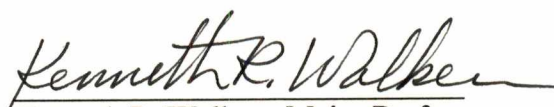
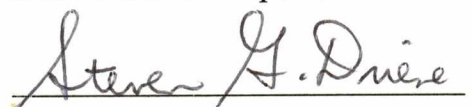
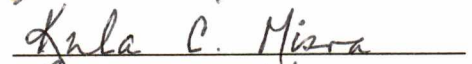
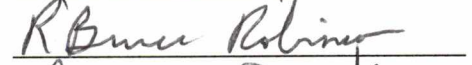
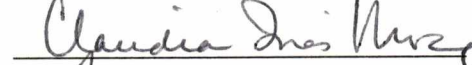
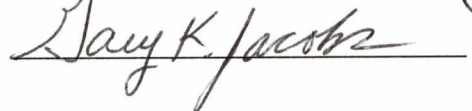


To the Graduate Council:

I am submitting herewith a dissertation written by J. Lincoln Foreman entitled "Petrologic and Geochemical Evidence for Water-Rock Interaction in the Mixed Carbonate-Siliciclastic Nolichucky Shale (Upper Cambrian) in East Tennessee." I have examined the final copy of this dissertation for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Doctor of Philosophy, with a major in Geology.


Kenneth R. Walker, Major Professor

We have read this dissertation and
recommend its acceptance:

Accepted for the Council:


Vice Provost
and Dean of The Graduate School

PETROLOGIC AND GEOCHEMICAL EVIDENCE FOR WATER-ROCK
INTERACTION IN THE MIXED CARBONATE-SILICICLASTIC NOLICHUCKY
SHALE (UPPER CAMBRIAN) IN EAST TENNESSEE

A Dissertation
Presented for the
Doctor of Philosophy
Degree
The University of Tennessee, Knoxville

J. Lincoln Foreman

May 1991

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ABSTRACT

Intraclastic limestone, thin fossiliferous limestone, and oolitic limestone interbedded with fissile shale are characteristic of the Nolichucky Shale (Upper Cambrian) in the Whiteoak Mountain thrust sheet, eastern Tennessee. These lithologies are interpreted as slope and basinal units deposited in water depths that may have exceeded 250-300 m. Intraclastic limestone debris flow deposits and skeletal and oolitic calcarenite turbidites indicate that nearly all carbonate is allochthonous, derived from up-slope and platform shallow-water environments. Storms were a dominant factor in initializing down-slope transport from shallower-water settings.

Polymictic limestone conglomerate and shallow-water-derived calcarenite composed of ooids, peloids, and fossils, indicate that depositional slopes were steep enough at times to permit considerable transport (in excess of 20 km) from up-slope areas. During times of increased siliciclastic sedimentation, carbonate production was suppressed and the basin margin had a gentle ramp morphology. Algal and oolitic shoals built up close to sea level when siliciclastic input was reduced, resulting in the progradation of carbonate facies into the basin. Depositional slopes at the basin margin were steepened to the point that gravity sediment flows were effective in transporting shallow water carbonate lithologies into the basin.

A variety of ooid cortical fabrics are preserved in shallow-water and deep-water oolitic limestone from the Nolichucky Shale. Petrographic and geochemical evidence suggest that these ooids were originally high-magnesium calcite fibrous radial ooids. Unequivocal evidence for originally aragonitic and bimineralic ooids is lacking. The ooids are now low-magnesium calcite ooids with approximately 1.5 mol% MgCO_3 . Iron and manganese concentrations in these ooids are 2-3 orders of magnitude greater than predicted and observed for unaltered Modern marine calcite.

The elevated iron and manganese concentrations in the Nolichucky ooids reflect stabilization and progressive alteration in the presence of pore fluids modified by interaction with interbedded shale. Fibrous radial ooids with relatively good original fabric preservation and sparry radial ooids with poor original fabric preservation commonly occur in the same thin sections. This can be explained by different degrees of water-rock interaction which was probably controlled by small-scale variations in microporosity and permeability. Increased water-rock interaction resulted in aggrading neomorphism of fibrous radial ooids to sparry radial ooids and associated increases in Fe^{2+} and Mn^{2+} in the sparry radial ooids.

The Nolichucky Shale from the southern Appalachian foreland thrust system experienced a three-stage history of calcite vein formation from oldest to youngest: bed-normal, bed-parallel, and bed-oblique fracturing, respectively. Freezing temperatures from primary fluid inclusions in calcite veins demonstrate that all veins precipitated from Na-Ca-Cl brines with ca. 27 wt% NaCl equivalent salinities.

The Whiteoak Mountain thrust sheet contains four coeval bed-normal vein sets trending 015° , 055° , 090° , and 320° . These pre-tectonic veins are cut by syn-tectonic bed-parallel slickensided veins. Fluid inclusion thermometry from calcite veins indicates precipitation at 80-110 $^{\circ}\text{C}$ for all bed-normal veins. These temperatures correspond to burial depths of 2.4-3.6 km, which were most likely attained during the late Mississippian to early Pennsylvanian. At that time, horizontal stress directions were controlled by both Alleghanian and Ouachitan orogenic stresses. The interaction of these two stress fields may account for the simultaneous development of the four coeval bed-normal sets. In this model, the 055° and 320° sets formed parallel and perpendicular to the Alleghanian front, respectively, whereas the 015° and 090° sets formed sub-perpendicular and sub-parallel to the Ouachitan front, respectively. Subsequently, bed-parallel veins formed during thrusting at 110 $^{\circ}\text{C}$ and 3.6 km burial depth, indicating that local thrust-wedge thickness did not exceed the original maximum burial thickness.

Isotopic compositions from interbedded limestone and authigenic carbonate in calcareous shale show a parallel decrease in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ with increasing depth in the Nolichucky Shale, indicating extensive water-rock interaction between these lithologies during authigenic calcite precipitation. Prior to fracturing, fluid isotopic compositions, and carbonate minerals precipitated from them, were dominated by local rock compositions, reflecting a stratified fluid system and lateral fluid flow. In contrast, isotopic compositions from calcite veins in the Nolichucky Shale have a fairly constant composition with depth. Bed-normal fracturing enabled cross-formational fluid flow along open fractures, vertical mixing of vein-forming fluids, and possible extraformational introduction of Na-Ca-Cl brines into the Nolichucky Shale. During calcite vein precipitation, there was little isotopic exchange between fluids and adjacent wall-rocks resulting in vein calcites that are not in isotopic equilibrium with the adjacent wall-rocks.

Cross-formational fluid flow along open fractures occurred before local Alleghanian thrusting because the vein calcites are deformed by Alleghanian twinning and are offset by bed-parallel calcite veins that have Alleghanian displacements. Pressure-corrected homogenization temperatures for the bed-normal and bed-parallel calcite veins indicate that they formed at near maximum sedimentary burial depth of 3.6 km in advance of thrusting. These veins lack evidence for inclusion deformation caused by increased burial, which suggests that thrusting did not subject the Nolichucky Shale to a maximum burial depth in excess of 3.6 km.

Homogenization temperatures of primary fluid inclusions in a single bed-oblique calcite vein average 157 °C and pressure-corrected temperatures exceed 200 °C. These temperatures are significantly greater than maximum burial temperatures of 110-150 °C calculated for the Nolichucky Shale. These elevated temperatures indicate that during later thrusting, fluids hotter than the ambient rock temperature migrated through the Nolichucky Shale. However, these hotter fluids were unable to re-equilibrate older, lower temperature fluid inclusions or reset time-temperature indicators including conodont color alteration

indices, illite crystallinity values, and clay-mineral assemblages. These hot fluids may reflect large-scale fluid migration out of the Alleghanian hinterland driven by tectonic overthrusting.

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CHAPTER 1

INTRODUCTION

OBJECTIVES AND ORGANIZATION

The Nolichucky Shale is an ideal unit for diagenetic studies of mixed carbonate-clastic sequences. In the eastern Tennessee study area, the limestone/shale ratio varies across strike from 0.30 in the Whiteoak Mountain thrust sheet to approximately 0.70 in the Copper Creek thrust sheet to the southeast. Though the overall lithologic makeup of the Nolichucky Shale is different in these two thrust sheets, similar lithologies occur in both. Oolitic limestone present in both thrust sheets can be examined for petrographic and geochemical differences which may be related to the degree of water-rock interaction between the interbedded limestone and shale during burial diagenesis.

The following hypotheses concerning the diagenesis and nature of water/rock interaction during burial in the Nolichucky Shale (Upper Cambrian), east Tennessee were tested in this dissertation. Hypothesis statements are in italics and the chapter in which each is addressed in parentheses.

1 (Chapter 2). *Environmental conditions associated with the deposition of oolitic limestone in the Oak Ridge area (Whiteoak Mountain thrust sheet) were significantly different than those associated with the deposition of previously studied oolitic limestone from the Interstate 75 outcrop (Copper Creek thrust sheet), 10 miles north of Knoxville. The greater proportion of shale and intraclastic limestone in drill core from the Whiteoak Mountain thrust sheet would suggest that water depths in this thrust sheet were significantly greater than in the Copper Creek thrust sheet.* If this hypothesis is correct,

then the ramp to shallow intracratonic basin model of Markello and Read (1981, 1982) for Nolichucky Shale deposition in Tennessee must be reevaluated. In addition, differences in depositional setting between the Whiteoak Mountain and Copper Creek thrust sheets must be identified prior to resolving differences in Nolichucky diagenetic fabrics between the two thrust sheets. Detailed examination of drill core and outcrop samples along with subsurface lithologic correlation of units in the Nolichucky were undertaken to evaluate this hypothesis.

2. (Chapter 3). *The variety of diagenetic fabrics seen in ooids in the Nolichucky Shale are the result of different original mineralogies.* Sandberg (1983) suggests that during the Cambrian the preferred abiotic carbonate mineralogy changed from aragonite and high-magnesium calcite (HMC) to calcite with generally lower magnesium. The cross-over from aragonite and HMC (pre-early to middle Cambrian) to calcite-facilitating seas (middle to late Cambrian) may have occurred as early as the earliest middle Cambrian or as late as the latest late Cambrian (Sandberg, 1983). Since the exact timing of the Cambrian shift is uncertain, it is possible that all three varieties of abiotic marine carbonate were present in the Nolichucky Shale (Dresbachian) oolitic limestone prior to burial. Petrographic (conventional and cathodoluminescent microscopy) and geochemical techniques (trace elements and stable isotopes) were used to infer the original mineralogy of Nolichucky ooids.

3. (Chapter 3). *Coarse radial ooids (sparry radial) represent greater diagenetic alteration of originally fibrous radial high-magnesium calcite ooids; this alteration occurred during the stabilization of HMC to diagenetic low-magnesium calcite and was accompanied by an uptake of iron and manganese.* The radial arrangement of calcite in Nolichucky ooids is preserved in the majority of ooids, however, the general shape and width of crystals is variable resulting in three varieties of radial ooids: fibrous radial ooids, fibrous radial/sparry radial ooids, and sparry radial ooids. Of these three ooid types, only fibrous radial ooids are known to occur in ooid-forming marine environments suggesting the other

fabrics are the result of diagenetic modification. Petrographic and geochemical evidence is presented in Chapter 3 to investigate this hypothesis.

4. (Chapter 4). *The majority of bed-normal veins in the Nolichucky Shale on the Oak Ridge Reservation (ORR) represent filled joints formed by removal of overburden in the Mesozoic and Cenozoic eras.* Studies of bed-normal joints in sedimentary rocks from the central Valley and Ridge and the Plateau (Engelder, 1985) indicate that unloading joints formed by the erosion of overlying rocks following the formation of the Alleghanian fold and thrust belt in the late Paleozoic. Determining the timing and formation mechanism of these joints is required before an understanding of ancient and modern fluid flow through fractured rocks can be gained. Field and core studies of joints and veins from the ORR were undertaken to determine the orientation and relative timing of different joints and veins with respect to each other and the regional Alleghanian-age structure. Fluid inclusions in the veins were examined to determine temperature, pressure, and fluid composition during vein formation. The temperature data were integrated into a decompacted burial history curve for the Nolichucky Shale to constrain timing of the veins. Finally, a mechanism for vein formation was considered in terms of the regional stress history of the area.

5. (Chapter 5). *Prior to bed-normal fracturing, the Nolichucky Shale experienced dominantly horizontal fluid flow with little cross-formational flow; subsequent to fracturing, fluid pathways were developed which allowed for vertical mixing and homogenization of fluids in the fracture fluid conduits.* The formation of open fractures in the Nolichucky Shale provided discrete pathways for fluid flow, unlike the tortuous flow paths between grains and cements characteristic of matrix fluid flow. The interbedded limestone and shale which characterizes the Nolichucky Shale may have limited vertical cross-formational fluid flow prior to fracturing. Fluid inclusions and stable isotopic analyses from interbedded limestone and shale carbonate and vein calcite were used to determine if vertical compositional gradients existed which supported this hypothesis.

6. (Chapter 5). *The fluid inclusion and isotopic data from authigenic carbonate in the Nolichucky should reflect water-rock interaction between diagenetic fluids and interbedded shale.* The preponderance of shale in the Nolichucky Shale in the ORR area suggests that fluids expelled from interbeds of this lithology may have been important in the burial history of the Nolichucky, especially during later burial when surface-derived fluid would have less access to the unit. Vein calcite isotopic compositions and fluid inclusion trapping temperatures were used to calculate the equilibrium composition of the vein-forming fluid. This composition was compared to theoretical compositions of fluids in equilibrium with interbedded shale lithologies at temperatures determined from fluid inclusions in vein calcite. This approach is particularly useful for evaluating the importance of fractures as conduits for the circulation of meteoric water to depths of several kilometers in sedimentary basins. Isotope studies have demonstrated that meteoric water may be circulated to great depths in sedimentary basins (Longstaffe, 1987; Rye and Bradbury, 1988).

This dissertation will examine the depositional and diagenetic history of the Nolichucky Shale starting with deposition (Chapter 2), early and intermediate-stage burial diagenesis (Chapter 3), and finally late-stage burial diagenesis (Chapters 4, 5). Particular emphasis has been put on oolitic lithologies as they are one of the most common limestone lithologies in the Nolichucky Shale. Also, there is little question as to the environment of marine ooid formation; this cannot be said of the intraclastic limestone lithologies which are also abundant in the Nolichucky Shale. From a practical standpoint, oolitic limestone is relatively coarse-grained, which aids in petrographic observations and permits microsampling for detailed geochemical characterization.

STUDY AREA AND GENERAL STRATIGRAPHY

The Nolichucky Shale is exposed in northeast-trending outcrop belts in east Tennessee as the result of uplift and erosion following the late Paleozoic Alleghanian Orogeny (Figure 1.1a). This deformation was responsible for the formation of the Valley and Ridge Province in which the Nolichucky Shale occurs. This dissertation focuses on the Nolichucky Shale within Rodgers (1953) central phase of the Conasauga Group in the vicinity of Oak Ridge, Tennessee (Figure 1.1b). The Nolichucky Shale is preserved in the Whiteoak Mountain thrust sheet and the Copper Creek thrust sheet. In these thrust sheets the Conasauga Group consists of alternating carbonate-dominated and shale-dominated members and reflect the interfingering of shale and siltstone lithologies from the west and northwest with carbonate units from the east and southeast (Figure 1.1b). The stratigraphic nomenclature for the Conasauga Group is shown in Figure 1.2.

Though previous studies have investigated the paleoenvironmental aspects of the Nolichucky Shale, in southwest Virginia (Markello, 1979; Markello and Read, 1981, 1982), and east Tennessee (Hasson and Haase, 1988; Weber, 1988), the diagenetic history of the unit has not been considered.

The reason diagenetic studies in the Nolichucky Shale had yet to be undertaken may have been that a fundamental understanding of its depositional history was required before its diagenetic history (which is strongly dependent on depositional setting) could be deciphered. The controls of depositional setting on later diagenesis of the Nolichucky Shale are evaluated in subsequent chapters with appropriate citations to previous work.

Good exposures of the Nolichucky Shale are rare in east Tennessee as a result of its argillaceous nature and ease with which it weathers. However, shallow (<300 m) boreholes through the Nolichucky on the U.S. Department of Energy Oak Ridge Reservation (ORR, Figure 1a) provide nearly continuous cored intervals through the

		Central E. TN	NE. TN; SW. VA; NW of Pulaski Fault	NE. TN; SW. VA; SE. of Pulaski Fault	S. Central TN	NE. AL	NW. GA
Upper Cambrian	Trempealeau	Copper Ridge Dolostone	Copper Ridge Dolostone	Conocho- cheague Limestone	Copper Ridge Dolostone	Copper Ridge Dolostone	Copper Ridge Dolostone
		Maynardville Limestone	Honaker Dolostone	Elbrook Dolostone	Ketona Dolostone	Ketona Dolostone	Ketona Ds.
	Dresbachian	Nolichucky Shale					
		Maryville Limestone			Undiff. Conasauga Group	Undiff. Conasauga Group	Undiff. Conasauga Group
Middle Cambrian	Conasauga Group	Rogersville Shale	Honaker Dolostone	Elbrook Dolostone	Ketona Dolostone	Ketona Dolostone	Ketona Ds.
		Rutledge Limestone					
		Pumpkin Valley Sh.			Undiff. Conasauga Group	Undiff. Conasauga Group	Undiff. Conasauga Group
Lower Cambrian		Rome Fm.	Rome Fm.	Shady Ds.	Absent	Absent	Rome Fm.
		Shady Ds.	Shady Ds.				
		Chilhowee Group	Chilhowee Group	Chilhowee Group			

Figure 1.2. Generalized stratigraphic nomenclature of Cambrian sedimentary rocks in east Tennessee and nearby areas. Figure taken from Weber (1988). Detailed references to nomenclature also provided by Weber (1988).

Nolichucky and other Cambrian and Ordovician units. These cored intervals provide excellent sample material for geochemical studies since surface weathering effects such as carbonate dissolution and oxidation of ferroan carbonates are significantly reduced in the shallow subsurface. Detailed lithologic logs of the Nolichucky Shale in borehole GW134 which provided a complete coverage of the unit, are presented in Appendix A. The lithologies sampled in this borehole and two others (GW139, GW140) were correlated with other boreholes in the Whiteoak Mountain and Copper Creek thrust sheets through the use of natural gamma ray logs. Because cored boreholes have matching geophysical logs, the response of these logs to lithology can be readily correlated by comparing the logs with drill core. This is the preferred method for subsurface correlation (Cant, 1983) and permitted a better understanding of the three dimensional distribution of lithologies in the Nolichucky Shale than could be achieved previously using outcrops alone.

METHODS

Hand samples and 48x76 mm thin sections from outcrops of the Nolichucky Shale were prepared by Weber (1988) as part of his dissertation on the depositional history of the Nolichucky and overlying Maynardville Limestone. These were available for this dissertation in addition to thin sections prepared for this study from 6.4 cm diameter drill core from the Oak Ridge Reservation. All samples were oriented with respect to stratigraphic-up and those selected for fracture study were marked to show strike and dip.

Though cores from the ORR were not oriented in the borehole with respect to magnetic north, the bedding dip direction and a line perpendicular to it could be identified easily in core samples because bedding intersects the borehole axis at an angle of approximately 50 °. The line perpendicular to the dip direction is equivalent to bedding strike and is assumed to be equal to the strike of bedding in outcrop (055 °) in the

Whiteoak Mountain thrust sheet on the ORR. Here outcrop bedding strike is nearly constant, varying by less than ± 5 degrees between sampled wells; outcrop bedding in the Whiteoak Mountain sheet dips to the southeast at an average dip of 50° . Fracture sets identified from drill core using this method have identical orientations to fracture sets measured in outcrop as part of this study, and by Lemiszki and Hatcher (1990).

Samples analyzed for stable isotopes were slabbed with a water-cooled saw and polished to assist in identifying constituents for sampling. Microsamples were obtained with a microscope-mounted microdrill capable of drilling rock constituents about 1 mm in width. The microdrill setup and common acid extraction line used to extract microsamples were built by myself with assistance from K.R. Walker, based on discussions with K.C. Lohmann (University of Michigan). The microsampling facilities are presently maintained by the Carbonate Research Group at the University of Tennessee. Details concerning the stable isotopic and other geochemical techniques used in this dissertation are presented in the chapters that follow.

GENERAL SIGNIFICANCE OF STUDY

The study of water-rock interaction during burial diagenesis is important because the majority of ancient sedimentary rocks have spent most of their post-depositional histories under burial diagenetic conditions (Choquette and James, 1987). Current topics of research with regard to burial diagenesis include recognizing evidence for, and controls of, fluid flow in the deep subsurface, sources of subsurface fluids, controls on subsurface brine compositions, and mass transfer of dissolved constituents between burial fluids and surrounding rocks (Burst, 1969; Cathles and Smith, 1983; Hutcheon, 1983; Foscolos, 1984; Galloway, 1985; Bredehoeft and others, 1988).

The mixed carbonate-siliciclastic nature of the Nolichucky Shale permits detailed study of fluid flow and diagenetic reactions between interbedded carbonate and clastic

lithologies. Though numerous examples from the literature report interactions between interbedded sandstone and shale undergoing clay dehydration and compaction (Magara, 1976; Boles and Franks, 1979; Neglia, 1979; Bredehoeft and others, 1988), few studies consider interbedded *limestone* and shale undergoing dehydration and compaction during burial. Isotopic compositions of carbonate minerals in interbedded limestone and shale can provide information on the mass transfer of carbonate between these lithologies. Because the isotopic composition of oxygen from these two lithologies differs by as much as 10-15 per mil (Hoefs, 1980), contributions from each to authigenic carbonate minerals should be readily distinguishable in the isotopic composition of the authigenic carbonate. Data presented in Chapter 5 of this dissertation demonstrates that the oxygen isotopic composition of authigenic carbonate in shale and calcite veins reflects contributions from an isotopically light silicate reservoir in addition to heavier oxygen from limestone with compositions similar to those in the Nolichucky Shale.

Fluid inclusion data and isotopic compositions of vein calcite in the Nolichucky allows fluid compositions responsible for fracture-occlusion to be estimated and compared to published fluid compositions from other sedimentary basins (e.g. Carpenter and others, 1974; Morton and Land, 1987). These compositions can be used to evaluate burial brine formation mechanisms such as dissolution of evaporites in the subsurface by burial fluids, and membrane filtration through shales (Hanor, 1987). Also, the role of smectite dehydration in providing high fluid pressures necessary for hydraulic fractures can be evaluated by examining the composition of fluids responsible for fracture occlusion. In the case of Nolichucky calcite veins, inclusion fluids are too saline to have been derived solely by removal of clay interlayer-water during clay dehydration. Several studies have demonstrated that fluids released during clay dehydration are relatively fresh (Burst, 1969; Galloway, 1985).

The effects of overthrusting and tectonic fracturing on basin-wide fluid flow and water-rock interaction may be evaluated in the Nolichucky Shale because Alleghanian

thrusting has imparted clearly defined deformation fabrics on Nolichucky lithologies and veins. These fabrics include twinning of vein calcite crystals, polished surfaces in shale units representing zones of localized shear, and slickenline-lineations on bed-parallel calcite veins. These deformation fabrics permit timing of fracture-occlusion and diagenetic mineral growth in surrounding host rocks (both important controls on fluid flow and water-rock interaction) with respect to the Alleghanian Orogeny. The effects of mountain-building associated with thrusting and resultant change in basin-wide fluid flow have been demonstrated in young, sedimentary basins including the Alberta basin (Longstaffe, 1987) and Paris basin (Korotchansky and Mitchell, 1972). In these basins, meteoric waters are able to circulate to depths of 2-3 km driven by elevated topography and increased hydraulic head (Galloway, 1985). Fracture porosity facilitates the downward migration of meteoric waters into these sedimentary basins. In the case of Nolichucky calcite veins, isotopic and fluid inclusion evidence for vein-precipitating fluid compositions can be used to evaluate the importance of meteoric water during vein precipitation. Meteoric waters are characteristically isotopically depleted in oxygen and have low salinities (Galloway, 1985; Sheppard, 1986). If meteoric waters were involved in the precipitation of Nolichucky calcite veins, the relatively heavy oxygen isotopic compositions and high salinities of Nolichucky vein-precipitating fluids, as described in Chapter 5 of this dissertation, requires that meteoric fluids experienced considerable chemical interaction with rocks in the deep subsurface prior to precipitating vein calcites.

REGIONAL SIGNIFICANCE OF STUDY

The relationship of veins to the development of Alleghanian structures is clear from field observations (e.g. Kilsdonk and Wiltschko, 1989; Lemiszki and Hatcher, 1990). However, conditions of vein formation in the southern Appalachians, in terms of depth, pressure, and temperature, have not been reported previous to this study. Pressure,

temperature, and burial depth estimates for calcite veins can be combined with cross-cutting relationships between the veins and other deformation fabrics to constrain environmental conditions in the lithosphere prior to, during, and subsequent to thrusting. This is a relatively recent approach and has been used with success in the Pyrenees (Rye and Bradbury, 1988) and central Appalachians (Srinistava and Engelder, 1990).

Determinations of vein-forming conditions and fluid compositions are possible because of the integrated approach of fluid inclusions and stable isotopes from vein calcites.

Nolichucky calcite veins formed at near-maximum burial depths of 2.4-3.6 km from Na-Ca-chloride brines at temperatures of 80-110 °C. Several veins have formation temperatures greater than maximum ambient temperatures presumably experienced by the Nolichucky and suggest thrusting may have facilitated the upward migration of hotter burial brines into the Nolichucky. The role of thrusting in effecting large-scale migration of sedimentary basin burial fluids and its ability to concentrate economic deposits adjacent to fold and thrust belts is discussed by Oliver (1986).

Previous uses of fluid inclusions for diagenetic studies in the East Tennessee area have been limited primarily to Mississippi Valley-type (MVT) mineralization (Taylor and others, 1983; Kesler and others, 1989). The compositions of East Tennessee MVT mineralizing fluids (Kesler and others, 1989) and Nolichucky vein-precipitating fluids are quite similar in terms of cation/chloride ratios, however, East Tennessee MVT fluids are about 75% less saline than Nolichucky vein-precipitating fluids (see Chapter 5 of this dissertation). The similarities between these fluids has led some researchers to favor a similar age for both East Tennessee MVT mineralization and Nolichucky vein-precipitation (Walker, pers. comm., 1990). However, a similarity in fluid compositions is not an adequate criterion in itself to establish a genetic link between these two events. Data presented in Chapter 4 of this dissertation suggest that vein formation in the Nolichucky Shale from the Whiteoak Mountain thrust sheet is Late Mississippian to Pennsylvanian in age.

The age of fracture formation in the Nolichucky Shale is of interest to researchers studying Recent groundwater flow on the Department of Energy Oak Ridge Reservation (ORR). On the Oak Ridge Reservation, the Nolichucky Shale is one of the most utilized geologic formations for the disposal of hazardous waste material. Though it is generally recognized that open fractures are a significant conduit for groundwater flow, the age and causative mechanisms for these fractures are poorly understood (Dreier, pers. comm., 1989). Some models for fracture-formation can be interpreted to suggest that bed-normal fractures in the Nolichucky Shale are the result of unloading stresses and may be relatively young (Engelder, 1985). Conditions of fracture formation and occlusion presented in Chapter 4 of this dissertation indicate that most filled fractures in the Nolichucky Shale are Paleozoic in age, thereby ruling out the possibility that they formed recently by unloading. The fact that open fractures (joints) occur in outcrop with orientations identical to filled fractures (calcite veins) in drillcore from shallow boreholes suggests that the joints and veins are genetically related. Open fractures in outcrop can form by dissolution of vein calcites in the presence of acidic groundwaters and is a likely mechanism for the fractures seen in Nolichucky Shale outcrops on the ORR.

This study constitutes a part of a much larger project underway at the Carbonates Research Laboratory, University of Tennessee, Knoxville. That project involves the comparison of the complete diagenetic histories of the Cambrian Conasauga Group and the Middle Ordovician Chickamauga Group within a sequence-stratigraphic framework. Similar lithologies are seen in both sequences though original marine carbonate mineralogies may have been different. Sandberg (1983) and Wilkinson and others (1985) suggest that Early and Middle Cambrian oceans precipitated predominantly aragonite and magnesium calcite whereas Ordovician seas were dominated by magnesium calcite precipitation. Fabric and geochemical evidence from Nolichucky ooids presented in Chapter 3 of this dissertation do not demonstrate with certainty that aragonite was a significant component of these marine grains. It is likely that the cross-over from aragonite

to calcite-facilitating seas occurred prior to Nolichucky deposition and may be represented in the underlying Maryville Limestone or other lower Conasauga unit.

CHAPTER 2

SLOPE AND BASINAL CARBONATE DEPOSITION IN THE NOLICHUCKY SHALE
(UPPER CAMBRIAN), EAST TENNESSEE: EFFECT OF CARBONATE
SUPPRESSION BY SILICICLASTIC DEPOSITION ON BASIN-MARGIN
MORPHOLOGY

Study Area

Continuous 6.4 cm diameter core was obtained from shallow boreholes drilled to 300 m below the surface on the U.S. Department of Energy Oak Ridge Reservation (ORR, Figure 1a). The present study is based on detailed examination of core from four boreholes (GW134, GW139, GW140, and Joy-2), cursory study of several other cores, and detailed outcrop examination (Figure 2.1a). The Oak Ridge Reservation contains a portion of the Appalachian foreland Whiteoak Mountain (WOM) and Copper Creek thrust sheets. Strata dip to the southeast at approximately 50 ° (King and Haase, 1987), and range in age from Lower Cambrian (Rome Formation) to Middle Ordovician (Chickamauga Group).

In the Oak Ridge area, there is a minimum displacement on the Copper Creek thrust of 14 km (Lemiszki and Hatcher, 1990). Therefore, the minimum palinspastic separation

Note: This chapter is a paper in press and reflects valuable contributions from the secondary authors: Foreman, J.L., Walker, K.R., Weber, L.J., Driese, S.G., and Dreier, R.B., Slope and basinal carbonate deposition in the Nolichucky Shale (Upper Cambrian), east Tennessee: Effect of carbonate suppression by siliciclastic sedimentation on basin-margin morphology, in Harris, P.M., and Lomando, T.J., eds., Mixed Carbonate-Clastic Sequences-A Core Workshop: Society of Economic Paleontologists and Mineralogists Core Workshop, in press.

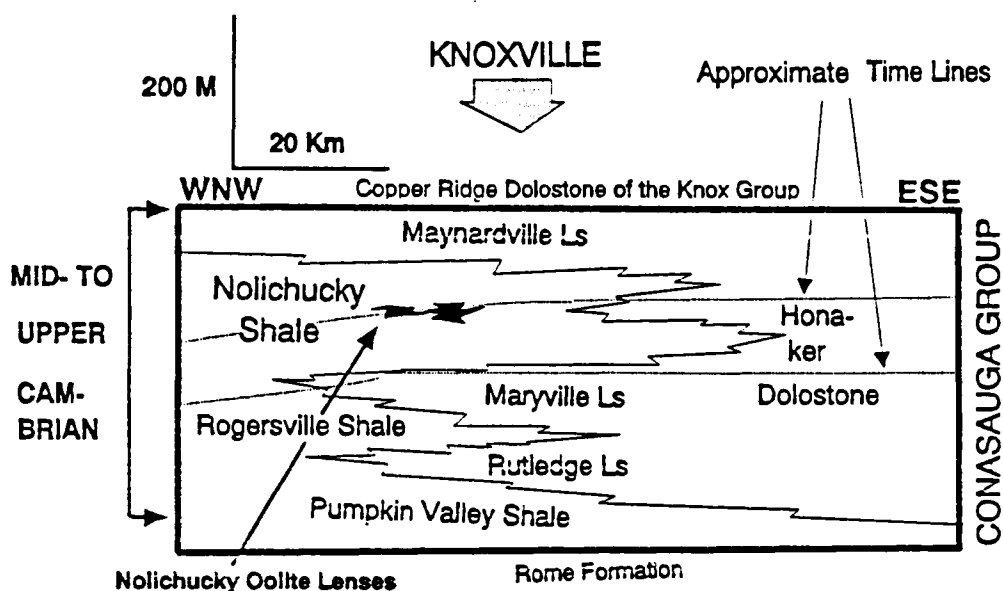
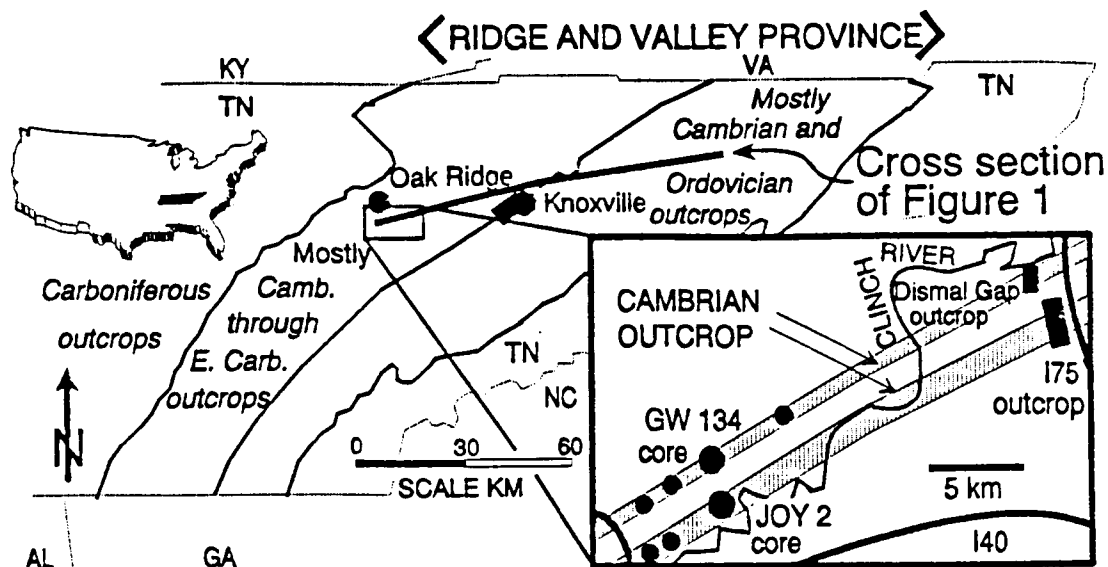


Figure 2.1. Locality and general stratigraphy of study area. A. Map of outcrops (filled rectangles) and cored boreholes (filled circles) examined. Northwest and southeast striped bands in inset figure correspond to Conasauga Group outcrops in the WOM and Copper Creek thrust sheets, respectively. From SW to NE in the WOM sheet, the boreholes are GW137, GW117, GW134, and GW130. B. Highly-generalized cross-section oriented nearly perpendicular to depositional strike for Conasauga Group members in the vicinity of Knoxville, TN. Approximate locations of shallow-water oolite from the Copper Creek thrust sheet indicated. Modified from NW-SE cross-section of Rodgers (1953).

between boreholes in the WOM sheet (GW134, GW139, and GW140) and the Copper Creek sheet (Joy-2) is approximately 20 km. A similar palinspastic separation is expected between the I-75 and Dismal Gap outcrops (Figure 2.1a). In the vicinity of Oak Ridge, structural strike and depositional strike are nearly parallel (Weber, 1988). Consequently, significant facies changes can be observed between the WOM and Copper Creek sheets.

Previous Studies

Rodgers (1953) recognized a northwestern, central, and southeastern phase for the Conasauga Group (Middle to Upper Cambrian) characterized by shale and siltstone, shale and limestone, and dominantly limestone and dolostone, respectively. The study area lies in the central Conasauga phase (Rodgers, 1953; Weber, 1988). Here, southeastward-thickening carbonate units of the Rutledge Limestone, Maryville Limestone, and Maynardville Formation, interfinger with northwestward-thickening siliciclastic units of the Pumpkin Valley Shale, Rogersville Shale, and Nolichucky Shale (Figure 2.1b). The Nolichucky Shale (Dresbachian) is approximately 150 m to 175 m thick in the WOM and Copper Creek thrust sheets, respectively (Weber, 1988). The paleogeography and distribution of Conasauga Group lithofacies in eastern Tennessee has been summarized in very general terms by Hasson and Haase (1988).

In southwest Virginia and northeast Tennessee, the Nolichucky has been described in detail by Markello (1979), and Markello and Read (1981, 1982). In eastern Tennessee, Weber (1988) examined the Nolichucky Shale and Maynardville Limestone, with particular emphasis on outcrop exposures and drill core from the Copper Creek thrust sheet. Similarities in the Nolichucky Shale are recognized in both Virginia and Tennessee. These include: 1) deposition in an intracratonic basin, 2) a east-to-west trend from shallow peritidal limestone to deeper water limestone, and 3) an abundance of storm-generated

features in the deeper water limestone (Markello and Read, 1981, 1982; Weber, 1988). However, significant differences also exist in the sedimentology of the Nolichucky Shale in southwestern Virginia and eastern Tennessee. This study focuses on the most basinward lithologies of the Nolichucky Shale that contrast significantly with the up-slope storm-dominated units of the Nolichucky in the Copper Creek thrust sheet described by Weber (1988).

LITHOFACIES DESCRIPTION AND INTERPRETATION

Gross Vertical Trends

The vertical trends described here are based on observations in core from borehole GW134 in the WOM thrust sheet (Figure 2.1a). Similar trends are also observed in core from other boreholes in the same thrust sheet. In general, intraclastic limestone interbedded with shale and calcareous siltstone are the dominant lithologies in the lower Nolichucky Shale (Figures 2.2, 2.3). Allochthonous oolitic and skeletal packstone and grainstone interbedded with shale are the dominant lithologies in the middle Nolichucky (Figures 2.4, 2.5). Laminated peloidal packstone, lime mudstone, and shale are the most abundant lithologies towards the top of the Nolichucky Shale (Figure 2.6). Thrombolitic limestone overlying oolitic limestone and lime mudstone occur in the uppermost Nolichucky (Figure 2.6). On a formation scale, there is an upward progression towards shallower-water lithofacies, though numerous shallowing-upward (less shale upward) and deepening-upward (more shale upward) sequences are superimposed on this overall trend (Figure 2.7). These shallowing and deepening intervals can be correlated between wells in the WOM sheet (Bear Creek Valley, Figure 2.8) with confidence; correlations between wells in the Copper Creek sheet are more difficult, as are correlations between the two thrust sheets.

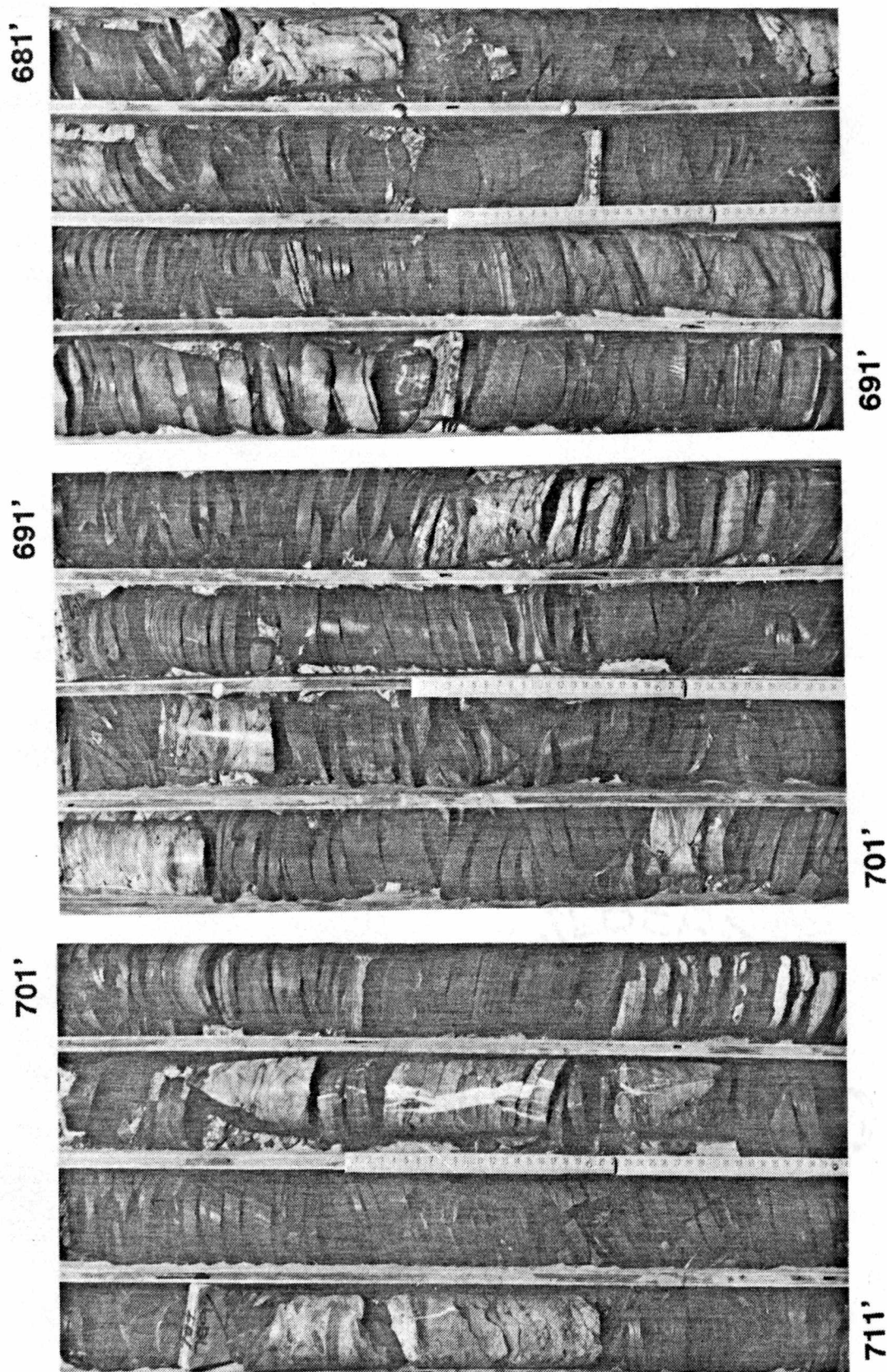


Figure 2.2. Lowermost 30 ft (9 m) of the Nolichucky Shale in 2.5 in. (6.4 cm) diameter core from borehole GW134. Intraclastic limestone conglomerate is dominant carbonate lithology; Calcareous siltstone and very fine-grained skeletal calcarenite occur in 690-688 ft interval. Numbers correspond to core footage below ground surface. In this and subsequent figures, stratigraphic-up is towards top of figure.

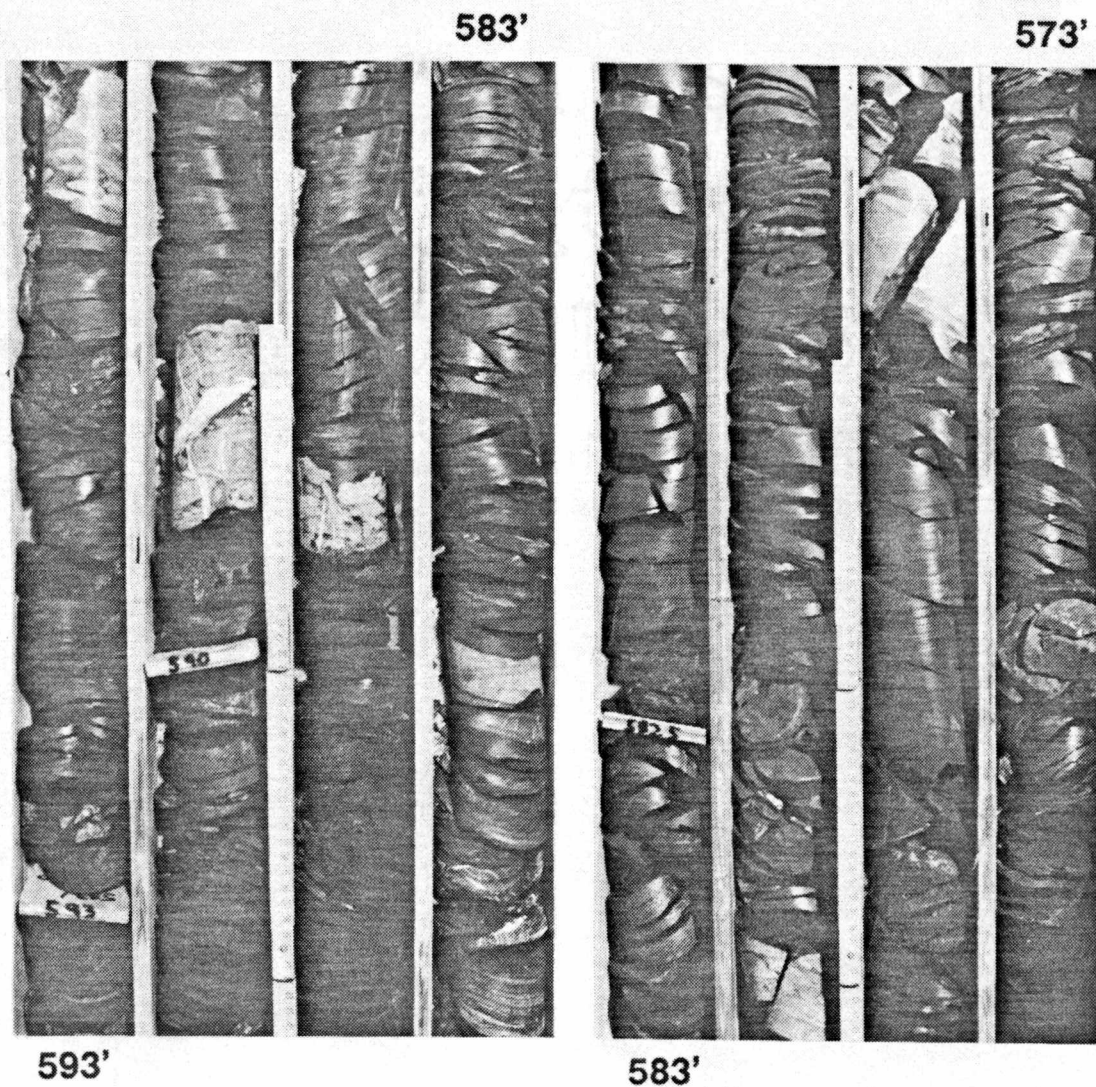


Figure 2.3. Intraclastic limestone debris flow deposits (light-colored intervals) and black shale. Fine-grained calcarenite at 584.5 ft with sharp top and base. Glauconite abundant in skeletal limestone calcarenite at 575.5 ft. GW134, 593-573 ft.

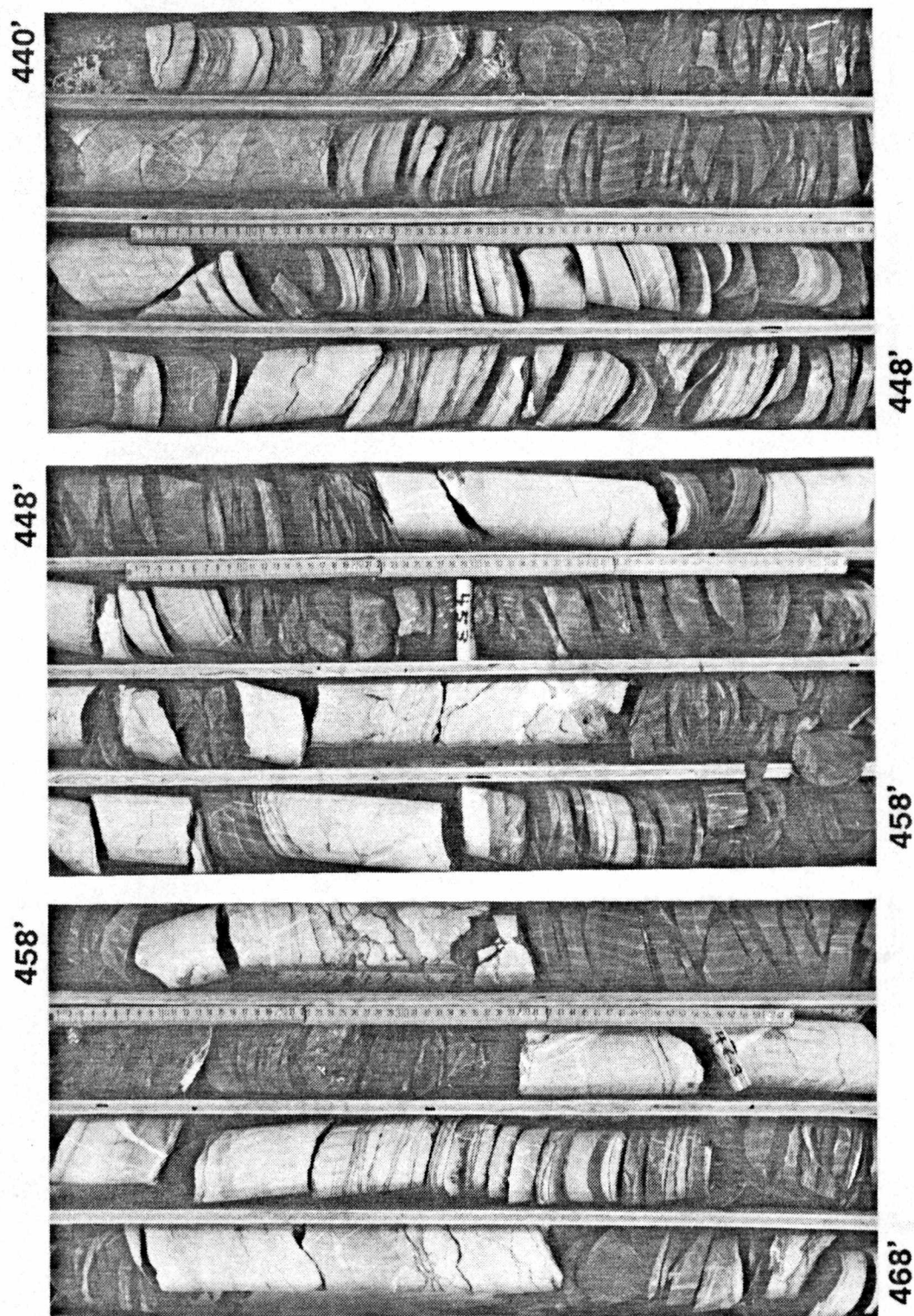


Figure 2.4. Middle Nolichucky Shale characterized by fining-upward sequences of thick oolitic and skeletal limestone overlying calcareous siltstone and skeletal calcarenite. Well developed fining-upward sequences at 464-462 ft. Intraclastic limestone at 458.5 and 454 ft. GW134, 468-440 ft.

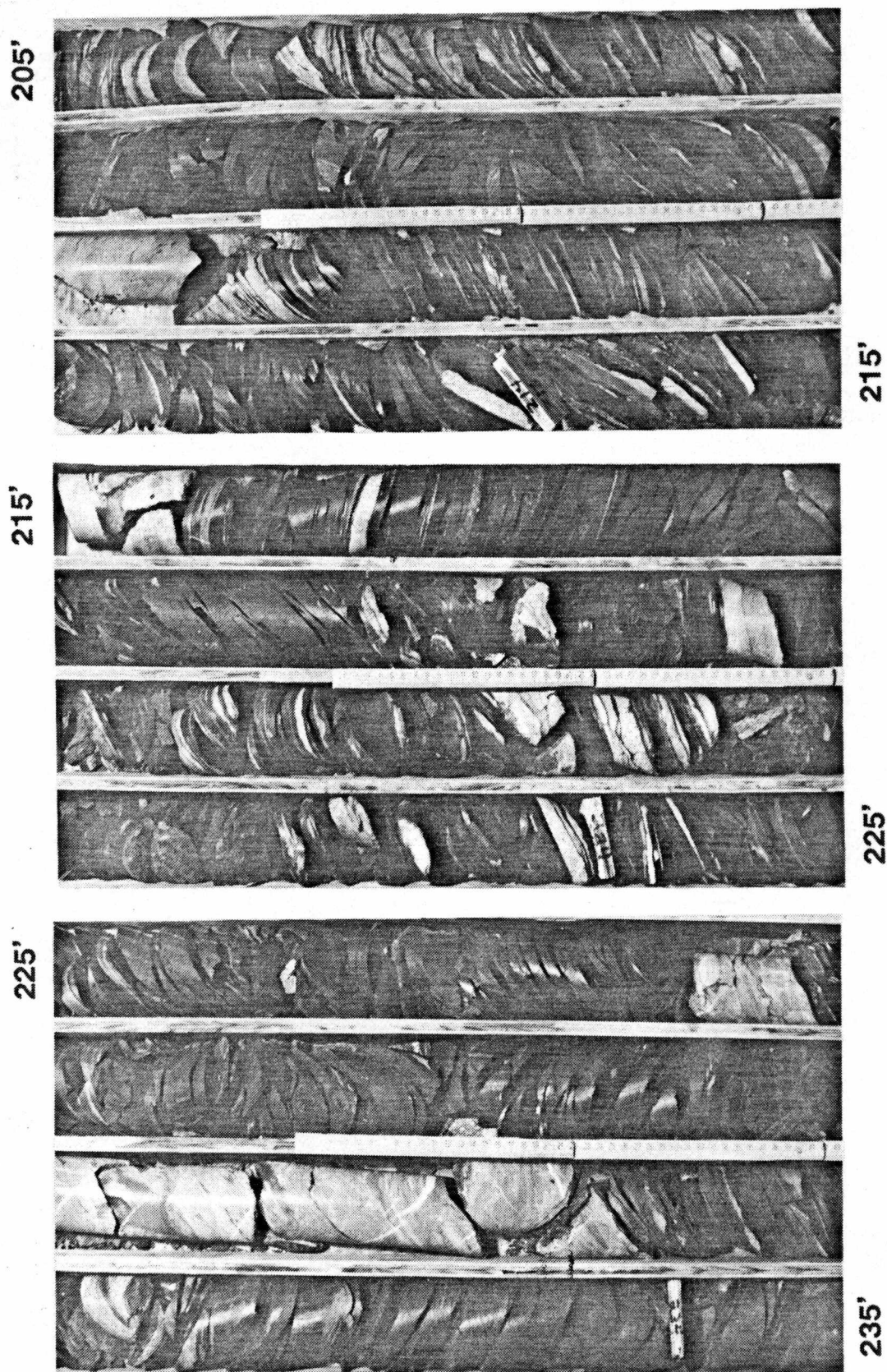


Figure 2.5. Dark brown-black shale with thicker intraclastic limestone and thin skeletal calcarenite beds (light-colored). Intraclastic limestone clasts include lime mudstone and laminated peloidal limestone. GW134, 235-205 ft.

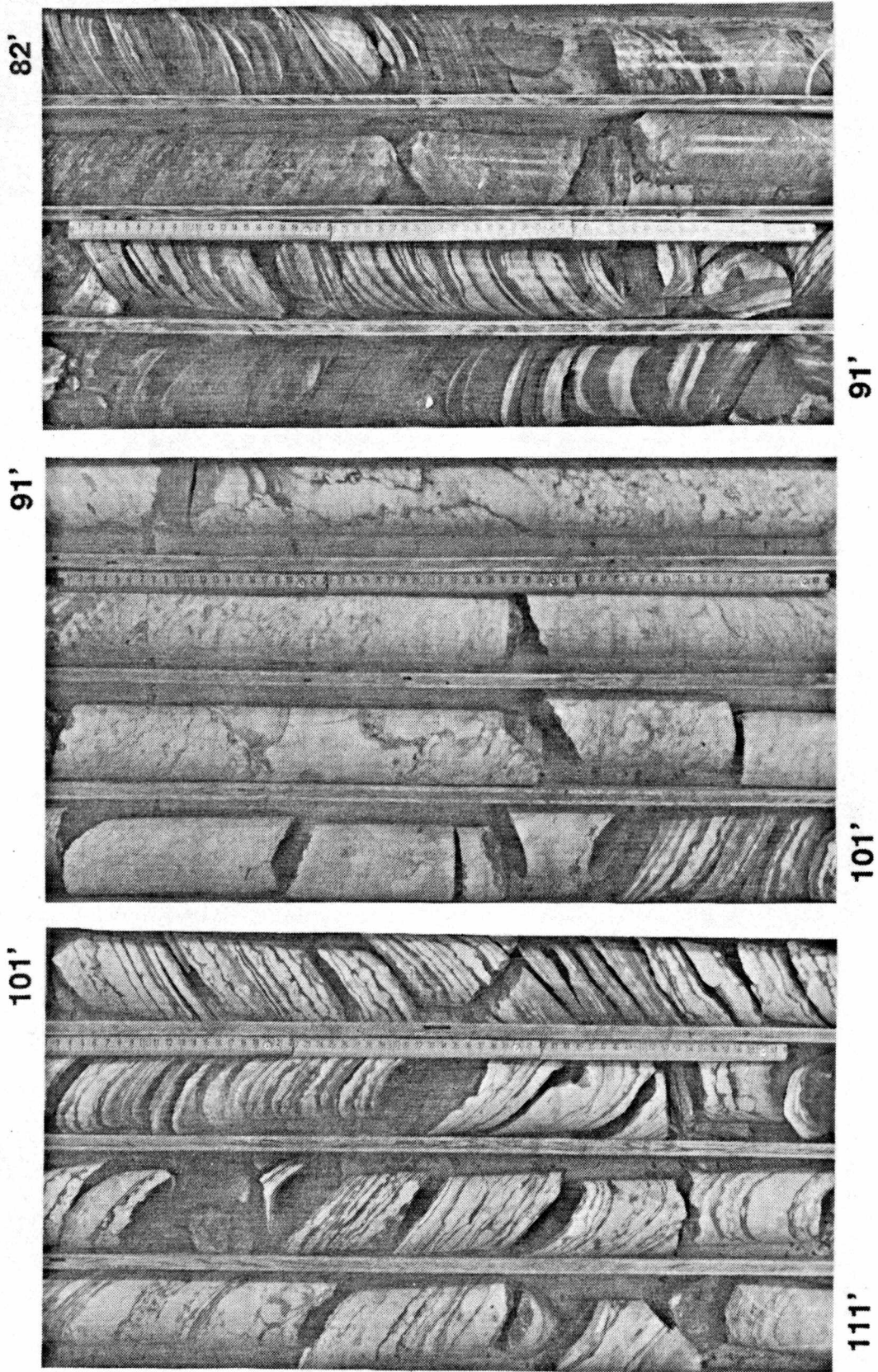


Figure 2.6. Wavy-bedded lime mudstone("ribbon-rock", 111-101 ft) grading up into thrombolitic limestone (100-91 ft). Oncolitic limestone from 87.5-84 ft. Bedding-parallel lime mud-filled burrows common from 107-101.5 ft. GW134, 111-101 ft.

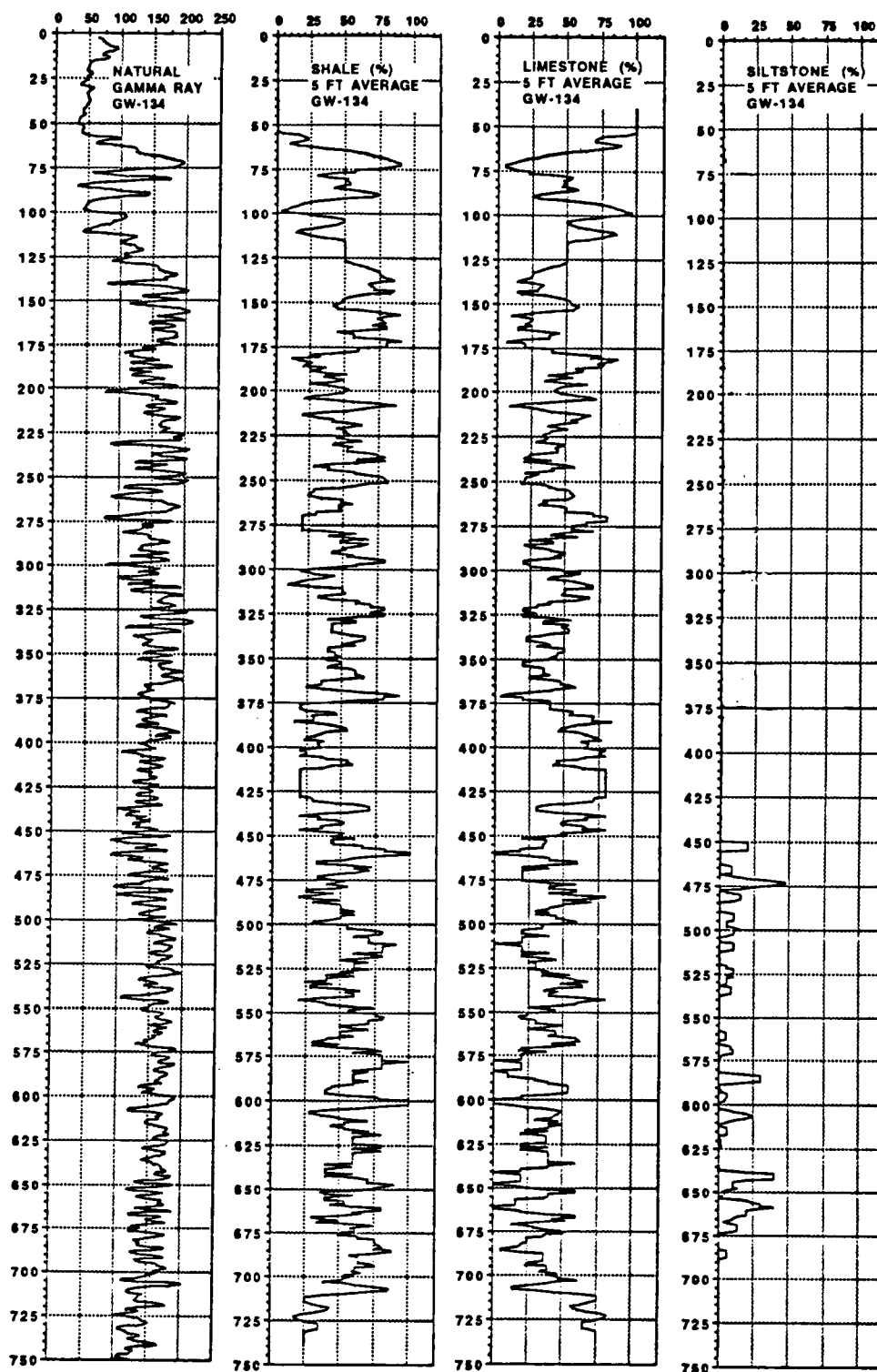


Figure 2.7. Natural gamma ray log for GW134 with corresponding shale, limestone, and siltstone relative abundance from detailed core descriptions. Lithologic logs were produced from semi-quantitative descriptions of core at 1 ft sample intervals. Upper Nolichucky contact at 55 ft; lower contact at 710 ft. Evidence of shallowing-upward sequences seen in gamma ray log at 500-450 ft, 375-300 ft, and 175-50 ft.

