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# Holocene lateral channel migration and incision of the Red River, Manitoba, Canada

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**Abstract**

The Holocene evolution of the shallow alluvial valley occupied by the Red River was investigated at two successive river meanders near St. Jean Baptiste, Manitoba. A transect of five boreholes was sited across the flood plain at each meander to follow the path of lateral channel migration. From the cores, 24 wood and charcoal samples were AMS radiocarbon dated. The dates from the lower half of the alluvium in each core are interpreted to represent the age of the lateral accretion deposits within the flood plain at the borehole sites. The ages of these deposits increase progressively from ~ 900 to 7900 and 1000 to 8100 cal years B.P. along each transect, respectively, from the proximal to distal portions of the flood plain. At the upstream meander, the average rate of channel migration was initially 0.35 m/year between ~ 7900 and 7400 cal years B.P., then decreased to 0.18 m/year between ~ 7400 and 6200 cal years B.P., and subsequently varied between 0.04 and 0.08 m/year. Net channel incision of the river since 8100 cal years B.P. is estimated to have ranged between 0.4 and 0.8 m/ky. The pre-6000-years-B.P. interval of greater channel migration is hypothesized to reflect a higher phase of sediment supply that was associated with the establishment of the river system on the former bed of glacial Lake Agassiz. Since 1000 years B.P., the outward migration of the meanders has caused a gradual enlarging of 0.7–2% in the cross-sectional area of the shallow valley at the two meanders. When considered proportionally over timescales of up to several centuries, the widening of the valley cross-section is very low to negligible and is deemed an insignificant factor affecting the modern flood hazard on the clay plain.


*Keywords:* Red River; Flood plain; Lateral channel migration; Channel incision; Sediment supply; Alluvial chronology

## 1. Introduction

Studies investigating fluvial geomorphic change over timescales of many thousands of years have provided insights into the long-term temporal character and controls on modern rivers. As a general context to the contemporary alluvial landscape, many major rivers in temperate, mid-latitude areas during the late Quaternary experienced a general transition from braided to meandering planforms because of changes in discharge and sediment supply regimes associated with deglaciation (e.g., Starkel, 1991; Knox, 1995, and references therein). Lower intensity perturbations in fluvial activity have continued during the Holocene and have been driven by variations in climate, shifts in vegetation ecozones, base level and tectonic changes,

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anthropogenic effects and/or local factors (see Knox, 1983, 1995, 2000; Starkel, 1991). Studies of geomorphic change over longer timescales have also provided insights into processes where geomorphic development is not apparent over the short term (years, decades; see, for example, Brakenridge, 1985; Tinkler et al., 1994). Consideration of long-term variations in the intensity and spatial pattern of fluvial processes have aided the interpretation of North American archaeological records (see Bettis, 1995). Of increasing importance, however, is the consideration of longer term geomorphic change to modern environmental issues or geological hazards. Such studies have proved insights into, for example, flood frequency–magnitude and climate change (Knox, 1993; Nott and Price, 1999), factors controlling bank-breach floods (Jiongxin, 2001), paleohydrology (O'Connor et al., 1994) and the occurrence of prehistoric river impoundments by landslides (Brooks and Hickin, 1991; Reneau and Dethier, 1996).

The Red River in Manitoba, Canada (Fig. 1), experiences extreme spring floods that form a broad, shallow flood zone and rise and fall slowly over a period of up to 4–6 weeks. A key geomorphic factor contributing to the flood hazard is that the river occupies a shallow, low-gradient alluvial valley that has insufficient capacity to contain large flows (LeFever et al., 1999; Brooks and Nielsen, 2000). During such floods, river waters overtop the margins of the valley and spread across the adjacent flat clay plain. For example, the flood of April–May 1997 (see Rannie, 1998; Todhunter, 2001), the largest Red River flood in southern Manitoba since 1852, inundated an area of about 2000 km$^2$ and up to 40 km wide between the Canada/USA border and Winnipeg (Fig. 1; data from Manitoba Natural Resources). An important consideration is whether geomorphic processes relating to the development of the shallow alluvial valley are altering the cross-sectional area sufficiently to affect the contemporary flood hazard on the clay plain. While unlikely to be relevant at annual to decadal timescales, the change in valley cross-sectional area may be significant at a century timescale and thus represent a background process that could be either aggravating or easing the long-term trend of the flood hazard.

This paper summarizes the results of a borehole investigation into the alluvial deposits of the shallow valley occupied by a sinuous section of the Red River near St. Jean Baptiste, southern Manitoba (Fig. 1). The objectives are to: (i) document the geomorphic

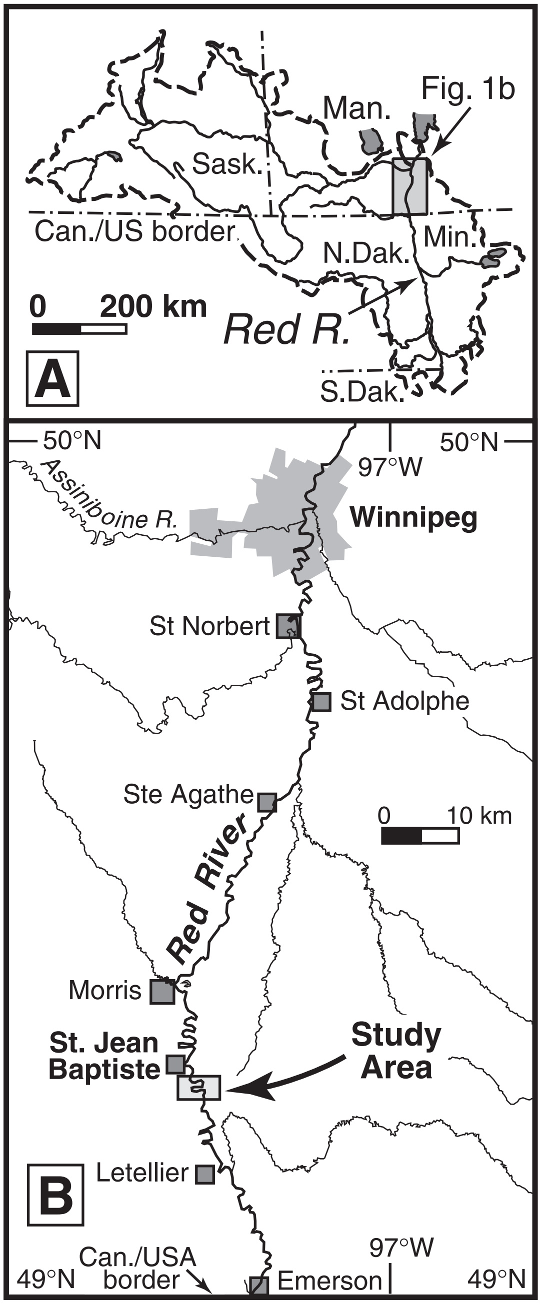


Fig. 1. Maps showing (A) the Red River drainage basin and (B) the location of the study area near St. Jean Baptiste, Manitoba.

evolution of this valley by reconstructing lateral channel migration and channel incision since the early–mid Holocene; (ii) identify the major control(s) on this evolution; (iii) assess the significance of the resulting change in valley cross-sectional area to the contemporary flood hazard on the clay plain.

## 2. Red river

The north-flowing Red River drains an area of 290,000 km$^2$ (Fig. 1A) that traverses the flat and gently sloping clay plain of the Red River valley (Upham, 1895). The clay plain is composed of gla-


Fig. 2. Oblique aerial photographs of the Red River showing (A) the irregular meandering planform of the river and (B) the downstream meander at the study site where the 99RR1 transect of boreholes are located. The shallow alluvial valley occupied by the river is approximately the width of the meander belt, but is imperceptible from the flat clay plain in both photographs. The view in (A) is downstream looking toward the study area meanders that are located in the background (marked by arrow), while (B) is looking west across the alluvial valley.

ciolacustrine sediments that aggraded within glacial Lake Agassiz during the late Pleistocene and early Holocene (see Teller and Clayton, 1983). In Manitoba, the river became established on the lake bed between 7800 and 8200 $^{14}C$ years B.P., as the lake waned and receded northward (Fenton et al., 1983; Teller et al. 1996). The river has since eroded a shallow valley into the clay plain, up to 15 m deep and 2500 m wide, that contains the genetic flood plain of the river. Reflecting the northward drainage and development on the lake bed, the river does not occupy a valley eroded by melt water under a glacially influenced or glacial-lake-influenced hydrological regime.

