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# FLUVIAL TERRACES OF THE LITTLE RIVER VALLEY, ATLANTIC COASTAL PLAIN, NORTH CAROLINA

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## ABSTRACT

**An optically-stimulated luminescence (OSL) and radiocarbon chronology is presented for fluvial terraces of the Little River, a tributary to the Cape Fear River that drains 880 km² of the Sandhills Province of the upper Coastal Plain of North Carolina. This study differs from previous work in the southeastern Atlantic Coastal Plain in that numerical age estimates are provided for all terraces in the valley of mappable extent by direct dating of fluvial sediments. The Little River valley contains a floodplain and five fluvial terraces with average heights above modern river bed level that range from 3.0 m (T1) to 29.0 m (T5b). Dating indicates the floodplain has a late Holocene (1.3 ± 0.3 ka) to historical age while terraces range in age from 9.9 ± 2.0 ka (T1) to 94.0 ± 15.9 ka (T5b). Age separation of the six fluvial surfaces is corroborated by distinct differences in soil morphology and chemistry. Terrace heights above modern river level and terrace ages indicate a long-term net incision rate of 0.29 mm/yr during the last 100 ka. This rate is nearly an order of magnitude higher than late Pleistocene uplift rates reported for the Cape Fear River valley in the 1980s, based on age estimates for the Wando Formation. However, the 0.29 mm/yr rate is consistent with OSL and radiocarbon dates reported from terraces in the adjacent Pee Dee River valley. Together, these data refine the ages assigned to fluvial facies of the Wando terrace and suggest it is composed of multiple fluvial deposits with a wide range of late Pleistocene ages. The long-term net incision rate of the Little River is consistent with the range of neotectonic uplift rates reported for this region. Scrutiny of the inter-terrace ages indicates that Little River terraces may reflect short-term aggradation in response to periodic climate-mediated increases in sediment supply that is compensated by long-term incision in response to neotectonic uplift.**

## INTRODUCTION

Fluvial deposits are of great interest to geomorphologists because they provide information about river behavior over long timescales and may contain evidence of past river responses to external forcing mechanisms, such as tectonics, climate change, eustasy, or human impacts (Jacobson et al., 2003; Knighton, 1998). If fluvial deposits can be dated and placed in a chronological framework, then potential drivers and rates of channel change may be evaluated.

The southeastern Atlantic Coastal Plain is composed of unconsolidated, sandy marine sediments, and rivers of this province typically form wide valleys capable of accommodating lateral channel migration while at the same time preserving former deposits. As a result, river valleys in the Coastal Plain commonly contain a relatively well-preserved record of late Quaternary fluvial deposits. This stands in contrast to river valleys in the adjacent Piedmont province, where resistant igneous and metamorphic rocks confine rivers to narrow valleys and inhibit preservation of complete depositional records. Numerous studies have reported age estimates for fluvial terraces in the southeastern Atlantic Coastal Plain (Howard et al., 1993; Leigh et al., 2004; Leigh and Feeney, 1995; Leigh, 2006; Leigh, 2008; Markewich et al., 1987; Markewich et al., 1988; Thom, 1967), but relatively few workers have provided a chronology that encompassed all the terraces present in

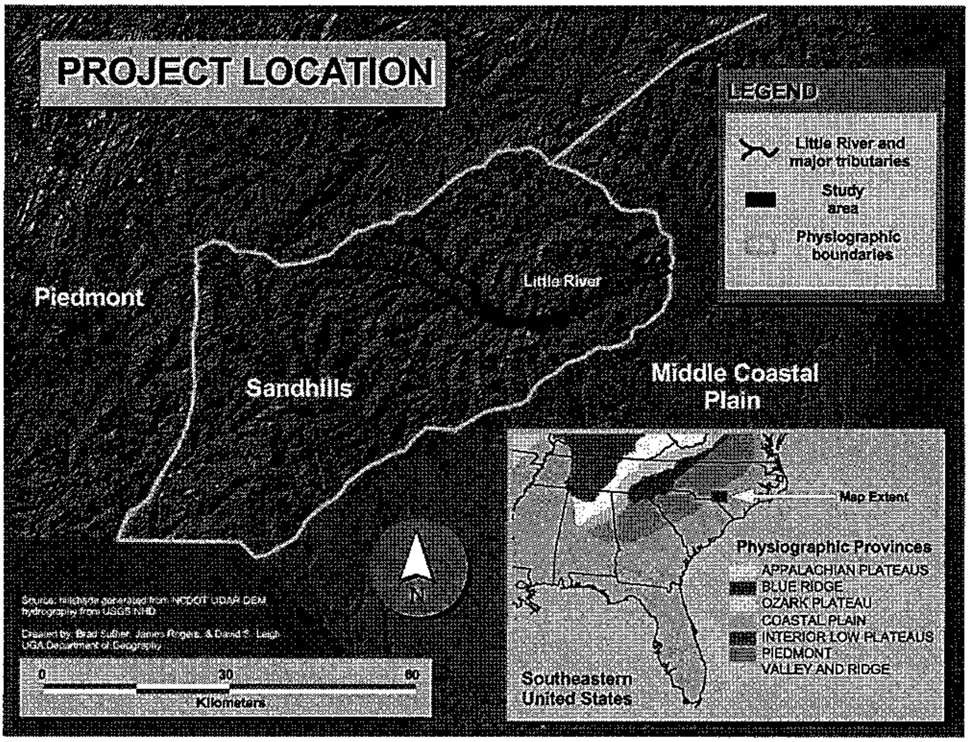



**Figure 1. Shaded relief map showing the location of the study area in the Sandhills province of the upper Coastal Plain of North Carolina. Inset shows location of study area within the southeastern United States.**

a given valley (Owens, 1989; Soller, 1988). Thus, one objective of this paper is to establish a chronology for fluvial terraces of the Little River, a tributary of the Cape Fear River located in the upper Coastal Plain of North Carolina. This study differs from previous work in the region in that numerical age estimates are provided for all fluvial terraces in the valley by direct dating of fluvial sediment using optically-stimulated luminescence (OSL). Another objective is to present a long-term net incision rate for the Little River for the last 100 ka based on terrace age estimates. The incision rate documented by this study is significant in that it differs from uplift rates reported for terraces of the Cape Fear River (Soller, 1988) and provides insights into incision over late Quaternary timescales in the region.

## STUDY AREA

The Little River is a tributary to the Cape Fear River and drains 880 $km^2$ of the Sandhills province of the upper Coastal Plain of North Carolina (Figure 1). The Sandhills constitute a highly dissected landscape of deeply-weathered, sandy, quartz-rich soils that formed from marine and fluvial deposits of late Cretaceous to Tertiary age (Horton & Zullo, 1991). Surficial geologic mapping indicates that uplands within the Little River basin are comprised of interbedded kaolinitic clays and clayey sands of late Cretaceous age that are overlain in various locations by unconsolidated Tertiary sands and gravels known as the Pinehurst Formation (Conley, 1962). Triassic sandstone and siltstone, felsic tuffs, and late Cretaceous fine sands and clays of marine origin outcrop along the valley sides of the upper Little River and its major tributaries (Conley, 1962). The segment of valley studied is approximately 17 km long and 2 – 3 km wide and is located in the vicinity of the Fort Bragg Military Reservation (Figure 2). The valley contains a floodplain and terraces

