Rapid and widespread response of the Lower Mississippi River to eustatic forcing during the last glacial-interglacial cycle

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ABSTRACT

The Lower Mississippi Valley provides an exceptional field example for studying the response of a continental-scale alluvial system to upstream and downstream forcing associated with the large, orbitally controlled glacial-interglacial cycles of the late Quaternary. However, the lack of a numerical chronology for the widespread Pleistocene strata assemblage known as the Prairie Complex, which borders the Holocene floodplain of the Lower Mississippi River, has so far precluded such an analysis. Here, we apply optically stimulated luminescence (OSL) dating, mainly on silt-sized quartz from Prairie Complex strata. In total, 27 OSL ages indicate that the Prairie Complex consists of multiple allostratigraphic units that formed mainly during marine isotope stages 7, 5e, and 5a. Thus, the aggradation of the Prairie Complex is strongly correlated with the sea-level highstands of the last two glacial-interglacial cycles. Fluvial incision during the sea-level fall associated with the MIS 5a–MIS 4 transition extended as far inland as ~600 km from the present-day shoreline, testifying to the dominant downstream control of fluvial stratigraphic architecture in the Lower Mississippi Valley. In addition, the short reaction time of the Lower Mississippi River suggests that large fluvial systems can respond much more rapidly to allogenic forcing than is commonly believed.

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INTRODUCTION

The response of continental-scale alluvial systems to sea-level and climate change has received wide interest (e.g., Schumm, 1993; Shanley and McCabe, 1994; Blum and Price, 1998; Blum and Törnqvist, 2000; Törnqvist et al., 2000, 2003; Goodbred, 2003; Amorosi and Colalongo, 2005; Busschers et al., 2007; Rittenour et al., 2007; Amorosi et al., 2008), not only to understand how large sediment-dispersal systems may behave under rapidly changing environmental conditions, especially in the coastal zone, but also to improve our capability to make predictions about the ancient rock record. Continental-margin fluvial systems are often found to be under the influence of both upstream (climate) and downstream (sea level) controls, in addition to tectonic subsidence (e.g., Blum and Price, 1998; Törnqvist et al., 2000). Fluvial stratigraphic architecture (sensu Blum and Törnqvist, 2000) is primarily a product of the interaction of these two external forcing mechanisms (reviews of the extensive literature on this subject are provided by Shanley and McCabe [1994] and Blum and Törnqvist [2000]).

Within this context, two key issues are rarely addressed by means of field-based research. The first issue concerns the relative role of sea-level and climate change in the development of a continental-scale fluvial system throughout a glacial-interglacial cycle. This question has proven challenging since the incomplete records of most fluvial systems typically do not capture a full glacial-interglacial cycle, and if they do, chronological control is often insufficient (Blum and Törnqvist, 2000).

The second issue relates to the rapidity of fluvial response to external forcing. This problem has been investigated mostly by means of theoretical studies (see, e.g., the review by Paola, 2000). Bull (1991) separated fluvial response time into two components: the reaction time, which represents the duration from the onset of the perturbation to the initial fluvial adjustment, and the relaxation time, which represents the duration from the initial adjustment to the final achievement of a new equilibrium. Paola et al. (1992) defined the fluvial equilibrium time (τ), comparable to the relaxation time of Bull (1991), and suggested that fluvial response depends on the relationship between τ and the periodicity of the external forcing (T). This concept was further developed by Swenson (2005) and Swenson and Muto (2007), who showed that the fluvial response can significantly lag behind external forcing when τ is comparable to or larger than T. According to this theory, most large fluvial systems would be relatively insensitive to late Quaternary allogenic forcing (Métivier and Gaudemer, 1999; Castelltort and Van Den Driessche, 2003), although Törnqvist (2007) pointed out that exceptions to this rule exist. Flume experiments that focused on the effect of base-level changes have produced mixed results. Koss et al. (1994) and van Heijst and Postma (2001) suggested that the initial fluvial adjustment could significantly lag behind sea-level change, while Heller et al. (2001) and Strong and Paola (2008) found an almost instantaneous fluvial response. Clearly, theoretical and experimental studies need to be compared with field-based investigations.

We address these two issues by an investigation of the late Quaternary strata of the Lower

GSA Bulletin; May/June 2012; v. 124; no. 5/6; p. 690–704; doi: 10.1130/B30449.1; 12 figures; Data Repository item 2012110.

Mississippi Valley. This setting provides a relatively well-described stratigraphy (e.g., Fisk, 1944; Autin et al., 1991; Saucier, 1994), along with high-resolution digital elevation data. Most importantly, the Lower Mississippi Valley provides a rare—perhaps the only—continental-scale fluvial sedimentary record near a land surface that potentially spans the entire last glacial-interglacial cycle. Therefore, it offers an exceptional field example to study the relative role of upstream versus downstream controls over a 10^5 yr time scale.

We investigated the surface geomorphology, stratigraphic architecture, and geochronology of the late Quaternary fluvial strata of the Prairie Complex (cf. Autin et al., 1991) in the Lower Mississippi Valley with high-resolution digital elevation data, sediment cores, and quartz optically stimulated luminescence (OSL) dating. Episodes of fluvial aggradation and incision on the order of tens of meters were constrained by OSL ages and compared with independent late Quaternary proxy records to address the questions outlined here.

STUDY AREA AND RESEARCH HISTORY

The Lower Mississippi Valley has attracted several generations of researchers with an interest in fluvial morphodynamics and sedimentary geology (e.g., Fisk, 1944; Fisk and McFarlan, 1955; Autin et al., 1991; Saucier, 1994; Blum et al., 2000; Rittenour et al., 2007). The late Quaternary record of the Lower Mississippi Valley includes a series of braid belts that formed during the last glacial (Rittenour et al., 2007) plus an older, but chronologically poorly constrained unit known as the Prairie Complex, which occupies large portions of the valley margins (Autin et al., 1991; Saucier, 1994). Saucier (1994) synthesized previous work (Fisk, 1938, 1939, 1940; Saucier, 1967; Smith and Saucier, 1971) and mapped the majority of the Prairie Complex sediments as floodplain and coastal-plain deposits (Fig. 1). The Prairie Complex in Louisiana was recently renamed as the Prairie Allogroup and subdivided into lower-rank allostratigraphic units (e.g., Heinrich, 2006). However, since similar mapping does not exist for areas farther north, we adopt the Prairie Complex nomenclature here, following Autin et al. (1991).

There is general agreement that the Prairie Complex consists of multiple allostratigraphic units and was formed by a meandering fluvial system similar to the Holocene Lower Mississippi River, at a sea level considerably higher than the lowstand of the Last Glacial Maximum (Autin et al., 1991; Saucier, 1994; Autin, 1996; Autin and Aslan, 2001). The present understanding of the Prairie Complex chronostratigraphy is mostly inferred from stratigraphic correlation and a limited number of ^{14}C ages. Loess stratigraphy has been used to assign ages to allostratigraphic units in the northern Lower Mississippi Valley (e.g., Rutledge et al., 1996; Blum et al., 2000), but many loess sheets pinch out toward the south (Autin et al., 1991). The Prairie Complex is veneered by Peoria Loess, which constrains its age to older than 25 ka (Bettis et al., 2003).

H.N. Fisk and his associates (e.g., Fisk, 1938, 1944, 1951; Fisk and McFarlan, 1955; Bernard and LeBlanc, 1965) assigned a mid-Wisconsinan age to the Prairie Complex, corresponding approximately to marine isotope stage (MIS) 3. Most early ^{14}C measurements of Prairie Complex deposits, particularly those on well-

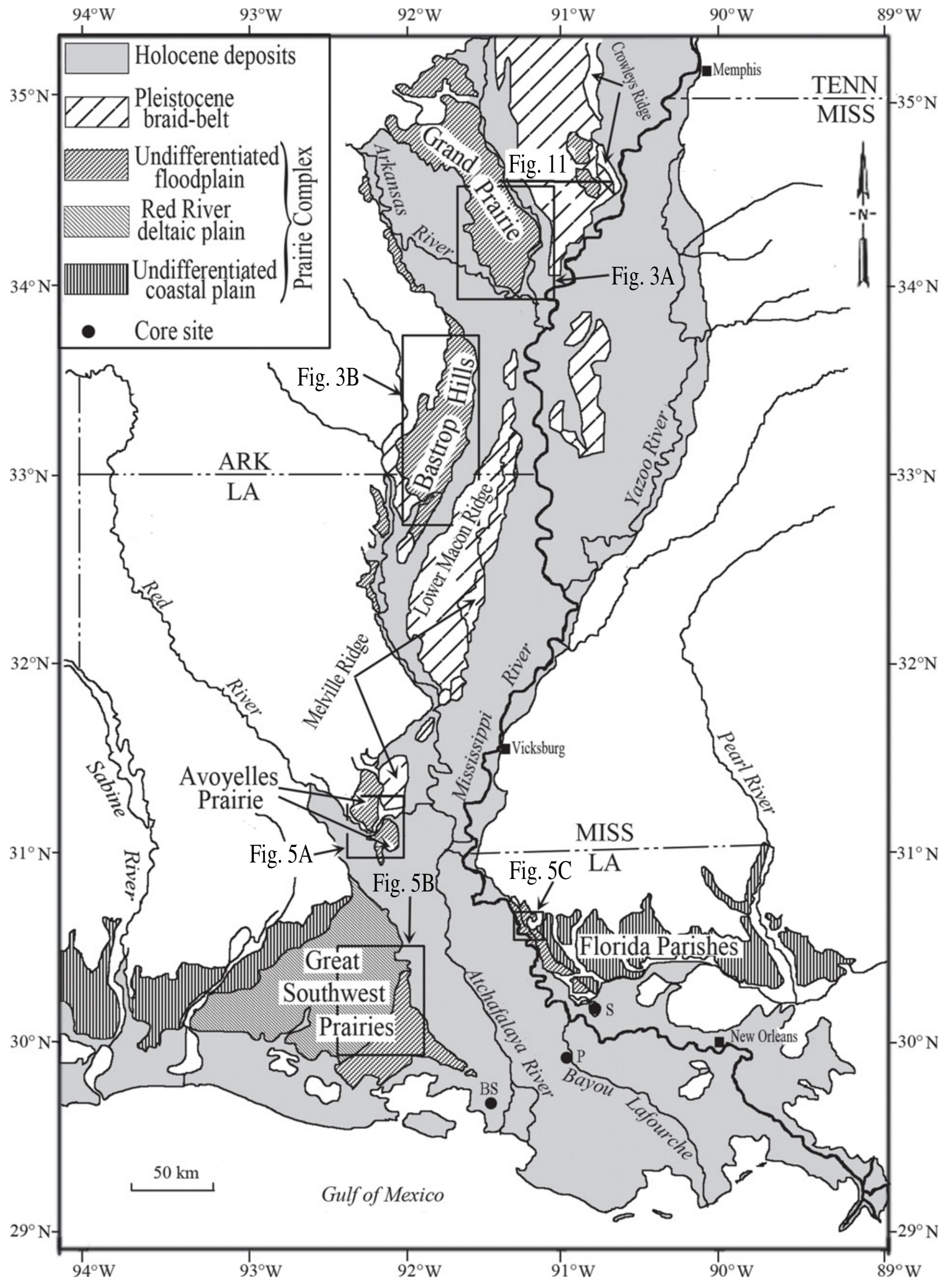


Figure 1. Near-surface occurrence of Holocene deposits, Pleistocene braid-belt deposits, and the Prairie Complex in the Lower Mississippi Valley and coastal Louisiana (modified from Saucier, 1994). The filled circles are core sites (S—Sorrento I, P—Paincourtville I, and BS—Bayou Sale IV). The locations of other core sites with optically stimulated luminescence (OSL) data are shown in Figures 3 and 5.

