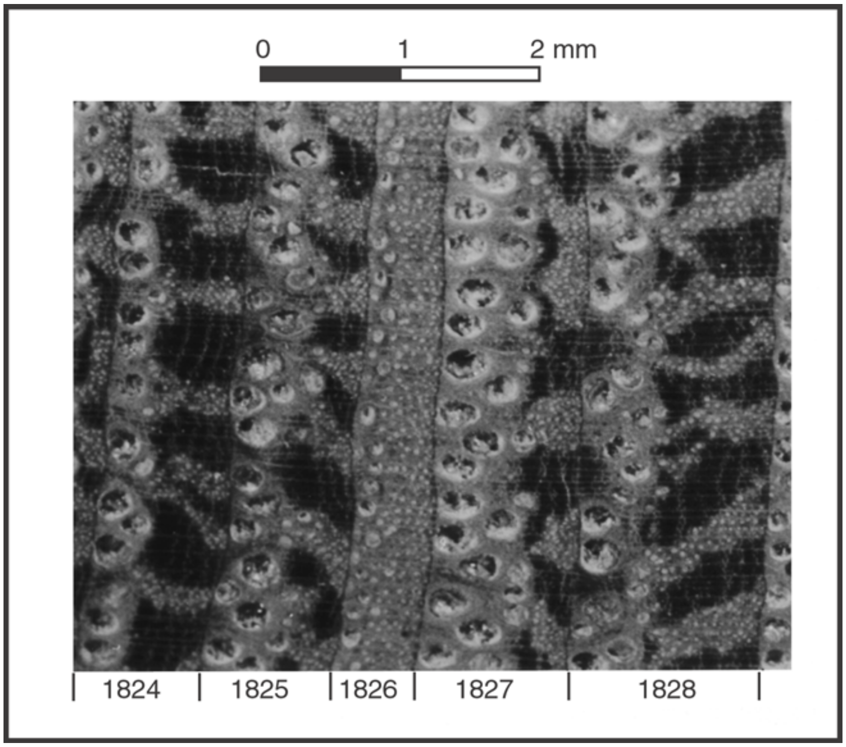

Fig. 1 *Photograph of the 1826 flood ring in a bur oak tree (Q. macrocarpa) (after St. George and Nielsen, in press). The flood signature character is evident by the size of the earlywood vessels which are strikingly smaller than those of the adjacent tree-rings. Although in this case the width of the 1826 tree-ring is narrower than the adjacent rings, this is not diagnostic of a flood ring.*

A network of tree-ring sites was developed for the Red River valley, Manitoba (St. George and Nielsen, in press). Samples were collected from living oak trees collected at sixteen sites along the Red River, oak timbers recovered from nearly a dozen local historical buildings and Euro-Canadian archaeological sites, and subfossil oak logs recovered from the river banks of the Red and Assiniboine rivers that were exposed during low flow stages (Fig. 2). The combined tree-ring record for the southern Manitoba includes 398 cross-dated trees and extends from AD 1286 to 1999 (St. George and Nielsen, in press).