Within Manitoba, the Red River is a single-channeled, meandering river with an irregular sinuosity and a low average valley gradient of 0.0001 (Figs. 1B and 2). The flood plain and the contemporary riverbanks are composed predominantly of silt alluvium (Brooks, submitted for publication), and thus, the river represents an example of a mud-dominated stream. A curvilinear pattern of ridge and swale topography commonly marks past channel positions across the valley bottom and reveals that the flood plain has been formed by the lateral migration of meanders. This topography also reveals no obvious change in meander geometry, which would be indicative of a significant alteration in the discharge and/or sediment supply regimes or a threshold response in the river planform (see Schumm, 1977). Sequential aerial photographs and the comparison of mid-19th and late-20th century maps reveal no significant lateral channel migration along the river.

At Emerson, Manitoba (Fig. 1B), 28 km upstream of the study area and with a contributing drainage area of 104,000 $km^2$, the mean annual discharge of the Red River is 98 $m^3/s$, and the lowest and highest mean monthly flows are 22 (February) and 362 $m^3/s$ (April), respectively (Water Survey of Canada data, 1912–1995). Bankfull discharge, defined as a flow with a 2-year return period, is approximately 600 $m^3/s$. The annual hydrograph typically consists of a freshet flow in April or May arising from snowmelt runoff, while secondary peaks generated by rainfall can occur between June and October. The flood of record at Emerson occurred on 27–28 April 1997, peaking at 3740 $m^3/s$ (data from Manitoba Water Resources). The suspended sediment load of the river consists of over 90% silt and clay, regardless of sediment concentration and river discharge (Glavic et al., 1988). The annual suspended sediment load at Emerson varies between 0.4 and 1.5 million tonnes with minimum and maximum sediment concentrations ranging from 10–20 (December to February) to 500–1900 mg/l (April to July), respectively (1978–1986 data; Glavic et al., 1988).

Postglacial crustal rebound resulting from the isostatic depression of the landscape by the Laurentide Ice Sheet during the late Pleistocene has affected the regional landscape of the eastern Prairies. This is readily exemplified by the differential tilting of Lake Agassiz shorelines in Manitoba, eastern North Dakota and western Minnesota (Upham, 1895; Johnston, 1946; Teller and Thorleifson, 1983). Isobase maps defined by the deformed shorelines indicate that greater uplift occurred toward the NE in the direction of Hudson Bay (Johnston, 1946; Teller and Thorleifson, 1983). Although the rates of uplift have diminished over the Holocene (see Tackman et al., 1998), regional tilting is active in southern Manitoba, as revealed by contemporary lake gauge and geodetic data (Lambert et al., 1998; Tackman et al., 1999).

## 3. Methods

Data for this study were obtained by coring the flood-plain alluvium of the Red River across the valley bottom at two successive river meanders located about 2.5 and 3.5 km S–SE (upstream) of St. Jean Baptiste, Manitoba (Figs. 1B, 2B and 3). These sites were chosen because of the relatively broad width and locally uniform morphology of the meander belt, easy access onto the flood plain via cultivated fields and a preserved ridge and swale pattern on the flood plain. These meanders are typical of the Red River in southern Manitoba, notwithstanding local variations in channel sinuosity. At the study area, the channel sinuosity is 2.8, the valley gradient is 0.00006 and the bankfull channel is about 75 m wide and 6–9.5 m deep. The boreholes were positioned in two transects (99RR1 and 99RR3) along the east and west sides of the valley bottom, with each transect consisting of five boreholes (label A to E in Fig. 3). They were sited to follow the path of lateral migration of the channel, as revealed by the ridge and swale

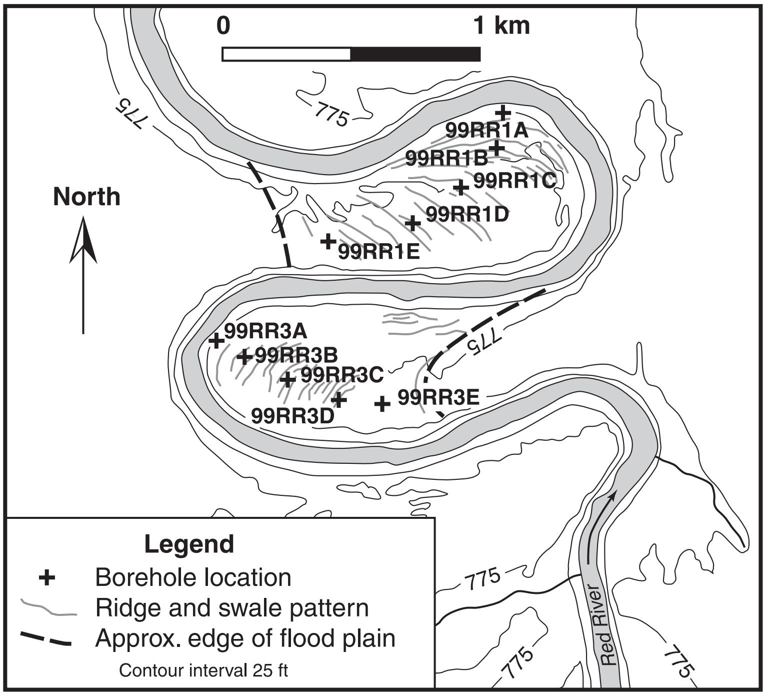


Fig. 3. Map showing the locations of the 10 boreholes along the two transects and the ridge and swale pattern across the flood plain.

pattern. The heights of the boreholes along each transect were surveyed relative to an arbitrary datum. Brooks et al. (2001) described in detail the coring and core logging methods.

Twenty-four wood and charcoal samples were submitted to Beta Analytic (Miami, FL) for AMS radiocarbon dating. These samples were selected carefully to favour "fresh-looking" samples to avoid dating reworked materials. The radiocarbon ages were calibrated to calendar years by Beta Analytic following Talma and Vogel (1993) and using the calibration data set of Stuiver et al. (1998).

## 4. Core stratigraphy

The cores contain two distinct units: a lower unit of glaciolacustrine deposits that aggraded within Lake Agassiz and an upper unit of flood-plain alluvium (Fig. 4). The presence of glaciolacustrine deposits indicates that the entire vertical sequence of the alluvial deposits is represented in the 10 cores. As summarized from Brooks (submitted for publication), the Red River alluvium in the 10 cores ranges from 15.3 to 21.6 m thick and is composed primarily of silt. The deposits consist of three main lithofacies; cosets of massive silt, thick (>0.4 m) massive silt and deformed and disrupted silt. Sand beds and interlaminated sand and silt form relatively minor deposits, principally within the lower half of the alluvium. Thin beds of medium–coarse sand and pea gravel are present in two cores within the lower meter of the alluvium. Authigenic $CaCO_3$ precipitate and Fe-oxide mottling are present in the alluvial deposits and generally increase in vertical extent and lateral concentrations from the younger (A) to older (E) cores. $CaCO_3$ precipitate is present predominantly in the upper half of the alluvium and forms irregularly shaped, whitish agglomerates up to 2 cm across and occasionally nodules. The Fe-oxide mottling is present in the lower half of the alluvium. The mottles are irregularly shaped, generally several millimeters in scale and tend to be developed in zones of slightly finer silt sediments.

