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The ups and downs of the Missouri River from Pleistocene to present: Impact of climatic change and forebulge migration on river profiles, river course, and valley fill complexity

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ABSTRACT

The Missouri River is a continent-scale river that has thus far escaped a rigorous reporting of valley fill trends within its trunk system. This study summarizes evolution of the lower Missouri River profile from the time of outwash in the Last Glacial Maximum (LGM) until establishment of the modern dominantly precipitation-fed river. This work relies on optically stimulated luminescence (OSL) dating, water-well data, and a collection of surficial geological maps of the valley compiled from U.S. Geological Survey EDMAP and National Science Foundation Research Experience for Undergrads projects. Mapping reveals five traceable surfaces within valley fill between Yankton, South Dakota, USA, and Columbia, Missouri, USA, that record two cycles of incision and aggradation between ca. 23 ka and ca. 8 ka. The river aggraded during the LGM to form the Malta Bend surface by ca. 26 ka. The Malta Bend surface is buried and fragmented but presumed to record a braided outwash plain. The Malta Bend surface was incised up to 18 m between ca. 23 ka and ca. 16 ka to form the Carrolton surface (ca. 16 ka to ca. 14 ka). The Carrollton surface ghosts a braided outwash morphology locally through overlying mud. Aggradation followed (ca. 14 ka to ca. 13.5 ka) to within 4 m of the modern floodplain surface and generated the Salix surface (ca. 13.5 to ca. 12 ka). By Salix time, the Missouri River was no longer an outwash river and formed a single-thread meandering pattern. Reincision at ca. 12 ka followed Salix deposition to form the short-lived Vermillion surface at approximately the grade of the earlier Carrolton surface. Rapid aggradation from ca. 10 ka to ca. 8 ka followed and formed the modern Omaha surface (ca. 8 ka to Present). The higher Malta Bend and Omaha profiles are at roughly the same grade, as are the lower Carrolton and Vermillion surfaces. The Salix surface is in between. All surfaces converge downstream as they enter the narrow and shallow bedrock valley just before reaching Columbia, Missouri. The maximum departure of the profiles is 18 m near Sioux City, Iowa, USA, at ~100 km downstream from the James Lobe glacial input near Yankton, South Dakota. Incision and aggradation appear to be driven by relative changes in input of sediment and water related to glacial advance and retreat and then later by climatic changes near the Holocene transition. The incision from the Malta Bend to the Carrolton surface records the initial breakdown of the cryosphere at the end of the LGM, and this same incisional event is found in both the Ohio and Mississippi valleys. This incisional event records a "big wash" that resulted in the evacuation of sediment from each of the major outwash rivers of North America. The direction and magnitude of incision from the LGM to the modern does not fit with modeled glacioisostatic adjustment trends for the Missouri Valley. Glaciotectonics likely influenced the magnitude of incision and aggradation secondarily but does not appear to have controlled the overall timing or magnitude of either. Glaciotectonic valley tilting during the Holocene, however, did likely cause the Holocene channel to consistently migrate away from the glacial front, which argues for a forebulge axis south of the Missouri Valley during the Holocene and, by inference, earlier. This is at least 200 km south of where models predict the Holocene forebulge axis. The Missouri Valley thus appears to reside in the tectonic low between the ice front and the forebulge crest. The buffer valley component of incision caused by profile variation could explain as much as 25 m of the total ~40 m of valley incision at Sioux City, Iowa. The Missouri Valley also hosted a glacial lobe as far south as Sioux City, Iowa, in pre-Wisconsinan time, which is also a factor in valley excavation.

INTRODUCTION

The Missouri River (Fig. 1) is assumed to be a glacial-front system that periodically carried outwash sediments (Bluemle, 1972; Ruhe, 1983; Blum, 2019). This inference is anchored mostly on the position of the bedrock valley (Bluemle, 1972; Ruhe, 1983) and associations of the valley with adjacent loess deposits (Frye et al., 1948; Frye and Leonard, 1952; Ruhe, 1983; Forman and Pierson, 2002). Direct records of development for the Quaternary river within its valley, however, are very limited (Guccione, 1983; Holbrook et al., 2006a). Bedrock valley trends argue that the Missouri River drainage evolved throughout the Pleistocene in response to advancing ice sheets and was alternately shaped by precipitation and meltwater sources (Flint, 1947; Bluemle, 1972; Kehew and Teller, 1994; Catto et al., 1996). Lack of detailed mapping and dating within the valley stratigraphy, however, limits direct inference of the late glacial history for this continental-scale river. The degree to which glacioisostatic adjustment from glacial loading and unloading further impacted valley position and depth also remains unknown.

This study summarizes the late Pleistocene to Holocene evolution of the lower Missouri River and will provide the first evidence of aggradation and incisional trends within the valley proper. Detailed mapping and optically stimulated luminescence (OSL) dating of fluvial valley fill strata are used to reconstruct pattern and profile for the Missouri River from the Last Glacial Maximum

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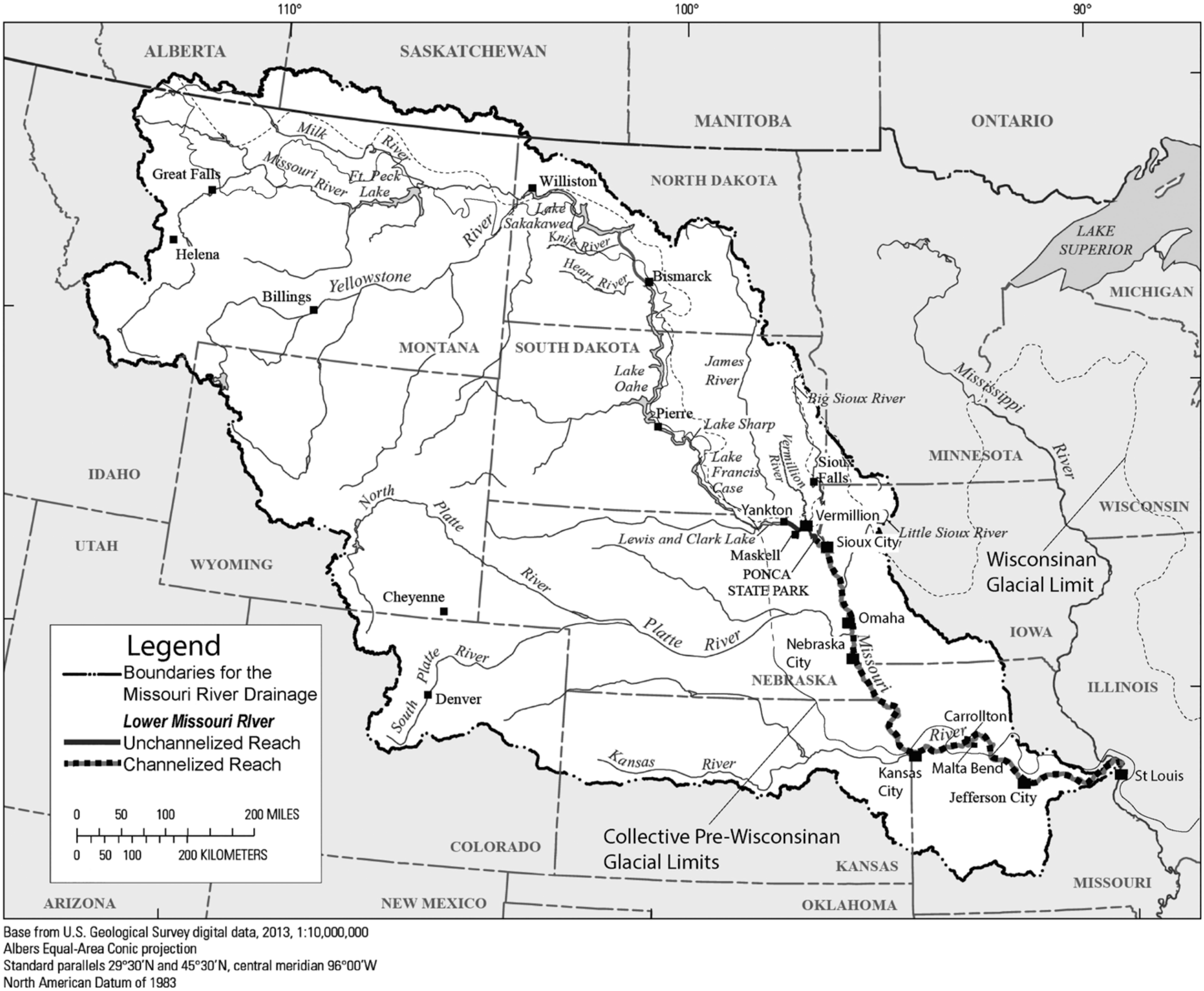


Figure 1. Map shows Missouri River drainage and locations of rivers and towns (modified from Rus et al., 2015). Glacial limits are from Soller and Reheis (2004) and Mickelson and Colgan (2003).

(LGM) to the present. Glacial systems are generally typified by highly variable sediment and water input related to climate extremes during glacial cycles and are thus typified by large aggradation and incision cycles (e.g., Mack, 2007; Howard et al., 2008; Counts et al., 2015). Similarly, some amount of uplift likely occurred across the Missouri River Valley as flexural loading of Pleistocene ice sheets triggered a forebulge south of glacial fronts, and uplift occurred in areas depressed by ice during rebound when ice sheets retreated. It is thus possible that some component of recent valley incision cycles reflects denudation processes driven by glaciotectonic cycles. This study defines and quantifies the incision/aggradation pattern and migration history of the Missouri River in its lower valley and tests these trends for the relative impacts of glacial/interglacial climatic change and glaciotectonics as drivers.

CONTINENTAL GLACIATION AND THE MISSOURI RIVER VALLEY: A REVIEW OF PRIOR WORK

Overview

The Missouri River developed the current general west-east drainage pattern sometime in the early Pleistocene (Pre-Illinoian), but its drainage patterns likely shifted over repeated glaciations and only settled into the current location during the Wisconsin stage (Bluemle, 1972; Kehew and Teller, 1994; Catto et al., 1996; Galloway et al., 2011; Fildani et al., 2018; Blum, 2019). The valley likely received outwash beginning at some unknown point approaching the Last Glacial Maximum (LGM, 26.5–19 ka; Palacios et al., 2020) and continued over early parts of a subsequent retreat marked by brief stadial and interstadial episodes (Fig. 2). In general, ice volumes in North America during the LGM were relatively constant or slowly advancing until an initial phase of deglaciation that started at ca. 20.5 ka and continued to ca. 18 ka (Dyke, 2004; Palacios et al., 2020). Ice volumes were relatively constant from ca. 18 ka to 16.5 ka until deglaciation resumed from ca. 16.5 ka to 12.5 ka. After a brief stable period from ca. 15 ka to 14.5 ka., melting resumed from ca. 14.6 ka to 14.0 ka associated with the beginning of

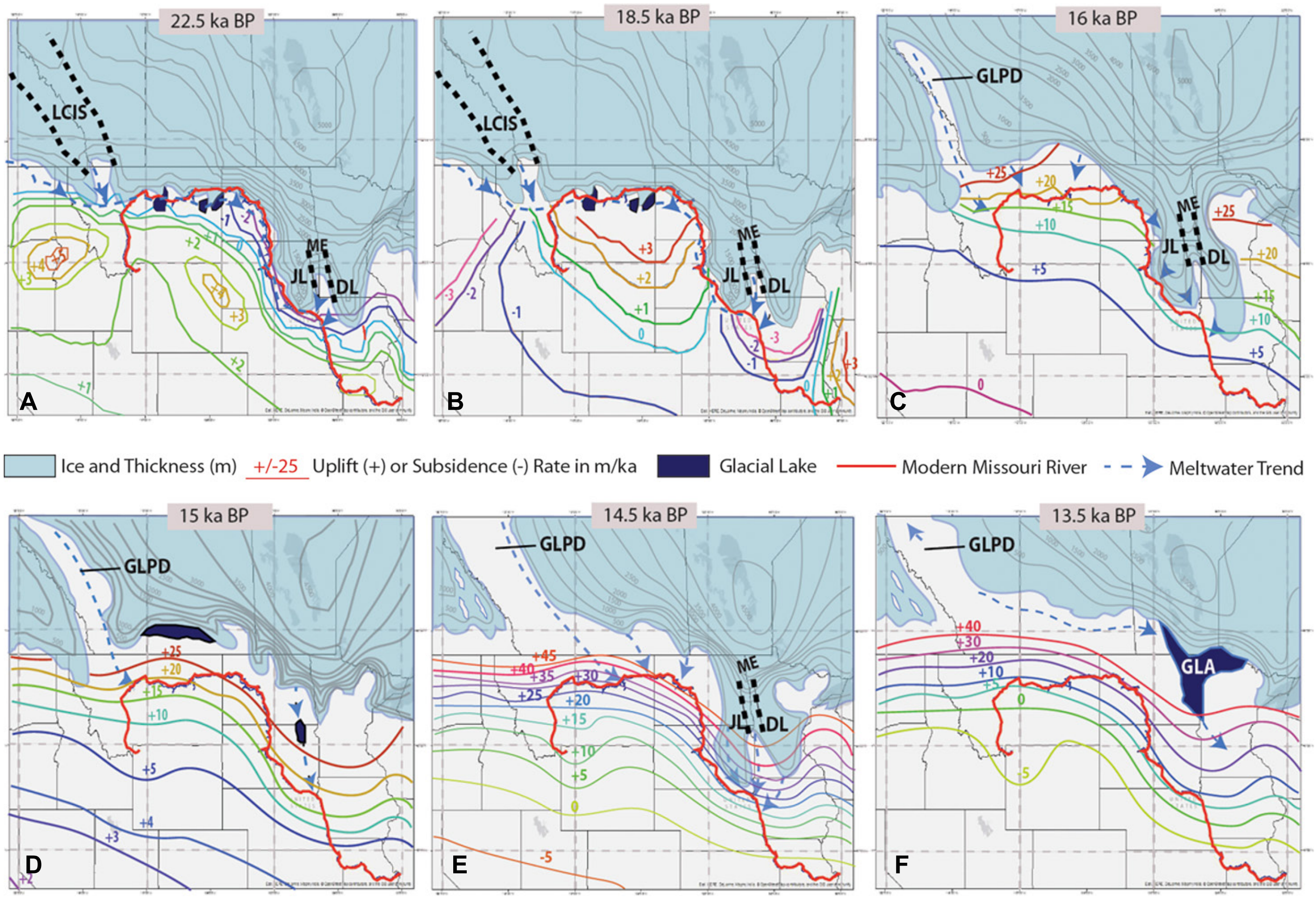


Figure 2. Proposed drainage, ice thickness, and glacioisostatic adjustment (GIA) rates are given for the Missouri Valley at (A) 22.5 ka, (B) 18.5 ka, (C) 16 ka, (D) 15.5 ka, (E) 14.5 ka, and (F) 13.5 ka. Glacial ice fronts are modified from Dyke (2004). Glacial Isostatic Adjustment (GIA) and ice thicknesses are modified from the ICE-6G (VM5a) model of Peltier et al. (2015) and Argus et al. (2014). Values for GIA contours are modeled rates for vertical movement, reflect the difference between topographic elevations generated by the model over the millennium centered at the expressed time, and are presented in units of m/ka. Positive numbers represent rebound, and negative numbers represent subsidence. Ice thicknesses are contoured in meters. Drainage pathways into the Missouri River are extrapolated from literature review. LCIS—Laurentide-Cordilleran Ice Saddle, ME—Missouri Escarpment, JL—James Lobe, DL—Des Moines Lobe, GLPD—Glacial Lake Peace drainage area, GLA—Glacial Lake Agassiz.

the Bølling – Allerød (14.6 ka to 12.9 ka; Palacios et al., 2020) warm period (Dyke, 2004; Lambeck, 2004; Lambeck et al., 2014). Glaciation resumed from 12.9 ka to 11.5 ka during the Younger Dryas (12.9 ka to 11.7 ka; Palacios et al., 2020) cool period (Dyke, 2004; Lambeck, 2004; Lambeck et al., 2014). Melting resumed in the early Holocene from 11.5 ka but slowed at ca. 8.2 ka and again at 6.7 ka as melting came to an end (Lambeck et al., 2014). These larger trends of ice advance and retreat do not consistently reflect advance and retreat of the lobes and sublobes impacting the Missouri River drainage because of localized effects (Dyke et al., 2002; Lundstrom et al., 2009; Lambeck et al., 2014). These effects and variations are considered in more detail below.