identified materials, yielded apparently infinite ages of >30 ka (McFarlan, 1961). Saucier (1968, 1974, 1981) argued that the Prairie Complex was probably deposited prior to the valley cutting of the last glacial, and inferred a Sangamonian age (MIS 5e). Autin et al. (1991) proposed a mid-Wisconsinan age for the youngest allostratigraphic unit of the Prairie Complex based mostly on the correlation of paleosols and tentative correlations with a few ^{14}C dated terraces (Alford et al., 1985; Autin et al., 1988). They assumed that the oldest allostratigraphic unit is of Sangamonian age. Saucier (1994) postulated that most of the Prairie Complex strata were formed during MIS 5. More recent analysis of paleosols in the Avoyelles Prairie region (Fig. 1) by Autin and Aslan (2001) inferred that the youngest allostratigraphic unit of the Prairie Complex was formed during MIS 3. Mange and Otvos (2005) and Otvos (2005) dated the coast-parallel Prairie Complex sediments using thermoluminescence and OSL, yielding ages of MIS 5 to MIS 3. Unfortunately, they provided no details regarding either the dating procedure or the stratigraphic context of dated samples. Heinrich (2006) suspected that the results of Otvos (2005) may have been affected by bioturbation. In conclusion, currently existing age estimates for the Prairie Complex cover a wide time span and largely preclude an understanding of the main driving forces of the evolution of the Lower Mississippi River during the last glacial-interglacial cycle.

Allogenic forcing of the Lower Mississippi Valley during the last glacial-interglacial cycle has been profound: A large glacio-eustatic sea-level signal in the Mississippi Delta has affected its downstream reaches, whereas large and sometimes catastrophic fluctuations of upstream sediment supply and water discharge have occurred, particularly during deglaciation (e.g., Knox, 1996; Brown and Kennett, 1998; Licciardi et al., 1999; Teller et al., 2002). There are two contrasting schools of thought about the predominant controls of Lower Mississippi Valley evolution. The classic, Fiskian view (e.g., Fisk, 1944), which has profoundly influenced sequence-stratigraphic models, assumes that sea-level change alone drives large-scale fluvial processes throughout the Lower Mississippi Valley. This view has been subsequently challenged by workers who noted that sediment and water discharge variability, primarily due to climate change, leaves a conspicuous imprint throughout much of the Lower Mississippi Valley (Saucier, 1994; Blum et al., 2000; Rittenour et al., 2007). Under this latter scenario, sea-level influence might have been limited to the Mississippi Delta only (Saucier, 1994, 1996).

The chronology of late Quaternary fluvial strata holds the key to resolving this controversy. The advent of OSL dating, notably by means of the single-aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000, 2003; Wallinga, 2002; Wintle and Murray, 2006; Rittenour, 2008) provides new opportunities to tackle this problem. The Lower Mississippi Valley braid belts were successfully OSL dated to the last glacial by Rittenour et al. (2005), providing the basis to associate braid-belt formation with glacial advance and retreat (Rittenour et al., 2007). Despite this important progress, the chronology of the Prairie Complex constitutes the critical remaining element to unlock the evolution of the Lower Mississippi Valley during an entire glacial-interglacial cycle.

METHODS

Field Sampling

Sediment cores from the Prairie Complex were collected with a Giddings soil probe and a Geoprobe. The Prairie Complex was divided into five geographic regions (Fig. 1) in accordance with previous work (Autin et al., 1991; Saucier, 1994; Autin, 1996). All the allostratigraphic units described by Autin et al. (1991) and Autin (1996) were sampled. In addition, late Holocene overbank deposits were sampled in the Mississippi Delta (Fig. 1) with an Edelman auger and a 1-m-long gouge. Sediment description and subsampling occurred in the field after extruding cores from the Giddings soil probe or Geoprobe. In total, 53 cores, 2.4–17.7 m in length, were taken for this study (see Table DR1 for the location of dated cores[1]). Soil texture following the U.S. Department of Agriculture (USDA) classification, sedimentary structures, Munsell colors, and weathering features were logged in the field (Mateo, 2005). Stratigraphic interpretation of the core logs followed the Lower Mississippi Valley loess distribution model described in Autin et al. (1991) and USDA soil survey data for loess (Soil Survey Staff, 2002); for fluvial deposits, data from Autin et al. (1991), Saucier (1994), Autin (1996), and high-resolution digital elevation data were consulted. We used extruded cores for OSL sampling. OSL samples were taken from core segments without surface fractures, and they were wrapped with two layers of aluminum foil to avoid further light exposure, plus an extra layer of plastic film to minimize the loss of pore water during transport and storage.

[1]GSA Data Repository item 2012110, core details, OSL dating results, sedimentary logs, and evidence of bioturbation and Mississippi River discharge in the last 10 yr, is available at http://www.geosociety.org/pubs/ft2011.htm or by request to editing@geosociety.org.

OSL Dating

The OSL samples (Table DR2 [see footnote 1]) were processed in the laboratory under subdued yellow-red light. The outer rim of the core samples (1–2 cm in thickness) was sliced off and used for water content and natural radioactivity measurements. The remaining material was treated following conventional procedures for sample preparation (Mauz et al., 2002) to obtain quartz grains in the size ranges of 4–11 µm, 11–15 µm, 100–200 µm, and 150–250 µm (Table DR2 [see footnote 1]). Fine silt-size quartz (~2 mg) pipetted onto 10-mm-diameter aluminum disks and sand-size quartz mounted on an ~2-mm-diameter area on stainless-steel disks were measured with either a Risø DA-15 automatic luminescence reader equipped with 41 blue light-emitting diodes (LEDs) or a Risø DA-15 B/C equipped with 21 blue LEDs (470 ± 30 nm in both cases) for stimulation. Infrared stimulation was performed with a laser diode emitting at 830 ± 10 nm. The luminescence emissions were detected through an optical filter (Hoya U340, 7.5 mm) transmitting at 320–390 nm.

A standard SAR protocol (Murray and Wintle, 2000) was used to obtain equivalent doses (D_e) for each sample. Preheating tests (Murray and Wintle, 2000) validated the suitability of a preheat treatment at 220 °C for 10 s (Fig. 2A). The reliability of the SAR protocol was examined with a dose recovery test (Murray and Wintle, 2003). A mean dose recovery ratio of 1.00 ± 0.06 (1σ, $n = 37$) was obtained. The postinfrared (IR) OSL ratio (Duller, 2003) was used to check for feldspar contamination. The dose responses of the samples are best described with the sum of an exponential and a linear function (Fig. 2B) and show excellent intra-aliquot reproducibility (Shen and Mauz, 2011). They show no evidence of OSL saturation up to a 500 Gy laboratory dose (cf. Rittenour et al., 2005; Murray et al., 2008). OSL component-resolved analysis was used to show that the fast component dominates the OSL signal in the initial part of the decay curve (Shen and Mauz, 2009).

The natural radioactivity of the samples was measured with a high-resolution, low-level gamma-spectrometer (Table DR2 [see footnote 1]). Several samples (both Holocene and from the Prairie Complex) show a deficit of ^{226}Ra over ^{234}Th in the ^{238}U decay chain. It is assumed that this secular disequilibrium is caused by the loss of ^{226}Ra and has prevailed since sediment deposition. Therefore, the dose rate contribution from the ^{238}U decay chain of these samples is divided into two portions. One portion is represented by the short-lived

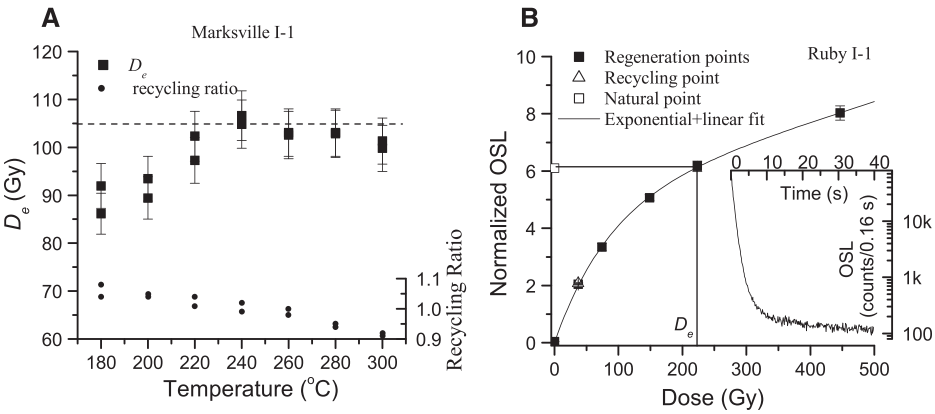


Figure 2. (A) Example of a preheating test of Prairie Complex fine silt–size quartz, showing an equivalent dose (D_e) plateau between 220 and 300 °C. The dashed line marks the D_e of this sample. The recycling ratio of the preheating test is shown at the bottom. (B) Example of the dose response of Prairie Complex fine silt quartz. The dose response is best described by the sum of an exponential function and a linear function. The inset is an optically stimulated luminescence (OSL) decay curve of the natural signal for the same sample.

radioisotopes that are in secular equilibrium with ^{226}Ra measured by ^{214}Pb and ^{214}Bi. The other portion is represented by the long-lived radioisotopes measured by ^{234}Th in excess of ^{226}Ra. It is further assumed that the excess did not change significantly during burial. The natural dose rate was calculated according to Adamiec and Aitken (1998) and by applying a quartz *a*-value of 0.03 (Mauz et al., 2006) and an alpha attenuation factor of 31% for 11–15 µm quartz grains (Bell, 1980). For the sand-sized quartz, an internal dose rate of 0.03 ± 0.02 Gy/k.y. was used (Grün and Fenton, 1990; Vandenberghe et al., 2008). Beta attenuation factors were taken from Mejdahl (1979). The contribution of cosmic radiation was calculated following Prescott and Hutton (1994). For samples that show secular disequilibrium in the ^{238}U decay chain, the calculated dose rates are 4%–12% lower than those calculated by using the ^{234}Th activity and 1%–5% higher than those calculated by using the ^{226}Ra activity if secular equilibrium is assumed (cf. Olley et al., 1996; Guibert et al., 2009).

