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INTERPRETATION OF THE DEPOSITIONAL ENVIRONMENTS AND SOFT-SEDIMENT DEFORMATION IN THE UPPER TANGLEWOOD MEMBER (UPPER ORDOVICIAN) OF THE LEXINGTON LIMESTONE, CENTRAL KENTUCKY, U.S.A.

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INTERPRETATION OF THE DEPOSITIONAL ENVIRONMENTS AND SOFT-SEDIMENT DEFORMATION IN THE UPPER TANGLEWOOD MEMBER (UPPER ORDOVICIAN) OF THE LEXINGTON LIMESTONE, CENTRAL KENTUCKY, U.S.A.

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DISSERTATION

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A dissertation submitted in partial fulfillment of the requirements for the degree of Doctorate of Philosophy in the College of Arts and Sciences at the University of Kentucky

By
Dibya Raj Koirala
Lexington, Kentucky

Director: Dr. Frank R. Ettensohn, Professor of Geology
Lexington, Kentucky
2017

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The upper Tanglewood Member is the final member of the Lexington Limestone and is well-known for its soft-sediment deformation. This study has confirmed the carbonate-shoal-complex origin of the unit, and detailed study shows that its development took place during five small-scale, sequence-like, fining-upward cycles related to eustasy and tectonics. Four lithofacies are represented in the unit. Facies analysis of each cycle shows that the thickest and coarsest part of each cycle corresponds to previously uplifted basement-fault blocks; the occurrence of thick, coarse facies on the same fault blocks suggests that the blocks continued to experience uplift due to Taconian far-field forces generated on the eastern margin of Laurentia. The upper Tanglewood Member includes six deformed horizons that can be traced into equivalent parts of the Clays Ferry and Point Pleasant formations. Concurrence of four lines of evidence, suggested by Ettensohn et al. (2002d) for interpretation of seismites, indicates that the widespread horizons of deformation are seismogenic in origin. Reactivation of basement structures due to Taconian far-
field forces probably induced seismicity on the intra-platform carbonate complex so as to produce soft-sediment deformation.

Petrographic investigation indicates that most of the cements in the upper Tanglewood limestones appear to be late diagenetic, fresh-water phreatic cements. Comparing the petrography of deformed and undeformed portions of the same horizon showed no significant differences in terms of cementation, indicating that cementation occurred primarily after deformation. The primary impact of deformation on the microstructure of the unit was the randomization of grain fabric and the increased presence of broken intraclasts.

KEY WORDS: Upper Ordovician Lexington/ Trenton Limestone, Intra-platform shoal complex, Eustasy, Taconian Orogeny, Seismogenic soft-sediment deformation, Small-scale cyclicity
INTERPRETATION OF THE DEPOSITIONAL ENVIRONMENTS AND SOFT-SEDIMENT DEFORMATION IN THE UPPER TANGLEWOOD MEMBER (UPPER ORDOVICIAN) OF THE LEXINGTON LIMESTONE, CENTRAL KENTUCKY, U.S.A.

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Table 1.1 Subjective scheme for rating occurrence and relative intensity of seismogenic deformation
Chapter 1—Introduction

Purpose

The Ordovician Lexington Limestone (Trenton Limestone in subsurface) was initially named by M.R. Campbell in 1898. Since then, the study of this unit has been a continuous pursuit, and many interesting findings have emerged in every step of the study. In the beginning, the Lexington Limestone was treated as a uniform, layer-cake unit. However, U.S.G.S. geological mapping from the 1960’s to the 1990’s changed our concepts about the unit. In fact, detailed geological mapping showed that the Lexington Limestone was not uniform or tabular in nature as previously thought, but instead was an irregular facies mosaic, especially in upper parts of the unit in central Kentucky (Figs. 1.1 and 1.2). Beyond central Kentucky in the subsurface, the unit is more tabular and uniform in thickness and is called the Trenton Limestone; the Lexington and Trenton occur across a large rectangular area in east-central United States called the Lexington Platform. In central Kentucky, the Lexington Limestone is greater than 100-m thick and may be as young as early Edenian in age (Fig. 1.1) (Cressman, 1973; Shaver, 1985), whereas in surrounding parts of the Lexington Platform, its thickness is 15 – 60 m, and it is no younger than mid-Chatfieldian age (Fig. 1.1) (Shaver, 1985). During mapping, it was realized that in central Kentucky, the additional thickness of the Lexington composed a series of shallow-water carbonate shoal complexes. Ettensohn (1991) called this shallow-water carbonate shoal complex the Tanglewood buildup (Fig. 1.1). Beyond central Kentucky, horizons equivalent to the Tanglewood buildup are represented by the deeper water Clays Ferry, Kope or Point Pleasant formations (Fig. 1.1). Hence, because the limestones of the Tanglewood buildup are unique across the Lexington Platform, one purpose of this study is to enhance our understanding about
the tectonic and eustatic controls on the origin of these carbonate-shoal complexes in central Kentucky by studying the well-defined upper carbonate shoal complex known as upper tongue of the Tanglewood Member (Figs. 1.1 and 1.2). As in other parts of the Lexington Limestone, the upper Tanglewood Member also contains numerous deformed horizons which can be traced across long distances. Some of the deformed horizons in lower and middle parts of Lexington Limestone have been interpreted to represent seismogenic deformation (seismites) by various researchers (e.g., Rast and Moshier, 1990; Pope et al., 1997; Rast et al., 1999; Jewell and Ettensohn, 2004; McLaughlin and Brett, 2004). The deformed horizons in the upper Tanglewood Member contain numerous deformational structures, such as ball-and-pillow structures, diapirs, contorted bedding, folding and homogenized beds. Therefore, another purpose of this study is to interpret this soft-sediment deformation through the study of structures by mapping the intensity of deformation and through thin-section analysis of equivalent deformed and undeformed horizons.

The older idea that Lexington Limestone is a tabular, “layer-cake” unit actually reflects the fact that members in the lower part of today’s Lexington Limestone (members to the level of the Brannon) are indeed extensive and more or less layer-cake-type units. It was this lower part of today’s Lexington Limestone that was called the Lexington Limestone until 1960, and this lower part of the unit is still called the Lexington or Trenton Limestone in the subsurface beyond the central Kentucky Bluegrass area. However, the more shaley upper parts of the today’s Lexington Limestone were formally placed in the Cynthiana Formation (Nosow and McFarlan, 1960). In the early 1960’s during the U.S.G.S. geological mapping project, it was not hard to realize that the upper part of Lexington Limestone was not, in fact, a tabular or layer-cake-type unit as previously thought, but instead was a rather complex facies-mosaic (Fig. 1.1) developed
around bodies of calcarenites and calcirudites, called the Tanglewood Member (e.g., Black et al., 1965; Cressman 1973, Cressman and Karklins, 1970; Black and Cuppels, 1973). Therefore, older layer-cake names were abandoned for upper parts of the Lexington Limestone, and new names like the Tanglewood were proposed. Moreover, Cressman (1973) realized that this complex facies mosaic was largely restricted to Bluegrass region of central Kentucky, where the formation was up to three times thicker than elsewhere in the subsurface. Beyond the Bluegrass area, the Lexington Limestone, especially in the subsurface, is thinner and contains only the lower part of the unit that is exposed in central Kentucky. Overlying the thinner Lexington/Trenton beyond central Kentucky are shaley units that include the Maquoketa, Clays Ferry, Kope and Point Pleasant formations.

In 1991, Ettensohn came to realize that the distribution of thicker Lexington Limestone, which comprised a complex facies mosaic of carbonates, largely coincided with basement structures. He suggested that the upper part of the Lexington Limestone was a structurally controlled carbonate shoal related to reactivation of basement faults by Taconian far-field forces. With the above situation in mind, the following are the more specific goals of this research:

1) To divide the upper Tanglewood (Figs.1.1 and 1.2) into meaningful subdivisions that can be correlated across long distances;

2) To interpret upper Tanglewood lithofacies in terms of modern depositional environments;

3) To understand the evolution of the upper Tanglewood Member;

4) To clarify the relationships between local basement structures and individual upper Tanglewood lithofacies;

5) To interpret the nature of soft-sediment deformation in the upper Tanglewood Member;
Fig. 1.1 Interpretive stratigraphic column of the Tanglewood buildup. Interpretive stratigraphic column showing the nature of the Tanglewood buildup and the position of the upper tongue of the Tanglewood Member of the Lexington Limestone in central Kentucky, relative to the position of the Millersburg Member or Clays Ferry Formation at the base of the unit. The subtle unconformity at the base of the Lexington Limestone represents the breakup of the Blackriverian carbonate platform and the inception of the Taconic tectophase of the Taconian Orogeny (adapted from Ettensohn et al., 2002b).
6) To understand possible relationships between soft-sediment deformation and basement structures; and

7) To compare equivalent deformed and undeformed horizons by thin-section analysis.

The main objective of this study centers around solving the above problems, which are accomplished herein by the evaluation of previous studies and collection of additional information, which is analyzed in view of modern carbonate depositional environments, far-field tectonics, structural influence, and event stratigraphy. Such a study has not been carried out in detail for the upper Tanglewood Member to date, and hence, from that perspective, this study is especially critical in understanding the unique stratigraphic position of the upper Tanglewood Member of the Lexington Limestone.

Methods

Depositional Environment

Facies Association

During the progress of this study, 45 sections were measured and described in the field. The distribution of upper Tanglewood Member (Fig.1.2) was mapped by describing 38 sections in the field and seven cores. However, cores were not taken into account for the analysis of soft-sediment deformation. Geological quadrangle maps were used wherever no exposures were available. The underlying and overlying members were noted in all descriptions. Sedimentological and stratigraphic descriptions of the sections were completed using strip-log
format (e.g., Fig. 1.3) to investigate lateral and vertical facies changes. Sedimentary facies were recognized on the basis of their textures, sedimentary structures, paleocurrent patterns, and fossil content. I use the term facies association, which reflects a combination of two or more facies in units, which are considered to represent broader environmental complexes.

Fig. 1.2 Location map of the study area. Location map of the study area showing the surface extent of the Lexington Limestone (Cressman, 1973) and the distribution of the upper tongue of the Tanglewood Member in both surface and subsurface (cross-hatched) relative to central Kentucky counties and the Lexington Limestone, based on measured sections (red, green, and black dots), core data (black triangles), and geologic quadrangle maps. Notice the prominent basement structural trends; tick marks on likely downthrown sides (after Ettensohn, 1992; Ettensohn et al., 2002a, 2004; A = unnamed fault system; B = Georgetown-Gratz fault system; C = Centerville fault system; D = Lexington fault system; E = Vanceburg-Ironton fault system).
This approach was used in order to distinguish facies associations, interpret the processes which formed them, and reconstruct analogous environmental interpretations in which such a facies association could develop. The Grabau (1903) classification is used to describe megascopic lithologic textures. Thus, an upper Tanglewood lithology might be described as low-angle, cross-laminated calcarenite with thin shale partings.

**Fig. 1.3 Measured section from Georgetown, KY.** Measured section from Georgetown, KY (38° 16’ 15.52” N, 84° 33’ 09.03’’), with its corresponding exposure along highway I-75. Note the prominent fining-upward cycles and deformed horizons (DH). Notice the upper shaley parts of each cycle. The five noted cycles and deformed horizons are present across the distribution of the upper Tanglewood Member, providing a means of correlation. The Millersburg Mbr., as well as the lowermost cycle (Cycle 1) and the upper part of Cycle 5, are not shown in the figure.

**Temporal correlation**

In order to understand the nature and distribution of depositional environments, it was necessary to have some means of correlation across the unit. Only in this way would it be
possible to demonstrate how depositional environments changed across the distribution of the unit. To establish temporal correlations across the unit, the presence of sequence-like cycles and deformational event horizons were used for correlation, assuming that they represent two types of contemporaneous events that occur across the entire upper Tanglewood Member.

**Sequence-like cycles**

Cyclic sedimentation, which is apparent throughout most of the Lexington Limestone, has resulted in five small-scale, sequence-like cycles in upper Tanglewood Member (Fig. 1.3). Inasmuch as the cyclicity apparent in the upper Tanglewood Member is allocyclic, mainly contributed by eustatic sea-level variations that were experienced nearly everywhere (e.g., McLaughlin and Brett, 2004), these cycles have correlative value for chronostratigraphy. Each small-scale, sequence-like cycle represents a relatively conformable succession of genetically related beds that is bounded by combined erosion and transgressive surfaces (Fig. 1.4). Each cycle of sedimentation is characterized by a fining-upward succession, deposited contemporaneously over a depositional surface during a marine transgressive phase and is commonly identified and separated from other cycles by erosive/transgressive surfaces at their tops and bottoms (Fig. 1.4). The grain size of the sediments in each cycle changes abruptly on either side of the surface. This abrupt change in the grain size is associated with the change of energy at the sediment-water interface caused by marine transgression and regression. This concept of abrupt change in sediment grain size is helpful in identifying the individual bounding surfaces of each cycle. Another important aspect is that even if lateral changes in depositional environments are present, we can still easily identify respective fining-upward cycles in the different environments, because the effecting marine flooding events were contemporaneous.
across the entire shoal complex and beyond (e.g., Busch and Rollins, 1984; Busch and West, 1987).

**Fig. 1.4 Cycles used for chronostratigraphic correlation in two different exposures.** Chronostratigraphic correlation in two different exposures using cyclicity. A–Frankfort (38° 20' 35.00” N, 84° 50' 55.20”W); B–Cynthiana (38° 24’ 2.50”N, 84° 17’ 33.30”W). Red lines represent cycle boundaries.

**Deformed horizons**

Even though each of the sequence-like cycles noted above represents a short-term event of perhaps 100,000 years duration, even shorter-term evidence of virtually instantaneous events of
perhaps a few hours or days is present in some of the cycles. These events are represented by six
deformed horizons (Figs. 1.3 and 1.5), and by virtue of the fact that they were nearly
instantaneous and widespread, they can be used as aids in correlation (e.g., Algeo and Brett,
1999; McLaughlin and Brett, 2004). Many such deformed horizons have been recorded
throughout the Lexington Limestone, and some have been interpreted to represent seismites (e.g.,
Jewell and Ettensohn 2004; Ettensohn et al., 2002b). Similarly, other upper Ordovician soft-
sediment deformation features exposed in Kentucky have also been interpreted as seismogenic
deformation (Algeo and Brett, 1999; McLaughlin and Brett, 2004), and I have also interpreted
these particular deformed horizons in upper Tanglewood Member of Lexington Limestone as
seismites (Chapter 6). Their widespread distribution beyond the Lexington Limestone into Clays
Ferry and Point Pleasant equivalents indicates that these may serve as marker horizons for
chronostratigraphic correlation. All sections, however, do not exhibit the six deformed horizons.
Therefore, establishing the position of deformed beds relative to the sequence-like cycles was
important. This was done by observing several upper Tanglewood sections that contained all
cycles and all six deformed horizons. One of these sections (Fig. 1.3) was then designated as a
reference section, and all other sections were compared with it and correlated accordingly. Use
of deformed horizons was especially useful for correlation where the bounding surfaces for
cycles were not apparent due to erosion from above. Altogether, 38 sections and seven cores
(Appendix A), both within and outside the shoal area (Fig. 1.2), were measured, described and
correlated in this way (Figs. 1.3, 1.4).

Thirty-eight exposures that contain the upper Tanglewood or laterally equivalent units
were measured and described from 13 counties in central Kentucky and surrounding areas (Fig.
1.2). Horizons of soft-sediment deformation in the upper Tanglewood and equivalent Clays Ferry
and Point Pleasant formations were defined on the basis of the cycles that contain them. Similarly, these deformed horizons may be used to correlate from the upper Tanglewood Member into equivalent parts of deeper-water units like the Clays Ferry, Kope and Point Pleasant formations.

![Fig. 1.5 Exposure of the upper Tanglewood from Nicholas County.](image)

**Fig. 1.5 Exposure of the upper Tanglewood from Nicholas County.** Exposure of the upper Tanglewood from Nicholas County, KY (38° 24’ 10.32”N, 84° 0’ 51.84”W), showing a single deformed horizon, 2.4-m (8-ft) thick.

**Soft-sediment deformation**

Lowe (1975) has broadly classified the causes of soft-sediment deformation into three processes: hydroplastic deformation, liquefaction, and fluidization, each of which leaves distinct features in the sedimentary record (Table 1.1). Using these features, the predominant deformation processes responsible for generating each horizon in the upper Tanglewood were
characterized accordingly. On the basis of minimum fluidization or pore-fluid velocity ($U_0$), these deformational processes can be described based on certain features. At lower energies with pore-fluid velocities below $U_0$ (less than the minimum fluidization velocity, $U_0$), hydroplastic deformation (Fig. 1.6) is the most dominant type of deformation; it is characterized by simple contortion through the folding or faulting of beds and preserves primary laminae and bedding. As the energy increases and pore-fluid velocity approaches the minimum fluidization velocity ($U_0$), liquefaction (Fig. 1.7) is the predominant process; it is characterized by diapiric intrusions in the form of vertical piping, sedimentary dikes, or sand volcanoes. Fluid flow is normally characterized by laminar flow and fine-grained sediments separate in a rising column of water. As the energy increases further and pore-fluid velocity is at or greater than the minimum fluidization velocities ($U_0$), fluidization of sediments becomes the predominant process. Fluidization, in which fluid flow is turbulent, is characterized by nearly complete destruction of primary laminae and homogenization of the unit (Fig. 1.8).

Fig. 1.6 Hydroplastic deformation from the Point Pleasant Formation. Convolute bedding in the upper Tanglewood from Cycle 2 equivalent in the Point Pleasant Formation, exposed along US Highway 1159 in Bracken County (38°43'54.95"N, 84° 6'3.62"W).
Fig. 1.7 Liquefied bed from Nicholas County. Liquefied bed and diapirs from Nicholas County (38° 24' 26.01" N, 84° 0' 46.43" W) in Cycle 1.

Fig. 1.8 Homogenized beds. Homogenized beds within Cycle 2 in Anderson County along US Highway 62 (38° 3' 30.90" N, 84° 55' 15.30" W). Note the thickening and thinning of the deformed horizon.
When soft-sediment deformation results from earthquakes, it is due to a sudden, but temporary, increase of pore-fluid velocity which is normally proportional to earthquake energy released through cyclic loading (Lowe, 1975; Ettensohn et al., 2002b). Therefore, assuming uniform lithology, the energy released by an earthquake tends to decrease further away from the epicentral area, and the most intense deformation will be around or near the epicenter (Ettensohn et al., 2002a). Each of the three deforming processes described above requires successively higher pore-fluid velocities, and hence energy input. Accordingly, the three deforming processes noted above should reflect the energy released in space and time so that the most energy-intense process, fluidization, should occur close to the epicenter and the least energy-intensive processes, hydroplastic deformation, should occur more distal to the epicenter. In between these two extremes, features related to liquefaction should be common. Hence, mapping the distribution of deformation intensity in possible seismogenic horizons using the features should show a concentric pattern of deformation types around a likely epicentral area, which is characterized by the most intense deformation, fluidization. Effectively, such maps are isoseismal bands of intensity about an epicenter and can be used to approximate the epicentral areas of ancient earthquakes.

In order to map the distribution of deformation types, a nine-point rating scheme of increasing deforming intensity (Table 1.1) was developed for use in the field (Jewell, 2001; Ettensohn and Stewart, 2003; Jewell and Ettensohn, 2004). It was impractical to map the extent of each numbered degree of deformation intensity, because deformation intensity represented by any given number on the table varies tremendously from place to place. Therefore, isoseismal maps were prepared by mapping the three major types of deformation, simple folding and faulting (1 – 3), diapiric intrusion (4 – 6), and homogenization (7 – 9). Because variation may be
great, even at the outcrop scale, it was commonly necessary to map the predominant type of deformation at any one exposure.

<table>
<thead>
<tr>
<th>Major Types of Deformation</th>
<th>Simple Folding/Faulting</th>
<th>Diapiric Injection</th>
<th>Homogenization</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Thinning and thickening of beds; simple folds, contortion, pull-apart, brecciation, thin and sporadic</td>
<td>4 Sporadic presence of penetrating diapirs</td>
<td>7 Some original bedding remains; widespread channel-like bodies, intrusions or pockets of fluidization</td>
<td></td>
</tr>
<tr>
<td>2 Horizons with channel- or pillow-like, flow-cell deformation; longer expanses deformed but still sporadic</td>
<td>5 Thicker deformed horizons with moderate occurrence of diapirs</td>
<td>8 Moderately thick horizons (~1 m) of largely homogenized sediment; few remnants of original bedding persist</td>
<td></td>
</tr>
<tr>
<td>3 Thicker, persistent horizons of simple fold-and-pillow-type deformation without penetrating flame structures</td>
<td>6 Diapirs widespread and persistent throughout interval; local pockets of fluidization associated with diapirs</td>
<td>9 Thick horizons (~2 m) of completely homogenized sediment; rounded clasts floating in structureless matrix; widespread</td>
<td></td>
</tr>
</tbody>
</table>

General characteristics

Primary structures and bedding preserved; simple folding, thinning and thickening; faulting

Much original bedding preserved; vertical water-escape structures and redistribution of grains; laminar flow

Large-scale destruction of bedding; major homogenization; turbulent flow

| Dominant process | Hydroplastic Deformation | Liquification | Fluidization |

*Table 1.1 Subjective scheme for rating occurrence and relative intensity of seismogenic deformation (from Jewell, 2001; Ettensohn and Stewart, 2003; Jewell and Ettensohn, 2004).*

**Paleogeographic, Paleoclimatic and Tectonic Framework**

By Late Ordovician time, the continent of Laurentia, which would later become North America, sat on the equator (Fig. 1.9), with two bordering oceans: the Panthalassic Ocean to the north and west and the Iapetus Ocean to the south and east (Scotese et al., 1991). During much of Paleozoic time, southern and eastern Laurentia were active continental margins. Evidence for three major orogenies on these margins, containing more than a dozen recognizable tectophases, is recorded within Ordovician to Pennsylvanian strata (Ettensohn, 1991, 1992, 2008). Among these strata, Upper Ordovician strata are particularly widespread and well-preserved, recording the first orogeny to affect the southeastern margin of Laurentia after an approximately 500
million-year hiatus (Drahovzal et al., 1992). In Kentucky, these Upper Ordovician strata are well-exposed in various sections and formations. The Lexington Platform, a part of Laurentia, was a rectangular depositional platform on the margin of the Appalachian basin or Taconian foreland basin (Fig. 1.10) (Keith, 1989). This platform resulted from the breakup of larger Blackriverian platform during the inception of Taconic tectophase of the Taconian orogeny to the east (Ettensohn et al., 2002a).

Fig. 1.9 Paleogeographic map of east-central Laurentia during Late Ordovician time (adapted from Scotese et al., 1991).
The earlier Blackriverian carbonate platform was an extensive carbonate platform with a uniform pattern of subtropical carbonate deposition. By Rocklandian time (late Sandbian, early Chatfieldian, mid-Caradoc), this uniform pattern of deposition halted when this extensive platform broke apart along the margin of a nearly 1000-km-long corridor (Fig.1.10), known as the Sebree Trough (Ettensohn et al., 2002a). The development of the Sebree Trough was coeval with the onset of major tectonism and subsidence in northern parts of the Taconic foreland basin. The coincidence of Sebree Trough development with this phase of tectonism suggests that the subsidence of this intracratonic basin was associated with far-field tectonic forces and reactivation of older structures (Ettensohn et al., 2002a). The Sebree Trough corridor reflects development of two largely independent carbonate-platform sequences on either side of an intervening area of nondeposition that is recognized by a pronounced thinning of Trenton Limestones along a regional unconformity. This unconformity is extensive across the Galena Shelf, appears to fall in section along the northwest margin of the corridor where corrosion and nondeposition were greatest, and rises southeastwardly onto the Tanglewood buildup of the Lexington Platform to become the sub-Sulfur Well unconformity noted by Cressman (1973) (Keith and Wickstrom, 1993; Hohman, 1998; Ettensohn et al., 2002a). As a result, the Lexington Platform separated from other parts of the old Blackriverian Platform, and the deeper-water Sebree Trough (Fig. 1.10) developed along its western and northwestern boundaries (Ettensohn, 1999; Ettensohn et al., 2002a, 2004).

By Edenian time (early Katian; mid-Caradoc), a thick accumulation of deeper-water dark shales, or interbedded fine-grained, argillaceous carbonates and shales, on the most corroded part of the unconformity, appeared to form a trough separating the shallow-water Galena carbonate
shelf to the northwest from the shallow-water Trenton carbonate shelf to the northeast and from
the shallow-water Lexington carbonate platform to the southeast. Therefore, the Lexington
Platform and Sebree Trough (Fig. 1.10) reflect the structural and stratigraphic differentiation of
the older, Blackriverian carbonate platform during Rocklandian-to-early Edenian (mid-Caradoc,
early Katian) parts of the Taconic tectophase of the Taconian Orogeny.

Fig. 1.10 Paleogeographic map of east-central Laurentia during Late Ordovician time (early Katian,
mid-Caradoc; Chatfieldian, late Shermanian). The Lexington Platform, shown at approximately 25° S
latitude, dipped abruptly to the northwest into the Sebree Trough and to the southeast into the Taconic
foreland basin. Note the great distance from the Taconic highlands on the northwestern margin of the
Lexington Platform (>400 km) where the study area is located. Study area within dashed box (adapted
from McLaughlin and Brett, 2004).
Central Kentucky, which lies on the former Lexington Platform, was situated about 20 to 25 degrees south latitude (Fig. 1.10), about 600-km southwest of a mountainous region called Taconica along the eastern coastline of Laurentia (Scotese et al., 1990). This mountainous area developed from the collision of the Laurentian continent margin with an island arc when Baltica moved toward Laurentia. The collision resulted in mountainous regions situated at the Virginia and New York promontories, called Blountia and Taconica respectively (Ettensohn, 1991).

