History of the Missouri River Valley from the Late Pleistocene to Present: Climatic vs. Tectonic Forcing on Valley Architecture

by

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First off, I'd like to thank my advisor Dr. John Holbrook whose assistance made this work possible. I would also like to thank my thesis committee, Dr. Xiangyang (Cheyenne) Xie and Dr. Mike Slattery, for taking time to guide me throughout this process. Special thanks to all of the previous Mud Rangers who participated in the Missouri River Education Mapping program and helped create the database used in this study. Lastly, I would like to thank Andy Wickert whose assistance proved very useful in this study.

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Chapter I: Introduction

The Missouri River, the longest river in North America (Fig. 1) is generally regarded as a glacial-front system that periodically carried outwash sediments (Bluemle 1972; Ruhe 1983). Record of development for the Quaternary river within its valley are very limited (Holbrook, et al., 2006; Guccione, 1983). Most inferences regarding the Pleistocene aged Missouri River are anchored on adjacent loess deposits (Forman and Pierson, 2002; Frye, et al., 1948; Ruhe, 1983) which allows for very little indication on how the Missouri River Valley was constructed. The Missouri River drainage pathways were ever changing throughout the Pleistocene (Flint 1947; Bluemle 1972; Kehew and Teller 1994; Catto et al., 1996) and wide swings in glacial patterns created highly variable sediment and water discharge rates resulting in complex valley architectures. In addition, glacial isostatic adjustment from glacial loading and unloading may have been large enough to cause tectonic forcing of river profiles by uplift or subsidence which may have also helped to create the Missouri River Valley architecture. Lack of detailed mapping and dating within the valley, however, makes it difficult to discern what factors contributed most heavily to the overall construction of one of the largest river valleys in North America.

This study summarizes what is known to date of the late Pleistocene history of the Missouri River and will provide the first evidence of aggradation and incisional histories within the valley. Detailed mapping and optically stimulated luminescence dating techiniques used in this study allowed for the reconstruction of the fluvial history of the Missouri River Valley from the late Pleistocene to the present based on valley-fill strata. In addition, this study will explore relationships between glacial cycles and responses of the Missouri River to tectonic and climatic drivers.

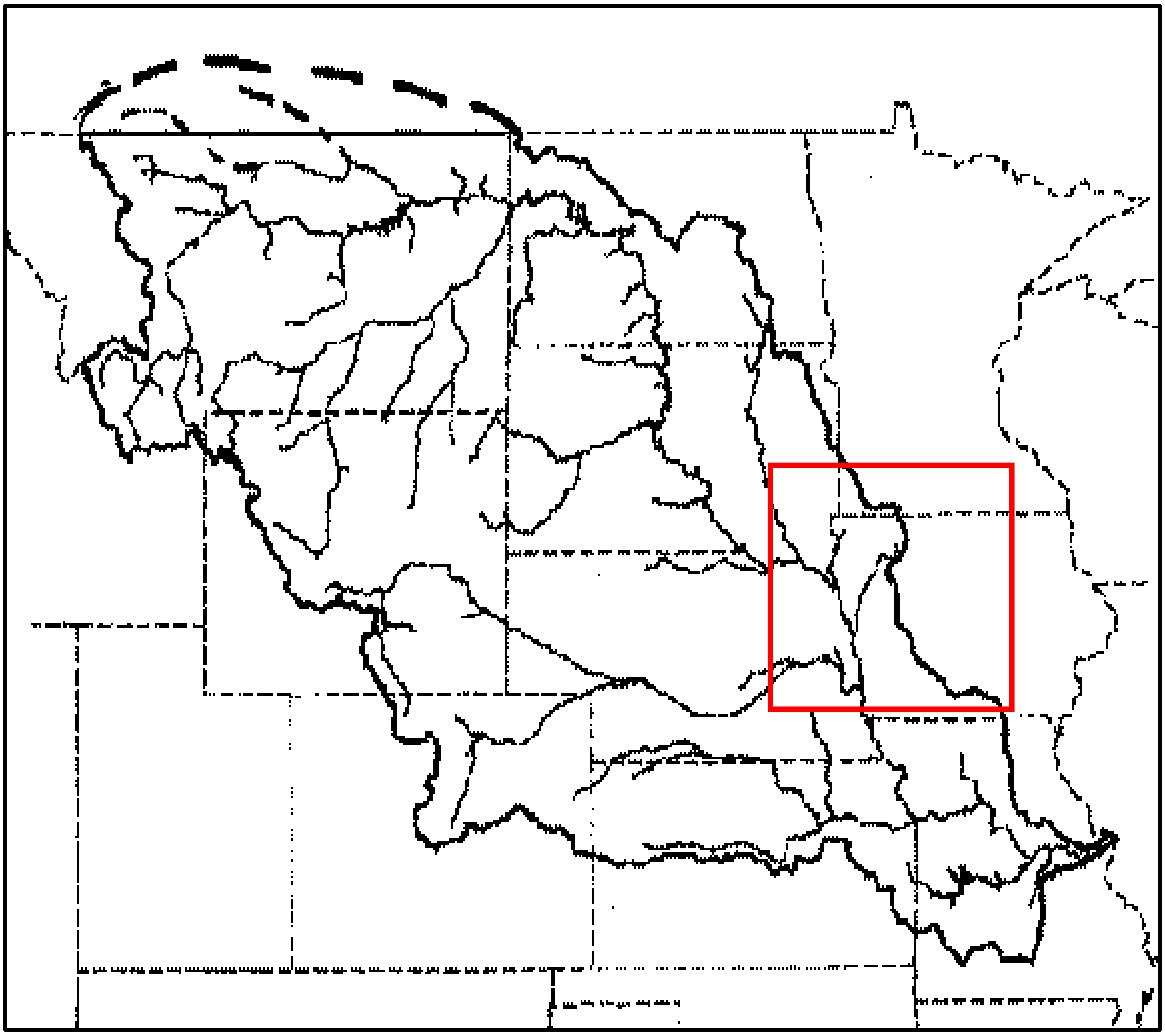


Figure 1 - Drainage are of the Missouri River Valley. Study area is outlined with red box. Image sourced from https://www.eol.ucar.edu/projects/gcip/dm/documents/map_99_lsa_nw/section2.html.

I.1 Valley Architecture Construction

This study uses dates and elevations of buried channel surfaces within the Missouri River valley to sequence fluvial incision/aggradation cycles, and to test valley responses against glacial events. River systems are known to respond to changes in ratios of water and sediment supply by incision and aggradation. When water supply dominates sediment supply, rivers will incise (Blum and Tornqvist, 2000). Conversely, when sediment supply dominates water supply, rivers will aggrade. This process can cut and fill valleys. Holbrook et al. (2006) described valleys cut and filled by these processes as “buffer valleys”, and the long term vertical range of incision and aggradation define the ” buffer zone”. In such valleys, total depth of the valley is controlled by the highest amount of incision caused by upstream input cycles (Holbrook et al., 2006). Glacial systems are generally typified by wide swings in sediment and water input related to climate extremes during glacial cycles and are thus strong candidates for buffer valley development. Denudation valleys are caused by incision into tectonic uplifts, and can also be a product of glacial cycles. As an area uplifts, rivers will incise in order to stay at equilibrium, thus creating an incised valley relative to the adjacent uplifted areas (Holbrook and Bhattacharya, 2012). In this case, incision should be relative to the amount of uplift, and later subsidence should cause aggradation. It is possible that some amount of uplift occurred in the Missouri River valley as flexural loading of Pleistocene ice sheets would have caused a “forebulge” to uplift areas south of glacial fronts. Additionally, uplift should occur in subsided areas during unloading of ice sheets. It is possible that parts of the Missouri River valley were influenced by both or either buffer or denudation processes driven by glacial cycles.

In order to distinguish possible responses of the Missouri River Valley to glacial fed “buffer valley” or a tectonically driven “denudation valley” incisional modes, upstream input cycles and forebulge movement must be constrained and compared to valley incision and fill cycles. Such examination requires knowledge of glacial history within North America and more specifically changes in ice margins volumes through time. For this particular study examinations are narrowed to the time period since the Last Glacial Maximum (LGM), and further limited to the history of the Cordilleran and Laurentide Ice Sheets, which would have contributed the majority of melt water to the Missouri River valley.

I.2 Continental Glaciation and The Missouri River Valley

Existing evidence argues that the Missouri River was glacially shaped during the Pleistocene, periodically damned by glaciers, and frequently carried glacial outwash (Bluemle 1972; Ruhe 1983; Guccione 1983; Hill and Feathers 2002). The Missouri River developed the current general west-east drainage pattern sometime in the early Pleistocene (Pre-Illinoian), but its glacial drainage patterns were shifting until the latest Wisconsin stage (Bluemle 1972; Galloway et al., 2011; Kehew and Teller, 1994; Catto et al., 1996). Glaciation and subsequent drainage patterns since the Late Glacial Maximum (LGM 29-21 ka BP.) was cyclical, with multiple brief stadial and interstadial periods. In addition, multiple lobes and sublobes did not advance and retreat in phase with the global glacial patterns (stadial vs interstadial) reflected in eustatic sea-level curves constructed from O^{18} isotopes (Dyke et al., 2002a; Lundstrom et al., 2009; Lambeck et al., 2014).

Eustatic sea-level provides a generalized time line of global glacial history since the LGM (Fig 1.3). The LGM ranges from around 29-20.5 ka BP during which ice volumes were relatively constant or slowly advanced (Lambeck et al., 2014). Deglaciation occurred from 20.5 ka BP until 18 ka BP. Ice volumes were relatively constant from 18-16.5 ka BP. Deglaciation continued from 16.5-12.5 ka BP with a brief static event from 15-14.5 ka BP, and accelerated sea-level rise from about 14.5-14.0 ka BP associated with the Bølling−Allerød warm period (Lambeck et al., 2014; Dyke 2004). Sea-level ceased to rise from 12.5-11.5 ka BP during the Younger Dryas cool period (Dyke 2004; Lambeck et al., 2014). Sea-level rose from 11.5 ka BP through the Holocene, but slowed at about 8.2 ka BP and again at 6.7 ka BP consistent with the end of glaciations (Lambeck et al., 2014). Global ice volumes and sea-levels, while important to the overall trend of glaciations, do not directly control the drainage history of the Missouri River valley because advances and ablations can be localized and out of sync with global patterns. Additionally, drainage is not only dependent on ice margins but also paleo-topography as well as the presence of glacial lakes, all of which are considered in more detail below.

I.3 Pre-LGM

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 final rerouting during the Late Wisconsin. This is evidenced by maps of till and loess deposits. Pre-illionoian "Nebraskan" tills occur within the Missouri River Valley in multiple locations including northcentral Missouri, western Missouri, Iowa, Nebraska, and Kansas (Guccione 1983; Bayne et al., 1971; Hallberg, 1980a,b, cited in Guccione, 1983; Reed and Dreeszen, 1965 cited in Guccione, 1983). Flint (1947) and Warren (1952) dated diversions of the White River into the Missouri River drainage system around Chamberlain, South Dakota to be Pre-Wisconsin – either "Kansan" or "Illinoian". The presence of the Loveland Silt in multiple locations within the Missouri River Valley argues the river carried glacial outwash during Illinoian glaciations (Forman & Pierson, 2002; Forman et al., 1992). Evidence for rerouting of the Missouri River in North Dakota is reported by Bluemle (1972) as occurring in the early Wisconsin when ice advances blocked wide west-east and southwest-northeast trending valleys, and created southward trending ice margin shaped valleys. Before the Late Wisconsin advance to the LGM, during the Mid-Wisconsin interstadial (approximately 32-35 ka BP), the area which was drained by the Missouri River was ice-free (Hill, 2002). At this time, the ice margin approximated the Canadian Shield (Dyke et al., 2002). Recent glacial diversions of the Missouri River headwaters into the lower valley occurred throughout the Late Wisconsin (Bluemle 1972; Kehew and Teller 1994; Catto et al., 1996).

