# Deciphering Holocene sea-level history on the U.S. Gulf Coast: A high-resolution record from the Mississippi Delta


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## ABSTRACT

**Published Holocene relative sea-level (RSL) curves for the U.S. Gulf Coast are in mutual conflict, with some characterized by a smooth RSL rise akin to widely accepted eustatic sea-level curves versus others, including several recent ones, that are characterized by a conspicuous "stair-step" pattern with prolonged (millennium-scale) RSL stillstands alternating with rapid (meter-scale) rises. In addition, recent work in Texas and Alabama has revitalized the notion of a middle Holocene RSL highstand, estimated at 2 m above present mean sea level.**

**An extensive sampling program in the Mississippi Delta (Louisiana) focused on the collection of basal peats that accumulated during the initial transgression of the pre-existing, consolidated Pleistocene basement. We used stable carbon isotope ratios to demonstrate that many of these samples accumulated in environments affected by frequent saltwater intrusion in the <30 cm zone between mean spring high water and mean sea level, and we selected plant macrofossils that were subjected to AMS $^{14}C$ dating. Nearly 30 sea-level index points from a ~20 km$^2$ study area on the eastern margin of the delta suggest that RSL rise followed a relatively smooth trend for the time interval 8000–3000 cal yr B.P., thus questioning the occurrence of major RSL stillstands alternating with abrupt rises. Given the narrow error envelope defined by our data set, any sea-level fluctuations, if present, would have amplitudes of <1 m.**

**Although a true middle Holocene highstand never occurred in the Mississippi Delta, the high level of detail of our time series enables a rigorous test of this hypothesis. Correction of our data set for a hypothetical tectonic subsidence rate of 1.1 mm yr$^{-1}$ (assuming a constant subsidence rate compared to the tectonically relatively stable adjacent coast of Texas) leads to sea levels near 2 m above present during the time interval 6000–4000 cal yr B.P. However, this model also implies a RSL position near –2 m around 8000 cal yr B.P., which is inconsistent both with data of this age from Texas, as well as with widely accepted sea-level data from elsewhere. We therefore conclude that a middle Holocene highstand for the U.S. Gulf Coast is highly unlikely, and that the entire area is still responding glacio-isostatically, by means of forebulge collapse, to the melting of the Laurentide Ice Sheet.**



**Keywords: sea-level change, Holocene, Mississippi Delta, Gulf Coast, Quaternary geology.**


## INTRODUCTION

Although Holocene sea-level studies have been carried out on the United States Gulf Coast since the early days of $^{14}C$ dating, the relative sea-level (RSL) history of this area remains strikingly controversial. The compilation of Holocene sea-level data by Pirazzoli (1991) highlights the large variability of reconstructed patterns of RSL rise for the northern Gulf of Mexico. Many Holocene RSL curves from the Mississippi Delta and the surrounding U.S. Gulf Coast exhibit a conspicuous "stair-step" pattern of short periods of rapid RSL rise alternating with stillstands or even RSL falls (e.g., Curray, 1961; Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Penland et al., 1991; Thomas and Anderson, 1994; Fig. 1), in many cases with durations on the order of millennia. These RSL curves commonly show vertical differences of 5–10 m and are usually out of phase, with high rates of RSL rise in some of them coinciding with stillstands in others (e.g., compare the two RSL curves from the Mississippi Delta in Fig. 1). Rapid changes in rates of RSL rise are generally associated with major meltwater pulses, such as have been documented for the last deglaciation (Fairbanks, 1989; Edwards et al., 1993; Bard et al., 1996). Indeed, the inferred Gulf of Mexico features have been linked to sudden collapses of marine sections of the Antarctic Ice Sheet (Anderson and Thomas, 1991), but they are difficult to reconcile with widely used "eustatic" sea-level data (e.g., Edwards et al., 1993; Bard et al., 1996) that show more continuous, but gradually decreasing, rates of RSL rise during the Holocene.

Recent work in Texas and Alabama (Morton et al., 2000; Blum and Carter, 2000; Blum et al., 2001, 2002) has provided a case for one or more middle to late Holocene RSL highstands,


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at least 2 m above present mean sea level, revitalizing numerous earlier reports of such phenomena throughout the Gulf of Mexico (e.g., Behrens, 1966; Holmes and Trickey, 1974; Stapor, 1975; Tanner et al., 1989; Stapor et al., 1991; Donoghue and Tanner, 1992; Tanner, 1992; Walker et al., 1995; Donoghue et al., 1998). While the cause of a higher than present middle Holocene RSL is uncertain, conceivable driving mechanisms could include equatorial ocean siphoning (the transfer of ocean water to offshore collapsing forebulge areas; Mitrovica and Peltier, 1991) and/or hydro-isostasy (a result of the transfer of viscous mantle material from the oceans to continents due to ocean-water loading; Bloom, 1967; Walcott, 1972). If true, this would have major consequences for geophysical models that predict a dominant role of glacio-isostatic adjustment (forebulge collapse) throughout the U.S. Gulf Coast, resulting in continuous Holocene RSL rise (e.g., Mitrovica et al., 1994; Peltier, 1998, 2002a; Mitrovica and Milne, 2002). Nevertheless, these modeling studies also hint at a possible mechanism for a middle Holocene highstand in the westernmost Gulf of Mexico, which might be attributed to hydro-isostatic uplift of Mexico due to its proximity to ocean-water loads in the Gulf as well as the Pacific Ocean. In the meantime, Blum et al. (2002) suggested that a late Holocene RSL fall might reflect true glacio-eustatic change, associated with readvance and thickening of the Antarctic and Greenland Ice Sheets. However, the notion of Holocene RSL highstands remains a highly controversial issue as witnessed by vigorous discussions in the literature (e.g., Donoghue et al., 1998; Otvos, 1999, 2001), as well as recent work by Rodriguez et al. (2004) in eastern Texas, providing evidence for middle Holocene estuarine conditions well offshore of the present shoreline.

The Mississippi Delta has never provided evidence for Holocene RSL fall, which is most readily explained by considerable tectonic subsidence rates due to lithospheric loading by the deltaic depocenter. Nevertheless, the Mississippi Delta provides an excellent testing ground for the occurrence of Holocene RSL highstands. Compared to the adjacent Gulf Coast, obtaining high-density, continuous records of Holocene RSL change is more feasible and can thus eliminate potential problems associated with data sets, like many of the ones recently published, that exhibit considerable observational gaps. The occurrence of a middle Holocene RSL highstand elsewhere along the Gulf Coast should be echoed in the Mississippi Delta by a sudden decrease in the rate of RSL rise.

Classical studies of RSL rise in the Mississippi Delta were carried out by McFarlan (1961) and Coleman and Smith (1964). Although these investigations were at the forefront of sea-level research in their time, conceptual, methodological, and technological advances provide considerable scope for new work. More recent studies in this area have been scarce and have mainly focused on the central parts of the Mississippi Delta, where modern rates of RSL rise, as determined from tide-gauge records, can be up to 2 cm $yr^{-1}$ (Penland and Ramsey, 1990). A major problem in the central Mississippi Delta is the difficulty of finding sites not undergoing rapid and spatially highly variable compaction of underlying strata. Since the muddy and organic Holocene succession can be as much as 70 m thick in this area (e.g., Roberts et al., 1994), this is a serious point of concern. Age-depth plots of sea-level indicators obtained in such settings therefore show vertical scatter as high as 8 m (Penland et al., 1991), heavily overprinting the roles of eustasy, isostasy, and tectonism.

The study by Coleman and Smith (1964) also reveals considerable vertical scatter (up to 2 m) of samples of similar age. This may well be related to the fact that these authors had to rely on age measurements of inherently inaccurate bulk peat samples in a time when the analytical precision of $^{14}C$ measurements was lower. By analogy, recent work on the age of Mississippi River subdeltas (Törnqvist et al., 1996) has shown that chronologies established in the 1960s have sometimes overestimated ages by up to 2000 $^{14}C$ yr. This can partly be attributed to new technological developments, notably the advent of $^{14}C$ dating by accelerator mass spectrometry (AMS). In addition, Coleman and Smith's (1964) results may also have been influenced by differential compaction (cf. Pirazzoli, 1991), although these authors downplay that factor. The present study reports on a large set of new RSL data from the Mississippi Delta that is relatively insensitive to compaction and records the interplay of eustasy, isostasy, and tectonic subsidence during a substantial part of the Holocene.

