CHAPTER 4. LATE HOLOCENE SEA LEVELS AND THE CHANGING ARCHAEOLOGICAL LANDSCAPE OF ST. CATHERINES ISLAND

DAVID HURST THOMAS

Extending from Outer Banks (North Carolina) to Cape Canaveral (Florida), the configuration of the low-lying barrier islands of Georgia Bight has been conditioned by a mixed regimen of wave and tidal energy (Hubbard et al., 1979; Frey and Howard, 1986). The major embayment centers near St. Helena Sound (South Carolina), where tidal ranges approach 3 m, with a diminishing of tidal range both northward and to the south. The continental shelf along this area is gently inclined, approximating a 2 percent gradient (meaning a change of 1 m of depth for every 20 km of distance) and constant to about 95 km offshore. About 18,000 years ago, the shoreline was roughly 100 m below the present level and the shoreline about 100 km seaward of its present location (Miller, 1998: 43).

This shifting sea level has had a dramatic role in defining the shape of the Georgia/Florida coastline because, given the shallow inclination of the continental shelf, even relatively minor shifts in sea level were accompanied by significant horizontal displacement of the shoreline. In general, a rapid rising sea level will destabilize a coastal ecosystem, and with sea-level rise averaging about 1 cm/year, the coastlines of the Early Holocene must have been remarkably unstable (Colquhoun et al., 1981; Davis, 1997: 157–158). About cal 5500 B.C. (7000 B.P.), the rate of rise slowed to about 3 mm/year; but even at this rate, the shoreline at Sarasota (Florida), for instance (with an offshore shelf gradient averaging about 1:1000), would still have moved about 300 m/century, a migration still too rapid to permit the formation of large and relatively stable barrier islands.

As the sea-level rise along the Georgia Bight slowed and approached present levels, roughly cal 3000 B.C., the coastal Georgia landscape came to look quite similar to that of today (DePratter and Howard, 1977; Oertel, 1979; Colquhoun et al, 1980; Howard and Frey, 1980; Miller, 1998: 39; Booth et al., 1999a, 1999b). Such relative stability provided an opportunity for waves, tides, and longshore transport to mould a coastline into the complex mix of barrier islands, inlets, estuaries, and marshes that today define the Atlantic coast off South Carolina, Georgia, and Florida. Stream gradients were reduced and stabilized, and the coastal biota was essentially modern, in a landscape considerably wetter than before.

As the Atlantic Ocean approached modern levels, new barrier islands formed, prograded, eroded, and reformed along the Georgia shoreline (DePratter and Howard, 1980). The presence of this offshore beach ridge system, some of it "welded" onto Pleistocene island cores, caused the barrier islands to grow seaward, typically assuming the characteristic butterfly, "double island" configuration still evident on Wassaw, Ossabaw, and St. Simons Islands (see chap. 9). Behind the barrier islands, bays gradually filled, fostering the formation of the extensive salt marsh system, with its tidal creeks and estuaries.

MODELS OF LATE HOLOCENE SEA-LEVEL RISE IN THE GEORGIA EMBAYMENT

Despite the relative stability in Late Holocene sea levels and associated landforms, some significant (if less pronounced) fluctuations were yet to come (Fairbridge, 1961; DePratter and Howard, 1980: 33; Brooks et al., 1989: 96; Miller, 1998: 39)—and these changes had serious implications for foragers living along the Georgia Bight.

THE GEORGIA COASTLINE

Crusoe and DePratter (1976: 2) suggest that large oyster beds did not develop behind the barrier islands of coastal Georgia until the rising sea level flooded the previously freshwater lagoons, sometime between cal 3700 B.C. and 2100 B.C. (5000 and 4000 B.P.). The oldest recognizable shorelines date to cal 2800 B.C.–1700 B.C. (4500–3700 B.P.) and St. Simons ceramics are typically associated with these surfaces. Particularly notable are the numerous Late Archaic shell rings that characterize the Georgia Bight (Waring and Larsen, 1968; DePratter, 1975; Marrinan, 1975; Russo, 1996; Sassaman and Ledbetter, 1996). Speaking specifically of the Georgia coast, Howard and DePratter (1980: figure 15) suggest that shell rings existed on both sides of the barrier islands; the surviving shell rings tend to occur on the estuarine side of Pleistocene barrier islands, but with those on the seaward side have likely eroded away altogether.

