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Abstract

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ADDITIONAL EDIACARAN FOSSILS FROM THE LATE PRECAMBRIAN CAROLINA TERRANE, SOUTH-CENTRAL NORTH CAROLINA

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ABSTRACT

Additional representatives of the late Precambrian Ediacaran biota from Stanly County, North Carolina, are presented, described, and compared with previously documented body fossils from the same area. These representatives suggest a greater diversity in geometry of *Pteridinium carolinaensis* than was previously indicated in North Carolina, and refine the stratigraphic interval and depositional situation in which fossils of the Ediacaran biota are known to occur in Stanly County.

INTRODUCTION

The Carolina terrane has generally been ignored from the perspective of paleontology because of relatively poor quality exposures, or because most geologic studies have been directed toward the igneous, metamorphic, and tectonic history of the area, or because of the assumption that the terrane's meta-volcanic or meta-sedimentary units were non-fossiliferous. Between 1856, when Emmons (1856) mistakenly identified spherulites from rhyolitic flows of the Uwharrie Formation (Figure 1) as *Palaeotrochis* and 1973, when St. Jean (1973)

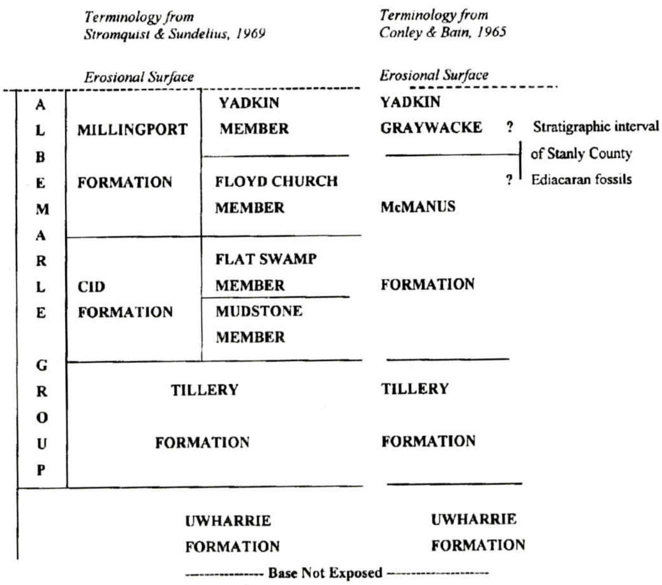

Figure 1. Generalized stratigraphic column for the Carolina terrane, south-central North Carolina

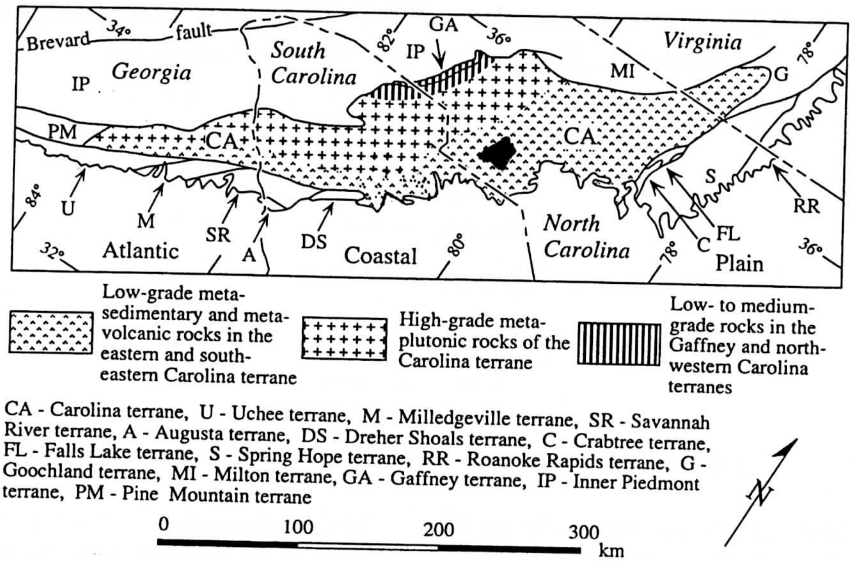

Figure 2a. Extent and areal relationships of Carolina terrane, illustrating the generalized distribution of low-grade metasedimentary and metavolcanic rocks, and location of Stanly County. Modified from Secor and others, 1999.

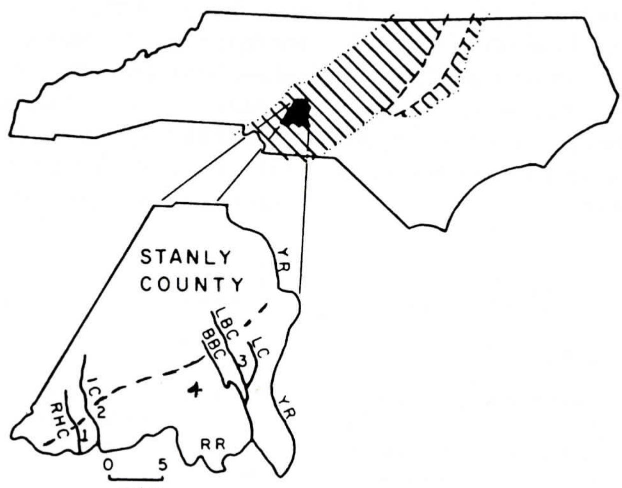

Figure 2b. Location map of Stanly County, North Carolina body fossil localities. Pattern approximates outcrop of Carolina terrane. RHC = Rock Hole Creek; IC = Island Creek; BBC = Big Bear Creek; LBC = Little Bear Creek; LC = Long Creek; RR = Rocky River; YR = Yadkin River; 4 = Oakboro. Numbers indicate body fossil locations with 4 being Gleaning Mission Church at Oakboro, North Carolina. Dashed line approximates map contact of McManus Formation (right) and Yadkin Graywacke (left). Modified from Gibson and others (1984).

described body fossils from Stanly County, North Carolina, there were no documented fossil finds in the Carolina terrane.

Since 1973, several fossil finds have been reported from the North Carolina (Figure 2) part of the Carolina terrane. These include a small number of body fossil representatives of the Ediacaran biota (St. Jean, 1973; Gibson and others, 1984; Gibson and Teeter, 2000) and typical late Precambrian ichnofossils (Gibson, 1989), all from Stanly County, North Carolina. Stromquist and Sundelius (1969) described possible coalified algae, also from Stanly County, and Cloud and others (1976) reported *Vermiforma antiqua* from the Durham, North Carolina area. Seilacher and others (2000) questioned the validity of *Vermiforma antiqua* as biologic traces. Koeppen and others (1995) reported euconodonts from the Jacobs Creek quarry, near Denton, in Davidson County North Carolina, and euconodonts, bryozoan zooecial steinkerns, and a gastropod steinkern from a quarry near Asheboro, North Carolina.

In the South Carolina part of the terrane, an Atlantic Province Cambrian trilobite assemblage has been reported (Maher and others, 1981; Secor and others, 1983) that includes a large number of individuals. Bourland and Rigby (1982) described sponge spicules from Cambrian age units of the Carolina terrane in South Carolina.

Given the areal extent of the Carolina terrane (Figure 2), this is not an impressive fossil record, although large numbers of trilobite specimens have been recovered from South Carolina (Secor and others, 1983). The Ediacaran body fossils described herein add to the knowledge base of the Precambrian biota of the Carolina terrane.

GEOLOGIC SETTING

The Carolina terrane extends from central Virginia southwestward to central Georgia (Figure 2). Terrane stratigraphy is a complex of volcanosedimentary (Gibson and Teeter, 1984), tuffaceous volcanic (Harris and Glover, 1983), intrusive igneous (McConnell and Glover, 1982), and marine-shelf rocks (Secor and others, 1983) within the Appalachian orogen (Williams and Hatcher, 1982). These lithofacies have been altered to some degree by thermal and dynamic metamorphic processes. The tectonic relationships of the terranes in the Appalachian orogen are not always clear (Secor and others, 1997; Kish and Black, 1982).

There are extensive portions of the Carolina terrane in which original stratigraphic features have been destroyed by post-depositional igneous and metamorphic activity (Russell and others, 1985). However, the Carolina terrane in south-central North Carolina was generally subjected to only lower greenschist facies grade metamorphism (Stromquist and Sundelius, 1969), so that primary and secondary sedimentary features are frequently visible in outcrop (e.g., Gibson, 1985). The combined effects of this low-grade metamorphism and tectonism in south-central North Carolina resulted in welded bedding surfaces, and in many areas, very closely spaced, non-bedding parallel cleavage patterns along which there is preferential breakage. This geologic history combined with deeply weathered bedrock, and extensive soil/vegetation cover, reduces the potential for preservation of entire body and trace fossils on exposed bedding surfaces.

STRATIGRAPHY

In south-central North Carolina, the Carolina terrane includes four generally recognized units (Figure 1). The oldest unit is the Uwharrie Formation, composed of 4,572-6,096 meters of felsic volcanoclastic rocks (Stromquist and Sundelius, 1969). The Albemarle Group (Conley and Bain, 1965) appears to be concordant (conformable?) on the Uwharrie and includes the Tillery, McManus and Yadkin formations (Conley and Bain, 1965).

Typically, the Tillery Formation in south-central North Carolina is dominated by thinly laminated siltstone and claystone couplets (Conley, 1969). The lower part of the formation also includes cross-bedded sandstones and subaqueous debris flows (Gibson and Teeter, 1984). The upper part is more tuffaceous, with thin, water-laid tuff layers of sand-sized to fine pebble-sized lithic and crystal clasts interbedded with the dominant thinly laminated claystone and siltstone. This formation has an approximate thickness of 1,524 meters (Stromquist and Sundelius, 1969), which may be exaggerated by thrust faults (Huntsman and Dockal, 1985).

Overlying the Tillery in a broadly gradational contact is the McManus Formation, as described by Conley and Bain (1965). This formation is some 3,048 - 4,877 meters thick and is composed of a lower, massively bedded mudstone member of the Cid Formation (Stromquist and Sundelius, 1969) and an upper, more thinly bedded siltstone-mudstone member of the Floyd Church Formation of Milton (1984). The clay and silt are of volcanic origin and numerous layers of volcanic tuff and flows, as well as devitrified glass shards, are evident in the McManus Formation (Stromquist and Sundelius, 1969).

Koeppen and others (1995) reported euconodonts from the Jacobs Creek quarry in Davidson County, just north of Stanly County, and indicated that the collecting site was in the mudstone member of the Cid Formation of Stromquist and Sundelius (1969), or the lower McManus Formation of Conley and Bain (1965). Based on this fossil evidence, Koeppen and others (1995) assigned a late Cambrian or Ordovician age to the unit, and indicated that the fossil data pointed toward previously undocumented stratigraphic and structural discontinuities being present in the section.

The Ediacaran fossils indicate a late Precambrian age for the upper part of the McManus Formation in Stanly County. Yet, the late Cambrian to Ordovician age suggested by Koeppen and others (1995) for the Cid Formation (Lower McManus) in Davidson County, is for a unit that is generally considered to be stratigraphically lower in the McManus Formation (Figure 1) than the late Precambrian Ediacaran fossils, and hence, should also be older. The interpretation of Koeppen and others (1995) suggests a structural discontinuity between the lower and upper McManus units, perhaps west of the Ugly Creek Thrust of Dockal and Huntsman (1990).

Overlying the McManus Formation is the Yadkin Graywacke. This formation is composed of about 914 meters (Stromquist and Sundelius, 1969) of medium- to coarse-grained, compositionally and texturally immature volcanic sandstones and interbedded siltstones. The original thickness of the Yadkin Graywacke is unknown as the upper surface is the present erosion surface. Field evidence indicates an interfingering contact between the McManus and the Yadkin formations. Sedimentary structures within the lower Yadkin suggest shallow marin deposition (Gibson and Teeter, 1984) that ap pears to grade upward into non-marine facies.

EDIACARAN BODY FOSSILS OF SOUTH-CENTRAL NORTH CAROLINA

St. Jean's (1973) report of fossils (Figure 3 from Stanly County, North Carolina, was th first published record of the occurrence of bod fossils in the Carolina terrane. These fossils ar preserved on a water worn boulder of McManu Formation lithology that was recovered from Is land Creek, west-northwest of Oakboro, Nort Carolina. Gibson and others (1984) describe additional body fossil specimens from Littl Bear (Figure 4) and Rock Hole (Figure 5 creeks. The Little Bear Creek specimen is pre-

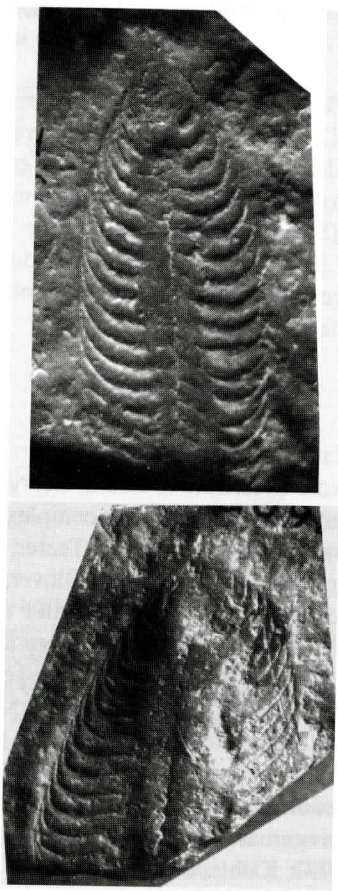

Figure 3. ***Pteridinium carolinaensis*** **collected from Island Creek, Stanly County locality (from St. Jean, 1973).**

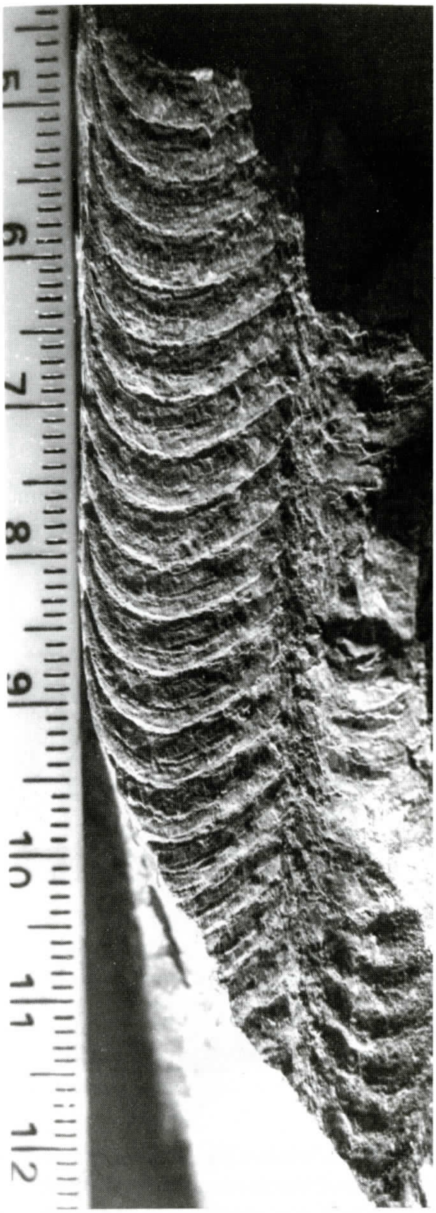

Figure 4. ***Pteridinium carolinaensis*** **from Little Bear Creek, Stanly County locality. Scale in centimeters.**

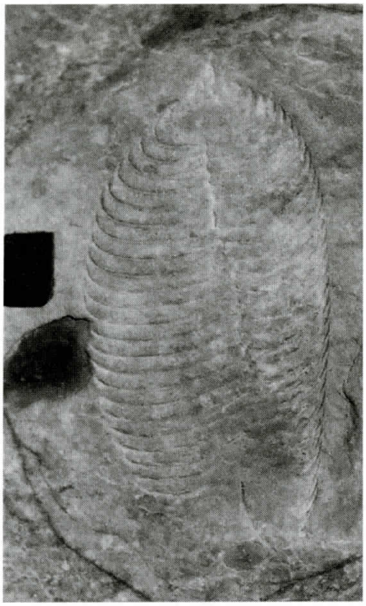

Figure 5. ***Pteridinium carolinaensis*** **from Rock Hole Creek locality. Note duplication of pointed ends of individual segments on right side of specimen reflecting the presence of a third wing. Scale block on left is 1 centimeter long.**

served on a quarried boulder of McManus Formation lithology that was part of a 150-year-old chimney (Teeter, 1984) constructed of the native stone that written records indicate was quarried within 600 meters of the chimney location. The Rock Hole Creek specimen is preserved on a water-worn stream boulder, similar to the Island Creek specimens.

The fossils described in this paper were collected in 1987 from boulder-sized bedrock slabs uncovered during site preparation for a construction project at the Gleaning Mission Church, Oakboro, Stanly County, North Carolina. The slabs on which these fossils were preserved were moved approximately 40 meters from the construction site, to the edge of the church parking lot. Though arguably not "in place," the rock slabs and fossils are as close to "in place" as any Late Precambrian body fossils so far discovered in south-central North Carolina.

SPECIMENS FROM GLEANING MISSION CHURCH SITE

Mr. Ruffin Tucker, an amateur paleontologist, discovered an Ediacaran body fossil in the construction debris at the Gleaning Mission Church, in early 1986. This specimen (Figure 6) is preserved as an impression in a 0.5 cm thick layer of water laid crystal/lithic tuff, a common

Figure 6. *Pteridinium carolinaensis* from Gleaning Mission Church locality. This specimen was collected by Mr. Ruffin Tucker in 1986. Note ichnofossil, *Planolites sp.*, to right of base of *Pteridinium*. Pencil body is 0.8 centimeters wide.

Figure 7. *Pteridinium carolinaensis* from Gleaning Mission Church locality. Note difference in overall shape as compared to specimens illustrated in Figures 3 – 6; presence of central stalk across which segments are offset, and presence of groove broadening from apex, which is indication of the presence of a third wing. Scale is in centimeters.

secondary lithofacies throughout the dominant mudstone lithology of the upper McManus Formation. That specimen approximates the overall outline, size, and general geometry of *Pterindinium carolinaensis* specimens previously described from Stanly County (St. Jean, 1973, Gibson and others, 1984).

Subsequently, two additional body fossils (Figure 7 and Figure 8) were recovered from the same site. These specimens do not correspond in size or overall outline geometry to those specimens previously described from this area.

Description

The two additional specimens (Figure 7 and Figure 8) from the Gleaning Mission Church locality are composed of petaloids or vanes made up of curved segments like the previously described (Gibson and others, 1984) forms. The incomplete body fossils (Figure 7 and Figure 8) are preserved as very low relief, concave molds on bedding surfaces within the typical mudstone lithology of the McManus Formation. The body fossil in Figure 7 preserves parts of two broad surfaces of the organism, as a mainly concave mold of one surface, and a shallowly

Figure 8. *Pteridinium carolinaensis* from Gleaning Mission Church locality. Not as well preserved as the specimen in Figure 7. Note difference in overall shape as compared to *Pteridinium* illustrated in Figures 3 – 6 and presence of central groove/stalk across which segments are offset. Scale is in centimeters.

convex mold of part of the other surface. The change from the concave to convex mold preservation occurs at the irregular fracture. The other specimen (Figure 8) is not as well preserved, but illustrates the same internal and external geometry as that of the specimen in Figure 7.

The specimens from the Gleaning Mission Church locality (Figures 7 and 8), though incomplete, have lengths of 18 and 10 cm, respectively, and maximum widths of 6 and 4.5 cm, respectively. The maximum width measurement occurs at about 80 and 90 percent, respectively, of the distance from the apical end to the projected basal end. Descriptively, the outline geometry of these fossils is similar to that of the flame of a candle, which broadens abruptly from the wick at the base and then tapers to the apex of the flame.

There is an overall, bilateral symmetry to the fossils (Figures 7 and 8), caused by only two of the three "wings," "vanes," or "petaloids" (terms used in various descriptions of these and similar Ediacaran fossils) being visible on bedding surfaces. This bilateral symmetry is across a narrow, central, segmented ridge. Individual segments of the wings are arranged in a staggered or echelon pattern across this central ridge. Distally from the central ridge, the individual segments curve slightly toward the apex, becoming more strongly curved toward the margins of the fossils. Each segment decreases in width distally and terminates as a point. The re-curved, pointed distal ends of the segments overlap to form a relatively smooth outline to the fossil. At least 30 segments are preserved on individual vanes.

A shallow groove broadens systematically from the apex toward the projected base of both specimens (Figure 7 and 8), but is most obvious in Figure 7, due to better preservation. The geometry of this groove suggests the presence of a third wing or vane similar to that visible on other *Pteridinium carolinaensis* specimens (e.g., Figure 5). The size of the shallow groove indicates that the third vane or wing is subordinate in size to vanes preserved on the bedding surface. A previously described specimen from Rock Hole Creek (Figure 5) appears to have near equality in size of each of the three wings or vanes. This characteristic may be a result of a difference in age of the organism or stages of growth, as these two specimens (Figures 7 and 8) are composed of more segments than the more ovate specimens previously described (Gibson and others 1984), and may be older or more mature organisms.

Comparison

The general description of *Pteridinium carolinaensis* is that it is an elongate, straight to gently curved, parallel-sided to gently tapered, leaf-like structure (Narbonne and others, 1997). As generally preserved on bedding surfaces, *Pteridinium carolinaensis* exhibits a bilateral symmetry with segmented "wings," "petaloids" or "vanes" joined at a midline or commissure with a marked zigzag pattern formed by the alternating arrangement of individual segments in opposing "wings." Individual segments in each wing are re-curved toward what is probably the distal end of the organism, narrowing toward the margin of the fossil, and terminating as a point. A third wing is frequently reported, and also appears attached at the commissure with segments producing a zigzag commissure line (Jenkins, 1985). *Pteridinium carolinaensis* described by St. Jean (1973) and Gibson and others (1984), as well as the specimen from the Gleaning Mission Church discovered by Mr. Tucker show similar characteristics, being generally ovoid, and having offset segments forming a zigzag line of commissure.

The more obvious differences between these specimens (Figure 7 and Figure 8) and those previously described (Figures 3, 4, and 5) from Stanly County are the overall size and outline geometric shape. These incomplete (Figure 7 and Figure 8) specimens are 18 cm and 10 cm long as compared to 3.3 and 4.4 cm for the Island Creek specimens, 7.5 cm for the Little Bear Creek specimen, and 7.75 for the Rock Hole Creek sample. In terms of outline shape, the Island Creek, Rock Creek and Little Bear Creek *Pteridinium carolinaensis* specimens are preserved as generally elliptical forms, while the two specimens (Figure 7 and Figure 8) from the Gleaning Mission Church site are broadly ovate (Figures 7 and 8), broadening abruptly from the incomplete base and then tapering in the direction of segment curvature, toward the apex. Individual segments are re-curved with the pointed, distal ends of the segments overlapping to create relatively smooth, non-curved margins.

Another characteristic that is distinctive in these reported specimens compared to previously described specimens, is the ridge-like feature along the axis (Figure 7) rather than a zigzag commissure noted in previously described specimens. This ridge-like feature is no present in the Rock Hole Creek or the Islan Creek specimens. It may be present in the Littl Bear Creek specimen. This ridge-like featur may represent a central stem or stalk. Acros this ridge-like feature, proximal ends of th curved segments are arranged in an offset o echelon pattern (Figure 7 and Figure 8), charac teristic of *Pteridinium carolinaensis* (e.g., Fig ure 5).

INTERPRETATION OF DEPOSITIONAL ENVIRONMENT

The Ediacaran fossils (St. Jean, 1973, Gibson and others 1984) and trace fossils (Gibson, 1989) previously described from the Carolina terrane of south-central North Carolina have been recovered from the upper part of the McManus Formation, a unit interpreted as representing a relatively low energy depositional environment that was subjected to episodic higher energy depositional events (e.g., Dockal and Huntsman, 1990). Occasional thin layers of crystal/lithic tuff within the mudstone represent surges of higher energy deposition, e.g., that lithology in which is preserved the elliptical *Pteridinium carolinaensis* (Figure 6).

