# Katian GSSP and Carbonates of the Simpson and Arbuckle Groups in Oklahoma

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**TABLE OF CONTENTS & GPS COORDINATES** pg. #

## INTRODUCTION

The Arbuckle Mountains region of Oklahoma is characterized by a large inlier of faulted and folded rocks of Precambrian and Paleozoic age. Precambrian and Cambrian basement rocks and early Paleozoic carbonates are overlain by westward dipping Pennsylvanian and Permian strata and by Cretaceous sediments of the Western Interior Seaway. The Arbuckle Mountain region of Oklahoma contains one of the best and most continuous exposures of late Cambrian to Devonian aged strata in all of the midcontinent (nearly 3350 m or 11,000 feet; Ham 1969), most of which is highly fossiliferous. This incredible sequence of rocks has generated substantial interest within the geologic community, with several books (e.g., Sprinkle 1982; Johnson 1991), field guides (e.g., Ham 1969; Fay et al. 1982a; Johnson et al. 1984; Fay 1989; Ragland and Donovan 1991; Cardott and Chaplin 1993; Suneson 1996), and journal papers (e.g., Goldman et al. 2007; Carlucci et al. 2014) devoted to the geology of the area.

The collision of Gondwana (Yucatan terrane) with Laurentia and the development of the Ouachita Mountains during the Pennsylvanian and Permian uplifted the carbonate strata that are the focus of this trip. Exposure of these subsurface rocks has had not only scientific impact, but economic repercussions as well. The Arbuckle (latest Cambrian; Stage 10 to Floian) and Viola groups (Katian) are mined for cement-producing materials, dolomite, and commercially viable crushed stone. Quartz arenites in the Simpson Group are mined for silica and even the Precambrian basement rock (Tishamingo Granite) is quarried for building materials. The oil and natural gas resources stored in subsurface extensions of the Simpson Group in Arbuckle region strata have long engendered substantial interest. Highly porous sandstones of the Bromide, Tulip Creek, and McLish formations and fractured carbonates of the Viola and Arbuckle Groups are well known petroleum reservoirs.

This guidebook was written for the 2015 International Symposium on the Ordovician System (ISOS) as a synopsis of the recent work (e.g., Goldman et al. 2007; Carlucci et al. 2014, forthcoming work for the ISOS meeting) on Ordovician-Silurian rocks of south-central and south-eastern Oklahoma. This new research and past studies (e.g., Harris 1957; Longman 1976; Longman 1982a, b; Fay et al. 1982a; Fay et al. 1982b) underscore the scientific importance of this region. The global stratotype section and point for the Katian Stage of the Upper Ordovician Series is examined on this trip. The first appearances of important graptolites, conodonts and chitinozoans in that section are crucial for worldwide chronostratigraphic correlation. Vertical and lateral facies changes of the Simpson Group demonstrate the variety and intricacy of sedimentary cycles and the importance of updating depositional models with sequence stratigraphic data. Carbonate facies of the Arbuckle Group are of general interest to all geologists, as they demonstrate a wide variety of sedimentary structures and fabrics that were deposited in tropical epeiric seas. Arbuckle Group carbonates show a variety of peloidal, oolitic, fossiliferous, stromatolitic, and brecciated facies that provide important insights into the depositional history of the "Great American Carbonate Bank" (Taylor et al. 2012). Simply put, these deposits are an exceptional natural laboratory for the sedimentary geologist. Siliciclastic deposits are also common in the Simpson and Arbuckle Groups, with shoreface sands and

siltstones forming "bookends" to formation boundaries.  The scientific importance of the Arbuckle region also extends into the realm of structural geology, where geologic cross sections (Fig. 1) of the Ardmore Basin, Arbuckle Anticline, and Washita Valley demonstrate overturned strata, extensive reverse faulting, and a series of major synclines and anticlines at a variety of scales.  Pennsylvanian age tectonic features are just another example of why the Arbuckle Mountains is an excellent natural laboratory for field geologists.  We hope to convey some of that importance to the attendees of this 2015 ISOS pre-meeting field trip.

## Tectonic Setting of the Arbuckle Mountain Region

The Arbuckle Mountains contain a core of Precambrian and Cambrian basement rocks that are uniquely exposed in the southwest portion of the region.  These basement rocks are well-studied and represent some of the same igneous provinces exposed in the Wichita Mountains of Oklahoma.  Precambrian basement extends through the subsurface between the Wichita and Arbuckle Mountains (Ham 1969) and underlies much of the deformed area associated with the uplifts.  Precambrian basement rocks (approximately 1.3 bya) in the Arbuckles are represented by the Tishamingo and Troy granites, which vary from coarse to fine-grained, and are rich in microcline and biotite (Taylor 1915).  These are overlain by the Cambrian Colbert rhyolite group, which consists of extrusive rhyolite flows and tuffs, together with beds of agglomerate and sills of diabase (Finnegan and Hanson 2014; Hanson et al. 2014).  Details of the basement rocks of Oklahoma, including their petrology, distribution, and origin were most recently discussed by Pucket et al. (2014).

The uplift of the Cambrian and Ordovician strata in the Arbuckles is associated with the Ouachita Orogeny (Viele and Thomas 1989), a mountain building event in the Pennsylvanian and Permian caused by collision of  the Yucatan terrane, which is part of present day Mexico, but was at that time attached to Gondwana, with the Laurentian craton.  The Ouchitan tectonic system is extensive, and ranges from Alabama (Black Warrior Basin) through Arkansas and Oklahoma (Wichitas and Arbuckles), and then southwest into Texas (Llano, Marathon, Solitario Uplifts).  As a result of Ouchita tectonics, the Arbuckle mountain region exposes a series of fold and thrust belt structural features, such as the Arbuckle Anticline (Figs. 1, 2), Mills Creek Syncline, Ardmore Basin, and Washita Valley.  A detailed analysis of all these structural features is beyond the scope of this field trip, but there are some important details to note.

The most intensely deformed part of the region is the Arbuckle Anticline (Figs. 1, 2), a faulted anticline that is overturned to the north.  The faulted portion of the Arbuckle Anticline contains a graben that is filled with Pennsylvanian synorogenic molasse sediments (Collings Ranch Conglomerate).  The Collings Ranch Conglomerate is structurally deformed into a synclinal fold, indicating that deformation continued after deposition.  The core of the anticline consists of the Cambrian Colbert Rhyolite Group, with Upper Cambrian and Ordovician carbonates flanking the rhyolites.  The fault axis is located near the East Timbered Hills region, offsetting volcanics on either side of the fold.  Just south of the Arbuckle Anticline lies the Ardmore Basin, a downwarped remnant of the Southern Oklahoma Aulacogen (Brewer et al.
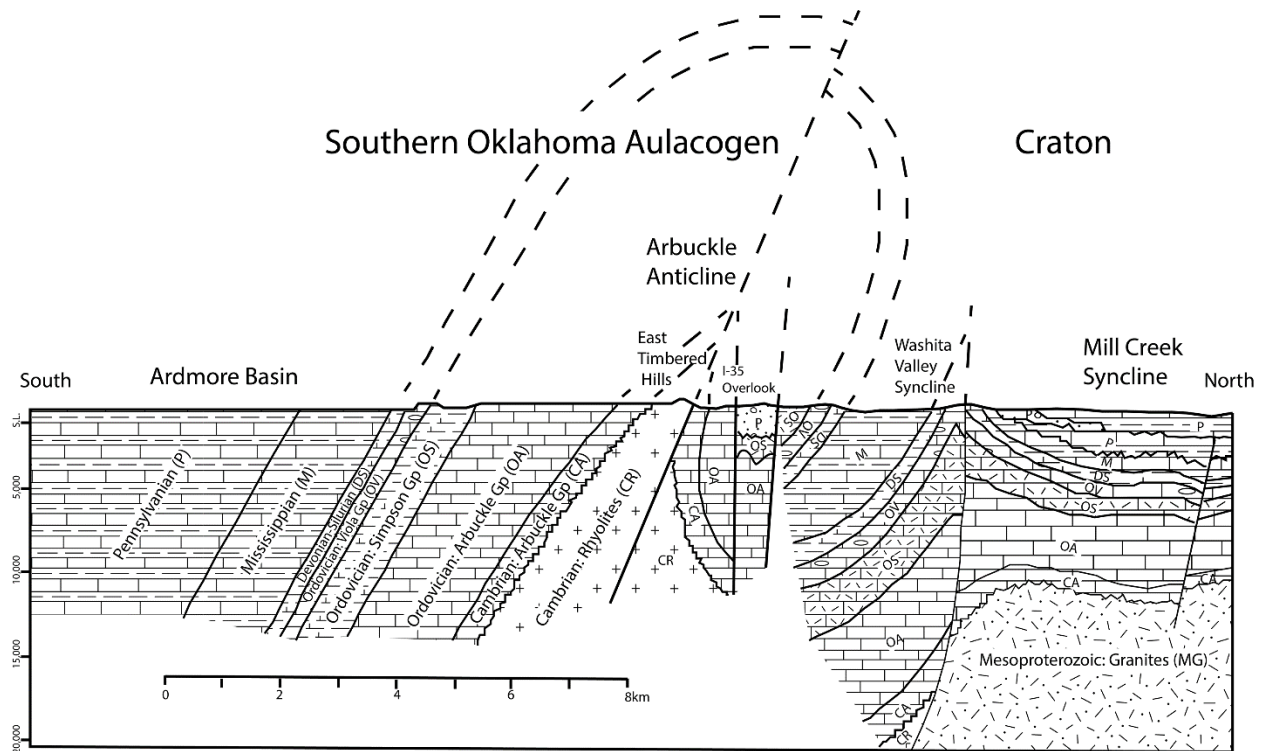
**TEXT-FIGURE 1.—Structural cross-section of the Arbuckle Mountains region along I-35, between Davis and Ardmore (modified from Ham 1969). Line of cross section shown in Figure 2.**

1989), which includes over 10,000 meters of Cambrian – Pennsylvanian-age strata. The overall structure is of a large, faulted syncline, punctuated by smaller anticlines. Mississippian and Pennsylvanian rocks in the Ardmore Basin dip between 45-90° to the northeast, and include the Hoxbar, Deese, Noble Ranch, and Dornick Hills Groups (Suneson 1996). Much of these deposits consist of conglomerates and shales that record the extensive input from the adjacent Ouchita event. As part of the larger Southern Oklahoma Aulacogen system, the Ardmore Basin is central to our discussion of the Cambrian-Ordovician carbonates in central Oklahoma.

**Formation of the Southern Oklahoma Aulacogen**

Approximately 550 million years ago, the granitic basement rocks (Tishamingo Granite) of central Oklahoma begun to undergo extensional stress associated with the rifting of Iapetus and development of a failed continental rift (Southern Oklahoma Aulacogen, [SOA], or Anadarko Basin). The faults bounding the SOA were northwest trending, and the structure extended in a zone across south-central Oklahoma, and into the panhandle of Texas. The Washita Valley fault zone, for example, was a rift-forming normal fault that separated the subsiding aulacogen in the south from the craton to the north (Fig. 2). During the Middle Cambrian, continued extension led to the infilling of the SOA with the volcanics of the Colbert Rhyolite Group. Suneson (1996) noted that this igneous activity was concentrated in southern
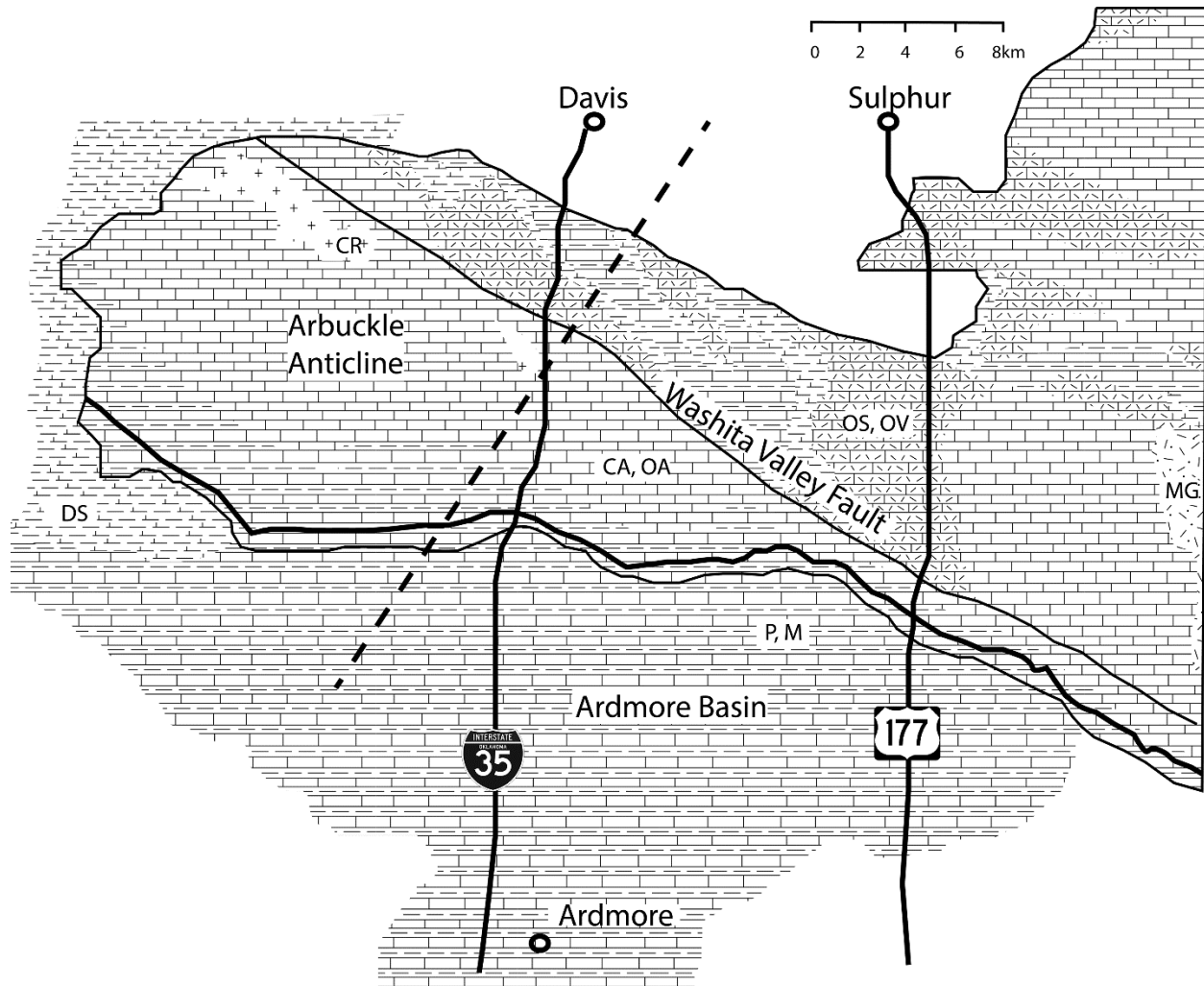
**TEXT-FIGURE 2.—Generalized geologic map of the Arbuckle Mountains region between Davis and Ardmore (modified from Ham 1969). Abbreviations correspond to units shown in Figure 1, dotted line is the line of cross section.**

Oklahoma because the basement rock had been weakened by faulting associated with the rifting. Cooling and contraction of the Cambrian rhyolites potentially led to additional subsidence that allowed the Southern Oklahoma Aulacogen to be a major depocenter in the Ordovician (Suneson 1996). Between the Late Cambrian and Early Devonian, subsidence of the aulacogen allowed for extensive accumulation of marine limestones, sandstones, and shales (e.g., Arbuckle and Simpson Groups). Cambrian and Ordovician strata of Oklahoma were deposited in a broad epeiric sea (Oklahoma Basin) that extended across most of the state (Johnson 1991; Carlucci et al. 2014). The Oklahoma Basin intersected the margins of the SOA, where deeper water sedimentation was dominated by carbonate interbedded with sandstone and shale. Most authors (e.g., Longman 1982a, b; Johnson 1991) have considered the Southern Oklahoma Aulacogen (SOA) to be the depocenter of the Oklahoma Basin, as subsidence rates and sediment thicknesses are considerably higher than in shallow-ramp to platform environments outside the SOA. To the

north, the Oklahoma Basin was bordered by the stable Arbuckle platform (Longman 1982b), which was a desert region that likely supplied wind-blown sand deposited as sheets into the SOA. In the early to middle Cambrian, the craton was deeply eroded. Input of siliciclastics into the SOA was temporarily suspended during a major transgression in the late Cambrian which established the broad epeiric sea across vast areas of Oklahoma, and facilitated deposition of the Arbuckle Group (St. John and Eby 1978; Johnson et al. 1984). In the lower to middle Whiterockian, Simpson Group deposition began when the carbonate shelf that bordered the SOA was exposed (McPherson et al. 1988). Wind-blown sand was reworked and mantled the carbonate platform (Johnson et al. 1988), and was eventually overlain by marine shale and carbonate.

After the initial infilling of the aulacogen in the Ordovician, subsidence developed again in the Late Devonian to Mississippian, accumulating thick deposits of marine shale (e.g., Woodford, Delaware Creek, and Goddard shales). In the Pennsylvanian and Permian, uplift of the entire region led to the development of many previously mentioned tectonic features (including the Arbuckle Mountains themselves), and an angular unconformity between Ordovician-Mississippian and Pennsylvanian strata.

**Paleozoic Stratigraphy of the Arbuckle Mountains**

The Paleozoic stratigraphic succession of the Arbuckle Mountains (Fig. 3) comprises a thickness of more than 3,000 meters (10,000') of sediments ranging in age from Early Cambrian to Pennsylvanian, recording some 200 million years of geologic time. It is arguably one of the most complete and thickest Cambrian-Ordovician successions in central North America. The following summary provides an overview of stratigraphy and facies of this classic succession and it incorporates the litho- and biostratigraphic research of many previous workers, most notably Taff (1902), Edson (1927), Decker (1931, 1933, 1935, 1941), Decker and Merritt (1931), Hendricks et al. (1937), Wengerd (1948), Cardott and Chaplin (1963), Amsden (1967), Ham (1969), Ham and Amsden (1973), Amsden and Sweet (1983), Sprinkle (1982), Fay et al. (1982a, b), Longman (1982a, b), Finney (1986, 1988), Fay (1989), Ethington et al. (1989), Derby et al. (1991), Wilson et al. (1991), Johnson (1992, 1997), Amati and Westrop (2004, 2006), Goldman et al. (2007), Leslie et al. (2008), Bergström et al. (2010), Rosenau et al. (2012); Carlucci et al. (2012, 2014). Average stratigraphic thicknesses given by Fay (1989) are used in the following descriptions.

At the base of the succession resting unconformably upon Proterozoic basement is the Colbert Rhyolite, now dated radiometrically as early Cambrian age (525 Ma). This basal igneous rock is nonconformably overlain by the Upper Cambrian Timbered Hills Group; the latter is comprised of about 73 meters (240') of arkosic, glauconite-bearing Reagan Sandstone and 32 m (105') of Honey Creek Limestone. Above is a thick succession (2,073 m; 6800') of predominantly massive, shallow-water, dolomitic and frequently cherty carbonates with a variety of sedimentary structures indicative of peritidal to shallow subtidal deposition. The Arbuckle Group strata are roughly dated on the basis of trilobites and other megafossils as being of Late

## Arbuckle Mountains Along I-35N

| Global Series | Global Stages | N. American Series | N. American Stages | Group | Formation | Field Trip Stop # |
|---|---|---|---|---|---|---|
| Silurian | Rhuddanian | Cincinnatian | Richmondian-Gamachian | Lower Hunton Gp | Cochran Limestone | 6 |
| Upper Ordovician | Hirnantian | Cincinnatian | Richmondian-Gamachian | Lower Hunton Gp | Keel Limestone | 6 |
| Upper Ordovician | Hirnantian | Cincinnatian | Richmondian-Gamachian | Lower Hunton Gp | Sylvan Shale | 6 |
| Upper Ordovician | Katian | Mohawkian | Chatfieldian | Viola Gp | Welling Fm | 8 |
| Upper Ordovician | Katian | Mohawkian | Chatfieldian | Viola Gp | Viola Springs Fm | 4, 7, 11, 12 |
| Upper Ordovician | Sandbian | Mohawkian | Turinian | Viola Gp | Bromide Fm | 4, 7, 11, 12 |
| Middle Ordovician | Dapingian-Dariwillian | Whiterockian | Rangerian-Chazyan | Simpson Group | Tulip Creek Fm | 11 |
| Middle Ordovician | Dapingian-Dariwillian | Whiterockian | Rangerian-Chazyan | Simpson Group | McLish Fm | |
| Middle Ordovician | Dapingian-Dariwillian | Whiterockian | Rangerian-Chazyan | Simpson Group | Oil Creek Fm | 9 |
| Middle Ordovician | Dapingian-Dariwillian | Whiterockian | Rangerian-Chazyan | Simpson Group | Joins Fm | |
| Lower Ordovician | Tremadocian-Floian | Ibexian | Starsian-Tulean-Black Hillsian | Upper Arbuckle Group | West Spring Creek Fm | 2 |
| Lower Ordovician | Tremadocian-Floian | Ibexian | Starsian-Tulean-Black Hillsian | Upper Arbuckle Group | Kindblade Fm | 2 |
| Lower Ordovician | Tremadocian-Floian | Ibexian | Starsian-Tulean-Black Hillsian | Upper Arbuckle Group | Cool Creek Fm | 1 |
| Cambrian | Stage 10 | Sunwaptan – Skullrockian | | Lower Arbuckle Group | Signal Mountain Fm | |
| Cambrian | Furongian | Sunwaptan | | Lower Arbuckle Group | Royer Dolomite | |
| Cambrian | Furongian | Sunwaptan | | Lower Arbuckle Group | Fort Sill Limestone | 10 |

## Southeast Oklahoma, Atoka County Black Knob Ridge Section

| Global Series | Global Stages | N. American Series | N. American Stages | Formation | Field Trip Stop # |
|---|---|---|---|---|---|
| Upper Ordovician | Katian | Mohawkian | Chatfieldian | Polk Creek Shale | 3 |
| Upper Ordovician | Katian | Mohawkian | Chatfieldian | Bigfork Chert *GSSP* | 3 |
| Upper Ordovician | Sandbian | Mohawkian | Turinian | Womble Shale | 3 |

**TEXT-FIGURE 3.—Stratigraphic nomenclature of the Paleozoic as exposed in the Arbuckle Mountains and Black Knob Ridge regions. Data compiled from Derby et al. (1991), Stitt (1991), Johnson and Cardott (1992), Bauer (1994), Johnson et al. (1994), Babcock et al. (2005), Goldman et al. (2007), Bergström et al. (2008), Rosenau et al.(2012).**

Cambrian (Furongian) and Early Ordovician (Tremadocian-Floian) age.  The Cambrian portion of the Arbuckle Group shows a three-fold division with lower and upper dark bluish gray, massive carbonates of the Fort Sill Limestone (47 m; 155') and Signal Mountain Formation (126.5 m; 415').  They are separated by a thick, pinkish to ocherous yellow dolostone of the Royer Formation (219 m; 717').  The Signal Mountain Formation may span the Cambrian-Ordovician boundary.