The Pre-LGM Missouri Valley

The Missouri River carried glacial outwash prior to the LGM, potentially as early as the early Pleistocene, and was diverted multiple times during glacial advances up until the most recent rerouting during the Late Wisconsin. Pre-Illinoian tills, originally mapped collectively as "Nebraskan" tills, extend over multiple locations within the Missouri River drainage, including northcentral and western Missouri, USA; Iowa, USA; Nebraska, USA; and Kansas, USA (Reed and Dreeszen, 1965; Bayne et al., 1971; Hallberg, 1980a, 1980b; Guccione 1983) (Fig. 1). Flint (1947) and Warren (1952) dated diversions of the White River Valley into the Missouri River drainage system around Chamberlain, South Dakota, USA, as Pre-Wisconsin. The presence of the Illinoian Loveland Silt in multiple locations along the Missouri River Valley argues that the river carried glacial outwash during Illinoian glaciations (Forman et al., 1992; Forman and Pierson, 2002). Blumele (1972) argued for rerouting of the Missouri River in North Dakota, USA, in the early Wisconsin when ice advances blocked wide west-east and southwest-northeast–trending valleys, creating a southward-trending, ice-margin Missouri Valley. During the mid-Wisconsin interstadial (ca. 32–35 ka), the Missouri drainage was ice-free (Hill and Feathers, 2002), and the ice margin approximated the area of the Canadian Shield (Dyke et al., 2002). Additional glacial diversions of the modern Missouri River headwaters into the lower valley continued throughout the late Wisconsin (Bluemle, 1972; Kehew and Teller, 1994; Catto et al., 1996).

The LGM-Holocene Missouri Valley

Laurentide ice margins reached the LGM position around 21.4 ka (Fig. 2A), coinciding with the last eustatic global sea level minimum (Dyke, 2004; Lambeck et al., 2014). Ice margins advancing toward the LGM position pushed south of the Missouri Escarpment, a major drainage divide that separated flow between the Missouri River and the Mississippi/Lake Agassiz system to the east (Dyke, 2004; Lambeck et al., 2014). LGM drainage of the southwestern section of the Laurentide ice sheet was thus down the Missouri River (Kehew and Teller, 1994). The LGM ice margin blocked tributaries of the Missouri River in several locations forming glacial lakes (Fig. 2). The Loma sublobe blocked the Missouri River north of Highwood Mountains, Montana, USA, forming glacial Lake Great Falls (Fig. 2B) (Hill and Feathers, 2002). Other glacial lakes including glacial lakes Musselshell and Jordan formed around Fort Peck, Montana (Colton et al., 1961). It is possible these glacial lakes acted as sediment traps and may have affected the downstream sediment budget of the lower Missouri River Valley. The most proximal sources of water and sediment from glacial drainage for the lower Missouri Valley were the James Lobe, which intruded the Missouri Valley near Yankton, South Dakota, and the Des Moines Lobe, which supplemented drainage to the Missouri Valley through the Big Sioux River tributary flowing along the modern Iowa/South Dakota boarder (Lundstrom et al., 2009) (Figs. 1 and 2).

Deglaciation began toward the end of the LGM around 20.5 ka and continued until ca. 16.5 ka (Dyke, 2004; Lambeck et al., 2014) (Fig. 2C). The glacial ice front retreated northward, and between 17 ka and 16 ka (Fig. 2C) the ice saddle between the Cordilleran and Laurentide ice domes began ablating (Dyke, 2004). By ca. 16 ka, the Maskwa drainage corridor reorganized into the Buffalo corridor in southern Saskatchewan, sending southwestern Laurentide ice directly down the James Lobe tributary of the Missouri River (Ross et al., 2009). This collectively increased the Missouri River Valley drainage area significantly, though the potential for sediment traps in the Laurentide-Cordilleran ice saddle and the lower Missouri Valley remains a possibility.

Drainage from the James and Des Moines lobes to the Missouri River changed significantly between 15.8 ka and 12.8 ka, beginning at and during the Bølling-Allerød global warm period. Accelerated warming caused complete melting of the James and Des Moines Lobes by 15.0 ka (Fig. 2D) (Dyke, 2004). The presence of white spruce (*picea glauca*) around the South Dakota/Nebraska border at 15 ka supports the disappearance of the James Lobe and the presence of a non-tundra landscape at this place and time (Yansa, 2006). During this brief disappearance of the James Lobe, drainage still continued through the James River Valley because ice from the Red River Lobe blocked the eastward flow of meltwater from the western glacial front toward Lake Agassiz and the Mississippi River drainage (Kehew and Teller, 1994). By 14.5 ka, the James Lobe had readvanced back to the Yankton and Richland, South Dakota, area (Fig. 2E), approximating its previous southernmost boundary (Fig. 2B) (Lundstrom et al., 2009). It is unknown if a glacial lake was present during this brief disappearance of the James Lobe, although it is possible that glacial lakes similar to Lake Souris and Lake Dakota existed as reported in the later James phase of drainage by Kehew and Teller (1994). By 13.9 ka, the James and Des Moines Lobes retreated from the area for good (Fig. 2F) (Dyke, 2004). Glacial Lake Agassiz began forming around 14.3 ka (Lepper et al., 2011), and drainage of the southwestern Laurentide Ice Sheet soon switched (13.9 ka) eastward, flowing into the Lake Agassiz drainage and then down the Minnesota and Mississippi Rivers (Clayton and Moran, 1982). These ice models show the capture of the James spillway by Lake Agassiz around 13.9 ka (Fig. 2F), which marks the end of the Missouri River as a glacial drainage system.

Glacial Isostatic Rebound and Forebulge Migration in and Near the Missouri Drainage

The lower Missouri Valley sits in an area modeled to include migration of a glacial forebulge and co-subsidence along the ice front. Magnitude and rates of this glacial isostatic adjustment (GIA) are dependent on the total amount of ice loaded onto the crust and the viscosity of the mantle upon which the ice is loaded (e.g., Peltier, 2002). Isostatic adjustment since the LGM is both observed in the modern and modeled over the past across North America (e.g., Tushingham and Peltier, 1991; Mitrovica et al., 1993; Clark et al., 1994; Peltier, 2002; Peltier, 2004; Sella et al., 2007; Argus et al., 2014; Peltier et al., 2015; Wickert et al., 2019). The models are based on input values of mantle viscosity, ice-thickness, and position of glacial ice fronts and evolve as more data become available. Modeling efforts are also tuned using constraints from very-long-baseline interferometry (VLBI) (James and Lambert, 1993; Mitrovica et al., 1993, Argus et al., 1999), GPS (Sella et al., 2007), and absolute gravity measurements (Larson and van Dam, 2000; Lambert et al., 2001). GPS observations by Sella et al. (2007) suggest that the current "0" line, which divides glacial subsidence (i.e., forebulge collapse) from glacial rebound (i.e., recovery from ice-front subsidence), extends through the Great Lakes and the northern extremities of the Missouri drainage. However, GIA boundaries and elevation trends migrate over time. GIA can impact fluvial channels by tectonically modifying channel slopes, affecting vertical profiles as well as lateral migration trends (cf. Holbrook and Schumm, 1999; Schumm et al., 2000; Holbrook et al., 2006b).

This study utilizes the ICE-6G (VM5a) model of GIA since the LGM by Argus et al. (2014) and Peltier et al. (2015) to approximate the location, magnitude, and migration of glacial subsidence and uplift relative to the Missouri Valley since the LGM (Fig. 2). This model is used as an example in this study because it represents a recent and thorough model that shows the expected timing and magnitude of GIA throughout the lower Missouri Valley. The model also represents the expectations for local migration of a forebulge deforming the surface by magnitudes of 10^1 m, which is a scale that should have impacted channel slope. Thus, it offers an opportunity to test if the Missouri River fluvial record reflects the trends expected from formation and migration of a forebulge of this magnitude within this part of the continent.

METHODS

This study integrates data from surficial geologic maps, water-well logs, and OSL dating to define fluvial allounits, determine their ages, and construct longitudinal profiles for Missouri Valley stratigraphic units and surfaces. Surficial geologic maps help define the distribution and characteristics of surficial units, which are then dated using OSL. Subsurface data from water wells are used to map units not exposed at the surface, and these surfaces are also dated from separate drill holes using OSL. Both maps and subsurface data are used together to develop elevation profiles for unit surfaces.

The data used here represent an analysis of information collected over several years of research during multiple student projects as well as additional new information that was collected specifically for this study. Surficial geologic maps developed here are compilations from a series of 61 7.5′ geologic quadrangles of the Missouri Valley produced by undergraduate and graduate students of the "Big Muddy Expedition" (Alexandrowicz et al., 2011; Amadi et al., 2011; Anderson et al., 2011a, 2011b; Avdeev et al., 2011; Baak et al., 2011; Brown et al., 2011; Burt et al., 2011; Carlin et al., 2011; Carritt et al., 2011; Caster et al., 2011; Cordova et al., 2011a, 2011b, 2011c; Dolde et al., 2011; Egyed et al., 2011; Emenhiser et al., 2011a, 2011b; Farley et al., 2011; Garrett et al., 2011; Ghimire et al., 2011; Hildebrant et al., 2011; Jobe et al., 2011; Kennedy et al., 2011a, 2011b; Kliem et al., 2011a, 2011b; Leddy et al., 2011; Macklin et al., 2011; Main et al., 2011; Markson et al., 2011; Meyers et al., 2011; Moreno et al., 2011a, 2011b, 2011c, 2011d; Newman et al., 2011; Noah et al., 2011; Nzewunwah et al., 2011a, 2011b; Owinyo et al., 2011; Pagan et al., 2011; Peterson et al., 2011; Radakovich et al., 2011a, 2011b; Reed et al., 2011; Rios et al., 2011; Tanksley et al., 2011; Trimble et al., 2011; Allen et al., 2013; Baylor et al., 2013; Cloos et al., 2013; Gose et al., 2013; Long et al., 2013; Rienstra et al., 2013; Shelley et al., 2013; Wagner et al., 2013; Adams et al., 2014; DuBose et al., 2014; Woodworth et al., 2014a, 2014b). These maps were produced as part of a research and educational mapping program funded by grants from the USGS EDMAP program and the National Science Foundation Research Experience for Undergrads program and performed over the summers of 2004–2012 under the supervision of one of the authors (J. Holbrook). These maps are compiled in Kashouh (2012) and Anderson (2014) and were published individually and are available through the University of South Dakota Press (see http://mri.usd.edu/pubsfor full compendium). These maps identify allounits (e.g., channel fills, bars, splays, etc.) using surficial characteristics visible in aerial photos, topographic maps, digital elevation models, and field traverses and were tested for lithofacies characteristics with Dutch-augered boreholes (after Berendsen and Stouthamer, 2001). Core samples from the Dutch augers were taken at 10 cm intervals and logged for Munsell color, texture, oxidation state, and organic traces (after Soil Survey Division Staff, 1993). These data were supplemented during this study with public state water-well data from South Dakota, Nebraska, Iowa, Kansas, and Missouri that were mined from the archives of these respective geological surveys and used to map and trace subsurface valley fill units. Subsurface data are integrated into representative cross sections and longitudinal profiles over discrete representative valley lengths dispersed over the lower valley. Additionally, this project includes 53 OSL samples (Tables 1 and 2), which were mostly collected and processed during the Big Muddy Expedition, but with some collected specifically for this study. Crosscutting relationships were combined with OSL dating to develop relative and numerical age sequences for surface and subsurface allounits.

OSL Sampling and Dating Procedure

OSL ages for this study were measured on quartz collected from sand intervals of point bars and channel fills that are generally covered by finer-grained loess, levee, and overbank deposits. Sample sites were selected from landforms recording major shifts in allounit deposition as determined from cross-cutting relationships and elevation changes. Samples were each recovered from various depths in logged boreholes using a suction-coring device formed from an opaque sampling tube mounted to the point of a hand auger with an intervening check valve. All holes were opened with Dutch augers before sampling except the two deepest holes near Elk Point, South Dakota, which were opened with a motorized truck-mounted auger before sampling. Information recorded in the field for samples included lithology, elevation, burial depth, and latitude and longitude.

Lab processing was performed at University of Nebraska at Lincoln. Sample preparation was performed under amber light. Lab processing techniques included isolating sand-sized particles (90–150 μm and 150–250 μm) via wet sieving techniques. Sand-sized particles were then treated with HCL and hydrogen peroxide to remove carbonates and organics. Samples were then floated in a 2.7 g/cm^3 sodium polytungsate solution to separate quartz and feldspar grains from heavier minerals. Feldspars were then removed from quartz grains by treating samples with 48% HF acid for 75 min followed by 47% HCL for 30 min. The samples were then resieved to remove feldspar fragments leftover from the HF acid treatment. Quartz grains were then mounted on the innermost 2–5 mm of a 1 cm disk using a silicone-based spray (Silkospray). Optical measurements were carried out using Riso Automated OSL Dating System Models TL/OSL-DA-15B/C and TL/OSL-DA-20 with blue and infrared diodes and used Single Aliquot Regenerative Dose techniques (Murray and Wintle, 2000). Early background was subtracted using methods from Ballarini et al. (2007) and Cunningham and Wallinga (2010). Preheat plateau tests between 180° C and 280° C were used to find appropriate preheat and cutheat temperatures. Growth curves were then examined according to Wintle and Murray (2006) to determine if samples were below saturation. A minimum of 50 aliquots were used unless samples did not contain enough sand-sized grains. Individual aliquots were constantly measured for large errors compared to the average, and aliquots determined to be unacceptable compared to the average were left out when averaging aliquot data sets. Central Age Model (Galbraith et al., 1999) was used for D_e unless D_e distributions indicated the Minimum Age Model (Galbraith et al., 1999) was more appropriate according to a decision table by Bailey and Arnold (2006).

ALLOSTRATIGRAPHIC UNITS AND SURFACES

Mapping revealed five alloformations with five corresponding upper surfaces (Figs. 3 and 4). Optically stimulated luminescence results define the ages of these surfaces (Tables 1 and 2; Figs. 3 and 5). The bedrock surface flooring the valley fill is also mapped and profiled. Physical characteristics and age relationships of each alloformation and surface are discussed in detail below.

Malta Bend Alloformation

The Malta Bend Alloformation is the oldest unit correlated regionally within this study and formed during the LGM. The Malta Bend Alloformation aggraded to its current elevation by at least ca. 26 ka, lasted until at least ca. 23 ka, and can be correlated from the upper to the lower part of the lower valley. It is named for the town of Malta Bend, Missouri, which sits locally on the Malta Bend terrace. The top of this alloformation, the Malta Bend surface, is defined as the first fluvial deposits under the Peoria Loess. This unit is found, dated, and described from auger holes in two locations: Herman, Nebraska, and the type area near Malta Bend, Missouri (Fig. 5). One pre-LGM age of 31.5 ± 2.2 ka was recovered from an alluvial channel within the valley north of Vermillion, South Dakota, that predates the Malta Bend Alloformation elsewhere. This age confirms that rivers existed in the upper part of the lower Missouri Valley by this time. This channel could not be found elsewhere and was not correlated. The possibility that this is an early part of the Malta Bend surface, rather than a separate and earlier surface, cannot be excluded or confirmed with available data.