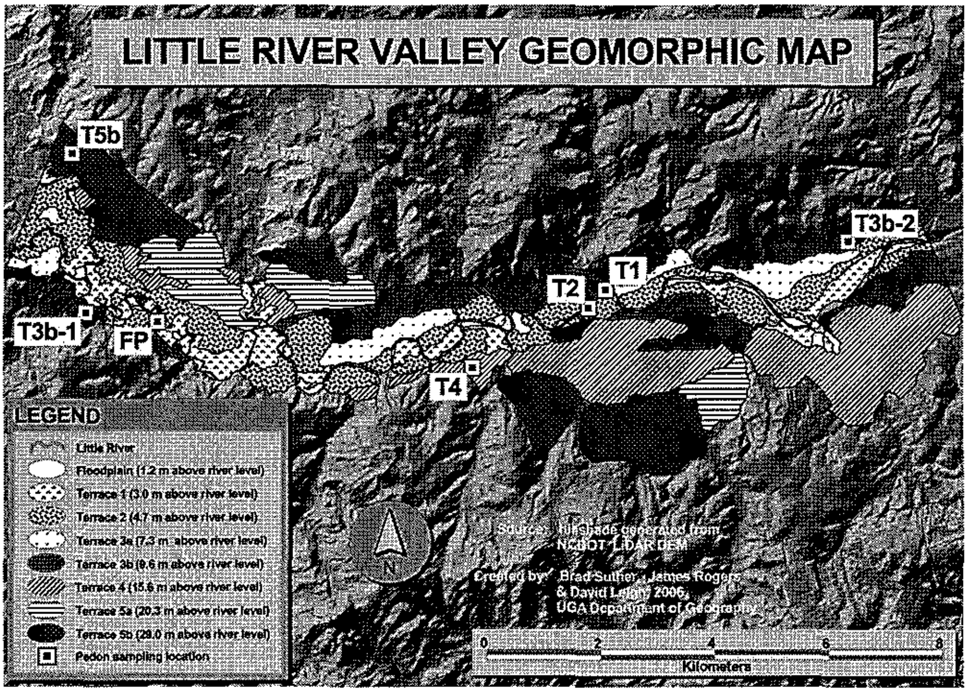



**Figure 2. Geomorphic map of the study area. Vertical exaggeration of shaded relief is 3X.**

with soils of varying degrees of development that are mapped as ranging from Entisols (Typic Fluvaquents) on the floodplain to Ultisols (Arenic and Typic Hapludults) on higher terraces by the Natural Resource Conservation Service (NRCS) Soil Survey (Hudson, 1984; Wyatt, 1995). The majority of the study area is forested and is dominated by longleaf pine (*Pinus palustris*) savanna, though residential and agricultural landuses are present. The climate is humid subtropical. Average annual precipitation is 117.9 cm, and the average daily temperature is 16.9 °C, with an average January temperature of 6.1 °C and an average July temperature of 27.4 °C (NCDC, 2004). The modern channel has a sandy bed and meandering pattern, with an average slope of 0.00057, a width of 10 to 25 m, and sinuosity of 1.59. The downstream boundary of the study area is located approximately 25 river kilometers upstream of the confluence with the Cape Fear River and is approximately 130 km inland from the present head of the Cape Fear River estuary.

## METHODS

The terrace chronology presented in this paper was established to provide a geomorphic and chronologic framework for a soil chronosequence study that evaluated changes in soil properties over time by comparing soils developed on terraces of differing ages (Suther, 2006). Geomorphic mapping was accomplished by interpretation of hillshade and slope derivatives of 2 m and 6 m pixel edge Light Detection and Ranging (LIDAR) digital elevation models provided by the Department of the Army and the North Carolina Department of Transportation (2006), respectively, and US Geological Survey 7.5' topographic quadrangles, followed by field-verification based on visual inspection of the landscape and examination of soil and stratigraphic profiles. Locations of the soil and stratigraphic descriptions and OSL ages presented in this paper were selected to facilitate the soil chronosequence study and were situated to ensure comparability between terrace soils and to reduce the effects of local drainage conditions and topography on soil development.

Because this research was conducted as a component of a soil chronosequence study and a comprehensive characterization of the fluvial history of the Little River was beyond its scope, extensive subsurface investigation to determine the alluvial architecture of the valley was not conducted. Descriptions of soils, sediments, and stratigraphy of terrace deposits were made in backhoe pits at each OSL sampling location using standard NRCS terminology (Soil Survey Division Staff, 1993). Soil descriptions of hand auger samples for a minimum of four additional profiles per terrace were conducted to account for soil variability and are reported elsewhere (Suther, 2006).

## OSL Dating

### Principles of OSL Dating

OSL dating estimates the time elapsed since sediment was last exposed to sunlight (Stokes and Walling, 2003). After sediment is buried, ionizing radiation from the decay of naturally occurring radionuclides (U, Th, $^{K}40$) in surrounding sediments, and to a lesser extent from cosmic rays, results in the eviction of some electrons in quartz (and other silicate minerals) from their ground state (Aitken, 1998). Evicted electrons become trapped within imperfections in the crystal lattices of these minerals and accumulate over time ($\geq 10^5$ years) until they are exposed to an amount of light or heat energy sufficient to stimulate their return to ground state (Forman et al., 2000). Return to ground state results in emission of excess energy in the form of a photon (luminescence) that is proportional to the total amount of ionizing radiation to which the sample was exposed during burial (Stokes and Walling, 2003). If the irradiation necessary to produce a given luminescence signal can be reconstructed in the laboratory (equivalent dose), and if an annual dose rate from surrounding sediments and cosmic ray contributions can be estimated, a sample age that approximates time of deposition may be calculated by the following equation (modified from Forman et al., 2000)

$$\text{age (ka)} = \text{equivalent dose (Gy)} / \text{dose rate (Gy/ka)} \qquad (1)$$

Where

$$\text{dose rate} = (aD_\alpha W + D_\beta W + D_\gamma W) + D_c \qquad (2)$$

$W$ = water content factor

$a$ = alpha radiation attenuation coefficient

$D_\alpha$ = alpha radiation

$D_\beta$ = beta radiation

$D_\gamma$ = gamma radiation

$D_c$ = cosmic radiation

The single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000) was used to determine equivalent dose. In this procedure, following measurement of an aliquot's natural luminescence signal, each sample aliquot is subjected to a series of irradiation–pre-heat–stimulation cycles that enable construction of a regenerated growth curve that relates luminescence signal intensity to irradiation dose for each aliquot. The regenerated growth curve is then used to determine an aliquot's equivalent dose. The equivalent dose is the irradiation dose that the growth curve predicts is required to produce the natural luminescence signal observed for a given aliquot. During irradiation–pre-heat–stimulation cycles, the effects of repeat irradiations on luminescence (recuperation effects) are evaluated by determining whether the luminescence intensity generated by a given irradiation dose remains consistent. If necessary, corrections for changes resulting from recuperation effects are incorporated into the equivalent dose estimate.

An underlying assumption of OSL dating is that a sample's luminescence signal is zeroed (bleached) prior to burial (Stokes and Walling, 2003). OSL dating of alluvium may be confounded by partial and/or inhomogeneous bleaching of grains during fluvial transport and deposition (Murray et al., 1995; Olley, 1998, 1999; Stokes and Walling, 2003). However, a number of studies indicate that reliable OSL age estimates can be obtained from fluvial environments (Colls et al., 2001; Leigh et al., 2004; Rittenour et al., 2003, 2005; Srivastava et al., 2001; Stokes et al., 2001; Wallinga et al., 2001). Equivalent dose frequency distributions for Little River samples were inspected for asymmetry indicative of partial bleaching (Olley et al., 1998, 1999). In general, distributions were approximately Gaussian in form, suggesting relatively uniform bleaching of grains. Therefore, the mean equivalent dose was used to calculate OSL age estimates. All ages are reported with two standard deviation error estimates (± 2 sigma).

## OSL Dating Procedures

Samples for OSL dating were obtained for at least one terrace per terrace level within the valley (Figure 2). Samples were collected by driving a 4 cm diameter heavy gauge PVC tube laterally into the face of each pit below the depth of bioturbation. C horizon material (unweathered alluvium) was sampled where possible. Sample tubes were sealed immediately after sampling to prevent exposure to light. Sample preparation and handling for OSL dating were carried out in subdued red-light conditions. Five centimeters of sediment were removed from each end of the PVC sample tubes for dose rate estimation. Luminescence measurements were made on samples from the central section of the PVC cylinder that was least likely to have been exposed to sunlight during sampling. All samples were treated with 10% HCl and 30% $H_2O_2$ to remove carbonates and organic matter. Samples were sieved to extract the 150-170 μm-size fraction. Quartz and feldspar grains were separated by density with Na-polytungstate (ρ=2.58 g $cm^{-3}$). The quartz fraction was etched using 48% HF for 80 min followed by 36% HCl for 30 min to remove the outer surface affected by alpha radiation. The quartz grains were mounted on stainless steel discs using Silkospray™.

Light stimulation of the quartz was achieved using a RISØ array of blue LEDs centered at 470 nm. Detection optics comprised two Hoya 2.5 mm thick U340 filters and a 3 mm thick Schott GG420 filter coupled to an EMI 9635 QA photomultiplier tube. Measurements were taken with a RISØ TL-DA-15 reader. A 25-mCi $^{90}Sr/^{90}Y$ built-in source was used for sample irradiation.

The single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000) used to determine equivalent dose involved a five-point regenerative dose strategy with three dose points to bracket the equivalent dose, a fourth zero dose to test for recuperation effects, and a fifth repeat dose, usually of the smallest change correction incorporated in the SAR protocol. All measurements were made at 125°C for 100 s after a pre-heat to 220°C for 60 s. For all aliquots, the recycling ratio between the first and the fifth point ranged within 0.95-1.05. Data were analyzed using the ANALYST program of Duller (1999).