Organic material is present throughout the alluvium as small (< 1 mm) disseminated organic mate-

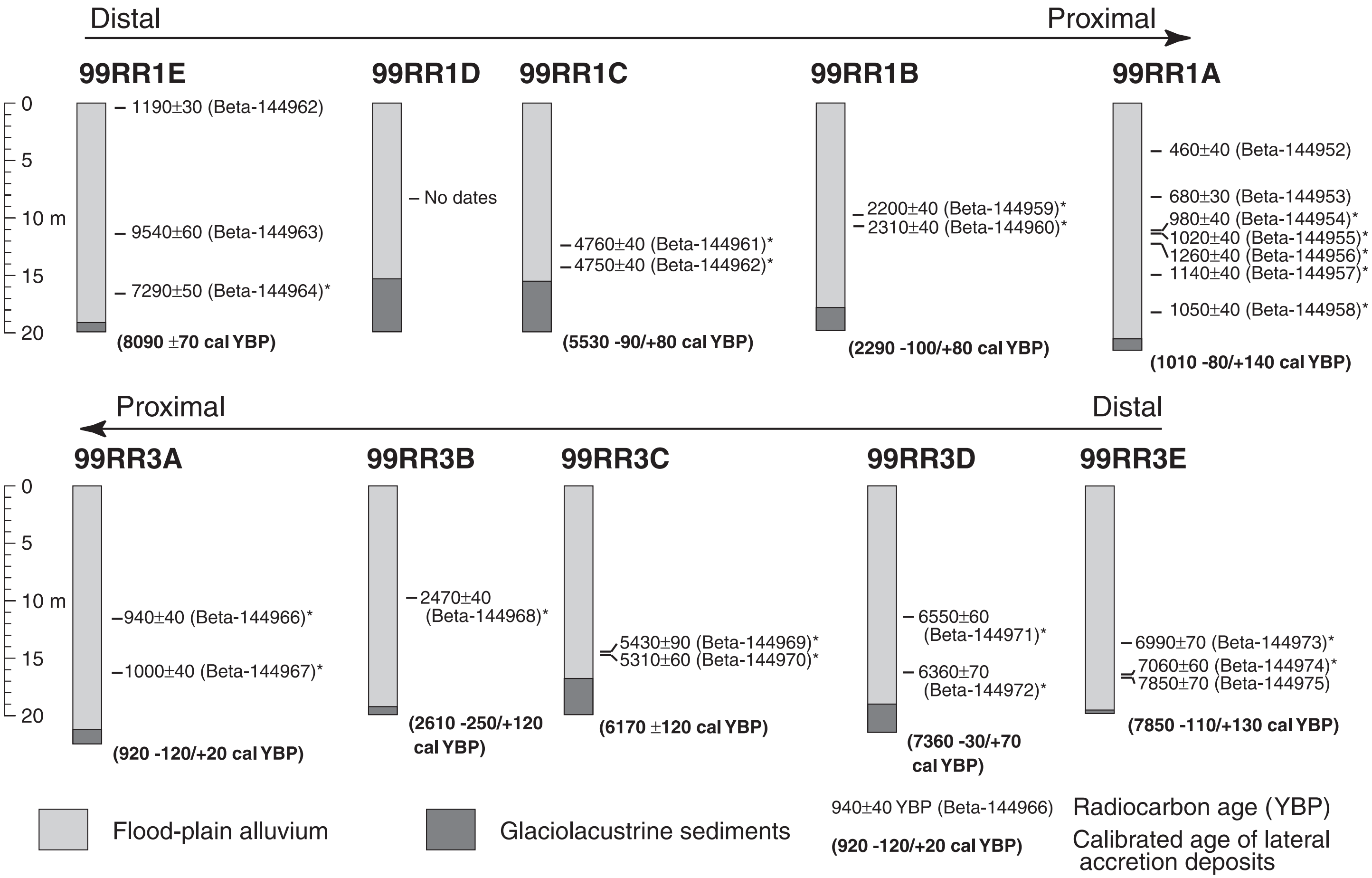


Fig. 4. Summarized lithostratigraphic diagram of the cores showing the depth locations of the 24 AMS radiocarbon ages. The boreholes are arranged from the distal to proximal areas of the flood plain with the sequencing intended to reflect the relative ordering at successive river meanders. The age at the bottom of each core (except 99RR1D) is the representative calibrated age of the lateral accretion deposits, as explained in the text (also see Table 2). The dates used in this determination of this representative age are flagged with an asterisk (*).

rial, fine charcoal and to a lesser extent, shell fragments. Whole shells, larger charcoal and wood fragments are present infrequently. Buried paleosols, several centimeters thick, occur in the upper 2 m of 99RR1D, 99RR1E, 99RR3B, 99RR3D and 99RR3E. All but one of the 24 organic samples submitted for

Table 1
Listing of AMS radiocarbon ages from the boreholes

| Borehole | Location (latitude/longitude) | Depth (m) | Sample number | Material | Radiocarbon age ($^{14}$C years B.P.) | $^{13}$C/$^{12}$C ratio (‰) | Calibrated age and age range at 1$\sigma$ (cal years B.P.) |
|---|---|---|---|---|---|---|---|
| 99RR1A | 49°14.8′N 97°18.5′W | 4.11 | Beta-144952 | wood | 460 ± 40 | − 26.1 | 515 (500–525) |
| | | 8.16 | Beta-144953 | probably *Betula*[a] | 680 ± 30 | − 29.1 | 655 (650–665) |
| | | 11.19 | Beta-144954 | deciduous wood[a] | 980 ± 40 | − 29.0 | 925 (910–940) |
| | | 11.37 | Beta-144955 | wood | 1020 ± 40 | − 27.8 | 940 (925–955) |
| | | 12.25 | Beta-144956 | *Salix*[a] | 1260 ± 40 | − 27.4 | 1185 (1165–1260) |
| | | 14.97 | Beta-144957 | deciduous wood[a] | 1140 ± 40 | − 28.8 | 1055 (980–1075) |
| | | 18.24 | Beta-144958 | deciduous wood[a] | 1050 ± 40 | − 27.0 | 950 (935–970) |
| 99RR1B | 49°14.7′N 97°18.6′W | 9.88 | Beta-144959 | deciduous wood[a] | 2200 ± 40 | − 29.0 | 2160, 2260, 2300[b] (2140–2300) |
| | | 10.78 | Beta-144960 | deciduous wood[a] | 2310 ± 40 | − 26.1 | 2340 (2325–2350) |
| 99RR1C | 49°14.6′N 97°18.7′W | 12.27 | Beta-144961 | wood | 4760 ± 40 | − 28.0 | 5480, 5530, 5575[b] (5465–5585) |
| | | 14.19 | Beta-144962 | *Salix*[a] | 4750 ± 40 | − 26.5 | 5475, 5540, 5570[b] (5465–5585) |
| 99RR1E | 49°14.5′N 97°19.1′W | 0.41 | Beta-144963 | charcoal (wood) | 1190 ± 30 | − 24.5 | 1080 (1065–1165) |
| | | 11.38 | Beta-144964 | wood | 9540 ± 60 | − 26.1 | 10,755, 10,985, 11,030[b] (10,710–10,870 and 10,935–11,080)[c] |
| | | 16.68 | Beta-144965 | wood | 7290 ± 50 | − 28.4 | 8060, 8090, 8110[b] (8020–8160) |
| 99RR3A | 49°14.3′N 97°19.5′W | 11.62 | Beta-144966 | deciduous wood[a] | 940 ± 40 | − 26.1 | 910 (790–925) |
| | | 16.22 | Beta-144967 | deciduous wood[a] | 1000 ± 40 | − 26.4 | 930 (915–945) |
| 99RR3B | 49°14.3′N 97°19.4′W | 9.78 | Beta-144968 | wood | 2470 ± 70 | − 27.1 | 2490, 2645, 2700[b] (2360–2730) |
| 99RR3C | 49°14.2′N 97°19.2′W | 14.40 | Beta-144969 | *Salix*[a] | 5430 ± 90 | − 25.1 | 6270 (6175–6300) |
| | | 14.77 | Beta-144970 | *Salix*[a] | 5310 ± 60 | − 27.5 | 6020, 6070, 6105[b] (5990–6185) |
| 99RR3D | 49°14.2′N 97°19.1′W | 11.40 | Beta-144971 | *Salix*[a] | 6550 ± 60 | − 25.7 | 7440 (7425–7485) |
| | | 16.15 | Beta-144972 | *Salix*[a] | 6360 ± 70 | − 27.0 | 7275 (7240–7330) |
| 99RR3E | 49°14.2′N 97°18.9′W | 13.68 | Beta-144973 | *Salix*[a] | 6990 ± 70 | − 26.3 | 7805 (7725–7865 and 7900–7920)[c] |
| | | 16.44 | Beta-144974 | wood | 7060 ± 60 | − 27.6 | 7865, 7900, 7920[b] (7815–7945) |
| | | 16.69 | Beta-144975 | deciduous wood[a] | 7850 ± 70 | − 27.8 | 8610 (8565–8715) |

[a] Identified by C. Keith (written communication, 1999).
[b] Multiple ages represent separate intercepts of the mean radiocarbon age with the calibration curve.
[c] Two age ranges represent separate intercepts of the mean radiocarbon age at 1$\sigma$ with the calibration curve.

radiocarbon dating yielded post-glacial Lake Agassiz ages, which is consistent with a fluvial origin of the deposits (Table 1).