In Herman, Nebraska, the Malta Bend channel belt deposits are dated beneath 7 m of Peoria Loess on a high terrace within the valley. This loess-covered terrace was also recognized previously by Dahl (1961) and has a land surface of between 5 m and 10 m above the modern floodplain. This places the Malta Bend channel surface here roughly on grade with the modern river floodplain (Figs. 3 and 4). The transition between the Peoria Loess and the Malta Bend Alloformation is distinguished by a change in grain size and color. The Peoria Loess is a fine, massive, well-oxidized, silty loam (10 yr 5/4) while the underlying Malta Bend Alloformation is capped by less than 1 m of dark (10 yr 5/1) clay with lighter gray mottles

TABLE 1. OPTICALLY STIMULATED LUMINESCENCE AGES COLLECTED THROUGHOUT STUDY AREA AND TWO RADIOCARBON DATES FROM NZEWUNWAH (2003)

Sample number	UNL lab number	Alloformation	Dose rate (Gy/ka)	No. of aliquots	De (Gy)	Optical age (ka)
GN/JBA1	UNL3894	Pre-Last Glacial Max.	1.39 ± 0.09	50	43.89 ± 1.47	31.5 ± 2.2
HE/JBA2B	UNL3896	Malta Terrace	2.55 ± 0.15	55	67.51 ± 3.37	26.4 ± 2.1
VE/MH57	UNL2126	Carrolton (Braided)	1.56 ± 0.08	24	26.27 ± 0.18	16.8 ± 1.0
EN17	UNL2123	Carrolton (Braided)	2.21 ± 0.12	26	30.97 ± 0.90	14.0 ± 1.0
CW41	UNL853	Carrolton (Braided)	1.70 ± 0.08	34	26.12 ± 0.55	15.32 ± 0.91
CW42	UNL848	Carrolton (Braided)	1.71 ± 0.08	31	25.78 ± 0.68	15.04 ± 0.93
CW/JBA3	UNL3895	Carrolton (Braided)	2.14 ± 0.14	60	30.76 ± 0.41	14.4 ± 1.0
EP31	UNL2127	Salix (Meandering)	1.91 ± 0.12	27	25.66 ± 0.83	13.4 ± 1.0
SLX50	UNL2496	Salix (Meandering)	2.03 ± 0.10	44	24.92 ± 0.27	12.2 ± 0.7
EN14	UNL2125	Vermillion	1.97 ± 0.11	26	22.06 ± 0.52	11.2 ± 0.8
VE/MA49	UNL2120	Vermillion	2.08 ± 0.11	33	21.28 ± 0.46	10.2 ± 0.7
NB48	UNL1061	Omaha (Meandering)	2.22 ± 0.10	16	17.77 ± 0.27	8.02 ± 0.43
SL45	UNL2491	Omaha (Meandering)	1.93 ± 0.09	31	14.05 ± 0.47	7.29 ± 0.48
SL47	UNL2492	Omaha (Meandering)	2.12 ± 0.10	32	13.90 ± 0.27	6.56 ± 0.39
SL3	UNL2501	Omaha (Meandering)	2.21 ± 0.10	41	14.56 ± 0.50	6.59 ± 0.42
CE64	UNL850	Omaha (Meandering)	2.18 ± 0.09	35	12.89 ± 0.35	5.92 ± 0.34
CE1	UNL852	Omaha (Meandering)	1.84 ± 0.09	36	9.55 ± 0.29	5.18 ± 0.33
CE60	UNL858	Omaha (Meandering)	2.26 ± 0.14	34	10.86 ± 0.32	4.81 ± 0.35
ONA33	UNL2493	Omaha (Meandering)	2.32 ± 0.14	29	10.86 ± 0.47	4.68 ± 0.37
HA21	UNL1065	Omaha (Meandering)	1.94 ± 0.07	25	9.22 ± 0.30	4.77 ± 0.28
CW54	UNL862	Omaha (Meandering)	1.85 ± 0.08	28	8.19 ± 0.24	4.43 ± 0.27
CO3	UNL3745	Omaha (Meandering)	2.42 ± 0.14	51	10.54 ± 0.25	4.35 ± 0.28
CW60	UNL849	Omaha (Meandering)	1.99 ± 0.11	31	8.31 ± 0.19	4.18 ± 0.28
CE59	UNL851	Omaha (Meandering)	2.05 ± 0.09	27	8.13 ± 0.19	3.98 ± 0.22
BU32	UNL2124	Omaha (Meandering)	2.15 ± 0.11	34	7.79 ± 0.14	3.62 ± 0.24
HA12	UNL1064	Omaha (Meandering)	1.71 ± 0.09	25	6.42 ± 0.14	3.76 ± 0.23
NB3	UNL1070	Omaha (Meandering)	1.77 ± 0.09	40	6.44 ± 0.15	3.65 ± 0.24
CW61	UNL855	Omaha (Meandering)	2.42 ± 0.11	30	8.65 ± 0.38	3.57 ± 0.48
JE55	UNL2128	Omaha (Meandering)	2.37 ± 0.09	42	8.28 ± 0.14	3.49 ± 0.19
CE26	UNL856	Omaha (Transitional)	2.14 ± 0.11	29	7.30 ± 0.17	3.41 ± 0.21
NB49	UNL1060	Omaha (Transitional)	2.37 ± 0.13	29	8.07 ± 0.30	3.41 ± 0.25
NB46a	UNL1062	Omaha (Meandering)	2.27 ± 0.14	28	7.37 ± 0.22	3.25 ± 0.24
LE4	UNL1071	Omaha (Transitional)	2.05 ± 0.10	24	6.60 ± 0.20	3.23 ± 0.21
CW62	UNL854	Omaha (Transitional)	1.92 ± 0.10	30	5.67 ± 0.14	2.95 ± 0.19
CW13	UNL1069	Omaha (Transitional)	2.21 ± 0.12	29	6.28 ± 0.20	2.85 ± 0.20
MS50	UNL624	Omaha (Meandering)	1.87 ± 0.09	45	5.18 ± 0.70	2.77 ± 0.42
BNC7	UNL2819	Omaha (Meandering)	1.44 ± 0.07	50	3.93 ± 0.10	2.72 ± 0.15
CW63	UNL859	Omaha (Transitional)	1.92 ± 0.09	29	5.03 ± 0.10	2.62 ± 0.15
MS49	UNL623	Omaha (Transitional)	2.15 ± 0.10	45	4.98 ± 0.54	2.32 ± 0.30
MB23	UNL861	Omaha (Braided)	2.38 ± 0.09	27	5.52 ± 0.22	2.32 ± 0.14
OSW59	UNL2833	Omaha (Meandering)	2.05 ± 0.11	57	4.45 ± 0.09	2.17 ± 0.12
CE61	UNL860	Omaha (Meandering)	2.21 ± 0.09	24	4.73 ± 0.16	2.14 ± 0.13
NC15	UNL3742	Omaha (Meandering)	1.82 ± 0.08	52	3.89 ± 0.17	2.13 ± 0.15
BNC53	UNL2818	Omaha (Meandering)	1.63 ± 0.08	51	3.22 ± 0.12	1.98 ± 0.13
TK44	UNL2824	Omaha (Braided)	1.80 ± 0.09	62	3.53 ± 0.06	1.95 ± 0.10
SD1	UNL3739	Omaha (Meandering)	1.94 ± 0.09	55	3.79 ± 0.15	1.95 ± 0.12
LS34	UNL2821	Omaha (Meandering)	1.47 ± 0.07	55	2.78 ± 0.10	1.89 ± 0.11
LE30	UNL1063	Omaha (Braided)	1.87 ± 0.11	26	3.53 ± 0.12	1.88 ± 0.14
AL39	UNL2499	Omaha (Meandering)	2.75 ± 0.10	27	4.90 ± 0.24	1.78 ± 0.12
LE16	UNL1066	Omaha (Transitional)	1.67 ± 0.08	20	2.95 ± 0.21	1.77 ± 0.16
NC15A	UNL3743	Omaha (Meandering)	1.82 ± 0.09	56	2.96 ± 0.16	1.63 ± 0.12
OSW5	UNL2494	Omaha (Meandering)	1.70 ± 0.09	35	2.65 ± 0.14	1.56 ± 0.12
EP25	UNL2119	Omaha (Meandering)	2.59 ± 0.09	27	3.95 ± 0.18	1.53 ± 0.11
LS30	UNL2825	Omaha (Braided)	2.01 ± 0.10	61	2.88 ± 0.08	1.44 ± 0.08
SD11	UNL3740	Omaha (Braided)	2.23 ± 0.12	58	3.16 ± 0.13	1.42 ± 0.10
LE6	UNL1067	Omaha (Transitional)	1.82 ± 0,10	23	2.31 ± 0.07	1.27 ± 0.09
BNC54	UNL2822	Omaha (Braided)	2.21 ± 0.08	50	2.62 ± 0.04	1.19 ± 0.05
AL12	UNL2490	Omaha (Braided)	2.82 ± 0.12	34	2.84 ± 0.25	1.01 ± 0.10
MB15	UNL857	Omaha (Braided)	2.28 ± 0.09	27	2.16 ± 0.07	0.95 ± 0.06
PO6	UNL2121	Omaha (Braided)	2.37 ± 0.11	53	2.25 ± 0.12	0.95 ± 0.08
TK20	UNL2823	Omaha (Braided)	1.81 ± 0.07	50	1.69 ± 0.10	0.93 ± 0.07
SL50	UNL2830	Omaha (Braided)	2.55 ± 0.10	54	2.13 ± 0.10	0.83 ± 0.05
TK25	UNL2824	Omaha (Braided)	1.71 ± 0.07	61	1.35 ± 0.07	0.79 ± 0.05
TK2	UNL2832	Omaha (Braided)	2.12 ± 0.08	59	1.63 ± 0.10	0.77 ± 0.05
BU30	UNL2122	Omaha (Braided)	2.25 ± 0.08	42	1.66 ± 0.10	0.74 ± 0.06
SL49	UNL2834	Omaha (Braided)	2.43 ± 0.09	58	1.66 ± 0.11	0.68 ± 0.05
OSW58	UNL2497	Omaha (Braided)	2.66 ± 0.11	37	1.80 ± 0.15	0.68 ± 0.6
SSC138	UNL2498	Omaha (Braided)	2.26 ± 0.08	37	1.52 ± 0.19	0.67 ± 0.09
TKN44	UNL2820	Omaha (Braided)	2.15 ± 0.08	63	1.39 ± 0.08	0.65 ± 0.04
OSW12	UNL2495	Omaha (Braided)	3.26 ± 0.11	43	1.91 ± 0.15	0.59 ± 0.05
SSC137	UNL2500	Omaha (Braided)	2.73 ± 0.09	47	1.46 ± 0.12	0.46 ± 0.06
SLX57	UNL2827	Omaha (Braided)	1.91 ± 0.07	66	0.82 ± 0.04	0.43 ± 0.03
*OB62	UNL985	Omaha (Braided)	1.72 ± 0.09		2.47 ± 0.11	1.44 ± 0.11
*OB64	UNL986	Omaha (Braided)	2.06 ± 0.09		2.67 ± 0.11	1.30 ± 0.09
*OB66	UNL987	Omaha (Braided)	2.23 ± 0.09		2.59 ± 0.10	1.13 ± 0.08
*OB82	UNL988	Omaha (Braided)	2.06 ± 0.08		3.02 ± 0.11	1.47 ± 0.09
†MB25	AMS ^{14}C	Malta Bend	Peat/seed		Calibrated	23.58 ± 0.08
†MB25	AMS ^{14}C	Malta Bend	Peat/seed		Calibrated	24.458 ± 0.09

Notes: Error ranges are expressed as 1σ standard deviations. Locations for dates are in Figure 5.
*Holbrook et al. (2006a).
†Nzewunwah (2003).

TABLE 2. OPTICALLY STIMULATED LUMINESCENCE DATA FOR SAMPLES COLLECTED THROUGHOUT THE STUDY AREA

Sample number	Burial depth (m)	H_2O* (%)	K_2O (%)	U (ppm)	Th (ppm)	Cosmic (Gy)
GN/JBA-1	7.3	29.9	1.48	1.30	3.40	0.09
HE/JBA-2B	3.5	33.0	2.27	3.24	10.22	0.14
VE/MH-57	7.2	19.0	1.52	1.2	3.99	0.09
EN-17	7.3	27.0	2.03	2.5	7.80	0.09
CW-41	6.1	17.56	1.78	0.9	3.7	0.10
CW-42	4.9	19.09	1.76	1.1	3.7	0.11
CW/JBA-3	9.0	38.2	1.97	3.66	7.03	0.08
EP-31	6.4	28.0	1.90	1.9	5.98	0.10
SLX-50	4.1	25.8	1.89	2.25	6.09	0.13
EN-14	6.0	24.5	1.89	1.9	6.15	0.10
VE/MA-49	8.3	26.3	2.17	1.9	5.62	0.08
NB48	2.4	13.8	1.95	1.5	6.1	0.16
SL-45	2.35	21.9	1.90	1.66	4.26	0.16
SL-47	1.58	21.5	2.14	1.62	4.67	0.18
SL-3	2.9	20.0	1.82	3.17	4.52	0.15
CE-64	4.2	12.65	2.05	1.3	5.1	0.12
CE-1	5.0	19.28	1.90	1.1	4.3	0.11
CE-60	2.7	31.27	2.28	2.0	8.1	0.15
ONA-33	3.28	34.2	2.15	3.37	7.57	0.14
HA21	2.4	4.6	1.64	1.0	3.9	0.15
CW-54	2.9	15.02	1.86	1.0	3.5	0.15
CO-3	1.6	28.4	2.54	2.14	6.93	0.18
CW-60	3.2	27.29	1.90	1.8	6.7	0.14
CE-59	1.7	12.72	1.70	1.3	6.4	0.17
BU-32	4.0	23.6	1.92	2.2	7.06	0.13
HA12	2.3	22.4	1.83	1.0	3.4	0.16
NB3	2.2	24.4	1.81	1.2	4.6	0.16
CW-61	1.9	12.98	2.24	1.5	5.7	0.17
JE-55	4.5	7.5	1.94	1.8	5.85	0.12
CE-26	3.9	20.58	2.00	1.6	6.9	0.13
NB49	2.4	26.8	2.26	2.0	8.5	0.16
NB46A	4.0	31.6	2.08	3.2	6.8	0.13
LE4	1.3	20.7	1.75	1.2	4.6	0.18
CW-62	3.0	21.83	1.66	1.6	7.1	0.15
CW13	3.2	26.6	2.00	2.0	8.7	0.14
MS-50	3.3	27.3	1.52	1.5	7.3	0.14
BNC-7	2.6	21.9	1.32	1.30	3.72	0.16
CW-63	1.8	14.14	1.89	1.0	3.6	0.17
MS-49	2.0	27.0	1.81	1.7	7.5	0.16
GP-23	2.0	6.54	2.14	1.3	4.3	0.16
OSW-59	2.1	27.9	1.76	2.36	7.57	0.17
CE-61	3.0	11.53	2.02	1.3	5.2	0.14
NC-15	1.7	20.0	1.89	1.19	3.67	0.18
BNC-53	4.7	22.8	1.67	1.23	3.89	0.12
TK-44	3.7	20.8	1.78	1.46	4.06	0.14
SD-1	2.7	19.5	1.92	1.37	4.97	0.16
LS-34	1.5	20.3	1.59	0.83	1.95	0.18
LE30	4.4	29.7	1.88	1.6	6.6	0.12
AL-39	2.7	8.8	2.21	3.10	4.44	0.15
LE16	2.9	20.1	1.77	0.8	3.6	0.16
NC-15A	2.2	23.3	1.90	1.31	4.19	0.17
OSW-5	1.25	24.9	1.74	1.41	3.46	0.19
EP-25	4.4	5.1	2.07	2.0	6.07	0.12
LS-30	2.9	23.9	2.08	1.67	4.38	0.15
SD-11	2.0	28.7	2.05	2.62	7.59	0.17
LE6	3.2	27.0	1.60	1.6	7.7	0.14
BNC-54	0.7	8.2	1.77	1.60	5.23	0.20
BL-19	0.3	26.3	2.22	3.08	7.39	0.22
AL-12	1.75	17.1	2.18	3.51	8.10	0.17
GP-15	3.3	5.48	1.92	1.3	5.3	0.14
PO-6	4.3	20.9	2.02	3.0	6.35	0.13
TK-20	1.7	10.4	1.59	1.20	3.64	0.18
SL-50	1.7	12.0	2.00	2.43	7.07	0.18
TK-25	2.1	13.0	1.52	1.07	4.21	0.17
TK-2	1.5	7.1	1.55	1.78	5.77	0.18
BU-30	2.8	6.6	1.90	1.6	4.69	0.15
SL-49	1.8	11.7	2.07	1.92	5.92	0.17
OSW-58	0.85	16.5	2.13	3.69	4.69	0.20
SSC-138	2.48	7.7	1.73	1.46	7.43	0.16
TKN-44	1.8	10.2	1.66	1.92	5.77	0.17
OSW-12	1.05	5.9	1.61	4.03	13.99	0.19
SSC-137	1.5	4.6	1.95	3.08	4.84	0.18
SLX-57	1.1	6.4	1.59	1.16	3.90	0.19
*OB62	5.0	20.1	1.70	1.1	4.7	0.11
*OB64	3.7	13.0	1.87	1.3	5.2	0.13
*OB66	3.5	7.3	1.99	1.2	4.7	0.14
*OB82	3.7	6.5	1.82	1.1	4.3	0.13

Note: Ages for samples are in Table 1 and locations are in Figure 5.
*Holbrook et al. (2006a).

and manganese concretions above fine channel-belt sand. There is a pedogenically mixed zone of ~50 cm in the uppermost part of the Malta Bend Alloformation, where fine silt-to-sand of the Peoria Loess mixes with clays from beneath. No distinct paleosol is developed here beyond this evidence for mixing. An OSL date of 26.4 ± 2.1 ka from the Herman, Nebraska, location comes from the sands of the Malta Bend Alloformation just beneath this mixing zone and the underlying clay. Older Roxanna (55 ka to 27 ka) (Leigh and Knox, 1993; Rodbell et al., 1997; Markewich et al., 1998) and Loveland (190 ka to 120 ka) (Forman and Pierson, 2002) loess deposits present in the surrounding hills are absent on the terrace.