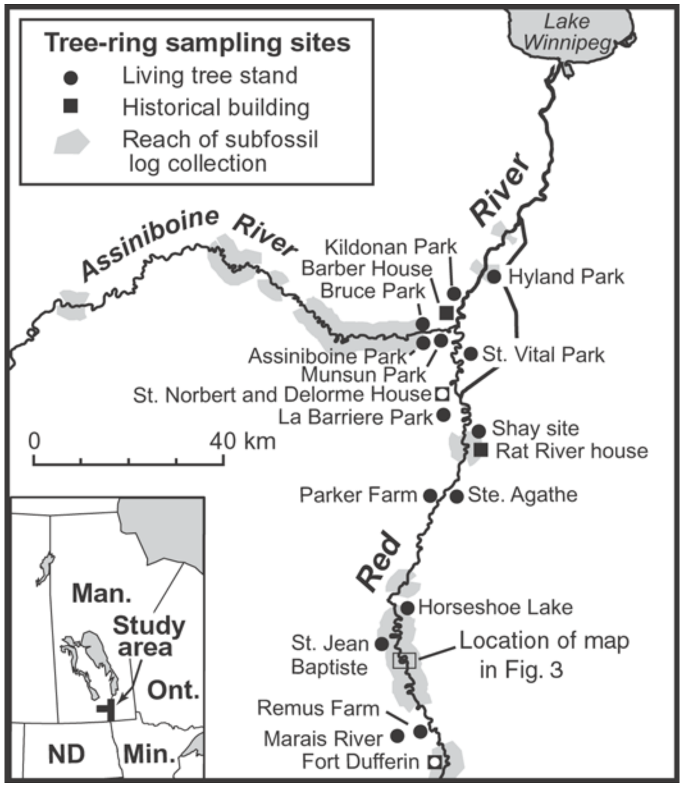

Fig. 2 *Map showing the oak tree (Q. macrocarpa) sampling sites along the Red and Assiniboine rivers in southern Manitoba (after St. George and Nielsen (submitted)).*

Flooding in the lower Red River valley since AD 1648: Tree rings provide a record of flooding for the lower Red River (LRR), between Winnipeg and Morris, that extends from AD 1648 to 1999 (St. George and Nielsen, submitted). Flood rings document seven high-magnitude floods during that period: 1997, 1979, 1950, 1852, 1826, 1762 and 1747. The five most recent flood rings are coincident with known high-magnitude floods, but the signatures in 1747 and 1762 predate local instrumental and historical flood records and represent previously unknown floods. The frequency of occurrence and anatomical development of flood signature in the 1826 ring suggest that the Red River flood of that year was exceptional and was the most severe flood to have occurred in at least the last 352 years. Despite comments by an early fur trapper who suggested flooding was unusually extensive in 1776 (Ross, 1856), there is no tree-ring evidence for such an event (St. George and Nielsen, submitted).

Assiniboine and upper Red River floods during the last 500 years: Flood rings are also contained in tree-ring samples collected along the Assiniboine River and the subfossil logs collected along the banks of the Red River between Morris, Manitoba and the Canada-US border. Subfossil logs collected north of the Canada-US border likely originated from the upper portion of the Red River basin in Minnesota and North Dakota and floated into southern Manitoba prior to burial. Although information for the upper Red River is derived from relatively few trees, its flood record extends from AD 1997 to 1448 and contains flood rings in AD 1762, 1747, 1741, 1727, 1726, 1682, 1658, 1538 and 1510. Two floods in the upper Red River (URR) record also affected the LRR, but the limited number of samples from the URR makes it difficult to determine if the other events were the result of localized flooding along a tributary or upper reaches of the Red River, or from high magnitude flooding along the lower Red River. Most of the post-1648 floods in the URR record are interpreted to have been localized events since they did not produce flood-rings along the northern portion of the Red River. Although the interpretation of the Assiniboine River samples is also limited by the number of sub-fossil trees recovered, tree-ring flood signatures suggest that the Red and Assiniboine basins, which have markedly different hydroclimates, both experienced spring flooding in 1826, 1538 and 1510. The Assiniboine and Red River floods of 1826 also coincided with severe frost damage recorded in *Q. stellata* and *Q. alba* in the south-central United States (Stahle, 1990), which implies that unusual spring weather in that year extended throughout central North America.