Pteridinium is one of the most widespread and characteristic representatives of the Ediacaran Biota and is generally thought to have inhabited shallow, high-energy marine environments (e.g., Fedonkin, 1981; Fedonkin, 1983; Glaessner and Wade, 1966). The *Pteridinium* fossils so far recovered from the Carolina terrane are incomplete, lacking the proximal end stalk or holdfast attachment, and are not associated with shallow water, high energy facies. This incomplete nature of the *Pteridinium* fossil remains, preserved in relatively low energy facies of the Carolina terrane, supports the premise (Jenkins, 1985) that the organisms preserved in relatively low energy facies were forcibly removed from their living positions by strong water currents or wave action and transported by water currents to the site of deposition. The two dimensional trace fossils preserved on the same bedding surfaces as the body fossils (Figures 6 and 9), substantiate the premise that either slow rates of deposition or episodic depositional events that permitted repopulation of the substrate by the trace-making organism, and that the incomplete body fossils had been transported from a higher energy environment to their place of deposition.

In Stanly County, North Carolina, the lower part of the Yadkin Graywacke contains primary sedimentary features indicative of a near shore, high-energy, aqueous depositional environment (Gibson and others, 1990). The upper part of the underlying McManus Formation is interpreted as representing an essentially equivalent, but deeper water, lower energy deposit (Gibson and Teeter, 1984).

The contact between the high energy, nearshore marine Yadkin Graywacke and the underlying, lower energy, deeper water McManus Formation is an interfingering contact. Evidence of this contact is found only a few meters up section from the Gleaning Mission Church location. The vertical proximity of the Ediacaran body fossils so far recovered from Stanly County to this stratigraphic contact, and to the map contact between the Yadkin and McManus formations, supports the supposition that the body fossils found are allochems in the McManus Formation, having been forcibly removed from their higher energy, shallow water life positions (represented by the lower Yadkin Graywacke) and transported down gradient by water currents and deposited in a lower energy environment (McManus Formation).

The recovery of three specimens from a contiguous area of approximately 180 square meters also suggests that there was probably a substantial population of sessile, bethonic Ediacarian organisms living in or on the sand substrate in shallow, high-energy marine environments in reasonably close proximity to the site of deposition and preservation. The sedimentary features preserved in the lower Yadkin Graywacke support the interpretation that the basal portion of that unit is a shallow marine, high energy deposit, and suggest a potential biofacies source of the *Pteridinium* fossils found in the lower energy McManus Formation.

SUMMARY

The recovery of additional specimens of *Pteridinium carolinaensis* from the Gleaning Mission Church site in Stanly County, North Carolina, broadens the knowledge of this member of the Ediacaran biota, documenting size and geometric outline differences. These fossils also further clarify the stratigraphic interval of the Carolina terrane in south-central North Carolina in which the Ediacaran biota is likely to occur, and suggest that additional specimens are likely to be preserved in the marine facies of the Yadkin Graywacke.

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PLEISTOCENE ENCROACHMENT OF THE WATEREE RIVER SAND SHEET INTO BIG BAY ON THE MIDDLE COASTAL PLAIN OF SOUTH CAROLINA

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ABSTRACT

In Sumter County, South Carolina, an inland sand sheet east of the floodplain of the Wateree River has encroached into Big Bay and may have buried other Carolina bays completely. The encroachment provides an opportunity to study relationships between the development of dunes and Carolina bays, oriented oval depressions that are common on the south Atlantic Coastal Plain. The sand sheet was derived from the Wateree River floodplain, probably when it was much more sparsely vegetated than at present. A drill hole at the leading edge of the sand sheet, where it encroaches into the western side of Big Bay, revealed clays of the Duplin Formation overlain by 4.5 m of organically enriched bay-fill sediments (>48,000 radiocarbon yr B.P.) and 4.5 m of eolian sand. Pollen assemblages from the lower part of the bay fill indicate that the adjacent terrestrial vegetation was open and dominated by grasses. Oak and hickory suggest warm conditions, perhaps Oxygen Isotope Stage 5 (134–75 ka B.P.), an interval that spans the Sangamon interglacial to the early Wisconsinan, or early Oxygen Isotope Stage 3 (65–32 ka B.P.), during the Wisconsinan. Pollen assemblages from the middle of the bay fill indicate an open, pine-dominated terrestrial community, plausibly associated with cool climate, perhaps Oxygen Isotope Stage 4 (75–65 ka B.P.), a period of early Wisconsinan glacial advance. Buried soil horizons in the upper half of the bay fill suggest that encroachment of the sand sheet was episodic, with a major episode occurring before 48,000 radiocarbon yr B.P. Conifer macrofossils suggest relatively dry conditions in the basin at that time; dry climate is also requisite for massive transport of sand. This major episode may thus have occurred under cool, dry climate in Oxygen Isotope Stage 4 or under

warm, dry climate early in Oxygen Isotope Stage 3. Archaeological evidence indicates that redistribution of sediment on the sand sheet continued at least intermittently into the Holocene, although activity may have slowed after 4000–3000 yr B.P.

INTRODUCTION

Distinctive eolian landforms occur on some of the sandy surfaces in the interior of the south Atlantic Coastal Plain. Carolina bays, found from northern Florida to the Delmarva Peninsula (Johnson, 1942; Prouty, 1952), are oval, oriented basins formed by the action of wind on water ponded in upland depressions (Thom, 1970; Kaczorowski, 1977). In Georgia and the Carolinas, inland dunes developed on the northeast sides of some southeasterly flowing streams (Johnson, 1959; Colquhoun, 1965, figure 3; Thom, 1970; Carver and Brook, 1989; Nystrom and others, 1991; Markewich and Markewich, 1994). The dune fields commonly are located downwind of river floodplains exhibiting relict braided channel patterns. In some areas, dunes have encroached into bays, and bays into dunes (e.g., Thom, 1970).

Bays and dunes can provide information on the history of Quaternary climate and geomorphic processes (Ivester, 1999) and of Quaternary ecological processes (e.g., Taylor and Brooks, 1994; Taylor and others, 1999) on the Coastal Plain. In the lower Great Pee Dee Valley of South Carolina, Thom (1970) inferred that dunes and bays developed synchronously, probably during the last glacial stage and early Holocene (40–7 ka B.P.) in periods of cool pluvial conditions with strong southwesterly winds and high water tables. Based on their work on dunes in Georgia, Leigh and Ivester (1998) invoked cool, dry climate during the full-glacial period of dune activity. Ivester (1999) refined this argument, adding inferences that winters were windy and springs were moist. For the period 19–13 ka B.P., Watts (1980) inferred cool, dry climate from the pollen record of White Pond, a Carolina bay in central South Carolina.

Soller and Mills (1991) argued that bays must have formed roughly 200–100 ka B.P. or earlier and that both dunes and bays may have been created in multiple generations. The older features may have been reshaped during successive episodes of suitable climate. Based on a thorough review of the literature and numerous dates for inland sand dunes in Georgia, Ivester (1999) identified at least two periods of dune activity in the southeastern Coastal Plain. For the most recent period of major dune activity in Georgia, dates cluster between 30–15 ka B.P., during the Wisconsinan glacial maximum. Earlier dates, obtained from optically stimulated luminescence of quartz sand, range from >140–65 ka B.P. Bays and dunes displaying well-preserved morphology may have formed or been reactivated during late Pleistocene-Holocene time (e.g., Markewich and Markewich, 1994; Soller and Mills, 1991; Brooks and others, 1996; Grant and others, 1998; Ivester, 1999).

In Sumter County, South Carolina, dunes of an inland sand sheet encroached into, and may have buried, many Carolina bays, providing an opportunity to link the development of bays and dunes in an area interior to the locale studied by Thom (1970). Our main objectives were to determine the time of encroachment of the sand sheet, to determine whether the encroachment coincided with or closely followed formation of the bays, and to infer environmental contexts for these episodes. We obtained stratigraphic data from a drill hole at the leading edge of the sand sheet where it had buried the western side of Big Bay, the largest of the bays in the study area. Bay-fill sediments beneath the sand sheet yielded some material suitable for radiocarbon dates. They also yielded some identifiable pollen, woody peat, and other fossil materials. We measured orientations of both dunes and bays in the area to infer directions of prevailing winds during their formation. We used archaeological data from concurrent studies (Brooks and others, 1998; Cable and Cantley, 1998) to determine whether movement of sediment on the sand sheet continued into the Holocene. The results outline a history of eolian episodes, plausibly beginning during the earlier period of dune activity (>140–65 ka B.P.) determined by Ivester (1999). Future coring in the basin and on the sand sheet will help to resolve details of

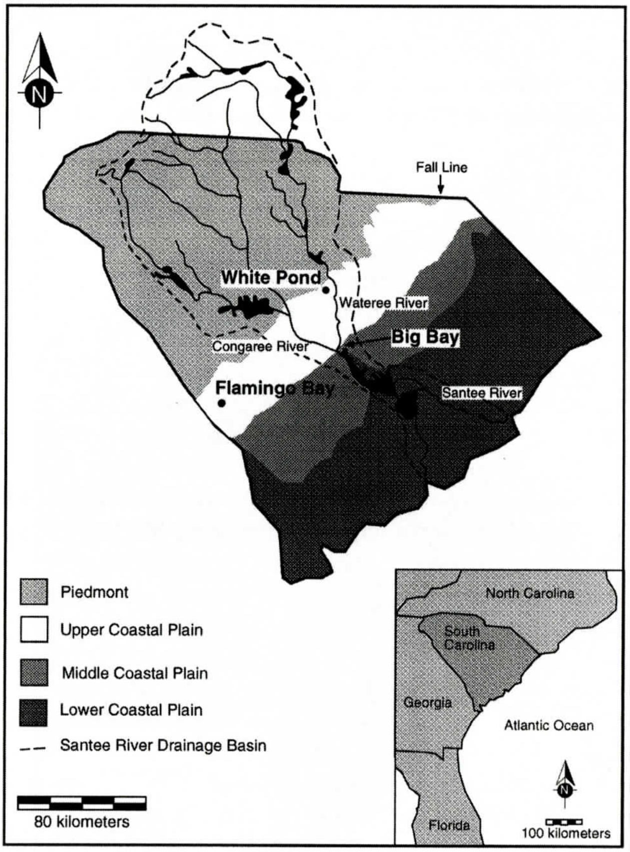

Figure 1. Study region (modified from Sexton, 1999, figure 1). Big Bay is located east of the Wateree River on the Middle Coastal Plain of South Carolina.

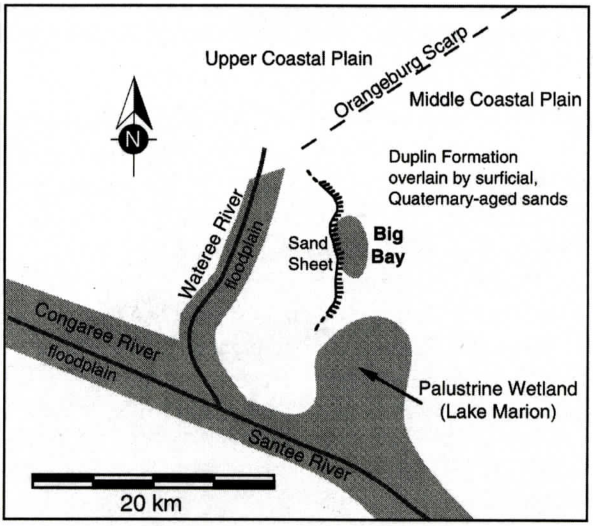

Figure 2. Geomorphic setting of Big Bay.

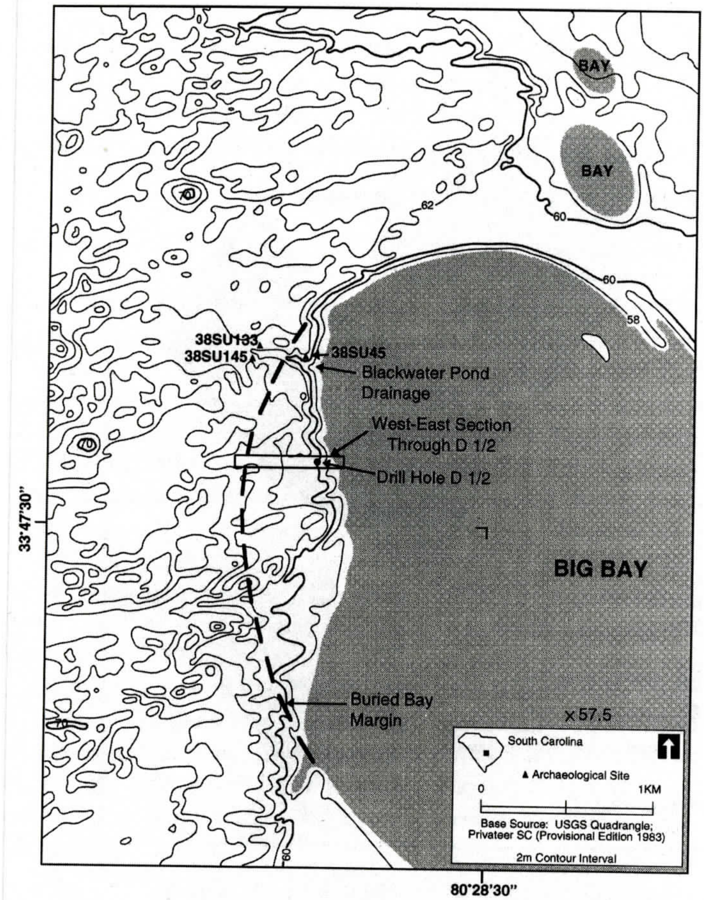

Figure 3. Big Bay and the eastern margin of the Wateree River-to-Big Bay sand sheet. Note the encroachment of the sand sheet into the west side of Big Bay. Note also the SW-NE trending of the sand sheet dunal topography and the NW-SE orientation of the bays immediately east of the leading edge of the sand sheet.

chronology and process.

GEOLOGIC SETTING

The project area is located on the United States Air Force Poinsett Electronic Combat Range in Sumter County, South Carolina. Big Bay lies seaward of the Orangeburg Scarp at the leading (eastern) edge of a large sand sheet (Johnson, 1959; Colquhoun, 1965; Pitts and others, 1974: Sheets 108, 109, 117, and 118; Willoughby, 1997a, 1997b). The sand sheet is near the floodplain of the Wateree River, 10 km west of Big Bay (Figures 1 and 2). Braided stream deposits at the confluence of the Piedmont-draining Wateree and Congaree Rivers were probably the immediate source of sediments for the sand sheet (Brooks and others, 1998). The toe of the Orangeburg Scarp is the boundary between the Middle and Upper Coastal Plain, and it marks the landward extent of the oldest Pliocene marine transgressive-regressive sequence (Lower Pliocene, Duplin Formation) preserved at the surface in the central Carolinas (Soller and Mills, 1991; Ward and others, 1991; Nystrom and Willoughby, 1992; Campbell, 1998; Willoughby and others, 1999). Carolina bays are abundant on upland surfaces outside the sand sheet but are absent from the sand sheet itself, which may have buried and obscured

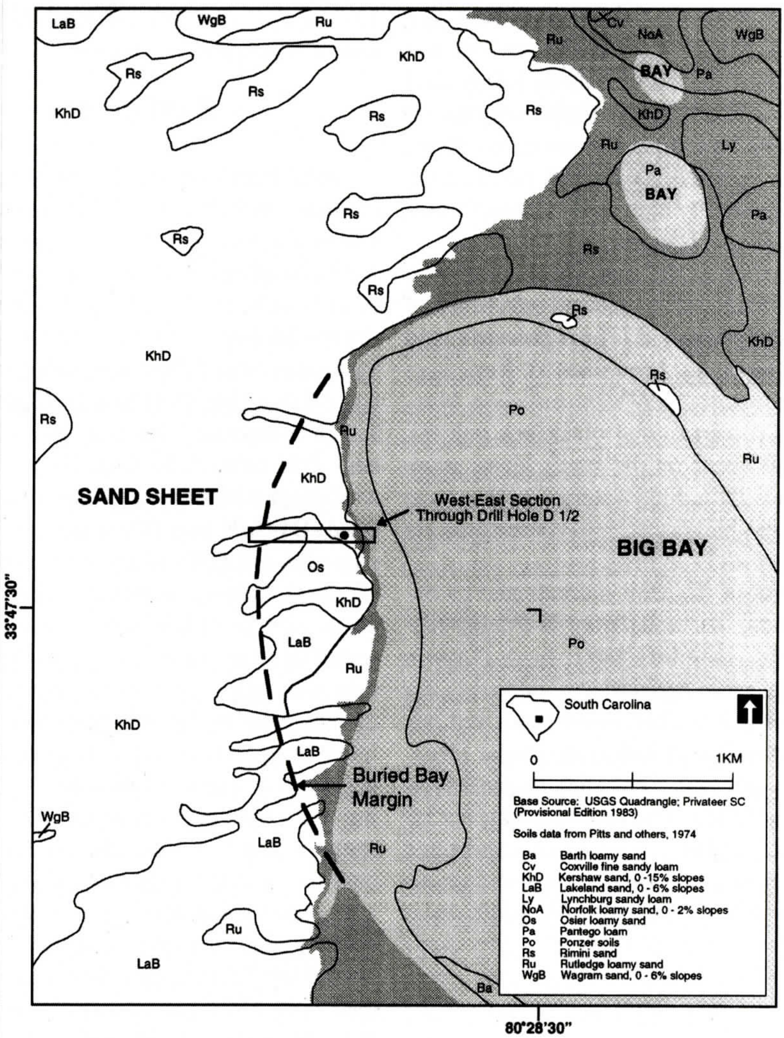

Figure 4. Soil associations of Big Bay and the eastern margin of the Wateree River-to-Big Bay sand sheet (same base map and scale as Figure 3).

some bays.

Big Bay, approximately 5 km long and 3 km wide, is the largest of the bays in the project area. The oval outline of the bay is interrupted by the sand sheet, and elevation decreases abruptly (~3 m) into the bay (Figure 3).

The soils of the sand sheet are mainly Kershaw and Lakeland sands with patches of Wagram and Rimini sands and Osier loamy sand (Pitts and others, 1974; Figure 4). The Kershaw sand, predominant at the study site, occurs on crescent-shaped and parallel ridges in nearly level to strongly sloping, deep, loose, excessively drained, sandy Coastal Plain sediments. Kershaw sand has very low organic content, very rapid permeability, slow runoff, and a very low available water capacity. Texture throughout the A-AC-C profile is sand to coarse sand to a depth of 2.03 m (80 in; standard depth for soil characterization). The C horizon is poorly developed and typically has less than 5 percent fines. Lakeland sand is similar with coarse to fine sand throughout the profile; the C horizon has 5 to 10 percent fines.

The soils of the interior of Big Bay consist mainly of Ponzer soil surrounded by a bay-margin border of Rutledge loamy sand (Pitts and others, 1974; Figure 4). The Ponzer soil is a very dark brown muck underlain by loamy sand or sandy loam. The Rutledge loamy sand has an A horizon with high organic content; the C horizon is sand. Permeability is rapid, but drainage is usually impeded by high water tables. Consequently, ponding is frequent.

The vegetation of the sand sheet adjacent to Big Bay is dominated by pine and oak. Much of the pine, mainly longleaf (*Pinus palustris*) and slash (*P. elliottii*), has been planted. Xeric species such as turkey oak (*Quercus laevis*), bluejack oak (*Q. incana*), and blackjack oak (*Q. marilandica*) occur on the sand ridges; more mesic species including water oak (*Q. nigra*) and mockernut hickory (*Carya tomentosa*) are found in the troughs.

The interior of Big Bay is dominated by hydric vegetation. The understory is a dense tangle of shrubs including fetter-bush (*Lyonia lucida*), titi (*Cyrilla racemiflora*), inkberry (*Ilex glabra*), blueberries (*Vaccinium* spp.), red bay (*Persea borbonia*), sweetbay magnolia (*Magnolia virginica*), and vines such as greenbrier (*Smilax laurifolia*). The sparse overstory includes loblolly pine (*Pinus taeda*) and pond pine (*P. serotina*), pond cypress (*Taxodium ascendens*), and scattered stands of white cedar (*Chamaecyparis thyoides*).

METHODS

Paleowind Directions for Bays and Dunes

To estimate direction of winds prevailing during the formation of the bays and dunes, we measured the orientations of 40 distinctive dunes near the leading edge of the sand sheet and of 22 Carolina bays east of the sand sheet on the Privateer Quadrangle, SC, 7.5 Minute Series (Topographic), 1983, US Geological Survey. We followed the methods used by Carver and Brook (1989) in their comparison of paleowind directions in the formation of bays and inland dunes in Georgia and the Carolinas. We calculated the paleowind direction as normal to the long axis for Carolina bays (Kaczorowski, 1977) or parallel to the presumed main direction of deposition for dunes.

Stratigraphy

Preliminary coring with a Dutch gouge auger included samples along a transect extending ~50 m into the wetland area of the bay, immediately east of drill hole D1/2 (Figure 3). Samples at drill hole D1/2 were taken using a truck-drawn Model B-34 Mobile Drill Rig with 10.16-cm (4-in) diameter, solid-stem augers and a fishtail bit. Drill hole D1/2, described below in composite form, consisted of three adjacent drill holes. The first, D1, delineated the stratigraphy and depth ranges; the second and third, D1-A/B and D2, were used to retrieve samples in 1.52-m (5-ft) auger sections (flights). During sample retrieval, the auger stems were pulled, sampled, cleaned, and reinserted into the ground every 5 ft. The threading rate was held equal to the penetration rate, thereby keeping the sediments stratigraphically intact by prohibiting up-hole extrusion. The auger was then extracted without rotation, producing a continuous column for the collection of samples. Samples were taken immediately adjacent to visually obvious stratigraphic boundaries, and their depths were recorded in meters below surface (m BS). Sediment contaminants encountered during up-hole extraction from the drill hole walls were carefully removed from the outer surfaces of the samples.

Microfossils, Carbon Content, and Radiocarbon Dates

Slide smears from all samples were examined for the presence of microfossils at 200X with a compound microscope by Evelyn Gaiser (Savannah River Ecology Laboratory). Pollen was analyzed by Jean Porter (University of Georgia) following standard procedures described by Faegri and Iversen (1989). Woody peat stem fragments were identified by Gary Crites (University of Tennessee); radiolarian microfossils, by Maria Poli (University of South Carolina). Organic carbon content was measured by loss-on-ignition according to

Table 1. Calculated paleowind directions for dunes and bays.

Feature	Frequency	Paleowind Direction (North Azimuths) Mean	Range
All dunes combined	40	259°	221°–288°
Parabolic dunes	3	267°	263°–270°
Infilled u/v-shaped parabolic dunes	30	260°	228°–288°
Barchan dunes	1	267°	-
Linear (seif, longitudinal) dunes	6	246°	221°–270°
Bays	22	241°	232°–248°

Dean (1974) by Jean Porter. Radiocarbon dating was performed by Beta Analytic, Inc. (Miami, Florida) using Accelerator Mass Spectrometry (AMS).

RESULTS

Paleowind Directions for Bays and Dunes

The orientations of the dunes and bays indicate prevailing winds generally from the southwest during formation (Table 1). On average, the wind direction was 18° more westerly for dunes than for bays, and the range was greater for dunes (221°–288°) than for bays (232°–248°). Although dunes were formed by prevailing winds from the southwest, many show evidence of modification by winds from the northwest. For example, leading edges or "noses" of some parabolic dunes are bent to the southeast. The orientations of linear or seif dunes, which are typically formed by winds coming from two different directions during the year (e.g., Ahlbrandt and Fryberger, 1982), also suggest paleowinds from both the southwest and northwest. Dunes are readily modified by shifting winds, and the fluidity of their forms may explain their broader range of orientations, in comparison with bays. This fluidity also introduces the possibility for greater error in measured orientations and inferred paleowind directions.

Stratigraphy

In the wetland area of the bay immediately east of drill hole D1/2, white quartz sand with negligible clay was recovered with the Dutch gouge auger at ~80 cm below the bay sediment surface; it probably represents the top of the sand sheet (Figure 5). Drill hole D1/2 was drilled to a depth of 10.61 m BS, revealing three distinctly different depositional packages: the Duplin Formation at the base; an intervening, organically enriched bay-fill sequence; and the overlying sand sheet (Figures 5 and 6).