Initially, during Blackriverian (mid Caradoc; late Sandbian) time, collision in the southern Appalachian region produced the Blountian highlands (Shanmugam et al., 1982; Walker, 1984), during which the Blackriverian High Bridge Group was deposited in central Kentucky. Later, during Shermanian time, collision with the Amonoosuc Island arc in the northern Appalachian region emplaced an accretionary wedge onto the Laurentian margin, forming the Taconic allochthon or Taconica (Rowley and Kidd, 1981). This tectonic event reflects the Taconic tectophase of Ettensohn (1991), and it was during this event that the Lexington Limestone was deposited.

The Lexington Limestone and its Trenton equivalents are commonly separated from the underlying Blackriverian, very shallow, and open-marine to peritidal carbonates by a subtle unconformity (Fig. 1.1) that probably represents shallowing and Taconic bulge movement during inception of the Taconic tectophase (Ettensohn 1991; Ettensohn et al., 2002a).

The Lexington Platform exhibited excellent conditions for the deposition of carbonates during Late Ordovician time. It was situated in the Late Ordovician subtropics with a warm, semiarid climate (Fig. 1.10). The long and narrow Sebree Trough on the western side of the platform was connected to the open Ouachita Sea from which nutrient-rich waters upwelled,
providing abundant essential nutrients for the carbonate-secreting organisms on the Lexington Platform (Ettensohn et al., 2002a; Ettensohn, 2010). In contrast, the Taconic foreland basin developed on the eastern side of the platform during the Taconic Orogeny. This foreland basin acted as a clastic sink for the sediments derived from the Taconic Mountains, thereby promoting carbonate production in the absence of clastic influx on the Lexington Platform.

**Structural Framework**

The central Kentucky area coincides with the Jessamine Dome, a broad, irregular structure with strata dipping roughly 3.75 to 5.62 meters per km to the west and somewhat less to the north and east (Cressman, 1973); it is a structural culmination on the axis of the Cincinati Arch. Two major normal fault systems, the Kentucky River and West Hickman-Byan Station fault systems, transect the Jessamine Dome and intersect in the Little Hickman Quadrangle near the apex of the dome (Fig. 1.11 and 1.12). Besides these two major faults, many other northwest-trending normal faults and grabens have been reported in this region (Black and Haney, 1975). The Kentucky River and Irvine-Point Creek fault zones are among several ancient features in the region that have basement precursors (Fig. 1.11) that were active during Cambrian (Webb, 1969) and Ordovician times (Ettensohn, 1992a; Ettensohn and Kulp, 1995; Rast and Goodman, 1995; Ettensohn et al., 2004).
**Fig. 1.11 Basement structures in east-central United States.**  Map of east-central United States showing the distribution of major Keweenawan, Grenvillian, and Iapetan basement structures of the Appalachian foreland basin and adjacent parts of the foreland relative to the New York (N) and Virginia (V) promontories. The stippled area represents dark-shale infill of the Sebree Trough and adjacent Appalachian lows that supported the upwelling of deep, cold, nutrient-rich waters during the Chatfieldian (late Sandbian–early Katian) reactivation of basement structures when Taconic tectonism was concentrated at the New York promontory. The Tanglewood buildup is one of several examples of uplift and structural inversion during the Taconic tectophase (adapted from Ettensohn et al., 2002a).

The structural reactivation that generated the Lexington Platform also apparently gave rise to the Jessamine Dome through the reactivation of basement structures (Borella and Osborn, 1978; Ettensohn and Kulp, 1995; Ettensohn et al., 2002a). However, the Cincinnati Arch was apparently not active during deposition of the Lexington Limestone (Weir et al., 1984). Stratigraphic evidence suggests fault reactivation and uplift in the Jessamine Dome area during Middle and Late Ordovician time (Borella and Osborne, 1978; Weir et al; 1984). The reactivation of these basement faults was probably related to bulge movement associated with the Taconic tectophase ongoing to the east of the Lexington Platform.
Fig. 1.12 Major fault systems in central Kentucky. Map of central Kentucky illustrating the major fault systems and possible basement lineaments. A = Georgetown-Gratz Fault system; B = unnamed fault; C = unnamed fault; D = Brumfield Fault zone; E = Lexington Fault system; F = unnamed fault; G = unnamed fault; H = West Hickman-Bryan Station fault system; I = Centerville Fault; J = unnamed fault; K = unnamed fault; L = unnamed fault; M = Irvine-Paint Creek Fault system; N = unnamed fault; O = Kentucky River Fault system (adapted from Clepper, 2011).
As already suggested, the Kentucky lithosphere is characterized by many zones of structural weakness related to basement faults (Denison et al., 1984; Black, 1986; Shumaker, 1986; Ettensohn et al., 2002b). Most of these basement faults are related to Keweenawan extension (~1.1 Ga), Grenville compression (1.0 Ga), or Iapetan rifting (0.74 Ga) (Ettensohn et al., 2002b). The Keweenawan and Iapetan rifts are oriented basically southwest-northeast (Stark, 1997; Ettensohn et al., 2002b), whereas Grenville structures are largely oriented north-south or northwest-southeast (Fig. 1.11). These structures on the southeastern margin of Laurentia were most active during early Late Ordovician time (late Rocklandian-early Edenian; late Sandbian-early Katian; mid-Caradoc). During this period, collision of an island arc on northern parts of this margin (period of Taconic Tectophase) transmitted horizontal stress cratonward as far-field forces on and near the Lexington Platform which uplifted and tilted parts of the platform (Ettensohn et al. 2002b). Since the Lexington Limestone was deposited contemporaneously with the Taconic orogeny, structural or tectonic control on the distribution of facies in the Lexington Limestone due to the reactivation of basement faults resulting from the far-field forces, has been suggested (Ettensohn, 1992; Ettensohn and Kulp, 1995). The Nashville and Jessamine domes were also apparently fault-bounded structures that experienced their first uplift at this time as indicated by rapid facies changes (Borella and Osborne, 1978; Weir et al., 1984).

Locally developed, tidal- and wave-influenced grainstones in the upper Lexington Tanglewood Member are restricted to the Jessamine Dome area, supporting the uplift of the dome area at this time. Rapid facies and thickness changes in the upper Tanglewood Member are observed across parts of the Kentucky River and Irvine-Point Creek Fault zones, indicating that these faults were also active during Ordovician time (Cressman, 1973; Pope and Read, 1997a, b; Ettensohn et al., 1986).
Of course, all of the fault reactivation is probably related to far-field forces generated during the Taconian Orogeny. This Taconian orogeny, which occurred when several island arcs collided with the southeastern margin of Laurentia, has very important implications for the Lexington Limestone, because the deposition of the Lexington Limestone coincided with the Taconic tectophase of the Taconian Orogeny. The Taconian Orogeny is divided into two tectophases, the Blountian and Taconic (Kay and Colbert, 1965). The first tectophase, the Blountian tectophase, reflects convergence at the Virginia and Alabama promontories and is represented by the rocks of the High Bridge Group, which are part of an extensive Blackriverian carbonate platform (Keith, 1988, Jewell, 2001, Ettenson et al., 2002b). After Blountian convergence, convergence shifted northeastward and took place at New York promontory (Kay and Colbert, 1965, Ettenson, 1991), where the Taconic tectophase was focused (Ettenson, 1991).

During the Taconic tectophase, the Martinsburg foreland basin was formed in northern parts of the Appalachian Basin, while on its northwestern margin the old Blackriverian carbonate platform collapsed and the new Lexington platform was established (Ettenson, 1999, 2004; Jewell, 2001; Ettenson et al., 2002a,b). In Kentucky, the Taconic tectophase coincided with the deposition of dominantly subtidal carbonates on a ramp that sloped northeastwardly into the Martinsburg foredeep, and passed northwestwards into the Sebree Trough (Cressman, 1973; Ettenson et al., 2002b)

The Sebree Trough was a narrow, linear, bathymetric trough that developed during Late Turinian to early Chatfieldian time above the failed late Precambrian-Early Cambrian Reelfoot rift and other Keweenawan structures. This trough extended from the southern margin of the Lawrentian craton in central Arkansas northeastward into southern Illinois before turning
abruptly eastward, and continuing on as parts of the Rough Creek graben in western Kentucky (Kolata and Nelson, 1991). The rift system is considered to be a precursor to the proto-Illinois Basin and had a major influence on basinal depositional systems throughout most of Paleozoic time (Kolata and Nelson, 1997).

**Previous Study**

**Lexington Limestone**

Campbell first named the Lexington Limestone in 1898. Since then, there have been many changes in nomenclature. Most early studies of the Lexington Limestone stratigraphy were based on paleontological evidence because of the scarcity of exposure.

Foerste (1913) identified and listed the fossils contained within each member of the Lexington (Trenton) Formation. Miller (1915) determined that the Cynthiana Formation (upper part of present-day Lexington Limestone) was a member of the Cincinnatian Series, and that it was overlain by the Eden Shale (now the Clays Ferry Formation) and underlain by the Perryville Limestone. At that time, the upper Tongue of the Tanglewood Member would have been included in the Cynthiana Formation. Subsequent research on the Lexington Limestone was performed by McFarlan (1938, 1943), McFarlan and White (1948), and Nosow and McFarlan (1960), all of whom identified various new members in the unit.

In the 1960’s, detailed mapping of Kentucky was begun by the U.S. Geological Survey in cooperation with the Kentucky Geological Survey. Much information regarding lithology, relationship among members, and relationships between the Lexington Limestone and
surrounding units was produced as a result of this mapping project. The preliminary results of this study were published by Black and MacQuown (1965), Black et al. (1965), and Cressman and Karklins (1970).

Cressman (1973), on the basis of U.S. Geological Survey mapping, divided the Lexington Limestone into eleven members, which include, in ascending order: the Curdsville Limestone Member (mostly calcarenite), the Logana Member (calcsiltite and shale), the Grier Limestone Member (fossiliferous, shaley, nodular calcarenite), the Perryville Member (calclutite), the Brannon Member (calcsiltite and shale), the Sulphur Well Member (shaley, bryozoan-rich calcarenites), the Tanglewood Limestone Member (calcarenite/ calcirudite), the Devils Hollow Member (gastropodal calcirudite and calcilutite), the Greendale Lentil, the Stamping Ground Member, and the Millersburg Member (all of which include nodular, fossiliferous limestone and shales). Later, the Strodes Creek Member (calcsiltites and interbedded dark shales) was added by Black and Cuppels (1973), when it was recognized that it was a mappable unit within the Millersburg Member.

Subsequently, the Lexington Limestone has been heavily studied giving emphasis to the nature of individual members (Hrabar et al., 1971; Mackey, 1972, Conkin and Dasari, 1986; Ettensohn et al., 1986; Ettensohn, 1992; Ettensohn and Kulp, 1995; Kasl, 2001; Jewell, 2001; Ettensohn et al., 2002a,b,c; Jewell and Ettensohn, 2004; Ettensohn et al., 2004), as well as to the possibility of structural control on deposition (Ettensohn and Kulp, 1995; Ettensohn et al., 2002a, b, 2004).

In particular, Ettensohn and coworkers have shown that parts of the Lexington Limestone reflect a structurally related carbonate buildup (e.g., Ettensohn 1992; Ettensohn and Kulp, 1995) that resulted from structural inversion (Ettensohn et al., 2004). Thus, they related these effects to
far-field forces emanating from the Taconic tectophase of the Taconian Orogeny (Ettensohn et al., 2002a; Ettensohn and Lierman, 2015)

Most recently, Clepper (2011) studied the lithostratigraphic and paleoenvironmental framework of the Upper Ordovician Lexington Limestone in the Bluegrass region of central Kentucky. She completed detailed distribution maps of every member and established the sequence stratigraphy of the Lexington Limestone.

**Soft-sediment deformation and seismites**

Soft-sediment deformation is attributed to the overpressuring of pore fluids in an unconsolidated sediment, resulting in the modification, and often destruction, of primary sedimentary structures along with the formation of new secondary sedimentary structures (Lowe, 1975). A variety of mechanisms can initiate such conditions, including bedform shear, loading, slumping, seismicity, and many other processes (Allen, 1986; Obermeier, 1996; Owen, 1996; Jones and Omoto, 2000; Owen and Moretti, 2011). The major sedimentary structures formed by soft-sediment deformations include convoluted bedding, folding, ball-and-pillow structures, flame structures, injectites, sand volcanoes, and many other less common features.

One of the major problems in interpreting soft-sediment deformation is differentiating what caused the deformation. There have been many attempts to differentiate seismogenic deformation from non-seismogenic deformation. The term ‘seismite’ was originally coined by Seilacher (1969) to describe sedimentary beds that had undergone deformation following disturbance by earthquakes. Various researchers (e.g., Ettensohn et al., 2002a; Jewell and Ettensohn, 2004; Sims, 1975; Wheeler, 2002) have suggested criteria to differentiate seismogenic deformation from other types.
Sims (1975) and Wheeler (2002) have given six criteria to interpret soft-sediment deformation as seismogenic deformation, which include (1) presence of sharp, but non-erosive contacts of the deformed interval with undeformed rocks lying both above and below; (2) synchronicity, or demonstration that the deformation is confined to a relatively short (narrow) and correlatable stratigraphic horizon; (3) intensity zoning of the deformation, such that within any laterally consistent and mappable deformed horizon there is a systematic increase in the size and intensity of deformation toward a possible epicentral area; (4) the size and nature of the soft-sediment deformation is comparable with soft-sediment deformation of known seismic origin; (5) a suitable tectonic setting in which an earthquake is capable of generating large-scale and laterally persistent soft-sediment deformation; and (6) a suitable depositional setting likely to experience seismic shaking and sediment properties capable of producing soft-sediment deformation structures, such as correct grain size, lack of cementation and compaction.

Ettensohn et al. (2002a), on the other hand, have suggested concurrence of four lines of evidence to interpret seismogenic deformation, which include: (1) deformation consistent with seismogenic origin; (2) widespread distribution in temporally and stratigraphically constrained horizons; (3) a pattern of increasing frequency or deformation intensity toward likely epicentral areas; and (4) the ability to exclude other likely causes. There are many examples, some of which are briefly discussed below.

For instance, Ezquerro et al. (2015) have interpreted part of the liquefaction and fluidization in Late Pliocene–Early Pleistocene palustrine/lacustrine sediments of central-eastern Iberian Chain (eastern Spain) as seismically induced structures. Their interpretation was based on the structural and seismogenic framework of the area, the detailed morphological study of soft-sediment deformation structures, and sedimentological evidence from core data.
Wallace and Eyles (2015) similarly identified seismites within Ordovician–Silurian carbonates and clastics of southern Ontario, Canada. The key criteria used to interpret seismites were the prominence and size of the structures, compared to other deformation structures produced by storm activity in surrounding strata; their presence in widespread stratigraphically constrained (correlatable) horizons in core and outcrop; and most importantly, their similarity to commonly reported deformed horizons that are widely accepted as seismites in regionally correlative siliciclastic and carbonate strata in closely adjacent areas of the U.S.A.

Hilbert-Wolf et al. (2016) have identified balloon-shaped inflation structures in Cretaceous East African Rift strata, which were assumed to have been generated due to gas release with a subsequent water-escape phase during a localized seismic event. The two new types of soft-sediment deformation that they reported from the Cretaceous Namba Member include (1) balloon-shaped inflation structures and (2) associated linear surface fractures with linked sandstone splays.

Mugnier et al. (2011) reported seismites in the Pliocene to Pleistocene fluvio-lacustrine sediments of the Kathmandu Basin of Nepal. The soft-sediment deformation features have been linked to earthquakes because of their synchronicity with historic seismic events or dike initiation.

The lateral extent of the deformation is considered to be a key factor for recognizing in the sedimentary record (Sims, 1975). Nonetheless, Moretti and Ronchi (2011) have interpreted liquefaction features with limited lateral extent as seismites in the Pleistocene fluvio-lacustrine deposits of the Neuquen Basin (northern Patagonia) of Argentina. The limited lateral extent of
the analyzed seismite beds was explained by the disappearance of the sandy fluvial/marginal-
lacustrine facies to which the deformation was constrained.

Early Paleozoic soft-sediment deformational features in eastern North America are
abundant and widespread. Seismites are widely regarded as geological event deposits of great
importance for the understanding basin-scale tectonics and evolution. However, most of our
knowledge about seismites comes from case studies in Europe and North America, primarily
from Tertiary or Quaternary clastic sediments deposited in terrestrial or marginal-marine
settings. In contrast, there have been very few studies on carbonate seismites in cratonic shallow-
marine settings like Lexington Limestone.

Rast and Moshier (1990) first suggested that most of the deformed horizons in the
Lexington Limestone are seismites. Since then, further work has been ongoing by various
researchers to confirm those interpretations. Generalized interpretation of deformed horizons in
the Lexington Limestone as seismites has already been made by Pope et al. (1997a, b), Rast et al.
(1999), Ettensohn and Stewart (2002), Ettensohn et al. (2002a, 2004), and Jewell and Ettensohn
(2004), but none of these workers has specifically examined the upper tongue of the Tanglewood
Member, which contains some of the most deformed carbonates in the Lexington Limestone.

To reduce the ambiguity between seismogenic and non-seismogenic deformation,
Wheeler (2002), Ettensohn et al. (2002b), and Jewell and Ettensohn (2004) suggested four
concurrent lines of evidence to interpret seismites, which include 1) deformation consistent with
a seismogenic origin, 2) deformation with widespread distribution, 3) deformation that can be
temporally or stratigraphically constrained across this distribution, and 4) deformation that
shows systematic increase in frequency and intensity toward a likely epicentral area.
Subsequently, Ettensohn et al. (2002c) showed just such a pattern of changes for three deformed
horizons in the Brannon Member of the Lexington Limestone, suggesting that the horizons were seismogenic in origin.

The Cane Run Bed of the Lexington Limestone also possesses up to three horizons of soft-sediment deformation that form a distinctive event bed; time-equivalent horizons of different lithology are also deformed. Jewell and Ettensohn (2004), on the basis of the concurrence of four lines of evidence, again suggested seismogenic origin for each horizon. Their interpretation was bolstered by the facts that the deformation contains random fold axes, crosses facies boundaries, is associated with a periodically reactivated basement structural lineament, and crosscuts undeformed beds both above and below.

Although the Cane Run Bed and the included deformation are restricted to the Grier Member, Jewell (2001) showed that equivalent horizons of deformation also occur in coeval parts of the Perryville (Cornishville Bed), Brannon, and Tanglewood (lower tongue) members.

McLaughlin and Brett (2004) also studied various deformational features, such as convolute bedding/lamination, dish structures, mudstone diapirs, boudins and brecciated fabrics, among others, from the Upper Ordovician rocks of north-central Kentucky and interpreted the deformed horizons as seismites, mainly on the basis of the widespread nature of multiple deformation horizons that were attributed to the reactivation of basement faults.
Chapter 2—Cyclicity in Upper Tanglewood Deposition

Introduction

Repetition of different facies within a sequence gives rise to cyclicity in sedimentary rock successions. Such repetitive patterns reflect the repetition of specific environments and the characteristic sedimentation that accompanies them. Depending upon the duration of deposition, sedimentary cycles have been divided into categories, ranging from first to sixth order. Many researchers (e.g., Vail et al., 1991; Carter 1998) have described such sedimentary cycles as ranging from several hundred million years (1st order) to less than twenty thousand years (6th order) in duration. Cyclicity comprises a wide range of cycle sizes, from thin-bedded couplets to meter-scale phenomena to very thick (macro-scale, 100 m) sedimentary sequences. Some of the smaller-scale, discyclic or quasiperiodic features were apparently caused by autogenetic processes operating in the basin itself, whereas most of the medium- to large-scale sequences resulted from regional (e.g., tectonism, climate) or global processes (e.g., eustasy). Therefore, cyclicity can be described in terms of autocyclic or allocyclic on the basis of its origin.

According to Beerbower (1964), autocycles are produced by processes that only take place within the basin of deposition. Tides and storms are examples of the types of processes that create autocycles. Responses to autocyclic processes tend to be local and may range from millimeter-scale ripple migration to regional-scale events such as delta switching. Unlike allocyclic processes, autocyclic processes tend to be relatively instantaneous geologic events that are random in both time and space, and they contain few interregional feedback mechanisms. As a result, autocyclic processes are aperiodic. Because effects are local and mainly involve changes...
in energy, autocyclic processes generally result in changes in physical sedimentology, and, more often than not, they do not appear to result in significant changes in chemical sedimentology.

Allocyclic processes are normally regional in extent, and they are brought on by factors that act across an extensive area, mainly due to changes in relative sea level. A relative change in sea level can be either global or regional or even localized. Eustatic sea-level rise generally refers to a global rise in sea level, such that the resulting changes should be seen on all continents at about the same time. Global or eustatic sea-level changes may be caused by changes in the volume of water in an ocean or by changes in the volume capacity of the ocean basins. Changes in the volume of water in ocean basin can be caused by climatic factors such as glacial accretion or wastage and the amount of water stored in lakes or aquifers. On the other hand, changes in the volume of ocean basins are commonly related to sea-floor spreading rates.

Regional sea-level changes are typically caused by tectonic activity related to deformational loading in an orogeny or due to fault movement, either of which may generate uplift or subsidence. Sometimes, regional tectonism may also exert influence on the global sea level. In contrast, localized sea-level fluctuation typically reflects relative sea-level change related to local fault movements that cause nearby land subsidence or uplift, but it has no significant impact on global sea-level change.

Relative sea-level (e.g., Fig. 2.1) is measured between the sea surface and a local moving datum, such as basement or a surface within the sediment pile. Relative sea-level rises are mainly due to subsidence and eustatic sea-level rise. Eustasy and subsidence rate together control the amount of space available for sediment accumulation, termed accommodation space.
**Fig. 2.1 Vail sea-level curve showing major cycles of relative sea-level changes.** Note the two large first-order cycles, as well as the second- and third-order cycles that comprise them. The curve at the approximate time of upper Tanglewood deposition is shown by the arrow. Numbers at the left of the column are the approximate ages in millions of years (Ma) of period boundaries (adapted from Vail et al., 1977). The age at the arrow is approximately 453 Ma (Ogg et al., 2008).

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<th>GEOLOGIC PERIODS</th>
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<th>RELATIVE VARIATION OF SEA LEVEL</th>
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Numbers at the left of the column are the approximate ages in millions of years (Ma) of period boundaries (adapted from Vail et al., 1977). The age at the arrow is approximately 453 Ma (Ogg et al., 2008).
Tectonic processes, which are believed to cause changes in the volumetric capacity of the basins, are comparatively slower than climatic processes. Therefore, long-term changes in sea level are commonly attributed to tectonic processes, whereas shorter-term changes are mostly attributed to changes in the volume of water in the ocean due to climatic factors (Haq and Schutter, 2008).

Orders of Magnitude in Sedimentary Cycles

Earth scientists have been interpreting changes in sea level on the basis of stratigraphic data for more than a century. The stratigraphic record is a composite of superimposed sedimentary cycles of various orders of magnitude. There are basically six orders of cycles used to create sea-level curves to represent sea-level changes. These cycle orders can broadly be divided into macro-scale cycles and field-scale cycles.

Macro-scale cyclicity

Macro-scale cycles normally cannot be observed in a single outcrop. Cycles of the first and second order are macro-scale cycles (Fig. 2.1), comprising successions of considerable thickness (100 m up to several km). These cycles represent long time periods, usually between 10 Ma and more than 100 Ma. Long-term tectonic processes usually create these kinds of cycles. Therefore, crustal extension, lithospheric flexure, ocean spreading, subduction, thermal contraction as well as other tectonic processes can have important impact on the development of macro-scale cycles. These cycles are distinguished by long persisting onlap of coastal sediments onto continental margins (Watts et al., 1982; Sheridan, 1987). Even though tectonic processes generally operate at regional scales, they can exert influence on a global scale because regional
processes change the total volume of the ocean (Vail et al. 1991). In contrast, depositional cycles of higher orders are caused by more local changes in ocean-water volume (glacio-eustatic cycles). First-order continental-flooding epochs or second-order transgressive-regressive facies cycles are examples of macro-scale cycles.

**First-order cycles**—With a duration in the range of 100—200 million years, first-order cycles have been related to the changes in the volume of ocean basins due to supercontinent formation and dispersal (Vail et al., 1977). During supercontinent formation, the sea-floor spreading rates decline and mid-oceanic ridges subside; therefore, ocean volume increases, which ultimately lowers global sea level. Similarly, high rates of sea-floor spreading during continental dispersal lower the volume of ocean basins, thereby increasing global sea level. The upper Tanglewood study interval is included in part of a Paleozoic first-order cycle about 300 Ma years long and is part of an Ordovician–Silurian secondary cycle that is about 69 Ma long (Fig. 2.1).

**Second-order cycles**—Second-order cycles of sea-level change are related to changes in sea-floor spreading rates (Vail et al., 1977). These cycles are produced as a result of change in the volume of oceanic ridges and have durations of 10 to 80 million years (Mitchum, 1977).