Identifying specific depositional environments in the alluvium is complicated by the lack of obvious vertical changes in texture or lithofacies within the deposits. Based on the topography and geomorphology of the flood plain, the presence of organic matter in the deposits and the character of modern sedimentation along the riverbanks, the alluvium is interpreted to be composed of overbank deposits from 0 to 2–3 m depth, oblique accretion deposits from 3–5 to 8–12 m depth and oblique accretion/channel deposits below 8–12 m (Brooks, submitted for publication). Lag deposits composed of thin beds of medium to coarse sand or pea gravel can be present locally in the lower 1 m of the alluvium.

## 5. Flood-plain chronology

The radiocarbon ages are depicted stratigraphically in Fig. 4. At least one date was obtained from each of the cores with the exception of core 99RR1D. Twenty-one dates are from the lower halves of the alluvial unit within the cores (Fig. 4). The within-core variability of these dates can be assessed from core 99RR1A from which seven organic samples were dated, five from the lower half of the alluvial unit (Fig. 4). The radiocarbon ages of these latter samples range from 980 ± 40 to 1260 ± 30 $^{14}$C years B.P. (Fig. 4). Although some transposition in the vertical sequencing occurs (Fig. 4), the ages are statistically indistinguishable at an error range of 2$\sigma$. Geomorphically, this similarity of age indicates that vertical sedimentation over the lower half of the alluvial unit was relatively rapid and probably occurred over a few hundred years.

This consistency of radiocarbon ages in the lower halves of the alluvial unit is duplicated by the pairs of dates from cores 99RR1B, 99RR1C, 99RR3A, 99RR3C and 99RR3D and from two of the three dates in core 99RR3E (Fig. 4). Two dates, 9540 ± 60 and 7850 ± 70 $^{14}$C years B.P. from cores 99RR1E and 99RR3E, respectively, are anomalously old relative to the other dates at nearby depths. The organic materials that yielded these dates are interpreted to have been reworked and thus unrepresentative of the age of the encapsulating sediments. The 9540 ± 60 $^{14}$C years B.P. age clearly is too old since it predates the final recession of glacial Lake Agassiz in southern Manitoba, while the 7850 ± 70 $^{14}$C years B.P. age is substantially older than two other dates sampled at approximately similar depths within core 99RR1E.

Excluding the two anomalously older dates, the radiocarbon dates from the lower half of the alluvium within a given borehole are interpreted to represent the age of the lateral accretion deposits at that location of the flood plain. The age of these deposits thus reveals a past position of the inner bank of a meander at a specific time, as the deposits would have aggraded immediately adjacent to the river channel. To better

Table 2
Summary listing of the single or averaged calibrated radiocarbon ages representing the age of the lateral accretion deposits at each borehole

| Borehole | Number of dates used in determining the average age[a] | Mean age (cal years B.P.) | 1$\sigma$ error | Age range at 1$\sigma$ (cal years B.P.) |
|---|---|---|---|---|
| 99RR1A | 5 | 1010[b,c] | − 80/+ 140 | 930–1150 |
| 99RR1B | 2 | 2290[b] | − 100/+ 80 | 2190–2370 |
| 99RR1C | 2 | 5530[b] | − 90/+ 80 | 5440–5610 |
| 99RR1D | no data | – | – | – |
| 99RR1E | 1 | 8090[d] | ± 70[e] | 8020–8160 |
| 99RR3A | 2 | 920 | − 120/+ 20 | 800–940 |
| 99RR3B | 1 | 2610[d] | − 250/+ 120 | 2360–2730 |
| 99RR3C | 2 | 6170[b] | ± 120 | 6050–6290 |
| 99RR3D | 2 | 7360 | − 30/70 | 7330–7430 |
| 99RR3E | 2 | 7850[b,f] | − 110/+ 130[e] | 7740–7980 |

[a] See Table 1 for listing of calibrated radiocarbon ages.

[b] Where there are multiple calibrated ages for an individual radiocarbon date (see Table 1), this date is represented by the average of these ages in the determination of the overall calibrated average age of the lateral accretion deposits in this borehole.

[c] The calibrations of the radiocarbon ages 460 ± 40 and 680 ± 30 $^{14}$C years B.P. from this borehole are not included in the average calibrated age.

[d] Based on the average of the intercepts on the calibration curve from a single radiocarbon date (see Table 1).

[e] Where the calibration of a radiocarbon date has produced two separate ranges of error at 1$\sigma$ (see Table 1), the overall maximum and minimum of these ranges is used in the determination of the age range at 1$\sigma$ for the overall calibrated average age for the lateral accretion deposits in this borehole.

[f] The calibration of the radiocarbon age 7850 ± 70 $^{14}$C years B.P. from this borehole is not included in this average calibrated age.

compare the radiocarbon ages between the boreholes, the dates were calibrated as mentioned above (Table 1). Where a core contains two or more dates, the calibrated ages were averaged to attain a representative age for the lateral accretion deposits in that borehole (Table 2). All further reference to the age of the flood plain is made to the calibrated (and averaged, where applicable) ages of the lateral accretion deposits in the boreholes, unless otherwise specified.

Three radiocarbon dates are from the upper half of the alluvial unit. Two of the three are from core 99RR1A, and the ages decrease toward the top of the core (Fig. 4). The depth and age of the two dates reveal that the sedimentation rate decreases toward the top of the core. The third date, 1190 ± 30 $^{14}$C years B.P., is from core 99RR3E and was sampled at a depth of 0.41 m (Fig. 4). It indicates that the distal areas of the flood plain experience a low rate of sedimentation, which is consistent with the presence of buried paleosols in cores 99RR1D, 99RR1E, 99RR3B, 99RR3D and 99RR3E.

## 6. Lateral channel migration

The calibrated ages of the flood plain increase sequentially along the two transects, becoming older from the A to E cores (Fig. 4; Table 2). This ordering traverses the proximal to distal portions of the flood plain based on the ridge and swale pattern. The ages of the two most distal cores (99RR1E and 99RR3E) are only ~ 1000 $^{14}$C year younger than the final recession of glacial Lake Agassiz, which occurred at ~ 8000 $^{14}$C years B.P. (~ 8900 cal years B.P.; Teller et al., 1996). This, in combination with the sequential ages of the cores, indicates that the two study area meanders have undergone only a single sequence of expansion, an interpretation that is consistent with there being only a single pattern of ridge and swale topography across the flood plain (Fig. 3).

The temporal pattern of the channel migration was reconstructed by interpolating the positions of the inner bank at 1000-years-B.P. intervals between the calibrated age of the lateral channel deposits and using the pattern of the ridge and swale topography as a

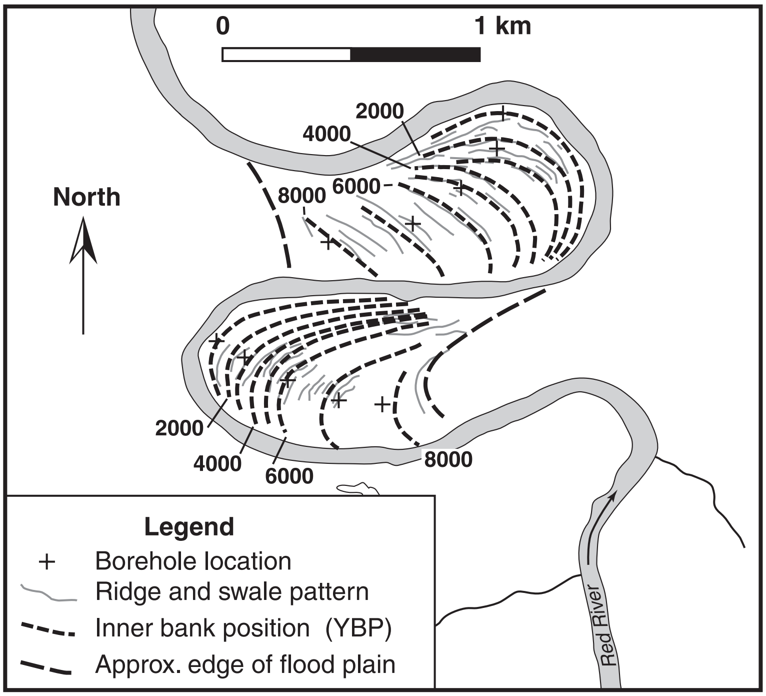


Fig. 5. Isochrones depicting the location of the inner bank of the meanders across the study area at intervals of 1000 years B.P.