Red River floods during the last millennium

Millennium-scale paleoflood records possibly are preserved in the lake bottom sediments of the southern basin of Lake Winnipeg and in the basins of small lakes located on the Red River floodplain. To investigate this in Lake Winnipeg, 15 sediment cores, each about 1.5 m long, were collected from five sites in Lake Winnipeg in August 1999 (Lewis et al., 2001). Radiocarbon dating, paleomagnetic, radiochemical and palynological data reveal an intact stratigraphy in the cores that spans the past 1100 yr. Over this period, sedimentation rates are estimated to have averaged 1.4-1.8 $mm.a^{-1}$, and changes in pollen reveal gradual shifts in regional climate. In the upper portion of the cores, increases in weed pollen, macrofossil abundance, net grain size, concentrations of Cd, Cu, Hg, Pb, P, U and Zn, as well as changes in organic matter and isotopic properties, record anthropogenic and possibly climatic impacts over the past century. With respect to a paleoflood record, the most promising data are textural analyses, which contain fifteen to twenty strata, 1-5 cm thick, of abruptly higher silt content. The age of several of these beds in the upper portion of the cores corresponds closely with the age of known historic Red River floods, while the presence of additional beds lower in the cores may represent evidence of older floods. The origin of the beds currently is being assessed to determine conclusively whether their formation relates to Red River floods, or possibly other mechanisms, such as, dust or lake storms.

Similar to the Lake Winnipeg work, the deposits of three small lakes on the Red River floodplain, Horseshoe Lake and Lake Louise, Manitoba, and Salt Lake, North Dakota, were cored and analyzed for a flood record (Medioli, 2001). Chronological, textural and palynological data reveal a marked increase in sedimentation rates in all three lakes following the introduction of western agricultural practices. Detailed analyses of the texture, mineralogy and geochemistry have been completed, and reveal significant changes in lake chemistry and biota that are attributed to the effects of modern agriculture, for example, increases in phosphorus and nitrogen (Medioli, submitted). Analysis examining downcore changes in diatoms and thecamoebians is underway and may reveal evidence of past flood events. This work is based on the application of a biostratigraphic flood signature that has been defined using flood sediments sampled from the flood zones immediately following the 1997 and 1999 floods (Medioli and Brooks, submitted).

Discussion

The tree-ring paleoflood record along the lower Red River (LRR) contains three periods during which there is a grouping of Red River high-magnitude floods: the mid 1700s, the early to mid-1800s and the latter half of the 20th century. Conversely, the record also indicates that the LRR experienced prolonged intervals with little to no extreme flooding, particularly between 1648-1746, 1763-1810 and 1862-1949. This irregular occurrence of flooding, which extends from several decades to nearly a century, implies that the risk of flooding changes over time. While the interpretation of past changes in flood frequency must be conditional due to the limitations of the paleoflood record, both the paleoflood and historical-instrumental records demonstrate that there have been long intervals without any extreme Red River floods. Statistical analysis of the instrumental/historical flood series for the Red River indicates that high flows over the last 200 years have been clustered (Booy and Morgan, 1985) and non-stationary (Burn and Goel, 2001). Therefore, the risks of future flooding for Winnipeg and other communities in the Red River valley would be better estimated using techniques that account for non-stationarity and non-randomness introduced by climatic and landscape changes.

GEOMORPHIC PROCESSES

Floodplain coring and chronology

A key geomorphic factor contributing to the Red River flood hazard is that the shallow, low gradient valley formed by the river has insufficient capacity to contain large flows (Brooks and Nielsen, 2000). To assess the relevance of fluvial geomorphic changes (erosion or aggradation) within this valley to the flood hazard on the adjacent clay plain, a borehole investigation of the floodplain was undertaken at two successive river meanders located near St. Jean Baptiste (Fig. 3; see Brooks et al. (2001) for details). Boreholes were cored in two transects (99RR1 and 99RR3) on opposite sides of the valley bottom, with each transect consisting of five boreholes (Fig. 3). The boreholes were sited to follow the path of lateral migration of the channel, as revealed by a ridge and swale pattern on the floodplain (Fig. 3).