**Field-scale cyclicity**

Field-scale cycles can generally be observed in a single exposure. Field-scale sequences represent the typical outcrop cycle, several meters to tens of meters thick. Many of these field-scale marine cycles are interpreted today as representing global and relative sea-level changes on the order of 100’s of thousands to several million years in duration.
Third-order cycles—These cycles have durations of 1 to 10 million years. This order of sea-level change is not fully understood. Originally it was thought that glacio-eustacy controlled these sea-level changes, but glaciers form and retreat far too rapidly, in only tens of thousands of years. So, these cycles are too long to attribute to glacio-eustasy. Similarly, this scale of cycle is too short to attribute to most tectonic events that affect the volumes of basins.

Fourth-order cycles—Fourth-order cycles represent eustatic variations within the larger eccentricity cycles of the Milankovitch band. These cycles have durations of 200 to 500 thousand years. The so-called parasequences of the Vail nomenclature as well as some of the coal cycles in cyclothems fall into this category.

Fifth-order cycles—This order of sea-level variation is probably caused by the shorter periodicities of the Milankovitch frequency band. This cycle has durations of 10-200 thousand years.

Sixth-order cycles—Sea-level changes with higher frequencies than the Milankovitch variations (less than 10 thousand years) are sixth-order cycles. These are typically local autogenic cycles that relate to specific conditions within a basin. Conditions differ from basin to basin, and hence, must be defined for each basin.

The expression of all these cyclic phenomena in sedimentary sequences is significantly modified locally by various depositional environments. It is the aim of sequence stratigraphy to divide the stratigraphic record into genetically related units via their sequence boundaries and to investigate the processes which have formed them.
Late Ordovician Sea-level Curves

The Ordovician Period saw the highest sea levels of Paleozoic time (Fig. 2.1), and the lowest continental relief. Sea level rose more or less continuously throughout the Early Ordovician time, levelling off somewhat then continuing to rise during the rest of the period (Fig. 2.1). Although local episodes of sea-level decline occurred, overall sea-level rise continued until early Maysvillian time when a global decline began (Fig. 2.2). Early parts of Late Ordovician time record the highest sea levels (Fig. 2.2) of the Paleozoic (e.g., Miller et al., 2005; Haq and Schutter, 2008; Ross and Ross, 1995). According to Ross and Ross (1995), the base of the Upper Ordovician sequence in North America is marked by a major transgression that shifted sediments from siliciclastics to interbedded carbonates and dark shales.

Following the beginning of an overall transgression in the late parts of Middle Ordovician time, it appears that sea levels continued to rise up to middle parts of the Katian (early Maysville time) (Fig. 2.2). After mid-Katian time, sea level falls for the rest of the Ordovician, a decline which was related to the rapid onset of glaciation on Gondwana during Hirniantian time.

Relatively few sea-level curves are available for Ordovician time. More recent sea-level curves covering the Ordovician Period come from Avalonia (Woodcock, 1990), Laurentia (Fig. 2.2) (Ross and Ross, 1992, 1995), western Gondwana (Herdia and Beressi, 1995), and Baltoscandia (Nielson, 2004; Dronov, 2005). The most detailed Ordovician curve is based on Balloscandia (Nielson, 2004), and therefore, does not necessarily reflect global patterns (Fig. 2.2).
Fig. 2.2 Late Ordovician sea-level curve. Late Ordovician sea-level curve showing a comparison of North American (from Ross and Ross, 1992, 1995), and Baltoscandian (from Nielson, 2004) sea-level curves. The North American curve is adjusted approximately to the timescale used in Nielson (2004), Clepper (2011). Inferred correlations between the highstand and lowstand events are shown by the light-colored tie lines; less certain correlations also include question marks. The red box indicates the part of the Nielson curve that correlates with the Lexington Limestone (adapted from Clepper, 2011).

Several transgressive-regressive sequences are present in the lower part of the Blackriverian (Turinian) Stage (Ross and Ross, 1995). Throughout middle parts of the stage, sea level remained relatively high with several fourth- and fifth-order shallowing-upward carbonate cycles. The end of the Turinian Stage, however, is marked by two abrupt drops in sea level, the second of which exposed large areas of the shelf (Ross and Ross, 1995). The Shermanian (upper Chatfieldian) was a time of relatively high sea levels. Shermanian curves show numerous, generally well-defined fourth- and fifth-order depositional sequences. The top of the Shermanian, or base of the Eden, is usually a subtle regional unconformity at the top of a shoaling-upwards, shallow-marine carbonate sequence (Pope and Read, 1997b) and represents an abrupt shallowing.
Cincinatian-age rocks reflect four major transgressions and regressions of short duration; Holland (1993) and Patzkowsky and Holland (1993) recognize many smaller, fifth-order transgressions.

The highest Ordovician third-order depositional sequence in North America includes the time between the Turinian and the end of the Ordovician (Figs. 1, 2). This depositional sequence is internally more complex than those in the Lower and Middle Ordovician and reflects ramp-like deposits that are the result of repeated, abrupt major sea-level fluctuations. This is the pattern that suggests glacioeustatic sea-level fluctuations (Ross and Ross, 1995).

**Sea-Level Curve for the Lexington Limestone**

Clepper (2011) has developed a sea-level curve for the entire Lexington Limestone of central Kentucky (Fig. 2.3). The curve shows two third-order transgressive sequences, each divided into several fourth-order sequences. According to Clepper (2011), the end of her first third-order cycle is reflected in highstand system-track deposits represented by the Brannon tongue of the Tanglewood Member. Her final third-order cycle in the Lexington begins at the Sub-Sulphur Well discontinuity and extends upward into the Clays Ferry Formation (Fig. 2.3), which represents the highstand-system track. Each of the two third-order sequences contains several fining-upward, fourth-order cycles or parasequences. The deeper or fining-upward parts of each cycle are easily recognized in the curve. In the lower third-order curve, they include, in order of deposition from the base of the Lexington, the Logana Member, the Macedonia Bed of the Grier Member, the Cane Run Bed of the Grier Member, and the Brannon Member. These deep-water deposits all grade up into shallow-water, open-marine Grier deposits, with the
exception of the Brannon, which grades upward into the shoal complexes of the Brannon tongue of the Tanglewood. The second third-order sequence also contains four regressive parasequences. The first three parasequences all begin with the deposition of a tongue of the Clays Ferry Formation which grades upward into the shallow-water, open-marine deposits of the Millersburg Member. The final parasequence in the Lexington Limestone begins with a tongue of the Clays Ferry that grades upward into the shoal complex of the upper tongue of the Tanglewood Member, which is the subject of this study.

Overall, the sea-level curve used for this study is consistent with sea-level curves from other studies on Late Ordovician sea-level (Pope and Read, 1997b; Nielsen, 2004), confirming the idea that, while the Lexington Limestone is locally regressive in nature, it represents a largely deepening trend that reflects a probable response to flexural subsidence and transgression.

**Meter-Scale Cyclicity during Upper Tanglewood Deposition**

The cyclicity described previously in the Lexington Limestone represents largely third-to fourth-order cyclicity (Clepper, 2011), but cycles of higher order also exist. Cyclicity was studied in detail in upper Tanglewood Member in order to explain the smaller-scale cyclicity (meter-scale) present within the unit and to see if it could possibly be used for correlation. Upper Ordovician carbonates of the upper Tanglewood Member of Lexington Limestone in central Kentucky exhibit well-developed meter-scale cycles in most exposures (Figs. 2.4 and 2.5). Each cycle reflects a fining-upward or deepening trend from calcirudite/calcarenite to shales or mudstones. This trend indicates that these cycles are bounded by combined transgressive and
regressive surfaces above and below. Normally, these cycles begin with coarse-grained calcarenite and grade upward into shales and mudstones, but, in some places, cycles begin with calcirudite and grade upward into calcisiltites. Each cycle is composed of more than one facies or facies association.

**Fig. 2.3 Lexington Limestone sea-level curve.** Sea-level curve used for this study of the Lexington Limestone Formation. The blue line shows third-order sequences. The black line shows fourth-order sequences. The approximate positions of the different members and beds of the Lexington Limestone and the three discontinuities within the Lexington Limestone are shown. T = Turinian; R = Rocklandian; SBD = sub-Brannon discontinuity; SSWD = sub-Sulphur Well discontinuity; SMD = sub-Millersburg discontinuity (adapted from Clepper, 2011).
Repetitive, shallowing-upward stratal units are usually known as cycles. Van Wagoner introduced the term parasequence in 1985 for shallowing-upward cycles that are genetically related and bounded by marine flooding surfaces or other correlative surfaces. It is widely accepted fact today that deltaic parasequences or beach parasequences show coarsening-upward cycles, whereas tidal-flat parasequences show fining-upward cycles. In both cases, they have shoaling-upward features.

Parasequences form when the rate of generation of accommodation exceeds the rate of sediment supply to the basin. A relatively abrupt rise in relative sea level (caused by sediment compaction, tectonic subsidence or eustasy) generates the flooding surface that terminates the parasequence, and the cycle repeats. For example, in deltaic environments, sediments get coarser and coarser upward due to delta progradation; then delta switching occurs and because of subsidence and compaction of sediments with little sediment influx, a marine flooding surface develops. This is one cycle of deposition. Over time, another cycle will form when the delta switches back to the same place.

A periodic parasequence has regional continuity and forms in response to deposition during a global sea-level cycle. An episodic parasequence has limited lateral extent and forms in response to (for example) tidal-flat migration or delta switching.

The standard definition of a parasequence, which exhibits shallowing-upward cycles, cannot be applied to the cycles of the upper Tanglewood Lexington Limestone. In fact, we cannot define Lexington high-resolution cycles in terms of shallowing-upward cycles. Rather, the unit is defined by deepening-upward cycles. The boundaries of the upper Tanglewood cycles, the sharp bases of limestones, are transgressive surfaces superimposed on a regressive erosional
surfaces. Therefore, I believe that it is a small-scale sequence, because it has both a regressive surface and a transgressive surface (superimposed on regressive surface). A sequence is defined as a conformable succession of genetically related strata bounded by subaerial unconformities or their correlative conformities. Such “unconformities” in the upper Tanglewood are represented by subtle discontinuity surfaces just below upper and lower limestone beds (bases of yellow triangles in Fig. 2.4). Therefore, I use the term “sequence-like cycle” to represent high-resolution cycles in the upper Tanglewood (Fig. 2.4).

**Fig. 2.4 Upper Tanglewood Member in Nicholas County.** Upper Tanglewood Member in Nicholas County (38°25'07.95" N, 83°59'56.71" W), showing five sequence-like, fining-upward cycles in yellow color. The solid red lines are marine flooding surfaces.
Fig. 2.5 Relative sea-level curve on the basis of texture and associated cyclicity. Relative sea-level curve on the basis of texture and associated cyclicity in the Nicholas County section (38°25'07.95" N, 83°59'56.71" W). Solid red line is a fourth-order, relative, sea-level curve; black line is a composite fifth-order, relative, sea-level curve.
The Origin of Meter-Scale Cyclicity in the Upper Tongue of the Tanglewood Member

The origin of meter-scale cyclicity in carbonate successions has been a controversial topic for many decades, and many researchers have attempted to describe the cause of cyclic facies repetition in the geologic record (e.g., Fisher, 1964; Goldhammer et al., 1990; Bosence et al., 2009; Laya et al., 2013). Three basic mechanisms have been suggested: autocyclic, tectonics and eustacy. Autocyclic processes (e.g., Tucker and Garland, 2010; Schlager, 2005) reflect internal-basin controls and are considered to be the result of normal sedimentary mechanisms. Tides and storms are examples of the types of mechanisms that create autocycles. Autocycles show limited stratigraphic continuity. On the other hand, the other two processes are allocyclic processes and reflect controls outside the basin due to tectonics and eustasy (Milankovitch rhythms/orbital forcing).

Some researchers (e.g, Grotsch, 1996) have shown that the cyclic successions are the result of long-term eustatic variations in sea level, and they have attempted to correlate cyclic successions on a global scale. In contrast, other researchers (e.g, Bosence et al, 2009) have described cyclic successions as the products of both allocyclic and autocyclic processes operating at the same time. In addition to the traditional variation of sea level and/or tectonics as allocyclic explanations for meter-scale cycles, Laya et al. (2013) have suggested that climatic changes are also very important in generating cycles. High-frequency cycles, however, have been commonly attributed to Milankovitch cycles related to eccentricity, obliquity and precession cycles (e.g., Weber et al, 2001; Schwarzacher, 1993). Because shallow-marine carbonate environments are environmentally sensitive to low-amplitude sea-level changes, many
researchers have performed cyclostratigraphic analysis in such settings (e.g., Lever, 2004; Schwarzacher, 2005).

Cyclicity in the upper Tanglewood Member is identified on the basis of detailed description and comparison of many outcrops, because use of any single outcrop is insufficient to draw conclusions about cyclicity. Clepper (2011) developed a sea-level curve for the entire Lexington Limestone (Fig. 2.3) in which she identified third- to fourth- order cycles. Therefore, in the upper Tanglewood Member (smaller subset in the Lexington), the basic building blocks of the cyclicity would be expected to be 4th- or 5th-order parasequences (Fig. 2.5).

By definition, parasequences are typically upward-shoaling successions that are bounded by flooding surfaces, which form in response to the relatively rapid rise of sea level. Normally, shallow-water, shoreline, shelf and platform deposits display shallowing-upward cycles, which are characterized by changes in composition, grain-size, fossils and microfacies. While most parasequences record shoaling-upward conditions, they do not all necessarily display coarsening-upward sediments, as is the case for tidal-flat parasequences. In the definition of parasequences, little explanation is given for the possibility of transgressive deposition, but transgressive or deepening-upward cycles do exist (e.g., Turner et al., 2012). Obviously, the origin of transgressive cycles must be somewhat different than shoaling-upward cycles, but in the definition of parasequences, transgressive deposits are inadequately addressed (Arnott, 1995).

One mechanism for explaining such fining-upward cyclicity is tidal cyclicity. While discussing tidal autocyclicity, it is necessary to understand about the progradation of tidal flats over the subtidal carbonate factory, thereby halting sediment generation until sufficient water depth is attained by slow subsidence to regenerate carbonate production. However, the upper Tanglewood Member was deposited as a shoal, which never produced tidal-flat sediments,
because most of the time, it remained submerged during deposition. It might have been exposed above sea-level occasionally during low tide, but it did not produce extensive tidal-flat sediments. Moreover, the upper Tanglewood cycles can be correlated across tens of kilometers, even into the deeper-water Kope and Clays Ferry formations. Therefore, the autocyclic model involving tidal-flat progradation must be discarded as a plausible mechanism for upper Tanglewood cycles.

Repeated episodic subsidence of the basin is another mechanism that can generate asymmetric meter-scale, deepening-upward carbonate cycles (Cisne, 1986; Hardie et al; 1991). However, examples due to episodic subsidence normally come from tectonically active settings (Yeats, 1978; Bull and Cooper, 1986), and they are poor analogous for successions in ancient passive margins or cratonic areas like the Lexington Platform. Moreover, as I will suggest later, upper Tanglewood shoal deposits are apparently related to structural uplift along basement fault zones. Therefore, it is unlikely that they would have experienced episodic subsidence at the same time.

In contrast, however, glacio-eustatic sea-level oscillations could explain the meter-scale, fining-upward cycles in a simple and convincing manner (e.g., Grotzinger, 1986; Goldhammer et al., 1987; Jenette and Pryor, 1993; Kim and Lee, 1998). Sea-level fluctuations due to orbitally forced Milankovitch cycles resulting in the decay and formation of glaciers with periods of ~20, 41, 100, and 400 kyr have been well-documented throughout the Quaternary (Hays et al.; 1976). Also, ancient meter-scale cycles have been increasingly attributed to eustatic sea-level oscillations in the Milankovitch band (Grotzinger, 1986; Herbert and Fischer, 1986; Goldhammer et al; 1987; Koerschner and Read, 1989, Erlick and Read, 1991; Osleger and Read, 1991; Jennette and Pryor, 1993). I believe that the cyclicity observed in upper Tanglewood
Member (Figs. 2.3–2.5) is related to allocyclicity, mainly eustatic fluctuation, and has correlative value for the reasons described below.

Long-term subsidence of the craton is an inherent, autocyclic possibility, but such rates are slower than the production rates of carbonate sediments on shallow-marine platforms, which allow thick successions of platform carbonates to accumulate. Eustatic sea-level changes, on the other hand, may recur repeatedly at high frequencies and at rapid rates, which can outstrip carbonate-production rates. The resulting stratigraphic record of such rapid repetitions would be one of cyclic accumulation, which is what we see in the upper Tanglewood succession (Fig. 2.4). Moreover, the deposition of the Tanglewood Member, and specifically the upper Tanglewood Member, has been related to episodic movement on local basement faults. In fact, Ettensohn et al. (2002a, 2004) have described repeated reactivation of these faults during Lexington deposition, and the upper Tanglewood Member shoal complex has been described as a local regressive sequence deposited during an ongoing transgression elsewhere because of uplift on reactivated basement faults (Ettensohn et al., 2004). Because of this, the upper Tanglewood Member tongues into deeper water units like the Clays Ferry, Kope and Point Pleasant formations in all directions (Fig. 1.1). Hence, it could be argued that the cyclicity was related to episodic uplift on the structures, such that the lack of any substantial regressive phase after marine flooding surfaces might be related to sudden intermittent uplift of the shoal area due to fault reactivation. On the other hand, the absence of regressive phases in the cycles may simply reflect that fact that they were eroded during the transgressive phase and lack the same preservation potential. Added to this is the fact that the same type and number of cycles are present in deeper-water lithofacies equivalents of the upper Tanglewood (based on biostratigraphy and correlation with deformational event horizons) in the Clays Ferry and Point
Pleasant formations (Fig. 1.1), which were deposited beyond the influence of basement structures. Hence, the same small-scale sequence-like cycles, with a transgressive phase followed abruptly by a low-stand phase without any intervening regressive phase, are present in deep- and shallow-water settings, both on and away from the structurally controlled platform, strongly pointing to eustatic causes.

I concur with several other workers in the area (e.g., Jennette and Pryor, 1993; Holland et al., 1997; McLaughlin and Brett, 2004) that the cycles are most likely eustatic in nature. Such meter-scale cycles have been linked to the Milankovitch orbital rhythms (e.g., Tucker and Garland, 2010), but the possibility of glacio-eustasy cannot be ignored. Workers like Pope and Read (1997), Pope and Steffen (2003), Saltzman and Young (2005) and Elrick et al. (2013) have suggested that Late Ordovician glaciation may have begun by early Katian (late Caradoc; Chatfieldian) time and continued to climax in latest Ordovician (Hirnantian) time. Using biostratigraphy (Sweet, 1979) correlated into an absolute time scale (Cooper and Sadler, 2012), I estimate that the upper Tanglewood represents a period of deposition no more than 500 Ka in length. If we assume that each fifth-order cycle was approximately equal in length, then each of the five cycles (Fig. 2.5) represents about 100 Ka in duration. Not only is 100 Ka about the duration of the eccentricity cycle (Tucker and Garland, 2010), but is also about the length of some possibly analogous Pleistocene glacial events (Pillans and Gibbard, 2012), which may also reflect orbital rhythms. So whether upper Tanglewood cycles (Fig. 2.5) reflect purely orbitally driven climate rhythms, glacio-eustacy, or some combination, I am uncertain, but what I can say is that they are about 100 Ka in duration and are most likely “eustatic” in nature.
Chapter 3—Cross Sections

Introduction

For this study, a total of 45 outcrops and 9 cores were measured (Fig. 1.2) and described (see Appendix A). The locations of the outcrops and cores were plotted on a map, and chronostratigraphic correlations were performed on the basis of cyclicity and deformation horizons across different parts of the upper Tanglewood Member with the goal of covering as much of the member as possible. As a result, a grid of seven section lines was developed (Figs. 3.1–3.8). Correlations among the strata along these lines allowed for the observation of spatial variations in facies associations and thicknesses.

Section line A–A’

Section line A-A’ is a north-south-trending line on the western flank of the upper Tanglewood shoal complex. This line extends from just north of Lawrenceburg in Anderson County along US Highway 127, continuing northward of Frankfort in Franklin County to Monterey in Owen County (Figs. 3.1 and 3.2). The line cuts across the main part of the Upper Tanglewood shoal complex, showing apparently greater thicknesses to the south and abrupt thinning near the Franklin-Owen county line (Fig. 3.2).

Clearly, sections in the northern parts are thinner compared to those in the southern sections, although the two southern sections are incomplete. Similarly, the sections to the north, as a whole, are more shaley in comparison to those in the south. The second section from the left is near Lawrenceburg (38° 3' 30.90" N, 84° 55' 15.30" W) and is the coarsest section in the study
area, containing fewer shale intervals than anywhere else in the study area. Moreover, the southern sections are underlain by the Clays Ferry Formation whereas the northern sections are underlain by the Millersburg Member. Both the northern and southern sections are overlain by the Clays Ferry Formation.

*Fig. 3.1 Section lines. Map showing the locations of the section lines used in this study.*
In general, the first cycle is of similar thickness, about 1.5 feet (0.45 m) in all sections. This cycle contains a deformed bed, which is absent in some sections. Cycle 2 is greatest in thickness in the Lawrenceburg section (Anderson County) and along this line attains a thickness of 15 feet (4.57 m). In this section, this cycle is more massive and coarse-grained than anywhere else in the entire upper Tanglewood. However, the same cycle is only 2 feet (0.6 m) in thickness in the Peaks Mill section (38° 18' 29.40" N, 84° 50' 44.20" W) (second from the northern end) of Franklin County. Though this cycle has two distinct deformed beds in most sections of the upper Tanglewood, the same two horizons were not observed near Peaks Mill or in the Lawrenceburg section. In both of these sections, only a single trace of deformation is apparent. Similarly, Cycle 3 gradually becomes thinner and thinner toward the north. It also contains two distinct deformed
beds, which are also absent in the Lawrenceburg and Peak Mill sections. Cycle 4 is not complete in the southern sections, but it is more shaley than the other cycles in the northern sections where it is complete. This cycle does not contain any deformed beds. The last cycle (Cycle 5) begins with coarse-grained massive beds followed by more muddy sediments and a final bed of coarse-grained calcarenite. Though there is one deformed bed along other section lines in Cycle 5, this deformed horizon is not observed in this section line. Overall, this section is less shaley in the lower cycles (2, 3) and in upper Cycle 5 and more shaley in Cycles 1 and 4.

Section line B–M–N–B’

This section line (Figs. 3.1 and 3.3) is a southeast-northwest-trending line through the middle part of the upper Tanglewood shoal complex. This line extends from an upper Tanglewood equivalent in the Clays Ferry Formation from Clays Ferry in northern Madison County to north of Georgetown in Scott County along Highway I-75. In the middle part around Georgetown, the section is thicker than in sections from the northern and southern parts. The upper Tanglewood equivalent in the Clays Ferry Formation is more shaley than the more typical upper Tanglewood Member as shown in Figure 3.3. However, the upper Tanglewood equivalent in the Clays Ferry Formation still appears to be quite distinct from the more typical Clays Ferry Formation, because the proportion of calcarenite to shale is higher in the upper Tanglewood equivalent.

Like other sections in other lines, Cycle 1 is more or less of equal thickness along this line, about 1.5-feet thick (0.45 m). Cycle 1 contains one deformed horizon in the northernmost section which is not observed in other sections. Cycle 2 contains two distinct deformed beds elsewhere except in the northernmost section, where only one deformed bed is observed. This
cycle is thinner in northern parts than in southern parts of the section line. Likewise, Cycle 3 is also thinner in the northern and southern boundary sections. Though this cycle contains two deformed horizons in other section lines, only one is observed in this section line, particularly in the southernmost section. In northern parts, deformed horizons are not observed in Cycle 3.

Compared to other columns, Cycle 4 is more shaley and does not contain any deformed beds. However, the northernmost section contains more massive, coarse-grained calcarenites than elsewhere along the section line. Except where the section is incomplete and in the northernmost section, only one very distinct deformed horizon is observed in Cycle 5. In this cycle, lower and upper parts contain more calcarenites and calcirudites, whereas middle parts tend to be more shaley.

**Section line C–C’**

This correlation line is also a north-south trending line on the eastern flank of the upper Tanglewood shoal complex (Figs. 3.1 and 3.4). This line extends from just east of Winchester in Clark County to northern parts of Nicholas County. The southern section near Winchester is very well-exposed along I-64, whereas northern sections are all well-exposed along US Highway 68 (1455). The section just east of Winchester is on the southern margin of the upper Tanglewood shoal complex. To the south of that section, the upper Tanglewood grades southward into the Clays Ferry Formation. The upper Tanglewood gets gradually thicker toward the north in this section line. Moreover, the section at the boundary of the upper Tanglewood with the Clays Ferry Formation at Winchester is far more shaley than its counterparts in other sections.
Fig. 3.3 Correlation among sections along line B–B’.

