Seismic Evaluation of Differential Tectonic Subsidence, Compaction, and Loading in an Interior Basin

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ABSTRACT

Quantitative analysis of depth-converted reflection times defines long-term differential motion across individual structures in a central Appalachian interior basin known as the Rome trough. Differential motion decreases exponentially with time. Rotation about a hinge defining the trough's west margin reached approximately 37% of total displacement in about 63–78 million years (m.y.). Displacement across the trough's faulted east margin occurred more rapidly and reached 37% of the total in 13–51 m.y. A major fault in the interior of the trough developed rapidly with 37% of total displacement reached in from 16 to 23 m.y. Longer term rotation across the west margin may be due to its participation in the overall subsidence of the craton during the Paleozoic. The time spanned by the formation of the East-Margin and Interior faults was restricted to the Cambrian in the northern part of the area, but to the south, movement along the East-Margin fault continued through the Middle Ordovician.

The general effects of differential compaction and loading for a single lithology model are computed from the standard compaction and Airy isostasy equations. Depth-dependent compaction requires that thicker strata over a hanging wall or subsiding fault block undergo greater compaction than occurs in thinner strata over the footwall or structurally high areas. Seismic interpretation of trough structures does not reveal the presence of compaction faults or of long-term differential compaction. The observations suggest that in this interior basin differential compaction of major stratigraphic intervals was nearly complete prior to deposition of later sequences. Local isostatic compensation of differential loads across faults or fault blocks requires movement along near-vertical crust-penetrating faults and abrupt thinning by differential amounts across the base of the crust. This possibility seems unlikely within the context of current models of intracratonic extension. Analysis indicates that differential thickening of strata across trough structures portrayed in reactivation history diagrams defines long-term tectonic movement rather than a mixture of tectonic, compaction, and load-related motion. The analysis also suggests that estimates of the time at which a given horizon entered the oil or gas window and estimates of the total depth reached by a horizon during subsidence may also be in error if simple depth-dependent compaction corrections are used.

INTRODUCTION

Modeling the tectonic development of sedimentary basins has become an integral part of exploration activities in the past two decades. Information about the timing of deformation relative to the timing of hydrocarbon migration is critical to determine if hydrocarbons will be trapped within a given structure. Deformation history of individual extensional structures within a basin is the net result of tectonic and thermomechanical processes responsible for basin formation, combined with lithosphere loading, sediment compaction, and variations of these factors throughout the basin (Ungerer et al., 1984).

Models of basin development have become increasingly complex. Sleep (1971) noted that an exponential thermal-cooling model accurately portrays subsidence histories of the Atlantic continental margin and several basins within the interior of the North American craton. The model derived by Sleep (1969) was initially used to explain mid-ocean ridge topography as the result of thermal cooling of isostatically compensated hot materials intruded along the ridge axis. McKenzie (1978) noted that simple thermal models require significant surface

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erosion of thermally uplifted regions to account for the elevated Moho observed beneath them. Using the North Sea as an example, McKenzie (1978) noted that significant erosion has not been observed and suggests that stretching may have initially thinned the crust. Sclater and Christie (1980) presented a detailed analysis in support of initial continental stretching followed by thermal subsidence. Long-term activity across fault zones and differentially subsiding fault blocks or margins is observed in seismic data. Yet, in the absence of detailed well-bore data required to specify porosity, density, lithology, and water depth across such structures, apparent and continued displacement resulting from compaction and loading (respectively) cannot be estimated accurately and separated from tectonic subsidence. In this paper, differential motion across structures of the central Appalachian Rome trough (Figures 1, 2) is measured from seismic profiles. An extensive normal fault (Figure 2) known as the East-Margin fault (Shumaker, 1993) defines the east margin of the trough along its length. With local exception, subsidence across the western margin of the trough occurs through rotation about a hinge zone (Figure 2) known as the Wilson

MAJOR BASEMENT STRUCTURE BENEATH THE STUDY AREA

Extension during the Early Cambrian deformed the Precambrian foreland basement in the central Appalachian study area. The extensional basement complex formed by this event (Figure 2) is known as the Rome trough (McGuire and Howell, 1963) or the Eastern Interior aulacogen (Harris, 1978). The Rome trough is part of a larger system of grabens (Shumaker, 1986a, b; Thomas, 1991, 1993) that formed during the opening of the Early Paleozoic Iapetus ocean. Several wells drilled in the area along with limited seismic data confirm the presence of the trough and a thick sequence of synrift rocks (e.g., Shumaker, 1986a, b; Allen, 1988; Read, 1989; Ryder et al., 1992; Ryder, 1992; Shumaker and Wilson, 1996).

Interpretations of four seismic lines (Figure 1) define major basement structure beneath the study area (Figure 2). An extensive normal fault (Figure 2) known as the East-Margin fault (Shumaker, 1993) defines the east margin of the trough along its length. With local exception, subsidence across the western margin of the trough occurs through rotation about a hinge zone (Figure 2) known as the
Ohio-West Virginia hinge zone (Ryder et al., 1992). In the southern part of the area along line 1 (Figures 2, 3) the trough forms a simple asymmetrical graben; however, to the north the trough widens and is subdivided by one or more interior faults. Minor internal faults are ubiquitous within the trough interior and are not represented at this scale.