NOLICHUCKY SHALE BEAR CREEK VALLEY

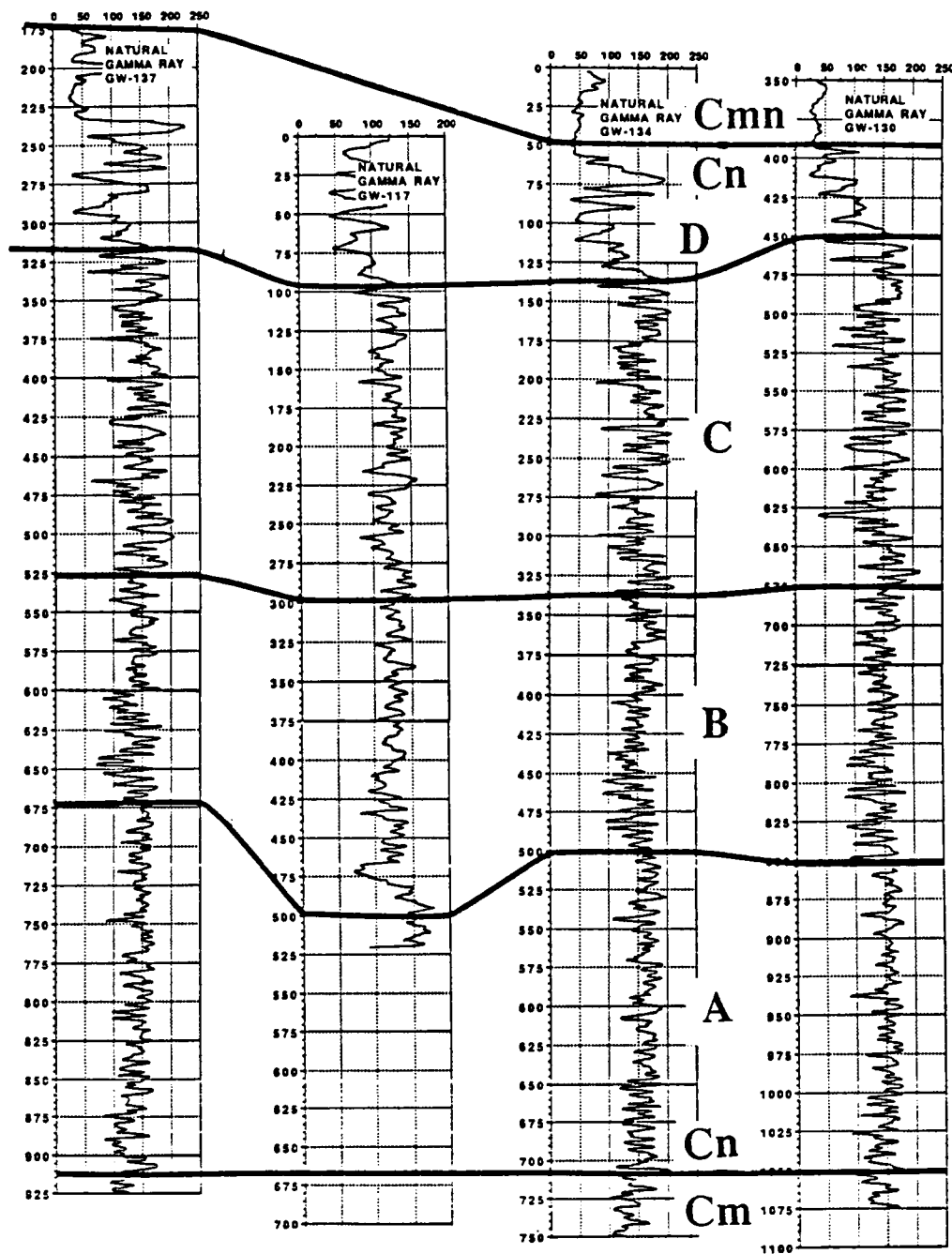


Figure 2.8. Natural gamma ray log correlations in the Nolichucky for boreholes in WOM thrust sheet. A,B,C, and D interval correlations based on baseline gamma ray shifts and correlation of single beds. Southwest (GW137) to northeast (GW130) transect is parallel to strike. Borehole locations in Figure 2.1.

Carbonate Lithologies and Interpretation

Intraclastic limestone conglomerate: description.

Intraclastic limestone (flat-pebble) conglomerate is one of the most common Nolichucky limestone lithologies in the GW134 core, especially in the lower third of the unit (709-575 ft, Figure 2.2). Intraclastic limestone beds are typically less than 0.30 m thick and have sharp lower and upper contacts. Some intraclastic limestone beds can be correlated between boreholes over a distance of at least 6 km along strike. Across-strike correlations between the Joy-2 and GW134 boreholes are tenuous, however, several intraclastic limestone intervals in the lower Nolichucky may be correlative in both wells. Clast size-sorting in individual beds is generally lacking, and the fabric can be either matrix-supported (Figures 2.9a, 2.10a) or clast-supported (Figures 2.9b,d). In a single bed, the intraclastic limestone may be clast-supported at the base and matrix-supported at the top of the bed (Figure 2.9c). Shale, fine skeletal debris, peloids, and more rarely ooids, are common constituents of the matrix (Figures 2.9a, 2.11a). Clasts are tabular-shaped and clast orientations range from nearly parallel to bedding (Figures 2.9a,d), to nearly perpendicular to bedding (Figure 2.9b). It is not uncommon to see different clast orientations in the same bed over a short distance. Clasts may project into overlying shale (Figure 2.10b). Imbricated clasts (long axis parallel to transport direction and dipping against the transport direction; Walker, 1979), are observed in some intraclastic limestone (Figures 2.9a,b, 2.10b), however, in core it is commonly difficult to be sure of the maximum clast axis orientation.

Intraclastic limestone may be either monomictic or polymictic. Laminated peloidal packstone, trilobite and echinoderm packstone, oolitic packstone and grainstone, and lime

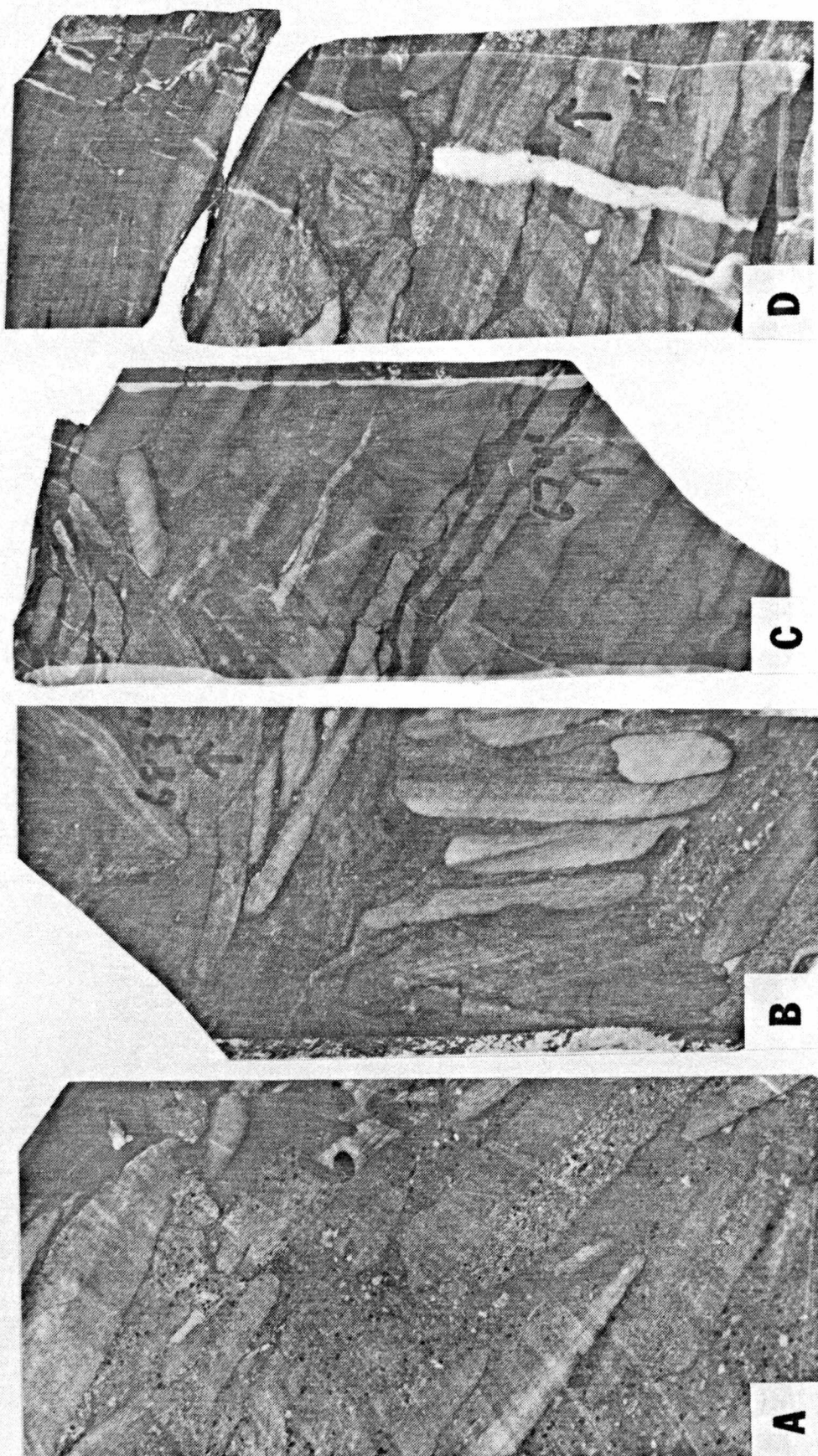


Figure 2.9. Intraclastic limestone in slabbbed core (GW134). A. Matrix-supported conglomerate with peloidal and skeletal limestone clasts. Dark specks are glauconite. Width of photo-1.75 in., 676 ft. B. Edgewise limestone conglomerate with clasts of lime mudstone (light-colored) and laminated siltstone and laminated peloidal limestone. Width of photo-2.5 in., 683 ft. C. Pseudo-breccia at base consisting of laminated peloidal limestone interbedded with shale, overlain by matrix-supported intraclastic limestone with lime mudstone, peloidal limestone, and calcareous siltstone clasts. width of photo-2.5 in., 674 ft. D. Clast-supported intraclastic limestone consisting of laminated peloidal limestone clasts, overlain by shale with stringers of glauconite. width of photo-1.75 in., 705 ft. Labels on core are positioned at base of core in this and subsequent figures.

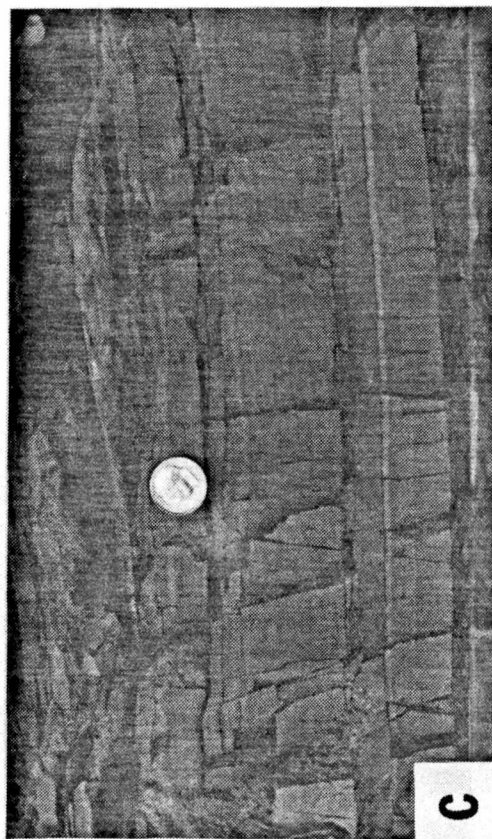
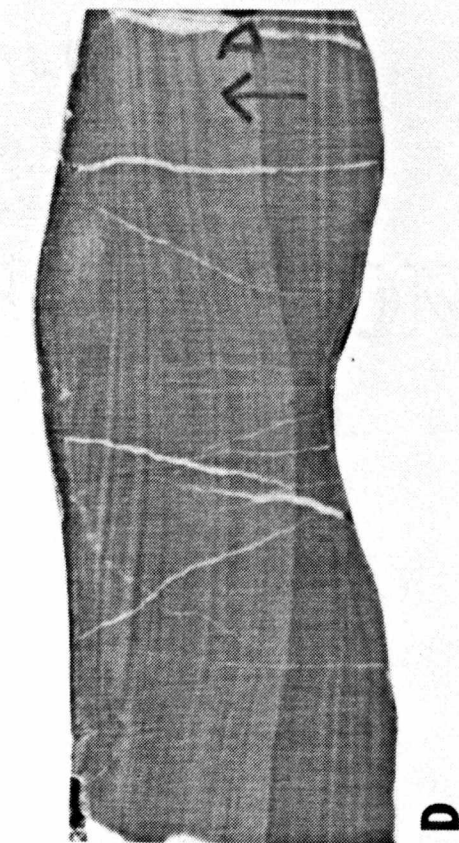
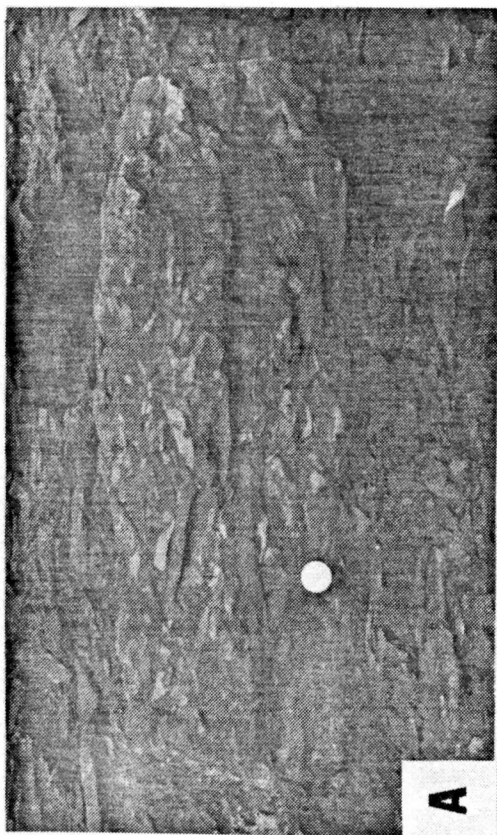


Figure 2.10. Outcrop photos from Dismal Gap, TN (see Fig. 2.1 for exact location). A. Matrix-supported polymictic intraclastic limestone near base of Nolichucky. Clasts include calcareous siltstone, peloidal limestone, and lime mudstone. Dime coin for scale in all photos. B. Intraclastic limestone bed with "fanned clasts" (near coin). Most clasts are calcareous siltstone. C. Micro-hummocky cross-stratified siltstone with climbing ripple above coin. D. Slabbed bed of micro-hummocky cross-stratified siltstone showing scour of underlying parallel-laminated siltstone. Bed is 5 cm thick.

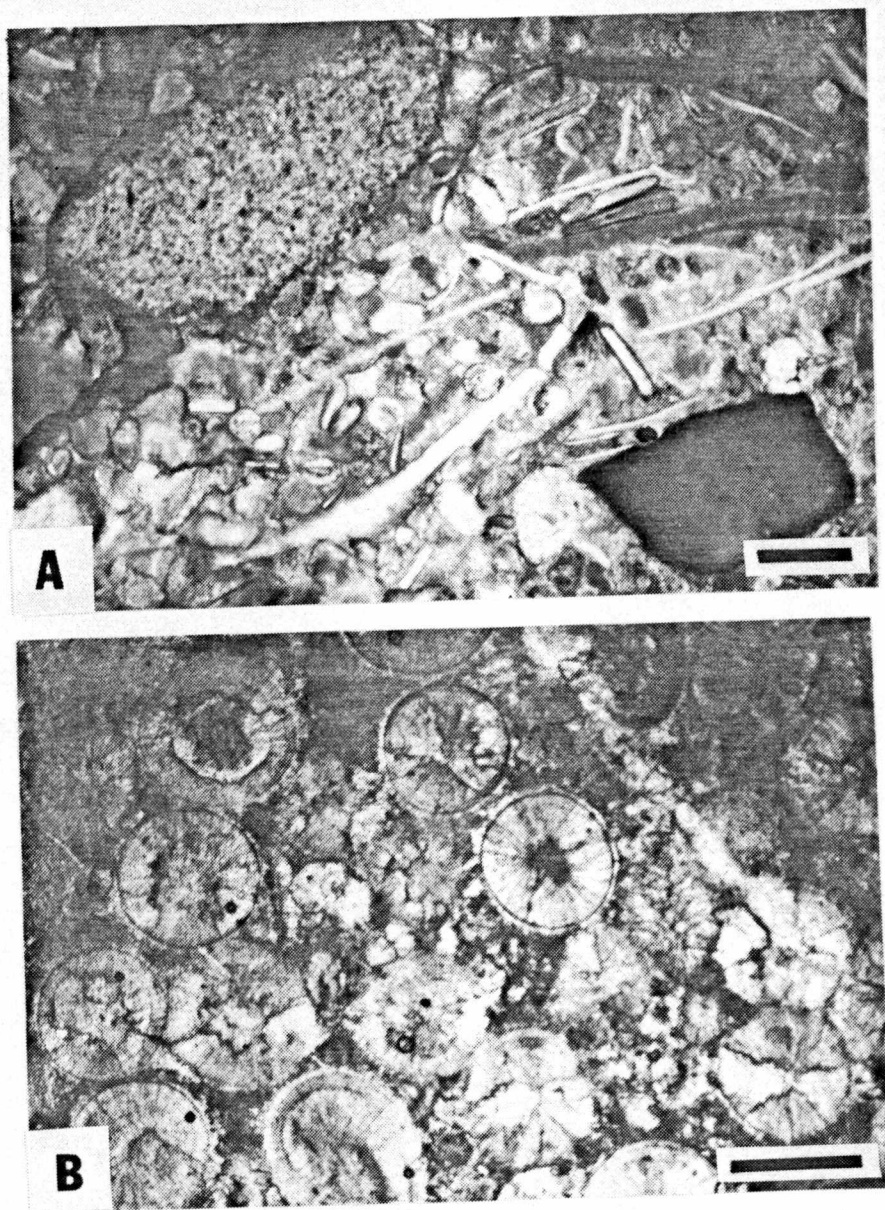


Figure 2.11. Photomicrographs of core from WOM sheet. **A.** Skeletal packstone containing trilobites, ooids, echinoderms, and clasts of peloidal limestone (upper left) and algal? lime mudstone (lower right), 385 ft, GW134. Scale-1 mm. **B.** Oolitic packstone showing radial-fibrous ooids, and peloids in matrix, 1195 ft, GW140. Scale-1 mm.

mudstone are the most common clast varieties. Monomictic clasts are commonly similar in nature to underlying interbeds of limestone or calcareous siltstone, suggesting a local origin (Figure 2.9c). Polymictic limestone conglomerate is most abundant at the base of the Nolichucky Shale with clasts consisting of lime mudstone, fossiliferous and peloidal packstone, and calcareous siltstone. From the base of the Nolichucky (709 ft, GW134) up to approximately 575 ft (GW134), monomictic conglomerate becomes more abundant at the expense of the polymictic conglomerate. Clasts in the monomictic conglomerate are predominantly laminated calcareous siltstone, similar in appearance to underlying siltstone which is interbedded with shale. Lime mudstone intraclasts and fossiliferous and peloidal intraclasts become more common above 575 ft (GW 134). The lithologies making up these clasts also become more abundant above 575 ft. In the lower part of the Nolichucky, intraclastic limestone is commonly underlain and overlain by shale (Figures 2.2, 2.3). Above this interval, intraclastic limestone usually occurs at the top of coarsening-upward sequences (Figure 2.4). Across-strike trends in the nature of intraclastic limestone conglomerate, from the WOM sheet to the Copper Creek sheet (west-to-east), include: 1) beds are thicker to the east, 2) lime mudstone and skeletal and peloidal limestone clasts are more abundant to the east, and 3) calcareous siltstone clasts almost completely disappear to the east.

Intraclastic limestone: interpretation

The intraclastic limestone conglomerate has many characteristics of debris flows. These include: 1) chaotic internal arrangement of clasts, 2) clasts which are commonly matrix-supported, 3) upper surface containing clasts which project into overlying lithologies, and 4) sheet-like to channel morphology. Numerous examples of intraclastic limestone conglomerate have been attributed in the literature to sediment gravity flows

(e.g., Middleton and Hampton, 1973; Cook and Taylor, 1977; Hopkins, 1977; Kozar and others, 1986). Intraclastic limestone is a diagnostic indicator of carbonate slope and basin deposition (McIlreath and James, 1979; Cook and Mullins, 1983; Mullins, 1986). The association of intraclastic limestone with shale, absence of evidence for shallow-water origin (mudcracked lithologies, fenestral limestone, algal limestone), and similarities to published descriptions of intraclastic limestone, favor a slope to basin setting for the limestone conglomerate in the Nolichucky Shale.

Oolitic limestone: description

Oolitic packstone is restricted to the upper two-thirds of the Nolichucky (above 500 ft in GW134, Figure 2.7). Individual beds rarely exceed 0.3m in thickness. This lithology is usually part of coarsening-upward packages (Figure 2.4) and is the coarsest limestone in these packages with the exception of intraclastic limestone. Grading in individual oolitic beds is commonly poorly developed (Figure 2.12a) however, oolitic limestone may fine upward from intraclastic limestone (Figure 2.12b) to ripple-laminated wackestone caps (Figures 2.12c,d). In thin section, oolitic packstone consists of radial-fibrous to radial-concentric ooids in a peloidal matrix (Figure 2.11b). Ooid nuclei are typically peloids and micritized skeletal debris. Trilobite and echinoderm fragments are also typical ooid nuclei. Ooids with detrital quartz nuclei do not occur. This indicates that detrital quartz was absent in shallow-water ooid-forming environments.

Oolitic packstone and grainstone is more abundant, and individual beds are thicker in the Copper Creek thrust sheet (I-75 outcrop) than in cores from the WOM sheet. Two oolitic limestone bodies approximately 6 m thick have been recognized from the middle Nolichucky in this outcrop (Weber, 1988). Herringbone cross-stratification is clearly visible in these bodies. Interparticle porosity is nearly completely occluded by fibrous calcite cement of marine origin (Foreman and Weber, 1989).

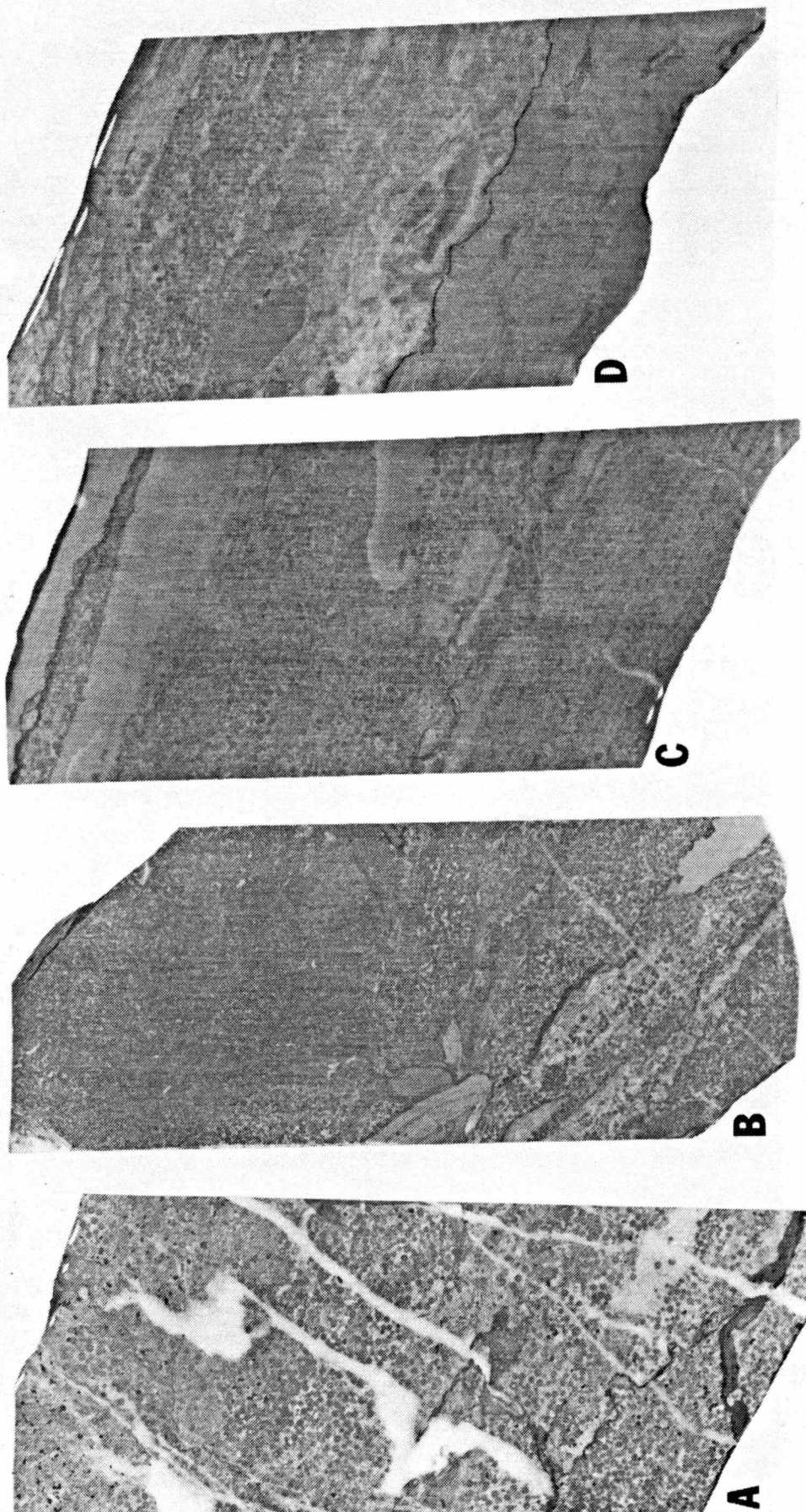


Figure 2.12. Oolitic limestone in slabbled core from WOM sheet. A. Oolitic packstone composed of oolitic limestone clasts and individual ooids. Bed-normal veins cut from lower left to upper right. width of photo-2 in., 440 ft, GW134. B. Oolitic packstone with lime mudstone and oolitic limestone clasts fining-upward into ooid calcarenite. width of photo-2. in., 183 ft, GW139. C. Ooid calcarenite with clasts at base fining-up into ripple-laminated? lime mudstone. Calcarenite scours into underlying laminated ooid-peloidal packstone. width of photo-1.75 in, 1026 ft, GW140. D. Skeletal limestone with intracasts grading up into ooid packstone and lime mudstone (top). width of photo-1.75 in., 1111 ft, GW140.

Oolitic limestone: interpretation

Oolitic limestone lithologies observed in the WOM thrust sheet are interpreted to be allochthonous for the following reasons: 1) beds are less than 0.3 m thick and have sheet-like morphologies, 2) oolitic limestone is associated with slope and basin lithologies (intraclastic limestone and shale), 3) bi-directional cross-stratification, typical of shallow water ooid-forming environments, is absent in the WOM cores, and 4) the oolitic limestone shows internal fining-upward fabric. Oolitic limestone is interpreted to be graded calcarenite, the carbonate-equivalent of a siliciclastic turbidite. Carbonate turbidite lithologies are characterized by: 1) well-defined bedding (Figures 2.4a,b), 2) sharp bases that can be parallel to underlying beds or scour them (Figures 2.12b,c), 3) absence of sole marks and load structures, and 4) presence of all or part of the Bouma sequence (Figures 2.12b,d; McIlreath and James, 1979). Oolitic calcarenite beds commonly exhibit division A (intraclasts and ooids grading up into ooids) and division B (horizontally laminated ooid packstone) of the Bouma sequence; occasionally, division C (ripple laminated ooid wackestone and lime mudstone) cap the calcarenite. Shale overlies the calcarenite and is equivalent to division E of the Bouma sequence (fine-grained sediment deposited by waning turbidity currents and hemipelagic sedimentation. There is a nearly continuous gradation between graded oolitic calcarenite and oolitic debris flow and grain flow deposits (Kozar and others 1986). This gradation is not unusual because many debris flows can be traced down-slope into turbidite lithologies (McIlreath and James, 1979; Walker, 1979). Oolitic calcarenite is interpreted to have formed by turbidity currents that transported up-slope ooid sand sheet deposits into the basin. Ooids in the up-slope sand sheets were swept off of the shallow water oolite shoals by storm currents (Weber, 1988).

Fossiliferous-peloidal limestone: description

Beds of fossiliferous and peloidal packstone and wackestone, present throughout the Nolichucky Shale, range in thickness from 6 cm to less than 1 cm. In general, thicker beds are characterized by sharp tops and bases (Figures 4, 13a,c), whereas thinner beds have sharp bases and diffuse tops (Figures 2.4, 2.11b,d). Internally, grading may be absent or fining-upward (Figures 2.12d, 2.13a-d). Thicker beds in the lower Nolichucky commonly have abundant glauconite. In thin-section, glauconite is seen to replace peloids and less commonly, fossils. Thinner beds are laterally discontinuous in core and have lenticular morphologies; inclined current-ripple laminations can be recognized in many of these beds (Figures 2.4, 2.13b). Fossils include echinoderms, trilobites, and rare phosphatic brachiopods.

Horizontally-laminated peloidal packstone with minor fossil debris is an important constituent in the Nolichucky Shale and comprises one of the most common types of clasts as well. The laminae collectively form many fining-upward sequences in individual beds. In Copper Creek thrust sheet outcrops, this lithology commonly exhibits micro-hummocky cross-stratification (Weber, 1988).

Fossiliferous-peloidal limestone: interpretation

Thicker fossiliferous and peloidal limestone beds have many similarities to the oolitic calcarenite interpreted as carbonate turbidite deposits. The horizontally-laminated peloidal limestone is interpreted to be a calcarenite with only division B of the Bouma sequence represented. Some fossiliferous grainstone beds, consisting of chaotic elongate trilobite and echinoderm fragments, were probably deposited as debris or grain flows. Thin beds of fossiliferous limestone grade upward from a fossil lag into ripple-laminated wackestone and mudstone. This lithology is commonly laterally discontinuous and resembles starved

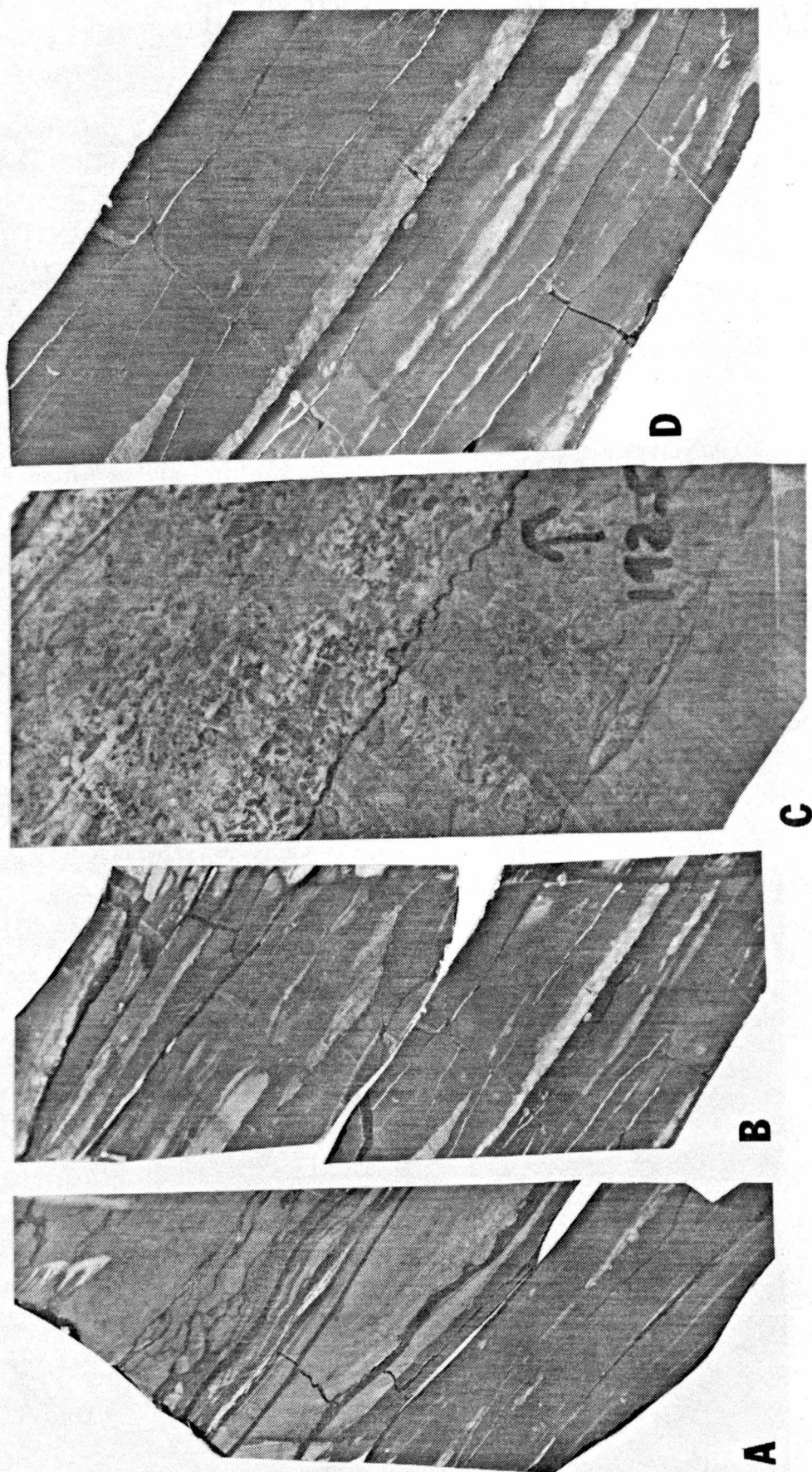


Figure 2.13. Fining-upward skeletal calcarenite beds in shale from slabbed core (GW134). A. Skeletal packstone grading up into lime mudstone. width of photo-2.0 in., 210.5 ft. B. Skeletal packstone grading up into lime mudstone with diffuse tops. Ripple-laminated beds in upper half of photo may be starved ripples. Width of photo-2.0 in., 211 ft. C. Horizontally-laminated skeletal packstone calcarenite with lime mud clasts overlying skeletal wackestone (bottom of photo). Width of photo-1.75 in., 192.5 ft. D. Enlarged area from lower half of Fig. 14b showing diffuse bed tops and horizontal burrow filled with lime mud in shale (center of photo, just under middle skeletal bed). Width of photo-1.75 in.

ripples. Similarities between the oolitic calcarenite and fossiliferous-peloidal calcarenite favor a slope environment for the latter lithology. Graded calcarenite is a common slope lithology; the nature of the constituent grains depends on the type of sediment available in up-slope settings (McIlreath and James, 1979).

Lime mudstone: description

Light to medium gray lime mudstone interbedded with shale is most common in the upper Nolichucky (Figure 2.6). Individual beds are less than 2.5 cm thick and have irregular tops and bases (Figures 2.6a,b). Lime mudstone with interbedded shale exhibits a "ribbon-rock" appearance. The proportion of lime mudstone to shale increases towards the top of the Nolichucky Shale and increases across strike from west to east. Lime mudstone grades up into thrombolitic limestone indicating shallow-water deposition at the top of the Nolichucky (Figure 2.6b). In thin section, lime mudstone consists of micrite and faint outlines resembling peloids. Laminations are rarely observed. Evidence for bioturbation includes burrow mottles, and micrite and fine fossil debris filling bed-parallel burrows (Figure 2.6a). Lime mudstone occurs as discrete beds or as fining-upward caps on fossiliferous peloidal limestone and oolitic limestone (Figures 2.9c,d, 2.13a,b). Soft-sediment deformation features are common in the lime mudstone beds, exemplified by wavy and contorted bedding, and psuedo-breccias formed by injection of shale into semi-rigid lime mudstone layers. These disrupted mudstone layers are likely sources of mudstone clasts seen in some of the intraclastic limestone beds.

Lime mudstone: interpretation

A deep-water origin for intervals of lime mudstone interbedded with shale is favored over a shallow-water lagoonal setting for the following reasons: 1) absence of features

typically associated with lagoonal lime mudstone such as fenestrae and desiccation cracks, and 2) interfingering of this lithology with deep water lithologies such as shale, and fossiliferous-peloidal limestone turbidite beds. An exception to the second line of evidence is present at the top of the Nolichucky where lime mudstone grades into thrombolitic limestone; the thrombolitic limestone is one of the most shallow-water lithologies observed in the Nolichucky Shale (Weber, 1988).

The lime mudstone interbedded with shale probably represents suspension fallout of fine-grained carbonate derived from shallower waters. Storms were a likely mechanism for putting fine-grained carbonate in suspension and spreading it over a large area (Weber, 1988).

Siliciclastics

Shale: description

In the WOM thrust sheet, shale is the dominant lithology in the Nolichucky (approximate shale/limestone=1.75). Shale intervals range in thickness from approximately 3 m to less than 2.5 cm. Thick, shale intervals can be correlated between wells in the WOM and Copper Creek thrust sheets with a high degree of confidence. Dark grey shale is the most common, however, black, dark green, olive green, and brown shale are also present. The darkest shale occurs in intervals where other lithologies are the least abundant (Figure 2.4). Adjacent to limestone interbeds, shale color changes to dark green over a thickness of several centimeters. The shale maintains a fairly uniform texture over this interval, suggesting the cause of the color alteration may have been diagenetic. The shale has good fissility, especially in outcrop. Horizontal laminae are well preserved and are defined in thin-section by the parallel alignment of clay mineral flakes and less

commonly, intercalated glauconite and quartz-rich layers which are usually less than 1-2 mm thick.

Illite, chlorite, and kaolinite are the dominant clay minerals in the Nolichucky Shale (Lee and others 1987; Weber, 1988). Other minerals include quartz, potassium feldspar, pyrite, glauconite, ferroan calcite, and ferroan dolomite; organic matter is less than 2.0 wt% (Lee and others, 1987). Dolomite, the dominant carbonate mineral in the shale, occurs as euhedral rhombs with iron-depleted cores and iron-rich rims. The most common fossil debris observed in the shale consists of phosphatic inarticulate brachiopods and trilobite fragments.

Shale: interpretation

The association of shale with allochthonous deep-water lithologies supports a deep-water origin for the shale in the Nolichucky. Features in the shale that suggest deposition under disaerobic or anaerobic conditions include: 1) lack of bioturbation during deposition of the shale, 2) presence of pyrite indicating low Eh conditions (Berner, 1980), and 3) occurrence of glauconite, also indicating Eh conditions which "are not particularly oxidizing" (Odin and Matter, 1981). In the WOM, shale represents the background sedimentation with deposition of fine-grained material by hemipelagic fallout. This sedimentation was interrupted by allochthonous carbonate debris flows (intraclastic limestone) and carbonate turbidites (oolitic and fossiliferous-peloidal limestone). The rapid deposition of these allochthonous carbonates commonly resulted in soft-sediment deformation. Contorted bedding, in-place pseudo-breccias, and shale injected into overlying carbonate beds are examples of rapid loading features in the Nolichucky. Exceedingly slow deposition and perhaps current winnowing, were responsible for the concentration of glauconite in thin layers in the shale.

Calcareous siltstone: description

Calcareous siltstone is most common in the lower third of the Nolichucky Shale (Figure 2.7). Laterally, calcareous siltstone is absent in the Copper Creek thrust sheet, and fairly common throughout the lower half of the Nolichucky at Dismal Gap (WOM thrust sheet, Figure 2.1a). Calcareous siltstone beds are thicker and more abundant at Dismal Gap than in any of the drill core from the ORR. Calcareous siltstone becomes more abundant to the northwest across strike (Rodgers, 1953). In core, beds are usually less than 10 cm thick and have sharp bases and tops (Figure 2.4c). Calcareous siltstone occurs in coarsening-upward sequences where it usually occurs at the base of the sequence. Carbonate turbidite and intraclastic limestone debris flows typically overlie this lithology. In core, this lithology can be easily mistaken for laminated fine-grained peloidal limestone. Internally, calcareous siltstone is finely laminated; phosphatic brachiopod fragments and coarse mica flakes are interlayered with angular quartz silt and feldspar. These interlayers define fining-upward packages, are laterally continuous, and resemble micro-hummocky cross-stratification (MHCS) in outcrop (Figures 2.10c,d). The calcareous siltstone beds have erosive bases and commonly have current-rippled tops (Figures 2.10c-d). Bed-parallel burrows, filled with calcareous siltstone, are common at the base of many siltstone beds. These burrows do not penetrate more than 1-2 mm into the underlying shale.

The carbonate component of the siltstone reflects detrital and diagenetic contributions. Microspar between silt grains may indicate recrystallization of detrital lime mud. Occasionally, coarser carbonate resembling peloids was observed. However, most of the carbonate is diagenetic. Euhedral ferroan dolomite, similar to that observed in the shale, is the most abundant carbonate phase. Poikilitic ferroan calcite is less common.

Calcareous siltstone: interpretation

The internal horizontal laminae in the calcareous siltstone have the morphology characteristic of micro-hummocky cross-stratification (MHCS) (Figure 2.10d). Clearly, the number of different mechanisms proposed for MHCS suggests that they are multi-genetic. A discussion of the different models for this sedimentary structure is not appropriate for this paper. However, Weber (1988) has interpreted laminated peloidal limestone in the Nolichucky from Copper Creek thrust sheet outcrops, containing MHCS, to represent storm-related deposition. Similar structures in the calcareous siltstone may suggest these are also storm deposits. However, current-ripples at the top of many calcareous siltstone beds (Figures 2.10c,d), indicates a dominance of uni-directional flow during the formation of these ripples. The laterally continuous nature of the laminae in the siltstones (Figure 2.10c) and absence of shale partings in the laminated siltstone may imply that deposition occurred during a single event. The ripples at the top of many siltstone beds suggests that some of this deposition occurred under uni-directional flow rather than bi-directional flow. This type of deposition may represent a form of antidune stratification in turbidites (Prave and Duke, 1990). Though hummocky cross-stratification is usually considered an indicator of shoreface and shelf environments (Prave and Duke, 1990), a deep-water origin for the calcareous siltstone micro-hummocky cross-stratified units is favored because of its association with deep-water shale and carbonate turbidites and debris flows. These siltstone units are tentatively classified as incomplete BCE Bouma turbidite deposits.

The lateral trends observed in the calcareous siltstone that indicate a northwest source for the siltstone are the following: 1) siltstone beds become thicker and more abundant to the northwest, 2) absence of siltstone in shallow-water carbonates to the east, and 3)

current-ripple orientations on the tops of some siltstone beds which indicate a southeast-directed current (Figure 2.11c).

DISCUSSION

Slope-to-Basin Depositional Model

Pronounced lithologic differences recognized in the Nolichucky Shale, between outcrop and core in the WOM and Copper Creek sheets indicate that water depths were considerably deeper to the northwest. The most obvious difference is the increased amount of shale and siltstone at the expense of carbonate in the WOM sheet. All of the carbonate lithologies described previously are thinner in the WOM sheet than the Copper Creek sheet to the east. In contrast to several massive oolitic limestone and thrombotic limestone intervals in the Copper Creek sheet, interpreted to be autochthonous shallow-water deposits (Weber, 1988), limestone deposition in the WOM sheet is the result of down-slope movement of shallower-water carbonate by turbidity currents and debris flows. The intraclastic limestone debris flows are thicker to the east (Copper Creek sheet), and clasts are dominated by shallower-water lime mudstone and fossiliferous limestone; calcareous siltstone clasts are absent in intraclastic limestone in the Copper Creek thrust. Calcareous siltstone and shale become more abundant to the northwest, suggesting a northwest source for the siliciclastic sediment.

Estimated Water Depths

With the exception of micro-hummocky cross-stratified siltstone, which may have formed under the influence of storm currents, the majority of Nolichucky deposition in the WOM sheet probably occurred below normal storm wave base. This is indicated by the

absence of oscillation-ripples on the tops of carbonate beds and the sharp contact between the carbonate (and siltstone) interbeds and overlying shale (Kreisa, 1981; Weber, 1988). Bottom waters were disaerobic to anaerobic and rarely mixed with overlying oxygenated waters as indicated by: 1) absence of bioturbation expressed by fissile shale and laminated siltstone and carbonate, 2) abundance of pyrite in the shale and some limestone units, and 3) abundance of glauconite in the shale and limestone lithologies.

Approximate minimum water depth in the WOM sheet can be calculated using a minimum palinspastic separation between up-slope lithologies in the Copper Creek and down-slope lithologies in the WOM sheet. Assuming a palinspastic separation of 20 km and a depositional slope of 1.0° , water depths in the WOM sheet would have been 365 m greater than in the Copper Creek thrust sheet. A depositional slope of 1° is very conservative considering the abundance of intraclastic limestone debris flow deposits and calcarenite turbidite deposits. The presence of glauconite also provides estimates of minimum water depths. Present-day marine sediments contain glauconite at water depths of 50-500 m, however, glauconite is most common on the upper slope and outer shelf between 200-300 m water depth (Odin and Matter, 1981). Clearly, water depths of several hundred meters existed in the Nolichucky basin during deposition.

Contrast with the Nolichucky in southwestern Virginia

Water depths of several hundred meters or more, estimated for Nolichucky lithologies in the vicinity of Oak Ridge, Tennessee, are an order of magnitude greater than published estimates for the Nolichucky in southwestern Virginia (several meters to a few tens of meters; Markello and Read, 1981, 1982). There are some major lithologic differences in the Nolichucky Shale between the two areas. First, a middle limestone member, consisting of shoaling-upward cyclic algal limestone, is present in Virginia

(Markello and Read, 1981, 1982) but absent in Tennessee. Shallowing-upward carbonate sequences are also recognized in the middle Nolichucky in the WOM sheet (Figures 2.8, 2.9). However, these consist of allochthonous peloidal and oolitic calcarenite, and intraclastic limestone, indicating deposition from turbidity currents. The presence of this middle limestone member in the Nolichucky in Virginia forced Markello and Read (1981, 1982) to constrain maximum water depths in the intracratonic basin. Second, polymictic conglomerate deposits composed of upslope lime mudstone and downslope peloidal limestone and calcareous siltstone occur in the Nolichucky in the WOM sheet, particularly at the base of the Nolichucky. This intraclastic limestone indicates considerable mixing of lithologies during transport. In contrast, the majority of conglomeratic limestone in Virginia contains clasts with similar lithologies to the underlying beds; the similarity of clasts to the underlying lithologies indicates limited down-slope transport on extremely low depositional slopes (Markello and Read, 1981). Finally, cyclic carbonate peritidal facies have not been observed from the Nolichucky in eastern Tennessee (Weber, 1988), in contrast to well-developed cyclic carbonate units in Virginia (Markello and Read, 1981; Read, 1989).

Stratigraphic Model for the Conasauga Group and Nolichucky Shale

Based on analysis of ORR cores, and outcrop studies by Simmons (1984), Kozar (1986), Weber (1988), and Srinivasan (in prep.), a preliminary, predictive, depositional model for the Conasauga Group in eastern Tennessee has been developed (Figure 2.14). This model was briefly outlined by Walker and others (1990). The model is based on several key observations concerning the depositional regime and several assumptions about overall depositional patterns.

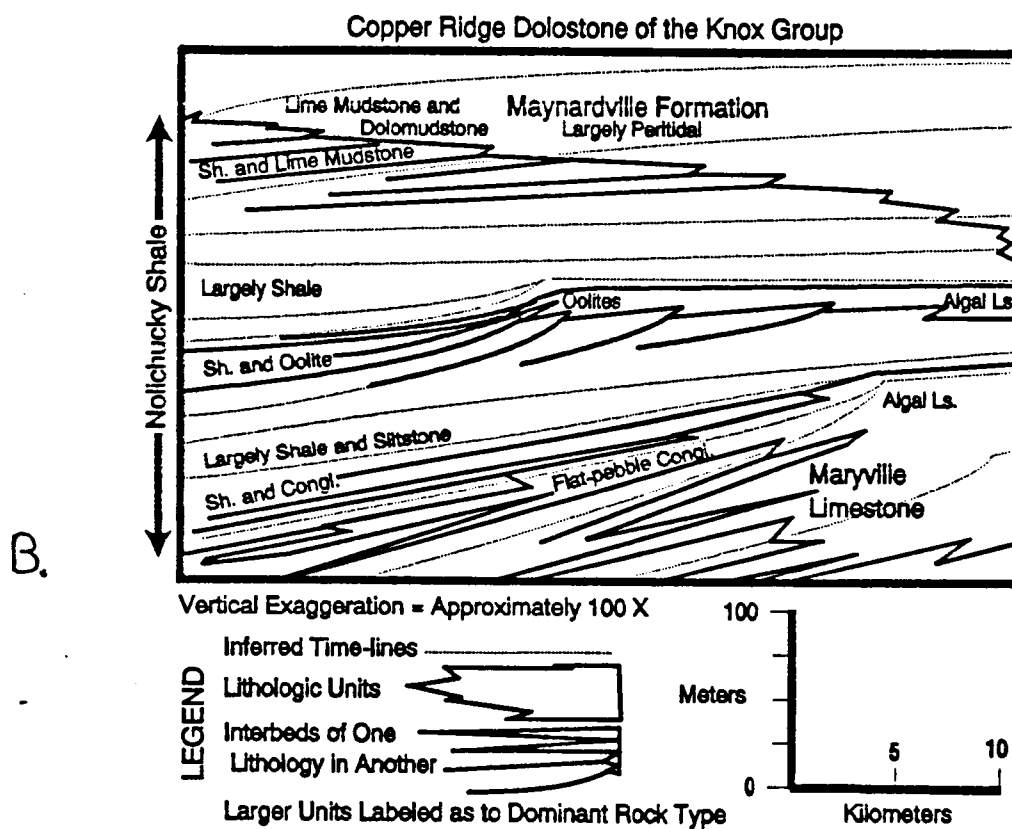
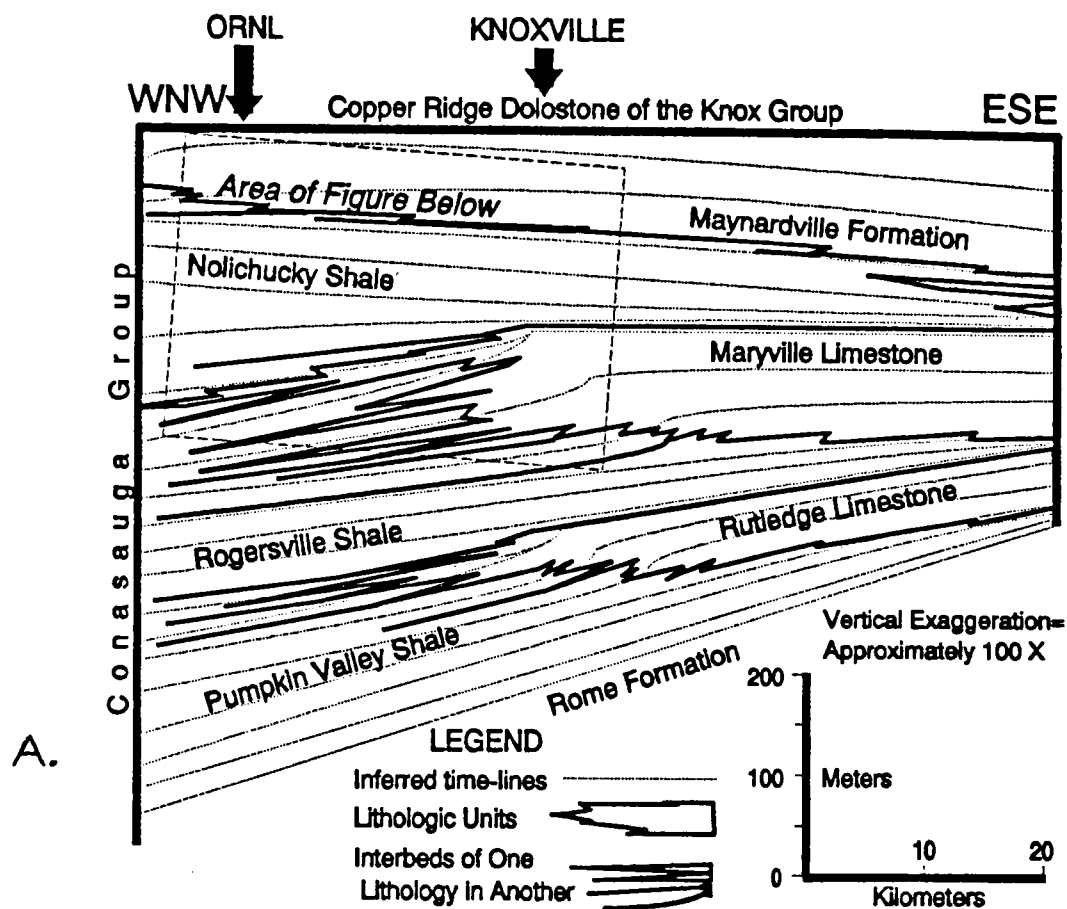
The key observations are as follow:

1. The basin in which the Rogersville and Nolichucky shale formations were deposited varied in water depth from about 100 meters to at least 250 meters (perhaps substantially more).
2. The carbonate units of the Rutledge, Maryville, and the Maynardville limestone represent progradation of shallow-water platform environments basinward.
3. The lower parts of these carbonate units represent deposition on a very gently sloping ramp at times of the shallowest water depths in the basin.
4. The upper parts of these carbonate units, with the exception of the Maynardville, correspond to times of slope steepening, platform marginal algal buildups, and emplacement of debris-flow, flat-pebble conglomerate and/or carbonate turbidites in the basin (Srinivasan, in prep.)
5. Thinner restricted carbonate intervals in the dominantly shale units (e.g., the Nolichucky) represent times of more minor progradations of platform environments (such as oolite shoals) toward the basin, as well as periods of ramp steepening and minor deepening in the basin (see Figure 2.14b).

The several assumptions involved in construction of the model are as follow:

1. Times of encroachment of shale onto the upper ramp and platform represent increase in the input of siliciclastic sediment from the northwestern source region.
2. This shale encroachment led to suppression of carbonate production, a retreat of carbonate environments eastward up-slope, and eventually, establishment of more ramp-like conditions in the area of transition from basin to platform.
3. Times of reduced siliciclastic input led to increased carbonate production in the transition zone, the progradation of shallow water environments toward the basin, and eventually a gradual steepening of the margin (the distally steepened ramp of Read, 1982).

Figure 2.14. Preliminary model of Conasauga Group stratigraphic relationships. **A.** Model for the entire Group. Dotted lines are time-lines inferred from stratigraphic relationships; solid lines enclose formation-level units. Datum is the inferred time-line at the top of the Maryville Limestone which is equivalent to a level about 10 m above the base of the Nolichucky Shale in the ORR area. The part of the succession enlarged in Figure 2.14b is shown by the dashed rectangle. Note gradual progradation of each carbonate unit toward the basin, and the concomitant conversion of the basin-platform transition from gentle ramp to steeper slope. See text for additional details. **B.** Stratigraphic model of the Nolichucky Shale and adjacent units in the ORR area, and the inferred transition to the carbonate platform to the east-southeast. Datum is the inferred time-line near the middle of the Nolichucky. Note the smaller scale platform progradation at the level marked by development of oolite bodies in the transition zones (Copper Creek sheet, I-75 outcrop). Also note how the lithologic succession in the Nolichucky near the ORR (lithologies labeled near left edge of diagram) is in close agreement with the model predictions for the whole Group. See text for further discussion. From Walker and others (1990).



Using these observations, assumptions, and unit thickness trends along a west-southwest to east-northeast transect, a cross-section was constructed from the ORR area to northeast Tennessee (Figure 2.14). The dotted lines are inferred time-lines (paleobathymetric profiles). Although they are based on sedimentological evidence, these time-lines are likely more accurate temporally than are biostratigraphically based correlations. In any event, the latter generally agree with the time relationships shown. The time-lines reflect the alternate deepening and shallowing of the basin corresponding to alternate steepening and flattening, respectively, of the transition to the carbonate platform to the east. The datum for the cross-sectional model for the Group (Figure 2.14a) is the inferred time-line at the top of the Maryville Limestone in the area of transition between platform and basin (equivalent to a level about 5-10 meters above the base of the Nolichucky in the basin area).

Figure 2.14b shows the Nolichucky-equivalent part of the Conasauga Group enlarged with greater lithologic detail shown (this part of the sequence is shown on Figure 2.14a as a dashed rectangle). The datum here is an inferred time-line (paleobathymetric profile) near the middle of the Nolichucky. This figure indicates how the concepts of the model also can be used to explain observations on a smaller scale. Within the shale-dominated parts of the sequence, minor carbonate progradations led at times to establishment of a steeper slope, and emplacement of allochthonous carbonate into the basin. Such a situation existed when the middle Nolichucky was being deposited. Oolite shoals developed on the ramp about 14 km east of the ORR area (I-75 outcrop). Simultaneously, ooid turbidites were emplaced in the latter area by transportation down the steepened slope. The lithologic composition of the Nolichucky in ORR cores is well-explained by logical extension of the larger-scale model.

The controlling influence on the deposition of the entire sequence (Figure 2.14a) was changes in siliciclastic input to the basin from its western side, and the effect of increased

input as a suppressor of carbonate production in the transition from basin to platform on the east. This resulted in the alternate establishment in the transition zone of true ramp-like geometry, and steepened, slope-like geometry. This scenario is the same promulgated by Walker and others (1983), in their "Carbonate Suppression Model" for the development of carbonate to terrigenous-clastic sequences. The control on siliciclastic input is not known, but likely involved sea-level changes. If this is true, suppression of carbonate deposition may have also involved exposure of the platform. This and other hypotheses derived from the model are presently being testing by members of the Carbonate Research Laboratory at the University of Tennessee.

CONCLUSIONS

Interbedded limestone, siltstone, and shale of the Nolichucky Shale are the product of slope and basin deposition. Vertical trends indicate an overall shallowing sequence punctuated by numerous smaller-scale shallowing-upward and deepening-upward sequences. Dark shale, interbedded with intraclastic limestone and calcareous siltstone, are characteristic of the Nolichucky in the Whiteoak Mountain sheet and represent basinal facies. Maximum water depths in the basin probably ranged from 100-300 m. These water depths are an order of magnitude greater than those estimated for the Nolichucky basinal lithologies in southwestern Virginia by Markello and Read (1981, 1982). Slope facies include shale interbedded with calcarenite, consisting of allochthonous shallow-water carbonate, and intraclastic limestone. Upper slope facies are predominantly lime mudstone interbedded with shale, and calcarenite consisting of shallow-water carbonate grains.

Intraclastic limestone debris flow deposits and calcarenite turbidite lithologies in the Nolichucky Shale imply that depositional slopes were steep enough to allow gravity-transport of shallow-water carbonate at least 20 km into the intrashelf basin. Depositional

slopes were probably the greatest during deposition of the lower Nolichucky, evidenced by abundant polymictic intraclastic limestone with up-slope-derived lime mudstone clasts and down-slope-derived laminated peloidal limestone and calcareous siltstone. Storms were an important factor for mixing allochthonous carbonate with hemipelagic shale in the intracratonic basin.

CHAPTER 3

PETROGRAPHIC AND GEOCHEMICAL EVIDENCE FOR MODIFICATION OF ORIGINAL FIBROUS RADIAL HIGH-MAGNESIUM CALCITE OIDS WITH INCREASING WATER-ROCK INTERACTION: NOLICHUCKY SHALE (UPPER CAMBRIAN), EAST TENNESSEE

INTRODUCTION

Purpose

The widespread occurrence of marine ooids in the geologic record makes them attractive subjects for studies ranging from detailed facies analysis and hydrodynamic interpretations, to changes in ocean water chemistry through time. Though mechanisms of marine ooid formation are somewhat controversial, it is clear that the range of environmental conditions for ooid formation are considerably limited. Modern and ancient examples illustrate that well agitated, shallow subtidal settings are required for symmetrical ooid growth (Bathurst, 1975). Oolitic limestone is a common lithology of the Nolichucky Shale (Upper Cambrian) in east Tennessee in association with shallow subtidal peloidal packstone and algal limestone. This oolite occurs as lenticular shoal-like bodies with many features characteristic of *in situ* ooid formation. However, oolitic limestone also occurs as thin calcarenite sheets interbedded with intraclastic limestone and shale, reflecting redeposition in a deep-water basinal setting. One of the objectives of this paper is to compare the depositional and diagenetic fabrics of these contrasting oolitic limestone lithofacies.

Ooids from the Nolichucky Shale exhibit fibrous radial, sparry radial, and monocrystalline cortical fabrics. This paper intends to demonstrate that the majority of the ooid fabrics reflect burial diagenetic alteration of originally high-magnesium calcite (HMC) ooids. Some of the ooid fabrics suggest aragonite and bimineralic precursor mineralogies, but petrographic and geochemical evidence do not support this unequivocally. Finally, this paper will present a model that accounts for differences in the present chemistry of shallow-water ooids versus deep-water resedimented ooids.