Unlike other sections in other section lines, Cycle 1 attains a maximum thickness of 10 feet (3 m) along this line in Nicholas County (38° 21’ 57.48” N, 84° 2’ 20.46” W) and a distinct deformed horizon in Cycle 1 is observed in all sections except in the southernmost section near Winchester. Cycle 2, in contrast, is more or less of equal thickness everywhere along this line and contains deformed horizons in all sections except in the Bourbon section (38° 16’ 20.24” N, 84° 10’ 22.62” W), which is second from the southernmost section. In the most northern section (38° 24’ 52.65” N, 84° 0’ 15.07” W), two distinct deformed beds are obvious, but in other sections, only one thick deformed horizon is distinct. Cycle 3 contains one or two deformed horizons and contains more calcarenites than the other cycles in this column. The ratio of coarse-
grained sediments to clay is much higher in this cycle. Cycle 4 increases in thickness toward the northern part of the line. This cycle contains no deformed horizons and is more shaley in comparison to the other cycles. In Cycle 5, only one deformed horizon is present, and it is thick and very distinct. The lower and upper parts of this cycle are more calcarenitic and calciruditic, whereas middle parts are more shaley. This cycle is overlain by the Kope Formation in the northern two sections and by the Clays Ferry Formation in the southern two sections.

**Section line A’–B’–O–P–C’**

This section line is east-west-trending on the northern flank of the upper Tanglewood shoal complex (Fig. 3.1). This line extends from Monterey in Owen County to northern parts of
Nicholas County (38° 24' 52.65" N, 84° 0' 15.07" W). Thickness of the sections increases toward the east.

The lower cycle, Cycle 1, is thinner than the other cycles and contains one or two deformed horizons, but deformation is absent in the middle two sections. Cycle 2 is thicker, reaching 10 feet (3 m) in the westernmost Monterey section, where it contains two thick, distinct deformed horizons. In the central part, however, only one deformed horizon is observed. Eastern parts of the section are more shaley than western counterparts. Cycle 3 is more massive and the proportion of calcarenite to shale is higher than in the other cycles. Cycle 3 contains one to two deformed horizons, but in the middle three sections, no deformed horizons are observed. Cycle 4 is more shaley in nature and contains no deformed horizons. Its thickness is maximum in the eastern section, where it reaches a thickness of 17 feet (5.18 m), but it thins progressively to the west, where it is only 4-feet (1.21-m) thick. The uppermost cycle, Cycle 5, is only 1.5-feet (0.45-m) thick in the most western section, whereas it is about 8-feet (2.4-m) thick in easternmost section. Though this cycle contains one deformed horizon at the base in the other section lines, no deformed horizons are observed along this line. In the same way as other section lines, lower and upper parts of this cycle are more massive and calcarenitic, whereas middle parts are more shaley.

Section line A–N–P–C’

This section line extends more or less diagonally from the southwest corner of the shoal complex in Anderson County to the northeast corner in northern Nicholas County (Figs. 3.1, 3.6).
As in the other section lines, Cycle 1 is thinner than the other cycles, and it exhibits one or two, thin, deformed horizons at the northeast and southwest ends of the section. Cycle 2 is thicker in the southwest end of the section line and progressively thins toward the northeast. This cycle contains two distinct deformed horizons in every section along this line. In Cycle 3 of this correlation line, the proportion of calcarenite/calcirudite to shale is much higher, and this cycle contains mainly cross-bedded calcarenite and massive calcirudite with occasional shale partings. It contains two distinct deformed horizons in the northeast and southwest ends of the section. The thickness of Cycle 4 is greater in northeast corner. Like other sections in the other
correlation lines, this cycle is more shaley. That means the proportion of calcarenite/calcirudite
to shale is the lowest of any other cycle. This cycle also does not contain any deformed horizons.

Like other sections in other correlation lines, the lower and upper parts of Cycle 5 are
more massive and contain a higher proportion of calcirudite/calcarenite. In only one of the
sections along this line is a distinct and thick deformed horizon observed in lower parts of the
cycle.

**Fig. 3.6** Correlation among sections along line A–C'.
**Section line A–B–C**

This section line more or less parallels the southern flank of the shoal complex from Anderson County in the west to Clark County in the east (Figs. 3.1 and 3.7). This section line also crosses the Clays Ferry Formation at the type locality in Madison County.

In this section line, Cycle 1 is thinner than the other cycles, and it exhibits two thin deformed horizons on the western end of the section. However, no deformed horizons are present in this cycle in other sections, including the Clays Ferry Formation type locality. Cycle 2 is thicker on the western and eastern ends and thinner in the middle of the section line at the Clays Ferry type locality. This cycle contains two distinct deformed horizons in every section along this line except at the eastern end in Clark County, where only one deformed horizon is observed. At the western end of this correlation line, Cycle 3 contains mainly cross-bedded calcarenite and massive calcirudite with two distinct deformed horizons. In contrast, the remaining sections along this section line contain much higher percentages of shale; the thickness of this cycle is also less and only one deformed horizon is observed (Fig. 3.7). The thickness of Cycle 4 is less on the eastern end. Like other sections in other correlation lines, this cycle is more shaley, which means that the proportion of calcarenite/calcirudite to shale is the lowest of any other cycle. This cycle also does not contain any deformed horizons. Cycle 5 is not observed on the western end. However, in the other parts, the lower and upper parts are more massive and contain a higher proportion of calcarenite. In only one of the sections (type locality of Clays Ferry Formation in Madison County) along this line is a thin deformed horizon observed in lower part of the cycle (Fig. 3.7).
Overall, the upper Tanglewood equivalent within Clays Ferry Formation does not appear to be any different than the upper Tanglewood at the eastern end in Clark County in terms of the proportion of shale to calcarenite. However, it is more shaley than its counterpart at the western end of this section line.

**Fig. 3.7 Correlation among sections along line A–B–C.**

**Section line B–P–T**

This section line runs nearly north-south from the Clays Ferry Formation in Madison County to the Point Pleasant Formation in Bracken County (Figs. 3.1 and 3.8). This section line,
therefore, contains the upper Tanglewood equivalent in both the Clays Ferry and Point Pleasant formations, as well as upper Tanglewood in the shoal complex. While observing the section line, it appears that the upper Tanglewood equivalent in the Point Pleasant Formation is more massive and less shaley than its counterpart in the Clays Ferry Formation. The Point Pleasant Formation appears more or less identical with the upper Tanglewood in the shoal complex in terms of the ratio of shale and calcarenite (Fig. 3.8). Cycle 4 is more massive, thicker and contains less shale in the Point Pleasant Formation than in its counterparts in other sections along this section line.

Fig. 3.8 Correlation among sections along line B–T.
Summary

As shown on Figures 3.1 and 3.6, correlations that pass from the middle part of the upper Tanglewood Member in a southwest-to-northeast direction (section line A–C’) clearly show that the thickness of the shoal is somewhat less in the middle part of the section line than in the southwest and northeast corners. Similarly, in the northeast corner, facies associations are characterized by low-angle cross bedding with occasional shale partings. Correlation along the western flank (A–A’) (Figs. 3.1 and 3.2) gives us the clear impression that this side was deposited as a typical shoal and beach with little reworking, especially in the middle of the section because of its massive nature and lesser mud content. However, shale content gradually gets higher in the same horizon as we go toward the northwest and southeast corners at the boundary of the upper Tanglewood Member with the Clays Ferry or Kope formations. Despite that, the fourth cycle is deeper everywhere on the western flank. So, it can be argued that during the fourth cycle of deposition, the western flank of the shoal had the deepest conditions, whereas during the third cycle of deposition, it had the shallowest conditions, which during shoal development may have been periodically exposed at low tide. Also, while observing the section line B-P-T (Figs. 3.1 and 3.8), it is clear that the Point Pleasant Formation is very similar to typical upper Tanglewood Member in terms of shale-to-calcarenite ratio. Therefore, there is a possibility that the Point Pleasant was also deposited as a typical shoal in a shallow-water environment, unlike the Clays Ferry Formation, which was deposited in a deeper-water environment.
Chapter 4—Facies Association

Introduction

Initially, the distribution of the upper Tanglewood Member (Fig. 1.2) was mapped by describing sections in the field and core data. Where exposures were unavailable, distribution was mapped using published geological quadrangle maps. In order to examine vertical and lateral facies changes across the unit distribution, stratigraphic columns were drawn up for each exposure and core (Fig. 1.3) (Appendix A). Each columnar section shows overlying and underlying units where visible, as well as predominant carbonate grain sizes using the Grabau classification (Grabau, 1903), sedimentary structures, paleocurrent patterns, and fossils. I use the term “facies association” (e.g., Santos et al., 2015) to include a combination of two or more facies, which comprise broader depositional complexes. The approach used here has been to distinguish facies associations, interpret the processes which formed them, and reconstruct analogous local environments in the context of the geology of the Tanglewood buildup. In the following section, four discrete lithofacies are described from the upper Tanglewood facies association, each of which is interpreted in terms of depositional environment.

My hypothesis, based on previous interpretations (e.g., Hrabar et al., 1971; Cressman, 1973), is that the Upper Tanglewood Member represents a shoal complex. Within such a complex, different kinds of environments existed, depending on water depth and related-energy conditions (e.g., Aigner et al., 2007; Major et al., 1996). The shallowest portion of a shoal is completely different than the deepest part of the shoal in terms of sediment texture and structure, though we call all these environments collectively a shoal complex. Although much of the
Lexington Limestone has been described in terms shelf- or ramp-type environments (e.g., Ettensohn et al., 2002c, d), the concept of shoal-complex environments as I use them, is different than the typical shelf environment, in that an isolated shoal is surrounded by deeper waters in all directions. However, linear shoals do exist in nature. In fact, an isolated shoal complex is characterized by environments and facies that occur in the same order in all directions, because they are controlled by depth. In other words, the facies association develops in a more or less concentric pattern (Fig. 4.1). In contrast, carbonate shelves or ramps typically have linear shorelines, and facies develop in one direction as depth increases gradually toward offshore areas (Fig. 4.2). Hence, shelves or ramps typically have linear or unidirectional patterns of facies. So the common terms typically applied to describe ramp- or shelf-carbonate environments, such as offshore, upper shoreface, and foreshore have been replaced in this study by the terms offshoal, upper shoalface and foreshoal, to differentiate them from carbonate-ramp and shelf environments (Fig. 4.1). Using this approach, I have distinguished four main lithofacies in the upper Tanglewood shoal-complex association, each of which is interpreted to represent a depositional environment; these environments range from shoaltop/beach to deeper offshoal facies (Figs. 4.1, 4.3). For context, the deeper offshoal facies was probably no deeper than 30–50 m (e.g., Brett et al., 2015). Each of the lithofacies and corresponding environmental facies is briefly described below.

**Massive calcirudite/calcarenite lithofacies**

This facies is composed of massive, light-gray, medium- to thick-bedded (10-100 cm) fossiliferous calcirudite and coarse-grained calcarenite (Figs. 4.3, 4.4); shale partings are largely absent. The common fossils primarily comprise brachiopods, bryozoans, crinoids, stromatoporoids and corals. Although most of the fossils are broken and abraded, massive, in-
situ, unbroken colonies of bryozoans, corals or stromatoporoids occur locally. The calcirudites or calcarenites are poorly sorted and occasionally display low-angle, planar cross-bedding with wave ripples, but more commonly, the unit is massive or planar-bedded. The lithofacies is gradational at its tops and bottoms and is best developed in Cycle 3. Because of its massive beds, the facies is more resistant than the surrounding facies, causing it to be very conspicuous in outcrop (Fig. 4.4).

![Figure 4.1](image)

**Fig. 4.1** Generalized environmental model for carbonate sedimentation on a shoal complex. Generalized environmental model for carbonate sedimentation in the upper tongue of the Tanglewood Member.

**Lithofacies interpretation**

This lithofacies is best developed in Cycles 2, 3, and 4, where it is typically no more than a meter thick. Its best development is in the outcrop area in southwest Anderson County, where
its thickness is in excess of three meters (Fig. 4.4). The abundance of calcirudites and coarse-grained calcarenites, as well as the absence of muds, suggests a very shallow, high-energy setting. Megaripples with wave lengths of 0.6 m reinforce the notion of a high-energy setting. The presence of low-angle, planar crossbeds and wave ripples suggests that the area was reworked periodically by wave processes. For this lithofacies, a shoal-top or foreshoal setting between high and low tides, in which coarse skeletal sands and larger fossil debris were moved back and forth by passing waves, has been interpreted. Ephemeral beaches apparently developed but were short-lived, as no long-term exposure features were developed. Communities of tabulate corals, stromatoporoids and bryozoans developed locally on deeper parts of the shoal, but were ultimately buried by migrating sands. We interpret the massive calcirudite/calcarenite facies to represent a shoal-top/foreshoal setting (Figs. 4.1 and 4.3).

**Fig. 4.2** Generalized cross-sectional profile of the beach and nearshore zone. Generalized cross-sectional profile of the beach and nearshore zone along a ramp, showing also the principal zones of wave activity (adapted from Boggs, 2001).
Fig. 4.3 Four major lithofacies and their interpreted depositional environments. Four major lithofacies and their interpreted depositional environments in the upper tongue of the Tanglewood Member.

**Massive calcirudite/calcarenite lithofacies:**
Thick, massive beds of fossil-fragment calcirudite and coarse-grained calcarenite with no to few shale partings; low-angle, planar cross beds; current ripples and megaripples; occassional massive, in-situ communities of corals, stromatoporoids and bryozoans.

**Interpretation:** Shoal-top/foreshoal environments.

**Crossbedded calcarenite and shale lithofacies:**
Thin to medium beds of medium-grained, cross-bedded or planar-bedded calcarenite with shale partings; bidirectional crossbedding with current and wave ripples; local lags of coarse fossil debris; parts of fining-upward sequences.

**Interpretation:** Foreshoal/upper shoalface environments with wave and tidal reworking.

**Nodular to wavy-bedded calcarenite lithofacies:**
Thin to medium, nodular, irregularly bedded with abundant shale intercalations; crude grading in some beds; abundant fragmented fossils and bioturbation.

**Interpretation:** Lower shoalface environmental setting; above storm wave base but below normal wave base.

**Calcisiltite and shale lithofacies:** Fine-grained calcarenites and calcisiltite with abundant shale intercalations; crude graded bedding; bioturbation present, but body fossils are rare.

**Interpretation:** Offshoal to platform-lagoonal environments, but above storm wave base.
Fig. 4.4 Massive calcirudite/calcarenite lithofacies in Anderson County. Massive calcirudite/calcarenite lithofacies in Anderson County, KY (38° 05' 45.45" N, 84° 55' 09.66" W). Red lines represent marine-flooding surfaces. Exposed here are the parasequences 2, 3 and 4. Parasequences 2 and 3 are largely composed of the massive calcirudite/calcarenite lithofacies developed in high-energy conditions.

Crossbedded calcarenite and shale lithofacies

Medium-grained, crossbedded or planar-bedded calcarenite with occasional shale partings is the main constituent of this lithofacies (Fig. 4.3). Bioclastic grains are moderately to well sorted, but lags of coarse, poorly sorted, fossil debris also occur. The lithofacies is characterized by unimodal, planar crossbedding, as well as bidirectional cross-bedding (herringbone cross-bedding), and both wave and current ripples are present (Fig. 4.3). Medium to thin bedding characterizes the facies, which is the most common facies in the upper Tanglewood. Typically, it occurs as parts of fining-upward sequences.
Lithofacies interpretation

This lithofacies shows a combination of wave and tidal features. Planar cross-bedding (Figs. 4.1 and 4.3), dipping in the offshoal direction, is common in this lithofacies and probably represents wave swash in a shoal environment along a low-angle to sub-horizontal depositional surface in the foreshoal/upper shoalface area (e.g., Clifton, 2006). Bidirectional cross-bedding also indicates that the tidal currents in this zone had significant influence. Prominent shale partings on many of the cross beds (Figs. 4.1 and 4.3) apparently represent slack-water deposition during high tides or deposition on deeper parts of the foreshoal. Lags of coarse, fragmented fossils in this zone might indicate that fossil debris was reworked and transported from the shoal in an offshoal direction where it was concentrated by waves and tides in low areas. Breaking waves and tides apparently generated local currents that kept the shoals supplied with reworked sediment. The crossbedded calcarenite and shale lithofacies is interpreted to represent a foreshoal/upper-shoalface setting (Fig. 4.1).

Nodular to wavy-bedded calcarenite lithofacies

This lithofacies is predominantly composed of white to gray, nodular- to wavy-bedded, medium- to fine-grained, skeletal calcarenites, separated by abundant shale partings or thin beds of fossiliferous shale (Fig. 4.3). Where bioturbation has not been severe, traces of crude grading are common in calcarenite beds. This facies commonly weathers to a characteristic rubble. Fragmented bryozoans, brachiopods, crinoids, corals, gastropods and clams are common. Many of the fossils are highly broken and abraded. Wave ripples are less common, but bioturbation is pervasive and has given rise to the nodular or wavy appearance of bedding.
Lithofacies interpretation

Although some wave activity is still apparent, it is relatively rare in this lithofacies, meaning that deposition largely occurred below fair-weather wave base. Though lesser amounts of wave energy were present, storm activity apparently generated periodic, high-energy conditions that left crudely graded tempestites. In this deeper, quiet setting, benthic communities were also established, but they were mostly short-lived because of storm agitation and disruption. In their study of Lexington Limestone environmental settings, Ettensohn et al. (2002c, d) described similar facies as having been deposited in an open-marine, intermediate-ramp setting, above storm wave base but below normal wave base. Walker (1984) has interpreted similar facies associations to have been deposited between a fair-weather wave base and a storm wave base. Although the sediments in this setting experienced far less influence from shorter-period fair-weather waves, they did experience repetitive storm activity (e.g., Galloway and Hobday, 1996). So the presence of alternating calcarenites and shales represents episodic storm activity, whereas the nodular appearance of bedding probably represents burrowing activity. The nodular to wavy-bedded calcarenite lithofacies is interpreted to represent a lower shoalface setting, below normal wave base, but clearly above storm wave base (Fig. 4.1).

Calcisiltite and shale lithofacies

This lithofacies is composed of thin, regular-bedded calcisiltites and fine-grained calcarenites with abundant interbedded shale (Fig. 4.3). Although the facies is less common, it can be found occasionally in the very upper or very lower parts of some upper Tanglewood sections. Crude graded bedding and parallel laminae occur locally, but ripples and crossbedding do not occur. Although bioturbation is present, fossils are rare and largely fragmental.
Lithofacies interpretation

The predominance of fine-grained lithologies and lack of traction-related sedimentary structures suggests a deeper open-marine setting, still above storm wave base, but in largely low-energy conditions where the fallout of fine, suspended materials predominated during fair-weather periods (e.g., Clifton, 2006). Those fine-grained limestone beds that are present apparently represent backflow deposition of distal tempestites (e.g., Aigner, 1982). Ettensohn et al. (2002c, d) described similar Lexington Limestone environments as deep-ramp, open-marine settings, whereas Santos et al. (2015) described similar lithologies as offshore to lower-shoreface facies represented by quiescent periods and episodic sediment supply. The scarcity of benthos in the facies may reflect soft bottoms. Moreover, this kind of deeper open-marine environment appears to have characterized the later stages of many cycles throughout the upper Tanglewood shoal complex. The calcisiltite and shale lithofacies is interpreted to represent an offshoal to platform-lagoonal setting with depths in the 30–50-m range (Fig. 4.1).

Facies distribution in time across the shoal complex

Assuming that each cycle as correlated (Figs. 1.3, 1.4 and 3.1-3.8) is largely coeval across the shoal, we determined the predominant facies (Fig. 4.3) at each measured section in each cycle and mapped the distribution of dominant facies, cycle by cycle (Fig. 4.5). Because the upper part of Cycle 5 is commonly eroded away in most sections, making it difficult to interpret, that cycle has not been analyzed in Figure 4.5.

The mapping in Figure 4.5 shows that the major shoal areas in the upper Tanglewood Member were concentrated in the southwestern and northeastern parts of the Tanglewood shoal
complex. Moreover, superposition of basement fault trends (Figs. 1.2, 1.11 and 1.12) on the facies maps shows that the coarsest shoal facies in each cycle generally coincide with structural trends (Fig. 4.5), suggesting a relationship. After a regional flooding event represented by a tongue of the Clays Ferry Formation or of the Millersburg Member (Fig. 1.1), which everywhere underlies the upper Tanglewood Member, small shoal complexes began to develop on top of or near the basement fault trends (Fig. 4.5, no. 1); intervening platform areas were occupied by offshoal to platform-lagoonal environments. During times 2 and 3 (Fig. 4.5, nos. 2 and 3), these two complexes continued to enlarge, even beyond their respective fault blocks, while their associated shoal facies spilled out across the entire upper Tanglewood platform. Time 3 represents the maximum extent of the shoals and the shallowest-water conditions on the platform. Cycle 3 in most places across the platform is represented by the thickest and coarsest calcirudite/calcarenite bodies in the entire member (Figs. 1.3, 4.3 and 4.4).

In contrast to previous cycles, however, lithofacies in Cycle 4 reflect the deepest-water conditions on the platform and may represent the resumption of regional transgression across the platform area (Fig. 1.3). Shoal, beach and foreshoal facies disappeared from the northeastern shoal, whereas only foreshoal and lower shoalface environments persisted in the southeastern shoal area. Interestingly, at this time the remaining platform area did not totally revert to an offshoal/platform-lagoonal setting as it had been in time 1. Instead, most of the platform area between structures A and D remained a lower-shoalface area, except for a small area in northern Scott County, where foreshoal facies abruptly developed. The recurrence of thicker, coarser lithofacies in Cycle 5 (Figs. 1.3 and 1.4) suggests another brief period of shallowing before final Clays Ferry inundation, but because much of the cycle is commonly destroyed by erosion, it has not been examined closely.
Fig. 4.5 Facies Map. Maps showing the distribution of predominant facies (Fig. 4.3) for the first, second, third, and fourth cycles on the basis of the outcrop observations and measured sections.
Chapter 5—Soft-Sediment Deformation

Introduction

One of the unusual aspects of the upper Tanglewood Member is the abundance of soft-sediment deformation throughout the unit, and what caused this deformation is yet another problem in the upper Tanglewood. However, interpreting the cause of such deformation is not easy, because post-depositional, soft-sediment deformation can be produced by various causes, including storms, rapid loading of sediments, tsunamis, slumping of sediments, and earthquake shocks, among others. In particular, many attempts have been made to differentiate seismogenic deformation from other non-seismogenic deformation, especially since several instances of seismogenic deformation have been interpreted previously from other parts of the Lexington Limestone (Ettensohn et al., 2002b, c; Jewell and Ettensohn, 2004). So, my hypothesis here is that the deformation in the upper Tanglewood is also most likely seismogenic in origin. The following parts of this chapter will discuss the characteristics of seismogenic deformation and compare them with features observed in the upper Tanglewood.

Seismogenic soft-sediment deformation is mainly produced due to sudden, greatly increased pore pressure in water-saturated silts or fine-grained sands, confined below by impermeable layers (Ettensohn et al., 2011). Therefore, two factors primarily control seismogenic deformation: presence of susceptible sediments and seismic waves of sufficient intensity to generate enhanced pore pressure. As originally defined by Seilacher (1969), seismites are re-deposited sediments as a result of disturbance and pervasive modification by earthquakes. Together with turbidites and tsunamites, seismites are widely regarded as
geological event deposits of great importance to the understanding of basin-scale tectonism and evolution.

Many researchers interested in soft-sediment deformation have attempted to identify parameters that can be used to link a sedimentary deposit to a seismic event, because similar deformational features are produced by causes other than seismic events, such as gravitational instability, impact of storm waves, circulation of groundwater flow, swelling due to salt or mud intrusion and meteorite impacts (e.g., Wheeler, 2002). Thus, relating a sedimentary record with soft-sediment deformation to a particular triggering mechanism is still a highly complex task. In some instances, researchers can only suggest a seismogenic interpretation based on the geological context in which the deposits are found (e.g., Montenat et al., 2007; Mugnier et al., 2011). Though many sedimentary features have been related to seismogenic deformation in the past three decades, even more examples from a wider array of geological settings are still required in order to enhance their recognition in the geological record. This is particularly needed, taking into account that seismites described in the literature have been interpreted from a wide variety of deformation structures.

Because of the possibility of so many origins for soft-sediment deformation, in order to successfully interpret it, it is necessary to exclude as many causes as possible so that only one cause is likely. For a seismitic interpretation, this means excluding all possible causes except a structural/tectonic cause. For example, the post-Miocene-to-present tectonic history of northeastern Brazil suggests that the most likely cause of soft-sediment deformation in the area was the reactivation of Precambrian and Cretaceous faults (Rossetti et al., 2010). Similarly, exclusion of all other causes across the central part of the Lexington Platform suggested that most soft-sediment deformation in the Lexington Limestone is related to its tectonic history and
the reactivation of basement faults (Ettensohn et al., 2002a, b, c; Jewell and Ettensohn, 2004; McLaughlin and Brett, 2004).

The Lexington platform, on which the upper Tanglewood Member was deposited, was an intracratonic platform, which experienced coeval far-field tectonic forces related to Taconic tectonism on the eastern margin of Laurentia. This tectonism apparently caused the reactivation of basement faults (Ettensohn, 1991; Ettensohn et al., 2002a, b, c; Jewell and Ettensohn, 2004), and the soft-sediment deformation recorded in Lexington Platform has been interpreted to be linked to earthquakes induced by the reactivation of these faults due to Taconic far-field forces.

Based on these interpretations, there has been a growing interest in the study of epicontinental seismicity in the last two decades, in terms of triggering mechanisms, epicentral areas, and recurrence intervals (e.g., Jewell and Ettensohn, 2004; McLaughlin and Brett, 2004; Ettensohn et al., 2010).