The Lower Ordovician portion of the Arbuckle Group itself averages nearly a mile in thickness (1653 m; 5422') and includes in ascending order, the Butterly Dolostone (90 m; 297'), McKenzie Hill Formation (274m; 900'), Cool Creek Formation (396 m; 1300'), Kindblade Formation (430 m; 1410'), and West Spring Creek Formation (462 m; 1515'). These carbonates are exposed on both the north and south flanks of the Arbuckle Mountains, although they are disturbed by faulting in some areas. The Cool Creek Formation (stop 1) displays excellently preserved peritidal indicators including stromatolites, oncoids, and flat-pebble conglomerates. Large stromatolites are also typical of the upper beds of the West Spring Creek Formation slightly below its contact with the Simpson Group. Overall, the Arbuckle Group is the expression of a long ranging gradually subsiding passive margin of Laurentia, the "Great American carbonate bank".

### Simpson Group

The fully exposed Middle-Upper Ordovician strata unconformably overlying the upper part of the Arbuckle Group are assigned to the Simpson Group, Viola Group and Sylvan Shale. These strata are the primary focus of this trip and hence are discussed in somewhat greater detail.

The widespread Simpson Group, named for exposures near the village of Simpson, presently called Pontotoc (Taff 1902), is a highly fossiliferous, mixed carbonate and siliciclastic succession about 732m (2400') thick that ranges in age from Dapingian to Sandbian (Fig. 3). The interval is divided into five formations, each of which, except for the basal Joins, has been defined as starting with a lower submature quartz-rich sandstone, overlain by shales and then limestones (Decker and Merritt 1931). Altogether, this suggests a lower lowstand to early transgressive sand to shale succession with a maximum flooding within the lower shales, and an abruptly upward shallowing and "cleaning" upward succession.

The basal Joins Formation (up to 90 m thick) commences with a thin basal conglomerate that records a transgressive lag of carbonate clasts derived from erosion of the underlying West Spring Creek Formation. This unit marks the overspreading of the Sauk-Tippecanoe megasequence boundary (or Knox unconformity, which is locally of relatively small magnitude). The conglomeratic beds are overlain by thin, micritic limestones and shales with a low diversity fossil fauna, but yielding diagnostic conodonts that are assignable to the *Histiodella altifrons* to lower *H. sinuosa* conodont zones (Bauer 2010). Decker and Merritt (1931) note that these beds also contain common specimens of the graptolite *Didymograptus artus* indicating a Chazyan (late Dapingian- early Darriwillian) age, which is consistent with the conodont biostratigraphy.

The Oil Creek Formation, named for Oil Creek 14 miles SW of Sulfur, Oklahoma, is the thickest unit of the Simpson Group ranging from more than 91 m (300') to over 328 m (1075') near Spring Creek at the Daube Ranch. It comprises a basal sand which thickens eastward from a feather edge in western localities to over 175 m in the eastern Arbuckles. It locally oversteps the truncated Joins Formation to the north and rests directly on the West Spring Creek. This basal sandy interval is overlain by a thick succession of coarse bioclastic limestones (echinoderm pack- and grainstones) and shales. Beds of intraformational conglomerate are numerous as are

hardgrounds, many of which show encrusting bryozoans and pelmatozoan holdfasts.  A moderately diverse fauna (~35 species) includes ramose trepostome bryozoans, orthid and clitambonitacean brachiopods (*Clitambonites*, *Dinorthis*), gastropods *(Lecanospira*, *Liospira*, *Maclurites*), small bivalves, nautiloids, and leperditians.  Trilobites (especially *Pliomerops*) and rhombiferan cystoid plates are also abundant and well preserved.  Megafaunas indicate an early Chazyan (Darriwillian) age.  Bauer (2010) assigns the Oil Creek Formation to the mid Darriwilian  *Histiodella sinuosa* to *H. holodentata* conodont zones.

The McLish Formation named for McLish Ranch near Bromide, Oklahoma is about 102 to 162 m (335 to 533') thick and sharply, locally unconformably, overlies the Oil Creek Formation. The McLish comprises a basal unit (Burgen Member) of up to 12 m (40') of hard to uncemented quartz sand overlain in turn by thin greenish shales (6 m; 20') and earthy brownish limestones that pass upward into dense micritic, fenestral dove gray limestones with minor dolostones and greenish shales.  Overall, the unit is not highly fossiliferous, but the lower brown limestones yield sponges, gastropods, a limited diversity of brachiopods (*Rafinesquina*, S*trophomena*, *Zygospira*), small bivalves, and a thin zone of cystoid plates (*Palaeocystites*).  Bergström (1971) and Bauer (1987) correlate the McLish formation with the *Cahabagnathus friendsvillensis* conodont Zone indicating a middle Chazyan (Darriwillian) age.

The Tulip Creek Formation (83-120 m; 271-394') is named for Tulip Creek near Springer, Oklahoma, and again comprises a basal sandstone (Wilcox Member), overlain by a poorly exposed interval of soft shales and thin-bedded limestones.  A limited fauna of some 20 species includes abundant plates of crinoids and cystoids and the typically Blackriveran brachiopods *Pionodema* and *Dalmanella*, a few gastropods, bivalves and nautiloids.  Rare conodonts, assignable to the *Cahabagnathus friendsvillensis* Zone (= *Pygodus serra* North Atlantic conodont Zone), suggest a late Chazyan age (Bergström 1971; Bauer 1987).

The uppermost unit of the Simpson Group is the Bromide Formation named for exposures in quarries near the small town of Bromide, which is itself named for naturally carbonated waters derived from aquifers in the Simpson Group.  The Bromide is richly fossiliferous and has been studied in detail from the standpoint of stratigraphy and paleontology (especially its rich echinoderm faunas; see Sprinkle 1982).  Decker and Merritt (1931) listed nearly 100 species including diverse brachiopods (28), bryozoans (13 species), trilobites (12), gastropods (11), cephalopods (8), algae, tetradiid corals and echinoderms. The latter have been intensively studied in the interim and now more 61 genera in 13 classes are recognized making the Bromide one of the most diverse Ordovician faunas (Sprinkle 1982).  Trilobite diversity in the Bromide has also been refined upwards (e.g., Shaw 1974; Carlucci et al. 2012; Carlucci and Westrop 2014) in recent years.  The Bromide Formation is notably cyclic on several scales and the documentation of this cyclicity is a major theme of recent papers (Carlucci et al 2014; Carlucci and Westrop 2014) and of this field excursion.

The Bromide is traditionally divided into two members, the Mountain Lake and Pooleville (Cooper, 1956), both named for exposures along Spring Creek on the Daube Ranch (formerly Johnston Ranch) west of Ardmore.  Recent study of the Bromide (Carlucci et al. 2014)

has led to recognition of a third, basal sandstone member, redefinition of the other members and interpretation of three depositional sequences probably equivalent to the M2, M3, and M4 of Holland and Patzkowsky (1996). Each commences with coarse sandstone or skeletal grainstone that passes upward into a shaly interval and then into a progradational shale to thinly bedded wacke- and packstone, and in some cases peritidal lime mudstone highstand facies.

As with other Simpson Group formations, the Bromide commences with widespread quartz sandstone which thickens toward the northeast, although originally included in the Mountain Lake Member this sandstone unit is sufficiently distinctive that it warrants separate designation and Carlucci et al. (2014) named this the Pontotoc Member. As is the pattern with the other formations, the sandstone passes upward into a greenish gray, chloritic shale with thin fossiliferous limestones. A ~1m shale thin limestone interval overlying a distinctive thick packstone ledge has yielded a prolific echinoderm fauna. This "Lower Echinoderm Zone" (Fay and Graffham 1969) has been identified in 16 localities in the Arbuckles (Sprinkle 1982) and more tentatively in the Criner Hills. More than 6,000 complete echinoderm thecae have been collected. More than half of those collected are *Hybocrinus*, another quarter the paracrinoid *Platycystites*, and in total some 30 other genera, including paracrinoids, rhombiferans, crinoids, edrioasteroids, asteroids and stylophorans.

This shale and packstone succession passes upward abruptly into an interval of thick grainstones and shales of the upper Mountain Lake Member. To the north, near Fittstown this package of echinoderm skeletal grainstones passes laterally into thicker sandstone and sandy limestone package comparable to the Pontotoc Member, but more restricted in aerial distribution. The overlying thick upper Mountain Lake Member is comprised of thinly bedded limestones and shales that yield additional brachiopods, bryozoans, echinoderms and trilobites, including the isoteline *Vogdesia*. These beds closely resemble a main mass of the interval assigned to the Pooleville to the south in the Criner Hills, which we have inferred, actually to be coextensive with the Mountain Lake Member.

A sharply based package of thick echinoderm grainstones occurring about 75 m above the base of the Bromide is identifiable at all localities and we have used it to redefine the base of the Pooleville Member. It is overlain by a greenish shaly interval identified previously in the Arbuckles as the upper Echinoderm Zone, which has yielded several thousand echinoderms, mostly the paracrinoid *Oklahomacystis*. As with the lower Echinoderm zone this interval has been widely correlated through some 15 locations in the Arbuckles. This interval is overlain by a shallowing upward succession of dense, burrow mottled lime mudstones and wackestones, typical of the Pooleville Member. The lateral continuity of both Lower and Upper Echinoderm Zones, which represent highstand facies of small-scale cycles is simply an example of the overall continuity of many units within the Bromide and strongly suggests an allocyclic (eustatic) influence on their formation.

The upper several meters consist of fenestral, microbially laminated micrites with thin shales and desiccation cracks. This interval, the Corbin Ranch submember contains thin clay beds some of have been shown to be K-bentonites and tentatively identified as the Deicke and

Millbrig beds, the most widespread bentonites in the Ordovician.  Thus, this interval is identified as equivalent to very similar and extraordinarily widespread micritic facies of the upper "Blackriveran" in the Mississippi Valley (Plattin Fm) Appalachian Basin (Lowville Fm), Kentucky (Tyrone Fm), Tennessee (Carters Fm), and elsewhere.  It is clearly of Turinian (upper Sandbian) age based upon conodonts, as well as bentonites.   These indicators also indicate correlation of the Corbin Ranch with the upper portion of the Womble Shale in the allochthonous deep water facies of the Ouachitas, as at the Katian stratotype section at Black Knob Ridge in Atoka.

A most important new discovery of our research is that the Pooleville Member appears to be truncated progressively southward into the Oklahoma Aulacogen, such that the overlying Viola Group rests on progressive lower Pooleville beds and ultimately on strata of the upper Mountain Lake Member.  This pattern suggests, either, that there was a reversal of topography following deposition of the Bromide or that other process of deeper ramp corrosion, erosion and bypass thinned and/or removed upper Bromide strata prior to deposition of the Viola Springs. This pattern requires further study.

### *Viola Group and Sylvan Shale*

Throughout much of the Arbuckles, the Viola Group, named for exposures near Viola, Oklahoma (Taff 1903), is a ~213 m (700') thick succession of resistant, ridge-forming, dark gray, but whitish weathering limestones.  The Viola is very widespread with representation as far north as South Dakota and as far west as the subsurface of Colorado (Wengerd, 1948).  In places it rests unconformably on strata as old as the Arbuckle Group at a disconformity that may represent the M4-M5 sequence boundary of Holland and Patzkowsky (1996).  This unit represents much of the Katian Stage, roughly the Chatfieldian and Cincinnatian of North American terminology. The majority of the Viola Group is assigned to the Viola Springs Formation, thin- to medium-bedded, sparsely fossiliferous, commonly cherty, laminated calcisiltites with thin, dark shaly partings. Bedding can be notably hummocky (HCS?) and nodular in some portions of the formation.  Thin shaly beds yield graptolites, cryptolithine trilobites and small brachiopods at many levels.  The upper  ~25-30 m of medium to thick-bedded packstone and grainstone is assigned to the Welling Member; it yields a diverse benthic fauna of bryozoans, brachiopods and trilobites. The base of the Welling may represent a disconformity, perhaps a mid-Richmondian unconformity.

A rather different facies aspect of the Viola Springs Formation is represented in the northern Arbuckles at Fittstown (also termed Highway 199 or Murray Lane in the literature, stop 4 here). Here, the lowest 80 cm is distinctively set off as a non-cherty interval with shaly wackestone, rich in graptolites of the upper *Climacograptus bicornis* Zone.  The next 15 m of the Viola Springs Formation (member 4 of Wengerd 1948) is more typical cherty, thin-bedded, and buff weathering.  Higher beds are less cherty and include pelmatozoan rich pack and grainstones (20 m, member 3 of Wengerd) overlain by light gray weathering, slightly argillaceous, highly fossiliferous wacke- and packstones (members 1 and 2).  These beds yield graptolites at some levels, indicative of the *Diplacanthograptus spiniferus* Zone as well diverse bryozoan, coral,

brachiopod and trilobite faunas.  The latter have recently been studied in detail by Amati and Westrop (2006).  Amati (2014) recognized two distinctive faunas, separated by a relatively abrupt transition at about 32 meters above the base of the Viola Springs Formation at Fittstown. Wengerd (1948) was able to trace members throughout the Arbuckle Mountains, despite changes in thickness and facies.

The Viola Springs strata yield a succession of graptolites that have permitted a detailed biostratigraphy (e.g., Finney 1986, 1988).  Although there is some ambiguity, the Viola Springs at the Fittstown section is dated from the *C. bicornis* Zone (uppermost Sandbian) to the Katian *D. spiniferus* Zone (see further discussion of Stop 4).  In sections along I-35 the succession ranges upward to the *Amplexograptus manitoulinensis* of the upper Cincinnatian (Richmondian). Recent studies of carbonate carbon isotopes have also revealed a series of positive isotopic excursions that have tentatively been identified as the GICE and Kope Excursion (Bergström et al. 2010). The upper Cincinnatian (Richmondian) Waynesville excursion was tentatively identified in the Welling Formation.

Based upon graptolite biostratigraphy, the Viola Springs is approximately equivalent to the Bigfork Chert of the allochthonous succession in the Ouachitas as at Black Knob Ridge (see Stop 4). The latter shows facies similarities with the distal Viola Springs, including a succession of thinly bedded chertified limestones and nodular cherts and minor shales. The sparse benthic fauna is typified by cryptolithine trilobites at some levels.

The Viola Group is conformably overlain by some 180 m (600') of dark fissile mudrock (Wengerd 1948), assigned to the Sylvan Shale (stop 6). The latter has been dated as late Katian on the basis of graptolites of the *Styracograptus tubuliferus* to *Dicellograptus ornatus* zones (Dworian 1990).  It is approximately equivalent to the widespread Shale in the Ouachitas and the Mannie Shale of the Appalachian Basin.  It is overlain sharply and probably unconformably by thin widespread oolitic limestone of the Keel Formation in Oklahoma and laterally equivalent Noix Oolite in the Misssissippi Valley both of Hirnantian age and probably representing initial transgression following the late Ordovician glacioeustatic lowstand.

### Post-Ordovician Strata

Silurian and Lower Devonian shaly carbonates are assigned to the Hunton Group, which totals 40 to 70 m (130-230'). The Silurian succession in the Arbuckles includes the Llandovery age Cochrane and early Wenlock Clarita carbonates, each about 4 m thick, separated by a thin but widespread shale tongue, and the overlying Henryhouse Formation 58 m (191') of mixed siliciclastics and limestones. Locally, these rocks are highly fossiliferous with well-documented brachiopod, trilobite and echinoderm faunas (Amsden 1975).

Devonian strata are thin and highly incomplete but include the Lower Devonian (Lochkovian) Harragan Formation,~8 m (25') thick in the southern Arbuckles and the overlying Bois d'Arc, 2.5m (8') thick, both marly, slightly cherty limestones and shales noted for diverse brachiopod and trilobite assemblages.  These beds are unconformably overlain by the Upper Devonian (Famennian) Woodford Shale, black shales and cherts, up to 85 m (280') thick.  The

Lower to Middle Mississipppian consists of a relatively thin carbonate succession (Sycamore Limestone, 67 to 113m m, 380').  It is overlain by thick Middle and Upper Mississippian siliciclastics (the Delaware Creek Shale and the Chesterian Goddard Formation).  Both are comprised of shales and sandstones, together totaling more than half a mile in thickness (Fay 1989).  This shift toward much thicker siliciclastic sediments records the onset of early phases of reactivation in the Anadarko Basin (Oklahoma Aulacogen) and uplift of the Wichita Mountains to the south.  Finally, the Pennsylvanian Collings Ranch Formation consists of more than 900 m (3000') of reddish, coarse polymictic conglomerate, representing Ouachita syntectonic molasse sediments.

## FIELD TRIP STOPS

Figure 3 shows the Upper Cambrian to lower Silurian stratigraphy with important global and North American stage and series boundaries.  The majority of the field trip stops (3, 4, 7, 8, 11, and 12) focus on deposits that formed during the Sandbian-Katian (Upper Ordovician) stages, including the Global Standard Stratotype and Point (GSSP) and auxillary section for the base of the Katian (Goldman et al. 2007), and a newly established reference section for the Bromide (Carlucci et al. 2014).  Other stops expose strata above the Katian (Hirnantian), and below the Sandbian (Dapingian-Dariwillian, Tremadocian-Floian), and even into Stage 10 of the Upper Cambrian.  Field trip stops in relation to important city and county boundaries, and major highways are shown in Fig. 4.

## DAY 1
## Stop 1: Turner Falls Overlook and the Cool Creek Formation

Turner Falls, a well-known attraction for visitors to central Oklahoma, lies just west of US Highway 77 in the East Timbered Hills region (see Fig. 1). Stop 1 is located at an overlook looking back towards the recreational area of Turner Falls.  The falls are interesting from a geological perspective because they are building outward along an accreting travertine precipice, rather than eroding inwards like most waterfalls (Ham 1969).  In the Pleistocene and continuing into the present day, Honey Creek has been down cutting into Arbuckle Group limestone and storing calcium carbonate in solution that is re-precipitated onto the precipice. The falls are currently in balance between travertine development and mechanical erosion from down cutting, although, during the middle Pleistocene, it appears that Honey Creek cut a deep wedge into its depositional platform (Ham 1969).