Sonic logs from deep wells in the area (see Shumaker and Wilson, 1996) were used for velocity control. A constant velocity function derived from the sonic logs was used to convert reflection traveltime to depth. There are a limited number of sonic logs in the region, but comparison between them of the average velocities for major stratigraphic intervals does not reveal significant velocity variation. Line 1 (Figure 3) is a reprocessed single-fold seismic line, whereas lines 2, 3, and 4 are multichannel (optimal 24-fold) Vibroseis™ data. The quality of the data shown in line 1 is typical of the quality of data available for this study. Interpretations are based on the most coherent seismic reflections observed in the data. Although different interpretations of these lines are possible, the major features of the interpretations are not expected to vary significantly. Identification of reflections on line 1 is based on synthetic seismic correlation (Shumaker and Wilson, 1996). The strata associated with the reflections shown on line 1 range in age from Early Cambrian to middle Mississippian (Figure 4). Depth conversion of major reflections across line 1 (Figure 5) reveals that the trough, at regional scale, is a simple asymmetrical graben. The western limb gradually subsided relative to the Ohio-West Virginia hinge zone that lies off the line to the northwest (Figure 2). The East-Margin fault forms the eastern boundary of the trough, although minor structures are observed farther east in the footwall on line 1 (Figure 3). Minor structural irregularity is observed within the trough (Figure 3). A low-relief structural high is present across the trough in the Silurian Rose Hill through Devonian Huron reflections (Figures 3, 5). Depth-converted seismic traveltimes reveal a structural rise of approximately 100 m in the Devonian Onondaga reflector west of the East-Margin fault into the interior of the trough. Structural relief observed on contour maps of well log data presented by Gao and Shumaker (1996) of the Onondaga Limestone in this area also reveal approximately 100 m of structural rise into the trough. This general agreement between seismic and well log derived structure further supports the absence of significant velocity anomalies. Uplift of the trough interior is interpreted to have produced the structural high. This uplift may have occurred in the Late Devonian since the interval between the Devonian Onondaga and Mississippian Greenbrier reflections (Figures 3, 5) is thinner across the trough than it is to the east of the trough margin; however, because the area lies along the distal edge of detached structures produced during the Alleghany orogeny, that disharmony could be

Figure 2—Contour map of acoustic basement across the Rome trough derived from lines 1 through 4.
produced by detachment and thickening of the section just east of the trough margin (Shumaker and Wilson, 1996).

Line 2 (Figures 2, 6A) extends across an unfaulted portion of the western trough margin. West margin collapse is largely rotational about the well-defined Ohio-West Virginia hinge zone. Approximately 1.2 km of offset is observed across the East-Margin fault. Interior faults form a double-step across which the basement drops about 0.9 km to the northwest. Lower and Middle Cambrian sections thicken significantly eastward toward the Interior and East-Margin faults. The Interior fault observed on line 2 is interpreted to extend northward across line 3 (Figures 2, 6B). On line 3, the east margin of the trough is cut by two normal faults, which have a combined displacement of nearly 2 km. Faulting along the Interior and East-Margin faults is accompanied by fault-block rotation. Upper level structures on the east end of line 3 are detached above a decollement in the Ordovician Martinsburg Shale. These detached structures formed during the Alleghany orogeny. Along line 4 (Figures 2, 6C), the western margin of the trough also lacks major basement faults. Normal faults similar to those that subdivide the interior of the trough to the south are more numerous along line 4. Maximum depths of around 6 km (subsea) observed on line 1 to the south increase northward to more than 8 km along line 4.

The basement interpretation presented here (Figure 2) is intended as a generalized view of major trough structures in the study area and differs from previous interpretations of the trough (e.g., Kulander and Dean, 1993). For example, the interpretation of Kulander and Dean (1993) suggested that offset along the East-Margin fault decreases steadily from about 1.5 km to less than 0.5 km southwest to northeast across the area; however, lines 1–4 reveal that displacement along the East-Margin fault increases from approximately 1 km along line 1 to 1.7 km along line 3 (Figure 6). Farther to the north along line 4, the displacement decreases to approximately 1 km. Harris (1978), Donaldson and Shumaker (1981), and Kulander and Dean (1993) described the west margin of the trough...
trough as a continuous normal fault through the area. Lines 1–4, however, do not reveal significant faults on the west flank of the trough. Line 2 reveals Ryder’s Ohio-West Virginia hinge zone (Figure 2) across which the west margin rotates down into the trough. Line 3 extends high up onto the west margin of the trough but does not cross the hinge. Normal faults marking the western edge of the trough may be concentrated along the Ohio-West Virginia hinge zone, which is observed only on line 2 (Figure 6A). Tegland (1977), Swimm (1986), and R. W. Beardsley (unpublished seismic) showed local faults along the west margin; however, the interpretations presented here (Figure 6) do not support the presence of a continuous west margin fault. West margin relief appears to be accommodated largely through rotation about the Ohio-West Virginia hinge zone.

Figure 4—Depth, traveltime, and interval velocity are presented for the Paleozoic section of the Appalachian foreland. Major reflections observed in the seismic data of the foreland are noted with an asterisk (taken from Shumaker and Wilson, 1996).
SEISMIC ANALYSIS

Quantitative analysis of individual structures was done using reactivation history diagrams. A schematic representation (Figure 7) of the major structural features of the trough (Figure 5) is used to illustrate construction of the reactivation history diagram. Deposition of sediments across an active normal fault (e.g., southeastern end of Figure 7) produces thicker sediment accumulations over the hanging wall and thinner accumulations over the footwall. In the absence of strike-slip movement, increased thickness within any given interval across the fault will be related to vertical displacement across the fault during deposition of that interval. In Figure 7, for example between points C and D, the thickness of the basal unit 1 thins from 880 m over the hanging wall to 250 m over the footwall. The thickness difference is 630 m. An identical relationship exists for strata deposited across the area deformed by rotation on the northwest flank of the graben (Figure 7). Unit 1 (Figure 7) thickens in the downdip direction between points A and B. The thickness difference across the northwest flank is also 630 m.