Unlike modern seismicity, much of our information about ancient seismicity comes from soft-sediment deformation, but much of this deformation has often been systematically overlooked or misinterpreted. Carbonates have also been systematically ignored in most attempts to understand soft-sediment deformation. However, in view of the fact that soft-sediment deformation is not a unique response to seismicity, this complexity requires a distinct series of criteria to confirm the likelihood of seismogenic origins for soft-sediment deformation. Wheeler (2002) and Ettensohn et al. (2002a) have proposed the concurrence of four criteria as a test for seismogenic origins. These criteria include: 1) presence of deformation consistent with a seismogenic origin; 2) deformation that is widespread; 3) deformation that can be temporally or stratigraphically constrained across its distribution; and 4) deformation that is zoned across its distribution, showing systematic increases in frequency, size or intensity toward a likely
epicentral area. The greater the concordance of these criteria, the more likely a seismogenic origin is. Although concurrence of the four criteria in soft-sediment deformation almost certainly does indicate a seismogenic origin, the co-occurrence of these criteria is rare and wholly dependent on past depositional environments and modern occurrence. Hence, the four criteria cannot be used as universal seismogenic indicators, and in the absence of such indicators, we must continue to rely upon the integration of tectonic, structural, and paleoenvironmental clues for substitute corroboration (Ettensohn et al., 2010).

Most of our knowledge about seismites comes from case studies in Europe and North America, primarily from Tertiary or Quaternary clastic sediments deposited in terrestrial and marginal-marine settings. However, there have been very few studies on carbonate seismites in cratonic shallow-marine settings like Lexington Limestone. Therefore, this study of soft-sediment deformation is potentially of great importance.

**Description**

Six deformed horizons have been observed throughout the upper Tanglewood Member. However, all six deformed horizons are not observed in every section. The nature of deformation also varies from section to section. In some sections, a single horizon is intensely deformed, whereas in other sections the same horizon is only slightly deformed, and equivalent horizons in other sections may not be deformed at all. Deformational features within the upper Tanglewood Member are confined to Cycles 1, 2, 3 and 5. Nowhere in Cycle 4 have deformed horizons been observed. Up to two deformed horizons are present in Cycle 1, but in most sections, only one deformed horizon is observed (Fig. 5.1). Cycles 2 and 3 may have up to two deformed horizons (Fig. 5.2), whereas Cycle 5 may have only a single deformed horizon at its base (Fig. 5.1).
Fig. 5.1 Measured section of upper Tanglewood Member in Nicholas County. Measured section of upper Tanglewood Member along US Highway 68 in Nicholas County (38°21'57.48"N, 84°2'20.46"W), showing the deformed horizons (DH) in various cycles. Note the deformed horizon within Cycle 1. The numbers in parasequences next to each deformed horizon (DH) reflect the intensity of deformation (See Table 1.1).
Fig. 5.2 Measured section along US Highway 68 in Nicholas County. Measured section along US Highway 68 in Nicholas County (38°24'10.32"N, 84° 0'51.84"W), showing the presence of two deformed horizons (DH) within Cycles 1, 2, and 3. The numbers in parasequences next to each deformed horizon (DH) reflect the intensity of deformation (See Table 1.1).
Deformed sediments

Deformation in the upper Tanglewood has been observed in carbonate sediments ranging in size from calcisiltites to calcirudites, but the most intense deformation is typically confined to the alternating calcarenite and shale lithofacies, whereas the simplest deformation of hydroplastic deformation with simple contortion and folding (Fig. 5.3) is commonly restricted to the coarse-grained carbonates of the massive calcarenite/calcirudite lithofacies. Various deformational structures are described below with their interpretation.

Fig. 5.3 Exposure of the massive calcarenite/calcirudite facies. Exposure of the massive calcarenite/calcirudite facies of the upper Tanglewood along US Highway 127 in Anderson County (38° 5′45.60″N, 84°55′9.90″W), showing hydroplastic deformation of simple contortions within Cycle 2.
**Ball-and-Pillow structures**—Ball-and-pillow structures are the most widely observed deformational features throughout the upper Tanglewood (Fig. 5.4); these structures have also been described as saucer structures (e.g., McLaughlin and Brett, 2004). These structures are masses of layered sediment which have become detached and sunk into underlying shale (Fig. 5.4). So, these structures are normally found in carbonate beds that overlie shale. They are normally hemispherical or kidney shaped and range in length from a few centimeters to several decimeters.

*Fig. 5.4 Ball-and-pillow structures with a diapiric intrusion.* Ball-and-pillow structures with a diapiric intrusion of mud in the upper Tanglewood deformed horizon within Cycle 2, exposed along US Highway 68 in Nicholas County (38°24′26.01″N, 84°0′46.43″W).

**Interpretation**—Ball-and-pillow structures result from the application of a physical shock to partially unconsolidated sediment. This shock causes periodic disruptions to occur in partially unconsolidated sediment layer at points of weakness, which induces instability (McLaughlin et al., 2004). Individual lobes break off at points of weakness and quickly settle
downward into the soft unconsolidated muds below. These lobes displace mobile muds upwards, usually as diapirs along the margins of the lobes. The ‘pillows’ that are produced typically retain much of their original bedding composition.

**Convolute bedding/lamination**—This kind of sedimentary deformation structure has been observed widely in deformed horizons within the upper Tanglewood Member. These structures have normally been observed within the larger ball-and-pillow structures described above. These structures are formed by complex folding or intricate crumpling of beds or laminae into irregular, generally small-scale anticlines and synclines (Fig. 5.5).

Convolute laminae are normally restricted to a single horizon, while similar horizons above and below the affected bed show no evidence of deformation. Convolute bedding is most commonly observed in alternating layers of fine-grained calcarenite and shale. Deformation ranges from simple, broad, concave-upward synforms with narrow antiforms to complex recumbent and overturned folds with random vergence directions. Beds containing convolute laminae commonly range in thickness from about 5 cm to 30 cm.

**Interpretation**—Convolute bedding/lamination, also known as “load convolution” (Dzulynski and Smith, 1963), is a complex type of structure involving hydroplastic deformation, which disrupts, but does not destroy, the original bedding/lamination. Convolute bedding/lamination occurs due to partial liquefaction and loss of strength in sediments associated with dewatering processes (Lowe, 1976; Collinson, 1994). Elevation of pore pressures in sediment layers during sudden episodes of consolidation (e.g., earthquake shaking) is the most commonly proposed mechanism for the formation of convolute bedding/lamination. Cyclic loading, resulting from pounding by large waves or seismic shocks (Maltman, 1994), can cause temporary pore-pressure elevation. This elevated pore pressure results in the loss of grain-to-
grain contacts and temporary loss of strength. The overpressured sediment may then become liquefied. Interbedded silt and mud tend to show more complex convolutions than coarser-grained materials, in which dish and pillar structures are typically more prevalent (Dzulynski and Smith, 1963).

**Fig. 5.5 Convolute bedding in an upper Tanglewood deformed horizon.** Convolute bedding in the upper Tanglewood deformed horizon within Cycle 2, exposed along US Highway 68 in Nicholas County (38°24'26.01"N, 84° 0'46.43"W).

**Sand or mud volcanoes/diapiric intrusion**—Sand and mud volcanoes are intrusions of mudstone or sandstone into the overlying sedimentary sequence. Siliciclastic mudstone diapirs typically occur between ball-and-pillow structures (Fig. 5.4). The large mudstone diapirs are clearly differentiated from normal semi-tabular shales. These mudstone diapirs often contain scattered fossil fragments and small calcarenitic boudins at their tops. Smaller diapiric tongues of
mud, which are called flame structures, also occur locally. It is very common for mud diapirs to terminate in ‘mushroom-like’ tops that expand laterally above the edge of the adjacent ball-and-pillow structures (Fig. 5.4).

**Interpretation**—Injection diapirs result from a combination of fluidization phenomena and hydro-fracturing (or hydraulic “jacking”) (Lowe, 1975, 1976; Cosgrove, 1995). The upward movement of a diapir is relative to the surrounding sedimentary rocks. Diapirs are initiated by unequal loading of a layer of material of equivalent viscosity. Foundering layers of denser sediment may trigger the upward movement of initially underlying liquefied/fluidized thixotropic mud and contained particles (Sorauf, 1965). Upward displacement of pore water generates fluidization in vertical channels or pipes in overlying sediments. Clastic dikes will generally develop only when the sediment above the source layer has low permeability or significant cohesion due to a higher content of fine-grained sediments. In such situations, the cohesive sediments lying above are deformed by brittle failure. Most of the escaping fluidized pipes follow such fractures during upward migration (Owen, 1995).

**Boudins and brecciated fabrics**—Boudins are nodule-like structures, a centimeter to decimeter in diameter, composed of calcisiltite or fine-grained calcarenite (Fig. 5.6). They reflect the unequal competencies of the layers forming the boudins and adjacent layers during the sediment flow accompanying deformation. Where this flow becomes less streamlined and more irregular, the nodules may exhibit a more brecciated fabric. Boudins most commonly occur in small numbers associated with ball-and-pillow structures and mudstone diapirs and are made up of the same beds that compose the larger structures. However, in some exposures boudins occur in very large numbers to the exclusion of saucers and diapirs (Fig. 5.6). These massive to pseudo-bedded intervals closely resemble nodular shaley limestones, but gradually grade
laterally into ball-and-pillow structures and mudstone diapirs. Additionally, pseudo-bedded intervals show little to no evidence of bioturbation, and may display random orientation of ellipsoidal nodules.

**Fig. 5.6 Brecciated-to-nodular fabric.** Brecciated-to-nodular fabric exposed in upper Tanglewood equivalent within Point Pleasant Formation within Cycle 2 along US Highway 1951 in Bracken County (38°45'9.85"N, 84° 7'15.81"W).

**Interpretation**—Boudins are the end product of deformation of thin alternating mud and carbonate silt/sand beds (McLaughlin and Brett, 2004). Autoclastic breccias result from the brecciation of indurated beds (in many cases early lithified carbonates), alternating with liquefied thixotropic layers. During the deformation event, thin interbedded mud and silt/sand layers are mobilized to form convolute bedding. If the upward velocities of fluidized mud are sufficiently high, individual silt/sand layers may rapidly attenuate, forming small boudins that are further
rounded by the fluid movement of the mud matrix (Rast et al., 1999). In the process, the boudins may even get transported a very short distance. Large numbers of boudins forming brecciated fabrics seem to represent an extreme end member for intensely deformed rhythmite and calcarenite facies (Mclaughlin and Brett, 2004).

**Homogenization/fluidization**—Fluidization reflects the nearly complete destruction of any sedimentary structures, mainly bedding, or the homogenization of a complete sequence of layers so that only a few nodules or fragments of calcarenites and calcisiltite appear to be floating in a mud matrix. This kind of homogenization is observed extensively in Anderson, Nicholas (Fig. 1.8) and Bracken counties. Bedded sequences greater than 6-feet (1.92 m) (Fig. 5.7) thick have been completely homogenized in the Lawrenceburg section (Anderson County), and another bedded sequence almost 4-feet (1.2 m) thick has been homogenized in the Nicholas County section. Local pockets of homogenized beds with limited lateral extent are observed here and there in any deformed section, but extensively homogenized, thick horizons are observed only in the Anderson and Nicholas county sections. In the same way, there are homogenized deformed horizons in upper Tanglewood equivalents in the Point Pleasant Formation, which are extensively exposed in Bracken County.

**Interpretation**—Homogenized beds form by both the upward and lateral flowage of fluidized sediments of thin alternating mud and carbonate silt/sand beds during the intense cyclic shaking of sediments (Lowe, 1975). In this situation, muddy or non-lithified sediments behave as fluids, and coarse-grained partially lithified calcarenites break into fragments, undergo erosion by moving fluids, and eventually try to settle into the mud or fluidized matrix (Lowe, 1975). Commonly, however, upward-moving fluid forces resist any settling and the fragments are held upward in suspension.
A fluidized or homogenized bed consists of a fluid-solid mixture that exhibits fluid-like properties. The bed can be considered to be a heterogeneous mixture of fluid and solid that is effectively represented by a single bulk density. When fluid is forced vertically through a bed of cohesionless granular solids, the sediment porosity and packing remain essentially unchanged at the lowest flow velocities, and the fluid escapes by seepage. As fluid velocity is progressively increased, a critical value is reached at which downward gravitational forces on the grains are balanced by the upward fluid drag. This is termed the incipient or minimum fluidization velocity (Lowe, 1975). At higher flow rates, the bed expands rapidly, porosity increases, and the sediment framework ceases to be grain-supported and becomes fluid-supported; the sediment is then said to be fluidized. If fluid discharge continues to increase, the fluid force will eventually exceed gravitational forces on the individual dispersed particles, and the sediment becomes fully suspended. The fluidizing agent may be either gas or liquid. Gas fluidization generally involves the rise of discrete gas bubbles which tend to coalesce as they rise and burst upon reaching the sediment surface.

**Load structures**—Load structures of various sizes and dimensions are quite obvious in the deformed horizons of upper tongue of the Tanglewood Member (Fig. 5.7). Normally, it is mostly the coarse-grained calcarenites that appear to have sunk into muddy sediments.
Interpretation—Load structures form in response to gravitational instabilities created by the deposition of relatively dense sediment on a relatively less dense sediment substrate. If the gravitational forces deriving from such instabilities exceed the shearing strength of both sedimentary layers, loading occurs.

Most load structures are the result of pre-consolidation soft-sediment deformation. Many, however, particularly those formed within the body of a single sedimentary unit, develop in response to loss of strength associated with fluidization or liquefaction. Load structures occur less commonly within individual sedimentary units than at lithologic boundaries. Two basic types of internal loading are observed: (1) loading of more compacted sediment into less compacted sediment, both having similar composition and texture; and (2) loading in response to compositional or textural density difference. During consolidation, the former occurs commonly in association with homogeneous fluidization or liquefaction. Discrete masses of already
consolidated, grain-supported sand within fluidized zones sink slowly to replace the underlying sand removed by fluidization. These subsiding sediment bodies are typically defined by the presence of dish structures along their lower margins.

**Thinning and thickening of beds**—Deformed beds typically show inconsistency in thickness across very short distances (Fig. 5.8). In the Lawrenceburg (Anderson County) section, the deformed bed (Fig. 1.8) is up to 5.5-ft. (1.67-m) thick in one place, but in moving a distance of about 20 meters, it drops to the thickness of 3 ft. (0.91 m). Thinning and thickening of beds are quite obvious in other sections as well.

![Fig. 5.8 Thinning and thickening of a deformed horizon.](image)

*Fig. 5.8 Thinning and thickening of a deformed horizon. Thinning and thickening of a deformed horizon in Cycle 2 of upper Tanglewood Member exposed along US Highway 127 in Anderson County (38° 3’ 16” N, 84° 55’ 9.02” W).*
**Interpretation**—Soft-sediment deformation occurs in sediments that are incompletely consolidated, but even the same layer may be unconsolidated in some areas and consolidated in other nearby areas. So, a deformational horizon may thicken and thin depending upon whether or not parts of the unit were consolidated. In thinned portions, much more of the unit was consolidated and not subjected to deformation. In thickened areas of deformation, much more of the unit was consolidated.

**FRACTURES**—Vertical fractures are quite obvious in deformed horizons and may cut the deformed horizons and abruptly terminate in adjacent undeformed horizons, either above or below the deformed horizon. Some of these fractures have actually become faults that appear to have displaced the beds by small amounts (Fig. 5.9). Normally, these fracture structures are associated with convolute laminae (Fig. 5.9). The fractures die out in both upward and downward directions and hence can be described as growth faults or fractures.

*Fig. 5.9 Fracture developed across a deformed horizon. Fracture developed across deformed horizon in Cycle 2 which is exposed along US Highway 127 in Franklin County (38°14'53.88"N, 84°51'12.11"W).*
**Interpretation**—Such structures are probably post-depositional and post-consolidation. They seem to occur across deformed horizons in places where consolidated deformation instabilities (deformed laminae in Fig. 5.9) are later accommodated by short fractures or faults.

**Cross-cutting deformation**—Crosscutting relationships between deformed and undeformed rocks are obvious in many exposures (Fig. 5.10). Although deformed rocks are commonly found interbedded with undeformed rocks, the deformed beds may appear to cut the undeformed beds at subtle angles from above and below at many exposures.

*Fig. 5.10 Cross-cutting relationship. Fluidized horizon (inside green boundary) of upper Tanglewood Member in Cycle 2 completely truncating lower beds(red boundary) and partially truncating upper bed exposed along US Highway 127 in Anderson County (38° 3' 15.16" N, 84° 55' 9.02" W).*
**Interpretation**—Cross-cutting relationships between deformed and undeformed rocks at subtle angles are especially diagnostic of a seismogenic origin (Davenport and Ringrose, 1989; Rast et al., 1999). Cross-cutting relationships across underlying layers could be due to intrastratal or normal erosion, but the crosscutting of overlying rocks is difficult to explain where the possibility of onlap is ruled out (Ettensohn et al., 2002d). Ettensohn et al. (2002d) have pointed out that seismically augmented pore pressure can only influence susceptible sediments, which because of the vagaries in cementation or clay content may occur at oblique angles to the normal bedding, thus appearing to have been truncated from below or above (Fig. 5.11).

![Fig. 5.11 Cross-cutting deformation. An exaggerated, schematic diagram at outcrop scale showing a deformed horizon that crosscuts parts of three beds at a low angle. Stippling represents flow rolls and contortion in calcisilites or fine-grained calcarenites; dashed lines represent shaley sediments, and parallel lines represent undeformed beds (from Ettensohn et al., 2002b).](image)

**Soft-Sediment Deformation in Equivalent Parts of the Clays Ferry and Point Pleasant Formations**

On the basis of cycle-stacking patterns, deformed horizons in the upper Tanglewood Member can be correlated with similar horizons in the Clays Ferry and Point Pleasant formations. The upper Tanglewood sequence-like cycles and deformation horizons are virtually identical to those in the Clays Ferry, the major difference being that the Clays Ferry equivalents
are more shaley in composition than upper Tanglewood counterparts. Four deformed horizons can be observed in equivalent parts of the Clays Ferry Formation (Fig. 5.12).

Fig. 5.12 Measured section in the type Clays Ferry Formation. Measured section of the type Clays Ferry Formation, showing four deformed horizons in the upper Tanglewood equivalent along US Highway 25 (type locality of Clays Ferry Formation) in Madison County (37° 52' 51.50" N, 84° 20' 26.00" W).
Similarly, there are plenty of exposures of upper Tanglewood equivalents in the Point Pleasant Formation of Bracken County, Kentucky and Clermont County, Ohio (Figs. 5.5 and 5.6). These exposures also contain deformed horizons, which, on the basis of cycle-stacking patterns, can be correlated with deformed horizons in the upper Tanglewood Member. These deformed horizons also show the same kind of deformational structures as does the upper Tanglewood, including ball-and-pillow structures, fluidization, diapiric intrusion, folding and contortion, among others.
Chapter 6—Interpretation of Soft-Sediment Deformation

Introduction

Seismogenic, soft-sediment deformation is controlled by two factors: the presence of susceptible sediments (generally water-saturated silts or fine sands) confined below by impermeable layers (e.g., Obermeier et al., 1990) and a sudden input of energy sufficient to generate enhanced pore pressure (e.g., Sieh, 1982; Holzer et al., 1989). Allen (1977, 1984) has pointed out that poorly compacted (i.e., loose) water-saturated silts and finer sands are the most susceptible to liquefaction. In the case of thixotropic muds (that is, time-dependent viscous muds that are thick and viscous under static conditions but become thin and less viscous over time when shaken, agitated, or otherwise stressed), deformation may occur in unconfined circumstances, especially if the triggering event is abrupt and rapid. On the other hand, for partially consolidated and cohesive sediments, burial below relatively impermeable layers is essential to generate the overpressuring necessary for deformation; otherwise, deforming pore fluids escape at the surface and stress is dissipated (Jewell and Ettensohn, 2004). A 2-to-3-meter-thick confining layer of sediments can impede seismogenic deformation because the resulting confining pressure is greater than any seismically augmented pore pressure (Obermeier et al., 1990). Therefore, there is a good reason to assume that most marine seismogenic deformation occurred at or below, but not too far below, the sediment-water interface.

Locating Likely Epicentral Areas

Assuming uniform sedimentary conditions and no local effects, such as local basin shape, sediment thickness, sediment type, and the nature of the involved structures at depth, increases in
pore-fluid pressure and velocity should be proportional to the earthquake energy released through cyclic loading. Moreover, the resulting deformation should reflect the amount and location of the energy release, with the greatest energy release, and hence the most intense deformation, occurring near the epicenter. Based on Lowe’s (1975) classification and interpretation of water-escape structures, three forms of deformation, each requiring successively higher pore-fluid velocities, and hence energy input, are possible in deformed horizons. Lowe (1975) related deformation processes to energy input via the minimum fluidization velocity ($U_o$), and similar relationships have been noted elsewhere between the severity of earthquakes and resulting soft-sediment deformation (e.g., McCalpin, 1996). At lower energies and resulting pore-fluid velocities well below $U_o$, hydroplastic deformation is the most common form of deformation and is characterized by the simple contortion or folding of beds with the preservation of primary lamination or bedding; elutriation of finer grains (separation of particles based on their size, shape and density) is minor. At higher energies and a pore-fluid velocity approaching $U_o$, liquefaction becomes predominant and is characterized by vertical piping, dikes, or sand volcanoes that initiate destruction of primary lamination and bedding; flow structures are laminar and elutriation of finer grains becomes more prevalent. At very high energies with resulting pore-fluid velocities that exceed $U_o$, fluidization predominates and is characterized by complete destruction of primary lamination or bedding and homogenization of the unit; flow structures are turbulent and elutriation of fine grains is ubiquitous. Hence, individual deformation types may reflect differing energy inputs. As a result, if individual deformed horizons can be characterized in exposure by the predominance of one of the three types of deformation, and if the types of deformation can be mapped in temporally constrained horizons, then point sources of energy input of the type encountered in seismicity...
can be suggested by areas of more intense deformation and a roughly concentric pattern of decreasing-energy deformation bands surrounding it (Ettensohn et al., 2000b, c, 2001; Jewell and Ettensohn, 2004). Such isoseismal bands of intensity are generally concentric about the epicenters of modern quakes and the area of most intense deformation. The area of greatest intensity of deformation is inferred to be the epicentral area (e.g., Reiter, 1990). In short, mapping the distribution of deformation types in well-constrained, deformed horizons has the potential for locating ancient epicentral areas, as has been suggested by Ettensohn et al. (2002b, c) and Jewell and Ettensohn (2004).

**Approximating Paleoearthquake Magnitude**

If deformation can be linked to a seismogenic origin, then the minimum-magnitude earthquake required for liquefaction is generally thought to be M=5 or greater (Carter and Seed, 1988; Ambraseys, 1988; Obermeier et al., 1990). Based on this limit, there are two methods to approximate the magnitude of such an earthquake. The first method is determining the areal extent of the deformation. Assuming uniform sedimentation and bed thickness, and by determining the distance between the epicentral area and the farthest extent of deformation, an approximation of earthquake magnitude can be made. The greater the extent of deformation, the greater the magnitude of the earthquake. In ancient seismites, both the likely epicentral area and extent of deformation can be determined by mapping types of deformation as noted above, provided adequate temporal and stratigraphic control is available (Ettensohn et al., 2000, 2001, 2002b, c; Jewell and Ettensohn, 2004). Moreover, according to Obermeier et al. (1990), in any given deformed horizon, the number and size of dewatering structures increase toward the
epicentral area, and the number and size of these structures are mappable parameters. Once the epicentral area is located, epicentral distance to the farthest deformation can be measured and plotted on established curves (Ambraseys, 1988; Obermeier et al., 1990) to provide an approximate earthquake magnitude (Fig. 6.1). Similar methods have been employed in the Lexington Limestone by Pope and Read (1992) and Ettensohn et al. (2002b, c).

![Graphical relationship between earthquake magnitude and epicentral distance](image)

**Fig. 6.1** Graphical relationship between earthquake magnitude and epicentral distance. Graphical relationship between earthquake magnitude and epicentral distance to the farthest liquefaction features (adapted from Obermeier et al., 1996).

A second method is based on statistical correlations among magnitude, fault type, fault-rupture length, and surface displacement from historical earthquakes (Bonilla et al., 1984). At least the fault type and either rupture length or displacement must be known. These parameters
are more difficult to collect from the fossil record, but if seismites can be related to reactivation of a given fault, and facies changes on either side of the fault may allow for a change-in-depth approximation that can serve as a proxy for surface displacement, then this method is available for estimating magnitude.

### Ioseismal Map for the Deformed Horizons

For this study, a total of 45 outcrops were measured (Figs. 6.2 - 6.7) and described from the upper Tanglewood Member, Point Pleasant Formation, and Clays Ferry Formation. In order to map the distribution of deformation types noted by Lowe (1975), a rating scheme was necessary to classify types of deformation in the field. In Table 1.1, I present a nine-point scale of increasing deformational intensity from Jewell and Ettensohn (2004). Although mapping individual numbered intervals has proven to be impractical because of extreme variation from locality to locality, we have been successful in mapping distribution of the three major types of deformation: hydroplastic deformation (1-3), liquefaction (4-6), and fluidization (7-9). Ioseismal maps for each deformed horizon are described below relative to their occurrence within the five temporally constrained chronostratigraphic cycles.