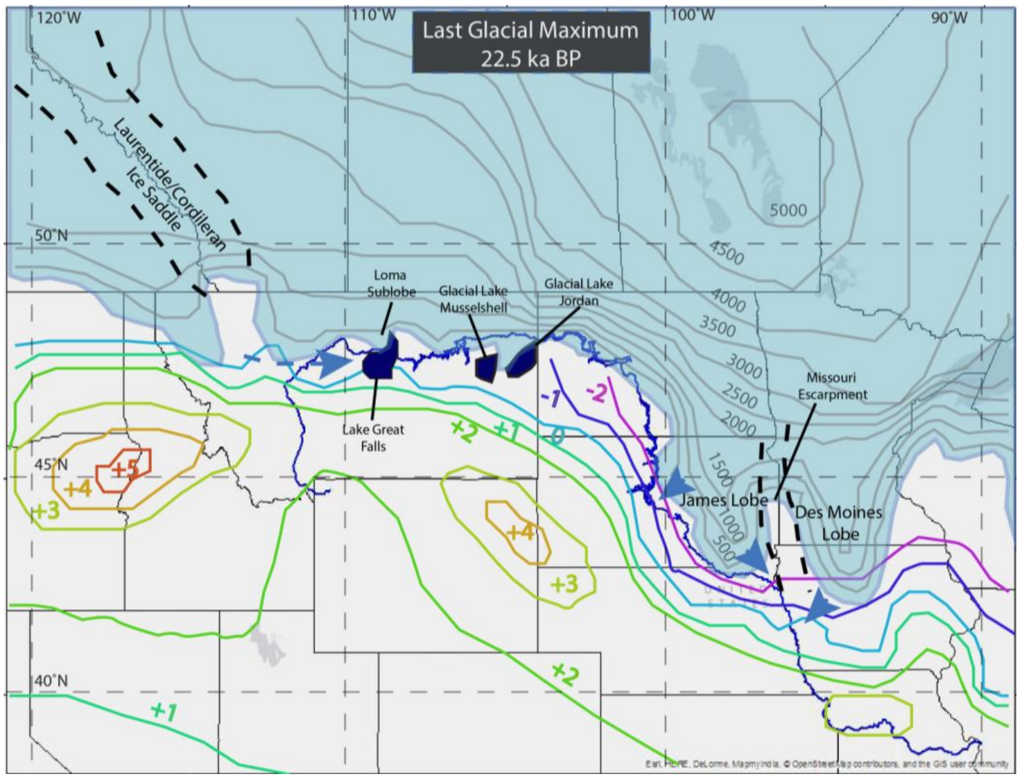


Figure 1.2A – 22.5 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

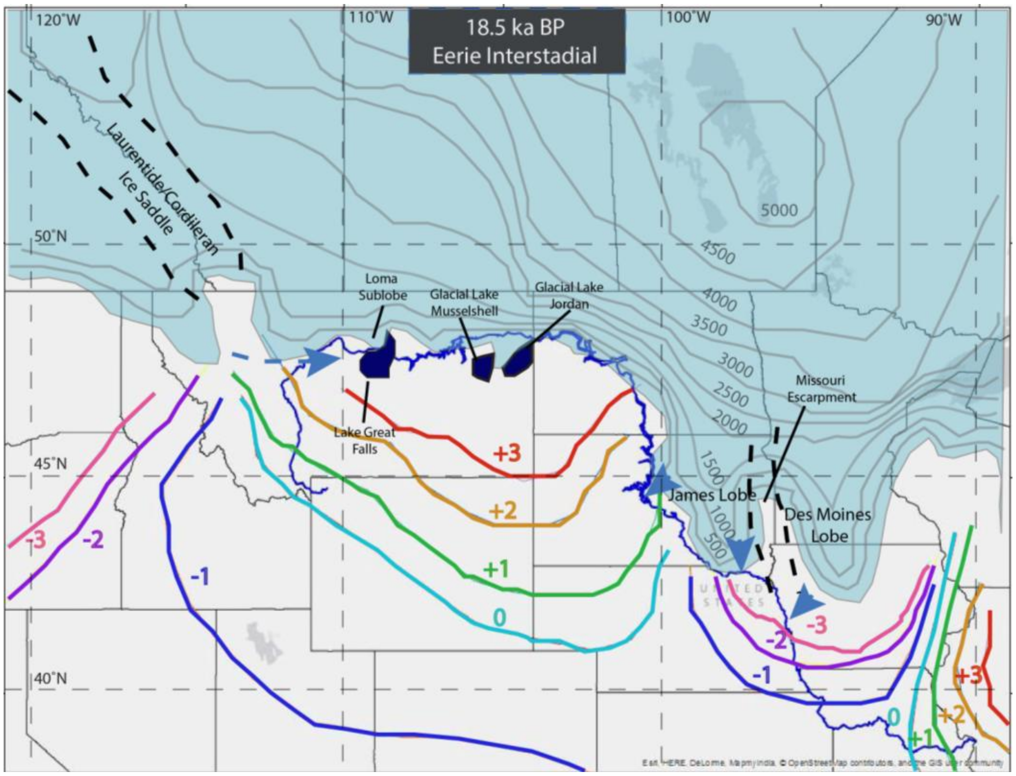


Figure 1.2B – 18.5 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

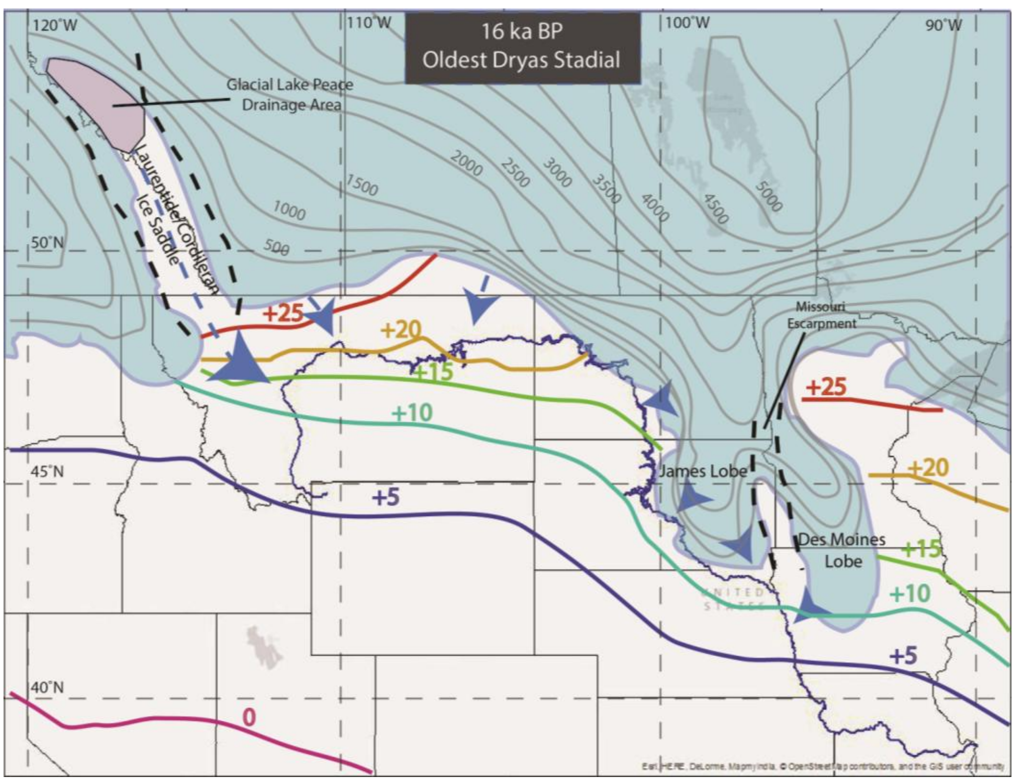


Figure 1.2C – 16 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

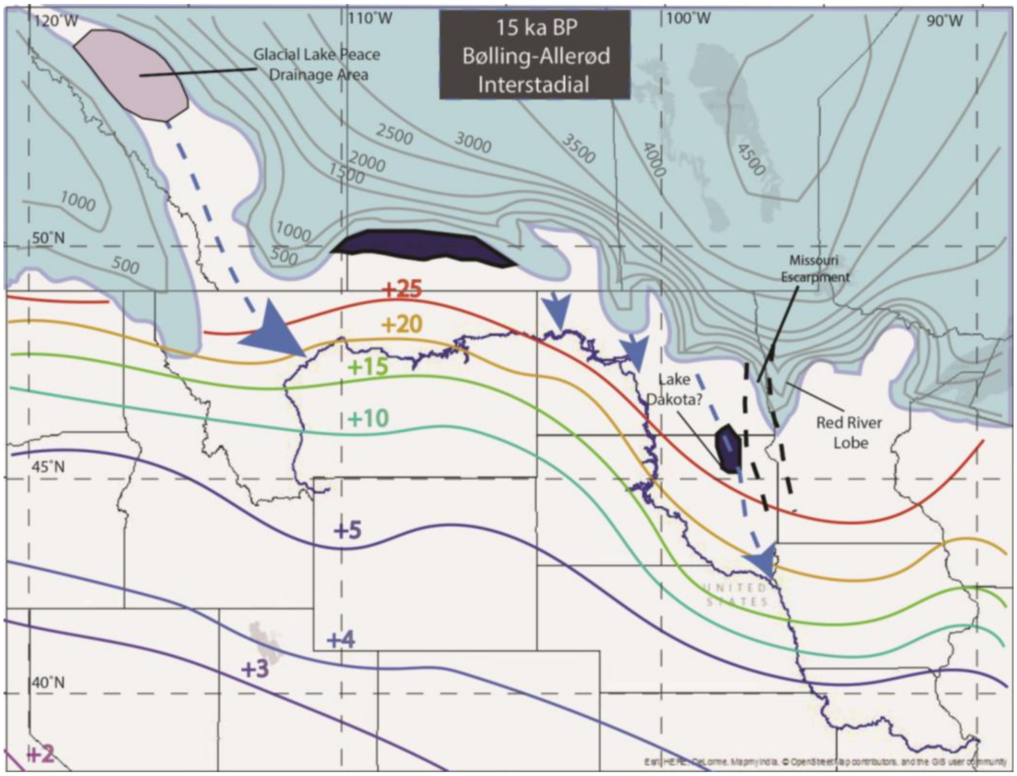


Figure 1.2D – 15 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

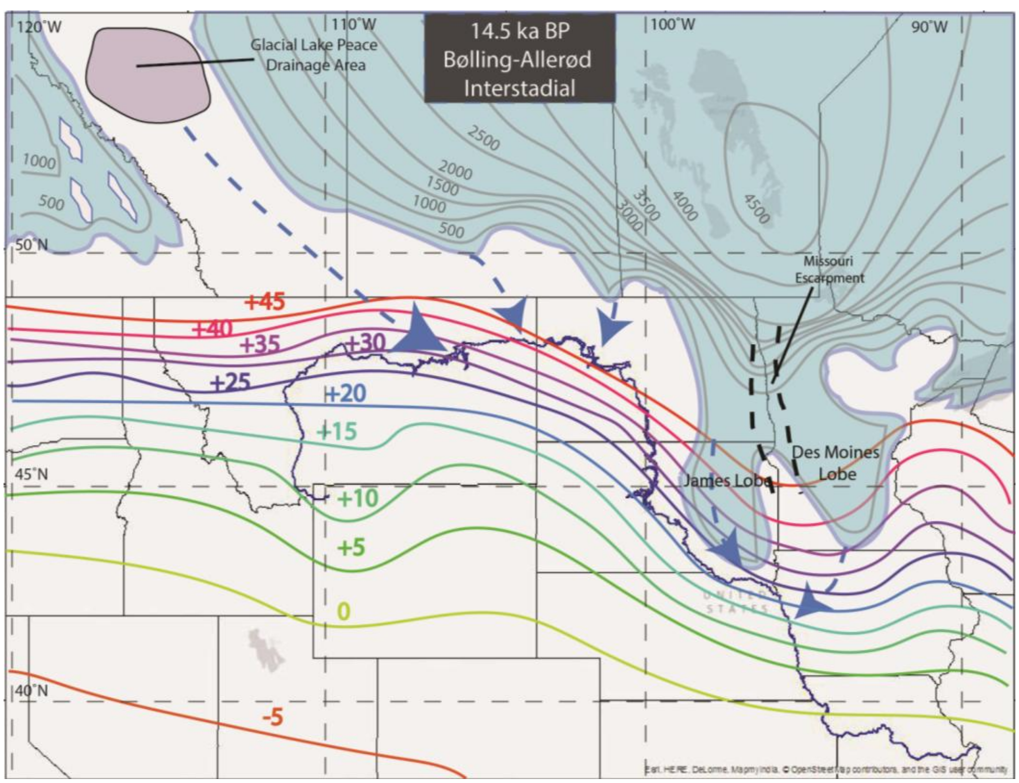


Figure 1.2E – 14.5 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

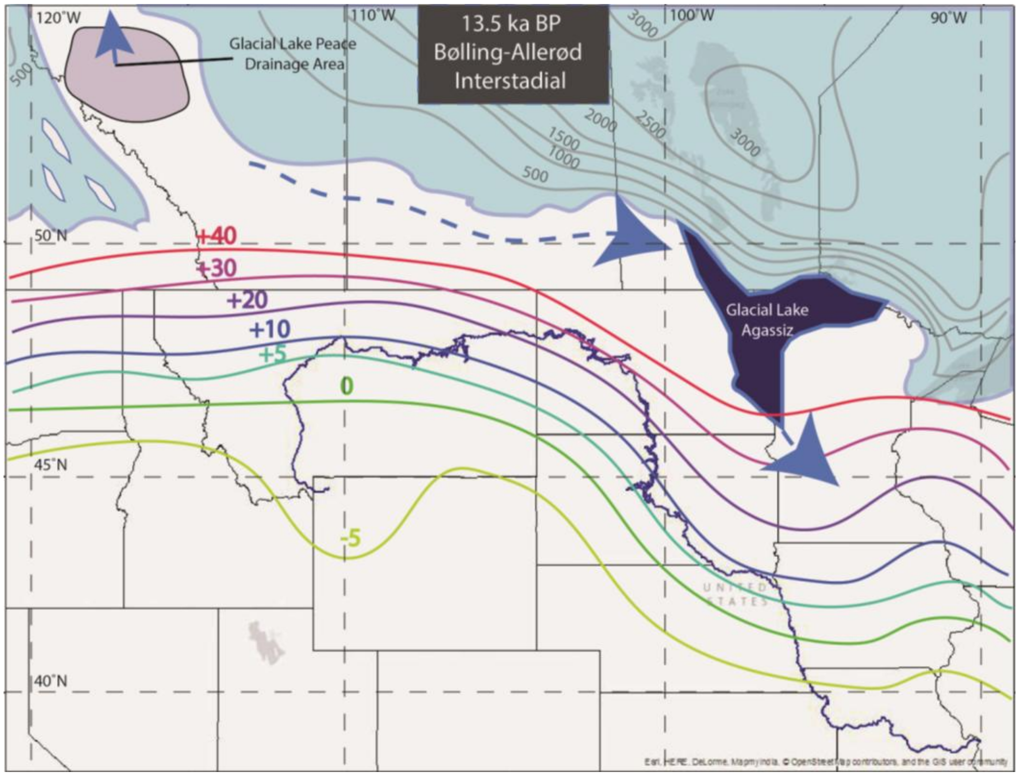


Figure 1.2F – 13.5 ka BP Laurentide Ice Sheet margins from (Dyke, 2004) and Peltier (2014).

Figure 1.2 A-F - Figures show proposed drainage, ice thickness, and GIA rates at 22.5 ka BP. 18.5 ka BP, 16 ka BP, 15.5 ka BP, 15 ka BP, 14.5 ka BP, and 13,5 ka BP. Glacial ice fronts are modified from Dyke (2004). GIA and ice thicknessis are modified from Peltier et al., (2014) and Argus et al., (2014). GIA countours shown are in meters/ka. Positive numbers represent rebound and negative numbers represent subsidence. Ice thicknesses are contoured in meters. Drainage pathways into Missouri River are taken from literature reviews (see 1.4).

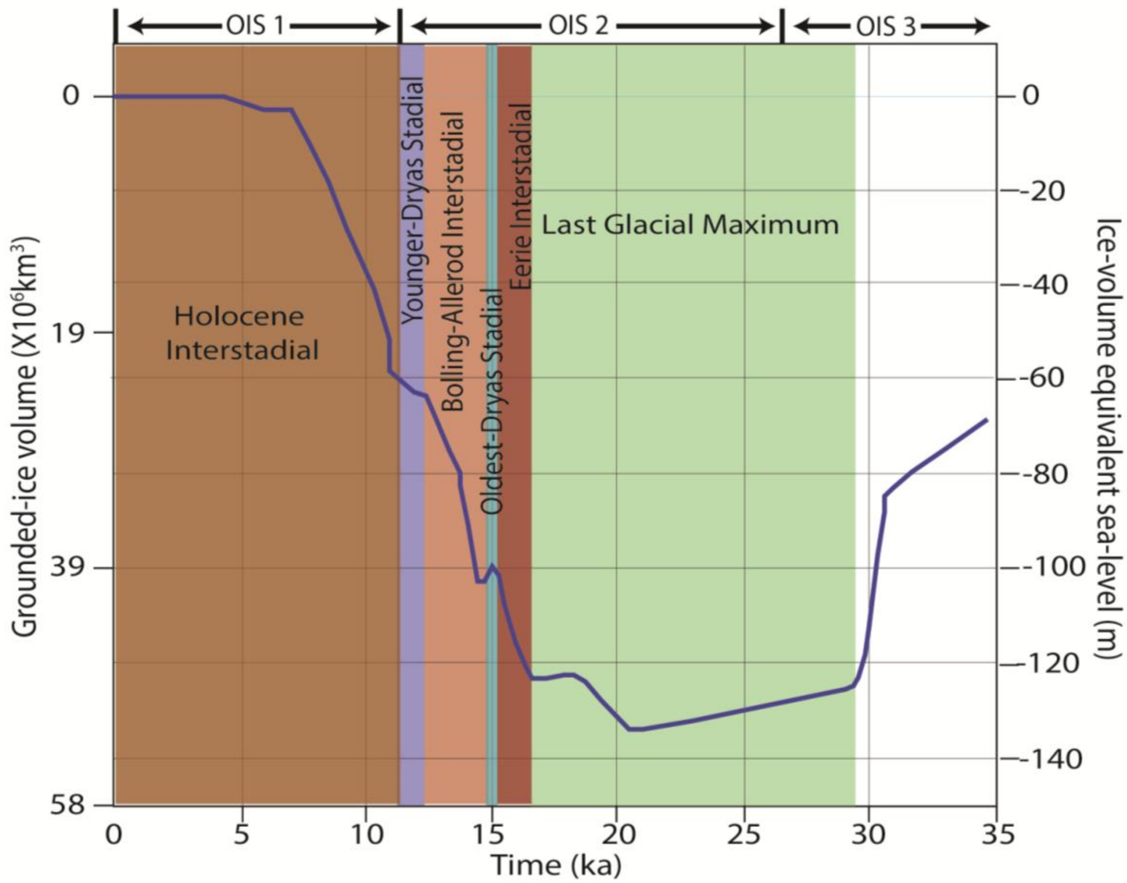


Figure 1.3 – Ice volumes and equivalent sea-level measurements through Late Wisconsin. Modified from Lambeck (2014). Interstadial and stadial events as well as oxygen isotope stages also shown for reference.

I.4 LGM-Holocene

Laurentide ice margins grew from the Middle Wisconsin interstadial margin until they reached their maxima (LGM) around 21.4 ka BP (Fig 1.2a, 1.3) during the last eustatic global seal level minimum (Dyke 2004; Lambeck 2014). The LGM ice margins pushed south of the Missouri Escarpment, a major drainage divide which separated flow between the Missouri River and Lake Agassiz to the east. Thus, drainage of the southwestern section of the Laurentide ice sheet was down the Missouri River (Kehew and Teller, 1994). The LGM ice margin blocked tributaries of the Missouri River in several locations forming glacial lakes. The Loma sublobe blocked (Fig 1.2a) the Missouri River north of Highwood Mountains, Montana forming glacial

Lake Great Falls (Hill et al., 2002). Other glacial lakes including glacial lakes Musselshell and Jordan were 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 were the James Lobe, with the Des Moines lobe supplementing drainage to the Missouri Valley through the modern day Little Sioux River tributary.

Deglaciation began at the end of the LGM around 20 ka BP and continued until approximately 16.5 ka BP (Lambeck 2014; Dyke 2004). By 17 ka BP (Fig. 1.2c), the glacial ice front retreated northward sufficiently in the Montana and North Dakota region to allow flow through temporary blockages (Dyke 2004). In addition, around this time between 17-16 cal ka BP (Fig. 1.2c), the Cordilleran-Laurentide ice saddle began ablating significantly (Dyke 2004). The Lake Peace drainage area began flowing southward towards the Mississippi drainage area around 17 ka BP (Catto et al., 1996) and by approximately 16 ka BP, the Maskwa drainage corridor reorganized into the Buffalo corridor in southern Saskatchewan sending southwestern Laurentide ice drainage directly down the James Lobe (Ross et al., 2009). All of these factors coupled would have added significant amounts of drainage area for the Missouri River Valley, though it is unknown as to what type of sediment traps were present at this time between the Cordilleran-Laurentide ice saddle and this study area.

Significant changes to the James Lobe and associated drainage to the Missouri River occurred during the Bølling-Allerød global warm period between 15.8-12.8 ka BP. Before the onset of the Bølling-Allerød warm period around 15.8 ka BP, the James and Des Moines lobes approximated their LGM margins (Fig 1.2c) (Dyke 2004). Accelerated warming caused complete melting of the James and Des Moines Lobes by 15.0 ka BP (1.2d) (Dyke 2004). The presence of white spruce (*picea glauca*) around the South Dakota/Nebraska border dated to 15,000 ka BP supports the disappearance of the James Lobe and the presence of a non-tundra landscape to this place at this time (Yansa 2006). During this brief disappearance of the James Lobe, drainage continued through the James River Valley due to ice from the Red River Lobe blocking eastward flow (Kehew and Teller 1994). By 14,500 ka BP, the James Lobe had readvanced back to the Yankton and Richland, SD area, approximating its previous southernmost boundary (Fig 1.2e) (Lundstrom 2009). It is unknown as to whether a glacial lake was present during this brief disappearance of the James Lobe, although it is possible 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 BP, the James and Des Moines Lobe were melted (Fig. 1.2f) (Dyke 2004). Glacial Lake Agassiz began forming around 14.3 cal ka BP (Lepper et al., 2011), and drainage of the southwestern Laurentide Ice Sheet soon switched (13.9 ka BP) into the eastward flowing Lake Agassiz drainage towards the Minnesota and Mississippi Rivers (Clayton and Moran 1982). The capture of the James spillway by Lake Agassiz around 13.9 ka BP marks the end of the Missouri River as a glacial drainage system.

I.5 Glacial Isostatic Rebound and Forebulge Migration

Glacial isostatic adjustment (GIA) is caused by the loading and unloading of large volumes of ice onto the continental crust. Amounts and rates of 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. Isostatic adjustment since the LGM is well documented (Tushingham and Peltier, 1991; Mitrovica et al., 1993; Peltier, 2002; Peltier, 2004; Argus et al., 2014; Peltier et al., 2014). The main differences in models are based on input values of mantle viscosity, ice-thickness, and glacial ice fronts, which each evolve with ever improving data. Modeling efforts are also tuned according to constraints from very-long-baseline interferometry (VLBI) (James and Lambert, 1993; Mitrovica et al., 1993, Argus and Peltier, 1999), GPS, and absolute gravity measurements (Larson et al., 2000; Lambert et al., 2001; Sella et al., 2007). GIA modeling efforts are of particular interest for this study due to the possibility of isostatic adjustment affecting regional slopes of the Missouri River Valley since the LGM which may affect vertical profiles as well as lateral migration of channels (Holbrook et al., 2006; Holbrook et al., 2012; Peakall, 1995). GPS observation by Sella et al., (2007) suggest the current "0" line which divides subsidence from rebound extends through the Great Lakes, north of this study area. However, in order to understand fluvial relationships with GIA, rates, location, timing, and magnitudes of GIA need to be constrained since the LGM. This study ultilizes the most recent modeling efforts (Argus et al., 2014; Peltier et al., 2014). Ice-thickness and topographic measurements through time were reported from this model (Argus et al., 2014; Peltier et al., 2014). In order to estimate modeled GIA rates through time, this study simply subtracts differences in topography through time to get an estimate of rebound and subsidence rates, timing, and location for intervals of interest. Modified contours from these models are presented in figures 1.2a-1.2f and are used to draw relationships between fluvial processes and GIA.

Chapter II: Methods

This study builds on undergraduate and graduate mapping efforts of the Big Muddy Expedition, funded by of the USGS EDMAP program and the National Science Foundation Research Experience for Undergrads performed from 2004-2012. Students mapped floodplain allounits (e.g., channel fills, bars, splays, etc.) over 7.5' quadrangles in groups of two. Allounits were mapped by first identifying surficial characteristics from aerial photos, topographic maps, digital elevation models, and field traverses (c.f. Holbrook et al., 2006a,b). Interpretations based on surficial characteristics were then tested by confirming associated characteristic lithofacies successions through the Dutch Auger hand-drilling method (c.f. Berendsen and Stouthammer 2001). Core samples from the Dutch Augers were taken at 10 cm intervals and logged for Munsell color, texture, oxidation state, and organic traces. USDA Field criteria were applied to ascertain sediment texture according to the USDA texture triangle (Soil Survey Division Staff, 1993). Crosscutting relationships were combined with OSL and ^{14}C dating techniques to develop numerical and relative age constraints of allounits.

The resulting database produced collectively by these studies includes 49 7.5' quadrangle maps of surficial Holocene allounits (see http://mri.usd.edu/pubsfor full compendium); 53 OSL samples (Table 1 and 2), and 2 ^{14}C dates. This database formed the spine of the dataset used in this study, but was supplemented with public state water well data from SD, NE, IA, KS, and MO. Lithology descriptions from these water wells helped to gain better control on surfaces beyond hand-auger depths and assisted in the construction of a longitudinal profile of Quaternary surfaces.