The water content of the Prairie Complex sediments has changed over time (Autin and Aslan, 2001). It is assumed that water content has decreased from an initial value comparable to that of the Holocene sediments to the measured value. To account for this variation, the water content used for age determination for the Prairie Complex sediments is the average water content (0.23 ± 0.10) of the Prairie Complex sediments (0.20 ± 0.07, 2σ, 21 measurements) and their Holocene counterparts (0.26 ± 0.07, 2σ, 9 measurements, 3 of which are reported in Table D2 [see footnote 1]). For the Holocene samples, the measured water content was used.

The D_e values, the natural dose rates, and the OSL ages for all samples are listed in Table DR2 (see footnote 1). The aliquots used for D_e calculation all show (1) no feldspar contamination, (2) recycling ratios within 1 ± 0.1, and (3) recuperation < 0.05 (Murray and Wintle, 2003). The SAR protocol measurement commonly produces multiple D_e estimates from a number of subsamples (also called aliquots) for each sample. To obtain a D_e for each sample for age calculation, three statistical approaches, an arithmetic mean, a minimum age model (MAM), and a central age model (CAM; Galbraith et al., 1999), were considered to analyze the associated D_e distribution. The MAM is designed for samples affected by insufficient OSL signal resetting prior to deposition that are measured with sand-size quartz, while the other two are used for sufficiently bleached samples. For samples measured with fine silt-size quartz (4–11 µm and 11–15 µm), the arithmetic mean D_e and CAM D_e of accepted aliquots are statistically identical, and the former was used for age calculations. For samples measured with sand-size quartz (100–200 µm and 150–250 µm), the MAM and the CAM were considered to take the error of individual D_e into account. The statistical procedure described by Arnold et al. (2007) was used to select between the CAM and the MAM, resulting in the MAM for samples Marksville I-3 and Paincourtville I-4 and I-5 and the CAM for samples Mt. Pleasant I-1 and St. Landry I-1.

PRAIRIE COMPLEX GEOMORPHOLOGY, STRATIGRAPHY, AND OSL CHRONOLOGY

Grand Prairie

The topography of the southern part of the Grand Prairie region (Fig. 3A) shows well-preserved meander belts, similar in size to the modern Arkansas River. This surface has been interpreted as a relict Arkansas River alluvial fan (Saucier, 1967, 1994). Core data (Fig. 4A; Fig. DR1A [see footnote 1]) reveal that the Grand Prairie is covered by several meters of loess. Yellowish-red (2.5–7.5 YR in hue) Arkansas River deposits, 5–7 m in thickness, underlie the loess. The Arkansas River deposits are in turn underlain by yellowish (2.5 Y in hue) Mississippi River deposits (Fig. DR1A [see footnote 1]; Saucier, 1994). These two units are separated by a weakly developed paleosol.

The Arkansas River deposits have an OSL age of 77 ± 8 ka (Indiana Spur I-1), while the Mississippi River deposits provide two mutually consistent OSL ages of 85 ± 8 ka (Indiana Spur I-2) and 88 ± 10 ka (Van I-1).

Bastrop Hills

The topography in the Bastrop Hills region (Fig. 3B) is strikingly different from that in the Grand Prairie. The Bastrop Hills are mostly higher in elevation and heavily dissected. There are hardly any geomorphic features that reveal depositional environments of near-surface strata. Relative elevation and landscape dissection suggest that the Bastrop Hills may be older than the Grand Prairie, although the two regions exhibit a similar stratigraphy (Fig. 4B; Saucier, 1994).

Three OSL ages for the Arkansas River deposits are 118 ± 10 ka (Selma I-1), 140 ± 16 ka (Log Cabin I-1; Fig. DR1B [see footnote 1]), and 136 ± 15 ka (Twin Oaks I-1; Fig. DR1C [see footnote 1]). The stratigraphically deeper Mississippi River deposits are dated to 213 ± 11 ka (Selma I-2).

Avoyelles Prairie

Two terrace levels can be recognized on opposite sides of the Red River in the Avoyelles Prairie region (Fisk, 1940; Autin et al., 1991). The higher terrace northwest of the Red River has surface elevations at ~30 m above modern sea level, while the lower terrace to the southeast is at ~23 m. In turn, the lower terrace is ~10 m higher than the adjacent Holocene floodplain. The size of the well-preserved meander loops on the lower terrace suggests that it was

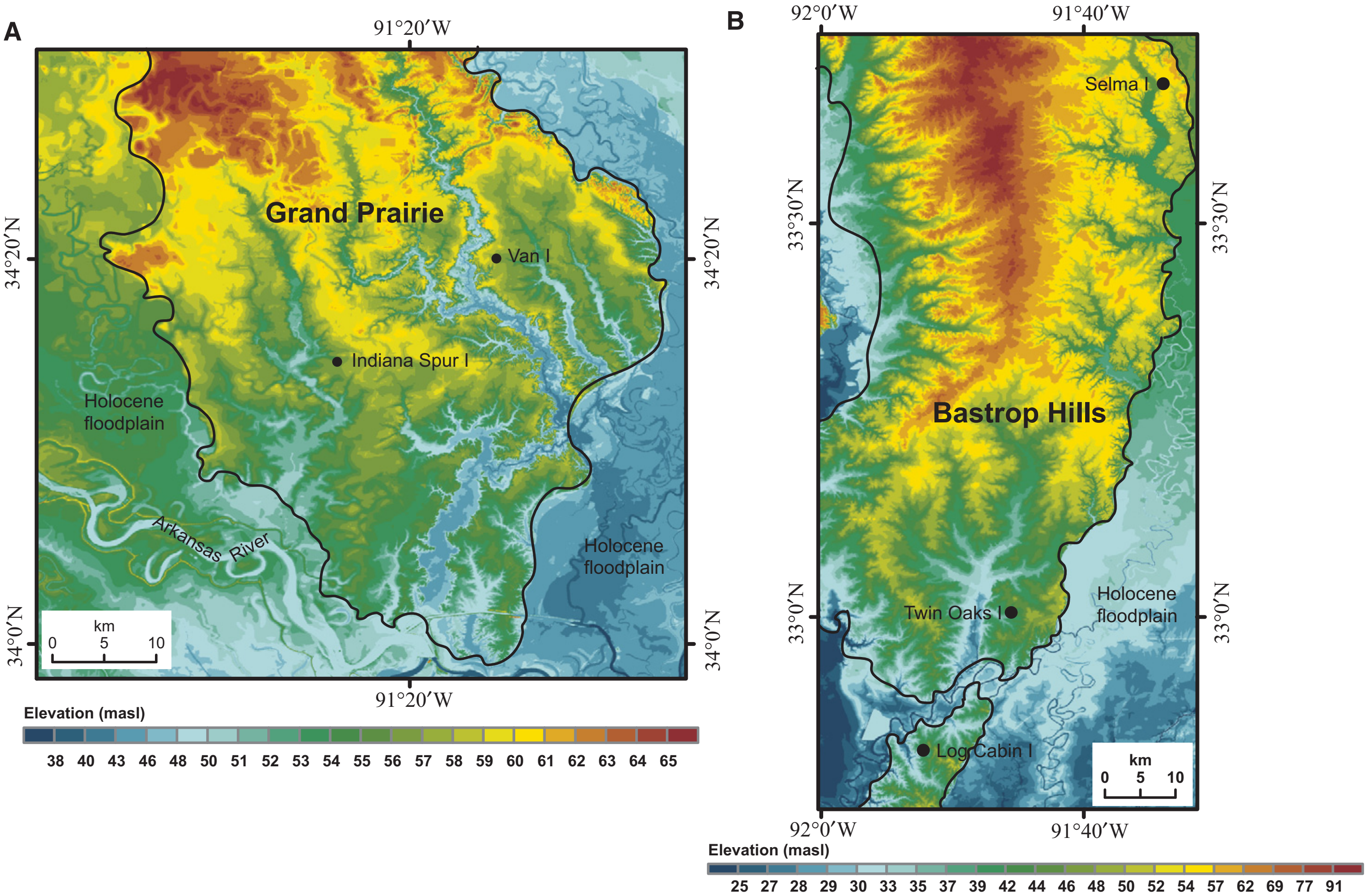


Figure 3. Digital elevation maps (DEMs) of (A) the southern Grand Prairie region and (B) the Bastrop Hills region, both outlined by the solid lines following Saucier (1994). The locations of optically stimulated luminescence (OSL) dated cores are indicated by filled circles. Data are from the 1/3 arc-second National Elevation Data set (http://seamless.usgs.gov/; Gesch, 2007). masl—meters above sea level.

formed by the Mississippi River (Fisk, 1940; Saucier, 1994; Autin and Aslan, 2001). On the higher terrace, a smaller relict meander belt can be identified (Fig. 5A), suggesting that it was formed by the Red River, the Arkansas River, or even a smaller river.