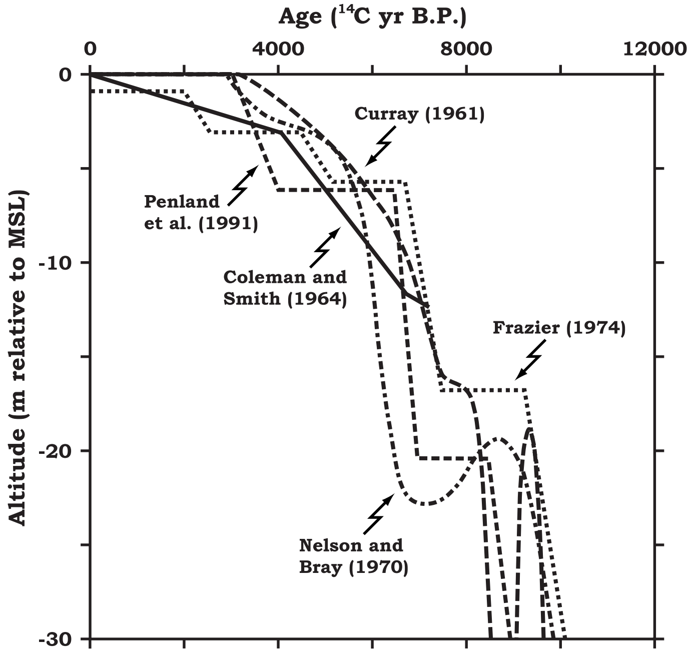



**Figure 1. Examples of relative sea-level curves for the Mississippi Delta and adjacent Gulf Coast by Curray (1961), Texas; Coleman and Smith (1964), western Mississippi Delta (their upper curve); Nelson and Bray (1970), Texas/Louisiana border area; Frazier (1974), Texas/Louisiana; and Penland et al. (1991), central Mississippi Delta.**

## APPROACH

An approach that has so far hardly been followed in the Mississippi Delta (in fact, rarely along the northern U.S. Gulf Coast) is the use of basal peats as sea-level indicators. The few exceptions in coastal Louisiana include the studies by Gould and McFarlan (1959) and McFarlan (1961), each of which reported a limited number of basal-peat ages. The basal-peat methodology is specifically suitable for submerging coastal settings and has proven extremely powerful on the Atlantic Coast of North America (e.g., Bloom and Stuiver, 1963; Redfield, 1967; Belknap and Kraft, 1977; Van de Plassche et al., 1989; Gehrels and Belknap, 1993; Gehrels et al., 1996; Kearney, 1996; Gehrels, 1999; Shaw and Ceman, 1999; Bratton et al., 2003; Donnelly et al., 2004), as well as in northwestern Europe (e.g., Jelgersma, 1961; Van de Plassche, 1982, 1995; Denys and Baeteman, 1995; Kiden, 1995; Shennan et al., 2000; Kiden et al., 2002; Shennan and Horton, 2002). The rationale of this approach is that basal peat overlies a relatively consolidated (commonly Pleistocene) basement experiencing negligible compaction (cf. Kaye and Barghoorn, 1964), and can be related to a paleo-(ground)water level (GWL). Provided that the GWL is closely related to mean sea level (MSL) or mean high water (MHW), accurate sea-level indicators can be obtained if the altitude and age of the basal peat is carefully measured. Obviously, the assumption of a compaction-free subsurface is crucial, and recent work (Kooi and De Vries, 1998) suggests that hydrodynamic compaction (i.e., expulsion of groundwater) due to sediment loading is an ongoing process even in deeply buried strata (e.g., Tertiary clays underlying the southern North Sea Basin), although at rates orders of magnitude slower than those reported for the Holocene succession in the Mississippi Delta.

A particularly appealing aspect of the Gulf of Mexico is the microtidal regime, which largely eliminates a major source of uncertainty that is prominent in many of the regions where basal-peat dating has been carried out so far. As demonstrated by Van de Plassche (1980, 1995), spatially nonuniform distortion of the tidal wave in inshore deltaic settings, known as the "flood-basin effect," may severely complicate the interpretation of basal peats as sea-level indicators. Nevertheless, basal-peat dating in mesotidal and macrotidal environments has been used successfully to quantify differential glacio-isostatic movements, both in the northeastern United States (Gehrels and Belknap, 1993; Gehrels et al., 1996) and in northwestern Europe (Denys and Baeteman, 1995; Kiden, 1995; Shennan et al., 2000; Kiden et al., 2002; Shennan and Horton, 2002). Comparative analysis of large numbers of sea-level (SL) index points from Great Britain by Shennan and Horton (2002) has shown that basal peats are particularly likely to provide consistent RSL records.

A serious limitation of bulk basal peats as sea-level indicators is the difficulty to interpret them ecologically to establish their indicative meaning with respect to sea level (see, e.g., the discussion by Smith and Coleman, 1967). Whereas freshwater peat may form well above MSL, peat that accumulates under more saline conditions has a more tightly constrained indicative meaning. Nevertheless, studies in the Rhine-Meuse Delta (The Netherlands) have shown that in a ~50-km-wide coastal zone, GWL as determined exclusively from freshwater peat has been almost horizontal and close to MSL or MHW, in particular during the past ~6000 cal yr (Van Dijk et al., 1991; Van de Plassche, 1995; Cohen, 2003).

Botanical macrofossils have long been recognized as important paleoecological indicators in sea-level research (Behre, 1986). Törnqvist et al. (1998) demonstrated the potential of improving the accuracy of basal peats as indicators of GWL by $^{14}C$ dating carefully selected plant macrofossils (e.g., fruits, seeds, twigs) by AMS. This allows 3–4 orders of magnitude smaller samples (<1 mg carbon) to be dated than the conventional technique that is based on bulk peat samples. Besides improving dating accuracy (Törnqvist et al., 1992), plant macrofossils also enable the selection of specific paleoecological indicators, and, thus, provide a more straightforward relationship with GWL. In addition, both indicator taxa and macrofossil assemblages (representing plant communities) may be used to reconstruct the salinity, thus further constraining the sea-level relationship (Behre, 1986). Of particular interest, the use of macrofossils rather than bulk peats also opens new perspectives for dating relatively organic-poor clastic deposits (e.g., humic clays), as long as they contain in situ plant remains. Apart from that, stable carbon isotope ratios ($\delta^{13}C$) have proven extremely useful to estimate the salinity regime of vegetation and sediments in modern Mississippi Delta wetlands (DeLaune, 1986; Chmura et al., 1987), providing another tool for using basal peats as sea-level indicators (Chmura and Aharon, 1995).

## STUDY AREA

We carried out our investigation in a ~20 km$^2$ study area on the eastern margin of the Mississippi Delta where widespread basal peat overlies a consolidated Pleistocene basement. Early studies (e.g., Saucier, 1963; Kolb and Van Lopik, 1966) have indicated that this Pleistocene basement occurs at relatively shallow depth (<15 m below the surface) in the Lutcher-Gramercy area, halfway between Baton Rouge and New Orleans, Louisiana (Fig. 2). Saucier (1963) further identified a fluvial drainage pattern that was excavated during the RSL lowstand of the last glacial. This would potentially enable the collection of basal-peat samples along relatively steep altitudinal gradients in a small area, which is essential to minimize the possible role of differential crustal movements within our study

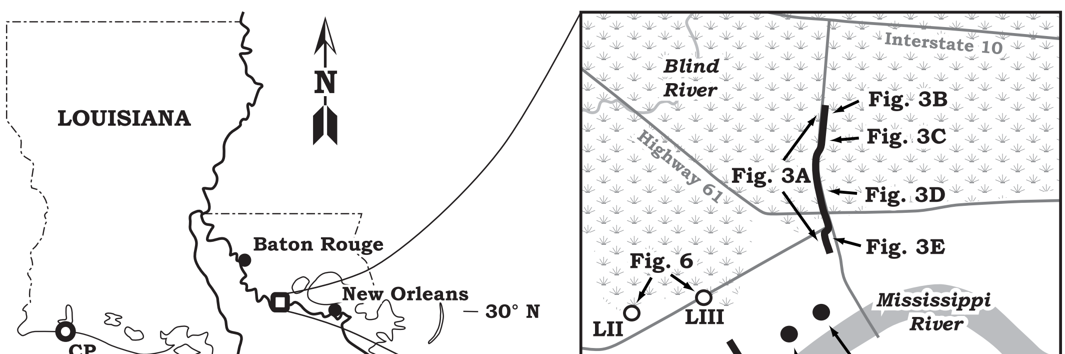



**Figure 2. Location of the study area with position of cross sections (Figs. 3 and 4) and $^{14}C$ dated cores (Fig. 6). Most sampling sites were located in agricultural land bordering the Mississippi River; access to swamp areas was obtained along one of the few roads. Louisiana state map shows location of tide-gauge stations (Table 1). C—Cocodrie; CP—Calcasieu Pass; GI—Grand Isle East Point; SWP—Pilot Station SW Pass; LII—Lutcher II; LIII—Lutcher III.**

area. More detailed mapping by Blaauw and Meijneken (1998) has confirmed the presence of a distinct, erosional fluvial morphology of the Pleistocene surface in the Lutcher-Gramercy area, interpreted as a small paleovalley system occupied by a tributary to the trunk paleovalley of the Mississippi River, southwest of our study area (Fisk and McFarlan, 1955; Saucier, 1994). We focused our data collection such to maintain at least a few kilometers distance to the nearest subsurface salt dome (the Hester dome with top of salt at >2 km depth; Halbouty, 1979). There is no known evidence for active growth faults within our study area.

## METHODS

Field-data collection consisted of hand drilling using an Edelman auger and a 1-m-long gouge (Oele et al., 1983). We used a 3-cm-diameter gouge for reconnaissance drilling and a 6-cm-diameter gouge for obtaining core samples, typically 15 cm in length. Land surface elevations of all core sites in our cross sections (Figs. 3 and 4) were surveyed with a TOPCON GTS-4B electronic total station and tied to National Geodetic Survey (NGS) benchmark BJ3747 (30º03′20″ N, 90º39′31″ W). The elevation of this benchmark, as well as temporary benchmarks close to our sampling sites, was measured with a differential Global Positioning System (GPS).

The GPS data were collected using three Trimble 4700 receivers and Compact L1/L2 antennas with groundplanes. Measuring sites were set up at or near the benchmarks using tripods with tribrachs and optical plummets. The sites were occupied for 6–8 h each; the base station (near the BJ3747 benchmark) was occupied for three successive days (6–8 h per day). All vertical measurements are tied to the BJ3747 benchmark, anchored by a 14.6 m long, stainless steel rod and classified as stability category B (defined as "probably hold position/elevation well"). The orthometric height of the BJ3747 benchmark was determined by differential leveling and adjusted by NGS in February 1994; its elevation is +4.814 m with respect to the North American Vertical Datum (NAVD) 88. After data processing with Trimble Geomatics Office, the network was adjusted using the NAVD 88 orthometric height of BJ3747, rather than the ellipsoid height, and this elevation was propagated through the network. The local geoid separation at each temporary benchmark was determined using GEOID99.