II.1 OSL Dating

OSL ages for this study are taken from quartz collected from point bars, channel fills, and terraces, generally covered by finer-grained deposits. Sample recovery used an opaque sampling tube mounted to the point of a hand auger to form a suction-coring device. Information recorded in the field for samples included elevation, burial depth, and latitude and longitude. All OSL dates are cross-referenced with crosscutting relationships where possible to assure they are consistent with landform successions and prior dates. Lab processing was performed at University of Nebraska at Lincoln by Dr. Ron Goble. Sample preparation was performed under amber light. Lab processing techniques included isolating sand sized particles (90-150μm 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 separated from quartz grains by treating samples with 48% HF acid for 75 minutes followed by 47% HCL for 30 minutes. The samples were then resieved to remove feldspars leftover from the acid wash. Quartz grains are 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 a Riso Automated OSL Dating System Models TL/OSL-DA-15B/C and TL/OSL-DA-10 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 180º and 280º 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 size 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 set. Central Age Model (Galbraith et al. 1999) were 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).

II.2 Longitudinal Profile

Cross sections showing Pleistocene surfaces were drawn from water well data and previously drilled boreholes throughout the entire lower valley (Yankton, SD to Colombia, Missouri) in order to create longitudinal profiles of alloformation surfaces. Construction of a longitudinal profile is done by combining public water well data to expand the known range of surfaces previously identified and dated from surficial mapping. Strip logs from public water wells provide lithology at depths which hand augering cannot reach. Lithology descriptions from these water wells are not as detailed as hand auger logs and generally record signifant textureal breaks only, but these data are more extensive. Alloformation surfaces are traced longitudinally by correlating between valley cross sections constructed from combinations of hand augered boreholes and water well records. Borehole data were spatially organized in ArcGIS using 30m Digitial Elevation Models.

Chapter III: Results

Missouri River Valley fill (Figs 3.2-3.10) records at least two cycles of aggradation and incision since the last glacial maximum and reveals five surfaces traced longitudinally from Yankton, SD to Colombia, MO (Fig 3.10). These surfaces reflect major swings in river gradients throughout the late Pleistocene and early Holocene. Starting within LGM, the Malta Bend Alloformation terrace records channel aggradation to elevations close to modern floodplain levels. The LGM Malta Bend Alloformation is incised to form the Carrolton surface around 16 ka BP. Between 16-13 ka BP, the Missouri River profile aggraded close to modern floodplain levels (Salix Alloformation), but again incised between 12-10 ka BP (Vermillion Alloformation) down to the previous Carrolton Alloformation grade . Finally, the river aggraded back to modern floodplain levels between 10-8 ka BP, where the river has migrated laterally with only minor vertical adjustment over the past 8 ka (Omaha Alloformation). Optically stimulated luminescence results are reported in Table 1 and 2, while locations of OSL boreholes are shown in Figs. 3.2-3.5.

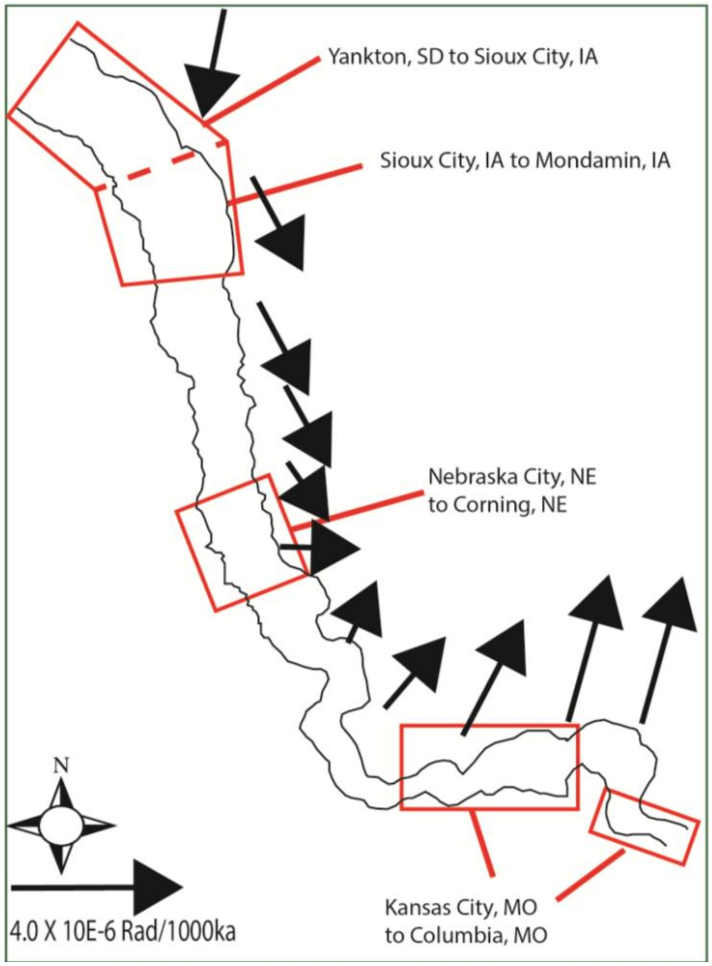


Figure 3.1 – Areas targeted for development of composite cross sections for the Missouri River Valley. Valley tilt vectors from GIA model of Argus et. al, (2014) and Peltier et al., (2014) are illustrated with black arrows. Valley tilt vectors represent the rotation since 8 ka BP. Lengths of arrows are consistent with magnitude of valley tilt. Valley tilt direction and magnitudes are taken from contours of topography differences over the past 8 ka from Argus et. al, (2014) and Peltier et al., (2014). Vector directions are drawn perpendicular to countours and magnitudes were determined by contour values over the past 8 ka.

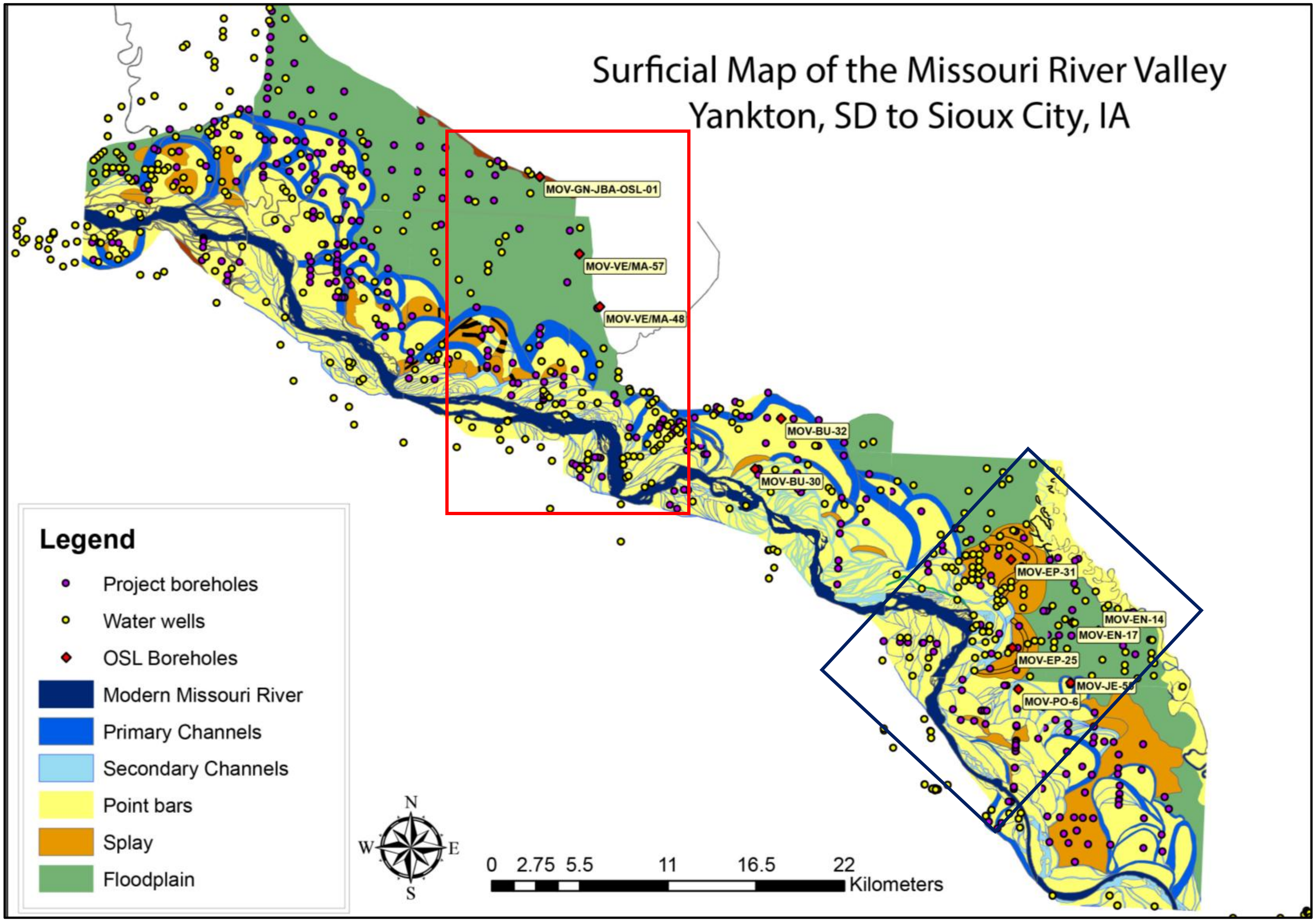


Figure 3.2 – Surficial map of the Missouri River with location of Vermillion composite cross-section (red box) and Elk Point composite cross-section (blue box).

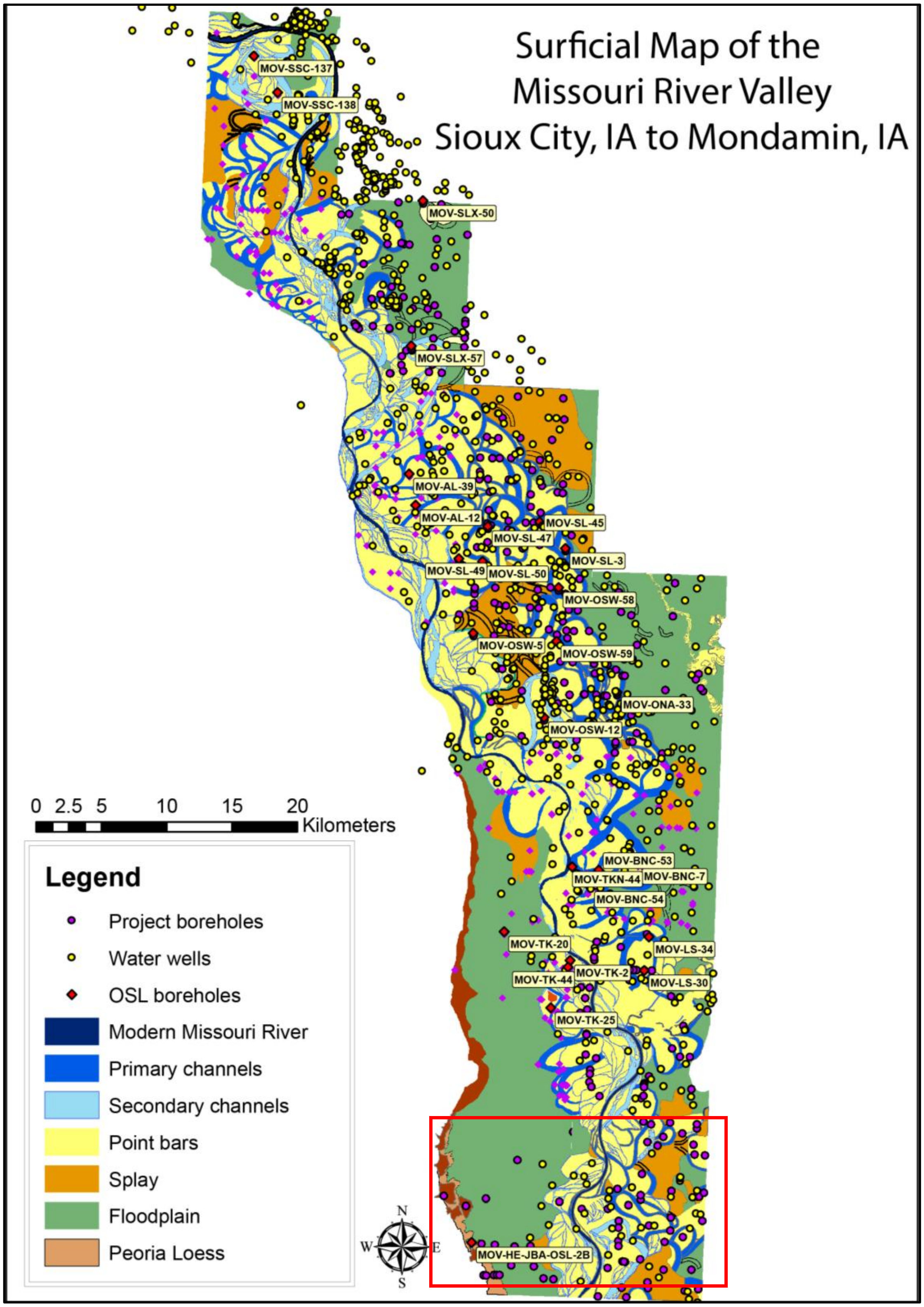


Figure 3.3 - Surficial map of the Missouri River with location of Herman composite cross-section (red box), Sioux City, IA to Mondamin, IA.

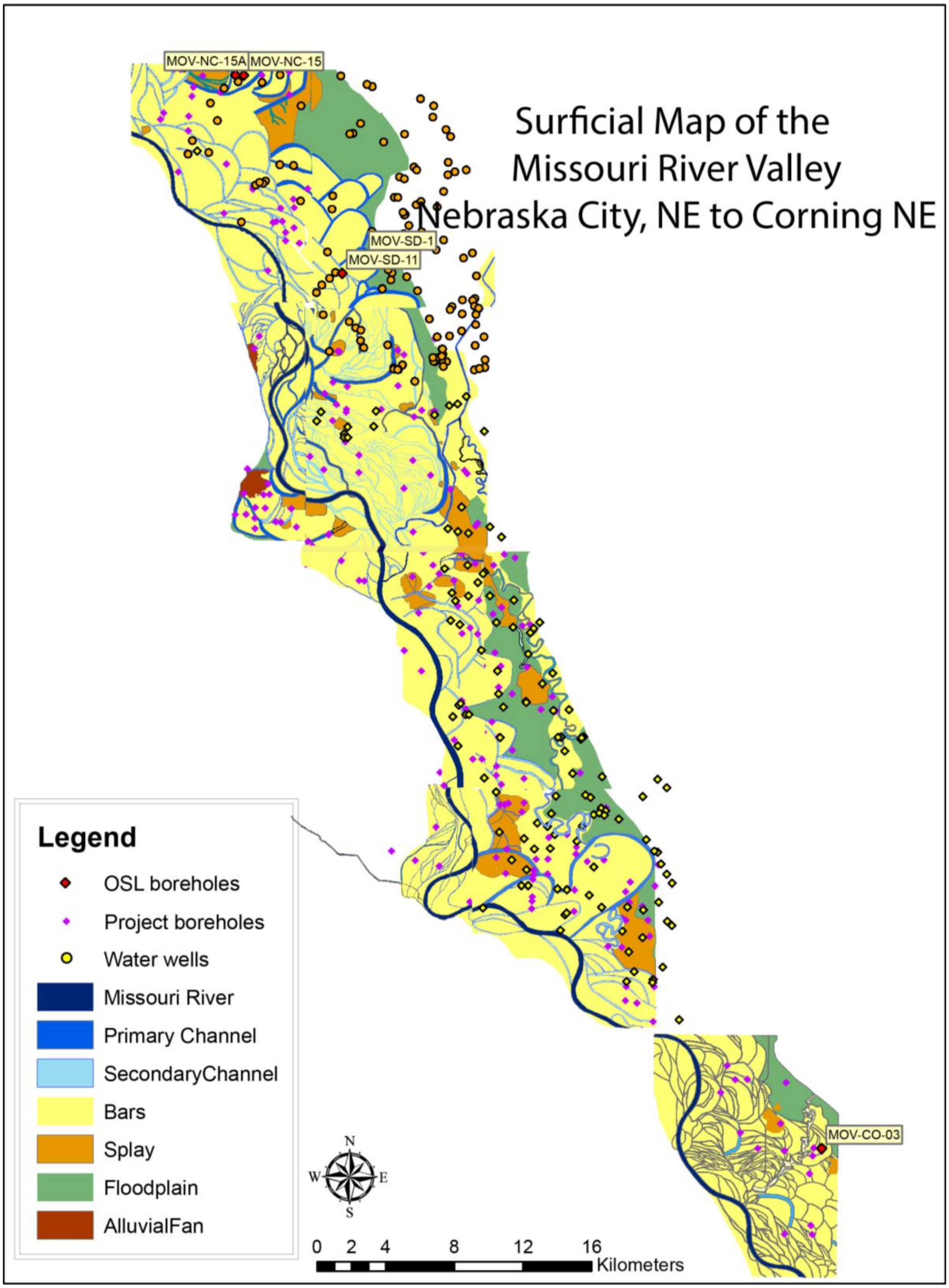


Figure 3.4 - Surficial map of the Missouri River with location of cross-sections. Nebraska City, NE to Corning, NE.

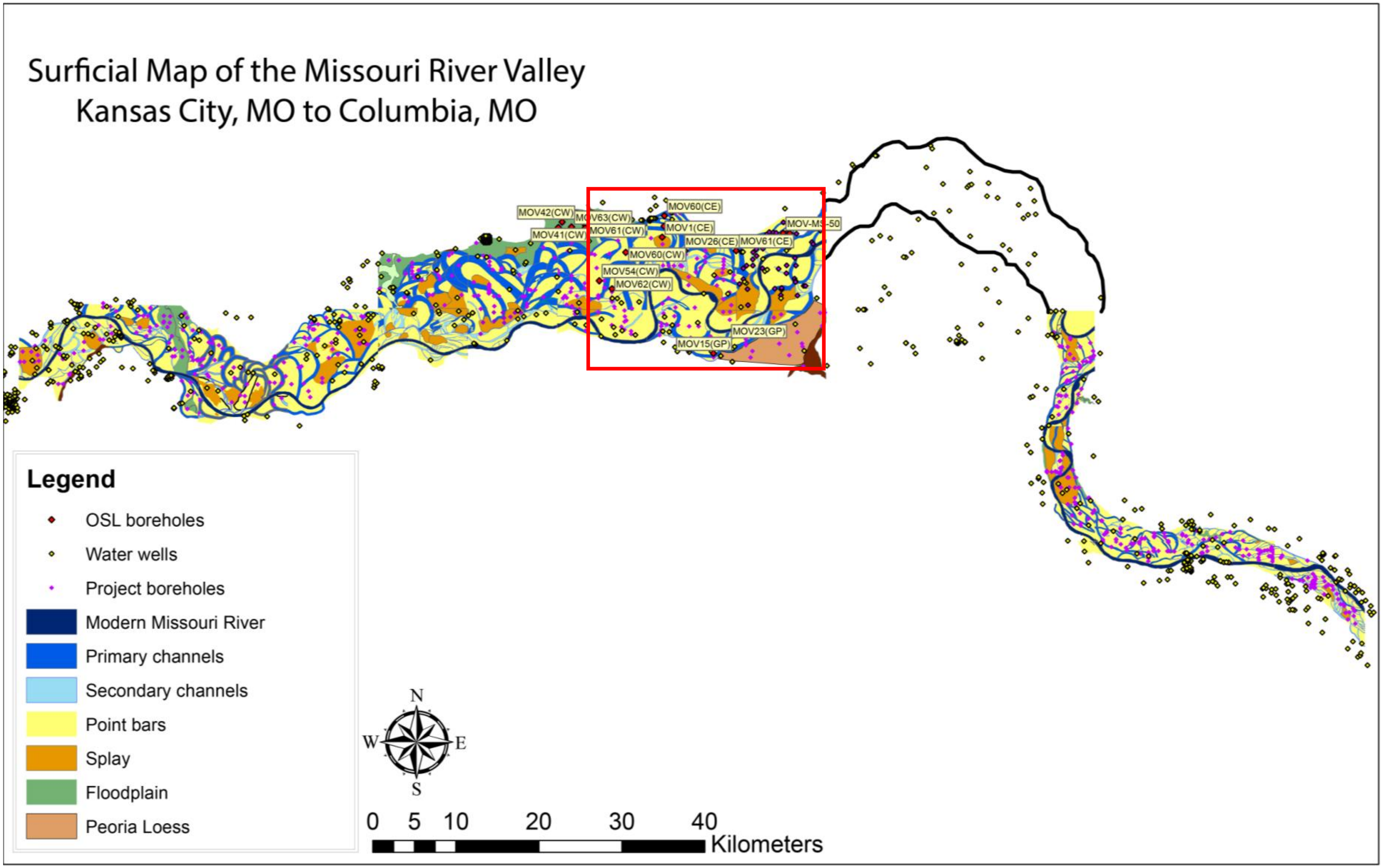


Figure 3.5 - Surficial map of the Missouri River with location of Malta Bend composite cross-section (red box). Kansas City, MO to Columbia, MO.