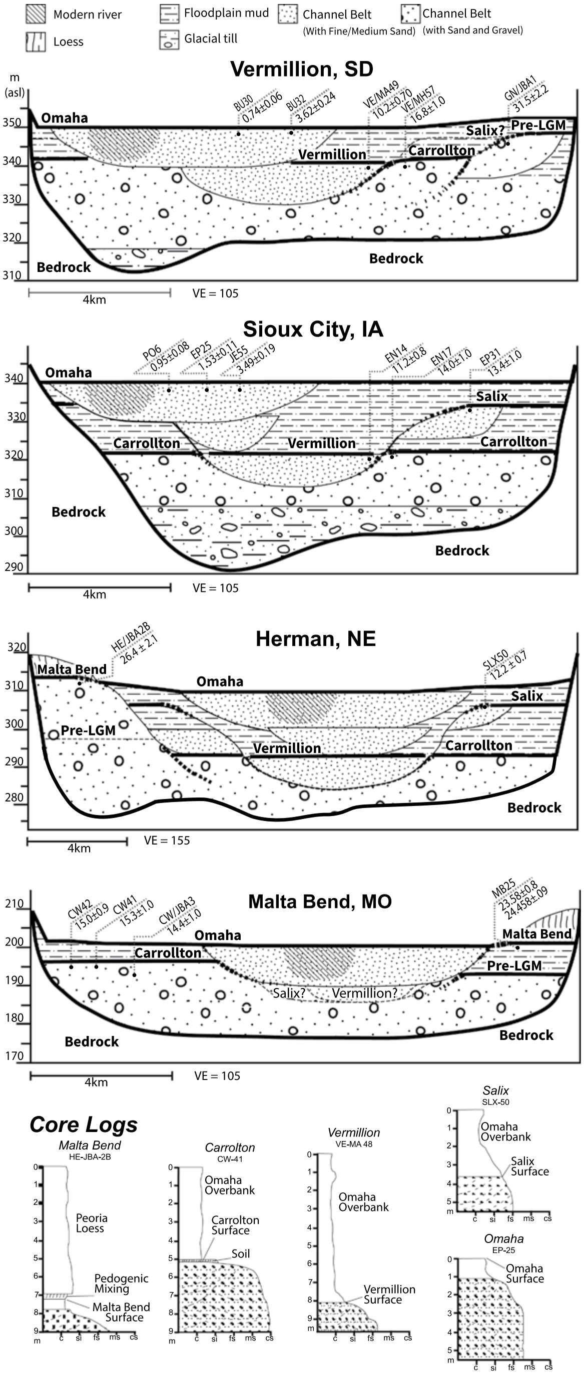


Figure 3. Representative cross sections and vertical sections of the lower Missouri River Valley show thicknesses of allounits and elevations of mapped surfaces. Each cross section is an amalgam of data representative of the thickness, elevations, and relative positions of surfaces in the valley reach near the location indicated for the cross section rather than a cross section beneath a discrete line. Locations of dates in cross sections are in Figure 5, and data for dates are in Tables 1 and 2.

The Malta Bend Alloformation in the Malta Bend area of Missouri is a thick floodplain deposit that formed lateral to channel belts, which are no longer found within the valley and likely were eroded by later incision events. In the Malta Bend quadrangle of Missouri, the top of the Malta Bend Alloformation is ~11 m below a high terrace bench covered in Peoria Loess. The upper surface of the Malta Bend Terrace is between 7 m and 15 m above the modern floodplain. Like in the Herman, Nebraska, location, the contact between the Malta Bend Alloformation and overlying Peoria Loess is approximately on grade with the modern floodplain, and the higher topography of this terrace is fully the thickness of the Peoria Loess alone (Fig. 3). The upper portion of the Malta Bend Alloformation was described by Nzewunwah (2003) as laminated organic clayey silt, bioturbated by roots, with pale brown mottles and less common yellow-brown mottles. In the MOV-MB-25 borehole (Figs. 3 and 5), organic material from two closely spaced thin peaty layers were dated within the Malta Bend Alloformation near the contact with the overlying Peoria Loess. The two accelerator mass spectrometry (AMS) C^{14} samples from this layer yield 23.58 ± 0.08 ka and 24.458 ± 0.09 ka calendar dates, respectively (Nzewunwah, 2003). This correlates closely to the OSL date from the Herman, Nebraska, area and dates to the LGM, when the Malta Bend channel system aggraded to elevations similar to the modern floodplain in both the northern and southern parts of the lower Missouri Valley.

Carrolton Alloformation

The Malta Bend Alloformation was incised and the Carrolton Alloformation was subsequently aggraded sometime between the early (ca. 23 ka) and post (ca. 16 ka) LGM. The Carrolton surface sits atop coarse sand and gravel of the Carrolton Alloformation and beneath fine floodbasin strata where it is not locally incised by channel-belt deposits of younger alloformations. The Carrolton surface ranges widely in depth along its longitudinal profile (Figs. 3 and 4) and is named for the town of Carrolton, Missouri. In the northernmost study area, just west of Vermillion, South Dakota (Figs. 3 and 4), the Carrolton surface is 7.0 m below the modern floodplain and formed at 16.8 ± 1.0 ka (Table 1). Farther downstream, just west of Sioux City, Iowa (Figs. 3 and 4), the Carrolton surface is 18.0 m below the modern floodplain and is 14.0 ± 1.0 ka. The Carrolton Alloformation comprises sandy channel-belt deposits at these two northern locations and directly underlies a thick unit of younger backswamp gleyed clay (Gley 2 4/10 bg) of the Omaha Alloformation. Carrolton channel-belt deposits here are well oxidized (5 yr 4/2) and range from medium to coarse sands capped by ~1 m of loam to silty loam. Farther south near Carrolton, Missouri (Figs. 3 and 4), this surface is much shallower and covered by only 4–5 m of younger backswamp muds. These Carrolton channel-belt deposits are capped by a thin (15 cm), poorly developed loamy paleosol with root traces. The paleosol consists of a mixture of gray silt and clay that grades downward into fine sands. The fine sands are heavily mottled with Mn nodules and Fe concretions. Channel deposits range from medium to coarse, well-oxidized sand with some fine gravels that are poorly sorted and well rounded. The surface near Carrolton has OSL ages at three adjacent locations on bar deposits (15.0 ± 1.0 ka and 15.1 ± 1.0 ka) and on sand below a 4 m channel fill (14.4 ± 1.0 ka). The Carrolton Alloformation formed by ca. 16 ka and lasted until ca. 14 ka. Water wells track the Carrolton Surface throughout the valley (Fig. 5)

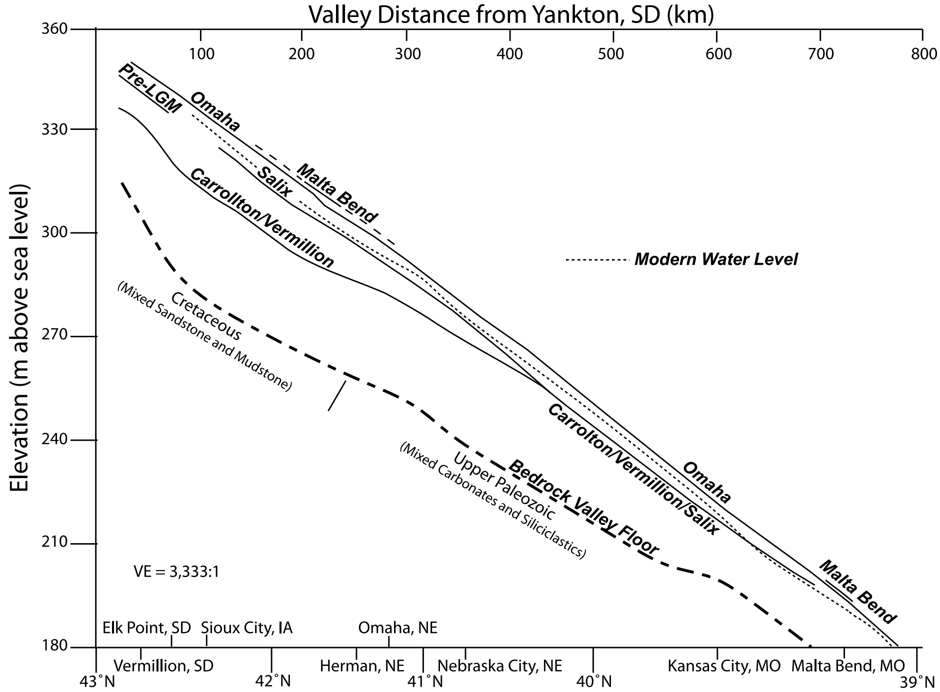


Figure 4. Longitudinal profiles show surfaces mapped within the Missouri Valley from Yankton, South Dakota, to Glasgow, Missouri. Profiles are based on surficial mapping and subsurface tracing using water wells. Locations of auger holes and water wells are in Figure 5. Bedrock extents are from King, Beikman, and Edmonston (1994). LGM—Last Glacial Maximum.

and show deepening around Sioux City, Iowa, that gradually shallows farther toward the south. Carrolton surface topography is visible through the thinner overlying mud sections near Yankton, South Dakota, and Carrolton, Missouri, and reveals a braided river morphology at both localities.

Salix Alloformation

The Salix Alloformation marks a period of aggradation in the latest Pleistocene and is buried by up to 6 m of floodplain fines on the northeastern and eastern sides of the valley (Fig. 5). The Salix surface imprints through overlying floodbasin fines to the modern surface and reveals a single-channel meandering morphology with definable point bars and wrapping abandoned channel fills. Point bars of Salix meander loops are dated in two locations. Near Elk Point, South Dakota, the point bar sand deposits are 6.4 m below the modern floodplain surface and are 13.4 ± 1.0 ka. Farther south near Salix, Iowa (Fig. 5), point bar sand deposits are 4 m below the modern floodplain and are 12.2 ± 0.7 ka. Point bar deposits at both locations are fine and well-oxidized sand with abundant manganese and iron stains. Colors range from 10 yr 4/3–10 yr 9/3 with a 10 yr 8/6 mottle. Water wells support tracing of this surface longitudinally to just south of Nebraska City, Nebraska. Here, the Salix surface grades into and becomes indistinguishable from the Carrolton and Vermillion surfaces, which both shallow by this distance to the approximate grade of the Salix surface (Figs. 3 and 4).

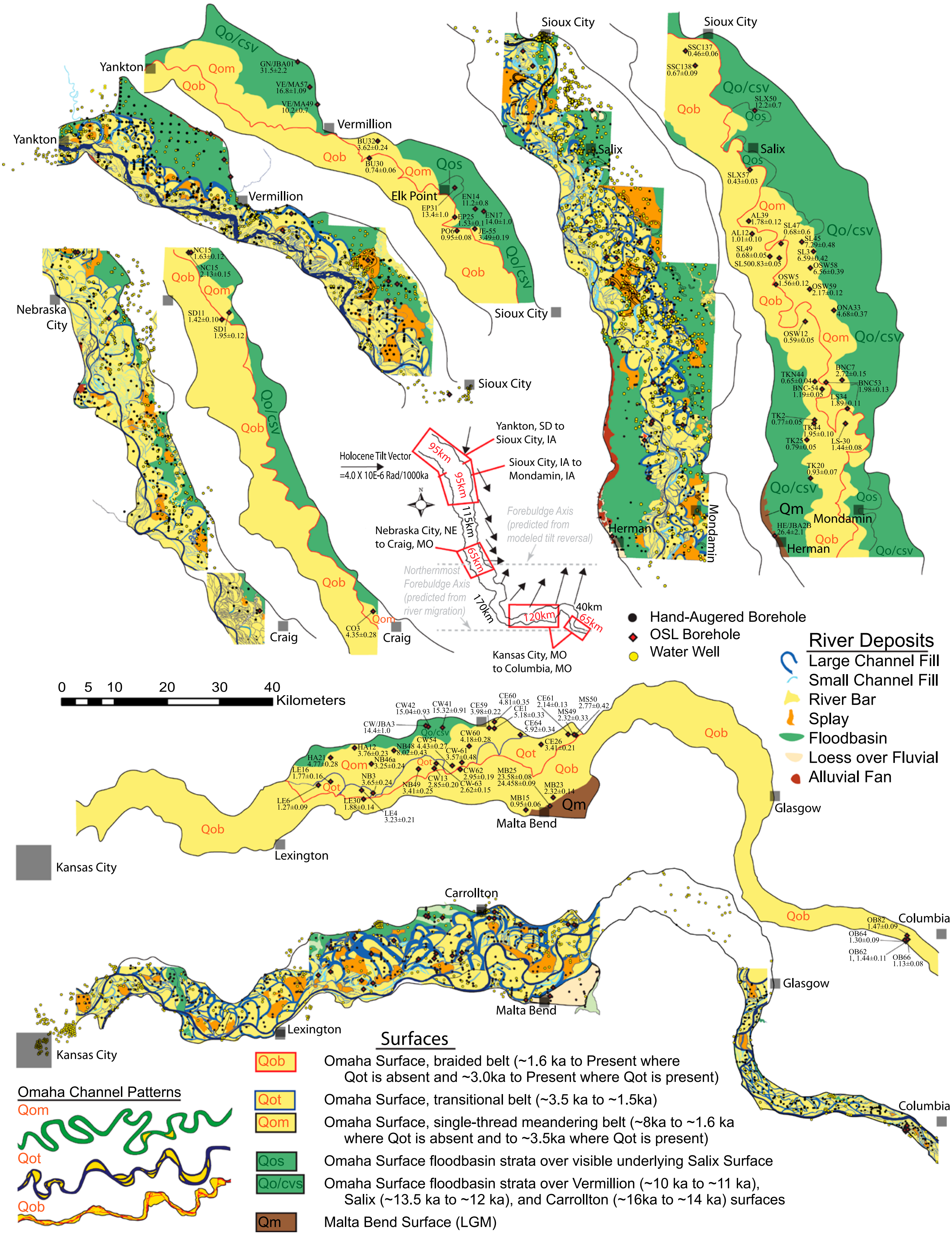
Vermillion Alloformation

The Vermillion Alloformation incises below the Salix surface during the Pleistocene/Holocene transition. The Vermillion Alloformation constitutes fine- to medium-grained channel-belt deposits and is buried by floodbasin fines and younger channel-belt deposits of the Omaha Alloformation. In the type area near Vermillion, South Dakota, this alloformation is incised into the Carrolton surface and is 8.3 m below the modern floodplain (Figs. 3 and 5). Here, the upper Vermillion deposits are fine to medium, well-oxidized channel-belt sands (5 yr 4/2) and are 10.2 ± 0.7 ka. Farther down dip near Elk Point, South Dakota, the Vermillion surface is significantly deeper at 18.3 m and again incised into the Carrolton surface. Sandy channel-belt deposits here are optically dated at 11.2 ± 0.8 ka (Figs. 3 and 5). The incision of the Salix surface to form the Vermillion surface begins at ca. 12 ka and approximates the Holocene/Pleistocene boundary. The duration of the Vermillion surface is more uncertain than the other surfaces, lasting any part of the 5000 years between the earliest possible abandonment of the Salix surface at ca. 12.9 ka to the initiation of the Omaha surface at ca. 8 ka. The error ranges of its current ages actually overlap, so the duration could be as short as decades. Given that the 5000 year gap over which the Vermillion surface forms has to incorporate both the deep incision from the Salix surface and then even larger aggradation to the Omaha surface, and that the Vermillion surface is confined to a narrow incision with comparatively little lateral reworking, it most likely was a short-lived surface. A duration between ca. 10 ka to ca. 11 ka seems reasonable and is consistent with the available ages. The Vermillion profile is within a couple of meters in elevation to the Carrolton surface and is indistinguishable in water wells where ages, detailed grain size, and sub-meter scale resolution are not available. The profiles of the Carrolton and Vermillion surfaces are thus mapped here as merged (Fig. 4).