In the East Timbered Hills surrounding Turner Falls, exposures of the Colbert Rhyolite Group (see line of cross section on Figure 2) form a scenic vista.  The Turner Falls overlook provides a perfect opportunity to see exposures of the some of the early core deposits of the Arbuckles, overlain by younger Cambrian limestone.  The overlook is located just south of the
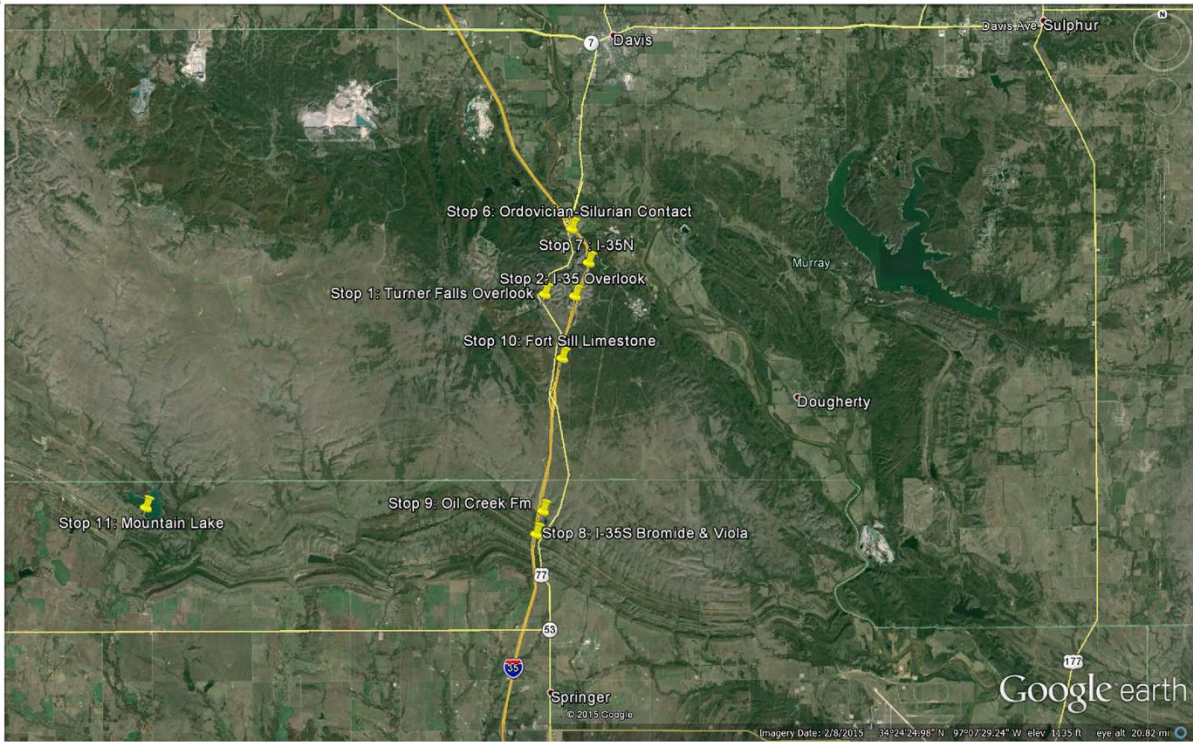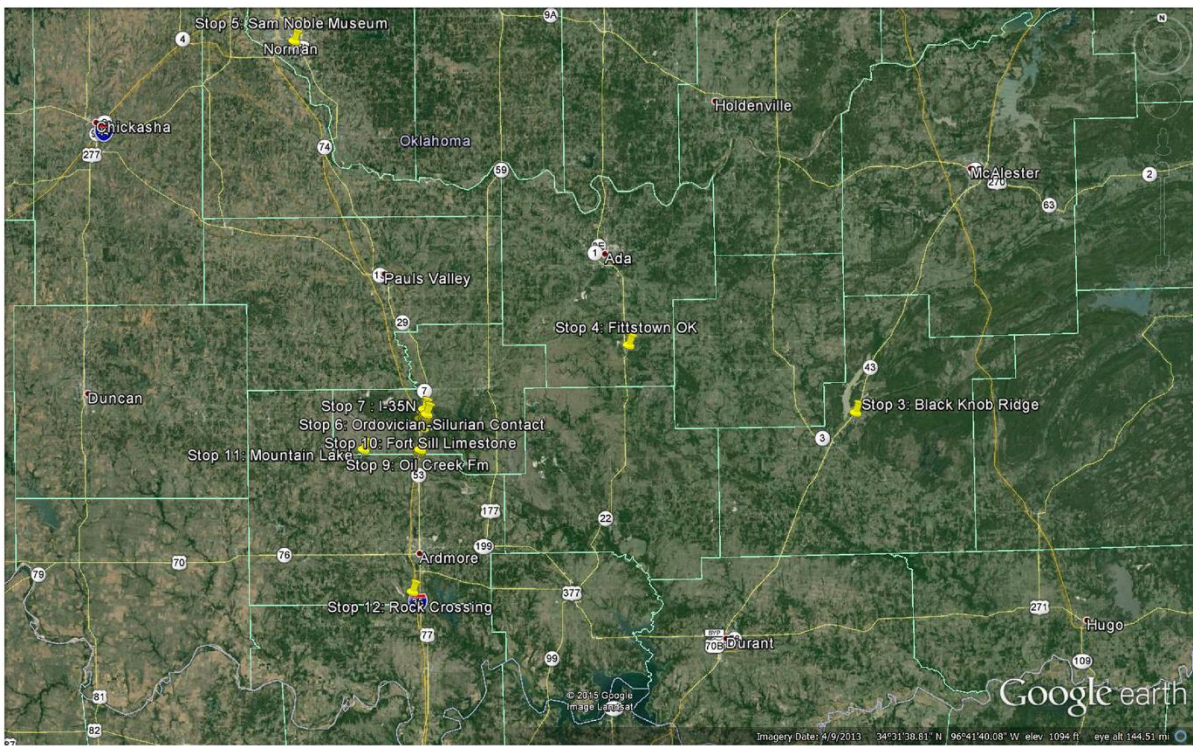
**TEXT-FIGURE 4.—Field trip stops (1-12) with county, town, and state boundaries. a, close up view of all localities in the Arbuckle Mountains. b, broad view, including localities outside the Arbuckles.**

Washita Valley Fault (Fig. 2), immediately to the southwest is a hill that forms the rhyolitic core of the Arbuckle Anticline. Between the core of the anticline and Turner Falls are folded and faulted strata of the Arbuckle Group. The overlook exposes the Cool Creek Formation of the Arbuckle Group (Fig. 3), which is separated by the Mackenzie Hill Formation (exposed across the valley) by a fault trace that parallels the Honey Creek valley (Cardott and Chaplin 1993). Many of these fault traces are present in the East Timbered Hills, and they represent splays from the Washita Valley or Chapman Ranch fault zones. As noted earlier, the Washita Valley fault zone is a normal fault that likely developed during early rifting and formation of the Southern Oklahoma Aulacogen in the late Precambrian and early Cambrian.

The Lower Ordovician (Tremadocian) Cool Creek Formation is one of the lower units of the Ordovician portion of the Arbuckle Group (Fig. 3). Arbuckle Group deposition varies from subtidal to supratidal, and took place on a low gradient carbonate ramp on the southern edge of the North American craton (Wilson et al. 1991). The Arbuckle Group consists of eight formations (some omitted from Fig. 3) that total nearly 2400 m (8,000 feet) in the SOA, with extensive dolomitization across most of the units. Wilson et al (1991) and Cardott and Chaplin (1993) characterized the shallowing upward cycles (Fig. 5) that are prevalent in much of the Arbuckle Group. They identified common components in the sub- inter- and supratidal environments of individual cycles (idealized succession shown in Figure 6). The top of each cycle is typically disconformable, and overlain by thin transgressive marine beds that represent backstepping prior to successive parasequence development.

Across from the gift shop at stop 1, is a roadcut through the Cool Creek Formation that shows a series of facies and sedimentary structures associated with peritidal cycles. Facies at stop 1 include: stromatolitic and thrombolitic boundstone, intraformational mud-supported breccias, oolitic grainstone, bioturbated mud- wacke- packstone, heterolithically bedded units, and chert-rich mud- and boundstones. Figure 5 shows a series of facies through a portion of a shallowing upward cycle. At the base, is an intraformational breccia of lime mudstone and algal boundstone clasts. This is directly overlain by a thin algal boundstone and then a heterolithically bedded unit of micrite and quartz silt. At the top of the cycle is a micrite with chert nodules and bands, which likely represent evaporative conditions (see explanation below). The boundstone above the chert facies likely represents the start of a new cycle, and a switch to lower intertidal or subtidal conditions.

Bedded and nodular evaporates (anhydrite) have been identified in the Cool Creek Formation in the subsurface (St. John and Eby 1978), although in outcrop these same facies have been extensively replaced by chert. Halley and Eby (1973) and St. John and Eby (1978) both suggested that hypersaline conditions were common during Cool Creek Formation deposition, based on a number of indicators, including: syndepositionally broken ooids, length-slow chalcedony, and high-relief stromatolites that occupy lower intertidal or subtidal conditions devoid of grazers. There is also direct evidence of "vanished evaporites" in the Cool Creek Formation. St. John and Eby (1978) discovered evidence of pseudomorphs and microscopic molds in all chert nodules they thin sectioned. SEM studies also revealed evidence of very small
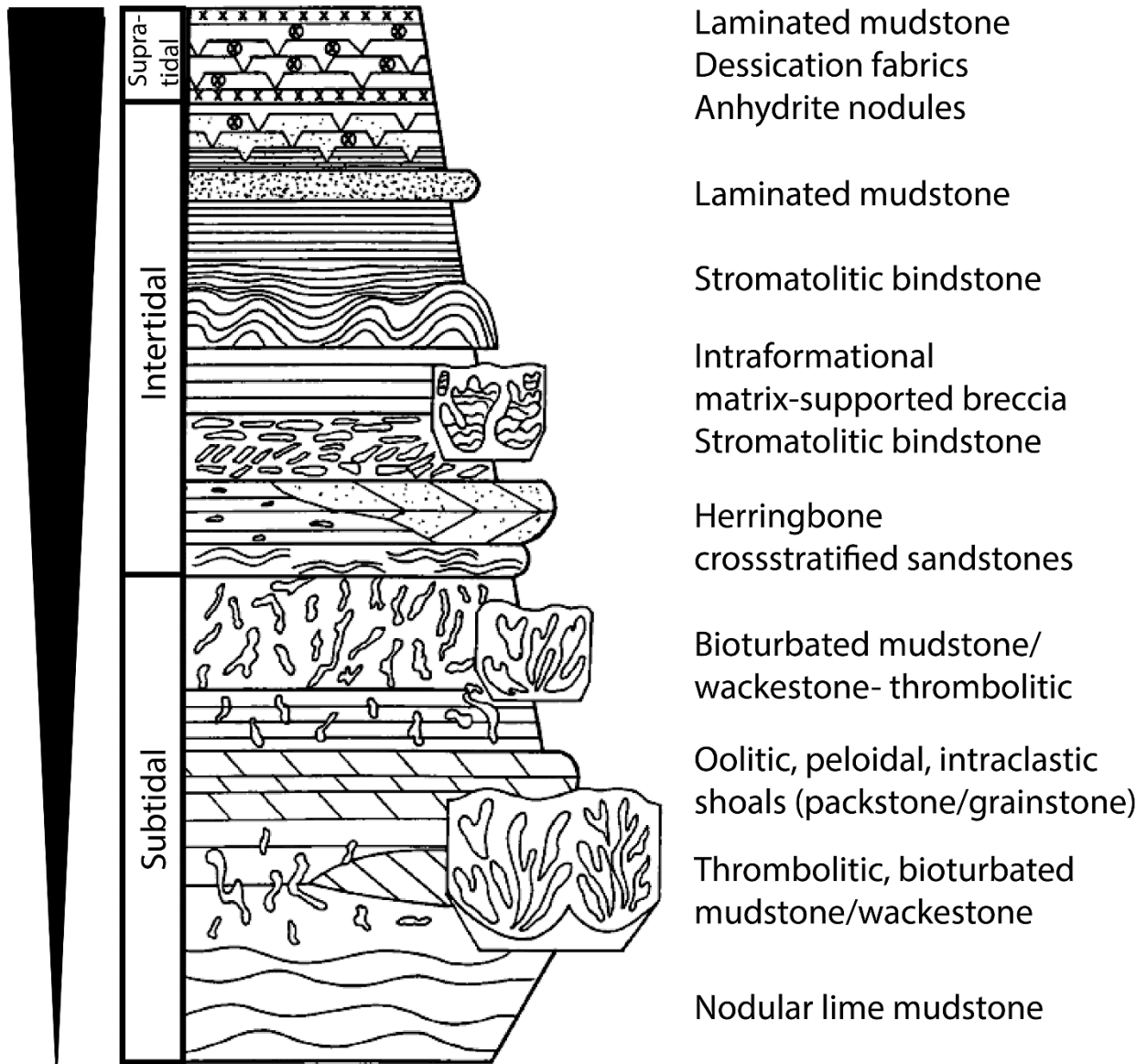
**TEXT-FIGURE 5.—Outcrop photograph from the Cool Creek Formation (stop 1), across from the gift shop at the Turner Falls overlook. See text for explanation of facies. GPS: 34°25'36.44''N, 97° 8'42.32''W.**

anhydrite and celestite crystals that escaped silicification (Ragland and Donovan 1986). Macro-scale evidence of replacement of evaporites includes solution-collapse breccias in multiple intervals of the Cool Creek (St. John and Eby 1986).

### Stop 2: I-35 Overlook, Kindblade Formation, Collings Ranch Conglomerate

The scenic view at stop 2 shows the Bromide Formation (see Johnson et al. 1984, fig. 8) exposed just to the north along I-35N, in fault contact with exposures of the northward dipping strata of the upper Arbuckle group (Kindblade and West Spring Creek formations) exposed immediately to the south.

Limestone beds of the Kindblade Formation (Arbuckle Group) are exposed at the scenic I-35 overlook at stop 2. The Kindblade is the middle unit of the Upper Arbuckle Group (Fig. 3), and was deposited in supra to subtidal marine environments. Loch (2007, fig 1) identified two separate facies belts: hypersaline supratidal to shallow subtidal marine limestones, and more distal, faunally diverse, open marine limestones. Supratidal to shallow subtidal deposits are

Laminated mudstone
Dessication fabrics
Anhydrite nodules

Laminated mudstone

Stromatolitic bindstone

Intraformational
matrix-supported breccia
Stromatolitic bindstone

Herringbone
crossstratified sandstones

Bioturbated mudstone/
wackestone- thrombolitic

Oolitic, peloidal, intraclastic
shoals (packstone/grainstone)

Thrombolitic, bioturbated
mudstone/wackestone

Nodular lime mudstone

**TEXT-FIGURE 6.—Idealized shallowing upward cycle from the Cool Creek Formation (Arbuckle Group) (modified from Wilson et al. 1991; Cardott and Chaplin 1993). Figure 5 shows facies from a portion of this cycle.**

dolomitized across much of the state (Ross 1976; Loch 2007). In comparison to the underlying Cool Creek Formation (stop 1), the Kindblade represents a return to more "open" marine conditions, with comparatively fewer sedimentary structures indicative of evaporative conditions (e.g., pseudomorphs and oolites).

**TEXT-FIGURE 7.—Limestone dissolution megabreccia in the Kindblade Limestone, stop 2 (GPS: 34°25'34.26"N, 97° 8'4.29"W).**

The outcrop consists of thin- to thick-bedded subtidal marine wackestones, packestones, and locally stromatolitic boundstones.  Oolitic and peloidal packstones often form sharp boundaries with silt-laminated, digitate boundstones.  Fay (1989) described three distinct facies associations that can be seen in this outcrop, which likely record a larger–scale regressive cycle. Lower deposits are characterized as thin to medium bedded, bioturbated lime mudstones, with thin skeletal grainstones interpreted as tempestites.  The middle association consists of oolitic wackestone and packstones, interbedded with lime mudstone.  Fay (1989) and Loch (2007) interpreted this unit as being shallower than the lower unit, possibly representing transport of ooid shoal allochems into a more restricted, lagoonal environment.  The absence of oolitic grainstones (expected in a high energy shoal) is consistent with this hypothesis.  The uppermost association consists of thick-bedded lime mudstones with dolomitic partings, stromatolitic boundstones, and a decrease in faunal diversity.  Most authors (Fay 1989; Osleger and Read 1991: Loch 2007) consider this lithology consistent with restricted circulation.  Therefore, lagoonal deposits are a likely explanation, but the Kindblade probably did not reach intertidal conditions.

Another interesting feature of the Kindblade Limestone at this exposure is a megabreccia (Fig. 7) with a framework of randomly oriented boulders in an otherwise undeformed sequence of strata.  Tapp (1978) showed that insoluble clay layers encased the boulders and were nearly

$90^{\circ}$ to layering. This strongly suggests that the megabreccia formed by dissolution of limestone along joints during the Pennsylvanian, and subsequent collapse of the boulders during Ouchita uplift.  Thus, it is interpreted as both a karst and tectonic feature. Stop 2 also exposes one of the best examples of the Pennsylvanian Collings Ranch Conglomerate (CRC) in the Arbuckles.  The CRC is an interesting unit because it is a coarse orogenic product that records evidence of the faulting, folding and uplift in the late Pennsylvanian.  The CRC is a limestone boulder conglomerate (polymict and grain-supported in most exposures) that was deposited in an area trending from the NW-SE as the Arbuckle Mountains were still rising (Cardott and Chaplin 1993).  The unit crops out in the northern portion of the Arbuckle Anticline along the Washita Valley Fault zone.  The conglomerate is unconformable with the underlying Bromide and Viola Groups (seen in stop 7 of this trip), and is preserved as a NW-SE trending, fault-bounded graben. Provenance studies (e.g., Nick and Elmore 1990) have shown that the limestone clasts mostly originate from various formations within the upper and lower Arbuckle Group, with sandstone and limestone clasts derived from the Simpson Group present in small quantities.  The lack of rhylolitic clasts suggests that the core of the Arbuckles was not exposed during Pennsylvanian erosion, or possibly that the drainage area did not include the Cambrian Colbert Rhyolite (Cardott and Champlin 1993).

Individual cycles of poorly sorted, graded, and grain-supported clasts are well exposed in outcrop at stop 2.  These deposits are considered by most authors to have formed as debris and mud flows in an intermontane alluvial fan complex, with both sheet and channel geometries. Cardott and Champlin (1993) suggested that matrix-starved intervals with lenticular geometries represent a more proximal braided stream environment, suggestive of rapid channel shifting.