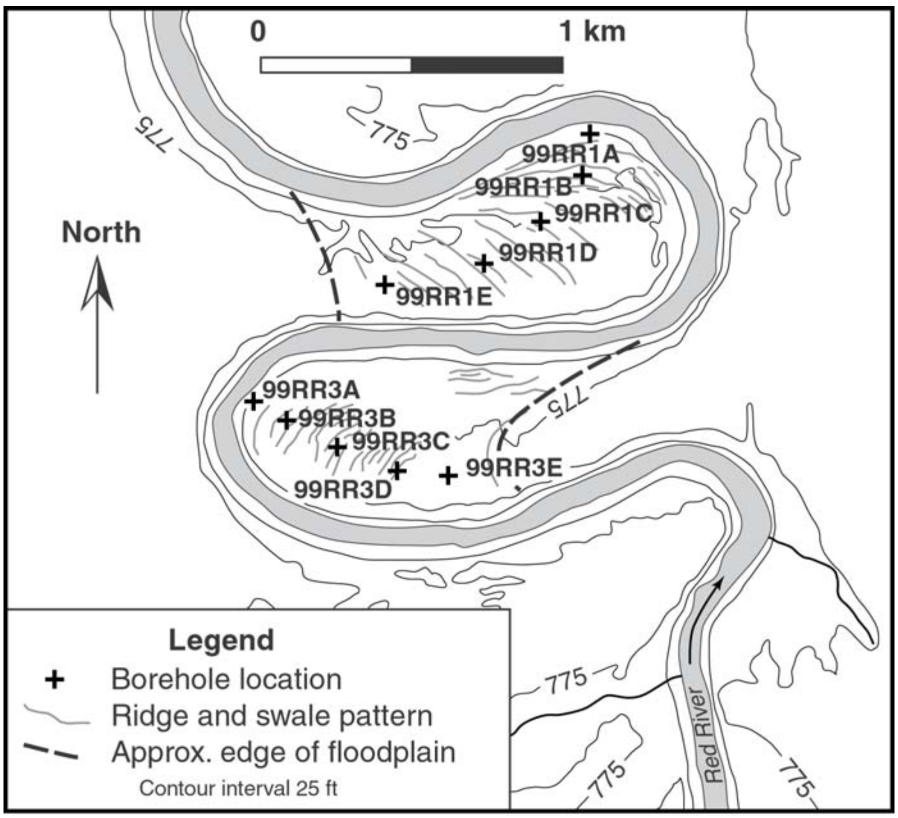

Fig. 3 *Map showing the borehole locations across the floodplain (see Fig. 2 for general location: after Brooks, submitted-a). The boreholes are located in two transects and are sited to approximately follow the path of lateral channel migration across the valley bottom. The pattern of channel migration is depicted by a pattern of ridge and swale topography preserved on the floodplain surface.*

The floodplain alluvium in the ten cores ranges in thickness from 15.3 to 21.6 m and consists primarily of silt that differs markedly in textural and structural from the glaciolacustrine deposits of the clay plain (Brooks, submitted-a). Twenty-four radiocarbon dates from nine boreholes provide chronological control on the floodplain deposits (Fig. 4; Brooks, submitted-b). Nineteen of the dates are interpreted to be representative of the age of encapsulating lateral accretion deposits in the lower portion of the alluvium. As the lateral accretion deposits would have aggraded immediately adjacent to the channel, the ages reveal past positions of the river channel over the past 8000 yr BP.