Omaha Alloformation

The Omaha Alloformation includes the aggradational and surficial floodplain strata that postdate the abandonment and burial of the Vermillion surface. The Omaha Alloformation encompasses the Holocene, except potentially some part of its first 2000 years. Numerous ages throughout the lower valley of the oldest surficial channel-belt deposits show the river everywhere aggraded to its modern profile from the Vermillion profile by ca. 8 ka (Tables 1 and 2). Surficial maps (Figs. 3 and 5) show that the river appears to have laterally migrated extensively with only minor elevation shifts ($\leq\sim$2 m) since aggrading to the current level (Holbrook et al., 2006a; Holbrook and Allen, 2021).

The Missouri River has a migration trend over the last 8000 years of Omaha deposition that is consistently away from the position of the LGM ice front. In all reaches where the valley has an east/west component, the channel belts have consistently migrated from the northeast side of the valley toward the southwestern side (Fig. 5). In the reaches of Yankton, South Dakota, to south of Sioux City, Iowa; Nebraska City, Nebraska, to Craig, Missouri; and Kansas City, Missouri, to Columbia, Missouri, older Pleistocene deposits are preserved on the north/northeastern side of the valley, and younger Holocene channel-belt deposits are preserved toward the south/southwestern side of the valley. The ages of Holocene strata within the channel belt also progressively young toward the south, and the river currently



Yankton
Vermillion
Elk Point
Sioux City
Salix
Herman
Mondamin
Nebraska City
Craig
Kansas City
Lexington
Malta Bend
Carrollton
Glasgow
Columbia
Qo/csv
Qom
Qob
Qos
Qot
Qm
Holocene Tilt Vector
=4.0 X 10E-6 Rad/1000ka
Yankton, SD to Sioux City, IA
Sioux City, IA to Mondamin, IA
Nebraska City, NE to Craig, MO
Kansas City, MO to Columbia, MO
Forebuldge Axis (predicted from modeled tilt reversal)
Northernmost Forebuldge Axis (predicted from river migration)
95km
115km
65km
170km
120km
40km
0 5 10 20 30 40 Kilometers
Hand-Augered Borehole
OSL Borehole
Water Well
River Deposits
Large Channel Fill
Small Channel Fill
River Bar
Splay
Floodbasin
Loess over Fluvial
Alluvial Fan
Surfaces
Qob Omaha Surface, braided belt (~1.6 ka to Present where Qot is absent and ~3.0ka to Present where Qot is present)
Qot Omaha Surface, transitional belt (~3.5 ka to ~1.5ka)
Qom Omaha Surface, single-thread meandering belt (~8ka to ~1.6 ka where Qot is absent and to ~3.5ka where Qot is present)
Qos Omaha Surface floodbasin strata over visible underlying Salix Surface
Qo/cvs Omaha Surface floodbasin strata over Vermillion (~10 ka to ~11 ka), Salix (~13.5 ka to ~12 ka), and Carrollton (~16ka to ~14 ka) surfaces
Qm Malta Bend Surface (LGM)
Omaha Channel Patterns

Figure 5. Surficial geologic maps show the Missouri River Valley fill. Reaches represented by each map are indexed in the inset with the valley bedrock boundary, and the distances of valley mapped and left unmapped are both included in this index. Also included are the tilt rates of the valley and forebulge axis projected from the ICE-6G (VM5a) model of Peltier et al. (2015) and Argus et al. (2014) for the past 8000 years along the valley length. Maps are paired for each reach. In each case, the map on the left records a composite of individual quadrangle maps produced by students from the Big Muddy Expedition with locations of auger and water well data, and the map on the right is a simplified map of allounits with locations of dates. Further information about the dates is in Tables 1 and 2. Representative Omaha channel patterns are from Holbrook and Allen (2021).

resides along the most southern valley wall in each of these reaches (Fig. 5). Only in the one mapped valley reach perpendicular to the LGM ice front (the southern half of the Sioux City, Iowa, to Herman, Nebraska, segment) does this pattern break and the Holocene channel belt and modern river resides near the valley center (Fig. 5).

The river pattern also changes in coincidence with this lateral shift (Holbrook et al., 2006a; Holbrook and Allen, 2021). The Missouri River upstream from northern Missouri had a meandering morphology between 8 ka and 1.6 ka and then switched to a braided pattern around 1.6 ka (Fig. 5; Kashouh et al., 2011; Kashouh, 2012; Holbrook and Allen, 2021). The meander deposits of the braided river (see Holbrook and Allen, 2021) are consistently on the South or Southwestern side of the valley compared to the older, single-thread meandering river deposits (Fig. 5). The valley reach between Kansas City, Missouri, and Malta Bend, Missouri, shows a transition from meandering to braided pattern as well but is slightly amended from the pattern and timing observed farther upstream (Fig. 5). The river east of Kansas City, Missouri, was single-thread meandering from ca. 8 ka to 3.5 ka, but then formed a transitional meandering-to-braided pattern from ca. 3.5 ka to 1.5 ka before turning fully braided meandering thereafter (Holbrook et al., 2006a; Holbrook and Allen, 2021). The single-thread to transitional to fully braided deposits transition systematically southward, recording progressive southward lateral migration of the river over this pattern change (Fig. 5). The change from the transitionally braided to fully braided patterns is also gradual, starting at 3.0 ka, and is associated with $\sim$2 m of incision (Fig. 5). The shift from single-thread meandering to a braided meandering pattern is much more abrupt upstream of Kansas City, Missouri, and appears to have started and completed within two centuries. The river was artificially channelized back to a single-thread meandering form in all reaches south of Ponca State Park, Nebraska, over the mid-1900s but retains its approximate braided meandering form from there upstream to the southernmost dam at Yankton, South Dakota (Fig. 1; Jacobson et al., 2009).

DISCUSSION

Climate and the Filling of the Missouri River Valley

Missouri River Valley fill (Fig. 3) records at least two cycles of aggradation and incision since the LGM and reveals five surfaces traced longitudinally from Yankton, South Dakota, to Colombia, Missouri (Fig. 4). These surfaces reflect major swings in river gradients throughout the late Pleistocene and early Holocene and each mark the top of an alloformation by the same name. The Malta Bend surface records channel aggradation to elevations close to modern floodplain levels during the LGM. The LGM Malta Bend Alloformation is incised to form the Carrolton surface by around 16 ka. Between ca. 14 ka and 13.5 ka, the Missouri River profile aggraded close to modern floodplain levels (Salix surface) but again incised this alloformation at around 12 ka. Post Salix incision cut down to approximately the grade of the previous Carrollton surface by ca. 11 ka to initiate the Vermillion surface. The river aggraded back to modern floodplain levels between ca. 10 ka and ca. 8 ka, where the river has migrated laterally with only minor vertical adjustment over the past 8 ka (Omaha surface).

Cycles of incision and aggradation, and shifts in river pattern, correspond to major changes in climate trends. This section explores potential causal relationships between climate events since the LGM and these valley-fill trends.

LGM Malta Bend Surface

The Malta Bend Alloformation aggraded to current floodplain elevations in response to maximum advancement of glaciers into the Missouri River drainage (Figs. 3 and 6). Timing for initial aggradation of the Malta Bend Alloformation is presently undated and unknown. The date of ca. 26 ka from the top surface of the Malta Bend Alloformation, however, shows this unit had aggraded to its highest elevation, roughly equal to the modern Omaha floodplain surface, by the beginning of the maximum Wisconsinan glacial advancement. Higher sediment input to the valley due to advancement of glaciers into the drainage is likely the cause for the elevated Malta Bend profile. The glacial front entered the Missouri drainage proximal to the Missouri River during the LGM and even dammed the river locally (Colton et al., 1961; Hill and Feathers, 2002). The Missouri River received runoff from the main Laurentide ice sheet in the upper drainage (e.g., Kehew and Teller, 1994) and directly into the lower valley via the James and De Moines Lobes (e.g., Lundstrom et al., 2009). Ablation rates during the colder periods when the Malta Bend terrace formed were likely slow and therefore could have contributed to low water inputs relative to sediment. This higher sediment-to-water ratio could explain aggradation toward a relatively high profile (Fig. 4) like that of the Malta Bend surface (cf., Blum and Tornqvist, 2000; Holbrook et al., 2006c).

The high profile observed on the Malta Bend surface of the Missouri Valley is characteristic of the other major rivers draining the glacial front during the LGM as well and likely reflects a common cause. The highest LGM-to-recent terraces in the glacial-fed Ohio River (Terrace T4; Counts et al., 2015) and lower Mississippi River (Ash Hill and Sikeston Ridge terraces; Rittenour et al., 2007) valleys both formed during the LGM contemporaneously with the Malta Bend surface (Fig. 7). A high LGM terrace is not preserved in the central Mississippi bedrock valley but was once present (Hajic, 1991; Hajic et al., 2011; Carson et al., 2019), as is indicated by a flight of down-stepping terraces recording incision from a higher LGM surface that is preserved in tributary valleys as the St. Charles Terrace Group (Fig. 7). The Missouri, Mississippi, and Ohio Rivers are the three main interior rivers that drained the southern interior boundaries of the continental ice sheets during the LGM. The consistently high LGM terraces in these valleys likely reflect a common association of high sediment to water ratios delivered across the LGM glacial front. The pattern of the rivers forming the Malta Bend Alloformation is obscured beneath thick loess deposits. The equivalent T4 (Counts et al., 2015) and Ash Hill/Sikeston Ridge (Rittenour et al., 2007) surfaces, however, reveal braided "outwash" morphology. The Malta Bend channels were likely braided as well, but this is unconfirmed.

Carrollton Surface (ca. 16 ka to ca. 14 ka) and the Big Wash

Collapse of the Laurentide Ice Sheet after the LGM triggered incision of the Malta Bend Alloformation and the establishment of the much lower Carrollton outwash plain. The Missouri River incised up to 18 m locally between the

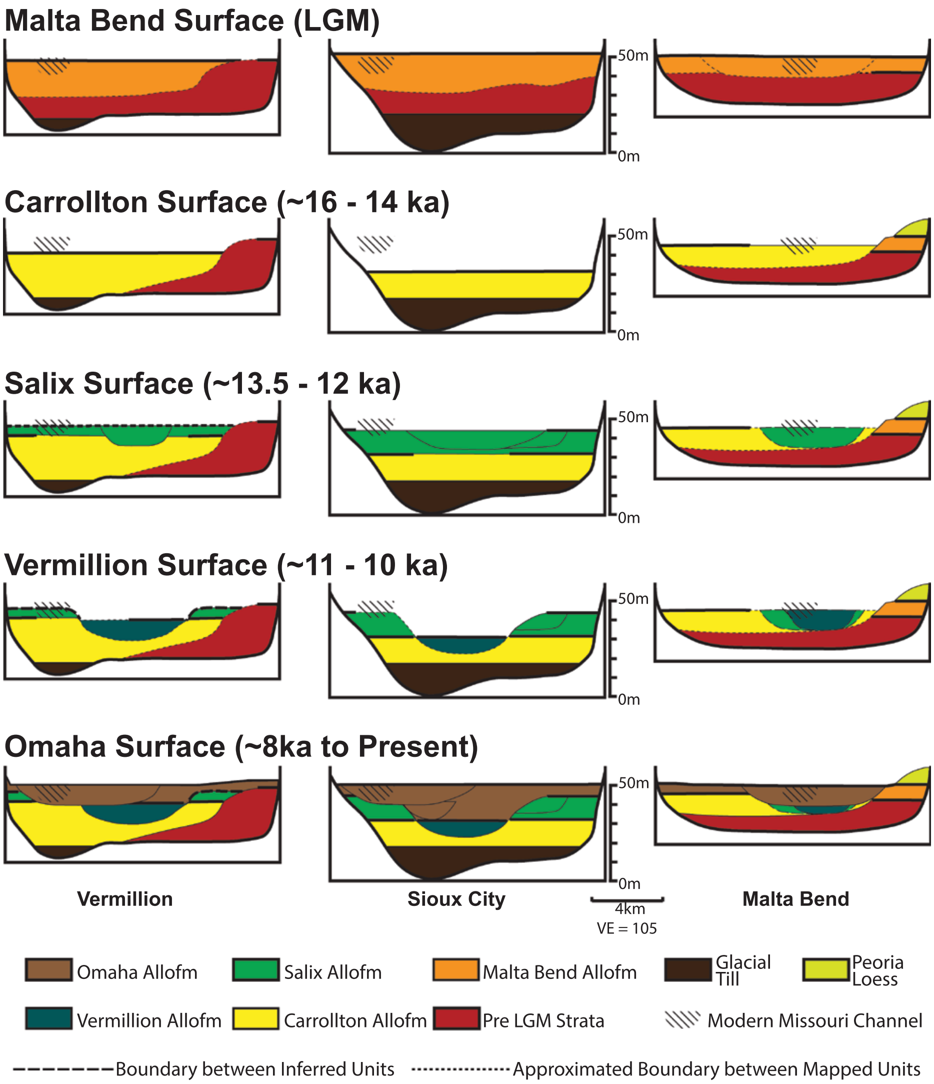


Figure 6. Record of incision and aggradation in the Missouri Valley since the Last Glacial Maximum (LGM) is shown for the Vermillion, South Dakota; Sioux City, Iowa; and Malta Bend, Missouri, Valley reaches. In each case, the valley evolution is derived from the cross sections in Figure 3.

youngest Malta Bend dates of ca. 23 ka (LGM) and the oldest Carrollton age of 16.8 ± 1.09 ka (Figs. 3 and 6). Rivers of the Carrolton surface continued to spread widely throughout the valley after this time leaving only small portions of the Malta Bend Alloformation preserved in a couple of inlets where the geometry of the valley was optimal for protection of these strata from erosion. Most of the younger parts of the Malta Bend Alloformation toward the center of the valley were likely eroded, and incision of the Carrollton surface probably occurred toward the later part of the age-dated interval.