Duplin Formation

The deepest deposit in drill hole D1/2 is the Duplin Formation. The location immediately seaward of the Orangeburg Scarp (Figures 1 and 2), the stratigraphic position and elevation above present mean sea level (Figures 5 and 6), the nature of the sediments (Figure 6), and the fragmentary radiolarian microfossils (Figure 6) indicate a shallow shelf marine depositional environment (Cronin and others, 1984; Owens, 1990; Colquhoun and others, 1991; Soller and Mills, 1991; Ward and others, 1991; Campbell, 1993, figure 1; D.J. Colquhoun, 1997, personal communication). This formation has been mapped in detail in the vicinity of Big Bay (Nystrom, 1992; Willoughby, 1997a, 1997b). The radiolarian microfossils were too fragmentary for further identification. The age of the Duplin Formation is late Early Pliocene (~3.4–3.0 Ma, Ward and others, 1991; Campbell, 1998).

Willoughby (personal communications, 1997–2001; 1997a; 1997b) identified the Duplin Formation in the vicinity of Pinewood at the south end of Big Bay at elevations approaching ~70 m amsl (top) to 47–50 m amsl (base). The Duplin Formation at Big Bay (51 m amsl and below) lies within the deepest part of this range. Thus, the abrupt boundary between the Duplin Formation and the overlying bay fill (Figures 5

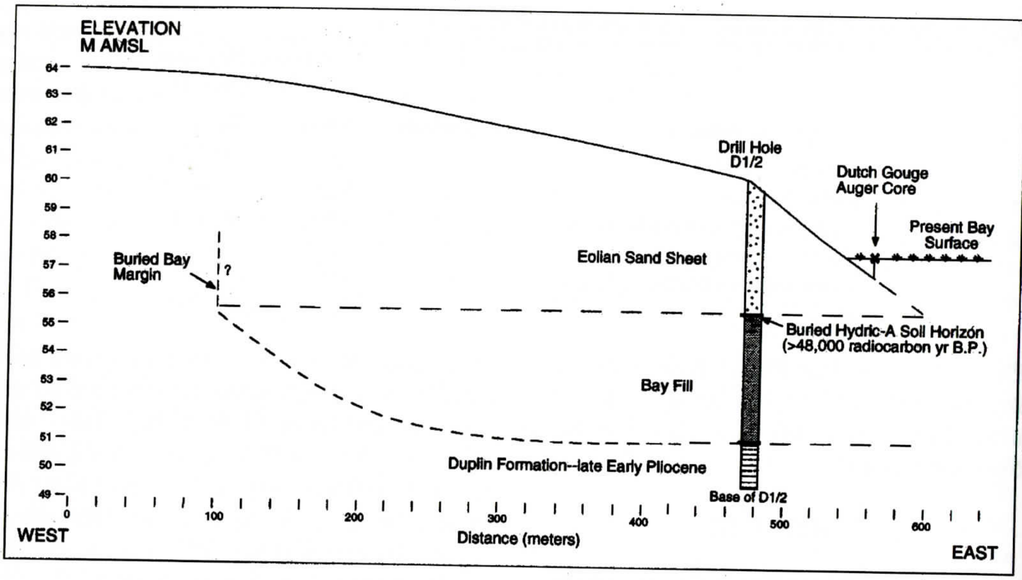

Figure 5. West-east schematic section through the western margin of Big Bay at drill hole D1/2 (see Figure 3 for location).

and 6) is likely an erosional unconformity resulting from scour into the upper portion of the formation during bay development.

Bay-Fill Sequence

The bay-fill sequence consists of distinctive beds, most of which are dominantly quartz sand (Figure 6). Exceptions are the upper 5 cm of D1-A-1, which was immediately below the sand sheet and contained peat (conifer stem fragments, probably *Pinus*), and D2-1, which was at the base of the sequence and consisted of sandy mud. Abrupt boundaries (Figure 6) indicate changes in the depositional environment. The lighter hues and sparser organic material of the lower beds (below D1-B-4) suggest unvegetated deeper-water habitats; their color and texture approximate the soil profile of Rutledge loamy sand. The Rutledge loamy sand in the basin was probably derived from sand introduced by alluvial or colluvial processes, with eolian processes also contributing on the western side of Big Bay. Among the upper bay-fill beds, the darker colors, microfossils, and abrupt upper boundaries of D1-B-4 and D1-B-3 suggest that they may have developed as Hydric-A soil horizons during earlier periods of surface stability. These were probably vegetated, shallow-water habitats. The stratigraphic position of D1-A-1 immediately beneath the sand sheet, along with its color, texture, and biotic content, indicate that it was a Hydric-A soil horizon.

At 57.5 m amsl, the modern bay surface is approximately 6.55 m above the base of the bay-fill sequence. Thus, the basin of the bay was 6.55 m deeper at this location before filling began. At least some of this depth probably represents scour into the underlying Duplin Formation during early, high-energy, open-water phases of bay development (Thom, 1970; Kaczorowski, 1977). The 6.55-m depth is not excessive for a bay that is 5 x 3 km in size. The large area and depth of the bay may be consequences of the relatively flat terrace (Duplin Formation), which would have provided little obstruction for erosion caused by wind-generated currents in ponded water.

Sand Sheet

No internal stratification was observed in the sand sheet (Figures 5 and 6). The soil profile is consistent with Kershaw sand, in which the C horizon is essentially pedogenically unmodified. Only the base of the sand sheet, whose abrupt lower boundary represents a discontinuity with the underlying bay-fill sequence, was

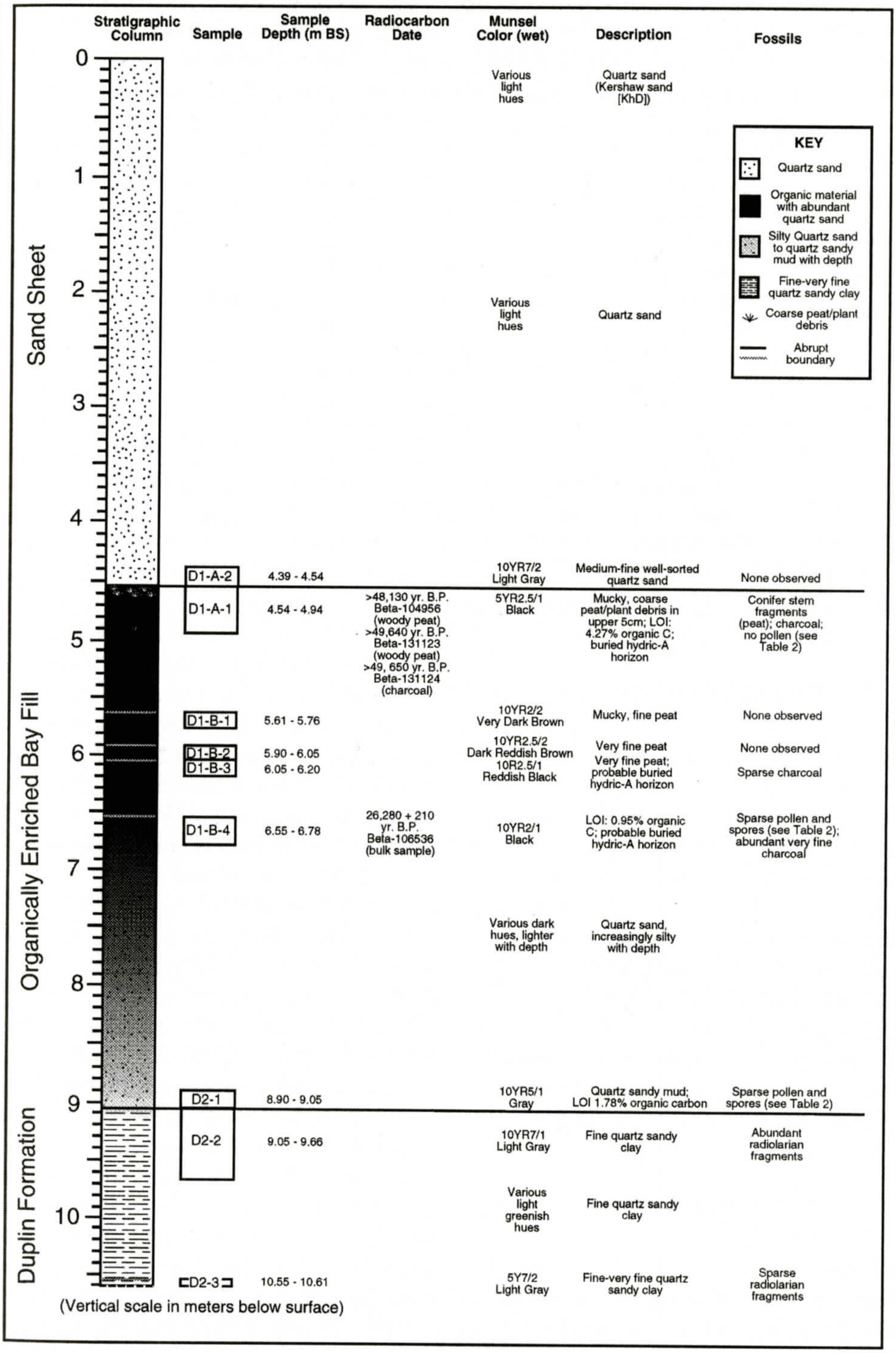

Figure 6. Drill hole D1/2 composite stratigraphic section (see Figure 3 for location). Surface elevation at D1/2 is ~60 m amsl.

Table 2. Pollen analyses of D1-B-4 and D2-1 samples from drill hole D1/2 (see Figure 3 for location).

	D1-B-4 (6.55–6.78 m)		D2-1 (8.90–9.05 m)	
Pollen type	Number	Per cent	Number	Per cent
Pinus	66*	40.2%	3	1.9%
Quercus	-	-	5	3.1%
Carya	5	3.1%	3	1.9%
Ulmus-type	1	0.6%	-	-
Fraxinus	2	1.2%	2	1.2%
Nyssa	-	-	1	0.6%
Liquidambar	-	-	1	0.6%
Ilex	-	-	3	1.9%
Salix	1	0.6%	-	-
Ericaceae	1	0.6%	-	-
Asteraceae	31	18.9%	9	5.6%
Poaceae	36	21.9%	61	37.7%
Unknown	-	-	6	3.7%
Unidentifiable	21	12.8%	68	41.9%
TOTAL	164	100%	162	100%

*55% broken

sampled. It consists of clean, medium-fine, well-sorted quartz sand (Figure 6).

Radiocarbon Dates

AMS radiocarbon dates of the bay-fill sequence were obtained on bulk sediment from D1-B-4 and on conifer stem fragments and charcoal from the upper 5 cm of D1-A-1 (Figure 6). The younger date obtained for the deeper sample was probably caused by contamination or other problems associated with dating of bulk soil material. The three samples from D1-A-1, two of large woody peat fragments and one of charred wood fragments, yielded consistent results. According to the radiocarbon dates, these samples are too old to be assigned finite ages (e.g., Stone and Brown, 1981; Rosholt and others, 1991).

Pollen

Samples from D1-B-4 and D2-1 in the bay-fill sequence were submitted for pollen analysis because preliminary observations indicated that pollen was present in these samples only. A sample from D1-A-1, immediately beneath the sand sheet, was also submitted but, as expected, no pollen was found. Identifiable pollen was dominated by *Pinus* (pine), Poaceae (grasses), and Asteraceae (asters and other composites) in D1-B-4 and by Poaceae in D2-1 (Table 2).

The well-preserved late Pleistocene-Holocene pollen at White Pond (Figure 1) in Kershaw County, South Carolina (Watts, 1980), provides a basis for interpreting the environmental conditions represented by the sparser pollen at Big Bay. White Pond is a Carolina bay located about 50 km north-northwest of Big Bay in the Sand Hills region of the Upper Coastal Plain. From the pollen assemblages, Watts defined a *Pinus-Picea*-Herb zone dominated by northern pines from 19,100–12,800 yr B.P., a *Fagus-Carya* zone dominated by deciduous trees from 12,800–9,550 yr B.P., and a *Pinus-Liquidambar* zone in which dominance gradually shifted from oak to pine from 9,550 yr B.P. to the present. He inferred cold, dry climate for the earliest interval, which spans the glacial maximum at ~18,000 yr B.P., cool moist climate for the middle interval, and warmer, essentially modern conditions during the latest interval, which spans most of the Holocene.

The tree pollen from both Big Bay assemblages almost certainly predates the White Pond record. No typically boreal taxa occurred in the two pollen inventories of Big Bay; all of the taxa occur in the modern vegetation of Big Bay.

The sparse tree pollen of the lower Big Bay sample (D2-1) includes *Liquidambar* and *Nyssa*, taxa restricted to the Holocene assemblages at White Pond. The pine-dominated assemblage in the upper Big Bay sample (D1-B-4) lacks species that would associate it distinctively with pine-dominated assemblages of either the Holocene or full glacial periods at White Pond or at sites in southeastern North Carolina (Frey, 1951; Whitehead, 1964). Pollen of two *Pinus* species associated with cooler or "northern" climates in southeastern North America (Whitehead, 1964; Watts, 1980) are typically smaller than pollen of other species. Too few intact grains were present in D1-B-4 to yield an adequate size frequency distribution, but the pine pollen seemed to be of the larger types, thus not diagnostic of either warm or cool climate.

High proportions of herbs and grasses in the Big Bay pollen samples (Table 2) suggest prairie-like or savanna-like vegetation with sparse or patchy trees. Such conditions would have been likely on the advancing sand sheet to the west of the bay. For comparison, values for White Pond are generally <5% grasses and ~5% herbs in the *Pinus-Picea*-Herb and *Fagus-Carya* zones, and 5–10% grasses and traces of herbs in the *Pinus-Liquidambar* zone (Watts, 1980, figure 2). The upper Big Bay pollen sample contains a substantially higher proportion of herbs of the family Asteraceae than the lower sample. Watts (1980) associated the Asteraceae-dominated herb assemblage of the full glacial *Pinus-Picea*-Herb zone with cool, dry conditions at White Pond. He also noted its similarity to an assemblage from Lake Annie, Florida (Watts, 1975), that may have been present when dunes were active around the lake.

DISCUSSION

Formation and Infilling of the Bay

The late Early Pliocene age of the underlying Duplin Formation sets an upper limit for the age of the bay. The bay was probably shaped by the action of wind on ponded water (Thom, 1970; Kaczorowski, 1977) in an upland depression. Its evolution probably involved lateral expansion and orientation of the basin through shoreline erosion, as well as scour into the Duplin Formation (Figure 5). Both bays and inland dunes in the vicinity of Big Bay formed under prevailing westerly winds. The bays, whose orientations were probably fixed earlier than those of the dunes, indicate more southerly winds. The sands for dune formation most likely were swept from sparsely vegetated, braided channels of the Wateree River floodplain to the southwest.

The great thickness (~4.5 m) of the bay-fill sediments beneath the sand sheet suggests an extended period of sedimentation. The higher proportions of silt and clay and the absence of peat in the lower portion of the sequence (below 6.78 m BS) indicate open-water habitats, which would have permitted the high-energy, water-driven erosional processes required to shape the basin. The depositional environment changed substantially at D1-B-4 (6.55–6.78 m BS), and the D1-B-4 surface may represent a transition from open water to wetland vegetation. Abrupt stratigraphic boundaries in the upper half of the bay fill suggest that encroachment of the sand sheet was episodic. At the top of the bay-fill sequence (D1-A-1), the conifer stem fragments in the woody peat indicate a terrestrial or wetland forest. The absence of pollen from the upper portion of the bay-fill sediments implies more severe desiccation than at the lower levels (D1-B-4 and D2-1), where pollen was present.

The possible Hydric-A soil horizons at D1-B-4 and D1-B-3 suggest intervals of surface stability or, at least, slower sedimentation. However, the predominance of siliciclastic material in the bay-fill sediments indicates frequent influxes of sediment, probably mainly as slopewash (alluvial, colluvial, or both) from the basin slopes or as eolian input in advance of the sand sheet. Occasional episodes of high water may also have occurred. During such episodes, sediments eroded from the shoreline may have been redistributed by currents within the basin.

The vegetation, possibly pine forest, at the D1/2 drill hole location was buried by an encroaching wedge of sandy sediment. The sharp contact between the sand and the organic bay fill, as well as the good preservation of the peat, suggest that burial occurred rapidly, perhaps during a drier period when surface sediments were more mobile. The steepness of the leading edge of the sand sheet just east of the D1/2 drill hole (see Figure 5) also suggests deposition into a vegetated environment rather than into open water.

Chronology and Paleoenvironment

The radiocarbon dates for the buried peat horizon (D1-A-1) place a major period of dune activity on the sand sheet at >48,000 radiocarbon yr B.P. This minimum date for Big Bay is not inconsistent with Ivester's (1999) earlier period (>140–65 ka B.P.) of dune activity in Georgia. A period of dry climate may have been conducive to the massive transport of sand into Big Bay. The presence of conifer macrofossils in the buried peat indicates little or no water in that area of the basin, due to dry climate, infilling, or both. If the climate was cool and dry, the activity at Big Bay could plausibly have occurred during an early Wisconsinan glacial advance, perhaps during Oxygen Isotope Stage 4 (75–65 ka B.P.; we use the dates summarized by Rapp and Hill, 1998, for oxygen isotope stages). Successive burial of soil horizons in the upper portion of the bay fill suggests that movement of the sand sheet was episodic.

A few clues about conditions preceding the migration of the sand sheet into the bay are given by the pollen assemblage at D1-B-4. The terrestrial habitat adjacent to the bay was plausibly open, perhaps pine savanna. The high proportion of Asteraceae indicates cool, dry climate or dune habitat (Watts, 1975, 1980). The transition from open water to wetland habitat, suggested by the possible A horizon at this level, may represent the effect of drier climate or infilling.

At the base of the bay-fill sequence, the sediments indicate open-water habitat, and the pollen (D2-1) suggests grass-dominated terrestrial vegetation. Frey (1952) reported a similar list of tree taxa, including *Liquidambar* and *Nyssa*, from the Horry Clay near Myrtle Beach, South Carolina; that formation is believed to be of Sangamon interglacial age (Smiley and others, 1991).

If the tree pollen indicates a warm climate, then infilling of the bay may have begun during a warm interval, perhaps early Oxygen Isotope Stage 3 (65–32 ka B.P., a weak interstadial period of the Wisconsinan; Morrison, 1991) or Oxygen Isotope Stage 5 (135–75 ka B.P., Sangamon interglacial and early Wisconsinan). Obviously, it would be necessary to infer the earlier time if encroachment of the sand sheet occurred during Oxygen Isotope Stage 4. We note that pollen records from Florida, extending back to 70 ka, indicate wide fluctuations in environmental conditions during the Wisconsinan (Watts and Hansen, 1988; Grimm and others, 1993; Jacobson and others, 1999).

Holocene Activity of the Sand Sheet

The archaeological record in the upper 1 m of the sand sheet indicates that movement of sediment on the sand sheet continued into the Holocene. Archaeostratigraphic data were obtained from three archaeological sites (38SU45, 38SU133, 38SU145) in the Blackwater Pond drainage at the leading edge of the sand sheet where it encroaches into Big Bay (Figure 3; data from Brooks and others, 1998). Buried occupation surfaces and their ages, as inferred from temporally diagnostic artifacts, are depicted in stratigraphic section in Figure 7.

The vertical separation of the early and mid-Holocene occupation surfaces suggests periods of sand sheet stability, punctuated by periods of reactivation that resulted in burial of the occupation surfaces. Rapid burial is indicated by discontinuous or polymodal vertical distributions of artifacts and *in situ* cultural features, rather than more uniform distributions that develop when surface accretion is very slow with low rates of sediment input (Brooks and Sassaman, 1990; Brooks and others, 1990; Brooks and others, 1998). However, the depositional surfaces mainly follow existing contours, suggesting that the shapes of landforms were not greatly modified.

The very shallow burial of later occupation surfaces indicates that the sand sheet was essentially stabilized about 3000yr B.P. The considerable lateral variation in depth of burial of the

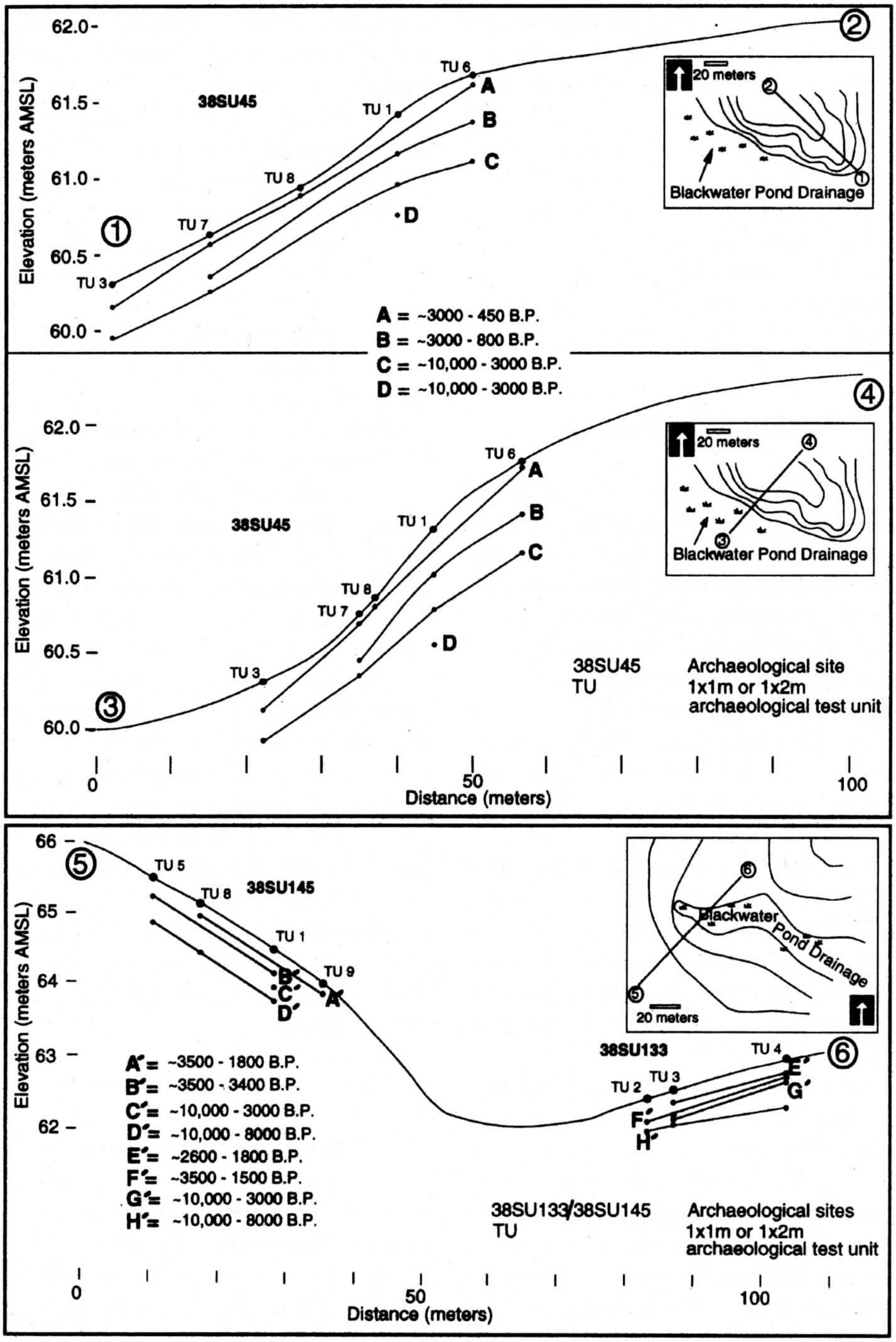

Figure 7. Archaeostratigraphic data from the Blackwater Pond drainage (see Figure 3 for location). Upper panel depicts NW-SE and NE-SW cross-sections of the archaeological site 38SU45 landform. Lower panel depicts NE-SW cross-section of the landforms containing archaeological sites 38SU145 and 38SU133. The depths, surface trends and minimum potential age ranges of buried surfaces, inferred from backplotting archaeostratigraphic data against a common baseline, document Holocene-era activity of the sand sheet.

more recent occupation surfaces indicates that the depositional processes were operating at a small spatial scale (i.e., slopewash on the order of tens of meters) and were, most likely, anthropogenically-induced (Brooks and others, 1998). The A-AC-C profile typical of the Kershaw sand soil indicates weak pedogenic development, which may be attributable both to inherited sediment characteristics (i.e., deep, loose sand with a low content of fines) and to this comparatively recent stabilization of the landform.