The stratigraphy of the lower terrace was described and interpreted by Autin and Aslan (2001). It is capped by over 2 m of Peoria Loess (Fig. 4C). Underneath the Peoria Loess, a 2–4-m-thick yellow clay bed is found, which was interpreted as a Mississippi River overbank deposit that formed after the abandonment of the meander belt. Autin and Aslan (2001) described wedge-shaped cracks filled with Peoria Loess embedded within the yellow clay. Underlying the yellow clay, there are increasingly sandy natural-levee and channel-belt deposits. Sample Marksville I-1 was taken from the yellow clay and was dated to 35 ± 3 ka, i.e., older than the Peoria Loess. Sample Marksville I-2 from the natural-levee facies has an OSL age of 88 ± 9 ka. Sample Marksville I-3 from the channel-belt facies provides an anomalously young age of 6 ± 3 ka.

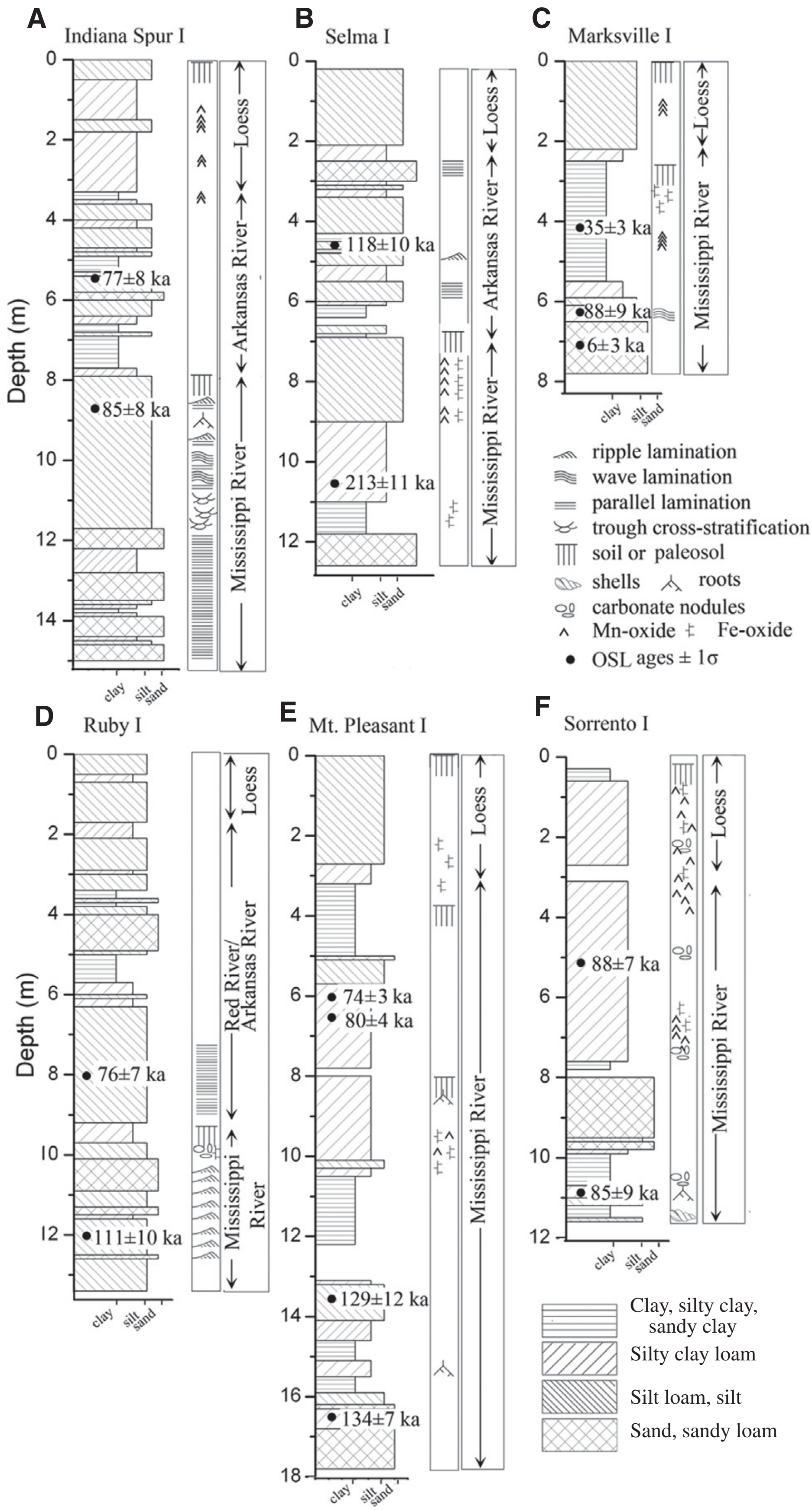
The higher terrace is mantled by ~2 m of loess (Fig. 4D). Sediments underlying the loess are reddish in color (2.5–5 YR in hue), suggesting a Red River or Arkansas River origin, in agreement with the geomorphic characteristics. The reddish sediments are underlain by yellowish brown sand and silt-rich sediments that may testify to a Mississippi River source. A paleosol containing carbonate nodules ~1 cm in diameter separates these two units. The Red River–Arkansas River sediments were dated to 76 ± 7 ka (Ruby I-1); the underlying strata have an age of 111 ± 10 ka (Ruby I-2).

Great Southwest Prairies

Two types of relict fluvial channels can be identified in the Great Southwest Prairies (Fig. 5B). In the northwest, numerous northeast-south-west–oriented channel belts were formed by the Red River when it discharged directly into the Gulf of Mexico. These have been mapped previously as the Red River deltaic plain (Saucier, 1994). In the southeast, meander loops with identifiable oxbows and point bars, comparable in size to those of the modern Mississippi River, are known as the Lafayette meander belt (Fisk, 1940; Saucier, 1994; Kinsland et al., 2008).

The Red River deltaic plain is covered by up to 4 m of Peoria Loess (Fig. 6). The Peoria Loess is underlain by 1.5–6 m of gray to brown clay and silty clay loam, interpreted as overbank deposits of the Mississippi River. Red River overbank deposits have also been observed at some locations (G. Kinsland, 2009, personal commun.). The uppermost fluvial strata are heavily mixed with the overlying Peoria Loess by shrink-swell cracking and bioturbation such as root penetration and burrowing (Fig. DR2 [see footnote 1]). Silty to sandy Red River channel-belt deposits underlie the overbank deposits. Two OSL ages

Figure 4. Sedimentary logs of selected optically stimulated luminescence (OSL) dated cores from the Prairie Complex. The gaps in the sedimentary logs indicate no core recovery. Core locations are shown in Figure 3A for Indiana Spur I, Figure 3B for Selma I, Figure 5A for Ruby I and Marksville I, Figure 5C for Mt. Pleasant I, and Figure 1 for Sorrento I.



for the overbank deposits are 31 ± 3 ka (Carencro II-1) and 52 ± 4 ka (Carencro I-1). One OSL age for the Red River channel-belt deposits is 91 ± 9 ka (Carencro II-2).

Relict Mississippi River channel belts of the Lafayette meander belt crosscut the Red River deltaic plain (Fig. 5B). The channel-belt deposits are overlain by 2–4 m of Mississippi River overbank deposits, which are in turn veneered by 2–5 m of Peoria Loess (Fig. 7). The overbank deposits, along with those in the Red River deltaic plain, could be a southward continuation of the yellow clay in the Avoyelles Prairie region (Mateo, 2005). The sandy loam or sandy channel-belt deposits are underlain by Red River deltaic deposits (Fig. 7). The overbank deposits have an OSL age of 44 ± 3 ka (New Iberia I-1). The channel-belt deposits have two OSL ages of 80 ± 8 ka (Delcambre I-1) and 52 ± 4 ka (St. Landry I-1, Fig. DR1D [see footnote 1]), while the stratigraphically deeper Red River deltaic deposits have an OSL age of 104 ± 10 ka (New Iberia I-2).

Farther south, Prairie Complex strata dip below the Holocene deltaic plain (Fisk, 1939). Core Bayou Sale IV from this setting (Fig. 1) recovered the top of Prairie Complex deposits at 18 m below sea level. A sample taken at 20.2 m below sea level (Bayou Sale IV-4) has an OSL age of 71 ± 6 ka.

Florida Parishes

The Prairie Complex in the Florida Parishes on the east side of the Lower Mississippi Valley has been described as Mississippi River channel-belt deposits (Autin et al., 1988) and Mississippi River back swamp (Saucier, 1994). Light detection and ranging (LIDAR) imagery does not reveal relict channel belts of Mississippi River size (Fig. 5C). Therefore, they are more likely overbank deposits.

Core Mt. Pleasant I was collected closely behind a bluff along the Mississippi River that was originally described by Lyell (1849) and studied in detail by Autin et al. (1988). Two paleosols were observed in this core (Fig. 4E). The first paleosol separates the Peoria Loess