**DAY 2**
**Stop 3: Black Knob Ridge, Katian GSSP, Womble Shale, Bigfort Chert, and Polk Creek Shale.**

The Upper Ordovician rocks that crop out in the Ouachita Mountains of west-central Arkansas and southeastern Oklahoma are composed primarily of graptolite-rich shales associated with deep-water limestones and cherts (Ethington et al. 1989). These strata were deposited in the deep marine environment of the Ouachita Foredeep off the southern margin of Laurentia (Finney 1988). The rich graptolite faunas have been used traditionally to correlate these rocks with other Upper Ordovician successions in North America and around the world. A correlation chart for the main lower Upper Ordovician graptolite zonal successions discussed in this field guide is provided below.
In southeastern Oklahoma Upper Ordovician strata are exposed along Black Knob Ridge, a low narrow ridge at the extreme western end of the Ouchita Mountains (Hendricks et al. 1937; Finney 1988).  The units exposed along Black Knob Ridge are, in ascending order, the Womble Shale, Bigfork Chert, and Polk Creek Shale. The base of the Ordovician succession is in fault contact with the Pennsylvanian Atoka Formation and the Silurian age Blaylock Sandstone
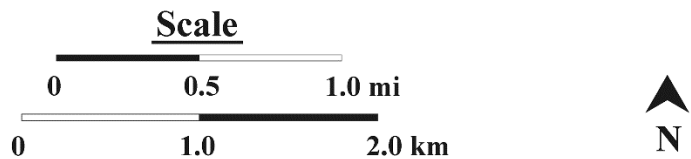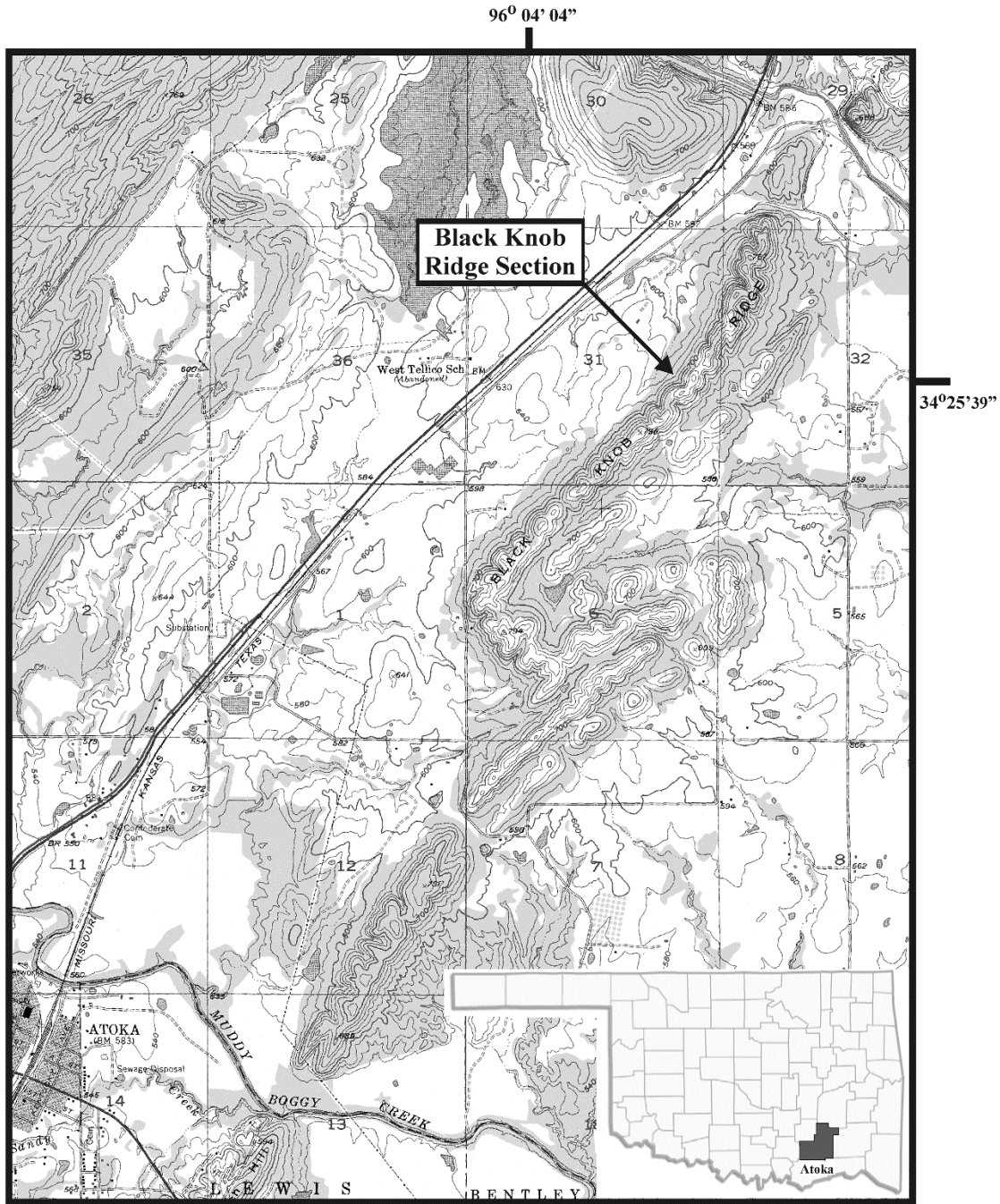
**TEXT-FIGURE 8.—Locality map for the Black Knob Ridge Section. The section is located 5 kilometers north of the town of Atoka, SW1/4, Section 31, T. 1S, R. 12E, Atoka County, Oklahoma; 34° 25' 39.08" N, 96° 04' 3.78" W. From Goldman et al., 2007.**

disconformably overlies the top of the sequence (Ethington et al. 1989).  An excellent exposure of the Womble to Polk Creek succession occurs on a hill slope approximately 5 kilometers north of the town of Atoka, SW1/4, Section 31, T. 1S, R. 12E, Atoka County, Oklahoma; 27˚ 25.9' N, 96˚04.5' W (Figure 8). This exposure, which we refer to as the Black Knob Ridge (BKR) section (Figure 9A, B), extends along strike for several hundred meters, is readily accessible, contains a continuous graptolite succession across the *Climacograptus bicornis – Diplacanthograptus caudatus* zonal boundary, and yields biostratigraphically important conodonts and chitinozoans.

The International Subcommission on Ordovician Stratigraphy (ISOS) recommended that the first appearance of *Diplacanthograptus caudatus* at the Black Knob Ridge section be used as the GSSP of the middle stage of the Upper Ordovician Series. The ISOS also recommended the

| Global Stages | Laurentia (Oklahoma & Eastern USA) | | Pacific Faunal Province (General) | Australia | Laurentia (Scotland) |
|---|---|---|---|---|---|
| **Katian** | *Amplexo. manitoulinensis* | | *Styraco. tubuliferus* | *Dicello. gravis* | *Pleuro. linearis* |
| | *Geniculo. pygmaeus* | | | *Dicrano. kirki* | |
| | *Diplacantho. spiniferus* | *D. spiniferus* | *Diplacantho. caudatus* | *Diplacantho. spiniferus* | *Dicrano. clingani* |
| | | *O. rudemanni* | | | |
| | *Diplacantho. caudatus* | *Corynoides americanus* | | *Diplacantho. lanceolatus* | |
| **Sandbian** | *Climaco. bicornis* | | *Climaco. bicornis* | *Ortho. calcaratus* | *Climaco. bicornis* |

**Table 1.—Correlation of Upper Ordovician biozones discussed in this field guidebook. Key references used in constructing this chart are Finney (1986), Goldman et al (2007), Riva (1969, 1974), VandenBerg and Cooper (1992), and Zalasiewicz et al. (1995).**

designation Katian Stage for this stage, a name derived from the nearby Katy Lake (now drained) near the southern end of Black Knob Ridge (Bergström et al. 2006). These decisions were ratified by the ICS in 2006 (Bergström et al. 2006).

At the BKR section, approximately 20 meters of dark, graptolite-rich Womble Shale are exposed. The lower 15 meters is composed of blocky, white to chocolate brown weathering mudstone, interbedded with fissile black shale, siliceous limestone, and bedded chert. Within this unit, approximately 15 meters below the base of the Bigfork Chert is a distinct channel-like feature that is filled with 80 cm of soft, fissile, dark gray shale with abundant specimens of *Dicellograptus* sp. (Figure 9C). This feature might be interpreted as the M3-M4 sequence boundary of Holland and Patzkowsky (1996).  About five meters below the Bigfork Chert the
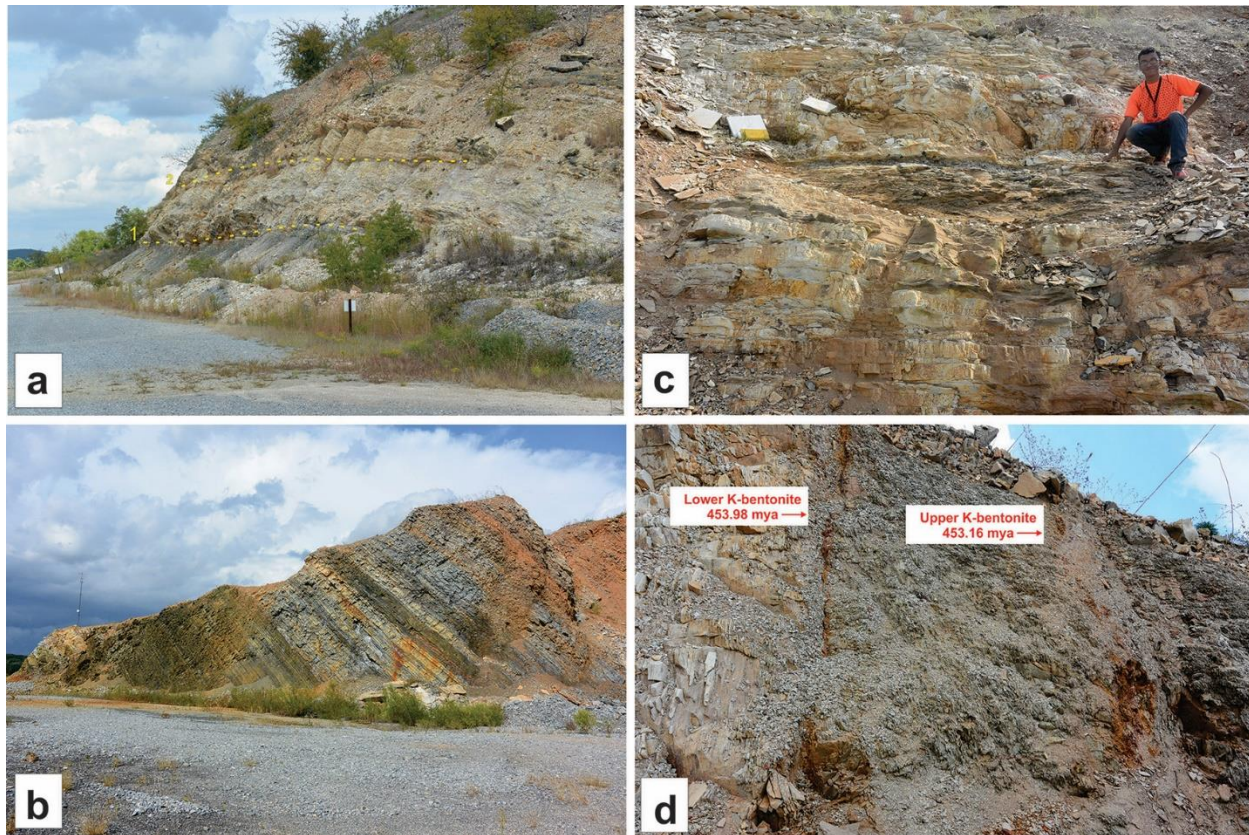
**TEXT-FIGURE 9.—The Black Knob Ridge Section. a, The Upper Womble Shale and Bigfork Chert at the Black Knob Ridge Section. 1, The contact between the two units is placed at the first organic-rich, siliceous shale that forms a prominent ledge. This bed is extremely rich in graptolites and conodonts. 2, The base of the *Diplacanthograptus caudatus* Zone. The yellow dashed line is 4.0 meters above the base of the Bigfork Chert and marks the FAD of *D. caudatus*. b, Steeply dipping uppermost Womble Shale and Bigfork Chert. Up-section is to the right. c, Channel in the upper Womble Shale. d, Distinctive shale package at the top of the Womble Shale. This 3.65 meter intervals contains two K-bentonites with ages of 453.98 and 453.16 mya, respectively (Sell et al. 2013).**

mudstone becomes more fissile, thinner bedded, and platy. These beds are in turn overlain by a distinct 3.65 meter package of soft, friable shale that has a 4 cm K-bentonite at its base and another 10 cm K-bentonite at 1 meter above the unit base (Leslie et al. 2008; Figure 9D herein). These beds have zircon phenocrysts that yielded U/Pb ages of 453.98 Ma and 453.16 Ma, respectively (Sell et al., 2013). These age data suggest that in addition to their correct biostratigraphic position (upper *C. bicornis* Zone), the Womble K-bentonites at the BKR section are potentially the Deicke and Millbrig K-bentonites (Sell et al. 2013). These dates also fit well with the most recent age of the base of Katian, 453 Ma, the date given in the latest Ordovician time scale by Cooper et al. (2012). It is possible that this shale package correlates either with the upper Pooleville Member of the Bromide Formation at Highway 99, which also contains two K-bentonites (Leslie et al. 2008), or the lowermost 0.8 meters of the Viola Springs Formation at that same section (see below).

Conformably overlying the Womble Shale are approximately 145 meters of Bigfork Chert (Figures 9). The contact between the two units appears gradational. The base of the Bigfork Chert is a 0.5 meter interval of hard, splintery black shale that contains abundant conodonts and chitinozoans. The Bigfork Chert is composed of nodular and bedded chert, and siliceous mudstone intercalated with black shale and siliceous limestone. Because limestone beds are absent in the shale below and above the Bigfork Chert, its upper boundary is placed at the last limestone bed in the section (Finney 1988). The limestone is medium bedded, siliceous, fine- to coarse-grained skeletal calcarenites. Fossils include graptolites, conodonts, chitinozoans, sponge spicules, inarticulate brachiopods, and radiolarians (Hendricks et al. 1937) with skeletal fragments of pelmatozoans and brachiopods (Finney 1988). Decker (1935) noted an abundance of the trilobite *Cryptolithus* in some beds of the Bigfork Chert at the Stringtown Quarry (BKR); this trilobite is also typical of the correlative Viola Springs Formation. There is no evidence of a depositional break within the lower Bigfork Chert at the study section.

The Polk Creek Shale overlies the Bigfork Chert with an apparently conformable contact. Although the Polk Creek has not been measured at the BKR section, Hendricks et al. (1937) measured 43 meters at the Atoka city trash dump and Dworian (1990) recorded 32 meters from a locality along Black Knob Ridge south of the stratotype section. Decker (1935) correlated the Polk Creek Shale with the Sylvan Shale (see Stop 6) and showed that both had a similar graptolite fauna, including *Dicellograptus complanatus* that indicates a late Katian age.

The upper Womble shale contains an abundant graptolite fauna that is referable to the *Climacograptus bicornis* Zone. Diagnostic elements of this fauna include *C. bicornis*, *C. bicornis tridentatus*, *C. cruciformis*, *Orthograptus whitfieldi*, *O. calcaratus* ssp., *Archiclimacograptus modestus*, *Dicranograptus spinifer*, *D. contortus*, *D. arkansasensis*, *Normalograptus brevis*, and *Nemagraptus gracilis*. The transition between the *C. bicornis* Zone and the underlying *N. gracilis* Zone has not yet been found at Black Knob Ridge.

*Climacograptus bicornis*, *C. bicornis tridentatus*, *Archiclimacograptus modestus*, and *Dicranograptus arkansasensis* range upward into the lowermost 3.1 meters of the Bigfork Chert. *Orthograptus quadrimucronatus* makes its first appearance 3.2 meters above the base of the Bigfork Chert. The base of the Katian Stage of the Upper Ordovician Series is placed at the FAD of *Diplacanthograptus caudatus*, 4.0 meters above the base of the Bigfork Chert (Figure 10). At this horizon, several taxa diagnostic of the *D. caudatus* Zone first appear. These are, *Orthograptus pageanus*, *Neurograptus margaritatus*, and *Corynoides americanus*. *Dicranograptus hians* was found 2.0 meters higher up. *Diplacanthograptus spiniferus* and *Climacograptus tubuliferus* debut at 9.8 and 52.5 meters, respectively, above the base of the Bigfork Chert (Finney 1986 and personal communication). Characteristic graptolite species from the *Climacograptus bicornis* and *Diplacanthograptus caudatus* zones are illustrated in Figure 11.

Conodonts have been known from the Black Knob Ridge since Hendricks et al.'s (1937) description of the geology at this section. Harlton (1953) also reported the occurrence of conodonts at BKR. However, as Repetski and Ethington (1977), and Ethington et al. (1989) reported, these early studies did not identify the conodonts, and their stratigraphic occurrences
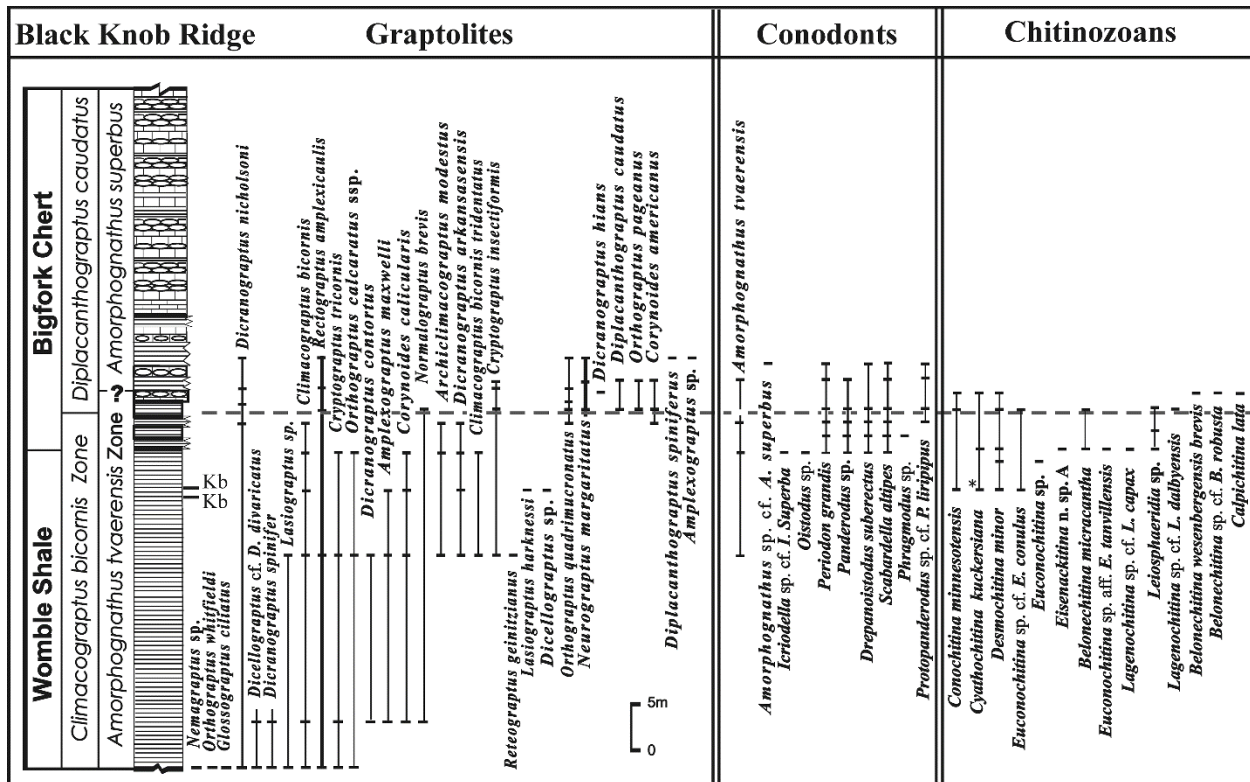
**TEXT-FIGURE 10.—Stratigraphic column with a range chart of graptolites, conodonts, and chitinozoans for the Black Knob Ridge section. The GSSP for the base of the Katian Stage is placed at the first appearance of *D. caudatus* four meters above the base of the Bigfork Chert. The *C. bicornis - D. caudatus* zonal boundary occurs high in the *Amorphognatus tvaerensis* conodont Zone. From Goldman et al., 2007.**

were not documented precisely. Bradshaw (1974) identified a conodont fauna of Midcontinent aspect from the Bigfork Chert at BKR. She reported occurrences to the genus-level, and identified a fauna of *Panderodus*, *Belodina*, *Drepanodus*, *Oistodus*, and *Phragmodus* from siliceous limestone beds. More recently, Krueger (2002) reported on the occurrence of conodonts from the Stringtown Quarry approximately 3 kilometers north of the BKR section. The fauna that Kreuger reported from the limestone beds is also of Midcontinent aspect.

Unfortunately, the former limestone beds in the lower part of the Bigfork at the BKR section are completely silicified and cannot be dissolved for conodonts. There are also well-preserved conodonts on the dark shale bedding planes in both the Womble Shale and Bigfork Chert that are typical of the North Atlantic Fauna. The uppermost Womble Shale contains an abundant, low diversity but biostratigraphically important conodont fauna that includes elements of *Amorphogathus tvaerensis* and *Icriodella* cf. *I. superba* (Figure 10). The presence of these species demonstrates that the uppermost Womble at BKR is within the *B. alobatus* Subzone of the latest *A. tvaerensis* Zone (Bergström 1982). It is younger than the *B. gerdae* Subzone fauna reported elsewhere from the Womble by Repetski and Ethington (1977; Ethington et al. 1989). The uppermost Womble Shale at BKR also contains *Periodon grandis*, *Drepanoistodus*
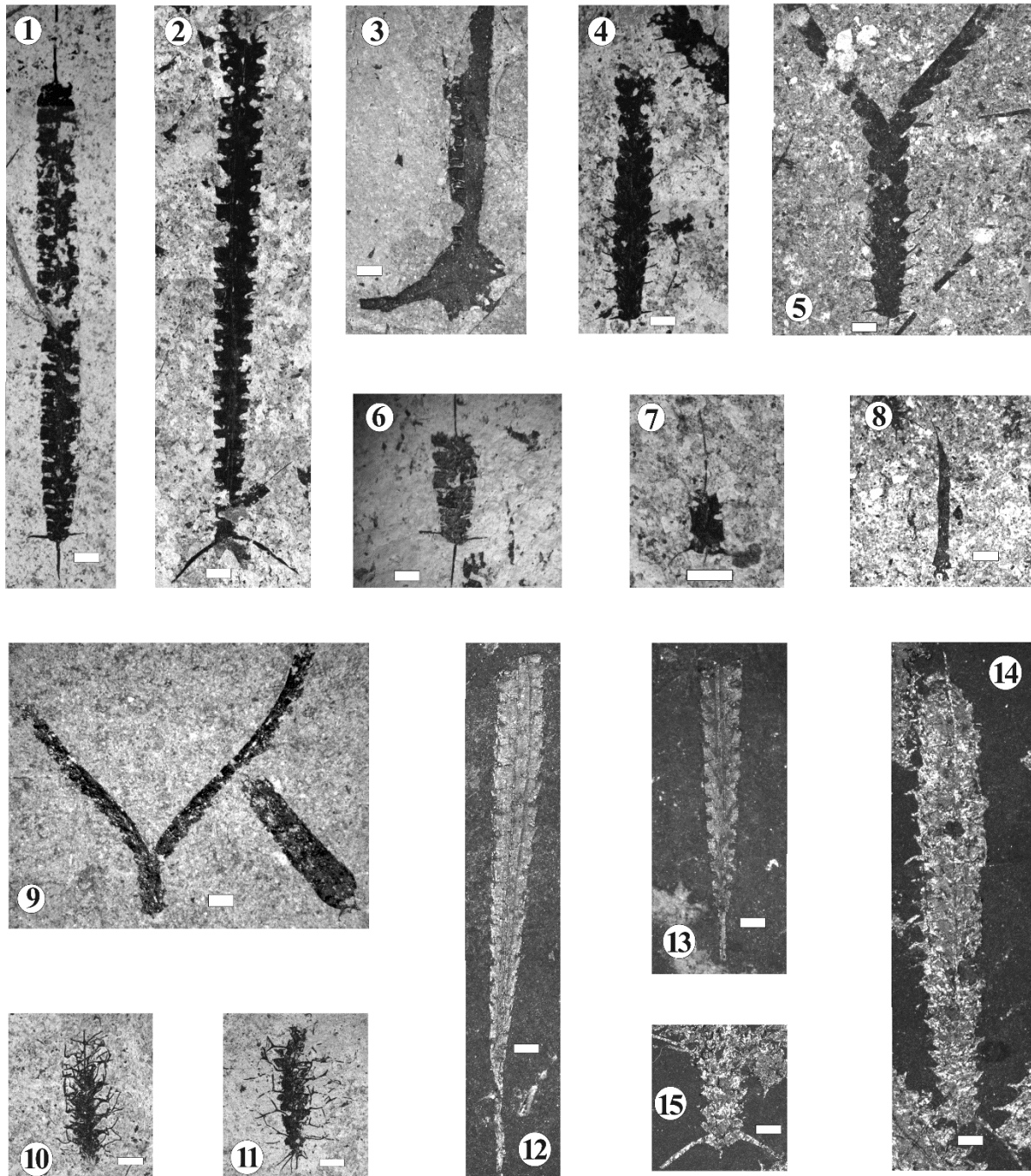
**TEXT-FIGURE 11.—Graptolites from Black Knob Ridge. 1–8, graptolites from the** *Climacograptus bicornis* **Zone. 1, 6,** *Archiclimacograptus modestus***; 2, 7,** *Climacograptus* **bicornis; 3,** *Climacograptus bicornis tridentatus***; 4, 5,** *Dicranograptus spinifer* **(=** *D. nicholsoni longibasalis***); 8,** *Corynoides calicularis***. 9–15, graptolites from the** *Diplacanthograptus caudatus* **Zone. 9,** *Dicranograptus hians* **and** *Cryptograptus insectiformis***; 10, 11,** *Neurograptus margaritatus***; 12, 13,** *Diplacanthograptus caudatus***; 14, 15,** *Orthograptus pageanus.* **Scale bar on each photograph is 1 mm.**

*suberectus*, *Dapsilodus* sp. aff. *D. mutatus*, *Oistodus* sp., and *Panderodus* sp. (Goldman et al. 2007).