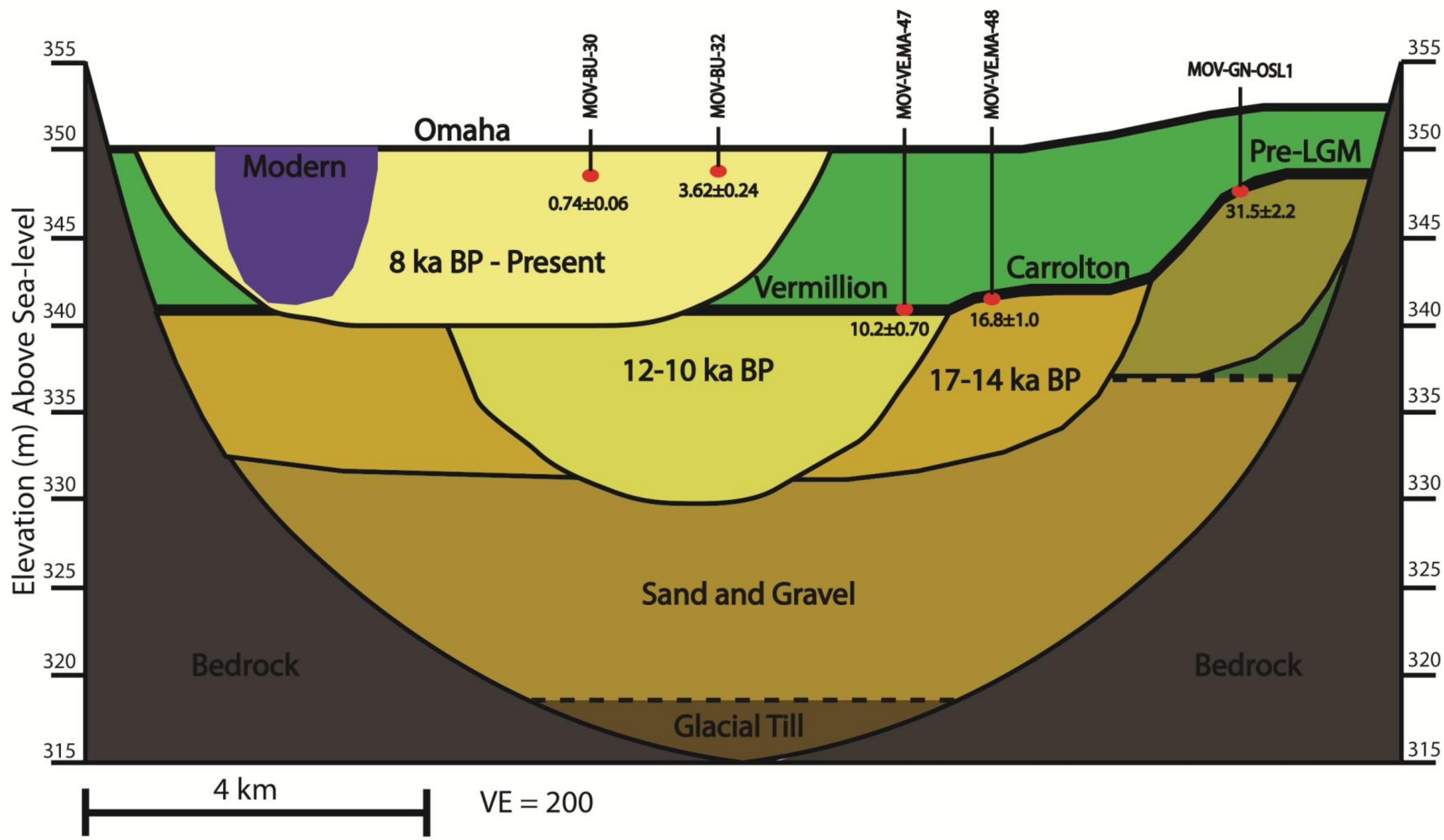


Figure 3.6 – Vermillion area composite cross-section.

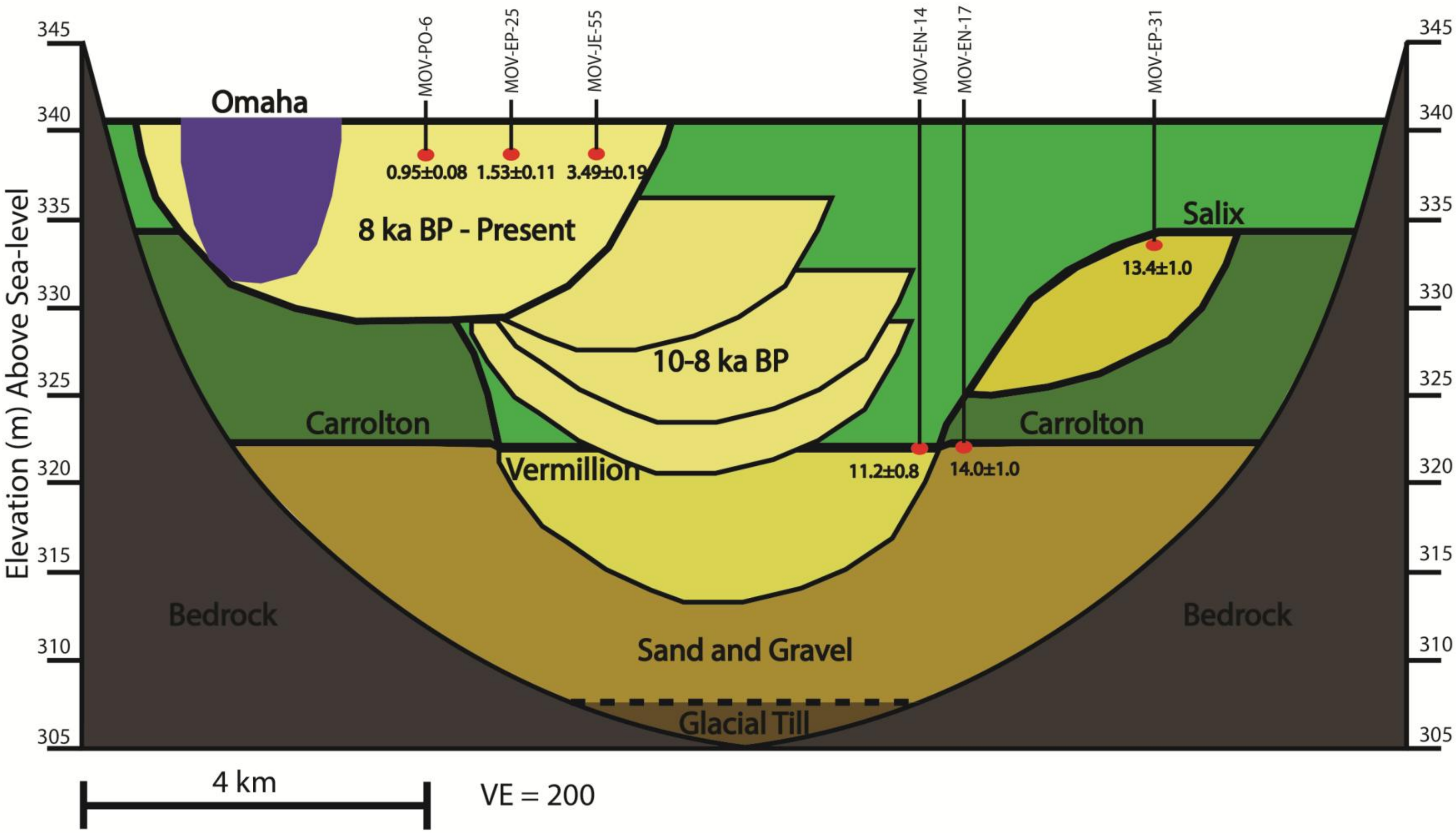
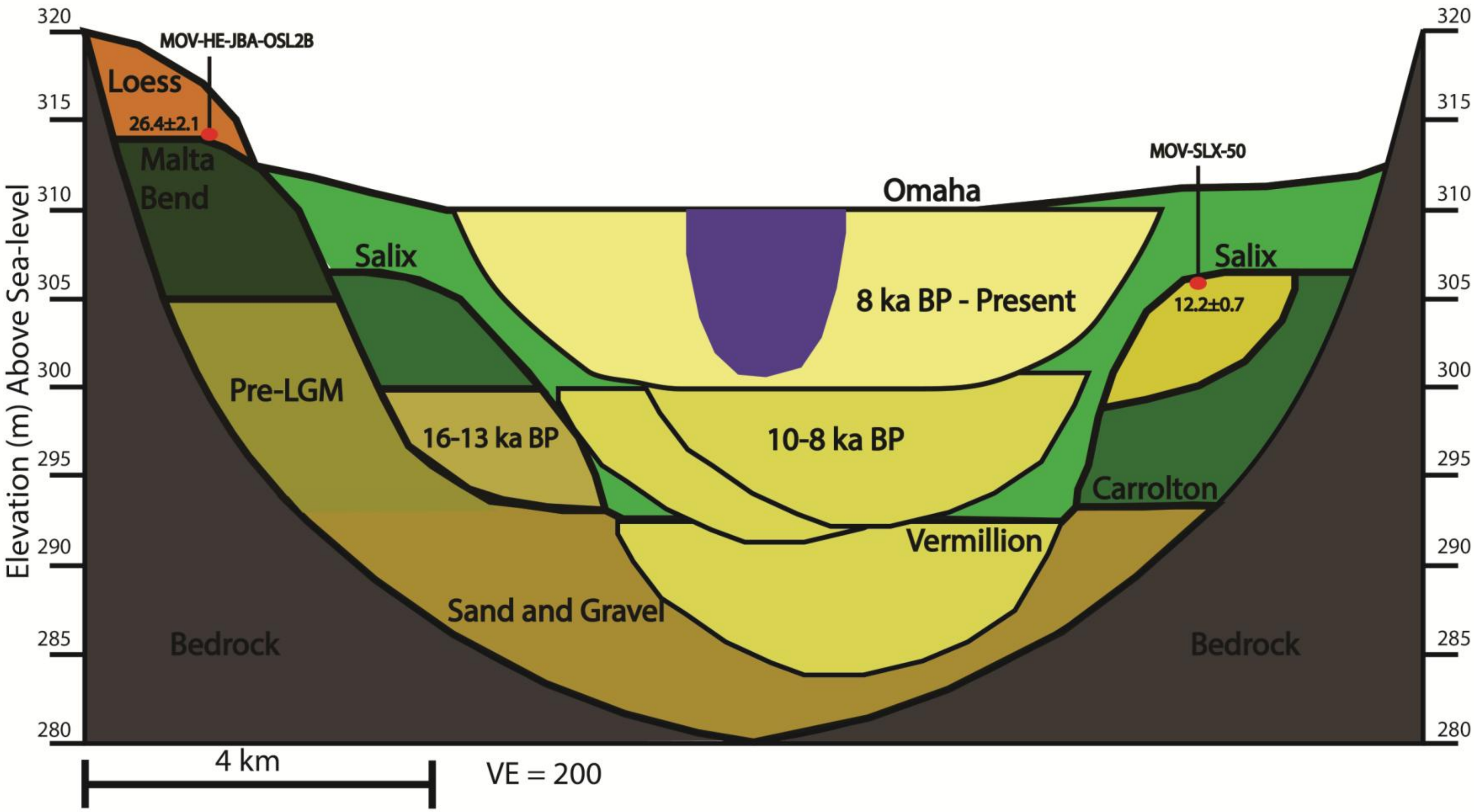


Figure 3.7 – Elk Point area composite cross-section.



Elevation (m) Above Sea-level
320
315
310
305
300
295
290
285
280
MOV-HE-JBA-OSL2B
Loess
26.4±2.1
Malta Bend
Salix
Pre-LGM
16-13 ka BP
Omaha
8 ka BP - Present
10-8 ka BP
Vermillion
Sand and Gravel
Bedrock
MOV-SLX-50
Salix
12.2±0.7
Carrolton
Bedrock
4 km
VE = 200

Figure 3.8 – Herman area composite cross section.

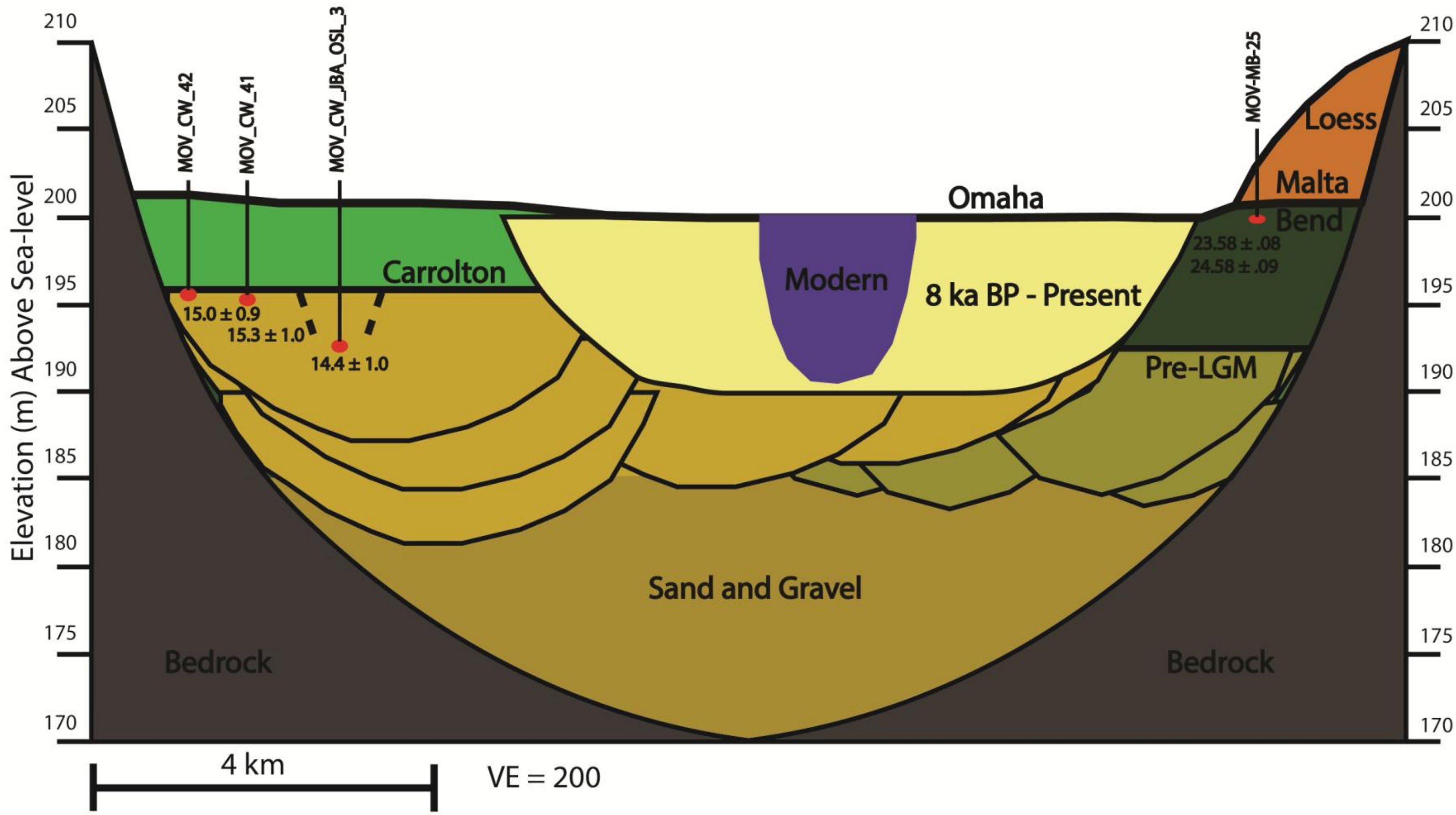


Figure 3.9 – Malta Bend area composite cross-section.