Regional Processes

Nystrom (1993) stated that "it is likely that strong southwesterly winds have been a factor in modifying the surface of the Coastal Plain, including the formation of Carolina bays, from the late Miocene (or possibly earlier) until recent time." Both the bays and dunes in the vicinity of Big Bay formed under conditions when atmospheric circulation patterns generated prevailing westerly winds. The bays, whose orientations were probably fixed earlier than those of the dunes, indicate more southerly winds. Similarly, for bays and dunes in Georgia and South Carolina, Carver and Brook (1989) found that paleowind directions were on average about 10° more westerly for dunes than for bays. They were uncertain whether the deviation indicated different conditions during formation or whether it was an accident of sampling.

The early date for the shutdown of the high-energy, open-water phase of morphological evolution at Big Bay suggests that these phases were not contemporaneous among bays throughout the Coastal Plain. At Big Bay, this phase was completed at least before >48,000 radiocarbon yr B.P., and perhaps as early as the Sangamon interglacial. For bays near the modern coast of South Carolina in Horry and Marion counties, Thom (1970) inferred shutdown of the open-water phase around 7000 yr B.P. This change may have been associated with rising sea level, which was approaching modern elevations by then (Brooks and others, 1986). Brooks and others (1996) speculated that Flamingo Bay (Figure 1), a small Carolina bay in Aiken County, South Carolina, in the Upper Coastal Plain, became transformed from a shallow open-water lake to a vegetation-filled wetland pond during the mid-Holocene. In contrast, some bay lakes in North Carolina, such as Singletary Lake and Jones Lake, presently have open water with substantial stretches of shoreline exhibiting high-energy beach-face processes.

Holocene deposition on the Wateree River-to-Big Bay sand sheet has been studied only through archaeological investigations on the west side of Big Bay. At several other bays on the Upper Coastal Plain of South Carolina, archaeological records also indicate low rates of episodic eolian deposition during the Holocene. On the eastern rim of Flamingo Bay, for example, archaeological stratigraphy indicates that at least 50 cm of sand was deposited during the Holocene (Brooks and others, 1996).

At this point little can be said about the conditions that gave rise to the braided channel regime that existed in the Wateree River. However, if the dune sands were derived from the braided channels, as seems likely, then we can infer that the braided regime must have been established before encroachment of the sand sheet into Big Bay. In any event, further coordinated study of bays, dunes, and braided channels is likely to reveal much about the climatic, geomorphic, and ecological history of the Coastal Plain.

CONCLUSIONS

Several lines of evidence suggest some general conclusions about the timing and nature of processes influencing development of bays and dunes in the vicinity of Big Bay.

- Big Bay formed no earlier than the Early Pliocene on marine sediments of the Duplin Formation.
- Both bays and inland dunes in the vicinity of Big Bay formed under conditions when atmospheric circulation patterns generated prevailing westerly winds. The bays, whose orientations were probably fixed earlier than those of the dunes, indicate

more southerly winds.

- The sand sheet buried a vegetated area about 500 m inside the western edge of the bay more than 48,000 radiocarbon yr B.P. Big Bay and the braided floodplain of the Wateree River must have been fully formed well before this minimum date.
- Pollen indicates that the bay began infilling during a period when terrestrial vegetation adjacent to the bay was open and dominated by grasses. The tree taxa suggest a warm climate, not dissimilar from the modern climate, perhaps during early Oxygen Isotope Stage 3 (65–32 ka B.P.) or Oxygen Isotope Stage 5 (135–75 ka B.P., Sangamon interglacial and early Wisconsinan).
- Increasing sand content upsection in the bay-fill sediments suggests increasing contribution from the sand sheet to the west of the bay.
- Pollen from the middle of the bay-fill sequence indicates an open, pine-dominated terrestrial community, plausibly but not diagnostically associated with cool climate, perhaps during Oxygen Isotope Stage 4 (75–65 ka B.P.) or earlier.
- Buried soil horizons in the upper half of the bay-fill sequence suggest that encroachment of the sand sheet was episodic, with a final major episode occurring prior to 48,000 radiocarbon yr B.P. Conifer macrofossils suggest relatively dry conditions in the basin at that time; dry climate seems necessary to have permitted the massive transport of sand. If the climate was warm and dry, this event could have occurred during Oxygen Isotope Stage 3 (65–32 ka B.P., mid-Wisconsinan); if cool and dry, during Oxygen Isotope Stage 4 (75–65 ka B.P., an early Wisconsinan glacial advance).
- Archaeological evidence indicates that redistribution of sediment on the sand sheet continued at least intermittently into the Holocene, although activity may have slowed after 4–3 ka B.P.
- The early date for the shutdown of the high-energy, open-water phase of evolution at Big Bay suggests, in contrast with dates for some other bays, that these phases were not (and are not) contemporaneous among bays throughout the Coastal Plain.

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THE PALEOECOLOGY AND DEPOSITIONAL ENVIRONMENTS OF THE McCLELLAND SANDPIT SITE, DOUGLAS, GEORGIA

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ABSTRACT

Evidence from the Atlantic Coastal Plain of Georgia, South Carolina, and North Carolina shows that, beginning about 25,000 years ago and extending up until about 4,000 years BP, alternating periods of wet and dry climate typified the Southeast. Palynological evidence from a number of sites in several southeastern states suggests that prairie-like vegetation was dominant in this region during the mid-Holocene, particularly; this is believed to be indicative of arid conditions during that time. Differences in the paleoclimate and vegetation would have had significant impacts on the landscape morphology of the Southeast. Evidence of these physical, environmental, and biological changes has been preserved at the McClelland sandpit site in southeastern Georgia. The depositional events recorded at the site were active during the mid-Holocene dry period known as the Hypsithermal. The flora of the site was characterized by prairie, or southern pine savannah. Two generations of pine trees are preserved at the McClelland site. These were inundated with coarse, angular quartz sand beginning around 7000 years ago. We hypothesize that this may have been the result of a stream channel that became avulsed, spilling its coarse bedload across the landscape. The site was finally covered by an essentially structureless aeolian dune deposit, the accumulation of which is consistent with coeval dune activity across the Southeast during the Hypsithermal. The most recent period of landscape evolution, which resulted in the burial of worked flint flakes, might have been approximately 2000-3000 years ago.

INTRODUCTION

The southeastern United States has experienced dramatic changes in climate and landscape morphology since the last Pleistocene glaciation. Markewich and Markewich (1994) found, in a study of inland dunes on the Coastal Plain of Georgia and the Carolinas, that the climate of the late Pleistocene to middle Holocene was characterized by aridity. This was due to both global climatic influences and frequent severe droughts that, aided by strong west-southwesterly winds, made possible the formation of multiple dune deposits along major rivers. During the mid-Holocene Hypsithermal interval (8000-4000 years BP), a warm, arid climate expanded eastward from the Midwest. This was characterized by prevailing westerly winds that brought considerable dryness to most of the eastern United States. The Hypsithermal interval also saw an eastward expansion of prairie grasses from the Midwest, which correlated with the climatic trend of increased warmth and aridity that extended from the mid-continent to the coast (Delcourt and Delcourt, 1984). Work on the vegetation history of Florida (Watts, 1969) and Georgia (Watts, 1971) indicates that much of the Coastal Plain was dominated by pine, sclerophyllous oak, and isolated prairie communities, with scattered marshes, between 8500 and 5000 years BP. Watts (1971) postulates that the presence of the more xeric prairie grasses may have resulted from either drier atmospheric conditions or from a depressed (by as much as 12 m) water table. As ocean-surface temperatures rose to their present-day levels and the Gulf Stream moved closer to the Atlantic Coast, increased moisture visited the southern states in the form of more frequent tropical storms. Conifer forests replaced the prairie grasses, and, by ~4000 years BP, the vegetation of the Coastal Plain province took on its modern character (Delcourt and Delcourt, 1984).

The Coastal Plain of Georgia and the Carolinas was, for much of the Quaternary, characterized by physiographic features that could form only under dry, windy conditions. Carolina bays, elliptical bodies of water whose axes aligned themselves perpendicular to the prevailing wind direction, formed during the Pleistocene. Though the origins of bays have been attributed to a variety of causes (Bliley and Burney, 1988; Carver and Brook, 1989; Markewich and Markewich, 1994), an element of aeolian erosion and deposition seems to have characterized their genesis. Parabolic dunes, whose axes were also aligned with a prevailing southwesterly wind direction, developed along the northeastern banks of major rivers all across the Southeast (see Pickering and Jones, 1974, Markewich and Markewich, 1994, and Daniels and others, 1969, for examples). These dunes began forming as early as 15,000 years BP (Markewich and Markewich, 1994).

Evidence from numerous studies, including Booth and others (1999), indicates that sea level was as much as 4 to 5 m below its current elevation about 6000 years BP, having risen from a maximum lowstand of nearly 100 m. Areas currently occupied by extensive freshwater wetlands (e.g., Okefenokee and Dismal Swamps were essentially dry basins during that lowstand (Rich, 1984, Whitehead, 1972). The increased gradient between the Piedmont and the coastline would have altered river morphology such that the wide, meandering rivers now extant on the Coastal Plain may have had different morphologies during the mid-Holocene. The Savannah River, for example, incised its channel into and flowed out across the continental shelf (Lennane and Rich, 2000). The kinds of changes that affected large streams probably affected smaller, ephemeral streams, whose sediment load and discharge volume would have been significantly different under the influence of the drier climate, when grassy vegetation occupied interfluves and increased stream gradients existed.

The current study was undertaken because a number of tree stumps had been uncovered by sand mining operations in Douglas, Georgia, adjacent to Seventeen Mile Creek (Figure 1). We believed that a study of the trees and the sediments that entombed them would shed light on the natural history of the area, including the dynamics of an ancient stream. When radiocarbon dates showed that the trees had died during the Hypsithermal, we further realized that we had a unique opportunity to improve our understanding of the ecology and geology of this unusual period of time.

LOCATION AND FIELD RELATIONS

The study site, known as the McClelland sandpit, is located north of Douglas, Georgia, one-half mile (0.80 km) northwest of U. S. Highway 221 N. The city of Douglas is located in central Coffee County (Figure 1). Coffee County is situated in the southern region of the state, 100 km north of the Georgia-Florida border. Coffee County is part of the Atlantic Coastal Plain, and lies in part in the Bacon Terraces, Vidalia Upland, and Tifton Upland physiographic provinces of Clark and Zisa (1976; Figure 2). Douglas and its immediate vicinity occupy the Hazelhurst and Pearson Terraces, most recently described by Huddlestun (1988). The sandpit and it surroundings are dominated

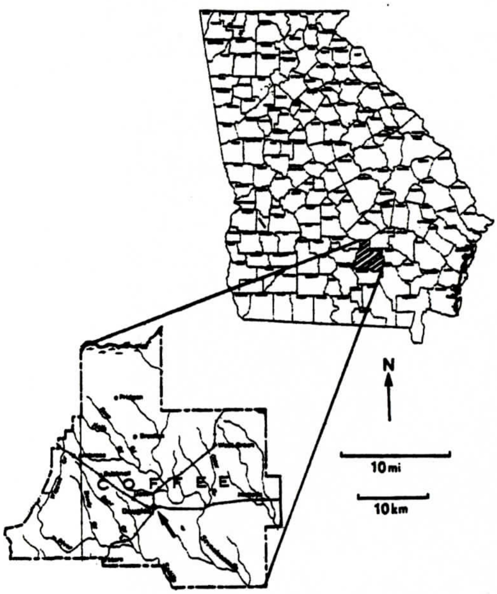

Figure 1. Location of the study area. Arrow indicates location of Douglas.

by the sandy soils and pine/scrub vegetation characteristic of the Coastal Plain, though swamp and marsh vegetation is found in areas of saturated soils. The UTM coordinates for the study site, herein referred to as the "McClelland Sandpit Site" are 17 E 327800 N 3490600

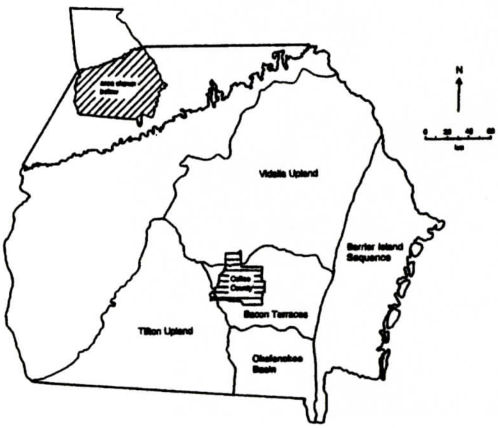

Figure 2. Location of Coffee County relative to the physiographic provinces of the immediate vicinity. The Bacon Terraces District consists of several subtle, moderately dissected marine(?) terraces that parallel the modern coastline. The McClelland Site developed on top of a terrace.

(Chris Trowell, personal communication, 1998). The site elevation is approximately 235 ft (71 m) above mean sea level (USGS, 1971).

The McClelland site is characterized by two distinct horizons of sand, the "upper sand" and "lower sand" bodies of this report (Figure 3). The upper sand body comprises the high walls of the pit. This upper sand is a white to yellow, medium- to fine-grained quartz sand that is free of organic debris. The upper sand is massive and structureless, though in one small area of the pit cross-bedded sands were seen during a field trip in June, 1998 (Figure 4). We interpret the high angle of dip in the beds to support their aeolian origin. The upper sand otherwise appears to be a homogeneous feature resulting from a single episode of emplacement. Localized indurated bands of clay-iron staining within the sand body probably represent fluctuations in groundwater levels over time, rather than being paleosols or other markers of separate depositional episodes (Andrew H. Ivester, personal communication, 1998). The upper sand also shows local evidence of fluvial incision, and contains fire-treated flint and chert flakes. The flakes we found were embedded in the high wall of the sandpit approximately 1.1 m below the ground surface. Most of these are interpreted to be waste flakes and are pink or yellow. The flakes are of various sizes, but average 2.5 cm long by 1.25 cm wide, with thicknesses of only a few millimeters; all show evidence of having been worked by human hands (Sue M. Moore, personal communication, 1999). Archaeological evidence from the Ocmulgee Big Bend and Satilla River areas indicates that Archaic-period peoples (8000 BCE-1000 BCE) utilized sand ridges adjacent to streams as temporary campsites during hunting expeditions (Snow, 1977). The presence of the flakes we found in the shallow subsurface of the upper sand suggests that, by the time of prehistoric human habitation at the site, most of the upper sand body had to have been emplaced, but that the sand was still mobile enough to cover the flint worker's debris.

Unlike the upper sand body at the McClelland site, the lower sand body is markedly heterogeneous, both laterally and vertically,

Figure 3. Upper (white) and lower (dark) sand bodies of the McClelland site are shown here. N the abundant tree stumps in the lower sand.

Figure 4. High angle cross-bedding in the upper sand, as seen in June, 1998. Note pick for sc (handle is 64 cm long).

throughout the pit. The lower sand is predominantly medium- to coarse-grained and is stained shades of brown and black. It contains significant quantities of organic matter in the form of clasts of wood and charcoal, and tree stum These stumps stand upright and are encased bedded sediment deposits. Most strikir though, is the fresh appearance of the woo

The trees at the site do not display the characteristics of long-dead, rotted and humified wood that would be expected of trees that had ceased to live several thousand years ago. Radiocarbon dates obtained for two of the trees show that they died between 5000 and 7000 years ago, yet the wood splinters just as a young, freshly-timbered tree would. The appearance of the wood leads us to postulate that these stumps were buried rapidly enough that they were not exposed to aerobic decay processes that would have broken down the wood's structure. The trees found in the lower sand are all upright, and the beds of sediment that comprise the lower sand all terminate in perpendicular contacts with the trees. One portion of the lower horizon, at the northern end of the site, consists of trees surrounded by finer-grained, clay-rich sediment beds that contain less organic debris than the rest of the lower sand.

METHODS

The layout of the McClelland sandpit was surveyed using pace-and-Brunton methods, as described in Compton (1985). Vertical distances (e.g. heights of trees, heights of pit walls) were measured with a steel tape measure, except in the case of the western edge of the pit; the height of the wall in that case was sighted using the clinometer on the Brunton compass, and then calculated using right-triangle relations (Compton, 1985). Errors of closure on the pace-and-Brunton traverses were corrected using the method outlined in Freeman (1992).

Fourteen trees and twenty-one sediment samples were examined in this study. Sediment samples were collected for sedimentological and palynological analyses, and were taken from each identifiable bed adjacent to each tree using a flat scraper, provided that the bed was thicker than about 10 cm (to ensure no contamination from adjacent beds during sampling). Wood samples were collected from Trees 1, 3, 4, and 7 for radiocarbon and morphological analyses. Sediment samples were also collected from the upper sand body, and were handled in the same manner as those collected from the lower horizon. Chert flakes from the high wall of the pit were collected and taken for identification to Dr. Sue M. Moore, Georgia Southern University Department of Sociology and Anthropology. The characteristics of the sand bodies as seen in the field, especially the clear, though superficial differences in shapes of the sand grains in the two layers of sand as seen in the sand pit, suggested that sedimentological analyses would prove useful in reconstructing the history of the site. Though we considered performing standard sieve analyses, we decided that a textural analysis of the sediment would provide us with important information relative to the provenance and maturity of the sand. Roundness appeared to be a particularly interesting variable since grains were often large and appeared to be angular, characteristics not typical of sediments being deposited in the area today. Though we searched for a standard method of preparing samples for shape analysis, we found none, so we devised our own method, which is a combination of palynological extraction techniques and those used to isolate heavy minerals and conodonts from matrix. Details are available in Zayac (2000).

In order to analyze individual sediment grains in the lab, the humic acid "cement" in the sediment samples had to be removed. To remove the organic material, sediment was treated with 1.78M KOH. Small quantities of sediment were mixed with the KOH, boiled, and repeatedly washed. In order to remove clays, samples were given a series of detergent baths to break up the clays and remove them from the samples.

Clay-containing sediment was placed in a test tube with a pinch of powdered laundry detergent and boiled vigorously for 10 minutes. Samples were centrifuged, then washed with distilled water. Sediment was then transferred to a soil-drying oven, set at 57 degrees C, and allowed to dry completely.

A preliminary analysis of the sediment showed that all samples contained heavy minerals. In order to determine the mineralogical identity of the sand-sized heavy minerals, a heavy-liquid separation was performed to extract the heavy-mineral component from the sediment. The separation apparatus consisted of

Table 1. Results of point-counting of heavy minerals, according to mineralogical identity. Percen ages given are out of 100% of the grain population counted.

Sample #	% Leucoxene + Rutile	% Ilmenite	% Zircon	% Other	Total grains counted
T1SA	1.00	32.50	38.00	28.50	200
T2SB	4.18	37.68	37.24	20.90	239
T3SC	12.86	21.43	34.28	31.43	70
T4S1	14.50	39.50	27.50	18.50	200
T4S2	13.97	18.44	33.52	34.07	179
T4S3	14.06	26.56	35.94	23.44	64
T5S4	4.00	29.50	42.00	24.50	200
T6S5	7.50	27.00	40.00	25.50	200
T7S6	11.00	20.50	40.00	28.50	200
T7S7	2.50	61.50	22.00	14.00	200
T7S10	10.71	22.62	36.31	30.36	168
T8S13	15.54	17.10	42.49	24.87	193
T9S11	10.00	30.50	39.00	20.50	200
T9S12	13.50	32.00	31.50	23.00	200
T11S16	15.56	24.44	31.11	28.89	45
T11S17	9.50	11.50	39.50	39.50	200
T12S18	16.38	16.38	24.14	43.10	116
T12S19	11.00	19.50	37.00	32.50	200
T13S20	12.50	16.50	35.50	35.50	200
T14S21	13.50	13.00	41.00	32.50	200
HL1S15	10.50	14.50	43.50	31.50	200
HWS8	4.00	31.00	33.00	32.00	200
HWS9	10.00	20.00	39.50	30.50	200
Total - all samples	9.74	25.87	36.31	28.08	4074

bromoform (**G** = 2.890) placed in a Pyrex funnel, which was clamped to a ring stand. A sediment sample was poured into the funnel and the grains were allowed to separate according to specific gravity relative to the bromoform. Heavy minerals (i.e., those with **G** > 2.890) settled to the bottom of the funnel, and lighter minerals (i.e., those with **G** < 2.890) floated atop the bromoform. Heavy mineral grains were washed with acetone and allowed to dry. The washing process was repeated for the light minerals.

Grain mounts were made of both light and heavy minerals, using glycerine jelly as the mounting medium. Both heavy-mineral and light-mineral fractions were point-counted in reflected light under a binocular microscope. Heavy minerals were counted for both mineralogical composition and roundness. Light minerals were counted only for roundness, though a qualitative assessment was made for each light mount as to the general mineralogical composi tion of the mount. The roundness classificatio system developed by Powers (1953) was used i visually assessing the roundness of sedimen grains. A maximum of 200 grains was counte on each slide, as recommended by Carve (1971). In cases of slides with less than 20 grains in the sample population (e.g., T7S7), th maximum number of grains to be found withou risking duplicate counting of grains was count ed.

Palynological analysis consisted of polle counts of sediment samples from Trees 1, 2, 3 4, 5, 8, and 9. Prior to counting, these sediment were treated with HF, HCl, and KOH to remov silicates, carbonates, and soluble organic mate rial. No acetolysis was done because we did no want to risk losing grains through unnecessar processing (the great age of the sediments sug gested that clearing them would be unnecessar in any case). At least two slides were prepare

Table 2. Results of point-counting of heavy minerals, based on roundness classification of Powers (1953). Percentages given are normalized to 100% of the grain population counted.

Sample #	% Very angular	% Angular	% Subangular	% Subrounded	% Rounded	% Very rounded	Total counted
T1SA	0.00	9.50	37.50	32.00	18.00	3.00	200
T2SB	4.00	26.50	32.50	27.00	10.00	0.00	200
T3SC	0.00	9.86	47.88	38.03	4.23	0.00	71
T4S1	0.00	7.00	37.00	39.50	16.50	0.00	200
T4S2	0.00	6.71	35.37	38.41	18.90	0.61	164
T4S3	0.00	14.06	29.69	39.06	17.19	0.00	64
T5S4	0.50	9.50	37.00	36.00	14.50	2.50	200
T6S5	0.00	0.50	39.00	51.50	9.00	0.00	200
T7S6	0.50	10.50	36.00	39.50	11.00	2.50	200
T7S7	0.50	15.50	23.00	30.50	18.00	12.50	200
T7S10	0.00	9.20	35.63	41.38	13.22	0.57	174
T8S13	0.00	5.50	35.00	51.00	8.00	0.50	200
T9S11	0.00	14.00	37.00	37.00	12.00	0.00	200
T9S12	0.00	2.00	38.50	50.00	9.50	0.00	200
T11S16	0.00	4.44	35.55	51.11	8.89	0.00	45
T11S17	0.00	2.50	36.00	48.00	12.00	1.50	200
T12S18	0.00	6.09	31.30	48.70	10.43	3.48	115
T12S19	0.00	1.50	30.00	56.50	12.00	0.00	200
T13S20	0.50	6.00	38.00	43.00	11.00	1.50	200
T14S21	0.00	1.50	25.00	51.00	17.50	5.00	200
HL1S15	0.00	2.00	38.50	34.00	21.00	3.50	200
HWS8	0.00	16.00	41.00	32.00	11.00	0.00	200
HWS9	0.00	3.50	39.00	45.50	12.00	0.00	200
Total - all samples	0.30	7.91	35.38	41.51	13.14	1.76	4033

from each treated sample. Counts of pollen grains on each slide were conducted until at least 300 identifiable grains had been seen. Simple percentages were then calculated based on the total number of grains counted for each slide.

RESULTS

Mineralogical Analyses

The results of point-counting of sediment slides are given in Tables 1, 2, and 3. The most common mineral encountered in the heavy mineral suite was zircon, which comprised 36.3% of the total heavy minerals counted. Least common were leucoxene and rutile. The heavy-mineral grains were most commonly subrounded and, least commonly, very angular. All light-mineral fractions were dominated by quartz. Also common throughout the light-mineral populations were small, tabular, anhedral grains of muscovite, K-spar and plagioclase grains, which display varying degrees of weathering, and soft, white kaolin (?) clasts, which probably represent a cohesive weathering product of what were once feldspar grains. The most common light-mineral grain shape was subrounded; the least common shape was very angular.