guide to position isochrones (Fig. 5). The isochrones reveal that the two meanders have extended outward and rotated downvalley over time. The spacing of the isochrones, however, also reveals that the rate of lateral channel migration was greater prior to about 6000 years B.P. than afterward. This trend is well illustrated at the upper meander, where the 99RR3 transect of boreholes closely follows the meander growth across the flood plain, the lateral pattern of meander growth is simple, and all five boreholes contain radiocarbon-dated organic matter. At the downstream meander, no dates were obtained from borehole 99RR1D, thus, the position of the 6000-, 7000- and 8000-years-B.P. isochrones have been interpolated from the ages and spacing of boreholes 99RR1E and 99RR1C. Nevertheless, the migration pattern at both meanders reveals that approximately half of the flood plain formed between ~ 6000 and 8900 cal years B.P. (the final recession of glacial Lake Agassiz), which represents about one-third of the post-Lake Agassiz time interval.

The average rates of channel migration for the upstream meander were calculated using ages and straight-line distances between the 99RR3 transect boreholes (Table 3 and Fig. 6A). The error ranges shown for the migration rates are based on the 1$\sigma$ uncertainty of the calibrated radiocarbon ages for the lateral accretion deposits (Table 3). All of the rates along the 99RR3 transect are considered to reasonably approximate the maximum rate of migration at the different time intervals because the borehole locations generally follow the path of maximum meander growth (or the erosional axis of the meander; see Hickin, 1974).

As shown in Fig. 6A, the average rates of channel migration along the upstream meander were initially 0.35 m/year between ~ 7900 and 7400 cal years B.P., then decreased to 0.18 m/year between ~ 7400 and 6200 cal years B.P. and varied between 0.04 and 0.08 m/year since ~ 6200 cal years B.P. Significantly, Fig. 6A indicates that the upstream meander has been migrating laterally at 0.04 m/year after ~ 900 cal years B.P. This implies that the river presently is undergoing long-term lateral migration at a low rate, despite there being insignificant change in the channel position between mid-19th and late-20th century maps or in sequential aerial photographs. Similarly, a low rate of lateral channel migration occurred post-1000 cal years B.P. along the downstream meander, as indicated by the average rate of migration between borehole 99RR1A and the riverbank (Table 3).

The "apparent" rates of lateral channel migration between the boreholes along the 99RR1 transect are also listed in Table 3. However, except for the rates between boreholes 99RR1B, 99RR1A and the modern riverbank, the meander growth is poorly represented by these rates compared to the upstream meander. Reasons for this are that the meander growth has occurred in two phases within a single sequence of meander expansion (see Hickin, 1978); more impor-

Table 3
Average rates of lateral channel migration between the 99RR1 and 99RR3 transect boreholes and the contemporary river bank

| Locations | Distance (m) | Ages of bank–borehole or borehole–borehole (cal years B.P.)[a] | Difference in averaged age (cal years B.P.) | Upper and lower difference in age range at 1$\sigma$ (cal years B.P.)[a] | Average rate of channel migration (m/year) | Range of average rate of channel migration at 1$\sigma$ (m/year) |
|---|---|---|---|---|---|---|
| Bank to 99RR1A | 40 | 0[b]–1010 | 1010 | 930–1150 | 0.04 | 0.03–0.04 |
| 99RR1A to 99RR1B | 152 | 1010–2290 | 1280 | 1040–1440 | 0.12 | 0.11–0.15 |
| 99RR1B to 99RR1C | 208 | 2290–5530 | 3240 | 3070–3420 | 0.06 | 0.06–0.07 |
| 99RR1C to 99RR1E | 541 | 5530–8090 | 2560 | 2410–2720 | 0.21 | 0.20–0.23 |
| Bank to 99RR3A | 38 | 0[b]–920 | 920 | 800–940 | 0.04 | 0.04–0.05 |
| 99RR3A to 99RR3B | 127 | 920–2610 | 1690 | 1410–1930 | 0.08 | 0.07–0.09 |
| 99RR3B to 99RR3C | 191 | 2610–6170 | 3560 | 3320–3930 | 0.05 | 0.05–0.06 |
| 99RR3C to 99RR3D | 215 | 6170–7360 | 1190 | 1040–1380 | 0.18 | 0.16–0.21 |
| 99RR3D to 99RR3E | 171 | 7360–7850 | 490 | 310–650 | 0.35 | 0.26–0.55 |

[a] The errors at 1$\sigma$ are listed in Table 2.
[b] The age of the contemporary river bank is taken as 0 cal B.P. (i.e., A.D. 1950).

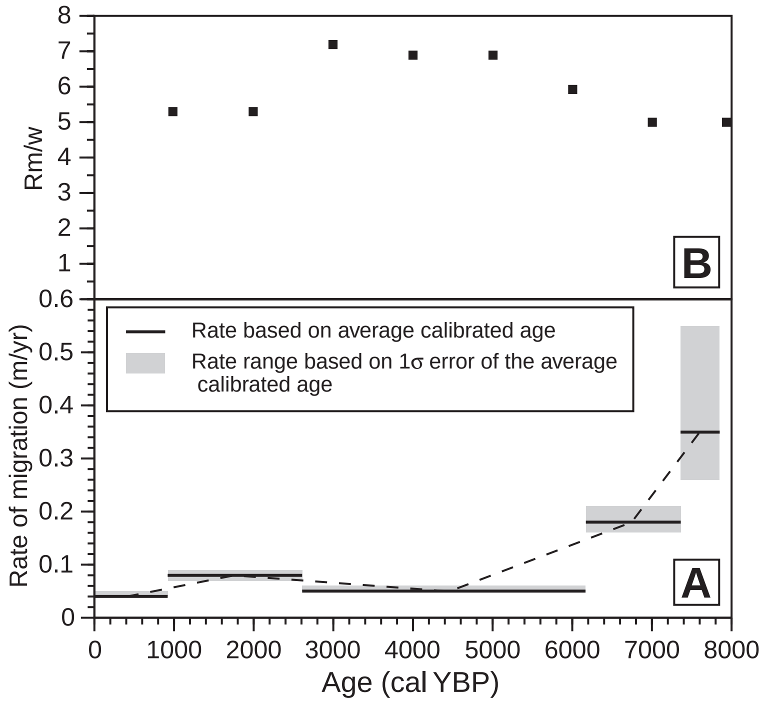


Fig. 6. (A) Average rates of lateral channel migration between the locations of 99RR3 boreholes and contemporary river bank. The time interval depicted for each rate is based on the representative calibrated radiocarbon ages for the boreholes (Table 2), while the error zones reflect the 1$\sigma$ uncertainty in these calibrated ages (see Table 3). (B) Plot of Rm/$w$ ratio at 1000-years-B.P. intervals for the 99RR3 (upstream) meander, based on the isochrones in Fig. 5.

tantly, boreholes 99RR1E, 99RR1D and 99RR1C are poorly aligned with the path of maximum meander growth, and borehole 99RR1D lacks radiocarbon-dated organic material.