The conodont fauna from lowermost Bigfork Chert at BKR consists of *A. tvaerensis*, *Periodon grandis*, *Protopanderodus* cf. *P. liripipus*, *Drepanoistodus suberectus*, *Dapsilodus* sp.aff. *D. mutatus*, *Phragmodus* sp., and *Panderodus* sp. This fauna is nearly identical to that from the upper Womble, with the exception of relatively abundant specimens of *P.* cf. *P. liripipus*. Of particular interest is the occurrence of two specimens of *Amorphognathus* sp. approximately 5.7 meters above the base of the Bigfork Chert (Goldman et al., 2007). These specimens are morphologically very similar to *A. superbus*, but unquestionable identification is not possible from the material at hand.

The biostratigraphically significant conodonts known from BKR suggest that the *Climacograptus bicornis - Diplacanthograptus caudatus* zonal boundary is located in the *B. alobatus* subzone of the *Amorphognathus tvaerensis* conodont zone (Goldman et al., 2007). This correlation is consistent with the graptolite – conodont zonal relationships described from Europe and eastern North America by Bergström (1971, 1986) and Goldman et al. (1994).

**Stop 4: Bromide Formation, Viola Springs Formation, Katian Auxiliary GSSP**

The HW 99/377 section has a long history of study (e.g., Decker 1933; Wengerd 1948), and the locality is known in the older literature as "Murray Lane". The section encompasses the upper Sandbian and lower Katian succession, with the boundary lying somewhere within the lower 20 m of the Viola Springs Formation. It was designated as an auxiliary stratotype for the base of the Katian (Goldman et al. 2007) because it yields diverse shelly faunas in addition to graptolites and conodonts. Young et al. (2005) demonstrated the presence of a positive carbon isotope excursion that they equated with the GICE excursion of the Upper Mississippi Valley, although, as discussed below, this correlation has been disputed (Westrop et al. 2012). The underlying Bromide Formation is less fossiliferous, although a recent sequence-stratigraphic framework (Carlucci et al. 2014) facilitates correlation with the classic localities farther to the south (e.g., Sutherland and Amsden 1959; Shaw 1974).

*Overview of the succession*

*Bromide Formation*. The HW 99 locality includes the type section of the basal Pontotoc Member of the Bromide Formation (Carlucci et al. 2014), which comprises a succession of cross-bedded to rippled sandstones and siltstones. In platformal sites like HW 99, it represents the TST of the lowest of three depositional sequences identified in the Bromide (Carlucci et al. 2014). The overlying Mountain Lake Member (HST of Sequence 1; Sequence 2) is poorly exposed, but Fay et al. (1982b) provide a detailed description of exposures from an adjacent quarry that is now almost completely covered. Higly fossiliferous, greenish gray shales and thin packstones of the "lower Echinoderm Zone" are still exposed, though largely covered, along the northwest side of the roadcut, southwest of crossing of the gully, which separates the roadcuts into two sectors. These shales yield abundant plates of hybocrinids, rhombiferans and other
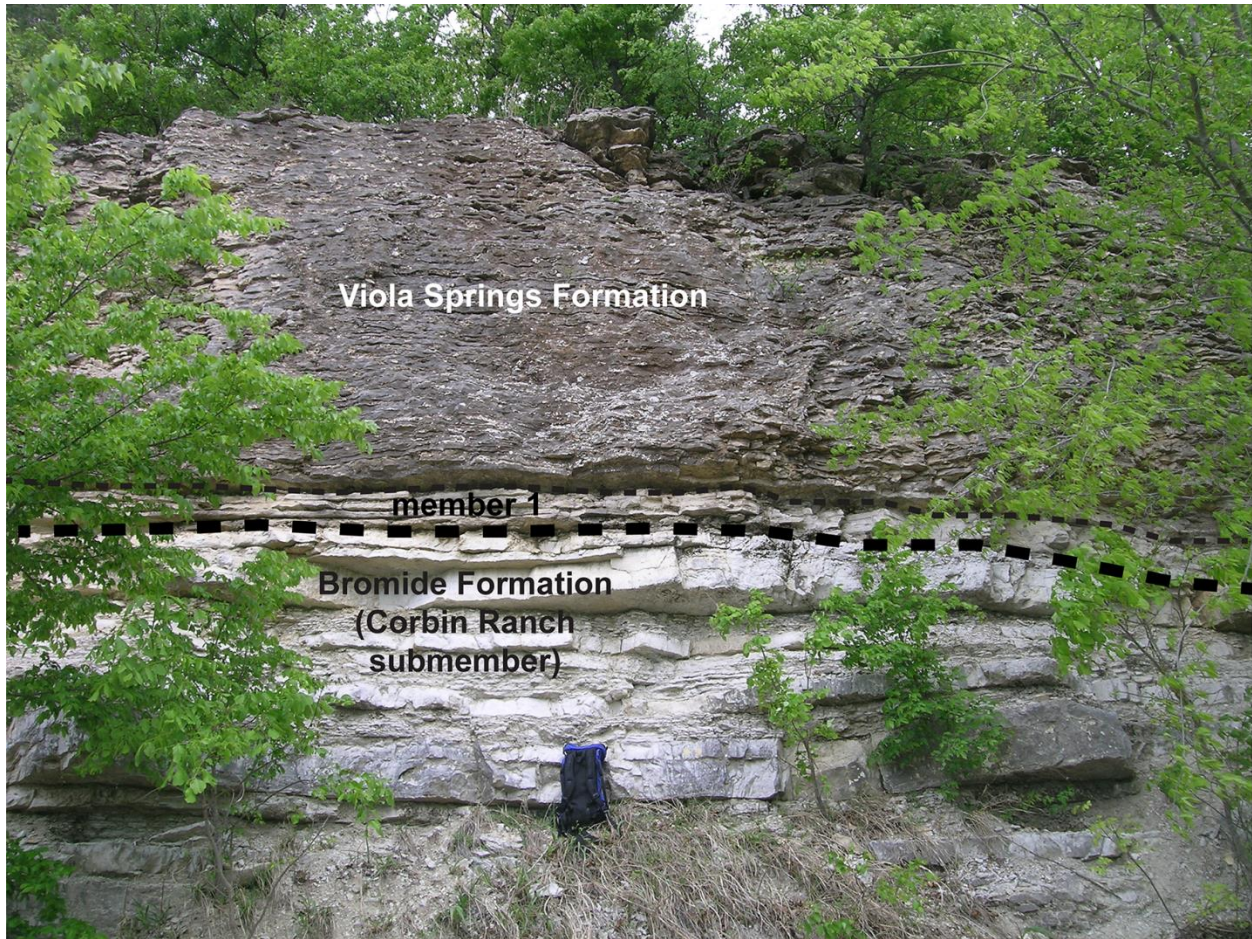
**TEXT-FIGURE 12.—Outcrop photograph (stop 4) showing the position of "member 1" of the Viola Formation (to be formally named elsewhere) relative to the peritidal cycles of the Corbin Ranch Submember of the Bromide Formation. GPS: 34°34'11.11"N, 96°37'52.31"W.**

echinoderms as well as diverse brachiopods and bryozoans. An overlying bundle of sandstones and sandy limestones crops out along the same cut just SW of the gully and area of covered section.

The youngest strata assigned to the Bromide Formation, fully exposed NE of the gully, represent the Pooleville Member (HST of Sequence 3). The lower Pooleville is for the most part a sparsely fossiliferous succession of wackestone and calcisiltite that is overlain by peritidal carbonates of the Corbin Ranch submember (Amsden, in Amsden and Sweet 1983) (Figure 12). The latter includes the rhynchonelliform brachiopod *Ancistrorhynchia* (Amsden, in Amsden and Sweet 1983), and rare sclerites of the trilobite *Bathyurus* (e.g., Ludvigsen 1978); the latter is indicative of the traditional Blackriveran stage of eastern North America. The Pooleville is also noteworthy because bentonites at 4.1 and 11.3 m below the top of the Corbin Ranch Member have been correlated with the well-known Millbrig and Deicke K-bentonites (Rosenau et al. 2012). However, these bentonites have not yet been geochemically "fingerprinted", and interpretation rests largely upon stratigraphic position.

*Viola Springs Formation*.  The Viola Springs Formation begins with a thin (~ 0.8 m) package of mostly deep subtidal, graptolitic wackestone and shale that has a sequence-like architecture (sequence 4 of Carlucci et al. 2014).  It will be named formally elsewhere and is simply designated "member 1" herein (Figure 12).  At the base, the TST is recorded by a condensed interval of packstone and grainstone that includes rare sclerites of unnamed species of *Cryptolithus* (Amati 2014) and *Flexicalymene* (Amati 2004) that indicate a "Trentonian" age, and a cm-thick clay (Decker 1933) that may represent a bentonite.  The overlying HST comprises about five beds of wackestone and shale that yield an upper Sandbian graptolite fauna.  Previous workers (e.g. Young et al. 2005; Leslie et al. 2008) have agreed that the base of the Viola Springs is the M4–M5 sequence boundary that has been correlated widely across eastern North America.  Member 1 likely represents the M5A of the Cincinnati region (Carlucci et al. 2014).

Member 1 is truncated by an irregular unconformity surface that is succeeded by about 15 m of cherty, laminated lime mudstone, wackestone and calcisiltite, with a sparse fauna of cryptolithine and isoteline trilobites, and graptolite fragments (Amati 2014, appendix 2).  A major lithofacies change to coarse, bryozoan-rich, bioclastic grain- and rudstone (Amati and Westrop 2006) occurs at about 18 m above the base of the Viola Springs, although the contact with the underlying cherty limestone is not exposed. This high-energy bioclastic facies presumably represents the TST of a depositional sequence that we suspect is correlative with sequence M6 of eastern North America.

About 40 m above the base of the Viola Springs, the succession shifts to deeper subtidal, trilobite-brachiopod bioclastic pack- and rudstone interlayered with wackestone (high diversity wackestone–rudstone facies of Amati and Westrop 2006).
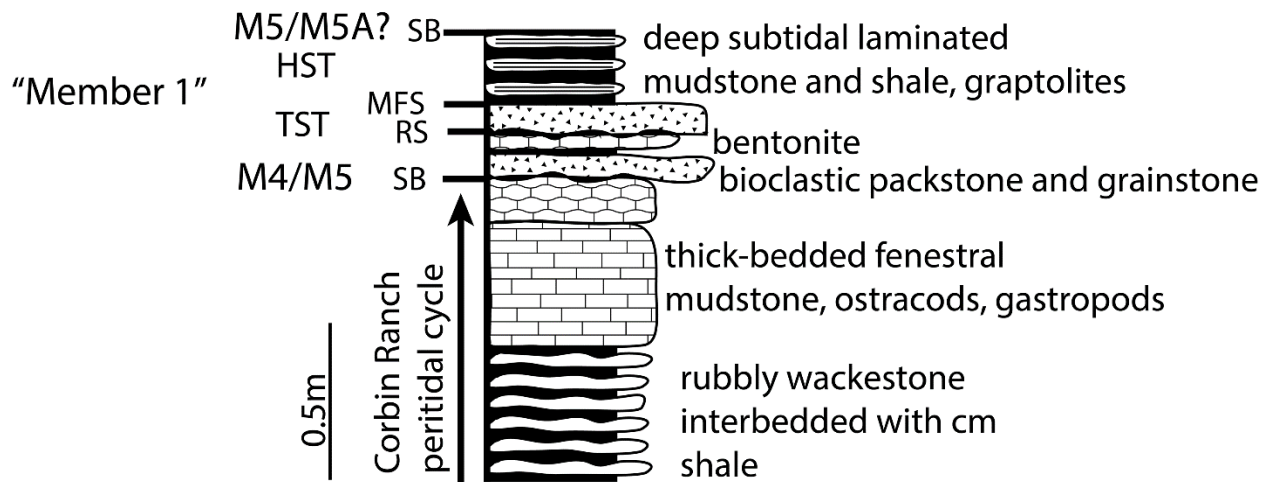


**TEXT-FIGURE 13.—Stratigraphic column of "member 1" and inferred position of sequence boundaries, maximum flooding surfaces, ravinement surfaces, and Mohawkian depositional sequences (M4, M5, M5A).**

### *Age and correlation of the HW99 section*

Recent assessment of the conodont biostratigraphy of the HW 99 section (Young et al. 2005) follows graphic correlation by Sweet (1983, 1984). The Bromide and lower Viola Springs (including member 1) falls within the *Phragmodus undatus* Zone (Sandbian). The bulk of the Viola Springs has been assigned to the *Plectodina tenuis* Zone, with only the uppermost few meters representing the *Belodina confluens* Zone (Young et al. 2005, fig. 3). Thus, the conodont biostratigraphy is consistent with the interpretation of the unconformable contact of the Bromide and Viola Springs as the base of the M5 depositional sequence. However, the conodont faunas would also suggest that the entire Viola Springs at HW 99 lies within sequence M5, and that the carbon isotope excursion is correlative with the GICE (Bergström et al. 2010, fig. 4). It also implies that the coarse bioclastic TST at about 18 m records a lower rank sequence (4th order?) within M5.

The trilobite and graptolites also support the conclusion that the base of the Viola Springs (member 1) is of late Sandbian age, and correlative with the base of M5. However, in the upper half of the section, the graptolite and trilobite faunas indicate a significantly younger age than is suggested by the conodonts. The trilobites belong to the *Bumastoides* and *Thaleops* biofacies of Amati and Westrop (2006), and include a number of species (e.g., Amati and Westrop 2004; Amati 2014; Swisher et al. 2015) that are known to occur in the Kimmswick Limestone in the St. Louis, Missouri, area. There the Kimmswick rests unconformably on strata, including the Guttenburg Limestone, that surely record the GICE (Metzger and Fike 2013), and the trilobite faunas are no older than M6, and may extend into correlatives of the C1 sequence (Swisher 2015). Elements of the Kimmswick trilobite fauna enter the HW 99 succession in the coarse bioclastic grain- and rudstone facies at about 18 m above the base of the Viola Springs (Amati 2014), and indicate that this is in fact the local expression of a TST at the base of sequence M6. Moreover, if the age of the Viola Springs suggested by the trilobites is correct, then there are obvious implications for the identity of the carbon isotope excursion present in the upper half of the section. It cannot be the GICE, and may in fact the same excursion that Bergström et al. (2010) identified as the "Kope excursion" in the lower part of the Viola Springs at the I35 south section (Westrop et al. 2012). A correlation chart illustrating the Upper Ordovician graptolite zones, conodont zones, and important event and chemostratigraphic marker horizons in the principal localities discussed in the text is shown in Table 2 below.

The differing perspectives on the age of the upper half of the Viola Springs at HW 99 demand that either the conodont or the trilobite faunas have diachronous first appearances. They can be reconciled only if the conodont index species have relatively late (stratigraphically young) first appearances relative to the trilobites or, alternatively, species of the Kimmswick trilobite fauna appear relatively earlier in Oklahoma. This problem is still not fully resolved, but graptolite faunas seem to favor diachroneity of conodont species.

| Series | Stage | N. American Sequences and Graptolite Zones | | | North Atlantic Conodont Zones | | N. American Midcontinent Conodont Zones |
|---|---|---|---|---|---|---|---|
| | | Sequences | Pacific Province | Eastern N.A. | | | |
| Upper Ordovician | Katian | C5 | Dicellograptus complanatus | Dicellograptus complanatus | Amorphognathus ordovicicus | | A. grandis |
| | | C4 | Styracograptus tubuliferus | Amplexograptus manitoulinensis | Amorphognathus superbus | | |
| | | C3 | | | | | O. robustus |
| | | C2 | | G. pygmaeus | | | O. velicuspis |
| | | C1 | Kope δ¹³C Excursion | Diplacanthograptus spiniferus | | | B. confluens |
| | | M6 | Diplacanthograptus caudatus | | | | |
| | | M5 | | O. ruedemanni | | | |
| | | | | Corynoides americanus | | | |
| | Sandbian | M4 | GICE δ¹³C Excursion | | Amorphognathus tvaerensis | B. alobatus | P. tenuis |
| | | M3 | Millbrig k-bentonite | | | | P. undatus |
| | | | | | | | B. compressa |
| | | M2 | Climacograptus bicornis | Diplograptus foliaceous (= D. multidens) | | B. gerdae | E. quadridactylus |
| | | M1 | | | | | P. aculeata |
| | | | Nemagraptus gracilis | Nemagraptus gracilis | | B. variabilis | C. sweeti |
| | | | | | | P. anserinus | |

**Table 2.—Correlation chart illustrating the Upper Ordovician graptolite zones, conodont zones, and important event and chemostratigraphic marker horizons discussed in the text. Key references used in constructing this chart are Finney (1986), Goldman et al (2007), Bergstrom et al. (2010) and Webby (2004).**

Graptolites are unevenly distributed through the Viola Springs Formation, and the boundary between the *Climacograptus bicornis* and *Diplacanthograptus caudatus* zones is poorly constrained (Goldman et al. 2007). The lowermost 0.4 meters of the Viola Springs Formation (Member 1) contain *Climacograptus bicornis*, *Dicranograptus semispinifer* (= *D. nicholsoni longibasalis* Ruedemann), *Normalograptus brevis*, *Corynoides calicularis*, and *Hustedograptus* sp., a fauna that that belongs to the Upper Sandbian *C. bicornis* Zone. Interestingly, collections at 0.45 and 0.55 m contain graptolites that indicate the strata are uppermost Sandbian in age. The 0.45 m collection contains the graptolites *Lasiograptus* n.sp. A. and *Rectograptus intermedius*, species that also occur in the 75NY-2 drill core from eastern New York State (Roloson 2010) and outcrops from the Gaspé Peninsula, respectively (Riva, pers. com.). These taxa occur in an interval that postdates the LAD of *C. bicornis*, but just before the FAD of *D. caudatus*. A collection at 0.55 m contains *Orthograptus quadrimucronatus*, a graptolite generally considered to be restricted to Katian age strata. In some Chinese sections this species has been reported in upper Sandbian strata but generally (and in North America) its range is Katian. Hence, the graptolite fauna in Member 1 of the Viola Springs Formation at Highway 99 indicates that these beds belong to the uppermost Sandbian Stage.