Dendrological Analyses

Radiocarbon dates of 6870 +/- 50 years (USGS WW 2180) and 5050 +/- 50 years (USGS WW 2181) were obtained for Trees 4 and 7, respectively. Wood samples from the McClelland site were initially determined to be of the genus *Pinus* (William Stern, personal communication to Frankie Snow, 1998). Subsequently, samples from the trees were analyzed by Newsom, who verified the identifications as the genus *Pinus* sp. Individual species of pine

Table 3. Results of point counts for roundness. Percentages are normalized to 100% of total light-mineral grains counted.

Sample #	% Very angular	% Angular	% Subangular	% Subrounded	% Rounded	% Very rounded	Total counted
T1SA	0.00	6.17	45.67	38.33	9.33	0.50	600
T2SB	0.17	15.50	33.17	32.33	17.67	1.17	600
T3SC	0.00	2.00	37.33	55.17	5.50	0.00	600
T4S1	0.00	3.83	38.83	50.83	6.50	0.00	600
T4S2	0.00	1.50	38.33	52.17	8.00	0.00	600
T4S3	0.00	4.67	28.50	62.33	4.50	0.00	600
T5S4	0.00	4.33	40.00	51.17	4.50	0.00	600
T6S5	0.00	0.33	41.33	52.17	6.17	0.00	600
T7S6	0.00	5.00	36.67	52.50	5.50	0.33	600
T7S7	0.25	29.43	38.65	23.69	7.98	0.00	401
T7S10	0.00	1.83	36.17	54.00	8.00	0.00	600
T8S13	0.00	4.67	37.50	49.00	8.67	0.17	600
T9S11	0.00	3.50	42.17	50.33	4.00	0.00	600
T9S12	0.17	0.67	25.83	67.33	6.00	0.00	600
T11S16	0.00	4.00	36.17	53.00	6.67	0.17	600
T11S17	0.00	1.50	33.33	55.83	9.33	0.00	600
T12S18	0.00	7.00	27.67	54.17	11.17	0.00	600
T12S19	0.00	0.67	29.83	60.67	8.83	0.00	600
T13S20	0.00	2.50	39.33	47.17	10.17	0.83	600
T14S21	0.00	1.17	36.83	52.50	8.00	1.50	600
HL1S15	0.00	1.17	43.50	47.17	8.17	0.00	600
HWS8	0.00	4.00	42.00	48.33	5.67	0.00	600
HWS9	0.00	1.83	36.00	56.83	5.33	0.00	600
Total - all samples	0.02	4.30	36.70	51.14	7.63	0.21	13601

are not separable by wood anatomical characteristics; however, they can be classified according to specific anatomical groups within the genus. The McClelland trees belong to the *Taeda* wood anatomical group; this group includes all southern pines, such as longleaf (*P. palustris*), slash (*P. elliottii*), and loblolly (*P. taeda*). Growth rings in the wood samples are underdeveloped, which may indicate either that the preserved trees at the site are actually root structures, or that the trees were juveniles (i.e., less than 15-25 years old) when they died. The excavation activities at the McClelland site, as well as our limited sampling abilities, precluded more detailed analysis of the preserved trees, which might have clarified this uncertainty.

Palynological Analyses

The results of the pollen analysis are shown in Table 4 and Figure 5. All pollen samples were collected from the lower sand of this report. A total of 1647 grains was counted. Among these, 39 taxa were identified, including 36 extant taxa of pollen and spores, as well as one taxon of probable algal cyst (*Pseudoschizaea*) and two extinct or extirpated genera (*Engelhardia-Momipites* and *Pterocarya*). These latter genera are rare within the sample population, leading us to conclude that these grains were reworked from older, possibly Miocene, strata. Of the 39 taxa present, only 10 were present in any sample in amounts greater than or approximately equal to 1%. The maximum values for these taxa are as follows: *Pinus*, 31.2% (derived chiefly from Trees 8 and 9); Poaceae, 21.7%; Asteroideae, 12.8%; *Ilex*, 5.8%; *Corylus*, 5.0%; *Quercus*, 2.5%; *Pseudoschizaea*, 2.2%; *Myrica*, 1.7%; TCT (Taxodiaceae-Cupressaceae-Taxaceae), 0.97%; and *Engelhardia-Momipites*, 0.97%.

The results of the pollen analyses are pre-

Table 4. Relative abundance of palynomorph taxa from seven trees samples in the McClelland sand pit. Totals are normalized to 100% of 1647 grains. * algal cyst; **Extinct/extirpated.

Taxon	Relative abundance	Taxon	Relative abundance
Alnus	0.48	*Ostrya-Carpinus*	0.12
Ambrosia	0.78	*Pinus*	31.20
Asteroideae	12.80	*Platanus*	0.06
Betula	0.30	Poaceae	21.70
Carya	0.48	*Polygonum*	0.06
Castanea	0.60	*Pteridium*	0.06
Corylus	5.00	*Pseudoschizaea**	2.20
Cyperaceae	0.78	*Quercus*	2.50
Cyrillaceae	0.12	*Sagitarria*	0.06
Ericaceae	0.55	*Salix*	0.12
Ilex	5.80	*Sphagnum*	0.60
Juglans	0.06	TCT	0.97
Labiatae	0.06	*Typha*	0.06
Liquidambar	0.12	*Ulmus*	0.06
Magnolia	0.12	Umbelliferae	0.30
Myrica	1.70	*Viburnum*	0.12
Nymphoides	0.24	*Woodwardia*	0.55
Nyssa	0.12	*Urtica*	0.06
Osmunda	0.60	*Pterocarya***	0.06
		*Engelhardia-Momipites***	0.97
		unknowns	7.10

sented here by two means; combined sample data represent the relative abundances as calculated from the total of 1647 grains (Table 4 and Figure 5), whereas upper and lower tree data show relative abundances of samples from the two forest horizons (Figure 5 only). The combined analyses listed in Table 4 provide a general characterization of the flora present at the McClelland site over the period of approximately 1800 years represented by the buried forests. Combining the analyses allows the results presented here to be compared to results published elsewhere (e.g., Rich and Pirkle, 1994). Those data were derived from samples that were collected over intervals of a foot or more (at least 18 cm) using jet drills and augers. Those samples were, thus, homogenized over that range of depths. The combined analyses we show here emulate the results derived from such mixed sample depths. The combined analyses have the obvious disadvantage of concealing any differences that may exist between the pollen/spore floras of what we interpret to be two forests that are of very different ages. Trees 2, 3, 4, and 5 were all buried in the lowest level of sand that covers the bottom of the pit, and these trees occupy a single horizon that is approximately 6870 years old (i.e., the age of Tree 4). These samples are designated "lower trees" in Figure 5, which is a graphical representation of the pollen counts performed for the McClelland site. Trees 1, 8, and 9 were all enclosed in a more friable white sand which clearly overlies the humate-cemented sands that occupy the very bottom of the pit, and stand at the same horizon as Tree 7 (dated at approximately 5050 years BP). These latter samples represent the vegetation that grew at the site at that time, and are identified in Figure 5 as the "upper trees."

INTERPRETATION

The coarse, angular nature of the siliciclastic sediments surrounding the McClelland trees (i.e., lower sand) precludes their having been deposited by wind; the particles are far too coarse to have been transported by aeolian saltation or suspension. The lower sand also lacks

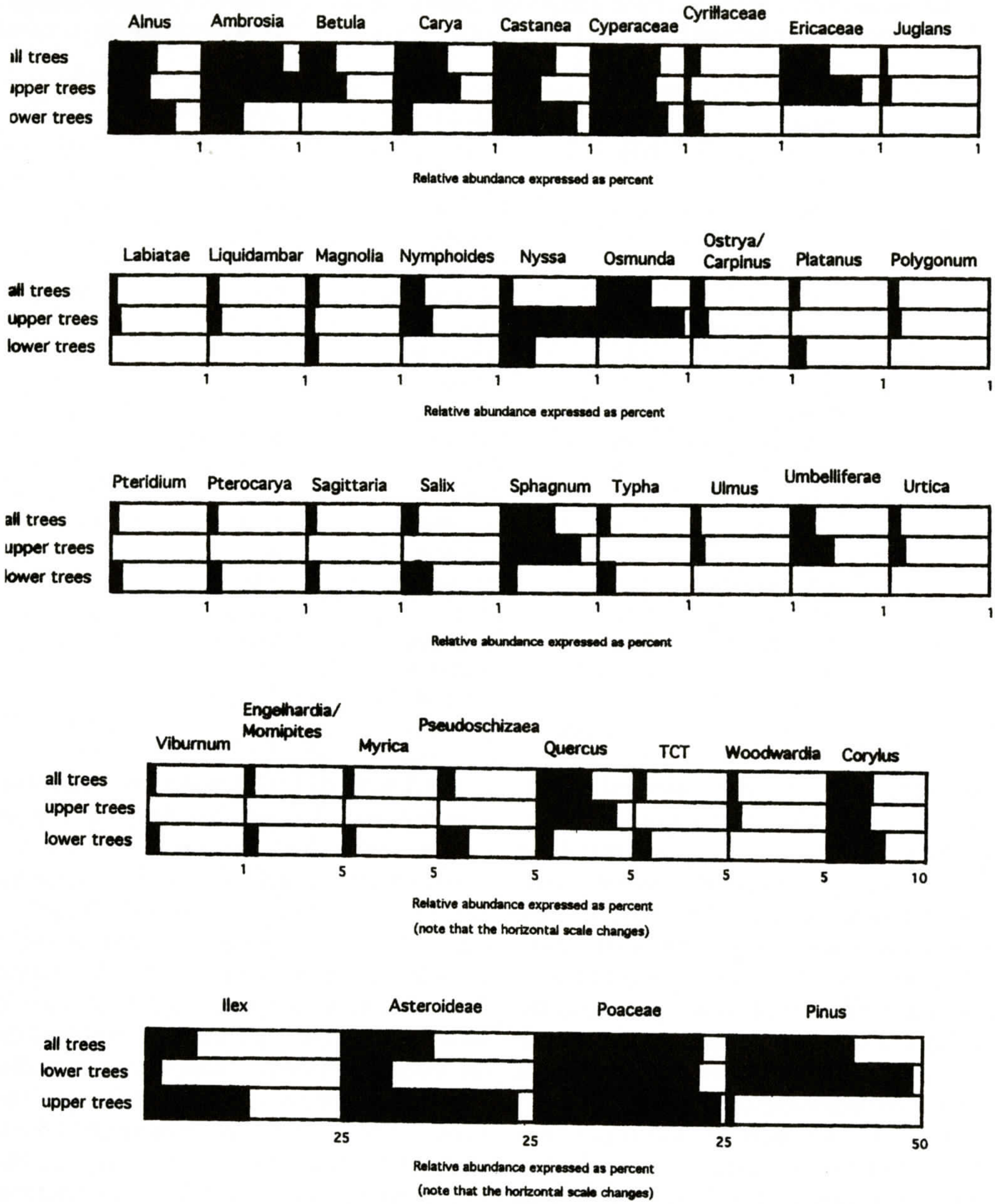

Figure 5. Relative abundances of pollen. Upper bars show abundances as calculated in all tree samples. Middle bars show abundances for only upper trees. Lower bars show abundances for only lower trees.

evidence for either a marine-transgressive or a lacustrine depositional environment. This leaves us to consider some form of fluvial depositional environment to account for the presence of the lower portion of the McClelland sandpit deposit. In light of the field and sedimentological evidence gathered from the site, we propose that a fluvial system, such as a nearby creek, was responsible for the burial of the McClelland trees. Further, we believe that this stream system experienced either a channel avulsion or a levee breach, with an accompanying crevasse splay. Crevasse splay deposits consist of sand to fine gravel and spread laterally in thin sheets for, sometimes, a considerable distance (D. G. Smith, 1983). Once a levee has been breached, splays commonly accrete vertically and prograde laterally with each successive flooding or discharge event. Ordinarily dry environments thus become inundated with water and clastic material. A crevasse-splay could have been responsible for the burial of the McClelland trees in such coarse sediment, and over a sufficiently long time that the trees were not killed instantly. Crevasse splays can also be created by channel avulsion (N. D. Smith and others, 1989). Such an event would lead to more rapid and more voluminous deposition of coarse material, since the flow of the channel itself (and, consequently, of its coarse bedload) would be directed along the path of the avulsion.

In the context of the burial of the McClelland trees, an avulsed channel would have delivered the requisite coarser sediments and deposited them in a planar fashion. The avulsed-channel hypothesis has the added advantage of providing for burial of the trees more rapidly than would have occurred during an intermittent event, such as a splay formed merely by a levee breach during high water. We believe that the trees in the sandpit represent two generations of forest that were buried over time by an avulsed stream. The finer-grained deposits at the north end of the sandpit may have been at the distal end of the area of deposition. Alternatively, the north area of the sandpit might have been the floodplain beyond the influence of the prograding sediments--either of these scenarios would account for the finer grain size surrounding those trees.

The upper sand body was probably emplaced near the end of the Hypsithermal, as outlined in Markewich and Markewich (1994). As explained earlier in the paper, dune activity was common across the Southeast during this period, and the fine, structureless sands overlying the McClelland trees probably represent active dune formation.

The palynological composition of the McClelland-site sediment samples and the identity of the wood samples indicate that the plant community growing at the site during the time of sediment deposition must have been a pine savannah or, perhaps, a prairie. The savannah, or prairie interpretation is drawn from the unusually large percentages of grass and insect-pollinated composite pollen (Poaceae, with a maximum of 21.7%, and Asteroideae, with a maximum of 12.8%, respectively). Baker and Waln (1985) point out that palynologists since the time of Kapp (1970) have interpreted anomalously large amounts of grass and composite pollen to represent the presence of ancient prairies. They mention sites such as the Jinglebob vertebrate locality in southwestern Kansas, where combined grass and composite values range from 24-80% of the samples, and where vertebrate remains argue convincingly for a prairie ecosystem. In later work, Baker et al. (1989) observe that there are not many good modern pollen assemblage analogues for prairie sites because of the means by which samples have been collected. Still, abundances of grasses in the range of about 18-40%, and total composites in the range of 40-80% [our interpretation of their data and Figure 5.1 of Baker and Van Zant, 1980)] lead them to suggest the presence of ancient prairies at Lake Okoboji West, Iowa, between 5200 and 7700 years ago. In an analysis of Billy's Lake, Minnesota, Jacobson and Grimm (1986) interpret the presence of prairie vegetation between 8020 and 3440 years ago when sediments accumulated between 5 and 30% grasses, and 15 to about 40% total composites. They also mention the fact that shrub and hardwood pollen are uncommon, and there is minor pine (15% or less) and very little oak (10% or less). This latter fact is particularly important in our discussion of the McClelland site because pine and oak trees and, hence, their pollen, are currently found in abundance nearly everywhere in the southeastern states. Cohen (1975), for example, analyzed the palynological contents of peats from seven plant communities in the Okefenokee Swamp in order to "fingerprint" those surface environments palynologically. Among his average percentages for *Pinus* he found a range from 12.4 to 36.0% of the pollen totals, and for *Quercus* he identified a range of 4.2 to 10.3%. Our pine values are comparable to Cohen's, but the average abundance of oak is much less.

Differences in analytical techniques, particularly the method of determining "pollen sum," and the extent to which composites are differentiated can make direct comparisons of palynological analyses difficult to perform. That situation notwithstanding, we believe that, based on the palynological composition of the samples, the McClelland site and its environs may have resembled the modern Platte and Saskatchewan river basins, which feature coarse sandy-bedload channels surrounded by prairie vegetation (Brierley, 1996). Some shrubs, such as holly (*Ilex*) and hazelnut (*Corylus*), probably grew among the pines beyond and atop the banks of the McClelland-site stream, with the relative abundance of each changing over time. The area generally seems, however, to have been populated by herbaceous plants such as those that might today be found in the Midwest. Our interpretation of a Hypsithermal prairie flora for the Southeast is not new, but instead adds the McClelland site to the considerable list of sites that offer very good evidence for the presence of ancient prairies in the southeastern U.S. Watts (1969) was perhaps the first to recognize prairie elements in pollen profiles from the Southeast. He identified a distinct horizon of sediments within Mud Lake, in Marion County, Florida, that correlates generally with the Hypsithermal. His pollen zone M-2, also known as the *Quercus* zone, is dated from 8160 to 5070 years BP; this zone contains pollen that may have been produced by a community of dry oak forest with, perhaps, interspersed patches of prairie-like herbaceous vegetation and sage (*Artemesia*). In later work, Watts (1971) found zone M-2 strata within the deposits of Lake Louise, in Lowndes County, Georgia; these strata accumulated between 8510 +/- 100 years and 6710 +/- 140 years BP. Watts describes the pollen/spore flora as representing sclerophyllous oak-hickory vegetation, with an extensive representation of herbs such as those now found in prairies of the upper Midwest (i.e., grasses, sedges, ragweed and other composites, chenopods and amaranths, and pinweed). In their summary of a number of pollen studies from the Southeast, Delcourt and Delcourt (1985) state that the climate of the Hypsithermal, being both warmer and drier than modern conditions, allowed for the eastward expansion of prairie vegetation as far as southern Georgia and northern Florida. Watts and Hansen (1988) corroborate this interpretation in their summary of data from Lake Louise, Georgia, as well as Lakes Sheelar, Annie, and Tulane, which are found at various locations in Florida. Pollen zone M-2 is also present at these sites, demonstrating that the Hypsithermal dry oak/prairie flora is widely recognizable in the Southeast.

CONCLUSIONS

The depositional history recorded at the McClelland sandpit represents the mid-Holocene dry period known as the Hypsithermal. The flora of the site consisted of pine savannah or prairie with few oaks or other hardwoods, and groves of pine. The trees at the site were inundated with coarse, angular quartz sand beginning around 7000 years ago; this may have been the result of a local stream that avulsed its channel, spilling its coarse bedload across the landscape. The same thing may have been responsible for the burial of a second generation of trees at the McClelland site. The site was finally covered by a structureless aeolian dune deposit, the deposition of which is consistent with dune activity across the Southeast during the Hypsithermal. The entire site, as found today, was probably an active site of deposition no later that 2000-3000 BP, as suggested by buried flint chips that were found at the site about 1 meter below ground surface.

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A Study of Plagioclase-hosted Melt Inclusions in the Ordovician Deicke and Millbrig Potassium Bentonites, Southern Appalachian Basin

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ABSTRACT

Rhyolitic melt inclusions are found within plagioclase (An_{53-77}), K-feldspar, and quartz pyroclasts in the stratigraphic equivalents of the Ordovician Deicke and Millbrig potassium bentonites (K-bentonites) in the Southern Appalachian Basin. Rhyolitic melt inclusions are particularly abundant (>100 inclusions per thin section) in pyroclastic plagioclase within the Deicke K-bentonite at the Hinds Creek Quarry, Tennessee. The abundant host plagioclase grains are interpreted to be pyroclasts based on the presence of polysynthetic twinning, major element compositions, and blue-green cathodoluminescence. The mean concentrations of FeO (1.01 wt.%.), MgO (0.17 wt.%), and TiO_2 (0.21 wt.%) in the Deicke K-bentonite melt inclusions are higher than the mean concentrations of FeO (0.68 wt.%), MgO (0.05 wt.%), and TiO_2 (0.08 wt.%) in the Millbrig rhyolitic melt inclusions. The differences in the concentrations of FeO and TiO_2 between the Deicke and Millbrig melt inclusions are statistically significant at ≤1% level. Based on data gathered in this study, the Deicke and Millbrig-equivalent K-bentonites in the Southern Appalachian Basin can be distinguished from each other and correlated based on the chemical compositions of melt inclusions. The Ordovician K-bentonite melt inclusions also provide an estimate of the composition of the original magma now altered to illite/smectite.

INTRODUCTION

Over 20 distinct beds of altered tephra known as K-bentonites have been identified in Ordovician rocks in the Southern Appalachian Basin (e.g. Kolata and others, 1996). Early studies noted that K-bentonites were formed by the devitrification of volcanic ash and tephra, and each K-bentonite bed represents a geologically instantaneous event - a major explosive volcanic eruption (e.g., Nelson, 1921, 1922, 1926). The relatively wide areal extent makes K-bentonites extremely useful as isochrons for biogeographic, paleogeographic, paleoecologic, sequence stratigraphic, and sedimentological investigations. Recent studies have focused on the use of trace elements for stratigraphic correlation, petrology, and paleovolcanology, and on understanding the chemical character of fluid flow due to the Alleghanian Orogeny (e.g.

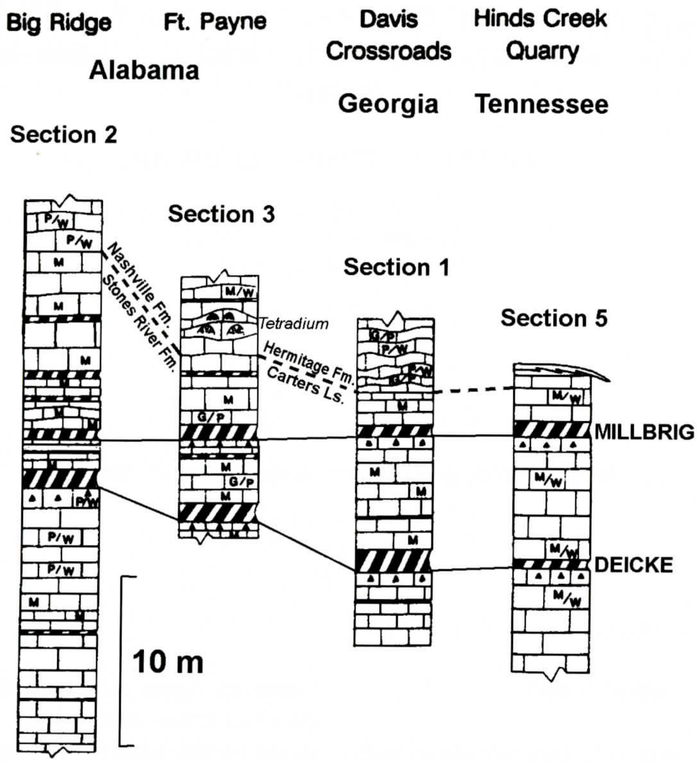

Figure 1: Stratigraphic columns of representative sections containing Deicke and Millbrig equivalent K-bentonites modified from Haynes (1994). M denotes lime mudstone. W denotes wackestone. P is packstone and G is grainstone. Samples studied petrographically in this study were collected from the basal 6 –10 inches that contain the highest amount of pyroclasts. Solid triangles denote bedded chert. The section number corresponds to the locations shown in Figure 2.

Haynes, 1992, 1994; Elliott and Aronson, 1987, 1993; Bergström and others, 1995; Haynes and others, 1995; Huff and others, 1992, 1996; Kolata and others, 1996). The oldest Ordovician K-bentonites in this region were found in lower Middle Ordovician (Chazyan) rocks that overlie the Knox Unconformity. The greatest numbers of beds were found in the Middle Ordovician Carters Limestone and Hermitage Formation. The two thickest and most prominent Ordovician K-bentonites are found near the Carters-Hermitage contact (Figure 1). These two K-bentonites are stratigraphic equivalents to the Deicke and Millbrig K-bentonites which have been traced eastward from the upper Mississippi Valley to the southern Appalachians, and northward to Canada (Huff and Kolata, 1990; Haynes, 1994; Kolata and others, 1987, 1996).

The physical aspects of the origin and tectonomagmatic setting of these eruptions have been discussed by Huff and others, 1996, and the geochemistry of these K-bentonites have been discussed in recent papers (Huff and Türkmenoglu, 1981; Huff, 1983; Kolata and others, 1987, 1996; Samson and others, 1988, 1989; Haynes, 1992, 1994; Huff and others, 1992; Delano and others, 1994; Bergstrom and others, 1995; Haynes and others, 1995, 1996; Haynes and Melson, 1997, Adhya and others, 1999). Alteration of groundmass and phenocrysts makes reconstruction of Paleozoic magmatic conditions difficult. The bulk of the tephra and vitric ash was altered subsequently to clay in open-system geochemical environments and later altered to illite-smectite by potassic diagenetic processes (Huff and Türkmenoglu, 1981; Elliott and Aronson, 1987, 1993; Hay and others, 1988; Haynes, 1994). The unaltered melt inclusions and phenocrysts in K-bentonites are used to determine the original composition of the parent magma, to estimate of the amount of SO_2 erupted, and to estimate the composition of the parent magma at the time of eruption (e.g. Delano and others, 1994; Kolata and others, 1996). Melt inclusions in quartz have been used also to correlate Middle - Late Ordovician K-bentonites within the Utica Shale in the Northern Appalachian Basin (Delano and others, 1994). The Millbrig-equivalent K-bentonite has been identified on the basis of the chemical compositions of melt inclusions and apatite phenocrysts in central Pennsylvania and Kentucky (Adhya and others, 1999). The purpose of this study is to use the chemical compositions and petrography of the melt inclusions to distinguish and correlate the Deicke and Millbrig equivalent K-bentonites in the Southern Appalachian Basin.