Meltwater increase as the ice sheets collapsed (e.g., Joyce et al., 1993; Fig. 2) is likely the cause of the Carrollton incisional phase. Meltwater was routed down the Missouri Valley during this time via the ice saddle between the Laurentide and Cordilleran ice sheets (Catto et al., 1996; Ross et al., 2009). The James Lobe fully collapsed between 15.5 ka and 15 ka, readvanced between 15 ka and 14.5 ka, and disappeared again by 14 ka, each time providing large quantities of meltwater input directly into the Missouri River via the James River tributary at Yankton, South Dakota (Dyke, 2004; Lundstrom et al., 2009). This event is recorded globally by the ca. 19 ka 1A0 meltwater pulse (Clark et al., 2004) and the ca. 14.5 ka 1A meltwater pulse (Carlson, 2009; Deschamps et al., 2012) that raised sea level ~10 m and ~15 m, respectively, in ≤500 yr, as well as a potential short-lived pulse at 16.8 ka in between (Williams et al., 2010). Gregoire et al. (2012) particularly suggested that the collapse and further separation of the Laurentide and Cordilleran ice sheets during the opening of the ice-free corridor was a primary contributor to the 1A pulse. The Carrolton incisional phase and following surface and alloformation correspond with these three meltwater pulses, and the Missouri River appears to have routed the meltwaters that came from the western interior part of the Laurentide and Cordilleran ice sheets during this period. The sediment-poor meltwater resulting from glacial collapse (cf. Mississippi Valley; Knox, 1996; Bettis et al., 2008) is the likely reason for the deep late-to-post LGM incision (between ca. 23 ka and ca. 16 ka) and stabilization (until ca. 14 ka) of the Carrolton surface. This incision was likely accentuated because the James Lobe discharged proximal to the valley, so the sediment-poor meltwater did not have an opportunity to erode and entrain sediment before entering the Missouri Valley. Sediment storage in glacial lakes farther upstream in the Missouri drainage (Colton et al., 1961; Kehew and Teller, 1994; Hill and Feathers, 2002) may have also contributed to low sediment to water ratios. The Carrolton surface ghosts through overlying muds near Yankton, South Dakota, and Carrollton, Missouri, and records a river pattern consistent with a braided outwash plain.

The breakdown of the Laurentide and Cordilleran ice sheets at the end of the LGM caused a "big wash" that incised alluvium and evacuated sediment from outwash valleys across North America. This incision is recorded by a common drop in profiles from the higher LGM terraces to lower post-LGM glacial outwash terraces in the central Mississippi River Valley (Savanna: Flock, 1983; Hajic, 1991; Rittenour et al., 2007), the lower Ohio River Valley (Braid belt T3; Counts et al., 2015), and the lower Missouri River Valley (Carrollton surface) (Fig. 7). Sediments stored in the valleys of each of these major glacial rivers during the LGM were evacuated by the heightened discharge of meltwaters associated with glacial collapse, with the Missouri Valley showing the largest impact of the three (Fig. 7). This episode of incision also impacted the tributaries of these continent-scale rivers by causing incised terraces and draining of backwater lakes throughout their respective drainages (see Hajic, 1991; Rittenour et al., 2007; Counts et al., 2015). Other glacial outwash systems like the Rhine (E1; Lämmermann-Barthel et al., 2009) and Columbia (Jarrett and Malde, 1987; Godsey et al., 2011) also show similar incision of a prior and higher LGM terrace to form a new lower terrace profile at this time. Menot et al. (2006) in fact argue for a major period of river incision throughout northern Europe at the close of the LGM associated with collapse of the Fennoscandinan and British-Irish ice sheet. This "big wash" is not strictly a phenomenon of the Mississippi River and its tributaries but is

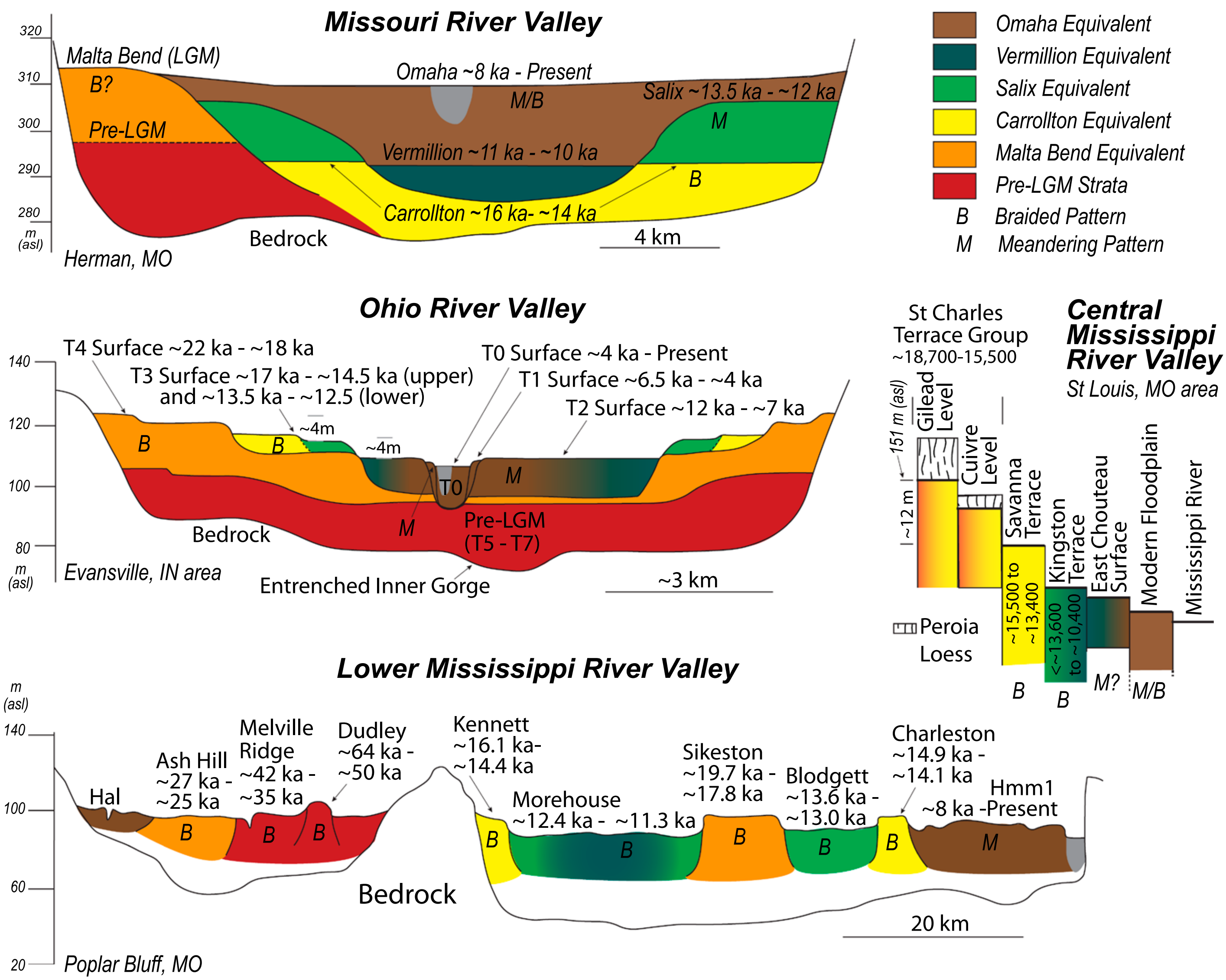


Figure 7. Representative cross sections compare the Missouri River, Ohio River, and Mississippi River Valley fills with respect to units of common age. The Missouri Valley cross section is the Herman, Nebraska, cross section from this study (Fig. 3). The Ohio Valley cross section is from Counts et al. (2015), the central Mississippi cross section is from Hajic (1991) and Bettis et al. (2008), and the lower Mississippi Valley cross section is from Rittenour et al. (2007). The relationship between terraces in the central and northern Mississippi Valley is not necessarily one-to-one (Knox, 1996; Bettis et al., 2008), is evolving, and not pursued in the current study.

instead the local expression of a larger global response to glacial collapse at the end of the LGM.

The sediment evacuated from glacial valleys of the Mississippi drainage during LGM glacial collapse should have caused a noticeable spike in sedimentation in the Gulf of Mexico. Based on average widths and depths of incision (Fig. 7), the Missouri River evacuated $\sim 8\times10^{10}$ m^3 of sediment over the reaches from Yankton, South Dakota, to Malta Bend, Missouri, and the Ohio evacuated $\sim 2.5\times10^{10}$ m^3 down dip of Louisville, Kentucky, USA, during the post-LGM Carrollton incision. This approximation does not include additional contributions from incised tributaries or from valley incision outside these reaches. The contribution from the central Mississippi Valley is unknown but likely of comparable magnitude because of comparable incision (Fig. 7). The Carrollton-equivalent surfaces in the lower Mississippi Valley are less than ~2 m lower than the LGM surfaces (Fig. 7). The lower Mississippi Valley thus did not store this sediment pulse and was likely even a minor contributor. This constitutes an increase in delivery to the Gulf of Mexico of at least 1×10^{11} m^3 of sediment over the big wash from the evacuation of the lower trunk valleys of the Ohio and Missouri Rivers alone. For perspective, the Holocene Mississippi discharge is estimated at $\sim 4\times10^8$ to 5×10^8 T/yr (Meade et al., 1990; Kesel et al., 1992; Blum and Roberts, 2009). Assuming the material removed from valleys was mostly equivalent to porous sand (density of ~2.0 T/m^3), this Holocene sediment discharge would equate to $\sim 2.25\times10^8$ m^3/yr of sediment removed from valleys, or 2.25×10^{11} m^3/ka. The total increase in sediment delivery to the Gulf of Mexico from the Carrollton episode of valley evacuation was thus at least on the order of millennial sediment delivery by the Mississippi River over the Holocene. This post-LGM increase also would likely have been in contrast to a comparable decrease in sediment delivery because of valley aggradation leading up to the LGM (also see Counts et al., 2015; Carson et al., 2019). Provided the incision lasted only a few millennia, it should have produced a substantial increase in sediment accumulation rate in the Gulf of Mexico. This increase should be particularly noticeable as the material evacuated from the valleys was likely more sand rich and should thus have impacted lithofacies distribution in addition to sediment volume rates. Recent examinations of the Gulf of Mexico source-to-sink system argue for the potential of such a post-LGM uptick in sediment delivery (Bentley et al., 2016; Fildani et al., 2018), but this increase is currently not quantified. This relationship between sediment storage in valleys and pulsed sediment delivery to the sink, however, constitutes an additional level of granularity to source-to-sink models that is also expressed elsewhere for late Pleistocene to Holocene river systems (e.g., the Ganges; Goodbred and Kuehl, 2000).

The Salix Surface (ca. 13.5 ka to ca. 12 ka) and the End of the Outwash Phase

The Salix surface aggraded to within 4 m of the modern Omaha floodplain surface between ca. 14 ka and ca. 13.5 ka (Figs. 3 and 6), and it records the end of the Missouri as a glacial outwash river. Salix aggradation took place during the Bølling-Allerød warm period, which coincided with the formation of Glacial Lake Agassiz, retreat of the Laurentide Ice Sheet beyond the Missouri drainage, and related rerouting of glacial meltwater at 13.9 ka into the Minnesota and Mississippi River drainage systems (Fig. 2). Rerouting of glacial outwash would have dramatically decreased water input into the lower Missouri River Valley. This loss of sediment-starved meltwater discharge is consistent with the aggradation of the Salix surface, as well as the change in river morphology from the braided outwash valley trains of the Carrollton surface to the single-thread meandering channels of the Salix surface. The abandonment of the braided Carrollton surface and establishment of the single-thread meandering Salix surface independently supports prior work that argued for diversion of glacial meltwater from the Missouri Valley at this time (see Clayton and Moran, 1982).

The Vermillion Surface (ca. 11 ka to ca. 10 ka) and Younger Dryas Incision

The Vermillion surface (ca. 11–10 ka) marks incision through the Salix surface and into the underlying Carrolton surface starting at ca. 12 ka (Figs. 4 and 7). Incision generated by this event was much narrower than the full valley and preserved fragments, laterally, of the Salix Alloformation (Figs. 3, 5, and 6). The Vermillion incision postdates and is unrelated to outwash events of the Laurentide Ice Sheet through the Missouri Valley. Climatic change in the Rocky Mountains and remaining upper Missouri River drainage likely accounts for the incision forming the Vermillion surface. The particulars of the climatic changes that drove this cycle are unconstrained at this time. The best hypothesis is that incision and establishment of the lower Vermillion surface records advance of valley glaciers and added snow pack in the Rocky Mountains that is associated with the coeval Younger Dryas. The cooling characterizing the Younger Dryas (12.9 ka to 11.7 ka) resulted in glacial advance after a period of Bølling-Allerød retreat in the mountainous parts of the Missouri River drainage in Montana and Wyoming (Licciardi and Pierce, 2018; Pierce et al., 2018; Dahms et al., 2018; Marcott et al., 2019; Palacios et al., 2020). Additional discharge from seasonal glacial ablation and snow melt during this cold phase are a potential explanation for the incision that resulted in the Vermillion surface.

The Omaha Surface (ca. 8 ka to Present) and Establishment of the Modern Floodplain

The second and final post-LGM aggradation phase of the Missouri River Valley (between ca. 10 ka and ca. 8 ka) ended with the establishment of the modern floodplain that constitutes the Omaha surface. The Missouri River migrated laterally and extensively over the past 8000 yr with only minor (~2 m) variation in elevation. The resulting meander-belt deposits and adjacent backswamp strata form the surficial parts of the Omaha Alloformation (Figs. 3 and 6). Highly amalgamated channel-belt sand deposits are typical of boreholes drilled through the Omaha Alloformation down to the Vermillion surface. This argues that aggradation between the Vermillion and Omaha surfaces was constrained within a narrow valley. This is consistent with extensive preservation of deposits from the earlier Carrolton and Salix surfaces along valley sides. The cause of the aggradation that generated the Omaha surface is uncertain but is likely tied to major climate shifts as the Younger Dryas ended and the warmer Holocene climate became fully established.

The Omaha surface incorporates a major shift from single-thread meandering to a braided meandering pattern during the late Holocene (Holbrook and Allen, 2021). Exposed channel deposits of the Omaha surface are single-thread meandering from their initiation at ca. 8 ka until ca. 1.6 ka when they become fully braided throughout the lower valley (Fig. 5). While tectonically driven slope increase can cause such pattern shifts (Holbrook and Schumm, 1999; Schumm et al., 2000), there is no evidence to support a slope-driven interpretation. This is because the shift is not local but is throughout the lower valley. The elevation differences between the meandering and braided surfaces are also minor, and isostatic rebound trends that might have increased slope in favor of braiding do not consistently predict such an increased slope along the longitudinal profile (Fig. 2 and next section). The most reasonable interpretation is that this pattern shift records a change in sediment and water discharge ratios related to a climatic change in the upper drainage that altered patterns throughout the lower valley. The cause of this shift is uncertain but does coincide with a major phase of increased glaciation during a Neoglacial in the hinterland drainage areas of the Rocky Mountains (reviewed in Solomina et al., 2015).

The transition to braided started earlier in the reaches east of Kansas City, Missouri (at ca. 3.5 ka), though the transition was not completed there until ca. 1.5 ka (Fig. 5; Holbrook et al., 2006a; Holbrook and Allen, 2021). Upstream of Kansas City, Missouri, this transitional phase is not present; thus, this earlier initiation appears to be related to changes caused by confluence of the Kansas River with the Missouri River at Kansas City, Missouri. The Kansas River records climate-related incision and erosion of the Newman terrace (up to 10 m; W.C. Johnson, 2020, personal commun.) by the Holiday terrace from 3.9 ka to 1.1 ka (Johnson et al., 2019). Introduction of additional bedload sediment related to this incision in the Kansas River Valley is the simplest explanation for the early start of braiding in the Missouri River east of Kansas City, Missouri.

Profiles of the Missouri River from LGM to Present

Longitudinal profiles of the Missouri River vary substantially over space and time (Fig. 4), diverging near Sioux City, Iowa, and converging just east of Malta Bend, Missouri. Impact of upstream climate variations and glacial input appears to have maximized near Sioux City, Iowa, where profiles show the most dispersion. Maximum profile divergence is ~100 km downstream of where the James Lobe discharged into the Missouri River via the James River at Yankton, South Dakota, and at approximately the same location where the Des Moines Lobe partly discharged into the Missouri River via the Big Sioux River near and upstream of Sioux City, Iowa (Fig. 4). Proximal input of glacial meltwaters to the Missouri River was, thus, likely the largest driver of this profile dispersion. Profiles merge progressively downstream as the impact of upstream climate controls dissipate with distance, closing substantially just downstream of Nebraska City, Nebraska, and finally merging after Malta Bend, Missouri (Fig. 4).