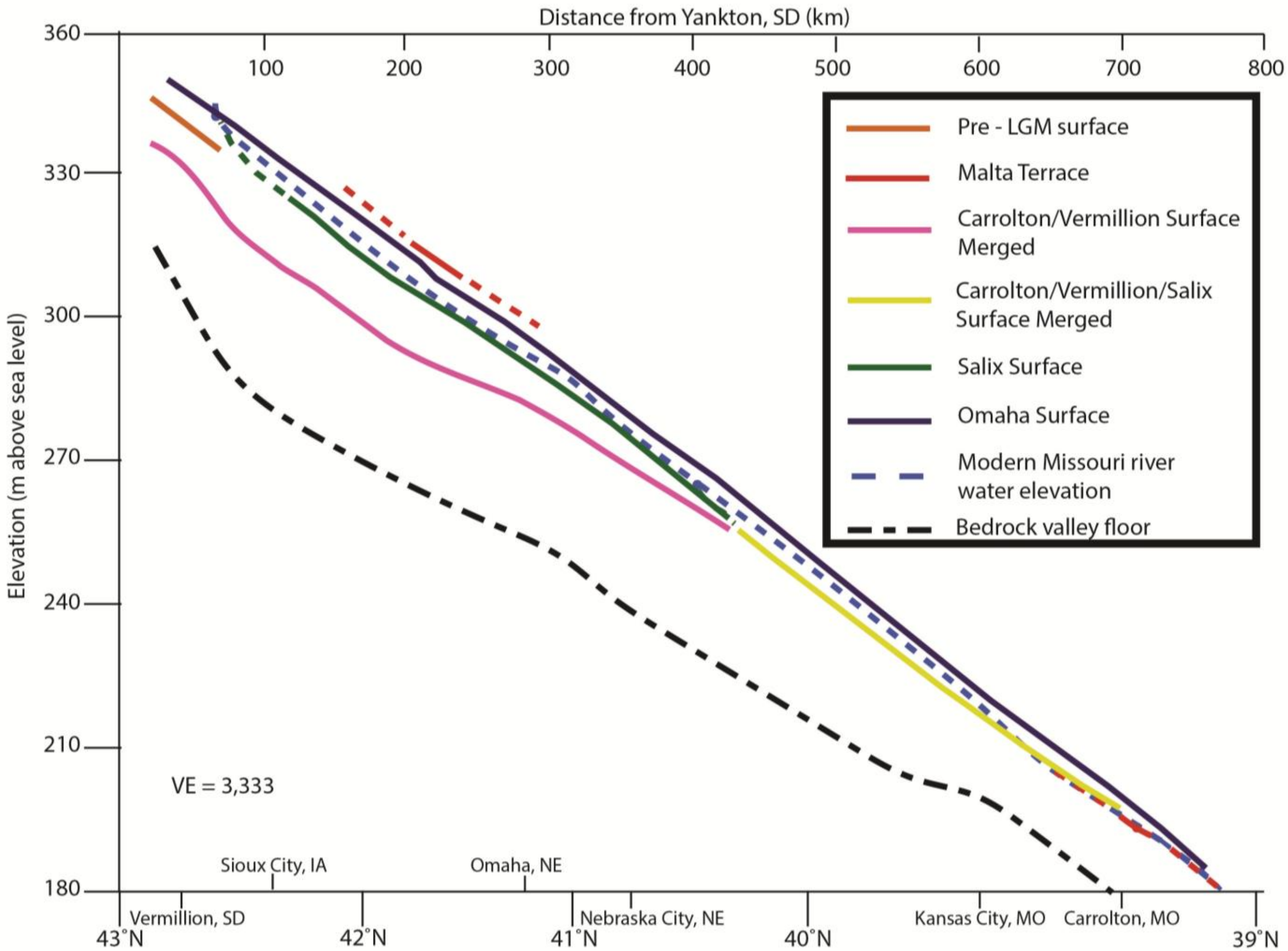


Figure 3.10 – Longitudinal profile of the Missouri River Valley. (Vermillion and Carrolton Surface are merged because resolution of data did not allow for separation of the two surfaces.) Vermillion, Carrolton, and Salix are all merged south of Nebraska City where separation of surfaces is below resolution.

Sample Number	UNL lab number	Alloformation	Dose rate (Gy/ka)	No. of aliquots	De (Gy)	Optical age (ka)
MOV-GN-JBA-OSL1	UNL3894	Pre-LGM	1.39±0.09	50	43.89±1.47	31.5±2.2
MOV-HE-JBA-OSL2B	UNL3896	Malta Terrace	2.55±0.15	55	67.51±3.37	26.4±2.1
MOV-VE/MH-57	UNL2126	Carrolton	1.56±0.08	24	26.27±0.18	16.8±1.0
MOV-EN-17	UNL2123	Carrolton	2.21±0.12	26	30.97±0.90	14.0±1.0
MOV-CW-41	UNL853	Carrolton	1.70±0.08	34	26.12±0.55	15.32±0.91
MOV-CW-42	UNL848	Carrolton	1.71±0.08	31	25.78±0.68	15.04±0.93
MOV-CW-JBA-OSL3	UNL3895	Carrolton	2.14±0.14	60	30.76±0.41	14.4±1.0
MOV-EP-31	UNL2127	Salix	1.91±0.12	27	25.66±0.83	13.4±1.0
MOV-SLX-50	UNL2496	Salix	2.03±0.10	44	24.92±0.27	12.2±0.7
MOV-EN-14	UNL2125	Elk Point	1.97±0.11	26	22.06±0.52	11.2±0.8
MOV-VE/MA-49	UNL2120	Elk Point	2.08±0.11	33	21.28±0.46	10.2±0.7
MOV-SL-45	UNL2491	Holocene (meandering)	1.93±0.09	31	14.05±0.47	7.29±0.48
MOV-SL-47	UNL2492	Holocene (meandering)	2.12±0.10	32	13.90±0.27	6.56±0.39
MOV-SL-3	UNL2501	Holocene (meandering)	2.21±0.10	41	14.56±0.50	6.59±0.42
MOV-CE-64	UNL850	Holocene (meandering)	2.18±0.09	35	12.89±0.35	5.92±0.34
MOV-CE-1	UNL852	Holocene (meandering)	1.84±0.09	36	9.55±0.29	5.18±0.33
MOV-CE-60	UNL858	Holocene (meandering)	2.26±0.14	34	10.86±0.32	4.81±0.35
MOV-ONA-33	UNL2493	Holocene (meandering)	2.32±0.14	29	10.86±0.47	4.68±0.37
MOV-CW-54	UNL862	Holocene (meandering)	1.85±0.08	28	8.19±0.24	4.43±0.27
MOV-CO-3	UNL3745	Holocene (meandering)	2.42±0.14	51	10.54±0.25	4.35±0.28
MOV-CW-60	UNL849	Holocene (meandering)	1.99±0.11	31	8.31±0.19	4.18±0.28
MOV-CE-59	UNL851	Holocene (transition)	2.05±0.09	27	8.13±0.19	3.98±0.22
MOV-BU-32	UNL2124	Holocene (meandering)	2.15±0.11	34	7.79±0.14	3.62±0.24
MOV-CW-61	UNL855	Holocene (transition)	2.42±0.11	30	8.65±0.38	3.57±0.48
MOV-JE-55	UNL2128	Holocene (meandering)	2.37±0.09	42	8.28±0.14	3.49±0.19
MOV-CE-26	UNL856	Holocene (transition)	2.14±0.11	29	7.30±0.17	3.41±0.21
MOV-CW-	UNL854	Holocene	1.92±0.10	30	5.67±0.14	2.95±0.19

62		(transition)				
MOV-50	UNL624	Holocene (transition)	1.87±0.09	45	5.18±0.70	2.77±0.42
MOV-BNC-7	UNL2819	Holocene (meandering)	1.44±0.07	50	3.93±0.10	2.72±0.15
MOV-CW-63	UNL859	Holocene (transition)	1.92±0.09	29	5.03±0.10	2.62±0.15
MOV-49	UNL623	Holocene (transition)	2.15±0.10	45	4.98±0.54	2.32±0.30
MOV-GP-23	UNL861	Holocene (transition)	2.38±0.09	27	5.52±0.22	2.32±0.14
MOV-OSW-59	UNL2833	Holocene (meandering)	2.05±0.11	57	4.45±0.09	2.17±0.12
MOV-CE-61	UNL860	Holocene (transition)	2.21±0.09	24	4.73±0.16	2.14±0.13
MOV-NC-15	UNL3742	Holocene (meandering)	1.82±0.08	52	3.89±0.17	2.13±0.15
MOV-BNC-53	UNL2818	Holocene (meandering)	1.63±0.08	51	3.22±0.12	1.98±0.13
MOV-TK-44	UNL2824	Holocene (meandering)	1.80±0.09	62	3.53±0.06	1.95±0.10
MOV-SD-1	UNL3739	Holocene (meandering)	1.94±0.09	55	3.79±0.15	1.95±0.12
MOV-LS-34	UNL2821	Holocene (meandering)	1.47±0.07	55	2.78±0.10	1.89±0.11
MOV-AL-39	UNL2499	Holocene (meandering)	2.75±0.10	27	4.90±0.24	1.78±0.12
MOV-NC-15A	UNL3743	Holocene (meandering)	1.82±0.09	56	2.96±0.16	1.63±0.12
MOV-OSW-5	UNL2494	Holocene (meandering)	1.70±0.09	35	2.65±0.14	1.56±0.12
MOV-EP-25	UNL2119	Holocene (meandering)	2.59±0.09	27	3.95±0.18	1.53±0.11
MOV-LS-30	UNL2825	Holocene (meandering)	2.01±0.10	61	2.88±0.08	1.44±0.08
MOV-SD-11	UNL3740	Holocene (Braided)	2.23±0.12	58	3.16±0.13	1.42±0.10
MOV-BNC-54	UNL2822	Holocene (Braided)	2.21±0.08	50	2.62±0.04	1.19±0.05
MOV-BL-19	UNL3744	Holocene (Braided)	2.49±0.14	35	2.70±0.13	1.08±0.08
MOV-AL-12	UNL2490	Holocene (Braided)	2.82±0.12	34	2.84±0.25	1.01±0.10
MOV-GP-15	UNL857	Holocene (Braided)	2.28±0.09	27	2.16±0.07	0.95±0.06
MOV-PO-6	UNL2121	Holocene (Braided)	2.37±0.11	53	2.25±0.12	0.95±0.08
MOV-TK-20	UNL2823	Holocene (Braided)	1.81±0.07	50	1.69±0.10	0.93±0.07
MOV-SL-50	UNL2830	Holocene (Braided)	2.55±0.10	54	2.13±0.10	0.83±0.05

MOV-TK-25	UNL2824	Holocene (Braided)	1.71±0.07	61	1.35±0.07	0.79±0.05
MOV-TK-2	UNL2832	Holocene (Braided)	2.12±0.08	59	1.63±0.10	0.77±0.05
MOV-BU-30	UNL2122	Holocene (Braided)	2.25±0.08	42	1.66±0.10	0.74±0.06
MOV-SL-49	UNL2834	Holocene (Braided)	2.43±0.09	58	1.66±0.11	0.68±0.05
MOV-OSW-58	UNL2497	Holocene (Braided)	2.66±0.11	37	1.80±0.15	0.68±0.6
MOV-SSC-138	UNL2498	Holocene (Braided)	2.26±0.08	37	1.52±0.19	0.67±0.09
MOV-TKN-44	UNL2820	Holocene (Braided)	2.15±0.08	63	1.39±0.08	0.65±0.04
MOV-OSW-12	UNL2495	Holocene (Braided)	3.26±0.11	43	1.91±0.15	0.59±0.05
MOV-SSC-137	UNL2500	Holocene (Braided)	2.73±0.09	47	1.46±0.12	0.46±0.06
MOV-SLX-57	UNL2827	Holocene (Braided)	1.91±0.07	66	0.82±0.04	0.43±0.03

Table 3.1 – OSL data collected throughout study area. 1 σ standard deviations are shown for ages.

Sample Number	**Burial Depth (m)**	**H_2O^*(%)**	**K_2O (%)**	**U (ppm)**	**Th (ppm)**	**Cosmic (Gy)**
MOV-GN-JBA-OSL1	7.3	29.9	1.48	1.30	3.40	0.09
MOV-HE-JBA-OSL2B	3.5	33.0	2.27	3.24	10.22	0.14
MOV-VE/MH-57	7.2	19.0	1.52	1.2	3.99	0.09
MOV-EN-17	7.3	27.0	2.03	2.5	7.80	0.09
MOV-CW-41	6.1	17.56	1.78	0.9	3.7	0.10
MOV-CW-42	4.9	19.09	1.76	1.1	3.7	0.11
MOV-CW-JBA-OSL3	9.0	38.2	1.97	3.66	7.03	0.08
MOV-EP-31	6.4	28.0	1.90	1.9	5.98	0.10
MOV-SLX-50	4.1	25.8	1.89	2.25	6.09	0.13
MOV-EN-14	6.0	24.5	1.89	1.9	6.15	0.10
MOV-VE/MA-49	8.3	26.3	2.17	1.9	5.62	0.08
MOV-SL-45	2.35	21.9	1.90	1.66	4.26	0.16
MOV-SL-47	1.58	21.5	2.14	1.62	4.67	0.18
MOV-SL-3	2.9	20.0	1.82	3.17	4.52	0.15
MOV-CE-64	4.2	12.65	2.05	1.3	5.1	0.12
MOV-CE-1	5.0	19.28	1.90	1.1	4.3	0.11
MOV-CE-60	2.7	31.27	2.28	2.0	8.1	0.15
MOV-ONA-33	3.28	34.2	2.15	3.37	7.57	0.14
MOV-CW-54	2.9	15.02	1.86	1.0	3.5	0.15
MOV-CO-3	1.6	28.4	2.54	2.14	6.93	0.18
MOV-CW-60	3.2	27.29	1.90	1.8	6.7	0.14
MOV-CE-59	1.7	12.72	1.70	1.3	6.4	0.17
MOV-BU-32	4.0	23.6	1.92	2.2	7.06	0.13

MOV-CW-61	1.9	12.98	2.24	1.5	5.7	0.17
MOV-JE-55	4.5	7.5	1.94	1.8	5.85	0.12
MOV-CE-26	3.9	20.58	2.00	1.6	6.9	0.13
MOV-CW-62	3.0	21.83	1.66	1.6	7.1	0.15
MOV-50	3.3	27.3	1.52	1.5	7.3	0.14
MOV-BNC-7	2.6	21.9	1.32	1.30	3.72	0.16
MOV-CW-63	1.8	14.14	1.89	1.0	3.6	0.17
MOV-49	2.0	27.0	1.81	1.7	7.5	0.16
MOV-GP-23	2.0	6.54	2.14	1.3	4.3	0.16
MOV-OSW-59	2.1	27.9	1.76	2.36	7.57	0.17
MOV-CE-61	3.0	11.53	2.02	1.3	5.2	0.14
MOV-NC-15	1.7	20.0	1.89	1.19	3.67	0.18
MOV-BNC-53	4.7	22.8	1.67	1.23	3.89	0.12
MOV-TK-44	3.7	20.8	1.78	1.46	4.06	0.14
MOV-SD-1	2.7	19.5	1.92	1.37	4.97	0.16
MOV-LS-34	1.5	20.3	1.59	0.83	1.95	0.18
MOV-AL-39	2.7	8.8	2.21	3.10	4.44	0.15
MOV-NC-15A	2.2	23.3	1.90	1.31	4.19	0.17
MOV-OSW-5	1.25	24.9	1.74	1.41	3.46	0.19
MOV-EP-25	4.4	5.1	2.07	2.0	6.07	0.12
MOV-LS-30	2.9	23.9	2.08	1.67	4.38	0.15
MOV-SD-11	2.0	28.7	2.05	2.62	7.59	0.17
MOV-BNC-54	0.7	8.2	1.77	1.60	5.23	0.20
MOV-BL-19	0.3	26.3	2.22	3.08	7.39	0.22
MOV-AL-12	1.75	17.1	2.18	3.51	8.10	0.17
MOV-GP-15	3.3	5.48	1.92	1.3	5.3	0.14
MOV-PO-6	4.3	20.9	2.02	3.0	6.35	0.13
MOV-TK-20	1.7	10.4	1.59	1.20	3.64	0.18
MOV-SL-50	1.7	12.0	2.00	2.43	7.07	0.18
MOV-TK-25	2.1	13.0	1.52	1.07	4.21	0.17
MOV-TK-2	1.5	7.1	1.55	1.78	5.77	0.18
MOV-BU-30	2.8	6.6	1.90	1.6	4.69	0.15
MOV-SL-49	1.8	11.7	2.07	1.92	5.92	0.17
MOV-OSW-58	0.85	16.5	2.13	3.69	4.69	0.20
MOV-SSC-138	2.48	7.7	1.73	1.46	7.43	0.16
MOV-TKN-44	1.8	10.2	1.66	1.92	5.77	0.17
MOV-OSW-12	1.05	5.9	1.61	4.03	13.99	0.19
MOV-SSC-137	1.5	4.6	1.95	3.08	4.84	0.18
MOV-SLX-57	1.1	6.4	1.59	1.16	3.90	0.19

Table 3.2 - OSL data collected throughout study area.

III.1 Malta Bend Alloformation

The Malta Bend Alloformation is the oldest surface correlated within this study and formed during the LGM. This surface is dated in Herman, NE (Fig 3.2), and in the type area on the Malta Bend quadrangle of Missouri (Fig 3.4). In Herman, NE, the Malta fluvial channel belt surface is located 7 meters below ground level on a high terrace within the valley, which lies between 5-10 meters above the modern floodplain and places the Malta Bend Alloformation on grade with the modern river floodplain (Fig 3.8). The Malta Bend Alloformation in the Malta Bend area is a thick floodplain deposit formed from a channel surface which no longer occurs within the valley and likely was eroded by later incision events. Deposits of the Malta Bend terrace underlie the Peoria Loess in both locations. In Herman, NE, the transition between the Peoria Loess and the Malta terrace is distinct due to change in grain size and color. The Peoria Loess is a fine, massive, well oxidized silty loam (10 yr 5/4) while the underylying Malta Bend Alloformation is a dark (10 yr 5/1) clay with lighter gray mottles and manganese concretions. There is a reworked zone of approximately 50 centimeters in the uppermost part of the Malta Bend Alloformation, where fine silts and sands of the Peoria Loess are mixed with clays from beneath. No distinct paleosol is developed here beyond this evidence for mixing. An OSL date (MOV-HE-JBA-OSL-2B) of 26.4±2.1 ka BP from the Herman, NE location comes from the contact of the Peoria Loess and the underlying Malta Bend alloformation. Older Roxanna (55-27 ka BP) (Leigh and Knox 1993; Rodbell et al., 1997; Marowich et al., 1993 in Rittenour et al., 2007) and Loveland (190-120 ka BP) (Forman and Pierson 2002) loess deposits of the surrounding area are absent. In the Malta Bend quadrangle of Missouri, the Malta Bend Alloformation occurs about 11 meters below a high terrace bench covered in Peoria Loess. The ground level where the Malta Bend Alloformation was drilled in the Malta Bend quadrangle is between 7-15 meters above the modern floodplain. Like in the Herman location, the contact between the Malta 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 3.9). 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 (Fig 3.9), a thin peaty layer was found within the Malta Bend Alloformation near the contact with the Peoria Loess. Two AMS C^{14} samples from this layer were dated to 23.58±.08 y.b.p and 24.458±0.09 y.b.p. (Fig 3.9) correlating closely to the OSL date from the Herman, NE area and dating the time when the Malta channel system aggraded to this elevation.

III.2 Carrolton Alloformation

Incision of the Malta Bend Alloformation between the LGM and 16 ka BP led to one of the deepest surfaces mapped in this study. The Carrolton surface ranges widely in depth along its longitudinal profile (Fig 3.10). In the northernmost area of study, around Yankton, SD (Fig 3.2), the Carrolton surface is 7.0 meters below the modern floodplain and dates to 16.8±1.0 ka BP (Table 3.1) Farther down dip, in the Elk Point quadrangle of SD, (Fig 3.2) the Carrolton surface is 18.0 meters below the modern floodplain and is dated to 14.0±1.0 ka BP. The Carrolton Alloformation is composed of sandy channel belt deposits at these two northern locations and directly underlies a thick unit of younger back swamp gleyed clays (Gley 2 4/10bg) that extends from the Carrolton to the modern surface where not interrupted by later channel-belt deposits. Carrolton channel belt deposits are well oxidized and range from medium-to-coarse sands (5y 4/2) which fine upwards into loams and silty loams. Farther south in the Carrolton West and East quadrangles of Missouri (Fig 3.5) this surface is much shallower, between 4.0-4.3 meters below the modern floodplain. Here, apparent bar deposits of the Carrolton surface are covered by younger backswamp muds. The Carrolton channel belt deposits are capped by a thin (15cm) poorly developed loamy paleosol. The paleosol consists of a mixture of gray silt and clay which grade into fine sands. The fine sands are heavily mottled with Mn nodules and Fe concretions.