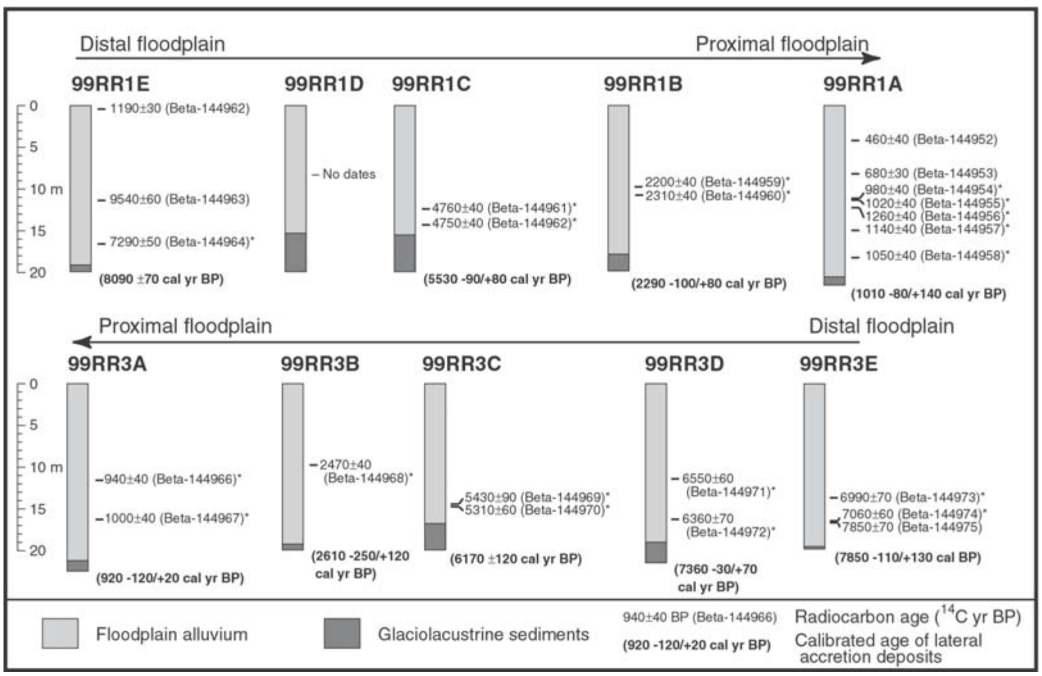

Fig. 4 *Summarized lithostratigraphic diagram of cores showing the depth locations of the 24 AMS radiocarbon ages (after Brooks, submitted-b). The boreholes are arranged from the distal to proximal areas of the floodplain with the sequencing intended to reflect the relative ordering at successive river meanders. The age at the bottom of each core (except 99RR1D) is the representative calendar age of the lateral accretion deposits in the cores. Where more than one date is present in the lower half of a borehole, the calendar ages have been averaged.*

Lateral channel migration and incision

In Fig. 5, past channel positions are depicted by 1000 yr BP isochrones. These reveal that the two meanders have extended outwards and rotated downvalley since 8000 yr BP in a single phase of lateral channel migration. The isochrone spacing along both meanders also shows that the rate of lateral channel migration was greater prior to 6000 yr BP than afterwards. Significantly, there has been appreciable lateral channel migration since 1000 yr BP, when the rate of channel migration averaged about 0.04 $m.a^{-1}$ at both meanders (Brooks, submitted-b).