The difference in interval thickness across a structure provides a measure of the structural displacement that occurred during the span of geologic time associated with deposition of that interval. Syndepositional subsidence and continued movement of trough structures actively influenced sedimentation (e.g., Figures 5, 6). Thickness differences that occur over several contiguous time intervals provide an historical record of structural development. Reactivation history diagrams (Wilson et al., 1994; Dominic and Wilson, 1995; Shumaker and Wilson, 1996) constructed from thickness differences portray historical activity across individual structures.

Thickness differences (d) measured from Figure 7 are listed in Table 1. The thickness of unit 1, for example, decreases 630 m (250 to 880 m) from the interior of the trough onto its flanks. Negative thickness differences imply subsidence of the trough relative to its margins. Positive thickness differences (e.g., those for units 4, 7, and 8 in Figure 7 and Table 1) indicate that the trough interior was uplifted during deposition of those units. The cumulative increases in thickness reach the present-day total of –985 m (Table 1), which also coincides with the present-day basement relief across both margins of the trough (Figure 7). To allow for exponential fitting (see discussion in following section), the cumulative difference (Table 1) is expressed relative to a maximum vertical displacement (d_o) of the trough interior relative to its margins. In this example, d_o is set equal to 1400 m (Table 1) but will vary from line to line. The cumulative differences are subtracted from the maximum value to obtain the adjusted cumulative differences (Table 1). The maximum value (d_o) is chosen to prevent values from becoming negative or zero and can be thought of as the maximum possible vertical displacement across a given structure.

The adjusted cumulative difference (Table 1) is graphed as a function of geologic time (Figure 8). This plot is referred to as a reactivation history diagram (RHD) to distinguish it from a subsidence history curve. RHDs are distinguished from subsidence curves in that they portray differential movement across individual structures rather than total subsidence at a given point. Differential movements represented by reactivation history diagrams are not corrected for compaction and loading effects. Compensation for compaction and loading are discussed following presentation of RHDs for the major structures of the trough. The issue of correction is
critical to interpretation of the RHD because non-tectonic displacements are expected throughout geologic time as a result of differential compaction and loading of different thickness sediment deposited across a fault or tilted fault block.

DIFFERENTIAL MOVEMENT INFERRED FROM RHD

RHDs for the west and east margins of the trough (Figures 9, 10, respectively) and for the Interior fault (Figure 11) portray differential motion that has occurred across the major structural elements defining the trough. Fowler (1990), Lerche (1990), and McKenzie (1978) model subsidence \( (d) \) as a thermal process that, in first-order approximation, varies exponentially as

\[
d = d_0 e^{(t/t_0)} - 1
\]  

In equation 1, \( d_0 \) and \( t_0 \) are constants. Although \( d_0 \) and \( t_0 \) are constants, their values may change from...
one data set to another within a basin. The constant $t_0$ represents the thermal time constant of the lithosphere.

$$t_0 = \frac{L^2}{\pi^2 \kappa}$$  \hspace{1cm} (2)

(Fowler, 1990) where $L$ is the thickness of the lithosphere and $\kappa$ is the thermal diffusivity. Although the mechanism or mechanisms responsible for trough formation are not known with certainty, differential motion across trough structures shown in the RHDs follows an exponential decay process and is modeled by equation 1. In the current application, equation 1 is solved in the form

$$d_0 + d = d_0 e^{(-t/t_0)}$$  \hspace{1cm} (3)
In this expression, $d_0$ is approximately equal to the maximum possible offset across the fault or rotating margin along this line. As noted, the adjusted cumulative differences (Table 1) are obtained by subtracting the cumulative difference from $d_0$ at each stage. In this example (Table 1), $d_0$ is 1400 m. The variable $d$ (Table 1) represents the change in thickness of each interval into the trough. Smaller values of $t_0$ are associated with more rapid differential subsidence than are larger values of $t_0$. Time constants derived from exponential fitting of the adjusted cumulative differences are negative, and the negative sign is shown explicitly in equation 3, which defines an exponential decay process. Time constants ($t_0$) were computed by fitting exponential functions only to the subsiding portions of the RHDs (Figures 10–13). At time $t = t_0$, the exponent in equation 3 equals -1 and the term on the right becomes $d_0/e$ or approximately 0.37$d_0$. Thus, in this application, $t_0$ represents the time it takes for 37% of the maximum possible displacement ($d_0$) across a structure to occur.

The time scale of the observations (Figure 8) extends from the early Cambrian–middle Mississippian or approximately 570–340 Ma. Growth of trough structures is assumed to have begun in the Early Cambrian. Time zero is taken as 570 Ma. The time constant ($t_0$) was calculated by fitting exponential functions to the seismic profiles across the trough’s unfa ulted west margin from 62.5 to 77.5 m.y. (Figure 9) and implies that approximately 37% of the total displacement across the margin occurred over a time interval of 62.5–77.5 m.y. These large time constants reflect long-term rotational collapse of the trough’s west flank throughout much of the Paleozoic. Time constants along the east margin range from 13 to 51 m.y. (Figure 10). The exponential curve fit to line 1 was derived from a 230 m.y. period extending from the Early Cambrian (570 Ma) into the Mississippian (340 Ma). On line 1 the motions of the east and west trough margins closely parallel each other. Slight inversion or tectonic uplift of the trough interior begins to occur along the east margin after 120 m.y. (or 450 Ma in the Middle–Late Ordovician). Inverse movement along the East-Margin fault on lines 2 and 4 is more pronounced than on