The basal strata of the St. Simons period shell rings can lie as much as 1 m below the present marsh surface (Waring, 1968a, 1968c; DePratter, 1975; Marrinan, 1975), suggesting that when they were occupied, sea level must have been (at least) 1–2 m below present. At Bilbo, for instance, the basal zone contains fiber-tempered ceramics, dating to cal 1700 B.C.–2200 B.C. (3700–4125 B.P.), overlain by a lens of gravel lag deposits and freshwater mussel shell (Waring, 1968c: 155; Williams, 1968: 330). Brooks et al. (1986) interpret this sequence as reflecting a changing sea level, with the fiber-tempered strata lying more than 1 m below modern levels and the mussel level reflecting a drop in sea level and a seaward movement of the saltwater system.

DePratter and Howard (1980: 34–35) hypothesized that roughly cal 1700 B.C. (3700 B.P.) the sea level rose, from about −1.5 m at to a highstand of approximately present MSL by cal 800 B.C. (3000 B.P.); then, approximately cal 200 B.C. (2500 B.P.), sea level dropped once again, to about −4 m MSL. DePratter (1977b: 11) argues for a hiatus in shellfishing between cal 800 B.C. and 400 B.C. (3000 and 2700 B.P.), due to a sea-level regression that reduced the potential of the estuarine shellfishery on the northern Georgia coast (see also DePratter and Howard, 1977, 1981). He further argues that sites established in the cal 900 B.C.–A.D. 300 (3100–2100 B.P.) range should actually be submerged or buried seaward of the present shoreline, if not completely destroyed by subsequent sea-level rise. By cal 350 B.C. (2600 B.P.), sea level arose again, and extensive oyster beds reformed (Crusoe and DePratter, 1976: 2).

To test their hypothesis, DePratter and Howard (1981) searched the coastal zone of Georgia for additional evidence of lowered sea levels in the Late Holocene. They located a number of pine, oak, and cedar stumps exposed in eroding creek banks. These tree stumps, deeply rooted in salt marsh deposits and often overlain by younger salt marsh deposits, tend to date cal 700 B.C. to cal 450 B.C. (3100–2400 B.P.).[1] Based on these data, DePratter and Howard (1981) identified a single highstand along the Georgia coast from cal 3200 B.C.–1250 B.C. (4500–3000 B.P.). These investigators likewise employed undifferentiated marsh peat as an indicator for a similarly long temporal span during which the fluctuation was supposed to have occurred, but they were unable to constrain that event further.

Such shifting sea levels influence the availability of certain aquatic resources near the estuary mouths. A generally rising sea level with corresponding estuarine expansion causes an increased dispersion of some resources (such as small intertidal oyster beds in the expanding tidal creek network), which can be related to a shifting inland of the intertidal zone with associated increase in salinity (Bahr and Lanier, 1981). Similarly, at the Refuge and Second Refuge sites (located 20 km upriver from Bilbo) Brooks et al. (1985: 300) note the presence of a basal freshwater peat (dating approximately cal 1700 B.C.), shifting toward saline conditions that corresponds to the initial (Refuge period) occupation dated to cal 900 B.C. (3100 B.P.). As sea levels rose, oysters became more common, until a return to freshwater conditons during the terminal Refuge period (and lasting from the Deptford through Irene periods). These data suggested to Brooks et al. (1985: 10) that at cal 800 B.C. (3000 B.P.), the sea levels stood −3 to −4 m below present MSL, rising once again, from cal 100 B.C. to A.D. 700 (2400–1700 B.P.).

The South Carolina Coastline

Colquhoun et al. (1981) have documented that roughly cal 9000 B.C. (10,000 B.P.), sea levels were about 9 m lower than present, rising rapidly until cal 2300 B.C. (4200 B.P.), or slightly earlier, to within 3–4 m of present sea-level position.