Study Area and Previous Work

Oolitic limestone of the Nolichucky Shale (Dresbachian) was examined in outcrop and drill core from the Copper Creek and Whiteoak Mountain thrust sheets in the vicinity of Knoxville and Oak Ridge, Tennessee (Figure 3.1). Nolichucky exposures in outcrop and drill core permitted transects parallel and perpendicular to depositional strike. The petrographic and environmental interpretations presented in this paper are the result of examining nearly 500 slabbed hand samples and approximately 200 48x76 mm thin sections prepared as part of dissertation research by Weber (1988) and Foreman (this dissertation) on depositional and diagenetic aspects of the Nolichucky Shale.

The Nolichucky Shale is one of five members of the Conasauga Group (Middle to Upper Cambrian, Figure 3.2). Rodgers (1953) recognized a western shale phase, a central shale and carbonate phase, and an eastern carbonate phase; the study area lies within Rodgers' central phase. Here, the Nolichucky (150-175 m thick) is characterized by interbedded dark calcareous shale, calcareous siltstone, intraclastic limestone, and oolitic limestone; near Oak Ridge, the Nolichucky contains approximately 35 percent limestone; limestone is the dominant lithology (75 percent) in the vicinity of Knoxville (Weber, 1988; Foreman and others, in press). The Nolichucky Shale is overlain by shallow subtidal, intertidal, and supratidal limestone and dolostone of the Maynardville Limestone (Weber, 1988), and

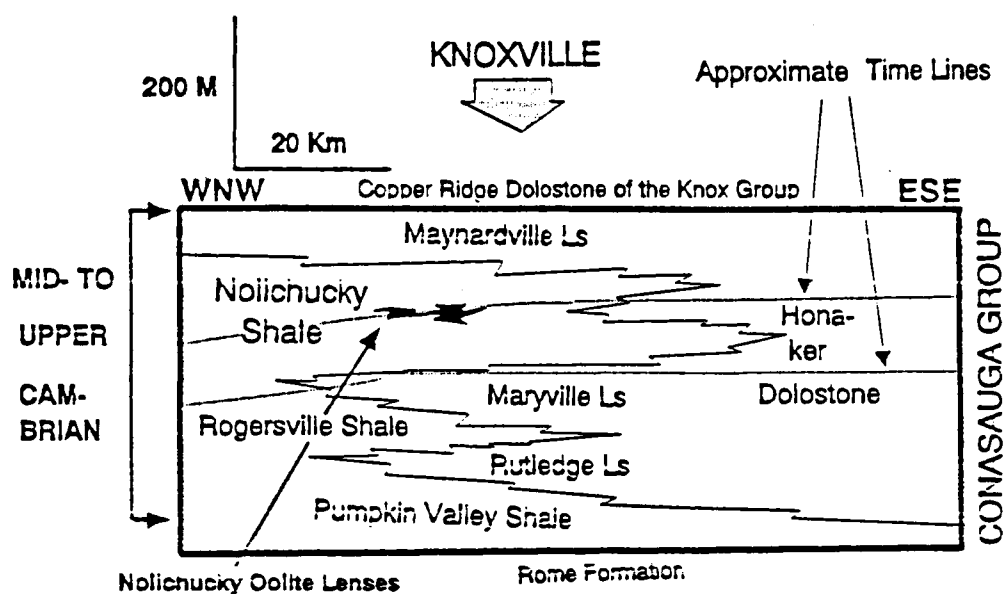
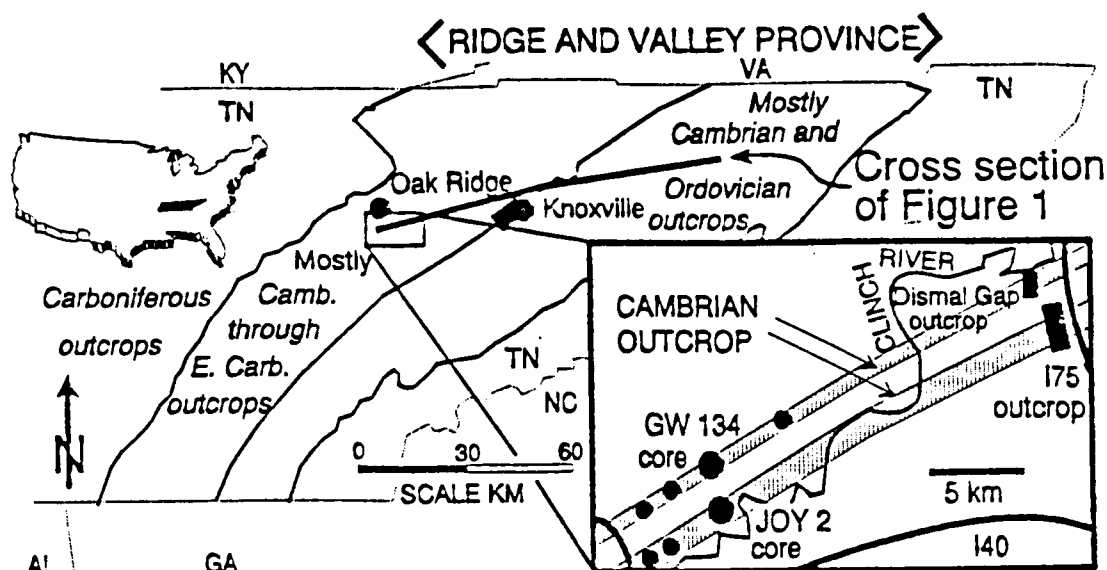


Figure 3.1. Locality and general stratigraphy of study area. A. Map of outcrops (filled rectangles) and cored boreholes (filled circles) examined. Northwest and southeast striped bands in inset figure correspond to Conasauga Group outcrops in the WOM and Copper Creek thrust sheets, respectively. From SW to NE in the WOM sheet, the boreholes are GW137, GW117, GW134, and GW130. B. Highly-generalized cross-section oriented nearly perpendicular to depositional strike for Conasauga Group members in the vicinity of Knoxville, TN. Approximate locations of shallow-water oolite from the Copper Creek thrust sheet indicated. Modified from NW-SE cross-section of Rodgers (1953).

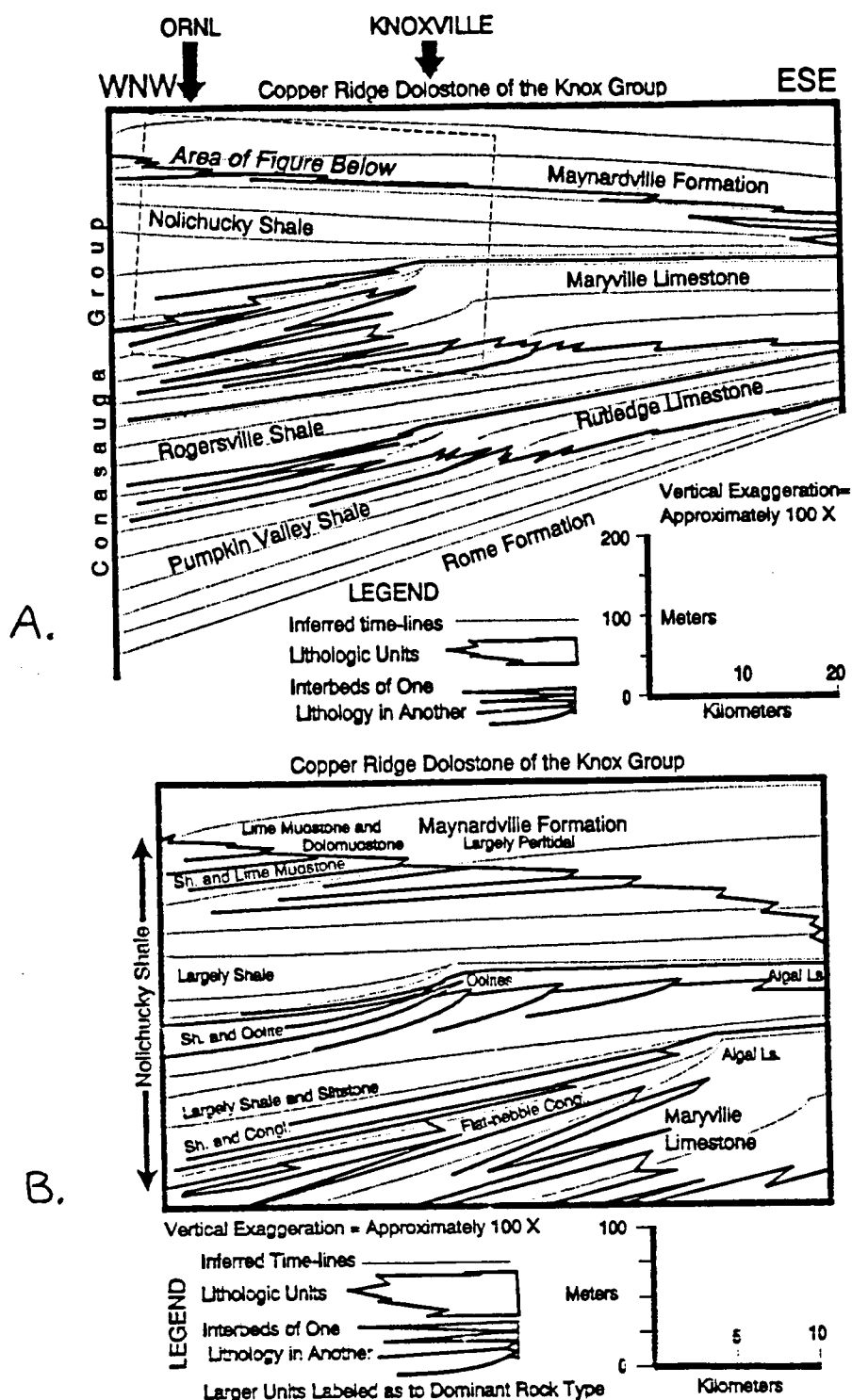


Figure 3.2. Stratigraphy of study area. A. Generalized stratigraphic section for the Cambrian Conasauga Group in east Tennessee showing predominantly carbonate units (Rutledge, Maryville, and Maynardville) extending out into deep-water shale dominated sequence (Pumpkin Valley, Rogersville, and Nolichucky). B. Simplified stratigraphic section for the Nolichucky Shale showing up-ramp and shelf oolitic limestone shoal deposits (Copper Creek thrust sheet, I-75 outcrop) shedding oolitic sands downslope into the basin (Whiteoak Mountain thrust sheet, Oak Ridge area).

overlies subtidal limestone and shale of the Maryville Limestone (Simmons, 1984; Kozar, 1986). Hasson and Haase (1988) discussed general thickness and lithologic trends for the Nolichucky Shale and other Conasauga Group members in east Tennessee.

A carbonate ramp to shallow intracratonic basin depositional model for the Nolichucky Formation in southwest Virginia was presented by Markello and Read (1981, 1982). However, modifications to this model are required in the Oak Ridge area where intraclastic limestone debris flow deposits, carbonate turbidite, and the absence of shallow-water limestone lithologies suggest steeper depositional slopes and deeper water conditions than those hypothesized by Markello and Read (Foreman and others, in press). A sequence stratigraphic model of the Conasauga Group in east Tennessee is presented in Kozar and others (1990; their Figure 1) and Walker and others (1990; reproduced here as Figure 3.2). Detailed environmental studies of oolitic limestone in the Nolichucky Shale by Weber and others (1987), Weber (1988), and Foreman and Weber (1989) recognized subenvironments in these lithologies that are closely analogous to modern ooid-forming environments and adjacent settings. This paper will focus on the deposition and diagenesis of lenticular oolite and adjacent ooid sand-sheets from the Copper Creek thrust sheet and deep-water ooid sand-sheet deposits associated with intraclastic limestone and shale from the Whiteoak Mountain (WOM) thrust sheet. Lithofacies in the two thrust sheets suggests that water depths in the Copper Creek thrust sheet ranged from several meters to several tens of meters (Weber, 1988), whereas depths in the Whiteoak Mountain thrust sheet may have been greater than several hundred meters (Foreman and others, in press).

Methods

Oriented samples from outcrop and drill core were prepared for 48x76 mm thin sections. Samples selected for isotopic analysis were slabbed in water-cooled saws to

prevent contamination from petroleum-based coolants. Thin sections were stained with a mixed stain of Alizaren Red S and potassium ferricyanide following the procedure of Dickson (1965, 1966). Thin sections were examined with conventional transmitted light microscopy and reflected light. Polished thin sections were also examined for cathodoluminescence with a Nuclide Lumniscopescope with a 14 KeV beam at 50 μ A in a helium atmosphere.

Element geochemistry was determined using a Cameca SX-50 automated electron microprobe. All analyses were standardized twice daily to carbonate standards provided by the Smithsonian Institution, described by Jarosewich and others (1983, 1987). Operating conditions for calcite were 30 sec. count time, 15 μ m beam spot, 10 nA beam current, and 15 KeV accelerating voltage. Detection limits in ppm for these conditions at 99 % confidence are Mg (150), Sr (450), Fe (500), and Mn (450). Analytical uncertainty, based on Poissons counting statistics, was determined to be $\pm 10\%$ relative, for Mg, Fe, and Mn. All analyses showed iron and magnesium above detection limits; manganese was below detection limits in less than 5 percent of the analyses and strontium was not detected in any analyses. Several samples were examined for microdolomite inclusions in calcite using back-scatter electron imaging with a 2 μ m step distance on the microprobe and various color image-processing procedures available with the Cameca SX-50.

Carbon and oxygen isotopic compositions were determined from microsamples (1 mg average) drilled with a microscope-mounted microdrill. These samples were extracted on a common acid extraction line at 55 $^{\circ}$ C using 104 % H_3PO_4 . The isotopic composition of the ooids from each sample were determined from 10-20 ooids necessary for each 1 mg sample. Powders were heated to 110 $^{\circ}$ C for 1 hr prior to extraction; no pre-treatment to separate calcite from coarse dolomite was required because this dolomite could be avoided with the microdrill. Of course, microdolomite inclusions could not be avoided with the microdrill. All isotopic analyses were conducted on a VG903 Stable Isotopic Ratio Analyser. In-house calcite standards were run as unknowns with the Nolicucky samples

and run against NBS-20 or in-house standards checked against NBS-20. The average precision for a single day's run was better than 0.2 per mil for both carbon and oxygen; repeat analyses of unknowns conducted eleven months later were off by less than 0.2 per mil for oxygen and 0.5 per mil for carbon.

RESULTS

Lithologic Description-Field Evidence

Oolitic limestone lithologies in the Nolichucky Shale can be subdivided into ooid build-ups and ooid sand-sheets on the basis of field characteristics which include bed thickness, morphology, and type of cross-stratification within beds. The ooid build-ups occur as lenticular bodies generally 5-10 m thick with abundant herringbone cross-stratification. Ooid sand-sheets are typically less than 0.2 m thick planar beds that may remain fairly constant in thickness along an outcrop or pinch out over 10-20 m. Cross-stratification is rarely observed in the ooid sand-sheets. When the two oolitic lithologies occur together, ooid sand-sheets are observed grading laterally into the lenticular ooid build-ups.

Two lenticular ooid buildups are well exposed in the Interstate 75 roadcut (I-75) 10 miles north of Knoxville, within the Copper Creek thrust sheet (Figure 3.1). They occur approximately 75 m below the overlying Maynardville Limestone and occupy a stratigraphic position about midway in the Nolichucky Shale (Figure 3.3). The lower ooid buildup is 6 m thick (Figure 3.3a) and has been examined in detail by Weber and others (1987), Weber (1988), and Foreman and others (1988, 1989). Underlying this buildup are several meters of interbedded shale, coarse intraclastic limestone, and oolitic limestone sand-sheets. The buildup was established on laterally discontinuous intraclastic limestone lobes that grade upward into well-sorted coarse-grained oolitic grainstone.

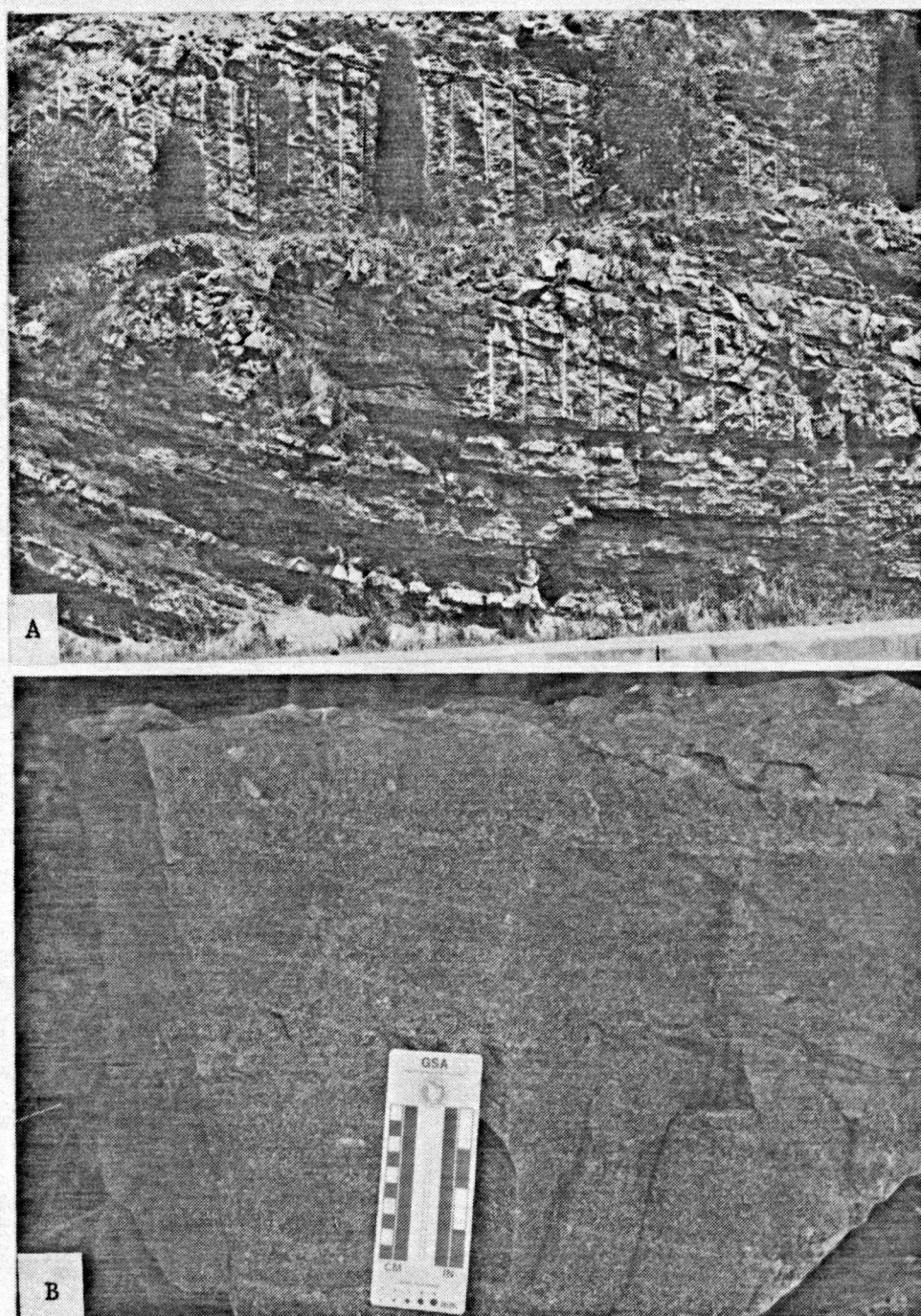


Figure 3.3. Outcrop and slab photos of Nolichucky Shale oolitic limestone. **A.** I-75 outcrop lenticular oolite shoal deposits bounded by shale and intraclastic limestone. Trees in background are about 5 m tall. **B.** Close-up of mobile shoal oolite from I-75 outcrop with well-developed herringbone cross-stratification.

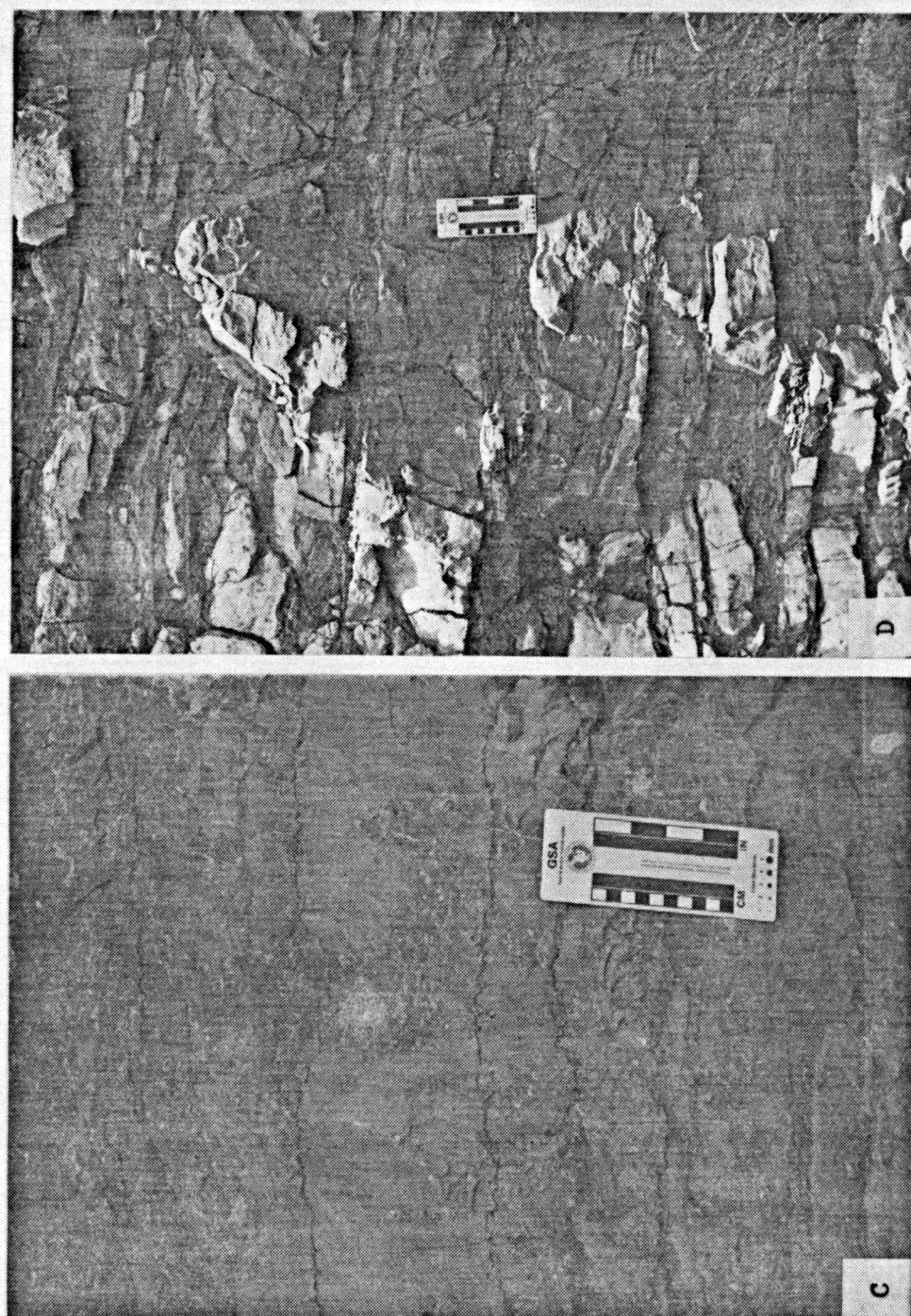


Figure 3.3 (continued). C. I-75 outcrop photo of bed-parallel stylolites in upper part of lenticular ooid shoal deposit. D. I-75 outcrop showing ooid sand-sheet deposits interbedded with shale.

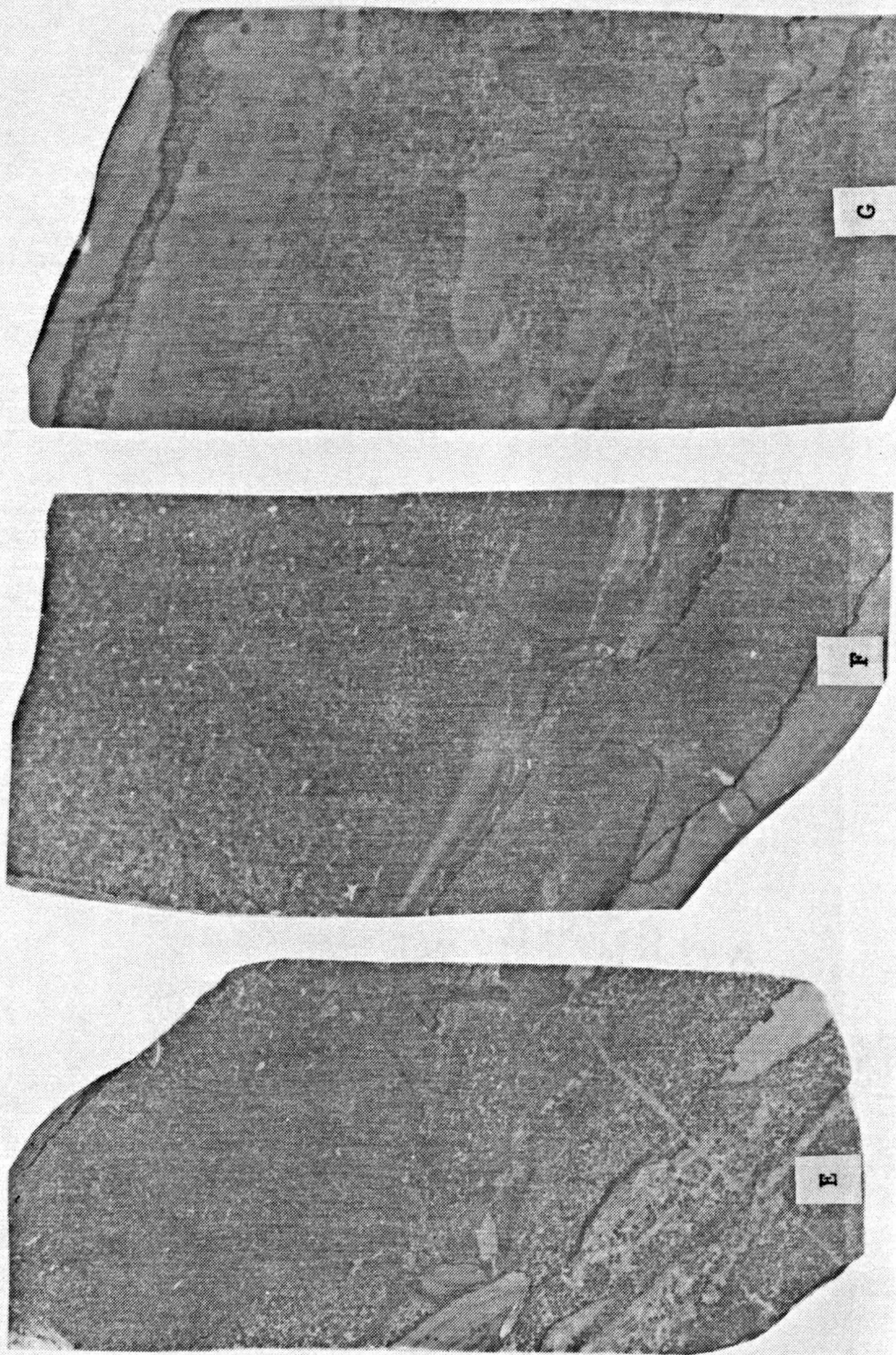


Figure 3.3 (continued). E. Slabbed 6.5 cm diameter core from borehole GW139 showing deep-water ooid calcarenite with intraclasts of ooid packstone at base. Width of slab is about 0.85 cm. Stratigraphic-up is at top of this and subsequent photos. F. Slabbed 6.5 cm diameter core from borehole GW134 of deep-water ooid calcarenite with ooids increasing in diameter upward. Width of slab about 0.6 cm. G. Slabbed 6.5 cm diameter core from borehole GW140 of deep-water ooid calcarenite overlain by a lime mudstone cap with faint ripple laminae. Lime mudstone clast (light-colored) occurs immediately to left of up-arrow. Width of slab is about 0.5 cm.

Bidirectional herringbone cross-stratification is common in the lower 4.5 m of this buildup (Figure 3.3b). Laterally discontinuous shale and lime mudstone interlayers that thicken over troughs also occur in this lower interval in addition to abundant stylolites. Stylolites typically occur along hardground surfaces that increase in abundance towards the top of this 4.5 m interval (Figure 3.3c). A prominent hardground 4.5 m above the base of the buildup was colonized by thrombolites. The remaining 1.5 m of this buildup consists of algal stromatolites developed on top of the thrombolites, and ooids and peloids between stromatolite heads. Ooids are the dominant grain type in this buildup though peloids and micrite increase in abundance in the upper 1.5 m. Ooid fabrics and diagenetic features will be described in detail below.

Ooid sand-sheet deposits occur in the I-75 outcrop (Figure 3.3d) and other outcrops in the Copper Creek thrust sheet. These are considered shallow-water sand-sheet deposits because of their association with other shallow-water deposits in this thrust sheet. In the Whiteoak Mountain (WOM) thrust sheet, ooid sand-sheet deposits are significantly different from shallow-water ooid sand-sheets in the Copper Creek thrust sheet. In the Copper Creek sheet, these beds are thicker and occur throughout the Nolichucky. At the I-75 outcrop (Copper Creek thrust sheet), these sand-sheet deposits grade laterally into the lenticular ooid grainstone buildups with abundant fibrous cement. In contrast, sand-sheet deposits in the WOM thrust sheet are thinner, consist of ooid-peloidal packstone, and typically occur as part of 0.5-1.0 m thick, upward coarsening packages (Figures 3.3e,f). These packages begin with calcareous siltstone that grades upward into the oolitic sand-sheet deposits, and are capped by intraclastic limestone. Internally, the oolite beds fine-upward from oolitic limestone clasts into graded horizontally laminated oolite, and capped by rippled-laminated wackestone (Figure 3g). These oolite sand-sheets do not interfinger with shallow-water lithologies, but are interbedded with basinal shale and allochthonous limestone units (Foreman and others, in press). Because of muddy matrix, cement is rare

in the ooid sand-sheet beds.

Petrography

Ooids in the Nolichucky Shale oolitic limestone units are strikingly similar to ooids in the Port au Port Group (Middle and Upper Cambrian) in western Newfoundland (Chow and James, 1987). Because of this similarity, we will use the fabric terminology of Chow and James when appropriate to describe Nolichucky ooids. Nolichucky ooids range in size from approximately 0.2 mm to 2 mm. There is a strong relationship between ooid diameter and nucleus type. The largest ooids typically have echinoderm nucleii or trilobite nucleii. Peloids are usually found in the center of smaller ooids. The following ooid varieties are recognized from the Nolichucky Shale: Fibrous radial, fibrous radial/sparry radial, sparry radial, and monocrystalline ooids. Fibrous radial/sparry radial ooids are transitional in fabric between fibrous radial and sparry radial ooids. All ooids take a potassium ferricyanide stain indicating iron (Fe^{2+}) in the calcite that probably exceeds 200 ppm (Oldershaw and Scoffin, 1967). Microprobe data presented in a later section confirm this observation.

Fibrous Radial Ooids- These ooids consist of fibrous calcite with long dimensions oriented perpendicular to the nucleii (Figures 3.4a,b). Maximum fiber widths rarely exceed 2-3 μm (Figure 3.5a). Typical nucleus types for these ooids are trilobites and peloids. Brown amorphous material (organic matter?) and pyrite are often found between adjacent fibrous crystals (Figure 3.4d). The outer laminae of the largest fibrous radial ooids typically have several concentric laminae defined by dark micrite (Figures 3.4a,c).

Fibrous radial ooids are dull to moderately luminescent. Concentric laminae occur as nonluminescent bands and are restricted to the outermost part of the ooid cortex. Ferroan baroque dolomite (nonluminescent) replaces fibrous radial ooids in the shallow-water ooid

Figure 3.4 (next page). Photomicrographs from Nolichucky shallow-water and deep-water oolite. Abbreviations presented here are used in subsequent figures, horizontal dimension in parentheses. Up direction towards top of photo. **A.** Shallow-water ooid sand-sheet with monocrystalline (M), fibrous radial (FR), sparry radial (SR) and baroque dolomite ooids (D). Pore spaces occluded by radial fibrous calcite cement. (6 mm). **B.** Shallow-water mobile shoal oolite with large cavity filled by fibrous radial calcite cement (F). Polygonal pattern (P) is transverse cut through fibrous cement (6.5 mm). **C.** Deep-water ooid calcarenite with fibrous radial and sparry radial ooids. Fibrous cement and peloids (p) occupy pore space (6.5 mm). **D.** Fibrous radial/sparry radial ooid (FS) with crystal rays bounded by organic? material and pyrite/ Renalcis fossil debris (R) surrounded by microcrystalline calcite spar (2 mm). **E.** Sparry radial ooids in deep-water ooid calcarenite. Fibrous calcite cement binds ooids (1.6 mm). **F.** Monocrystalline ooids replaced along rims by dolomite (arrow), fibrous cement between ooids. Shallow-water mobile shoal oolite (1.6 mm).

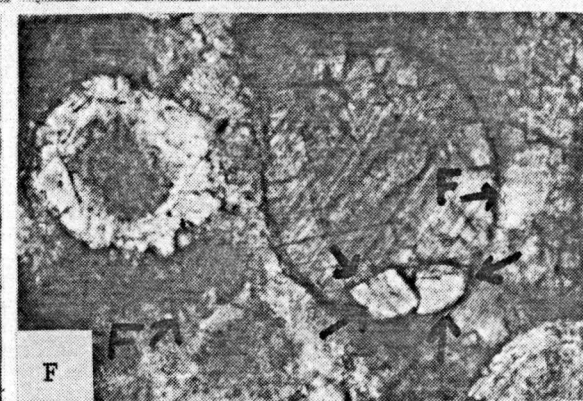
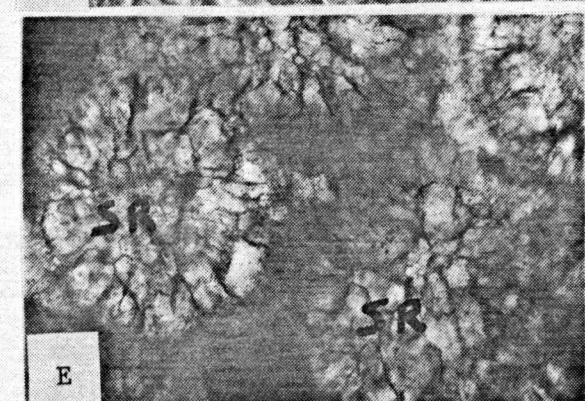
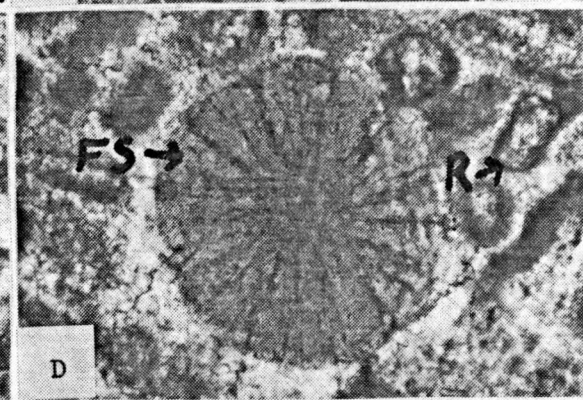
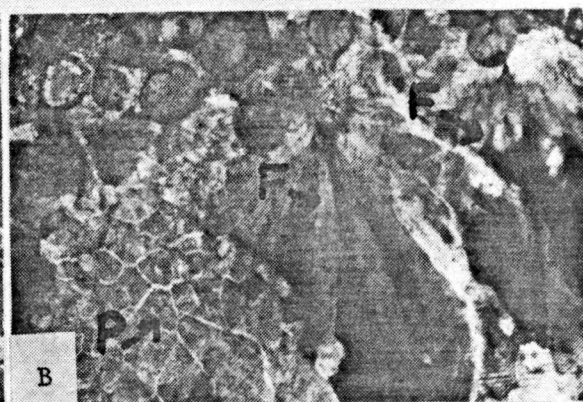
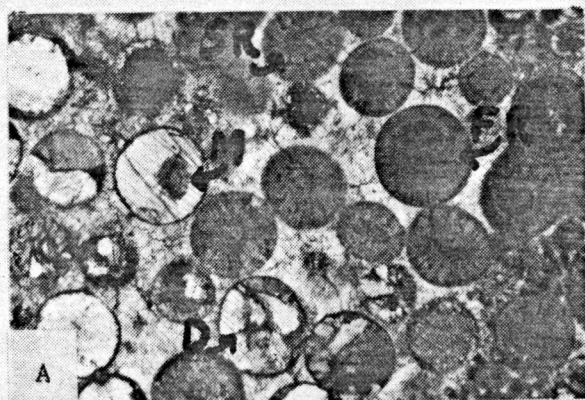
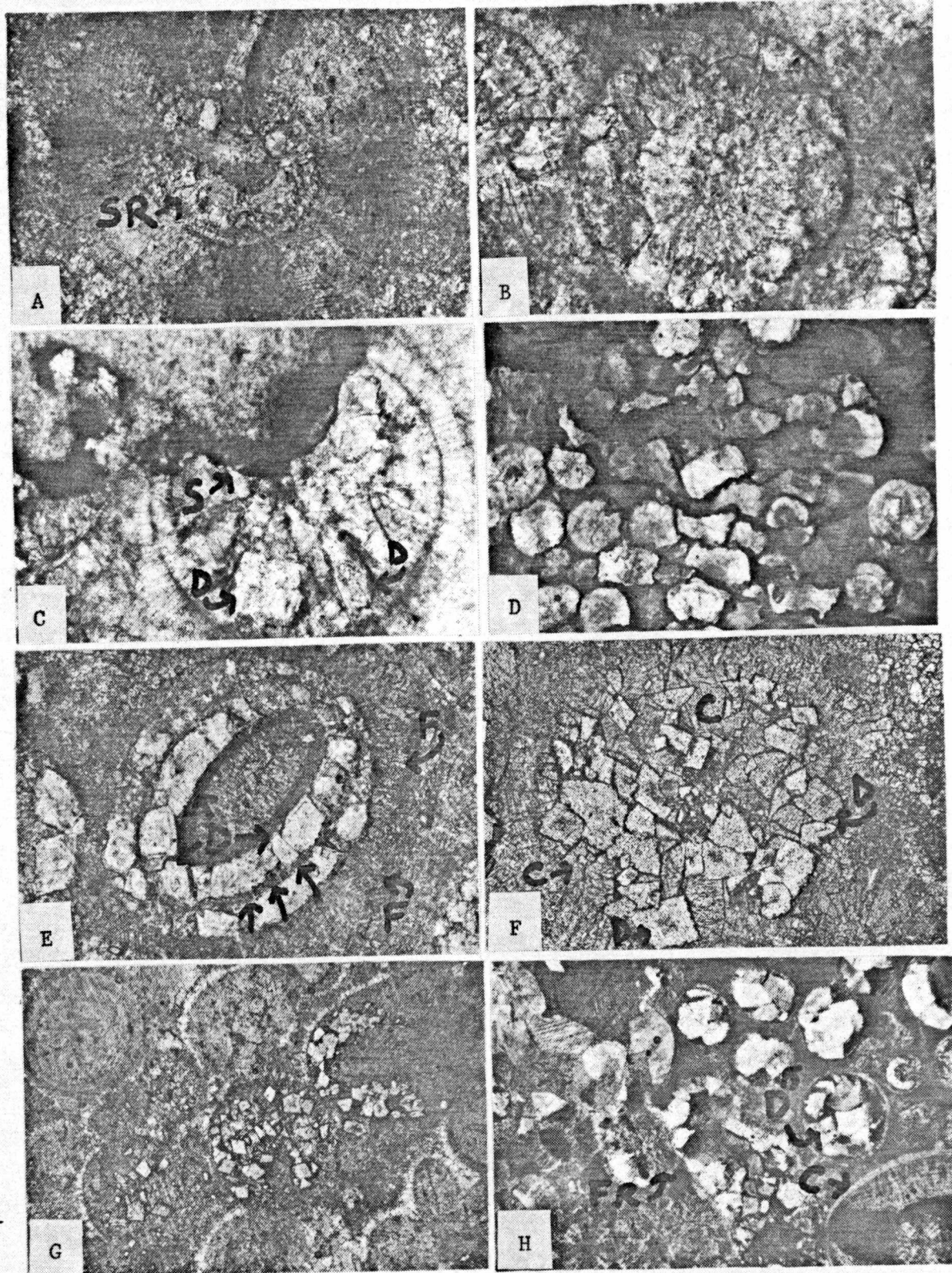


Figure 3.5 (next page). Photomicrographs from Nolichucky shallow-water and deep-water oolite. **A.** Coarse fibrous radial outer cortex with sparry radial inner cortex, echinoderm nucleus. Shallow-water stabilized shoal oolite (1.6 mm). **B.** Different degrees of alteration in ooid from deep-water ooid calcarenite. Outer portion has coarser sparry radial crystals than inner portion of ooid, unidentified nucleus (1.6 mm). **C.** Sparry radial ooid cut by bed-parallel stylolite (S). Dolomite (D) replacing ooid, isopachous fibrous cement (arrow). Deep-water ooid calcarenite (1.6 mm). **D.** Interpenetration of fibrous/sparry radial ooids along stylolites in deep-water ooid calcarenite packstone (1.6 mm). **E.** Fibrous/sparry radial ooid replaced by zoned dolomite (D) along inner and outer laminae. Coarse fibrous crystals preserved between replaced laminae (arrow), fossil making up nucleus filled by fibrous calcite, fibrous calcite cement abundant in this sample from shallow-water mobile shoal oolite (1.6 mm). **F.** Coarse sparry radial ooid replaced by zoned ferroan dolomite (D) and coarse ferroan calcite spar (C). Relict sparry radial crystals (arrows) visible in ooid, shallow-water stabilized shoal oolite (1.6 mm). **G.** Stabilized shoal oolite showing localized replacement of sparry radial and monocrystalline ooids. Fibrous radial and sparry radial ooids occur in close proximity to dolomitized ooids (6.5 mm). **H.** Shallow-water ooid packstone with partial replacement of sparry radial ooids (arrows) by coarse ferroan calcite spar and complete replacement of ooids by ferroan baroque dolomite (D). Well-preserved fibrous radial ooid in lower left (FR), compound ooid (C) in lower right (6.5 mm).



buildups and sand-sheets but this is somewhat less common in the deep-water ooid sand-sheets near Oak Ridge. The dolomite is often found in concentric bands, usually towards the ooid rim. Compared to other ooid varieties, fibrous radial ooids are less often replaced by dolomite. Fibrous radial ooids tend to be more abundant in packstone and wackestone lithologies. These lithologies are characteristic of the deep-water ooid sand-sheets and stabilized ooid shoal lithologies at the I-75 outcrop.

Sparry Radial Ooids- These ooids are distinguished from fibrous radial ooids by having coarser radiating calcite crystals. Sparry radial ooids may have maximum crystal widths of 10 μm resulting in ooids with only 15-20 individual wedge-shaped crystal rays making up the cortex (Figure 3.4e). Individual rays may extend from the nucleus to the rim, in which case crystal boundaries tend to be straight. These rays resemble radial fibrous calcite crystals described by Davies (1977) and Mazzullo (1980) and are usually in optical continuity with isopachous fibrous cements surrounding the ooids. Sparry radial ooids with narrower straight-sided crystals making up the cortex may be difficult to distinguish from fibrous radial ooids. These ooids are classified as fibrous radial/sparry radial ooids (Figure 3.4d).

Sparry radial ooids may also have crystals which do not span the full distance from nucleus to ooid rim. These crystals are usually cigar-shaped with irregular borders and are less turbid than coarser wedge-shaped rays just described. Cigar-shaped crystals do not have visible twin planes and they are usually not in optical continuity with isopachous fibrous calcite cement. Single sparry radial ooids are rarely observed with both wedge-shaped and cigar-shaped crystals. Peloids and trilobite fragments are the most commonly observed nucleus types in the sparry radial ooids.

Sparry radial ooids are more frequently replaced by ferroan baroque dolomite than are fibrous radial ooids, especially in the up-slope shallow-water localities. Sparry radial ooids with cigar-shaped crystals are more frequently dolomitized than are straight ray crystals

(Figure 3.5e). The dolomite often has crystal boundaries that follow the cigar-shaped crystals indicating that the crystals predate the dolomitization (Figure 3.5c). Pyrite is also common in dolomitized sparry radial ooids where it outlines the ferroan baroque dolomite and cigar-shaped calcite fibers. Sparry radial ooids are dull to moderately luminescent unless they are replaced by dolomite in which case they are nonluminescent. Faint, nonluminescent concentric laminae are often distinguished in these ooids.

Monocrystalline Ooids- These ooids are characterized by a single calcite crystal making up the cortex. This crystal is nearly always in optical continuity with an echinoderm grain in the ooid nucleus (Figure 3.4f). The echinoderm grain is more turbid than the surrounding crystal and is usually outlined by pyrite inclusions or microdolomite. Depending on how the thin section cuts these ooids, a nucleus may be faintly visible or absent. No evidence of dropped nuclei has been observed in Nolichucky ooids; the nucleus is almost always centered within the monocrystalline ooid.

These ooids have similar luminescence as the fibrous radial and sparry radial ooids. A relict fibrous radial fabric is apparent in many monocrystalline ooids and occurs as linear inclusions radiating away from the echinoderm nucleus. Faint concentric laminae can be seen in these ooids. They are often partially replaced by ferroan baroque dolomite which is nonluminescent (Figures 3.4f, 3.5g). Monocrystalline ooids are more common in the shallow-water lithologies, especially in grainstone units with abundant fibrous cement.

Cements- Radial fibrous calcite and bladed calcite are the most abundant cement types in the Nolichucky oolitic lithologies. These occur as isopachous coatings on ooids and commonly occlude all interparticle porosity. The cements are only lightly stained by potassium ferricyanide. Fibrous calcite cement is most abundant in the ooid buildups and shallow-water ooid sand-sheets (Figures 3.4a,b, 3.5e,f). Diffuse peloids packed between ooids in the deep-water ooid sand-sheet lithologies occlude interparticle porosity and limit fibrous calcite cementation (Figure 3.4c). Equant calcite is a rare second generation cement in the shallow-water lithologies, probably due to a lack of pore space following

cementation by fibrous calcite. Microcrystalline, equant ferroan calcite occludes porosity in deep-water oolitic sand lithologies. The calcite is uniform in size and typically grades into peloids. This fabric resembles *structure grumeleuse* described by Bathurst (1975) and probably reflects aggrading neomorphism of matrix micrite and peloids (Figure 3.4c).

Dolomite- The majority of the dolomitization in Nolichucky oolitic lithologies is restricted to ooids (Figures 3.5c,e,f). This fabric-selective dolomitization is more common in the up-slope shallow-water lithologies. Sparry radial ooids and monocrystalline ooids are replaced more often than any of the other ooid varieties (Figures 3.4f, 3.5c,e,f). In these ooids dolomite is clearly replacing calcite in the ooids though the dolomite crystals may be slightly corroded adjacent to the calcite (Figure 3.5c). All the dolomite is ferroan, with cloudy centers and clear rims (Figures 3.5f,g), and is typically baroque or saddle dolomite (Figure 3.5i; Radke and Mathis, 1980). Dolomite crystals do not extend outside the ooid cortex and often mimic the shape of concentric laminae in dolomitized radial ooids (Figure 3.5e).

Compaction- Mechanical compaction of fibrous radial ooids would result in fractures that would most likely follow the fibrous crystal margins; concentric laminae often are spalled off by mechanical compaction (Wilkinson and Landing, 1978). These features are rare in the Nolichucky oolitic units because contact between ooids during burial was limited by either fibrous cement or micrite matrix. Flattened ooids formed by crushing of oomolds are not present in the Nolichucky oolitic lithologies. Oomolds result from wholesale dissolution of the ooid interior resulting in a moldic porosity stage which would obliterate any original radial fibrous cortical fabric.

Evidence for chemical compaction, namely bed-parallel stylolites, is common in the oolitic packstone units in the Nolichucky, often resulting in sutured contacts between adjacent ooids (Figure 3.5d). However, stylolitization is rare in fibrous calcite-cemented grainstone from the shallow-water lithologies. Stylolites cross-cut all the ooids described previously (Figure 3.5c). Bed-normal calcite veins are fairly common in the oolitic

lithologies and cross-cut ooids and stylolites. Fluid inclusion evidence indicates these veins formed at maximum burial depths of about 3.6 km in the Whiteoak Mountain thrust sheet (Foreman and Dunne, in press). Details concerning the paragenetic sequence of oolitic fabrics will be presented in a later section.

Geochemistry

Trace elements- Trace element data for the Nolichucky ooids are presented in Figures 3.6 and 3.7 to illustrate differences between the two depositional settings, and fibrous radial and sparry radial ooids, respectively. The Mg vs Fe, and Mg vs Mn plots include likely starting compositions for the ooids (stippled pattern) in order to evaluate the hypothesis that the ooid fabrics are the result of different degrees of water-rock interaction. Examining chemical data in terms of water-rock interaction models in diagenetic systems is discussed in detail by Meyers and Lohmann (1985), Banner (1986), Banner and others (1988), Meyers (1989), and Banner and Hanson (1990). It is clear from Figure 3.6 that ooids from both shallow settings (Figures 3.6a,c) and deep settings (Figures 3.6b,d) have lost considerable Mg after deposition (justification for the estimate for original MgCO_3 content will be provided in the discussion). The trend of magnesium content with respect to iron and manganese is constant in both settings though the shallow-water ooids have slightly more magnesium ($\text{MgCO}_{3\text{shallow-water}} = 1.67 \pm 0.19$ mol%; $\text{MgCO}_{3\text{deep-water}} = 1.17 \pm 0.27$ mol%). Using the Student's t-test, it can be demonstrated that the magnesium content of shallow-water ooids is greater than that of deep-water ooids at the 99% confidence interval. The concentration of both manganese and iron are greater in the deep-water ooids compared to the shallow-water ooids ($\text{Mn}_{\text{shallow-water}} = 873 \pm 264$ ppm; $\text{Mn}_{\text{deep-water}} = 2273 \pm 594$ ppm; $\text{Fe}_{\text{shallow-water}} = 3586 \pm 1053$ ppm; $\text{Fe}_{\text{deep-water}} = 4273 \pm 1161$ ppm; Figure 3.6e,f). From the Student's t-test, $\text{Mn}_{\text{deep-water}} > \text{Mn}_{\text{shallow-water}}$ at the 99% confidence interval and $\text{Fe}_{\text{deep-water}} > \text{Fe}_{\text{shallow-water}}$ at the 90% confidence interval. In

shallow-water ooids, manganese is invariant with respect to iron (Figure 3.6e) whereas manganese and iron are positively correlated in deep-water ooids (Figure 3.6f).

The chemical data from deep-water ooids were used to determine if a relationship exists between ooid fabric and chemistry. The MgCO_3 content of fibrous radial ooids ($\text{MgCO}_3 = 1.18 \pm 0.24$ mol%) is not greater than that of sparry radial ooids ($\text{MgCO}_3 = 1.22 \pm 0.21$ mol%) at the 90% confidence interval. Figure 3.7 demonstrates that sparry radial ooids are more enriched in manganese and iron with respect to fibrous radial ooids ($\text{Mn}_{\text{fibrous radial}} = 1977 \pm 522$ ppm; $\text{Mn}_{\text{sparry radial}} = 2529 \pm 559$ ppm; $\text{Fe}_{\text{fibrous radial}} = 4139 \pm 1391$ ppm; $\text{Fe}_{\text{sparry radial}} = 4608 \pm 1033$ ppm). From the Student's t-test, $\text{Mn}_{\text{sparry radial}} > \text{Mn}_{\text{fibrous radial}}$ at the 99% confidence interval, and $\text{Fe}_{\text{sparry radial}} > \text{Fe}_{\text{fibrous radial}}$ at the 95% confidence interval. Though considerable scatter exists in the data, there appears to be a greater covariance between iron and manganese in the sparry radial ooids in comparison to the fibrous radial ooids (Figures 3.7e,f). Similar trends are suggested for fibrous radial and fibrous/sparry radial ooids in the shallow-water oolite though the trends are not as well developed.

Isopachous radial fibrous calcite cement is fairly abundant in the shallow-water oolite and its composition can be compared to the composition of coexisting shallow-water ooids. Magnesium concentrations are similar in shallow-water ooids and fibrous cement ($\text{MgCO}_{3\text{shallow-ooids}} = 1.67 \pm 0.19$ mol%; $\text{MgCO}_{3\text{fibrous cement}} = 1.52 \pm 0.72$ mol%; Figures 3.6a, 3.8a). Iron is more abundant in the shallow-water ooids when compared to fibrous cement ($\text{Fe}_{\text{shallow-ooids}} = 3586 \pm 1053$ ppm; $\text{Fe}_{\text{fibrous cement}} = 1803 \pm 880$ ppm; Figures 3.6c, 3.8b) and is significant at the 99% confidence interval. The manganese data for fibrous cement must be viewed with caution because values less than 450 ppm are near or below the detection limit for the microprobe. The invariant trend between Fe and Mn for fibrous cements (Figure 3.8c) is similar to that for shallow-water ooids (Figure 3.6e).

Echinoderms from the shallow-water oolite are similar to the shallow-water ooids

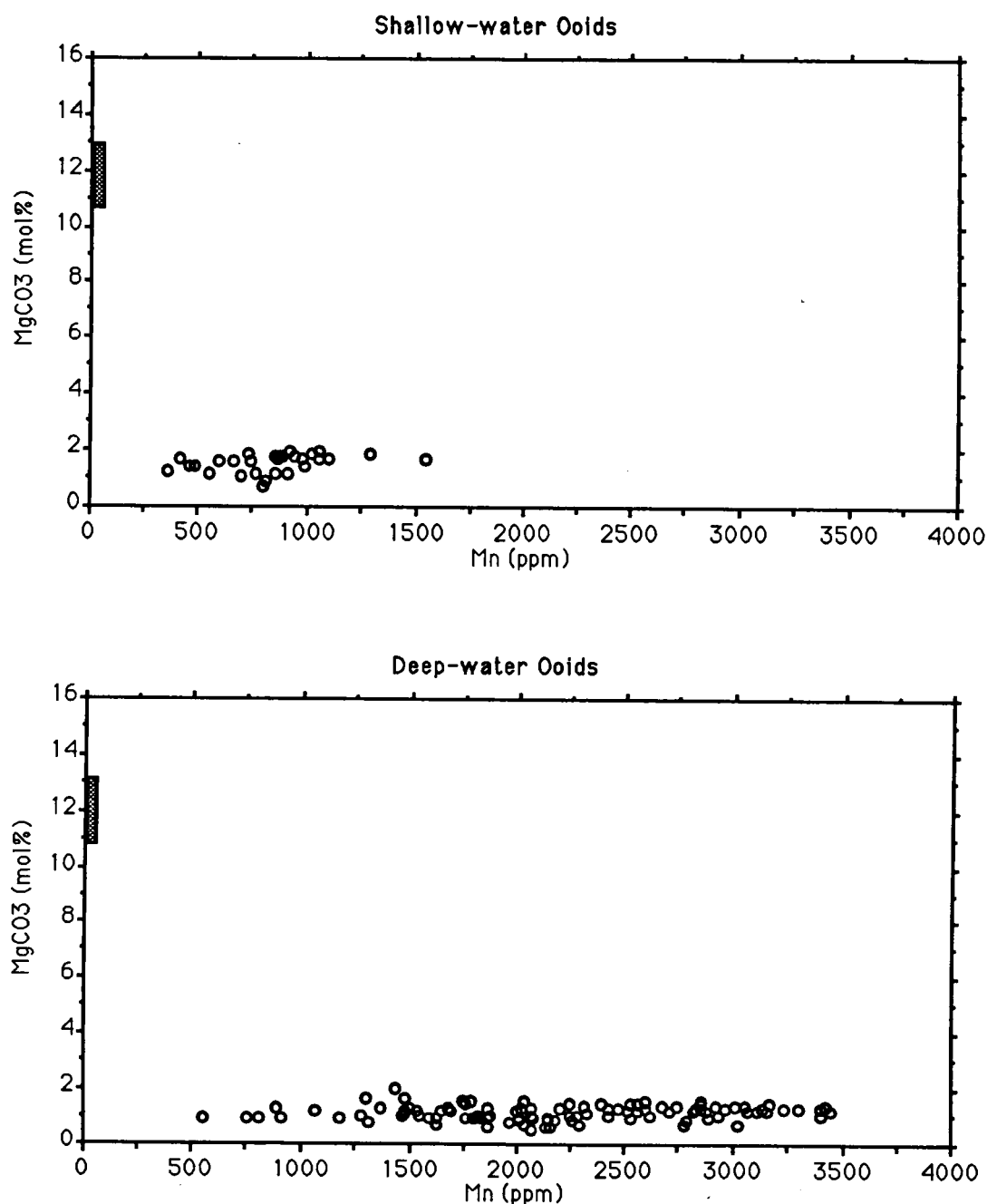
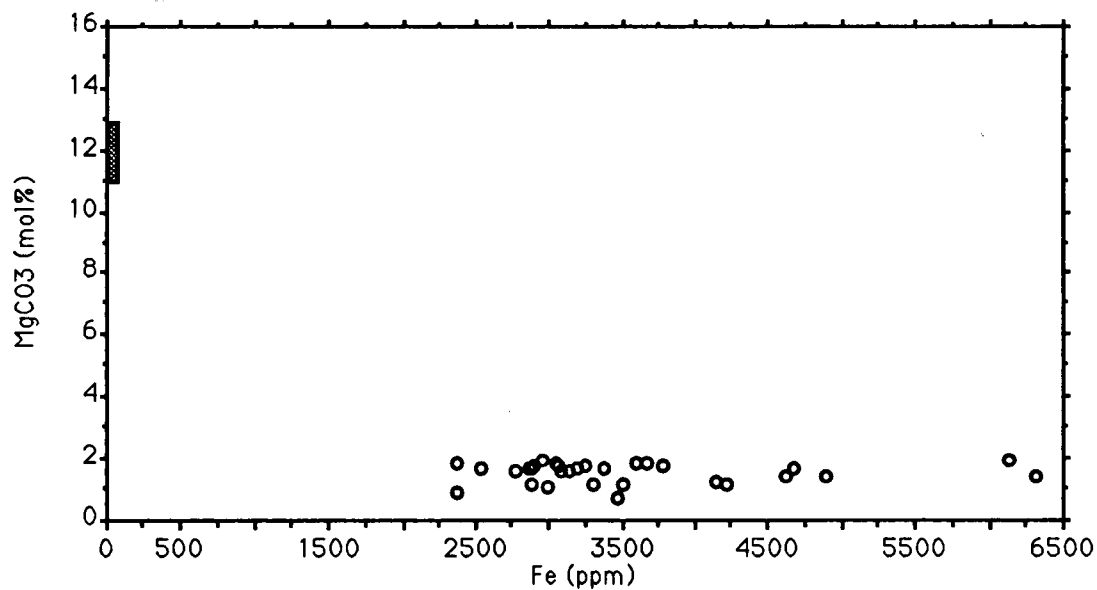


Figure 3.6. Electron microprobe data for shallow and resedimented deep-water ooids. Included are fibrous radial, fibrous/sparry radial, sparry radial, and monocrystalline ooids. Hypothesized original manganese, iron, and manganese content for this and subsequent figures shown in filled rectangle. A. Magnesium with respect to manganese for shallow-water ooids. B. Magnesium with respect to manganese for deep-water ooids. See text for summary of statistical tests.

Shallow-water Ooids



Deep-water Ooids

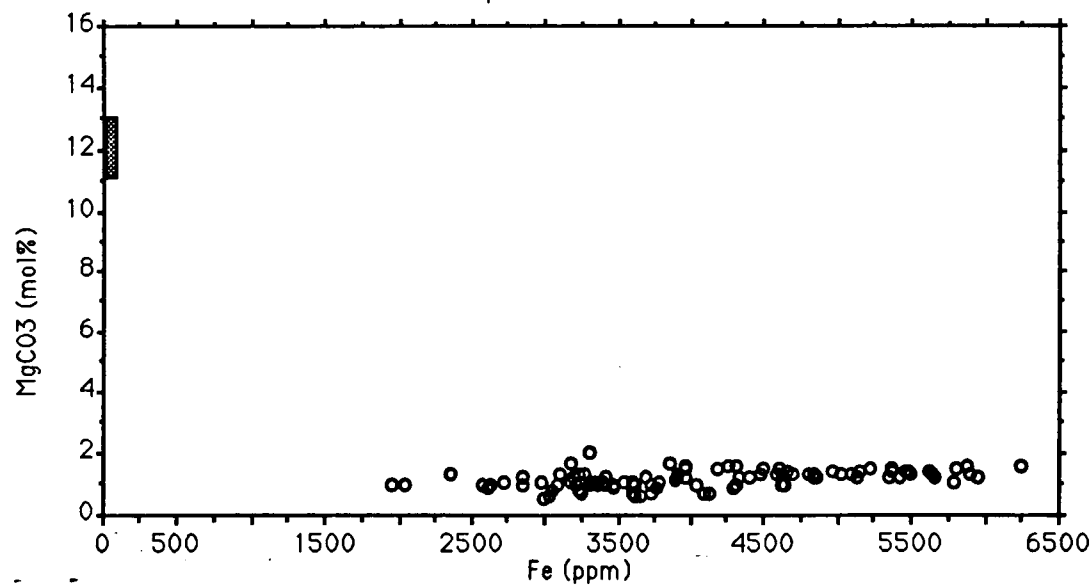


Figure 3.6 (continued) C. Magnesium with respect to iron for shallow-water ooids. D. Magnesium with respect to iron for resedimented deep-water ooids.