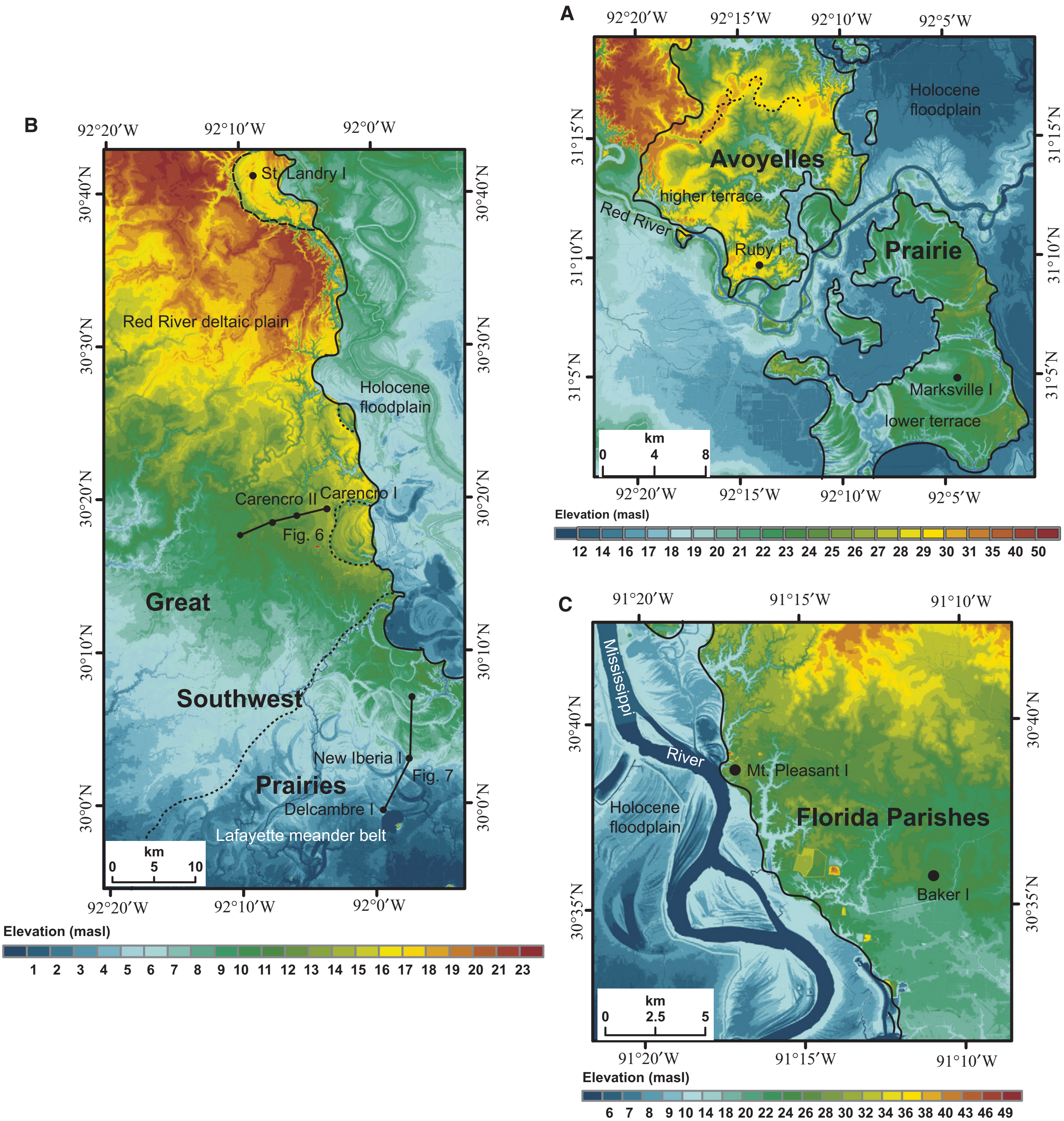


Figure 5. Light detection and ranging (LIDAR) images of the Prairie Complex. (A) Avoyelles Prairie region. The two terraces are outlined by solid lines. The dotted lines on the higher terrace delineate relict river channels. (B) Great Southwest Prairies region. The solid line separates the Prairie Complex to the southwest from the Holocene floodplain to the northeast; the dashed lines show where the Lafayette meander belt crosscuts the Red River deltaic plain. (C) Florida Parishes region. The solid line separates the Prairie Complex to the northeast from the Holocene floodplain to the southwest. Data were provided by the Louisiana State University CADGIS Research Laboratory, Baton Rouge, Louisiana, 2010 (http://atlas.lsu.edu).

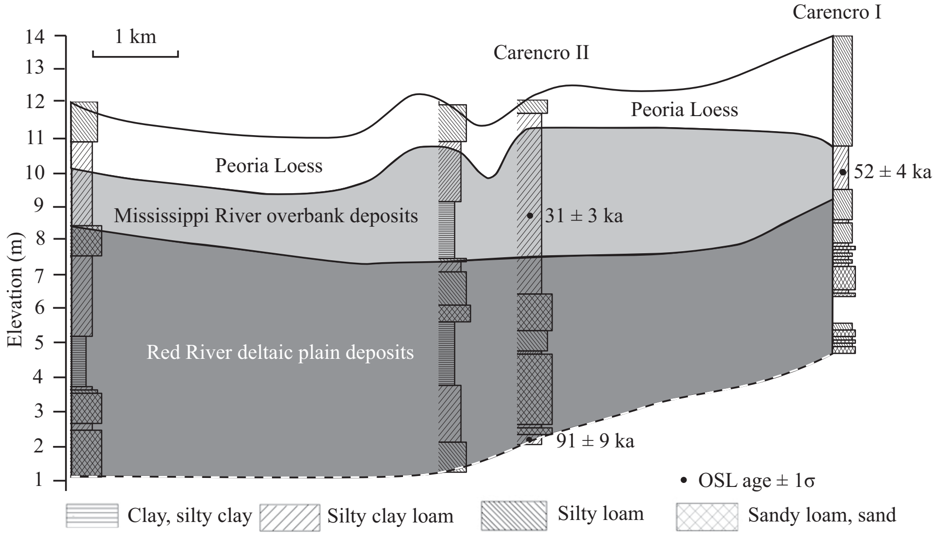


Figure 6. Cross section from the Red River deltaic plain (see Fig. 5B for location) showing the stratigraphic relationships among Red River deltaic plain deposits, Mississippi River overbank deposits, and Peoria Loess. See Figure 4 for grain-size information of each sedimentary unit. OSL—optically stimulated luminescence.

from the underlying fluvial sediments at 3–4 m depth, and the second paleosol at ~8 m depth divides the fluvial sediments into two units. Autin et al. (1988) described a third paleosol in the Mt. Pleasant bluff, just below the maximum depth of our core. Core Sorrento I (Fig. 4F) recovered a similar stratigraphy to that above the second paleosol in core Mt. Pleasant I. However, *Rangia* sp. shells indicative of a brackish lagoonal environment were found near the base of the core at ~10.5 m below sea level. Core Baker I (Fig. DR1E [see footnote 1]) has a deeply weathered surficial layer that was correlated to the lowermost paleosol at the Mt. Pleasant I site by Autin (1996).

At Mt. Pleasant I, the Mississippi River deposits above the second paleosol provide two OSL ages of 74 ± 3 ka (Mt. Pleasant I-1, weighted mean of the two ages for this sample) and 80 ± 4 ka (Mt. Pleasant I-2), while those below are 129 ± 12 ka (Mt. Pleasant I-3) and 134 ± 7 ka (Mt. Pleasant I-4) in age. At Sorrento I, the Mississippi River sediments yield ages of 88 ± 7 ka (Sorrento I-1) and 85 ± 9 ka (Sorrento I-2). At Baker I, the Mississippi River sediments provide an OSL age of 206 ± 14 ka (Baker I-1), supporting the stratigraphic correlation of Autin (1996).

RELIABILITY OF OSL AGES

OSL Signal Resetting

Adequate OSL signal resetting is a fundamental requirement in OSL dating (e.g., Olley et al., 1998; Wallinga, 2002; Rittenour, 2008). Stokes et al. (2001) showed that the residual OSL of fluvial sediments tends to decrease with increasing transport distance. Therefore, sediments in the Lower Mississippi Valley are expected to exhibit a relatively small residual OSL due to a river length of up to several thousand kilometers.

We investigated OSL signal resetting in independently dated Holocene overbank deposits along Bayou Lafourche (Fig. 1). OSL ages of three samples show good stratigraphic consistency and are in agreement with the ^{14}C age constraint provided by an underlying peat bed (Fig. 8; Törnqvist et al., 1996). The D_e distribution of the sand-size quartz (Fig. 9A) is bimodal. The MAM age of the sandy fraction is slightly younger than the age obtained from the silty fraction of the same sample (Table DR2, samples Paincourtville I-4 and I-5 [see footnote 1]). This suggests that although a residual OSL signal cannot always be ruled out, it probably accounts for no more than a few hundred years in age equivalent. This is supported by a number of recent studies (Jain et al., 2004; Rittenour et al., 2005; Rowland et al., 2005) that showed a similarly small residual OSL signal for late Holocene Mississippi River channel-belt deposits. Considering that Prairie Complex sediments constitute depositional environments similar to these Holocene counterparts (Autin and Aslan, 2001), they are likely to have equally small residual OSL signals.

For the Prairie Complex deposits, sample Mt. Pleasant I-1 was dated using both fine silt-size and sand-size quartz. The D_e distribution of the sand-size quartz (Fig. 9B) is slightly positively skewed, and insufficient bleaching is probably not the main reason for this (cf. Murray and Funder, 2003). The silt-size and sand-size fractions of the sample produce indistinguishable OSL ages (Table DR2 [see footnote 1]), confirming that the OSL ages are not affected by insufficient bleaching.

Other Sources of Error

Secular disequilibrium in the ^{238}U decay chain as discussed previously may add <10% uncertainty, which is within the uncertainties of the OSL ages. Most of the OSL ages (Table DR2 [see footnote 1]) are consistent not only in terms of stratigraphic order, but they also show good reproducibility within a stratigraphic unit. For

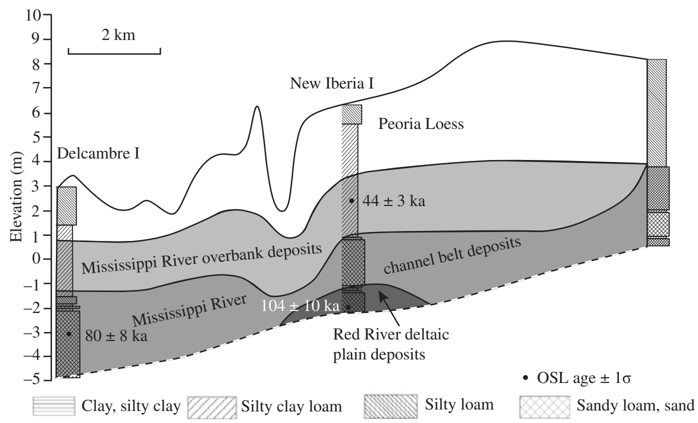


Figure 7. Cross section from the Lafayette meander belt (see Fig. 5B for location) showing the stratigraphic relationships among Red River deltaic plain deposits, Mississippi River deposits, and Peoria Loess. See Figure 4 for grain-size information of each sedimentary unit.