Because sea level provides the hydrological base level for both surface and groundwater, this eustatic lowering of sea level exerted a great influence on the freshwater hydrological regimen of the Georgia Bight. Whereas numerous freshwater wetlands survive on the Lower Coastal Plain of the southeastern United States—the best-known examples including the Everglades and Big Cypress Swamp in Florida, Georgia's Okefenokee Swamp in Georgia, and the Dismal Swamp (Virginia)—Brooks et al. (1989: 91) suggest that prior to the Early/Middle Holocene, most of these present wetlands and lakes were dry.

The hydrological threshold for peat formation was surpassed about cal 3700 B.C. (5000 B.P.), suggesting contemporaneous local rise in relative sea level to within 3.5 m or so of the present elevation (Brooks et al., 1989: 91, figure 5.1). After this time, sea-level change primarily influences wetland–estuarine development and biotic shifts in climax forest communities, documenting a change from drier to wetter conditions. This was a time of tremendous increase in the number and area of peat depositing wetlands of the Lower Coastal Plain, and low moor (marsh or swamp) peat formation. "Thus, the direct influence of sea level as a base-level control acting upon the freshwater hydrologic regime in lowland, coastal areas appears to be considerable" (Brooks et al., 1989: 92).

The relative sea-level stability after cal 3700 B.C. (5000 B.P.) reflects considerably slower and lower magnitude changes compared with conditions during the Early Holocene, but conditions along the South Carolina coastline are by no means static. The earliest known shell midden deposits along the South Carolina coast occur at cal 2400 B.C. (4200 B.P.), providing firm evidence for initial development of existing estuarine systems by this time. Large shell middens dating cal 2300 B.C.–cal 800 B.C. (4200–3000 B.P.), Stalling and Thom's creek period sites are generally located in the seaward areas of estuaries, usually adjacent to major channels (Brooks et al., 1989: 94). Many of these deposits have been heavily eroded by subsequent sea-level rise.

Based on data gathered between Winyah Bay and the Savannah River, Colquhoun and Brooks (1986) documented a highstand about cal 2100 B.C. (4000 B.P.), and it is possible that some midden sites established during the cal 1800 B.C.(3800 B.P.) regressive interval may have been completely submerged and/or buried under more recent deposits.

Between cal 800 B.C. and A.D. 1500 (3000–800 B.P.), Brooks et al. (1989: 94, fig. 5.2) note a general trend for shell middens to move inland and be more widespread, correlating with sea-level rise and associated estuarine expansion, suggesting "that estuarine systems on the South Carolina Coast have changed relatively little, either areally or in general configuration, in response to sea level over the last 2,000 years."

More recent research by Gayes et al. (1992) have refined this picture somewhat. Noting that Colquhoun and Brooks (1986) and DePratter and Howard (1981) have previously suggested that a highstand in relative sea level occurred about cal 2000 B.C. (4000 B.P.), Gayes et al. (1992) observe that the resolution of these studies was constrained by poor preservation of deposits and methodological constraints.[2] Using high-resolution data on foraminiferal zonation and closely space vibracores, Gayes and his associates have defined the relative sea-level fluctuations within Murrells Inlet, a small tidal-marsh inlet on the north coast of South Carolina.

Specifically, Gaye et al. (1992: 159, fig. 6) have determined that the late Holocene highstand began with a transgressive, 2-m rise in sea levels between cal 3300 B.C. and 2300 B.C. (5300 and 4300 B.P.). This was

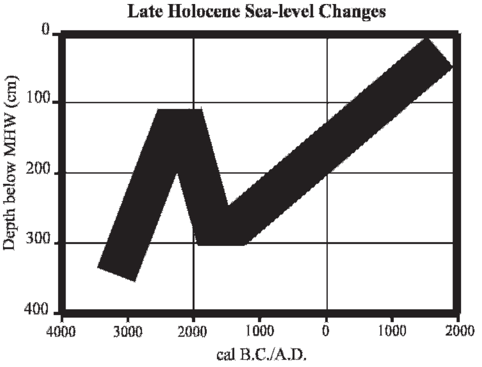

Fig. 4.1. Sea-level fluctuations extrapolated from radiocarbon evidence obtained from Murrells Inlet on the northern coast of South Carolina (after Gayes et al., 1992: fig. 6).

followed by a regressive phase, during which sea levels fell 2 m from cal 2300 B.C. to 1600 B.C. (4300 B.P. to 3600 B.P.). The rate of both rising and falling sea level during this period was 50 cm/100 year (Gayes et al., 1992: 159; fig. 6). Since cal 1600 B.C. (3600 B.P.), sea levels have risen slowly and steadily at a rate of 10 cm/century (until the present).