Merging of the surface profiles east of Malta Bend, Missouri, corresponds to shallowing and narrowing of the bedrock valley. The Missouri River Valley narrows significantly from a wide ($\geq$11 km) alluvial valley to a narrow (<5 km) bedrock valley abruptly downstream from Malta Bend toward Glasgow, Missouri. Narrowing of the valley here is also accompanied by significant shallowing of the valley to 10–20 m below the modern floodplain surface as compared to 30–40 m at areas upstream (Figs. 3–5). Significant shallowing of bedrock valley depth suggests the river was unable to scour into bedrock in this southern area. Bedrock could have anchored the river, causing a buttress effect on the profiles that locked the basal river scour to the bedrock elevation (cf. Mackin, 1948; Holbrook et al., 2006c). An intriguing but untested hypothesis is that this is where the early Pleistocene river valley once diverged, and the shallower and narrower valley beginning abruptly here and extending from Glasgow to St. Louis, Missouri, records a young course that was only recently avulsed and incised. Regardless of the cause, the variations in stream power versus sediment load originating upstream were too dissipated at Malta Bend, Missouri, to cause a variable impact on post-LGM bedrock incision.

Interestingly, the two post-LGM cycles of incision and aggradation in the Missouri River are not replicated by the lower Ohio and central Mississippi Rivers, which both show uninterrupted incision from the LGM to modern (Fig. 7). The Missouri River has more dynamic incision and aggradation cycles and appears to be more climatically sensitive than the Ohio and Mississippi Rivers. This is probably related to its comparatively shorter glacial outwash stage and its contrasting and more climatically sensitive grassland-to-mountain post-glacial catchment. The Ohio and Mississippi systems hosted outwash past the Carrollton equivalents and into Salix and Vermillion equivalent Kingston/Morehouse/T3 deposits (Fig. 7). The post-glacial catchments (see Knox, 1996) of the Ohio (dominantly woodland) and Mississippi (dominantly woodland with grassland) were also likely less climatically sensitive and are not typified by alpine conditions like the Missouri drainage. These contrasting drainage conditions probably caused large variations in sediment and water ratios for the Missouri River as compared to the Ohio and Mississippi Rivers.

Because of their temporal mismatch and the distance from the confluence, the rises and falls of the Mississippi trunk profile is excluded as a cause for variations in Missouri River profiles. The profiles of the Missouri River merge well before the Mississippi confluence and do not match the profile variations of the central Mississippi River (see above and Figs. 3, 4, and 7). Furthermore, the southern anchoring point of the Missouri River profiles near Glasgow, Missouri, is approximately six backwater lengths upstream of the Mississippi confluence and well beyond the normal influence of downstream effects caused by confluence incision (cf., Blum et al., 2013).

Cycles of cut and fill over the valley reach from Vermillion, South Dakota, to northern Missouri are of sufficient magnitude to account for up to half of the depth of valley scour in this reach. The modern Omaha and Malta Bend surfaces sit ~30 m below the bedrock surface in the surrounding hills and ~35 m above the bedrock valley floor over this reach. When the 18 m of incision and the ~10 m of channel depth are considered together, the interval of incision and aggradation over most of this reach is up to 28 m. Valley erosion from climatically driven cycles of incision during just the Wisconsinan phase could thus account for nearly half of the 65 m of current bedrock valley depth over this reach. Around 40% of the incision generating the Missouri bedrock valley could thus be accounted for by glacially dominated climate cycles. The remaining valley depth is more likely attributable to regional incision of the North American interior by long-term down cutting driven by tectonic denudation. This climate-driven versus denudation-driven component of valley incision records the buffer vs. denudation components, respectively, of valley incision (cf. Holbrook et al., 2006c; Holbrook and Bhattacharya, 2012) for the Missouri River.

The Yankton to Sioux City Reach as an Ice-Carved Valley

The Missouri Valley from Yankton, South Dakota, to Sioux City, Iowa, is a relict glacial valley. The basal fill of the Missouri Valley is glacial till over this reach (Fig. 3). The valley from Yankton, South Dakota, to Sioux City, Iowa, also has a generally cylindrical form with relatively straight sides and truncated spurs (Fig. 8). This part of the Missouri Valley likely records glacial carving by a glacial lobe extending from the James Valley tributary near Yankton, South Dakota, the same valley that hosted the Wisconsinan James Lobe. Since these till deposits are overlain by pre-LGM and later channel deposits, and there is no other supporting evidence for a Wisconsinan advance as far as Sioux City, Iowa, these till deposits and the glacial carving of the valley record a pre-Wisconsinan glacial advance of presently unknown age.

Role of Isostatic Adjustment on Valley Filling

Cycles of aggradation and incision are better explained by climatic drivers than coeval glacioisostatic adjustments (GIA), but GIA did likely drive Holocene channel migration patterns and partially impacted incision. Isostatic adjustment from glaciotectonic forebulge migration and glacial rebound could theoretically account for slope variations that drove incision and aggradation as the crust was locally uplifted and subsided beneath the river profiles. The locations and magnitude of glaciotectonic uplift and subsidence changed through time and are only measured directly in recent times (see Sella et al., 2007). Gaps in the knowledge of earlier glaciotectonic shifts are instead infilled using models. The ICE-6G (VM5a) model for isostatic adjustment (Argus et al., 2014; Peltier et al., 2015) is used here to estimate representative changes in glaciotectonic topography through time as a result of glacial loading and unloading. Proposed uplift and subsidence rates are contoured over the length of the Missouri drainage by subtracting projected local elevations from models between time intervals of interest (Fig. 2). Modeling of topography and associated uplift rates suggest that fluvial records of incision and aggradation are inconsistent with local timing, direction, and magnitude of modeled GIA. Though tectonic uplift and subsidence likely influenced profiles, and do provide an explanation for the migration of Holocene channels consistently away from the glacial front, the climatic interpretations presented previously are the preferred explanation for cycles of aggradation and incision.

Impact of GIA on River Profiles

Patterns of predicted glaciotectonic uplift and subsidence across the Missouri Valley do not coincide with patterns of incision and aggradation within the valley during the late Pleistocene and Holocene transition. During the LGM, models suggest that subsidence rates in the northern section of the lower valley were on the order of 1–2 m/ka, while uplift rates at the southern area of this study at this time were 1–2 m/ka (Argus et al., 2014; Peltier et al., 2015). This time period is when the Malta Bend surface formed, which is approximately at modern stream levels in both areas. This profile does not show the differential incision and aggradation expected from a strictly tectonic cause and appears to follow the elevation trends of the current river, which is currently experiencing regional subsidence (see Sella et al., 2007). High uplift rates at ca. 16 ka during the time of the oldest dates for the Carrollton surface could account for a few meters of accentuated incision and help explain locally deep (18 m) incision near Sioux City, Iowa. This interpretation, however, is problematic as the model also predicts uplift rates here on the order of 15 m/ka for the following 2000 yr, when the Carrollton surface was still active. The Carrollton surface

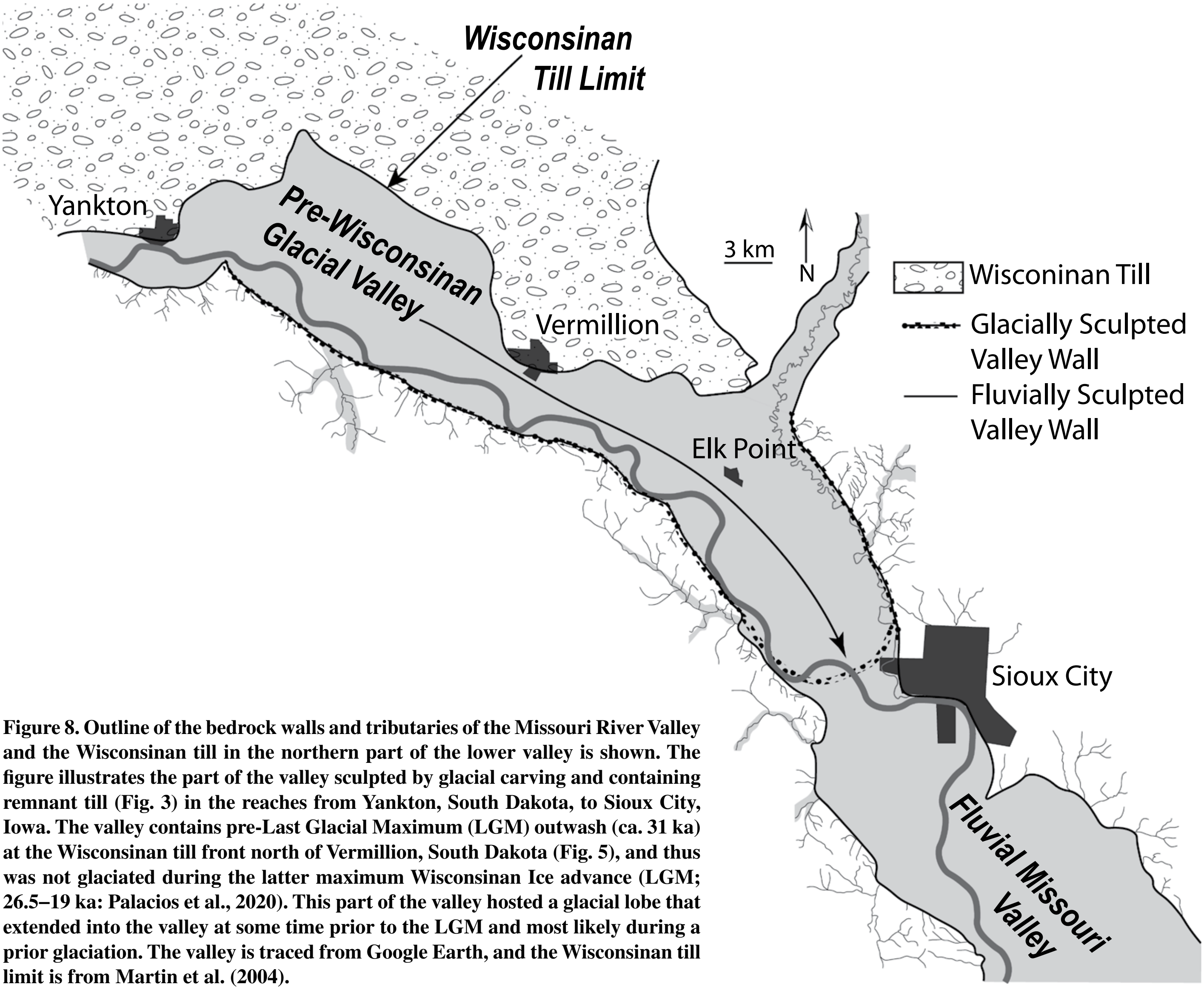


Figure 8. Outline of the bedrock walls and tributaries of the Missouri River Valley and the Wisconsinan till in the northern part of the lower valley is shown. The figure illustrates the part of the valley sculpted by glacial carving and containing remnant till (Fig. 3) in the reaches from Yankton, South Dakota, to Sioux City, Iowa. The valley contains pre-Last Glacial Maximum (LGM) outwash (ca. 31 ka) at the Wisconsinan till front north of Vermillion, South Dakota (Fig. 5), and thus was not glaciated during the latter maximum Wisconsinan Ice advance (LGM; 26.5–19 ka: Palacios et al., 2020). This part of the valley hosted a glacial lobe that extended into the valley at some time prior to the LGM and most likely during a prior glaciation. The valley is traced from Google Earth, and the Wisconsinan till limit is from Martin et al. (2004).

does not vary laterally by more than a couple of meters in elevation anywhere in the valley, and there is no location where the Carrollton surface shows the ~30 m of down-stepping incision from ca. 16 to ca. 14 ka that is expected if GIA were the driving cause of the low Carrolton profile. Between 14 ka to 13.5 ka, the model suggests that the study area experienced uplift rates on the order of 0–5 m/ka, which should have favored incision. However, this is also the time period in which the Missouri Valley aggraded up to 14 m above the Carrolton surface to initiate the higher Salix surface.

Available estimates for variations in water discharge versus glacioisostatic adjustment are consistent with the above observations that argue for preferential climate control on stream profiles. Since modeled GIA rates are nonuniform in space and time, and on scales up to 15 m/ka, normal expectations are that at least some of the higher local rates should have produced differential displacements between terraces that followed predicted GIA trends (see Holbrook and Schumm, 1999; Schumm et al., 2000). This, however, is not the case. The lack of correspondence between predicted GIA and terrace displacements suggests that the river either was not responding to GIA slope modification or the magnitudes of uplift and subsidence at times were not what the GIA model predicts.

A cursory examination of slope controls shows that even the apparently high GIA shifts predicted by the model would likely have played a subordinate role in profile trends of the Missouri River when compared to climatic change. Slope of the modern Missouri River from Sioux City, Iowa, to Kansas City, Missouri, in the reach most perpendicular to GIA ranges from 0.00019 to 0.00015 (Carlston, 1969). Slope changes relative to the current profile predicted by modeled GIA range between 4%/ka and 27%/ka from the LGM to the beginning of the Holocene (Table 3; Fig. 2). Estimated discharges from LGM to present specifically for the Missouri River are unavailable. Wickert (2016), however, determined that the discharge to the Gulf of Mexico by the collective Mississippi drainage climbed and then fell at least ~50%, and up to ~300%, between the LGM and the end of ice sheets in the Missouri drainage. Discharge then fell another ~50% in the following transition from the latest Missouri outwash phase into the Holocene. Incision and aggradation of river profiles is controlled both by stream power and sediment supply (Blum and Tornqvist, 2000; Holbrook et al., 2006c). Changes in sediment supply from the Missouri drainage are unconstrained for this

TABLE 3. SLOPE CHANGE RATES PROJECTED FROM GLACIAL ISOSTATIC ADJUSTMENT (GIA) IN REACH MOST PERPENDICULAR TO ICE FRONT BETWEEN SIOUX CITY, IOWA, AND KANSAS CITY, MISSOURI, EXPRESSED AS ABSOLUTE CHANGE AND AS A PERCENTAGE OF MODERN RIVER SLOPE

Age (ka)	Rate of slope change*	Slope change† (%/ka)
22.5	-9.2×10^{-6}/ka	5.6
18.5	-6.9×10^{-6}/ka	4
16	$+1.8\times10^{-5}$/ka	10.6
15	$+3\times10^{-5}$/ka	17.6
14.5	$+4.6\times10^{-5}$/ka	27
13.5§	$+3.3\times10^{-5}$/ka	19.4
13.5#	-2.1×10^{-5}/ka	12.3

*Based on ICE-6G (VM5a) GIA Model (Peltier et al., 2015; Argus et al., 2014) (after Fig. 2).
†Relative to the slope of the modern Missouri River channel averaged to 1.7×10^{-4} for the target reach after Carlston (1969).
§For the subreach from Sioux City, Iowa, to Omaha, Nebraska.
#For the subreach from Omaha, Nebraska, to Kansas City, Missouri.

time interval. These discharge and slope estimates, however, do give insight into stream power. Stream power ($\Omega = \gamma QS$) is the linear product of specific weight of water (γ), discharge (Q), and slope (S), and defines the energy available to the river to do work, such as incision. Assuming the variations in total Mississippi discharge are also representative of the Missouri component of discharge, the relative variation in stream power because of discharge ranged between twice and two orders of magnitude more than the predicted slope impact because of GIA. This helps explain why profiles do not follow the GIA signal. GIA probably impacted incision and aggradation by slightly muting or accentuating its magnitude, but this tectonic signal was apparently overwhelmed by the larger climate signal.