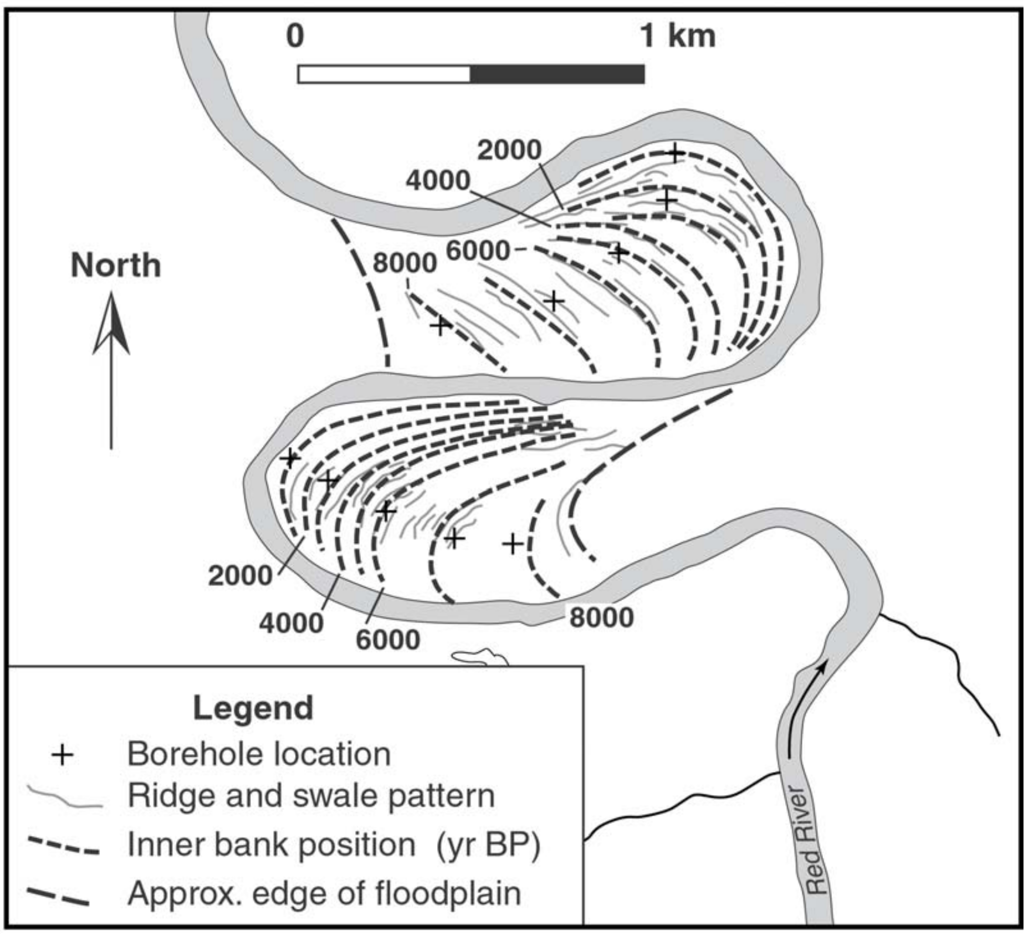

Fig. 5 *Map of the meander growth in the study area since 8000 yr BP (after Brooks, submitted-b). The positions of the 1000 yr BP isochrones have been extrapolated from the age of the lateral accretion deposits between the boreholes and placed using pattern of ridge and swale topography on the floodplain as a guide.*

Based on the difference in the depths of the base of the alluvium and the ages of the lateral accretion deposits between the 'A' and 'E' boreholes (Fig. 3), the average rates of incision at the two meanders are estimated to be 0.62 and 0.80 $m.kyr^{-1}$ for the two transects, respectively (Brooks submitted-b). These rates are marginally greater than those based on the difference in the floodplain topography, and probably are of the right order of magnitude for the river incision since 8000 yr BP.

Relevance to the flood hazard

The cross-sectional area of the valley at 1000 yr BP was estimated and compared with the modern valley to assess the significance of the lateral channel migration and incision to the flood hazard on the adjacent clay plain (Brooks, submitted-b). The valley cross-sections were measured across the apex of both valley meanders with the modern profile being based on a topographic map with a 5 ft (1.52 m) contour line interval (Red River Basin Investigation, 1951). The valley profile at 1000 yr BP was estimated by re-scaling the modern profile of both meanders to fit the position of the inner bank as defined by the 1000 yr BP isochrone. The same shape of channel was used in both profiles, thus the variation in area occurs over the floodplain portion of the valley bottom. The incision of the channel was ignored, as this rate is very low (see above).

The data reveal that the valley cross-section area (A) increased by about 2% and 0.7% at the upstream and downstream meanders, respectively, between 1000 and 0 yr BP (Brooks, submitted-b). This limited increase reflects the general low rate of lateral channel migration along both meanders, but more importantly, the very shallow depth of the floodplain portion of the valley bottom relative to the clay plain. Based on the ratio of A to wetted perimeter (P) of the profiles, the hydraulic radius (R) of the valley at 1000 and 0 yr BP decreases from 2.64 to 2.56 m and 2.01 to 2.00 m at the upstream and downstream meanders, respectively (Brooks, submitted-b). Assuming that the pre-settlement coefficient of roughness (n) and gradient (s) are constant over this interval at each meander, the change in R in the Manning equation produces a slight decrease in mean velocity (v_{mean}) of 2.1% and 0.6% at the upstream and downstream meanders, respectively (Brooks, submitted-b; Manning equation – $v_{mean} = n^{-1} \cdot s^{-0.5} \cdot R^{2/3}$). This indicates that the hydraulic efficiency of the channel at both meanders has, in fact, decreased slightly since 1000 yr BP, despite the enlargement of the valley cross-section.