### Deformation in Cycle 1

Cycle 1 has one distinct deformed horizon which is mainly concentrated in Nicholas and Franklin counties (Fig. 6.2). A distinct deformed horizon in this cycle is also observed in Scott County (38° 24' 3.20" N, 84° 34' 34.70" W) and at Monterey in Owen County (38° 25' 47.03" N, 84° 52' 6.41" W). In other sections, a deformed horizon was not observed in this cycle. Liquefied (rating scale 4-6) beds were observed in the Nicholas County sections (northeast corner of upper
Tanglewood buildup). The deformed horizon here is up to 0.4-m (1.5-ft) thick in these sections. In contrast, on the western flank of the upper Tanglewood buildup, both hydroplastically deformed and liquefied beds are observed. The isoseismal map prepared for this deformed horizon (Fig. 6.2) shows that the deformed zone is mainly concentrated between two basement faults A and B on the western flanks of the upper Tanglewood shoal complex. Clearly, there is a concentric pattern of increasing intensity of deformation from hydroplastic deformation to liquefaction in this block. On the northeast corner of the buildup, however, no such concentric pattern of increasing intensity of deformation is observed. However, the deformed zone is concentrated along basement fault lineament D.

Surprisingly, a liquefied bed was also observed in the Scott County section (38° 24' 3.20" N, 84° 34' 34.70" W) and in the Owen County section (38° 25' 47.03", 84° 52' 6.41"), which are located between basement faults A and C. However, the surrounding exposures showed no deformation in this region. Furthermore, no deformation has been noted from equivalent horizons in the Clays Ferry and Point Pleasant formations.

**Deformation in Cycle 2**

This cycle contains two distinct deformed horizons in most of exposures. The equivalent horizons in the Clays Ferry and Point Pleasant formations also display deformed horizons in this cycle (Figs. 5.12, 6.3 and 6.4). The Winchester area (Clark County), on the other hand, displays deformation in the upper horizon only (Figs. 6.3 and 6.4).
Fig. 6.2 Isoseismal map for the deformed horizon in Cycle 1. Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the deformed horizon in Cycle 1. A. =Anderson County, B. =Bracken County, C. = Clermont County, F. =Franklin County, H. = Harrison County, M. =Madison County, N. = Nicholas County, O. = Owen County, S. = Scott County, Cl. = Clark County.

The isoseismal map for the lower deformed horizon in this cycle is given (Fig. 6.3). The lower deformed horizon displays all kinds of deformation: hydroplastic deformation (1-3), liquefaction (4-6) and fluidization (7-9). There are five regions where the deformed horizons are displayed (Fig. 6.3). There are two sections around Lawrenceburg, Anderson County (38° 3' 30.90" N, 84° 55' 15.30" W and 38° 3' 15.16" N, 84° 55' 9.02" W), where this horizon displays
fluidized beds. Interestingly, the nearby exposure at 38° 5' 45.60" N, 84° 55' 9.90" W, very near to these fluidized sections, displays only hydroplastically deformed horizons. Because the hydroplastically deformed section, near the fluidized zone, is composed of massive calciruditic facies, it could be argued that the grain size of the sediment is an important influence on the type of the deformation. Otherwise, one would have expected at least a zone of liquefaction surrounding the fluidized zone. Instead, only a hydroplastically deformed zone is present without any liquefied zone, as we move outward from the fluidized zone. The deformed zone on the western flank of the shoal complex apparently cuts across the basement faults A and B in Franklin and Owen counties. The fluidized zone lies in the vicinity of basement fault B, whereas the liquefied zone lies on the vicinity of basement fault A (Fig. 6.3). In the northeast corner, the deformation lies between basement faults C and D, or more precisely near the basement fault D in Nicholas County (Fig. 6.3). A zone of deformation also occurs on the northern part of the Tanglewood buildup, north of basement fault C in Scott and Harrison counties. The equivalent horizon in the Clays Ferry Formation at the type locality (37° 52' 51.50" N, 84° 20' 26.00" W) is also deformed (Fig. 5.12). The equivalent horizons in the Point Pleasant Formation around Bracken and Clermont counties also display hydroplastically deformed and liquefied beds in this cycle (Fig. 6.3) and occur in the vicinity of basement fault E.

The isoseismal map for the upper deformed horizon in this cycle is provided in Figure 6.4. In the same way as the lower deformed horizon, this horizon also displays several kinds of deformation: hydroplastic deformation (1-3), liquefaction (4-6) and fluidization (7-9), with fluidized horizons on the northeast margin, in the vicinity of basement fault D. An equivalent horizon in the Point Pleasant Formation in Clermont County also displays fluidized beds, which occur in the vicinity of basement fault E. Unlike the lower deformed horizon, this horizon also
contains liquefied beds on the southeast corner around the Winchester area of Clark County. Similarly, equivalent horizons in the Clays Ferry Formation at the type locality (37° 52' 51.50" N, 84° 20' 26.00" W) are also liquefied (Fig. 5.12). The deformed region in the northeast corner clearly shows the concentric pattern of increasing intensity of deformation toward the center. In the same way, on the western flank of the upper Tanglewood shoal complex, this pattern is more or less observed but not as distinct as in the northeast corner. Here, locally fluidized beds are observed, but these beds are not persistent into surrounding sections of Anderson County, which are very near to it. It appears that in the Lawrenceburg section (38° 3' 30.90" N, 84° 55' 15.30" W), the horizon is fluidized, but just 200-m away from that section (38° 3' 15.16" N, 84° 55' 9.02" W) the beds are liquefied. Therefore, these two sections were mapped as occurring in the liquefied zone. The orientation of deformation on the western flank clearly cuts across the two basement faults A and B. Similarly, two other deformed zones are concentrated between basement faults A-C and C-D. One deformed zone in the Winchester area, Clark County, however, lies outside of basement fault blocks. The deformed zone in the equivalent Point Pleasant Formation lies in the vicinity of basement fault E.

**Deformation in Cycle 3**

Cycle 3 also contains up to two deformed horizons. Unlike Cycle 2, these deformed horizons are exposed only in very few exposures. Both the lower and the upper deformed horizons display all kinds of deformation: hydroplastic deformation (1-3), liquefaction (4-6), and fluidization (7-9) (Figs. 6.5 and 6.6). Overall, this cycle is more massive and is interpreted to have been deposited during the lowest sea-level conditions of all the cycles in the upper Tanglewood Member.
Fig. 6.3 Isoseismal map for the lower deformed horizon in Cycle 2. Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the lower deformed horizon in Cycle 2. A.=Anderson County, B.=Bracken County, C.=Clermont County, F.=Franklin County, H.=Harrison County, M.=Madison County, N.=Nicholas County, O.=Owen County, S.=Scott County, Cl.=Clark County.

The isoseismal map for the lower deformed horizon in this cycle is shown in Figure 6.5. The lower deformed zone is scattered sporadically in five different areas. Two deformed zones occur in the vicinity of basement fault D on the northeast margin. In this area, the beds display hydroplastic deformation and liquefaction. A concentric pattern of hydroplastically deformed and liquefied beds is observed in this area. On the western flank of the Tanglewood shoal,
however an irregular pattern of hydroplastic deformation, liquefaction and fluidization is displayed. The deformed zone cuts across both basement faults A and B. The Monterey section in Owen County (38° 25' 47.03'' N, 84° 52' 6.41'' W), which lies between the basement faults A and C, exhibits a fluidized horizon. Due to the lack of exposure in the surrounding area, it is not shown to have been encircled by hydroplastically deformed and liquefied zones.

Fig. 6.4 Isoseismal map for the upper deformed horizon in Cycle 2. Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the upper deformed horizon in Cycle 2. A.=Anderson County, B.=Bracken County, C.=Clermont County, F.=Franklin County, H.=Harrison County, M.=Madison County, N.=Nicholas County, O.=Owen County, S.=Scott County, Cl.=Clark County.
Similarly, two sections with fluidized horizons are observed in the Lawrenceburg sections in Anderson County (38° 3’ 30.90“ N, 84° 55’ 15.30“ W and 38° 3’ 15.16“ N, 84° 55’ 9.02“ W). This zone, lying in the vicinity of basement fault B, is also not encircled by either hydroplastically or liquefied zones. Liquefied zones lying in between the basement faults A and B are encircled by zones of hydroplastic deformation, showing the typical concentric pattern of increasing deformational intensity. The sections in Georgetown (Scott County), Winchester (Clark County) and equivalent horizons in the Clays Ferry Formation at its type locality (37° 52’ 51.50“ N, 84° 20’ 26.00“ W) (Madison County) are not deformed. A liquefied horizon is also observed in one exposure of equivalent parts of the Point Pleasant Formation in Clermont County (Fig. 6.5), which lies in the vicinity of basement fault E.

The isoseismal map for the upper deformed horizon in this cycle is shown in Figure 6.6. In this horizon, deformed beds have a sporadic but widespread distribution, including equivalent horizons in Clays Ferry and Point Pleasant formations. In the northeast corner of the upper Tanglewood complex, deformation mainly lies on and near basement fault D. Here the fluidized zone is encircled by a liquefied zone. Hydroplastic deformation is not obvious in this area due to the lack of exposure in surrounding areas. Another deformed zone lies beyond the basement fault D, in the Winchester and Clays Ferry areas of Clark and Madison counties, respectively, where hydroplastic deformation and liquefied beds are present.

A deformed zone on the western flank of the Tanglewood complex has an irregular distribution pattern with two areas of deformation, north of basement fault A and south of basement fault B, showing small fluidized horizons. The fluidized beds on the south side of the basement fault B are isolated from each other with no lower intensity of deformation surrounding the area, which is partly due to lack of exposure. However, there is one section (Fig. 6.6), lying
on the basement fault B that shows a hydroplastically deformed horizon. The fluidized zone in Monterey (38° 25' 47.03" N, 84° 52' 6.41" W) (Owen County), north of basement fault A, is accompanied by liquefied beds in the surrounding area.

**Fig. 6.5 Isoseismal map for the lower deformed horizon in Cycle 3.** Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the lower deformed horizon in Cycle 3. A.=Anderson County, B.=Bracken County, C.=Clermont County, F.=Franklin County, H.=Harrison County, M.=Madison County, N.=Nicholas County, O.=Owen County, S.=Scott County, Cl.=Clark County.
Equivalent horizons in the Point Pleasant Formation are also deformed in Bracken and Clermont counties (Fig. 6.6). Hydroplastic deformation and liquefaction of beds are observed there, lying in the vicinity of basement fault E.

**Fig. 6.6** Isoseismal map for the upper deformed horizon in Cycle 3. Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the upper deformed horizon in Cycle 3. A.=Anderson County, B.=Bracken County, C.=Clermont County, F.=Franklin County, H.=Harrison County, M.=Madison County, N.=Nicholas County, O.=Owen County, S.=Scott County, Cl.=Clark County.
**Deformation in Cycle 4**

No deformational features were observed in Cycle 4. This cycle represents deep-water conditions across the shoal complex, and the sediments here are all too fine-grained and clay-rich to support deformation.

**Deformation in Cycle 5**

Because Cycle 5 is not observed in most exposures of the upper Tanglewood Member due to erosion, deformed horizons are absent in most exposures, making it difficult to interpret. However, there are three exposures of Cycle 5 in the upper Tanglewood Member where fluidized and liquefied beds are observed at the base of the cycle (Fig. 6.7). Fluidized beds in the upper Tanglewood Member are observed in the vicinity of basement fault D in Nicholas County (38° 21' 57.48" N, 84° 2' 20.46" W). Although the sections are complete in this region, the surrounding exposures do not show any deformation in this cycle. There are two exposures in the upper Tanglewood Member in Scott County, where deformed horizons are observed between basement faults A and C.

The deformation in this cycle is, more importantly, displayed in the equivalent Point Pleasant Formation around Bracken and Clermont counties and at the type locality of the Clays Ferry Formation in Madison County. The deformation in the equivalent horizons of the Point Pleasant Formation shows widespread distribution (Fig. 6.7), and liquefaction is the primary type of deformation. In one exposure, however, a fluidized horizon is also observed in equivalent parts of the Point Pleasant Formation (Fig. 6.7). The deformed equivalents in the Point Pleasant Formation occur in the vicinity of basement fault E.
Fig. 6.7 Isoseismal map for the deformed horizon in Cycle 5. Distribution of deformation mapped by deformational processes (different patterns, see Table 1.1) for the deformed horizon in Cycle 5. A. =Anderson County, B. =Bracken County, C. = Clermont County, F. = Franklin County, H. = Harrison County, M. = Madison County, N. = Nicholas County, O. = Owen County, S. = Scott County, Cl. = Clark County.

Discussion: Case for Seismogenic Deformation

Introduction

The term “seismite” was first coined by Seilacher in 1969 as a genetic or interpretive term for post-depositional, soft-sediment deformation produced by earthquake shaking.
However, it was later realized that the characteristics he used to interpret seismites are not unique to seismogenically deformed sediments (Seilacher, 1991). There have been many cases in which deformed beds with characteristics similar to those used by Seilacher to define seismites have been interpreted to have been produced by slope-induced mass movement, rapid depositional loading, dewatering, and storm overpressuring on marine bottoms, among other causes. Therefore, many other criteria have been proposed for differentiating seismogenic deformation from other non-seismogenic deformation. Here, we have used the concurrence of four lines of evidences, suggested by Ettensohn et al. (2002d), Ettensohn and Stewart (2003), and Jewell and Ettensohn (2004) to interpret seismites because these criteria can be reasonably determined in the ancient geologic record (Ettensohn et al., 2002d, Jewell and Ettensohn, 2004, Ettensohn and Stewart, 2003). These criteria include: deformation that is consistent with a seismogenic origin, widespread occurrence of deformation in horizons that are temporally and stratigraphically constrained, deformation that shows systematic increase in frequency or intensity toward a possible epicentral area, and the ability to exclude other possible causes. The greater the number of these criteria that can be met, the more likely a seismogenic origin is (e.g., Greb et al., 2002; Jewell and Ettensohn, 2004). As an example, deformation in older Lexington units, including the Brannon Member and Cane Run Bed, has already been interpreted to be seismites based on these criteria (Ettensohn et al., 2002d, Jewell and Ettensohn, 2004, Ettensohn and Stewart, 2003). The present study would further bolster these criteria, if the present deformation can be reasonably interpreted to be of seismogenic origin.

Other factors that can be used in making even stronger cases include occurrence of deformation in a presently or formerly suitable tectonic setting (active seismic region), presence of a potentially originating faults (Sims, 1975), deformation that crosscuts normal stratification
(Rast et al., 1999) and well-constrained deformation that crosses local and regional facies boundaries (Ettensohn et al., 2002d). Here, we will try to validate these criteria in order to interpret whether these deformed upper Tanglewood horizons are seismites.

The upper Tanglewood Member occurs sporadically throughout most of central Kentucky (Figs. 1.2 and 1.4). It attains its maximum thickness of 12.8 m (42 ft.) in Nicholas County (38°24'52.65"N, 84° 0'15.07"W). On the southern margin of its distribution in Fayette and Woodford counties, exposures are not available. Therefore, it is difficult to interpret the nature of deformation there, but in rest of the area, plenty of exposures are present. However, near the northern margin of upper Tanglewood distribution, in Scott and Harrison counties, the upper Tanglewood contains a lesser number of deformed beds with lesser intensities of deformation. Northeastern and western parts of the upper Tanglewood shoal complex contain intense deformation with higher numbers of deformed horizons. There are also plenty of exposures of the Point Pleasant Formation with upper Tanglewood equivalents around Bracken and Clermont counties. In these exposures, however, lower parts of the upper Tanglewood equivalents are not exposed extensively, but the uppermost deformed horizon (deformation in Cycle 5) is well displayed with a range of deformational intensities.

**Evidence for Seismogenic Deformation**

Soft-sediment deformation occurs in six distinct and separate horizons of the upper Tanglewood and laterally equivalent parts of the Point Pleasant and Clays Ferry formations. Evaluation of how these six deformed horizons correlate from exposure to exposure and from
unit to unit was based on the position of these horizons in the chronostratigraphic cyclic framework already described.

**Deformation Consistent with Seismogenic Deformation**

Interpretation of the seismites from the lower Lexington Limestone has already been made by various researchers (e.g., Pope et al., 1997; Rast et al., 1999; McLaughlin and Brett, 2004; Ettensohn et al., 2002b, c; Jewell and Ettensohn, 2004). Comparison of upper Tanglewood deformation with the ancient carbonate, soft-sediment deformation inferred to be of seismogenic origin (e.g., Pratt, 1998; Ettensohn et al., 2002c; Jewell and Ettensohn, 2004) shows very strong similarities, including examples of folds, simple contortions and homogenization (Figs. 1.8 and 5.3–5.8).

**Widespread Distribution**

The deformation structures are confined largely to two facies: lower shoalface (alternating layers of calcarenites and shale) and foreshoal and upper shoalface (calcarenites with thin shale partings). These facies are easily positioned within the meter-scale, chronostratigraphic cyclic framework. These meter-scale cycles not only isolate individual deformed beds from one another, but are widely traceable across the study area. The tracing of meter-scale cycles across the study area has created a high-resolution chronostratigraphic framework, which has permitted regional mapping of individual deformed beds. Widespread distribution of well-constrained deformation horizons is very important to interpret seismogenic origins, because it can rule out the possibility of more localized processes such as slumping and sediment loading. The deformed horizons in the upper Tanglewood are widely distributed as shown in Figures 6.2–6.7, and can also be traced into the equivalent Clays Ferry and Point
Pleasant formations. The total distribution of these deformed horizons covers an approximate area of 3750 km$^2$ for the upper deformed horizon in Cycle 2. Because the deformation crosses facies boundaries into the Clays Ferry to Point Pleasant formations, localized processes of deformation such as slumping, artesian effects and loading of the sediments can be ruled out. Cross-facies distribution of deformation can be particularly important, because it indicates that deformation is not related to any single depositional process.

**Deformation Showing a Systematic Increase in Intensity of Deformation**

Zoning of deformation that shows a systematic increase in intensity toward a likely epicentral area is a very important indicator of seismicity, because such zoning in an outwardly decreasing manner is a hallmark of seismicity (Obermeier, 1998; Wheeler, 2002). However, certain limitations apply when using this criterion (e.g., Woolery, 2008). Based on the rationale of Lowe (1975) and the mapped intensity and thickness of deformation in six deformed horizons, zoning is clearly present with probable epicentral areas located in each case. In most cases, probable epicentral areas are located near segments of basement faults, suggesting that these segments may have been the sites of ancient fault rupture. However, in the case of the uppermost deformed horizon in Cycle 5, the most intense deformation appears to be located in two different areas, making it difficult to interpret.

The lowermost deformed horizon (Fig. 6.2, in Cycle 1) has the least areal extent of deformation among the six, and this lesser extent could be due to its limited presence at the base of most exposures. Fluidization, the most energy-intensive deformation process, was not observed in this deformed horizon, suggesting that early Tanglewood seismicity was not strong enough to cause the fluidization of sediments or that the sediments were too coarse. Liquefaction is the highest energy deformation process observed in this horizon, which is best developed in
the northeast corner in the vicinity of basement fault D and on the western flank of the shoal complex between the two basement fault segments A and B. However, based on the extent and thickness of liquefaction in these areas, the main epicentral areas appear to have been near the northeast corner of the upper Tanglewood shoal complex. The more local occurrence of deformation near fault segment A may reflect a secondary reaction to the initial rupture near fault D.

The lower deformed horizon in Cycle 2 (Fig. 6.3) is more extensive in distribution than the lowermost horizon in Cycle 1. Fluidization is present in only two exposures on the western flank near the basement fault B. Liquefaction-type deformation is also present in equivalent parts of the Clays Ferry and Point Pleasant formations. Zoning of the deformation is present in three areas: the northeast corner along the vicinity of basement fault D, in the middle part of the northern upper Tanglewood boundary (between basement faults A and C), on the western flank (between basement faults A and B), and in the Point Pleasant Formation near fault E. Though the fluidized horizon on the western flank, near basement fault B, does not show a liquefied horizon in the surrounding exposures, which is supposed to have happened for zoning of the deformation resulting from seismogenic origin, the intensity of deformation here, and only here, indicates that the epicentral area must have been near this area. The lack of liquefaction near the fluidized beds could be related to two factors: the lack of exposures in the surrounding area, or the facies in surrounding exposures may have been too coarse-grained (calcirudite), which is less prone to deformation than the finer-grained, lower shoalface (alternating layers of calcarenites and shale) facies. Very coarse-grained facies have more porosity with lesser ability to maintain high pore pressures.
The upper deformed horizon in Cycle 2 is the most extensive in distribution among the six deformed horizons in the upper Tanglewood Member, also exhibiting deformation in laterally equivalent horizons of the Clays Ferry and Point Pleasant formations (Fig. 6.4). The extent and intensity of deformation in this horizon suggest that the seismicity during this episode was the greatest experienced by the shoal complex. Fluidization is present in this horizon in the northeast corner of the Tanglewood complex and in the Point Pleasant Formation (Fig. 6.4). Because the northeast corner of the upper Tanglewood shoal complex along basement fault D shows the characteristic zoning of deformation from hydroplastic deformation to fluidization, it appears that this area was near the ancient earthquake epicenter on fault D.

The lower deformed horizon in Cycle 3 has a very scattered pattern of deformation (Fig. 6.5), but no deformation is present in the southeast corner of the upper Tanglewood buildup and in the equivalent horizon of the Clays Ferry Formation at the type locality. Liquefied beds are present in only one exposure of the Point Pleasant Formation. In contrast, the northeast corner has a zoned distribution of deformation with the highest intensity being liquefaction, which is concentrated on the northern parts of basement fault D. On the western flank of the upper Tanglewood shoal complex, fluidization occurs in two separate exposures, which are too far apart to be directly connected. One occurrence lies north of the basement fault A and another lies on the south side of basement fault B (Fig. 6.5). Zoning is more or less displayed in the first while it is not observed in the second. This could be related to the fact that only limited exposures are available in the area, and those that are nearby are composed of mainly massive, coarse-grained calcirudite, which is less prone to deformation. On the basis of the thickness of the fluidized horizons in the two exposures on the southwest corner of the shoal complex, it is
assumed that the southwest corner, probably near fault segment A, represents the ancient epicentral area.

The upper deformed horizon in Cycle 3 (Fig. 6.6) shows a more or less similar pattern of scattered deformation except that the deformational zone in this part of the cycle also includes the southeast corner of the shoal complex, as well as the equivalent horizon in the Clays Ferry Formation. Zoned deformation with liquefaction and fluidization is present in the northeast corner (Fig. 6.6). In fact, fluidization is observed in three separate areas in this horizon. However, on the basis of the lateral extent of deformation, as well as of the thickness of the deformed horizons and zoning of the deformation, the northeast corner of the buildup appears to have been the epicentral area (Fig. 6.6).

In most exposures of the upper Tanglewood, the last cycle, Cycle 5, has been eroded away. Therefore, the deformed bed at the base of this cycle cannot be observed in most exposures. However, in the northeast corner, most sections are complete, and in this area, a fluidized horizon is observed in one exposure without any zoning of deformation intensity. Very extensive deformation of this horizon is present in equivalent parts of the Point Pleasant Formation (Fig. 6.7), where mostly liquefied horizons are observed. Because there is no zoning of deformation in the upper Tanglewood due to lack of exposures, it is difficult to interpret a possible epicentral area in the upper Tanglewood. However, the extensive deformation of this horizon in the Point Pleasant Formation does display some indication of zoning where fluidized and liquefied beds are observed. Therefore, it is more likely that this area in the equivalent Point Pleasant Formation on fault segment E represents the likely ancient earthquake epicenter. Overall, the zoning of deformation intensity in most of the deformed horizons clearly indicates a seismic origin for the deformation. One of the limitations in applying this criterion, however, is
the heterogeneity of field conditions. Actually, subsurface bedrock geometry can have a sizable effect on the local earthquake ground motions (e.g., Woolery, 2008). In fact, site effects are thought to be primarily responsible for the disproportionate amount of shaking. Woolery (2008) has shown that such concentric zonation of increasing the intensity of ground shaking did not occur for the July, 1980, northeastern Kentucky earthquake. Therefore, I do not rely wholly on this criterion to interpret seismogenic deformation.

**Excluding Other Causes**

As for excluding other causes, load-induced (i.e., nonseismic) liquefaction could be one factor for producing soft-sediment deformation. Load-induced deformation occurs in situations where thicker and coarser-grained sediments are deposited over thinner and finer-grained sediments. Normally, due to such loading, hydroplastic deformation to liquefaction could be expected. The pattern of deformation in the upper Tanglewood is shown in Figure 6.8. Nearly all of the deformed horizons are overlain by fine-grained sediments, and hence, it is difficult to call on load-induced deformational phenomenon alone to explain the deformation, as there was no sufficient load to produce the widespread and highly deformed horizons seen in the upper Tanglewood.

In addition to loading by coarser, denser sediments, storms may also produce soft-sediment deformation through bottom pressure differential during wave oscillation (Kraft, 1985a; Okusa, 1985) and through the basal drag force of bottom storm currents (Lowe, 1976; Orange and Breeh, 1992). The upper Tanglewood represents a carbonate-shoal complex, which was deposited above the storm wave base. Therefore, the possibility of storm effects on the sediments cannot be ruled out. However, if storms had been the main factor in producing deformation, we should have observed it everywhere in the upper Tanglewood shoal complex,
but it does not occur everywhere. Rather, the deformation appears to be concentrated around basement faults. Moreover, sedimentary structures indicative of the oscillation necessary to produce bottom pressure differentials are relatively uncommon in the upper Tanglewood. It is also difficult to explain the concentric patterns of deformation intensity and the repetition of intense deformation along the fault segments through a storm agency. In addition, storm-generated sedimentary structures such as flow structures, sole markings, parting lineations, and hummocky cross-stratifications are not present in the deformed beds.