Similarly, changes in channel pattern during Missouri Valley filling are better interpreted as climate driven and are unlikely to be caused principally by GIA. A change in pattern from single-thread meandering to braided tends to require increased stream power because of increased slope or discharge and/or an introduction of coarser bed load, and the converse is also true (see Holbrook and Allen, 2021). Changes between single-thread meandering and braided patterns in the Missouri Valley do not follow these trends. The switch from braided to meandering from the Carrolton to the Salix surface came with an increase in slope, as measured by both the comparatively higher Salix profile (Fig. 4) and contemporary slope changes predicted by GIA (Table 3). Slope changes taken in isolation should have favored a meandering Carrollton pattern and a braided Salix pattern, yet the opposite is observed. The change from a single-thread meandering to a braided meandering pattern during formation of the Omaha surface came with no significant change in slope. The switch from the braided Carrolton to the single-thread meandering Salix surface does include a decrease in gran size, but this change is better explained by the coincident shift from a glacial to a precipitation-fed system instead of a GIA source. Changes in channel pattern are better explained by changes in discharge trends because of climate, which are coupled in at least the Salix case with a climate-driven change in sediment load. The impact of GIA on both pattern and profile is thus minimal compared to that of climate. GIA was not of sufficient magnitude to leave an imprint on either architecture or lithofacies of Missouri Valley fill that was detectable by current methods (see Holbrook and Miall, 2020).

Wickert et al. (2019) note a deep (up to 64 m) and narrow inner gorge incised below the normal bedrock floor of the Mississippi Valley that they interpret as bedrock incision driven by forebulge migration in the early Pleistocene. No such inner gorge or related deep profile is apparent in the bedrock of the reaches examined in the Missouri River Valley (Figs. 3 and 5) nor is reported in the literature. This suggests the forebulge incision accumulated in the Mississippi Valley early in the Pleistocene did not happen in the Missouri Valley or is small and confined to unexamined valley reaches. Incision of this deep gorge in the Mississippi River Valley is interpreted to reflect knickpoint migration and head cutting because the river crossed over the forebulge, increasing channel slope downstream of the crest (Wickert et al., 2019). Our data thus do not show either post-LGM impact of a similar forebulge crossing on river profiles or a Mississippi-style legacy of valley incision in general from earlier forebulges in the bedrock profile.

Position of the Forebulge Axis from River Response

Tilting of the North American continent because of glacial loading and forebulge uplift impacted trends in Holocene river migration and likely controlled the location of the Missouri Valley as well. Holocene tilting caused the river to migrate consistently away from the glacial front. The position of the lower Missouri Valley also appears to reflect a long-term trend of river pinning between the relatively higher topography of a forebulge crest to the south and an ice front to the north.

The Missouri River has migrated consistently away from the LGM glacial front within its valley over the Holocene (Fig. 5), a trend which argues for a glaciotectonic imprint on channel migration. The Holocene Omaha channel belt is consistently on the valley side away from the glacial front in all reaches where the lower Missouri Valley and the LGM glacial front are parallel to oblique (Fig. 5). In the one valley reach perpendicular to the glacial front (the lower half of Sioux City, Iowa, to Herman, Nebraska, reach) the belt is in the middle of the valley and shows no preferred position. This preferential belt positioning indicates that the Omaha channel belt nowhere established in, or later approached, the northern/glacial side of its valley. Glaciotectonics thus impacted the position of the Missouri River well before establishment of the Omaha channel belt at ca. 8 ka and therefore within or before early Holocene time. Glaciotectonics continued to impact lateral migration of the river thereafter. The Missouri River has migrated consistently away from the glacial front within the Omaha channel belt in all parallel-to-oblique reaches of the lower Missouri Valley. From Yankton, South Dakota, to Sioux City, Iowa (Fig. 5), the river has migrated from the northeast to the southwest side of the valley progressively over the past 8000 yr and is currently hugging the valley wall on the southwestern side. Additionally, in the southern area where the valley trend is nearly west to east, between Kansas City and Malta Bend, Missouri, the river has similarly migrated north to south over the past 8000 yr. These two reaches are roughly parallel and south of the glacial front. In the upper half of the Sioux City, Iowa, to Herman, Nebraska, reach, and the Nebraska City, Nebraska, to Craig, Missouri, reach, where the valley is oriented southeast and oblique to the glacial front, the river again migrated southwest and away from the glacial front within the Omaha belt (Fig. 5). In the lower half of the Sioux City, Iowa, to Herman, Nebraska, reach, where the valley is perpendicular to the glacial front, the Missouri River currently occupies a more medial position within the Omaha belt.

The consistent migration trend of the Holocene Missouri River away from the position of the LGM glacial front is best explained by lateral valley tilting during glacial rebound and forebulge relaxation. Lateral tilting of valleys commonly affects migration of rivers over time toward the down-tilted side (Bridge and Leeder, 1979; Nanson, 1980; Leeder and Alexander, 1987; Peakall, 1995; Holbrook and Schumm, 1999; Schumm et al., 2000). Peakall (1995) showed that valley tilting between the order of 7.5×10^{-4} rad/ka yr and 4.6×10^{-7} rad/ka yr causes lateral channel migration, and higher rates tend to cause channel avulsion. Valleys on the glacial side of the forebulge should rotate and tilt toward the forebulge during glacial rebound, causing rivers to migrate away from the glacial front and toward the forebulge axis in valleys parallel or oblique to these trends (Fig. 9). The Holocene pattern of migration away from the LGM glacial front is extremely consistent in the

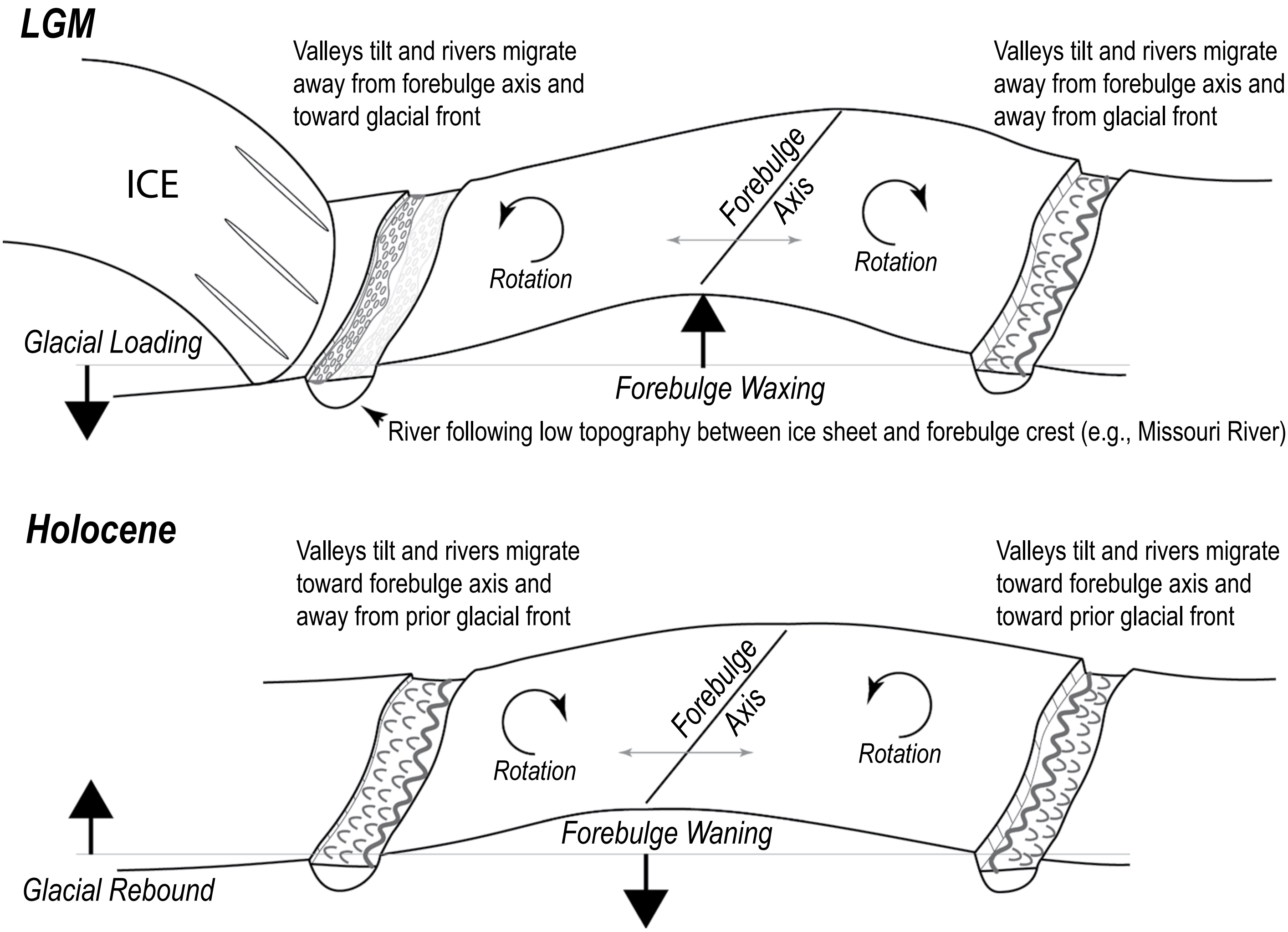


Figure 9. Diagrams show predicted patterns of tilting and river response to glacial loading and forebulge growth during the Last Glacial Maximum vs. glacial rebound and forebulge relaxation later during the Holocene.

lower valley, and magnitudes of GIA tilting are consistent with lateral migration (Fig. 5). These migration trends are unlikely to be coincidental and unrelated and are best explained by GIA-related tilting.

The pattern of channel migration away from the glacial front is only partly consistent with the GIA model, however, and argues for a forebulge axis at least 200 km south of where the model predicts (Fig. 5). Topographic differences from the ICE-6G (VM5a) GIA model indicate that valley tilting over the last 8000 yr of the Holocene was between 1.7×10^{-6} and 3.8×10^{-6} rad/ka along the Missouri Valley (Fig. 5), which is well within the range that tends to prompt channel migration. The directions of modeled valley tilt, however, are mostly but not completely consistent with observed river migration trends (Figs. 5 and 9). In the Yankton, South Dakota, reach to just south of Sioux City, Iowa, the valley trends NW-SE, and river migration occurred from NE-SW. The modeled direction of valley tilt is N-S and of sufficient trend and magnitude to explain observed lateral migration. North of Herman, Nebraska, the direction of valley tilting is parallel to the valley orientation and should not have impacted channel lateral migration. Again, the lack of preferred channel migration in this reach is consistent with the model (Fig. 5). From Nebraska City, Nebraska, to Columbia, Missouri, the modeled valley tilt direction is toward the north, which is opposite from the direction of channel migration over the Holocene. It is possible the tilting here is localized, and the ICE-6G (VM5a) model is too large to capture this event. The area of divergence from the model, however, is large. The lateral migration could also be working against valley tilting, but this would go against known mechanisms of channel migration (reviewed in Holbrook and Schumm, 1999; Schumm et al., 2000). The simplest solution is that the forebulge axis was actually farther south throughout the Holocene. The pattern of channel migration is more consistent with a forebulge axis centered south of Kansas City, Missouri, during the Holocene, and thus tilt orientation changed at least 200 km farther south than the ICE-6G (VM5a) model predicts (Figs. 2 and 5).

This more southern placement of the forebulge axis places the lower Missouri River Valley within the low trough between the topographically higher ice front to the north and forebulge crest to the south. Since the forebulge should migrate south as the ice front approaches and north as the ice front retreats (Fig. 2), positioning of the forebulge crest south of the valley in the Holocene should apply throughout late Wisconsinan glaciation and likely for earlier glacial episodes as well. Long-term placement of the valley north of the forebulge crest is consistent with the lack of a deep inner gorge in Missouri Valley bedrock. Wickert et al. (2019) argued that the deep inner gorge in the Mississippi Valley formed because of knickpoint incision as the river traversed the forebulge crest. If the Missouri River never crossed the forebulge crest, it had no need to develop a similar inner gorge. Such a long-term front-to-forebulge pinning would explain the current path of the lower Missouri Valley. Geodetics does not pin the location of the modern relict forebulge axis, but the region from the Great Lakes to the Gulf of Mexico is currently undergoing GIA-related subsidence (Sella et al., 2007). This is consistent with a wide settling forebulge and offers broad possibilities for placing a relict and dissipating Holocene axis over much of the mid-drift of the conterminous United States. Since Holocene valleys parallel and south of the relaxing forebulge axis should tilt north, and those to the north should tilt south (Fig. 9), similar mapping of Holocene channel migration trends like that conducted here provides an avenue for eventually isolating the forebulge axis and potentially constraining its migration trends.

CONCLUSIONS

(1) The Missouri River rapidly incised and aggraded up to 18 m in two climate-driven cycles over ca. 15,000 yr or less that spanned from the LGM to ca. 8 ka. Stabilization after each incision and aggradation event promoted lateral migration of the river and construction of a fluvial surface above a corresponding alloformation with the same name. The LGM Malta Bend surface stabilized by ca. 26 ka. The Malta Bend surface was then incised up to 18 m, and a new and lower Carrollton braidplain surface initiated between ca. 23 ka and ca. 16 ka. Subsequent aggradation between ca. 14 ka and ca. 13.5 ka resulted in stabilization of the meandering Salix surface at elevations ~4 m lower than both the modern floodplain and the prior Malta Bend surface. Renewed incision at ca. 12 ka, and the Holocene boundary, back to Carrolton surface elevations initiated the Vermillion surface. Renewed aggradation in the early Holocene starting at ca. 10 ka settled at levels approximating the prior LGM Malta Bend surface by ca. 8 ka. Lateral migration and overbank flooding of the Missouri River since ca. 8 ka formed the Omaha surface, which persists to today as the modern river floodplain. During formation of the Omaha surface, the Missouri River was single-thread meandering from ca. 8 ka until it switched to the recent braided meandering pattern at ca. 1.6 ka. The change from single-thread to braided meandering was transitional east of Kansas City, Missouri, spanning between ca. 3.5 ka and ca. 1.5 ka, but was abrupt elsewhere.

(2) Longitudinal profiles of surfaces are anchored in the lower Missouri Valley near Glasgow, Missouri, where each profile approaches a common elevation. The largest separation of profiles (up to 18 m) is ~100 km downstream of Yankton, South Dakota, near and south of Sioux City, Iowa. This records a point of coalescence and maximizing of the impacts of climatic change on profiles from the glacial inputs near Sioux City that dissipated downstream toward Glasgow.

(3) The Missouri Valley from Yankton, South Dakota, to Sioux City, Iowa, contains glacial till at its base and has a U-shaped morphology. This part of the Missouri Valley records carving by a glacial lobe that followed the valley as far as Sioux City, Iowa, that predated Wisconsinan glaciation.

(4) The building of a high LGM terrace, followed by deep incision to a much lower post-LGM terrace in the millennia prior to ca. 16 ka, is observed in the Missouri, Ohio, and Mississippi Valleys. This records a "big wash" of valley sediments associated with breakdown of the cryosphere at the end of the LGM and meltwater pulses that scoured and evacuated sediment from valleys in each of the major rivers draining the southern LGM glacial front. The magnitude of this increased sediment pulse to the Gulf of Mexico should have approached the size of millennial Holocene sediment input rates. This phenomenon is not unique to North America.

(5) River profiles reconstructed from Pleistocene terraces do not match the directions or magnitude of existing glacioisostatic adjustment models. Though forebulge migration and glacial rebound probably amplified or dampened incision and aggradation locally, this tectonic impact on river profiles appears to be on the scale of a few meters or less and was overwhelmed by the climate signal in the Missouri Valley.

(6) Migration of Missouri River channels is consistently away from the LGM glacial front at all locations, which argues for valley tilting consistently south and away from the front throughout the Holocene. This provides evidence that the Holocene Missouri Valley resided fully in the rebound area of the Laurentide ice sheet. The Holocene river thus appears pinned between the LGM glacial front and the forebulge topographic crest along its length, a condition that was probably typical of the valley throughout its development.

(7) Channel migration trends agree with calculated magnitudes for tilting from forebulge modeling throughout the valley but partly disagree in direction. River migration patterns argue for a forebulge axis centered south of Kansas City, Missouri, during the Holocene and thus at least 200 km south of where the model predicts. Possible reasons for the more northern axial position in the model are underestimates in ice thickness, inaccuracies in assumed mantle strength, and/or local tectonic anomalies the model does not currently capture.

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