METHODS

Samples Studied and Thin Section Preparation

Ordovician K-bentonites identified as equivalent to the well-known Deicke and the Millbrig K-bentonites were previously collected from various localities within the Southern Appalachian Basin (e.g. Haynes, 1994; Elliott and Aronson, 1993). Deicke K-bentonites at Hinds Creek Quarry, Tennessee and Harper Mountain, Kentucky, and Millbrig K-bentonites at Davis Crossroads, Georgia, Big Ridge, Alabama, Ft. Payne, Alabama, and Overton County, Kentucky were selected for petrographic study (Figure 2). With the exception of Ft. Payne, two thin sections were studied from each K-bentonite. Locality information is provided by Verhoeckx-Briggs (1997).

Bulk K-bentonite samples were impregnated with standard thin section epoxy in an argon atmosphere under approximately 10 bars pressure for 24 hours. Once the epoxy hardened, the sample was trimmed, ground and polished as per standard thin sectioning procedures. Design and construction of special apparatus for pressure impregnation of K-bentonites as well as subsequent cutting, grinding and polishing of thin sections for this study were carried out in the Department of Mineral Sciences at the Smithsonian Institution (Haynes and Gooding, unpublished data).

Petrographic and electron microprobe study

Thin sections were examined using conventional polarized light, reflected light, and cathodoluminescence (CL) microscopy to identify melt inclusions, pryoclastic minerals, and matrix minerals. The percentages of minerals reported are based on 540 counts per thin section. The slides were not stained so that we could use the thin sections for electron microprobe analyses. We used an ELM-3.2 Series 033 cold cathode luminoscope with a beam setting of 17 kV and 0.5 mA for the CL analyses. Pyroclastic plagioclase showed very strong blue-green luminescence while diagenetic albite exhibited weak luminescence because it has a highly ordered crystal lattice and contains low concentrations of trace elements (Kastner and Siever, 1979; Marshall, 1988;). CL analyses were used to determine more accurately the distribution of calcite, quartz, authigenic albite, and pyroclastic plagioclase in the thin sections.

The concentrations of major elements were measured in minerals and melt inclusions using the JEOL JXA-8600 Superprobe electron microprobe at the University of Georgia. In plane light, many melt inclusions were seen beneath the surface of the thin section. However, we analyzed inclusions that were exposed at the surface of the thin section, and this requirement constrained the overall number of inclusions that could be studied. The electron microprobe was operated at 15 kV accelerating voltage and 5 nA beam current. The beam width was defocused to either 5 or 10 μm depending on the size of the melt inclusion being analyzed. Counting times were from 20 to 30 seconds. A select set of well-characterized minerals, natural glasses and synthetic glasses were analyzed to determine optimum count times, beam current settings, and calibration for all major elements. The data were reduced on-line using the Noran

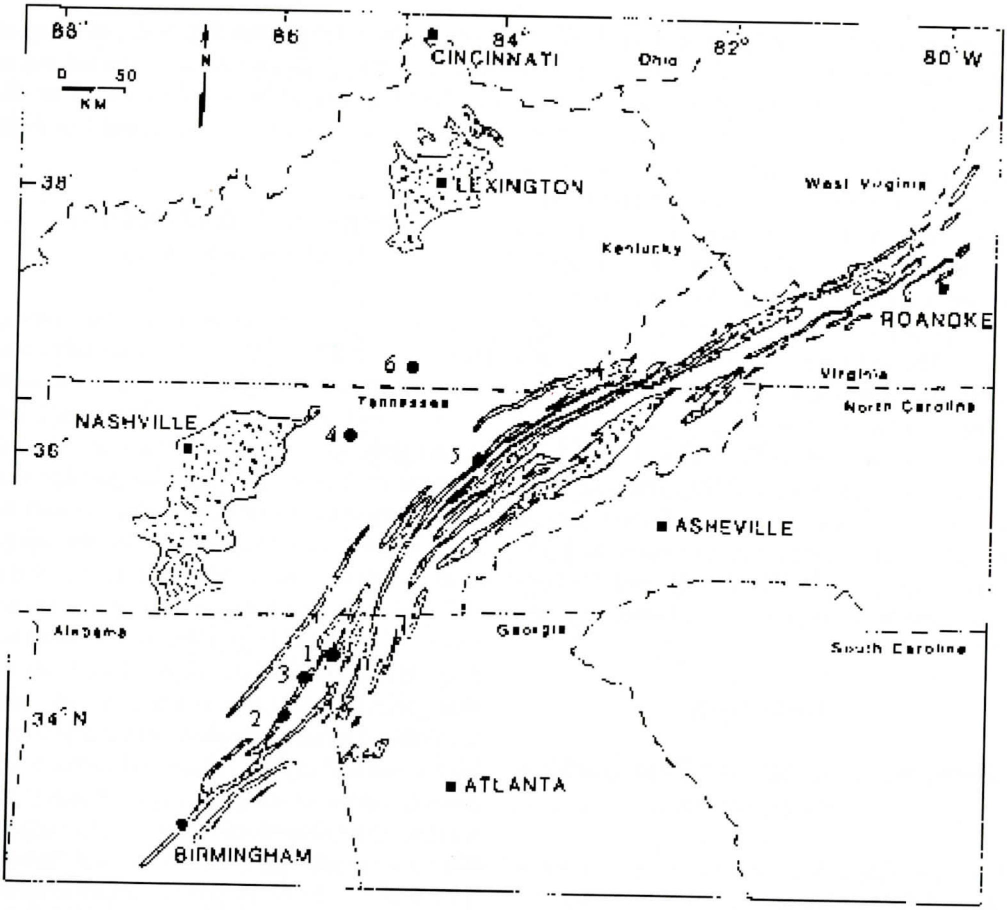

Figure 2: Map showing locations of K-bentonites (closed circles) analyzed in this study modified from Haynes (1994): 1, Davis Crossroads, GA (Haynes, 1994, locality 22); 2, Gadsden, AL; 3 (Haynes, 1994, locality 25), Ft. Payne, AL (Haynes, 1994. locality 24); 4, Overton County, TN (Elliott and Aronson, 1993); 5, Hinds Creek Quarry, Anderson County, TN (Haynes, 1994, locality 18); 6, Albany County, KY (Elliott and Aronson, 1993).

TASK Program and $\Phi(\rho Z)$ corrections (Bastin et al., 1984; Reed, 1993). A rhyolite glass standard USNM 72854 VG568 was analyzed repeatedly as a reference standard to monitor analytical precision (Jarosewich and others, 1980). We obtained very good agreement between our measured analyses of this standard glass (N = 6) and the accepted values. The total for major element oxides was between 98.18 and 100.81 wt.%. The difference from the accepted values was < 1 rel.% for the major element oxides Al_2O_3 and SiO_2, and < 6 rel.% for the minor element oxides whose concentrations are between 1 and 5 wt.%. In particular, the mean Na_2O concentration determined was within 2.7 rel.% of the nominal value, showing that alkali mobility was not a problem under our operating conditions.

Student t-test of the difference of means and F-tests of variance were performed to determine whether elemental compositions were significantly different between the Deicke and Millbrig melt inclusions (Davis, 1973). The Students T-statistic t was calculated using equations 3.23 and 3.24 in Davis (1973) as shown below, where X is the mean of a population, n is the number of measurements for a given population, s^2 is the variance of a population which is the square of the standard deviation:

$$t = \frac{(X_1 - X_2)}{sp\sqrt{\left[\frac{1}{n_1} + \frac{1}{n_2}\right]}}$$

$$sp = \frac{\sqrt{[(n_1 - 1)s_1^2 + (n_2 - 1)s_1^2]}}{(n_1 + n_2 - 2)}$$

The F statistic is given as $F = \frac{s_1^2}{s_2^2}$.

RESULTS

As shown from our data (Table 1) and previously by Haynes (1994), the Deicke and Millbrig K-bentonites have distinct mineral compositions in terms of their pyroclastic, authigenic, and cementing minerals. The Deicke K-bentonite has more authigenic albite and calcite cement, and lower amounts of quartz and biotite relative to the Millbrig K-bentonite. Given the similarity of the mineral percentages measured in this study relative to the mineral percentages measured by Haynes (1994), we are confident that we examined representative samples of the Deicke and Millbrig K-bentonites. However, the amounts of calcite cement measured in the Deicke and Millbrig K-bentonites in this study are much higher compared to the amounts measured by Haynes (1994). Calcite cement is highly visible from CL analyses and it may have been underestimated by examination in plane light microscopy.

The plagioclase grains hosting the melt inclusions are magmatic in origin based on the presence of polysynthetic twinning, and major element compositions. The magmatic plagioclase luminesces a vivid bright blue-green, consistent with a magmatic origin (Kastner and Siever, 1979; Marshall, 1988). Most K-bentonites collected in this study contain a mixture of luminescent and non-luminescent plagioclase. In particular, the Millbrig K-bentonite at the Davis Crossroad section contains both luminescent calcic plagioclase grains, and luminescent calcic plagioclase with authigenic non-luminescent feldspar rims. The magmatic plagioclase hosts range from An_{53} to An_{77}, with oxide totals between 99.30 and 100.85 wt.% (Verhoeckx-Briggs, 1997).

Melt inclusions occur in pyroclastic quartz, K-feldspar, and plagioclase in the K-bentonites examined in this study. The inclusions are isotropic and gray – white in plain light, and most do not contain gas bubbles or fluid phases (Figure 3). Inclusions are generally rectangular, 5-50 µm long, with rounded corners. In the two Deicke K-bentonites, the melt inclusions are mostly in plagioclase. In the four Millbrig K-bentonites, the melt inclusions are mostly in quartz, K-feldspar, and less abundantly in plagioclase. The Deicke K-bentonite from the Hinds Creek Quarry contains an unusually large number of melt inclusions in plagioclase (> 100 per thin section), and they are present in clusters with the long axes of the inclusions aligned in the same direction. In all other thin sections, the melt inclusions are much less abundant.

We analyzed melt inclusions from one Deicke K-bentonite (Hinds Creek Quarry) and three Millbrig K-bentonites (Overton, KY, Big Ridge, AL, and Davis Crossroads, GA). Data

Table 1: Mineral Abundances of the Deicke and Millbrig K-bentonites. Cc - calcite, Pl - plagioclase, Ab - Albite, Kf - K-feldspar, Qz - quartz, Bi - biotite, Op - opaques, Cly - clay, Acc - accessories.

Mineral	Cc	Pl	Ab	Kf	Qz	Bi	Op	Cly	Acc
Deicke (this study)	16	34	3	8	< 1	1	2	33	4
Deicke non-clay minerals (this study)	24	50	5	12	< 1	1	3	-	5
Deicke (Haynes, 1994)	1-5	1-15	0-20	5-80	<1	<1	< 1-5	-	0-10
Millbrig (this study)	13	19	2	3	6	13	2	40	2
Millbrig non-clay minerals	22	32	3	5	10	22	3	-	3
Millbrig (Haynes, 1994)	< 1	1-15	0-20	5-60	20-45	15-50	<1	-	< 1

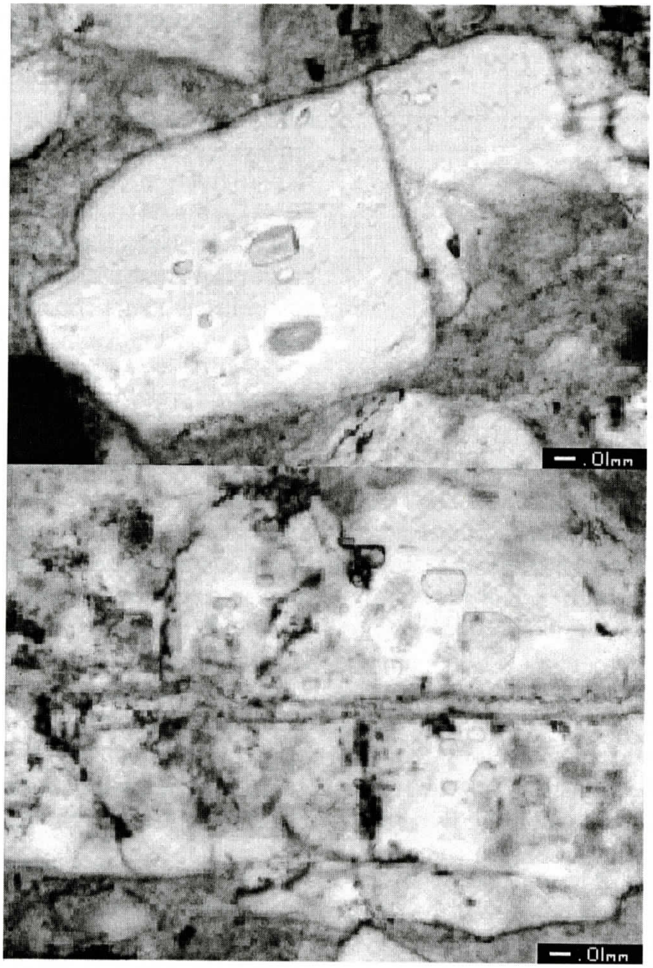

Figure 3: Rhyolitic melt inclusions in plagioclase from the Hinds Creek Quarry. The plagioclase pyroclasts are surrounded in a matrix of illitic clay and calcite. Plagioclase is heavily weathered in the lower photomicrograph.

from the Deicke K-bentonite at Hinds Creek Quarry are listed in Table 2, along with analyses of melt inclusions from the Ordovician Otsquago K-bentonite (Delano and others, 1994), plagioclase-hosted rhyolite melt inclusions from Pinatubo dacite (Rutherford and Devine, 1996), and USNM rhyolite glass (Jarsosewich and others, 1980). Analyses of the melt inclusions in quartz, K-feldspar, and plagioclase are given for three Millbrig K-bentonites in Table 3. The analytical totals for melt inclusions from the Deicke and Millbrig K-bentonites range from 91 wt.% to 97 wt.%, which are low but not unusual compared to previous studies of rhyolite melt inclusions (Delano and others, 1994; Hanson and others, 1996; Kolata and others, 1996). Rutherford and Devine (1996) also note a low total (93.64 wt.%) for plagioclase-hosted rhyolite melt inclusions in Mt. Pinatubo dacite (Table 2). The chief explanation for such low totals is the presence of dissolved volatiles that are not determined in the electron microprobe analyses.

Na_2O concentrations for the Deicke and Millbrig melt inclusions are fairly low, while at the same time K_2O values are generally high (Table 2). Normative calculations (Table 4) indicate 3 – 4% corundum for the melts, and Mertzbacher and Eggler (1984) suggest that normative corundum greater than about 0.9 wt.% results from volatile loss during electron

Table 2: Electron Microprobe Analyses of Plagioclase hosted Rhyolite Glass Inclusions from the Deicke K-bentonite at Hinds Creek Quarry, Otsquago K-bentonite (Utica Shale), Pinatubo Dacite and of the USNM rhyolite glass

Hinds Creek Quarry	SiO_2	Al_2O_3	TiO_2	Cr_2O_3	FeO	MgO	MnO	CaO	Na_2O	K_2O	F	Cl	Total
	73.22	12.35	0.19	0.00	0.89	0.18	0.09	0.82	1.59	3.76	0.06	0.07	93.22
	73.75	11.90	0.20	0.00	0.78	0.18	0.00	0.68	0.81	2.93	0.10	0.03	91.36
	73.83	12.08	0.21	0.00	1.09	0.17	0.00	1.20	3.35	0.80	0.04	0.07	92.84
	75.42	12.18	0.23	0.00	1.13	0.21	0.00	1.27	2.48	2.95	0.16	0.08	96.11
	75.74	11.79	0.16	0.01	0.98	0.19	0.06	1.07	2.31	4.59	0.03	0.07	97.00
	74.86	11.71	0.17	0.00	0.96	0.16	0.03	1.10	1.96	4.43	0.06	0.04	95.48
	74.55	11.62	0.28	0.04	1.09	0.20	0.04	1.20	2.38	4.49	0.14	0.06	96.09
	74.27	11.72	0.20	0.04	0.96	0.19	0.05	1.20	2.40	4.43	0.07	0.04	95.57
	74.77	11.85	0.24	0.00	0.99	0.16	0.05	1.19	1.66	3.98	0.14	0.09	95.12
	74.62	11.82	0.16	0.00	1.02	0.17	0.00	1.20	1.53	3.92	0.04	0.07	94.55
	75.15	12.12	0.26	0.04	0.95	0.16	0.00	1.11	2.39	4.44	0.08	0.06	96.76
	75.58	12.38	0.25	0.00	0.91	0.18	0.04	1.06	1.45	0.60	0.16	0.06	92.67
	75.82	12.13	0.28	0.00	1.03	0.16	0.03	1.23	1.50	3.68	0.03	0.07	95.96
	75.02	11.53	0.20	0.13	1.05	0.17	0.01	1.06	1.72	4.04	0.08	0.09	95.10
	73.48	12.80	0.23	0.00	1.25	0.12	0.11	1.57	1.42	1.61	0.14	0.04	92.77
	75.69	11.74	0.19	0.08	1.09	0.20	0.00	1.15	1.59	4.14	0.12	0.03	96.02
	73.67	12.12	0.20	nd	1.14	0.15	0.00	1.18	1.58	3.81	0.06	0.15	94.06
	74.96	12.38	0.23	nd	0.41	0.16	0.00	0.76	0.76	2.36	0.02	0.00	92.04
	74.20	11.98	0.20	nd	1.15	0.14	0.11	1.25	0.91	1.28	0.06	0.05	91.33
	73.81	12.20	0.25	nd	0.97	0.17	0.11	1.35	1.64	1.14	0.07	0.01	91.72
	75.45	11.91	0.17	nd	1.25	0.17	0.00	1.07	1.09	3.36	0.05	0.11	94.63
	74.04	12.68	0.19	nd	1.05	0.17	0.06	1.11	1.30	1.32	0.09	0.00	92.01
	76.12	11.96	0.12	nd	1.23	0.18	0.00	1.03	0.59	3.03	0.05	0.02	94.33
	74.85	11.81	0.20	nd	0.91	0.10	0.03	1.13	1.53	0.65	0.05	0.21	91.47
	74.90	11.66	0.24	nd	0.91	0.16	0.07	1.03	1.55	4.49	0.06	0.18	95.25
Mean	74.71	12.02	0.21	0.02	1.01	0.17	0.04	1.12	1.66	3.05	0.08	0.07	94.14
S.D.	2.32	1.21	0.05	0.04	0.22	0.03	0.04	0.65	0.61	1.39	0.04	0.06	2.98
Otsquago K-bentonite Mean (n=11)	73.44	11.49	.21	nd	1.82	.243	.065	1.81	3.26	2.24	nd	.256	94.83
Pinatubo Dacite	72.46	12.19	.14	nd	.78	.21	.06	1.09	3.87	2.84	nd	.128	93.64
USNH Mean (n = 6)	77.15	12.14	0.1	nd	1.16	0.03	0.01	0.43	3.65	4.9	0.04	0.1	99.71
S.D.	0.97	0.13	0.02	nd	0.09	0.02	0.01	0.04	0.16	0.07	0.05	0.08	
Accepted	76.71	12.06	0.12	nd	1.23	<0.1	0.03	0.5	3.75	4.89			99.52
Relative %	0.57	0.66	16.67		5.69		66	14	2.67	0.20			.19

Notes: Otsquago K-bentonite mean is from Hanson and others (1996); Pinatubo Dacite analysis is from Rutherford and Devine (1996); remaining data are from this study.

bombardment. However, our analyses of the reference standard show that alkali mobility is not an issue for this set of analyses.

The SiO_2 in the melt inclusions ranges from 71 to 76 wt.%, and indicates a rhyolitic composition (Tables 2 and 3). The average SiO_2 concentration (74 wt.%) of the melt inclusions from the Deicke K-bentonite at Hinds Creek Quarry is very similar to that (73 wt.%) of the melt inclusions from the K-feldspar, plagioclase, and quartz in three Millbrig K-bentonites (Table 3). As shown in Tables 2 and 3, the average silica concentrations of Deicke and Millbrig melt inclusions are also comparable to the average sil-

Table 3: Major element analyses of melt inclusions from the Millbrig K-bentonites.

Locality	Host	SiO_2	Al_2O_3	TiO_2	FeO	MgO	MnO	CaO	Na_2O	K_2O	F	Total
Big Ridge	Plagioclase	71.05	12.05	0.02	0.60	0.03	0.03	0.77	2.49	4.70	0.00	91.74
Big Ridge	K-feldspar	72.35	11.61	0.06	0.53	0.05	0.03	0.73	1.78	3.23	0.12	90.49
Big Ridge	Quartz	73.15	12.10	0.13	1.02	0.09	0.03	0.99	2.07	3.77	0.03	93.38
Big Ridge	Quartz	73.45	12.77	0.11	0.52	0.00	0.06	0.96	1.94	3.88	0.13	93.82
Big Ridge	K-feldspar	74.51	12.09	0.03	0.50	0.06	0.00	0.76	1.39	4.51	0.00	93.85
Davis Crossroads	Quartz	74.25	11.80	0.09	0.46	0.03	0.14	0.66	0.95	4.52	0.00	92.90
Overton, KY	Quartz	71.50	11.86	0.10	1.00	0.05	0.00	0.91	1.72	4.18	0.04	91.36
Overton, KY	Quartz	73.06	11.76	0.08	0.82	0.08	0.02	0.82	2.00	4.70	0.02	93.36
Average (n=8)		72.92	12.01	0.08	0.68	0.05	0.04	0.83	1.79	4.19	0.04	92.61
S.D.		1.22	0.36	0.06	0.23	0.03	0.05	0.12	0.46	0.52	0.05	1.26

Table 4: CIPW Norms for melt inclusions.

	Deicke (mean)	Millbrig (mean)	Otsquago K-bentonite	USNH Rhyolite Glass
SiO_2	74.71	72.92	73.22	77.15
TiO_2	0.21	0.08	0.21	0.1
Al_2O_3	12.02	12.01	11.49	12.14
FeO	1.01	0.68	1.82	1.16
MgO	0.17	0.05	0.243	0.03
MnO	0.04	0.04	0.065	0.01
CaO	1.12	0.68	1.81	0.43
Na_2O	1.66	1.79	3.26	3.65
K_2O	3.05	4.19	2.24	4.90
Totals	94.00	92.60	94.6	99.58
Quartz	50.13	44.16	40.67	35.44
Corundum	3.93	3.02	0.41	0.05
Orthoclase	18.03	24.79	13.24	28.69
Albite	14.05	15.15	27.59	30.89
Anorthite	5.56	4.12	8.98	2.13
Hypersthene	1.73	1.13	3.22	1.74
Magnetite	0.16	0.11	0.29	0.19
Ilmenite	0.40	0.15	0.40	0.19
Totals	94.00	92.60	94.60	99.58

ica concentrations of rhyolite melt inclusions in quartz from the Middle Ordovician Otsquago Creek K-bentonite in the Utica Shale (Hanson and others, 1996) and in plagioclase from a dacite flow erupted at Mount Pinatubo (Rutherford and Devine, 1996).