## 7. Channel incision

The depth of contact between the alluvium and glaciolacustrine deposits in the boreholes is depicted in Fig. 7. The heights of the boreholes along each transect are scaled to the topography of the flood plain relative to the height of the E boreholes. In Fig. 7, the depth of alluvium-glaciolacustrine contact varies between 15.7 and 23.5 and 17.0 and 25.1 m along the 99RR1 and 99RR3 transects, respectively. At the two oldest boreholes (99RR1E and 99RR3E), the depth of contact is between 19 and 20 m; whereas in the adjacent D and C cores, the contact is up to several meters shallower (15.7–18.7 m deep). The depths in the E boreholes indicate that the main phase of river incision into the clay plain occurred prior to the lateral migration of the channel past the locations of the D and C boreholes. If this incision had still been underway, the contact in the E boreholes would be distinctly shallower than the D (and possibly the C) boreholes, with the contacts thereby delineating a buried slip-off slope. Based on the age of the lateral accretion deposits in the oldest borehole (99RR1E), the main phase of channel incision was completed between ~ 8900 and 8100 cal years B.P. This period of incision is consistent with chronostratigraphic data from Winnipeg (Fig. 1B), where wood, buried at 10.7 m depth in alluvium near the mouth of the Assiniboine River valley, produced a radiocarbon age of 7490 ± 80 $^{14}$C years B.P. (or ~ 8300 cal years B.P.; Nielsen et al., 1993).

Along both transects, the net depth of contact deepens by 4.4 and 5.6 m between boreholes 99RR1E and 99RR1A and 99RR3E and 99RR3A, respectively

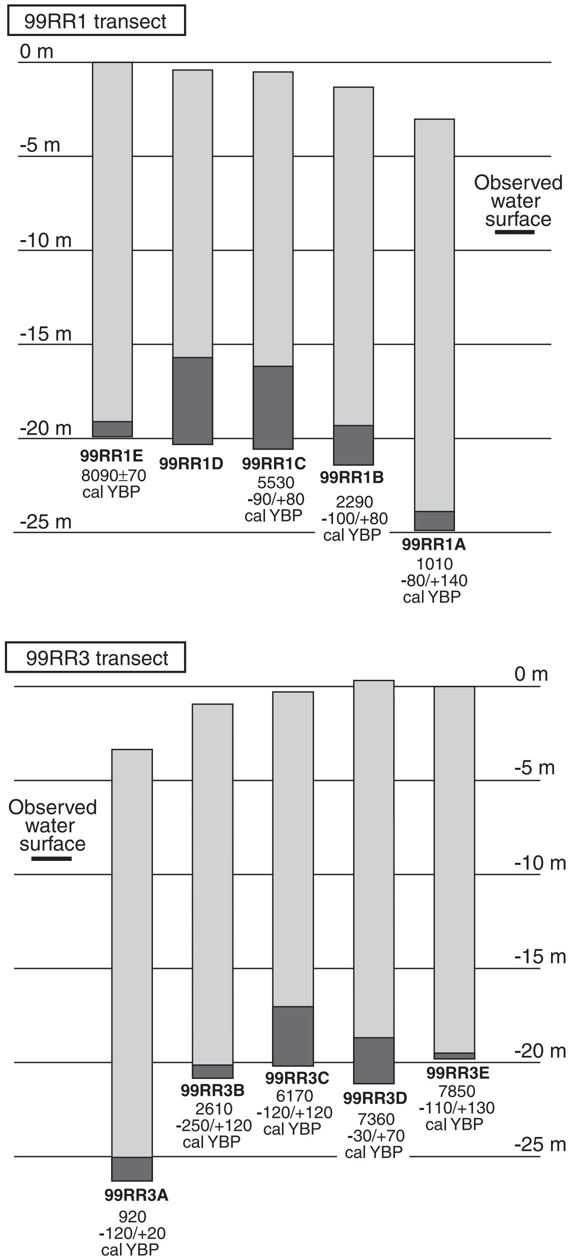


Fig. 7. The vertical positions of the boreholes along the 99RR1 and 99RR3 transects relative to the heights of the "E" boreholes. Also shown is the level of the observed river surface on the day of surveying. The depths to the alluvium/glaciolacustrine contacts are based on the core logs. The representative calibrated ages of the lateral accretion deposits within the boreholes (where available) are shown below each borehole label.

(Table 4). This deepening probably represents the gradual net incision of the river channel into the clay plain, since it also corresponds approximately to the slope of the flood plain (Table 4). However, along both transects, the depth of contact is up to 3.4 m shallower within the D and C boreholes than in the E boreholes, suggesting that perhaps there has been a period of channel aggradation. However, the lack of backswamp areas on the distal areas of the flood plain in the study area (or elsewhere along the river), which would be indicative of valley bottom aggradation, do not support this interpretation. The shallowing of the contact may instead reflect a decrease in the maximum depth of scouring between boreholes rather than the aggradation of the channel. Whether this was due to, for example, a climatically controlled change in the flow regime or the transitory formation of deep scour holes as the channel migrated laterally is not known.

The average rates of incision between the E and A boreholes are 0.62 and 0.81 m/ky for the 99RR1 and 99RR3 transects, respectively, based on the difference in the depth of contact and the calibrated flood plain ages (Table 4). These rates are marginally greater (40–60%) than those rates based on the difference in the flood-plain topography between the E and A boreholes (0.42 and 0.49 m/ky for the 99RR1 and 99RR3 transects, respectively; Table 4). The flood-plain topography may indicate that the average rate of incision was greater in the post-6200–5500-cal-years-B.P. interval because the slope of the flood plain is greater between the C and A than the E to C boreholes

Table 4
Rates of lateral channel incision

| Transect | Between boreholes | Δ calibrated age (cal years)[a] | Δ depth of alluvial/ glaciolacustrine contact (m)[b] | Δ topography of flood plain (m)[b] | Rate of incision (m/ky) |
|---|---|---|---|---|---|
| 99RR1 | E and A | 7080 | 4.4 | na | 0.62 |
| | E and A | 7080 | na | 3.0 | 0.42 |
| | E and C | 2560 | na | 0.5 | 0.20 |
| | C and A | 4520 | na | 2.5 | 0.55 |
| | C and B | 3240 | na | 1.7 | 0.53 |
| 99RR3 | E and A | 6930 | 5.6 | na | 0.81 |
| | E and A | 6930 | na | 3.4 | 0.49 |
| | E and C | 1680 | na | 0.3 | 0.18 |
| | C and A | 5250 | na | 3.1 | 0.59 |

[a] See Table 2.
[b] See Fig. 7.

along both transects (Table 4). These rates range from 0.55 to 0.59 m/ky (99RR1 and 99RR3 transects, respectively) between the C to A boreholes, and from 0.20 to 0.18 m/ky for the E to C boreholes (Table 4). Based on both the difference in depth of contact and the flood-plain topography, 0.4–0.8 m/ky is likely the right range of magnitude for the post-8100-cal-years-B.P. rate of lateral incision of the Red River channel into the Lake Agassiz clay plain.

## 8. Discussion

### 8.1. Long-term controls on lateral channel migration

The flood plain in the study area represents a continuous and substantial temporal record of the fluvial geomorphic activity. The flood-plain chronology indicates that the rate of lateral channel migration was faster prior to about 6 ka than afterward (Figs. 5 and 6). The long duration of this trend suggests that the controlling factor is environmental rather than a local perturbation, particularly because of a lack of local candidate mechanisms for the latter.

Changes in the intensity of Holocene fluvial activities can relate to variations in climate (e.g., Brakenridge, 1980; Knox, 1983, 1995, 2000; Starkel, 1995). Work by Knox et al. (1981) and Knox (1984, 1985) in the driftless area of Wisconsin in the Upper Mississippi Valley (UMV), which is relatively nearby to southern Manitoba, indicates that rates of fluvial erosion and deposition varied in response to three general Holocene climatic phases. These variations are attributed to long-term fluctuations in the recurrence of large floods. Notably in the period between ~ 7500 and 6000 $^{14}$C years B.P. (~ 8300–6800 cal years B.P.), Knox's work found that the erosion and deposition of alluvial sediments was concentrated on the flood plains and alluvial fans of small tributaries of less than about 10 km$^2$. Little evidence of alluviation was found along the flood plains of larger trunk streams (Knox et al., 1981). The flood-plain stability of the trunk streams is attributed to there being a generally low frequency of larger floods in a climate that was warmer and drier than present. Between ~ 6000 and 4500 $^{14}$C years B.P. (~ 6800–5200 cal years B.P.), climate was cooler and moister resulting in a higher frequency of larger floods and, correspondingly, the larger trunk rivers of the UMV experienced "intensified" lateral migration (Knox, 1985).