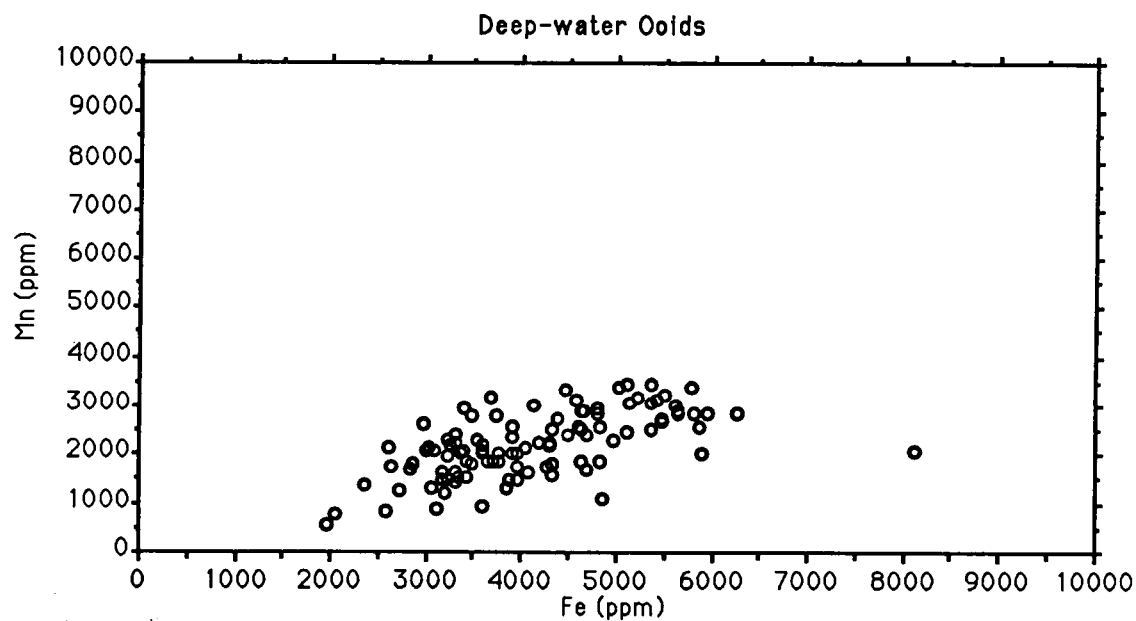
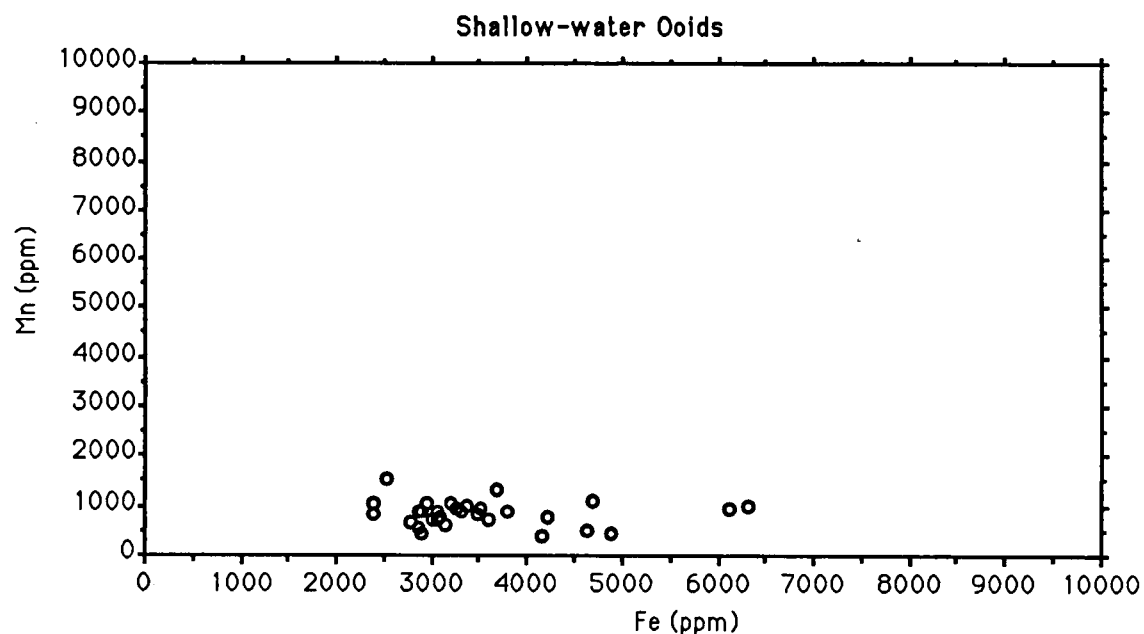


Figure 3.6 (continued) E. Iron with respect to manganese for shallow-water ooids. Mn is invariant compared to Fe. F. Iron with respect to manganese for resedimented deep-water ooids. Fe and Mn covary in contrast to shallow-water ooids.

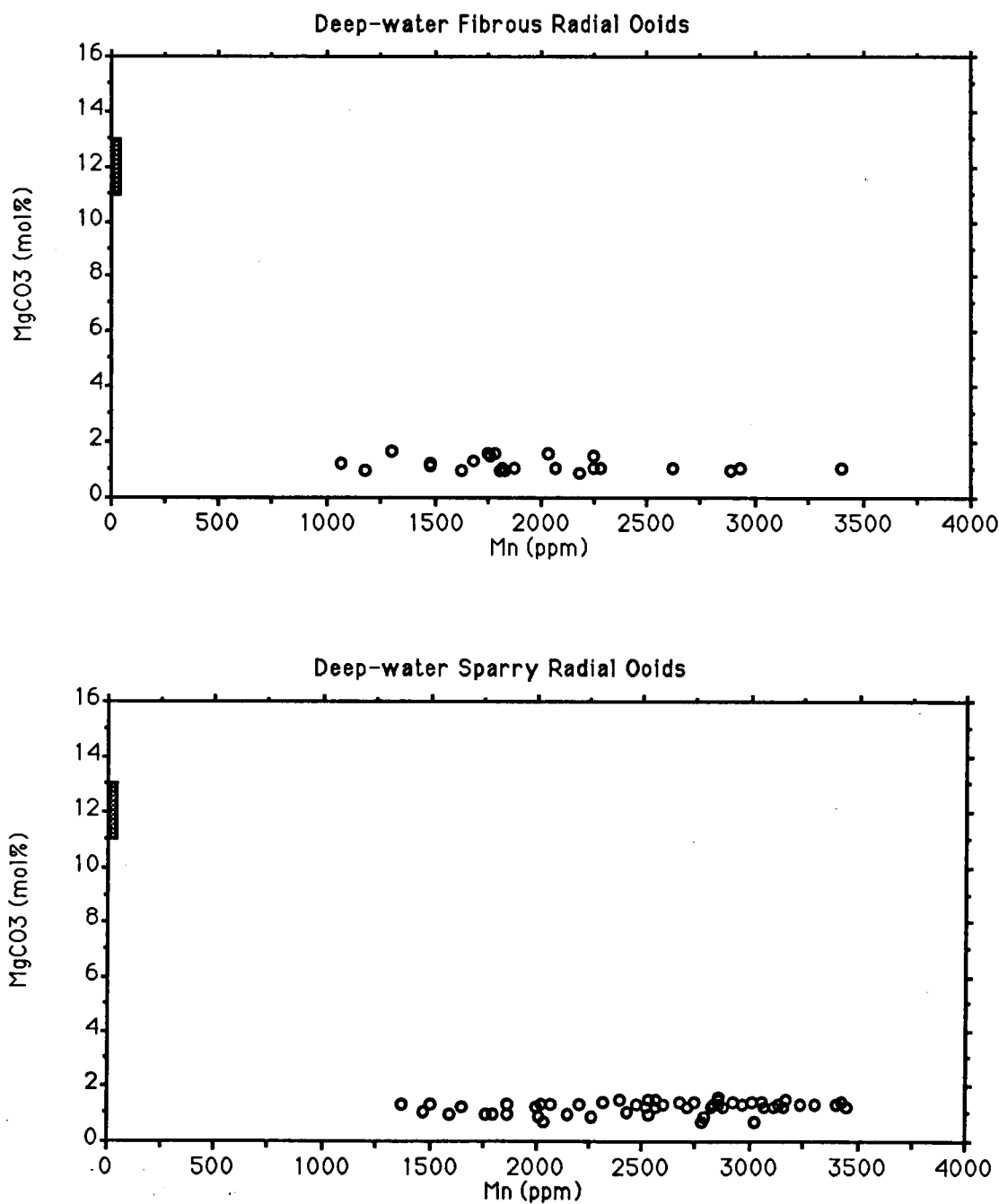


Figure 3.7. Microprobe data for deep-water fibrous radial ooids compared to deep-water sparry radial ooids. A. Magnesium with respect to manganese for fibrous radial ooids. B. Magnesium with respect to manganese for sparry radial ooids. See text for summary of statistical tests.

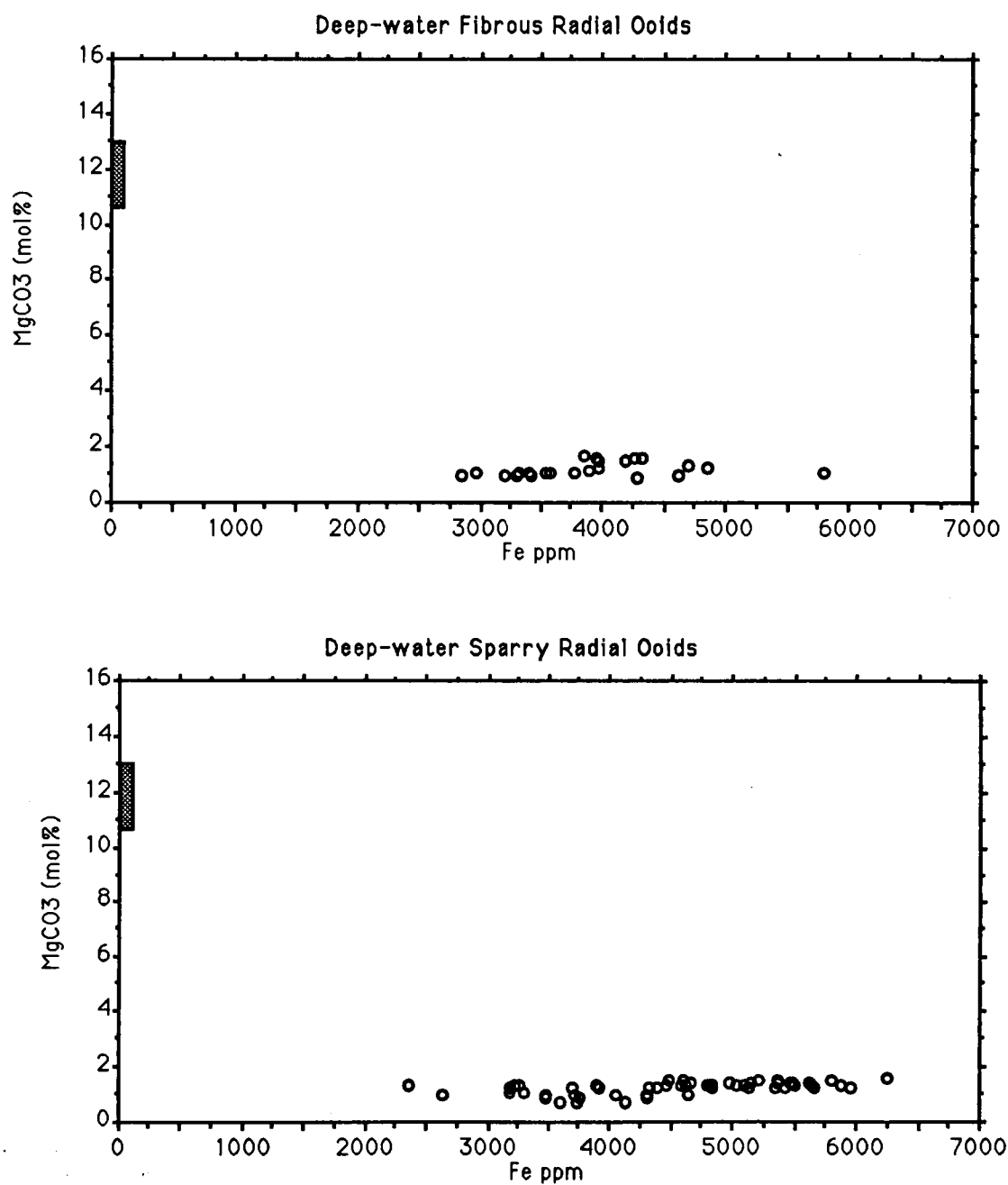


Figure 3.7 (continued) C. Magnesium with respect to iron for fibrous radial ooids. D. Magnesium with respect to iron for sparry radial ooids.

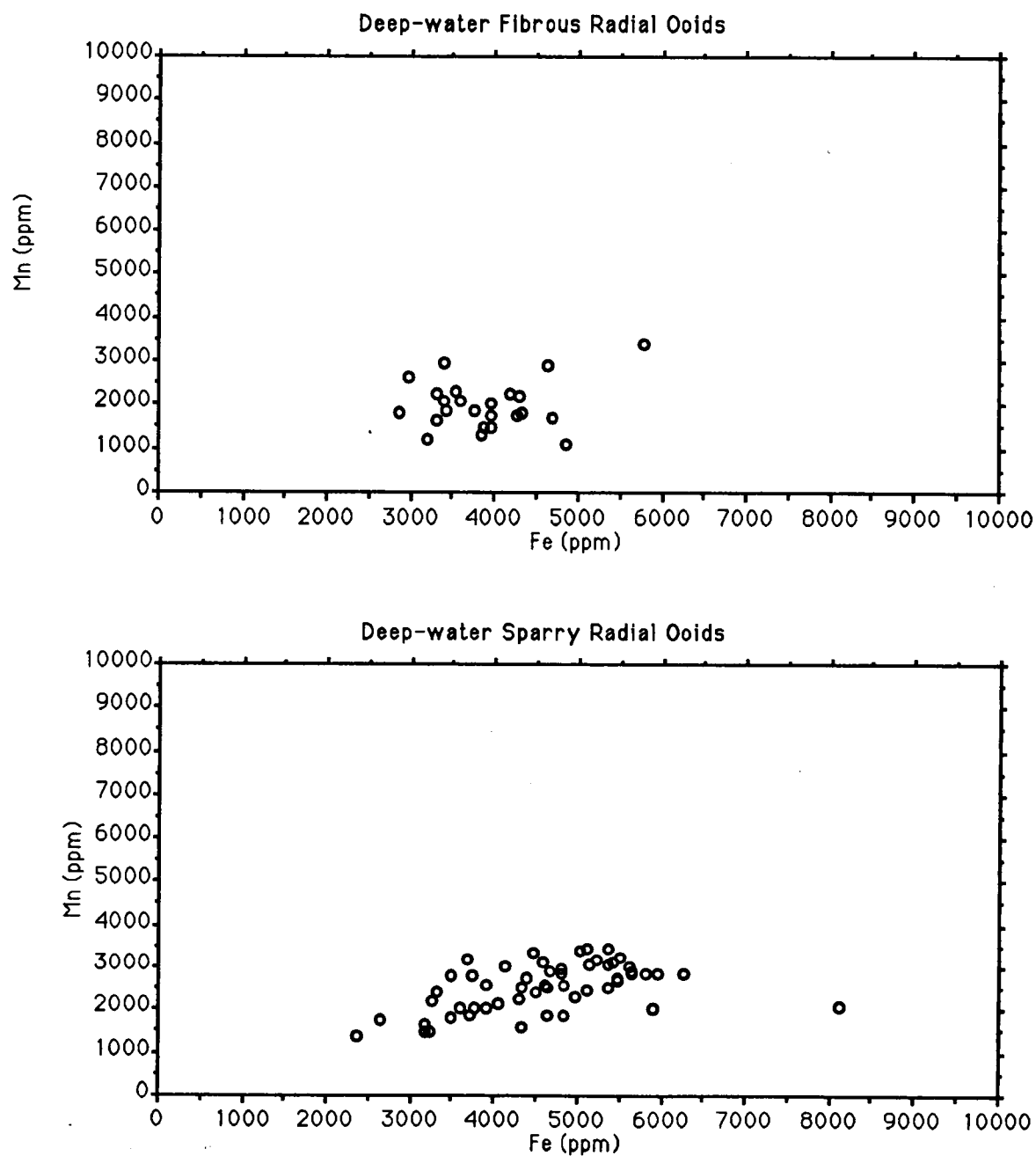


Figure 3.7 (continued) E. Iron with respect to manganese for fibrous radial ooids. F. Iron with respect to manganese for sparry radial ooids. Behavior is similar for both types of ooid in the deep-water setting.

with respect to iron and manganese ($\text{Fe}_{\text{echinoderms}} = 3171 \pm 600$ ppm; $\text{Mn}_{\text{echinoderms}} = 972 \pm 206$ ppm; Figure 3.9). The Fe vs Mn plot is also similar to the ooids and cements. Magnesium is significantly greater in the echinoderms ($\text{MgCO}_3_{\text{echinoderms}} = 2.85 \pm 1.28$ mol%) compared to all other calcite phases analyzed, though there is considerable variability in the magnesium content. This variability is attributed to the presence of 5-15 μm microdolomite rhombs in the echinoderm grains, identified using back-scatter electron imaging on the Cameca microprobe. Monocrystalline ooid coatings on echinoderm nuclei have a uniformly low magnesium content (2.0 mol% MgCO_3) and no recognizable microdolomite.

Stable Isotopes- The isotopic composition of fibrous radial and fibrous radial/sparry radial ooids from shallow-water oolite and deep-water oolite are shown in Figure 3.10. Radial fibrous calcite cement from the shallow-water oolite and dolomitized ooids from the shallow-water oolite are also included in this figure. With the exception of one sample, all the undolomitized shallow-water ooid analyses are significantly heavier than the deep-water ooids with respect to oxygen and carbon. The shallow-water fibrous cement compositions lie in the field for deep-water ooids. Shallow-water ooids dolomitized by ferroan baroque dolomite also occur in the compositional field for deep-water ooids.

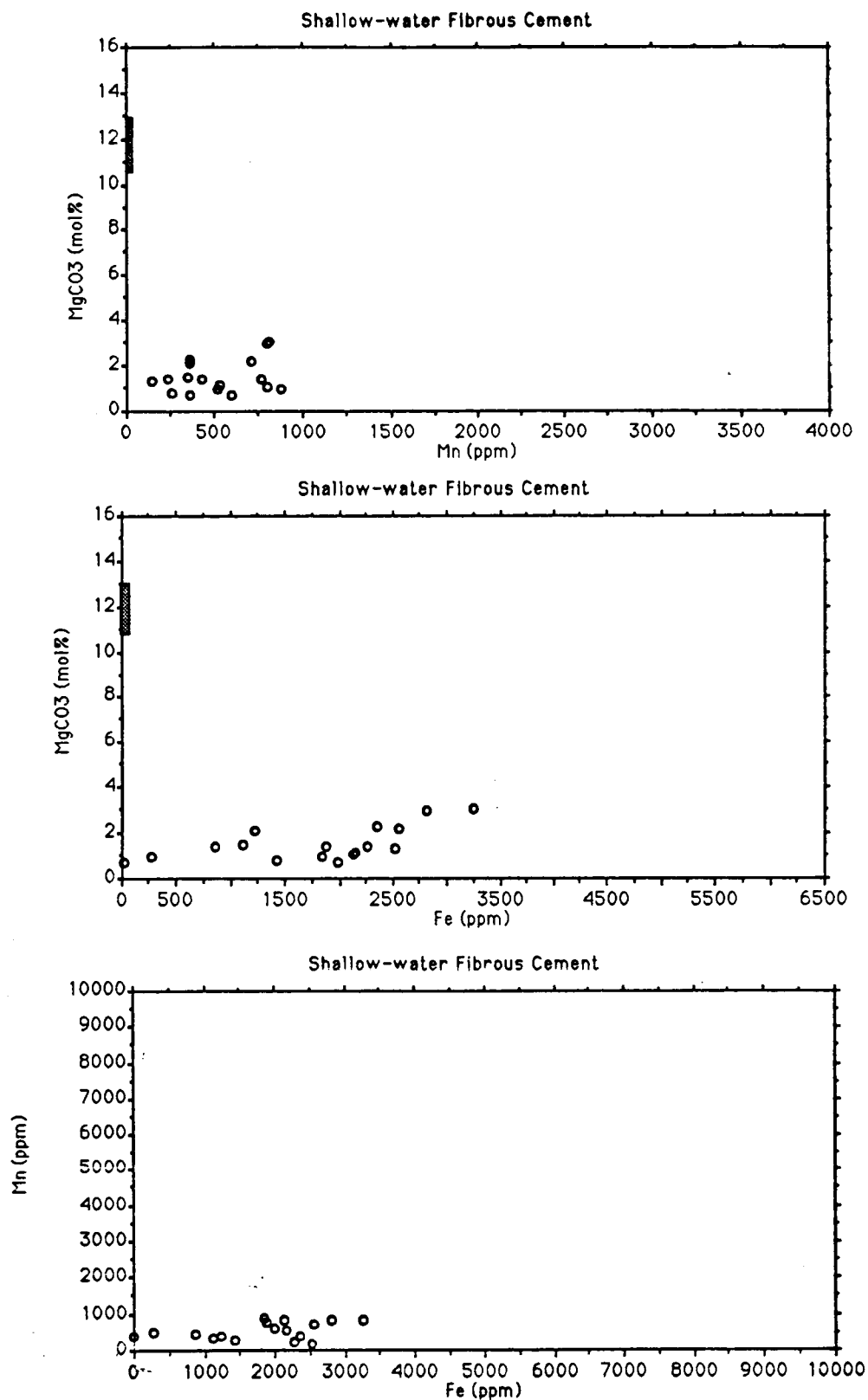


Figure 3.8. Electron microprobe data for shallow-water radial fibrous calcite cement. **A.** Magnesium with respect to manganese, almost half the analyses are below the manganese detection limit (450 ppm). **B.** Magnesium with respect to iron. **C.** Iron with respect to manganese. Behavior is similar to shallow-water ooids (Figure 3.6e).

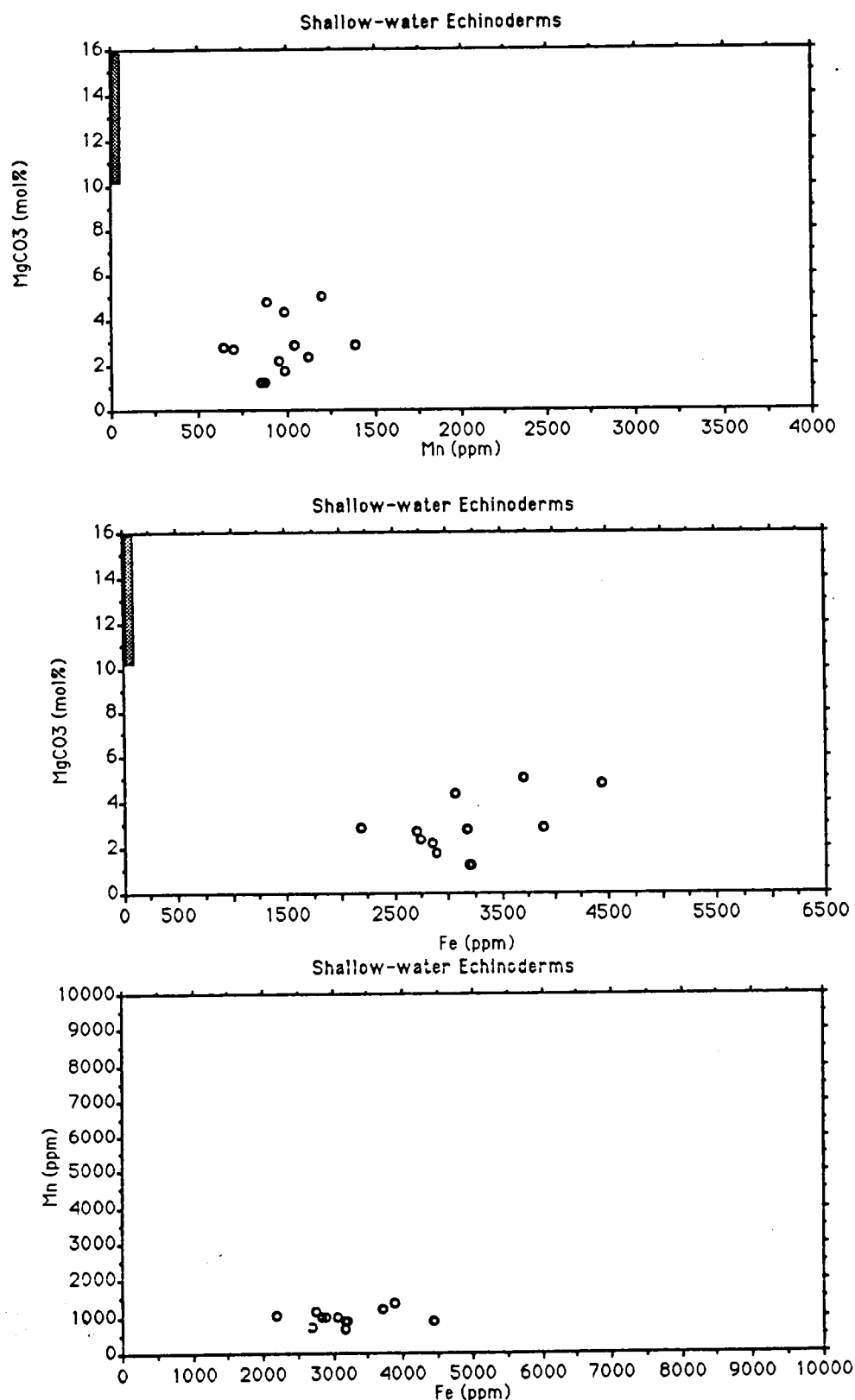


Figure 3.9. Electron microprobe data for shallow-water echinoderm debris. **A.** Magnesium with respect to manganese. Microdolomite is common in echinoderms and explains high magnesium values. **B.** Magnesium with respect to iron. **C.** Iron with respect to manganese. Behavior similar to shallow-water ooids (Figure 3.6e) and shallow-water fibrous cement (Figure 3.8c).

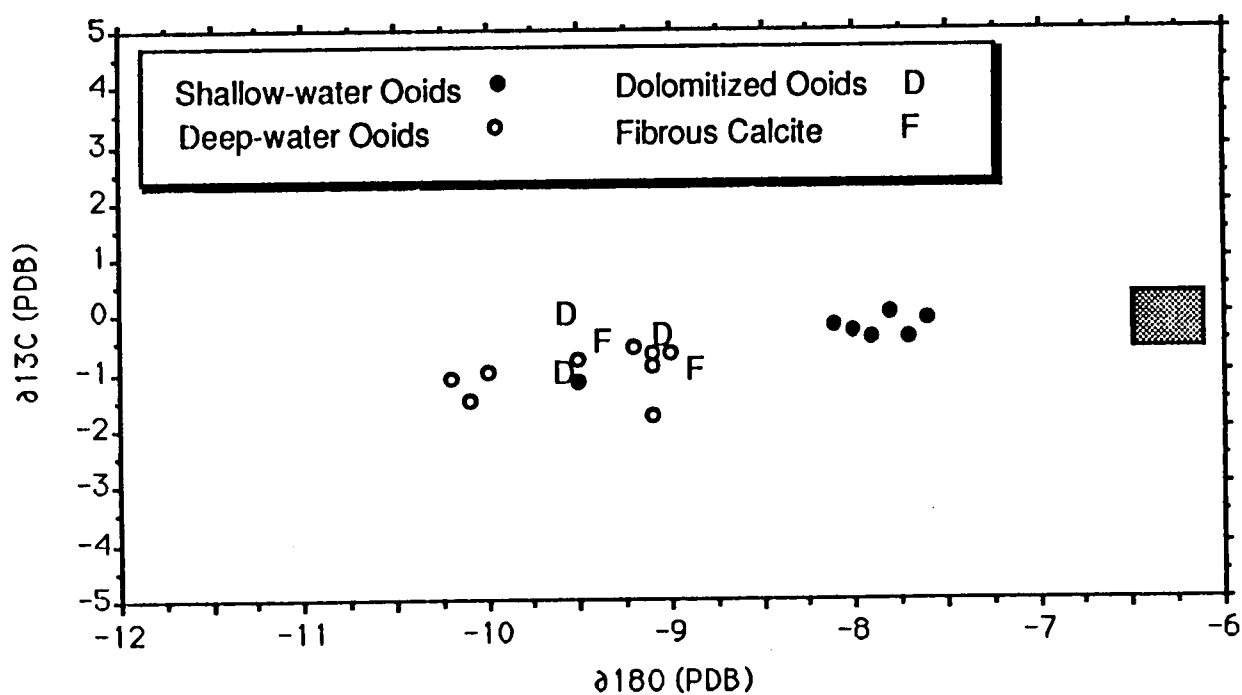


Figure 3.10. Stable isotopic data for shallow-water fibrous radial and fibrous/sparry radial ooids (filled circles), shallow-water fibrous cement (F), shallow-water baroque dolomite ooids (D), and deep-water fibrous radial and fibrous/sparry radial ooids (open circles). Filled rectangle corresponds to hypothesized original isotopic composition of Nolichucky radial fibrous ooids.

DISCUSSION

Environmental Interpretation

The lenticular oolitic buildup described from the Interstate 75 outcrop is interpreted to represent an oolite shoal where ooids were actively forming (Foreman and Weber, 1989). Herringbone cross-stratification in the lower portion of the buildup and the winnowed nature of the oolite indicate deposition under bidirectional currents and oscillating wave action. These conditions are characteristic of modern ooid shoals in areas of high current and wave agitation (Loreau and Purser 1973; Hine, 1977; Harris, 1979). The herringbone cross-stratification probably formed as a result of strong tidal currents that kept ooids in nearly continuous motion on the seafloor. This motion resulted in the formation of well-formed spherical ooids. The ooid grainstone comprising the lower portion of the I-75 oolite is considered to be analogous to the mobile ooid sand shoal subenvironment of the Joulters Ooid Shoal described from the Bahamas by Harris (1979). Modern mobile ooid shoal environments have water depths which are generally less than 2 m (Harris, 1979).

Oolitic packstone and grainstone beds which grade and thicken laterally toward the I-75 mobile ooid shoal lithofacies are interpreted as sand-sheets representing ooids transported off the ooid shoal by strong currents and storm events (Weber, 1988; Foreman and others, in press). The poorer sorting, characteristic of these beds, indicates less frequent current and wave agitation. Environments represented by these units were not likely to be sites of active ooid formation and these lithologies were probably deposited below normal wave base .

The ooid packstone capping the mobile ooid shoal facies at I-75 is texturally similar to

sediment present in the stabilized ooid sand-flat environment from the Joulters Ooid Shoal (Harris, 1979). In modern carbonate environments stabilized sand-flats occur in shallower water that is less agitated than the mobile shoal environment. The lack of current and wave agitation limits the degree of winnowing resulting in poorly sorted sediment with a significant amount of carbonate mud. Stabilized ooid sand-flats in the Bahamas are colonized by sea grasses and algal mats. In the I-75 oolite body, a hardground occurs at the contact between the underlying ooid grainstone mobile shoal lithofacies and the stabilized sand-flat. This hardground is colonized by thrombolites and algal stromatolites (Weber, 1988). Surrounding and overlying the thrombolites and stromatolites is oolitic packstone with oncoids and poorly formed ooids (Weber, 1988). This lithology is interpreted as a stabilized ooid sand-flat. Water depths in the stabilized sand-flat environment from the Joulters Ooid Shoal reported by Harris (1979) are generally less than 1 m. Shale and intraclastic limestone is in sharp contact with the stabilized sand-flat and suggests the ooid shoal was drowned by a relative rise in sea level.

The oolitic packstone interbeds in fissile shale from the Whiteoak Mountain thrust sheet (WOM) near Oak Ridge are markedly different from the ooid grainstone and packstone beds at I-75 which grade laterally into the mobile ooid shoal lithofacies. Oolitic limestone beds are thinner in the Oak Ridge area and typically occur between calcareous siltstone and intraclastic limestone debris flow deposits as part of coarsening-upward sequences (Foreman and others, in press; Walker and others, in press). The oolitic interbeds in the WOM sheet are interpreted as allochthonous deposits derived from up-slope shallower-water settings. The oolite interbeds are graded calcarenite, the carbonate-equivalent of a siliciclastic turbidite, and contain an incomplete Bouma sequence of division A (intraclasts and ooids grading up into ooids), division B (horizontally-laminated ooid packstone), and occasionally division C (ripple-laminated ooid wackestone cap, Foreman and others, in press). A carbonate turbidite interpretation is reasonable for these lithologies since shallow-water features recognized in the I-75 oolite, including algal limestone and

oncooids in the stabilized sand-flat lithofacies, are absent in the WOM thrust sheet. Also, thick oolite bodies indicative of active oolite shoals have not been observed in the Whiteoak thrust sheet. Water depths during deposition of the oolitic packstone interbeds in the Oak Ridge area were probably on the order of 100-300 m (Foreman and others, in press). Clearly the ooids in these units were resedimented from shallow-water settings typical of those represented by the I-75 oolite.

Original Ooid Mineralogy

The original mineralogy of Nolichucky ooids can be inferred from the petrography and geochemistry of Holocene and relict ooids in the oceans today. Unfortunately, we can not be certain that Cambrian ocean chemistry and ooid mineralogy were identical to modern oceans. In fact, the Middle Cambrian is believed to correspond to the cross-over from aragonite and magnesium calcite to magnesium calcite and calcite as the preferred nonskeletal carbonate mineralogy (Sandberg, 1983). In light of this, it is not inconceivable that all three polymorphs may have coexisted in Middle Cambrian seas. None the less, both textural and geochemical evidence exist that support a high-magnesium calcite mineralogy for fibrous radial ooids described from the Nolichucky Shale.

Ooids with a fibrous radial cortex are rare in modern oceans; the dominant ooid cortical fabric in oceans today is a tangential fabric composed of acicular aragonite. Four of eight reported occurrences of radial fibrous cortices are aragonitic ooids from hypersaline environments (Great Salt Lake, Persian Gulf, Red Sea, and Lizard Island). The remaining occurrences are high magnesium fibrous radial ooids from Baffin Bay, the Florida continental shelf, the Amazon continental shelf, and the Great Barrier Reef (Land and others, 1979; Major and others, 1988; Milliman and Barretto, 1975; Marshall and Dravies, 1975; respectively). Of the latter group, only the Baffin Bay ooids are in water shallow-enough to argue against them being relict ooids (Land and others, 1979).

As demonstrated by Sandberg (1975, 1983), and many others, aragonitic ooids, regardless of their cortical fabric, are rarely preserved in the geologic record. These ooids either are dissolved, forming an intermediate oomoldic porosity stage before being filled by low-magnesium calcite (LMC), or replaced at the microscopic level by LMC (calcitization). In both cases coarse sparry calcite obliterates the original cortex. Exposure to meteoric waters or deep ocean water (below the aragonite compensation depth) are the most likely diagenetic settings for the transformation of aragonite to calcite to occur. In contrast, ooids, cements, and skeletal carbonates with a HMC original mineralogy will lose the majority of their original magnesium during stabilization but still retain excellent fabric detail (Towe and Hembelen, 1976).

Though monocrystalline ooids and some coarse sparry radial ooids from the Nolichucky may suggest calcitized aragonitic ooids, the absence of dropped nuclei and crushed ooids argues against an aragonitic precursor (Wilkinson and Landing, 1978; Sandberg, 1983; Wilkinson and others, 1984; Wilkinson and others, 1985). Many of the monocrystalline ooids have a relict radial fibrous fabric defined by inclusions parallel to the original fibrous crystals (Figure 3.5). These fibrous crystals nucleated on echinoderm substrates which were most certainly HMC (Meyers and Lohmann, 1977; Lohmann and Meyers, 1978; Richter and Fuchtbauer, 1978). Similar crystal growth was described from HMC fibrous radial ooids with echinoderm nuclei from the Great Barrier Reef as evidence for precipitation of HMC on the echinoderm substrate (Marshall and Davies, 1975). Back-scatter electron imaging of Nolichucky monocrystalline ooids identified areas 5-10 μm in size of elevated magnesium concentrations in echinoderm nuclei and ooid cortices and are interpreted to be microdolomite. The presence of microdolomite supports a HMC precursor mineralogy for these ooids (Lohmann and Meyers, 1977; Meyers and Lohmann, 1978).

Trace element analyses from the Nolichucky ooids support the petrographic interpretation that the fibrous radial, sparry radial, and monocrystalline ooids were

originally HMC. Elevated magnesium, iron, and manganese, and low strontium concentrations are characteristic of stabilized high magnesium calcites (termed diagenetic low-magnesium calcite, dLMC; Brand and Veizer, 1980). Magnesium concentrations for modern aragonitic ooids and calcitized aragonitic ooids rarely exceed 1 mol% MgCO_3 and generally are less than 0.5 mol% MgCO_3 (Land and others, 1979; Tucker, 1985; Major and others, 1988). All Nolichucky ooid varieties have greater than 1.5 mol% MgCO_3 and less than 2.5 mol% MgCO_3 which supports a magnesium calcite precursor. Iron and manganese in Nolichucky ooids, greater than 1000 ppm and 500 ppm, respectively, are more than two orders of magnitude greater than reported values for modern marine precipitates. These elevated values can be explained by stabilization of HMC to dLMC in the presence of Fe^{2+} and Mn^{2+} and are typical in many ancient limestones (Richter and Fuchtbauer, 1978; Brand and Veizer, 1980; Veizer, 1983; Popp and others, 1986). The uptake of iron and manganese at the expense of magnesium in the Nolichucky ooids will be discussed below. Strontium concentrations in Nolichucky ooids are below the detection limit of 450 ppm. Strontium depletion is expected during stabilization of HMC and calcitization of aragonite (Brand and Veizer, 1980; Veizer, 1983). Initial higher strontium concentrations for aragonitic ooids, 10,000-12,000 ppm in Holocene ooids (Land and others, 1979; Major and others, 1988), causes these ooids to retain an elevated strontium content in excess of 1000 ppm following calcitization (Davies, 1977; Mazzullo, 1980; Tucker, 1985). The Sr^{2+} in Nolichucky ooids is clearly less than 450 ppm and argues against an aragonite precursor.

The combination of petrographic and trace element data strongly suggests an original magnesium calcite mineralogy for Nolichucky ooids. However, the question remains as to the amount of magnesium in the ooids prior to stabilization. Since the magnesium distribution coefficient is less than 1.0, the calcite cannot gain Mg^{2+} during stabilization (Brand and Veizer, 1980; Veizer, 1983) and therefore the present Mg^{2+} content indicates a minimum original concentration of 2.5 mol% MgCO_3 . Even though the controls on Fe^{2+}

and Mn^{2+} incorporation in calcite are not completely understood, several examples from the literature indicate that 500-1000 ppm Fe^{2+} and 100-250 ppm Mn^{2+} can be substituted into calcite which has lost about 4-6 mol% MgCO_3 (Major and others, 1988; Meyers, 1989). The amount of iron and manganese in Nolichucky ooids is several times greater than these concentrations and suggests the Nolichucky ooids lost at least 5-10 mol% MgCO_3 . With a present composition of 2.5 mol% MgCO_3 and assuming a loss of 5-10 mol% MgCO_3 , the Nolichucky ooids probably had an original composition of at least 7.5-12.5 mol% MgCO_3 . By comparison, the four occurrences of Recent fibrous radial high-magnesium calcite ooids have magnesium contents of 12.0-15.0 mol% MgCO_3 . These compositions agree well with the theoretical composition of 13.7 mol% MgCO_3 for abiotic carbonate precipitated from normal seawater at 25 °C (Fuchtbauer and Hardie, 1976).

Dolomitization of many of the fibrous radial and sparry radial ooids is limited to portions of the cortices bounded by concentric laminae (Figure 3.5e). This fabric is suggestive of bimineralic ooids and is expected to occur along the outermost laminae of the ooids (Wilkinson and others, 1984). The outermost laminae in larger fibrous radial calcitic ooids may be favorable sites for aragonite precipitation due to higher fluid shear and resulting lower $p\text{CO}_2$ (Wilkinson and others, 1984). Close examination of many of these fabrics in Nolichucky fibrous radial and sparry radial ooids reveals that dolomitized laminae do not occur preferentially along the outer laminae and that there is no relationship between this style of dolomitization and ooid size (Figures 3.5e,g). In addition, fibrous radial crystals extend beyond these dolomitized laminae and often are in optical continuity with fibrous crystals present between the nucleus and dolomitized laminae (Figure 3.5e). Therefore, it cannot be stated unequivocally that the dolomite is replacing aragonite or calcitized aragonite. Instead, a monomineralic HMC precursor is favored for Nolichucky ooids. In contrast, Cambrian fibrous radial and sparry radial ooids from Newfoundland commonly show evidence for ooid compaction including spalled outer laminae and

elephantine ooids and are interpreted as bimineralic ooids (Chow and James, 1987). These fabrics are absent in the Nolichucky oolite lithologies.

Evidence for Water-Rock Interaction

Several geochemical trends are present: 1) slightly greater iron and manganese in deep-water resedimented ooids relative to the shallow-water ooids, 2) less magnesium in the deep-water ooids compared to shallow-water ooids, and 3) more negative $\delta^{18}\text{O}$ for deep-water ooids compared to shallow-water ooids. Assuming the same starting composition for resedimented ooids in the deep-water calcarenite, and shallow-water ooid shoal, the geochemical differences can be explained by the degree of water-rock interaction, differences in fluid composition during stabilization, or a combination of the two (Brand and Veizer, 1980, 1981; Veizer, 1983; Meyers, 1989). It is unlikely that during transport ooids were segregated on the basis of composition, so it is reasonable to assume similar initial compositions for deep and shallow-water ooids. Environmental differences on the seafloor cannot explain the geochemical differences either because water depths varied by less than several hundred meters suggesting temperature differences of only 10-15 °C (Foreman and others, in press). Resedimented HMC ooids and skeletal debris usually undergo little stabilization at depths of less than 200 m (Marshall and Davies, 1975; Milliman and Barretto, 1975). An exception to this is the Florida continental shelf where resedimented HMC ooids in 40 m water depth have 500-1000 ppm and 100-250 ppm Fe^{2+} and Mn^{2+} , respectively. However, these ooids were exposed to meteoric water during the Pleistocene and may have reacted with these waters prior to the subsequent sea level rise. None the less, the concentrations of iron and manganese has been interpreted to indicate the incorporation of these elements during loss of about 4 mol% MgCO_3 on the seafloor (Major and others, 1988).

If stabilization occurs in deeper ocean waters, the oxygen isotopic composition of the dLMC should be equal to or heavier than shallow-water HMC and dLMC as a result of cooler temperatures in deeper water (O'Neil and others, 1969; Gomberg and Bonatti, 1970; Land and Goreau, 1970). However, resedimented Nolichucky ooids are depleted in ^{18}O by an average of 1.6 per mil relative to shallow-water ooids (Figure 3.10). Assuming the same seawater composition at the two localities, the deep-water ooids indicate stabilization at temperatures 13 $^{\circ}\text{C}$ *higher* than in shallow-water ooids using the calcite-water fractionation factor of O'Neil and others (1969).

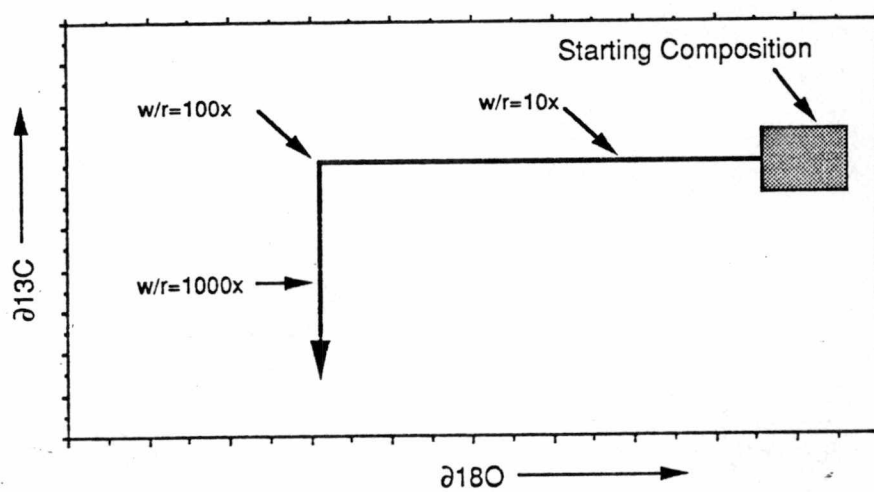
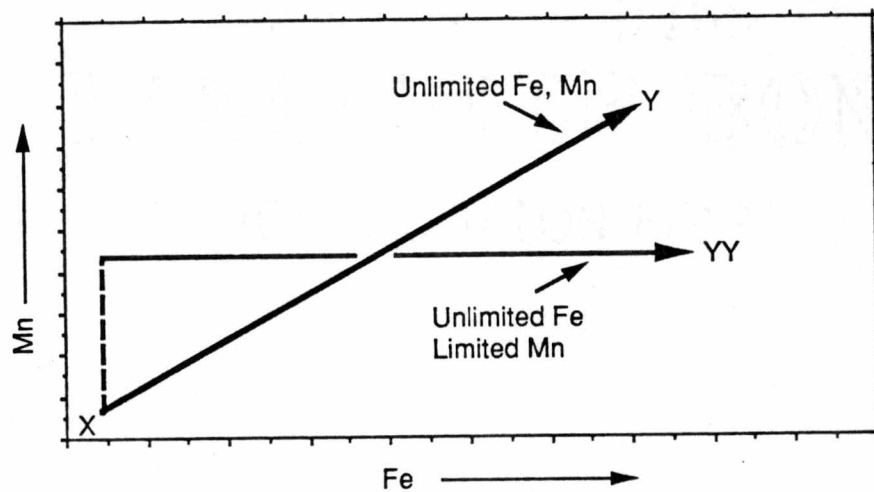
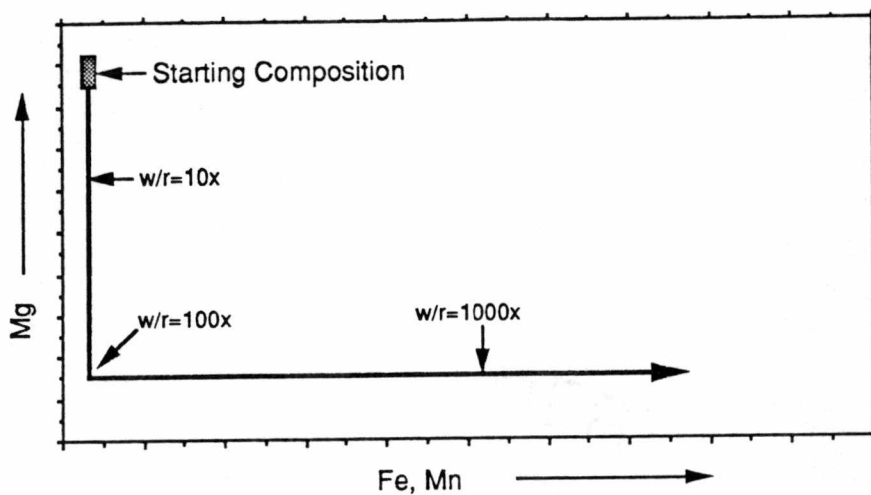
The isotopic evidence from the Nolichucky ooids is inconsistent with stabilization in normal marine water. Therefore, the isotopic data suggest one of three possibilities to explain the difference in isotopic composition of deep vs. shallow-water ooids: 1) the deep-water ooids interacted with more isotopically depleted meteoric water, 2) the deep-water ooids interacted with more isotopically depleted modified marine waters, or 3) the deep-water ooids were stabilized at higher temperatures than shallow-water ooids. The fact that the sedimentological and petrographic evidence does not support significant meteoric interaction in either the shallow-water or deep-water oolitic lithologies rules out the first possibility. If pore waters in the deep-water limestone were strongly controlled by interaction with surrounding clay minerals, it is possible for these fluids to be as much as 7 per mil more depleted in ^{18}O compared to normal seawater based on equilibrium fractionation factors between water, quartz and clays (Foreman and others, in review). HMC undergoing stabilization in these waters would tend to be more depleted in ^{18}O than shallow-water carbonates. However, depleted isotopic compositions could also result from stabilization in normal seawater at elevated temperatures. This is unlikely in the Nolichucky since the thickness of overlying Cambrian and Ordovician rocks increase toward the shallow-water locality (Copper Creek thrust sheet) and should have resulted in higher burial temperatures and lighter $\delta^{18}\text{O}$ for ooids from the shallow-water locality. This is supported by CAI (Epstein and others, 1977; Harris, 1979) and preliminary burial

curves for the east Tennessee area (Walker and Foreman, 1990; unpubl. data). Therefore, the more negative $\delta^{18}\text{O}$ from the deep-water ooids cannot be explained by differences in stabilization temperatures. Based on this evidence, greater interaction between the deep-water ooids and seawater modified by interaction with interbedded shale is the mechanism believed to account for the isotopic differences between the two localities. As mentioned previously, the limestone/shale ratio is significantly greater in the deeper-water setting and individual oolitic limestone beds are much thinner.

The trace element data also agree with the interpretation that deep-water ooids experienced greater interaction with shale-derived fluids. The lower Mg^{2+} from deep-water ooids can be explained by lower Mg/Ca ratios in pore waters during stabilization. Marine fluids in contact with clay minerals will exchange Mg^{2+} for Fe^{2+} resulting in a lower Mg/Ca fluid ratio and increase in Fe^{2+} in the pore water (Drever, 1971; Sayles and Menheim, 1975). This cation exchange mechanism may also explain the higher iron content in deep-water vs. shallow-water ooids. Initially however, most of the exchanged iron will be incorporated preferentially in pyrite rather than the dLMC (Berner, 1980). Though the exchange of Mg^{2+} for Mn^{2+} on clays has not been documented, this mechanism may also explain the higher Mn^{2+} of deep-water ooids. Several examples from the literature note the positive relationship in limestone between Mn^{2+} in calcite and the abundance of clay minerals (Brand and Veizer, 1980; Johnson, 1988).

The trace element and isotopic data suggest that deep-water ooids underwent stabilization in the presence of marine water modified by interaction with interbedded shale. Is it possible that this same fluid was also responsible for stabilization of shallow-water ooids? Less water-rock interaction between the shallow-water ooids and these fluids could result in a dLMC with more Mg^{2+} and less Fe^{2+} and Mn^{2+} , and more positive $\delta^{18}\text{O}$ compared to the deep-water ooids. This is illustrated in Figure 3.11 based on water-rock interaction models of Brand and Veizer (1980, 1981), Veizer (1983), Banner (1988), Meyers (1989) and Banner and Hanson (1990). For the same fluid composition,

Figure 3.11 (next page). Ideal reaction paths for high-magnesium calcite (HMC) during stabilization to diagenetic low-magnesium calcite (dLMC) in terms of degree of water-rock interaction. Arrows point in direction of increasing water-rock interaction. Concepts of model discussed in text. Based on Brand and Veizer (1980, 1981), Veizer (1983), Banner (1988, 1989), Meyers (1989), and Banner and Hanson (1990). **A.** Loss of magnesium with respect to iron and manganese uptake and water/rock ratio. Nolichucky ooids, cements, and echinoderms have experienced enough water-rock interaction that magnesium is invariant. Water-rock ratios are hypothetical. **B.** Iron and manganese uptake with respect to water/rock ratio. X corresponds to HMC iron and manganese concentration. Reaction path X-Y is expected with unlimited supply of Fe and Mn. This path is exhibited by deep-water ooids (Figure 3.6f, 3.7f). Reaction path X-YY indicates initial incorporation of Mn preferentially to Fe (dashed line, Fe availability limited by high redox) followed by constant Mn and increasing Fe. This suggests decreasing redox with a limited supply of Mn during water-rock interaction. Reaction path X-YY seen in shallow-water ooids, cements, and echinoderms. **C.** Isotopic exchange accompanying stabilization from HMC to dLMC. Isotopic compositions of resedimented deep-water ooids may be explained by greater water-rock interaction compared to shallow-water ooids (Figure 3.10) assuming the same fluid was involved with stabilization in both localities. Similarities in $\delta^{13}\text{C}$ for both localities suggests isotopic composition of dissolved carbon was similar to stabilizing phases. Other explanations are also possible (see text for additional details).



increasing water/rock ratio will shift the composition of the dLMC towards more Fe^{2+} and Mn^{2+} , if these elements are available in the pore fluid. Carbon and oxygen isotopic compositions will tend to become more depleted also.

At the present time, we are unable to distinguish with total confidence the effects of water-rock interaction and fluid compositions to explain the difference between dLMC in deep and shallow-water ooids. Nevertheless, the proximity of deep-water oolitic lithologies to interbedded shale supports the interpretation that these ooids experienced greater interaction with fluids modified by interaction with the shale.

Diagenetic Alteration of Ooid Fabrics

The variety of ooid fabrics in the Nolichucky Shale reflect varying degrees of diagenetic alteration since the petrographic and geochemical evidence indicates Nolichucky ooids were probably originally fibrous radial HMC. The present fabrics may be due to modification associated with stabilization of the HMC or due to later burial alteration as suggested by Chow and James (1987) for Cambrian ooids from Newfoundland. Many studies document that stabilization of HMC to dLMC may retain extremely fine fabric detail (Land, 1967; Gomberg and Bonatti, 1970; Towe and Hembelen, 1976; Schlager and James, 1978). In some cases, following stabilization, dLMC may undergo recrystallization resulting in coarser, more equant crystals (Sandberg, 1975; Towe and Hembelen, 1976). However, this is the exception rather than the rule (Richter and Fuchtbauer, 1978; Wilkinson and others, 1985).

Fibrous radial and sparry radial ooids from the same samples in deep-water oolitic lithologies near Oak Ridge were compared in terms of their Mg^{2+} , Mn^{2+} , and Fe^{2+} to determine if a relationship existed between ooid chemistry and degree of alteration from an original fibrous radial fabric (Figure 3.7). Sparry radial ooids have greater Mn^{2+} and

Fe^{2+} compared to fibrous radial ooids, but there is no difference in the amount of Mg^{2+} in these ooids. Referring to Figure 3.11, these compositions suggest sparry radial ooids have seen greater water-rock interaction than fibrous radial ooids, resulting in a coarsening of the original fibrous radial fabric after stabilization to dLMC.

Fibrous radial and sparry radial ooids can occur within several millimeters of each other in the same sample. Can immediately adjacent ooids can experience difference water-rock ratios during diagenesis? Experimental dissolution studies of Mississippian dLMC ooids by Donath and others (1980) indicates that fluid flow in a cemented oolite becomes channelized along partially dissolved ooids resulting in a texture where altered ooids occur immediately adjacent to unaltered ooids. In their study, calcite dissolution was typically localized along concentric laminae or along the interface between the ooid exterior and drusy calcite cement surrounding the ooid. Alteration was facilitated by increased fluid flow in these ooids (Donath and others, 1980) and greater water/rock ratios than in adjacent ooids. In light of that study, it is concluded that it is possible for Nolichucky oolitic lithologies to have experienced different degrees of water-rock interaction on the scale of a thin section. A varying abundance of organic material and crystal defects in fibrous radial cortices probably resulted in variable microporosity and permeability in single ooids and between different ooids (Donath and others, 1980). Increased permeability and porosity in the ooids would result in greater fluid flow and higher water/rock ratios. The petrographic and geochemical evidence (Figure 3.7) suggest that higher water/rock ratios were accompanied by increased recrystallization of the ooid cortex resulting in a coarser fabric (aggrading neomorphism) and increased iron and manganese uptake.

Paragenetic Sequence

From petrographic evidence, which includes cross-cutting relationships and superposition of diagenetic products in thin sections, a paragenetic sequence of diagenetic

events in Nolichucky oolitic lithologies can be determined. This is presented in Figure 3.12 and is divided into early, intermediate, and late burial stages. The transition from early to intermediate burial is arbitrarily chosen as corresponding to the cessation of early diagenetic framboidal pyrite formation; precipitation of ferroan baroque dolomite marks the onset of the late burial stage.

Based on studies of modern marine sediments and pore waters, the formation of early pyrite in Nolichucky basinal sediments (Oak Ridge area) was probably limited to the first several meters to tens of meters of burial (Berner, 1971, 1980). Pore water temperatures were likely to be similar to that of the overlying seawater, ca. 15 °C assuming a water depth of 200 m (Foreman and others, in press) and a surface temperature of 25 °C.

Minimum and maximum temperatures can be constrained for the onset of late burial diagenesis in the Nolichucky Shale based on the observation that baroque dolomite is cross-cut by twinned bed-normal calcite veins which also cut across bed-parallel stylolites. Baroque dolomite is generally thought to form at temperatures ranging from 60-130 °C (Radke and Mathis, 1980). The curved crystal faces characteristic of baroque dolomite suggest precipitation above a critical roughening temperature (approximately 50 °C) at which anhedral crystals are favored thermodynamically (Gregg and Sibley, 1984). Corrected fluid inclusion temperatures for bed-normal calcite veins in the Oak Ridge area reflect near maximum burial conditions for the Nolichucky Shale of 80-110 °C, corresponding to depths of approximately 2.4-3.6 km (Foreman and Dunne, in press). The Nolichucky Shale did not experience significantly greater burial as a result of thrusting during the late Paleozoic Alleghanian Orogeny (Foreman and Dunne, in press). The estimates for maximum burial temperatures agree well with conodont color alteration indices from overlying Ordovician limestone (CAI=1.5-2.0, ca. 60-110 °C, Epstein and others 1977, Harris 1979) and illite crystallinity values (Weaver Index=1.2, ca. less than 150 °C) from underlying Conasauga shales (Weaver, 1984).

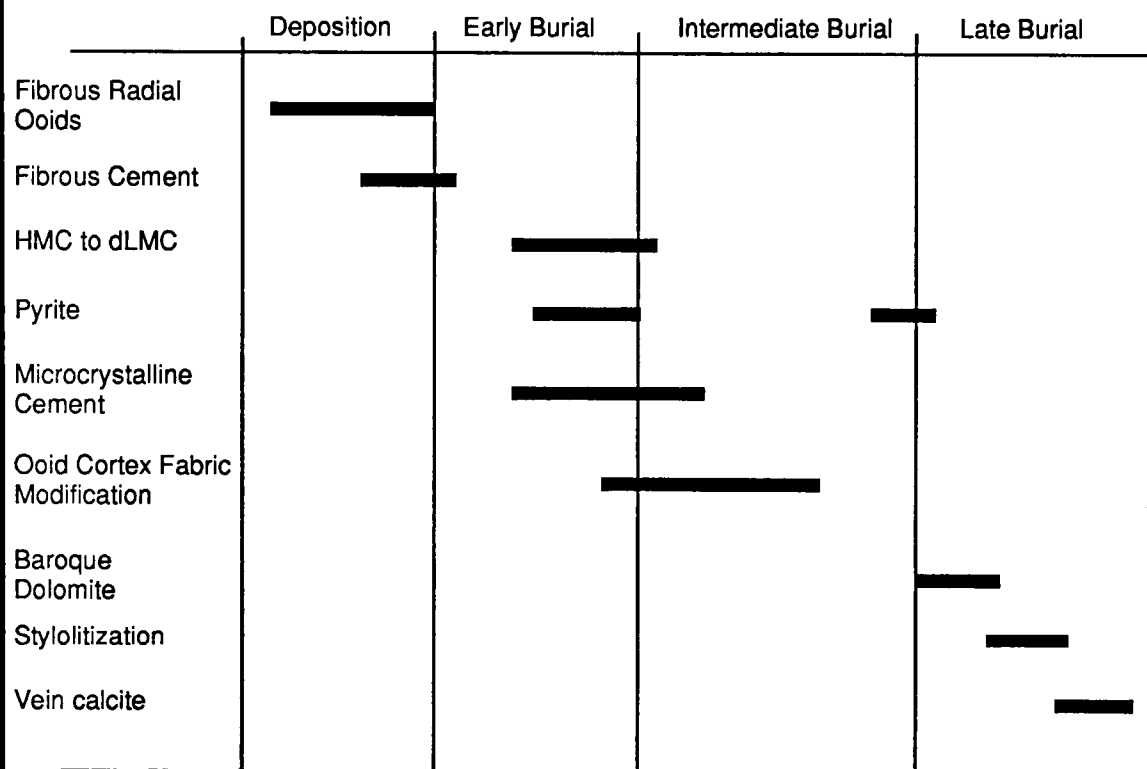


Figure 3.12. Simplified paragenetic sequence for Nolichucky oolitic lithologies (shallow and deep-water). Burial conditions for each burial stage are outlined in text. A model for ooid cortex fabric modification is presented in Figure 3.14.

Based on these data, the following conditions are suggested for burial diagenesis in the Nolichucky Shale: early burial- less than 25 °C, less than several tens of meters depth; intermediate burial- 25-80 °C, tens of meters to 2.5 km depth; and late burial- 80-110 °C, 2.4-3.6 km depth.

A decompacted burial curve for the Nolichucky Shale (Figure 3.13) indicates that the Nolichucky experienced burial conditions corresponding to the intermediate burial stage for most of its pre-thrusting history. During this stage, the Nolichucky oolitic lithologies underwent the greatest diagenetic alteration (Figure 3.14). Stabilization of HMC fibrous radial ooids to dLMC fibrous radial ooids probably occurred relatively early in the intermediate burial stage. Following stabilization, continued water-rock interaction of these dLMC ooids with iron and manganese-rich waters resulted in a progressive coarsening of ooid cortices with increasing uptake of iron and manganese (Figure 3.14). Although Chow and James (1987) do not present geochemical data for the Newfoundland ooids, their petrographic evidence led them to conclude that a similar scenario of progressive fabric alteration of dLMC ooids during burial had occurred in their samples. The fact that sparry radial ooids are replaced by ferroan baroque dolomite and cut by bed-normal veins indicates that this neomorphism of the dLMC ooid cortices occurred before the late burial stage.