Equivalent dose measurements were made on single aliquots of 9.6 mm diameter. Typically 15 to 25 aliquots per sample were analyzed. The dose rate calculation relied on the thick source ZnS (Ag) alpha counting technique for elemental concentration of uranium and thorium. Potassium was measured by ICP90, with a detection limit of 0.01%, using the Sodium Peroxide Fusion technique at the SGS Laboratory in Toronto, Canada. The cosmic ray Gamma contribution was assumed to be 150 ± 30 μGy/yr, as recommended for sediments located below an altitude of 1000 m between latitudes 0° - 40° (Prescott and Stephan, 1982). All sample ages were calculated using assumed pore-water content of 15 ± 5% for the in-situ sediment.

## Incision Rate Calculations

A net incision rate for the time interval represented by the entire terrace sequence was calculated using the equation

$$IR_{net} = E_{T5b} - E_{mb} / {}_{e}A_{T5b} \quad (3)$$

Where:

$IR_{net}$ = net incision rate
E = elevation
T5b = upper boundary of channel lag sediments of highest terrace
mb = bed of modern river
${}_{e}A$ = estimated age

Net incision rates between intervals of terrace deposition were calculated in a similar manner, by dividing the difference in elevation of the top of channel lag sediments between individual terrace sampling locations by the difference in terrace mean age estimates. The upper boundary of channel lag sediments was chosen as the vertical datum for incision rate calculations because it reflects the natural grade of the river during the time of sediment deposition, it is a laterally traceable surface, and because, where gravels are present, it typically coincides with the depth of refusal when coring. The upper boundary of channel lag in stratigraphic profiles was typically indicated by a pronounced increase, relative to overlying sediments, in the percent weight of >2 mm particles accompanied by a predominantly sand-sized (0.063 – 2.0 mm) matrix. Grain size data were obtained for each profile in vertical increments of 25 cm or finer, respecting soil horizon boundaries, using the hydrometer (Gee and Bauder, 1986) and sieve (Ingram, 1971) methods. The elevation of the upper boundary of channel lag was calculated by subtracting the depth to lag from the ground surface elevation at the sampling location. Ground surface elevations were determined from LIDAR data with a vertical accuracy of ±25 cm (NCDOT, 2006). In nearly all instances, OSL samples were obtained from fluvial sands from stratigraphic positions immediately above or within lag deposits, providing age estimates that approximate lag deposition. Three of the four age estimates obtained for T3b (TU5@190, T32A160, and T32A180) were in close stratigraphic association with channel lag deposits that were encountered at an elevation of 7.1 m and 6.7 m above the modern river bed at the T3b-1 and T3b-2 sampling locations, respectively. The average of the statistical means of these age estimates (59.7 ka), and the average elevation of the top of channel lag at the T3b-1 and T3b-2 sampling locations (6.9 m), were used to approximate T3b age and elevation, respectively, for the purpose of incision rate calculations.

## RESULTS

The Little River valley contains a modern floodplain and at least five mappable late Quaternary fluvial terraces (Figure 2). Surface morphology, the soil-stratigraphic profiles associated with OSL sampling locations, and OSL age estimates are described for each terrace below. Morphological attributes of each terrace are summarized in Table 1. OSL age estimates and supporting data are presented in Table 2. Table 3 provides soil profile descriptions of each OSL sampling location.

### Terrace Morphology

The floodplain is the alluvial surface being constructed by the modern regime of the Little River and has an average elevation of 1.2 m above the modern river bed. Average height above the modern river bed for the terrace treads ranges from 3.0 m (T1) to 29.0 m (T5b) (Table 1). Terraces 3 and 5 each have two components (denoted a and b), for which the "a" component constitutes a lower and in some instances more eroded surface and the "b" component constitutes a surface that is higher and in some cases more well-preserved. The floodplain and lower terraces are dominated by indeterminant ridge and swale topography from which former river channel patterns are difficult to discern. However, some T1, T2 and T3a surfaces exhibit scrollwork and paleochannel scars clearly indicative of former meandering channels, while others exhibit interwoven ridge and swale topography indicative of former braided channels. Well expressed scrollwork and paleochannel scars are only evident on terraces in the most upstream segment of the study area, while braided patterns are only found in the downstream segment of the study area (Suther, 2006). This variation in terrace morphology may indicate that the study area represents a transition zone along the Little River with respect to past changes in river regime and channel pattern or that individual terraces contain relatively short lived temporal variations in channel pattern. With the exception of one T3b surface that contains indeterminant ridge and swale topography, depositional topography is not apparent on higher terraces (T3b – T5b). In general, terraces exhibit increasing fluvial dissection with increasing height above river level.

**Table 1. Elevation, extent, and morphology of terraces within the study area.**

| Landform | Avg. height above river bed (m)[a] | Proportion of study area (%)[b] | Terrace treads[c]: Tread morphology | Degree of dissection |
|---|---|---|---|---|
| Floodplain | 1.2 | 1.4 | flat to indeterminant ridge & swale | none |
| Terrace 1 | 3.0 | 9.1 | indeterminant ridge & swale (some surfaces scrolled, others braided) | none |
| Terrace 2 | 4.7 | 19.1 | indeterminant ridge & swale (some surfaces scrolled, others braided) | none to slight |
| Terrace 3a | 7.3 | 6.4 | indeterminant ridge & swale (some surfaces possibly braided) | slight |
| Terrace 3b | 9.6 | 16.0 | flat with no depositional topography (one surface contains indeterminant ridge & swale) | slight to moderate |
| Terrace 4 | 15.6 | 23.2 | where tread is preserved, flat with no depositional topography (on eroded surfaces only remnant ridges remain) | moderate to high |
| Terrace 5a | 20.3 | 9.0 | where tread is preserved, flat with no depositional topography | moderate |
| Terrace 5b | 29.0 | 15.9 | where tread is preserved, flat with no depositional topography (on eroded surfaces only remnant knobs remain) | moderate to high |

[a]Average height of landform surface above modern river bed based on LIDAR DEM data.
[b]Proportion of areal extent of study area represented by each landform.
[c]Qualitative description of terrace tread morphology is based on visual inspection of LIDAR DEM data. Dissection was qualitatively assessed based on the number of intermittent and perennial streams, as mapped on USGS 7.5' topographic quadrangles, that originate on each landform.

## Chronology

The modern floodplain contains historical vertical accretion sands and silts deposited since Euro-American settlement of the Little River valley (approximately 200 yr BP) that overlie prehistoric vertical accretion sands and silts of mid- to late- Holocene age. Soils in the floodplain sampling locality consist of Entisols that typically have A-C-Ab-C' profile sequences and contain multiple buried A horizons. The contact between historical and prehistoric sediment at the floodplain sampling location occurs at a depth of 100 cm. The total thickness of floodplain alluvium was not measured and extends below the maximum observed depth of 220 cm. Assuming that deposition of historical sediment initiated 200 yr BP yields an average historical sedimentation rate of 5 mm/yr for the Little River at this site, which is in agreement with historical sedimentation rates documented in other river valleys in the southeastern United States (Leigh and Webb, 2006; Lichtenstein, 2003; Oppenheim, 1996; Costa, 1975; Happ, 1945).

An OSL sample from prehistoric floodplain vertical accretion sediments at a depth of 200 cm within a C''' horizon yielded an age of 1.3 ± 0.3 ka (LR@200). This sample contained concentrations of uranium and thorium that exceed

**Table 2. Optically-stimulated luminescence (OSL) dates and supporting data.**

| Sample | Mean age (ka) | Landform | Depth (cm) | U (ppm) | Th (ppm) | K (%) | Mean Equivalent Dose (Gy) | Dose rate (Gy/ka) |
|---|---|---|---|---|---|---|---|---|
| LR@200 | 1.3±0.3 | floodplain | 200 | 23.9±0.7 | 85.6±30.7 | 0.96 | 14.5±1.1 | 10.9±2.1 |
| 31CD475-T2 | 9.9±2.0 | terrace 1 | 90-120 | 10.8±1.1 | 17.9±10.6 | 0.37 | 36.3±1.9 | 3.7±0.7 |
| 31CD475-110 | 17.4±4.2 | terrace 2 | 110 | 3.9±1.4 | 14.2±4.9 | 0.42 | 37.2±5.0 | 2.1±0.4 |
| TU5@90 | 40.0±9.9 | terrace 3b-1 | 90 | 2.0±0.6 | 6.9±2.1 | 0.12 | 42.4±7.4 | 1.1±0.2 |
| TU5@190 | 55.2±15.2 | terrace 3b-1 | 190 | 1.7±0.3 | 3.4±1.1 | 0.20 | 46.9±11.5 | 0.9±0.1 |
| T32A160 | 72.7±13.1 | terrace 3b-2 | 160 | 1.6±0.3 | 4.1±0.9 | 0.18 | 63.0±9.1 | 0.9±0.1 |
| T32A180 | 51.3±12.2 | terrace 3b-2 | 180 | 2.60±1.0 | 9.90±3.5 | 0.21 | 74.1±8.0 | 1.4±0.3 |
| T4-161 | 74.6±10.4 | terrace 4 | 161 | 4.0±0.8 | 12.6±2.6 | 0.30 | 145.6±6.4 | 2.0±0.3 |
| T5-135 | 94.0±15.9 | terrace 5b | 135 | 3.8±0.7 | 5.4±2.8 | 0.10 | 123.1±6.6 | 1.3±0.2 |