Fig. 6.8 Measured section along US Highway 68 in Nicholas County. Section, measured along US Highway 68 in Nicholas County (38°21'57.48"N, 84° 2'20.46"W), showing the nature of deformed horizons. Note the uppermost, thick deformed horizon overlain by shale and rhythmites (alternating layers of calcarenite and shale facies). These facies seem incapable of inducing sufficient load to cause fluidization in the thick underlying horizons. The same case applies to the lowermost deformed horizon.
Slope-induced mass movement is another alternative mechanism to cause soft-sediment deformation, but such movement is more typical of the steeper, unstable slopes encountered in channels or near point sediment sources like deltas (Jewell and Ettensohn, 2004). Neither channels nor point sources are associated with the upper Tanglewood, Clays Ferry or Point Pleasant formations. Local channels probably did occur within the shoal complex, but local channels could not account for such widespread pattern of deformation. The possibility of higher slopes near the faults cannot be ruled out, but deformation can also be observed distant (>50 km) from the faults.

Other Evidences

Aside from concurrence of the four criteria noted above, a few other associations also point to seismic origins for the deformation. The most important of these is that the most intensely deformed horizons are concentrated around basement faults. The upper Tanglewood is already known to be a structurally controlled buildup related to reactivation of basement faults by Taconian far-field forces (Ettensohn et al. 2002a, b). Thus, seismic activity is most likely related to reactivation of local basement faults. Based on the presence of Ordovician synsedimentary faulting (Black and Haney, 1975) and the coincidence of many Lexington facies boundaries with structural lineaments (e.g., Ettensohn et al., 1986, 2004; Ettensohn, 1992; Ettensohn and Kulp, 1995; Clepper, 2011), synsedimentary reactivation of structures in the area must have been common and is an accepted means of generating facies changes and deformation (Koirala et al., 2016). Therefore, the Tanglewood buildup occurred in a suitable distal tectonic setting to experience seismicity.

The subtle crosscutting of deformed and undeformed beds (Fig. 5.10) is especially diagnostic of a seismogenic origin (Davaport and Ringrose, 1989; Rast et al., 1999). Crosscutting
relationships involving only underlying sediments could be due to erosion, but such relationship that involves overlying rocks is difficult to explain where the possibility of onlap can be ruled out (Ettensohn et al. 2002d). Ettensohn et al. (2002d) have also pointed out that seismically augmented pore pressure can only influence susceptible sediments, which, because of the vagaries in cementation or clay content, may occur at oblique angles to the normal bedding, thus appearing to have been truncated from below or above (Fig. 5.10).

Hence, based on agreement among criteria supporting a seismogenic origin, exclusion of the other most plausible mechanisms for deformation, and favorable comparison with experimentally derived and modern seismicity, we conclude that Late Ordovician, upper Tanglewood deformation is most likely of seismogenic origin.

Estimating Paleoearthquake Magnitude for the Deformed Horizons

If the soft-sediment deformation can be attributed to seismogenic causes, then the minimum-magnitude earthquake required to produce deformation is 5 or larger (Carter and Seed, 1988; Ambraseys, 1988; Obermeier et al., 1990). This number then represents the minimum magnitude of the earthquakes that generated the deformed horizons in the upper Tanglewood Member. Other methods have also been suggested by various researchers to estimate the magnitude of paleo-earthquakes. For example, Ambraseys (1988) has proposed the magnitude-bound method, in which distance from the epicenter to the farthest observed deformation is compared to a curve developed from historical observations (Fig. 6.1) to estimate moment magnitude for paleo-earthquakes. This method is based on the idea that for the shallow-focus earthquakes (<50 km), the greater the magnitude of earthquakes, the greater the energy input,
and as a result, the wider the distribution of deformation from the epicenter, given that uniform sedimentary conditions exist (Youd and Perkins, 1978; Ambraseys, 1988).

Epicentral distances were measured from Figures 6.2—6.7 for the deformed horizons of the upper Tanglewood relative to their apparent epicenters. Comparison to the curve (Obermeier, 1998) (Fig. 6.1) gives the following moment magnitudes: for the deformed horizon in Cycle-1, 7—8; for the lower deformed horizon in Cycle-2, 7.2—8; for the upper deformed horizon in Cycle-2, 7—8; for the lower deformed horizon in Cycle-3, 7—8; for the upper deformed horizon in Cycle-3, 7—7.7; and for the deformed horizon in Cycle-5, 7—7.5.

The magnitude-bound method to estimate the moment magnitude of paleo-earthquakes has certain limitations in applying it to the upper Tanglewood depositional setting. This method was developed for siliciclastic sediments in terrestrial settings. Therefore, exactly how this method may be applied to carbonate-rich sediments in marine setting is not certain. This method does not take into account the effects of the overlying water column or the greater density of carbonates. Because fluidization velocity ($U_0$) is directly proportional to the density of the phases involved (Lowe, 1975), it is possible that deformation in carbonates requires more energy than that involved in deforming siliciclastics. Moreover, this method assumes that liquefaction was the major deforming process (Obermeier, 2003). Hence, this method may not be accurate in the upper Tanglewood where hydroplastic deformation is the predominant deforming process except for the upper deformed horizon in Cycle 2. In addition, various factors, including varying site effects and the lack of exposures more distant from the deformed sections, may have also generated the above results.

No such direct surface manifestation of earthquakes, such as surface rupture or displacement of one block relative to the other, has been observed in the field. However, the
surface faults in the area might be one indicator of former earthquakes. An empirical relationship between magnitude and surface-rupture length for shallow-focus (<40 Km), continental interplate or intraplate earthquakes with magnitudes greater than 4.5 is provided by Well and Smith (1994). The equation is given below, where SRL=Surface Rupture Length

\[ M = 5.08 + 1.16 \times \log (SRL) \]

Using the equation above, we can calculate the required rupture length to produce the earthquake moment magnitude of 5, which is the minimum magnitude to produce deformation, to be 0.85 km. On the assumption that the rupture took place along the entire length of the subsurface faults, this method can be applied in our setting to estimate maximum possible earthquake magnitude that might have occurred during different phases of deformation.

For the deformed horizon in Cycle 1, the epicentral area lies along the fault line D (Fig. 6.2). Assuming the maximum surface rupture along the entire fault, which is about 80 km, the maximum possible earthquake is 7.28. Similarly, the maximum possible moment magnitude of a paleoearthquake for the lower deformed horizon in Cycle 2 is 7.47; for the upper deformed horizon in Cycle 2 is 7.28; for the lower deformed horizon in Cycle 3 is 7.34; for the upper deformed horizon in Cycle 3 is 7.28; and for the deformed horizon in Cycle 5 is 7.28. This method only gives the approximate maximum moment magnitude of the earthquake that could have hit the central Kentucky area based on the length of the exposed fault lines. Therefore, it can be assumed that no earthquake greater than of 7.5 moment magnitude has hit the area.
Why is some deformation so scattered?

Examination of the deformation map for each horizon (Figs. 6.2–6.7) indicates that in each horizon, a larger, central area of more intense deformation near the probable originating fault segment is detached from “outliers” of lesser deformation, commonly associated with other fault segments in the area. There were six major fault segments in the area, and deformation appears to have been separately concentrated along these fault lines. One possibility is that all deformation in any of these horizons is related to the movement on the same basement precursor fault in the epicentral area. Another possibility is that the earthquakes along one fault segment triggered other earthquakes on the nearby faults at the same or slightly later times. Since there is a growing realization that earthquakes on one structure can trigger other earthquakes on the same or nearby faults at nearly the same or later times through stress transfer (Stein et al., 1994; Harris et al., 1995; Yeats et al., 1997; Reilinger et al., 2000; Kilb et al., 2000), the possibility that those faults could have been reactivated at the same or later times cannot be ruled out. Although it cannot be verified for patterns of deformation in the upper Tanglewood, it is an interesting suggestion. As an alternative, other factors related to site effects (e.g., Reiter, 1990), such as local basin shape, sediment thickness, sediment type, and the nature of the involved structures at depth, could also explain the lack of deformation at different places in response to the same seismic event. Lack of exposures in the surrounding areas also makes it difficult to determine whether or not the central area of deformation is connected with the “outliers” of lesser deformation.
Chapter 7—Petrographic Study

Introduction

Sediment behavior under stress is an important phenomenon that should be tested by comparing the petrophysical properties in undeformed and deformed portions of the same sediment horizon, both in terms of preferred orientation of constituent grains and cementing materials. No such microscopic studies regarding deformed versus undeformed portions of soft-sediment-deformation horizons has currently been performed. In addition to diagenetic history and depositional-environment interpretation, this study will try to address some of the petrographic differentiation that occurs during soft-sediment deformation by analyzing petrographic variation between deformed and undeformed parts of the same horizon.

Sampling and Testing Methodology

During field studies, block samples (minimum 15cm×10cm×10cm) were collected from deformed and undeformed parts of the same horizon. The samples were carefully selected to avoid any signs of weathering and macroscopic heterogeneity (e.g., veins, fractures, etc.). All of the block samples were collected from fresh profiles. Then thin sections were prepared from these samples. While making thin sections, the slides were labelled to differentiate up and down orientations on the sample. For the purpose of distinguishing between calcite and dolomite, all sections were stained with Alizarin red-S.
Petrography

The upper Tanglewood limestones consists mainly of packstones and grainstones (Fig. 7.1 a, b).

Fig. 7.1(a) Photomicrograph of thin section (25X) under PPL. Photograph showing an upper Tanglewood limestone in thin section (25X) under PPL. Note the sparry-calcite cement between grains, syntaxial-rim cement around echinoderm grains, and allochems of bryozoan and brachiopod fragments, among others. The bryozoan zooarium in the upper right quadrant is infilled with phosphorite. This sample was taken from an undeformed horizon in Cycle 2 from Anderson County, KY (38° 05′ 45.45″ N, 84° 55′ 09.66″ W). The facies is of typical shoal facies. Note the preferential orientation of elongate grains. The grains and grain orientation are typical of the shoal facies.
Photomicrograph of the thin section (25X) in Fig. 7.1a under XPL. Photograph showing an upper Tanglewood limestone in thin section (25X) under XPL. Note the sparry-calcite cement between the grains, syntaxial-rim cement around echinoderm grains, and allochems of bryozoan and brachiopod fragments, among others. The bryozoan zooarium in the upper right quadrant is infilled with phosphorite. The sample was taken from an undeformed horizon in Cycle 2 from Anderson County, KY (38° 05′ 45.45″ N, 84° 55′ 09.66″W). The grains and grain orientations are typical of the shoal facies.

The major components are skeletal and non-skeletal grains, abundant sparry-calcite cement, and a few quartz grains. A matrix of micrite is rarely present, but micritization of the grains is abundant. The grain size also ranges from very fine-grained (silt-size) to coarse-grained (rudite-
size) depending upon location in the vertical and horizontal profiles. Equivalent horizons in the Clays Ferry Formation are finer-grained than their counterparts in the upper Tanglewood. The results of the microscopic studies are briefly discussed in the following sections.

**Characteristic grains (Allochem composition)**

**Skeletal grains**— Skeletal grains are mostly sand-size or larger (Fig. 7.1). Shell shapes are generally well-preserved. Recrystallization of original shell structures to sparry calcite is also quite common. The original microstructure of the grain, in most cases, has been obliterated by diagenesis, but the micritic envelope around the boundary maintains the original shape of the skeletal grain. Syntaxial-rim cements on crinoids are abundant (Fig. 7.1). The thin coatings are composed of dark micrite.

**Non-skeletal grains**— Non-skeletal grains are less abundant and mainly consist of peloids and lesser quantities of intraclasts (Fig. 7.2). Intraclasts are more common in deformed beds. Intraclasts containing fossils suggest that the winnowed platform-edge zone was very shallow and that the intraclasts were formed by reworking of semi-lithified sediments by moderate currents. Peloids are occasional in some samples and are surrounded by sparry calcite cement (Fig. 7.3). The presence of fecal pellets usually indicates low-energy, warm seas supersaturated with respect to CaCO₃. They may have been derived from deeper portions of the shoal complex.

**Diagenesis**

After deposition, the largely biochemical sediments were buried under successive layers of younger sediments. The increasing temperatures and pressures encountered during burial
bring about diagenesis of the sediments, leading to dissolution and destruction of some constituents, generation of some new minerals in the sediment, and eventually consolidation and lithification of the sediments into sedimentary rock (Boggs, 2009).

![Photomicrograph of intraclasts in thin section (100X). Photomicrograph showing an upper Tanglewood limestone in thin section (100X) under XPL. Note the intraclasts on upper right-hand quadrangle. The sample was taken from a deformed portion of the lower deformed horizon in Cycle 2 from Anderson County, KY (38° 3’ 30.90”N, 84° 55’ 15.30”W). Note the lack of well-defined grain orientation.](image)

**Fig. 7.2 Photomicrograph of intraclasts in thin section (100X).** Photomicrograph showing an upper Tanglewood limestone in thin section (100X) under XPL. Note the intraclasts on upper right-hand quadrangle. The sample was taken from a deformed portion of the lower deformed horizon in Cycle 2 from Anderson County, KY (38° 3’ 30.90”N, 84° 55’ 15.30”W). Note the lack of well-defined grain orientation.
Petrographic investigations on the basis of randomly selected samples indicate that the
diagenetic processes, which have modified the sequence of the upper Tanglewood Member of
the Lexington Limestone, consist of micritization, cementation, compaction, dissolution,
dolomitization and internal filling.

Fig. 7.3 Peloids and micrite coatings in a thin-section photomicrograph (100X). Photomicrograph
showing an upper Tanglewood limestone in thin section (100X) under XPL. Note the dark-brown rounded
grains (peloids) in the sample. Also, note the micrite coatings on some allochems. The sample was taken
from an undeformed portion of upper deformed horizon in Cycle 2 from Georgetown, KY (38° 16′ 15.52″
N, 84° 33′ 09.03″W). Note the preferential orientation of elongate grains.
**Micritization**

Micritization, as a type of relatively weak marine diagenetic process, takes place in disturbed or intermittently disturbed shallow-water environments. The process requires that carbonate grains remain on the seafloor for a long period of time rather than being quickly buried (Wei, 1995; Kabanov, 2000). Micritization is the process whereby the margins of carbonate grains are replaced by micrite at or just below the sediment/water interface. The process involves microbes attacking the outside of grains by boring small holes in them, which are later filled with micritic cement (Adams and MacKenzie, 1998). Endolithic algae, fungi, and bacteria play a major role in bioerosion by producing micritic envelopes, occurring within the phreatic zone (Bathurst, 1980). Repeated micritization may lead to destruction of the original grain structure. Micritization is obvious with a dark rim around skeletal grains such as brachiopods and bryozoans in the studied limestones (Fig. 7.1).

**Cementation**

Obvious signs of the effect of diagenesis are pore-filling cements which are commonly seen in the thin sections of carbonate rocks. Cementation is the process of precipitation of space-filling crystals between the grains (Adams and MacKenzie, 1998). The carbonate cementation process is initiated soon after deposition, during which intergranular pore space is progressively reduced, producing systematic changes in petrophysical properties. The cementation and diagenesis of carbonates can take place in many settings: in the marine environment during the deposition of the sediment, in the surficial vadose environment, and in the freshwater phreatic environment.
The marine phreatic environment, in most cases, is where carbonate sediments originate and begin their diagenetic history. The cementation rate is greatest in the active marine-phreatic environment where three conditions occur:

- pH increases above 9 due to photosynthesis and respiration of a reef biomass
- CO₂ degassing
- Waves, tides, or currents force water through pores (works best at shelf margins where buildup is present or along shoreface).

High-magnesium calcite or aragonite are the only cements that precipitate in the active marine phreatic zone. Both are unstable in Mg-deficient water regardless of whether it is marine, brackish, or fresh, and both tend to alter to low-Mg calcite, because most water is magnesium-deficient. Their common form is isopachous coatings on grains. Micritization of grains occurs in the active marine phreatic zone.

Cementation and dissolution of carbonates initiated in the marine realm continue during contact with fresh water, along with dewatering, grain orientation and grain breakage during earliest burial compaction. Vadose precipitation begins where water in the vadose zone becomes saturated with CaCO₃. Cementation is generally minor and reflects pore-water distribution. Meniscus cement precipitates, where water clings between grains in a meniscus form. Pendulous or microstalactitic cement precipitates where water droplets form underneath grains. No vadose cements were observed in the upper tongue of the Tanglewood, meaning that the unit was never exposed in its depositional setting.

The active freshwater phreatic zone is a terrestrial development. Meteoric water that enters the phreatic zone without passing through the vadose zone, is typically undersaturated
with respect to CaCO₃ but becomes saturated as it dissolves grains. Based on CaCO₃ saturation, the active freshwater phreatic zone may be subdivided into undersaturated and saturated zones. In the undersaturated zone, solution occurs, creating moldic or vuggy porosity. In an active, saturated, freshwater phreatic zone, extensive and rapid cementation occurs. The cement is typically an equant calcite spar that coarsens toward pore centers. Syntaxial overgrowths on echinoderm fragments are common, and these cements are common in the upper Tongue of the Tanglewood (Fig. 7.1).

The precipitation of cement is evident in the examined carbonate rocks. It is developed within both intergranular and intragranular pores as well as in fractures. Most of the cements in the upper Tanglewood limestones appear to be late diagenetic burial cements. Mainly two generations of sparry-calcite cementation are recognized in the upper Tanglewood limestones: freshwater phreatic cement between the grains and within the grains (intergranular and intragranular cements), and cementation in fractures.

The earliest generation of cement which is common in the marine environment is fine crystalline-rim cement around allochems. However, such cements are rare in the upper Tanglewood apparently due to later dissolution and reprecipitation in the phreatic environment. The second generation of cement is relatively large in size and fills the spaces between and within allochems and was largely produced by the dissolution of early aragonite cements (Fig. 7.4). It is characterized by equant, sparry calcite. Drusy calcite spar is also a characteristic pore-filling cement with an increasing crystal size toward the center of most pores. Syntaxial-rim cement is obviously seen on crinoids (Fig. 7.1), and this generation of cement is characteristic of the freshwater-phreatic environment as indicated by its geometry. A third generation of cement
occupies fractures (Fig. 7.5). Therefore, cementation in these rocks primarily reflects late-stage diagenesis after burial in the freshwater phreatic zone.

Fig. 7.4 Sparry calcite cement in a thin-section photomicrograph (100X). Photomicrograph showing an upper Tanglewood limestone in thin section (25X) under XPL. Note the abundant sparry-calcite cement. The sample was taken from an undeformed horizon in Cycle 2 from Anderson County, KY (38° 05' 45.45" N, 84° 55' 09.66"W). The example is typical of the shoal facies. The dark hole on the upward side of the slide is a cut in the thin-section chip to represent the up direction.
Compaction

Compaction processes have been observed in the studied limestones. Before cementation, compaction takes place dominantly by mechanical processes. With increase in overburden pressure, more stress is placed on the grain-to-grain contacts, and compaction starts to proceed by chemical dissolution at grain boundaries. Both mechanical and chemical compaction processes start under an overburden of tens to a few hundred meters. Meyers (1980) stated that most non-styloitic compaction is completed before burial by about 1980 m of overburden. Under a thicker overburden, a well-developed, thorough-going styloitic compression takes place. Both types of compaction are observed in the studied samples.

Mechanical compaction

Mechanical compaction is associated with dehydration, porosity reduction, and significant decrease of up to one quarter of the original thickness of sediments (Flügel, 2004). This process includes the repacking of allochemical components in carbonates. The parallelism of bryozoans can even be visible in hand specimens. The types of grain contacts produced by mechanical compaction in the studied limestones reflect point and tangential contacts.

Chemical compaction or pressure dissolution

Chemical compaction is a potentially important cause of the dissolution of calcium carbonates and the formation of stylolites, which penetrate all the studied rocks, apart from the allochems. Later, the dissolved calcium carbonate may be precipitated in suitable openings or fissures, forming calcite veinlets (Wanless, 1979). Progressive solution leads to alteration of the grain-to-grain contacts from the original point contacts, through planar (or tangential) contacts to interpenetration (concavo-convex), and sutural grain-to-grain contacts. Chemical compaction is a
very important diagenetic process in a burial environment. In addition, producing a pressure-dissolution fabric, such as stylolites, leads to the dissolution of grains and matrix, which is an important source of burial-calcite cement (Zhang et al., 2006). Chemical compaction processes within the upper Tanglewood limestone resulting in pressure solution are not obvious in thin section; only interpenetrating grains are obvious (Fig. 7.5).

Fig. 7.5 Internal fracture filling in a thin-section photomicrograph (100X). Photomicrograph showing an internal fracture filling by sparry calcite in an upper Tanglewood limestone in thin section under XPL (40X). The sample was taken from a ball-and-pillow structure in the upper deformed horizon of Cycle 2 from Georgetown, KY (38° 16’ 15.52” N, 84° 33’ 09.03” W). Note the random orientation of grains.
Internal filling

Internal filling (or fracture and cavity filling) is common in the studied rocks (Fig. 7.6). Fractures are predominantly filled with drusy and granular, mosaic, calcite cement, which indicates formation in a meteoric phreatic environment at the mesogenesis stage.

Fig. 7.6 Interpenetrating grains in a thin-section photomicrograph (100X). Photomicrograph showing interpenetrating grains in an upper Tanglewood limestone in thin section under XPL (25X). Note the grain contacts. The sample was taken from an undeformed horizon in Cycle 2 from Anderson County, KY (38° 05′ 45.45″ N, 84° 55′ 09.66″ W). The facies is of typical of the shoal facies.
The later diagenetic stage, or mesogenesis, involving increased burial and compaction, leads to grain reorientation, breakage and deformation of bioclasts, development of pressure-solution seams, micro-stylolites and stylolites. Formation of coarse sparry calcite cement (Fig. 7.6) probably developed as a result of re-precipitation following pressure dissolution.

**Dolomitization**

Dolomitization is a diagenetic process that converts limestones to dolostones through a microchemical process of calcium carbonate dissolution and dolomite precipitation. Dolomitization can change the rock fabric and the petrophysical properties significantly because the dolomite crystals are commonly larger than the replaced limestone grains. Dolomite cement systematically grows on dolomite crystal faces, reducing porosity. Dolomitization requires the addition of large quantities of magnesium through fluid flow. Dolomitization is quite obvious in most upper Tanglewood samples. Mainly euhedral crystals are present (Fig. 7.7). The source of extra magnesium probably came from the dissolution of high-magnesium cement, echinoderm grains, or from clays.

**Petrographic Difference between Deformed and Undeformed Portions of a Deformed Horizon**

In comparing the deformed and undeformed portions of the same horizon, it appears that the main difference lies in the preferred orientation of elongate grains. In undeformed portions of
the horizon, the elongate grains appear to be preferentially oriented; the grains are aligned parallel to the original depositional plane (Fig. 7.8).

![Photomicrograph showing dolomitization in an upper Tanglewood limestone in thin section under XPL (100X). Also, note the haphazard orientation of elongated grains (almost vertical to depositional plane or at least at significantly inclined angles to the depositional plane) (100X) relative to the depositional plane. The sample was taken from the deformed portion of an upper deformed horizon in Cycle 2 from Georgetown, KY (38° 16′ 15.52″ N, 84° 33′ 09.03″W). The indentation at the top side of the thin section is a cut in the thin-section chip to show the up direction. The dark-brown grains are phosphatized peloids and fossil fragments.]

**Fig. 7.7 Dolomitization in a thin section photomicrograph.** Photomicrograph showing dolomitization in an upper Tanglewood limestone in thin section under XPL (100X). Also, note the haphazard orientation of elongated grains (almost vertical to depositional plane or at least at significantly inclined angles to the depositional plane) (100X) relative to the depositional plane. The sample was taken from the deformed portion of an upper deformed horizon in Cycle 2 from Georgetown, KY (38° 16′ 15.52″ N, 84° 33′ 09.03″W). The indentation at the top side of the thin section is a cut in the thin-section chip to show the up direction. The dark-brown grains are phosphatized peloids and fossil fragments.
Fig. 7.8 Preferential orientation of elongated allochems in a thin-section photomicrograph. Photomicrograph showing the preferential orientation of elongated allochems parallel to the former depositional plane under XPL (100X). The sample is taken from an undeformed portion of the upper deformed horizon in Cycle 2 from Georgetown, KY (38° 16′ 15.52″ N, 84° 33′ 09.03″ W). The indentation on the top side of the photograph is a cut in the thin-section chip to show the up direction. The dark-brown grains are phosphatized peloids and fossil fragments.
Waves and storms typically transport sediment to a particular location and deposit it such that the elongate grains are preferentially oriented; compaction will only enhance this orientation. In deformed beds, however, the grains are aligned haphazardly or randomly. This is due to the fact that during deformation the original grain alignments get obliterated due to water escaping randomly through the sediment. More precisely, in some cases, the elongate grains appear to be aligned vertically to the depositional plane (Fig. 7.7). Such orientation can be attributed to the fact that during deformation, the water in this area was escaping vertically. If the water escapes in another direction, the grains will be aligned accordingly.

Another difference between deformed and undeformed sediments comes from the fact that intraclasts are more common in the deformed beds (Fig. 7.2). This is due to the fact that weakly cemented rock fragments or intraclasts become broken during deformation and were re-cemented during later stages of diagenesis. Dolomitization is common in deformed beds as well as in undeformed beds (Fig. 7.7).