The average concentrations for FeO and TiO_2 (Table 2, 3, and Figure 4) are significantly different between the Deicke and Millbrig melt inclusions based on the Student t-test of difference of means and F-tests of the similarity of variance at the ≤1% level of significance (Davis, 1973). The MgO concentrations of the Deicke melt inclusions are also greater than those of the Millbrig melt inclusions. A Students t-test for the differences in FeO and TiO_2 are, respectively, 3.65 and 6.34. These differences are greater than the t-test statistic 3.39 corresponding to 30 degrees of freedom at a level of significance of ≤1% (Davis, 1973). The analyses of the similarity of variance (F-test) show the variances from the means are similar at the ≤1% level. The difference in Deicke and Millbrig compositions is illustrated in Figure 4. While two Millbrig quartz-hosted inclusion compositions plot close to the Deicke melt in-

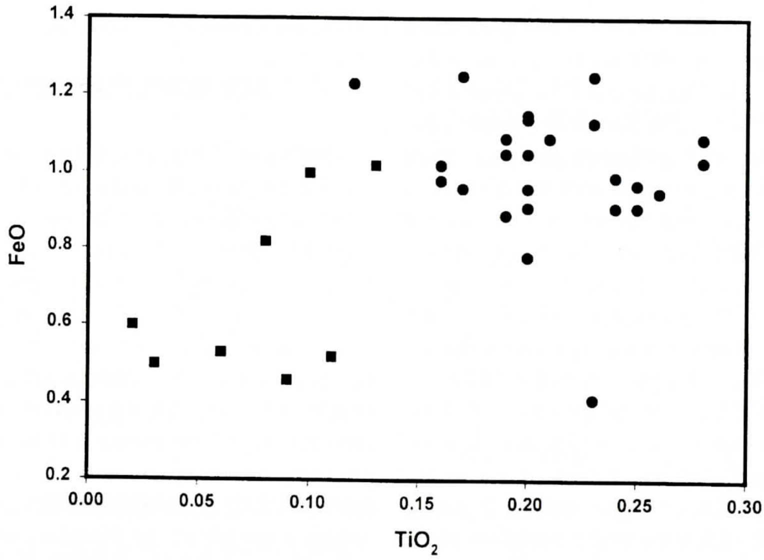

Figure 4: A plot of the FeO versus TiO_2 for melt inclusions from the Deicke (circles) and Millbrig (squares).

clusions, the analyses of the Deicke and the six remaining Millbrig melt inclusions are distinct from each other in this plot. The higher FeO and TiO_2 contents of the Deicke melt inclusions are consistent with the prevailing conclusion that the Deicke K-bentonite is derived from tephra having a more mafic magma source (Haynes, 1994). The FeO contents of the rhyolitic melt inclusions from Hinds Creek Quarry Deicke K-bentonite and the Millbrig K-bentonites are significantly lower than the FeO contents of the quartz-hosted Otsquago Creek K-bentonite (Tables 2 and 3). Thus, based on limited data, there are characteristic differences in the FeO and TiO_2 contents of melt inclusions between the Deicke and Millbrig K-bentonites. The FeO contents of the Deicke and Millbrig melt inclusions are lower than the FeO contents of the melt inclusions of the Otsquago Creek K-bentonite in the Utica Shale.

DISCUSSION AND SUMMARY

Cathodoluminesence analyses provide new insight regarding the calcite contents and the kinds of feldspars present in these K-bentonites. Calcite is the predominant cement in the Deicke K-bentonite as seen from CL data, with contents estimated to be as high as 22-24% (Table 1), which is much higher than past estimates of 1-5% by Haynes (1994). The source of much of the calcite is probably the enclosing Carters and Hermitage strata. It is also possible to distinguish pyroclastic plagioclase from diagenetic feldspars based on CL analyses. The differences in the overall mineralogy between the Deicke and Millbrig K-bentonites are also very apparent from the petrographic study and are consistent with previous results (e.g. Haynes, 1994).

Rhyolitic melt inclusions in plagioclase pyroclasts are not unusual features in volcanic rocks (e.g. Luhr and others, 1984; Irsara and others, 1993; Luhr and Melson, 1996; and Rutherford and Devine, 1996). The compositions of the rhyolite inclusions in Recent tephra are comparable to those of the plagioclase-hosted melt inclusions in Ordovician K-bentonites from this study. Rhyolitic melt inclusions are also found in middle to late Ordovician Utica Shale and in Millbrig-equivalent K-bentonites elsewhere in the Appalachian Basin (Delano and others, 1990, 1994; Hanson and others 1996; Kolata and others 1996; Adhya and others, 1999). The plagioclase-hosted melt inclusions in the Deicke and Millbrig K-bentonites are lower in FeO relative to the quartz-hosted rhyolite melt inclusions in the Ordovician Utica Shale (Tables 2 and 3), and the results in this study indicate the Deicke can be distinguished from the Millbrig in the Southern Appalachian Basin based on melt inclusion analyses. Most melt inclusions in this study do not contain fluid or gaseous phases, but the absence of gas and fluids in relatively large melt inclusions in tuffs and other pyroclastic deposits is not unusual (Roedder, 1979; Watson, 1984; Vogel and Aines, 1996). The low analytical totals (Tables 2 and 3) are due to the presence of volatiles.

Our results document the presence of abundant rhyolitic melt inclusions in plagioclase pyroclasts from Paleozoic K-bentonites of the Southern Appalachian Basin, and their usefulness in the identification and correlation of the Deicke and Millbrig K-bentonites. The melt inclusions within the calcic plagioclases (An_{53-77}) from the Deicke K-bentonite at Hinds Creek Quarry contain statistically distinct and higher amounts of TiO_2 and FeO compared to the rhyolitic melt inclusions from Millbrig K-bentonites (Figure 4). This difference is consistent with the conclusion that the Deicke K-bentonite is derived from tephra that was erupted from a more mafic source (Haynes, 1994). Mg and Fe were discriminating elements both here and in a previous study by Delano and others, 1994. The TiO_2 and FeO compositions of plagioclase hosts are not apparently a factor in explaining the differences in the compositions of the melt inclusions between the Deicke and Millbrig K-bentonites in this study (Verhoeckx-Briggs, 1997). Based on these data, we confirm the previous results of Delano and others (1994) and recent results of Adhya and others (1999), and we suggest that it is possible to identify and correlate the Deicke and Millbrig K-bentonites. Additional study is needed to test this emerging hypothesis that the Deicke and the Millbrig can be discriminated by melt inclusion chemistry (TiO_2, FeO, and possibly MgO) in the Southern Appalachian Basin. The Deicke K-bentonite may have been derived from a more mafic ash compared to the Millbrig K-bentonite based on the higher Mg and Fe contents of rhyolite inclusions, the higher anorthite contents in plagioclase, and the absence of quartz and biotite pyroclasts (Table 1).

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SAMPLE REGISTER[1]

1. Davis Crossroads, GA (Haynes, 1994, locality 22). Millbrig is ~0.5 meters thick. Bulk sample collected approximately half distance from base. Smithsonian Thin Section reference number: 116588-41.

2. Gadsden, AL (Haynes, 1994, locality 25). Millbrig is 0.75 meters thick. Bulk sample was collected from the basal 18 cm, which contains abundant pyroclastic grains. Smithsonian Thin Section reference number: 116588-47-2.

3. Ft. Payne, AL (Haynes, 1994. locality 24). This Millbrig K-bentonite is 25 cm thick, bulk sample collected. Smithsonian Thin Section reference number: 116588-48-2.

4. Overton County, TN (Elliott and Aronson,

1. The K-bentonite samples are keyed by number to locations given in Figure 2.

1993, locality 13). Millbrig K-bentonite is 77 cm thick in core 2-49-1 at 893-895.5 feet. Smithsonian Thin Section reference number: 116588-59.

5. Hinds Creek Quarry, Anderson County, TN (Haynes, 1994, locality 18). Deicke K-bentonite collected in outcrop, 25 cm thick. Bulk sample collected. Smithsonian Thin Section reference number: 116585-57.

6. Albany County, KY (Elliott and Aronson, 1993, locality 11). Deicke K-bentonite is 15 cm thick, bulk sample collected from core AA-6 at 1080 feet. Smithsonian Thin Section reference number: 116588-58.

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KYANITE COLOR AS A CLUE TO CONTRASTING PROTOLITH COMPOSITIONS FOR KYANITE QUARTZITES IN THE PIEDMONT PROVINCE OF VIRGINIA

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ABSTRACT

Kyanite quartzite at Willis Mountain in the central Piedmont Province of Virginia, contains very pale blue to almost white kyanite, whereas kyanite at nearby Baker Mountain is a deep blue to blue-green. We evaluated the possible compositional cause for this color difference by analyzing kyanite grains from both localities by electron microprobe. Willis Mountain kyanite is almost pure Al_2SiO_5, but Baker Mountain grains consistently contain measurable amounts of Cr and Fe (up to ~0.4 wt.% Cr_2O_3 and FeO). Thus, a combination of crystal field transition (Fe, Cr) and charge transfer ($Fe^{2+} \leftrightarrow Fe^{3+}$) mechanisms is probably responsible for the blue color at Baker Mountain.

The major element compositions from both localities are dominated by Si and Al, suggesting nearly pure quartz-kaolinite protoliths. As expected, levels of Cr are high at Baker Mountain (average ~600 ppm), but below detection limits at Willis Mountain. Concentrations of many other trace elements (Ni, Co, Zn, Ga, Nb, Y, Rb), as well as Mg, Ca, Na, and K are extremely low in all rocks. In contrast, Sn concentrations are uniformly high (10-26 ppm). Rocks from the two areas have different Zr concentrations and Zr/TiO_2 values. An interpretation for the origin of these deposits that is consistent with all our results involves metamorphism of volcanic rocks that were previously altered by severe leaching by hydrothermal fluids. The contrasting trace element signatures suggest that Baker Mountain rocks were derived from a mafic (basaltic) protolith, whereas the protolith at Willis Mountain was more evolved (possibly andesitic).

INTRODUCTION

Two localities in the central Piedmont Province of Virginia (Figure 1) have been sites of commercial kyanite mining, which began more than 60 years ago. The world's largest active kyanite mine is currently located at Willis Mountain in Buckingham County, but production at nearby Baker Mountain in Prince Edward County ceased many years ago. These deposits are similar in most respects, the dominant rock type being kyanite-bearing quartzite. However, kyanite crystals from the two localities show striking differences in color. Specifically, kyanite from Baker Mountain is a deep blue to blue-green, whereas at Willis Mountain it typically is pale blue to gray to almost white. This obvious color contrast was noted nearly 70 years ago by Watkins (1932) in an early report on the geology and economic aspects of the central Virginia "kyanite belt" (cf. Jonas, 1932).

The initial purpose of the present investigation was to evaluate these differences in color by determining kyanite compositions using an electron microprobe. Our results were somewhat surprising, and prompted us to obtain major and trace element analyses on selected whole-rock samples. The whole-rock data provide additional constraints on the nature of these rocks prior to metamorphism.

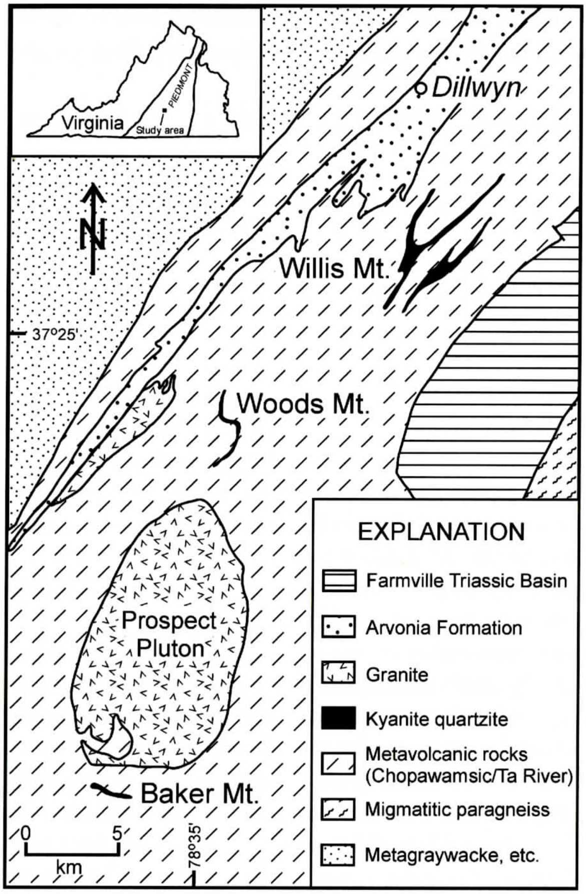

Figure 1. Geologic sketch map of the study area showing locations of the Willis and Baker Mountain kyanite quartzites. Minor amounts of kyanite quartzite also occur at Woods Mountain. Simplified from the geologic map of Virginia (Virginia Division of Mineral Resources, 1993).

GEOLOGIC SETTING

The kyanite quartzites occur in the central Piedmont Province of Virginia (Fig. 1), within what is currently known as the Chopawamsic Terrane (Horton and others, 1989). This terrane includes a variety of metavolcanic and metasedimentary rocks, as well as several granitic to tonalitic plutons. Geochemical data coupled with recent U-Pb zircon age determinations on both volcanic and plutonic rocks in this area indicate widespread arc magmatism in the middle Ordovician (Coler and others, 2000; cf. Pavlides, 1981).

Rocks in the immediate vicinity of the kyanite quartzites range from muscovite-feldspar-quartz schist to quartzofeldspathic biotite gneiss to amphibolite (Virginia Division of Mineral Resources, 1993). All of these rocks were originally mapped as part of the Chopawamsic Formation (e.g., Marr, 1980), and interpreted to represent a variety of predominantly metavolcanic rocks, with minor metasedimentary interlayers (e.g., Marr, 1992). On the current geologic map of Virginia (Virginia Division of Mineral Resources, 1993), the Chopawamsic Formation is now restricted to rocks west of the Arvonia Formation in the area of Figure 1. Similar, but typically more mafic metavolcanic rocks east of the Arvonia are thought to correlate with the Ta River Metamorphic Suite to the northeast (Rader and Evans, 1993; cf. Pavlides and others, 1982). The metasedimentary rocks (slate, schist, quartzite) of the Arvonia Formation (middle to late Ordovician) unconformably overlie Chopawamsic and Ta River rocks (Marr, 1980).

Several lines of evidence, including the presence of kyanite (± staurolite) in a variety of rocks and the overall abundance of amphibolite, indicate that the rocks of this area experienced metamorphism at amphibolite grade. Cochrane (1986) determined peak metamorphic conditions for biotite-amphibole gneisses in the vicinity of Willis Mountain to be ~600°C and ~6.5 kb.

Most previous studies have interpreted the kyanite quartzites of this region as a component within biotite and amphibole gneisses, and thus part of the Chopawamsic Formation or correlative rocks (Espenshade and Potter, 1960; Bennet, 1961; Brown, 1969; Gruen, 1982; Duke, 1983; Cochrane, 1986). However, Conley & Marr (1980) argued that the kyanite quartzites of Willis and Woods Mountains (about 15 km southwest of Willis Mountain; Fig. 1) should be correlated with the Arvonia Formation. In support of this interpretation, Conley and Marr (1980) described what they interpreted as relict sedimentary structures in the kyanite quartzites, including cross-beds and graded bedding, as well as local conglomeratic horizons. In addition, they suggested that the protolith for the quartzite was aluminous sandstone, and thus unlike any lithology in the dominantly volcanic-derived Chopawamsic Formation. Conley and Marr (1980) used these observations and interpretations, as well as similarities in stratigraphic sequence between the Arvonia in its type area and the kyanite-bearing rocks of this study to argue for their correlation.

PREVIOUS WORK

Origin of Kyanite Quartzites

Ideas about the origin of the kyanite quartzites in this region of Virginia have evolved from simple models involving isochemical metamorphism of aluminum-rich sedimentary rocks (Jonas, 1932; Jones and Eilertson, 1954) to models requiring severe hydrothermal alteration of a volcanic protolith followed by metamorphism (Good, 1981; Cochrane, 1986). Espenshade and Potter (1960) provided the first detailed field and petrographic investigation of these deposits, and considered them to be metamorphosed clay-bearing sandstones. Although Espenshade and Potter (1960) recognized the importance of hydrothermal alteration related to volcanism in the origin of many high-Al quartz-rich rocks in the southeastern U.S., they did not favor this mechanism for the Willis and Baker Mountain deposits, primarily because of their occurrence as "persistent layers at a restricted stratigraphic position" (Espenshade and Potter, 1960, p. 51). At about the same time, Bennet (1961) concluded that Willis Mountain and other deposits in the area represent metamorphism of extremely pure mixtures of quartz and kaolinite, which originated either through lateritic weathering or circulation of meteoric water. As noted above, Conley and Marr (1980) interpreted kyanite quartzites at Willis and Woods Mountain as metamorphosed aluminous sandstones. However, Gruen (1982) expressed skepticism regarding the relict sedimentary structures described by Conley and Marr (1980), and suggested that features resembling cross-beds may actually represent combinations of schistocity, mineral lineations, and intersecting fractures. Furthermore, Gruen (1982) argued that quartz "pebbles" in the alleged conglomerates are actually quartz porphyroblasts. Duke (1983) noted the relative rarity of these possible sedimentary features, and suggested that, if real, they might simply represent local winnowing of pyroclastic material.

More recent investigations have emphasized the probable importance of hydrothermal alteration of volcanic rocks to account for various high-aluminum quartz-rich rocks in the southeastern U.S., in which the aluminous mineral is kaolinite, pyrophyllite, or an alumino-silicate (e.g., Carpenter and Allard, 1982; Feiss, 1985; Schmidt, 1985). In these models, the aluminous rocks represent the product of severe hydrothermal (argillic) alteration by acidic fluids in a subvolcanic setting, such that all mobile elements have been stripped from the rocks (Schmidt, 1985; cf. Henley and Ellis, 1983). Therefore, the bulk compositions of these rocks reflect residual aluminum and silica enrichment, accompanied in some cases by addition of silica. The current Al-rich mineral is a function of metamorphic grade.

Good (1981) drew attention to the evidence for a volcanogenic origin for massive sulfides in the Willis Mountain area, and he seems to be the first to have suggested a purely volcanic origin for the kyanite quartzites. Marr (1992) offered a compromise interpretation involving initial deposition of aluminous sandstone, metamorphism to kyanite grade, and subsequent hydrothermal alteration. However, by analogy to modern volcanic systems, the severe hydrothermal alteration and leaching probably occurred concurrently with or shortly following volcanism, rather than long afterward (Carpenter and Allard, 1982). Gruen (1982) and Cochrane (1986) both advocated a volcanic origin for the protoliths of the kyanite quartzites in this region, and argued for hydrothermal alteration by acidic fluids as the mechanism for Al- and Si-enrichment. Cochrane (1986) proposed that a subvolcanic pluton provided the heat that drove the hydrothermal system, and that the hydrothermal fluids were at a temperature of 100°-200° C and a pH between 2 and 4.

Color in Kyanite

A number of studies have focused on the color-causing elements and mechanisms in kyanite [see Kerrick (1990) for a review]. The elements typically correlated with a blue to blue-green color include Ti (White and White, 1967; Rost and Simon, 1972), Fe (Faye and Nickel, 1969), and Cr (Cooper, 1980; Bosshart and others, 1982; Neiva, 1984; Gil Ibarguchi and others, 1991), either alone or in combination (e.g., Ghera and others, 1986). The primary mechanisms suggested to be responsible for inducing color include $Fe^{2+} \leftrightarrow Fe^{3+}$ charge transfer (Faye and Nickel, 1969), $Fe^{2+} \leftrightarrow Ti^{4+}$ charge transfer (Smith and Strens, 1976; Parkin and others, 1977), or crystal field transitions in Cr or Fe (Langer, 1976; Parkin and others, 1977). Thus, colorless kyanite tends to be relatively pure Al_2SiO_5, whereas more typical blue kyanite contains small amounts of one or more of these color-causing transition metals.

Most previous investigations have focused specifically on determining the element and mechanism responsible for kyanite color (including color zoning), with less attention to geologic setting or petrologic implications. Anomalously high Cr in kyanite is clearly related to bulk rock composition (e.g., Sobolev and others, 1968; Delor and Leyreloup, 1986). Gil Ibarguchi and others (1991) reported colorless Cr-poor kyanite inclusions in amphibole and blue, higher-Cr kyanite in the matrix of high-P ultramafic rocks. Gil Ibarguchi and others (1991) suggested that these varieties might provide evidence for kyanite growth under increasing pressure conditions, based on experiments by Seifert and Langer (1970). Cooper (1980) reported kyanite from white schists that was zoned with respect to Cr, up to a maximum of 2.88 wt.% Cr_2O_3, but did not offer an explanation for the zoning. In contrast to these earlier investigations, our study is somewhat unique in evaluating regional variations in kyanite color.

ANALYTICAL METHODS

Kyanite compositions were determined using a JEOL JXA-8900 SuperProbe at the University of Maryland, College Park. Operating conditions were 15 kV, 20 nA beam current, and a spot size that varied from 1-10 μm. All grains were analyzed for Si, Al, Ti, Fe, Mg, Mn, Ca, and Cr. In nearly every case, values for Ti, Mg, Mn, and Ca were below detection limits in kyanite from both localities. These detection limits, reported on a weight percent oxide basis, are approximately 0.04% (TiO_2), 0.04% (MgO), 0.02% (MnO), and 0.04% (CaO). Detection limits for FeO and Cr_2O_3 are approximately 0.01% and 0.02%, respectively. Counting times on the peak and background positions for Ti, Cr, and Fe were 40 and 20 seconds, respectively. Whole-rock samples were analyzed for major element oxides and trace elements by X-ray fluorescence (XRF) analysis at Washington University in St. Louis, using methods described by Couture and others (1993) and Couture and

Table 1. Modal mineralogy of kyanite quartzites

Locality*	BM	BM	BM	BM	BM	WM	WM	WM	WM	WM	WM
Sample	SD99-1	SD99-2	SD99-3	SD99-4	SD99-5	SD99-6	SD99-7	SD99-8	VK-1	VK-2	VK-3
quartz	58.1	45.4	55.0	54.2	9.2	62.1	59.0	63.4	71.0	52.8	70.2
kyanite	39.1	50.5	39.2	42.5	78.9	35.6	37.3	33.8	24.7	45.2	28.2
rutile	1.3	3.3	4.1	2.2	4.1	1.2	3.4	2.1	0.7	0.6	1.0
white mica	1.5	0.7	0.4	1.0	7.6	1.1	0.3	0.7	0.7	1.1	0.5
pyrite	--	--	--	--	--	tr	tr	tr	tr	tr	tr
topaz	--	--	--	--	--	--	--	--	1.2	0.3	0.1
zircon	--	--	--	--	tr	tr	tr	tr	0.1	tr	tr
other†	--	0.1	1.3	0.1	0.2	--	--	--	1.5	--	--
total points	821	756	727	815	828	845	826	767	806	650	819

* BM = Baker Mountain; WM = Willis Mountain
† = alteration products and rare Ba-Sr sulfates
tr = observed but not counted

Dymek (1996).

SAMPLING AND PETROGRAPHY

Samples from Willis Mountain (SD99-6, SD99-7, SD99-8, VK-1, VK-2, VK-3) were collected from fresh blasted rock at the mine. Kyanite in these samples varies in color from white to gray to pale blue. The Baker Mountain pit has now been reclaimed, and no similarly fresh samples are available. Most samples (SD99-1, SD99-2, SD99-3, SD99-4) were collected from a nearby wooded area that contains abundant float of the quartzite, and the rocks here are somewhat weathered. One sample (SD99-5) was collected at the former mine site, in the vicinity of a processing plant currently operated by the Kyanite Mining Corporation. Kyanite crystals in the Baker Mountain samples are typically blue to dark blue-green.

Modal abundances of minerals in eleven samples are listed in Table 1. A number of thin sections were cut with the intention of including as much kyanite as possible, so the reported amounts are probably not representative of a given sample in these cases. Nonetheless, the listed modes provide some indication of the nature and amounts of accessory minerals. Espenshade and Potter (1960) estimated that kyanite constitutes 10-30% of the quartzites in Virginia, and Cochrane (1986) found an average of 27.6% kyanite at Willis Mountain.