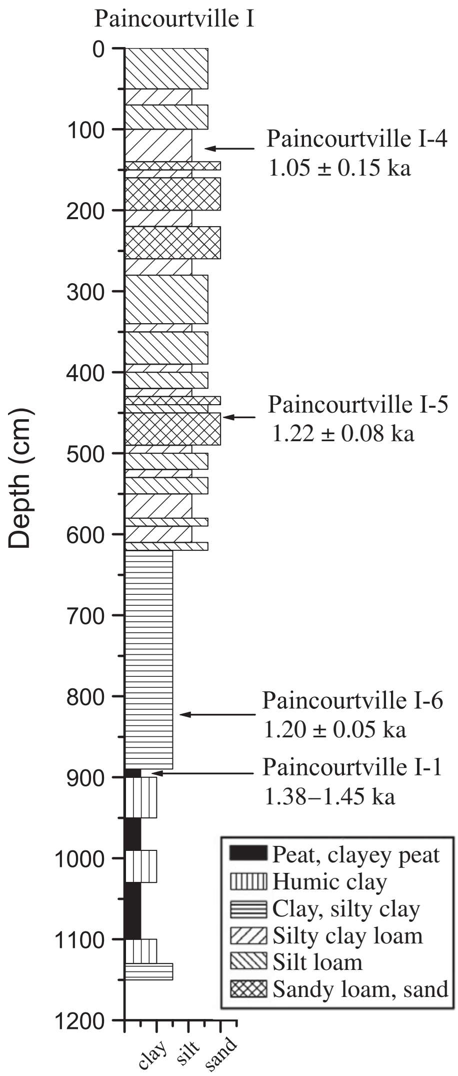


Figure 8. Optically stimulated luminescence (OSL) dating results for overbank deposits of Bayou Lafourche. A radiocarbon-dated underlying peat bed (Paincourtville I-1; Törnqvist et al., 1996) defines the maximum age of these overbank deposits. The calibrated radiocarbon age is expressed as before 2007, for comparison with the OSL ages.

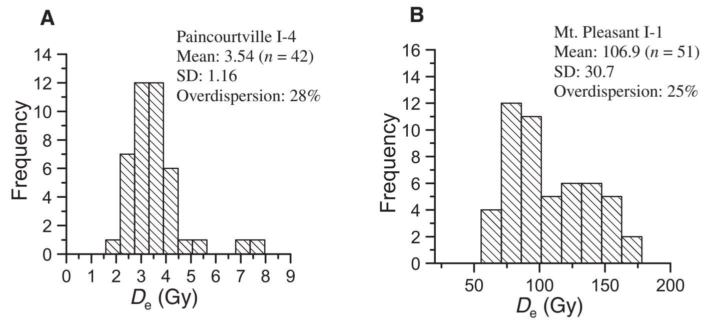


Figure 9. Histograms of D_e for OSL samples (A) Paincourtville I-4 and (B) Mt. Pleasant I-1. *n* = number of D_e estimations used for statistical analysis. SD = standard deviation. Overdispersion is calculated according to Galbraith et al. (1999). The bin width is SD/2. The D_e distribution of Paincourtville I-4 is bimodal while the D_e distribution of Mt. Pleasant I-1 is slightly positively skewed.

example, samples from the uppermost allostratigraphic unit of the Prairie Complex in cores Mt. Pleasant I and Sorrento I have comparable ages (Figs. 4E and 4F). Three samples of Arkansas River deposits taken from widely spaced localities in the Bastrop Hills region (Fig. 3B) also provide consistent OSL ages.

However, there are also problematic OSL ages, such as the anomalously young age (6 ± 3 ka) of sample Marksville I-3 (Fig. 4C). This problem probably arises from a portion of the dated grains being exposed to daylight. For all our samples, the core segments used for OSL dating were extruded in the field, and therefore the core surface (a few millimeters in thickness) was exposed to light. The inner portion of the cores (>1 cm from the surface) did not see sunlight in the field because the core segments were carefully selected so as to avoid surface fractures. Sample Marksville I-3 was taken from sandy point-bar deposits with very low cohesion. It is conceivable that this sample was fractured during transport, storage, and/or opening in the laboratory, leading to the mixing of exposed surface grains with unexposed interior grains. Sample St. Landry I-1 was taken from similar facies (Fig. 5B) and has an OSL age of 52 ± 4 ka. This sample may have experienced the same problem and may underestimate the true age of the Lafayette meander belt. Marksville I-3 and St. Landry I-1 are the only two samples that were taken from low-cohesion point-bar deposits. The rest of our samples were taken from cohesive silty deposits (Table DR2 [see footnote 1]) where the mixing process described here is unlikely to occur.

OSL ages of overbank deposits from Avoyelles Prairie (Marksville I-1) and the Great Southwest Prairies (Carencro I-1, Carencro II-1, and New Iberia I-1) range from 31 ± 3 to 52 ± 4 ka, i.e., significantly younger than the underlying channel belts (Figs. 4C, 6, and 7). We provide two possible explanations for this observation.

One possibility is that these OSL ages indicate Mississippi River overbank deposition during MIS 3 (cf. Autin and Aslan, 2001), on top of the underlying MIS 5 Prairie Complex surface. This would require the MIS 3 Mississippi River that formed the Lower Macon Ridge and Melville Ridge (Fig. 1; Rittenour et al., 2007) to aggrade to a level close to that of the MIS 5a Prairie Complex surface, allowing overbank flooding from this braided fluvial system to drape the MIS 5a Prairie Complex.

An alternative explanation is that OSL dating underestimates the age of the overbank deposits due to pedogenic and biogenic disturbance. It has been well documented that these overbank deposits are capped by a well-developed paleosol and are mixed with overlying materials (Miller et al., 1988; Schumacher et al., 1988; Autin and Aslan, 2001). This surface was subaerially exposed for at least thousands of years (Autin and Aslan, 2001), allowing reworking of the uppermost few meters. After burial by Peoria Loess, they were further mixed with the loess by physical and biological processes (Miller et al., 1988; Fig. DR2 [see footnote 1]). Since OSL dating estimates the time when the dated grains were last exposed to light, OSL ages of this unit are likely younger than the initial deposition, with the measured age depending on the degree of disturbance (cf. Bateman et al., 2003). Testing these explanations requires future work, for example, OSL dating of this unit using sediments with no evidence of disturbance, such as laminated deposits.

DISCUSSION

Prairie Complex Chronostratigraphy

The OSL ages confirm that the Prairie Complex consists of multiple allostratigraphic units, as inferred by previous studies based on paleosol correlation (Autin et al., 1988, 1991). We project the OSL ages of the Prairie Complex on a widely used global sea-level curve (Waelbroeck et al., 2002) (Fig. 10). The oldest dated allostratigraphic unit of the Prairie Complex formed during the MIS 7 interglacial and was preserved as Mississippi River deposits in the Bastrop Hills (Fig. 3B)

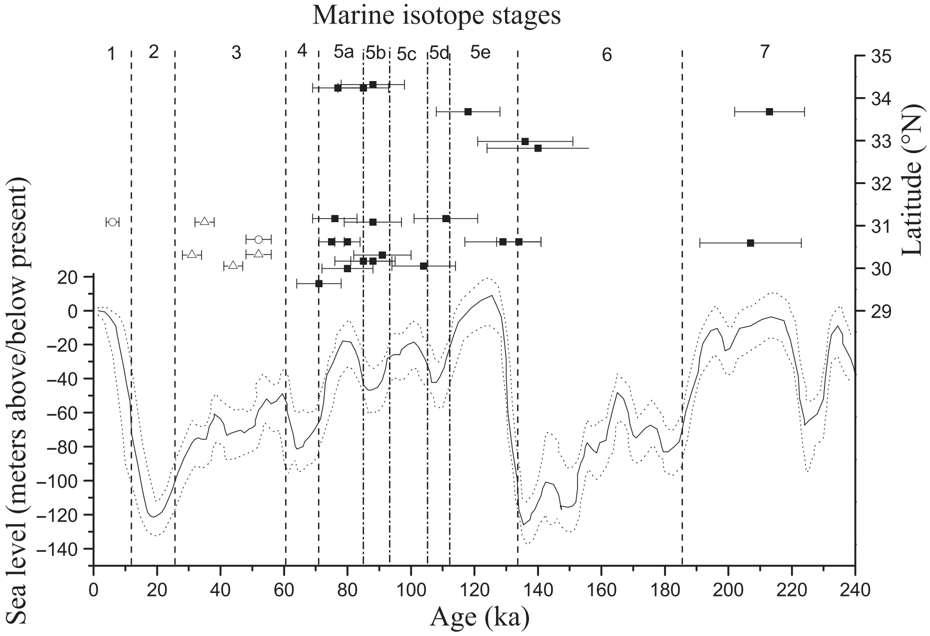


Figure 10. Comparison of optically stimulated luminescence (OSL) ages (with 1σ error bars) with a global sea-level curve (Waelbroeck et al., 2002). Marine isotope stages, substages, and their boundaries (vertical dashed lines) are also indicated (from Imbrie et al., 1984; Waelbroeck et al., 2002). The open circles represent OSL ages of samples that were likely exposed to light during sampling; open triangles represent OSL ages that may have been underestimated due to bioturbation and mixing with overlying Peoria Loess (see text for more details).

and at site Baker I in the Florida Parishes (Fig. 5C). Most of the remaining Prairie Complex was formed during the MIS 5 interglacial. The Arkansas River deposits in the Bastrop Hills and the stratigraphically deeper Mississippi River deposits at Mt. Pleasant in the Florida Parishes date to MIS 5e. The Mississippi River deposits below the higher terrace of the Avoyelles Prairie (Fig. 5A) were deposited around 111 ± 10 ka, interpreted as MIS 5c or MIS 5e. The Red River deltaic plain (Fig. 5B) deposits were dated 91 ± 9 to 104 ± 10 ka, corresponding to MIS 5a or MIS 5c. Prairie Complex deposits corresponding to MIS 5a are widely preserved and include the Grand Prairie (Fig. 3A), the Avoyelles Prairie, and the Lafayette meander belt (Fig. 5B), as well as the shallowest Mississippi River deposits in the Florida Parishes region. The Prairie Complex buried below the Mississippi Delta also provided a MIS 5a age. The Mississippi River meander belts near Rondo and Paragould, Arkansas, dated by Rittenour et al. (2007) at 77–91 ka (Fig. 11) were contemporaneous with the MIS 5a Prairie Complex.