<table>
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<th>Thickness Difference A to B $d$ (m)</th>
<th>Thickness Difference D to C $d$ (m)</th>
<th>Cumulative Difference $d - d_0$ (m)</th>
<th>Adjusted Cumulative Difference $d_0 = 1400$ m (m)</th>
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<tr>
<td>Unit 1 -630</td>
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<td>-630</td>
<td>770</td>
</tr>
<tr>
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<td>Unit 7 75</td>
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<tr>
<td>Unit 8 160</td>
<td>160</td>
<td>-985</td>
<td>415</td>
</tr>
<tr>
<td>Total -985</td>
<td>-985</td>
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</tr>
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*See Figure 7. $d = thickness difference. Adjusted cumulative differences are used to construct the reactivation history diagram.
line 1, whereas on line 3, interior uplift is restricted to the Early and Middle Devonian from about 375 to 350 Ma. The time constants computed for lines 2–4 range from 33 to 13 m.y. and are much less than that computed for line 1. The coupled response of the east and west margins along line 1 sets that area apart from the area to the north where differential motion across the east and west margins differ considerably through time.

Time constants of 16–23 m.y. were calculated for offsets occurring across the interior faults on lines 2–4 (Figures 2, 11). Time constants were computed for the initial 70 m.y. (Cambrian) during which normal offsets occurred. Uplift of the trough interior (inversion) began early in the Ordovician on line 2. On line 3 to the north, a slight amount of interior uplift began during the Early Silurian. On line 4 interior uplift occurred during the Middle–Late Cambrian. In general, normal offsets on the Interior fault appear to have been confined to the Cambrian period. Significant interior faulting is not observed on line 1 (see Figures 2, 3, 5).

Lack of longer term normal displacements on the East-Margin and Interior faults suggests they were decoupled from continued rotation of the west margin following the Cambrian. These observations suggest that rotational subsidence of the trough’s west margin continued in response to Iapetan shelf loading, whereas the East-Margin and Interior faults remained locked or served as zones of weakness along which slight reverse movements occurred.

Continued movement on the East-Margin fault of line 1 after the Cambrian is anomalous by comparison to that observed on lines 2–4. Gao (1994) reported similar reactivation along the east margin of the
trough just south of line 1. The analysis presented here in combination with Gao's observations suggests that normal offsets along the East-Margin fault north of line 1 ceased following the Cambrian but continued through the Silurian in the area south of line 2. These two areas may have experienced slightly different stress histories. Differences between these two areas are also observed in the form of significant differences in coal cleat trends between these two areas (Kulander and Dean, 1993). Crustal scale gravity models derived along lines 1-4 (Gurshaw, 1995; Morgan, 1996; T. H. Wilson, unpublished results) suggest that the crust thickens more than 10 km from north to south between these two areas.

The variations in $t_0$ provide a quantitative measure of differences in the rate of structural development within this area of the Rome trough. Because the analysis is made of differential motion across structural elements in the basin, it is not directly comparable to subsidence; however, the RHDs allow one to evaluate the degree to which individual structural elements participated in the overall subsidence of the basin. Previous studies of subsidence in the central Appalachians are limited. Ettensohn et al. (1992a) discussed the sedimentary-stratigraphic response and subsidence history of the Appalachian basin in Kentucky. Their study area included the western limits of the Rome trough. Ettensohn (1992b) interpreted the sedimentation history of the region within the context of lithosphere response to orogenic loading and subsequent relaxation. Sedimentation in the Kentucky foreland is interpreted by Ettensohn (1992b) to

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Figure 10—Reactivation history diagrams for the East-Margin fault of the Rome trough.

**Line 1**

$\begin{align*}
\text{Adjusted Cumulative Difference (m)} \\
\text{Age (Ma)}
\end{align*}$

$t_0=51.1\text{My} \\
r = 0.995$

**Line 2**

$t_0=23.1\text{My} \\
r = 0.998$

**Line 3**

$t_0=32.9\text{My} \\
r = 0.986$

**Line 4**

$t_0=12.8\text{My} \\
r = 0.995$
reveal a four-stage response to individual orogenic events, consisting of the initial cratonward migration of a peripheral bulge (Quinlan and Beaumont, 1984), foreland subsidence, relaxation and peripheral bulge reflection, and unloading. Goodman (1992) computed tectonic subsidence from two wells located in the Kentucky portion of the Rome trough. Computed tectonic subsidence reveals four distinct regional subsidence events associated with (1) initial Iapetan rifting, (2) Middle Ordovician Blountian and Middle–Late Ordovician Taconic orogenies and Silurian Salinic disturbance, (3) Devonian Acadian orogeny, and (4) Alleghanian orogeny.

Donaldson (1994) made subsidence computations within the Rome trough near its east margin, just north of line 1 (Figure 2). Donaldson used general lithological information from an Exxon deep well to evaluate subsidence along a seismic profile through the Granny Creek oil field in West Virginia; however, detailed well log data in the Rome trough area is limited, so Donaldson (1994) used North Sea compaction factors (Allen, 1990; Sclater and Christie, 1980) to decompact the section. Donaldson (1994) incorporated sea level variations from Hallam (1992) and paleobathymetry based on general facies interrelationships to sequentially remove or backstrip (Bond and Kominz, 1984) stratigraphic intervals represented in the seismic profile. Donaldson’s results revealed roughly uniform subsidence throughout the Paleozoic interrupted by a brief (∼5 m.y.) period of uplift within the trough during the Late Cambrian and equally brief period of increased subsidence during the Late Ordovician Taconic orogeny.

Reactivation history analysis presented previously (Figures 9–11) reveals additional information about movement along individual trough structures and accompanying regional subsidence; however, with the exception of subsidence directly associat-
ed with the trough-forming rifting, differential movement of trough structures does not appear to accompany individual Paleozoic orogenies. West margin subsidence continues into the Mississippian. The development of the East-Margin fault is quite varied. Both the East-Margin and Interior faults accommodated minor uplift of the trough interior following the Cambrian.