As regards the discharge capacity of the valley bottom (Q), although v_{mean} decreases after 1000 yr BP, A correspondingly increases slightly at each meander because of the valley widening. This widening compensates for the decrease in v_{mean} so that Q at each meander for 1000 and 0 yr BP are essentially identical, again, assuming n and s are constant at each meander (Brooks, submitted-b). The total discharge capacity of the valley cross-section therefore is unchanged, and it can be concluded that adjustments to the valley caused by lateral channel migration have caused negligible variation to the flood hazard on the clay plain since 1000 yr BP.

GEOLOGICAL PROCESSES

Glacio-isostatic uplift

Postglacial crustal rebound resulting from the isostatic depression of the land surface by the Laurentide Ice Sheet during the late Pleistocene, has affected the regional landscape of Manitoba. This is exemplified by the differential tilting up to the north-northeast of Lake Agassiz shorelines in Manitoba, eastern North Dakota, and western Minnesota (Teller and Thorleifson, 1983), and by contemporary crustal tilting indicated by geodetic and lake gauge data (Lambert et al., 1998; Tackman et al., 1999).

The modern elevations of tilted postglacial lake shorelines are being used to reconstruct the past decreases in river gradient arising from glacio-isostatic uplift. Temporal uplift of the river relative to the southern outlet of Lake Agassiz (located in South Dakota just south of the Red River basin) is being modeled by exponentially relaxing the relative uplift of the Lower Campbell beach level of Lake Agassiz at a rate that is consistent with the observed tilts of abandoned shorelines in Manitoba, specifically those of the Lake Winnipegosis basin (Lewis et al., 2000; Tackman et al., 1998). Preliminary estimates reveal that the river has lost about half its gradient since 8000 yr BP. The long-term significance of change in gradient to the flood hazard is currently being assessed.

LONG-TERM CLIMATIC CHANGE

Although intuition suggests that extreme floods would be more common under wetter climate conditions, specific climatic thresholds leading to shifts in regional flood hazards are difficult to determine. Annual (prior August to current July) precipitation since AD 1409 has been estimated from the Red River tree-ring network using regional bur oak ringwidth (St. George and Nielsen, in press). These data indicate that hydroclimate in southern Manitoba has been relatively stable over the last two hundred years, although interrupted briefly by pronounced wet intervals in the late 1820s and 1850s (Fig. 6). Prior to this, the reconstruction shows that the Red River valley experienced extremely dry conditions between AD 1670 and 1775, with below-normal precipitation occurring approximately two years out of three. Individual dry years in the Red River valley were usually associated with larger-scale drought across much of the North American interior (St. George and Nielsen, in press).