Figure 4.1 recapitulates the Gayes et al. (1992) formulation, upon which we will model the expectations for St. Catherines Island archaeology.

PROJECTING THE ARCHAEOLOGICAL RECORD OF ST. CATHERINES ISLAND

During the Late Holocene transgression, the landscape available to the St. Catherines Island forager blossomed, with high-ranking marine patches developing in close proximity to long-standing terrestrial patches, thereby minimizing transport costs from centrally placed residential bases.[3] But when the sea level dropped dramatically, as we believe it did, the estuarine oyster beds along the western margin of St. Catherines Island must have been heavily impacted. If patches of oyster beds survived at all, they did so at significantly diminished levels; any Late Archaic foragers exploiting this vastly reduced shellfishery would have created archaeological sites that are today either eroded away or buried beneath 2 m of more recently deposited salt marsh sediments.

These same fluctuating environmental constraints created a vastly different ecological setting on the oceanfront side of St. Catherines Island. A new barrier island formed offshore, protecting a vast, new saltwater marsh and providing foragers with an alternative source of salt marsh resources. The formation and subsequent disappearance of Guale Island and Guale Marsh likewise had a major impact on the behavior of St. Catherines Island foragers and the archaeological record they left behind.

Following the seminal work of DePratter and Howard (1980, 1981), Brooks et al. (1989), Colquhoun et al. (1981), and especially Gayes et al. (1992), we can now offer some hypotheses regarding the influence of Late Holocene sea-level changes on the archaeological record of St. Catherines Island.

CAL 3300 B.C.–CAL 2300 B.C. (5300–4300 B.P.)

During this interval, sea level rose 2 m (at a rate of about 50 cm/century; see Gayes et al., 1992: 159, fig. 6).

Whereas most barrier islands along the Atlantic and Gulf coast are comprised of quartz-sand barriers deposited during the Holocene, a restricted zone of the Georgia Embayment contains an extremely significant series of carbonate-rich Pleistocene remnants fronting the open Atlantic Ocean (Hayes, 1994). From Cape Canaveral northward, several east coast barriers along the Florida–Georgia shoreline are stabilized by the underlying Pleistocene-age formations, anchoring high foredunes that prevent overwashing and landward migration. Related coastal features, including tidal inlets, tidal flats, marshes, and back-barrier bays, reflect the tidal range and the tide/wage energy balance (Davis and Hayes, 1984; Davis, 1997: 158). Along the Georgia Embayment, such tidal inlets tend to be wide and deep, with extensive tidal flats and marshes that typify this mixed-energy coastline.

With the rising sea levels during the cal 3300 B.C.–2300 B.C. interval, we see the correlative growth of the estuarine and interisland marshlands. The juxtaposition of the high-ranking resources of the Pleistocene core (especially the mast crop and newly isolated white-tail deer herds) and the equally high-ranking saltwater marsh provided human foragers with a suddenly diverse and closely spaced set of marine and terrestrial patches. We expect that St. Catherines Island was occupied by Late Archaic foragers shortly after the island core separated from the mainland, circa cal 3000 B.C. Earlier human occupations of this landscape may well have occurred, but without the presence of saltwater marsh resources, foragers would have had to rely more heavily on terrestrial and lower ranked, open saltwater resources. Further, in the absence of debris from harvesting saltwater shellfish populations (i.e., shell middens), archaeological sites of this earlier era would be difficult to find (and in the absence of marine shell, the acid soils of the Pleistocene core would likely have dissolved most of the organics, leaving mostly lithic debris behind).