The nature of cement did not show any differences between deformed and undeformed beds. It appears that the cementation primarily took place after sediment deformation. Because most of the intact framework of the sediments appears to be preserved, it may be possible that the sediments underwent weak marine cementation before deformation during early diagenesis. This possibility is bolstered by the fact that the rock contains many fragmented intraclasts. Therefore, it may be more realistic to assume that the initial marine cements might have been obliterated during deformation and later stages of diagenesis. The main episode of cementation probably took place after sediment deformation when the sediment came into contact with the ground water during later uplift.
Conclusions

The petrographic study of these limestones has resulted in the following conclusions. Diagenetic modification of the upper Tanglewood limestones commenced with grain micritization. This was accompanied by equant isopachous-calcite cementation at the eogenesis stage. Dissolution then largely removed this rim cement. Precipitation of equant sparry-calcite cements was followed by recrystallization of micrite into microsparite with increasing burial depth. At the mesogenesis stage with increasing burial and compaction, grain orientation and fitted-fabric texture developed, and this in turn led to the development of pressure-solution seams, micro-stylolites and stylolites. Fracture development was followed by precipitation of drusy calcite cement. The final stage of cementation occurred during uplift and deep burial when equant sparry-calcite and syntaxial-rim cements were deposited. Most of this cementation occurred after deformation. Although deformation may have destroyed some of the very early isopachous cements, its main effect was to randomize sediment distribution and generate more fractured intraclasts.
Chapter 8—Summary

Discussion

The upper tongue of the Tanglewood Member of the Lexington Limestone is the uppermost member of the Lexington Limestone, which overall is a largely shallow-water, shoal-like, carbonate accumulation that developed across an uplifted part of the intracratonic Lexington Platform in central Kentucky. The Lexington Limestone developed coevally during the most intense phase of the Taconian Orogeny to the east, and all aspects of its deposition from formation of the platform to individual facies development suggest interaction with basement structures reactivated by Taconian far-field forces (e.g., Rast et al., 1999; Kolata et al., 2001; Ettensohn et al., 2002a; McLaughlin and Brett, 2004). Because the basement is inhomogeneous and rife with brittle deformation left over from older episodes of lithospheric tension and compression (Fig. 1.2), these dislocations remain zones of weakness that can be reactivated when exposed to new stresses (Ettensohn and Lierman, 2012). Hence, the apparent reactivation of such structures during deposition of the Lexington Limestone, where other causes are absent (Ettensohn et al., 2002a), is inferred to have been caused by large-scale Taconian compressional and tensional stresses, or far-field forces, impelled westward along a network of old basement dislocations during coeval orogeny. The coincidence of facies and thickness changes with basement structures throughout the Lexington Limestone, and in particular those noted in the upper Tanglewood Member, clearly shows the efficacy of these forces, even in areas as far removed from orogeny as the Lexington Platform.
Overall, the Lexington Limestone is a transgressive unit across the platform, but in mid-Chatfieldian time (Fig. 1.1), carbonates representing shallower parts of the unit abruptly transitioned into deeper water shales and interbedded, fine-grained limestones in units like the Clays Ferry, Kope, Point Pleasant, and Maquoketa formations. This transition coincides with a major sea-level high-stand at about 451.8 Ma (Cooper and Sadler, 2012). This deepening event was experienced across the Lexington Platform, except in the central Kentucky area, where shallow-water, high-energy carbonates persisted as the Tanglewood buildup into early Edenian time as a relatively small, intra-platform shoal complex (Ettensohn, 1992; Ettensohn and Kulp, 1995; Ettensohn et al., 2002a, 2004; Koirala et al., 2016). The fact that the buildup and its coarse skeletal sands coincided with basement structures led Ettensohn (1992) and Ettensohn and Kulp (1995) to conclude that it was structurally controlled and involved reactivated basement fault zones. Later work has shown that basement structures also probably controlled smaller-scale facies distribution within the buildup (Ettensohn et al., 2002a, 2004; Clepper, 2011). Although the Tanglewood buildup is composed of three tongues of the Tanglewood Member (Fig. 1.1), only the upper tongue is the subject of this study as it represents the final phase of Lexington shoal-complex deposition before the area was completely inundated by deeper waters. Upper tongue deposition was initiated near the beginning of Edenian time at about 451 Ma after a regional flooding event, represented by a tongue of the Clays Ferry Formation or Millersburg Member (Fig. 1.1), briefly inundated the area.

Previous work on the underlying middle tongue of the Tanglewood has shown that the structural blocks between basement structures A and B and between structures C and D had previously acted as blocks that experienced synsedimentary uplift due to Taconian far-field forces (Ettensohn et al., 2004). Ettensohn et al. (2004) demonstrated with isopach maps that on
these uplifted blocks, thick sequences of middle-tongue, shoalface calcarenites and beach facies in the Devils Hollow Member (Fig. 1.1) had developed. Because overall upper Tanglewood Member shoal formation appears to have been similarly concentrated between the same basement blocks, there is a good reason to support the idea that renewed synsedimentary uplift of the blocks along old growth faults probably contributed to upper Tanglewood shoal formation. It is also possible that relict topography from middle Tanglewood shoals provided nuclei for subsequent development of upper Tanglewood shoals, although in either of the above possibilities, reactivated basement structures seem to have provided the initial impetus for shoal development. Although evidence has accrued that several different basement faults were reactivated at different times and places during deposition of the Lexington Limestone in central Kentucky (Ettensohn, 1992; Ettensohn and Kulp, 1995; Ettensohn et al., 2002a, 2004; Clepper, 2011), by the time of the early Edenian transgression, most of the former Tanglewood buildup had been inundated by deeper seas, and only in the small shoal-complex area around and between the two uplifted fault blocks (Fig. 1.2) did shallow-water Lexington Limestone deposition continue. As Figure 4.5 shows, shoal deposition began on the two uplifted fault blocks and expanded toward the center of the shoal complex. Shoal deposition had largely ended by the deposition of sequence four, although the fact that lower shoalface deposition persisted between structures A and D (Fig. 4.5, no. 4) may indicate inactivity on structure C with some interaction between structures A and D.

The lower central parts of the shoal complex, which included mostly offshoal/platform lagoonal or lower shoalface settings (Fig. 4.5), apparently supplied slightly deeper, stable bottoms that were bathed in cold, nutrient-rich waters which upwelled from the Sebree Trough (Ettensohn et al., 2002a; Ettensohn, 2010) (Figs. 1.10 and 1.11). In this setting, prolific
bryozoan-brachiopod and echinoderm communities, whose debris formed the shoals, apparently
developed as a “carbonate factory,” which through storms and tides, “fed” the shoals. In fact, the
shoal complex developed astride the trade-wind belt and a major storm belt (e.g., Marsaglia and
Klein, 1983; Ettensohn, 2010) so that wave and current agitation generated by the pervading
trade winds, as well as by long-term swell events and storm surges, would have continually piled
up skeletal debris from the “carbonate factory” to grow beach and shoalface environments. Tides
and related currents would have only enhanced the distribution of skeletal debris and the growth
of these shoals (Hrabar et al., 1971). The similar roles of prevailing winds, storm surges and tides
in building modern, isolated shoal complexes in a geographic setting like that of the Tanglewood
were noted by Lloyd et al. (1987), Morgan (2008) and Wanless and Dravis (2008) in discussing
the origin of shoals on the Caicos Platform.

Development of the new shoal complex took place during five, widespread, fining-
upward cycles related to eustasy. Four lithofacies that represent respectively, shoal-top/foreshoal,
foreshoal/upper shoalface, lower shoalface, and offshoal/platform-lagoonal environments (Figs.
4.1 and 4.3), are represented in the shoal complex. Facies analysis of each cycle shows that the
thickest and coarsest part of each cycle corresponds to previously uplifted basement fault blocks,
which probably experienced uplift during deposition, although the possibility of a precursor
topography along which shoal accretion occurred cannot be excluded. The occurrence of thick,
coarse facies on the same fault blocks suggests that the blocks continued to experience uplift into
shallow waters, where fine elastic sediment could be winnowed away and the growth of
carbonate-producing organisms supported; at the same time, tides, waves and storms
redistributed skeletal sediments produced on deeper parts of the shoal complex. Eustasy
apparently controlled cyclicity and some of the larger regional flooding events. However, some
of the same events coincided with major Taconian tectonic events (Ettensohn and Lierman, 2015), and the fact that facies coincide with local structural elements (Fig. 4.3) suggests that Taconian far-field forces may have also influenced facies distribution on the shoal complex through synsedimentary basement growth-faulting. Either uplift on these structures halted or sea-level rise and clastic influx outpaced any uplift, but by about 500 ka into Edenian time, the upper Tanglewood, intra-platform shoal complex was completely inundated by deeper waters, reflected in the shales and fine-grained carbonates of the Clays Ferry, Kope or Point Pleasant formations. Despite final inundation by deeper waters, the persistence of a carbonate shoal complex for about 500 ka in the upper tongue of the Tanglewood Member shows that tectonic far-field forces can have significant influence on the development of carbonate depositional environments in distal, intracratonic settings.

Soft-sediment deformation occurs in six distinct and separate horizons of the upper Tanglewood and laterally equivalent parts of the Point Pleasant and Clays Ferry formations. Evaluation of how these six deformed horizons correlate from exposure to exposure and from unit to unit was based on the position of these horizons in the cyclic framework.

Although soft-sediment deformation of the type described from the upper Tanglewood Member of the Lexington Limestone in central Kentucky is not unique, the concurrence of four major lines of evidence, including deformation consistent with a seismogenic origin, widespread distribution in temporally and stratigraphically constrained horizons, a pattern of increasing frequency or deformation intensity toward a likely epicentral area, and the ability to exclude other likely causes, points very strongly toward a seismogenic origin. This interpretation is bolstered by the facts that the deformation crosses facies boundaries, contains random fold axes and no glide planes, shows crosscutting relationships with overlying and underlying undeformed
On the basis of inferred intensity of deformation, each deformed upper Tanglewood horizon and its equivalents can be shown to have had a probable epicentral area connected to a specific segment of structural lineaments known as unnamed lineament (A), the Georgetown-Gratz fault system (B), the Centerville fault system (C), the Lexington fault system (D), and the Vanceburg-Ironton fault system (E). Overall, the preponderance of evidence supports a seismogenic origin for the six, deformed horizons. Interpretations relative to earthquake clustering, site effects and magnitude may also apply, but are less certain.

Examination of the distribution map for each deformed horizon (Figs. 6.2—6.7) shows that in each horizon there is a larger, central area of more intense deformation near the probable originating fault segment, and ‘‘outliers’’ of lesser deformation. It is possible that all deformation in any of the six horizons could have been related to the movement on the same basement precursor fault in the epicentral area. It is also possible that the deformation in a particular region might have been related to the movement on the nearby basement fault and the deformation in another region might have been related to the movement on another fault at the same time or later. Because there is a growing realization that earthquakes on one structure can trigger other earthquakes on the same or nearby faults at nearly the same or later times through stress transfer (Reilinger et al., 2000; Kilb et al., 2000), the possibility of repeated fault movement, so as to produce differential deformation, cannot be ruled out. Because deformed horizons in the outliers were apparently nearly coeval with equivalent horizons in the central area, but occur near other faults with basement precursors, one of the convincing suggestions is that those faults might have been reactivated at the same or later times. Alternatively, varying site factors (e.g., Reiter, 1990), such as local basin shape, sediment thickness, sediment type, and
the nature of the involved structures at depth, could have worked together to produce differential deformation in different places in response to the same seismic event.

Because petrographic comparison between deformed and undeformed portions of the same horizon showed no difference in final cementation, it appears that cementation primarily took place after sediment deformation. Inasmuch as most of the relict framework of sediments appears to be intact in the outcrops, it could be possible that the sediments might have experienced weak marine cementation before deformation during early diagenesis. This possibility is bolstered by the fact that the rock contains intraclasts which represent early cementation. Therefore, it could be more realistic to assume that the initial marine cements might have been obliterated during deformation or later stages of diagenesis, and the main episode of cementation probably took place after sediment deformation when the sediment came in contact with ground water due to uplift into a non-marine setting.

Given that other deformed horizons in the Lexington Limestone have previously been interpreted as seismites (e.g., Pope et al., 1997; Jewell, 2001; Ettensohn et al., 2002d; Ettensohn and Stewart, 2003; Jewell and Ettensohn, 2004), it appears that these deformed horizons are related to movement on specific basement fault zones (e.g. Jewell, 2001; Ettensohn et al., 2002d; Jewell and Ettensohn, 2004). This indicates the important fact that supposedly stable continental interiors are underlain by previously faulted blocks, and that some of these faults get reactivated during the craton-margin orogenies. Therefore, it is unsurprising that the upper Tanglewood seismites, which also occur in vicinity of basement faults, seem to have been reactivated coevally with the Taconian Orogeny. Hence, this study bolsters the fact that far-field forces can exert an important influence on the sedimentary record, even in supposedly stable cratonic settings like the Lexington Platform, when stresses from the craton-margin are transmitted toward craton
interior. We hope that the present study will enhance understanding of seismite formation in intra-cratonic carbonate-platform environments, which, in the present context, has not been adequately addressed.

**Conclusions**

In the process of completing this study, several significant findings were made:

1.) The upper tongue of the Tanglewood Member of Lexington Limestone represents the continued deposition of a shoal complex following deposition of middle tongue of the Tanglewood Member. Development of the new shoal complex took place during five small-scale, sequence-like, fining-upward cycles related to eustasy.

2.) The upper Tanglewood Member represents the final phase of Lexington shoal-complex deposition before the area was completely inundated by deeper waters.

3.) Upper tongue deposition was initiated near the beginning of Edenian time at about 451 Ma after a regional flooding event, represented by a tongue of the Clays Ferry Formation or the Millersburg Member, briefly inundated the area.

4.) Four lithofacies that represent respectively, shoal-top/fore-shoal, fore-shoal/upper shoalface, lower shoalface, and offshoal/platform-lagoonal environments, are represented in the upper tongue shoal complex.

5.) Facies analysis of each cycle shows that the thickest and coarsest part of each cycle corresponds to previously uplifted basement fault blocks; the occurrence of thick, coarse facies on the same fault blocks suggests that blocks continued to experience uplift into shallow waters,
where tides, waves, and storms redistributed skeletal sediments produced on deeper parts of the shoal complex.

6.) Eustasy apparently controlled cyclicity and some of the larger regional flooding events. However, some of the same events coincided with major Taconian tectonics events (Ettensohn and Lierman, 2015), and the fact that facies coincide with local structural events (Fig. 4.3) suggests that Taconian far-field forces may have also influenced facies distribution on the shoal complex through synsedimentary basement growth-faulting.

7.) The upper Tanglewood Member includes six deformed horizons, which can be traced into the partially equivalent Clays Ferry and Point Pleasant formations; the deformed horizons are characterized by ball-and-pillow structures, mud or sand volcanoes, simple contortion and folding of beds, and fluidized beds, among others.

8.) Concurrence of four lines of evidence, suggested by Ettensohn et al. (2002d), Ettensohn and Stewart (2003), and Jewell and Ettensohn (2004) to interpret seismites, includes deformation that is consistent with a seismogenic origin, widespread occurrence of deformation in horizons that are temporally and stratigraphically constrained, deformation that shows systematic increase in frequency or intensity toward a possible epicentral area, and the ability to exclude other possible causes, and strongly suggests a seismogenic origin for upper Tanglewood deformation.

9.) Reactivation of basement structures due to Taconian orogeny going on the east might have induced seismicity on the intra-platform carbonate basin so as to produce soft-sediment deformation.
10.) Petrographic investigation indicates that the upper Tanglewood limestone consists mainly of skeletal grainstone; the major components are skeletal and non-skeletal grains, abundant sparry calcite cement and some quartz grains.

11.) The diagenetic processes which have modified the sequences of the upper Tanglewood Member of Lexington Limestone consist of micritization, cementation, compaction, dissolution, dolomitization and internal filling of fractures.

12.) Most of the cements in upper Tanglewood limestone appear to be late diagenetic, fresh-water phreatic cements that reflect the infilling of intra- and inter-granular porosity.

13.) Comparing the petrography between deformed and undeformed portions of the same horizon showed no significant differences in terms of cementation, indicating that cementation occurred primarily after deformation. However, preservation of relict structures both in outcrop and thin section showed that the unit had probably been weakly cemented before deformation.

14.) The primary impact of deformation on the microstructure of the upper Tanglewood grainstones was the randomization of grain fabric, the destruction of early cementation, and the increased presence of broken intraclasts.
Appendices

Milepost 128, I-75 (38°15'33.21"N, 84°33'3.75"W)
Woolcott across road

Woolcott across road (38°44'5.22"N, 84° 6'3.77"W)

0 ft
  Mud Silt Sand Rudite

5 ft

10 ft

15 ft

20 ft

25 ft

DH (8)

3

2
Millersburg (38°16'20.24"N, 84°10'22.62"W)
Clays Ferry (37°52'51.50"N, 84°20'26.00"W)

Typical Clays Ferry Formation

Upper Tanglewood Equivalent

DH (1)

DH (5)

DH (4)

DH (4)

Mud Silt Sand Rudite
Sadieville (38°23'7.02"N, 84°33'35.70"W)

Clays Ferry Formation

DH (6)

5
4
3
2
1
Lawrenceburg 1 (38° 3' 30.90" N, 84° 55' 15.30" W)

Clays Ferry Formation

DH (6)

DH (9)

DH (8)

DH (8)

DH (8)

DH (10)

4

3

2

1

Mud Silt Sand Rudite
Lawrenceburg 2 (38° 3'15.16"N, 84°55'9.02"W)

Clays Ferry Formation

Mud Silt Sand Rudite

DH (6)

Bioturbation

Fluidization DH (8)

DH (8)

1

2

3

4

5
Frankfort 4 (38° 9'55.17"N, 84°50'29.91"W)

- 25 ft
- 20 ft
- 15 ft
- 10 ft
- 5 ft

- 0 ft

Mud  Silt  Sand  Rudite

Nodular (Millersburg Member)
Frankfort 9 (38° 9'43.70"N, 84°57'26.00"W)

Clays Ferry Formation

Millersburg Member

DH (1)

Mud Silt Sand Rudite

1 2 3 4 5
Frankfort 14 (38°18'29.40"N, 84°50'44.20"W)

Clays Ferry Formation

DH (6)

DH (5)

DH (1)

Millersburg Tongue

Millersburg Member

Mud, Silt, Sand, Rudite
Chilo (38°48'3.82"N, 84° 9'33.03"W)

Kope Formation

Mud  Silt  Sand  Rudite

0 ft  5 ft  10 ft  15 ft  20 ft
Nicholas County 8 (38°21'47.15"N, 84° 2'27.16"W)
Clays Ferry Formation

Mud  Silt  Sand  Rudite

Millersburg Member

Gratz Core (38°29'51.00"N, 84°58'37.00"W)
REFERENCES


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Elrick, M., Reardon, D., Labor, W., Martin, J., Desrochers, J., Pope, M., 2013. Orbital-scale climate change and glacioeustasy during the early Late Ordovician (pre-Hirnantian) determined from δ^{18}O values in marine apatite. Geology 41, 775–778.


Ettensohn, F.R., 2010. Origin of Late Ordovician (mid-Mohawkian) temperate-water conditions
on southeastern Laurentia: Glacial or tectonic? In: Finney, S., Berry, W. B. N. (Eds.), The
163–175.

Ettensohn, F.R., Kulp, M.A., 1995. Structural-tectonic control on Middle-Late Ordovician
deposition of the Lexington Limestone, Central Kentucky. In: Cooper, J.D., Droser,
M.L., Finney, S.C. (Eds.), Ordovician Odyssey: Short Papers for the Seventh
International Symposium on the Ordovician System. Pacific Section, Society for

Ettensohn, F.R., Lierman, R.T., 2012. Chapter 4, Large-scale tectonic controls on the origin of
Paleozoic, dark-shale, source-rock basins: examples from the Appalachian foreland-basin
region, eastern United States. In: Gao, D. (Ed.), Tectonics and Sedimentation:
Implications for Petroleum Systems. American Association of Petroleum Geologists
Memoir vol. 100, pp. 95 -124.

Ettensohn, F.R., Lierman, R.T., 2015. Using black shales to constrain possible tectonic and
structural influence on foreland-basin evolution and cratonic yoking: Late Taconian
Orogeny, Late Ordovician Appalachian Basin, eastern USA. In: Gibson, G.M., Roure, F.,
Manatschal, G. (Eds.), Sedimentary Basins and Crustal Processes at Continental Margins:
from Modern Hyper-Extended Margins to Deformed Ancient Analogues. Geological

Ettensohn, F. R., Stewart, A.K., 2002. Middle and Late Ordovician seismites from central
Kentucky. In: Ettensohn, F. R., Smath, M. L. (Eds.), Guidebook for geology fieldtrips in
Kentucky and adjacent areas, 2002 joint meeting of the North–Central Section and
Southeastern Section of the Geological Society of America, Lexington, Kentucky,
University of Kentucky, Lexington, Kentucky, pp. 129–150.

Ettensohn, F. R., Stewart, A. K., 2003. Middle and Late Ordovician seismites from central
Kentucky. In: Ettensohn, F. R., Smath, M. L. (Eds.), Guidebook for geology field trips in
Kentucky and adjacent areas, 2002 Joint Meeting of the North-Central Section and
Southeastern Section of the Geological Society of America. Kentucky Geological Survey
Guidebook 2, Series XII, pp. 130–151.

Ettensohn, F. R., Amig, B. C., Pashin, J. C., Greb, S. F., Harris, M. Q., Black, J. C., Cantrell, D.
and paleoenvironments of the bryozoan-rich Sulphar Well Member, Lexington Limestone

paleoearthquakes, Middle Ordovician Lexington Limestone, central Kentucky.


Obermeier, S. F., 1996. Use of liquefaction-induced features for paleoseismic analysis—An overview of how seismic liquefaction features can be distinguished from other features and how their regional distribution and properties of source sediment can be used to infer the location and strength of Holocene paleo-earthquakes. Engineering Geology 44, 1–76.


Pope, M. C. and Read, J. F., 1997b. High-resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian–Cincinnatian) foreland basin rocks, Kentucky and Virginia. American Association of Petroleum Geologists Bulletin, 81, 1866–1893.


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Integration of sedimentology, tectonics, stratigraphy and paleontology for paleoenvironment interpretation, soft-sediment deformation in sedimentary rocks, black shale, carbonate rocks, sequence stratigraphy, diagenesis in sedimentary rocks, petroleum geology

Research Experiences

Department of Earth and Environmental Sciences, University of Kentucky
• Soft sediment deformation, cyclostratigraphy, analysis of depositional environment of upper Ordovician carbonate shoal, petrography of carbonate rocks

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Petroleum Geology, Exploration Geophysics, Geochemistry, Sedimentary Geology, Structural Geology, Igneous and Metamorphic Petrology, Geomorphology, Physical Geology, Mineralogy, Hydrogeology, Geological Mapping, Engineering Geology, Stratigraphy, Paleontology, Historical Geology

Teaching Experience
Teaching assistant (University of Kentucky) 2012-Present
- Structural Geology and GIS (Spring, 2017)- lab and field teaching
- General University Physics (Fall, 2016)- lab teaching
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- Dinosaurs and Disasters (Fall, 2015)- grader
- Sedimentary Geology (Fall, 2014 and Fall, 2015)- lab teaching of thin section analysis of sedimentary rocks and occasional lecturing
- Mineralogy (Spring, 2015)- lab teaching of minerals in thin sections and hand specimens
- Structural Geology (Spring, 2014)- grader and occasional lecturer
- Fundamentals of Geology (Spring 2014)- instructor in recitation and lab
- Geological Mapping (six-week field work in Colorado, Summer, 2013)- instructor as a TA
- Introduction to Oceanography (Spring, 2013)- instructor in recitation
- Petroleum Geology (Spring, 2013)- occasionally lecturing a class in the absence of professor
- Earthquakes and Volcanoes (Spring, 2012)- instructor in recitation

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Publications
Eustatic and far-field tectonic control on the development of an intra-platform carbonate-shoal complex: upper tongue of the Tanglewood Member, Upper Ordovician Lexington Limestone, central Kentucky, U.S.A.
Dibya Raj Koirala, Frank R. Ettensohn, Marta L. Clepper

Tectono-metamorphic Evolution of the far-Eastern Nepal Himalaya
S.M. Rai, Sakai, D.R. Koirala · [...] · S. Ghimire

Seismogenic soft-sediment deformation (seismite) in the upper Tanglewood Member, Upper Ordovician Lexington Limestone, central Kentucky, U.S.A.
Dibya Raj Koirala, Frank R. Ettensohn

Conference Presentations
Sequence and event stratigraphy on a late Ordovician carbonate buildup: parasequences and seismites in the upper part of the Lexington Limestone (Edenian), central Kentucky, U.S.A.
4 November 2015, GSA Annual Meeting, Baltimore, MD

Eustatic and tectonic control on the distribution of carbonate shoal: Example from the upper tongue of the Tanglewood Member, upper Ordovician Lexington Limestone, Bluegrass region, central Kentucky, U.S.A.
21 September 2015, AAPG Eastern Section, Indianapolis, IN

Depositional environments in the upper tongue of the Tanglewood Member, upper Ordovician Lexington Limestone, Bluegrass region, central Kentucky, U.S.A.
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2014- : GSA Grant for Dissertation Work
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