COMPACCIÓN

The RHDs presented are not corrected for differential sediment compaction or loading amplification. The potential influences of these processes on the preceding results are examined in this section using generalized compaction models. The effect of differential sediment compaction is modeled stage-by-stage through time (Figure 12) using a single lithology. This model study is not intended to replicate the details of movement observed across a specific structure within the trough, but only to illustrate the general effect of depth-dependent compaction assumed in several compaction models (e.g., Baldwin and Butler, 1985). Model development employs the compaction model presented by Sclater and Christie (1980) in which porosity (φ) decreases with depth following the relationship

\[ \phi = \phi_0 e^{-cz} \]  

(4)

Here, \( \phi_0 \) is the initial porosity after deposition, \( \phi \) is the present-day porosity at depth \( z \), and \( c \) is a lithology-dependent compaction constant. During compaction, it is also assumed that the rock volume \( V_w + V_s + V_m \) equals \( V_w + V_s + V_m \), where \( V \) refers to volume and the subscripts \( w \) and \( s \) to water and matrix, respectively. Use of this relationship assumes that total grain volume remains unchanged and thus that significant diagenesis has not occurred (Angevine et al., 1990). Use of equation 4 allows one to express the uncompacted thickness of a given interval in terms of its compacted thickness.

An interval encountered in a well or seismic section between depths \( z_1 \) and \( z_2 \) will have original thickness \( z' \) (or \( z_2 - z_1 \), where \( z_1 = 0 \) at the surface) defined as

\[ z' = z_2 - z_1 = \frac{\phi_0}{c} (e^{-cz_1} - e^{-cz_2}) + \frac{\phi_0}{c} (1 - e^{-cz}) \]  

(5)

\( z' \) must be determined numerically. Given interval thickness \( z' \) immediately following deposition, it is possible to compute the mean sediment density of the interval and thus the loading effect of the original water-saturated sediment column. The process of unloading layers in this fashion is referred to as backstripping (e.g., Bond and Kominz, 1984). It is a simple matter to carry the computations in the forward direction and compute interval thickness \( (z_2 - z_1) \) after burial. To construct the forward model, take an interval of thickness \( z' \) at the surface and move its base to depth \( z_2 \). In this process, \( z_1 \) is calculated rather than \( z' \), and rearrangement of equation 5 yields

\[ z_1 + \frac{\phi_0}{c} e^{-cz_1} = \frac{\phi_0}{c} + z_2 + \frac{\phi_0}{c} e^{-cz_2} - z' + \frac{\phi_0}{c} e^{-cz} \]  

(6)

The terms on the right are all known. Substitution yields the constant

\[ a = \frac{\phi_0}{c} + z_2 + \frac{\phi_0}{c} e^{-cz_2} - z' + \frac{\phi_0}{c} e^{-cz} \]  

(7)

then substitution of equation 7 into equation 6 yields

\[ z_1 = a - \frac{\phi_0}{c} e^{-cz_1} \]  

(8)

and \( z_1 \) can be solved iteratively.

Equation 8 was used to generate a forward-compaction model (model I) across a fault that was active only during deposition of the basal sedimentary layer. Values for \( c \) and \( \phi_0 \) taken from Sclater and Christie (1980) for North Sea shale (5.1 ×10⁻⁹/cm and 0.63, respectively) were used. The initial stages of development in model I are shown in Figure 12. Two kilometers of throw is introduced across a basement fault during stage 1 deposition (e.g., deposition of the basal sedimentary layer). No additional fault offsets were introduced. During stage 2 and all subsequent stages, the hanging wall and footwall were dropped uniformly in 0.5 km steps per stage. Differential compaction of sediments across the underlying fault produces normal offset of the interface separating stage 1 and stage 2 sediments (Figure 12). The process is repeated, and during growth occurs across the fault as a result of differential compaction of sediments on either side of the fault. A total of nine stages was computed leading to the final stage shown in Figure 13A.

As the sediments of model I continued to be deposited, differential compaction led to increased offset of individual horizons crossing the fault. Interval thickness across the fault decreases as burial depth increases. The reactivation history diagram for the basement fault in model I (Figure 13B) suggests that the initial vertical displacement across the fault was only about 1.6 km. The fault appears to have been moved continually during the following stages (stages 2–9) until the maximum displacement of
2 km was finally reached. If the process of compaction were to continue to operate over long time periods as predicted by continuous porosity-depth relationships such as that defined by equation 4, we would expect to see the general effect of differential compaction represented in Figure 13A. If compaction actually follows equation 4, then the RHD (Figure 13B), which is uncorrected for compaction, misrepresents the timing of basement fault development because the entire 2 km of fault displacement occurred entirely during stage 1.

Although considerable compaction of the sediments in the trough must have occurred, syndepositional compaction of earlier unconformity bounded sequences does not appear to cause the systematic decreases in normal displacements of younger strata across the East-Margin or Interior faults as predicted by the depth-dependent compaction process. Model I (Figure 13A) also suggests that compaction effects will be greater if the thickness difference in the basal sequence deposited across the fault is greater. The anticipated effect is not observed in seismic data across the trough. For example, along line 1 thickening of the Rome Formation across the East-Margin fault is about 0.76 km. Total thickness of the Rome over the hanging wall is approximately 1.16 km and that over the footwall is 0.4 km. Additional normal displacement of 0.35 km occurred during the Middle Cambrian, followed by even greater normal displacement of 0.37 km during the Late Cambrian–Middle Ordovician (Rutledge through Trenton events of Figure 5). Thickening of the Rome across the East-Margin fault on line 2 increases to 1.2 km (0.44 km greater

Figure 12—The influence of sediment compaction across a fault developed during deposition of stage 1 sediments (A) is followed through successive stages of sedimentation (B–D).
than on line 1) and on line 3 to approximately 1.3 km; however, thickening over the hanging wall in the overlying Middle and Upper Cambrian section in these areas is only about 0.4 km, which is much less than the 0.76 km of thickening observed on line 1. If the thickening of shallow intervals on line 1 is due primarily to compaction, then thickening along lines 2 and 3 should have been greater than the thickening observed on line 1. The compaction effect should increase in proportion to increases in the thickness of the strata across the fault.