Sample processing involved ultrasound treatment of core slices (mostly 2–4 cm thick) and wet sieving over screens with mesh openings of 2, 1, 0.5, and 0.25 mm. The majority of useful botanical material occurred in the 0.5–2 mm size range and was examined and identified under the microscope with 7× to 50× magnification. Degraded wood remains were thin-sectioned in three anatomical planes (cross, radial, and tangential sections) using a steel microtome blade. The sections were mounted on glass slides in a glycerine solution and examined under magnifications ranging from 40× to 1000×. In the case of charcoal, specimens were fractured along the anatomical planes to reveal the cell structure, then identification proceeded as with the degraded wood. All macrofossils selected for dating were stored in refrigerated, acidified, distilled water.

Radiocarbon dating was carried out by means of AMS at the Robert J. Van de Graaff Laboratory, Utrecht University (Van der Borg et al., 1997), to enable the direct dating of botanical macrofossils. In addition, a few bulk paleosol samples were dated conventionally at the Centre for Isotope Research, University of Groningen. Stable carbon isotope ($\delta^{13}C$) measurements were routinely performed for all $^{14}C$ samples. We calculated weighted means for those samples where multiple subsamples were dated. Calibration of the $^{14}C$ data was performed with the Groningen CAL25 software (Van der Plicht, 1993), using the INTCAL98 data set (Stuiver et al., 1998) and a smoothed calibration curve (smoothing parameter $\sigma_s$ = 80; Törnqvist and Bierkens, 1994). For each SL index point we obtained the 2σ (95%) confidence interval of the calibrated probability distribution.

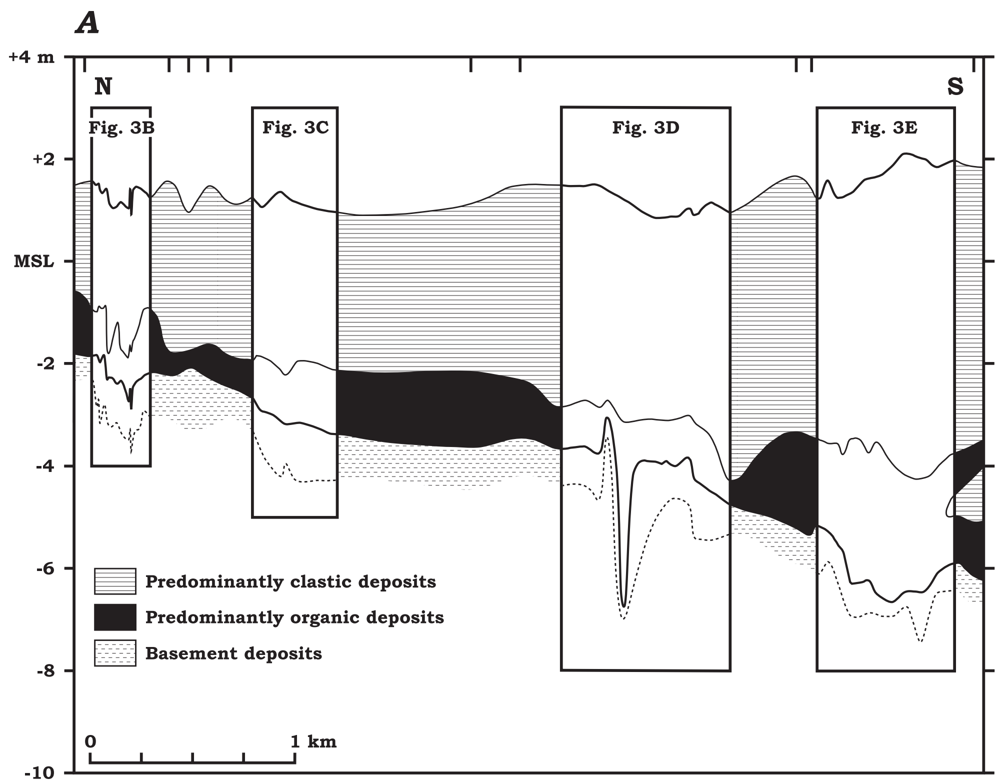
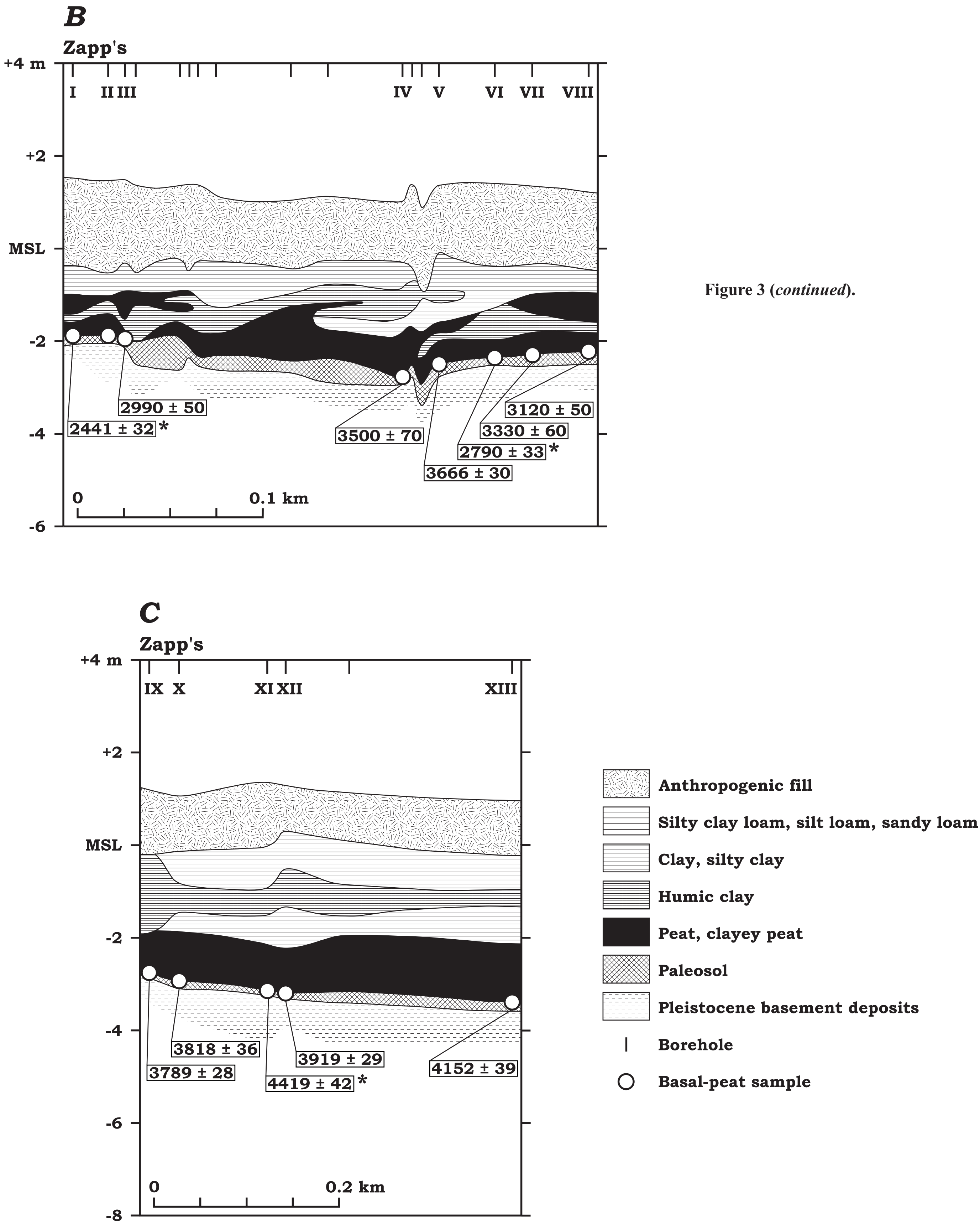



**Figure 3. Zapp's-Gramercy cross section. For location see Figure 2. (A) generalized overview cross section and (B-E) detailed cross sections in areas with considerable topography of the Pleistocene basement featuring the position of basal-peat samples. Sample ages are indicated in $^{14}C$ yr B.P. (those that were rejected are marked by an asterisk), with weighted mean ages in case of multiple subsamples (Tables 2 and 3). Note the generally straightforward relationship between sample altitude and age. *Figure continued on following pages.***