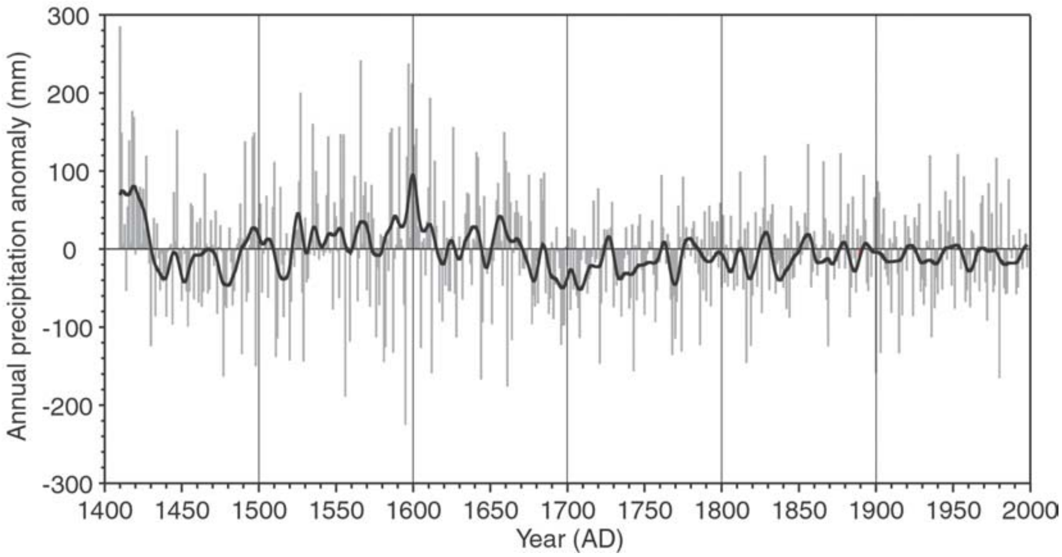

Fig. 6 *Reconstructed annual (August-July) precipitation at Winnipeg from AD 1409-1998, based on the width of tree rings in the Red River tree-ring network. Units are mean annual precipitation deviations from 1961-1990. Black line represents 15-year weighted running mean (after St. George and Nielsen, in press).*

Comparisons with limnological records from North Dakota and Minnesota suggest that multi-decadal fluctuations in regional hydroclimate have been consistent across the northeastern Great Plains during the last 600 years (St. George and Nielsen, in press). These paleoenvironmental records indicate that natural variation in this region can generate shifts in precipitation regimes that last for several decades and extend over several thousand square kilometres (St. George and Nielsen, in press). The hydroclimate record further indicates that past long-term changes in regional precipitation have varied by roughly 10% under 'natural climatic' conditions. This variability represents an obvious forcing factor influencing on the occurrence of extreme high and low river discharges at multi-decadal and centennial timescales. As a consequence, it seems that climatic case studies in regional drought and flood planning based exclusively on 20th century instrumental records may under-estimate the worst-case scenarios.

SUMMARY

The tree-ring paleoflood record for the lower Red River (LRR) extends from AD 1648 to 1999. Flood-rings provide evidence for seven high-magnitude floods: 1997, 1979, 1950, 1852, 1826, 1762 and 1747. Tree ring evidence suggests that the 1826 flood was the most severe flood to occur in at least the last 352 years. No tree-ring evidence has been found from a flood in 1776. The tree-ring flood record from the LRR contains three periods during which there is a grouping of Red River high-magnitude floods: the mid 1700s, the early to mid 1800s and the latter half of the 20th century. Tree-ring evidence indicates that there was spring flooding along both the Red and Assiniboine rivers in AD 1826, 1538 and 1510.

Two meanders near St. Jean Baptiste have extended outwards and rotated downvalley since 8000 yr BP in a single phase of lateral channel migration. Migration at both meanders since 1000 yr BP has averaged about 0.04 $m.a^{-1}$. The lateral incision of the channel is estimated to have averaged between 0.6 and 0.8 $m.kyr^{-1}$ since 8000 yr BP. The cross-sectional area of the valley bottom increased by about 2% and 0.7% at the two meanders between 1000 and 0 yr BP. The resulting change in the hydraulic radius of the valley bottom translates into a decrease of 2.1% and 0.6% in mean flow velocity (v_{mean}). This loss in v_{mean} at each meander is compensated for by the increase in cross-sectional area so that the total discharge capacity of the valley cross-section is unchanged. Overall, changes to the valley cross-section area caused by the lateral migration of the river have produced negligible variations to the flood hazard on the clay plain since 1000 yr BP.

Preliminary modeling of post-glacial crustal uplift indicates that the Red River in southern Manitoba has lost about half of gradient since 8000 yr BP.

Tree-ring data indicate that the hydroclimate in southern Manitoba has been relatively stable over the last two hundred years, but prior to this, hydroclimate

was more variable. Notably, conditions in the Red River valley were extremely dry between AD 1670 and 1775, with below normal precipitation occurring approximately two years out of three. The long-term hydroclimate record indicates that regional drought and flood planning based exclusively on 20th century instrumental records may under-estimate the worst-case scenarios.

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