More pronounced differences between the compaction model (Figure 13A) and actual seismic response are illustrated in the RHDs for the East-Margin fault on lines 2 and 4 (Figure 10). These RHDs are characterized by a period of rapid growth (initial 70 m.y.) somewhat similar to that observed in the RHD derived from the model compaction response (Figure 13B); however, instead of the continued thickening over the hanging wall predicted by the model (Figure 13B), thinning associated with gradual uplift of the hanging wall occurs. A close-up view of the seismic data across the fault on line 2 (Figure 14), for example, reveals subtle thinning to the west over the hanging wall in the Rose Run to Trenton and Trenton to Lockport intervals. In the forward compaction model (Figure 13) posttectonic compaction-related offset is about 20% of the total. In general, RHDs for lines 1–4 (Figures 9–11, respectively) are not similar to the RHD computed from the compaction model (Figure 13B). The observations suggest that the majority of sediment compaction occurred prior to deposition of the overlying sequences, thus diminishing or altogether eliminating later differential compaction across the deeper fault.

These observations provide a further example of how use of exponential or other continuous porosity-depth relationships in subsidence calculations for this area will lead to erroneous or irreconcilable portrayals of subsidence history. Given the present-day offset across the East-Margin fault on line 2 (Figures 6B, 14) of approximately 1.2 km, its decompacted thickness, for example, immediately following deposition should have been 1.96 km. If that were so, then 0.76 km of normal offset is missing from the East-Margin fault. The required 1.96 km of normal offset would have to have been followed gradually in time by 0.76 km of reversal to leave the present-day 1.2 km of offset. This reversal would have to have occurred in such a way as to leave the overlying strata unfaulfted.

Although unlikely, it is possible for the basin to evolve in such a way that these compaction-related offsets do not appear. The solution is illustrated in the sequence of stages shown in Figure 15. In this model (model II), the initial basement offset is 2.89 km (Figure 15A). As subsequent layers of sediment blanket the area, the hanging wall is forced to rise by an amount equal to the additional compaction of the basal package of sediments. Thus sediments deposited early in the history of model II (Figure 15B) cause the basal sequence to thin from 2.89 to 2.31 km. If the hanging wall did not rise, this added blanket of sediment would have produced an apparent normal offset of 0.58 km in the top of the basal sequence (Figure 15B). Thus as each blanket of sediment is deposited, reverse movement must occur along the fault by exactly the amount required to balance compaction of the basal interval resulting from the increased weight of overlying sediments. In the final present-day configuration (Figure 15C), there is 2 km of vertical displacement across the fault, and although there is no displacement in the overlying horizons, the fault moved vertically upward 0.89 km during deposition of the
overlying strata. The reactivation history diagram for this model (Figure 15D) erroneously suggests that the present-day offset of 2 km occurred early in the development of the basin and that no additional displacements occurred. This scenario (Figure 15) requires that the hanging wall subside more slowly than the footwall so that the vertical offset across the fault decreases over time at a rate that matches the rate of compaction of the sediments covering the hanging wall. In the following section, I consider the possibility that this fault reversal could occur through the differential isostatic adjustment.

LOADING

Total subsidence is a combination of tectonic and isostatic adjustments to loading. Tectonic subsidence during basin formation is associated primarily with thermal cooling or stretching and thinning of the crust and lithosphere along with thermal cooling. The isostatic component of total subsidence represents the additional subsidence of the lithosphere or crust resulting from the added weight of sediment deposited in the basin. The tectonic component of subsidence is computed using equation 9 (e.g., Sclater and Christie, 1980; Angevine et al., 1990)

\[
Z_i = S_i \frac{\rho_m - \rho_s}{\rho_m - \rho_w} + Wd_i - \Delta SL_i \frac{\rho_m}{\rho_m - \rho_w} \tag{9}
\]

In this equation, \(Z_i\) is the water-loaded basement depth after removal of the \(i\)th layer; \(S_i\) is the compaction-corrected sediment load; \(\rho_m\), \(\rho_s\), and \(\rho_w\) are the densities of the mantle, compacted sediment, and water, respectively; \(Wd_i\) is water depth; and \(\Delta SL_i\) is sea level rise or fall relative to present-day sea level. The tectonic component equals the water-loaded basement depth \((Z)\) after sediment has been stripped away. To determine how the tectonic and loading components of total subsidence must vary

![Close-up view of seismic data across the East-Margin fault along line 2.](image-url)
during the development of model II, layers in the model (Figure 15) were removed (top to bottom) and the unloaded basement depth was computed. We make the simplifying assumption that the depression created by faulting is completely filled by sediment and that \( Wd \) and \( \Delta S \) are zero. Under these assumptions, equation 9 reduces to

\[
Z_i = S_i \frac{\rho_m - \rho_s}{\rho_m - \rho_w}
\]  

(10)

Sediment density (\( \rho_s \)) is computed using equation 11 (after Sclater and Christie, 1980)

\[
\rho_s = \sum_i \left[ \phi_i \rho_w + \left( 1 - \phi_i \right) \rho_{sg} \right] z_i'
\]  