We suspect that when the Late Holocene transgression topped out about cal 2300 B.C., and that sea level stood approximately 1.25 m below contemporary Mean High Water. Following the behavioral ecology models developed in Part II of this volume, we suspect that shell midden deposits from the earliest portion of the St. Simons (Late Archaic) period should be clustered along the salt marsh margin, situated to minimize transport costs of both terrestrial and marshland resources.

CAL 2300 B.C.–CAL 1600 B.C. (4300 B.P.–3600 B.P.)

During this regressive interval, sea level dropped quickly, at a rate of about 50 cm/century, reaching approximately 3 m below current Mean High Water (Gayes et al., 1992).

During the span of only seven centuries, the sea level dropped about 2 m. Such a dramatic shift should have had significant consequences for the foragers of St. Catherines Island. The saltwater marshes along the estuarine (western) side of the Island must have been dramatically reduced, if not eliminated altogether; accordingly, we project that marshside settlements along the western side of St. Catherines Island should be dramatically reduced (or eliminated) during the cal 2300 B.C.–cal 1600 B.C. interval. If marsh remnants did survive in the estuary, then human settlements were likely moved to lower elevations to exploit the dwindling salt marsh resources; this means that along the western margin of St. Catherines Island, any archaeological evidence during this seven-century interval is likely submerged under a meter or more of marsh sediments that accumulated later (as the sea rose to approach modern levels).

But a different scenario pertains along the seaside (eastern) margin of St. Catherines Island. As noted previously (in chap. 3), the rising sea level of the early Holocene triggered a rapid westward transgression of off-shore barrier islands, eventually docking these barriers to the relic late Pleistocene landscape by cal 3000 B.C. or so. This is when the new offshore "Guale Island" formed along the northeastern margin of St. Catherines Island, effectively buffering the ocean front and fostering the development of an extensive, interisland marsh ("Guale Marsh") that evolved as the sea level rose. Although Guale Island was eventually overtopped by the still rising sea level, it must have provided a refuge salt marsh habitat, along the eastern shoreline of St. Catherines, for foragers previously exploiting the estuarine salt marshes along the western side of the island.

In short, then, we project that during the interval cal 2300 B.C.–cal 1600 B.C., St. Simons/Refuge period archaeological sites should:

- Disappear along the western margin of St. Catherines Island (because the salt marsh had either evaporated or significantly retreated, meaning the sites resulting from exploiting the remnant saltwater marshes would have been flooded by subsequent sea-level rise), and
- Appear on the northeastern Pleistocene core remnant (where such settlements fronted the former expanse of Guale Marsh).

We also suspect that numerous St. Simons period archaeological sites might have once accumulated further to the east, on Guale Island (but they disappeared when Guale Island eroded).

CAL 1600 B.C.–PRESENT (3600 B.P.–PRESENT)

Sea levels have risen slowly and steadily (at a rate of 10 cm/century) from a low-water mark of roughly 3 m below Mean High Water to present levels.

As the sea levels rose after cal 1600 B.C., foragers of late St. Simons and Refuge-Deptford periods (circa cal 1600 B.C. to perhaps cal A.D. 1) likely witnessed:

- A deterioration of marshland resources along the eastern margin of St. Catherines Island (due to the overtopping of Guale Island and disappearance of Guale Marsh), and
- A resurgence of estuarine marshlands along the entire western margin of the island.

Accordingly, the archaeological landscape of St. Catherines Island should contain fewer Late Archaic and Refuge-Deptford sites along the northeastern corner of the island, with a progressive reoccupation of the western shoreline, as the estuarine marshland came back to its previous productivity.

Throughout the remainder of this monograph, we will generate archaeological evidence to test the above hypothesis on the archaeological record of St. Catheirnes Island.

NOTES

1. To calibrate the data on buried stumps and freshwater peats discussed in DePratter and Howard (1981), we have employed the terrestrial conversion IntCal04 (Reimer et al., 2004).
2. Because the radiocarbon dates cited in Gayes et al. (1992) are expressed at "corrected radiocarbon ages (k.a.)" we have converted these estimates to simple A.D./B.C. expressions, *without* applying additional correction factors.
3. The theoretical framework supporting this suggestion is developed in subsequent chapters.