Recent paleolimnological evidence from within or immediately adjacent to the Red River basin indicates that the warmer–drier "hypsithermal" climatic interval began as early as ~ 10,700–9500 cal years B.P., intensified after ~ 9200–7900 cal years B.P. and ended at ~ 5400–4500 cal years B.P. (e.g., Bradbury et al., 1993; Laird et al., 1996; Valero-Garcés et al., 1997). This reveals that the period of higher lateral channel migration and flood-plain construction in the study area occurred entirely within a warmer–drier climate, in contrast to the trunk streams of the UMV. Also different is the lack of intensification of fluvial activity along the Red River following the shift to a cooler–moister climate after ~ 5400–4500 cal years B.P. While this nonsynchronicity in fluvial geomorphic activity and different response to climate may simply reflect variability between separate watersheds (see Taylor and Lewin, 1997), consideration of the geomorphic history of the Red River basin suggests that another factor(s) may control the temporal rates of lateral channel migration in the study area.

The factors controlling the rate of lateral migration ($M$) can be summarized as (following Hickin and Nanson, 1984):

$$M = f(\omega, b, G, h, Y_b) \tag{1}$$

where $\omega$ represents unit stream power (a slope-discharge product), $b$ is a parameter expressing planform geometry, $G$ is the rate of sediment supply, $h$ is outer bank (or valley side) height and $Y_b$ the erosional resistance of the outer bank material. Along the study area, $h$ has been relatively constant over the past ~ 8000 cal years B.P., as revealed by the depths of alluvium–glaciolacustrine contact in Fig. 7. Particularly noteworthy are the nearly identical depths of contact at boreholes 99RR3E and 99RR3B, which are situated at locations on the flood plain that experienced markedly different lateral migration rates (Fig. 5). The depth of scour at A boreholes along both transects are marginally deeper than elsewhere (Fig. 7), but this increase in depth does not correspond to a significant decrease in the lateral migration rate between the 99RR3B and 99RR3A boreholes (Fig. 6). $Y_b$ also has been constant because the valley side materials into which the channel has laterally eroded are composed of homogenous glaciolacustrine deposits of the clay plain.

Variations in $b$ have occurred as the meanders migrated laterally across the valley bottom, but these changes are deemed an insignificant control of the lateral migration rates. Expressing $b$ as a ratio of the radius of meander curvature (Rm) to channel width ($w$), Hickin and Nanson (1975, 1984) have shown that the rate of channel migration can vary significantly with changes in Rm/$w$. Along the 99RR3 meander, however, Rm/$w$ is similar (5.0–5.3) at 8000 and 7000 years B.P. and at 2000 and 1000 years B.P. (Fig. 6B) when the average rates of migration are relatively high and low, respectively (Fig. 6B). In addition, the rates of migration are similar throughout the entire post-5000-years-B.P. interval, despite a decrease in Rm/$w$ from 6.9–7.2 to 5.3 between 3000 and 2000 years B.P.

From the context of lateral channel migration, unit stream power ($\omega$) is primarily a function of discharge ($Q$) and gradient ($s$). Based on a modern estimate of the 2-year flow at Emerson and the average valley gradient between Emerson and Morris (Fig. 1B), $\omega$ for the Red River is about 5.1 W m$^{-2}$ (Table 5). Over the past 8 ky, both $Q$ and $s$ have not been constant. Regarding $s$, crustal rebound has caused the valley gradient of the Red River to decrease over the post-Lake Agassiz period. As summarized in Table 5, the Emerson and Morris reach has lost about 24%, 37%, 45% and 50% of $s$ at 8000 years B.P. by 6000, 4000, 2000 and 0 years B.P., respectively (0 B.P. = A.D. 1950). This implies that $\omega$ has been greater in the past, perhaps as great as 9.9 W m$^{-2}$ at 8000 years B.P. when $s$ was highest, assuming a constant $Q$ (Table 5). However, this estimate may be high as $Q$ undoubtedly was lower than present prior to 5400–4500 cal years B.P. when climate was warmer–drier in the drainage basin, as summarized above. Nevertheless, a range of 9.9–5.1 W m$^{-2}$ probably is reasonably representative of the magnitudes of $\omega$ since 8000 years B.P. These magnitudes are low in absolute terms; and variations within this range are not deemed to have caused significant change to the rate of fluvial erosion of the cohesive glaciolacustrine sediments that form the outer banks, regardless of whether $\omega$ was higher pre-6000 years B.P. than afterward.

The long-term change in sediment supply ($G$) through the study area is not known but can be assessed qualitatively by consideration of experimental drainage basin studies on sediment supply and the scenario that led to the establishment of the Red River in southern Manitoba. The river became established during the final northward recession of glacial Lake

Table 5
Temporal change in valley slope and unit stream power, Emerson to Morris section of the Red River

| Age (years B.P.) | Valley slope[a] | Unit stream power (W/m$^2$)[b] | % of unit stream power at 8000 years B.P. |
|---|---|---|---|
| 0 | 0.000064 | 5.1 | 50.8 |
| 2000 | 0.000070 | 5.5 | 55.1 |
| 4000 | 0.000080 | 6.2 | 62.7 |
| 6000 | 0.000097 | 7.6 | 76.2 |
| 8000 | 0.000127 | 9.9 | 100.0 |

[a] The valley slope is based on the mean gradient of the "normal" water surface (about 750 m$^3$/s) between Emerson and Morris (236 and 233.3 m asl, respectively, over a valley distance of 41.9 km), using data from Red River Basin Investigation (1951). The change in gradient was determined by estimating the past elevations of the start and endpoints of this water surface at 2000-year-B.P. intervals according to the equation:

$$E_t = E_p A(e^{t/\tau} - 1)\,(\text{Lewis et al., 2000, after Peltier, 1994 and Tackman et al., 1998})$$

where $E_t$ is elevation at time $t$; $E_p$ is present elevation (m); $t$ is age when elevation is desired (calendar years B.P.); $\tau$ is relaxation time (3500 years); $A$ is site amplitude factor of uplift, defined according to:

$$A = \text{Ru}/(e^{10503/\tau} - 1)\,(\text{after Lewis et al., 2000})$$

where Ru is site relative uplift since 10,503 cal years B.P. Relative uplift for each gradient endpoint was determined using the difference in present elevation of the Lower Campbell beach of Lake Agassiz for the Emerson and Morris locations and the southern outlet of the lake (288.11 m asl; Teller and Thorleifson, 1983). The relative uplift values are 33.71 and 28.06 m for Morris and Emerson, respectively, based on Lower Campbell shoreline elevations where Lower Campbell isobases project through Morris and Emerson (Teller and Thorleifson, 1983). The date of 10,503 cal years B.P. is the age of the Lower Campbell beach (Teller et al., 1996). The relaxation rate of 3500 follows Lewis et al. (2000). Valley slopes are shown to six significant figures to demonstrate a progressive, albeit slight, loss of slope occurring since 8000 years B.P.

[b] Unit stream power ($\omega$) calculated according to $\omega = Qs\gamma w^{-1}$, where $Q$ is discharge [~ 2-year flow (600 m$^3$/s) for Red River at Emerson station (1883–1885, 1892–1896, 1898–1912–1997; data from Manitoba Water Resources)], $s$ is valley gradient, $\gamma$ is the specific weight of clear water (9800 N/m$^3$) and $w$ is channel width (75 m).

Agassiz, which caused a progressive drop in river base level and allowed the north-flowing ancestral river system to extend onto and incise the lake bed in NE North Dakota, NW Minnesota and southern Manitoba. Studies on sediment supply from experimental drainage basins described by Schumm (1977) and Schumm et al. (1987) indicated that the rapid lowering of base level results in a dramatic increase in sediment yield rates at the mouth of the drainage basin. These rates decline rapidly over time, but are highly variable and eventually stabilize to form a quasi-equilibrium. Variability in the rates can produce intermittent secondary peaks in the sediment yield and is attributed to the storage and periodic flushing of alluvium from the tributary valleys within the experimental drainage basins. Overall, however, the net trend from the experimental work is that a drop in base level produces sediment yield rates that are initially high and generally wane rapidly over time.