The MIS 5a fluvial strata in the Lower Mississippi Valley are traceable, although not continuously, from near the present shoreline to ~700 km upstream. The elevation of the MIS 5a floodplain surface is plotted versus latitude in Figure 12. A third-order polynomial fit of the data reveals a convex long profile. We interpret this long profile as being distorted by crustal movements during the past ~80 k.y., featuring distinct downwarping in the most downdip portions due to deltaic sediment loading and compensational flexural uplift between 30° and 34.5°N (cf. Fisk and McFarlan, 1955). North of 34.5°N, distortion of the long profile appears to be small. These phenomena will be discussed in more detail in a later contribution.

It is generally accepted that global sea level during MIS 5e peaked at 3–9 m above modern sea level, while MIS 5a sea level may have been as much as 20 m lower (e.g., Lambeck and Chappell, 2001; Waelbroeck et al., 2002; Cutler et al., 2003; Hearty et al., 2007; Kopp et al., 2009). However, several studies (e.g., Ludwig et al., 1996; Muhs et al., 2002; Wehmiller et al., 2004; Coyne et al., 2007; Simms et al., 2009) have indicated that MIS 5a sea-level indicators around the North American continent occur within 10 m of present sea level, probably due to the fact that post–Last Glacial Maximum isostatic equilibrium has not yet been attained (Potter and Lambeck, 2003). The Prairie Complex chronostratigraphy supports a MIS 5a sea-level highstand close to that of the present, for two reasons. First, MIS 5a deposits occur stratigraphically higher than MIS 5e deposits at relatively low latitudes, as shown in core Mt. Pleasant I (Figs. 1 and 4E). Second, in core Sorrento I, *Rangia* sp. shells indicative of brackish conditions were found at 10.5 m below modern sea level, immediately below MIS 5a fluvial sediments (Fig. 4F).

Fluvial Response to External Forcing in the Lower Mississippi Valley

The correlation of Prairie Complex strata with sea-level highstands during MIS 7 and MIS 5 suggests that sea level exerts a strong influence on the evolution of the Lower Mississippi Valley, reaching farther upstream than

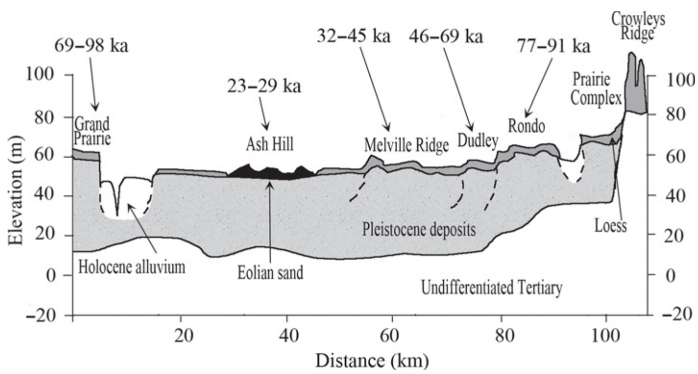


Figure 11. Cross section from the Grand Prairie to Crowleys Ridge (see Fig. 1 for location) with ages of key depositional features. Grand Prairie is part of the Prairie Complex. Rondo is a meander belt that has optically stimulated luminescence (OSL) ages comparable with those of the Grand Prairie. Dudley, Melville Ridge, and Ash Hill are braid-belt surfaces from the last glacial (modified from Rittenour et al., 2007).

what has been suggested (Saucier, 1994, 1996). This is likely due to the relatively low gradient of the Lower Mississippi Valley combined with the large sediment discharge of the Lower Mississippi River (Blum and Törnqvist, 2000). We discuss these issues in more detail later herein, with particular emphasis on major glacial-interglacial transitions.

MIS 6 to MIS 5e

Global eustatic sea level rose from at least 120 m below present at ca. 140 ka (Thomas et al., 2009) to 3–6 m above present sea level at the MIS 5e highstand by ca. 130 ka (Stirling et al., 1998; Esat et al., 1999; Bintanja et al., 2005; Hearty et al., 2007; Thomas et al., 2009). This sea-level rise induced fluvial aggradation and a valley slope reduction from the glacial period ~0.2 m/km (Rittenour et al., 2007) to ~0.1 m/km during sea-level highstand (Schumm, 1993).

The Mississippi River drainage basin experienced dramatic climate change around this transition, characterized by rapid warming (e.g., Zhu and Baker, 1995; Curry and Baker, 2000) and the disappearance of the Laurentide Ice Sheet. Sediment production in the formerly glaciated and proglacial areas would have been reduced due to the establishment of forest (e.g., Whitlock and Bartlein, 1997; Curry and Baker, 2000), prompting fluvial incision. Sharp et al. (2003) demonstrated that MIS 6 terrace fills were incised in the Missouri River drainage basin. Similarly, McKay and Berg (2008) showed that the Upper Mississippi River incised deeply into MIS 6 tills following glacial retreat.

The OSL ages for the Arkansas River deposits in the Bastrop Hills region and the stratigraphically deeper Mississippi River deposits at Mt. Pleasant range from 118 ± 10 to 140 ± 16 ka (Fig. 10), suggesting that the Lower Mississippi Valley was aggrading during the MIS 6-5e transition and into the MIS 5e sea-level highstand. Aggradation in the Bastrop Hills suggests that the upstream limit of the influence of sea-level rise is at least ~500 km from the present shoreline. Rittenour et al. (2007) arrived at a comparable distance by measuring the landward limit of coastal onlap (sensu Blum and Törnqvist, 2000) of the Holocene alluvium.

MIS 5a to MIS 4

Global sea level began falling after ca. 80 ka, following the MIS 5a sea-level highstand (Dorale et al., 2010). In the southern Lower Mississippi Valley, the youngest Prairie Complex chronostratigraphic unit was dated to 74 ± 3 to 88 ± 7 ka along the eastern margin of the Lower Mississippi Valley (Florida Parishes region) and to 71 ± 6 to 88 ± 9 ka along the western margin (Lafayette meander belt). The Mississippi River became detached from this floodplain surface (cf. Gibling et al., 2005) and was trapped in its newly incised valley following this sea-level fall. More than 600 km upstream from the present shoreline, geomorphic crosscutting relationships separate the MIS 5a Prairie Complex in the Grand Prairie region from the coeval Rondo meander belt (Rittenour et al., 2007; Fig. 11). The oldest fill terrace within this valley contains the Dudley braid belt, which formed as early as 69 ka (Rittenour et al., 2007), suggesting that the commencement of valley incision at this latitude may have coincided with the MIS 5a–4 sea-level fall.

Given its large horizontal and vertical scale, the widespread valley incision concurrent with the MIS 5a–4 sea-level fall was likely not due to autogenic processes (cf. Schumm, 1977; Best and Ashworth, 1997; Jerolmack and Paola, 2010). Instead, we argue that this incision was triggered by sea-level fall, for the following two reasons. First, the slopes of braid-belt long profiles from the last glacial (Rittenour et al., 2007) are generally significantly steeper than the MIS 5a Mississippi River long profile (Fig. 12), suggesting that these braid belts formed in a valley that deepens downdip, calling for a downstream control. Second, climate change during the MIS 5a–4 transition likely triggered fluvial aggradation rather than incision. Speleothem records from the Crevice Cave of Missouri (Dorale et al., 1998) and pollen and ostracod stratigraphy from the Raymond basin of southern Illinois (Curry and Baker, 2000) demonstrate that high-frequency climate variability during this transition may have increased catchment erosion (cf. Knox, 1996). Expansion of the Laurentide Ice Sheet also brought glacial outwash to the Upper Mississippi Valley by 75 ± 7 ka (McKay and Berg, 2008) and the northernmost Lower Mississippi Valley by 77 ± 4 ka (McVey, 2005). Consequently, sediment production increased in the Mississippi River drainage basin during this transition, yet Lower Mississippi Valley floodplains were incised.

This fluvial incision probably started at the shoreline of the MIS 5a highstand coastal prism (cf. Blum and Price, 1998; Talling, 1998; Törnqvist et al., 2000, 2003; Wallinga et al., 2004), where sea-level fall exposed a steep shoreface, although the shoreline likely remained on the continental shelf and did not drop below the shelf edge until MIS 2 (Törnqvist et al., 2006). Incision would have rapidly propagated upstream into the alluvial valley and caused the abandonment of the MIS 5a highstand floodplain. If the landward limit of sea-level influence is defined as the landward limit of incision caused by sea-level fall (Ethridge et al., 1998), sea-level influence extended at least ~600 km upstream from the present shoreline. We note that this value is roughly comparable to the distance inferred previously for the landward extent of the control of sea-level rise on fluvial aggradation.

Sea Level versus Climate Forcing

The debate about sea level versus climate forcing in the Lower Mississippi Valley has persisted for more than half a century (Fisk, 1944; Saucier, 1994, 1996; Blum et al., 2000; Rittenour et al., 2007), but no previous work possesses sufficient data to address this problem for a full glacial-interglacial cycle. The recently published chronology of braided-stream surfaces by Rittenour et al. (2007), combined with the present investigation, provides a unique new data set on continental-scale fluvial system development throughout the last glacial-interglacial cycle.

The Prairie Complex essentially represents a large-scale coastal/fluvial onlap feature that accumulated mainly during the rising limb and highstand of eustatic cycles (cf. Ross et al., 1995; Burgess and Allen, 1996; Martin et al., 2009) and was incised at the beginning of ensuing sea-level fall (cf. Heller et al., 2001; Törnqvist et al., 2003; Strong and Paola, 2008).

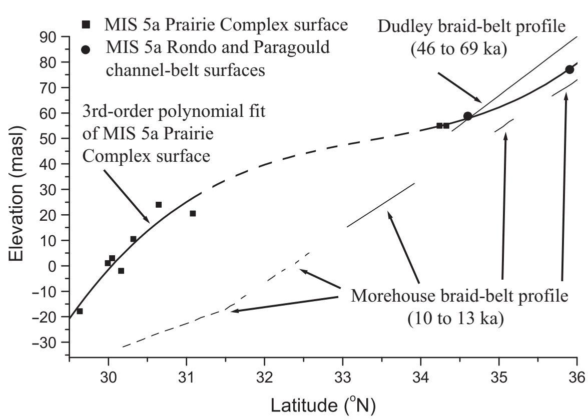


Figure 12. Long profiles of the last glacial Dudley and Morehouse braid-belt surfaces and the marine isotope stage (MIS) 5a Prairie Complex floodplain surface (masl—m above sea level). The surface elevation data of the MIS 5a fluvial strata (thickness of loess subtracted) are fitted with a third-order polynomial function. The Dudley and Morehouse long profiles and the Rondo and Paragould channel-belt data are from Rittenour et al. (2007).