Root traces occur in the top 10 centimeters of this horizon. Channel deposits range from medium-to-coarse well oxidized sands and some fine gravels which are poorly sorted and well rounded and are dated at two adjacent locations to 15.0±1.0 and 15.1±1.0 ka. An apparent younger channel did cut through the Carrolton surface (Fig 3.9) adjacent to these dated bar deposits. This channel fill is about 4 meters thick, consisting of silty clays (10yr 4/2) and the sandy basal part of the channel is dated to 14.4±1.0 ka BP. The Carrolton Alloformation lasted from at least approximately 16-14 ka BP and had constructional channel and bar topography. Water wells throughout the valley allowed for tracing of this surface (Fig 3.10) and shows extreme deepening around Elk Point, and gradual shallowing towards the south.

III.3 Salix Alloformation

The Salix Alloformation marks an apparent period of aggradation in the late Pleistocene and is mapped between 4-6 meters under the floodplain in the northeast and eastern sides of the Valley between Elk Point, SD, and Sioux City, IA. This surface "ghosts" through onto the modern surface and reveals a single-channel meandering morphology with definable point bars and wrapping abandoned channel fills (Fig 3.3). Point bars of Salix meander loops are dated in two locations. In the Elk Point quadrangle (Fig 3.2) of SD, the point bar sand deposits are 6.4 m below the modern floodplain and are dated to 13.4±1.0 ka BP. On the Salix quadrangle (Fig 3.3), point bar deposits are 4 m below the modern floodplain and are dated to 12.2±0.7 ka BP. Point bar deposits at both locations are fine and well oxidized sand with abundant manganese and iron stains. Colors range from 10yr 4/3 to 10yr 9/3 with a 10yr 8/6 mottle. Water wells allowed for tracing of this surface longitudinally to just south of Nebraska City, NE. Here the Salix surface grade becomes indistinguishable from the Carrolton and Vermillion which have both shallowed here to the approximate the grade of the Salix surface (Fig 3.10).

III.4 Vermillion Alloformation

The Salix Alloformation is incised around the Pleistocene/Holocene transition to form the Vermillion Alloformation. In the Vermillion quadrangle, the Vermillion surface is incised into the Carrolton surface and is 8.3 meters below the modern floodplain (Fig 3.2). Here, the Vermillion surface is fine-to-medium well oxidized channel-belt sands (5 yr 4/2) dated to 10.2±0.7 ka BP. Farther south, in the Elk Point quadrangle, the Vermillion surface is found significantly deeper at 18.3 meters, again incised into the predated Carrolton surface. Sandy channel-belt deposits here were optically dated to 11.2±0.8 ka BP (Fig 3.6). The Vermillion profile is close in elevation and grain size and commonly indistinguishable from the Carrolton surface where dates are not available. Resolution of the water-well data base is not sufficient to distinguish these units and the profile of the Carrolton and Vermillion surfaces are thus merged (Fig 3.10).

III.5 Holocene to Present (Omaha Alloformation)

Numerous dates throughout the valley of oldest surficial channel-belt deposits show the river aggraded to its modern profile from the Vermillion profile by 8 ka BP. Surficial maps are shown in figures 3.2-3.5. The river appears to have migrated laterally at this level with only minor elevation shifts since. While there appears to be minimal vertical incision or aggradation of the Missouri River over the last 8 thousand years (Holbrook 2006), the river has migrated significantly laterally. Figs. 3.11-3.14 show trends of movement from the latest Pleistocene to the present. In the stretch between Yankton, SD to Sioux City, IA, the channel belts have consistently moved from the northeast side of the valley towards the southwestern side (Fig 3.11). For this reason, Pleistocene deposits are preserved on the northwestern side of the valley. The river also changes pattern in conincidence with this lateral shift (Holbrook et al., 2006; Kashouh

2012). The Missouri river has a meandering morphology between 8-1.5 ka in the reach from Omaha north and then switches to a braided pattern around 1.5 ka BP (Kashouh 2012). The location of this braided channel belt is consistently on the southwestern side of the valley from Yankton, SD, to Sioux City, IA, allowing for the preservation of the 8-1.5 ka BP meandering on the north side of the valley only morphology. The same change in morphology is recorded between Sioux City, IA down to Kansas City (Kashouh 2012) (Figs 3.2-3.5). However, throughout this stretch there is not a consistent direction of movement across the valley. Instead, the meandering and braided morphologies are distributed throughout the valley and are commonly flanked east and west by the Pleistocene section. The southernmost section, however, from Kansas City to Columbia, MO, returns back to the same consistent southward movement across the valley as seen in the northernmost section between Yankton, SD and Sioux City, IA. This stretch shows a transition in pattern as well, but slightly amended to the pattern farther up dip. This river is meandering from approximately 8ka bp to 3.5 kabp, but then has a transitional meander to braided pattern from approximately 3.5 kabp to 1.5 kabp before turning fully braided thereafter (Holbrook, et al., 2006). The locations of the three pattern show a progression from the northern side of the valley towards the south, with Pleistocene deposits preserved in the northern side and modern braided systems on the southern side of the valley (Figs. 3.11-3.14).

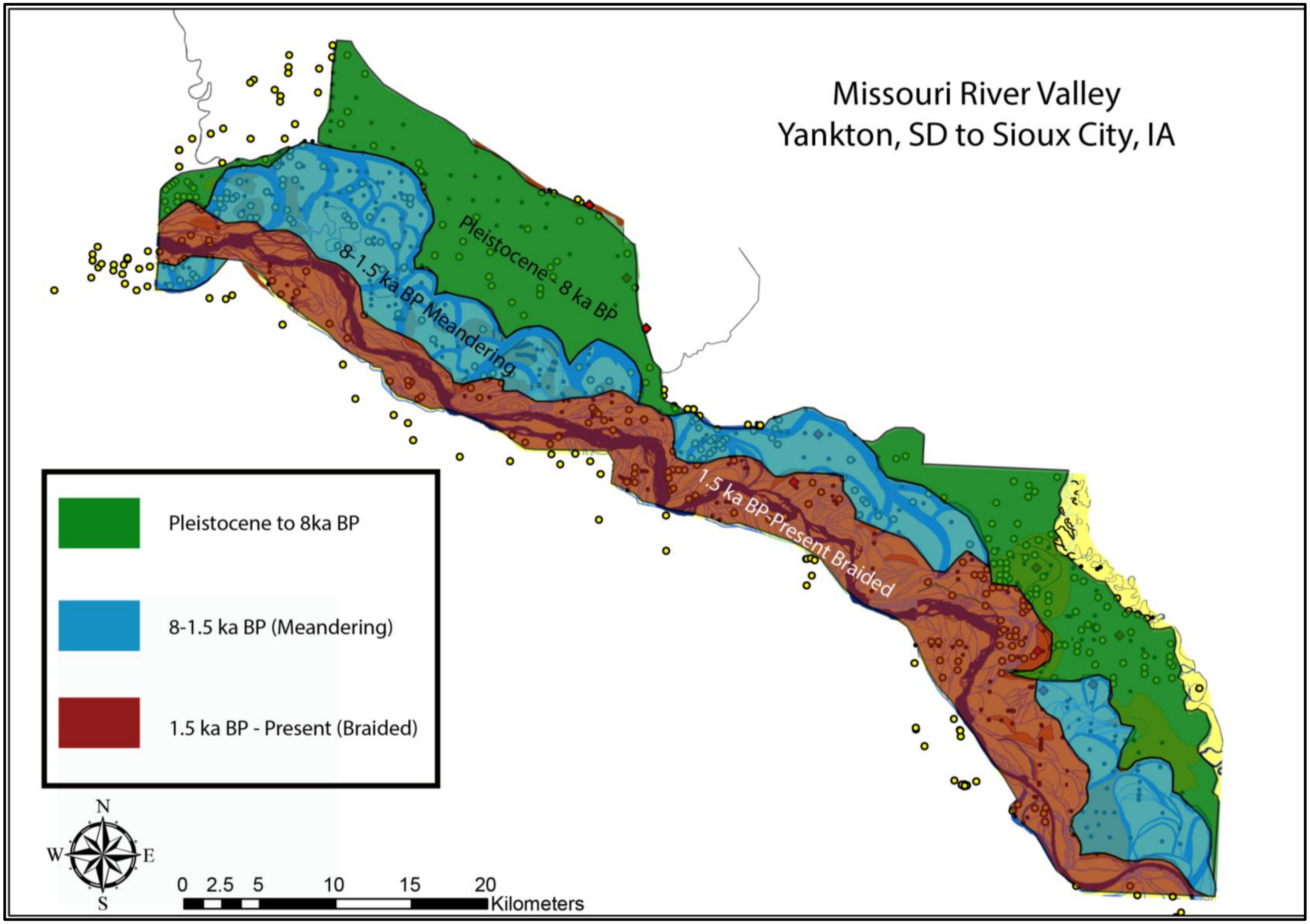


Figure 3.11 – Map showing lateral migration of Missouri River over past 8 ka.

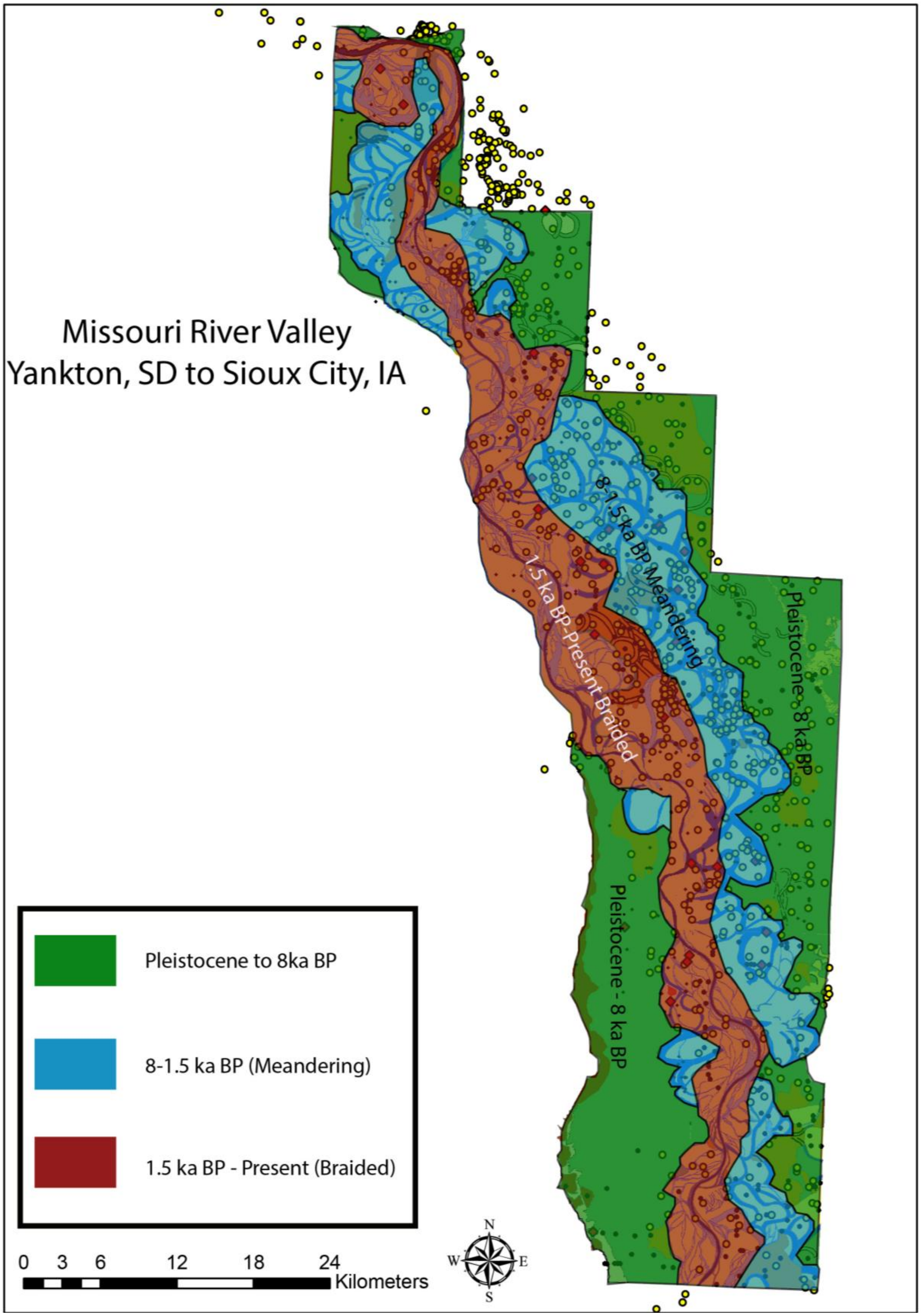


Figure 3.12– Map showing lateral migration of Missouri River over past 8 ka.

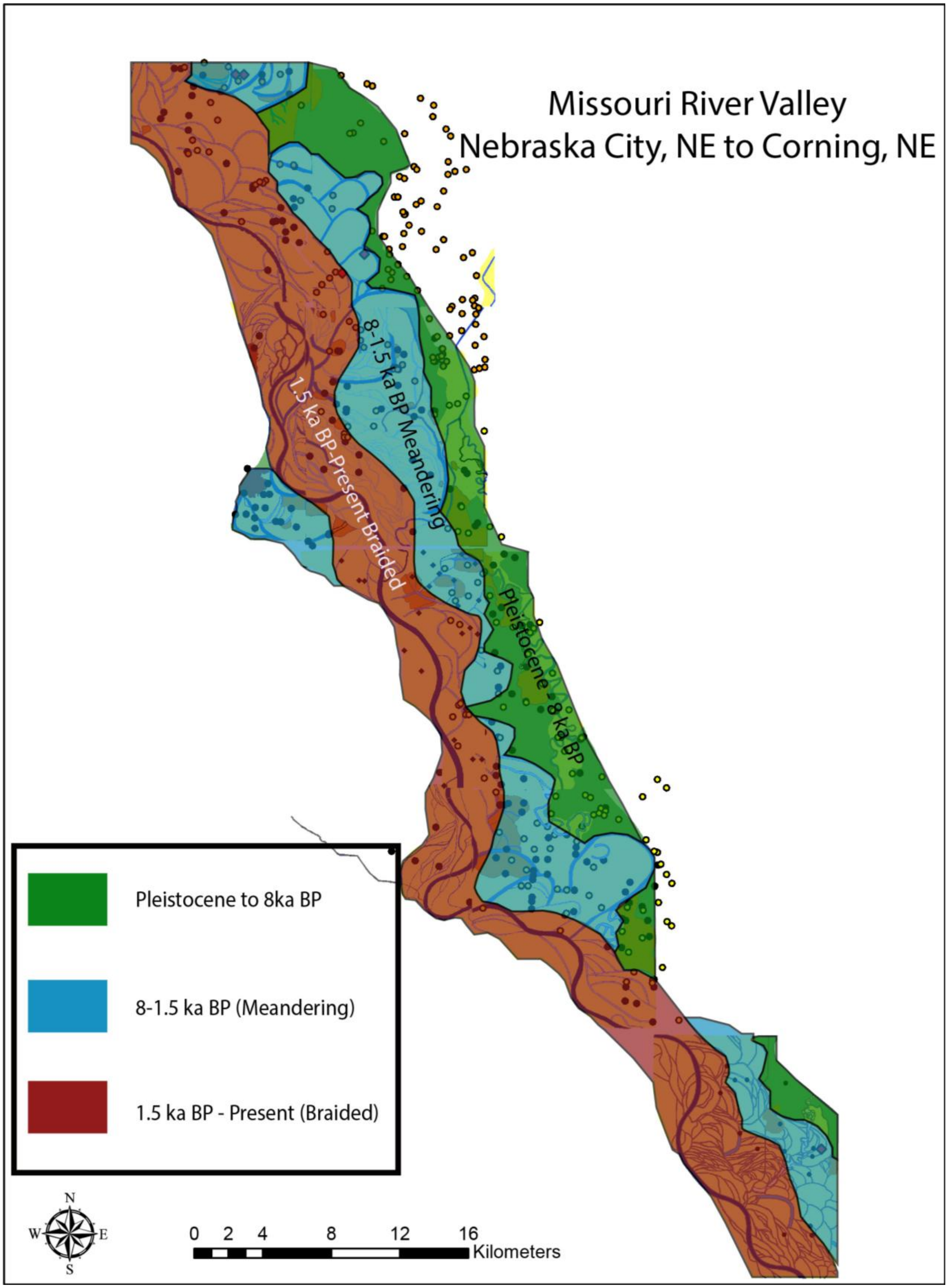


Figure 3.13 – Map showing lateral migration of Missouri River over past 8 ka.

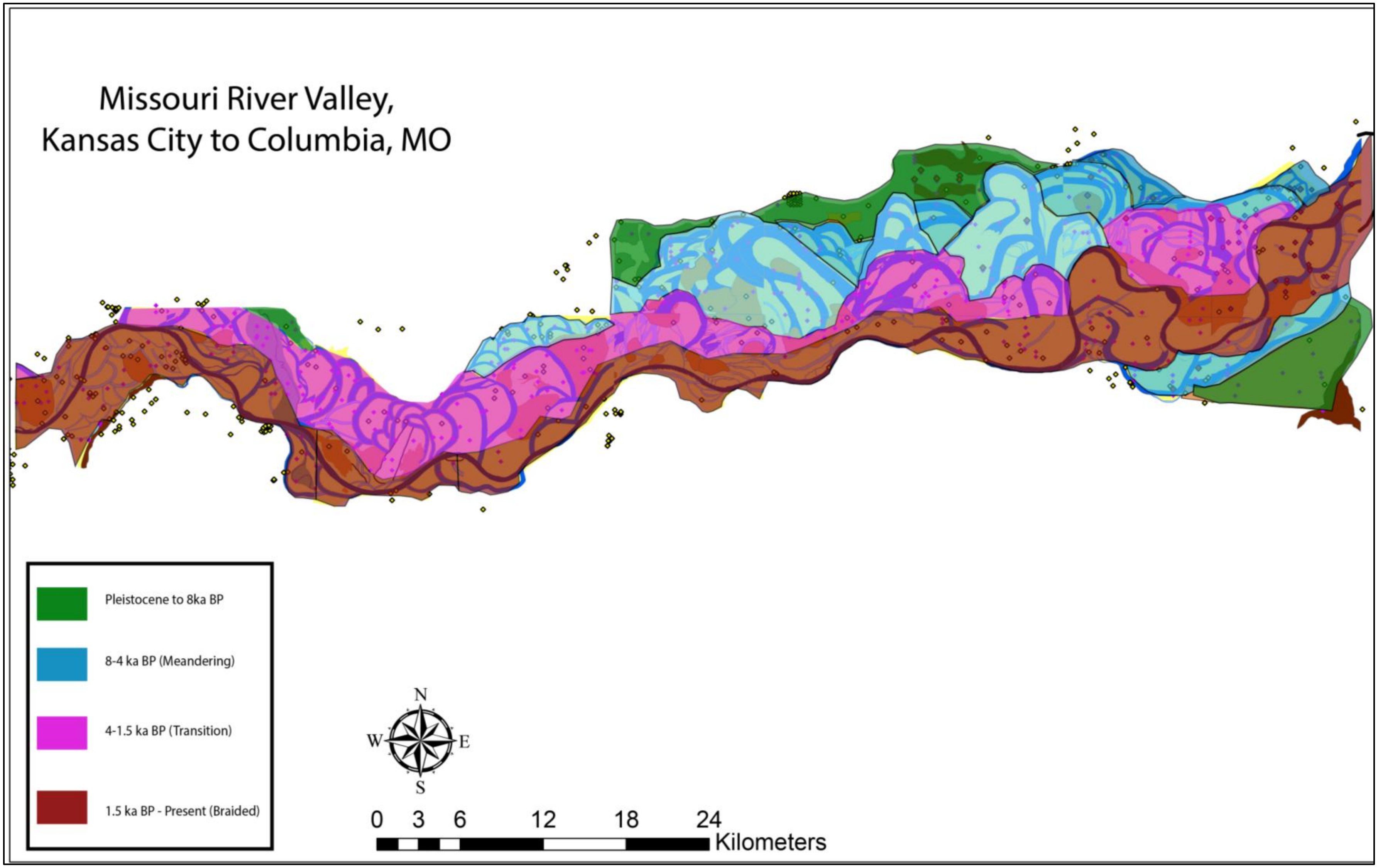


Figure 3.14 - Map showing lateral migration of Missouri River over past 8 ka.