Samples from Willis and Baker Mountains are petrographically similar, and consist dominantly of quartz and kyanite. Quartz grains are typically anhedral and range in size from <0.1 up to 10 mm in diameter. Subhedral to euhedral elongate grains of kyanite range to a maximum of about 4 cm long, and in many samples show some preferred orientation. Kyanite in all samples is colorless in thin section, and typically displays multiple twins parallel to {100} under crossed polars. Rutile is an abundant accessory, and ranges in size from <0.1 mm up to ~2.0 mm in diameter. It occurs as anhedral to subhedral interstitial grains, or as inclusions in quartz or kyanite. Rutile is consistently more abundant in the Baker Mountain samples, and this difference is reflected in higher whole-rock TiO_2 concentrations (Table 3). White mica is common, both as isolated subhedral grains up to ~8 mm long, and as masses adjacent to and within kyanite grains. Cochrane (1986) reported muscovite and paragonite in the Willis Mountain rocks, and Espenshade and Potter (1960) reported, in addition, fuchsite from Baker Mountain. Trace amounts of zircon are present in all Willis Mountain samples, and topaz was observed in two samples (VK-1 and VK-2). Although pyrite is widespread at Willis Mountain, it occurs only in trace amounts in our samples. Pyrite was not observed in the Baker Mountain samples collected by us, but is known to occur in rocks from this locality.

The only other mineral observed in our samples is a Ba-Sr sulfate (possibly a barite-celestite solid solution), which occurs in trace amounts at Willis Mountain. Additional miner-

Table 2. Representative analyses of kyanite

Locality*	BM	BM	BM	BM	BM	WM	WM	WM	WM	WM
Sample	SD99-1	SD99-1	SD99-1	SD99-2	SD99-3	SD99-6	SD99-7	SD99-7	SD99-8	SD99-8
SiO_2	37.25	36.83	36.70	37.49	37.21	36.96	37.28	36.87	36.69	36.50
Al_2O_3	62.99	63.07	62.61	62.48	62.97	63.77	63.31	63.27	63.65	63.62
Cr_2O_3	0.01	0.24	0.38	0.13	0.18	0.01	0.04	0.02	--	--
FeO	0.18	0.28	0.32	0.39	0.11	0.01	--	0.01	--	--
MgO	0.01	0.01	0.03	0.01	--	--	--	--	--	--
MnO	--	0.01	0.04	--	0.01	0.01	--	0.02	0.01	0.01
CaO	0.02	0.06	0.02	--	0.03	--	0.03	0.01	--	--
Total	100.56	100.50	100.10	100.51	100.51	100.76	100.73	100.23	100.38	100.17

* BM = Baker Mountain; WM = Willis Mountain
-- = below detection limit

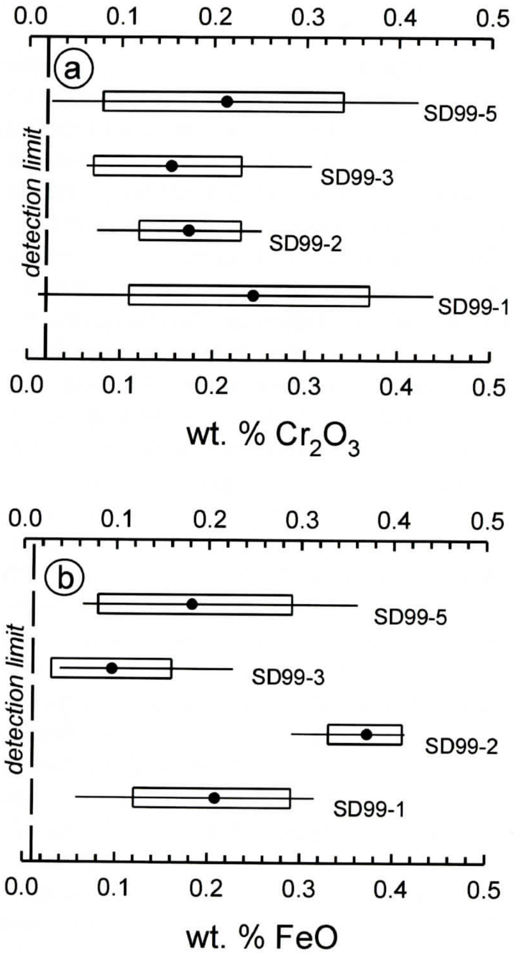

Figure 2. Plots illustrating concentrations of (a) Cr_2O_3 and (b) FeO in kyanite grains from Baker Mountain. On both plots, the solid circle represents the average composition, the box the standard deviation, and the line the total measured range.

als previously reported from these deposits include rare apatite, trolleite, lazulite, variscite, and garnet, among others (Espenshade and Potter, 1960; Mitchell and Fordham, 1987).

KYANITE COMPOSITIONS

Approximately 70 analyses of kyanite were obtained from seven samples, including four from Baker Mountain and three from Willis Mountain. On numerous grains, several points were analyzed to evaluate compositional heterogeneity. Representative compositions are listed in Table 2.

Most analyses of white to pale-colored kyanite from Willis Mountain show no detectable Ti, and only a few individual spots contain barely detectable Fe or Cr (Table 2). Thus, Willis Mountain kyanite is relatively pure Al_2SiO_5. Titanium was also not detected in most analyses of Baker Mountain kyanite, although a few analyses of areas immediately adjacent to rutile show slightly elevated TiO_2 (just above the detection limit). This result might be an artifact of beam overlap with rutile.

In contrast to Willis Mountain kyanite, the more strongly-colored kyanite from Baker Mountain consistently contains measurable levels of Cr and Fe, ranging to concentrations as high as 0.44 wt.% Cr_2O_3 and 0.41 wt.% FeO (as reported, but probably present as Fe_2O_3). Concentrations of Cr and Fe in Baker Mountain grains are illustrated in a series of "box-and-whisker" diagrams in Figure 2, where filled circles represent average concentrations, boxes indicate the standard deviation, and lines the total

Table 3. Whole-rock major and trace element compositions of kyanite quartzites (major elements in wt%, trace elements in ppm).

Locality	BM	BM	WM	WM	WM
Sample	SD99-1	SD99-3	VK-1	VK-2	VK-3
SiO_2	64.64	70.65	76.59	71.15	76.62
TiO_2	1.52	1.04	0.68	0.67	0.56
Al_2O_3	32.18	26.91	20.66	26.70	19.78
Fe_2O_3(T)	0.32	0.68	0.05	0.28	1.56
MnO	0.01	--	--	--	--
MgO	0.05	--	0.02	0.10	0.02
CaO	--	--	--	0.01	0.02
Na_2O	0.21	0.04	0.13	0.17	0.12
K_2O	0.07	0.08	0.03	0.03	0.03
P_2O_5	--	0.01	0.06	0.14	0.06
LOI	0.37	0.35	1.12	0.78	1.06
Total	99.37	99.76	99.32	100.03	99.81
Sn	26.1	25.5	22.5	12.4	10.0
Nb	<3	<3	9.8	9.5	7.1
Zr	64.5	47.9	186	193	148
Y	<3	<3	<3	<3	<3.6
Sr	<3.6	<1.4	222	446	128
Rb	<3.6	<2.3	<1.2	<1.7	<1.6
Pb	<4.5	9.3	10.8	24.5	20.9
Ga	11.3	11.3	<3.6	<1.5	<3.6
Zn	<1.8	<1.8	<2.8	<1.7	<2.0
Ni	<5.8	<9	<8.4	<4.3	<3.2
V	306	220	81.4	59.9	64.3
Cr	683	503	<42.6	<42.6	<42
Ba	<13.4	<21.3	<40.2	700	239
Co	<5.0	<5.1	<3.0	<3.5	<9.6

measured range. In virtually every analysis, concentrations of these minor elements are above the detection limits. Individual grains show some heterogeneity with respect to Cr and Fe content, although no obvious zoning patterns were detected. Concentrations of the two elements are not correlated, but there are no crystal chemical reasons for expecting such a correlation, assuming simple substitution of Cr^{3+} or Fe^{3+} for Al^{3+}. A plot of wt.% Cr_2O_3 + wt.% FeO vs. wt.% Al_2O_3 (not shown) does display a crude negative correlation, as expected, albeit with considerable scatter.

The levels of Fe in Baker Mountain kyanite were not unexpected, but the equally elevated Cr content was somewhat surprising. Given the abundance of kyanite, these levels of Cr suggested that whole-rock samples from Baker Mountain might contain unusually high Cr, especially for rocks that are otherwise dominated by quartz. For example, simple mass balance calculations assuming a minimum of 20% modal kyanite and an average Cr_2O_3 value of 0.2 wt.% (~1370 ppm Cr) indicate about 275 ppm Cr in the whole rock. To test this calculation, we analyzed whole rock samples from both localities to evaluate possible differences in Cr concentration (as well as other elements).

WHOLE-ROCK COMPOSITIONS

Two whole-rock samples from Baker Mountain and three from Willis Mountain were analyzed for major and trace elements, and the results are listed in Table 3. The major element compositions of all samples are dominated by SiO_2 and Al_2O_3, as expected, and variations in these oxides are clearly a function of the relative proportions of quartz and kyanite. The Baker Mountain samples contain higher TiO_2 (up to 1.5 wt.%) than those from Willis Mountain, consistent with higher modal rutile (Table 1). Levels of Fe_2O_3(T) are typically <1 wt.%, but sample VK-3 from Willis Mountain contains 1.69 wt.%. The Fe in Willis Mountain samples is probably a function of pyrite content (Table 1), given the Fe-poor nature of the kyanite. At Baker Mountain, at least some of the Fe is present in kyanite. Sodium occurs in small amounts ($\leq$0.21 wt.%) in several samples, and is probably present in the paragonite component of muscovite. All other major elements are present in negligible amounts in most samples.

As expected from the kyanite analyses, Cr concentrations are significantly different in rocks from the two localities. At Willis Mountain, levels of Cr are below the limit of detection by XRF (about 50 ppm at the Washington University lab). In contrast, both Baker Mountain samples have much higher Cr, with values of 500 and 680 ppm, respectively. All samples contain detectable V, but Baker Mountain samples have higher amounts (202-306 ppm) than Willis Mountain (60-81 ppm), consistent with differences in TiO_2 noted above. Concentrations of other transition elements, including Co, Ni, and Zn are below detection limits in all samples, as are concentrations of Y.

All samples contain relatively high amounts of Sn (up to 26 ppm), which probably is present in rutile (cf. Urusov and others, 1999). Concentrations of Zr are higher at Willis Mountain (148-193 ppm) relative to Baker Mountain (50-65 ppm) by about a factor of 3. The same may be true for Nb, although levels at Baker Mountain are below the detection limit. All Willis Mountain samples contain much higher Sr (up to 446 ppm) relative to Baker Mountain, and two samples contain high Ba (239 and 700 ppm), consistent with presence of the Ba-Sr sulfate mineral noted above. On the other hand, all samples contain very low Rb (<4 ppm in all cases).

A striking feature of the trace element data is the very low Ga concentration in all rocks. Because of charge and size similarities, Ga substitutes primarily for Al in minerals, and the two elements typically show a good positive correlation in most crustal rocks (Schroll, 1999). Thus, the low Ga levels (below the detection limit in many cases) in the kyanite quartzites are unexpected for rocks that contain such high Al. For example, given an average upper crustal Ga/Al value of 2.1 x 10^{-4} (Taylor and McLennan, 1985), these rocks should contain about 22 to 36 ppm Ga.

DISCUSSION

Kyanite Color: Elements and Mechanisms

A reasonable interpretation of our microprobe results is that Fe and Cr are primarily responsible for producing the blue to blue-green color of Baker Mountain kyanite. If Ti plays any role (cf. White and White, 1967), it must do so at levels that are below the detection limits of the microprobe procedure used in this study. However, we consider it likely that most, if not all, of the Ti in these rocks is present in rutile.

The presence of Fe and Cr suggests two possible mechanisms for the production of color in Baker Mountain kyanite. These mechanisms are: 1) crystal field transitions, which are possible in both elements; and 2) $Fe^{2+} \leftrightarrow Fe^{3+}$ charge transfer. Because both elements are present, a combination of these processes may be responsible for the particular color of a given grain. In addition, the fact that concentrations of Fe and Cr are not correlated implies that the specific color-causing mechanism may vary across a grain, although the net effect is to produce a similar color throughout.

Implications of Whole-Rock Compositions for Origin of the Deposits

As noted above, the origin of these deposits has been the subject of considerable debate, and our geochemical results permit us to evaluate this problem in some new ways. As a first observation, we note that the major element compositions are overwhelmingly dominated by Si and Al, with only minor Fe and Ti, and negligi-

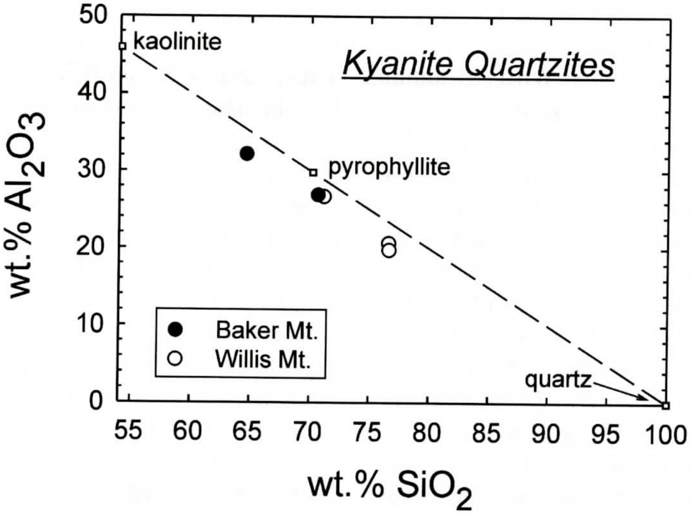

Figure 3. A plot of wt.% SiO_2 vs. wt.% Al_2O_3 in whole-rock samples of kyanite quartzite from Baker and Willis Mountains. Included for reference are the positions of end-member quartz, pyrophyllite, and kaolinite (plotted on an anhydrous basis).

ble Ca, Na, and K. These results rule out any kind of typical aluminous sandstone, i.e. arkose, as a protolith. On the other hand, the major element compositions are completely consistent with protoliths that contained variable mixtures of quartz and kaolinite, as illustrated on a plot of wt.% SiO_2 vs. wt.% Al_2O_3 in Figure 3. All samples fall near a line representing pure quartz-kaolinite mixtures, with small amounts of offset primarily due to additional Fe or Ti. Thus, development of nearly pure quartz-kaolinite protoliths seems to be a necessary precursor step in the formation of these deposits. Such protoliths potentially could develop in a number of ways, but our results suggest that they originated by severe hydrothermal alteration of volcanic rocks, as described by Schmidt (1985) for other high-Al deposits in the southeastern U.S., and by Henley and Ellis (1983) in general.

The surprisingly high concentration of Cr in the Baker Mountain rocks provides compelling evidence for a mafic igneous protolith in this case. The absolute amount of Cr in the Baker Mountain samples probably reflects some residual enrichment of unknown magnitude, assuming relative immobility of Cr in hydrothermal fluids. In principle, the severe alteration of shale, which may also contain moderate amounts of Cr [e.g. ~110 ppm in the average post-Archean shale of Taylor and McLennan, 1985], could result in a similar quartz-kaolinite, Cr-bearing precursor rock. However, known examples of advanced argillic alteration are restricted to areas of active volcanism, suggesting that a clastic sedimentary protolith is unlikely.

An interpretation involving a mafic igneous protolith at Baker Mountain obviously requires drastic modification of the original rock composition, including major increases in Si and Al, concomitant with almost complete removal of Mg, Fe, Ca, Na and K. In addition, most other transition elements found in basalts must have been stripped from the protolith, including Ni, Co, and Zn. Although extreme, such effects are exactly those produced in known or inferred areas of severe hydrothermal alteration by highly acidic fluids in subvolcanic settings (cf. Henley and Ellis, 1983).

Rocks that are clearly analogous to those at Baker Mountain have been reported from Archaean greenstone belts in Australia and South Africa. In western Australia, Martyn and Johnson (1986) and Ashley and Martyn (1987) described high-Cr quartz-andalusite rocks that show enrichments and depletions similar to those at Baker Mountain, including very low Ni. These authors interpreted the rocks to represent metamorphism of intensely altered komatiites, with which they are associated in the field. Ashley and Martyn (1987) calculated that the transformation from komatiite to quartz-andalusite precursor rocks involved volume decreases up to a factor of five, although in some cases Si had clearly been added rather than simply enriched residually. In South Africa, Schreyer and others (1981) described unusually aluminous corundum-fuchsite rocks, and also interpreted them as metamorphosed, severely altered komatiites. Schreyer (1987) later drew attention to some mineralogical and geochemical similarities between these rocks and the aluminous quartzites of the Carolina slate belt. In noting these analogous rocks, we do not mean to imply that the kyanite quartzites of this study originated as komatiites. Nonetheless, the examples described here illustrate the potentially extreme effects of hydrothermal alteration.

A mineralogical feature entirely consistent with the hydrothermal alteration hypothesis is the widespread presence of rutile in all rocks. Rutile is a characteristic mineral produced during the breakdown of other Ti-bearing minerals in porphyry alteration systems (Czamanske and others, 1981), and "its formation is virtually inevitable if this alteration is sufficiently severe" (Force, 1991, p. 49). Thus, the presence of rutile in the kyanite quartzites reflects the relatively immobile nature of Ti during such hydrothermal alteration, and accounts for the current levels of TiO_2 (and probably V) in these rocks. Schmidt (1985) also reported the consistent presence of rutile in other high-Al deposits in the southeastern U.S.

At least two other geochemical features of the rocks from Willis and Baker Mountains suggest that their protoliths were highly modified by hydrothermal fluids. First, we note that Sn values (10-26 ppm) are very high relative to average upper crust (~3 ppm), and similar to the concentrations in tin granites (Lehmann, 1990). Furthermore, the juxtaposition of high Cr and high Sn at Baker Mountain is particularly unusual because these elements typically occur in high concentrations in very dissimilar environments. One possible explanation for the high Sn values is that they also represent residual concentrations, although it is more likely that Sn was introduced by hydrothermal fluids, particularly if such fluids were fluorine-bearing (see below).

Second, the extremely low Ga/Al values displayed by these rocks require the action of some process that strongly fractionates Ga from Al. This same geochemical feature was reported by Schreyer and others (1981) in the corundum-fuchsite rocks noted above, as well as in some high-Al deposits of the southeastern U.S. (Schreyer, 1987). Schreyer and others (1981) rejected the possibility that this Ga/Al fractionation was due to intense lateritic weathering because bauxites do not show similar depletions in Ga. For this reason, in addition to other geochemical arguments, Schreyer and others (1981) concluded that such deposits cannot represent metamorphosed bauxites. Schreyer and others (1981) initially suggested that Ga/Al fractionation could be due to growth of some mineral (particularly alunite) that might reject Ga relative to Al, in a volcanic exhalative environment. However, Schreyer (1987) subsequently abandoned this idea, and noted that an adequate explanation for this anomaly remained to be determined. We speculate that this feature may be due to preferential leaching of Ga by fluorine-bearing fluids, because of the known tendency for Ga to complex with F (Schroll, 1999). Evidence for the action of fluorine-bearing fluids is clearly present in the rocks in form of trace amounts of topaz.

The different levels of Cr at Baker and Willis Mountains imply contrasting protolith compositions, assuming that both originated in a similar fashion as a consequence of severe hydrothermal alteration of volcanic rocks. Specifically, high Cr at Baker Mountain requires a mafic protolith, whereas the much lower Cr at Willis Mountain implies an intermediate to felsic protolith. In support of this interpretation, we note that Willis Mountain samples contain substantially higher amounts of the immobile element Zr, consistent with a more evolved protolith composition. Furthermore, values of Zr/Ti are higher in all Willis Mountain samples rel-

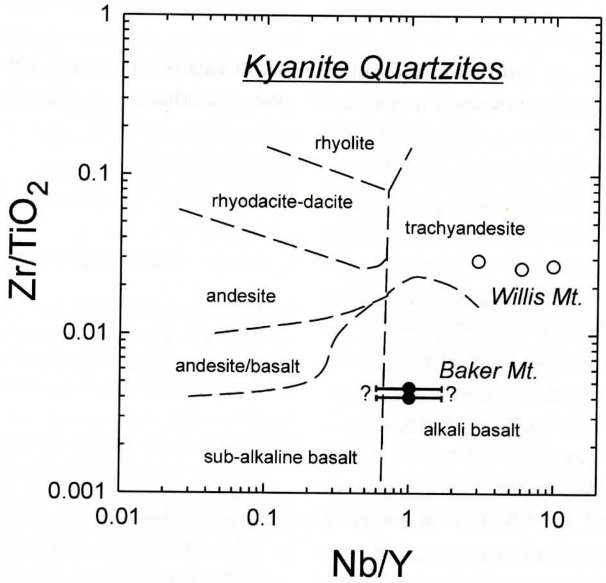

Figure 4. A plot of Zr/TiO_2 vs. Nb/Y in whole-rock samples of kyanite quartzite. Values for Nb/Y are not well-constrained in the Baker Mountain samples because of low concentrations, and an arbitrary value of 1 is used for plotting purposes. Error bars with question marks reflect this uncertainty. Similarly, the plotted positions for Willis Mountain samples reflect minimum Nb/Y values, based on the possible maximum concentration of Y (Table 3). Fields from Winchester and Floyd (1977).

ative to those from Baker Mountain. On the Zr/TiO_2 vs. Nb/Y discriminant diagram of Winchester and Floyd (1977), Willis Mountain samples plot in the field of trachyandesite, whereas Baker Mountain samples have basaltic ratios (Fig. 4). Because values of Nb and Y are low in all samples, and below the detection limits in many cases, accurate Nb/Y values cannot be calculated. For plotting purposes, an arbitrary value of 1 was used for Baker Mountain samples, and minimum values were calculated for Willis Mountain using measured Nb and possible maximum Y concentrations (Table 3). These uncertainties in Nb/Y preclude the use of Fig. 4 as a direct indicator of protolith rock type. Nonetheless, the contrast in Zr/TiO_2 values is a robust difference between the Baker and Willis Mountain samples. Furthermore, it is obvious from Fig. 4 that this ratio is relatively constant among the samples at each locality, suggesting that the original ratios have not been modified. At a minimum, these results clearly suggest contrasting mafic and felsic (or intermediate) protoliths at Baker and Willis Mountains, respectively. The higher Sr and Ba concentrations at Willis Mountain may also reflect these protolith differences, but it is unlikely that these are primary igneous values given the mobility of these elements in hydrothermal fluids.

Our interpretation for the origin of these deposits clearly differs from those of Espenshade and Potter (1960) and Conley and Marr (1980), who argued for aluminous sandstone protoliths. The presumed stratabound nature of the deposits was cited by Espenshade and Potter (1960) as evidence for a sedimentary origin. However, the quartz-andalusite rocks of Western Australia described above are strongly stratiform (Ashley and Martyn, 1987), which indicates that this characteristic does not necessitate a sedimentary origin. Furthermore, Gruen (1982) pointed out that kyanite quartzite in the Woods Mountain area (Fig. 1) is not everywhere conformable. In our opinion, the evidence described by Conley and Marr (1980) for relict sedimentary structures in these rocks is not compelling, and we agree with Gruen (1982) that other interpretations of the observed features apply [a view now shared by Marr (personal communication, 2001)].

Collectively, our results lead us to conclude that these kyanite quartzites originated in association with other metavolcanic rocks of this region, i.e., those of the Chopawamsic Formation and correlative rocks. Such rocks potentially represent unaltered varieties of the protoliths for the kyanite quartzites, and a more quantitative evaluation of our interpretations, including mass-balance calculations, could perhaps be made by determining whole-rock compositions of amphibolites, etc. near these deposits. Pavlides and others (1982) and Coler and others (2000) provided a few analyses of metavolcanic rocks in this region, but no previous geochemical study has focused specifically on rocks in the immediate vicinity of the kyanite quartzites.

In final support of our conclusions, we note that modern examples of the intense hydrothermal alteration effects like those advocated here occur most frequently in volcanic arc environments (Carpenter and Allard, 1982; Henley and Ellis, 1983), and the Chopawamsic Terrane probably developed in an arc setting (Horton and others, 1989). The driving force for such hydrothermal alteration is generally considered to be heat from subvolcanic plutons (e.g., Schmidt, 1985; cf. Cochrane, 1986). The ~458 Ma Prospect granite (Fig. 1) is a potential example of such a pluton, and it is only slightly younger than ~470 Ma metarhyolite just to the west of the study area (Coler and others, 2000). Thus, the presence of volcanic and plutonic rocks of similar age in this region is consistent with the development of the active hydrothermal systems required to produce the kyanite quartzite protoliths.

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