The following 20 meters of strata are devoid of graptolites but Finney's (1986) report of *Geniculograptus typicalis* at a horizon 25 m above the base of the Viola Springs and *Diplacanthograptus spiniferus* at 35 m, together with other species of the eastern Laurentian *D. spiniferus* Zone at higher levels (Goldman et al. 2007), is consistent with the upper half of the section being no older than M6. Graptolites collected in the upper half of the Highway 99 section include:

43 m – *Diplacanthograptus caudatus*, *Corynoides americanus*, *Cryptograptus insectiformis*

44 m – *Diplacanthograptus spiniferus*

45 m – *C. americanus*, *Cryptograptus insectiformis*

51 m – *G. typicalis*, *Orthoretiolites hami*, *D. caudatus*, *D. spiniferus*

65 m – *Orthograptus quadrimucronatus*

Additionally, the collection at 43 m contains the Baltic chitinozoan *Angochitina capillata*, a species restricted to the Nabala through Pirgu regional stages in Estonia (Goldman et al. 2007). If its stratigraphic range in North America is consistent with its range in Baltoscandia, then the occurrence of *A. capillata* at this level implies a correlation with the uppermost *D. spiniferus* Zone. No graptolite species that would indicate this interval should be assigned to the younger *Geniculograptus pygmaeus* Zone have been found at the Highway 99 section. This age interpretation fits with the carbon isotope excursion at Highway 99, previously referred to as GICE actually correlating with the Kope Excursion of Bergström et al. (2010).

## DAY 3
## Stop 5: Sam Noble Oklahoma Museum of Natural History

The Sam Noble Museum is both a state natural history museum as well as an academic unit of the University of Oklahoma.  Established in 1899, the Museum moved to a new building on the university campus in 1999, and this move consolidated the collections under one roof.  It is now one of the largest university-based natural history museums in North America.  Three paleontological collections house invertebrates, plants and microfossils, and vertebrates, respectively.

Coincident with the move to the new building, the Invertebrate Paleontology Collection expanded significantly with the acquisition of macrofossils from the Amoco Petroleum Company following its merger with BP.  This material represents the results of decades of work by the Amoco paleontology and biostratigraphy research group.  It includes numerous field samples from measured sections at localities throughout North America, including frontier regions such as Alaska and the Canadian Arctic. The Amoco collection is largely undescribed and represents a significant research resource.

With the incorporation of the Amoco donation and extensive material from the Oklahoma Geological Survey (T.W. Amsden's collections of Ordovician–Devonian brachiopods), the Invertebrate Paleontology Collection now comprises about one million specimens, including about 10,000 type and figured specimens.  Particular strengths include Cambrian and Ordovician trilobites and graptolites, Silurian and Devonian brachiopods, and Carboniferous brachiopods,

corals, and echinoderms.  Much of the collection can be searched via the web at http://samnoblemuseum.ou.edu/collections-and-research/invertebrate-paleontology/integrated-invertebrate-paleontology.

The exhibits at the Museum include an ancient life gallery that spans the Archean to Pleistocene.  Highlights include two Ediacaran dioramas, and a diorama based on the Upper Ordovician (Sandbian) of Oklahoma, as well as a series of dinosaur exhibits that include *Apatosaurus*, *Saurophaganax*, *Deinonychus*, *Tenontosaurus*, and *Pentaceratops*.

**Stop 6: Sylvan Shale, Keel and Cochrane limestones, Ordovician-Silurian contact**

The outcrop at stop 6 is located along HWY 77 on the west size of the I-35 interchange, across the road from Arbuckle Fried Pies. The base of the section is located to the south, where it exposes the uppermost Ordovician (Hirnantian) rocks in the Arbuckle Mountains, and its contact (Fig. 14) with the lowermost Silurian (Llandovery).  Progressively younger strata at this locality include the Sylvan Shale (Ordovician), Keel Formation (Ordovician), Cochrane Formation (Silurian), and Clarita Formation (Silurian). Additional exposed units to the north along HWY 77 are the later Silurian and Devonian strata of the Upper Hunton Group including the Henryhouse Formation, Haragan Formation, and Woodford Formation.

Our focus at stop 6 is the Chimneyhill subgroup of the Hunton (Keel, Cochrane, and Clarita), and the underlying Sylvan Shale.  The Sylvan Shale lies conformably on the Welling Formation (stop 7) of the Viola Group, and forms a recessive interval between the weathering resistant carbonates above and below it.  It formed as a widespread sheet of anoxic, fine siliciclastic silt and clay that extends westward into north Texas, eastward into the Ouchita Mountains region, and northward into the Mississippi Valley region (Playford and Wicander 2006).  The upper portion of the Sylvan (exposed at stop 6) is especially fissile and is lacking in carbonate fines, with the exception of a few marly interbeds and sandy dolomitic stringers.  Pyrite clusters are common, and provide further evidence of widespread reducing conditions prior to the deposition of the Chimneyhill subgroup.  The Sylvan Shale is generally considered to be of upper Katian (Ashgill or Richmondian) age on the basis of biostratigraphy (Amsden 1975).  Graptolites throughout the formation indicate an age no older than the *Dicellograptus complanatus* Zone, while chitinozoans also suggest an "Ashgill" age (Jenkins, 1970, 1971).  Some authors (e.g., Ham 1969) have argued that the Sylvan/Chimneyhill contact is disconformable, but Playford and Wicander (2006) stated that there is little biostratigraphic evidence to support that assertion.

The oldest unit of the Chimneyhill subgroup is the Keel Formation, exposed at stop 6 directly above the Sylvan Shale.  The Keel is a massive, fossiliferous oolitic packstone that contains both micrite matrix and spar cement, and may show silicified ooliths and flecks of glauconite in the matrix.  Brachiopods of the Keel Limestone are Hirnantian in age (Ham 1973), and those of the overlying Cochrane are early Silurian (Llandovery).  Much of the Llandovery is apparently missing at the contact, during which time there must have been regional erosion and truncation of the Keel (Amsden 1963; Ham 1973).  The Cochrane Limestone is the lowest unit of

**TEXT-FIGURE 14.—Ordovician-Silurian contact and formations exposed at stop 6 (GPS: 34°26'44.55''N, 97° 8'7.86''W). Brachiopods of the Keel Limestone are Hirnantian in age (Ham 1973), and those of the Cochrane are early Silurian (Llandoverian).**

the Silurian at stop 6, and is a thick-bedded, glauconitic limestone whose brachiopod fossils (*Microcardinalia protoplesiana*) reveal a late Llandovery (Aeronian to early Telychian) age the upper part that allow for correlation with the Silurian Blackgum Formation in eastern Oklahoma (Amsden, 1966; Ham 1973) and probably with the Sexton Creek Formation of Arkansas, Missouri and Illinois. The overlying thin shale above the Cochrane (Fig. 14), termed Prices Fall Member, is possibly a remnant of a widespread shaly interval termed 76 Shale in Arkansas and Illinois; this interval is a feather edge of latest Llandovery to early Wenlock (*Pterospathodus amorphognathoides* Zone) Osgood Shale of Indiana-Kentucky and records a major Telychian transgression. The overlying Clarita grainstones are likewise a local representative of lower Wenlock (Sheinwoodian) transgressive limestones.

**Stop 7: I-35N, Bromide Formation Reference Section**

Late Whiterockian through Mohawkian (Fig. 3) deposition in the Arbuckle Mountains and Criner Hills regions of Oklahoma is recorded by the mixed carbonate-siliciclastic Bromide Formation, the youngest unit of the Simpson Group.  Bromide deposition took place during a period of relative stability of the Oklahoma Basin (see Carlucci et al. 2014, fig. 1), where a shallow, tropical, carbonate-dominated sea bordered the SOA.  To the north, the Oklahoma Basin was bordered by the Arbuckle platform (Longman 1982b), which was a desert region that likely supplied wind-blown sand deposited as sheets in the basins. As with the majority of Simpson Group formations, carbonate production in the Bromide was periodically interrupted by siliciclastic sediment supply from the north and east.

The Bromide Formation is subdivided into three members and one submember.  The lowest unit is the Pontotoc Member (Carlucci et al. 2014), a massive, thick-bedded or rippled unit of shoreface sandstone.  The next unit (Mountain Lake Member) is a succession of quartz sandstone, interbedded sandstone and illitic-chloritic shale, and shale and limestone (mostly pelmatozoan packstone and grainstone).  The basal shaly interval above the Pontotoc Member comprises the "Lower Echinoderm" Zone of Sprinkle (1982), which is widely noted for high diversity assemblages of echinoderms, including crinoids, (hybocrinids, camerates, *Cleiocrinus*), paracrinoids, rhombiferan cystoids; these echinoderms have been well documented (Sprinkle, 1982).

The mixed carbonate-siliciclastic units are overlain by the Pooleville Member, a thick-bedded to massive unit of carbonate mud-, wacke-, and packstone.  The basal Pooleville is marked by a thick-bedded echinodermal grainstone, overlain by a shale and packstone rich interval, the "upper Echinoderm Zone" of Sprinkle (1982). The upper Pooleville is set off as a lithologically distinct submember called the Corbin Ranch, which is a fenestral, commonly dolomitic limestone that forms peritidal cycles just below the contact with the Viola Springs Formation.  Figure 15 summarizes the range of facies and inferred depositional environments in all units of the Bromide Formation.

*Biostratigraphy and Age of the Bromide*

Stop 7 is the newly established standard reference section (Carlucci et al. 2014) for correlation and discussion of depositional sequences in the Bromide Formation.  This outcrop is well preserved, nearly complete (except for the Pontotoc Member), and has previously been well-studied biostratigraphically (e.g., Bauer 1994).  Bauer (1994) suggested that the Bromide is Sandbian age (Fig. 3, late Whiterockian to Mohawkian of classical North American nomenclature) by comparison of the conodont fauna with the graphic correlation standards of Sweet (1984) and Bergström (1983). In the lower Mountain Lake, *Cahabagnathus sweeti* is indicative of the late Whiterockian, and is replaced by the Mohawkian conodonts *Baltoniodus gerdae* and *Eoplacognathus elongatus* in the Mountain Lake.  Bauer (1994) suggested that the conodont fauna of the intertidal Corbin Ranch submember includes forms that range from Blackriveran (Whiterockian) to Mohawkian, and could not constrain the age of the unit.  He

(1994, table 2) also noted that *C. sweeti* showed a strong proclivity for shallower water environments, while *B. gerdae* apparently favored deeper water environments. Such facies control on distribution has important implications for conodont-based correlation of the sections. Although *B. gerdae* occurs through the entire section down-ramp at RC in the Criner Hills, it is restricted to just four meters of the deepest portion of the upper Mountain Lake up-ramp at I-35N (Bauer 1994, table 1). The narrower stratigraphic distribution up-ramp is exactly what would be predicted from a hypothesis of facies control, so that differences in the first appearance of *B. gerdae* cannot be used as evidence that the deeper facies are stratigraphically younger down-ramp.

Independent estimates of the age of the upper Bromide strata in the Criner Hills (stop 12) relative to I-35N are inconclusive. Decker (1935, 1941) and Finney (1986) showed that graptolites from the upper Bromide (*Amplexograptus maxwelli*, *Dicellograptus flexuosus*) are diagnostic of the British *D. multidens* Zone. The *D. multidens* zone spans the Deicke, Millbrig, and Kinnekulle bentonites, and consequently, its duration is too long to provide a biostratigraphic test of the absence of Pooleville Member down-ramp (see stops 11 and 12 for more explantion). Interestingly, both Bauer (1994) and Rosenau et al. (2012) found the upper Sandbian condont, *Phragmodus undatus*, in the Pooleville at HWY 99 (far up-ramp, stop 4), but it has not been recorded from the down-ramp "Poolville" of the Criner Hills (stop 12). Although this could be yet another example of facies control, it might instead provide evidence that the "Pooleville" within the SOA has been miscorrelated with the true Pooleville, which is missing down-ramp.

The composite section comprises the north and south sides of I-35, and the median, which together expose a nearly complete succession of Bromide units. The completeness and preservation of the I-35N section allows it to serve as a model to characterize cyclicity at meter and decameter scales (Fig. 15) and correlate with other sections up- and down-ramp (Fig. 17).

### *Sequence Stratigraphy and Cyclicity*

Meter scale cycles in the Bromide Formation (Figs. 16, 18) consist of alternating small-scale or "micro"-TSTs and HSTs (meter-scale analogues of systems tracts) that form packages of condensed and more rapidly accumulated deposits (cf. Brett and Algeo 2001; Brett et al. 2008). In the Mountain Lake Member, micro-TSTs are amalgamated, nodular to tabular beds of poorly sorted bryozoan-echinoderm grainstone. Bases of micro-TSTs are marked by sharp facies offsets and scouring (merged transgressive/high-frequency sequence boundary surfaces), whereas flooding surfaces at their tops commonly include authigenic mineral crusts (iron oxide and pyrite) or a heavily bioturbated firmground (e.g., McLaughlin et al. 2008). HSTs are mostly shale, interbedded with thin nodular bryozoan and echinoderm packstone and grainstone.

**TEXT-FIGURE 15.—Facies descriptions, fossil content, and depositional interpretation of Bromide Formation lithologies (modified from Carlucci et al. 2014).**

| Facies | Description | Fossil content | Interpretation |
|---|---|---|---|
| Fenestral and Microbial Laminated Mudstone | White to dark grey, dense mudstone with fenestral cavities (Fig. 5d), microbial laminations, desiccation cracks, vertical tubes and oxidation marks. Cycles of well-bedded fenestral mudstone and thinner bedded rubbly wackestone that lack fenestrae are common. Wackestone beds have mm to cm scale shaly interbeds | Ostracods (e.g., *Leperditella*, *Ningulella*, *Ctenobolbina*), corals (e.g., *Tetradium*), conodonts, brachiopods (e.g., *Ancistrorhyncha*), gastropods, and rarely trilobites (e.g., *Bathyurus*) | Low energy intertidal to supratidal tidal flat |
| Lime Mudstone | Massive, has some small voids filled with calcite cement and contains well-preserved ostracod fossils | Ostracods | Restricted inner-ramp deposit likely from a shallow lagoon or tidal flat |
| Massive to Crossbedded Sandstone Facies | Detrital, fine to medium grained, well-sorted, well-rounded, frosted quartz arenite grains, with small percentages of clay minerals, feldspar and other rock fragments. Wave ripples, hummocky cross-beds, minor siltstone | Bryozoan and echinoderm | Upper shoreface to lower shoreface |
| Massive to Nodular Lime Mudstone | Medium to dark, grey, blue, or tan lime mudstone, massive or nodular bedded, peloidal (Fig. 3c), and heavily bioturbated. Recrystallization to microspar common in burrows and mottled features. Mudstone often interbedded with fossiliferous and thinner bedded wackestone | Ostracods, gastropods, tabulate corals (*Tetradium*, Fig. 5a), receptaculitids, and brachiopods | Shallow subtidal lagoon or protected inner-ramp |
| Interbedded Sandstone and Shale Facies | Detrital, fine grained, well-sorted, well-rounded, frosted quartz arenite grains, with shale interbeds. Abundant silt and bioclasts | Bryozoans and echinoderms | Transitional between shoreface and offshore deposits (inner-ramp), but likely still receiving sand from the lower shoreface |
| Rippled Calcisiltite and Grainstone | Couplets of white, thin-bedded (~2cm) wave rippled calcisiltite and medium bedded packstone or red-hued very coarse grainstone (Fig. 4a). | Pelmatozoans | High energy, detached shoreface deposited on a wave-dominated inner-ramp, during forced regression |
| Interbedded Grainstone | Thin to medium-bedded (cm scale), poorly sorted, lenticular grainstone (Fig. 4e) and mudstone interbedded with green to tan fissile shale. Articulated bryozoan horizons, some minor packstone | Crinoids, cystoids (e.g., *Diabolocrinus*, *Oklahomacystis*, *Pirocystella*), ramose and massive bryozoans (e.g., *Batostoma*, *Diamultites*), brachiopods (e.g., *Sowerbyella*, *Bellimurina*, *Valcourea*), and receptaculitids | At or below fair weather wave base, basinward of the lower shoreface but likely represent a broad bathymetric range, from upper middle-ramp to upper outer-ramp |
| Interbedded Grainstone, Rudstone and Shale | | Rudstone: brachiopods (e.g., *Sowerbyella*, *Strophomena*, *Plectoglossa*), trilobites (e.g., *Calyptaulax*, *Bumastoides*, *Thaleops*), ramose bryozoans, and cystoids | Subtidal, moderate energy; mid-ramp environment between storm and fair weather wave base, not subject to regular wave agitation |
| Thin Phosphatic Grainstone and Packstone | Below-wave base laminated mudstone and shale with a thin packstone unit at the base overlain by fine-grained bioclastic grainstone (Fig. 3f). Multiple irregular surfaces (Fig.5e) | Packstone: coral, gastropod and echinoderm debris; grainstone: diploographid graptolites and trilobites (e.g., *Cryptolithus*) | Thin transgressive lag that records the landward shift of the shoreline followed by inner-ramp shoal environments scoured by wave action |
| Bioclastic Grainstone and Packstone | Medium to dark, blue to grey, medium-bedded to thick-bedded or massive beds of rubbly, fossiliferous, argillaceous limestone. Packstone often interbedded with grainstone (Fig. 4b). Wavy or rubbly grainstone interbedded with planar beds. Authigenic mineral crusts of pyrite and iron oxide, condensation features | Trilobites (*Anataphrus*, *Failleana*), ramose bryozoans, and echinoderm debris (Fig. 3d) | High energy, inner-ramp shoal and mid-ramp packstone, subject to constant wave agitation, or frequent storm reworking (above storm wave base) |
| Wackestone with Rudstone Pavements | Light blue-grey, medium to thick-bedded, slightly nodular, tan-weathering wackestone. Packstone locally present. Beds are often capped by fossiliferous rudstone pavements (Fig. 5c) | Rudstone: brachiopods (e.g., *Sowerbyella*, *Strophomena*, *Plectoglossa*), trilobites (e.g., *Calyptaulax*, *Bumastoides*, *Thaleops*), ramose bryozoans, and cystoids | Between fair-weather and storm weather wave base in lower middle-ramp to outer-ramp |
| Packstone and Wackestone | Fossiliferous, amalgamated wackestone or packstone beds that are medium to coarse grained, with shales either absent or highly indurated. Condensed relative to F12. | Trilobites, strophomenid brachiopods, ramose bryozoans, and gastropods | Between fair-weather and storm weather wave base in lower middle-ramp to outer-ramp |
| Rhythmic Mudstone, Wackestone and Shale | Couplets of thin (3-10 cm), tan to grey, planar to nodular bedded mudstone and wackestone (Fig. 4c, d), with argillaceous sublaminated shale of varying thickness. Tempesites and obrution horizons common. Expands into the basin during sequence 2, correlates across study area. | Strophomenid brachiopods, trilobites (e.g., *Isotelus*, *Thaleops*, *Calyptaulax*, *Remopleuridies*), straight cephalopods, receptaculitids, echinoderms, gastropods (e.g., *Cyclonema*, *Eotomaria*) and ostracods (e.g., *Anisocyamus*, *Bromidella*). | Outer-ramp to basin, low energy, dysoxic, and typically below storm level wave base |
| Thin bedded Mudstone and Shale | Thin-bedded (~5cm), fine-grained lime mudstone, with very thin partings of shale and marls | Trilobites (e.g., *Lonchodomas*), straight cephalopods, graptolites and inarticulate brachiopods | |
| Fissile Illitic-Chloritic Shale | Dark greenish-grey, fissile, clay-rich, sublaminated green shale that weathers to a tan color. Clay component rich in illite-smectite and chlorite. | Echinoderms | Deltaic-fluvial fines transported onto the ramp and deposited below the action of storm waves. Upper occurrence represents a lowstand package of siliclastics that directly overlies a major sequence boundary |
| Laminated Mudstone and Shale | White to light grey lime mudstone interbedded with a fissile to marly shale. Mudstone fossiliferous, thinly laminated (Fig. 5f) and evenly bedded. Shale interbeds thicken upwards from 1 cm to 5 cm above the base of the Viola contact. | Diplograptid graptolites and trilobites (e.g., *Flexicalymene*, *Cryptolithus*) | Outer-ramp to basin environment below storm level wave base. Marks a major change in the trilobite fauna that characterizes sequence M5 of Holland and Patzkowsky (1996) elsewhere in eastern Laurentia, so that the ravinement surface beneath F8 is likely the M4/M5 sequence boundary |

# Meter-scale cycles
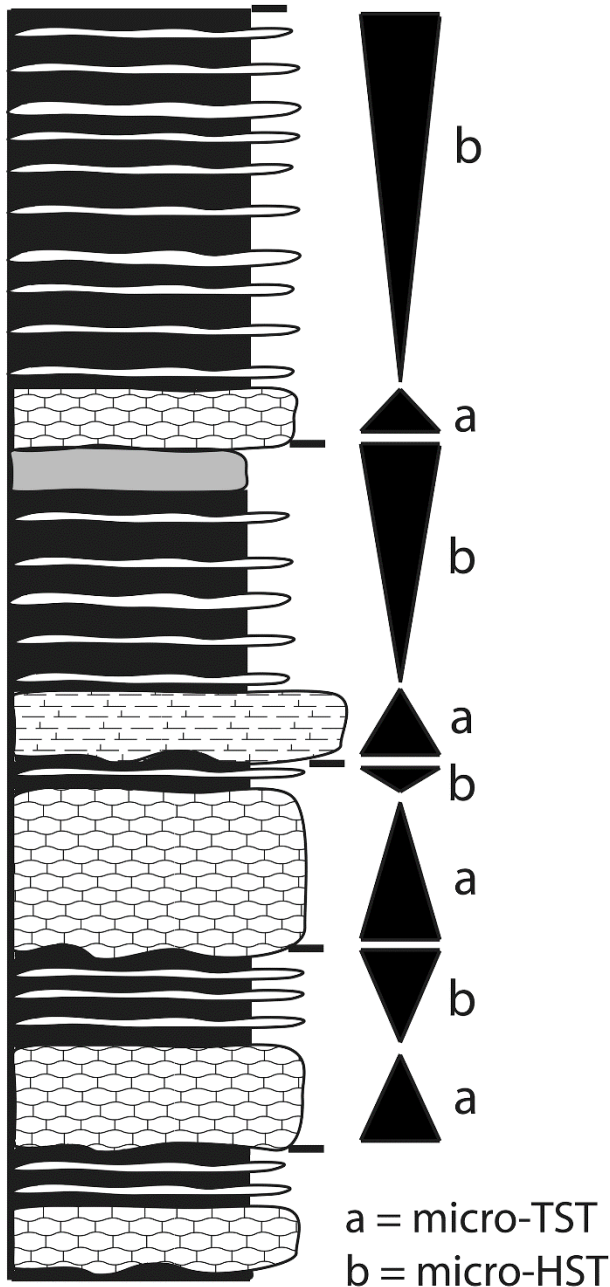


a = micro-TST
b = micro-HST

In the Pooleville Member, micro-TSTs are indurated wackestone and packstone and are shale-poor, show evidence of condensation upwards, poor sorting, pyritization of bioclasts, and flooding surfaces that are heavily stained with iron-oxide and pyritic or phosphatic mineral crusts. These mineralized surfaces also represent the highest degree of sediment starvation in the micro-sequence (maximum sediment starvation surfaces of Baum and Vail 1988; Brett et al. 2004) and are overlain by thinner, late TST shell hashes and/or micro-HSTs of aggradational to progradational, interbedded shale and limestone rhythms. Micro-HSTs are thinly and planar bedded near their bases, and gradually become more nodular as they decrease in siliciclastic content. Low-rank transgressive surfaces/high-frequency sequence boundaries sharply overlie the micro-HSTs, and are commonly irregular erosive disconformities.