Chapter IV: Discussion

IV.1 Incision and aggradation cycles in the Missouri River

Results from mapping and OSL dating argue for two full cycles of aggradation and incision since the LGM as recorded in longitudinal profiles of channel surfaces (Fig 3.10). The oldest surface recorded within this study spans the LGM (Malta Terrace) and records a river profile near the modern grade of the Missouri River. The Malta Terrace is covered by Peoria Loess which post-dates the LGM. The Peoria Loess on the Malta Terrace likely sourced as wind-blown silt from later (18 to 12 ka BP) outwash plains in the adjacent valley. The subsequent Carrolton surface records a time of incision throughout the valley. Rivers of the Carrolton surface appear to have spread widely throughout the valley only leaving small portions of the Malta Bend Alloformation preserved in a few valley inlets where the geometry of the valley proved optimal for protection from erosion. Incision stabilized to form Carrolton surface by approximately 16 ka BP. OSL and borehole resolution from this study cannot confirm the geomorphology of this surface during incision or after stabilization, nor the exact timing at which incision began. Instead this study confirms that the Missouri River had a significant drop in grade between the LGM and 16 ka BP, which coincides with the timing of initial breakdown of ice sheets feeding the Missouri River. Braided morphology is inferred from analogy with other major glacier draining river valley (Mississippi and Ohio) morphologies during this time period (Rittenour et al. , 2007 and Counts et al., 2015). However, the true morphology is masked by a thick cover and cannot be confirmed. Dating of the Carrolton surface gives a range of 16.8 to 14.1 ka BP. It is likely this surface was long lasting (several thousand years) thus allowing the extensive lateral movement recorded throughout the valley. The next surface (Salix) records a second aggradation event of the Missouri River, in which the profile aggraded to elevations near the modern river profile. River morphology of this surface is known to be meandering as the morphology “ghosts” through the thin backswamp cover. Only fragments of this surface are recorded owing to extensive erosion during later incision and lateral growth of the Vermillion surface in the early Holocene. The Vermillion surface (11-10 ka BP) was dated in two locations and is incised into the Carrolton surface. Incision generated by this event was more than likely much narrower than the full valley, allowing for quick incision (12-10 ka BP) back to the grade of the Carrolton surface without compete removal of either the Carrolton or the Salix deposits.

The final aggradation phase of the Missouri River Valley occurred between 10 and 8 ka BP., The Missouri River has migrated laterally extensively over the past 8 ka with only minor elevation change. Channel belt sand deposits are highly amalgamated with only minor floodplain mud preservation in boreholes drilled through the Holocene channel belt down to the Vermillion surface. This likely reflects that aggradation between the Vermillion and Holocene surfaces were constrained to a narrow valley. This is also consistent with the preservation of deposits from the earlier Carrolton and Salix surfaces. The interpretation of narrow valley is also consistent with the rapid aggradation from Vermillion to modern elevations as a narrower valley would require less sediment to fill. The Vermillion surface is defined in two small windows where the valley filled with floodbaisn mud.

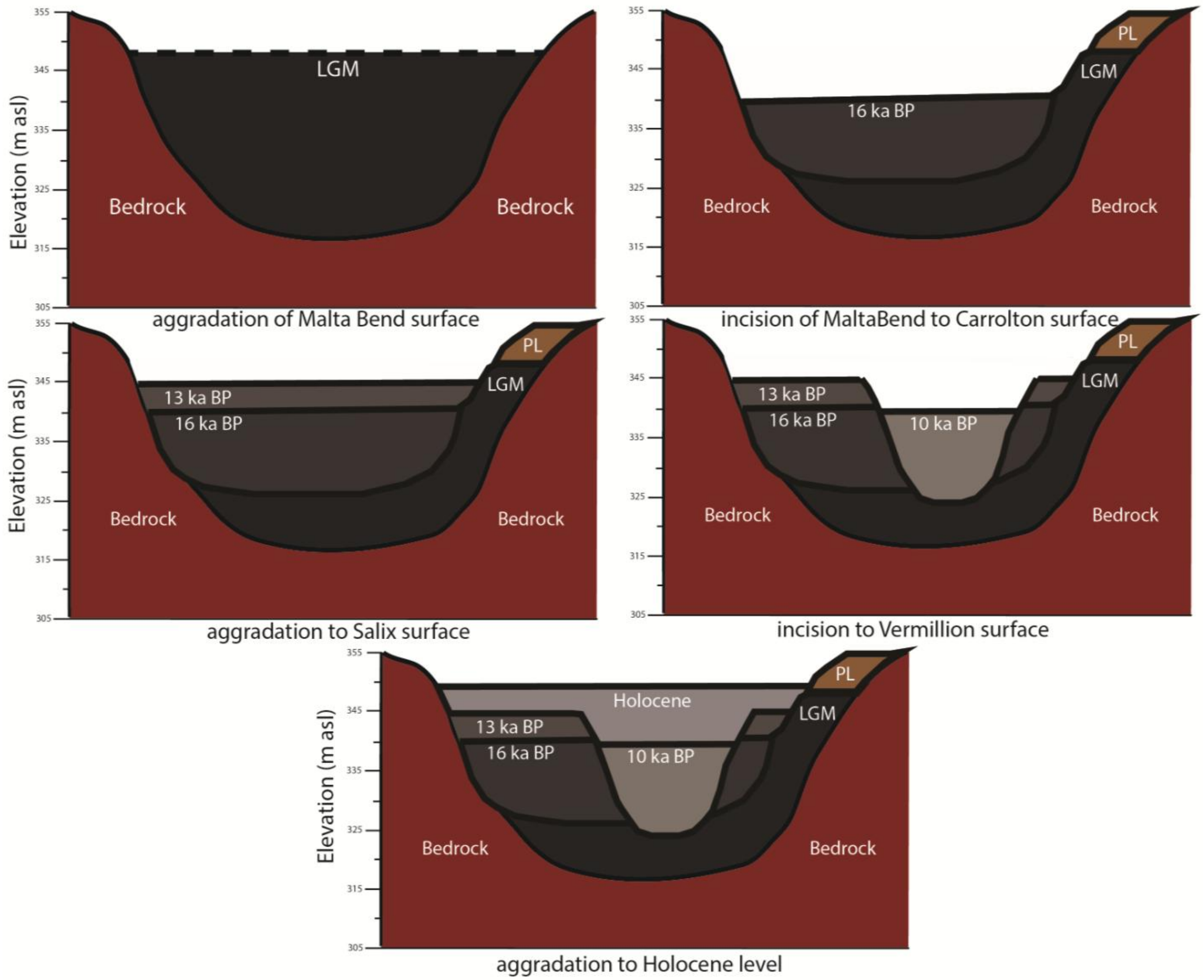


Figure 4.1 – Interpreted vertical movement of river profiles in the Vermillion area.

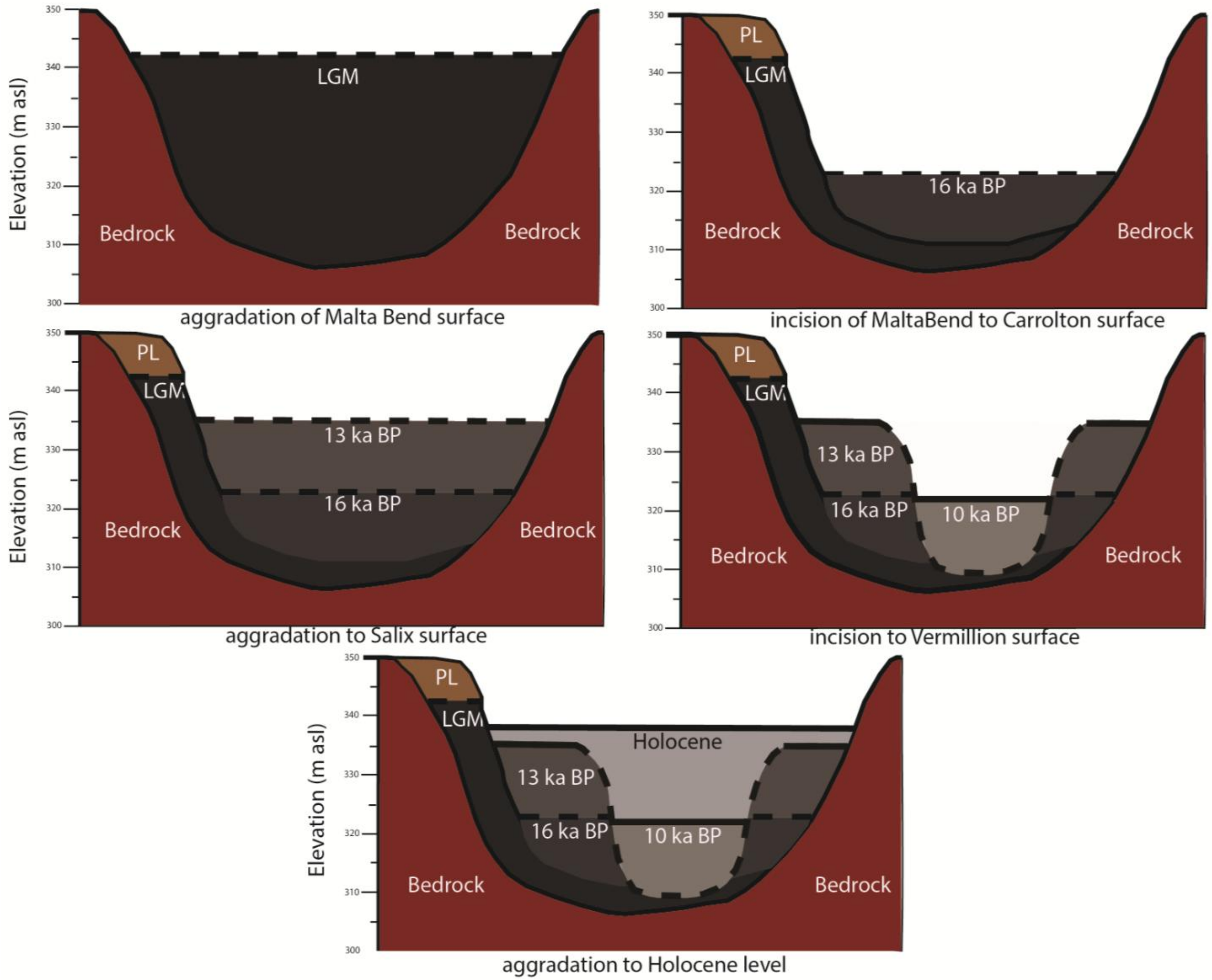


Figure 4.2 Interpreted vertical movement of river profiles in the Elk Point area.

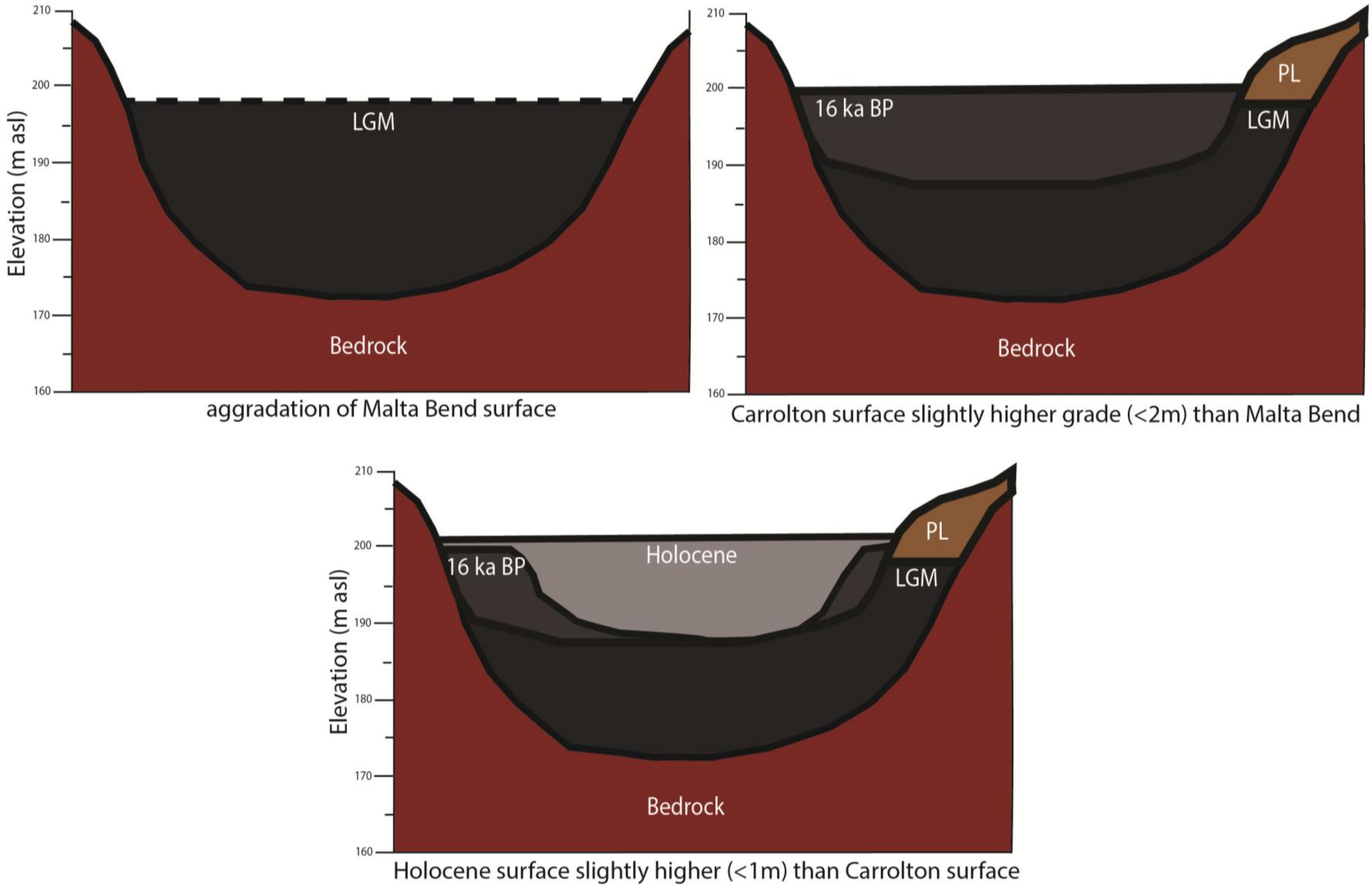


Figure 4.3 Interpreted vertical movement of river profiles in the Malta Bend area.

Water well data facilitates tracing of the five surfaces recognized in auger data throughout the Missouri River Valley (Fig 3.10), and reveals a significant difference of river grades longitudinally. In the Vermillion area, all five surfaces are recorded but show a relatively small difference in overall grade of all surfaces when compared to the Elk Point area farther downstream. In the Vermillion area, the largest difference in profiles is between the Holocene surface (8ka BP – present) and the Vermillion surface (10 ka BP) which is only a difference of around 8 meters. In the Elk Point area, however, the grade of the Malta Terrace and the Carrolton/Vermillion surface are separated by nearly 19 meters. This large difference in profiles wedges to a nearly uniform profile around the Malta Bend area of Missouri. This longitudinal shift in vertical profiles can be accounted by climate variations in accordance to the "buffers and buttresses" model for river profiles (c.f. Holbrook et al., 2006 and Holbrook & Bhattacharya, 2012). Upstream controls impact the overall river balance in transport capacity and sediment flux which can trigger phases of incision and aggradation in river valleys. The highest and lowest potential profiles from these sources over the valley duration of interest define the upper and lower "buffer profiles," The buffer profiles confine the highest and lowest profiles ultimately achieved over this time interval and the resulting preservation space for buffer-related valley fill (Holbrook et al., 2006). The overall preservation space for the time interval between the LGM and the present is the difference between the elevations of the Malta and the Carrolton/Vermillion surfaces. The Elk Point area records the largest swing in profiles and is therefore the area where upstream controls had the largest impact on profile variation.

The difference in elevation for vertical profiles of the Missouri River thins downstream, and profiles merge to within a few meters between Kansas City and Columbia, Missouri. Merging can be explained by either the downstream impact of upstream-controls, or by a possibile base-level buttress. The Missouri River Valley narrows significantly from a wide ($\geq$11km) alluvial valley to a narrow (<5km) bedrock valley abruptly down dip of the type area for the Malta surface at Glasgow, MO. Narrowing of the valley here is also accompanied by significant shallowing of the valley complex. South of Glasgow, MO, bedrock depth from water wells are reported anywhere from 30-70 feet below the modern surface compared to 105-130 feet below the surface in the Malta Bend area and areas up dip. Significant shallowing of valley depth suggests the river was unable to scour into bedrock in this southern area. Bedrock could have anchored the river and caused a buttress effect (c.f., Mankin, 1948), locking the basal river scour profiles to the bedrock elevation. Towards the north, no buttress effect is apparent and upstream controls appear to dominate, allowing for larger elevation swings in fluvial surfaces throughout the Pleistocene.

It is unlikely that the elevation of the trunk Mississippi River acted as a buttress for the tributary Missouri River in the Glascow, MO area. First, the Glascow location is ≈315 valley km and 6 backwater lengths up dip of the Mississippi confluence. This is considerably more than the distances of roughly one backwater length where downstream changes in base-level buttress elevation commonly impacts upstream profile elevation (Reviewed in Blum, et al., 2014). In addition, elevation changes of the Mississippi River and the Missouri River are out of phase. A high terrace known as the Savanna Terrace is reported by Flock (1983) that indicates aggradation between 16.5-15.6 ka BP of the Mississippi River at the Missouri River confluence location (Rittenour et al. 2007). The Missouri River is stable at and beyond this time and forming the Carrolton surface.