Thus, we find that sea-level control is indeed profound since it controls large-scale aggradation and incision associated with large-amplitude sea-level changes at an ~100 k.y. time scale driven by major glacio-eustatic transitions. This result in part reflects the relatively short fluvial reaction time of the Lower Mississippi Valley, discussed further later herein. However, details beyond this broad picture are considerably more complex. Rittenour et al. (2007) demonstrated the dominance of upstream-controlled sediment flux as a primary forcing factor controlling braided-stream terrace formation during sea-level fall and lowstand. Nevertheless, downstream control is also recorded during falling stage into lowstand, reflected by the progressive lowering of the elevation of these terraces (Fig. 11; also see figure 9 of Rittenour et al., 2007) during the last glacial. Collectively, these two investigations elucidate the significance of both upstream and downstream control of the Lower Mississippi Valley during a full glacial-interglacial cycle. The relative importance of upstream versus downstream control along depositional dip (cf. Shanley and McCabe, 1994) fluctuates not only in space but also in time, depending on the position within the glacio-eustatic cycle.

These findings are not unique to the Lower Mississippi Valley. Blum and Price (1998) and Blum and Aslan (2006) showed that the chronology and stratigraphic architecture of the Pleistocene Beaumont Formation and post-Beaumont valley fills of the Texas Gulf Coastal Plain demonstrate that large-scale fluvial aggradation and incision are related to 100 k.y. sea-level cycles, while the nature of paleovalley fills is punctuated by upstream forcing that typically operates over shorter time scales.

Rapidity of Fluvial Response to External Forcing

The profound fluvial response to sea-level change in the Lower Mississippi Valley testifies to the sensitivity of this large fluvial system to external forcing. The weighted mean age of the MIS 5a Prairie Complex OSL samples is 73–81 ka (2σ, $n = 6$) in the Florida Parishes and the Great Southwest Prairies regions (proximal to the present shoreline) and 73–93 ka (2σ, $n = 3$) in the Grand Prairie region (~600 km upstream from the present shoreline), while the MIS 5a–4 sea-level fall was initiated at 78–80 ka (Dorale et al., 2010). The fact that these three age ranges are indistinguishable has two implications. First, fluvial adjustment to the MIS 5a–4 sea-level fall in the downstream regions was essentially instantaneous, i.e., the fluvial reaction time (sensu Bull, 1991) was short. Second, this fluvial adjustment propagated at least ~600 km upstream, likely by means of knickpoint migration (Salter, 1993; Leeder and Stewart, 1996), within the time constraints that OSL dating is capable of resolving (~10 k.y.). Such a rapid fluvial reaction to external forcing contradicts theoretical studies (e.g., Leeder and Stewart, 1996; Castelltort and Van Den Driessche, 2003). This is probably due to the relatively large fluvial diffusivity (k) of the Lower Mississippi River, which was not fully captured in these studies.

The diffusivity can be bracketed using sediment discharge rate measurements from the modern Lower Mississippi River coupled to estimates of the channel slope during interglacial (modern) and glacial conditions. The first attempt to formally derive the diffusion equation and associated fluvial diffusivity for morphodynamic systems was presented by Paola et al. (1992). The main assumptions in this derivation include that the net effect of the hydrograph can be represented via a repeated characteristic flood, that the channel adjusts itself to provide a constant dimensionless shear stress for that flood, and that the flow is quasi-uniform such that the shear stress is proportional to the depth-slope product. With these assumptions in place, k can be estimated as

$$k = \frac{q_s}{S}, \tag{1}$$

where q_s is the bed-load sediment flux per unit width, and S is the channel slope.

We calculated an upper limit for the fluvial diffusivity in the Lower Mississippi Valley (k_{max}) using estimates of q_s during flood conditions over the Holocene and the slope of the modern Lower Mississippi River (~6.7×10^{-5}). We estimated q_s using measurements from the modern Lower Mississippi River during flood conditions reported by Nittrouer et al. (2008). During one flood event (Q_w = volumetric water discharge = 3.5×10^4 m^3/s), these authors measured a q_s of 5×10^{-4} m^2/s. To account for an ~50% reduction of q_s in the modern Mississippi River relative to average Holocene conditions (Blum and Roberts, 2009), we utilized a q_s value in our calculations for k_{max} that is double the value reported by Nittrouer et al. (i.e., $q_s = 1 \times 10^{-3}$ m^2/s). Utilizing Equation 1 with these values for q_s and S, we estimated a k_{max} of 5×10^8 m^2/yr. Next, a lower limit for the fluvial diffusivity (k_{min}) was estimated using the same q_s as for k_{max} (q_s is probably larger during glacial periods, as inferred earlier) and the slope of the Lower Mississippi River during glacials (~2×10^{-4}, close to the valley slope). These values result in an estimate for k_{min} of 1.6×10^8 m^2/yr.

Using a linear diffusivity model, Leeder and Stewart (1996) derived a k value of 6×10^5 m^2/yr for the Lower Mississippi Valley by assuming—limited by the absence of any data—that it took 100 k.y. for the fluvial incision induced by sea-level fall to migrate 350 km upstream. However, as we have shown here, fluvial incision following the MIS 5a–4 sea-level fall migrated as much as 600 km within 10 k.y. Inserting these findings in the Leeder and Stewart (1996) model, we find a k value of ~10^7 m^2/yr, which is in good agreement with our previous estimation.

Our estimated range of k for the Lower Mississippi Valley can be used to estimate the range of τ for the 1000-km-long Lower Mississippi Valley. A first-order approximation of τ is given by

$$\tau = \frac{L^2}{kF_F}, \tag{2}$$

where L is the length of the alluvial basin, and F_F is the average fraction of a year that a given river is in bankfull flood (Paola et al., 1992). In our analysis, we estimate that the Lower Mississippi River is in bankfull conditions (discharge >25.5×10^3 m^3/s; cf. Kesel, 1989) for one month of every year on average (Fig. DR3 [see footnote 1]). Equation 2 coupled to our estimates of k_{max} and k_{min} yields values for τ between 25 k.y. and 75 k.y. The value of τ is significantly larger than the expected reaction time based on the OSL ages, demonstrating that fluvial reaction in the Lower Mississippi Valley occurs on a time scale much shorter than τ. This is consistent with numerical modeling by Snow and Slingerland (1990), demonstrating that most fluvial adjustments occur well before the system achieves a new equilibrium after a perturbation.

The rapid fluvial response inferred from the OSL ages provides a test to the theory using the relationship between τ and T to predict fluvial response to external forcing (Paola et al., 1992; Swenson, 2005; Swenson and Muto, 2007). Given that the calculated values of τ for the Lower Mississippi Valley are below the ~100 k.y. periodicity sea-level forcing, the theory predicts a relatively rapid fluvial response (cf. Swenson and Muto, 2007), which is exactly what we observe.

Our new findings from the Lower Mississippi Valley demonstrate that a continental-scale fluvial system with a relatively large sediment discharge can react rapidly to external forcing. Similar observations were reported for the Ganges River, which responded quickly to early Holocene changes in Asian monsoon intensity (Goodbred, 2003). It is worth pointing out that the reaction time is more important than the equilibrium time in terms of its stratigraphic significance. The reaction time determines when a fluvial system starts adjusting to a perturbation

(Bull, 1991) and therefore defines the timing of coastal/fluvial onlap/offlap, as well as sequence-boundary formation.

CONCLUSIONS

This study provides, to our knowledge, the first comprehensive assessment of the evolution of a continental-scale alluvial system based on well-preserved strata representing a full glacial-interglacial cycle, and it allows the following inferences to be made:

(1) The widespread Pleistocene terraces (Prairie Complex) that flank the Holocene floodplain in the Lower Mississippi Valley contain sediments that were mainly formed during the MIS 7 and MIS 5 sea-level highstands.

(2) Sea level exerts a strong influence on the evolution of the fluvial strata in the Lower Mississippi Valley. Large-amplitude sea-level rise and fall prompted rapid and widespread fluvial aggradation and incision, respectively. The influence of sea-level change in the Lower Mississippi Valley can extend more than 600 km inland from the present shoreline.

(3) The fluvial strata in the Lower Mississippi Valley suggest that the relative importance of sea level and climate change fluctuates both in space and in time. Large-amplitude sea-level changes at a 100 k.y. time scale control large-scale fluvial stratigraphic architecture. During sea-level fall and lowstand, climate-controlled sediment flux can override sea level as the primary external forcing. Therefore, the fluvial architecture is a product of the interaction of downstream and upstream controls.

(4) The Lower Mississippi Valley is characterized by a large fluvial diffusivity and a relatively short (compared to the 100 k.y. cyclicity) equilibrium time, allowing it to respond surprisingly rapidly to large-amplitude late Quaternary glacio-eustatic sea-level change.

ACKNOWLEDGMENTS

This research was funded by award 39240-AC8 from the American Chemical Society–Petroleum Research Fund to Törnqvist. The International Association of Sedimentologists supported Mateo's field work in 2004 through the Postgraduate Grant Scheme. Logistical support for field work was extended by the U.S. Department of Agriculture, the National Resources Conservation Service, and the Louisiana Soil Survey. Jerry Daigle provided access to a Giddings soil probe for field sampling. Thurman Allen obtained permissions to work in the Bastrop Hills region. Pennington Oil Co. provided access to the Mt. Pleasant locality. Dennis Labatte allowed use of the field facilities at Poverty Point. Coring was made possible with the help of Thurman Allen, Rick McCulloh, L.E. "Doc" L'Herrison, and Carol Swain. Frank Evans, Klaus Behrle, and Rosemary Fannelli participated as field assistants as part of student internship projects at State University of New York–Brockport. Susan Packman is thanked for part of the optically stimulated luminescence sample preparation.

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Science Editor: Nancy Riggs
Associate Editor: Martin Kennedy

Manuscript Received 15 November 2010
Revised Manuscript Received 28 July 2011
Manuscript Accepted 29 July 2011

Printed in the USA