At a larger scale of accommodation space dynamics, Carlucci et al. (2014) documented three distinct, 3rd order depositional sequences that record the transition from siliciclastic to carbonate dominance during Bromide deposition. Portions of sequence 1 and the entirety of sequences 2 and 3 are exposed at stop 7. Sequence 1 is likely the last major depositional sequence of the Whiterockian (Dariwillian), and consists of massive, cross-bedded, and/or rippled sandstones (Pontotoc Member) that grade abruptly upward into sandy limestones, in what Carlucci et al. (2014) considered to be a transition from lowstand (LST) to early transgressive (TST) conditions.
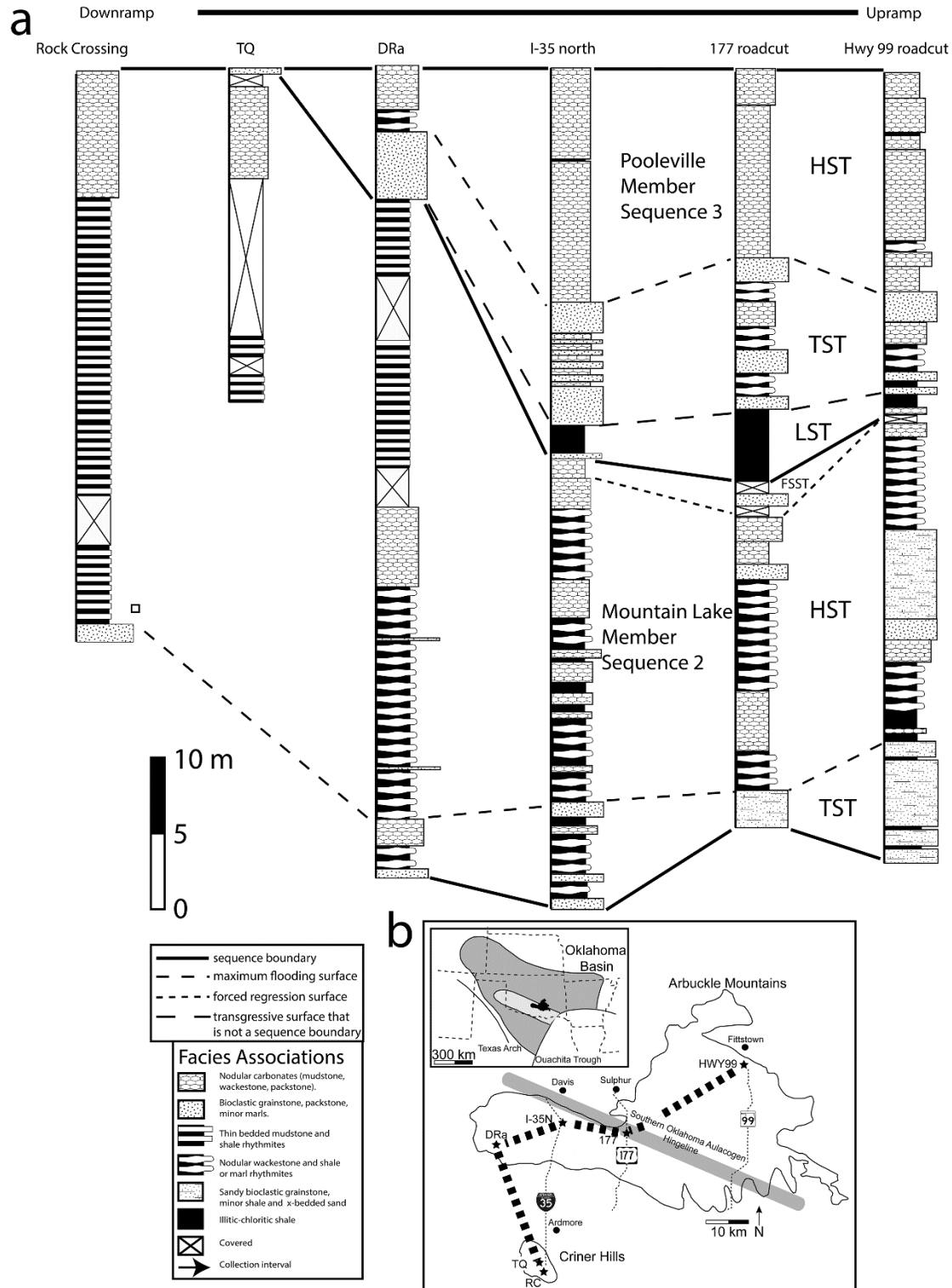
**TEXT-FIGURE 17.—Simplified summary of facies associations and depositional sequence architecture of the Bromide Formation (modified from Carlucci et al. 2014: Carlucci and Westrop 2015). a, correlation of stratigraphic columns based on sequence stratigraphic surfaces. b, Arbuckle Mountains map showing line of cross section and localities.**

**TEXT-FIGURE 18.—Outcrop example (stop 7) of meter scale cycles in the Mountain Lake Member of the Bromide Formation. See Figure 16 for interpretation. GPS: 34°26'9.21"N; 97° 7'46.88"W.**

Coarse, sandy bioclastic packstones and grainstones with condensation features represent TST deposition, and these are overlain at a maximum flooding surface by HST units of interbedded green fossiliferous shales ("Lower Echinoderm Zone"), rudstones, and grainstones. The sequence boundary is a sharp contact between the interbedded green shale and marly packstone, and a rippled and crossbedded, sandy crinoidal grainstone.

Complete description of the depositional architecture of sequences 2 and 3 is beyond the scope of this field guide; however, Figure 17 shows the facies associations contained within depositional sequences 2 and 3. These facies associations and the corresponding bounding surfaces are well exposed at I-35N, and participants can walk south along the highway and study the sequences. In brief, sequence 2 is composed of a basal sandy crinoidal grainstone and trilobite packstone succession interpreted as TST deposits. These are overlain by thick, marly limestone-shale rhythmites (HST) whose component meter scale cycles shallow upwards. Rapid facies change to a rippled calcisiltite and grainstone unit likely represents an interval of forced

regression, just prior to a major sequence boundary below the uppermost green shale of the I-35 outcrop. Sequence 3 deposition starts with an input of siliciclastic fines ("the big green shale" or Upper Echinoderm Zone) that likely formed as the coastline was exposed during sea level lowstand. These are overlain by a fossiliferous unit of rubbly, phosphatic packstones and grainstones with hematite-pyrite mineral crusts. This facies is interpreted as a carbonate shoal deposit that formed during a major marine transgression, and was mapped across the Arbuckles by Carlucci et al. 2014 (Fig. 17). Above the maximum flooding surface of the sequence 3 TST, predominantly carbonate mud- and wackestone deposition show a gradual shallowing upward trend to the fenestral beds of the Corbin Ranch Submember. The sequence boundary with the Viola Springs Formation (Sandbian-Katian boundary) directly overlies the fenestral beds at this locality.

Note that the thin sequence at the base of the Viola Springs Formation ("member 1" of the stop 4 discussion) is not present at the Bromide/Viola contact at stop 7, likely due to erosional truncation at the Bromide/Viola Springs unconformity.

**Stop 8: I-35S, Viola Springs and Welling Formation**

The I-35 South section comprises a pair of road cuts on the south flank of the Arbuckle Mountains that expose the uppermost Bromide, much of the Viola Springs, and the Welling formations. Amati (2014) recently published locality information, including a lithologic log of this section. The Bromide is represented by only a few meters of subtidal facies that includes massive wackestones with brachiopod and gastropod shell beds that likely record storm-influenced deposition. Cross sections through receptaculitids are commonly encountered in the uppermost beds. The peritidal facies of the Corbin Ranch submember, reduced to only a thin remnant at the I35N section relative to the type area near Fittstown (stop 4), is entirely missing at I-35S.

The Viola Springs Formation is exposed in two segments separated by a covered interval of about 75 m, although it is more complete on the north-bound lane. The base of the formation is a major unconformity marked by an irregular mineralized, rusty weathering surface that cuts out much of HST of Sequence 3 (Corbin Ranch, upper Pooleville Member). The discussion of the stratigraphy and facies follows Amati's (2014) description of the section. The basal two meters of the formation is a white-weathering ostracode-rich wackestone-packstone with interbedded black graptolite-bearing calcareous shale, and capped by another planar, mineralized surface. The succeeding 25 m is mostly thin-bedded, plane-laminated, cherty, lime mudstone with thin, dark gray, marly interlayers; bioturbation is variable, but generally low. Bioturbation increases higher in the section, and the lithology switches to burrow-homogenized wackestone to packstone with marly layers at about 55 m; debris of cryptolithine and isoteline trilobites, as well as graptolites can be seen commonly in the marly layers, although specimens are not easily collected.

Above the covered interval, the upper part of the section includes thicker-beddded limestone that forms meter-scale cycles with intervals of dark, strongly bioturbated lime

mudstone alternating with crinoidal grainstone. The section is capped by about 10 m of massive crinoidal grainstone of the Welling Formation. According to Amati (2014), the contact between the Viola Springs and Welling is gradational over about a meter of section.

### *Biostratigraphy*

*Graptolites*. Finney (1986, fig. 7) published a range chart (his section U) for graptolites. Most of his collections were made from the lower segment of the section. These show that the lower 20 m is characterized by an assemblage that includes *Diplacanthograptus spiniferus*, *Orthoretiolites hami* and *Geniculograptus typicalis*, taxa that indicate these beds belong to the *D. spiniferus* Zone. However, recent sampling by Goldman (unpublished) of the basal meter of the Viola Springs at this locality yields an assemblage with, among others, *Diplacanthograptus caudatus*, *Neurograptus margaritatus*, *Normalograptus brevis*, and *Dicranograptus hians*, which he interprets as earliest Katian (*D. caudatus* Zone). The presence of *D. caudatus* and *Dicr*anograptus *hians* indicates a correlation with the Australasian Stage Ea1, which is pre- *D. spiniferus* Zone. There is one juvenile specimen in the 3.0 – 6.0 cm collection that resembles *D. spiniferus*, but the identification is uncertain. Hence the lowermost Viola Springs Formation at the I-35 section is similar in age to its base at the Mountain Lake section, which also yields *D. caudatus* Zone graptolites (Finney, 1986).

The base of the *Geniculograptus pygmaeus* Zone is marked by the appearance of the eponymous species at 32 m above the base of the Viola Springs (Finney, 1986); this boundary correlates into sequence C1 of the Cincinnati region (Brett et al., 2004). Finney (1986, fig. 7) does not record any other species entering the succession above *G. pygmaeus*, although a few species extend from the lower to the upper segment of the section (e.g., *Orthograptus quadrimucronatus*).

*Conodonts*. Conodont data are limited for I35 South, but Bergström et al. (2010) recently showed the distribution of zones, albeit without any species range data. They placed the base of the Viola Springs within the *Belodina confluens* Zone, confirming the conclusions of Carlucci et al. (2014) that there is significant unconformity with the underlying Bromide Formation. The upper part of the formation is assigned to the *Amorphognathus superbus* Zone; the Welling Formation apparently lies in the overlying *A. ordovicicus* Zone.

On the basis of the conodont biostratigraphy, Bergström et al. (2010) concluded that a positive isotope excursion at the base of the Viola Springs at I-35S represented the "Kope excursion" of the Cincinnati region. They interpreted the excursion at the HW 99 section as the GICE, with the implication that this section is entirely older than the I-35S section (Berström et al., 2010, fig. 4). However, the trilobite biostratigraphy of HW 99 indicates that the interval containing the so-called "GICE" is younger than previously thought, and the positive excursion may well be the same excursion identified as the "Kope" at I-35S (Westrop et al., 2012). The graptolite data do not conflict with this hypothesis, which implies an overlap of about 25 m between the sections.

**DAY 4**
**Stop 9:  Oil Creek Formation (if time allows)**

This locality is an exposure through the 2$^{nd}$ oldest unit of the Simpson Group, the Oil Creek Formation.  The Oil Creek is well known in Oklahoma for its economic impact and industrial importance.  The basal unit of the Oil Creek is a thick (~ 350 ft, Ham 1973) sandstone that is comprised of well-sorted and rounded quartz sand.  This sand is over 99% pure silica, and companies such as the Pennsylvania Glass Sand Corporation (now known as U.S. Silica Corp) have been selling it for use in glass-making and as silica flour for over 90 years. Quartz sand in the lower Oil Creek is unique because there are few pressure solution contacts, quartz overgrowths, and only loose illite cement at grain boundaries (Ham 1973).  For these reasons, the sand is easily disarticulated in solution and processed into glass.  The cleanly washed and high porosity sands of the Oil Creek are also a prolific oil reservoir in southern Oklahoma, and have been extensively mapped into the subsurface.

The exposure of the Oil Creek at stop 9 is from the upper unit, which is primarily composed of sandy packstone, grainstone, and shale.  The transition from the lower unit is gradational, with bioclasts gradually becoming a more common framework component.  The sandy limestone facies as exposed here is heavily bioturbated and extremely fossiliferous, and contains a number of hardgrounds encrusted by bryozoan and echinoderm holdfasts, which are indicative of slowing rates of deposition relative to the lower sandstone unit.  Limestones are exceptionally fossiliferous and contain a diverse fauna of brachiopods, bryozoans, gastropods, nautiloids, and trilobites; cystoid plates are also common on some bedding planes. The lowest beds of the limestone unit contain stromatolitic boundstone and sand filled desiccation cracks (Brown 2003), and may record undocumented bounding surfaces that better document the nature of the transition.  Like the Bromide Formation (see stop 7) and other units in the Simpson Group, deposition in the Oil Creek evolves from primarily siliciclastic to mostly carbonate.  Ham (1973) stated that ascertaining the depositional environment of the Oil Creek was problematic because of the lack of diagnostic sedimentary structures.  One possibility is that the lower unit represents shoreface deposits with input from a regional desert on the Arbuckle Platform (Longman 1982), with limestone clasts becoming incorporated during a large-scale transgression.  Carlucci et al. (2014) noted that this type of model might be problematic for some units of the Simpson Group, because many of the sands thicken toward the SOA, as would be expected during lowstand normal regression.  The Oil Creek sands are thick and extensive mostly in eastern Oklahoma, though they do form a wedge toward the SOA (Schramm 1964, fig. 8), which might be consist with lowstand progradation.  In this scenario, it seems likely that wedge development during lowstand conditions was followed by a transgression at some point that allowed for the preservation of open marine fossils.

**Stop 10: Structural Features in the Fort Sill Limestone**

The Fort Sill Limestone (Fig. 19) is the lowest unit in the Arbuckle Group (Cambrian; Sunwaptan), primarily distinguished from the underlying Honey Creek Limestone based on the

**TEXT-FIGURE 19.—Folding in the Fort Sill Limestone, stop 10 (GPS: 34°24'29.76''N, 97° 8'20.85''W).**

prevalence of lime mud deposition in the Arbuckle Group (Ragland and Donovan 1991), and the loss of glauconite and siliciclastic sand. Therefore, the early Arbuckle Group represents a shift in depositional style, where the carbonate platform was able to retain most of the mud that its carbonate factory produced. Similar to other carbonate deposits of the Arbuckle Group, the Fort Sill was deposited in a shallow, low gradient epeiric sea. During the Cambrian and into the Early Ordovician, margins were passive with high rates of post-rift subsidence, and there was little relief in the lowland areas of Laurentia (Ragland and Donovan 1991). This unique setting allowed for accumulation of the incredibly thick deposits of tropical carbonates that characterize the Arbuckle Group.

The Fort Sill Limestone can be subdivided into three informal lithologic groups: a lower unit of thin bedded stromatolitic mudstones, middle unit of dolomitic siltstone-lime mudstone rhythmites, and an upper unit of massive algal boundstones (Ragland and Donovan 1991). The upper two lithologies are expressed at stop 10, but this locality is much better known for its structural features.

Along the north and south sides of I-35, the Fort Sill Limestone at stop 10 contains a series of fold sets that are offset by thrust faults. The fold sets take a variety of shapes, most commonly as box or polyclinal (sub-parallel hinge lines but non-parallel axial surfaces) folds (Tapp 1991). The folds at stop 10 all show a flexural-slip style of folding, where volume accommodation is preserved by layer-parallel slip between layers, rather than layer parallel flow ("flexural flow"). The folds at stop 10 show vergence to the north, and are generally considered to be evidence of compressional forces causing the Arbuckle uplift (Brown 1984: Tapp 1991) rather than strike-slip deformation (Wickham and Denison 1978). The structural features at the outcrop have been interpreted as part of a flower structure associated with the Chapman Ranch Fault (exposed near Turner Falls), part of the larger compressional forces of the Arbuckle Uplift (Brown et al. 1985), or as expression of volume accommodation in a larger-scale flexural-slip fold (Tapp 1991).

**Stop 11: Mountain Lake and Daube Ranch (DRa), Tulip Creek, Bromide, and Viola Springs Formation**

Ulrich (1911) first used the term "Bromide" for a unit that lay unconformably below the "Viola Limestone", but he did not propose any type locality or section. Edson (1927) proposed that the type locality of the Bromide be placed at the McLish Ranch, in the town of Bromide, Coal County, Oklahoma. The type sections of the Mountain Lake and Pooleville members of the Bromide as proposed by Cooper (1956) were placed at his Johnston Ranch locality, which is now owned by Sam Daube (referred to as DRa by Carlucci et al. 2014). Cooper defined the Mountain Lake to include a basal fine-grained quartz arenite (now called the Pontotoc Member), an overlying interbedded illitic-chloritic shale and sandstone, and an uppermost limestone and shale sequence at the top (Fay et al. 1982b). Cooper confined the Pooleville to the various limestones (occasionally interbedded with calcareous shales and marls) above the Mountain Lake, in particular near the SOA axis. Stop 11 at the DRa location is the type section of the Mountain Lake and Pooleville Members.

The section at stop 11 begins in the Tulip Creek Formation (Fig. 3) of the Simpson Group, exposed down-section from the Upper Humble Lake (a reservoir), with progressively younger strata towards the dam. Thick packages of green shale interbedded with massive limestones near the base of the section record the upper portion of the Tulip Creek, below its contact with the Pontotoc Member of the Bromide. Lower parts of the section may be overgrown during the summer, so visibility of the Pontotoc and lower Mountain Lake Members might be limited.