*Notes*: Sample water contents were assumed to be 15±5% by weight. Cosmic ray contribution was assumed to be 150±30 μGy per annum.

levels typically found in natural systems, which resulted in an unusually high dose rate estimate (Table 2), and suggests possible contamination of the sediments. Because overestimation of dose rate may result in underestimation of the true age of the sample, the authors regard this age with caution. However, age estimates of deposits associated with the next highest geomorphic surface in the terrace sequence, T1, which include a 9.9 ± 2.0 ka OSL date (this paper) and a 10.24 ± 0.03 ka radiocarbon date (Goman and Leigh, 2004, discussed below), coupled with a lack of pedogenic development in floodplain sediments (Table 3, this paper; Suther, 2006), provide context that suggests the prehistoric floodplain is Holocene in age, with a mid- to late- Holocene age being most likely. Assuming the floodplain profile was not truncated by erosion prior to the deposition of historical sediment, an age of 1.3 ± 0.3 ka would suggest that floodplain sedimentation at the sampling location proceeded at an average rate of 0.6 – 1.0 mm/yr between about 1 ka and the initiation of historical sedimentation. Though this value is not unreasonable, it is on the upper end of typical prehistoric sedimentation rates observed in river valleys of the southeastern United States (Leigh and Webb, 2006; Lichtenstein, 2003; Leigh, 1996).

Terrace 1 deposits exhibit roughly normal grading and consist of lateral accretion medium and coarse gravels and coarse sands that fine upward to vertical accretion sands. Soils in the T1 sampling locality consist of Inceptisols containing eluviated (E) horizons above incipient B horizons that lack clay coatings but have weak, medium subangular blocky structure and show signs of iron oxide accumulation. An early Holocene to terminal Pleistocene age (9.9 ± 2.0 ka, 31CD475-T2) is indicated for T1 deposits by an OSL sample taken from sandy lateral accretion deposits in a C1 horizon at a depth of 90 – 120 cm near the location of the T1 backhoe pit (description provided in Table 3). The calculated optical age estimate is in agreement with independent $^{14}C$ dates of peat sampled from the thalweg and point bar of an abandoned paleochannel on a T1 unit about 2.2 km west of the T1 OSL sampling location (Goman and Leigh, 2004). The basal age of peat was found to be 10.24 ± 0.03 ka (Beta-151615) in the thalweg of the paleochannel and 9.23 ± 0.19 ka (Beta-152729) on the point bar. Goman and Leigh note the discrepancy in age between the thalweg and point bar basal peats likely reflects the control of depositional setting on peat accumulation, with peat accumulating first in the low elevation thalweg and later above the point bar, which is at a slightly higher elevation.

Terrace 2 deposits are normally graded and consist of lateral accretion deposits of medium gravels and coarse sands that fine upward to vertical accretion sands. Soils in the T2 sampling locality are predominately Ultisols that contain eluviated (E) horizons underlain by argillic (Bt) horizons with moderate medium sub-

**Table 3. Soil profiles at OSL sampling locations. These profiles are representative of those characterized by Suther (2006) and are typical of soils found on the middle part of the upper surface of each respective terrace in the most well-drained and least eroded landscape position.**

| Landform | Horizon | Depth (cm) | Moist matrix color[a] | (>2 mm)[b] (%) | Sand[c] (%) | Silt[c] (%) | Clay[c] (%) | Structure[d] |
|---|---|---|---|---|---|---|---|---|
| FP | A | 0-7 | 10YR 4/2 | 0.0 | 85.9 | 9.6 | 4.5 | wk med gr |
| | C1 | 7-24 | 10YR 6/4 | 0.0 | 92.1 | 5.1 | 2.7 | wk med gr |
| | C2 | 24-37 | 10YR 5/4 | 0.0 | 84.6 | 12.1 | 3.2 | wk med gr |
| | A'b | 37-47 | 10YR 4.5/4 | 0.0 | 92.0 | 5.2 | 2.8 | wk med gr |
| | C'1 | 47-58 | 10YR 7/3 | 0.0 | 97.5 | 1.8 | 0.7 | sg |
| | C'2 | 58-78 | 10YR 6/6 | 0.0 | 95.0 | 3.3 | 1.7 | wk fn gr |
| | C'3 | 78-100 | 10YR 5.5/6 | 0.1 | 91.9 | 5.5 | 2.6 | wk med gr |
| | A''b | 100-114 | 10YR 5/4 | 0.1 | 94.0 | 4.2 | 1.8 | wk fn gr |
| | C''1 | 114-130 | 2.5Y 6/4 | 0.0 | 96.7 | 2.1 | 1.3 | wk fn gr |
| | C''2 | 130-170 | 10YR 5/4 | 0.0 | 95.3 | 3.2 | 1.5 | wk fn gr |
| | A'''b | 170-187 | 10YR 4/3 | 0.0 | 84.7 | 12.0 | 3.3 | mod fn gr |
| | C''' | 187-220+ | 10YR 4/4 | 0.0 | 78.5 | 15.1 | 6.4 | mod fn gr |
| T1 | Ap | 0-8 | 10YR 3/2 | 0.1 | 94.1 | 3.9 | 2.0 | wk fn gr |
| | E1 | 8-19 | 2.5Y 5/4 | 0.1 | 93.1 | 4.9 | 2.0 | wk fn gr |
| | E2 | 19-62 | 10YR 6/4 | 0.1 | 92.3 | 6.2 | 1.5 | wk fn gr |
| | Bw | 62-81 | 10YR 6/6 | 1.3 | 91.2 | 6.9 | 2.0 | wk med sbk |
| | BC | 81-88 | 10YR 5/8 | 5.4 | 94.4 | 4.1 | 1.5 | wk med gr |
| | C1 | 88-109 | 10YR 6/6 | 30.2 | 96.7 | 2.6 | 0.7 | sg |
| | C2 | 109-128 | 10YR 6/4 | 15.4 | 97.9 | 1.3 | 0.8 | sg |
| | C3 | 128-166 | 10YR 7/6 | 21.5 | 97.9 | 0.9 | 1.2 | sg |
| | C4 | 166-211+ | 10YR 7/6 | 45.3 | 97.7 | 1.3 | 1.0 | sg |
| T2 | A | 0-10 | 10YR 4/3 | 0.2 | 85.5 | 11.4 | 3.0 | wk fn gr |
| | EA | 10-20 | 2.5Y 5/4 | 0.1 | 87.7 | 9.8 | 2.5 | wk fn gr |
| | E | 20-60 | 2.5Y 6/4 | 1.1 | 83.7 | 12.2 | 4.1 | wk fn gr |
| | Bt | 60-100 | 7.5YR 5/7 | 1.6 | 77.9 | 10.7 | 11.4 | mod med sbk |
| | C | 100-125 | 10YR 6/6 | 8.3 | 96.1 | 1.1 | 2.8 | sg |
| | Csm | 125-140 | 5YR 5/8 | 33.0 | 93.8 | 1.2 | 5.0 | sg - cemented |
| | C' | 140-160+ | 2.5Y 6/3 | 49.2 | 94.1 | 2.0 | 3.9 | sg |
| T3b[e] (site 1) | Ap | 0-18 | 2.5Y 3/1 | 0.4 | 89.3 | 8.3 | 2.5 | wk med gr |
| | E1 | 18-39 | 2.5Y 6/4 | 0.6 | 90.8 | 7.3 | 1.9 | wk fn sbk |
| | E2 | 39-67 | 2.5Y 7/4 | 3.8 | 91.9 | 6.4 | 1.7 | wk fn sbk |
| | Bt | 67-98 | 10YR 5/6 | 6.6 | 79.5 | 4.7 | 15.8 | mod fn sbk |
| | C | 98-103 | 2.5Y 7/3 & 2.5Y 7/4 | 3.3 | 96.0 | 2.2 | 1.7 | sg |
| | Btb1 | 103-116 | 10YR 5/6 | 1.7 | 80.9 | 6.5 | 12.6 | mod fn sbk |
| | Btb2 | 116-140 | 10YR 7/4 (dep) | 2.3 | 68.6 | 13.6 | 17.8 | mod fn sbk |
| | Btb3 | 140-166 | 10YR 7/4 (con, dep) | 4.0 | 67.4 | 19.2 | 13.4 | mod fn sbk |
| | C' | 166-210 | 2.5Y 7/4 (con, dep) | 8.0 | 98.3 | 0.2 | 1.4 | sg |
| | 2C | 210-230+ | 2.5Y 7/1 & 5YR 6/4 | 0.0 | 11.1 | 46.6 | 42.2 | ma |

[a]Abbreviations: (con) = redox concentrations present, (dep) = redox depletions present.