(11)

where \( \phi_i \) is the mean porosity of the \( i \)th layer, \( \rho_w \) is the water density, \( \rho_{sg} \) is sediment grain density in the \( i \)th layer, and \( z_i' \) is the thickness of layer \( i \). Sediment grain density (\( \rho_{sg} \)) of 2.72 gm/cm\(^3\) is used. The average porosity \( \bar{\phi} \) of a layer is determined as follows (see Sclater and Christie, 1980)

\[
\bar{\phi} = \frac{\phi_0}{c} \left( e^{-cz_1} - e^{-cz_2} \right) \left( z_2' - z_1' \right)
\]  

(12)

where \( \phi_0 \) is the uncompacted sediment porosity, \( c \) is the lithology-dependent compaction constant, and \( z_1' \) and \( z_2' \) are the depths to the top and bottom, respectively, of the sequence at an earlier stage.

Equations 10–12 were used to backstrip layers from the model shown in Figure 15C. Sediment grain density (\( \rho_{sg} \)) of 2.72 gm/cm\(^3\) and mantle density (\( \rho_m \)) of 3.33 gm/cm\(^3\) were used for all layers. The time required to deposit each layer shown in the present-day configuration (Figure 15C) is assumed to increase exponentially toward the surface. Total and tectonic subsidence are plotted for the hanging wall (Figure 16A) and footwall (Figure 16B). The isostatic responses of the hanging wall and footwall parallel each other (Figure 16C). The differential isostatic adjustment remains constant (Figure 16D) with the exception of the initial episode of fault development. The results of the modeling indicate that differential subsidence across the fault and the forces driving reversal of the hanging wall must be tectonic in origin. The offset across the fault decreases from the maximum offset of 2.89 km during the initial stage of formation to the present-day offset of 2 km. This relative uplift is required to compensate for compaction of the overlying sediments with depth.

To put these results into perspective, the isostasy equation is solved for the extreme cases represented in models I and II. Isostatic balance requires thinning of the crust by an amount \( b_m \) in response to the addition of a sediment load of thickness \( b_s \), where

\[
b_m = b_s \frac{\rho_c - \rho_s}{\rho_m - \rho_c}
\]  

(13)

The density of the crust (\( \rho_c \)) is taken as 3 gm/cm\(^3\). The thinned crust is replaced by upper mantle with density, \( \rho_m \), which is taken as 3.33 gm/cm\(^3\).

Thinning was computed for sediment thicknesses corresponding to the initial hanging wall offsets of 2 and 2.89 km and final depths of 5 km for the hanging wall and 7 km for the footwall. The original thickness of the unloaded crust was assumed to be 45 km. The results (Figure 17) indicate that if the initial offsets of 2 km (model I) and 2.89 km (model II) are isostatically balanced, then subsediment crust must thin from 45 to 37.26 km and 34.78 km, respectively. After thinning, the base of the crust rests at 39.25 km in model I (Figure 17A) and 37.77 km in model II (Figure 17B). In the final configuration (Figure 17C), fault offset is 2 km and the crust has thinned to 25.9 km beneath the hanging wall and 29.9 km beneath the footwall. This places the base of the crust at 31 and 37 km beneath the surface on the footwall and hanging wall sides of the fault, respectively.

Isostatic compensation across individual faults requires that the fault extend almost vertically through the crust and that the lithosphere has no elastic strength. Recent deep-crustal seismic data has improved our understanding of extension tectonics. Current models of intracontinental extension suggest that the crust is divided into shallow brittle and deeper ductile regions (Strahler, 1998). Allmendinger et al. (1987) presented COCORP data across the Basin and Range Province in the western United States supporting the view that deformation of the lower crust is ductile. Listric normal faults that sole out into the deeper ductile regions of the crust (Miller et al., 1983; Echtler et al., 1994; Gibbs, 1984; Klgiefeld et al., 1984) or that sole-out into a low-angle crust-penetrating shear zone (Wernicke, 1981) are believed to accommodate extension in the upper brittle region of the crust. Individual faults in the shallow crust do not have crust-penetrating roots and are not isostatically compensated. Rather the extended region is isostatically compensated by regional scale crustal thinning as is shown by deep seismic profiling, for example, across the Upper Rhine graben (Echtler et al., 1994). Within this context, it is unlikely that the base of the crust will adjust
isostatically to individual offsets along trough structure as represented in Figure 17. The thickness differences across these structures occur over distances too small to produce Airy isostatic compensation.

**BURIAL CONDITIONS**

The presence of oil in a given reservoir interval partly depends on whether potential source rocks generated hydrocarbons, and this will depend, in part, on accurate assessment of source rock burial depth and the length of time spent at a certain depth. As an example, subsidence history curves corrected and uncorrected for compaction are compared in Figure 18. Sediment compaction must certainly occur, but the actual compaction process may follow a curve that stair-steps between the continuous compaction and no-compaction (uncorrected) predictions. There is an uncertainty in our knowledge of when the horizon at 4 km, for example, actually reached that depth. In the example (Figure 18) it could have reached that depth anytime between 57 and 95 m.y. after subsidence began. For a potential source rock just below the oil window (at 2 km in this example), there is an uncertainty of about 50 m.y. in the estimated time that the potential source rock actually passed through that depth. If rapid unroofing began around 275 m.y., then the continuous compaction prediction would indicate that the horizon spent very little time at that depth.
depth and is unlikely to have generated oil. The amount of time spent inside the oil window increases as the compaction rate is reduced.