In the middle Mountain Lake at DRa, there is an obvious expansion of the thickness of sequence 2 (Fig.17), in both the illitic-chloritic green shale, and shale-limestone rhythmite associations. A particularly thick, trench-forming shale interval is prominent at DRa (not shown on Fig. 17), and is interpreted as the lower green shale (lower Echinoderm Zone of Sprinkle 1982) of the Bromide Formation, which correlates up-ramp into the lower Mountain Lake

(Carlucci et al. 2014), rather than the Pooleville as done previously. This correlation is not intuitive at first, because there are lower shales at DRa that are poorly exposed (Fay et al. 1982b), but they are likely part of the lower sequence 1, which is dominated by grainstone and shales at all the Bromide sections.

Longman (1982) stated that the putative basinal deposits of the ''lower Pooleville'' within the SOA (e.g., at sections DRa, RC, stops 11 and 12) and those of the upper Mountain Lake along the hingeline (e.g., I-35 N, stop 7) shared the same characteristic bedding (compare Figs. 18 and 20). Evidence discussed by Carlucci et al. 2014 suggests that they are similar because the supposed "lower Pooleville" strata of DRa and RC are, in fact, correlative with the middle to upper Mountain Lake elsewhere. This hypothesis demands that the down-ramp expansion of the Bromide thickness (sequence 2, Fig. 17) applies only to the siliciclastic-rich Mountain Lake Member, whereas the predominantly carbonate Pooleville Member (sequence 3) is largely missing at these sections owing to erosional truncation. Evidence for down-ramp erosion of the Pooleville Member is shown on this field trip at stops 4, 7, 8, 10, and 11. There is an obvious truncation of the facies in our transect (partially shown here in Fig. 17), at the Bromide/Viola unconformity and down into sequence 3. "Member 1" of the Viola is removed from the basin between stop 4 and 7, the Corbin Ranch is removed between 7 and 8, and then the Pooleville is progressively removed into the aulacogen in the Criner Hills, from stops 8 to 10, and 11. The true Pooleville Member at DRa is only preserved on the other side of the Upper Humble Lake, and this correlation is further evidenced by the widespread, mappable grainstone that is always present at the base of sequence 3 (exposed at DRa along the ground in the wooded area by the dam). Indeed, this condensed grainstone unit is so consistently cut out from the basin that a thin wedge of it is preserved at the TQ locality (see Carlucci et al. 2014), just below the Bromide-Viola unconformity. Additional paleontological evidence includes closely comparable strophomenid brachiopod beds, horizons with straight cephalopods, and identical species of trilobites in the rhythmite packages in the I-35 N, DRa and RC sections (Carlucci and Westrop 2014).

Karim and Westrop (2002) described the taphonomy of the well-known '*Homotelus*' beds (*Vogdesia*) from the TQ locality in the Criner Hills. These same beds, previously referred to as belonging to the Pooleville Member in the Criner Hills, are exposed at stop 11 (Fig. 20) in a trench-forming limestone-shale rhythmite that overlies the green shale unit, and which clearly lies low within the Mountain Lake. These assemblages of articulated exoskeletons likely recorded behavioral aggregations that were preserved beneath storm-influenced 'mud dumps' (Karim and Westrop 2002; Brett et al. 2012). These obrution beds occur most commonly in the early HST of sequence 2 across the Bromide Formation. When siliciclastics are no longer sequestered near the coastline during early HST, rapid deposition of mud layers in mixed carbonate– siliciclastic systems leads to exceptional preservation. In the rhythmite-dominated HST deposits, sedimentary structures in the limestone beds include burrow-mottled fabrics, *Chondrites*, disrupted laminae, vertical burrows and strophomenid brachiopods and trilobites (including articulated *Vogdesia*, *Thaleops*, *Calyptaulax*, *Remopleurides*) that formed as obrution
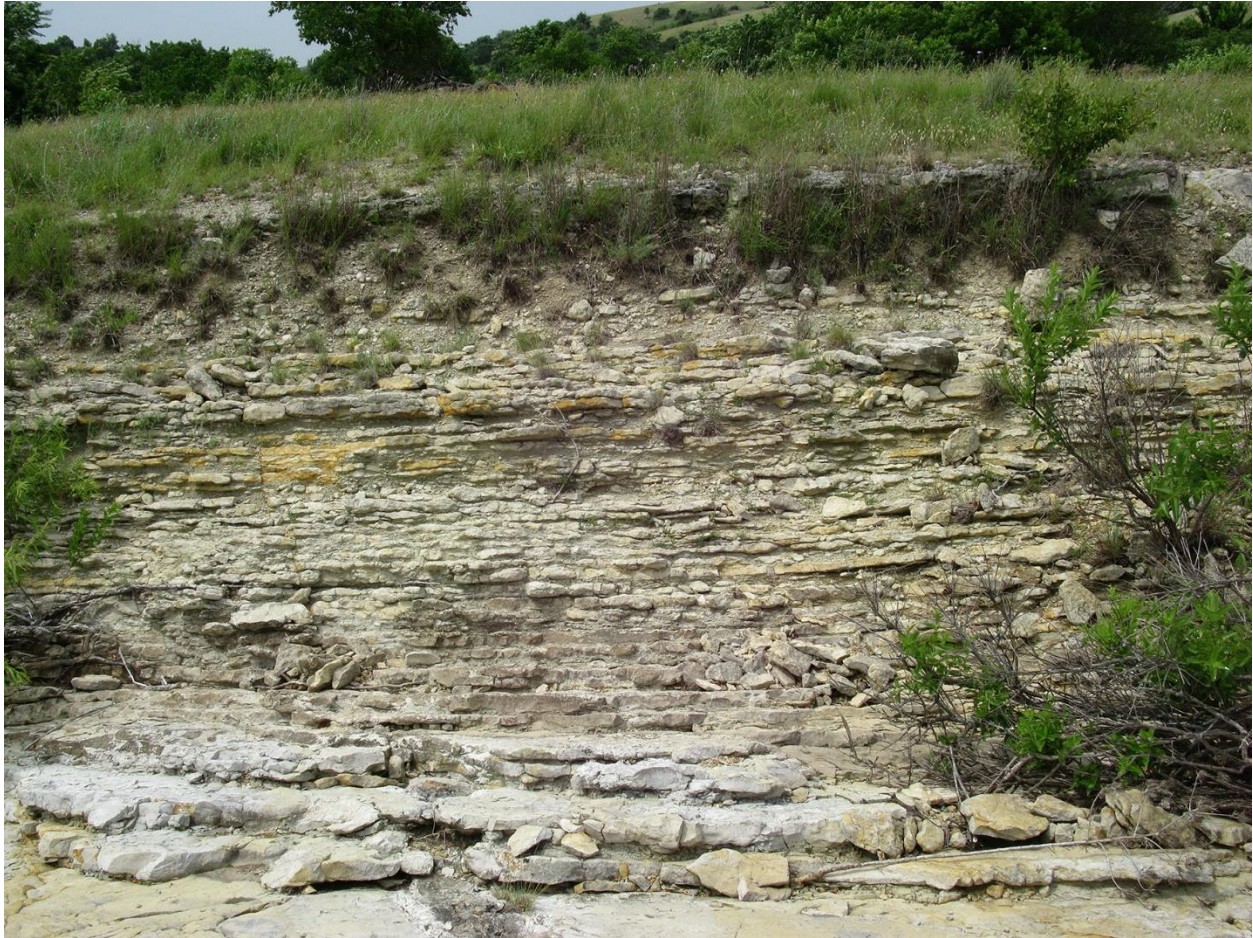
**TEXT-FIGURE 20.--Outcrop photograph from the DRa locality, showing limestone-shale rhythmites in the unit that produces the 'Homotelus' (*Vogdesia*) beds, interpreted as a behavioral aggregation smothered during a mud-dump event.  GPS: 34°21'56.66"N, 97°17'7.24"W.**

horizons below storm wave base (Karim and Westrop 2002). Trilobites in the sequence 2 obrution horizons lack epibionts as noted by Karim and Westrop (2002), whereas disarticulated specimens in shell pavements elsewhere are usually encrusted by various bryozoans and Cornulites tubes. This suggests the obrution-derived HST deposits were not exposed at the surface for long periods of time, and were probably smothered in place by mud blanketing events (Brett et al. 2012).

## Stop 12: Rock Crossing, Criner Hills Region, Bromide Formation

In the Criner Hills of southern Oklahoma, cuts along Hickory Creek (stop 12) are well-known for producing important trilobite fossils, including the holotypes of *Lonchodomas mcgeheei* (Decker 1931; Sutherland and Amsden 1953) and *Probolichas kristiae* (Carlucci et al. 2012).  Rock Crossing has a long history of study (e.g., Decker 1931; Decker and Merrit 1931; Sutherland and Amsden 1959; Longman 1982a, b; Fay et al. 1982; Carlucci et al. 2014), and

plays a pivotal role in any interpretation of Simpson Group strata because it lies southward of the inferred SOA basin axis.

East of D3265 Road, there is an exposure of rock that forms the creek bed around a meander of the Hickory Creek. The strata are assigned to the lower Bromide Formation (Mountain Lake) and correspond to sequence 1 of Carlucci et al. (2014). There is a hogback along the creek bank, developed in the Pontotoc Sandstone downstream of a waterfall that exposes the overlying sequence of packstone/grainstone, green shales, and bryozoan-rich beds (sequence 1 HST, sequence 2 TST of Carlucci et al. 2014). This succession is similar to that of the reference section at I-35N. Above these units at the top of the waterfall is the same limestone-shale rhythmite package ('*Homotelus* beds') that is exposed at stop 11. Trilobites such as *Vogdesia* are somewhat more difficult to find at this locality, but the unit is rich in large, articulated straight cephalopods, brachiopods, and other trilobites. Note how much thinner the interval between the basal sandstones and *Homotelus* beds is at this locality compared to stop 11.

Up-section towards the bridge, additional rhythmite packages are exposed, including the thin bedded, *Lonchodomas*-rich (Fig. 21) facies that is finely internally laminated and clearly formed below storm wave base (see Fig. 15). Facies form a shallowing succession towards the bridge, becoming more nodular and fossiliferous, before abruptly ending without any evidence of "normal" Pooleville deposition below the Viola contact. Carlucci et al. (2014) took this is as additional evidence for southward truncation of the M4/M5 sequence boundary, which is somewhat counterintuitive as the magnitude of the truncation appears to increase into the SOA basin. However, the pattern is extremely consistent from stops 4-7-8-10-11. First, "member 1" of the Viola is removed, then the Corbin Ranch, then part of the Pooleville, and then all of the Pooleville in the Criner Hills. Down-ramp facies change does not account for this pattern, because the same facies and stacking patterns typical of the Mountain Lake are consistently developed across the sections, with those at the top missing progressively southward, at the unconformity. The Viola Springs was deposited at the onset of a major tectonic phase of the Taconic Orogeny (Pope and Read 1997), and it is possible that far-field tectonics produced an inversion of topography of the SOA after the deposition of the Corbin Ranch submember, when shallowing was apparently still to the north, and prior to the start of Viola Springs deposition. This scenario is similar to the one proposed by Finney (1986, fig. 19) to explain an apparent earlier onset of Viola Springs deposition at the HWY 99 section, and we suggest that the uplifted southern region may have been beveled during a prolonged period of sea-level lowstand associated with the M4–M5 sequence boundary, which is a major break elsewhere in eastern Laurentia (e.g., Holland and Patzkowsky 1996). An alternative model for the removal of an older carbonate unit in a down-ramp direction is rapid subsidence of the Criner Hills region in association with pre-Viola Springs tectonics. Under this scenario, abrupt down warping of the strata into a corrosive environment below the pycnocline could possibly result in dissolution and erosion from internal waves and deep anoxic currents (see Pomar et al. 2012; Baird and Brett 1986).
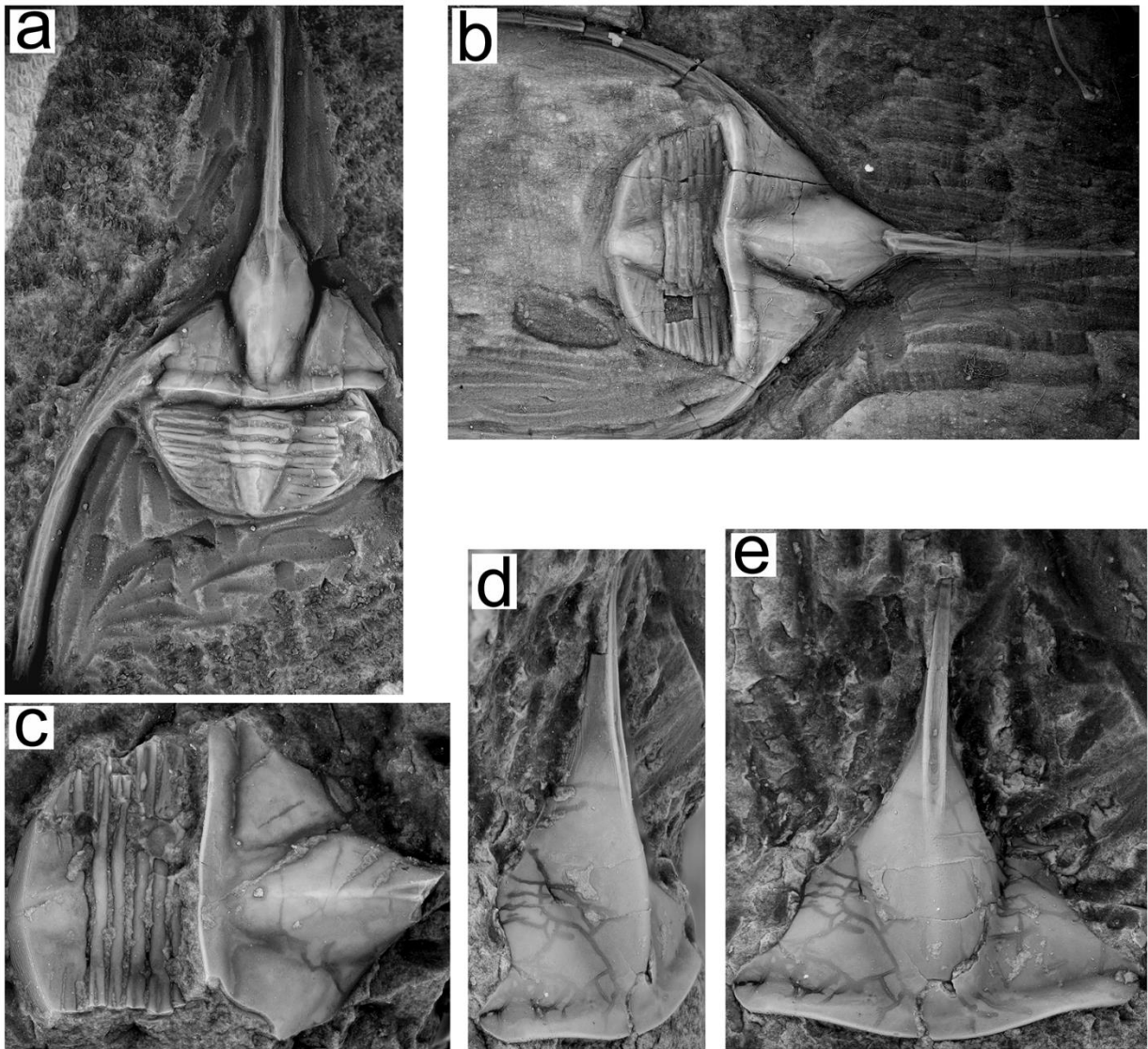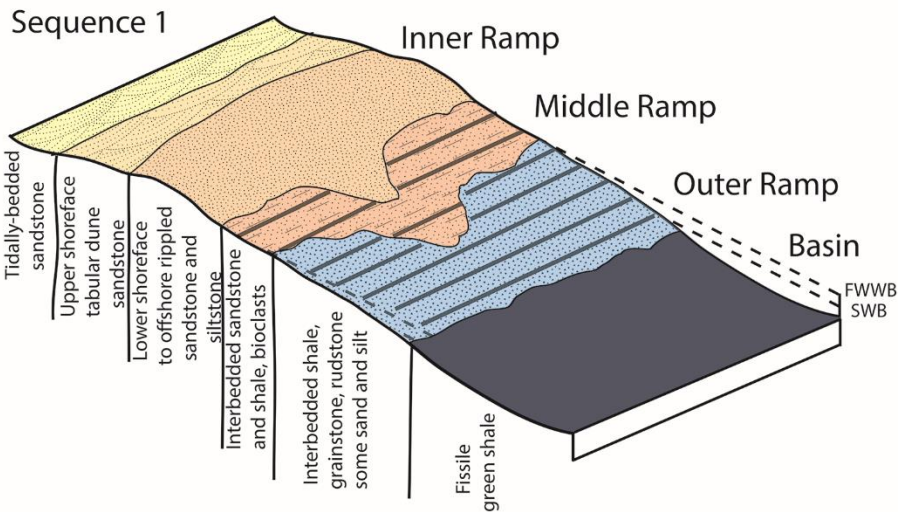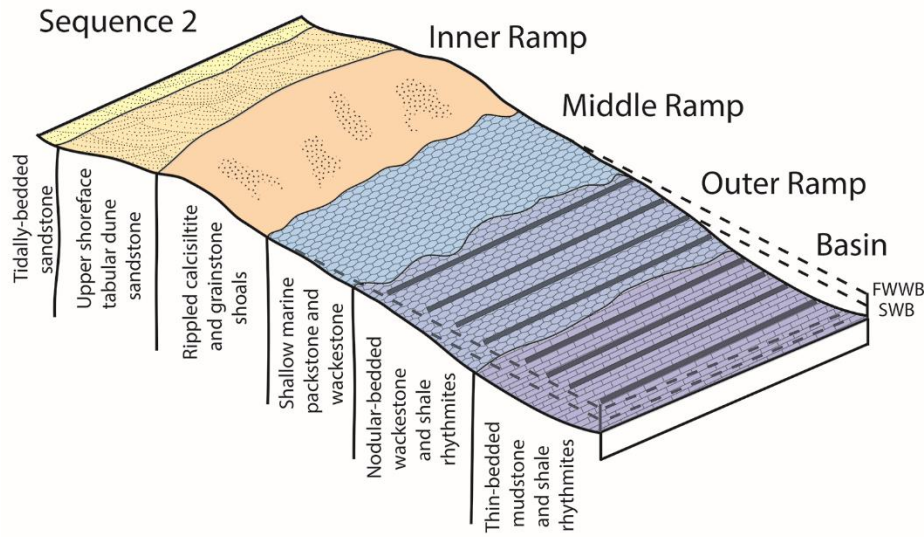
**TEXT-FIGURE 21.—*Lonchodomas mcgheei* (Decker 1931), a common fossil at stop 12 (Rock Crossing) approximately 15 meters below the Bromide/Viola contact. a, complete individual (OU 12531), x 4, dorsal view. b, topotype complete individual (OU 3448), x 4.2, dorsal view. c, damaged individual (OU 12533), x 8, dorsal view. d, e, cranidium (OU 12532), d, lateral view, x7 e, dorsal view, x6.7.**

The facies spectrum (Fig. 15) across the Bromide Formation suggests a more continuous and less dramatic down-ramp change in facies then expected. One implication is that localities such as Rock Crossing were likely deposited on a southern ramp that shallowed southward towards the Texas Arch. Intuitively, this makes sense because the succession in the lower Bromide on the northern ramp towards the SOA (I-35N, stop 7) is similar to the lower Bromide at Rock Crossing.

## Sequence 3

Inner Ramp

Middle Ramp

Outer Ramp

Basin

FWWB
SWB

Fenestral and microbial laminated mudstone

Nodular to massive wackestone tempestites

Bioclastic grainstone shoals and packstone

Nodular packstone

Thick bedded wackestone obrution pavements

N/A

## Sequence 2

Inner Ramp

Middle Ramp

Outer Ramp

Basin

FWWB
SWB

Tidally-bedded sandstone

Upper shoreface tabular dune sandstone

Rippled calcisiltite and grainstone shoals

Shallow marine packstone and wackestone

Nodular-bedded wackestone and shale rhythmites

Thin-bedded mudstone and shale rhythmites

## Sequence 1

Inner Ramp

Middle Ramp

Outer Ramp

Basin

FWWB
SWB

Tidally-bedded sandstone

Upper shoreface tabular dune sandstone

Lower shoreface to offshore rippled sandstone and siltstone

Interbedded sandstone and shale, bioclasts

Interbedded shale, grainstone, rudstone some sand and silt

Fissile green shale

**TEXT-FIGURE 22.—Depositional model of sequences 1-3 in the Bromide Formation. FWWB= Fair weather wave base, SWB = Storm wave base (modified from Carlucci et al. 2014).**

Rock Crossing was pivotal in developing a depositional model of the Bromide Formation through three third-order depositional sequences (Fig. 22). The model records the gradual transition from a siliciclastic dominated ramp (sequence 1), to a mixed siliciclastic-carbonate ramp (sequence 2), to a warm-water neritic carbonate ramp (sequence 3). The ramp profile in sequence 3 is preserved in most of Oklahoma, but not on the southern ramp of the SOA, and only partially in the center of the aulacogen.

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