[b]Percentages based on dry wt. of >2 mm fraction.

[c]Percentages based on dry wt. of <2 mm fraction.

[d]Abbreviations: wk = weak, mod = moderate, fn = fine, med = medium, cs = coarse, gr = granular, sbk = subangular blocky, abk = angular blocky, sg = single grained, ma = massive.

[e]OSL sample TU5@90 was obtained from a depth of 90 cm in this C' horizon about 1m away from the described profile, where the C' horizon occurs at a slightly shallower depth. Lower boundary of Little River alluvium is 210 cm.

**Table 3, continued. Soil profiles at OSL sampling locations. These profiles are representative of those characterized by Suther (2006) and are typical of soils found on the middle part of the upper surface of each respective terrace in the most well-drained and least eroded landscape position.**

| Landform | Horizon | Depth (cm) | Moist matrix color[a] | (>2 mm)[b] (%) | Sand[c] (%) | Silt[c] (%) | Clay[c] (%) | Structure[d] |
|---|---|---|---|---|---|---|---|---|
| T3b | Ap | 0-6 | 10YR 3/2 | 0.1 | 89.5 | 9.0 | 1.5 | wk fn gr |
| (site 2) | E1 | 6-42 | 2.5Y 5/4 | 0.2 | 89.6 | 8.6 | 1.8 | sg |
| | E2 | 42-73 | 2.5Y 6/4 | 0.3 | 88.3 | 9.4 | 2.3 | sg |
| | Bt1 | 73-100 | 10YR 5/8 (con) | 0.5 | 88.5 | 5.4 | 6.1 | wk med sbk |
| | Bt2 | 100-147 | 10YR 5/8 | 2.4 | 95.8 | 1.5 | 2.7 | wk fn sbk |
| | C & Bt | 147-170 | 2.5Y 7/3 (C)<br>10YR 5/8 (Bt) | 2.9 (C)<br>1.4 (Bt) | 99.1 (C)<br>96.8 (Bt) | 0.5 (C)<br>0.9 (Bt) | 0.4 (C)<br>2.3 (Bt) | sg (C)<br>wk med sbk (Bt) |
| | C1 | 170-190 | 10YR 7/8 (con, dep) | 5.5 | 98.6 | 0.3 | 1.1 | sg |
| | C2 | 190-227 | 10YR 7/6 (con,dep) | 7.3 | 98.5 | 0.5 | 1.0 | sg |
| | C3 | 227-235+ | 10YR 7/1 | 14.4 | 99.3 | 0.2 | 0.5 | sg |
| T4[e] | Ap | 0-11 | 2.5Y 3/1 | 0.1 | 87.7 | 10.8 | 1.5 | wk fn gr |
| | E | 11-54 | 2.5Y 6/4 | 0.5 | 87.0 | 11.4 | 1.6 | wk fn gr |
| | Bt | 54-90 | 10YR 5/8 | 1.9 | 85.6 | 9.1 | 5.3 | wk med sbk |
| | BE | 90-130 | 10YR 6/6 & 10YR 6/8 | 4.2 | 90.8 | 7.4 | 1.8 | wk med gr |
| | E' | 130-166 | 10YR 7/6 & 2.5Y 7/4 | 9.8 | 90.4 | 8.0 | 1.6 | sg |
| | B't1 | 166-204 | 7.5YR 5/8 (dep) | 27.1 | 84.6 | 5.7 | 9.7 | wk med sbk |
| | B't2 | 204-252 | 7.5YR 5/8 (con) | 6.2 | 82.8 | 4.7 | 12.5 | wk to mod med sbk |
| | B't3 | 252-287 | 10YR 7/6 (con, dep) | 2.4 | 74.2 | 6.5 | 19.3 | mod med sbk |
| | B't4 | 287-307 | 10YR 6/8 | 2.1 | 80.3 | 4.3 | 15.4 | mod med sbk |
| | B't5 | 307-347 | 10YR 7/1 & 10YR 7/2 | 0.2 | 51.4 | 20.5 | 28.1 | no data |
| T5b[f] | A | 0-15 | 2.5Y 3/2 | 0.0 | 90.0 | 7.5 | 2.5 | wk fn gr |
| | Ap | 15-40 | 2.5Y 4/2 | 0.1 | 89.0 | 9.3 | 1.8 | wk med gr |
| | E | 40-63 | 2.5Y 5/4 | 0.0 | 84.0 | 13.3 | 2.7 | wk fn gr |
| | Bt | 63-87 | 10YR 5/8 | 0.2 | 83.1 | 10.7 | 6.3 | wk med sbk |
| | E'1 | 87-113 | 10YR 6/6 (con, dep) | 0.2 | 87.8 | 9.7 | 2.5 | wk med sbk |
| | E'2 | 113-134 | 10YR 6/6 (con, dep) | 0.3 | 88.8 | 8.7 | 2.5 | wk fn gr to sg |
| | B't | 134-145 | 10YR 5/6 (con) | 0.4 | 82.4 | 11.6 | 6.0 | wk med sbk |
| | Btx1 | 145-180 | 10YR 5/8, 2.5Y 6/3, 2.5Y 6/2 (con, dep) | 0.7 | 77.4 | 14.6 | 8.0 | wk med sbk |
| | Btx2 | 180-256 | 2.5YR 4/8, 10YR 5/8, 10YR 6.5/1 (con, dep) | 1.3 | 70.3 | 8.1 | 21.6 | wk med sbk to ma |
| | Btx3 | 256-274 | 2.5YR 4/8, 5YR 5/8, 10YR 6/2, 10YR 6/1 (con, dep) | 6.9 | 72.0 | 10.5 | 17.5 | wk med sbk to ma |
| | Btx4 | 274-288 | 7.5YR 6/6 (con) | 7.6 | 75.2 | 11.8 | 13.0 | wk med sbk to ma |
| | Btx5 | 288-300 | 2.5Y 4/8, 7.5YR 6/8, 10YR 7/1, 10YR 7/2 (con, dep) | 11.5 | 68.6 | 11.5 | 19.9 | wk med sbk to ma |
| | Btx6 | 300-318 | 10YR 7/2 (con) | 20.5 | 61.1 | 12.5 | 26.4 | wk cs abk to ma |
| | | 318-360 | 10YR 7/2 (con) | 5.4 | 51.6 | 19.4 | 29.0 | |

[a]Abbreviations: (con) = redox concentrations present, (dep) = redox depletions present.

[b]Percentages based on dry wt. of >2 mm fraction.

[c]Percentages based on dry wt. of <2 mm fraction.

[d]Abbreviations: wk = weak, mod = moderate, fn = fine, med = medium, cs = coarse, gr = granular, sbk = subangular blocky, abk = angular blocky, sg = single grained, ma = massive.

[e]Description is composited from two pedons for illustrative purposes. Ap-B't2 horizons are taken from T4, pedon 1, and B't3-B't4 horizons are taken from T4, pedon 2 (Suther, 2006). Sampling limitations at pedon 1 prevented description of the entire profile. Lower boundary of alluvium is 307 cm.

[f]Lower boundary of alluvium is at 318 cm within the Btx6 horizon.

angular blocky structure and sand grains that are coated and bridged with clay. An age of 17.4 ± 4.2 ka (31CD475-110) is indicated for T2 by an OSL sample taken from sandy lateral accretion deposits in a C1 horizon at a depth of 110 cm. The complete alluvial thickness was not measured for T1 and T2 deposits, but based on descriptions that extend well into bedload sediment, it is greater than 211 cm for T1 and 160 cm for T2.

Deposits associated with two separate treads of T3b (T3b-1 and T3b-2) were dated. Ultisols have formed in sediments at both localities that typically have A-E-Bt1-Bt2-C profile sequences, with a third Bt or C & Bt horizon sometimes present. At the T3b-1 sampling locality, two allostratigraphic units are present, a lower unit (210 to 103 cm), interpreted as a truncated paleosol formed in lateral accretion sands; and an upper unit (103 to 0 cm), composed of a sand splay deposit overlain by lateral accretion sands that fine upward to vertical accretion sands and silts. The total thickness of alluvium at the T3b-1 locality is 210 cm. An OSL date of 40.0 ± 9.9 ka (TU5@90) was obtained from C horizon sand at a 90 cm depth associated with the sand splay deposit in the upper allounit. A date of 55.2 ± 15.2 ka (TU5@190) was obtained from sandy lateral accretion sediments in a C' horizon near the base of the lower allounit at a depth of 190 cm. The authors consider these ages to be in good agreement. Although in theory deposits at 90 cm and 190 cm could be the same age due to overlap in the 2-σ error of their age estimates, the allostratigraphy at location T3b-1 clearly indicates that sediments at 190 cm are older than those at 90 cm.

Terrace 3b-2 deposits exhibit normal grading and consist of lateral accretion coarse sands and fine gravels that fine upward into vertical accretion sands and silts. The total thickness of alluvium at the T3b-2 locality was not determined, but based on a description that extends well into bedload sediment, it is greater than 235 cm. At the T3b-2 locality, an OSL date of 72.7 ± 13.1 ka (T32A160) was obtained from sandy lateral accretion deposits in the C component of a C & Bt horizon at a depth of 160 cm, while a date of 51.3 ± 12.2 ka (T32A180) was obtained from sandy, lateral accretion deposits lower in the profile in a C1 horizon at 180 cm. The statistical means of these age estimates are inconsistent with their stratigraphic position. However, given the overlap in the 2-σ error of these estimates, it is possible that sediments at 160 cm are the same age as or slightly younger than those at 180 cm, as their stratigraphic positions indicate. Therefore, the authors regard them to be in rough agreement. Relative consistency in age estimates exists both within the individual T3b-1 and T3b-2 stratigraphic profiles and between the separate terrace units. This consistency lends support to the reliability of OSL age estimates presented in this study. Because T3a deposits tended to be less well preserved than those of T3b, they were not selected for dating.

Terrace 4 and T5b deposits consist of lateral accretion medium to coarse gravels and coarse sands that fine upward to vertical accretion sands and silts at the sampling localities. Soils on T4 and T5b are deeply weathered Ultisols and exhibit bisequal soil profiles with multiple argillic (Bt) horizons that are present to the base of and extend below the contact with underlying alluvial fill and/or Coastal Plain sediments that predate the deposition of T4 and T5b (Table 3; Suther, 2006). The total thickness of alluvium at the sampling locality is 307 cm for T4 and 318 cm for T5b. Because unweathered C horizon material was not available for T4 or T5b, OSL ages for both terraces were obtained from E horizons that contained about 90% sand by weight. An OSL sample taken from sandy, lateral accretion sediments at a depth of 161 cm in the E' horizon indicated an age of 74.6 ± 10.4 ka (T4-161) for T4 (Table 2). Although there is overlap in the 2-σ error in the age estimates for T4 and both dated units of T3b (samples TU5@190 and T32A160, Table 2), T4 is clearly older based on its greater height above river level (Table 1) and the degree of pedogenic development characteristic of T4 deposits (Table 3, this paper; Suther, 2006). Terrace 5b is estimated to be 94.0 ± 15.9 ka (T5-135) based on an OSL age obtained from sandy, vertical accretion sediments at a depth of 135 cm in the E'2 horizon. The 2-σ error of this date falls within the range of error of the T4 age estimate and two dates obtained from T3b (TU5@190 and T32A160, Table 2). However, the height of T5b above the modern river (Table 1) clearly indicates it is older than both T4 and T3b. Because T5a deposits were typically less well preserved than those of T5b, they were not selected for dating.

In summary, OSL age estimates indicate a mid- to late-Holocene age for prehistoric sediments of the active floodplain (≥1.3 ± 0.3 ka), an early Holocene age for T1 (9.9 ± 2.0 ka), and a late Wisconsin to terminal Pleistocene age for T2 (17.4 ± 4.2 ka). Age estimates for T3b sediments range from 40.0 ± 9.9 ka to 72.7 ± 13.1 ka (30.1 to 85.8 ka, considering the full breadth of 2-σ error of all T3b ages), indicating T3b may have been active from as early as OIS 5a through late OIS 3. Age estimates for T4 and T5b are 74.6 ± 10.4 ka (OIS 5a through OIS 4) and 94.0 ± 15.9 ka (OIS 5d – 5a), respectively. Although overlap exists in the 2-σ error of T3b, T4, and T5b ages, height above modern river level indicates T3b is the youngest of these terraces while T5b is the oldest, as the statistical means of age estimates suggest. Age separation of the six fluvial surfaces is also corroborated by a relative increase in pedogenic development of soils across the terrace sequence. From lower to higher surfaces, soils show a progressive increase in Bt horizon thickness, subsoil clay content, redness, and dithionite-extractable iron; a progressive decrease in the ratio of bases (CaO, $Na_2O$, $K_2O$, MgO) to alumina ($Al_2O_3$) and other resistant oxides in the bulk chemistry of the whole soil (<2 mm) fraction; and an increasingly mature clay fraction mineral assemblage (Suther, 2006). Despite the low precision and resolution of age estimates for older terraces, these data are significant in that they clearly indicate that the five mappable fluvial terraces within the Little River valley were deposited within the last 100 ka.

## Incision Rates

Little River terraces provide a record of fluvial incision since the time of their deposition, and terrace age estimates allow the calculation of net incision rates during the late Pleistocene and Holocene (Table 4). An age of 94.0 ± 15.9 ka for T5b indicates that the long-term net incision rate for the Little River in the vicinity of the study area has been about 0.29 mm/yr (0.25 – 0.35 mm/yr). If the full breadth of 2-σ error in OSL age estimates is considered, then it is possible that incision has been relatively constant during the last 100 ka. However, incision rates based on the statistical mean of OSL age estimates vary during this period and range from 0.65 mm/yr between the deposition of T5b and T4 (94.0 to 74.6 ka) to 0.07 mm/yr between the deposition of T3b and T2 (59.7 to 17.4 ka). A radiocarbon date from a depth of 246 cm obtained from basal peat overlying lag gravel in the thalweg of an abandoned paleochannel on T1 indicates that the Little River incised to its present bed elevation before 10.24 ± 0.03 ka (Goman and Leigh, 2004) and that the river has migrated laterally since that time.

It is important to note that OSL samples were obtained from soil chronosequence sampling locations rather than from locations in the valley spatially representative of the entire interval of aggradation recorded by each terrace. Age estimates presented here have neither the spatial nor temporal resolution sufficient to precisely bracket intervals of aggradation and abandonment of former floodplains. As a result, the incision rates reported above should be regarded as representing the long-term *net* incision that has resulted from the combined processes of incision and aggradation that have operated since the time of active fluvial sedimentation documented by the OSL date.

# DISCUSSION

## Comparisons with Previous Studies

A net incision rate of 0.29 mm/yr (0.25 – 0.35 mm/yr) for the Little River during the past 94.0 ± 15.9 ka is significant because it is nearly an order of magnitude higher than the uplift rate of 0.064 mm/yr (0.21 ft/1000 yr) reported by Soller (1988, p. A49) for the past 100 ka for the northwestern portion of the lower Cape Fear valley. Soller (1988) estimated uplift rates for the lower Cape Fear valley using a best-fit river

**Table 4. Net incision rates for the Little River.**

| Landform | Incision[a] (m) | Time interval[b] (ka) | O isotope stage | Net incision rate (mm/yr)[c] |
|---|---|---|---|---|
| T1 to modern river bed | 0.0 - 0.0 | 10.2 to 0.0 | early to late 1 | 0.00 |
| T2 to T1 | 4.0 - 0.0 = 4.0 | 17.4 to 10.2 | 2 to early 1 | 0.56 |
| T3b[d] to T2 | 6.9 - 4.0 = 2.9 | 59.7 to 17.4 | late 4 to 2 | 0.07 |
| T4 to T3b[d] | 14.4 - 6.9 = 7.5 | 74.6 to 59.7 | early to late 4 | 0.50 |
| T5b to T4 | 27.0 - 14.4 = 12.6 | 94.0 to 74.6 | 5b to late 4 | 0.65 |
| T5b to modern river bed | 27.0 - 0.0 = 27.0 | 94.0 to 0.0 | 5b to late 1 | 0.29 |

[a]Thickness of vertical incision based on difference in elevation above river bed of upper boundary of channel lag sediments between terraces. T3b lag elevation is an average of lag depth in the T3b-1 and T3b-2 profiles.
[b]Based on means of OSL dates (T2-T5b) and radiocarbon date (T1, Goman & Leigh, 2004).
[c]Rates were calculated by dividing incision thickness by the difference in terrace mean age estimates.
[d]Age estimate for T3b is an average of the statistical means of age estimates from OSL samples TU5@90, T32A160, and T32A180, which were in close stratigraphic association with lag sediments.

terrace longitudinal profiling technique and attributed incision of the Cape Fear River over the late Pliocene and Pleistocene to tectonic uplift associated with a series of local flexures that were superimposed on a gentle, persistent, regional uplift of the Cape Fear arch. Although the Little River incision rate and Soller's uplift rate were determined by different techniques, the discrepancy between them is nonetheless surprising given that the Little River is a major tributary to the Cape Fear River located on the southwestern limb of the Cape Fear arch in close proximity (~ 45 km) to Soller's study area. Because net incision rates calculated for most intervals between the deposition of dated Little River terrace sediments (Table 4) are also substantially higher than Soller's estimate, it seems unlikely this discrepancy results from underestimation of the age of T5b.

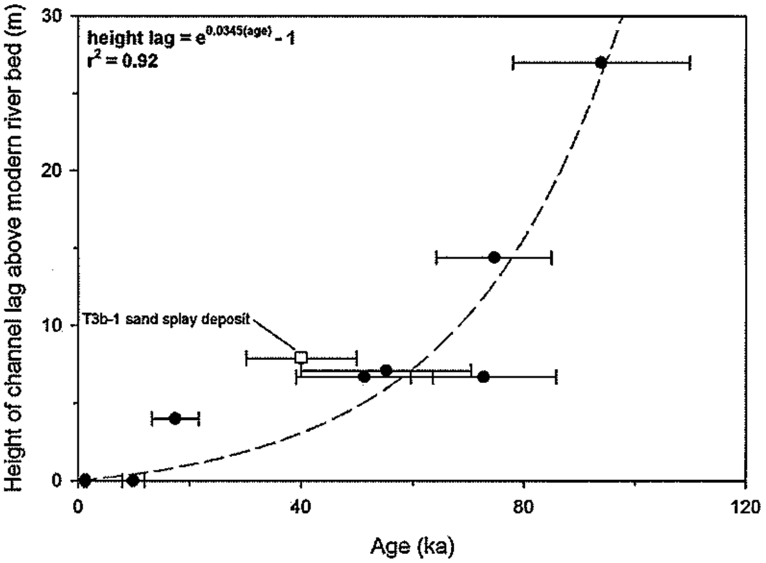
If uplift has been greater near the Piedmont, as Soller (1988) contends, it is possible the Little River, located in closer proximity to the Piedmont than Soller's study area, experienced more rapid local uplift than the lower Cape Fear River and responded with more rapid incision. Markewich (1985), working along the upper Cape Fear River near its confluence with the Little River, observed that the modern upper Cape Fear channel is incised 13 to 28 m beneath its lowest terrace. She suggested incision was in response to late Pleistocene uplift of the Sandhills upland north and west of Fayetteville, North Carolina (which includes the Little River catchment), along a flexure or zone of faulting that extends from the Rockfish Creek – Cape Fear River confluence northeast to Smithfield, North Carolina. Markewich estimated a 60 – 200 ka age for the lowest terrace of the upper Cape Fear River based on soil properties and two pieces of wood that yielded radiocarbon ages of >40 ka. Although incision rates calculated by the authors using a 200 ka age for this terrace (0.07 – 0.14 mm/yr) are lower than the Little River rate, rates based on the 60 ka minimum age for this terrace (0.22 – 0.47 mm/yr) are comparable to those of the Little River and illustrate how possible local variation in neotectonic uplift could account for the difference between the Little and Cape Fear River rates.

A more likely explanation for the discrepancy between Soller's uplift rate and the Little River incision rates is that the estimate of 100 ka for the Wando terrace, used to calculate Soller's rate for the Cape Fear valley during the late Pleistocene, is an overestimate. The Wando terrace, which also is present in the Pee Dee River valley, was not directly dated but assigned an age of 90 ka by Owens (1989). This age was determined by correlation to the Wando Formation, a marine deposit near Charleston, South Carolina, whose age is estimated by amino acid racemization and uranium-series dating of corals to be between 87 – 126 ka (McCartan et al., 1982; McCartan et al., 1980). However, Leigh et al. (2004, pp. 73 – 74) reported several OSL and radiocarbon dates that indicate the Wando terrace in the Pee Dee River valley was active ca. 13 – 22 ka. A younger age for this terrace is also reported by Thom (1967), who indicated it was active shortly prior to 17,000 and 36,000 $^{14}C$ yr BP based on radiocarbon dates from samples beneath dunes. Furthermore, geomorphic mapping by Leigh et al. (2004, p. 74) indicates that the Wando terrace in the Pee Dee valley is actually comprised of multiple terrace levels with different ages, based on cross-cutting relationships and differences in terrace elevations and surface morphologies. The Little River is a tributary to the Cape Fear, and some differences in drainage basin characteristics and external and internal drivers of aggradation and incision certainly exist between the two systems. However, it seems unreasonable that the valley of a major tributary such as the Little River would contain five late Quaternary fluvial terraces while the Cape Fear valley contains only one. Given this evidence and assuming that the Wando terraces in the Pee Dee and Cape Fear valleys are indeed correlative with one another, it is probable that the Wando terrace in the Cape Fear valley also contains deposits considerably younger than 90 – 100 ka. Overestimation of the age of the Wando terrace would result in an underestimation of the rate of incision since its deposition and could account for the discrepancy between the incision rates of the Little River and the Cape Fear River. Although data are insufficient to resolve correlations between Little River terraces and late Pleistocene terraces of the Cape Fear and Pee Dee Rivers, the findings of this study and Leigh et al. (2004) refine age estimates for the fluvial facies of the Wando Formation of Owens (1989) and indicate it is composed of multiple fluvial deposits with a range of late Pleistocene ages.

Mills (2000) compiled age-height data from 16 studies of terraces and cave sediments along rivers of the eastern United States ranging in age from Holocene to >10 Ma. After applying the scaling function of Gardner et al. (1987) in an effort to correct for the effect of measured time interval on calculated incision rate, he observed a 19 fold decrease in incision rate for a $10^3$ increase in measured time interval and concluded that incision may have increased during the late Cenozoic. The Little River incision rate is within the magnitude of incision rates calculated by the authors from terrace height-age pairs compiled by Mills from dated terraces <100 ka in the Appalachian and Coastal Plain Provinces (0.1 – 1.0 mm/yr) and suggests the Little River has experienced incision rates similar to other rivers of the eastern United States during this time period. However, incision rates calculated from ≥100 ka - >10 Ma deposits, which typically range from 0.01 – 0.10 mm/yr, are about an order of magnitude lower than the Little River rate (Mills, 2000). Because Mills summarized incision from rivers across the eastern United States over the past 10 Ma, possible local and temporal variations in tectonics and responses to past climate change and eustasy prevent definitive conclusions about the relationships between Little River incision and that of other rivers. However, considered in the context of the data compiled by Mills, the record of Little River incision may lend support to the contention that rivers in the eastern United States show a general trend of increasing incision through the late Cenozoic.

Incision rates determined from fluvial terraces preserved in unglaciated catchments of northwestern Europe allow comparison of the long-term net incision of the Little River with that of rivers in another temperate, passive mar-

**Figure 3. Height of upper boundary of channel lag above modern river bed at each OSL sampling location versus OSL age estimate (n=8). Floodplain and terrace 1 lag sediments typically occur at the approximate elevation of the modern river bed and are assigned a height of zero. Error bars give 2-sigma error of OSL age estimates. OSL sample TU5@90, associated with a sand splay deposit, was not included in the regression because it postdates lag sediments but is shown on the figure for illustrative purposes.**




gin setting that experienced substantial climatic change during the Quaternary. In general, the late Pleistocene long-term net incision rate of the Little River is nearly an order of magnitude greater than late Pliocene and Quaternary rates reported for rivers of northwestern Europe and southern England that lie outside the limits of the Pleistocene glaciations. Antoine et al. (2000) report net incision rates of 0.05 – 0.06 mm/yr during the past 1 Ma for the middle courses of the Seine River and Somme River valleys (northwestern France), while van den Berg (1994) reports a similar rate of 0.061 mm /yr over the past 2.4 Ma for the Maas River (The Netherlands). A comparable rate of 0.07 mm/yr is reported by Maddy (1997) for the upper course of the Thames River (southern England) during the past 1.8 Ma, although rates were slightly higher (0.13 – 0.14 mm/yr) during the past 440 ka for the Lower Thames (Maddy et al, 2000). According to these authors, terrace deposits reflect periods of aggradation that occurred during cold-climate, glacial or stadial conditions in response to high sediment yield under periglacial landscape conditions, whereas downcutting and terracing occurred rapidly in response to neotectonic uplift during interglacial or early glacial, transitional periods when climate was warmer and the sediment:discharge ratio was favorable to incision. Although the long-term net incision rate of the Little River is considerably greater than rates reported for northwestern European rivers, evaluation of Little River inter-terrace incision rates indicates that similar interactions between uplift and climate-mediated sediment supply may have contributed to the formation of the Little River terrace sequence (see below).

## Possible variation in Little River incision during the last 100 ka

If the 2-σ errors of age estimates of Little River terraces are considered, it is possible that the Little River incised at a relatively constant rate during the last 100 ka. However, incision rates based on mean ages indicate that incision has varied during this time (Table 4). Figure 3 depicts height above the modern river bed versus age for channel lag sediments at each OSL sampling location. Although sample size is too small to permit sound statistical analysis of this data set, nonlinear regression employing an exponential growth function was performed simply to illustrate the nonlinear relationship between height above river bed level and mean terrace age. This regression is presented to highlight the *possibility* of a nonlinear trend in incision over the last 100 ka and is not intended to serve as a quantitative model of river behavior. Variations in Little River incision may reflect the influence of external mechanisms, such as tectonics, eustasy, climate change, or glacioistostasy (Scott et al., 2009), or may reflect a "complex response" to changes with respect to aggradation or incision originating within a river system (Schumm, 1973). The temporal and spatial resolution of age estimates and information about the alluvial architecture of the Little River valley are insufficient to precisely bracket intervals of aggradation and abandonment of former floodplains. Although this prevents definitive conclusions about incision during the late Pleistocene, the Little River data allow for some interpretations and speculations, and possible forcing mechanisms are evaluated below.

The southeastern Atlantic Coastal Plain is located on a passive margin and is characterized by relatively slow rates of neotectonic uplift ranging from 0.01 to 1.8 mm/yr (Cronin, 1981; Marple and Talwani, 2000; Soller, 1988). The long-term net incision rate of the Little River falls within this range, and net incision over the past 100 ka may be in part a response to neotectonic uplift. Soller (1988) concluded that the spatial arrangement of terraces of the Cape Fear River resulted from tectonic uplift associated with a series of local flexures that were superimposed on a gentle, persistent, regional uplift of the Cape Fear arch. Soller did not report variation in uplift within the time interval represented by Little River terraces (late Pleistocene and Holocene). The Little River is a tributary to the Cape Fear located on the southwestern limb of the Cape Fear arch, and it seems reasonable that neotectonic activity played some role in the evolution of its present valley. However, the authors are unaware of any evidence that indicates the vertical or planimetric distribution of Little River terraces can be solely explained by incision in response to variations in uplift of the Cape Fear arch during the last 100 ka.

Eustasy can result in base level changes that force incision, and the base level of the Cape Fear River was lowered during glacial periods, particularly during the last glacial maximum at 22 ka when sea level was 125 m lower than today (Balsillie and Donahue, 2004). However, Leigh and Feeney (1995) and Leigh et al. (2004) observed that eustatic effects are not apparent more than 60 to 80 km inland from the modern shoreline in the Ogeechee River valley and the Altamaha and Pee Dee River valleys, respectively, and are limited to thin veneers of Holocene sediments deposited during sea level rise and valley backfilling that onlap and bury Pleistocene terraces graded to the slightly lower level of the formerly exposed continental shelf. Similar upstream limits of eustatic effects during the late Pleistocene have been reported by Blum and Aslan (2006) and Otvos (2005) in the Gulf Coastal Plain and are thought to be related to climate driven terrestrial sediment yield that overwhelmed and precluded upstream incision induced by lower sea level and restricted it to the outermost continental shelf and continental slope (Blum and Aslan, 2006; Leigh, 2008). The Little River, located some 130 km inland from the head of the estuary of the modern Cape Fear River, would have been insulated from base level effects during the late Pleistocene. This reasoning is consistent with the finding that the lowest incision rate calculated for the Little River (0.07 mm/yr) during the late Pleistocene spans the period of base level lowering associated with the last glacial maximum, between the deposition of T3b and T2 (59.7 to 17.4 ± 4.2 ka).

Recent studies have documented climate-driven changes in channel patterns and sedimentation styles in rivers of the southeastern Atlantic Coastal Plain (Leigh, 2008; Leigh, 2006; Leigh et al., 2004), and Little River data may indicate that environmental change influenced incision during the late Pleistocene. Net incision rates based on mean ages between the deposition of T3b and T2 (0.07 mm/yr, 59.7 to 17.4 ka, late OIS 4 to OIS 2) are comparatively lower than the rates between the deposition of T5b and T4 (0.65 mm/yr, 94.0 to 74.6 ka, OIS 5b to early OIS 4), T4 and T3b (0.50 mm/yr, early through late OIS 4), and T2 and T1 (0.56 mm/yr, 17.4 to 10.2 ka, OIS 2 to 1) (Table 4). The timing of possible variation in Little River incision shows a general correspondence to late Pleistocene climate-driven changes in river regime documented elsewhere in the southeastern Coastal Plain (Leigh, 2008). Leigh et al. (2004) documented braided channel patterns on terraces along seven Coastal Plain rivers, including the Cape Fear River. Optimal expression of braiding occurred between 17 – 30 ka (OIS 2), with braiding possibly occurring as early as 69 ka (OIS 4). Paleomeanders on terraces immediately predating braided terraces in the Pee Dee and Altamaha River valleys indicate rivers had meandering planforms prior to OIS 4 (Leigh et al., 2004; Leigh, 2008). Leigh et al. argue that braiding resulted from high sediment loads and bank instability related to the cooler, drier climatic conditions during OIS 2 through 4 that may have had a snowmelt runoff season that caused larger bankfull discharges than present. Channel patterns transitioned from braided to meandering at about 15 – 16 ka in response to global warming and regional increases in moisture that promoted a more densely vegetated landscape that resulted in reduced sediment yield and increased resistance of channel bank materials (Leigh, 2006).

Although Leigh et al. (2004) and Leigh (2006) do not present incision rates for Coastal Plain rivers during the late Pleistocene, given that Coastal Plain rivers appear to have been sandy, aggrading, braided systems during OIS 4 through 2, it is reasonable to suspect they would exhibit relatively low incision rates during this period. Little River incision rates during OIS 4 through 2 are comparatively lower than those exhibited during OIS 5 to 4 and OIS 2 to 1 and are consistent with this interpretation. Rapid incision to the elevation of the modern channel bed that occurred between the deposition of T2 to T1 (0.56 mm/yr, 17.4 to 10.2 ka) roughly coincides with the braided to meandering transition documented at about 15 – 16 ka for other Coastal Plain rivers that was followed by downcutting to the level of modern floodplains (Leigh, 2006).

Interestingly, incision accompanying terminal Pleistocene transitions from braided to meandering channels has also been documented for unglaciated catchments in the upper mid-latitudes of central and western Europe. Kozarski (1991) attributed the braided to meandering transition and associated downcutting of the Warta River (Poland) ca. 13,000 $^{14}C$ yr BP to decreases in sediment yield, bank erodibility, and stream power that resulted from climate-driven expansion of vegetation cover and changes in the river's discharge regime. Starkel (1991) also implicated climate-mediated changes in hydrologic regime and sediment supply as contributing factors to similar planform changes and downcutting that occurred along the upper Vistula River (Poland) ca. 13,000 $^{14}C$ yr BP but noted that incision was in part driven by lowering of base level following retreat of the Scandanavian ice sheet from the lower course of the Vistula's modern drainage. Similar climate-induced changes to sediment: discharge relationships also explain incision beneath the last glacial braidplain of the Maas River of northwestern Europe (Vandenberghe et al., 1994). In the above cases, several intervals of incision and aggradation, accompanied by transitional planform stages, have been documented between late Pleistocene braided systems and the establishment of Holocene systems with modern bed elevations and high sinuosity, modern-size meanders. Such transitional phases are not apparent in the Little River record, at least at its present resolution.

Little River incision between deposition of T5b and T4 (0.65 mm/yr, 94.0 to 74.6 ka, OIS 5b to OIS 4) may indicate river response to the climate of the last interglacial, when rivers may have had meandering patterns (Leigh et al., 2004) and when the climate of the southeastern United States was warmer and wetter than during the Wisconsinan (Watts, 1971). The relatively high incision rate between OIS 5 and 4 may reflect river response to more sediment-

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