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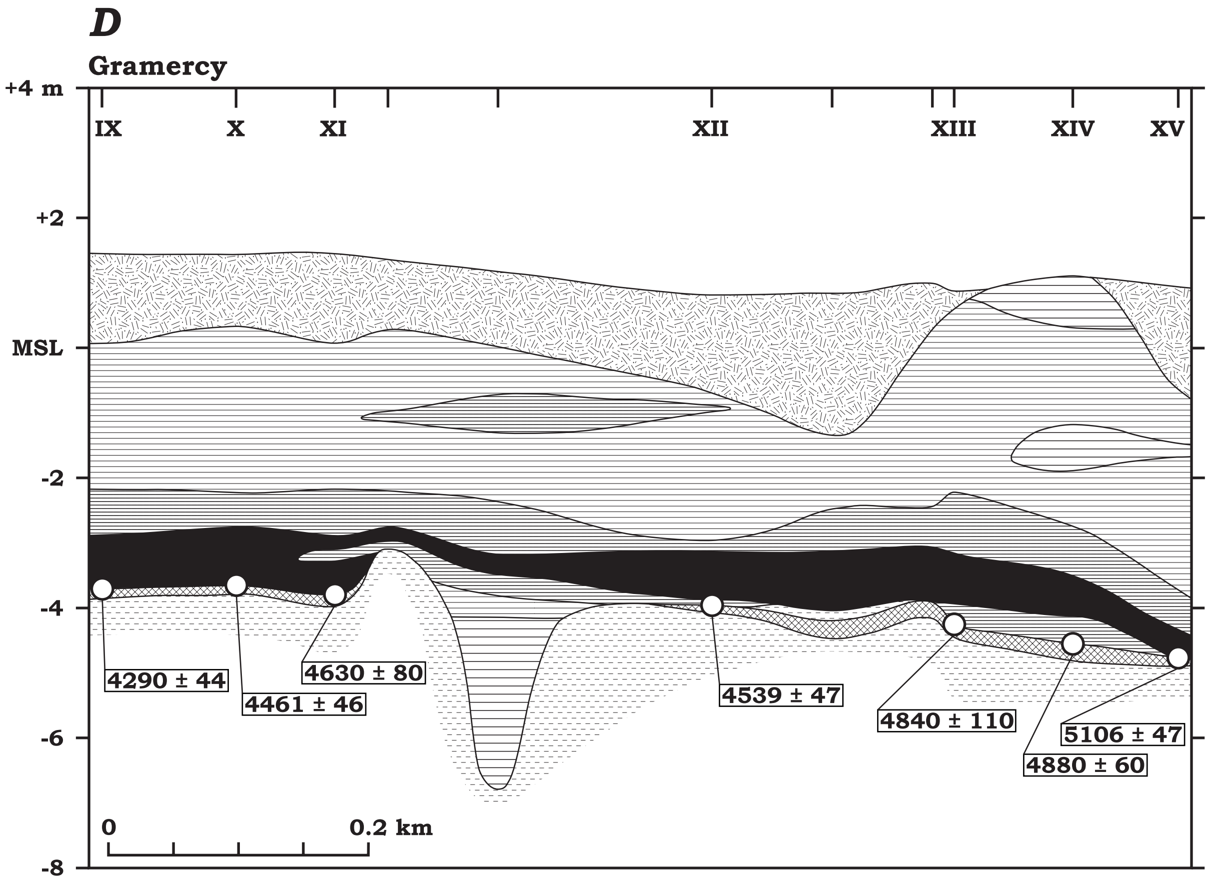


**Figure 3 (*continued*).**

We have quantified both the vertical (altitudinal) and the horizontal (age) error for all our SL index points. Altitudinal errors consist of a number of components (cf. Gehrels et al., 1996) that can be grouped into (1) errors associated with sampling, surveying, and GPS measurements; and (2) errors in the assessment of the vertical indicative range of the material dated (cf. Van de Plassche, 1986). We applied an error of –2 cm per meter depth to represent nonvertical coring, ±2 cm for measuring the sample depth in the gouge, ±3 cm for the total station surveys between GPS-measured benchmarks and sampling sites, and ±4 cm for the GPS measurements (±2 cm for the base station, ±2 cm for the stations that occupied temporary benchmarks). As a result, error bars increase with sampling depth. Previous studies (e.g., Van de Plassche, 1982) have suggested that basal peats may form anywhere between MSL and MHW. Obviously, this can only be firmly demonstrated if there are indicators for brackish or saline conditions. As will be discussed below, this is indeed the case for many of our samples. Given a spring tidal range for this part of the Gulf of Mexico that is always less than 60 cm (Table 1), we use a vertical indicative range of 30 cm. Allowing for the possibility that our samples may have formed anywhere between MHW and MSL, we have incorporated an error term of +30 cm, such to obtain error boxes that straddle the paleo-MSL for each SL index point. The age errors of our SL index points are defined by the 2σ (95%) confidence intervals of the probability distributions obtained after calibration of $^{14}C$ ages to calendar ages.

## RESULTS

### Stratigraphy

All our SL index points were obtained in two cross sections (Figs. 3 and 4), located ~4 km apart. In view of the objectives of this study, our analysis focuses primarily on the lowermost strata that have been depicted in the greatest detail in the cross sections. The basal unit throughout our study area is the highly consolidated, frequently oxidized, mud-dominated, Pleistocene Prairie Complex (Fisk, 1938; Autin et al., 1991) that was exposed at the land surface for a prolonged time period during the RSL lowstand of the last glacial. The majority of our boreholes terminate in an equally consolidated and up to several meters thick layer of Peoria Loess that blankets the Prairie Complex (Blaauw and Meijneken, 1998). Overlying this unit is an immature paleosol (Fig. 5), consisting of a dark gray A-horizon developed in silty material and enriched in highly decomposed organic matter (Entisol, suborder Aquent, representing a waterlogged environment; Soil Survey Staff, 1999). On average, the paleosol thickness increases upsection (e.g., compare Figs. 3B and 3E). In view of the immature nature of this paleosol, it is unlikely to be a result of pedogenesis during the prolonged subaerial exposure of late Pleistocene strata, as is confirmed by $^{14}C$ dating (see below). The thin, muddy unit that is found between the paleosol and Peoria Loess (Fig. 5) is interpreted as an initial, transgressive facies reflecting deposition during low-frequency storm events (e.g., hurricanes), prior to the establishment of a more permanently high groundwater table, controlled by sea level.

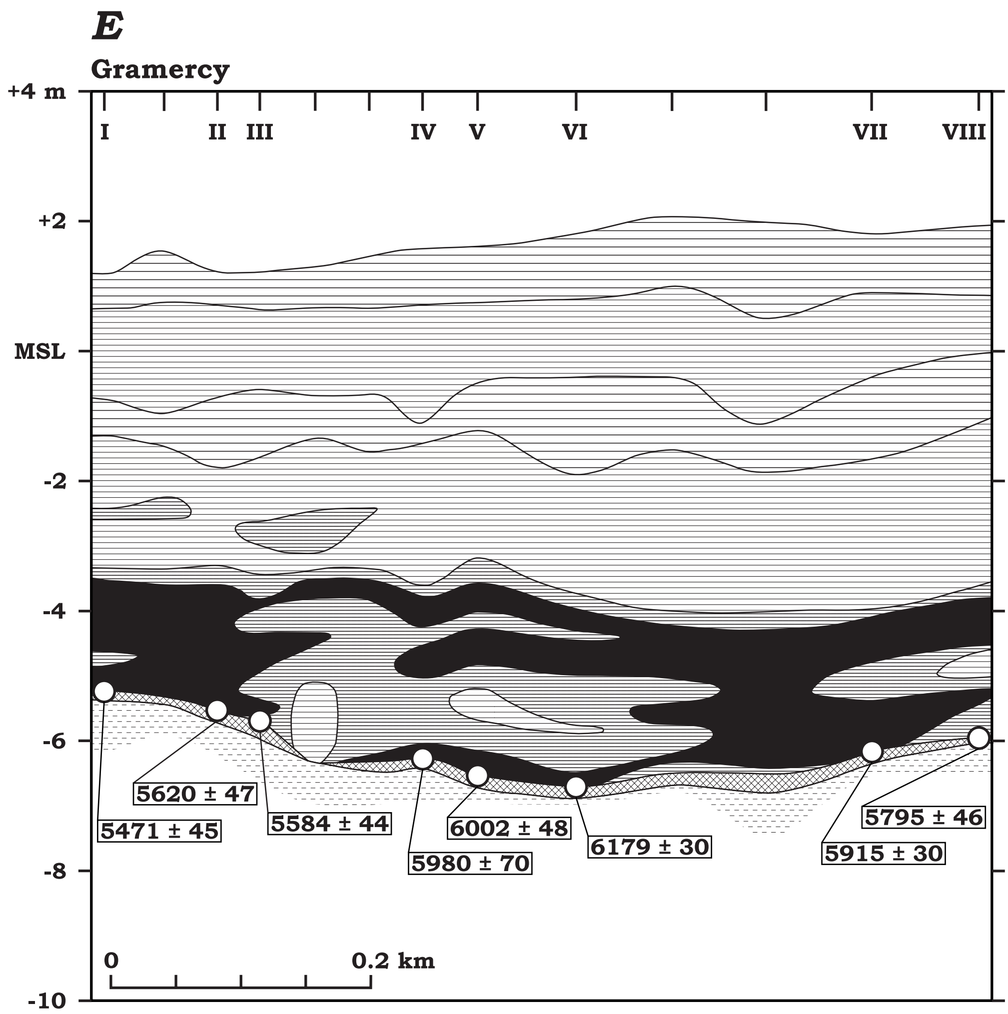


**Figure 3 (*continued*).**

Immediately overlying the paleosol is a dark brown basal peat bed (Fig. 5) that is commonly very clayey or silty. The organic-matter content of this unit varies considerably and is typically 25%–75% in terms of loss on ignition. In places, this unit consists of humic clay with values in the range 10%–25%. In most of the study area, the basal part of this organic unit is a herbaceous (marsh) peat, but at depths less than ~3 m below present MSL wood peat dominates. In several cases evidence was found for burrowing, presumably by crawfish, near the contact between the basal peat and the underlying A-horizon. We never encountered evidence for significant oxidation or pedogenesis within the basal organic strata, indicating that peat formation encroached on the underlying basement without interruption. Only a few sites show purely clastic deposits immediately overlying the Pleistocene basement, sometimes representing scour by small channels (presumably crevasse channels; Fig. 3D). The basal peat varies in thickness from a few decimeters to several meters and is in turn overlain by clastic strata that thicken rapidly to the south (toward the Mississippi River), interpreted as fluvial overbank deposits. Previous studies (Saucier, 1963; Törnqvist et al., 1996) have shown that these overbank deposits belong to the trunk channel belt of the St. Bernard subdelta (activated ~3600 $^{14}$C yr B.P.) and the Plaquemines-Modern subdelta (active since ~1300 $^{14}$C yr B.P.), respectively. In the Lutcher cross section (Fig. 4), the lowermost clastic deposits that immediately overlie the basal peat consist of distinctly different silty to sandy facies with abundant *Rangia cuneata* bivalves, indicating brackish conditions. This unit can be correlated with a widespread lagoonal deposit that was described in the Barataria Basin, south of our study area, by Kosters and Suter (1993).

**Radiocarbon Dating**

Conventional $^{14}$C ages were obtained from the A-horizon that caps the Pleistocene basement at two sites with distinctly different depth below MSL (Fig. 6, Tables 2 and 3). In both cases, three separate subsamples were dated, including the fraction of organic matter >180 µm, the alkali extract (humic acids), and the residue. For sample Lutcher III-1, the three fractions provided nearly identical results, while for sample Lutcher II-1 the residue was significantly older and likely contains reworked organic matter. Subsample Lutcher II-1c was therefore rejected. Weighted mean ages (Table 3) of 6685 ± 90 $^{14}$C yr B.P. at ~8 m below MSL, and 3470 ± 120 $^{14}$C yr B.P. at ~3 m below MSL provide conclusive evidence that this paleosol is a diachronous feature, associated with the initial rise of the GWL as an

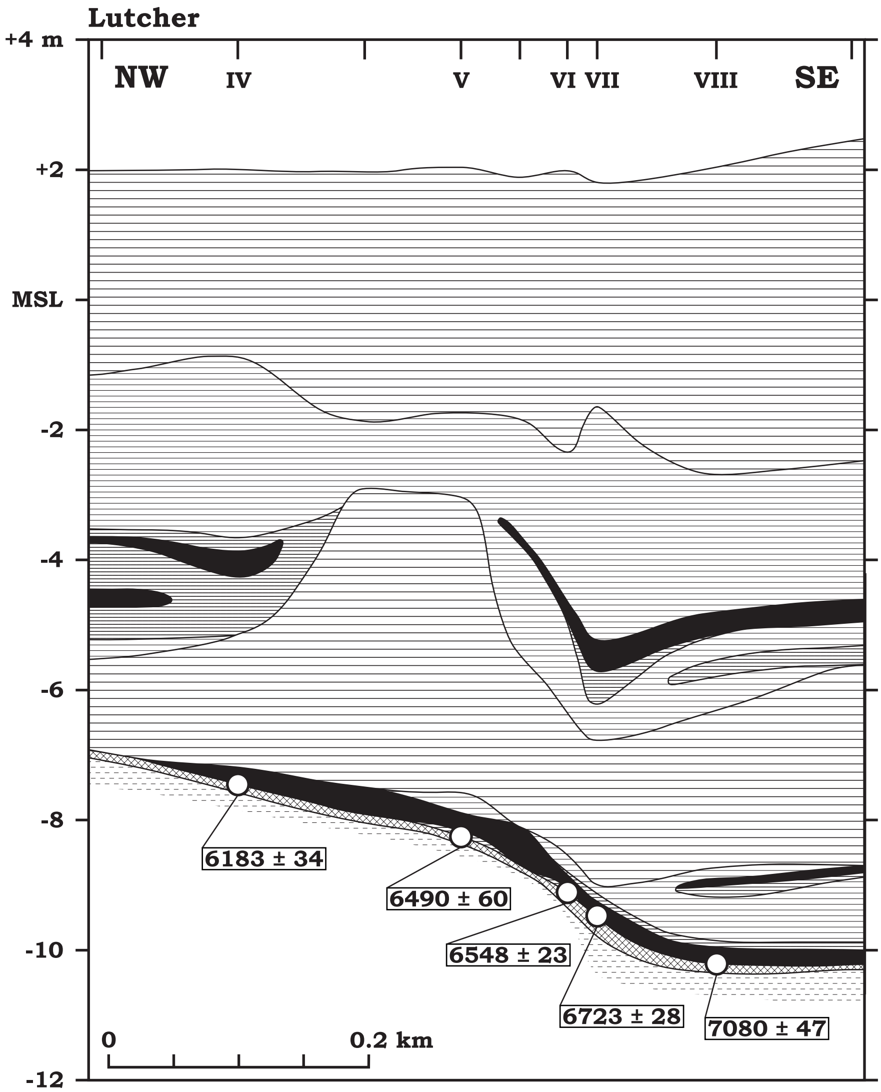


**Figure 4. Lutcher cross section. For location see Figure 2; for legend and further explanation see Figure 3.**

TABLE 1. TIDAL RANGE AT FOUR NOAA TIDE GAUGES IN LOUISIANA

| Tide gauge | Station ID | Latitude (N) | Longitude (W) | Mean tidal range (m) | Mean spring tidal range (m) |
|---|---|---|---|---|---|
| Pilot Station SW Pass | 8760943 | 28°55.5′ | 89°25.1′ | 0.38 | 0.41 |
| Grand Isle East Point | 8761724 | 29°15.8′ | 89°57.4′ | 0.32 | 0.34 |
| Cocodrie | 8762928 | 29°14.7′ | 90°39.7′ | 0.31 | 0.33 |
| Calcasieu Pass | 8768094 | 29°45.9′ | 93°20.6′ | 0.45 | 0.59 |

immediate response to RSL rise, and the transformation of the study area into a wetland.

Macrofossil analysis of the basal peat samples revealed a fair amount of variability as to the richness of datable materials. While in many cases elements could be used that are likely to have formed at the sampling site and have accumulated on the land surface (e.g., *Scirpus* spp. achenes), in numerous other cases we had to rely on charcoal that potentially could be either allogenic (i.e., washed in) or possibly could consist of underground (root) material. Therefore, we carried out comparative dating of multiple fractions in a number of samples. This shows that the majority of our data set consists of reproducible results for multiple fractions extracted from the same sample (e.g., samples Lutcher VI-1 and Lutcher VII-1). On

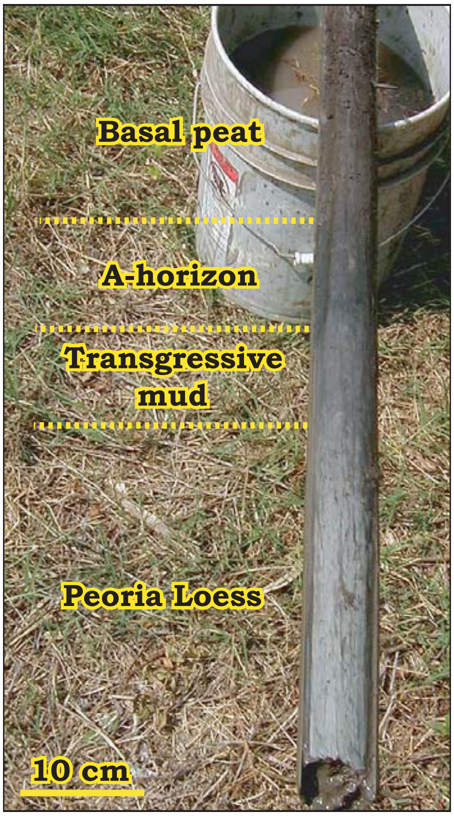


**Figure 5. Representative example of the contact between basal peat and the underlying Pleistocene basement (Peoria Loess). Note the characteristic, dark gray paleosol (A-horizon) immediately underneath the basal peat, in turn overlying a thin bed of transgressive mud (for further discussion see text section, Stratigraphy).**

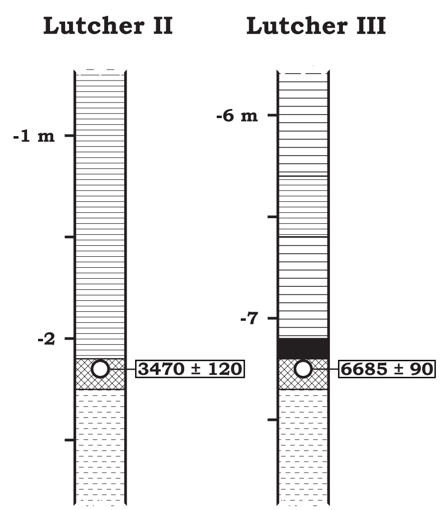


**Figure 6. Two sedimentary logs with $^{14}C$ dated A-horizons (after Blaauw and Meijneken, 1998). For location see Figure 2; for legend see Figure 3.**

the other hand, the consistency is somewhat less for younger samples that have accumulated more slowly. We rejected five of the macrofossil (sub)samples (Table 2) that were offset relative to the remainder of the data set by 400–1100 $^{14}C$ yr. Strikingly, three of these cases concern elements of *Nyssa aquatica*, suggesting particular problems with ages derived from this wetland tree species. Although we ensured that the wood samples did not consist of root material, all three outcomes are anomalously young. The most likely explanation for these spurious ages is that tree material penetrated into the subsurface, possibly due to falling trees during major storms. Despite the fact that we paid utmost attention to avoid collecting samples that were affected by burrowing, we also cannot exclude the possibility that some of the incompatible subsample ages result from bioturbation. Nevertheless, despite the few exceptions mentioned above, the dating of multiple fractions demonstrates that our $^{14}C$ data are mostly highly accurate. This is reflected by the generally robust age-altitude relationships of our basal-peat samples in the cross sections (Figs. 3 and 4).

### Stable Carbon Isotope Data

Many of the charcoal samples have $\delta^{13}C$-values of –12‰ to –16.3‰ (Table 2), indicative of plant tissue of $C_4$ taxa. Without exception, our charcoal samples are of herbaceous origin and appear to predominantly represent one grasslike taxon, likely *Spartina* spp. It is important to note that we have several examples of coexisting subsamples with contrasting $\delta^{13}C$-values, indicating plant communities with both $C_3$ and $C_4$ plants (samples Zapp's V-1, Gramercy V-1, Lutcher IV-1, Lutcher VI-1, and Lutcher VII-1). This implies that $C_4$ plant material may well be present in many of the samples where we only $^{14}C$ dated $C_3$ taxa, like *Scirpus* spp.

Chmura et al. (1987) reported mean $\delta^{13}C$-values for organic carbon-rich sedimentary material from saline (–16.2‰), brackish (–16.9‰), intermediate (–22.1‰), and fresh (–27.8‰) environments in the Mississippi Delta. Comparison with our data suggests that samples with $\delta^{13}C$-values of –19‰ to –22‰ were also likely affected, at least in part, by saline conditions. Strikingly, even a few samples (Zapp's V-1b, Zapp's XI-1, and Zapp's XII-1) that occur within the zone occupied by wood peat have heavy $\delta^{13}C$-values, suggesting that the transition to a freshwater, peat-forming environment was preceded by a short phase with saltwater intrusion.

In addition, one of the A-horizon samples (Lutcher III-1) exhibits relatively heavy $\delta^{13}C$-values (Table 2), indicating that the decomposed organic matter in this paleosol originated predominantly from $C_4$ plant material. The shallower paleosol sample Lutcher II-1 has $\delta^{13}C$-values suggesting soil formation in an environment intermediate between brackish and fresh.

### Relative Sea-Level Curve

The new RSL curve for the Lutcher-Gramercy area on the eastern margin of the Mississippi Delta (Fig. 7) consists of 29 error boxes, representing our SL index points. Collectively, the error boxes form an envelope that describes the pattern and rate of rise of MSL during the time interval 8000–3000 cal yr B.P., characterized by a generally smooth trend. We have taken a conservative approach in our assessment of altitudinal errors, as discussed above. For example, we used a vertical indicative range of 30 cm, but the present-day spring half-tidal range in coastal Louisiana is commonly considerably less (Table 1). However, we did not include possible effects of compaction that are difficult to quantify. Autocompaction (Kaye and Barghoorn, 1964) is likely negligible, given the fact that most of the samples are only 3–4 cm thick. Some compaction of the muddy paleosol that underlies the basal peat may have occurred, but the magnitude of this is most likely considerably less than the substantial errors we have assumed for the other uncertainties.

A remaining uncertainty that may require clarification in the future is the accuracy of the benchmark to which all the elevation measurements have been tied. This benchmark is anchored in a solid basement (either Prairie Complex or sandy channel deposits of the Mississippi River) at ~15 m depth, but NGS now assumes that no benchmarks in coastal Louisiana can be considered to be entirely stable. It is important to note, however, that our data set is internally consistent, and that future corrections would merely shift the RSL curve as a whole in a vertical sense. Since it is unlikely that any such correction would exceed a few decimeters, this would not affect our main conclusions.

The height of the error boxes varies between 58 and 75 cm, and in view of the consistency of our data, the envelope that circumscribes all the SL index points is only slightly thicker, and nearly always <1 m. This effectively sets the limit concerning the maximum possible amplitude of any more subtle sea-level fluctuations. In total, 14 out of 29 SL index points, scattered throughout the data set, provide direct stable carbon isotope evidence for saltwater influence (Fig. 7). There is no evidence for any systematic offset between SL index points with or without saltwater indicators. Given the $\delta^{13}C$-data discussed above, we assume a straightforward sea-level relationship for the entire data set, with

TABLE 2. LIST OF $^{14}C$ AGES WITH STABLE CARBON ISOTOPE DATA

| Sample name | UTM-coordinate (N) | UTM-coordinate (E) | Surface elevation (m) | Depth below surface (cm) | Material dated | $\delta^{13}C_{PDB}$ (‰) | Age ($^{14}C$ yr B.P.) | Lab number |
|---|---|---|---|---|---|---|---|---|
| Zapp's I–1 | 3331.990 | 723.160 | 1.51 | 337–340 | 1 fragment of *Nyssa aquatica* wood | –27.9 | 2441 ± 32† | UtC–11395 |
| Zapp's II–1 | 3331.970 | 723.160 | 1.46 | 330–334 | 8 charcoal fragments | N/A‡ | N/A‡ | N/A‡ |
| Zapp's III–1 | 3331.960 | 723.160 | 1.49 | 339–343 | 10 Cyperaceae achenes | –26.5 | 2990 ± 50 | UtC–11396 |
| Zapp's IV–1 | 3331.810 | 723.150 | 1.02 | 378–382 | 6 *Scirpus* spp. achenes | –28.0§ | 3500 ± 70 | UtC–11397 |
| Zapp's V–1a | 3331.790 | 723.150 | 1.39 | 384–388 | 5 charcoal fragments | –27.9 | 3603 ± 47 | UtC–11398 |
| Zapp's V–1b | 3331.790 | 723.150 | 1.39 | 384–388 | 5 charcoal fragments | –12.9 | 3712 ± 40 | UtC–11399 |
| Zapp's V–2 | 3331.790 | 723.150 | 1.39 | 381–384 | 1 *Nyssa aquatica* stone | –28.9 | 2539 ± 35† | UtC–11400 |
| Zapp's VI–1 | 3331.760 | 723.145 | 1.41 | 375–378 | 1 fragment of *Nyssa aquatica* wood | –28.7 | 2790 ± 33† | UtC–11401 |
| Zapp's VII–1 | 3331.740 | 723.145 | 1.39 | 366–369 | 5 *Rhynchospora* sp. achenes | –27.6 | 3330 ± 60 | UtC–11402 |
| Zapp's VIII–1 | 3331.710 | 723.145 | 1.24 | 344–347 | 1 *Rhynchospora* sp. achene | –31.2 | 3120 ± 50 | UtC–11427 |
| Zapp's IX–1a | 3331.190 | 723.100 | 1.16 | 390–394 | 30 *Scirpus* spp. achenes | –27.8 | 3728 ± 41 | UtC–11428 |
| Zapp's IX–1b | 3331.190 | 723.100 | 1.16 | 390–394 | 20 *Scirpus* spp. achenes | –27.7 | 3841 ± 38 | UtC–11429 |
| Zapp's X–1 | 3331.160 | 723.105 | 1.04 | 395–398 | 8 charcoal fragments | –21.1 | 3818 ± 36 | UtC–11430 |
| Zapp's XI–1 | 3331.085 | 723.045 | 1.34 | 447–450 | 7 charcoal fragments | –14.1 | 4419 ± 42† | UtC–11431 |
| Zapp's XII–1a | 3331.065 | 723.045 | 1.27 | 444–448 | 1 charcoal fragment | –12.3 | 3797 ± 40 | UtC–11432 |
| Zapp's XII–1b | 3331.065 | 723.045 | 1.27 | 444–448 | 6 charcoal fragments | –12.8 | 4059 ± 43 | UtC–11433 |
| Zapp's XIII–1 | 3330.820 | 723.030 | 0.94 | 430–434 | 5 *Rhynchospora* sp. achenes | –26.9 | 4152 ± 39 | UtC–11434 |
| Gramercy I–1 | 3328.560 | 723.250 | 1.16 | 635–638 | 7 charcoal fragments | –16.2 | 5471 ± 45 | UtC–11151 |
| Gramercy II–1 | 3328.500 | 723.220 | 1.22 | 672–676 | 4 charcoal fragments | –12.7 | 5620 ± 47 | UtC–11152 |
| Gramercy III–1 | 3328.470 | 723.230 | 1.25 | 691–694 | 12 *Scirpus* spp. achenes | –25.9 | 5584 ± 44 | UtC–11153 |
| Gramercy IV–1 | 3328.350 | 723.260 | 1.63 | 786–790 | 6 charcoal fragments, 1 *Scirpus* spp. achene | –16.3 | 5980 ± 70 | UtC–11154 |
| Gramercy V–1a | 3328.310 | 723.270 | 1.60 | 811–814 | 8 *Eleocharis* sp. achenes, 5 *Scirpus* spp. achenes | –27.8 | 6002 ± 48 | UtC–11155 |
| Gramercy V–1b | 3328.310 | 723.270 | 1.60 | 811–814 | >10 charcoal fragments | –19.3 | 5646 ± 50† | UtC–11156 |
| Gramercy VI–1a | 3328.240 | 723.300 | 1.80 | 845–851 | 28 *Scirpus* spp. achenes | –28.6 | 6101 ± 38 | UtC–11157 |
| Gramercy VI–1b | 3328.240 | 723.300 | 1.80 | 845–851 | 6 *Eleocharis* spp. achenes | –29.6 | 6313 ± 50 | UtC–11158 |
| Gramercy VII–1a | 3328.020 | 723.350 | 1.82 | 795–799 | 1 charcoal fragment | –15.0 | 5980 ± 50 | UtC–11159 |
| Gramercy VII–1b | 3328.020 | 723.350 | 1.82 | 795–799 | >10 charcoal fragments | –12.8 | 5878 ± 38 | UtC–11160 |
| Gramercy VIII–1 | 3327.940 | 723.370 | 1.89 | 784–787 | >10 charcoal fragments | –19.1 | 5795 ± 46 | UtC–11161 |
| Gramercy IX–1 | 3329.720 | 723.100 | 1.45 | 510–514 | 4 charcoal fragments | –24.6 | 4290 ± 44 | UtC–11435 |
| Gramercy X–1 | 3329.620 | 723.120 | 1.42 | 502–505 | >10 charcoal fragments | –20.3 | 4461 ± 46 | UtC–11436 |
| Gramercy XI–1 | 3329.550 | 723.150 | 1.45 | 520–523 | 6 *Scirpus* spp. achenes | –25.1 | 4630 ± 80 | UtC–11437 |
| Gramercy XII–1 | 3329.270 | 723.220 | 0.81 | 473–476 | 20 *Scirpus* spp. achenes | –27.9 | 4539 ± 47 | UtC–11438 |
| Gramercy XIII–1 | 3329.090 | 723.250 | 0.90 | 512–515 | 12 *Scirpus* spp. achenes | –28.0 | 4840 ± 110 | UtC–11439 |
| Gramercy XIV–1 | 3329.000 | 723.250 | 1.12 | 562–566 | 10 *Scirpus* spp. achenes | –28.2 | 4880 ± 60 | UtC–11440 |
| Gramercy XV–1 | 3328.930 | 723.290 | 0.95 | 570–573 | 21 *Scirpus* spp. achenes | –26.5 | 5106 ± 47 | UtC–11441 |
| Lutcher II–1a | 3326.100 | 717.940 | 1.7# | 380–390 | A–horizon (fraction >180 μm) | –23.8 | 3460 ± 140 | GrN–22112 |
| Lutcher II–1b | 3326.100 | 717.940 | 1.7# | 380–390 | A–horizon (alkali extract) | –23.1 | 3490 ± 230 | GrN–22378 |
| Lutcher II–1c | 3326.100 | 717.940 | 1.7# | 380–390 | A–horizon (residue) | –22.5 | 4110 ± 100† | GrN–22381 |
| Lutcher III–1a | 3326.620 | 719.900 | 1.0# | 820–830 | A–horizon (fraction >180 μm) | –15.7 | 6690 ± 270 | GrN–22113 |
| Lutcher III–1b | 3326.620 | 719.900 | 1.0# | 820–830 | A–horizon (alkali extract) | –14.3 | 6690 ± 150 | GrN–22380 |
| Lutcher III–1c | 3326.620 | 719.900 | 1.0# | 820–830 | A–horizon (residue) | –14.1 | 6680 ± 130 | GrN–22379 |
| Lutcher IV–1a | 3325.300 | 720.620 | 1.96 | 939–942 | 21 *Scirpus* spp. achenes | –27.0 | 6144 ± 48 | UtC–11142 |
| Lutcher IV–1b | 3325.300 | 720.620 | 1.96 | 939–942 | 1 charcoal fragment | –12.2 | 6220 ± 47 | UtC–11143 |
| Lutcher V–1 | 3325.150 | 720.700 | 2.00 | 1025–1028 | >10 charcoal fragments | –22.0 | 6490 ± 60 | UtC–11212 |
| Lutcher VI–1a | 3325.080 | 720.740 | 1.94 | 1104–1106 | 8 *Scirpus* spp. achenes | –26.7 | 6560 ± 46 | UtC–11144 |
| Lutcher VI–1b | 3325.080 | 720.740 | 1.94 | 1104–1106 | 1 charcoal fragment | –26.8 | 6519 ± 44 | UtC–11145 |
| Lutcher VI–1c | 3325.080 | 720.740 | 1.94 | 1104–1106 | 2 charcoal fragments | –16.0 | 6584 ± 49 | UtC–11146 |
| Lutcher VI–1d | 3325.080 | 720.740 | 1.94 | 1104–1106 | 8 charcoal fragments | –12.4 | 6537 ± 44 | UtC–11147 |
| Lutcher VII–1a | 3325.060 | 720.750 | 1.79 | 1124–1127 | 7 *Scirpus* spp. achenes | –27.8 | 6722 ± 41 | UtC–11148 |
| Lutcher VII–1b | 3325.060 | 720.750 | 1.79 | 1124–1127 | 1 charcoal fragment | –27.7 | 6700 ± 50 | UtC–11213 |
| Lutcher VII–1c | 3325.060 | 720.750 | 1.79 | 1124–1127 | 3 charcoal fragments | –12.0 | 6760 ± 60 | UtC–11149 |
| Lutcher VIII–1 | 3324.980 | 720.800 | 2.00 | 1219–1222 | 4 *Rhynchospora* sp. achenes | –27.3 | 7080 ± 47 | UtC–11150 |

†Rejected (see text).
‡No measurement due to insufficient material.
§Estimated.
#Surface elevation inaccurate (derived from topographic map).

TABLE 3. WEIGHTED MEAN $^{14}$C AGES OF SAMPLES FOR WHICH MULTIPLE SUBSAMPLES WERE DATED

| Sample name | Number of subsamples | Weighted mean $^{14}$C age |
|---|---|---|
| Zapp's V-1 | 2 | 3666 ± 30 |
| Zapp's IX-1 | 2 | 3789 ± 28 |
| Zapp's XII-1 | 2 | 3919 ± 29 |
| Gramercy VI-1 | 2 | 6179 ± 30 |
| Gramercy VII-1 | 2 | 5915 ± 30 |
| Lutcher II-1 | 2 | 3470 ± 120 |
| Lutcher III-1 | 3 | 6685 ± 90 |
| Lutcher IV-1 | 2 | 6183 ± 34 |
| Lutcher VI-1 | 4 | 6548 ± 23 |
| Lutcher VII-1 | 3 | 6723 ± 28 |

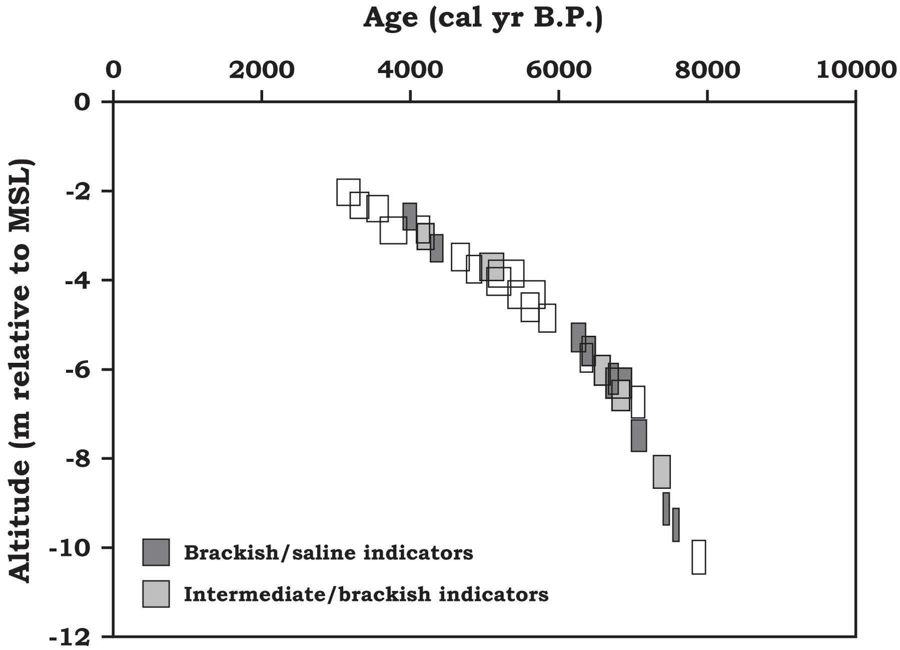


**Figure 7. Relative sea-level curve for the Lutcher-Gramercy area on the eastern margin of the Mississippi Delta. Each sea-level index point is represented by an error box, defined such to incorporate the rise of mean sea level with at least 95% probability.**

peat accumulation between MSL and MHW. The exceptionally low gradient of the GWL characteristic of deltaic environments (cf. Van Dijk et al., 1991; Van de Plassche, 1995; Cohen, 2003) makes it likely that the freshwater peats also formed within this narrow range.

## DISCUSSION

### Trend of Relative Sea-Level Rise

Our data set of 29 SL index points for the time interval 8000–3000 cal yr B.P. enables a test of the proposed "stair-step" pattern of RSL rise in the Mississippi Delta and surrounding Gulf Coast. The results (Fig. 7) suggest that RSL rose steadily throughout much of the Holocene and merely leave room for the possibility of relatively minor stillstands and accelerations. It is important to stress here that the basal-peat strategy is of particular use in settings with continuous RSL rise. RSL falls or prolonged stillstands would be reflected by unconformities (soils or oxidation horizons), neither of which have been encountered in our study area (Figs. 3 and 4). Conversely, it must be recognized that there is a limit with respect to the rate of RSL rise that peat formation can keep up with. Considering the fact that we have a laterally continuous, basal organic-rich unit, the existence of short-lived pulses of abrupt RSL rise is highly unlikely. Rapid RSL rises would have led to transgressive interruption of peat growth, most likely by a lagoonal deposit, and an associated landward shift of the peat-forming ecosystem.