Valley-long variations in profile are most likely related to swings in glacially driven transport capacity and sediment flux. Interestingly, this “buffer capacity” is not at its maximum in the northern most area of this study area, but instead narrows in the Vermillion area which suggests there was a delayed effect of upswings which were not maximized until the river reached the Elk Point area. This can most easily be explained by the location of the James Lobe throughout the Pleistocene (the most proximal source of glacially derived sediment and water) (Figs. 1.2a-e), which at times of glacial advances covered the Vermillion area, and spilled directly into the Elk Point area thus allowing for maximum incisions to occur in Elk Point. The fixing of the elevation for this sediment source at the elevation of the James Lobe outlet likely anchored the elevation of profiles during the glacial phase.

IV.2 Climate vs Glacial Isostatic Adjustment

Cycles of aggradation and incision evidenced through mapping invite speculation as to their cause. Two hypotheses for profile variations are proposed. First, isotatic adjustment in ground elevation could potentially account for slope variations that drove incision and aggradation as the continent was uplifted and subsided beneath the river profiles. This would argue for a denudation origin for river incision. Alternatively, glacially driven changes in sediment and water input could account for profile shifts and support a buffer model for valley incision and fill. Glacially driven shifts in sediment and water flux appears to be the more likely cause for profile shifts.

The most recent model of isostatic adjustment (ICE-6G(VM5a)) is referenced to estimate likely changes in topography through time as a result of glacial loading and unloading (Peltier et al., 2015 and Argus et al., 2014). Proposed uplift and subsidence rates are contoured over the length of the Missouri drainage by subtracting projected local elevations on models between time intervals of interest (Figures 1.2a-e). Modeling of topography and associated uplift rates suggests the record of fluvial records of incision and aggradation is not consistent with timing, direction, and magnitude of glacial isostatic rebound.

During the LGM, models suggest subsidence rates in the northern section are on the order of 1-2 meters/ka while the southern area of this study is uplifting at this time between 1-2 meters/ka (Peltier et al., 2015 and Argus et al., 2014). This time period is when the Malta Terrace was deposited which is a period of stability at and above 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 not experiencing these magnitudes of differential uplift and subsidence. The drainage-wide incisional event of the Carrolton surface occurred while the model suggests slight subsidence on the order of 0-3 meters/ka (Fig1.2b). GIA subsidence should have promoted river valley aggradation which does not fit with the timing or direction of incision to the Carrolton surface level. Between 16 ka BP to 13.5 ka BP, the model suggests this study area is experiencing strong uplift rates on the order of 10-30 m/ka, which should have favored incision. However, this is the time period in which the Carrolton surface was aggrading to the level of the Salix surface. Fluvial aggradation and incision is therefore not consistent with modeled uplift rates which show uplift during aggradation and incision during subsidence, which goes against previously proposed mechanisms of tectonic effects on valley profiles (Holbrook and Bhattacharya, 2012).

While the tectonic signal generated by GIA was likely an important factor in river incision and aggradation, it apparently was not the controlling factor, and was overwhelmed by other signals. The more likely control on profile was the climatic signal driven by glacial processes. The GIA valley tilting however likely did cause lateral movement of rivers at grade, which is addressed further in section 4.5.

Aggradation during the LGM is recorded by the Malta Terrace. At this time, the glacial front is very proximal to the Missouri River in most locations, and even damns the river in several locations (Hill et al., 2002, Colton et al.1961). Ablation rates during the time of the deposition of the Malta Terrace would have been slow, and therefore could have contributed to low water inputs relative to sediment and triggered the aggradation of the Malta Terrace. This appears to also be the case in the Ohio River Valley which also experienced aggradation at this time (T4 surface; Counts et al., 2015). Incision to the Carrolton surface occurred between 26.4 ka BP and 16.8 ka BP. This is consistent with the timing of accelerated collapse of the Laurentide Ice Sheet (Dyke et. al. 2004). In addition, meltwater was rerouted during this time and the ice saddle of the Laurentide and Cordilleran Ice Sheets were sent directly down the James Lobe. (Catto et al., 1996; Ross et al., 2009) Additionally, during this time, the James Lobe disappeared between 15.5-15 ka BP, readvanced between 15-14.5 ka BP, and disappeared again by 14 ka BP, providing large amounts of melt-water input directly into the study area (Dyke, 2004 and Lundstrom, 2009). This incision is therefore consistent with the timing of major amounts of outwash pumping through the valley. Additionally, there is evidence of incision in both the Mississippi (Kennet braid belt; Rittenour et al., 2007) and the Ohio River Valley (T3; Counts et al., 2015) during this same time, suggesting this fluvial incisional response to initial glacial collapse may be a phenomenon throughout North America. Aggradation up to the Salix surface occurred between 14 and 13.4 ka BP during the end of the Bølling-Allerød warm period. This timing of aggradation is consistent with the formation of Glacial Lake Agassiz, and retreat of the LIS beyond the Missouri Escarpment which caused rerouting of glacial melt 13.9 ka BP into the Minnesota and Mississippi River drainage systems. Rerouting of glacial outwash out of the Missouri River into Glacial Lake Agassiz would have dramatically decreased water input into the study area and is consistent with the aggradation of the Salix surface as well as the change in morphology from a presumed braided outwash to the Salix meandering channels. Later incision (12-10 ka BP) of the valley (Vermillion) cannot be associated with glacial drainage because the Missouri River was no longer carrying outwash at this time due to rerouting into Glacial Lake Agassiz at 13.9 ka BP. However, the valley would have still been responding to climate change in the headwaters of the Rocky Mountains and potentially carrying outwash from retreating Rocky Mountain glaciers. Changes in this headwater effect may have had a later response than that of the Laurentide Ice Sheet.

IV.3 Valley Tilting

Results of OSL dating and surficial mapping show a preferred migration direction over the past 8,000 years (Figures 3.11-3.14). In several sections of this study area, the preferred direction is perpendicular to the glacial front of the Last Glacial Maximum. From Yankton to Sioux City (Fig 3.11), the river has migrated from the northeast to the southwest side of the valley over the past 8,000 years, and is currently hugging the valley wall on the southwest side. This northeast to southwest migration trend is perpendicular to the local NW-SE trend of the LGM glacial front (Figures 1.2a and 3.1-3.5). Additionally, in the southern area where the valley trend is nearly west to east, between Kansas City and Malta Bend, the river has similarly migrated north to south over the past 8000yrs, which is also perpendicular to the local glacial front. Between Sioux City to Omaha (Fig 3.12), however, where the valley orientation is north-south, and the river is perpendicular to the LGM glacial front, no preferential migration of Holocene channels is observed except for in the southern portion where some preferential direction is seen towards the southwest.

The strong link between channel migration over the Holocene and orientation of the LGM glacial front argues that migration of Holocene channels is caused by valley tilting throughout the Holocene due to glacial isostatic adjustment. Lateral tilting of valleys is shown to affect migration of rivers over time toward the down-tilted side from theoretical computer modeling (Bridge and Leeder 1979), field studies (Nanson 1980, Leeder and Alexander 1987), and flume experiments (Peakall, 1995). Peakall (1995) showed that valley tilting between the order of 7.5 x 10^{-4} rad/1000 yr and 4.6 x 10^{-7} rad/1000 yr can cause lateral migration of channels. Topographic differences from the (ICE-6G(VM5a)) GIA model indicates valley tilting over the last 8,000 years on the orders between 1.7 x 10^{-6} and 3.8x10^{-6} rad/ka within this study area, consistent with experimental magnitudes (Fig 3.1). The directions, however, of valley tilt from the models are not completely consistent with observed river migration. In the northern area of the study, where the valley trends NW-SE and migration has occurred from NE-SW, the modeled direction and magnitude of valley tilt are modeled to be in a N-S direction and can explain lateral migration. In the central area from south of Sioux City, IA to Corning, NE, the direction of valley tilting is parallel to the valley orientation. The lack of preferred migration in this reach is consistent with the lack of preferred orientation of isotatic tilting throughout this area over the Holocene (Fig 3.1). Conversely, the southern section, between Kansas City and Columbia, has a valley tilt direction towards the north, which is the opposite direction of channel migration over the past 8 ka. If valley tilting is the cause of lateral migration in the northern and central sections of this study area, then it is possible the (ICE-6G(VM5a)) model is too large a scale to account for localized tilting to the south in the Kansas City to Columbia area, or that the forebulge was actually farther south throughout the Holocene than proposed by the (ICE-6G(VM5a)) and allowed for regional tilting to the south at lower latitudes. Alternatively, the lateral migration is working against valley tilting, which would go against proposed mechanisms of channel migration (Peakall 1995, Bridge and Leeder, 1979, reviewed in Holbrook and Schumm 1999, and Schumm, et al., 2000). It is of course also possible that valley tilting is not the mechanism which is driving Holocene migration. Alternative mechanisms include but are not limited to autocyclic processes and differential compaction, both of which have been shown to create preferential migration (Wells and Dorr 1987a, Mackey and Bridge, 1995). These more random alternatives however become difficult to infer againt such a consistent migration signal. The relationship between orientation of lateral migration during the Holocene suggest GIA has played a role in the migration of the Missouri River.

IV.4 Accomodation

External forcing, particularly base-level, is often argued to control valley architecture (Guccione, 1994; Posamentier and Vail, 1998; Blum and Tornqvist, 2000; Holbrook et al., 2006). Sequence-stratigraphic models particularly apply base-level control to predictive fluvial architecture (Shanley and McCabe, 1993; Wright and Marriot, 1993). These models propose sediment by-pass during low-stand, followed by amalgamation of fluvial channels during slow early base-level rise, and finishing with dispersed fluvial channels during transgression and high-stand. This predicted architecture is very similar to architecture within the Missouri River Valley. However, architecture of Missouri River fill is not driven by a change in a down-dip anchored base-level rise and fall. Instead, this architecture is driven by preferential preservation of amalgamated channel belts in the lower portions of the valley compared to the upper part of the valley over repeated cycles of aggradation and incision (Figure 4.4). In each example, the lower half of the valley has a high net-to-gross of channel-belt sediments compared with floodplain deposits of the upper half in which channel belts are isolated. OSL dating of surfaces within this area allowed for the understanding that the Missouri Valley architecture was created by multiple aggradation/incision events. (Fig 4.2 and 4.4c) The first major incision event (Carrolton) cleared most of the floodplain deposits of the previous LGM surface (Malta Alloformation). The Carrolton apparently was widespread throughout the lower portion of the valley. The Carrolton channels amalgamated is older amalgamated fluvial channel-belt deposits while removing older fluvial deposits of higher terraces that may potentially have been more mud rich. Rapid aggradation of rivers from the Carrolton surface to the Salix surface occurred between 15-13 ka BP and resulted in a profile rise of about 14 meters. High rates of aggradation compared with lateral migration allowed for significant amounts of floodplain deposits to be preserved through this time period. Later incision of the Salix surface by the Vermillion (12-10 ka BP) removed some of these backwater deposits. The valley of the Vermillion incision is assumed to be narrow, allowing for quick incision and aggradation back to modern levels around 8 ka BP. Final aggradation between 10-8 ka BP was quick and again allowed for more preservation of floodplain deposits due to aggradation outweighing lateral movement. The Missouri River channel belt has remained near the surface over the past 8 ka BP, and thus lateral movement throughout the valley has covered much of the very top layer of the valley with channel deposits and likely removed some of the older floodbasin deposits. The vertical space between the Holocene floodplain and the Vermillion surface, however, is dominated by floodplain fines just due to the nature of preservation during times of rapid aggradation.

The classic fluvial sequence stratigraphic model, which is commonly interpreted in the rock record, can be produced by multiple upstream-controlled incision/aggradation cycles with preferential preservation of lower channels and does not necessarily have to be generated by a one-cycle base-level rise driven by rise of a down-stream anchor like sea level. Miall (2014) interpreted a section of the Blackhawk Formation as being a product of differences in preservation potential within a buffer-valley system (Holbrook 2006), and are not necessarily due to a change in fluvial style brought on by down-stream anchored accommodation changes. The Missouri Valley provides a modern documentation of this alternative multi-cycle model of valley fill. This may be of economic importance in reservoir modeling because applying a classic fluvial sequence stratigraphic model to such a valley will underestimate the heterogeneity of the amalgamated channels at the base of the valley since these have been created by multiple scouring events as opposed to just one sequence.

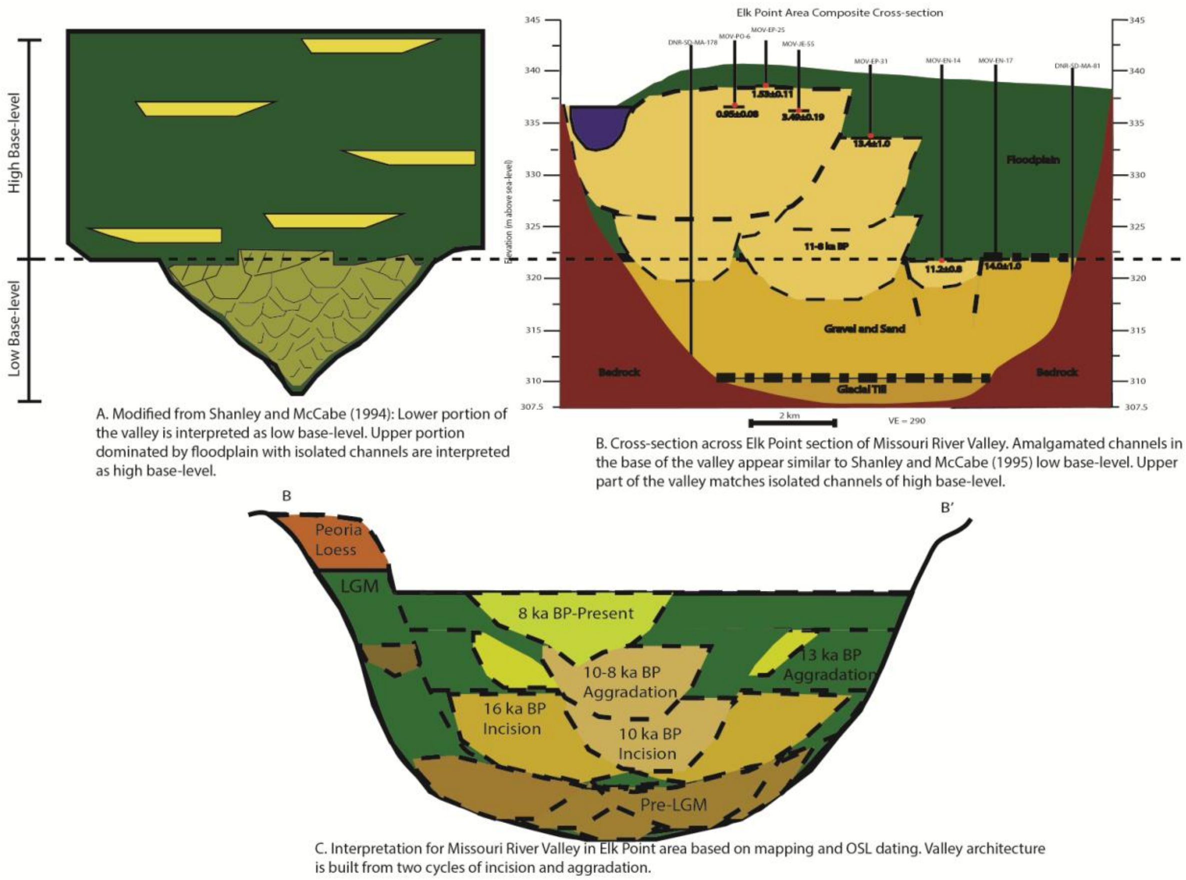


Figure 4.4 – Classic sequence stratigraphic fluvial architecture shown in 4.4a (modified from Shanley and McCabe, 1993). Figures A and B show similar amalgamation of channels in the lower portion of the valley and high preservation of floodplain deposits in upper portion of the valley. Figure 4.4c shows the timing of the surfaces which created this architecture.

Chapter V: Conclusions

1. Wide swings in aggradation and incision are recorded from the LGM through the early Holocene within the Missouri River Valley. Larger amounts of incision and aggradation are seen in the northern portion of this study area and taper to the south where recorded surfaces are all near the same grade. This is evidence for a larger

"buffer capacity" in the northern section of the Missouri River around Elk Point, SD which tapers towards a buttress towards Columbia, MO.

2. Aggradation near modern floodplain levels are seen during the LGM. Similar surfaces were noted in the Ohio and Mississippi River Valleys. Incision of the Carrolton surface is coincidental with breakdown of the Laurentide and Cordilleran ice sheets which also coincides with surfaces mapped in the Ohio and Mississippi River Valleys. These events could be widespread responses to the building and collapse of ice sheets during the Late Wisconsin.
3. Aggradation and incision events are not consistent with modeled GIA with respect to location, timing, direction, or magnitude of rebound and subsidence. This study suggests that climatic factors were more than likely the largest contributing factors for incision and aggradation.
4. Evidence for valley tilting over the last 8 ka years are interpreted from movement of Holocene channels of the Missouri River Valley. Movement is partly in agreement with calculated measurements of valley tilting from GIA modeling, with the main exception being in the southern section of this study area between Kansas City, MO to Malta Bend, MO. This section shows strong evidence for movement perpendicular to Wisconsin glacial fronts, but models indicate valley tilting in the opposite direction. This is most likely localized tilting to the south which may be at a smaller scale than GIA models account for which is more likely than the alternative hypothesis in which channel belts are moving away from valley tilting which goes against reported mechanisms for channel movement.

5. Fluvial assemblages within the Missouri River Valley have a similar construction to what is often associated with changes in accommodation caused by buttress movement. These particular assemblages are instead created by preferential preservation within the lower portion of the valley which appears to be “low accommodation” compared to the upper portion of the valley which is reworked during aggradation and incision.

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ABSTRACT

Fluvial History of the Missouri River Valley from Late Pleistocene to Present: Climatic vs. Tectonic forcing on Valley Architecture Construction

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Results from this study provide evidence for two cycles of aggradation and incision within the Missouri River Valley since the Last Glacial Maximum. Timing of these cycles were compared with drainage histories from deglaciation and glacial isostatic adjustment (GIA) from glacial loading and unloading. Comparison of the timing and magnitude of these cycles to modeled GIA rates and locations suggest these cycles were not caused by tectonic movement from GIA. Instead, incision and aggradation events fit better with glacial drainage patterns from the Laurentide ice sheet. This study also indicates that aggradation and incision events at the end of the LGM to 16 ka BP maybe a North American phenomenon as supported by similar responses in the Ohio and Mississippi River Valleys. While GIA does not support incisions and aggradation, this study does provide evidence for valley tilting affecting lateral migration of Missouri River Channel belts over the last 8 ka. A longitudinal profile created from Yankton, SD to Columbia, MO indicates the river valley has a higher “buffer” capacity around Elk Point, SD which thins to the north and the south. A possible “buttress” effect is indicated by very small vertical movement of river profiles in the Kansas City to Columbia, MO area. This buttress effect is most likely due to an inability for the Missouri River to incise into the narrow and shallow bedrock valley which develops around Arrow Rock, MO. Valley architectures created within the Missouri River Valley have a similar distribution of channels and floodplain fines as

models previously tied to down-stream anchored base-level controls. Findings within this study provide evidence for this architecture being built by preferential preservation within a climatically driven “buffer” system by multiple aggradation and incision events.