CONCLUSIONS

Fibrous radial and sparry radial ooids occur in shallow-water oolite shoal lithologies and resedimented deep-water ooid calcarenite from the Nolichucky Shale. Petrographic and geochemical evidence suggests that these ooids were originally high magnesium fibrous radial calcite ooids. Unequivocal evidence for originally aragonitic and bimineralic ooids is lacking. The ooids are now low-magnesium calcite ooids with approximately 2.5 mol% MgCO_3 . Iron and manganese concentrations in these ooids are 2-3 orders of

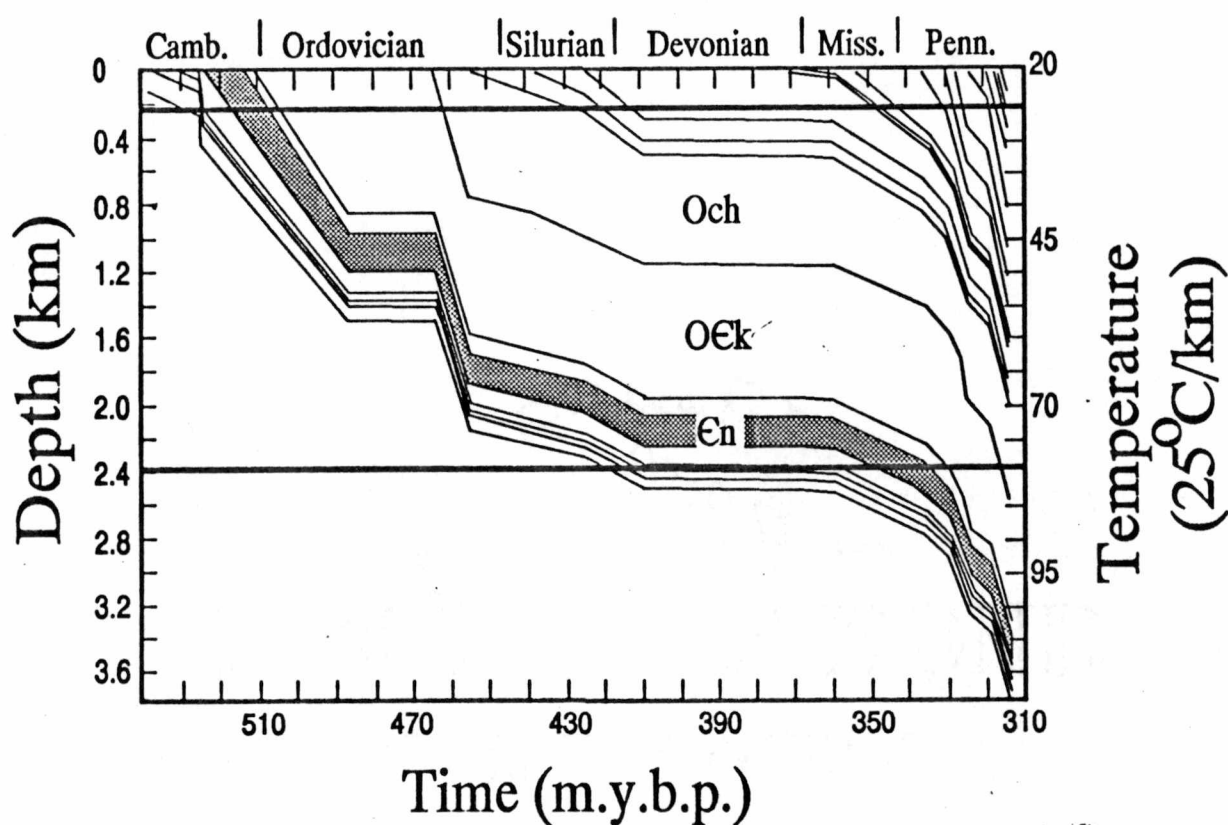


Figure 3.13. Decompacted burial curve for the Nolichucky Shale (stippled interval) in the Whiteoak Mountain thrust sheet. Temperatures are constrained by fluid inclusions in pre-thrusting veins, conodont CAI, and illite crystallinity. Upper and lower horizontal bold lines correspond to speculated maximum burial conditions for the early burial and intermediate burial stages, respectively. Maximum late burial conditions for the Nolichucky in the study probably did not exceed 125-150 °C and 4 km. See text for additional details. Figure modified from Foreman and Dunne (in press).

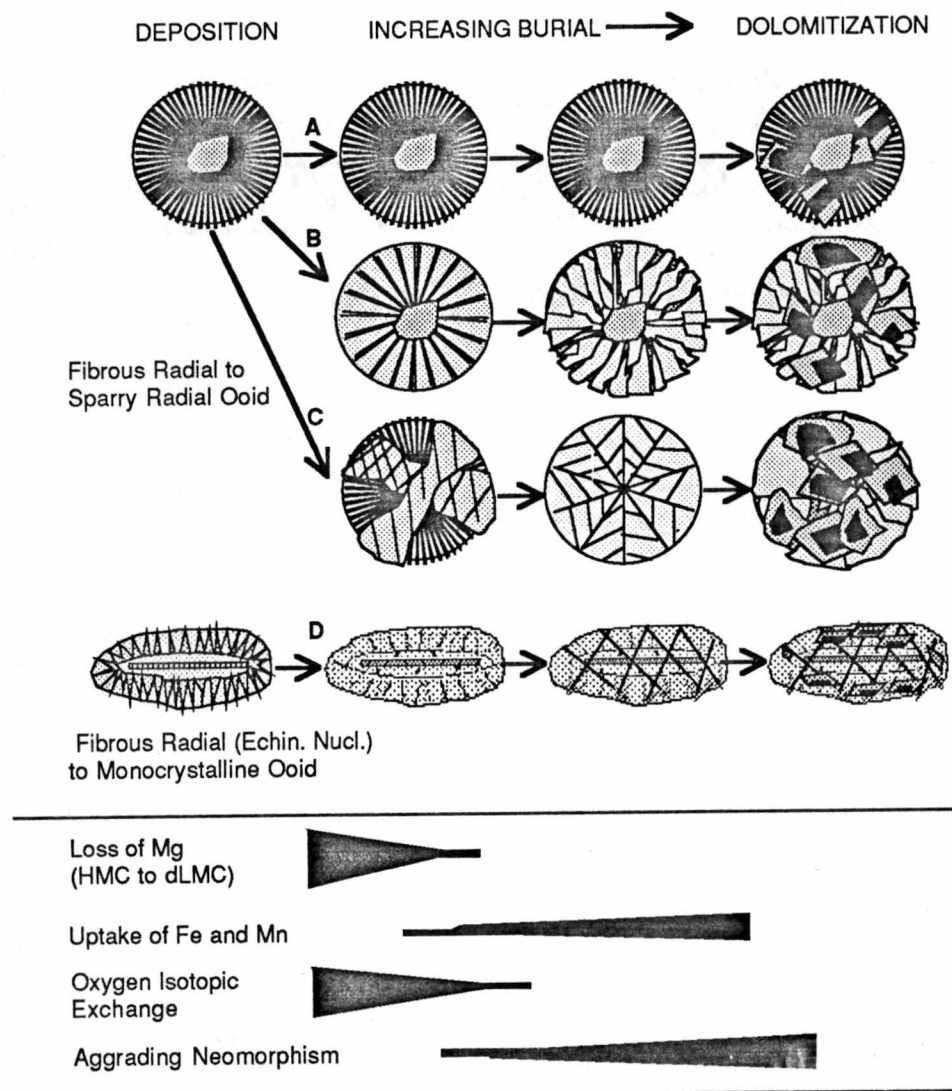


Figure 3.14. Idealized model for Nolichucky ooid cortex fabric modification during burial diagenesis and associated chemical changes (lower part of figure). All ooids are assumed to have originated as fibrous radial ooids. **Path A** indicates little modification of the fabric during diagenesis. **Path B** shows the formation of fibrous radial/sparry radial ooids. **Path C** shows the alteration path for the formation of sparry radial ooids. **Path D** explains the formation of monocrystalline ooids associated with echinoderm nuclei. All stages of fabric modification are seen indicating modification can be arrested at any time. The majority of oxygen isotopic exchange and magnesium loss during stabilization is believed to occur with little fabric alteration. Aggrading neomorphism (coalescence of fibrous crystals) and Fe and Mn uptake probably occur over a protracted period of water-rock interaction.

magnitude greater than expected for unaltered marine calcite. These concentrations reflect stabilization and progressive alteration in the presence of pore fluids modified by interaction with basinal shale. The occurrence of fibrous radial and sparry radial ooids in the same thin section can be explained by different degrees of water-rock interaction which was controlled by small-scale variations in microporosity and permeability.

Despite the general belief that low-magnesium calcite is relatively insensitive to burial alteration, ooids in the Nolichucky Shale show a variety of fabrics reflecting burial alteration of diagenetic low-magnesium calcite. Many of these fabrics might be confused for precursor aragonitic ooids or bimineralic ooids. Geochemical evidence is imperative for resolving the original mineralogy. It is clear that water/rock ratios can be quite variable in a limestone even over distances of several millimeters due to the influence of original fabric on fluid flow. Numerical modelling of water-rock interaction must take into account this characteristic of limestones which may be more typical than previously thought.

CHAPTER 4

CONDITIONS OF VEIN FORMATION IN THE SOUTHERN APPALACHIAN FORELAND: CONSTRAINTS FROM VEIN GEOMETRIES AND FLUID INCLUSIONS

INTRODUCTION

The great abundance of bed-normal joints and veins within foreland thrust sheets reflects the diverse mechanisms that can operate before, during, or after thrusting. Before thrusting, these fractures may develop in response to pre-existing regional stress fields unrelated to subsequent thrusting or as the result of tectonic stress buildup just prior to thrusting (Nickelsen and Hough, 1967; Dunne, 1986; Kilsdonk and Wiltchko, 1988). During thrusting, bed-normal and bed-oblique fractures are more locally distributed in response to buckle folding or passage through frontal and lateral ramps (Price, 1967; Srivastava and Engelder, 1990). Following thrusting, bed-normal fractures commonly develop in rocks that are nearly horizontal. These fractures develop during waning orogenic compression, during erosional unloading with geometries dictated by a residual tectonic stress or contemporary stress field, or during erosional unloading with geometries controlled by rock fabric (Engelder, 1985; Hancock and Engelder, 1989; Dunne and North, 1990). Consequently, a bedding-normal geometry does not uniquely constrain the

Note: This chapter is a paper in press and reflects valuable contributions from the secondary author: Foreman, J.L. and Dunne, W.M., Timing and conditions of calcite veining in the Nolichucky Shale (Upper Cambrian): Whiteoak Mountain thrust sheet, southern Appalachian foreland, east Tennessee: *Journal of Structural Geology*, in press.

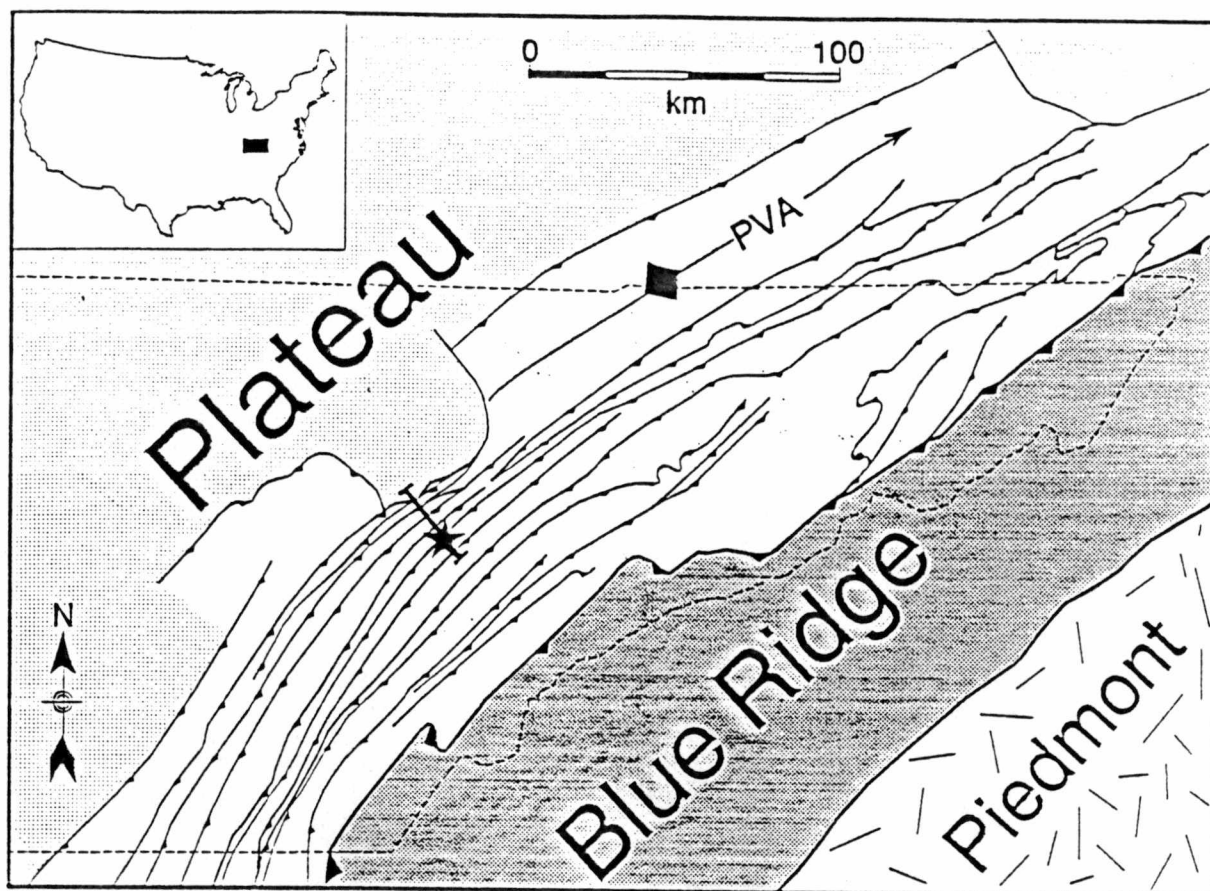
timing of these fractures with respect to thrust sheet formation. Other constraints such as crosscutting relationships of fractures, or environmental conditions (pressure, temperature, fluid composition) derived from geochemical data are required to constrain timing (Hancock, 1985; Kilsdonk and Wiltschko, 1988; Srivastava and Engelder, 1990). In this paper, the following topics will be discussed (1) the use of fracture analysis and fluid inclusion data to elucidate timing relationships between bed-normal veins, bed-parallel veins, and thrust sheet emplacement, (2) controls on bed-normal vein geometries in a regional context, and (3) the reliability of using bed-normal vein geometries in the study of thrust sheet development.

GEOLOGIC SETTING

Bed-normal and bed-parallel calcite veins from the Upper Cambrian Nolichucky Shale were examined in outcrop and drill core on the U.S. Department of Energy Oak Ridge Reservation (ORR) near Oak Ridge, Tennessee (Figure 4.1a). The Nolichucky Shale is about 175 m thick and consists of approximately 75% dark brown to dark green shale, and interbedded 0.3-1.0 m thick, limestone with less abundant calcareous siltstone.

The study area is located in the Whiteoak Mountain (WOM) thrust sheet of the southern Appalachian foreland thrust system (Figure 4.1a). The WOM sheet formed by the forelandward propagation of the thrust system during the Alleghanian Orogeny. Consequently, the WOM sheet overlies the younger Kingston thrust sheet and is overlain by the older Copper Creek thrust sheet (Figure 4.1b). The basal thrust of the WOM sheet is contained within the Lower Cambrian Rome Formation, the regional detachment in the foreland thrust system of the southern Appalachians (Rich, 1934; Rodgers, 1949). Balanced cross-sections across the WOM sheet (Milici, 1970; Woodward, 1985; J. Hopson pers. comm.), demonstrate that the sampled Nolichucky Shale is positioned above

a.



b.

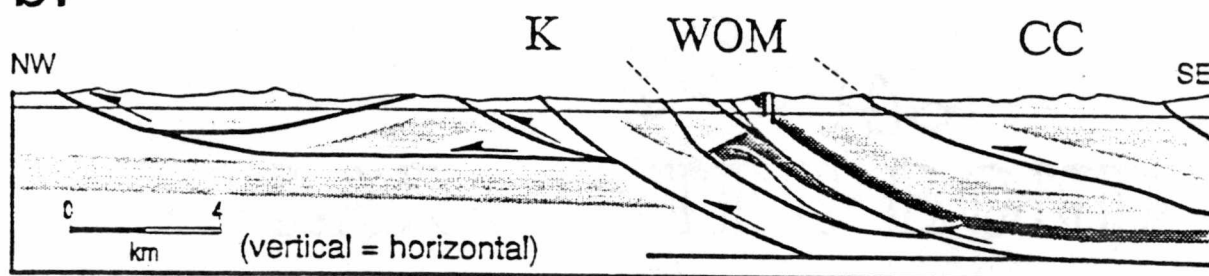


Figure 4.1. Structural map and balanced section. A. Oak Ridge Reservation (star) is located in the foreland of the Valley and Ridge Province (clear pattern), PVA-Powell Valley anticline, teathed lines-thrust faults, thin dashed line-Tennessee state boundary B. Balanced cross-section through the Whiteoak Mountain (WOM) sheet showing position of Nolichucky Shale (dark pattern) and Cambro-Ordovician Knox Group in light stipple. Vertical lines indicate approximate borehole locations. K-Kingston thrust sheet, CC-Copper Creek thrust sheet. Modified from Milici (1970), Woodward (1985), J. Hopson (pers. comm.).

a footwall ramp (Figure 4.1b) and only moved through the lower fault-bend fold during thrusting.

Within the study area, the Nolichucky Shale has an average orientation of 055/50SE. Folding within the Nolichucky Shale is rare in drill cores and involves small-scale folds with amplitudes of less than a meter. The only faulting observed is across rare shear veins with offsets of less than 2 cm.

FRACTURE GEOMETRY AND MORPHOLOGY

The filled-fractures that are the focus of this study are bed-normal and bed-parallel calcite veins. Though the veins are abundant in subsurface samples, they are uncommon in outcrops of the Nolichucky Shale. The absence of veins in outcrop has been attributed to pervasive groundwater flow and surface weathering effects (Dreier and others, 1988). Consequently, vein analysis was limited to core samples only. Bedding and vein orientations in three unoriented cores from less than 350 m depth were corrected with borehole deviation logs (King and Haase, 1987) and outcrop bedding orientation measurements (Lemiszki and Hatcher, 1990). Subsurface orientations correspond to outcrop measurements indicating that vein orientations in the shallow cores were successfully determined. Also, unmineralized fractures were avoided so as to not include coring-induced fractures (Kulander and others, 1979) in the data set. Core and outcrop fracture measurements are listed in Appendix C.

Bed-normal veins occur in four prominent sets (Figure 4.2). Orientations are 055° (parallel to present bedding strike), 320° (nearly perpendicular to present bedding strike), 015°, and 090° when bedding is restored to the horizontal. Kilsdonk and Wiltschko (1988) identified four similar sets in the Powell Valley anticline (Figure 4.1a), of the Pine Mountain thrust sheet, located less than 75 km north of Oak Ridge. They stated that the

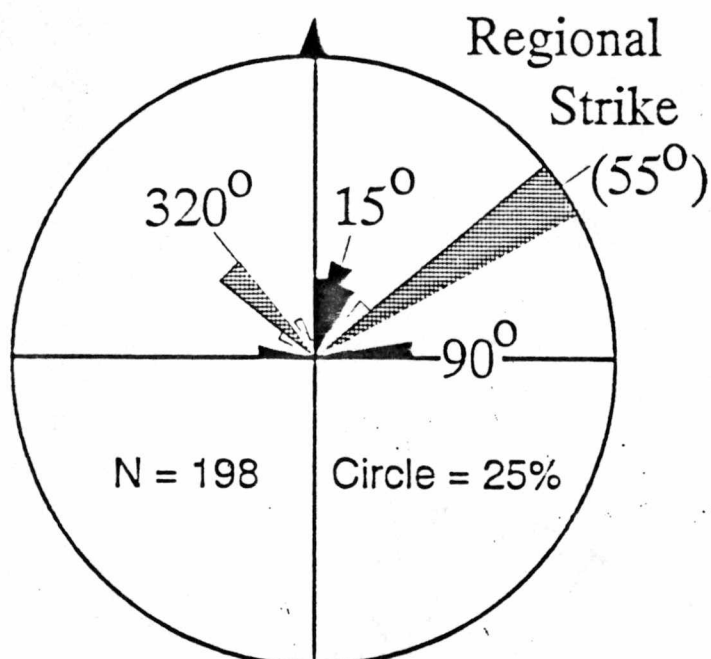


Figure 4.2. Rose diagram of azimuths of bed-normal veins after bedding has been rotated to horizontal. Light-stippled sets (055° and 320°) form orthogonal system to regional Alleghanian strike. Dark-stippled sets (015° and 090°) are sub-orthogonal to each other and oblique to Alleghanian strike.

strike-parallel (055°) and strike-perpendicular (320°) sets are younger than the 015° and 090° sets based on crosscutting relationships and the absence of 015° and 090° veins in post-Ordovician rocks. In contrast, the present study from the Middle Cambrian Nolichucky Shale found that the bed-normal vein sets mutually crosscut each other, indicating they formed coevally. Thirty-five bed-normal vein intersections in the cores show that the 015° and 090° sets lack consistent age relationships with respect to the 055° and 320° sets (Table 4.1).

Bed-normal calcite veins extend across the diameter of the drill core, commonly cut across bedding contacts between shales and interbedded limestones, and are typically 0.5 to 2 mm wide. They contain equigranular ferroan calcite and subordinate quartz and ferroan dolomite. Rarely, the calcite may be fibrous where the host wall rock changes from oolitic limestone to calcareous siltstone. Wall-normal displacements across the veins of transected ooids and fossils, and near wall-normal fibers (Figure 4.3a,b) indicate dilational Mode I fracturing (Engelder, 1982). The bed-normal veins are offset by bed-parallel slickensided veins in 49 of 50 occurrences. The slickenside lineations are parallel to Alleghanian thrusting direction, indicating they formed during thrusting. Calcite within both bed-normal and bed-parallel veins is twinned (Figure 4.3a) as a result of Alleghanian deformation (Wiltschko and others, 1985; Kilsdonk and Wiltschko, 1988; Craddock and van der Pluijm, 1989). Therefore, the four bed-normal sets are coeval Mode I fractures that formed before local Alleghanian thrusting and twinning.

Bed-parallel calcite slickensided veins are present at bedding surfaces between limestones and shale, but also occur within shale. They are commonly less than 1 mm thick with bedding-dip-parallel fibrous slickenlines on both vein margins. The veins are multilayered and when observed in bed-normal profiles parallel to transport, the layers have

Table 4.1. Relative ages of bed-normal calcite veins based on cross-cutting relationships. Table indicates number of times one set cross-cuts and is therefore younger than another. For example, The 090 ° was observed to cross-cut the 015 ° set three times (1st row, 3rd column) whereas the 015 ° set cross-cut the 090 ° set two times (3rd row, 1st column).

OLDER SET	YOUNGER SET			
	015 °	055 °	090 °	320 °
015 °	****	1	3	0
055 °	0	****	3	6
090 °	2	5	****	2
320 °	5	6	2	****

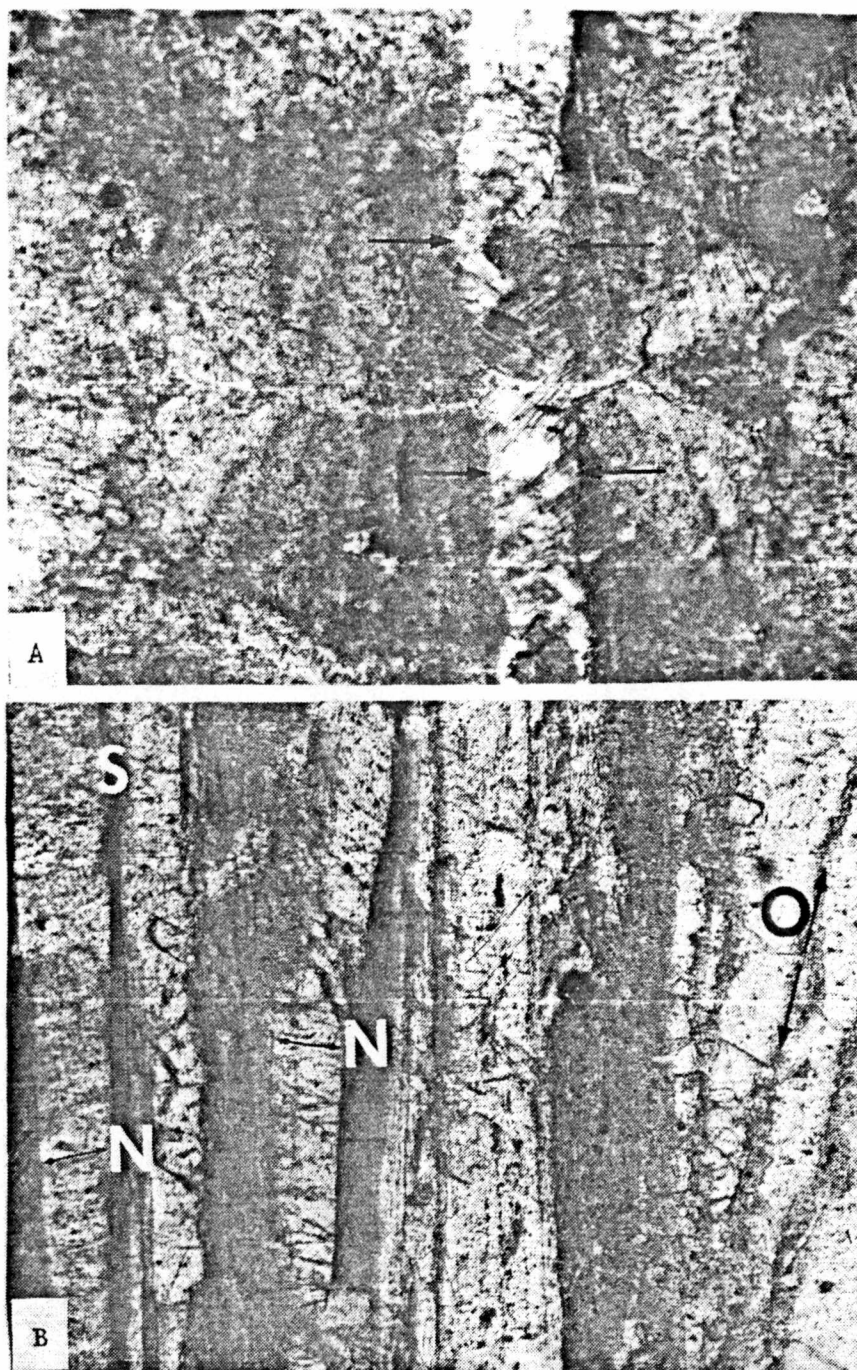


Figure 4.3. Photomicrographs of bed-normal and bed-parallel calcite veins. **A.** Bed-normal extension vein, with twinned equigranular calcite, showing wall-normal offset of radial fibrous ooid. Arrow shows direction of offset. Horizontal field of view is 1.0 mm. **B.** Bed-parallel composite vein containing almost wall-normal fibers (N), wall-oblique trails of shale wall fragments (O), and straight sharp vein boundaries with clay minerals where bed-parallel slip was concentrated (S). Horizontal field of view is 1.2 mm.

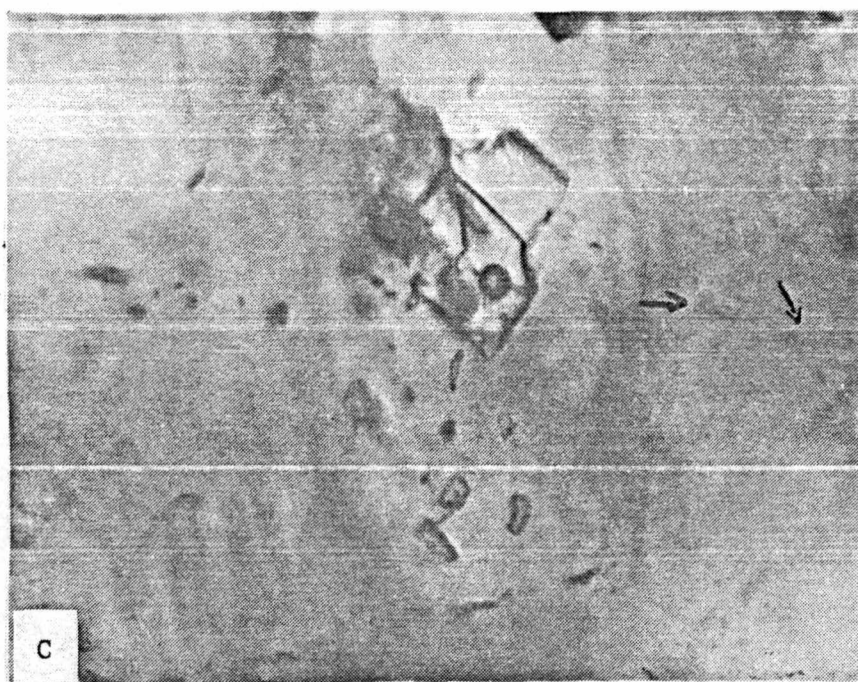


Figure 4.3 (continued) C. Large primary two-phase aqueous fluid inclusion with negative crystal shape in bed-normal vein calcite. Trend of secondary fluid inclusion trails marked by arrows. Horizontal field of view is 150 μm .

three calcite-grain textures (Figure 4.3): (1) Long, layer-parallel fibrous crystals, with uniform extinction and straight boundaries, against thin zones of fine-grained calcite and clay impurities, where bedding-parallel slip was concentrated (Stanley, 1990); (2) Granular, sometimes antitaxial fill, transected by strongly oblique wall-rock fragments, indicating oblique dilation (Cox, 1987); (3) Coarse, equant calcite fill, indicating simple dilation. Consequently, the bed-parallel veins are sites of multiple fracturing that accommodated motions ranging from bed-parallel slip to wall-perpendicular dilation. Locally, the veins are thickened by younger oblique-pinnate veins containing granular calcite that resulted from shear failure of the wall rock adjacent to the slickensided veins.

A set of younger tectonic veins has distinct characteristics: bed-oblique to bed-normal geometries, seldom penetrate more than one bed, cut the bed-parallel and bed-normal veins, may have shear offsets, and greater thickness to length ratios than the bed-normal and bed-parallel veins. Because they are not Mode I fractures and are clearly syntectonic, they are not the focus of this study, and hence, will not be discussed further.

FLUID INCLUSIONS

Methods

Fluid inclusions from twelve oriented vein-bearing samples were examined in doubly polished thin sections prepared using slow speed diamond saws and polishing equipment. Inclusions in crystal cleavage flakes spalled from several larger veins were also examined. Polished chips and unpolished flakes from the same veins had identical homogenization temperatures (T_H), demonstrating that preparation of the chips did not alter the inclusion characteristics.

The ease with which fluid inclusions in soft minerals such as calcite and fluorite

stretch, leak, and decrepitate, requires steps to prevent inclusion deformation during microthermometric analysis (Prezbindowski and Larese, 1987; Meunier, 1989). Inclusion stretching is an increase in inclusion volume, which is accomplished by deformation of the host crystal. Stretching results in T_H values that are greater than undeformed inclusions from the same population. When inclusions leak, fluid leaves the inclusions along microcracks or intersecting surfaces of the polished chip. The fluid loss also results in erroneously high T_H values. Decrepitation is catastrophic failure resulting in the loss of the inclusion content. No microthermometric measurements can be made from these empty inclusions.

The following steps were taken during microthermometric analysis to limit the possibility of unknowingly measuring deformed inclusions. First, only inclusions from the same field of view were measured at a time (generally three to five inclusions), before being replaced with a fresh chip from the same vein. By restricting measurements to inclusions within the same field of view, any sudden change in liquid/vapor ratios due to inclusion deformation can be observed, and removed from consideration. Second, heating runs were conducted before freezing runs to reduce the possibility of measuring homogenization temperatures of inclusions stretched by freezing (Meunier, 1989). Third, only inclusions with maximum dimensions greater than about 15-20 μm were used for the freezing runs because they provided the most reproducible results. Microthermometric analyses were reproducible to within ± 2.0 $^{\circ}\text{C}$. Repeat measurements were conducted with a blind over the thermocouple readout to limit observational bias.

Petrography

Liquid-rich two-phase (liquid + vapor) inclusions occur within bed-parallel and bed-normal calcite veins, have vapor/liquid ratios of 0.10 to 0.20, and do not contain daughter minerals. The maximum dimension of these inclusions ranges from less than 2 to 90 μm and average about 5 μm . Because of the cloudy nature of the calcites and small average

inclusion size, considerable effort was required to find inclusions suitable for reproducible microthermometric measurements.

Primary inclusions must be studied to determine precipitation temperature, fluid salinity and composition during crystal growth. In contrast, evidence from secondary inclusions indicate environmental and chemical conditions following crystal growth. These conditions may be significantly different from those associated with initial formation of the crystal. Thus, differentiating primary and secondary inclusions is extremely important (Roedder, 1984). Unfortunately, an unequivocal, primary origin for fluid inclusions is usually difficult to prove (Roedder, 1984). However, negative crystal inclusion shape, significant and somewhat random separation between adjacent inclusions, and the absence of inclusion trails cross-cutting crystal boundaries and growth zones (Figure 4.3c) suggest a primary origin (Roedder, 1984). Secondary inclusions form planes cross-cutting grain boundaries, are very small and irregularly-shaped, show evidence of 'necking-down', and have variable vapor/liquid ratios (Figure 4.3c) (Roedder, 1984). In this study, samples containing both primary and secondary inclusions yielded identical T_h values, suggesting similar conditions during trapping of both inclusion classes.

Heating Runs

Results from the heating runs for bed-parallel and bed-normal veins show a single population clustered at $T_h = 70 \pm 22$ °C (Figure 4.4). No significant difference in T_h exists between bed-normal veins and bed-parallel veins (Table 4.2). Although both bed-parallel and bed-normal veins have the same homogenization temperatures, pressure corrections presented in a later section indicate that bed-parallel veins formed under greater P-T conditions than bed-normal veins. Homogenization temperatures for the bed-normal and bed-parallel veins are shown in Figure 4.5.

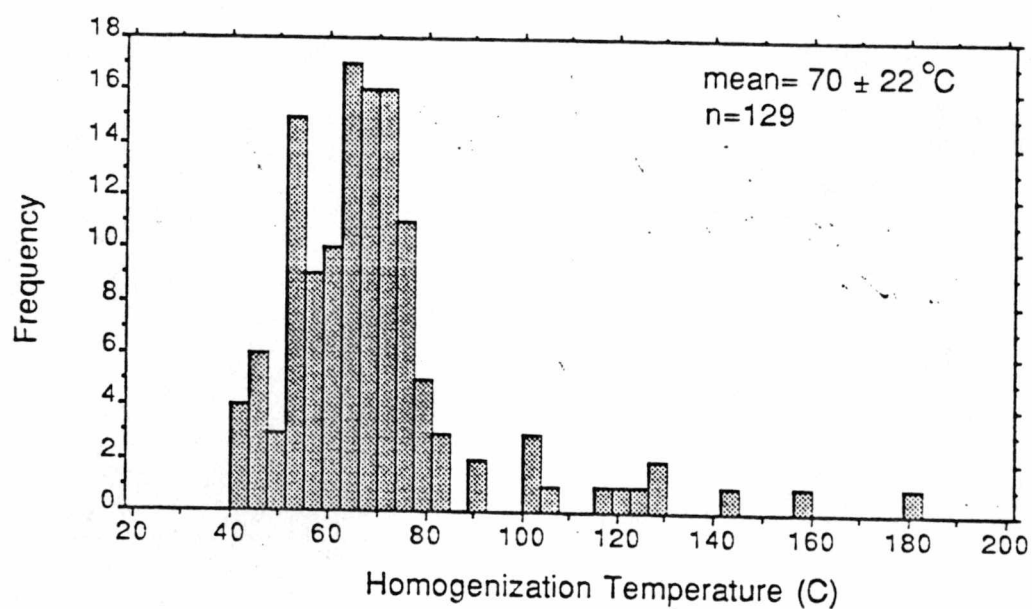


Figure 4.4. Frequency histogram of homogenization temperatures for all bed-normal and bed-parallel veins. Note clustering of homogenization values around 70°C .

Table 4.2. Summary of fluid inclusion data from calcite veins in the Nolichucky Shale on the Oak Ridge Reservation.

Vein Variety	Th	Median	N
bed-parallel	74.2 $\pm$ 34.8	61.1	31
bed-normal	67.6 $\pm$ 16.8	68.5	95
<u>bed -normal. subsets</u>			
055°	66.7 $\pm$ 16.1	67.4	69
320°	66.7 $\pm$ 10.9	68.4	27
015° & 090°	83.4 $\pm$ 24.3	83.7	6

Th- mean homogenization temperature $\pm$ 1 standard deviation

Median- median value for homogenization temperatures

N- number of inclusions measured

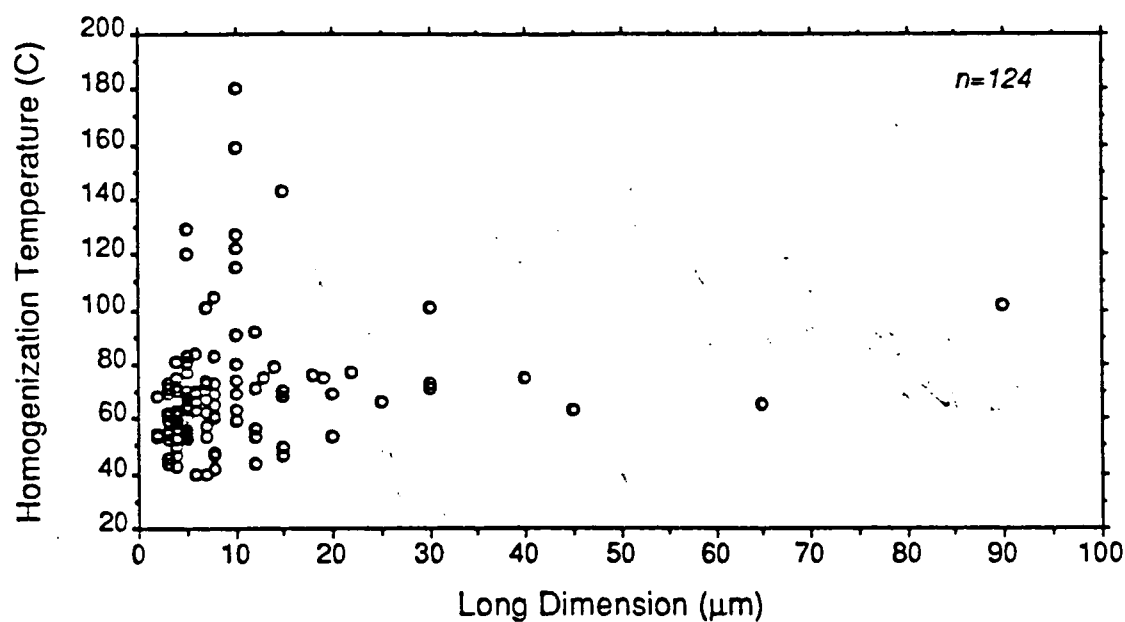


Figure 4.5. Homogenization temperature versus maximum dimension for fluid inclusions in bed-normal and bed-parallel calcite veins. Any apparent trends in data set are not supported by regression analysis. Lack of positive correlation between fluid inclusion long dimension and homogenization temperature suggests fluid inclusions have not been affected by stretching or leaking subsequent to their entrapment. See additional details in text.

Freezing Runs

The inclusion fluid composition can be estimated by observing the temperature at which phase changes occur during thawing of the inclusion from a frozen state (Shepherd and others, 1985). Inclusions required considerable undercooling (generally -70 to -90 °C) before becoming completely frozen, which is commonly the case for high salinity fluids. Freezing was recognized by a brown coloration that migrated across the inclusion, characteristic of NaCl-CaCl₂ fluids (Shepherd and others, 1985).

First melting or eutectic temperatures from inclusions in bed-parallel and bed-normal veins ranged from -57 to -48 °C (mean $T_e = -50 \pm 6$ °C, $n=22$), corresponding to the temperature at which liquid first co-exists with other solids, such as ice and hydrous salts, during thawing. These temperatures are in excellent agreement with published eutectic temperatures for the NaCl-CaCl₂-H₂O system (-55.0 °C Borisenko, 1977; -52.0 °C Crawford, 1981). With continued heating, the next phase transition was ice melting at a mean $T_m = -22 \pm 3$ °C ($n=28$). Further warming resulted in the melting of hydrohalite (NaCl·2H₂O) at a mean $T_{mh} = -8 \pm 4$ °C ($n=16$). Hydrohalite is distinguished from ice by its higher relief and tendency to remain fixed to the inclusion wall rather than the vapor bubble (Shepherd and others, 1985). Also, hydrohalite crystals are nearly volume-constant during temperature decrease, unlike ice crystals which rapidly increase in size. The results of freezing runs for bed-normal and bed-parallel calcite veins are summarized in Table 4.3.

The average inclusion melting behavior was used to estimate the fluid composition in a NaCl-CaCl₂-H₂O phase diagram (Figure 4.6), (Shepherd and others, 1985). The fluid calculated from this method (T_{mh} , Figure 4.6), has an average composition of 17.5 wt% NaCl and 9.5 wt% CaCl₂.

This composition is comparable to sedimentary basin formation waters recognizing that not all dissolved species in the inclusion fluid have been identified or quantified.

Table 4.3. Summary of freezing run results for fluid inclusions from calcite veins in the Nolichucky Shale on the Oak Ridge Reservation. Teut= first melting or eutectic temperature; Tm= last disappearance of ice; Tmh= last disappearance of hydrohalite ($\text{NaCl} \cdot 2\text{H}_2\text{O}$). 1sd= standard deviation. Number in parenthesis indicates number of inclusions.

<u>Vein Type</u>	<u>Teut</u>	<u>1sd</u>	<u>Tm</u>	<u>1sd</u>	<u>Tmh</u>	<u>1sd</u>
Bed Parallel	-46.1 (4)	5.6	-21.3 (5)	5.0	-11 (1)	-
Bed Perpendicular	-50.6 (18)	6.3	-23.1 (23)	2.7	-8.2 (15)	3.5

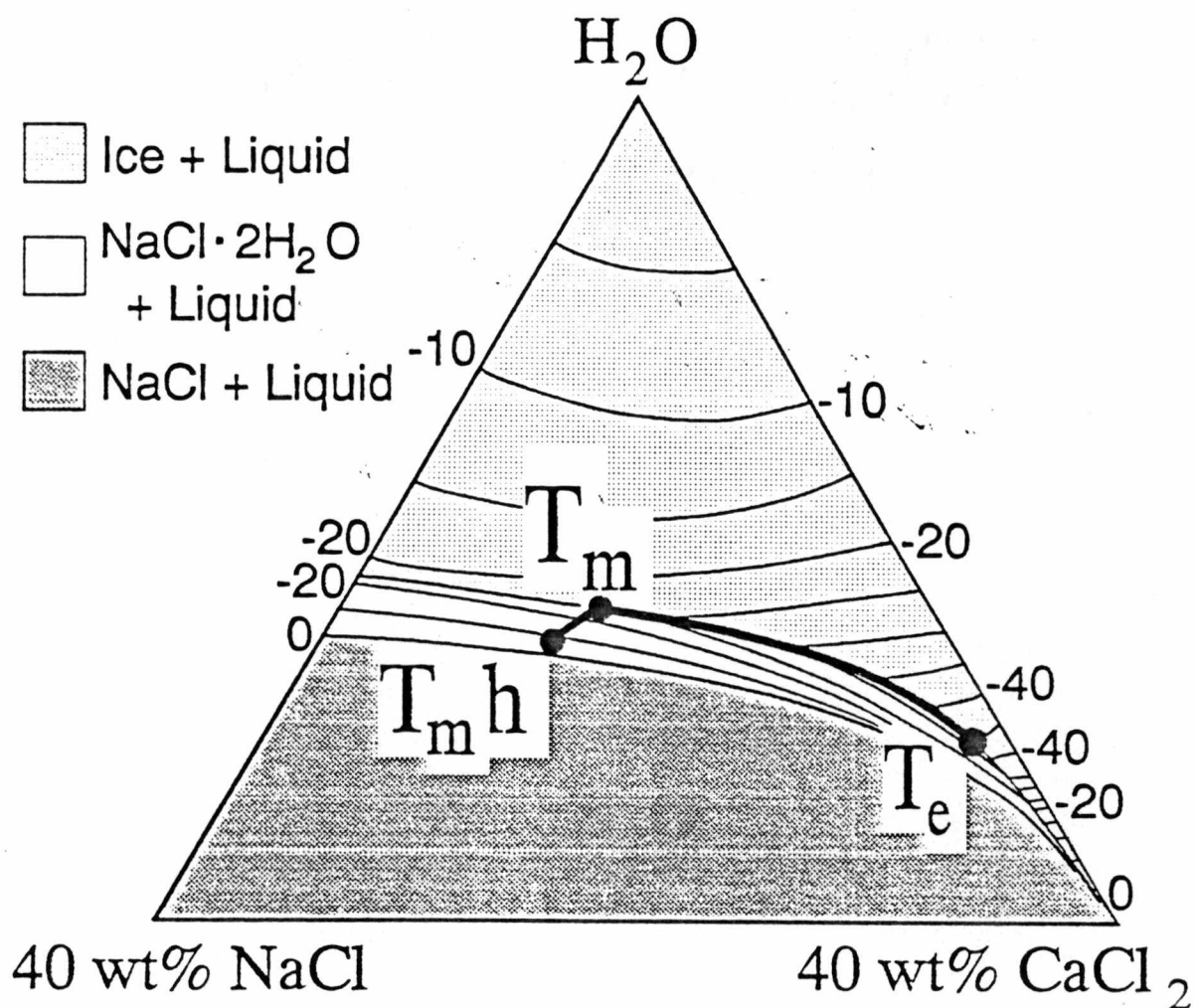


Figure 4.6. Three component phase diagram for the system H_2O -NaCl- CaCl_2 . Melting paths for all inclusions have been averaged and indicated by bold line. Fluid composition at T_{mh} corresponds to 17.5 wt% NaCl and 9.5 wt% CaCl_2 (about 27 wt% NaCl equivalent salinity). Isotherms are labeled in degrees Celsius. T_e = eutectic temperature, T_m = last ice melting temperature, T_{mh} = last hydrohalite melting temperature.

Estimates of inclusion fluid compositions (Figure 4.6) from the Nolichucky Shale veins (TDS=327,500 mg/l, Na=81,000 mg/l, Ca=40,000 mg/l, mol Na/Cl=0.62, and mol Ca/Cl=0.17), are similar to oil-field brines from the Cretaceous Paluxy Formation in central Mississippi (TDS=326,400 mg/l, Na=79,000 mg/l, Ca=34,000 mg/l, mol Na/Cl=0.61, mol Ca/Cl=0.15, Carpenter and others 1974), and brines from the Texas Gulf Coast Oligocene Frio Formation (TDS=105,000-280,000 mg/l, Na=23,000-40,000 mg/l, Ca=10,000-40,000 mg/l, mol Na/Cl=0.54-1.39, and mol Ca/Cl=0.23-0.44, Morton and Land, 1987). The weight ratios of Ca, Na, and Cl to TDS for the Nolichucky inclusion fluids (Ca/TDS=0.12, Na/TDS=0.24, and Cl/TDS=0.65) are remarkably similar to brines from central Mississippi (Ca/TDS=0.10, Na/TDS=0.24, and Cl/TDS=0.61) that probably formed from subareal evaporation of seawater (Carpenter and others, 1974). This correspondence is consistent with hypersaline conditions and minor evaporite formation following Nolichucky deposition in east Tennessee and southwest Virginia (Markello, 1979; Weber, 1988).

DISCUSSION

Pressure-temperature conditions during veining

With the exception of boiling fluids, homogenization temperatures underestimate inclusion trapping temperatures (Shepherd and others, 1985). The amount of correction applied to the homogenization temperature is a function of fluid pressure and fluid density at the time of fluid trapping (Shepherd and others, 1985). In this study, pressure corrections were determined by noting the P-T conditions at which an isochore, or constant fluid density curve (density=1.18 gcm⁻³, T_h=70° C for this model), intersected lithostatic and fluid pressure curves on a P-T diagram (Figure 4.7). This approach follows the

pressure correction technique of Srivistava and Engelder (1990) for bed-parallel and bed-normal veins in the central Appalachians, using the equation of state of Zhang and Frantz (1987) for P-T-X conditions in NaCl-H₂O systems.

A unique P-T value can be estimated for bed-parallel veins since vertical dilation during their formation should require that fluid and lithostatic pressure were equal (Gretener, 1979; Srivistava and Engelder, 1990). Using this argument, bed-parallel veins formed at about 110° C and 90 MPa (3.6 km) (Figure 4.7). In contrast, bed-normal veins with horizontal dilation do not have this pressure constraint, therefore, fluid pressure may have ranged from hydrostatic to lithostatic (Figure 4.7). Trapping conditions for inclusions in bed-normal veins ranged from 80-110° C and 25-90 MPa (2.4-3.6 km, Figure 4.7), reflecting the uncertainty in fluid pressure.

Post-entrapment changes to fluid inclusions

Inclusions in calcite have been demonstrated to stretch, leak, or decrepitate from exposure to temperatures greater than trapping temperatures. The increased magnitude and range of temperatures result from overheating during sample preparation and microthermometric analysis, or from geologic processes such as increased burial (Prezbindowski and Larese, 1987; Burruss, 1989; Meunier, 1989). An inclusion population deformed by overheating shows a positive correlation between inclusion size and homogenization temperature (Prezbindowski and Larese, 1987; Bodnar and Bethke, 1984). In this study, clustering of T_h values about a single mode (Figure 4.3), and the lack of correlation between inclusion size and T_h (Figure 4.5) demonstrates that the measured inclusions are undeformed.

Timing of veining

The timing of vein formation with respect to depth, temperature, and burial rate (Figure 4.8) was modeled using a decompacted burial curve (Sclater and Christie, 1980)

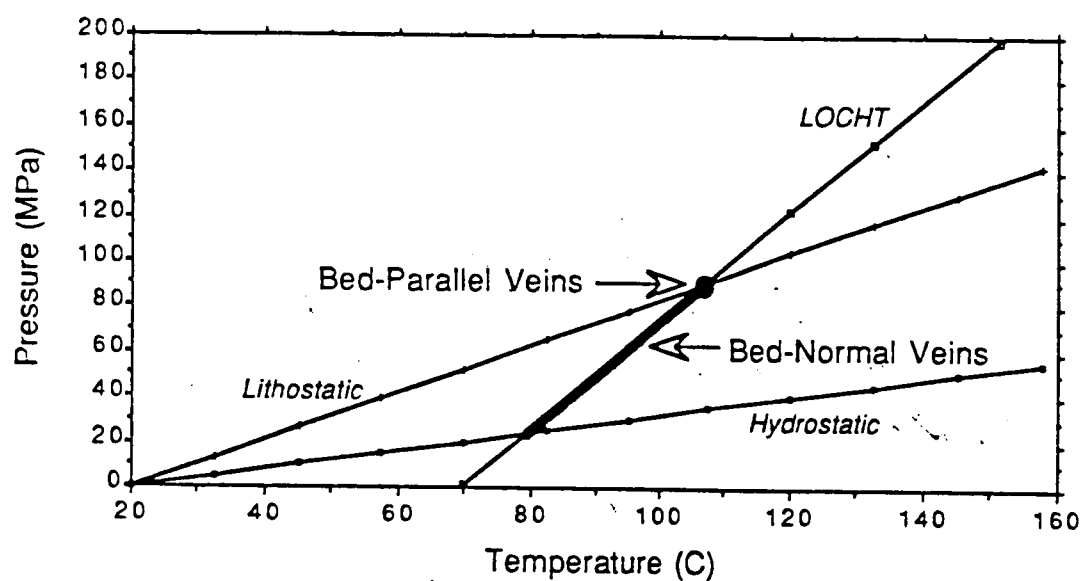


Figure 4.7. P-T diagram used to calculate inclusion trapping temperatures. Bold line segment, line of constant homogenization temperature (LOCHT), indicates calculated P-T conditions for bed-normal vein formation. Solid circle at lithostatic/LOCHT intersection shows calculated P-T conditions for bed-parallel vein precipitation. LOCHT calculated from equation of state of Zhang and Frantz (1987) with $T_h = 700^\circ\text{C}$ and salinity = 27 wt% NaCl equivalent. Lithostatic = lithostatic pressure gradient (26 MPa/km), hydrostatic = hydrostatic pressure gradient (10 MPa/km). geothermal gradient is assumed to be 25°C/km .

for Middle Cambrian to Pennsylvanian strata (Nolichucky Shale in black in Figure 4.8). Stratigraphic thicknesses in the Oak Ridge area were compiled from Hatcher and others (1989) and State of Tennessee Division of Geology maps (Swingle and Luther, 1964; Statler and Sykes, 1970). This burial curve illustrates the time period during which the Nolichucky Shale experienced temperatures equivalent to inclusion trapping temperatures, which are assumed to equal ambient burial temperatures.

The reliability of this burial curve (Figure 4.8), can be evaluated by comparing its predictions to time-temperature-sensitive indicators in the Nolichucky Shale and overlying rocks. Figure 4.8 predicts maximum burial temperatures of 75-90° C for Ordovician limestones (Och). These temperatures are consistent with conodont Color Alteration Index values (CAI=1.5-2.0) from the limestones, indicating 60-110° C temperatures (Epstein and others, 1977; Harris, 1979). Additionally, minerals indicative of the lower to middle shale-diagenesis zone (chlorite, kaolinite, mixed-layer illite/smectite with 20-30% smectite layers, and glauconite), are locally present in the Cambrian Conasauga shales (Lee and others, 1987) and indicate maximum burial temperatures of about 90-200° C (Weaver, 1984). Illite crystallinity values (Weaver Index=1.2) from these shales suggest maximum burial temperatures of less than 150° C (Weaver, 1984, p. 192). These temperatures are consistent with maximum temperatures of 110-125° C modeled by the burial curve for the Cambrian Conasauga Shales (Figure 4.8). These correlations support the interpretation that the Nolichucky Shale did not undergo significantly greater burial during local overthrusting.

Burial temperatures for the Nolichucky Shale during the Late Mississippian and Early Pennsylvanian (Figure 4.8) are in the range of temperatures indicated by the fluid inclusions from bed-normal veins. This correspondence suggests that bed-normal veining occurred after 350 my B.P. (Figure 4.8) and as much as 65 my prior to local Alleghanian thrusting (Secor and others, 1987).

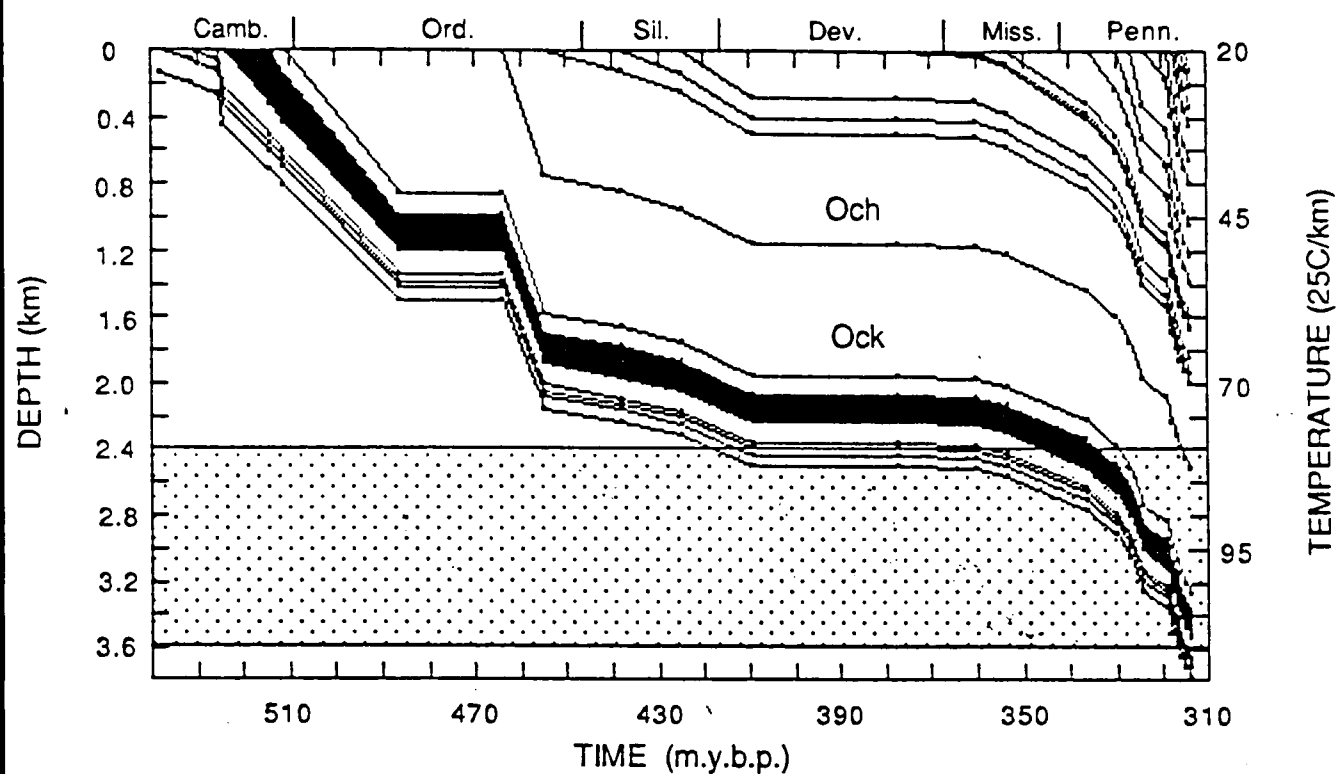


Figure 4.8. Decompacted burial curve for the Conasauga Group and younger strata in the Whiteoak Mountain thrust sheet. Dark pattern=Nolichucky Shale, Ock= Cambro-Ordovician Knox Group, Och=middle Ordovician Chichamauga Group. Stippled band in figure indicates range of corrected inclusion temperatures.

Bed-parallel veins formed contemporaneous with thrusting at a depth equivalent to maximum burial for the Nolichucky Shale (about 3.6 km, 110° C, Figure 4.8). However, the fluid inclusions in pre-tectonic bed-normal veins and tectonic bed-parallel veins did not experience significantly greater temperatures after entrapment because of a lack of inclusion deformation due to overheating (Figure 4.5). The absence of evidence for resetting of homogenization temperatures is consistent with the interpretation that veining in the Nolichucky Shale occurred close to maximum burial depths and temperatures. This interpretation is supported by numerous fluid inclusion studies from calcite collected at depth in oil fields where the homogenization temperatures reflect maximum burial temperatures (Prezbindowski and Larese, 1987; Burruss, 1989). Therefore, the Nolichucky Shale did not undergo significantly greater burial due to thrusting. This lack of greater burial may have resulted from the WOM thrust sheet being nearly at the tip of the thrust system (Figure 4.1b). In this position, local thrust wedge thickness, which is controlled by the rate of thrust sheet emplacement versus erosion rate, probably did not have to exceed the maximum sedimentary thickness prior to thrusting.

Regional stress history

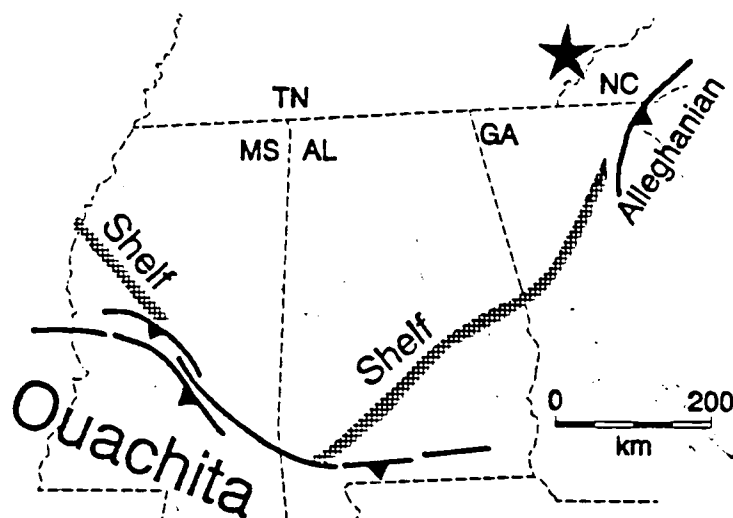
Crosscutting relationships, fracture geometry and abundance, fluid inclusions, and the use of a decompacted burial curve, demonstrate that the four bed-normal vein sets in the WOM cores formed: (1) before local thrusting that formed the WOM sheet; (2) under similar thermal conditions and from similar fluids; and (3) coevally during the late Mississippian and middle Pennsylvanian. At that time, bedding would have been nearly horizontal, and bed-normal veins would have formed as vertical fractures. The existence of four coeval bed-normal Mode I vein sets requires that the regional minimum principal stress fluctuated between four different horizontal directions because Mode I fractures form perpendicular to the minimum stress (Pollard and Aydin, 1988). Any proposed regional

stress regime during bed-normal veining has to account for this fluctuation in the horizontal stress directions. Two possible causes for the stress regime in the Oak Ridge area during bed-normal veining are considered: (1) passage of peripheral bulges and foreland basins; and (2) far-field orogenic stresses extending beyond the limits of thrusting.

First, the four sets could have resulted from the passage through Oak Ridge of two peripheral bulges and two foreland basins (Beaumont, 1981; Jordan, 1981) from the coeval Ouachita and Alleghanian orogenic loading events (Figure 4.9; Thomas, 1985). Both the bulges and basins could have caused outer-arc extension with vertical fracturing in response to flexure and basin sagging, respectively. Certainly, the Alleghanian peripheral bulge and foreland basin migrated through the study area because the site is presently in the Alleghanian thrust system. In contrast, the Ouachita bulge and basin probably did not reach Oak Ridge. Peripheral bulges commonly reach only 50-400 km beyond a thrust system with an upper limit of about 500 km (Stockmal and others, 1986; Fleming and Jordan, 1989), whereas Oak Ridge is nearly 600 km beyond the Ouachita deformation front (Figure 4.9). Also, Beaumont and others (1988) modeled the Ouachita orogenic loads as producing a bulge within 300 km of the Ouachita thrust system, which is considerably south of Oak Ridge. For these reasons, the Ouachita peripheral bulge and basin were not responsible for bed-normal fracturing in the study area.