Thus, we find no support for the presence of abrupt, meter-scale RSL rises alternating with RSL stillstands on the order of millennia (cf. Curray, 1961; Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Penland et al., 1991; Thomas and Anderson, 1994) and our evidence is more consistent with the classic studies of RSL change in the Mississippi Delta by McFarlan (1961) and Coleman and Smith (1964). In other words, our data are compatible with observations worldwide. Nevertheless, the envelope described by our error boxes leaves open the possibility for smaller scale RSL fluctuations with amplitudes of <1 m and possible frequencies on the order of centuries. However, for the time interval concerned we find no support for the existence of RSL fluctuations with amplitudes as high as several meters, such as have been reported elsewhere along the northern U.S. Gulf Coast (e.g., Holmes and Trickey, 1974; Stapor et al., 1991; Tanner, 1992; Walker et al., 1995). Our data also call into question the numerous claims of a direct causal relationship between RSL fluctuations and the dynamics of Mississippi River subdeltas (e.g., Frazier, 1974; Fairbridge, 1983; Penland et al., 1991; Tanner, 1992).

In our first analysis of this data set (Törnqvist et al., 2002) we fitted an exponential curve through the SL index points, yielding an $r^2$ of 0.994. However, visual inspection of the data suggests that RSL rise was particularly rapid prior to ~7000 cal yr B.P. (~3.5 mm yr$^{-1}$), then dropped rather abruptly to values of ~1.5 mm yr$^{-1}$. This is consistent with global assessments of eustatic sea-level change that have inferred an abrupt decrease (Fleming et al., 1998) or termination (Peltier, 2002b) of the glacial meltwater signal around this time. This reduction in the rate of RSL rise is represented in far-field regions by a RSL highstand.

### Middle Holocene Highstand

Clearly, our data provide no direct evidence for middle Holocene RSL positions higher than present. This was to be expected, in view of the location of the study area within the overall subsiding Mississippi Delta. However, our data enable a test of the highstand hypothesis, adhering to the assumption that tectonic subsidence rates in the study area were essentially constant. Since tectonic subsidence is mainly driven by the ongoing process of deltaic sediment loading, it is unlikely to exhibit major temporal variability over the relatively short timescale concerned. The following analysis constitutes a comparison of our data with the recently published RSL data from Texas (Blum et al., 2001), and it should be noted that for this comparison the commonly nonlinear glacio-isostatic and hydro-isostatic components are not taken into account. While hydro-isostasy is likely to be relatively small, glacio-isostatic responses in Louisiana and Texas have probably been fairly comparable, in view of the roughly similar distance to the Laurentide Ice Sheet.

We have corrected our RSL curve for a range of hypothetical subsidence rates, with the objective of reproducing a middle Holocene RSL highstand ~2 m higher than present. Correction for a linear subsidence rate of 1.1 mm yr$^{-1}$

(Fig. 8) indeed yields a highstand similar in altitude and timing to the one proposed for the central coast of Texas (Blum et al., 2001). However, the same correction also places our oldest SL index point (8000 cal yr B.P.) less than 2 m below present MSL. This is not only incompatible with virtually any widely used sea-level data from far-field regions (e.g., Edwards et al., 1993; Bard et al., 1996), but also with SL index points of similar age from Texas, located at ~9 m depth (Blum et al., 2001) and strikingly similar to our oldest data.

In addition, we have carried out a more direct comparison between our evidence and three SL index points from the Blum et al. (2001) data set that these authors considered most reliable (Fig. 9). While the oldest data points nearly coincide, the two RSL curves diverge dramatically over the subsequent ~1000 yr and become separated by 6–7 m. Considering that the coast of Texas is widely considered to be tectonically relatively stable, at least over the timescale of interest here (Paine, 1993; Blum et al., 2001), the implication would be that our study area would have experienced 6–7 m of uplift during this time interval, a scenario that must be considered extremely unlikely. Therefore, our data do not support the notion of a middle Holocene highstand on the U.S. Gulf Coast, similar to what has been proposed recently for eastern Texas (Rodriguez et al., 2004). This finding is also consistent with numerous studies in Florida (Scholl and Stuiver, 1967; Scholl et al., 1969; Robbin, 1984; Parkinson, 1989; Goodbred et al., 1998) and elsewhere on the northeastern U.S. Gulf Coast (Otvos, 1995, 2001). Comparison of our deepest SL index point with the oldest data from Texas from Blum et al. (2001) (Fig. 9) suggests that inland portions of the Mississippi Delta, like our study area, may be surprisingly tectonically stable. This is confirmed by the similarity of our RSL curve with the recent compilation for the Caribbean by Toscano and Macintyre (2003). Considering the general belief that the eustatic component of sea-level rise has been limited during the past 7000 cal yr (Fleming et al., 1998; Peltier, 2002b), the implication would be that the U.S. Gulf Coast is still responding glacio-isostatically, by means of forebulge collapse, to the melting of the Laurentide Ice Sheet.

Because our data set does not cover the past 3000 cal yr, we cannot entirely rule out the possibility of a RSL highstand during that time interval (cf. Otvos, 2001). Interestingly, Morton et al. (2000) described a range of coastal features in Texas that they interpret as indicating a RSL higher than present, with perhaps the most compelling feature being an elevated beach ridge $^{14}$C dated at ~1500 cal yr B.P. Most recently, a

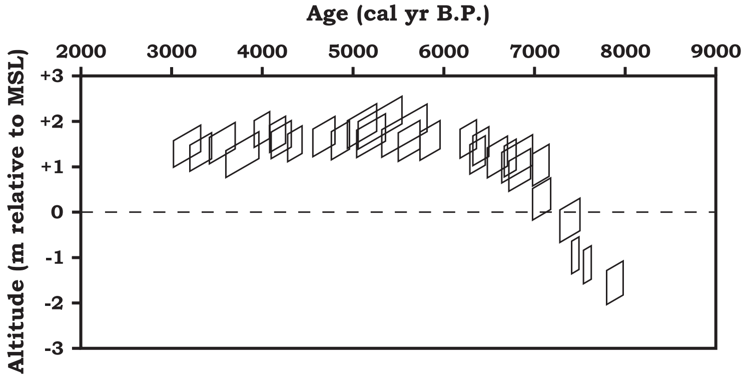


**Figure 8. Correction of the relative sea-level curve of Figure 7 for a hypothesized, constant tectonic subsidence rate of 1.1 mm yr$^{-1}$.**

study in southern Florida argued in favor of the presence, around 1000–2000 $^{14}$C yr B.P., of a RSL at least 0.5 m higher than present (Froede, 2002). Many of the presumed indicators of RSL highstands that have been presented previously (e.g., Behrens, 1966; Stapor et al., 1991; Tanner, 1992; Walker et al., 1995) also concern the late Holocene. Clearly, there is a need for continuous, high-resolution records of RSL change to test the hypothesis of one or more late Holocene RSL highstands.

It is increasingly clear that two schools of thought have evolved with respect to the U.S. Gulf Coast Holocene RSL history, with one that invokes one or more Holocene highstands and a second (including the present study) that advocates continuous submergence. It has been suggested recently (Blum and Carter, 2000; Blum et al., 2001, 2002) that such contrasting interpretations may relate to the distinctly different settings in which RSL data are collected (i.e., now-submerged contexts versus emergent landforms). In the case of the U.S. Gulf Coast, the majority of investigations that have favored the RSL highstand model involved the study of beach ridges and related coastal features. In this context, it is interesting to note that several investigations elsewhere have provided RSL data from coastal-barrier environments that closely corroborate basal-peat $^{14}$C records (e.g., Roep and Beets, 1988; Van de Plassche and Roep, 1989; Van Heteren et al., 2000). In other words, the idea that contrasting interpretations would be an inevitable result of methodological differences can be questioned.

A key challenge for future studies is to reconcile the apparently synchronous occurrence of elevated beach ridges and basal-peat formation well below present MSL. Although it has been postulated that elevated beach ridges in the Gulf of Mexico are unlikely to be the result of fluctuations in wave energy (e.g., Stapor et al., 1991), the possibility of a primarily climatic cause of these phenomena may deserve renewed attention.

## CONCLUSIONS

We present the first detailed RSL curve from the northern Gulf of Mexico based entirely on $^{14}$C ages of basal peat, a sea-level indicator that is increasingly recognized as one of the most powerful proxies in sea-level reconstruction. The overall trend of RSL rise is consistent with observations worldwide and calls into question the widely assumed presence of major RSL stillstands alternating with abrupt jumps, at least for the time interval 8000–3000 cal yr B.P. Despite the fact that our data set exhibits continuous submergence, the high level of detail of this time series enables a rigorous test of the hypothesis of a middle Holocene RSL highstand, ~2 m

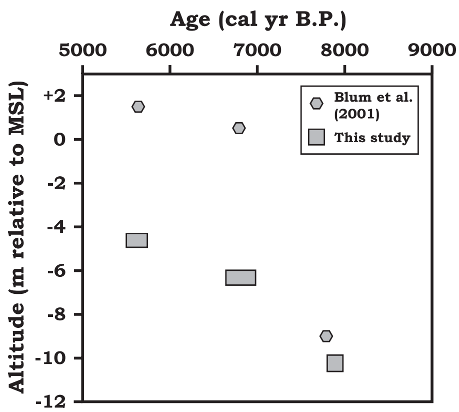


**Figure 9. Direct comparison of sea-level index points of similar age from the Mississippi Delta (this study) and central Texas (Blum et al., 2001).**

above present MSL, on the U.S. Gulf Coast. Assuming a constant tectonic subsidence rate, we demonstrate that such a scenario cannot be reconciled with our data. Nevertheless, our evidence cannot entirely rule out the possibility of a RSL highstand later during the Holocene.

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