This sediment yield model is considered generally applicable to the Red River in southern Manitoba because the basic process—the response of a fluvial system to a drop in base level—is similar. Although not located at the river mouth, the study area is situated well within the area of lake bottom exposed during the final recession of the lake. Thus, sediment transported by the river in response to the drop in base level would be routed through the study area reach, and the basic pattern of sediment yield summarized above should represent the sediment supply at this specific location. Such a hypothesis is consistent with the paraglacial sedimentation model described by Church and Ryder (1972) that generally applies to areas within or adjacent to formerly glaciated areas. This model characterizes fluvial sedimentation rates as being high during and immediately following a glaciation because of the fluvial reworking of glacial drift in the landscape. These enhanced rates decrease over time as the supply of sediment that is readily available to the fluvial system wanes but may last for several thousand years (Church and Ryder, 1972). In the Red River case, the high sedimentation rates reflect the former presence of a large glacial lake per se rather than the direct presence of glacial ice. However, the existence and ultimately the demise of Lake Agassiz was controlled by the Laurentide Ice Sheet which represents a fundamental connection to glaciation.

Flume experiments have demonstrated that the rate of lateral migration of a meandering channel can be increased by augmenting the sediment supply into the flume (Hickin, 1988). Based on the experimental sedimentation model and the hypothesized pattern of sediment yield, an initial period of high lateral migration rates should occur that wanes and eventually stabilizes. This indeed is the general trend in the lateral migration rates observed in Fig. 6A. The long-term change in sediment supply thus provides a reasonable explanation for the early phase of high lateral channel migration and flood-plain construction. An obvious question, however, is how an increased sediment supply specifically enhances lateral accretion along a mud-dominated river like the Red River as opposed to sand- or gravel-bed rivers. Whether or not the river experienced a complex response during the pre-6000-years-B.P. period that resulted in episodes of aggradation and degradation within the river valley, as described by Schumm (1977), is not apparent from the flood-plain cores or the basic morphology of the valley bottom.

### 8.2. Relevance of valley widening to the flood hazard

The radiocarbon ages from the flood-plain cores demonstrate that outward extension and downvalley rotation of the meanders has gradually widened the valley (Fig. 5). To assess the significance of this widening to the modern flood hazard on the adjacent clay plain, the cross-sectional area of the valley bottom at 1000 years B.P. was estimated and compared to that of the modern valley. The valley cross-sections were measured across the apex of both valley meanders. The modern valley profile is based on a 1951 map of the alluvial valley where relief is depicted at a 5-ft contour line interval (Red River Basin Investigation, 1951). The valley profile at 1000 years B.P. was estimated by rescaling the modern profile of both meanders to fit the position of the inner bank as defined by the 1000-years-B.P. isochrone. The modern channel shape was used in both sets of profiles; thus, the variation in area occurs over the flood-plain portion of the valley bottom. The rate of incision of the channel has been ignored, as this is low (see above).

The data reveal that the valley cross-section increased by about 2% and 0.7% at the upstream and downstream meanders, respectively, since 1000 years B.P. (Table 6). In absolute terms, this represents the addition of about 52 and 19 $m^2$ to the valley cross-sections (Table 6). The limited increase reflects the general low rate of lateral channel migration along both meanders, but more importantly, the very shallow depth of the flood-plain portion of the valley bottom relative to the clay plain. Hence, only a small amount of area was added to the valley cross-section since 1000 years B.P. as the river meanders widened the valley bottom. The valley widening produced a decrease in hydraulic radius—the ratio of the cross-sectional area to wetted perimeter of the profiles and a common hydraulic morphological parameter—of 2.64–2.56 and 2.01–2.00 m at the upstream and downstream meanders, respectively, between 1000 and present (Table 6). These decreases represent a change in hydraulic radius of about 3% and 0.5% (Table 6).

The section of river valley within the study area does not control the water surface of valley-full flows. Nevertheless, the low rates of lateral channel migration and incision along the study area meanders are considered to be typical of meanders along the Red River elsewhere in southern Manitoba. Thus, since 1000 years B.P., the river meanders have experienced only a slight change in valley cross-section and hence to the morphological parameter (i.e., hydraulic radius) affecting the discharge capacity of the valley. When considered proportionally over timescales of up to several centuries—timescales that are directly relevant to the modern flooding problem—the amount of widening of the valley cross-section is very low to negligible and the resulting change in total valley conveyance would be undoubtedly within the error of discharge measurement. Overall, therefore, valley widening is deemed an insignificant factor affecting the modern flood hazard on the clay plain, particularly compared to other variables such as, for example, climatic fluctuations (see Ashmore and Church, 2001; St. George and Nielsen, 2002) and changes in valley bottom roughness from agricultural land clearance.

## 9. Summary and conclusions

The radiocarbon age chronology of lateral accretion deposits at the study area span ~ 8100 cal years B.P. and reveals that the flood plain is the product of a single sequence of meander expansion. The general rate of channel migration has decreased over time and about half of the flood plain formed in the initial third of the post-Agassiz timescale. At the upstream meander, the average rates of channel migration were initially 0.35 m/year between ~ 7900 and 7400 cal years B.P., then decreased to 0.18 m/year between ~ 7400 and 6200 cal years B.P., and varied between 0.04 and 0.08 m/year since ~ 6200 cal years B.P. Both meanders in the study area have experienced about 0.04 m/year of lateral channel migration since ~ 1000 years B.P.

The interval experiencing the most rapid lateral channel migration coincides with the warmest–driest period of the Holocene in the eastern Prairies. The higher phase of Red River alluviation is interpreted to reflect an enhanced sediment supply to the study area arising from the development of the river system on the clay plain following the final recession of glacial Lake Agassiz from NE North Dakota, NW Minnesota and southern Manitoba. Other meandering river sys-

Table 6
Change in valley cross sectional area and hydraulic radius along the two study area meanders between 0 and 1000 years B.P.

| Meander (transect) | Time (years B.P.) | Valley cross-sectional area ($m^2$)[a] | % of valley cross-sectional area at 1000 years B.P. | Wetted perimeter (m)[a] | Hydraulic radius (m)[b] | % of hydraulic radius at 1000 years B.P. |
|---|---|---|---|---|---|---|
| Upstream | 0 | 2558 | 102 | 1000 | 2.56 | 103 |
| (99RR3) | 1000 | 2506 | 100 | 950 | 2.64 | 100 |
| Downstream | 0 | 2638 | 100.7 | 1321 | 2.00 | 100.5 |
| (99RR1) | 1000 | 2619 | 100 | 1300 | 2.01 | 100 |

[a] Measured with a Ushikata X-PLAN360d planimeter.
[b] Hydraulic radius = ratio of cross-section area to wetted perimeter.

tems also may have experienced a general decline in the long-term rates of lateral channel migration in response to waning post-glacial sediment supply, but such declines could have gone unrecognized because of a lack of sufficient dating control of the flood-plain deposits, the lack of preserved (or very fragmentary) older areas of the flood plain and/or the greater relative importance of other variables controlling channel migration.

The main phase of incision by the Red River into the Lake Agassiz clay plain occurred between ~ 8900 and 8100 cal years B.P. Since ~ 8000 cal years B.P., gradual incision of the river has occurred at a rate estimated to be between 0.4 and 0.8 m/ky.

The lateral migration of the river meanders is causing the gradual widening of the alluvial river valley. The resulting increase in valley cross-sectional area since 1000 years B.P. is about 2% and 0.7% at the two study area meanders, respectively. The change in cross-section translates into a decrease in hydraulic radius of about 3% and 0.5% at the two meanders. When considered proportionally over timescales of up to several centuries, the widening of the valley cross-section is very low to negligible and is deemed an insignificant factor affecting the modern flood hazard on the clay plain, especially compared to other variables (e.g., climate fluctuations and changes in valley bottom roughness from land clearance).

This study demonstrates that a low-energy, mud-dominated river can experience progressive, albeit slow, lateral channel migration—the basic evolution of which is consistent with other meandering river types. However, for the geomorphic behaviour to be fully displayed, the development of the flood plain needs to be examined over a timescale of relatively long duration (a "geomorphic" timescale of Hickin, 1983), which in the case of this study spans thousands of years.

When comparing the Red River alluvial history to other temperate, mid-latitude river systems, the late Quaternary history and geomorphic setting of the various systems needs to be considered carefully. In the case of the Red River in southern Manitoba, the mud-dominated river was established after the final recession of a large glacial lake; it developed on a flat, gently sloped clay plain that was the former lake bed, and its past discharge regime lacks a glacially influenced or glacial-lake-influenced stage.

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