The migration of the Alleghanian foreland basin could still have a role in the formation of the bed-normal veins, although, characteristics of foreland basins suggest that the migrating basin was not responsible. The passage of a foreland basin does not produce classical outer-arc extension by basinal sagging with joints and faults that are parallel to the depositional axis (Price, 1974), because foreland basins do not depress symmetrically. Instead, they are asymmetric with migrating depositional axes, subsidence and uplift, and basin-margins that migrate with respect to the depositional axes (Fleming and Jordan, 1990). The net result of these processes is not outer-arc extension but length-reduction by erosion during bulge movement from episodic thrusting or lithospheric relaxation, or

a. Late Mississippian



b. Middle Pennsylvanian

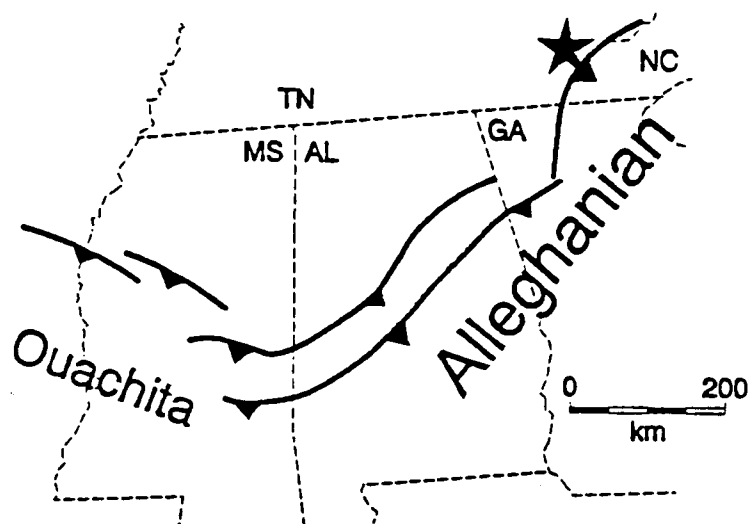


Figure 4.9. Orientation of major orogenic fronts in the Lower Mississippian and Middle Pennsylvanian. A. Reconstruction of Ouachitan and Alleghenian orogenic belts (thrust fault symbols) in the late Mississippian. B. Middle Pennsylvanian, the Alleghenian orogenic system has encroached on the Oak Ridge area. Modified from Fig. 5, Thomas (1985).

horizontal shortening during basin incorporation into the thrust system. Thus, the passage of the Alleghanian foreland basin would not have caused bed-normal fracturing at Oak Ridge.

Still, flexure during passage of the Alleghanian peripheral bulge could have caused bed-normal fracturing. Bulge flexure is a proposed mechanism for Mode I fracturing beyond a thrust system which produces fractures parallel to the thrust system (Turner and Hancock, 1990). Bulge flexures have amplitudes of 200-400 km, quarter wavelengths of 50-200 km (Beaumont and others, 1981; Fleming and Jordan, 1989), and arc-length increases of 0.0003-0.00005 %. Such small extensions do not unequivocally necessitate axial-parallel vertical fracturing during bulge migration. Therefore, flexure by the Alleghanian bulge could have formed the 055° vein set which is parallel to regional Alleghanian structural strike, but certainly would not have formed the remaining three coeval sets.

Consideration of far-field orogenic stresses as a cause of the coeval veining is based on the unusual location of Oak Ridge during the upper Paleozoic. At this time, it was proximal to the two coeval and non-parallel Ouachita and Alleghanian thrust systems (Figure 4.9). Vertical Mode I fractures beyond the limits of one thrust system are commonly orthogonal with one set parallel and the other set perpendicular to the trend of the system. This geometry reflects the control of the orogenic stress on the horizontal principal stress directions during nearly hydrostatic horizontal stress conditions (Stauffer and Gendzwill, 1987; Dunne and North, 1990). Considering the study area, the 055° and 320° sets are certainly parallel and perpendicular to the Alleghanian structural trend, respectively, and therefore, their orientation could have resulted from a horizontal stress geometry that was controlled by Alleghanian orogenic stress. The greater abundance of the 055° set over all other sets (Figure 4.2) may have resulted from the increasing proximity of the Alleghanian thrust system to the Oak Ridge area through time (Figure 9), or

alternatively, the passage of the Alleghanian peripheral bulge.

The coeval 015° and 090° bed-normal veins are subperpendicular and subparallel to the Ouachita thrust system, respectively, and their orientations may have been controlled by Ouachita orogenic stress. Such an interpretation would account for the fluctuations in the horizontal stress directions at Oak Ridge during the formation of the four coeval sets. The cause would have been fluctuations in the magnitude of the far-field Alleghanian and Ouachita orogenic stresses. This interpretation is preferred because it is consistent with the regional tectonics, fracture geometry, and fracture timing. However, it is still speculative because it is inferred from a single site and the 015° and 090° sets are not ideally orthogonal with respect to the Ouachita thrust system.

Vein timing relative to thrusting: Limitations of geometry

Srivastava and Engelder (1990) have clearly demonstrated a fracturing history for carbonate rocks passing through a ramp as part of the Nittany Mountain horse in the central Appalachians. Bed-parallel veining was followed by bed-normal veining parallel to bedding strike and then, bed-normal veining perpendicular to bedding strike. This sequence was demonstrated from crosscutting relationships of veins, reactivation of stylolites as fractures, and the results of their fluid inclusion analysis where younger bed-normal vein sets developed in cooler conditions, as a result of erosion contemporaneous with transport over the ramp. It is interesting to compare their results to the vein sets in the Whiteoak Mountain thrust sheet at Oak Ridge.

The same vein geometries as in the Nittany Mountain horse exist in the WOM thrust sheet, which was elevated to a similar height above a ramp but only passed through the lower fault-bend fold. The comparable elements in the WOM sheet are bed-parallel veins, bed-normal veining parallel to bedding strike (055° set), and bed-normal veining perpendicular to bedding strike (320° set). Yet, crosscutting relationships and fluid inclusion data demonstrate a different, dominantly pre-thrusting sequence. Coeval, pre-

thrusting bed-normal fracturing producing four coeval vein sets was followed by thrust-related bed-parallel veining. The difference in veining sequence for similar vein patterns in analogous structural settings, argues against using fracture geometries alone to reconstruct fracture history.

CONCLUSIONS

(1) Within the Whiteoak Mountain thrust sheet, the Nolichucky Shale contains four bed-normal vein sets that predate both bed-parallel slickensided veins and emplacement of the WOM sheet. Restoring bedding to a pre-thrusting attitude, the bed-normal vein sets trend 015° , 055° (regional strike direction), 090° , and 320° (sub-perpendicular to regional strike).

(2) Fluid inclusions indicate that bed-normal calcite veins precipitated at temperatures of 80 - 110° C, 2.4 - 3.6 km, and from highly saline NaCl-CaCl₂ brines. Bed-parallel veins formed during maximum burial at 110° C, 3.6 km, precipitating from brines of similar composition and salinity as bed-normal veins.

(3) Fluid inclusions and a decompacted Nolichucky burial curve suggest bed-normal veins formed between the late Mississippian and early Pennsylvanian during the onset of the Ouachitan and Alleghanian orogenies, but prior to movement of the WOM sheet. The trends of the bed-normal veins may have been controlled by far-field compressive stresses related to the orogenies, which resulted in four sets. The increased abundance of the 055° set may have resulted from passage of the Alleghanian peripheral bulge prior to thrusting. Bed-parallel slickensided veins formed contemporaneously with thrusting of the Whiteoak Mountain sheet.

(4) Comparison of these results with those of Srivistava and Engelder (1990) clearly illustrates that the geometry of bed-normal and bed-parallel veins, by itself, does not constrain the timing of fracture formation with respect to thrusting.

CHAPTER 5

ISOTOPIC AND FLUID INCLUSION EVIDENCE FOR WATER-ROCK
INTERACTION IN INTERBEDDED LIMESTONE AND SHALE DURING CALCITE
VEINING: UPPER CAMBRIAN NOLICHUCKY SHALE, EAST TENNESSEE.

INTRODUCTION

The factors controlling deep-burial fluid-rock interaction in sedimentary rocks are important research topics because many of the final properties of these rocks are a result of burial diagenetic processes. For example, hydrologic and hydrogeochemical characteristics of shallow uplifted sedimentary rocks (porosity, permeability, and mineral solubilities) are a result of previous fluid-rock interaction, much of it during deep burial. The successful prediction of economic deposits in sedimentary rocks is dependent on understanding the fluid system prior to and subsequent to emplacement of these deposits.

The research reported here is part of an ongoing research effort at the University of Tennessee into the comparative diagenesis of the Cambrian Conasauga Group and the Middle Ordovician Chickamauga Group of the southern Appalachian foreland fold and thrust belt (Foreman and Walker, 1989; Foreman and Weber, 1989; Steinhaff, and others, 1989; Foreman and Dunne, in press). In this paper, a model will be presented for fluid-rock interaction in the Nolichucky Shale (Conasauga Group, Upper Cambrian), east

Note: This chapter is a paper in review and reflects valuable contributions from the secondary authors: Foreman, J.L., Jacobs, G.K., Walker, K.R., and Dunne, W.M., Isotopic and fluid inclusion evidence for water-rock interaction in interbedded limestone and shale during calcite veining: Upper Cambrian Nolichucky Shale, east Tennessee: *Geochimica et Cosmochimica Acta*, in review.

Tennessee, using stable isotope and fluid inclusion data from calcite veins and host rocks. The evolution of the Nolichucky fluid system, from horizontally stratified with limited vertical fluid flow to cross-formational fluid flow along open fractures, will be related to the development of Alleghanian-age tectonic fractures. Evidence will also be presented to show that fracturing significantly effected the nature of isotopic exchange between burial fluids and surrounding rocks.

GEOLOGIC SETTING

Stratigraphy

This study reports on fluid inclusion and isotopic data from subsurface samples collected in the Nolichucky Shale (Upper Cambrian), at the U.S. Department of Energy Oak Ridge Reservation (ORR), near Oak Ridge Tennessee (Figure 5.1a). The Nolichucky Shale is one of five formations of the Conasauga Group (Middle to Upper Cambrian), which consists of alternating shale-rich and limestone-rich units (Figure 5.1b), deposited in an intracratonic basin (Hasson and Haase, 1988). The Conasauga Group occurs stratigraphically between thick siliciclastic and carbonate rocks of the Rome Formation (Lower Cambrian) and massive limestone and dolostone of the overlying Cambro-Ordovician Knox Group. In the study area, the Nolichucky Shale is 150-175 m thick, and is predominantly shale with 0.1-0.5 m thick interbeds of limestone and calcareous siltstone (Weber, 1988; Foreman and others, in press). Approximately 75 percent of the Nolichucky consists of illitic shale and calcareous siltstone, the remainder being oolitic, peloidal, skeletal, and intraclastic limestone (Figure 5.2). The amount of limestone decreases downward in the Nolichucky Shale from the overlying Maynardville Limestone (Figure 5.2). Directly underlying the Nolichucky Shale is the Maryville Limestone (Middle

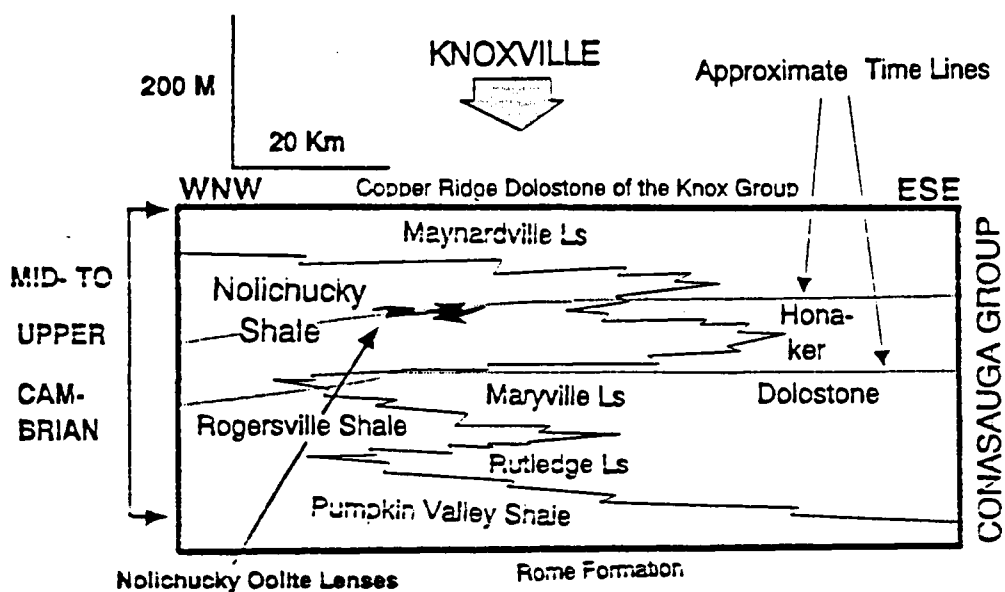
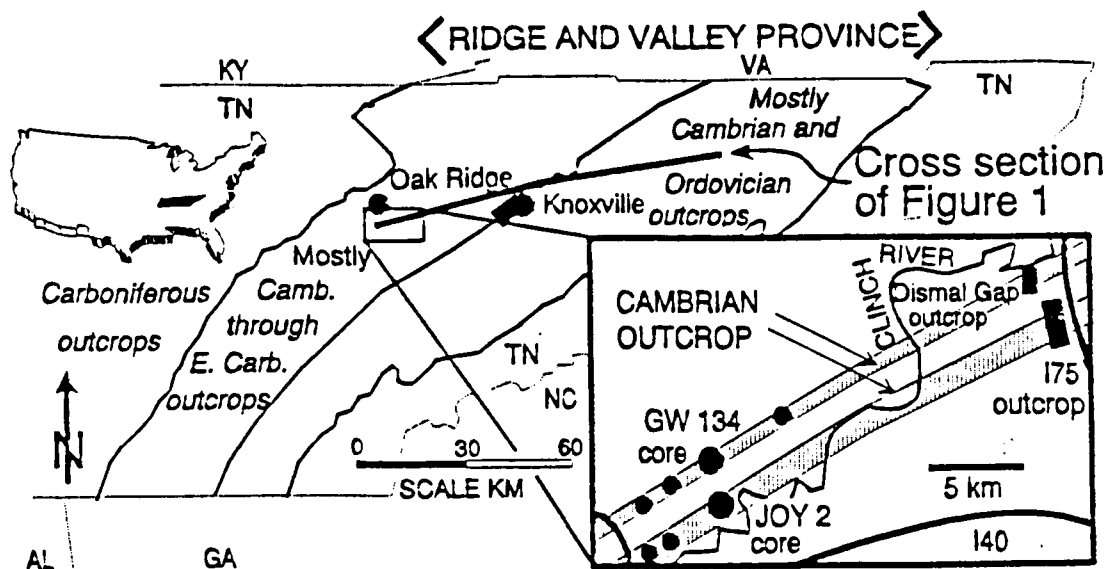


Figure 5.1. Locality and general stratigraphy of study area. A. Map of outcrops (filled rectangles) and cored boreholes (filled circles) examined. Northwest and southeast stripped bands in inset figure correspond to Conasauga Group outcrops in the WOM and Copper Creek thrust sheets, respectively. From SW to NE in the WOM sheet, the boreholes are GW137, GW117, GW134, and GW130. B. Highly-generalized cross-section oriented nearly perpendicular to depositional strike for Conasauga Group members in the vicinity of Knoxville, TN. Approximate locations of shallow-water oolite from the Copper Creek thrust sheet indicated. Modified from NW-SE cross-section of Rodgers (1953).

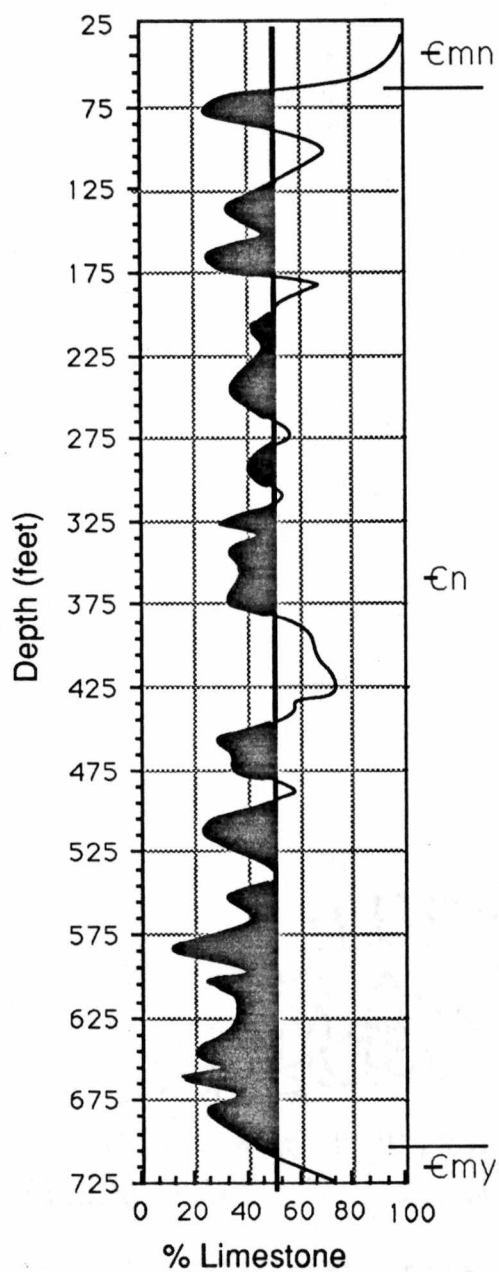


Figure 5.2. Relative abundance of carbonate and siliciclastic lithologies in the Nolichucky Shale, based on lithologic logging from Borehole GW134. Shading indicates intervals in which siliciclastic rocks (shale and siltstone) are more abundant than limestone.

Cambrian) which is 100-125 m thick. The Maryville Limestone is dominantly limestone with less shale than the overlying Nolichucky. Depositional fabrics in the Maryville Limestone indicate a shallow to deep subtidal origin (Kozar, 1986). The Maynardville Limestone (Upper Cambrian) overlies the Nolichucky, is 100 m thick, and consists of massive, stylolitic algal limestone and dolostone (Weber, 1988). Dolomitized oolitic-oncolitic and peloidal limestone is also present in the Maynardville, providing strong evidence for a shallow-subtidal to supratidal origin (Weber, 1988). Collapse breccias and evaporite pseudomorphs (calcite replacing anhydrite) in the Maynardville of nearby southwest Virginia provide good evidence that hypersaline conditions and evaporite formation occurred during Maynardville deposition in that area (Markello and Read, 1981).

During Nolichucky time, deposition of shale and siltstone in the intracratonic basin was punctuated by event-deposition related to storm processes (Weber, 1988; Foreman and others, in press). These storm events and resultant turbidite transport resulted in basinal deposition of allochthonous oolitic-peloidal sands and limestone intraclasts derived from up-slope settings. A southwest to northeast traverse across depositional strike shows a decrease in the shale/limestone ratio to the northeast indicating closer proximity to areas of carbonate sediment production (Figure 5.1b). Limestone beds become more oolitic and fossiliferous towards the northeast; shallow subtidal-intertidal algal stromatolitic and thrombolitic limestone also become more abundant to the northeast (Weber, 1988).

Structural Geology and Vein Geometry

The ORR is situated in the western Valley and Ridge province at the narrowest part of the Appalachian foreland fold-and-thrust belt. Two thrust sheets underlying the ORR formed during Late Paleozoic northwest-directed Alleghanian thrusting, the forelandward Whiteoak Mountain (WOM) thrust sheet and the hinterlandward Copper Creek thrust sheet.

Calcite veins are common in the Nolichucky Shale, however, averaged over its thickness, they constitute less than one volume per cent. Three types of calcite veins have been recognized in the Nolichucky within the WOM sheet on the basis of their orientation with respect to bedding, geometry, and relative timing. Bed-normal veins are the oldest vein sets and are cross-cut and offset in a direction parallel to tectonic transport by bed-parallel slickenside veins (Foreman and Dunne, in press). Slickenside lineations on the bed-parallel veins are oriented northwest-southeast and indicate movement to the northwest. Bed-oblique veins cross-cut the bed-normal and bed-parallel veins and are therefore the youngest set.

Cross-cutting relationships, combined with vein fabrics and geometries, can be used to date the different vein sets with respect to Alleghanian deformation within the WOM sheet. Bed-normal veins are twinned (Figure 5.3a) and offset in a northwest direction by bed-parallel veins and are clearly pre-Alleghanian (Foreman and Dunne, in press). The margins of bed-parallel veins have slickenside lineations consisting of fibrous calcite which are parallel to vein walls (Figure 5.3b). These fibers are aligned parallel to the Alleghanian transport direction (southeast-northwest) which suggests they are syn-tectonic. Bed-oblique veins are the youngest veins and have shear offsets compatible with a tectonic origin.

METHODS

Fluid Inclusions

Two-phase fluid inclusions occur within pre-tectonic bed-normal, tectonic bed-parallel, and tectonic bed-oblique calcite veins in the Nolichucky Shale. Microthermometric measurements were conducted to determine the temperature, salinity, and composition of

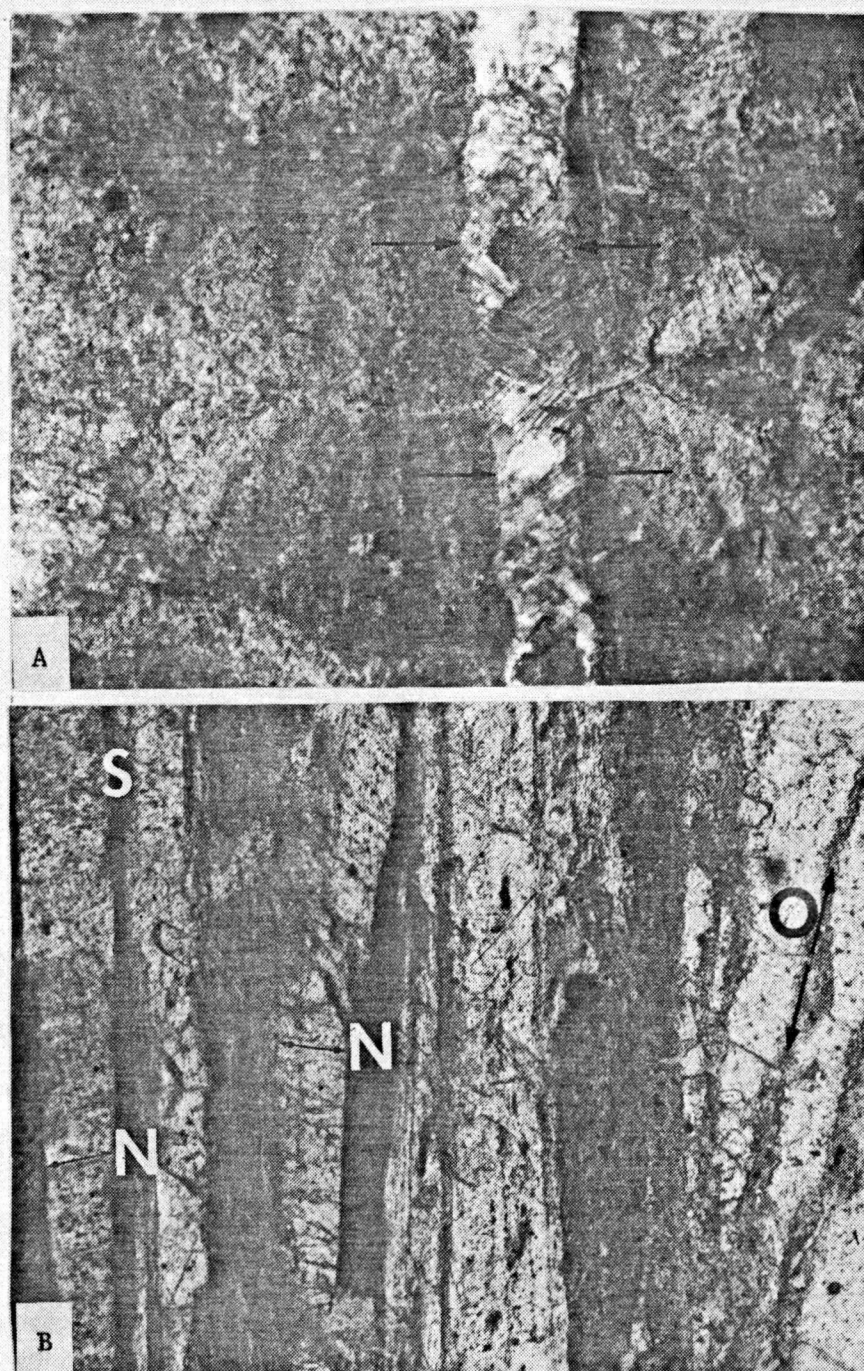


Figure 5.3. Photomicrographs of bed-normal and bed-parallel calcite veins. **A.** Bed-normal extension vein, with twinned equigranular calcite, showing wall-normal offset of radial fibrous ooid. Arrow shows direction of offset. Horizontal field of view is 1.0 mm. **B.** Bed-parallel composite vein containing almost wall-normal fibers (N), wall-oblique trails of shale wall fragments (O), and straight sharp vein boundaries with clay minerals where bed-parallel slip was concentrated (S). Horizontal field of view is 1.2 mm.

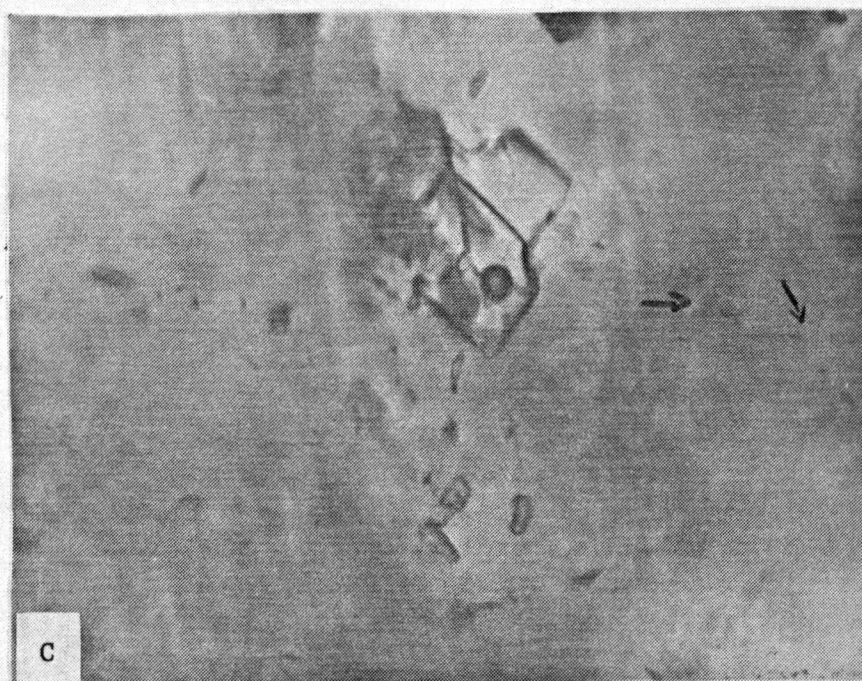


Figure 5.3 (continued) C. Large primary two-phase aqueous fluid inclusion with negative crystal shape in bed-normal vein calcite. Trend of secondary fluid inclusion trails marked by arrows. Horizontal field of view is 150 μm .

fluids responsible for calcite vein formation. Fluid inclusions were examined from twelve oriented, vein-bearing samples using standard preparation techniques for doubly polished chips (Roedder, 1982). Samples were collected from drill core obtained from three boreholes (GW134, GW139, GW140) in the Whiteoak Mountain thrust sheet on the ORR (Figure 5.1a). Whenever possible, cleavage flakes spalled from vein calcites were used for fluid inclusion study to reduce the possibility of damaging inclusions while preparing the doubly polished chips. Fluid inclusions from two veins were measured using both cleavage flakes and doubly polished thin sections. In each case, no significant difference in homogenization temperatures was noted, demonstrating that inclusions were not effected by sample preparation.

All heating and freezing measurements were conducted on a USGS-type modified heating-freezing stage manufactured by FLUID INC. During the microthermometric analyses, fluid inclusions were observed with a high resolution black-and-white monitor. Temperatures inside the stage were displayed on an LED thermocouple readout. Freezing runs were conducted after heating runs to reduce the possibility that inclusions stretched during freezing would produce erroneous homogenization temperatures (Meunier, 1989). Only inclusions visible within the same field of view were measured at one time and a new chip was used after the microthermometric measurements for these inclusions were completed. When possible, repeat analyses during heating and freezing were conducted on each inclusion. These repeat measurements were made with a blind over the thermocouple readout to reduce observational bias. Reproducibility was typically better than ± 2 °C.

Stable Isotopes

The carbon and oxygen isotopic composition of calcite veins was determined from micro-samples (0.2-1.5 mg) collected with a micro-drill. Eight veins sampled for isotopic

study were also examined for fluid inclusions resulting in a matched isotopic and fluid inclusion data set. Vein calcites were reacted with 104 % H_3PO_4 at 55 °C on a common acid micro-line and corrected to 25 °C based on corrections to in-house calcite standards reacted in the same group as the vein calcites. The CO_2 extracted was analyzed on a VG 903 stable isotope ratio mass spectrometer at the University of Tennessee Science Alliance Stable Isotope Laboratory. In-house calcite standards (HO-CC) run against NBS 20 (Solenhofen Limestone) were analyzed with the unknowns and the reproducibility was determined to be better than ± 0.2 per mil for both carbon and oxygen. Duplicate analyses for two samples conducted eleven months later produced results that were within 0.2 per mil for oxygen and 0.5 per mil for carbon. Additional isotopic analyses of limestone and calcite veins, determined by Haase and others (in prep.), from the Nolichucky Shale and other members of the Conasauga Group in the study area, are included in the present data set.

Calcite and dolomite occur in interbedded shale in a weight proportion of approximately 1:5. Carbon dioxide from the calcite and dolomite was extracted by timed stepwise reaction (Al-Aasm and others, 1990) from bulk shale samples (75-125 mg) crushed in an agate mortar, and then reacted with 104 % H_3PO_4 at 25 °C. The grain size of the shale prior to reaction with H_3PO_4 was tightly constrained by sieving (150-180 μm) to reduce errors resulting from grain size effects on reaction rates (Al-Aasm and others, 1990). The powders were dried at 120 °C for 2 hrs prior to reaction with H_3PO_4 . Because of the paucity of organic matter in the Nolichucky Shale (<2 wt%, Lee and others, 1987), no attempt was made to remove it from the powders prior to reaction with H_3PO_4 . Carbon dioxide was removed from the reaction vessels for isotopic analysis after 2 hrs (calcite), 6 hrs (calcite + dolomite), and 30 hrs (dolomite). Noncondensable gases were removed each time following the removal of CO_2 . The isotopic compositions of the CO_2 extracted after 6 hrs were generally intermediate in composition to that of CO_2 extracted after 2 and 30 hrs, and reflects the contributions of $\text{CO}_{2\text{calcite}}$ and $\text{CO}_{2\text{dolomite}}$. The calcite-

H₃PO₄ and dolomite-H₃PO₄ fractionation factors for oxygen were assumed to be equal. Uncertainties in the dolomite-H₃PO₄ fractionation factor at 25 °C are not likely to exceed 0.5 to 1.2 per mil with respect to the measured and true $\delta^{18}\text{O}$ values (Land, 1980; Rosenbaum and Sheppard, 1986).

RESULTS

Petrography

Peloidal-oolitic limestone and intraclastic limestone are the most abundant Nolichucky limestone lithologies in the study area (Weber, 1988; Foreman and others, in press). Framework grains and micrite matrix make up the majority of the limestone; cements filling interparticle porosity between grains comprise less than 10 vol% of the limestone. The majority of these cements are fibrous calcite, forming isopachous coatings on detrital grains and probably precipitated in a marine environment (Foreman and Weber, 1989). Pore-filling phases characteristic of meteoric or burial cements are rare in the ORR limestone indicating that nearly all of the depositional porosity in the limestone was occluded during shallow burial by marine waters.

Ferroan calcite and ferroan baroque dolomite are typically associated with burial diagenesis (Gregg and Sibley, 1984; Choquette and James, 1987). These minerals occur mostly as replacement phases in oolitic limestone in the study area. Both phases selectively replace ooids; a precursor aragonite or magnesium calcite mineralogy for these ooids may have facilitated their replacement in the presence of modified marine waters or burial fluids (Foreman and Weber, 1989). Similar examples of Upper Cambrian ooids containing evidence for similar precursor mineralogies have been recorded from Newfoundland (Chow and James, 1987).

The replacement of ooids by ferroan calcite and baroque dolomite occurred prior to stylolitization as evidenced by truncation of replaced ooids at stylolites. Stylolitization during burial compaction is the last diagenetic event prior to calcite vein formation in the limestone and probably resulted in the occlusion of any remaining porosity.

Stylolites are cross-cut by bed-normal calcite veins, clearly indicating that veining occurred after stylolitization. Bed-normal calcite veins consist of equidimensional ferroan calcite crystals that always contain at least one generation of twins (Figure 5.3a). These twins indicate that the bed-normal veins formed prior to Alleghanian deformation (Foreman and Dunne, in press). Bed-parallel veins consist of twinned equigranular ferroan calcite, fibrous ferroan calcite oriented normal to the vein wall, and fibrous ferroan calcite oriented parallel to the vein wall (Figure 5.3b). Fibrous calcite, oriented parallel to the vein wall, is commonly found on the outer margins of the bed-parallel veins.

Authigenic phases in Nolichucky shale lithologies, observed petrographically and with back-scatter electron imaging, include pyrite, ferroan calcite, ferroan dolomite, quartz overgrowths, and potassium feldspar overgrowths. Ferroan dolomite is the most abundant authigenic carbonate phase and consists of euhedral rhombs with iron zoning that can be observed petrographically through staining. The cores of these rhombs are typically non-ferroan whereas the rims are highly ferroan. These crystal morphologies are typical of dolomite formed at temperatures less than approximately 50 °C (Bathurst and Land, 1986). The age of authigenic calcite and dolomite formation in the shale relative to veining is equivocal because veins were not observed juxtaposed against these phases in the shale. However, calcite veins do cross-cut ferroan calcite and dolomite in limestone and siltstone which, because of these lithologies proximity to the shale, may suggests that vein formation post-dated the precipitation of these phases in the shale.

Fluid Inclusions

In order for analyses of fluid inclusions to accurately reflect conditions during crystal growth, primary fluid inclusions must be distinguished from those formed or modified following crystal growth (secondary and pseudosecondary inclusions; Roedder, 1984; Shepherd and others, 1985). The following features suggest a primary origin: 1) negative crystal inclusion shape, 2) significant and somewhat random separation between adjacent inclusions, and 3) absence of evidence for inclusions which cross-cut crystal boundaries or growth zones (Roedder, 1984). Evidence for secondary inclusions include: 1) inclusions distributed along planes that cross-cut crystal boundaries, 2) small and irregularly shaped inclusions, 3) inclusions that show evidence of 'necking-down', and 4) inclusions with variable vapor/liquid ratios (Roedder, 1984). Based on the criteria listed above, approximately 85% of the fluid inclusions observed in this study are primary. Figure 5.3c illustrates typical primary fluid inclusions examined. However, no significant difference in homogenization temperatures was observed between primary and secondary inclusions.

Inclusions in the calcite veins are liquid-rich, two-phase (vapor + liquid) inclusions with vapor/liquid ratios of 0.10 to 0.20 (Figure 3c). Daughter minerals were not observed in any inclusions. The maximum dimension of inclusions range from less than 2 μm to 90 μm and average 5 μm .

Homogenization temperatures (T_h) for the bed-normal and bed-parallel calcite veins suggest minimum precipitation temperatures of approximately 70 $^{\circ}\text{C}$ (Figure 5.4a; Appendix D). There is no significant difference in the T_h of bed-normal compared to bed-parallel veins (bed-normal $T_h = 67.6 \pm 16.8$ $^{\circ}\text{C}$, $n=95$; bed-parallel $T_h = 74 \pm 34.8$ $^{\circ}\text{C}$, $n=34$). A histogram of homogenization temperatures from the bed-oblique vein indicates two modes (70 and 180 $^{\circ}\text{C}$ with an average T_h of 157 ± 52 , $n=77$; Figure 5.4b). The 70 $^{\circ}\text{C}$ mode is in the temperature range of the bed-normal and bed-parallel veins whereas the

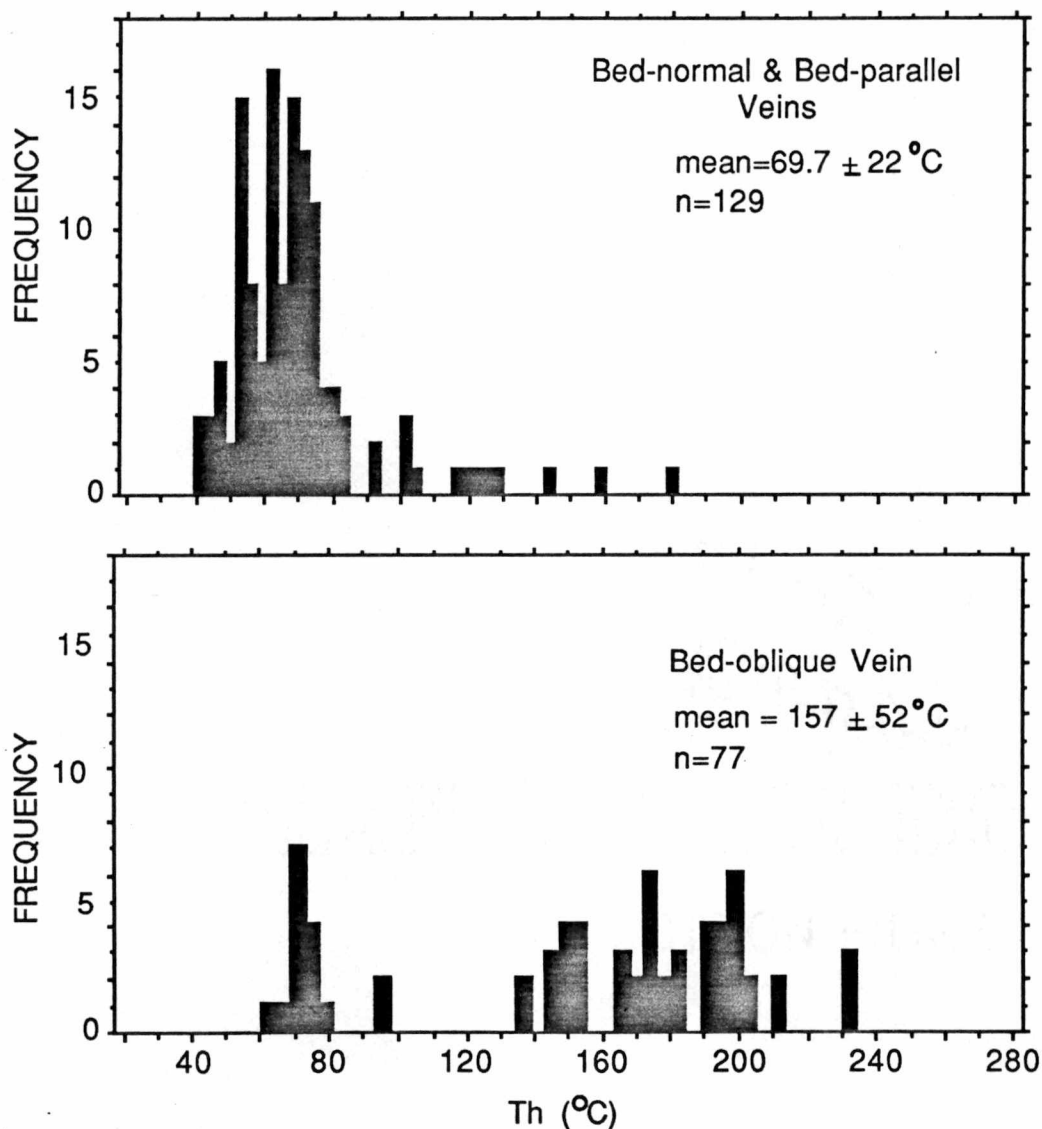


Figure 5.4. Homogenization temperatures from fluid inclusions in Nolichucky calcite veins. A. Frequency histogram of homogenization temperatures (T_h) from fluid inclusions in bed-normal and bed-parallel veins. B. Frequency histogram of homogenization temperatures (T_h) from fluid inclusions in single bed-oblique vein.

mode at 180 °C indicates a separate population of fluid inclusions characterized by higher homogenization temperatures. Re-examination of the rock billet from which this sample was prepared indicated that the polished chip consisted of a bed-normal vein included in the bed-oblique vein. The bed-normal vein had been engulfed by the bed-oblique shear vein during formation of the later vein.

Evidence against resetting of the inclusion homogenization temperatures by aquathermal expansion include: 1) the relatively close clustering of homogenization temperatures for the bed-normal and bed-parallel veins (Figure 5.4a), 2) separate inclusion populations in Figure 5.4b, and 3) absence of a positive correlation between inclusion size and T_h (Figure 5.5), . Aquathermal expansion and its adverse effects on the inclusions in these calcite veins would be expected if the inclusions were subjected to temperatures significantly greater than trapping temperatures, either as a result of increased burial or higher heat flow (Prezbindowski and Larese, 1987; Bodnar and Bethke, 1984).

Freezing runs were conducted on the fluid inclusions to estimate fluid composition and salinity during calcite vein formation. Inclusions in all veins required considerable undercooling (-70 to -90 °C) before becoming completely frozen, suggesting the inclusion fluids are saline brines (Shepherd and others, 1985). A brown discoloration was observed to sweep across the inclusions as they froze, a behavior typical of NaCl-CaCl₂ fluids (Shepherd and others, 1985). The first occurrence of liquid upon thawing (eutectic temperature, T_e) for bed-normal and bed-parallel veins average -50 ± 6 °C, (n=22). These eutectic temperatures are in excellent agreement with published eutectic temperatures for the NaCl-CaCl₂-H₂O system (-55 °C Borisenko, 1977; -52 °C Crawford, 1981). Eutectic temperatures for the bed-oblique vein averaged -43.2 ± 5.3 °C (n=5) and may reflect the presence of other dissolved salts such as potassium and magnesium chlorides. Ice melting was the next phase transition to occur in bed-normal, bed-parallel, and bed-oblique veins (bed-normal and bed-parallel veins: $T_m = -22.8 \pm 3.2$ °C, n=28; bed-oblique vein: $T_m = -27.2$

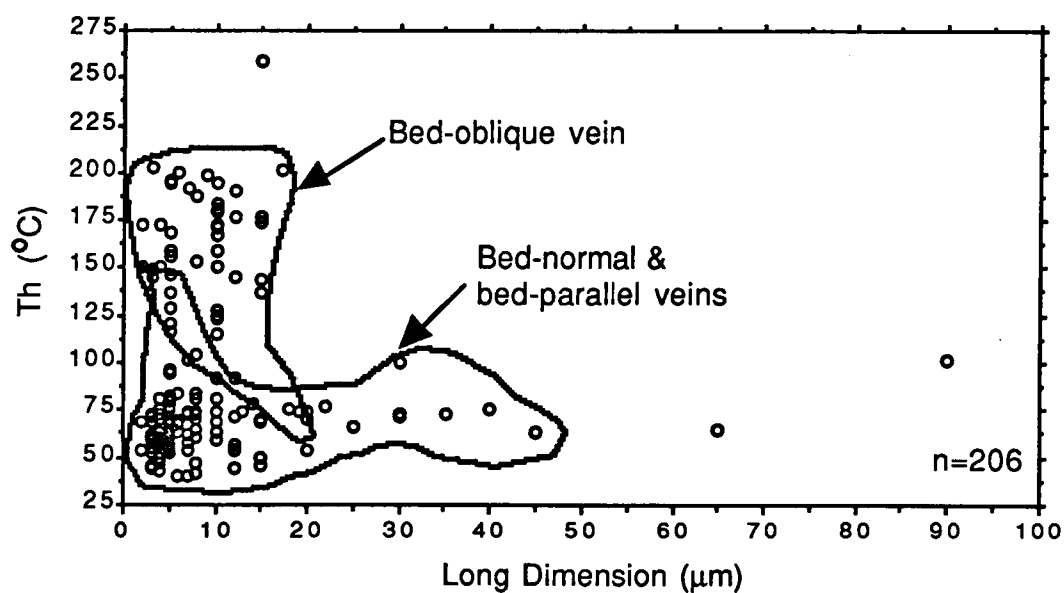


Figure 5.5. Homogenization temperature (T_h) as a function of long dimension of fluid inclusions for all calcite veins examined. The bed-normal and bed-parallel veins fall in the lower field whereas the upper field corresponds to inclusions from the single bed-oblique vein. Several outliers were not included in the fields.

± 2.2 °C, $n=9$), followed by hydrohalite ($\text{CaCl}_2 \cdot 2\text{H}_2\text{O}$) melting (bed-normal and bed-parallel veins: $T_{mh} = -8.4 \pm 3.5$ °C, $n=16$; bed-oblique vein: $T_{mh} = -8.1 \pm 1.9$ °C, $n=6$).

The eutectic temperatures for all inclusions and presence of hydrohalite demonstrates that the $\text{NaCl-CaCl}_2\text{-H}_2\text{O}$ system can be used to approximate the composition of these inclusion fluids based on their melting behavior (Shepherd and others, 1985). Hydrohalite is distinguished from ice by its higher relief and tendency to remain fixed to the inclusion wall rather than the vapor bubble (Shepherd and others, 1985). Also, hydrohalite crystals are nearly volume-constant during temperature decrease, unlike ice crystals which rapidly increase in size. The average melting path for these inclusions, constrained by the temperatures at which hydrohalite and ice melts, is shown on an $\text{NaCl-CaCl}_2\text{-H}_2\text{O}$ ternary phase diagram (Figure 5.6). The location of T_{mh} on the melting path in Figure 5.6, is used to calculate a final fluid composition using the Lever Rule (Shepherd and others, 1985). The calculated fluid composition is 17.5 wt% NaCl and 9.5 wt% CaCl_2 .

Fluid inclusion salinities are commonly reported in terms of NaCl wt% equivalent salinity on the basis of the freezing point depression (T_m). In mixed Ca-Na-K-Mg chloride solutions the relative error in estimating salinity using the freezing point method of NaCl equivalents is $< 5\%$ (Shepherd and others, 1985). The approximate NaCl wt% equivalent salinities for vein-forming fluids were calculated from the freezing point depression-salinity curve of Potter and others (1978) for NaCl solutions, extrapolated to 30 wt% NaCl equivalent salinity (Figure 5.7). The difference in average T_m for bed-normal and bed-parallel veins ($T_m = -22.8 \pm 3.2$, $n=28$) compared to the bed-oblique vein ($T_m = -27.4 \pm 2.2$, $n=9$) is significant at the 99% confidence interval and indicates that the single bed-oblique vein precipitated from fluids with approximately 2 wt% greater NaCl-equivalent salinity. There does not appear to be a relationship between homogenization temperature and salinity for the vein-forming fluids (Figure 5.7). The average fluid density for all the inclusions is approximately 1.18 g/cm^3 based on the final melting temperature of ice, and

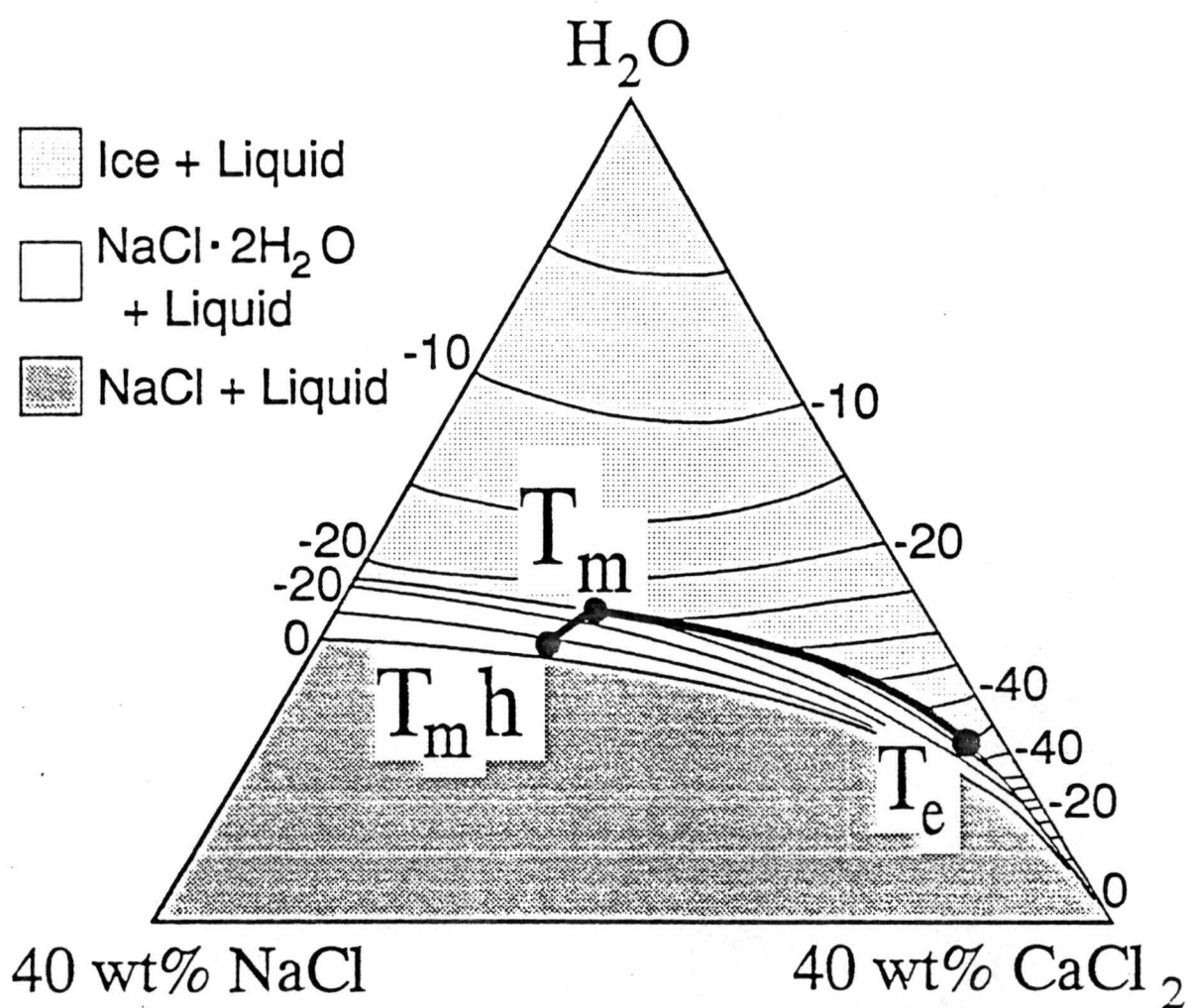


Figure 5.6. NaCl-CaCl₂-H₂O phase diagram with the average melting path of fluid inclusions for bed-normal and bed-parallel veins shown (T_e=eutectic temperature, T_m=ice melting temperature, T_mh=hydrohalite melting temperature).

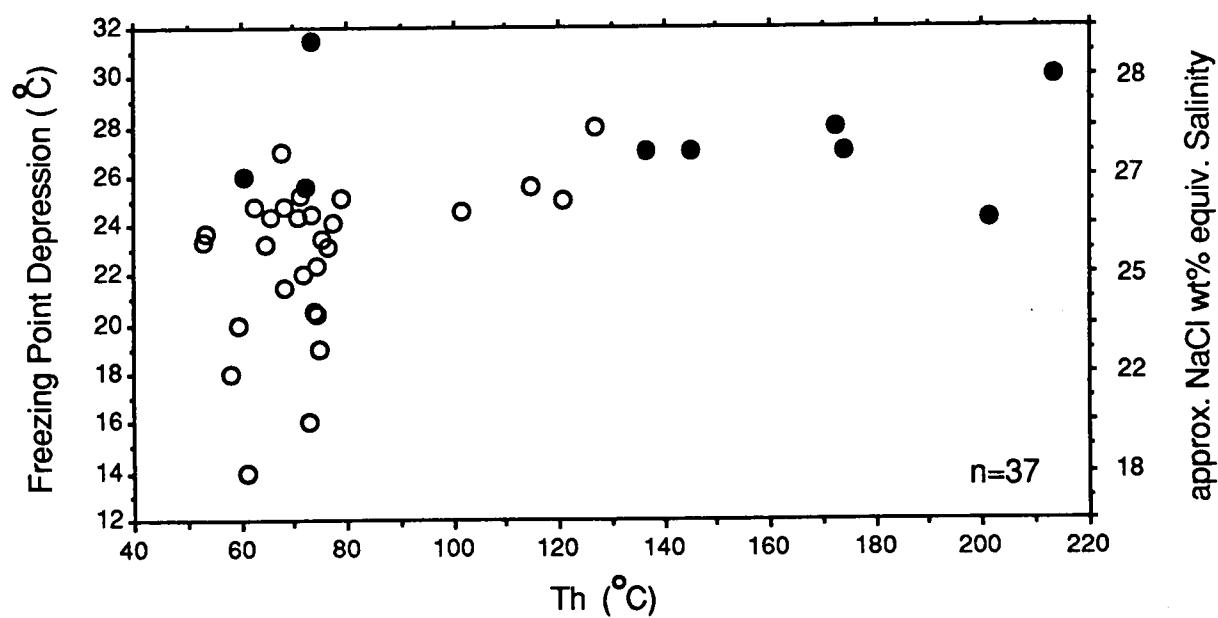


Figure 5.7. Freezing point depression (and corresponding salinity) as a function of homogenization temperatures (T_h) for bed-oblique vein (filled circles) and bed-normal and bed-parallel veins (open circles).

the Zhang and Frantz (1987) equation of state for the NaCl-H₂O system.

Stable Isotopes: limestone

Whole rock isotopic data for interbedded limestone (Haase and others, in prep.) in the Nolichucky Shale show that there is a general decrease in the whole rock $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ with increasing depth and distance from the overlying Maynardville Limestone (Figure 5.8). Limestone in the Maynardville is about 2.0 per mil heavier in ^{13}C than that at the base of the Nolichucky Shale. The whole rock $\delta^{18}\text{O}$ of Nolichucky limestone also decreases downward from the Maynardville/Nolichucky contact (Figure 5.8). The correlation coefficients for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ versus depth ($r=0.52$ and $r=0.58$ respectively, Figure 5.8) are significant at the 98% confidence level. However, linear regression analysis for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ versus depth ($r^2=0.27$ and $r^2=0.34$ respectively) demonstrates depth alone is unable to describe all the variation in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

Stable Isotopes: calcite and dolomite in shale

Calcite and dolomite extracted from bulk shale samples (Table 5.1) exhibit similar trends of decreasing $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ with increased depth in the Nolichucky Shale, corresponding to increased stratigraphic separation from the overlying Maynardville Limestone (Figure 5.9). The trend of isotopic compositions of calcite (2 hr extraction) with depth is similar to that of the interbedded limestone, however, the magnitude of the isotopic difference is greater. There is nearly a 3.0 per mil difference in $\delta^{13}\text{C}$ and a 5.0 per mil difference in $\delta^{18}\text{O}$ from the top to base of the Nolichucky (Figure 5.9a).

The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of dolomite (30 hr extraction) in the shale, show similar trends as the calcite, becoming lighter with increasing depth and stratigraphic separation from the

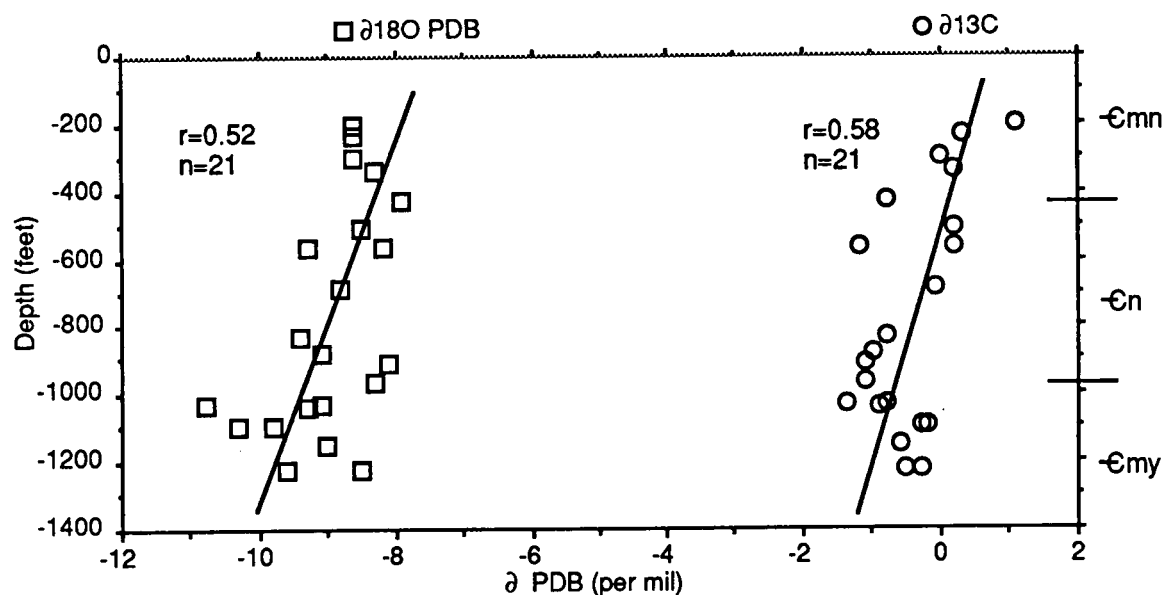


Figure 5.8. Whole rock isotopic composition of limestone from Borehole Joy-2. Formation abbreviations in this and following figures include Cm (Maynardville Limestone), Cn (Nolichucky Shale), and Cmy (Maryville Limestone). Isotopic data from Haase and others (in prep.). Width of symbols in this and following figures correspond to analytical error of ± 0.2 per mil for $\delta^{18}\text{O}$ (open squares) and $\delta^{13}\text{C}$ (open circles).

Table 5.1. Results of carbonate isotopic compositions in bulk Nolichucky shale lithologies, determined from stepwise-extractions.

¹ Sample	² Rxtn. time	³ Rxtn. time	⁴ δ ¹³ C	⁴ δ ¹⁸ O
GW134 297	120	120	-3.2	-11.34
GW134 297	300	420	-1.3	-10
GW134 297	1305	1725	-0.4	-8.66
GW134 297	1575	3300	-1.3	-10.31
GW134 267	120	120	-2.1	-10
GW134 267	290	410	-0.8	-9.07
GW134 267	1320	1730	-0.8	-9.07
GW134 267	1575	3305	-0.02	-7.32
GW140 1026	205	205	-3.3	-11.14
GW140 1026	360	565	-0.6	-9.18
GW140 1026	1275	1840	-0.4	-6.81
GW140 1026	1590	3430	-0.9	-7.94
GW140 1026	1665	5095	-5.01	-14.23
GW134 436	120	120	-2.07	-15.26
GW134 436	320	440	-1.46	-14.64
GW134 436	1510	1950	-1.13	-11.96
GW134 436	1260	3210	-4.11	-17.63
GW134 491	1420	1810	-0.62	-10.83
GW134 643	130	130	-3.36	-13.92
GW134 643	285	415	-1.69	-10.72
GW134 643	1360	1775	-2.06	-12.98
GW134 643	1375	3150	-0.95	-9.07
GW134 197	140	140	-0.51	-9.9
GW134 197	295	435	0.07	-8.97
GW134 197	1405	1840	-0.03	-8.45
GW134 360	125	125	-0.78	-10.72
GW134 360	290	415	-0.51	-9.9
GW134 360	1405	1820	-1.01	-10.52
GW134 360	1255	3075	0.47	-12.36

1. Sample number- well designation and downhole footage.
2. Reaction time with phosphoric acid (minutes).
3. Cumulative reaction with phosphoric acid (minutes).
4. Isotopic compositions with respect to PDB.

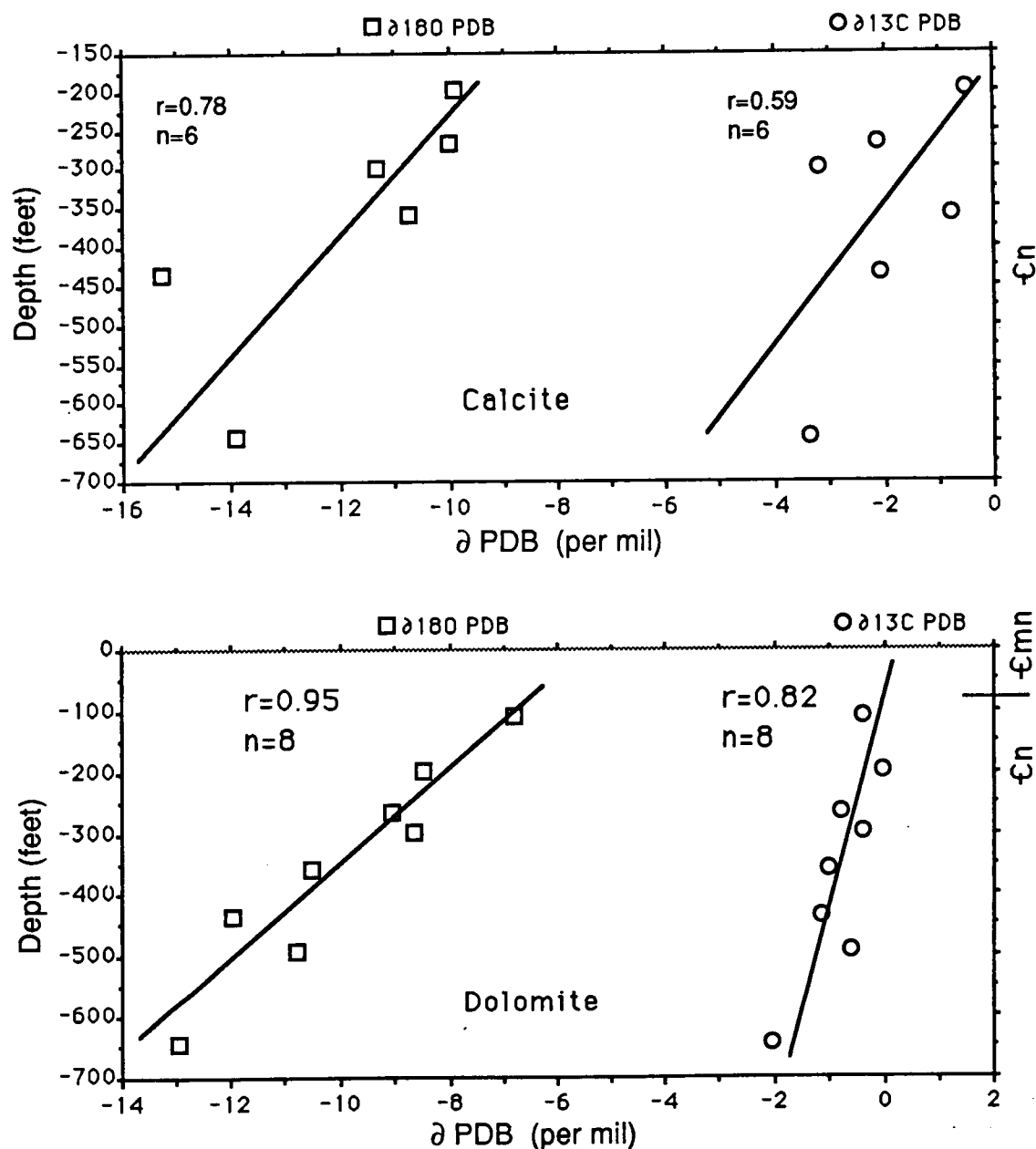


Figure 5.9. Isotopic compositions of carbonate from bulk shale samples. **A.** Isotopic composition of CO₂ extracted from shale after 2 hr (calcite) as a function of depth within the Nolichucky Shale. Borehole GW134. **B.** Isotopic composition of CO₂ extracted from shale after 30 hr (dolomite) as a function of depth within the Nolichucky Shale. The contact between the Nolichucky and overlying Maynardville is indicated at a depth of 90 feet. Borehole GW134.

overlying Maynardville Limestone (Figure 5.9b). The difference in both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ for dolomite with depth is comparable to that seen with the calcite isotopic analyses.

Stable Isotopes: vein calcites

Vein calcites in the WOM thrust sheet (Figure 5.10a) and Copper Creek (Figure 5.10b) thrust sheet are similar isotopically (Table 5.2). They show less variability with depth compared to limestone whole rock analyses (Figure 5.8) and considerably less variability with depth compared to calcite and dolomite within the shale (Figure 5.9). With the exception of two calcite analyses from the shale, the vein calcites are the most depleted with respect to ^{18}O (Figure 5.11). The $\delta^{13}\text{C}$ values for the vein calcites also are the lightest and show the widest range, with the exception of calcite extracted from the shale (Figure 5.11). The vein calcites are not in oxygen isotopic equilibrium with calcites in the limestone host rocks. The $\delta^{18}\text{O}$ for the vein calcites are 2.0 to 4.0 per mil lighter than the limestone.

DISCUSSION

P-T conditions during calcite vein formation

With the exception of boiling fluids, homogenization temperatures (T_h) will underestimate the true trapping temperature of fluid inclusions (Shepherd and others, 1985). A pressure correction must be applied to the homogenization temperature to determine true trapping temperatures. If fluid pressure during inclusion trapping can be independently determined, then the correction is relatively straightforward. Unfortunately,

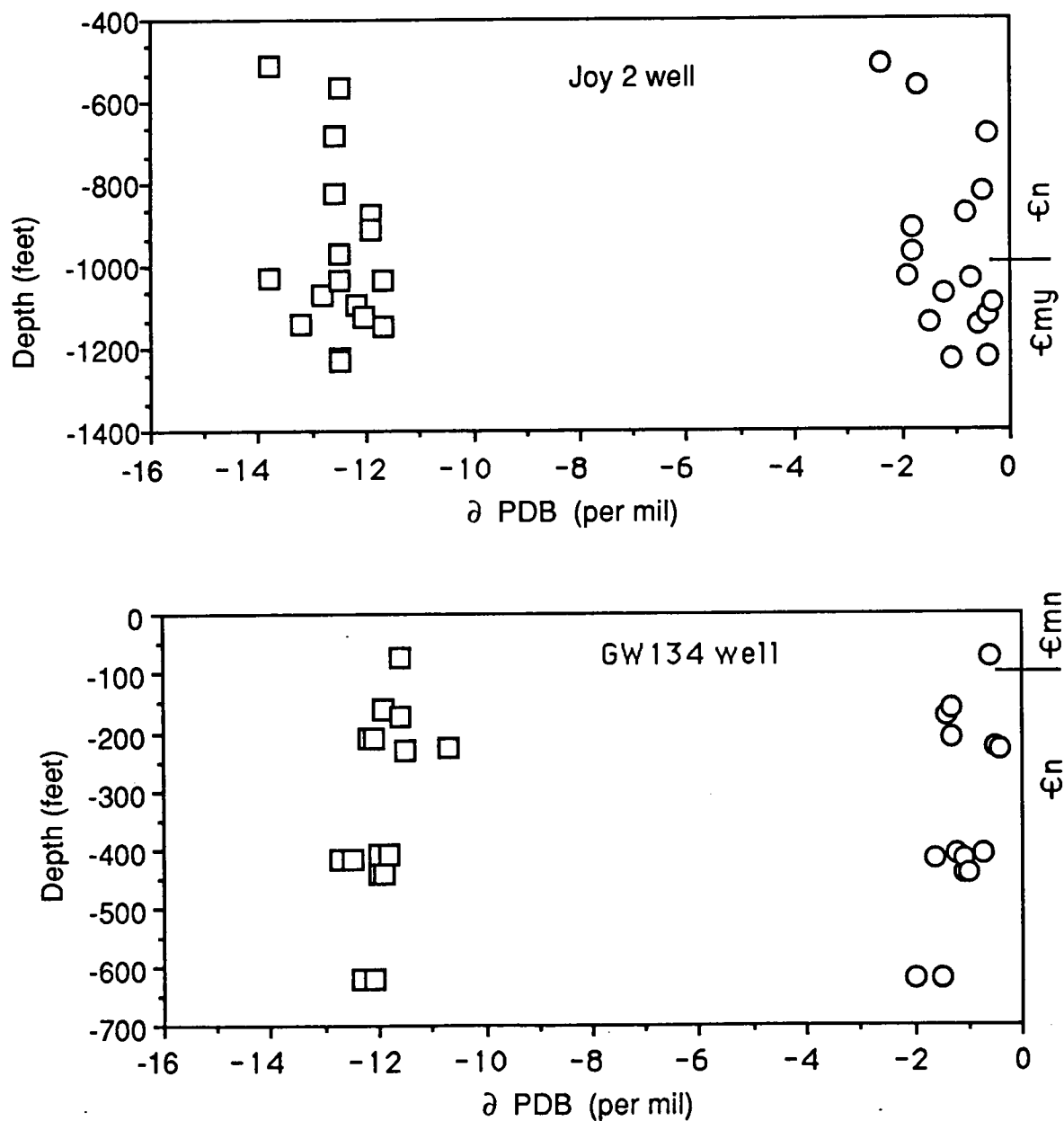


Figure 5.10. Nolichucky vein isotopic compositions. A. Isotopic composition of vein calcites as a function of depth. Figure includes data from bed-normal and bed-oblique veins. Borehole Joy-2. B. Isotopic composition of bed-normal and bed-parallel calcite veins as a function of depth. Borehole GW134.

Table 5.2. Isotopic composition of Nolichucky calcite veins.

WELL	¹ Sample	² Description	Wall Rock	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
GW134	76	bed perp vein	Shale	-0.6	-11.6
GW134	173.5A	bed perp vein	Shale	-1.4	-11.6
GW134	407	bed perp vein	Shale	-0.7	-11.8
GW134	407A	bed perp vein	Shale	-1.2	-12.0
GW134	416-1	bed perp vein	Limestone	-1.1	-12.5
GW134	416-2	bed perp vein	Limestone	-1.6	-12.7
GW134	434-1	bed perp vein	Limestone	-3.3	-16.1
GW134	440-1	bed perp vein	Limestone	-1.0	-11.9
GW134	440-2	bed perp vein	Limestone	-1.1	-12.7
GW134	440-3	bed perp vein	Limestone	-1.1	-12.0
GW134	440-4	bed perp vein	Limestone	-1.1	-12.7
GW134	621.5	bed par vein	Shale	-1.5	-12.1
GW134	621.5A	bed par vein	Shale	-2.0	-12.3
GW139	161-3	bed perp vein	Siltstone	-1.3	-11.9
GW139	180-1	bed perp vein	Siltstone	-2.0	-13.8
GW139	180-2	bed perp vein	Siltstone	-2.0	-13.4
GW139	210-1	bed perp vein	Siltstone	-1.3	-12.1
GW139	210-3	bed perp vein	Siltstone	-1.3	-12.2
GW139	228	bed perp vein	Limestone	-0.5	-10.7
GW139	229	bed oblq vein	Limestone	-0.3	-11.3
GW139	230	bed perp vein	Limestone	-0.4	-11.5

1 Sample numbers are equivalent to downhole footage. Samples with same footage indicate multiple veins.

2. bed perp.vein- bed perpendicular veins; bed par- bed-parallel veins; bed oblq vein- bed-oblique vein.

3. Isotopic compositions with respect to PDB.

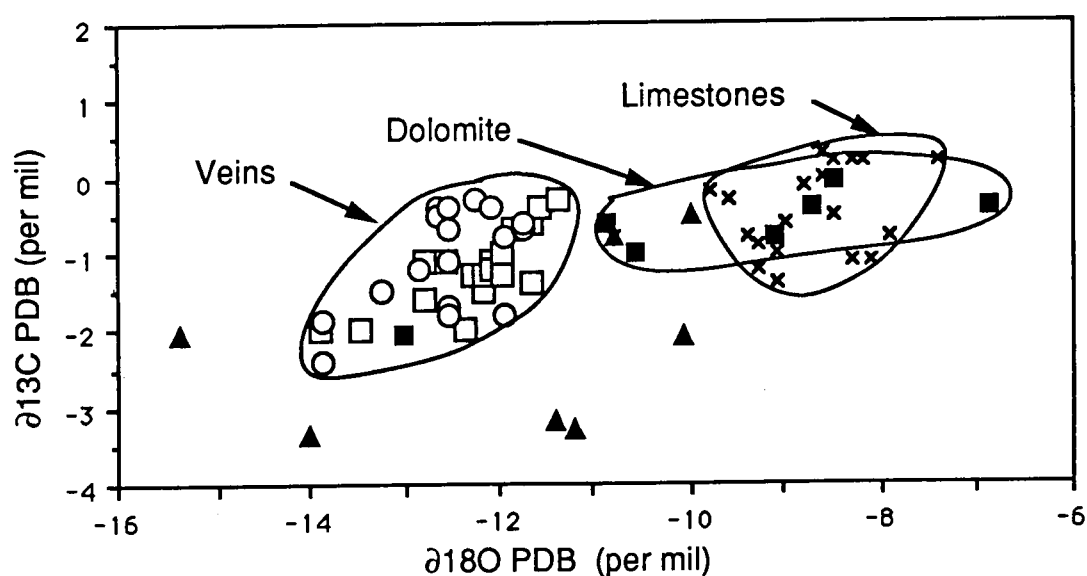


Figure 5.11. Isotopic composition of whole rock limestone, calcite veins, and carbonate from extracted from shale from the Nolichucky. Crosses and open squares correspond to whole rock limestone and veins, respectively, from Borehole Joy-2 (Haase and others in prep.). Open circles (veins), closed triangles (calcite extracted from shale), and closed squares (dolomite extracted from shale) are from Borehole GW134.

this is seldom the case, and other means must be used to correct the homogenization temperatures to reflect true trapping temperatures.

In this study, fluid pressure during the formation of calcite veins can be constrained using an understanding of the mechanisms of fracturing in the subsurface. In cases where bedding is nearly horizontal, the formation of bed-parallel extension fractures results in the 'lifting' of the overlying rock. In these instances, the fluid pressure must approach the lithostatic pressure (Gretener, 1979; Srivistava and Engelder, 1990). In contrast, during bed-normal extensional fracturing in rocks with bedding nearly horizontal, fluid pressures can range from hydrostatic to lithostatic because dilation is horizontal rather than vertical (Gretener, 1979; Srivistava and Engelder, 1990).

Pressure correction procedures used in this study follow those of Srivistava and Engelder (1990) for estimating conditions of bed-normal and bed-parallel vein formation in the central Appalachians. The range of possible temperatures and pressures during vein formation is determined by noting the minimum and maximum P-T conditions at which a fluid isochore (constant density line) intersects the hydrostatic and lithostatic pressure curves respectively. (Figure 5.12). An isochore plotted in P-T space illustrates the change in internal fluid pressure for an inclusion upon heating under conditions of constant inclusion volume and constant fluid density; the slope of the isochore is a function of the fluid density and must be known before a pressure correction is applied (Shepherd, 1985). The isochore shown in Figure 5.12 corresponds to a fluid with a density of 1.18 g/cm^3 (27 wt% NaCl equivalent salinity) having an homogenization temperature of 70°C (average T_h for bed-normal and bed-parallel veins). Lithostatic and hydrostatic pressure gradients of 26 MPa/km and 10 MPa/km respectively, and a geothermal gradient of 25°C/km , were assumed in order to plot the hydrostatic and lithostatic curves in Figure 5.12.

After pressure corrections, bed-normal veins are estimated to have formed at temperatures between $80\text{--}110^\circ\text{C}$ and fluid pressures of 25-90 MPa (hydrostatic to

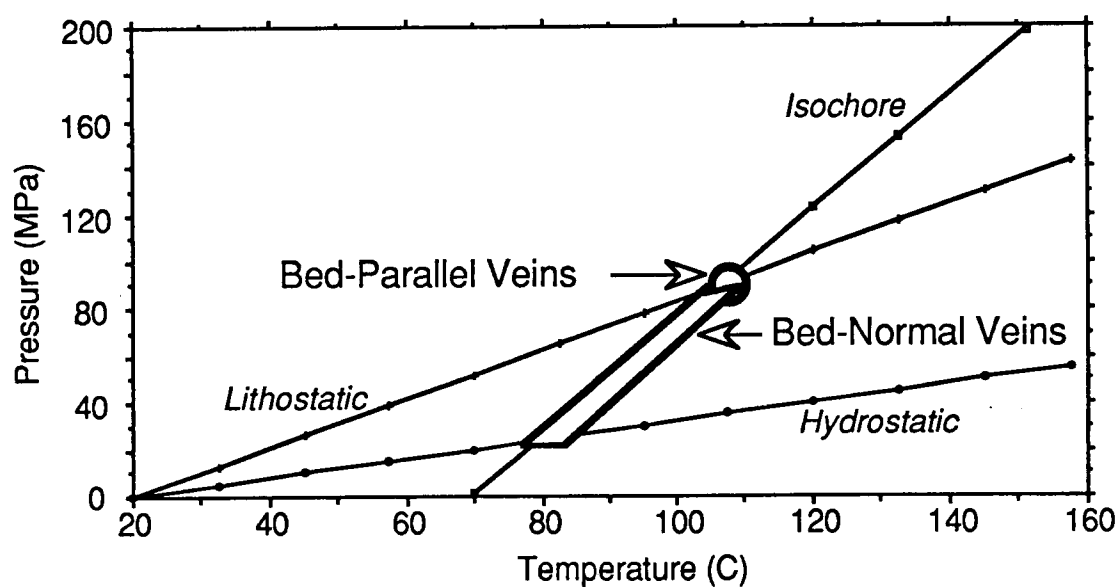


Figure 5.12. P-T diagram for calculating fluid inclusion trapping temperatures for bed-normal and bed-parallel veins. Open trapezoid indicates probable P-T conditions for bed-normal veins. Open circle corresponds to bed-parallel veins where fluid pressures equal lithostatic pressure. Isochore calculated from equation of state for NaCl-H₂O system (Zhang and Frantz, 1987) using fluid density of 1.18 gm/cm³ and $T_h = 70^\circ\text{C}$. Lithostatic gradient=26 MPa/km, hydrostatic gradient=10 MPa/km, geothermal gradient=25 $^\circ\text{C}/\text{km}$.

lithostatic pressure, respectively; Figure 5.12). Bed-parallel veins formed from fluids of approximately 110 °C and 90 MPa fluid pressure (lithostatic pressure; Figure 5.12).

Assuming both bed-normal and bed-parallel veins formed during increasing burial of the Nolichucky Shale, the higher temperatures for bed-parallel veins is expected because cross-cutting relationships clearly show that bed-parallel veins formed after bed-normal veins (Foreman and Dunne, in press). The absence of evidence for inclusion re-equilibration of fluid inclusion homogenization temperatures to higher temperatures indicates that bed-normal and bed-parallel vein formation occurred close to the maximum burial depth and temperature for the Nolichucky Shale (Figure 5.13, Foreman and Dunne, in press).

Published conodont alteration indices (CAI) for overlying Ordovician limestone (CAI=1.5-2.0, Harris and others, 1977; Harris, 1979), and illite crystallinity values for Conasauga Shale illite (Weaver Index=1.2; Weaver, 1985), demonstrates that the Nolichucky Shale did not experience temperatures in excess of 110-150 °C, respectively.

The pressure correction procedure for the bed-oblique vein is the same as that for the bed-normal veins, using an average $T_h=180$ °C, and fluid density of 1.18 g/cm³ (27 wt% NaCl wt% equivalent salinity). Pressure-corrected fluid temperatures range from 200 to 280 °C with fluid pressures of 70 to 250 MPa. This vein clearly formed from hotter fluids than the bed-normal and bed-parallel veins. Because evidence presented previously suggests that the Nolichucky Shale did not experience ambient temperatures much greater than approximately 110-150 °C, the fluid present during formation of the bed-oblique vein was hotter than the surrounding Nolichucky host rocks and probably originated from more deeply buried rocks. Fluid pressures in excess of approximately 100 MPa could not be produced by the 3.6 km of maximum overburden determined from the decompacted burial curve (Figure 5.13) suggesting that a major component of the calculated fluid pressure may be due to tectonic forces or aquathermal overpressuring.

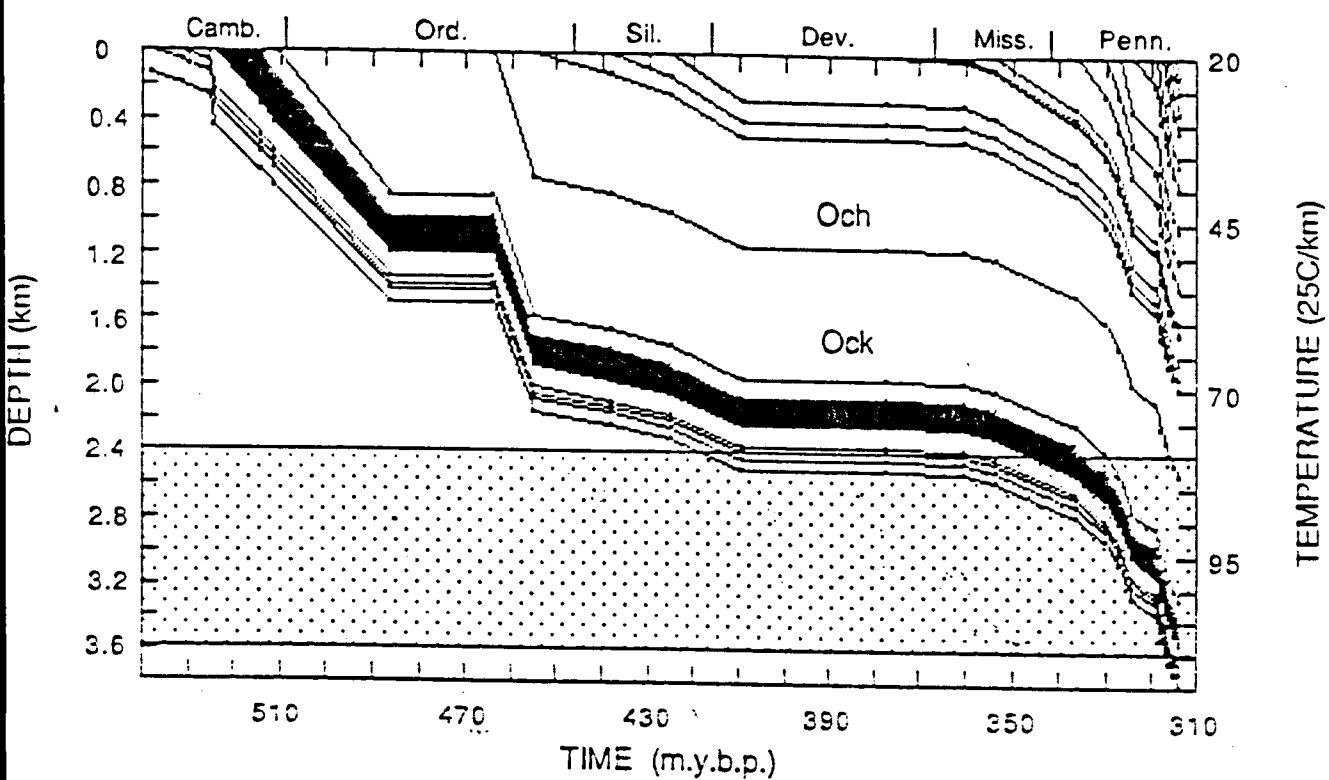


Figure 5.13. Decompacted burial curve for the Conasauga Group and younger strata in the Whiteoak Mountain thrust sheet, Oak Ridge Reservation. Dark pattern=Nolichucky Shale, Ock=Cambro-Ordovician Knox Group, Och=Middle Ordovician Chickamauga Group. Stippled band indicates probable range of vein calcite precipitation temperatures calculated from corrected average homogenization temperatures of bed-normal and bed-parallel veins.

Chemistry of vein-forming fluid

Subsurface brines form by a combination of processes including subaerial evaporation of seawater, subsurface dissolution of evaporites, and water-rock interactions during burial (e.g. albitization of plagioclase feldspar, dolomitization, and smectite dehydration; Hanor, 1987). Nolichucky vein-forming fluid compositions, estimated from fluid inclusion freezing data, can be compared to published analyses of sedimentary brines to evaluate possible mechanisms for the formation of the fluid that precipitated the calcite veins.

CaCl and NaCl brines from the U.S. Gulf Coast (Norphlet Formation brines and calcium-rich Frio Formation brines) are compared to calculated fluid compositions of inclusion fluids from Nolichucky calcite veins, and published compositions of inclusion fluid in sphalerite from the Mascot-Jefferson City MVT district in east Tennessee (Table 5.3). The Norphlet and Frio brines are among the most saline brines listed for these formations (Carpenter and others, 1984; Morton and Land, 1987). Using a composition of 17 wt% NaCl and 9.5 wt% CaCl₂, from the fluid inclusion freezing behavior (Figure 5.6), concentrations and mole ratios for Nolichucky inclusion fluids can be estimated assuming calcium, sodium, and chloride are the only constituents making up the total dissolved solids (TDS). The validity of this assumption is supported by the (Ca+Na+Cl)/TDS ratios for the other brines included in Table 5.3. Similarities in the composition of calculated Nolichucky inclusion fluids with the published brine compositions are striking (Table 5.3).

The two Gulf Coast brines in Table 5.3 have compositions which may be a result of subaerial evaporation of seawater, though water-rock interaction after burial probably was an important process in modifying the brine composition following extensive evaporation (Carpenter and others, 1974; Morton and Land, 1987). Morton and Land (1987) consider

dolomitization by the brine after extensive evaporation to be a viable means of decreasing the Na/Ca ratio in the Ca-rich Frio Formation brines. However, the absence of thick dolostone units in the Frio caused these authors to favor an external fluid source, where extensive water-rock interaction with evaporites or sediments experiencing albitization, could produce the observed Na/Ca ratios (Morton and Land, 1987). Thick accumulations of dolomite are present in the Rome Formation, Maynardville Limestone, and Knox Group, and sedimentological evidence for hypersalinity has been found in these three units (Churnet and Misra, 1981; Weber, 1988; McReynolds, 1989). The sedimentological evidence suggests that all of these units are potential sources of saline Ca-rich brines assuming the evaporation-dolomitization brine-forming model of Morton and Land (1987) is correct.

The brine-forming mechanism proposed by Morton and Land (1987) is one of many mechanisms suggested for the formation of brines (Hanor, 1987). In shale-rich sequences, membrane filtration may be a viable means of producing brines similar to those listed in Table 5.3. Several lines of evidence, including strontium isotopes and fluid inclusions, led Kesler and others (1989) to conclude that membrane filtration was a likely mechanism for generating the MVT mineralizing brines in east Tennessee. They believe enough shale was present in the Ordovician Sevier Shale basin to generate fluids responsible for all the MVT mineralization in east Tennessee. However, it is not clear that membrane filtration is capable of generating brines with high salinities (Kesler, 1989; Morton and Land, 1987) such as those characteristic of some MVT mineralizing fluids and the Nolichucky vein-forming brines.

Isotopic composition of vein-forming fluid

The isotopic composition of fluid in equilibrium with vein calcite at the time of precipitation can be determined from published calcite-water fractionation relationships

Table 5.3 Calculated fluid composition in calcite vein fluid inclusions and molar ratios compared to Gulf Coast formation waters and east Tennessee MVT fluids in sphalerite.

	¹ Nolichucky	² Norphlet Fm.	³ Frio Fm.	⁴ E. TN MVT
TDS	327,500	326,400	105,000-280,000	245,300
Cl	206,500	198,700	20,000-152,000	150,400
Na	81,000	79,000	23,000-40,000	59,00
Ca	40,000	34,000	10,000-40,000	27,500
Na/Cl	0.62	0.61	0.54-1.39	0.60
Ca/Cl	0.17	0.15	0.23-0.44	0.16
Na/Ca	3.52	3.42	1.74-4.00	3.74
Na/TDS	0.24	0.24	0.14-0.22	0.2
Ca/TDS	0.12	0.10	0.10-0.14	0.11
Cl/TDS	0.65	0.61	0.65-0.86	0.61
Σ /TDS	1.01	0.95	0.50-0.83	0.95

All concentrations are in mg/l. X/Cl expressed as mole ratios. X/TDS expressed as weight ratios. TDS=total dissolved solids. Σ /TDS=weight ratio (Na+Ca+Cl)/TDS.

¹Nolichucky concentrations calculated from freezing behavior assuming concentrations of species other than NaCl and CaCl₂ are minimal. Nolichucky fluid density=1.18.

²Brine compositions from Carpenter and others (1974).

³Ca-rich brines of Morton and Land (1987).

⁴Inclusion fluid composition, high salinity sphalerite, Mascot-Jefferson City; from Kesler and others (1989).

assuming precipitation temperatures can be constrained. The matched fluid inclusion and isotopic analyses for calcite veins in the Nolichucky Shale permits this calculation to be made for six bed-normal veins, one bed-parallel vein, and one bed-oblique vein because temperature is constrained from the pressure-corrected homogenization temperatures. These corrected temperatures were used with the calcite-water fractionation equation of O'Neil and others (1969) to calculate an upper and lower value for $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ assumed to be in equilibrium with the vein calcite during vein formation. The calculated water compositions for the eight veins form an envelope which ranges from $\delta^{18}\text{O}_{\text{H}_2\text{O}} = -3.5$ per mil SMOW for bed-normal veins to $\delta^{18}\text{O}_{\text{H}_2\text{O}} = 12.5$ per mil SMOW the bed-oblique vein. (Figure 5.14).

Sources of vein-forming fluids

Identifying the original isotopic composition and source of vein-forming fluids based on the fluid compositions presented in Figure 5.14 is hampered by two uncertainties: 1) The isotopic composition of early Paleozoic seawater is poorly understood, and as a result, estimates of meteoric water compositions, derived from the evaporation of Paleozoic seawater, are uncertain; and 2) It is difficult to quantify the amount of isotopic exchange between fluid and surrounding rock because of uncertainties in the original fluid composition and water-rock ratios. The hypothesis to be evaluated is that the isotopic compositions of the fluids responsible for veining were the result of water-rock interaction in the Nolichucky Shale prior to actual vein calcite precipitation. The null hypothesis states that the isotopic composition of these fluids could not have resulted from isotopic exchange and equilibrium between vein-precipitating fluids and host rocks in the Nolichucky Shale. If this is true, an extraformational fluid source must be considered.

Though considerable controversy exists concerning the $\delta^{18}\text{O}$ value of ancient oceans, for the last 2500 m.y. they have probably ranged from 0 to -3 per mil SMOW for $\delta^{18}\text{O}$

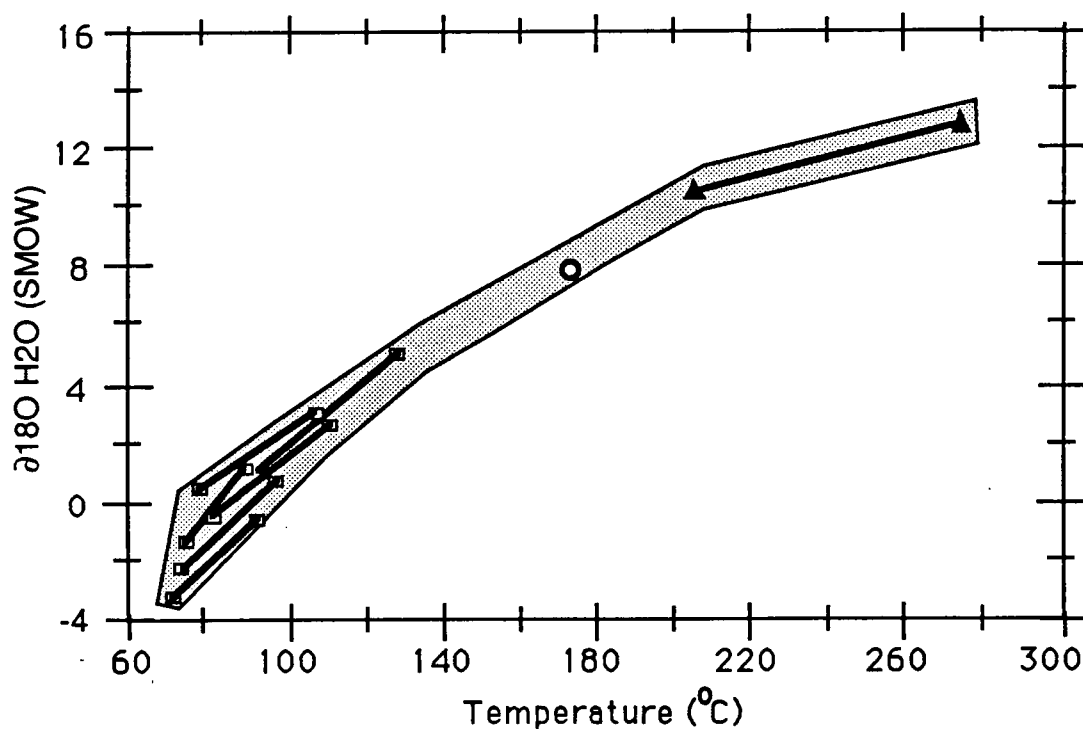


Figure 5.14. Isotopic composition of fluids calculated from corrected homogenization temperatures (horizontal axis) and vein calcite isotopic compositions, using the calcite-water fractionation relationship of O'Neil and others (1969). Open circle=bed-parallel vein; line segments from 75 to 130 $^{\circ}\text{C}$ correspond to bed-normal veins; highest temperature sample is for bed-oblique shear vein. Individual line segments reflect uncertainty in fluid pressure being equivalent to either lithostatic or hydrostatic pressure (Figure 5.12).

(Sheppard, 1986). Subaerial evaporation of seawater with these compositions would produce isotopically enriched fluids with isotopic compositions for $\delta^{18}\text{O}$ of approximately 0 to 3 per mil SMOW. In the subsurface, the interaction of these fluids with surrounding rocks would cause these fluids to shift towards heavier $\delta^{18}\text{O}$ (Sheppard, 1986) resulting in burial fluids with isotopic compositions greater than evaporated Cambrian seawater. Considering the lithologic nature of the Nolichucky Shale, it is reasonable to question how the interbedded limestone and shale intervals may have influenced the isotopic composition of the vein-forming fluid.

Mineral-water and whole rock-water fractionation equations can be used to estimate the isotopic composition of a fluid in equilibrium with interbedded limestone and shale in the Nolichucky Shale. Equilibrium fluid compositions for the limestone beds were calculated using an average bulk rock value for $\delta^{18}\text{O}=22.0$ per mil SMOW and the calcite-water fractionation equation of O'Neil and others (1969). A mineralogical composition of 50% illite/smectite (I/S=0.9), 40% quartz, 7% dolomite, and 3% calcite was chosen for Nolichucky shale interbeds based on XRD and thin section examination. A value of 15.0 per mil SMOW for $\delta^{18}\text{O}$ was used for illite and quartz based on data from Haase and others (in prep.) for the $<2\mu\text{m}$ fraction in the underlying Pumpkin Valley Shale. Values of $\delta^{18}\text{O}$ for calcite and dolomite in the shale were 18.7 and 21.0 per mil SMOW respectively. Fractionation factors used for the bulk shale fractionation equation are the following: quartz- H_2O (Friedman and O'Neil, 1977), illite/smectite- H_2O (Eslinger and others, 1979), dolomite- H_2O (Woronick and Land, 1985), and calcite- H_2O (O'Neil and others, 1969). Equilibrium isotopic compositions for vein-forming fluids were calculated using the calcite-water fractionation factor of O'Neil and others (1969), and $\delta^{18}\text{O}=18$ per mil SMOW for the vein calcite composition (Figure 5.10).

The equilibrium curve for the Nolichucky calcite veins (average $\delta^{18}\text{O}=18.0$ per mil SMOW) and the calculated vein-forming fluid compositions (Figure 5.14) lies between the

whole-rock Nolichucky limestone ($\delta^{18}\text{O}=22.0$ per mil SMOW) and whole-rock shale equilibrium curves (Figure 5.15). This result is in agreement with the hypothesis that the oxygen isotopic composition of the vein-forming fluid could have been the result of isotopic exchange between burial fluids and interbedded limestone and shale from the Nolichucky Shale.

The similarities in the $\delta^{13}\text{C}$ for interbedded limestone and vein calcites (Figure 5.11) also supports the hypothesis that the interbedded limestone were the major carbon reservoir during calcite vein formation. No evidence to support a significant carbon contribution from organic matter exists. Carbon derived from the degradation of organic matter under burial conditions is considerably lighter (-5 to -10 per mil) than carbon released by limestone dissolution (Curtis and Coleman, 1986).

The isotopic exchange model does not preclude the possibility that fluids external to the Nolichucky were involved in the formation of the calcite veins. Vein-forming fluids with similar isotopic compositions to those presented in Figure 5.14 could be produced within other interbedded limestone and shale units including the underlying Conasauga members, Rome Formation, or Ordovician Sevier Shale present less than 50 km palinspastically to the southwest in the Sevier Basin. The high temperatures indicated for one bed-parallel vein and the bed-oblique vein (Figure 5.14) suggests these fluids were hotter than the surrounding rock because the maximum burial temperature determined for the Nolichucky Shale is probably less than 110-150 °C (Figure 5.12; Harris and others, 1977; Weaver, 1985; Foreman and Dunne, in press). Such fluids could also be produced from sedimentary rocks buried at depths of 4 to 5 km assuming a 25 °C/km geothermal gradient and negligible heat loss during fluid migration. The Middle Ordovician shales of the Sevier Basin were buried to these depths before the end of the Devonian Period (Walker and Foreman, 1990).

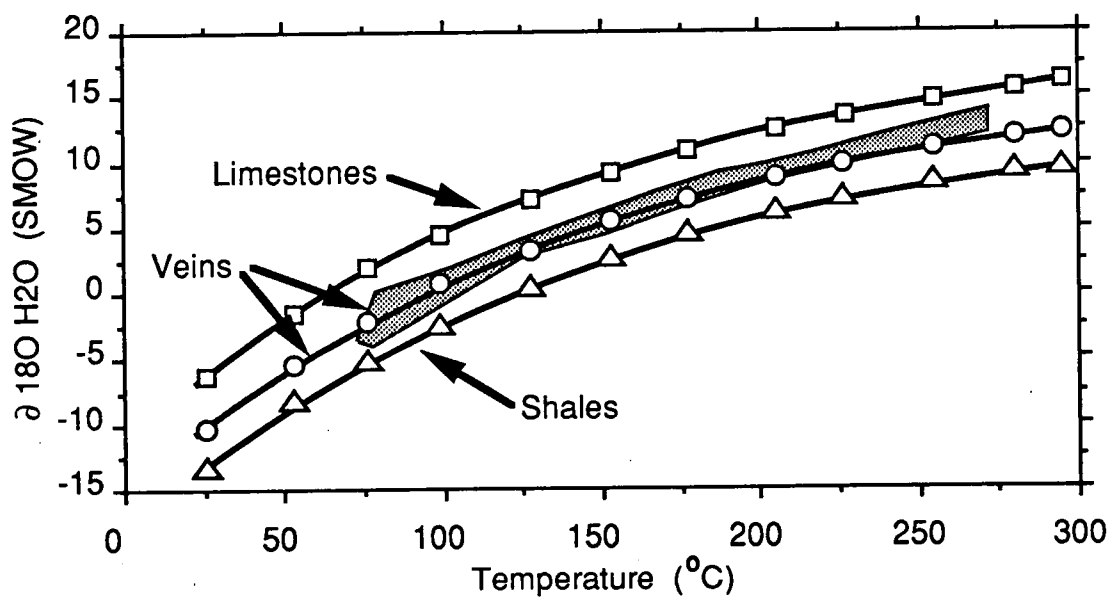


Figure 5.15. Calculated oxygen isotopic equilibrium curves for average whole rock limestone (upper curve, $\delta^{18}\text{O}=22$ per mil SMOW), average vein calcites (middle curve, $\delta^{18}\text{O}=18$ per mil SMOW), and average whole rock shale (weighted mineral-water fractionation expression discussed in text). Shaded field corresponds to matched isotopic-temperature data for eight veins presented in Figure 5.14.

Fluid flow during burial diagenesis

The preservation of isotopic compositional trends in limestone (Figure 5.8) and authigenic calcite and dolomite in the Nolichucky shale (Figure 5.9) would not be expected if pervasive cross-formational fluid flow and isotopic exchange occurred during burial. Instead, the data suggest that prior to fracturing, fluids in the Nolichucky may have been stratified with limited cross-formational fluid migration during precipitation of authigenic calcite and dolomite in the shale. The parallel isotopic compositional trends for limestone and authigenic shale-carbonate is to be expected if the interbedded limestone strongly influenced the composition of diagenetic fluids during authigenic calcite and dolomite formation in the shale. These trends indicate that a rock-dominated system (low water/rock ratio) existed during the precipitation of authigenic carbonates within the shale.

The vein-forming fluids have isotopic compositions that may reflect considerable isotopic exchange with Nolichucky limestone and shale beds (Figure 5.15). Assuming this is true, the absence of vertical compositional trends, as seen for the interbedded limestone and authigenic carbonates in the shale, can be accounted for by vertical mixing of fluids prior to or during vein calcite precipitation. This mixing would be greatly enhanced by vertical fracturing which channelized fluid flow along discrete pathways (fractures) and limited isotopic exchange with the surrounding host rocks.

CONCLUSIONS

The Nolichucky Shale experienced two contrasting episodes of fluid-rock interaction during the precipitation of authigenic carbonates in interbedded limestone and shale, and

calcite veins. The formation of fractures caused the Nolichucky to change from a rock-dominated system, to a fluid-dominated system in the immediate vicinity of open fractures during vein precipitation. Prior to fracturing, a horizontally stratified fluid system existed, during which time authigenic calcite and dolomite were precipitated in limestone and shale intervals. This earlier fluid system was characterized by extensive isotopic exchange between burial fluids and the interbedded limestone and shale. The fractures were filled by calcite precipitated from highly saline sodium and calcium chloride brines at temperatures of 80 to more than 200 °C. Fluid temperatures in excess of 110-150 °C were probably derived from deeper sedimentary rocks because maximum ambient burial temperatures for the Nolichucky probably did not exceed these values. The isotopic composition of the vein-forming fluid can be explained by isotopic exchange with limestone and shale with compositions similar to those in the Nolichucky Shale. Fractures facilitated mixing of fluids prior to vein formation and effectively limited the degree of isotopic exchange between these fluids and immediately surrounding host rocks during precipitation of the calcite veins. It is likely that similar processes have occurred in other foreland basin sedimentary sequences.

CHAPTER 6

CONCLUSIONS

1. Interbedded limestone, siltstone, and shale in the Nolichucky Shale are the product of slope and basin deposition. Vertical trends indicate an overall shallowing-upward sequence punctuated by numerous smaller-scale shallowing-upward and deepening-upward sequences. Maximum water depths in the basin probably ranged from 100-300 m and were an order of magnitude greater than those previously reported for the Nolichucky basin by Markello and Read (1981, 1982), and Weber (1988). A general stratigraphic model for the Conasauga Group that includes observations from this study is presented by Walker and others (in press).

2. Intraclastic limestone debris flow deposits and calcarenite turbidite lithologies imply that depositional slopes were steep enough to transport shallow-water carbonate at least 20 km downslope into the intrashelf basin. Depositional slope were probably greatest during deposition of the lower Nolichucky, evidenced by abundant polymictic intraclastic limestone. This part of the basinal Nolichucky Shale is equivalent in age to the uppermost part of the Maryville Limestone, further to the northeast and characterized by biohermal reef buildups (Walker and others, in press; Srinivasan, in prep.) These polymict conglomerate units contain up-slope-derived lime mudstone clasts and down-slope-derived laminated peloidal limestone and calcareous siltstone. Storms were probably an important factor for mixing allochthonous carbonate with hemipelagic shale in the intracratonic basin.

3. Petrographic and geochemical evidence suggests that Nolichucky ooids were originally fibrous radial high-magnesium calcite (HMC). The original magnesium content may have been about 12 mol% MgCO_3 , similar to modern radial fibrous ooids from Baffin Bay, Texas (Land and others, 1979). Unequivocal evidence for originally bimineralic or

aragonitic ooids is lacking. The ooids are now low-magnesium calcite (1.5 mol% MgCO_3) and show a variety of cortical fabrics interpreted to indicate progressive burial alteration from an originally fibrous radial cortex. Iron and manganese concentrations in these ooids are 2-3 orders of magnitude greater than reported for unaltered marine calcite. These high concentrations reflect stabilization and progressive alteration of the HMC ooids in the presence of pore fluids modified by interaction with basinal shale.

4. The occurrence of fibrous radial and sparry radial ooids in the same thin section can be explained by different degrees of water-rock interaction, that was controlled by small-scale variations in microporosity and permeability. The sparry radial ooids have compositions compatible with a greater degree of water-rock interaction compared to less-altered fibrous radial ooids.

5. In the Whiteoak Mountain (WOM) thrust sheet, four sets of bed-normal calcite veins are recognized in the Nolichucky Shale. These sets predate both bed-parallel slickensided calcite veins and emplacement of the WOM thrust sheet. Primary fluid inclusions in the bed-normal veins indicate they precipitated at temperatures of 80-110 °C (equivalent to 2.4-3.6 km burial depth) and from highly saline NaCl-CaCl₂ brines. Bed-parallel veins formed during maximum burial at 110 °C (3.6 km burial depth) and from brines of similar composition and salinity as bed-normal veins.

6. The fluid inclusions and a decompacted burial history curve for the Nolichucky Shale suggests bed-normal veins formed between the late Mississippian and early Pennsylvanian during the onset of the Ouachitan and Alleghanian orogenies, but prior to movement of the WOM sheet. The compressive stresses, and resultant far-field stresses related to the orogenies, may have controlled the trend of the bed-normal vein sets. Bed-parallel slickensided calcite veins formed contemporaneously with thrusting of the WOM thrust sheet.

7. The Nolichucky Shale experienced two contrasting episodes of fluid-rock interaction during the precipitation of authigenic carbonate in interbedded limestone and

shale, and calcite veins, respectively. Prior to fracturing, a stratified fluid system existed, during which time authigenic calcite and dolomite were precipitated in limestone and shale intervals. This earlier fluid system was characterized by extensive isotopic exchange between burial fluids and the adjacent interbedded limestone and shale. The isotopic composition of the authigenic carbonate in the shale was controlled by the compositions of adjacent limestone interbeds. This suggests that the limestone interbeds were the dominant carbon and oxygen reservoirs during authigenic carbonate precipitation. The formation of fractures caused the Nolichucky to change to a more open water-rock system in the vicinity of the fractures and immediately surrounding wall-rocks. During this period, vein-precipitating fluids underwent little isotopic exchange with surrounding wall-rocks resulting in calcite vein isotopic compositions that are similar throughout the Nolichucky Shale. It is likely that open fractures provided conduits for the introduction of extraformational fluids into the Nolichucky Shale though the isotopic compositions of the fluids suggest isotopic exchange with mixed carbonate-siliciclastic units similar in composition to the Nolichucky.

8. Several vein calcites have calculated fluid inclusion trapping temperatures in excess of 110-150 °C. Fluids responsible for the precipitation of these veins were probably derived from deeper sedimentary rocks because maximum ambient temperatures for the Nolichucky Shale probably did not exceed 110-150 °C. Calculated trapping temperatures for calcite veins that are less than these temperatures are compatible with an intraformational source for vein-forming fluids.

Future Work

The data and interpretations presented here on stabilization of originally high-magnesium calcite ooids could be expanded to provide additional details on original isotopic compositions of Upper Cambrian marine carbonates. Quantitative water-rock

interaction modelling could be attempted with these starting compositions. Phosphatic brachiopods that coexist with ooids in some Nolichucky Shale lithologies should be analyzed for their oxygen isotopic composition. Because marine phosphates are much more resistant to post-depositional isotopic exchange in comparison to marine carbonates, these brachiopods might provide a good estimate of original marine carbonate isotopic compositions.

Matched isotopic and trace element data for the same originally high-magnesium calcite component (e.g. large echinoderm fragments) would also assist in quantitative water-rock interaction modelling following the methods of Brand and Veizer (1980, 1981). Following extraction in H_3PO_4 for stable isotopes, the acid could be run for magnesium, iron, manganese, and strontium using flameless atomic absorption spectrometry or inductive coupled plasma source spectrometry (ICP). This analytical approach has already been shown successful for carbonates using the ICP (Coleman and others, 1989). Quantification of water-rock ratios and flux of burial fluids may aid in further understanding the compaction of clay-rich sediments during burial and the effects of the burial fluids on interbedded lithologies such as limestone. The Nolichucky Shale would probably be a good unit to continue this type of research.

The fluid inclusion work on calcite veins in the Nolichucky Shale provides a firm foundation for continuing this work on other vein populations in the Nolichucky Shale and other units in the Whiteoak Mountain thrust sheet and Copper Creek thrust sheet. The limited number of analyses on the bed-oblique vein set suggests upward migration of hot fluids, probably associated with actual thrust sheet emplacement. Additional data are required to substantiate this hypothesis and others concerning the source of fluids during veining and fluid migration mechanisms. The thicker calcite veins in the Nolichucky Shale are likely to have fluid inclusions large enough for analysis of fluids trapped in the inclusions.

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APPENDICES

APPENDIX A
LITHOLOGIC DATA FROM DRILL CORE GW134

Appendix A

Description of Data Columns

The following data represent description of drill core made at 1.0 foot intervals from the Nolichucky Shale based on the Handbook for Logging Carbonate Rocks (Bebout and Loucks, 1984) and put in spreadsheet form. A summary of columns follows:

<u>Column</u>	<u>Summary</u>
1	Dowhole footage from surface in feet.
2	% Limestone relative to shale and siltstone
3	% Siltstone relative to shale and limestone
4	% Shale relative to limestone and siltstone
5	Nature of contact with overlying bed (G-gradational; S-sharp)
6	Bed Type (IB-interbedded with planar contacts; WB-wavy-bedded with irregular contacts; XB-cross-bedded)
7	Bed thickness (VTK >4 in.; TK 1-4 in.; M 0.4-1 in.; TN 0.1-0.4 in.; VTN <0.1 in.)
8	% coated grains of total carbonate fraction
9	% intraclasts of total carbonate fraction
10	% fossil fragments of total carbonate fraction
11	% silt-size and smaller carbonate of total carbonate fraction
12	% detrital quartz
13	Dunham rock name (M-mudstone; W-wackestone; P-packstone; G-grainstone)
14	Limestone color (1st letter modifiers: L-light, M-medium, D-dark; 2nd letter or single letter-color: G-gray, T-tan, BN-brown, RBN-reddish brown, OL-olive, B-black, GN-green)
15	Shale color (color code same as for limestone color)
16	Additional notes (self-explanatory)

Table A.1- Digitized descriptive lithologic data from drill core GW134.

65	45	0	55	G	IB	M	0	0	0	0	80	20	M-W	G	VDG	-
66	40	0	60	G	IB	M	0	0	0	0	80	20	M-W	G	VDG	-
67	35	0	65	G	IB	TN	0	0	0	0	80	20	M-W	G	VDG	BPAR VUGS
68	30	0	70	G	IB	TN	0	0	0	0	80	20	M-W	G	VDG	BPAR VUGS
69	25	0	75	G	IB	TN	0	0	0	0	80	20	M-W	G	VDG	BPAR VUGS
70	10	10	80	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
71	5	5	90	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
72	3	2	95	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
73	3	2	95	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
74	5	5	90	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
75	10	10	80	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
76	20	15	65	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
77	30	30	40	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
78	45	45	10	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
79	5	5	90	G	IB	VTN	0	0	0	0	50	50	CS-M	G	VDG	-
80	100	0	0	G	-	VTN	60	0	0	0	20	20	0	W-P	MG	T SH- OX, SH?
81	90	0	10	G	IB	TN-VTN	60	0	0	0	20	20	0	W-P	MG	T SH- OX, SH?
82	20	0	80	G	IB	TN-VTN	60	0	0	0	20	20	0	W-P	MG	T SH- OX, SH?
83	20	0	80	G	IB	TN	30	0	0	0	30	40	0	P-W	LG	G-GN
84	20	0	80	S	IB	TN	30	0	0	0	30	40	0	P-W	LG	G-GN
85	85	0	15	G	IB-WB	VTN	60	0	0	0	40	0	0	P-G	LG	G-GN
86	85	0	15	S	IB-WB	VTN	60	0	0	0	40	0	0	P-G	LG	G-GN
87	50	0	50	G	WB	M	0	0	0	0	100	0	0	W	LG	Y-MG
88	50	0	50	G	WB	M	0	0	0	0	100	0	0	W	LG	Y-MG
89	0	0	100	G	-	-	0	0	0	0	100	0	0	-	-	DG
90	0	0	100	G	-	-	0	0	0	0	100	0	0	-	-	DG
91	35	0	65	G	WB-IB	M	0	0	0	0	100	0	0	W	LG	DG
92	40	0	60	G	WB-IB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
93	80	0	20	G	WB-IB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
94	80	0	20	G	WB-IB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
95	80	0	20	G	WB-IB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
96	80	0	20	G	WB-IB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
97	85	0	15	G	WB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
98	100	0	0	G	WB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG
99	100	0	0	G	WB	VTN	0	0	0	0	100	0	0	W	LG	LG-MG

Table A.1- Digitized descriptive lithologic data from drill core GW134.

100	100	0	0	G	WB	VTK	30	0	0	70	0	W-P	LG	-	THROMB?
101	100	0	0	S	WB	VTK	30	0	0	70	0	W-P	LG	-	THROMB?
102	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	W	LG	DG	SOFT SED DEFM?
103	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	W	LG	MG	SOFT SED DEFM?
104	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	W	LG	MG	SOFT SED DEFM?
105	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	W	LG	MG	SOFT SED DEFM?
106	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	SOFT SED DEFM?
107	55	0	0	45 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	SOFT SED DEFM?
108	60	0	0	40 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	SOFT SED DEFM?
109	75	0	0	25 G	-	VTK	80	0	0	20	0	P-G	MG	MG	SOFT SED DEFM?
110	80	0	0	20 G	-	VTK	100	0	0	0	0	G	LG-MG	-	TK BPER VNS
111	100	0	0	G	-	VTK	100	0	0	0	0	G	LG-MG	-	TK BPER VNS
112	100	0	0	G	-	VTK	100	0	0	0	0	G	LG-MG	-	TK BPER VNS
113	75	0	0	25 G	WB-IB	TN	0	0	0	100	0	M-P	LG	MGN-G	-
114	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M-P	LG	MGN-G	-
115	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MGN-G	-
116	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MGN-G	-
117	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MGN-G	-
118	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MGN-G	-
119	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MGN-G	-
120	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	-
121	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	-
122	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	-
123	50	0	0	50 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	-
124	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
125	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
126	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
127	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
128	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
129	50	0	0	50 G	WB-IB	TN-M	0	0	0	100	0	M	LG	MG	-
130	40	0	0	60 G	WB-IB	TN	0	0	0	100	0	M	LG	MG	-
131	30	0	0	70 G	WB-IB	TN	0	0	0	100	0	M	LG	LG-MG	-
132	25	0	0	75 G	WB-IB	TN	0	0	0	100	0	M	LG	LG-MG	-
133	25	0	0	75 G	WB-IB	TN	0	0	0	100	0	M	LG	LG-MG	-
134	25	0	0	75 G	WB-IB	TN	0	0	0	100	0	M	LG	LG-MG	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

135	25	0	75	G	WB-IB	VTN	0	0	0	100	0	M	LG	MG-DG	-
136	25	0	75	G	WB-IB	VTN	0	0	0	100	0	M	LG	MG-DG	-
137	25	0	75	G	WB-IB	VTN	0	0	0	100	0	M	LG	MG-DG	-
138	10	0	90	G	WB-IB	VTN	0	0	0	100	0	M	LG	MG-DG	-
139	10	0	90	G	WB-IB	VTN	0	0	0	100	0	M	LG	MG-DG	-
140	0	0	100	G	-	-	0	0	0	0	100	-	-	MG-DG	-
141	75	0	25	G	WB-IB	M	75	0	0	25	0	P-G	LG	MG-DG	-
142	70	0	30	G	WB-IB	M	75	0	0	25	0	P-G	LG	MG-DG	-
143	0	0	100	G	-	-	0	0	0	0	100	-	-	DG	-
144	0	0	100	G	-	-	0	0	0	0	100	-	-	MG-DG	-
145	0	0	100	G	-	-	0	0	0	0	100	-	-	MG-DG	-
146	0	0	100	S	-	-	0	0	0	0	100	-	-	MG-DG	-
147	85	0	15	G	XB	TK	80	0	0	20	0	P-G	LG	MBN	-
148	85	0	15	G	XB	TK	80	0	0	20	0	P-G	LG	MBN	-
149	50	0	50	G	XB	VTK	80	0	0	20	0	P-G	LG	MBN	-
150	20	0	80	G	XB	VTK	80	0	0	20	0	P-G	LG	MBN	-
151	20	0	80	S	XB	VTK	80	0	0	20	0	P-G	LG	MBN	-
152	90	0	10	G	XB	VTK	80	0	0	20	0	P-G	LG	MBN	-
153	90	0	10	S	-	VTK	35	35	0	0	0	P-G	MG	MBN-MG	SH W LS=MGN-DGN
154	75	0	25	G	-	VTK	35	35	0	0	0	P-G	MG	MBN-MG	SH W LS=MGN-DGN
155	10	0	90	G	IB	VTN	0	0	0	100	0	M	MG	MBN-RB	SH W LS=MGN-DGN
156	10	0	90	G	IB	VTN	0	0	0	100	0	M	MG	MBN-RB	SH W LS=MGN-DGN
157	10	0	90	G	IB	VTN	0	0	0	100	0	M	MG	MBN-RB	SH W LS=MGN-DGN
158	10	0	90	G	IB	VTN	0	0	0	100	0	M	MG	MBN-RB	SH W LS=MGN-DGN
159	10	0	90	G	IB	VTN	0	0	0	100	0	M	MG	MBN-RB	SH W LS=MGN-DGN
160	50	0	50	G	-	TN-M	60	10	20	10	0	P	G	RBN	SH W LS=MGN-DGN
161	50	0	50	G	-	TN-M	60	10	20	10	0	P	G	RBN	SH W LS=MGN-DGN
162	0	0	100	G	-	-	0	0	0	0	100	-	-	VDGN	SH W LS=VDGN
163	0	0	100	G	-	-	0	0	0	0	100	-	-	VDGN	SH W LS=VDGN
164	0	0	100	S	-	-	0	0	0	0	100	-	-	VDGN	SH W LS=VDGN
165	75	25	0	S	-	VTK	20	40	0	20	20	CS-P	G	VDGN	RBN
166	0	0	100	G	-	-	0	0	0	0	100	-	-	VDGN	BPER VNS IN SH
167	0	0	100	S	-	-	0	0	0	0	100	-	-	VDGN	BPER VNS IN SH
168	100	0	0	G	-	VTK	20	80	0	0	0	P-G	G	VDGN	BPER VNS IN SH
169	50	25	25	G	IB	VTK	0	0	0	50	50	W-CS	G	VDGN	BPER VNS IN SH

Table A.1- Digitized descriptive lithologic data from drill core GW134.

205	50	25	25	G	WB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
206	30	10	60	G	WB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
207	15	5	80	G	WB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
208	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN-DB	-
209	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN-DB	-
210	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN-DB	-
211	100	0	0	G	-	VTK	80	20	0	0	0	0	P	RBN-DB	CU PACKAGE
212	60	20	20	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	CU PACKAGE
213	40	20	40	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	FOLDED SH
214	60	20	20	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	FOLDED SH
215	80	0	20	G	-	M	0	0	60	40	0	P	G-ROSE	RBN-DB	-
216	80	0	20	G	WB:IB	TN	0	0	0	100	0	W	G	RBN-DB	-
217	40	0	60	G	WB:IB	TN	0	0	0	100	0	W	G	RBN-DB	-
218	10	0	90	G	WB:IB	TN	0	0	0	100	0	W	G	RBN-DB	-
219	85	0	15	G	-	M	60	40	0	0	0	P	G	RBN-DB	-
220	25	0	75	G	WB:IB	M	70	30	0	0	0	P	G	RBN-DB	-
221	25	0	75	G	WB:IB	M	80	20	0	0	0	P	G	RBN-DB	-
222	90	0	10	G	-	TK	80	20	0	0	0	P	G	RBN-DB	-
223	30	20	50	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
224	25	25	50	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
225	20	20	60	G	WB:IB	TN	0	0	0	50	50	W-CS	G	RBN-DB	-
226	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN-DB	-
227	95	0	5	S	-	TK	60	40	0	0	0	P	MG	RBN-DB	CU PACKAGE
228	45	45	10	G	IB	M	0	0	0	50	50	W-CS	MG	RBN-DB	CU PACKAGE
229	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN-DB	-
230	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN-DB	-
231	100	0	0	G	-	VTK	70	0	30	0	0	P-G	MG	RBN-DB	-
232	100	0	0	G	-	VTK	70	0	30	0	0	P-G	MG	RBN-DB	-
233	10	10	80	G	WB:IB	TN	0	0	0	50	50	W-CS	MG	RBN-DB	-
234	10	10	80	S	WB:IB	TN	0	0	0	50	50	W-CS	MG	RBN-DB	-
235	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN	-
236	100	0	0	S	XB	VTK	100	0	0	0	0	P-G	MG	RBN	DOL OIDS
237	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN	-
238	0	0	100	G	-	-	0	0	0	0	100	-	-	RBN	-
239	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

240	100	0	0	S	-	VTK	75	25	0	0	0	P	MG	RBN	-
241	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN	-
242	100	0	0	G	-	VTK	90	10	0	0	0	0	MG	VDBN	-
243	50	50	0	G	IB	TN	0	0	0	50	50	CS	MG	VDBN	-
244	30	30	40	G	IB	TN	0	0	0	50	50	CS	MG	VDBN	-
245	20	20	60	G	IB	TN	0	0	0	50	50	CS	MG	VDBN	-
246	5	5	90	G	IB	TN	0	0	0	50	50	CS	MG	VDBN	-
247	0	0	100	S	-	-	0	0	0	0	100	-	-	VDBN	-
248	100	0	0	S	-	VTK	0	0	50	50	0	0	MG	VDBN-B	PSUED CLASTS
249	0	0	100	G	-	-	0	0	0	0	100	-	-	VDBN-B	-
250	0	0	100	G	-	-	0	0	0	0	100	-	-	VDBN-B	-
251	0	0	100	G	-	-	0	0	0	0	100	-	-	VDBN-B	-
252	0	0	100	S	-	-	0	0	0	0	100	-	-	VDBN-B	-
253	90	0	10	G	-	VTK	100	0	0	0	0	0	G	GN-OLG	BPER&BPAR STYL
254	80	0	20	G	-	VTK	100	0	0	0	0	0	G	GN-OLG	BPER&BPAR STYL
255	20	20	60	G	WB:IB	TN	0	0	0	50	50	CS-W	G	GN-OLG	BPER&BPAR STYL
256	30	30	40	G	WB:IB	TN	0	0	0	50	50	CS-W	G	GN-OLG	BPER&BPAR STYL
257	45	45	10	G	WB:IB	TN	0	0	0	50	50	CS-W	G	GN-OLG	BPER&BPAR STYL
258	90	0	10	G	-	VTK	100	0	0	0	0	0	G	VDGN-R	-
259	90	0	10	G	-	VTK	100	0	0	0	0	0	G	VDGN-R	-
260	25	25	50	G	IB	TN	0	0	0	50	50	CS	G	VDGN-R	-
261	25	25	50	G	IB	TN	0	0	0	50	50	CS	G	VDGN-R	-
262	25	25	50	G	IB	TN	0	0	0	50	50	CS	G	VDGN-R	-
263	5	5	90	S	IB	TN	0	0	0	50	50	CS	G	VDGN-R	-
264	100	0	0	G	-	TK	0	100	0	0	0	0	G	VDGN-R	-
265	0	10	90	S	WB:IB	VTN	0	0	0	25	75	CS	G	RBN	ABNDT VNS LS
266	100	0	0	G	-	TK	0	80	0	0	0	0	G	RBN	-
267	50	0	50	G	-	TK	0	80	0	0	0	0	G	RBN	-
268	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN	-
269	100	0	0	G	-	VTK	80	20	0	0	0	0	G	RBN-OL	-
270	100	0	0	G	-	VTK	80	20	0	0	0	0	G	RBN-OL	-
271	100	0	0	G	-	VTK	80	20	0	0	0	0	G	RBN-OL	-
272	100	0	0	G	-	VTK	80	20	0	0	0	0	G	RBN-OL	-
273	0	0	100	S	-	-	0	0	0	0	100	-	-	RBN	-
274	100	0	0	G	IB	TK	85	15	0	0	0	0	G	RBN	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

275	25	75	0 S	IB	VTN	0	0	0	25	75 CS	G	RBN	-
276	100	0	0 S	-	VTK	100	0	0	0	0 P-G	G	RBN	CU PACKAGE
277	50	50	0 S	IB	TN	0	0	0	50	50 CS	G	RBN	CU PACKAGE
278	0	0	100 S	-	-	0	0	0	0	100 -	-	RBN	-
279	100	0	0 G	-	VTK	100	0	0	0	0 P-G	G	RBN	-
280	100	0	0 S	-	VTK	100	0	0	0	0 P-G	G	RBN	-
281	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN	-
282	0	0	100 S	-	-	0	0	0	0	100 -	-	RBN	-
283	100	0	0 G	-	VTK	100	0	0	0	0 P-G	G	-	ABNDT VNSLS
284	20	30	50 G	IB	TN	0	0	0	25	75 CS	G	-	ABNDT VNSLS
285	0	10	90 S	IB	TN	0	0	0	25	75 CS	G	RBN	-
286	85	15	0 G	-	VTK	100	0	0	0	0 P-G	G	-	-
287	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN	-
288	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN	-
289	40	40	20 G	IB	TN	0	0	0	50	50 CS	MG	RBN	-
290	75	0	25 S	-	M	0	100	0	0	0 P	MG	RBN	GLAUC WLS
291	85	0	15 G	-	VTK	100	0	0	0	0 G	MG	RBN	GLAUC WLS
292	50	0	50 G	-	VTK	100	0	0	0	0 G	MG	RBN	GLAUC WLS
293	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN	-
294	0	0	100 S	-	-	0	0	0	0	100 -	-	RBN	-
295	100	0	0 S	-	TK	100	0	0	0	0 G	MG	RBN	GLAUC WLS
296	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN-GN	OLGN WLS
297	0	0	100 G	-	-	0	0	0	0	100 -	-	RBN-GN	OLGN WLS
298	0	0	100 S	-	-	0	0	0	0	100 -	-	RBN-GN	OLGN WLS
299	25	75	0 G	IB	TN	0	0	0	50	50 CS	MG	RBN-GN	OLGN WLS
300	100	0	0 G	IB	VTK	100	0	0	0	0 P	MG	RBN-GN	V LRGE OOLDS
301	100	0	0 G	IB	VTK	100	0	0	0	0 P	MG	RBN-GN	V LRGE OOLDS
302	15	50	35 G	IB	TN	0	0	0	50	50 CS	MG	RBN-GN	-
303	10	25	65 G	IB	M	0	0	0	50	50 CS	MG	RBN-GN	-
304	80	0	20 G	-	VTK	100	0	0	0	0 G	MG	RBN-B	GLAUC WLS
305	70	0	30 G	-	VTK	100	0	0	0	0 G	MG	RBN-B	GLAUC WLS
306	30	0	70 G	-	VTK	100	0	0	0	0 G	MG	RBN-B	GLAUC WLS
307	0	100	0 G	-	TK	0	0	0	25	75 CS	MG	-	-
308	100	0	0 G	-	VTK	100	0	0	0	0 P-G	MG	RBN-DG	-
309	100	0	0 G	-	VTK	100	0	0	0	0 P-G	MG	RBN-DG	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

310	100	0	0	G	-	VTK	100	0	0	0	0	0	P-G	MG	RBN-DG	-
311	50	0	50	G	-	VTK	100	0	0	0	0	0	P-G	MG	RBN-DG	-
312	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
313	0	0	100	S	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
314	100	0	0	G	-	VTK	100	0	0	0	0	0	P-G	MG	RBN-DG	-
315	100	0	0	G	-	VTK	30	70	0	0	0	0	P-G	MG	RBN-DG	-
316	40	0	60	S	-	VTK	30	70	0	0	0	0	P-G	MG	RBN-DG	-
317	100	0	0	S	-	VTK	65	0	35	0	0	0	P-G	MG	RBN-DG	-
318	0	0	100	S	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
319	50	0	50	S	IB	M	30	35	0	35	0	0	W-P	MG	RBN-DG	-
320	0	0	100	G	-	M	0	0	0	0	0	0	100	-	RBN-DG	-
321	50	0	50	G	-	M	60	0	0	0	0	0	W-P	MG	RBN-DG	-
322	50	0	50	G	-	M	60	0	0	0	0	0	W-P	MG	RBN-DG	-
323	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
324	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
325	0	0	100	S	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
326	100	0	0	S	-	M	30	0	70	0	0	0	P-G	MG	RBN	-
327	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN	-
328	0	0	100	S	-	-	0	0	0	0	0	0	100	-	RBN	-
329	100	0	0	G	-	VTK	0	0	0	70	30	0	P	MG	RBN	-
330	75	25	0	G	IB	TK	0	0	70	30	0	0	P	MG	RBN	-
331	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN	-
332	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN	-
333	90	0	10	G	-	VTK	50	50	0	0	0	0	P-G	MG	RBN	-
334	75	25	0	G	IB	M	25	25	0	0	0	0	CS-P	MG	RBN	-
335	100	0	0	G	-	VTK	40	60	0	0	0	0	P-G	MG	RBN	-
336	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN	-
337	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN	-
338	30	30	40	G	IB	VTK	0	50	0	0	0	0	CS-P	MG	RBN-DG	B SH WLS
339	30	30	40	G	IB	VTK	0	50	0	0	0	0	CS-P	MG	RBN-DG	B SH WLS
340	50	0	50	G	IB	VTK	40	60	0	0	0	0	P-G	MG	RBN-DG	B SH WLS
341	0	0	100	G	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
342	0	0	100	S	-	-	0	0	0	0	0	0	100	-	RBN-DG	-
343	100	0	0	G	IB	TK	0	25	75	0	0	0	P-G	MG	RBN	DGN-B SH WLS
344	50	50	0	G	-	TK	0	0	25	0	0	0	P-CS	MG	RBN	DGN-B SH WLS

Table A.1- Digitized descriptive lithologic data from drill core GW134.

380	0	0	100 G	-	-	0	0	0	0	0	0	100	-	-	DG	BPER & BPAR VNS
381	100	0	0 G	WB:IB	TN	0	0	0	0	0	0	100	0 W-P	LG-MG	DG	BPER & BPAR VNS
382	80	0	20 S	XB	TK	0	0	0	0	0	0	100	0 W	LG	DG	-
383	0	0	100 G	-	-	0	0	0	0	0	0	100	-	-	VDG-B	BPAR VNS
384	90	0	10 G	XB	M	40	0	0	0	0	0	60	0 P	LG	MG	BPER & BPAR VNS
385	90	0	10 G	XB	M	40	0	0	0	0	0	60	0 P	LG	MG	BPER & BPAR VNS
386	90	0	10 G	XB	M	40	0	0	0	0	0	60	0 P	LG	MG	BPER & BPAR VNS
387	80	0	20 G	IB	M	0	0	0	0	0	0	100	0 W-P	LG	MG	CONTORT BDDNG
388	70	0	30 G	IB	M	0	0	0	0	0	0	100	0 W-P	LG	MG	CONTORT BDDNG
389	0	0	100 G	-	-	0	0	0	0	0	0	100	-	-	VDG-B	-
390	50	0	50 G	XB	TN	0	0	0	0	0	0	100	0 W	LG	GN-G	-
391	55	0	45 G	XB	TN	0	0	0	0	0	0	100	0 W	LG	GN-G	-
392	60	0	40 G	XB	M-TK	20	0	0	0	0	0	80	0 P	LG	MGN-G	BPER VNS
393	65	0	35 G	XB	M-TK	20	0	0	0	0	0	80	0 P	LG	MGN-G	BPER VNS
394	70	0	30 G	XB	M-TK	20	0	0	0	0	0	80	0 P	LG	MGN-G	BPER VNS
395	75	0	25 G	XB	M-TK	20	0	0	0	0	0	80	0 P	LG	MGN-G	BPER VNS
396	80	0	20 G	XB	M-TK	20	0	0	0	0	0	80	0 P	LG	MGN-G	BPER VNS
397	80	0	20 G	IB:XB	TN-M	0	0	0	0	0	0	100	0 W	LG	MG	BPER VNS
398	80	0	20 G	IB:XB	TN-M	0	0	0	0	0	0	100	0 W	LG	MG	BPER VNS
399	0	0	100 S	-	-	0	0	0	0	0	0	100	-	-	B	-
400	100	0	0 G	XB	TK	35	35	0	0	0	0	30	0 P-G	LG	-	BPER VNS
401	75	0	25 G	XB:IB	TN	0	0	0	0	0	0	100	0 W-P	LG	MG	BPER VNS
402	75	0	25 G	XB:IB	TN-M	0	40	0	0	0	0	60	0 W-P	LG	MG-DG	BPER VNS & STYLOS
403	75	0	25 G	XB:IB	TN-M	0	40	0	0	0	0	60	0 W-P	LG	MG-DG	BPER VNS & STYLOS
404	75	0	25 G	XB:IB	TN-M	0	40	0	0	0	0	60	0 W-P	LG	MG-DG	BPER VNS & STYLOS
405	75	0	25 G	XB:IB	TN-M	0	40	0	0	0	0	60	0 W-P	LG	MG-DG	BPER VNS & STYLOS
406	75	0	25 S	XB:IB	TN-M	0	40	0	0	0	0	60	0 W-P	LG	MG-DG	BPER VNS & STYLOS
407	100	0	0 S	-	TK	0	25	25	0	0	0	50	0 P	LG	-	BPER VNS & STYLOS
408	0	0	100 G	-	-	0	0	0	0	0	0	100	-	-	B	-
409	50	0	50 G	XB	TK	0	0	0	0	0	0	100	0 P	LG	MG	-
410	0	0	100 G	-	-	0	0	0	0	0	0	100	-	-	DG-B	-
411	80	0	20 G	XB	TK	75	0	0	0	0	0	25	0 P-G	LG	DG	-
412	80	0	20 G	XB	TK	75	0	0	0	0	0	25	0 P-G	LG	DG	-
413	80	0	20 G	XB	TK	75	0	0	0	0	0	25	0 P-G	LG	DG	-
414	80	0	20 G	XB	TK	75	0	0	0	0	0	25	0 P-G	LG	DG	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

415	80	0	20 G	XB	TK	75	0	0	25	0	P-G	LG	DG	-
416	80	0	20 G	XB	TK	75	0	0	25	0	P-G	LG	MG	BPER VNS
417	80	0	20 G	XB	TK	75	0	0	25	0	P-G	LG	MG	BPER VNS
418	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
419	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
420	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
421	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
422	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
423	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
424	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
425	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
426	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
427	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
428	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
429	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
430	80	0	20 G	IB;XB	M-TK	0	0	0	100	0	W	LG	MG-DG	ABDNT BPER VNS
431	40	0	60 G	-	VTN	0	0	0	100	0	W	LG	MG-DG	-
432	75	0	25 G	XB	M	75	0	0	25	0	P-G	LG	MG-DG	-
433	75	0	25 G	XB	M	75	0	0	25	0	P-G	LG	MG-DG	-
434	0	0	100 G	-	-	0	0	0	100	0	-	-	MG-DG	-
435	0	0	100 G	-	-	0	0	0	100	0	-	-	MG-DG	-
436	0	0	100 S	-	-	0	0	0	100	0	-	-	MG-DG	-
437	75	0	25 G	XB	M-TK	75	0	0	25	0	P-G	LG	MG	GLAUC IN LS
438	75	0	25 G	XB;IB	VTN-M	60	0	0	40	0	P	LG	MG	-
439	75	0	25 S	XB;IB	VTN-M	60	0	0	40	0	P	LG	MG	-
440	100	0	0 G	-	YK	85	0	0	15	0	P-G	LG	-	GLAUC IN LS
441	75	0	25 G	IB	TN-VTN	0	0	0	100	0	W-P	LG	GN	-
442	0	0	100 S	-	-	0	0	0	100	0	-	-	GN-DGN	-
443	85	0	15 G	XB	M	40	10	0	50	0	W-P	LG	GN-DGN	GN STYLO IN LS
444	0	0	100 G	-	-	0	0	0	100	0	-	-	DGN-B	-
445	80	0	20 G	XB;WB	TK	90	0	0	10	0	P-G	LG	MG	GLAUC IN LS
446	80	0	20 G	IB;WB	TN	0	0	0	100	0	W-P	LG	MG	-
447	80	0	20 G	IB;WB	TN	0	0	0	100	0	W-P	LG	MG	-
448	80	0	20 G	XB;WB	TK	90	0	0	10	0	P-G	LG	MG	GLAUC IN LS
449	80	0	20 G	IB;WB	TN	0	0	0	100	0	W-P	LG	MG	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.														
485	0	0	100 S	-	-	0	0	0	0	0	0	100	-	B
486	100	0	0 G	IB	TK	0	0	0	0	75	25 W-CS	G	-	-
487	100	0	0 G	IB	TK	0	0	0	0	75	25 W-CS	G	-	-
488	50	0	50 G	IB	TK	0	0	0	0	75	25 W-CS	G	-	BPER VNS COMMON
489	0	0	100 S	-	-	0	0	0	0	0	100	-	B	-
490	100	0	0 S	-	TK	100	0	0	0	0	0 P-G	G	-	BPER VNS COMMON
491	0	0	100 S	-	-	0	0	0	0	0	100	-	B	-
492	100	0	0 G	-	VTK	100	0	0	0	0	0 P-G	G	-	BPER VNS COMMON
493	0	50	50 G	IB	VTK	0	0	0	0	0	100 CS	G	B	BPER VNS COMMON
494	0	0	100 S	-	-	0	0	0	0	0	100	-	B	-
495	50	0	50 G	IB	TK	100	0	0	0	0	0 P-G	G	B	TN LS/SH INTBDS
496	50	0	50 G	IB	TK	100	0	0	0	0	0 P-G	G	B	TN LS/SH INTBDS
497	50	0	50 S	IB	TK	100	0	0	0	0	0 P-G	G	B	TN LS/SH INTBDS
498	75	25	0 G	IB	TN	0	0	0	75	0	25 P-CS	G	-	BPER VNS COMMON
499	0	0	100 S	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
500	100	0	0 G	-	VTK	60	0	0	40	0	0 P-G	G	B	BPER VNS COMMON
501	75	25	0 G	IB	M	45	0	0	25	0	30 P-CS	G	-	BPER VNS COMMON
502	0	25	75 G	IB	TN	0	0	0	0	0	100 CS	G	B	BPER VNS COMMON
503	0	0	100 G	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
504	0	0	100 S	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
505	100	0	0 S	-	TK	40	0	0	60	0	0 P-G	G	-	BPER VNS COMMON
506	0	0	100 G	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
507	0	0	100 G	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
508	0	0	100 S	-	-	0	0	0	0	0	100	-	B	BPER VNS COMMON
509	100	0	0 G	-	TK	50	50	0	0	0	0 P-G	G	-	GLAUC IN LS
510	0	50	50 G	IB	TN	0	0	0	0	0	100 CS	G	VDBN-B	DGN SH NEAR LS
511	0	0	100 G	-	-	0	0	0	0	0	100	-	VDBN-B	-
512	0	0	100 G	-	-	0	0	0	0	0	100	-	VDBN-B	-
513	0	0	100 G	-	-	0	0	0	0	0	100	-	VDBN-B	-
514	0	0	100 S	-	-	0	0	0	0	0	100	-	VDBN-B	-
515	100	0	0 S	-	TK	80	0	0	20	0	0 P-G	G	-	-
516	0	0	100 G	-	-	0	0	0	0	0	100	-	VDBN-B	-
517	0	0	100 G	-	-	0	0	0	0	0	100	-	VDBN-B	-
518	0	0	100 S	-	-	0	0	0	0	0	100	-	VDBN-B	-
519	100	0	0 S	-	M	60	0	0	40	0	0 P	G	-	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

520	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
521	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
522	100	0	0	0	G	-	TK	0	100	0	0	0	0	P	G	-	-
523	25	25	50	50	G	IB	TN	0	50	0	50	0	0	P-CS	G	MGN	CU SEQUENCE
524	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
525	25	25	50	50	S	IB:WB	M-TK	0	0	50	50	0	0	W-CS	G	MGN	PSUEDONODS
526	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
527	100	0	0	0	S	-	TK	0	80	20	0	0	0	P	G	-	-
528	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
529	75	25	0	0	S	IB	M	0	0	35	35	0	30	P-CS	G	-	CU SEQUENCE
530	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
531	100	0	0	0	S	-	M	0	0	100	0	0	0	P-G	G	-	GLAUC.
532	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
533	100	0	0	0	S	-	M	0	40	60	0	0	0	P	G	-	GLAUC.
534	100	0	0	0	G	-	TK	0	100	0	0	0	0	P	G	-	GLAUC.
535	40	40	20	0	G	IB	M-TN	0	33	0	33	0	34	W-CS	G	VDBN-B	-
536	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDBN-B	-
537	80	0	0	20	S	-	M	0	0	85	15	0	0	P	G	VDGN	GLAUC.
538	100	0	0	0	S	-	M	0	50	50	0	0	0	P	ROSE	-	SH CLR CHNG NR LS
539	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	VDGN	SH CLR CHNG NR LS
540	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	VDGN	-
541	50	0	0	50	G	-	M	0	0	0	100	0	0	W-P	ROSE	VDGN	-
542	50	0	0	50	S	-	M	0	0	0	100	0	0	W-P	ROSE	VDGN	-
543	100	0	0	0	G	IB	M	0	0	80	0	20	0	P-G	ROSE	-	-
544	100	0	0	0	G	-	M	0	10	0	70	20	0	P	ROSE	-	FOLD:BPFR STYL
545	100	0	0	0	S	-	M	0	10	0	70	20	0	P	ROSE	-	FOLD:BPFR STYL
546	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	B	-
547	0	0	0	100	S	-	-	0	0	0	0	0	100	-	-	B	-
548	100	0	0	0	S	-	M	0	0	40	60	0	0	P	G	-	GLAUC:PSUEDONODS
549	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	B	-
550	30	0	0	70	G	IB	M	0	0	40	60	0	0	P	G	VDGN	-
551	90	0	0	10	S	IB	M	0	0	40	60	0	0	P	G	VDGN	-
552	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	B	-
553	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	B	-
554	0	0	0	100	G	-	-	0	0	0	0	0	100	-	-	B	-

Table A.1- Digitized descriptive lithologic data from drill core GW134.

625	100	0	0 S	-	M	0	50	0	50	0	50	0 P	G	-	PSUEDONODS
626	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
627	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
628	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	B	-
629	100	0	0 S	-	M	0	100	0	0	0	0	0 P	G	-	-
630	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	B	-
631	100	0	0 S	-	M	0	50	0	50	0	50	0 P	G	-	-
632	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
633	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	B	-
634	100	0	0 S	-	TK	0	80	0	20	0	20	0 P	G	-	-
635	0	0	100 S	-	-	0	0	0	0	0	0	100 -	VDG-B	B	-
636	100	0	0 S	-	TK	0	80	0	20	0	20	0 P	G	-	-
637	0	0	100 S	-	-	0	0	0	0	0	0	100 -	VDGN	B	-
638	100	0	0 S	-	TK	0	80	0	20	0	20	0 P	G	-	-
639	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
640	0	100	0 G	IB	TN-M	0	0	0	0	0	0	100 CS	G	-	BPER VNS
641	0	100	0 G	IB	TN-M	0	0	0	0	0	0	100 CS	G	-	-
642	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
643	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	VDGN	-
644	100	0	0 G	-	VTK	0	0	60	40	0	40	0 P	G	-	-
645	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	VDGN	-
646	0	50	50 G	IB	TN	0	0	0	0	0	0	100 CS	G	B	BPER VNS
647	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	BPER VNS
648	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
649	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	B	-
650	0	20	80 G	IB	TN	0	0	0	0	0	0	100 CS	G	VDGN-B	-
651	100	0	0 S	-	VTK	0	100	0	0	0	0	0 P-G	G	-	-
652	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	VDGN-B	-
653	100	0	0 G	-	VTK	0	0	50	30	0	30	20 W-P	G	-	V TK BPER VNS
654	100	0	0 S	-	VTK	0	0	50	30	0	30	20 W-P	G	-	V TK BPER VNS
655	0	0	100 G	-	-	0	0	0	0	0	0	100 -	-	VDGN	-
656	0	0	100 S	-	-	0	0	0	0	0	0	100 -	-	VDGN	-
657	0	100	0 G	-	TN	0	0	0	0	0	0	100 CS	G	-	-
658	80	20	0 G	IB	TK	0	0	30	40	0	40	30 W-P	G	-	-
659	0	30	70 G	IB	TN	0	0	0	0	0	0	100 CS	G	VDGN	PSUEDONODS

Table A.1- Digitized descriptive lithologic data from drill core GW134.

695	30	0	70 G	WB	VTN	0	0	0	100	0 W	LG	GN-DG	VTN LS IN SH
696	15	0	85 G	WB	VTN	0	0	0	100	0 W	LG	GN-DG	VTN LS IN SH
697	75	0	25 S	-	M	0	100	0	0	0 P	LG	GN-DG	EDGCONGL
698	0	0	100 S	-	-	0	0	0	0	100 -	-	B	-
699	80	0	20 S	-	TK	0	100	0	0	0 P	LG	MG	-
700	0	0	100 G	-	-	0	0	0	0	100 -	-	B	-
701	75	0	25 G	IB	M	0	40	0	60	0 W-P	LG	MG	-
702	40	0	60 G	XB ₂ IB	TN	0	0	0	100	0 W	LG	DG	-
703	40	0	60 G	XB ₂ IB	TN	0	0	0	100	0 W	LG	DG	-
704	80	0	20 G	-	TK	0	100	0	0	0 P-G	LG	DG	BPER VNS
705	70	0	30 S	-	TK	0	100	0	0	0 P-G	LG	DG	-
706	0	0	100 G	-	-	0	0	0	0	100 -	-	B	-
707	0	0	100 G	-	-	0	0	0	0	100 -	-	B	-
708	0	0	100 G	-	-	0	0	0	0	100 -	-	B	-
709	0	0	100 G	-	-	0	0	0	0	100 -	-	B	-
710	75	0	25 G	IB	TK	0	50	0	50	0 P-G	G	DG	MDSTN/CHRT CLSTS/
711	75	0	25 G	XB	M	0	50	0	50	0 P-G	G	DG	-
712	75	0	25 G	XB	M	0	50	0	50	0 P-G	G	DG	-
713	75	0	25 G	XB	M	0	50	0	50	0 P-G	G	DG	-
714	75	0	25 G	WB	M	0	50	0	50	0 P-G	G	DG	-
715	75	0	25 G	WB	M	0	50	0	50	0 P-G	G	DG	-
716	75	0	25 G	WB	M	0	50	0	50	0 P-G	G	DG	-
717	75	0	25 G	WB	M	0	50	0	50	0 P-G	G	DG	-
718	45	0	55 G	WB	M	0	50	0	50	0 P-G	G	DG	-
719	45	0	55 G	WB	M	0	50	0	50	0 P-G	G	DG	TN LS IN SH
720	45	0	55 G	WB	M	0	50	0	50	0 P-G	G	DG	TN LS IN SH
721	85	0	15 G	IB	TN-M	0	20	0	80	0 W-P	LG	MG	TN LS IN SH
722	85	0	15 G	XB	TN-M	0	20	0	80	0 W-P	LG	MG	-
723	85	0	15 G	XB	TN-M	0	20	0	80	0 W-P	LG	MG	-
724	85	0	15 G	IB	TN-M	0	20	0	80	0 W-P	LG	MG	-
725	75	0	25 G	XB	TN	0	0	0	100	0 W	LG	DG	BPER VNS
726	75	0	25 G	IB	TN	0	0	0	100	0 W	LG	DG	BPER VNS
727	75	0	25 G	IB	TN	0	0	0	100	0 W	LG	DG	BPAR VNS
728	50	0	50 G	IB	VTN	0	0	0	100	0 W	LG	GN-G	FOLDED BEDS
729	50	0	50 G	IB	VTN	0	0	0	100	0 W	LG	GN-G	FOLDED BEDS

Table A.1- Digitized descriptive lithologic data from drill core GW134.

730	75	0	25	G	WB	M	0	10	0	90	0	W-P	LG	GN-DG	-
731	75	0	25	G	WB	M	0	10	0	90	0	W-P	LG	GN-DG	-
732	75	0	25	G	XB	TN	0	10	0	90	0	W-P	LG	GN-DG	-
733	75	0	25	G	XB	TN	0	10	0	90	0	W-P	LG	GN-DG	BPER VNS
734	75	0	25	G	XB	TN	0	10	0	90	0	W-P	LG	GN-DG	BPER VNS
735	75	0	25	G	XB	M	0	0	0	100	0	W	LG	GN-DG	-
736	75	0	25	G	XB	M	0	0	0	100	0	W	LG	GN-DG	-
737	75	0	25	G	IB	TN	0	0	0	100	0	W	LG	GN-DG	-
738	75	0	25	G	IB	TN	0	0	0	100	0	W	LG	GN-DG	-
739	75	0	25	G	IB	TN	0	0	0	100	0	W	LG	GN-DG	-
740	75	0	25	G	IB	TN	0	0	0	100	0	W	LG	GN-DG	-

APPENDIX B
ELECTRON MICROPROBE DATA

Electron Microprobe Data
Explanation of Data Columns

<u>Column</u>	<u>Explanation</u>
1	Phase analysed (BLD/EQCC-bladed/equant calcite cement; RF ECH NU-echinoderm nucleous on fibrous radial ooid; EQCC-equant calcite cement; FIBCC-fibrous calcite cement; RF COAT RI-fibrous radial ooid coating; RF-fibrous radial ooid; RF/RP-fibrous radial/prismatic radial ooid; MO-monocrystalline ooid; MO INT STD-monocrystalline ooid internal standard; RP RAY-prismatic radial ooid ray; RP-prismatic radial ooid; VEINCC-calcite vein)
2	Description of phase (same code as Column 1 or self-explanatory)
3	Thin section number (X-1: Beach Grove; GW139229 and GW139199: ORR cores)
4	wt% MgCO_3
5	mol% MgCO_3
6	wt% Mg
7	ppm Mg
8	wt% CaCO_3
9	mol% CaCO_3
10	wt% Ca
11	wt% MnCO_3
12	wt% Mn
13	ppm Mn
14	wt% FeCO_3
15	wt% Fe
16	ppm Fe
17	wt% Total

Table B.1 Electron Microprobe data for Nohelchuky Oxide, Concentrations, and Vein Calcities.

Analysis	Phase	T.S.	wt% MgCO ₃	mol% MgCO ₃	wt% Mg	ppm Mg	wt% CaCO ₃	mol% CaCO ₃	wt% Ca	wt% MnCO ₃	wt% Mn	ppm Mn	wt% FeCO ₃	wt% Fe	ppm Fe	wt% Total
RE COAT RI	Blk/Cr	X-1	1.8159	2.17	0.3231	5231	96.6175	97.12	38.6386	0.1834	0.0876	876	0.6342	0.3053	3053	99.2510
RE COAT RI	Blk/Cr	X-1	1.7520	2.08	0.5047	5047	96.9111	97.15	38.7760	0.1626	0.0776	776	0.7253	0.3495	3495	99.5512
RE COAT RI	Blk/Cr	X-1	1.8074	2.14	0.3206	5206	97.5599	97.21	39.0157	0.2078	0.0992	992	0.5269	0.2538	2538	100.1020
RE COAT RI	Blk/Cr	X-1	2.1257	2.49	0.6123	6123	97.8484	96.67	39.1511	0.2560	0.1223	1223	0.7178	0.3458	3458	100.9480
RE COAT RI	Blk/Cr	X-1	3.9997	4.78	1.1521	11521	91.6530	94.26	37.4724	0.1869	0.0893	893	0.9219	0.4441	4441	98.7614
RE COAT RI	Blk/Cr	X-1	1.8607	2.21	0.3360	5360	96.9679	97.10	38.7988	0.2008	0.0959	959	0.5906	0.2845	2845	99.6200
RE COAT RI	Blk/Cr	X-1	2.0007	2.40	0.3763	5763	95.7977	96.89	38.3103	0.2152	0.1123	1123	0.5701	0.2746	2746	98.6037
RE COAT RI	Blk/Cr	X-1	2.0007	2.40	0.3763	5763	95.7977	96.89	38.3103	0.2152	0.1123	1123	0.5701	0.2746	2746	98.6037
RE COAT RI	Blk/Cr	X-1	0.9881	1.19	0.2846	2846	96.5382	96.07	38.6268	0.1799	0.0644	644	0.6614	0.3186	3186	98.3679
RE COAT RI	Blk/Cr	X-1	2.7927	2.84	0.6855	6855	95.8068	96.46	38.3342	0.1348	0.0644	644	0.6614	0.3186	3186	98.3679
RE COAT RI	Blk/Cr	X-1	2.4089	2.90	0.6939	6939	94.7223	96.14	37.9003	0.2906	0.1388	1388	0.8054	0.3880	3880	98.2272
RE COAT RI	Blk/Cr	X-1	4.2659	5.04	1.2285	12285	94.3901	94.07	37.7673	0.2524	0.1205	1205	0.7600	0.3704	3704	99.6763
RE COAT RI	Blk/Cr	X-1	3.6662	4.37	1.0560	10560	94.4272	94.89	37.7822	0.2076	0.0992	992	0.6381	0.3074	3074	98.9321
RE COAT RI	Blk/Cr	X-1	2.7482	2.67	0.6476	6476	96.8313	96.72	38.7441	0.1452	0.0694	694	0.5984	0.2883	2883	99.2147
RE COAT RI	Blk/Cr	X-1	1.4491	1.73	0.4171	4171	96.9595	97.57	38.7954	0.2077	0.0992	992	0.5984	0.2883	2883	99.2147
RE COAT RI	Blk/Cr	X-1	1.0427	1.25	0.3003	3003	96.9538	98.01	38.7939	0.1834	0.0876	876	0.6600	0.3208	3208	98.8478
RE COAT RI	Blk/Cr	X-1	2.3782	2.85	0.6850	6850	95.8035	96.57	38.3229	0.2181	0.1012	1012	0.4500	0.2192	2192	98.8548
RE COAT RI	Blk/Cr	X-1	1.3713	1.65	0.3967	3967	96.9877	97.79	38.8067	0.1141	0.0545	545	0.5261	0.2534	2534	99.0052
RE COAT RI	Blk/Cr	X-1	0.4230	0.50	0.1218	1218	99.1397	99.39	39.6672	0.0003	0.0001	1	0.1277	0.0615	615	99.6906
RE COAT RI	Blk/Cr	X-1	1.0784	1.29	0.3106	3106	97.4734	98.18	39.4018	0.1246	0.0595	595	0.4829	0.2326	2326	99.1613
RE COAT RI	Blk/Cr	X-1	1.3684	1.63	0.3942	3942	97.3449	97.68	38.9496	0.2041	0.0975	975	0.5902	0.2843	2843	99.5076
RE COAT RI	Blk/Cr	X-1	1.0981	1.30	0.3163	3163	98.4732	98.13	39.4011	0.0935	0.0447	447	0.5608	0.2730	2730	100.2316
RE COAT RI	Blk/Cr	X-1	0.9582	1.16	0.2760	2760	97.2696	96.79	38.1193	0.1148	0.0532	532	1.2125	0.5841	5841	99.5532
RE COAT RI	Blk/Cr	X-1	1.3001	1.55	0.3745	3745	97.5962	97.96	39.0502	0.1985	0.0948	948	0.3695	0.1780	1780	99.4643
RE COAT RI	Blk/Cr	X-1	1.6856	2.04	0.4855	4855	97.3976	97.51	38.9707	0.0348	0.0166	166	0.5281	0.2544	2544	99.6462
RE COAT RI	Blk/Cr	X-1	2.0138	2.41	0.5855	5855	97.3366	97.27	38.9463	0.0836	0.0399	399	0.2822	0.1359	1359	99.2351
RE COAT RI	Blk/Cr	X-1	0.8590	1.02	0.2471	2471	98.6707	98.69	39.4801	0.0348	0.0166	166	0.3019	0.1454	1454	99.8669
RE COAT RI	Blk/Cr	X-1	1.5478	1.87	0.4458	4458	96.0066	97.61	38.4149	0.0349	0.0167	167	0.5250	0.2329	2329	98.1162
RE COAT RI	Blk/Cr	X-1	0.5696	0.68	0.1641	1641	97.9769	98.85	39.2025	0.1246	0.0595	595	0.4149	0.1998	1998	99.0859
RE COAT RI	Blk/Cr	X-1	1.8888	2.25	0.5441	5441	97.0920	97.26	38.8484	0.0761	0.0363	363	0.4004	0.2362	2362	99.5474
RE COAT RI	Blk/Cr	X-1	0.6052	0.72	0.1743	1743	99.2069	99.22	39.8147	0.0762	0.0364	364	0.0000	0.0000	0	100.1882
RE COAT RI	Blk/Cr	X-1	0.8394	1.00	0.2418	2418	98.8028	98.86	39.5320	0.1073	0.0512	512	0.0558	0.0269	269	99.8051
RE COAT RI	Blk/Cr	X-1	1.1161	1.34	0.3215	3215	96.8782	98.17	38.7629	0.1312	0.0499	499	0.3226	0.2317	2317	98.5480
RE COAT RI	Blk/Cr	X-1	0.9170	1.09	0.2642	2642	98.3389	98.38	39.3473	0.1661	0.0793	793	0.4425	0.2131	2131	99.8645
RE COAT RI	Blk/Cr	X-1	0.9159	1.16	0.2811	2811	97.0094	98.35	39.1255	0.1108	0.0529	529	0.4466	0.2152	2152	99.4428
RE COAT RI	Blk/Cr	X-1	1.1543	1.37	0.3325	3325	97.9345	98.19	38.1856	0.1592	0.0760	760	0.3907	0.1882	1882	99.6387
RE COAT RI	Blk/Cr	X-1	1.2154	2.07	0.4999	4999	97.2668	97.61	38.9264	0.0762	0.0364	364	0.2553	0.1230	1230	99.3536
RE COAT RI	Blk/Cr	X-1	1.2630	1.51	0.3644	3644	97.3216	98.22	39.0203	0.0727	0.0347	347	0.2315	0.1115	1115	99.0908
RE COAT RI	Blk/Cr	X-1	1.1569	1.36	0.3333	3333	99.0386	98.40	39.6213	0.0900	0.0430	430	0.1995	0.0865	865	100.4650
RE COAT RI	Blk/Cr	X-1	2.4720	2.93	0.7121	7121	96.9799	96.42	38.6507	0.1661	0.0793	793	0.5821	0.2804	2804	99.8182
RE COAT RI	Blk/Cr	X-1	2.5183	3.02	0.7341	7341	96.4791	96.25	38.6032	0.1695	0.0809	809	0.6736	0.3245	3245	99.8707
RE COAT RI	Blk/Cr	X-1	0.8247	0.99	0.2376	2376	97.4313	98.51	38.9842	0.1834	0.0876	876	0.3879	0.1844	1844	98.8222
RE COAT RI	Blk/Cr	X-1	0.6823	0.81	0.1966	1966	98.6249	98.88	39.4618	0.0554	0.0264	264	0.2952	0.1422	1422	99.6380
RE COAT RI	Blk/Cr	X-1	1.8341	2.19	0.5283	5283	96.6471	97.22	38.6704	0.1488	0.0711	711	0.5104	0.2555	2555	99.1604
RE COAT RI	Blk/Cr	X-1	1.1438	1.37	0.3295	3295	96.6200	98.18	39.0597	0.0762	0.0364	364	0.4706	0.2267	2267	99.2628
RE COAT RI	Blk/Cr	X-1	0.8892	1.07	0.2561	2561	96.5262	97.93	38.6221	0.1661	0.0793	793	0.4706	0.2267	2267	99.2628
RE COAT RI	Blk/Cr	X-1	1.2181	1.47	0.3509	3509	95.5829	97.47	38.2446	0.3695	0.1765	1765	0.7053	0.3397	3397	99.5528
RE COAT RI	Blk/Cr	X-1	1.2906	1.56	0.3716	3716	95.6970	97.34	38.2902	0.3664	0.1750	1750	0.8844	0.4264	4264	98.2378
RE COAT RI	Blk/Cr	X-1	1.3064	1.58	0.3763	3763	95.4016	97.32	38.1721	0.4252	0.2031	2031	0.8208	0.3954	3954	97.9530
RE COAT RI	Blk/Cr	X-1	1.2814	1.55	0.3691	3691	95.7189	97.34	38.2990	0.3733	0.1783	1783	0.8966	0.4318	4318	98.2701
RE COAT RI	Blk/Cr	X-1	0.8070	0.97	0.2325	2325	96.3147	97.65	38.5374	0.6052	0.2890	2890	0.9607	0.4628	4628	98.6876
RE COAT RI	Blk/Cr	X-1	0.8337	1.01	0.2401	2401	95.2332	96.83	38.1047	0.3802	0.1816	1816	2.0817	1.0028	10028	98.5289
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063	1.2035	0.5797	5797	99.4410
RE COAT RI	Blk/Cr	X-1	0.8193	1.01	0.2447	2447	96.6769	97.31	38.6823	0.4319	0.2063	2063				

APPENDIX C
VEIN AND JOINT ORIENTATION DATA

Appendix C
Explanation of Data Columns

<u>Column</u>	<u>Explanation</u>
1	Borehole number
2	Trend of intersection between bed-normal vein and bedding surface
3	no data
4	Borehole number
5	Trend of intersection between bed-normal vein and bedding surface
6	no data
7	Name of road adjacent to outcrop on ORR
8	Strike and dip of bed-normal joints in outcrop

Table C-1. Vein and joint orientation data from the Oak Ridge Reservation. All data is unrotated.

Well	Trend	Well	Trend	Outcrop	Strike, Dip
GW134	N 5.0 W,	GW139	N 5.0 W,	Bear Creek Rd.	N 6.0 W, 57.0 E
GW134	N 15.0 W	GW139	N 25.0 W	Bear Creek Rd.	N 6.0 E, 59.0 E
GW134	N 15.0 W	GW139	N 25.0 W	Bear Creek Rd.	N 5.0 E, 55.0 E
GW134	N 15.0 W	GW139	N 25.0 W	Bear Creek Rd.	N 89.0 W, 63.0 N
GW134	N 25.0 W	GW139	N 35.0 W	Bear Creek Rd.	N 80.0 W, 65.0 N
GW134	N 25.0 W	GW139	N 35.0 W	Bear Creek Rd.	N 20.0 E, 50.0 S
GW134	N 25.0 W	GW139	N 35.0 W	Bear Creek Rd.	N 42.0 E, 90.0 S
GW134	N 25.0 W	GW139	N 35.0 W	Bear Creek Rd.	N 5.0 E, 40.0 S
GW134	N 35.0 W	GW139	N 45.0 W	Bear Creek Rd.	N 85.0 E, 36.0 S
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	N 70.0 W, 35.0 E
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	N 5.0 E, 90.0 E
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 50.0 E, 80.0 S
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 46.0 E, 72.0 S
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 15.0 E, 64.0 W
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 15.0 E, 70.0 W
GW134	N 45.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 50.0 W, 45.0 N
GW134	N 55.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 45.0 W, 54.0 N
GW134	N 55.0 W	GW139	N 45.0 W	Bear Creek Rd.	N 65.0 W, 70.0 N
GW134	N 55.0 W	GW139	N 45.0 W	Bear Creek Rd.	N 71.0 W, 52.0 N
GW134	N 65.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 85.0 W, 77.0 W
GW134	N 65.0 W	GW139	N 45.0 W	Bear Creek Rd.	S 80.0 W, 69.0 W
GW134	N 65.0 W	GW139	N 55.0 W	Bear Creek Rd.	S 1.0 E, 87.0 W
GW134	N 75.0 W	GW139	N 55.0 W	Bear Creek Rd.	S 5.0 E, 77.0 S
GW134	N 85.0 W	GW139	N 55.0 W	Bear Creek Rd.	N 60.0 W, 58.0 N
GW134	N 85.0 W	GW139	N 55.0 W	Bear Creek Rd.	N 69.0 W, 40.0 N
GW134	N 85.0 W	GW139	N 65.0 W	Bear Creek Rd.	N 60.0 W, 69.0 N
GW134	N 85.0 W	GW139	N 75.0 W	Bear Creek Rd.	N 81.0 W, 50.0 N
GW134	N 5.0 E,	GW139	N 75.0 W	Bear Creek Rd.	N 35.0 W, 90.0 N
GW134	N 5.0 E,	GW139	N 75.0 W	Bear Creek Rd.	N 30.0 W, 67.0 N
GW134	N 5.0 E,	GW139	N 85.0 W	Bear Creek Rd.	N 45.0 W, 40.0 N
GW134	N 5.0 E,	GW139	N 85.0 W	Bear Creek Rd.	S 48.0 W, 46.0 N
GW134	N 5.0 E,	GW139	N 85.0 W	Bear Creek Rd.	S 35.0 W, 37.0 N
GW134	N 5.0 E,	GW139	N 85.0 W	Bear Creek Rd.	S 25.0 W, 61.0 N
GW134	N 5.0 E,	GW139	N 5.0 E,	Bear Creek Rd.	N 25.0 W, 90.0 N
GW134	N 5.0 E,	GW139	N 5.0 E,	Bear Creek Rd.	N 26.0 W, 90.0 N

Table C-1. Vein and joint orientation data from the Oak Ridge Reservation. All data is unrotated.					
GW134	N 5.0 E,	GW139	N 5.0 E,	Bear Creek Rd	S 35.0 E, 88.0 S
GW134	N 5.0 E,	GW139	N 15.0 E,	Bear Creek Rd	S 27.0 E, 85.0 S
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	N 88.0 W, 64.0 N
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	N 75.0 W, 55.0 N
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	S 0 W, 50.0 N
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	S 45.0 W, 40.0 N
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	S 20.0 W, 40.0 N
GW134	N 15.0 E,	GW139	N 15.0 E,	Bear Creek Rd	N 80.0 W, 50.0 N
GW134	N 25.0 E,	GW139	N 15.0 E,	Bear Creek Rd	N 75.0 W, 77.0 N
GW134	N 25.0 E,	GW139	N 15.0 E,	Bear Creek Rd	N 70.0 W, 52.0 N
GW134	N 25.0 E,	GW139	N 15.0 E,	Bear Creek Rd	S 0 E, 62.0 N
GW134	N 25.0 E,	GW139	N 25.0 E,	Bear Creek Rd	N 75.0 W, 74.0 N
GW134	N 25.0 E,	GW139	N 25.0 E,	Bear Creek Rd	N 76.0 W, 43.0 N
GW134	N 35.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 10.0 W, 65.0 N
GW134	N 35.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 15.0 W, 67.0 N
GW134	N 35.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 10.0 W, 65.0 N
GW134	N 35.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 25.0 W, 80.0 N
GW134	N 35.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 25.0 W, 72.0 N
GW134	N 45.0 E,	GW139	N 25.0 E,	Bear Creek Rd	S 69.0 W, 69.0 N
GW134	N 45.0 E,	GW139	N 45.0 E,	Bear Creek Rd	S 0 E, 60.0 N
GW134	N 45.0 E,	GW139	N 45.0 E,	Bear Creek Rd	S 80.0 W, 55.0 N
GW134	N 45.0 E,	GW139	N 45.0 E,	Bear Creek Rd	N 45.0 W, 90.0 N
GW134	N 45.0 E,	GW139	N 45.0 E,	Bear Creek Rd	S 40.0 E, 82.0 S
GW134	N 45.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 15.0 E, 60.0 S
GW134	N 45.0 E,	GW139	N 55.0 E,	Bear Creek Rd	N 80.0 W, 56.0 N
GW134	N 45.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 30.0 E, 85.0 W
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 27.0 E, 84.0 W
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 25.0 E, 80.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 20.0 E, 79.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 25.0 E, 84.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 30.0 E, 74.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 30.0 E, 73.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 23.0 E, 90.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 24.0 E, 80.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 30.0 E, 85.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 40.0 E, 89.0 S
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	N 35.0 W, 85.0 N
GW134	N 55.0 E,	GW139	N 55.0 E,	Bear Creek Rd	S 80.0 W, 42.0 N

Table C-1 - Vein and joint orientation data from the Oak Ridge Reservation. All data is unrotated.							
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	S 75.0 W, 38.0 N
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	N 85.0 W, 35.0 N
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	N 75.0 W, 75.0 N
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	S 0 E, 70.0 N
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	S 75.0 W, 50.0 N
GW134	N 55.0 E		GW139	N 55.0 E		Bear Creek Rd	S 75.0 W, 50.0 N
GW134	N 55.0 E		GW139	N 65.0 E		Bear Creek Rd	S 0 E, 82.0 N
GW134	N 55.0 E		GW139	N 75.0 E		Bear Creek Rd	S 75.0 W, 60.0 N
GW134	N 55.0 E		GW139	N 75.0 E		Bear Creek Rd	S 25.0 E, 67.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 40.0 E, 67.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 30.0 E, 90.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 75.0 W, 59.0 N
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 0 E, 80.0 N
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 30.0 E, 85.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 35.0 E, 85.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 34.0 E, 84.0 S
GW134	N 55.0 E		GW139	N 85.0 E		Bear Creek Rd	S 34.0 E, 86.0 S
GW134	N 65.0 E		GW139	N 85.0 E		Bear Creek Rd	S 27.0 E, 82.0 S
GW134	N 65.0 E					Bear Creek Rd	S 27.0 E, 77.0 S
GW134	N 65.0 E					Bear Creek Rd	S 85.0 W, 52.0 N
GW134	N 65.0 E					Bear Creek Rd	N 85.0 W, 48.0 N
GW134	N 65.0 E					Bear Creek Rd	N 82.0 W, 39.0 N
GW134	N 75.0 E					Bear Creek Rd	S 45.0 W, 55.0 N
GW134	N 75.0 E					Bear Creek Rd	S 70.0 W, 30.0 N
GW134	N 75.0 E					Bear Creek Rd	S 25.0 W, 50.0 N
GW134	N 75.0 E					Bear Creek Rd	S 25.0 W, 50.0 N
GW134	N 85.0 E					Bear Creek Rd	S 75.0 W, 72.0 N
GW134	N 85.0 E					Bear Creek Rd	S 75.0 W, 62.0 N
GW134	N 85.0 E					Bear Creek Rd	S 77.0 W, 60.0 N
GW134	N 85.0 E					Bear Creek Rd	S 75.0 W, 70.0 N
GW134	N 85.0 E					Bear Creek Rd	S 75.0 W, 85.0 N
GW134	N 85.0 E					Bear Creek Rd	N 50.0 W, 75.0 N
GW134	N 85.0 E					Bear Creek Rd	N 25.0 W, 88.0 N
GW134	N 95.0 E					Bear Creek Rd	N 35.0 W, 88.0 N
						Bear Creek Rd	N 40.0 W, 90.0 N
						Bear Creek Rd	N 45.0 W, 74.0 N
						Bear Creek Rd	S 20.0 E, 75.0 S
						Bear Creek Rd	S 40.0 E, 88.0 S

APPENDIX D
FLUID INCLUSION MICROTHERMOMETRIC DATA

Appendix D
Explanation of Data Columns

<u>Column</u>	<u>Explanation</u>
1	Sampled core borehole identification
2	Downhole footage from ground surface (in feet)
3	Vein type based on orientation with respect to bedding plane and bedding strike, respectively (bd par: bed parallel; bdblq/strk par: bed oblique and parallel to strike; bdperp/dip par: bed perpendicular and parallel to dip direction; bdperp/obq dip: bed perpendicular and oblique to dip direction; bdperp/strk par: bed perpendicular and parallel to strike)
4	Chip number
5	Homogenization temperature (°C)
6	Maximum dimension of inclusion (μm)
7	Eutectic temperature (°C)
8	Final ice melting temperature (°C)
9	Final hydrohalite melting temperature (°C)

Table D.1 Fluid inclusion data from calcite veins.

Well	Foot.	Vein Type	Chip	Th	Max. μm	Te	Tm	Tmh
GW134	622	bd par	2	158.5	10			
GW134	622	bd par	2	129.5	5			
GW134	622	bd par	2	143.3	15			
GW134	622	bd par	4	56.0	12			
GW134	622	bd par	5	65.0	5			
GW134	622	bd par	5	65.0	8			
GW134	622	bd par	5	63.0	6			
GW134	622	bd par	5	92.0	12			
GW134	622	bd par	5	180.3	10			
GW134	622	bd par	1	115.0	10		-25.5	-11
GW134	697	bd par	1	67.6	5			
GW134	697	bd par	2	81.0	5			
GW134	697	bd par	2	73.1	3			
GW140	1178	bd par	1	40.0	7			
GW140	1178	bd par	1	39.7	6			
GW140	1178	bd par	2	55.0	5			
GW140	1178	bd par	2	55.0	5			
GW140	1178	bd par	2	52.8	4			
GW140	1178	bd par	2	53.6	5			
GW140	1178	bd par	1	42.0	8			
GW140	1178	bd par	1	78.8	14	-52.8	-25.1	
GW140	1178	bd par	1	53.5	12	-46	-23.7	
GW140	1178	bd par	2	63.5	4			
GW140	1178	bd par	2	56.6	4			
GW140	1178	bd par	2	61.0	5	-39	-14	
GW140	1178	bd par	2	56.6	5			
GW140	1178	bd par	2	56.1	5			
GW140	1178	bd par	3	61.1	4			
GW140	1178	bd par	3	63.7	5			
GW140	1178	bd par	3	58.2	4	-46.6	-18	
GW140	1178	bd par	3	63.7	5			
GW139	229	bdoblq/strk par	3	149.5	3			
GW139	229	bdoblq/strk par	3	155.5	5			
GW139	229	bdoblq/strk par	3	258.0	15			
GW139	229	bdoblq/strk par	3	184.0	10			
GW139	229	bdoblq/strk par	3	146.0	5			
GW139	229	bdoblq/strk par	3	166.5	10			
GW139	229	bdoblq/strk par	3	176.0	12			
GW139	229	bdoblq/strk par	3	173.0	2			
GW139	229	bdoblq/strk par	3	136.5	15		-27	-9
GW139	229	bdoblq/strk par	3	145.0	12		-27	
GW139	229	bdoblq/strk par	3	172.5	10		-28	-10
GW139	229	bdoblq/strk par	3	174.0	15		-27	-6.5
GW139	229	bdoblq/strk par	3	137.0	5			
GW139	229	bdoblq/strk par	3	76.0	5			
GW139	229	bdoblq/strk par	3	96.0	5			
GW139	229	bdoblq/strk par	3	95.0	5			
GW139	229	bdoblq/strk par	3	180.0	10			
GW139	229	bdoblq/strk par	3	176.5	15			
GW139	229	bdoblq/strk par	5	81.0	8			

Table 2.1. Fluid inclusion data from calcite veins.

GW139	229	bdoblg/strk par	5	188.0	8			
GW139	229	bdoblg/strk par	5	195.0	10			
GW139	229	bdoblg/strk par	5	159.0	5			
GW139	229	bdoblg/strk par	5	154.0	8			
GW139	229	bdoblg/strk par	5	195.0	5			
GW139	229	bdoblg/strk par	5	195.0	5			
GW139	229	bdoblg/strk par	5	198.0	9			
GW139	229	bdoblg/strk par	5	192.0	7			
GW139	229	bdoblg/strk par	5	125.0	10			
GW139	229	bdoblg/strk par	5	171.0	10			
GW139	229	bdoblg/strk par	5	168.0	5			
GW139	229	bdoblg/strk par	5	150.0	10			
GW139	229	bdoblg/strk par	5	145.0	3			
GW139	229	bdoblg/strk par	5	150.0	2			
GW139	229	bdoblg/strk par	5	203.0	3			
GW139	229	bdoblg/strk par	5	151.0	4			
GW139	229	bdoblg/strk par	1	117.0	5			
GW139	229	bdoblg/strk par	2	196.0	5			
GW139	229	bdoblg/strk par	2	173.1	4			
GW139	229	bdoblg/strk par	3	73.6	20		-31.5	-9.4
GW139	229	bdoblg/strk par	1	200.2	6			
GW139	229	bdoblg/strk par	1	190.5	12			
GW139	229	bdoblg/strk par	1	201.3	17	-39	-24.2	
GW139	229	bdoblg/strk par	2	60.8	5	-40	-26	
GW139	229	bdoblg/strk par	1	72.5	35	-51.8	-25.6	-8.5
GW134	407	bdperp/dip par	1	59.0	3			
GW134	407	bdperp/dip par	1	57.5	7			
GW134	407	bdperp/dip par	1	73.7	7			
GW134	407	bdperp/dip par	1	77.0	5			
GW134	407	bdperp/dip par	2	70.8	3			
GW134	407	bdperp/dip par	2	72.4	4			
GW134	407	bdperp/dip par	2	83.0	5			
GW134	407	bdperp/dip par	2	69.0	3			
GW134	407	bdperp/dip par	2	80.6	4			
GW139	180	bdperp/dip par	1	47.5	8			
GW139	180	bdperp/dip par	1	46.7	8			
GW139	180	bdperp/dip par	1	46.1	15			
GW139	180	bdperp/dip par	2	62.0				
GW139	180	bdperp/dip par	2	65.0				
GW139	180	bdperp/dip par	2	73.0				
GW139	180	bdperp/dip par	2	75.0			-19	
GW139	180	bdperp/dip par	2	72.0			-22	
GW139	180	bdperp/dip par	4	68.4	5	-33.5	-21.5	
GW139	180	bdperp/dip par	4	55.0	3			
GW139	180	bdperp/dip par	4	53.8	5			
GW139	180	bdperp/dip par	4	70.0	6			
GW139	180	bdperp/dip par	1	75.0	4			
GW139	180	bdperp/dip par	1	83.3	8			
GW139	180	bdperp/dip par	1	62.7	4			
GW139	180	bdperp/dip par	1	68.2	2			
GW134	76	bdperp/obq dip	2	100.6	30			

Table p.1 - Fluid inclusion data from calcite veins.

GW134	76	bdperp/obq dip	2	122.7	10			
GW134	76	bdperp/obq dip	1	63.1	5			
GW134	76	bdperp/obq dip	1	83.7	6			
GW134	76	bdperp/obq dip	1	61.0	3			
GW134	76	bdperp/obq dip	1	69.4	6			
GW134	360	bdperp/strk par	1	62.3	3			
GW134	360	bdperp/strk par	1	64.3	5			
GW134	360	bdperp/strk par	1	61.8	5			
GW134	360	bdperp/strk par	1	54.7	3			
GW134	360	bdperp/strk par	1	52.9	3			
GW134	360	bdperp/strk par	1	53.5	2			
GW134	400	bdperp/strk par	1	79.9	5			
GW134	400	bdperp/strk par	1	80.4	10	-36.5		-8
GW134	400	bdperp/strk par	1	101.0	7			
GW134	400	bdperp/strk par	1	91.3	10			
GW134	440	bdperp/strk par	1	52.0	5			
GW134	440	bdperp/strk par	1	54.0	2			
GW134	440	bdperp/strk par	2	46.8	4			
GW134	440	bdperp/strk par	2	43.8	3			
GW134	440	bdperp/strk par	2	46.0	3			
GW134	440	bdperp/strk par	2	43.0	4			
GW134	440	bdperp/strk par	2	59.6	4			
GW134	440	bdperp/strk par	1	64.2	5			
GW134	440	bdperp/strk par	1	72.7	7			
GW134	440	bdperp/strk par	2	49.2	4			
GW134	440	bdperp/strk par	2	43.9	12			
GW134	440	bdperp/strk par	2	59.7	10	-20		
GW134	440	bdperp/strk par	2	67.6	7	-27		-15
GW134	440	bdperp/strk par	2	120.7	5	-25		-2.4
GW134	440	bdperp/strk par	1	67.4	5			
GW134	440	bdperp/strk par	1	69.8	4			
GW134	440	bdperp/strk par	1	66.7	6			
GW134	440	bdperp/strk par	1	74.5	7			
GW134	440	bdperp/strk par	2	126.8	10	-28		-12.5
GW134	440	bdperp/strk par	2	69.0	6			
GW134	440	bdperp/strk par	2	104.6	8			
GW134	440	bdperp/strk par	1	52.0	5			
GW134	440	bdperp/strk par	1	53.7	7			
GW134	440	bdperp/strk par	1	52.7	3			
GW134	440	bdperp/strk par	1	54.5	5			
GW134	440	bdperp/strk par	1	54.0	2			
GW134	440	bdperp/strk par	1	69.1	10			
GW134	440	bdperp/strk par	1	70.5	5			
GW134	440	bdperp/strk par	1	69.5	6			
GW139	228	bdperp/strk par	1	72.8	30	-45	-16	-6.5
GW139	228	bdperp/strk par	1	73.5	8	-56	-24.5	
GW139	228	bdperp/strk par	1	63.3	10			
GW139	228	bdperp/strk par	1	49.7	15			
GW139	228	bdperp/strk par	2	76.2	18	-50.2	-23.2	
GW139	228	bdperp/strk par	3	74.6	19	-55	-22.4	
GW139	228	bdperp/strk par	3	74.6	13	-52.5	-20.4	

Table D.1 Fluid inclusion data from calcite veins.

GW139	228	bdperp/strk par	1	75.4	40	-50.4	-23.5	-5.1
GW139	228	bdperp/strk par	1	77.2	22	-57.6	-24.1	-8.6
GW139	228	bdperp/strk par	1	70.5	6			
GW139	228	bdperp/strk par	1	68.6	5			
GW139	228	bdperp/strk par	1	70.1	5			
GW139	228	bdperp/strk par	2	65.0	65	-52.7	-23.3	-10.9
GW139	228	bdperp/strk par	2	60.0	8			
GW139	228	bdperp/strk par	2	62.3	7			
GW139	228	bdperp/strk par	1	61.0	8			
GW139	228	bdperp/strk par	1	71.3	12			
GW139	228	bdperp/strk par	1	70.6	15	-51.2	-24.3	-12
GW139	228	bdperp/strk par	1	72.8	8			
GW139	228	bdperp/strk par	1	69.4	8			
GW139	228	bdperp/strk par	2	101.8	90	-52.6	-24.6	-3.1
GW139	228	bdperp/strk par	3	71.3	30	-53.9	-25.2	-7.5
GW139	228	bdperp/strk par	4	68.5	15	-52.1	-24.8	-7.1
GW139	228	bdperp/strk par	4	63.0	45	-53	-24.8	-6.1
GW139	228	bdperp/strk par	4	69.6	20			
GW139	230	bdperp/strk par	1	66.0	25	-53.7	-24.4	-6.8
GW139	230	bdperp/strk par	2	53.3	20	-53.7	-23.4	-11.2
GW139	230	bdperp/strk par	1	74.0	10	-51.3	-20.5	
GW139	230	bdperp/strk par		64.4				
GW139	230	bdperp/strk par		10.4				

VITA

James Lincoln Foreman was born on April 2, 1960, in Coronado, California, just across the bay from San Diego, California. He lived in San Diego with his parents and four sisters where he attended grade school until January, 1974. He moved to Annapolis, Maryland with his mother and sisters in January, 1974 and attended Old Mill Senior High School until he graduated in May, 1978.

The following August, Lincoln entered the University of Maryland, College Park and graduated with a B.S. in Geology in May, 1983. In August, 1983, he enrolled in the Graduate Program at Bowling Green State University, Bowling Green, Ohio. In August, 1985, Lincoln left BGSU before finishing his M.S. thesis and worked for two months as a site geologist for the Department of Anthropology at the University of Pittsburgh. Realizing that this was not where his interests lay, he enrolled in the Doctoral program at the University of Tennessee, Knoxville, in January, 1986 and was awarded a Graduate Teaching Assistantship. He defended his Master's thesis at BGSU in April, 1986, and was awarded the M.S. degree in Geology from BGSU in May, 1986.

In April, 1988, Lincoln received a Doctoral Research Fellowship from the Environmental Sciences Division of the Oak Ridge National Laboratory under the supervision of Dr. Gary Jacobs. Lincoln directed much of his doctoral research towards understanding mechanisms of fracture formation and fracture occlusion in the Nolichucky Shale on the Oak Ridge Reservation. In May, 1991, he received the Doctor of Philosophy degree with a major in Geology from the University of Tennessee, Knoxville.

Lincoln is a member of the American Association of Groundwater Scientists and Engineers, American Association of Petroleum Geologists, Geological Society of America, National Water Well Association, Sigma Xi, and the Society of Economic Paleontologists and Mineralogists. Lincoln currently holds the position of Post-Doctoral Research Fellow at the Oak Ridge National Laboratory where he is involved in the hydrogeochemical characterization of Oak Ridge Reservation groundwaters.