Estimates of maximum burial depth are also affected by this potential source of error. The analysis of appetite fission track data presented by Roden (1992, 1999, personal communication) and Donaldson (1994) indicates that rocks currently at the surface in the West Virginia area of the Rome trough were once buried at depths of approximately 3 km. Rocks at various present-day depths throughout this area would have been buried at depths of 3 km and greater during the early Mesozoic. Estimates of the maximum depth reached by any given horizon depend on whether the estimates are corrected or uncorrected for compaction effects. Consider two horizons, one at present-day depth of 1.5 km and the other at 7 km (Figure 19A). If rocks currently at the surface are taken to a new depth of 3 km, what total depths will these two horizons actually reach? If compaction continues to operate on the already deposited rocks, then the new depths of these two horizons will be 3.94 and 9 km, respectively (Figure 19B). If, however, these strata do not undergo further compaction, then their new depths will be 4.5 and 10 km, respectively (Figure 19C). With continuous compaction, the shallow interval never descends into the gas window, and source rocks at this depth are potentially oil producing. Conversely, if further compaction is insignificant, then source rocks descend into the gas window.

CONCLUSIONS

Block rotation and vertical displacements through time across normal faults in the trough follow an exponential decay process. The exponential time
The time constant \( t_0 \), derived by exponential fitting, represents the time taken for 37% of the maximum possible offset to occur across a given structure. Differential subsidence across the western margin of the trough is characterized by time constants that range from 63 to 78 m.y. Time constants for the East-Margin fault increase from 13 to 51 m.y. northeast to southwest along the fault. Development of the Interior fault is consistently of short duration with time constants ranging between 16 and 23 m.y. Time constants for the west margin are in all cases larger than those for the East-Margin and Interior faults and indicate that rotational collapse of the west margin continued over a longer time period. The Interior fault developed quickly during the 70 m.y. time span of the Cambrian period. Development of the trough’s east margin is also a relatively short-term event confined primarily to the Cambrian period, with the exception of line 1 where normal displacements across the East-Margin fault continued through the Ordovician. Long-term subsidence of the trough is characteristic of the southern West Virginia segment of the Rome trough (Gao, 1994; Gao et al., 2000). The boundary between the southern and northern West Virginia segments of the trough occurs between lines 1 and 2 (Figure 2).

Understanding the cause for variation in the historical development of different trough structures requires further study. Long-term rotation of the trough’s west margin may reflect, in part, its participation in the overall subsidence of the Appalachian foreland. Differential subsidence across the west margin of the trough continued throughout most of the Paleozoic, whereas development of the East-Margin and Interior faults were relatively short-lived events restricted, in general, to the Cambrian period; however, uplift of the trough interior often occurred along the East-Margin and Interior faults following the Cambrian. This uplift may indicate that the East-Margin and Interior faults were passive participants in foreland subsidence of the Appalachian basin.

The import of the observations presented in this study to hydrocarbon exploration is related to their definition of the time periods over which developing structures are likely to have had an effect on potential reservoir intervals. Beardsley and Cable (1983) noted that recognition of the concept of reactivation of specific structures is critical to the development of frontier exploration working hypotheses. In the northern part of the study area, the East-Margin and Interior faults developed rapidly. Their impact on sedimentation was greatest during the Cambrian. Potential reservoirs may have

Figure 17—The loading effect of sediments for (A) model I, (B) model II, and (C) the final stage.

Figure 18—Hypothetical subsidence curves are shown for deep and shallow horizons. Curves are shown with and without the compaction correction. A horizon deposited at 95 m.y. is followed along corrected and uncorrected subsidence paths. Hypothetical locations of oil and gas windows are shown.
been created in these deeper intervals. Minor uplift occurring across the East-Margin and Interior faults later in the history of the basin may have created secondary fracture porosity in shallower intervals. The Interior and East-Margin faults represent potential deep exploration targets within the northern West Virginia area. In southwestern West Virginia the East-Margin fault was active over a longer period of time, and late-stage uplift and inversion are more pronounced (Gao, 1994; Gao et al., 2000); hence, the influence of this structure on reservoir formation in the southern West Virginia area extends through much of the Paleozoic (Gao, 1994; Gao et al., 2000; Yang, 1998). Long-term collapse across the west margin had a significant influence on sedimentation (Ryder et al., 1992; Ryder, 1992) and the creation of secondary fracture porosity along its length. The early formation of these structures places them in potential trapping positions early in the history of the basin. Continued movement across these structures through time influences current reservoir conditions in sequentially deposited depositional systems overlying these early structures (Beardsley and Cable, 1985).

Seismic interpretation and compaction modeling indicate that reactivation history diagrams accurately describe the history of vertical movements observed across individual structures within the Rome trough. The final configuration of layers in model II (Figure 15C) suggests that a fault with 2 km of vertical offset occurred early in the depositional history of the model. Model II (Figure 15C) is similar in appearance to the seismic response observed across line 2 (Figure 14). It seems unlikely that the East-Margin fault along line 2 evolved through the delicate match of tectonic displacement to compaction rate that was required to eliminate long-term compaction-related offsets across the fault in model II. Interpretation and modeling suggest that compaction of major stratigraphic intervals must terminate prior to deposition of subsequent intervals. Local isostatic adjustments of the lithosphere to differential loading across faults or rotating margins also appear unlikely to occur within the context of current models of intracratonic extension.

Observations and modeling suggest that reactivation history diagrams, when uncompensated for compaction and loading, accurately portray differential tectonic subsidence across Rome trough structures. The results suggest that the standard approach to backstripping may lead to erroneous results. Corrections for compaction and loading effects may lead to errors in the magnitude of estimated tectonic strain and portrayal of basin history. The application of these corrections must be considered basin by basin.


Wernicke, B., 1981, Low-angle normal faults in the Basin and Range province: nappe tectonics in an extending orogen:
