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New Mexico Geology, v. 29, n. 1 pp. 3-12, Print ISSN: 0196-948X, Online ISSN: 2837-6420.
https://doi.org/10.58799/NMG-v29n1.3

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Abstract

Sequence stratigraphy of the Lower Permian Abo member in the Robledo and Doña Ana Mountains of south-central New Mexico is established by correlation of closely spaced outcrops in a direction subparallel to the paleoshoreline. Sixteen to eighteen fourth-order sequences 3–15 m thick were associated with cutting of incised valleys. Interfluves between the incised valleys and inset fluvial terraces were capped by mature, polygenetic paleosols. The incised valleys were filled with lowstand fluvial red beds and, in some cases, by transgressive and highstand estuarine deposits, constituting a complete sequence. In other cases, transgression continued beyond infilling of the valleys by lowstand fluvial and transgressive estuarine deposits, such that marine facies were deposited across adjacent interfluves. Fifth-order cycles composed of meter- to centimeter-scale interbeds of shallow-marine limestone and shale-siltstone are also present in the Abo member and are similar to many of the cyclothems of Pennsylvanian and Early Permian age in the midcontinent of the United States. The fourth-order sequences and fifth-order cycles in the Abo member were primarily related to eustatic sea-level changes driven by waxing and waning of continental glaciers in Gondwana. Moreover, alternating arid and more humid climatic conditions associated with glacial cycles may have controlled carbonate versus siliciclastic deposition in the fifth-order cycles. A period of differential subsidence may be required to account for the presence of a 25-m-thick, highly fossiliferous "slope-forming interval" in the Robledo Mountains and its absence in the Doña Ana Mountains.

Introduction

Sequence stratigraphy is the current paradigm designed to explain the basin-scale arrangement of siliciclastic depositional systems and associated stratigraphic surfaces of erosion and nondeposition in terms of relative rise and fall of sea level (Fig. 1A; Van Wagoner et al. 1988, 1990). A "sequence" represents the sediment deposited between successive periods of relative sea-level lowstand and is bounded above and below by unconformities or their correlative seaward conformities. A type 1 sequence boundary is produced during a drop in sea level, when rivers on the coastal plain erode into underlying deposits as a result of lowering of their base levels and lengthening of their channels (Van Wagoner et al. 1988; Blum and Tornqvist 2000). This often results in a landscape of incised valleys separated by interfluves that may be sites of nondeposition and soil formation (Fig. 1B). The process of fluvial incision may be episodic, creating one or more inset terraces (Fig. 1B; Strong and Paola 2006). Sediment deposited during the period of sea-level fall and initial rise is referred to as the lowstand systems tract (Fig. 1A, LST).

When the rate of sea-level rise increases, sediments of the transgressive systems tract are deposited (Fig. 1A, TST). If incised valleys were cut during lowstand, they will be flooded by the rising sea, creating estuaries. The transgressive systems tract may be condensed compared to the other systems tracts, because of erosion beneath one or more marine ravinement surfaces. Near the crest of the sea-level curve, the rate of detrital influx to the shoreline may exceed the space made available by slow sea-level rise, causing the shoreline to prograde seaward, a process that may continue during subsequent slow sea-level fall. Sediment deposited during this interval of progradation and coastal offlap belongs to the highstand systems tract (Fig. 1A, HST). Deposition of the highstand systems tract proceeds until the rate of sea-level fall accelerates, at which time coastal-plain rivers once again begin to erode, creating the upper sequence boundary.

Lower Permian strata along the western margin of the Orogrande Basin of south-central New Mexico consist of interbedded terrestrial and marine rocks (Fig. 2; Kottlowski 1963), suggesting that this area occupied a position between the limit of coastal incision during sea-level lowstand and maximum onlap during marine transgression (cf. Blum and Tornqvist 2000). These strata in the Robledo and Doña Ana Mountains near Las Cruces are well suited for sequence stratigraphic analysis, because of the ability to correlate depositional facies and stratigraphic surfaces along canyon walls and between closely spaced outcrops in a direction roughly parallel to the paleoshoreline. This paper builds upon earlier sequence stratigraphic interpretations of the Abo member by Mack et al. (2003a) by presenting more detailed correlations in both mountain ranges and by interpreting small-scale, marine limestone-shale cycles in terms of sequence stratigraphy.

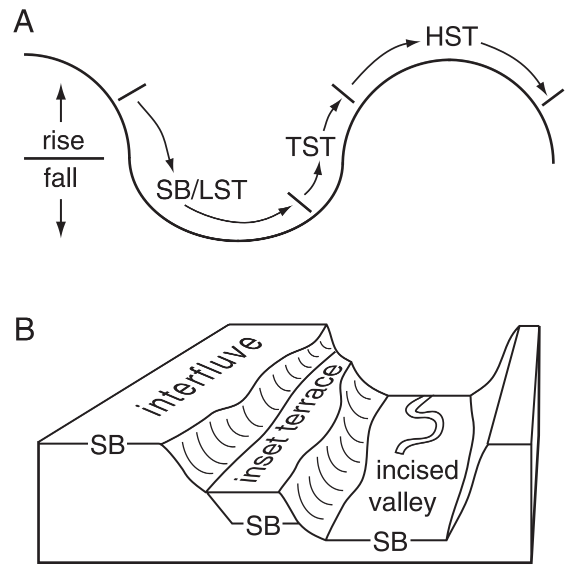


FIGURE 1—**A:** Schematic sinusoidal sea-level curve showing the probable time of erosion of a sequence boundary and deposition of systems tracts. SB = sequence boundary; LST = lowstand systems tract; TST = transgressive systems tract; HST = highstand systems tract. **B:** Incised valley cut during sea-level lowstand, showing interfluves and one inset fluvial terrace. SB = sequence boundary.

Geologic setting

Stratigraphy

In the Robledo Mountains, approximately 100 m of interbedded red beds, gray shale, and limestone were mapped as the Abo Tongue of the Hueco Formation by Seager et al. (in press), whereas similar strata in the Doña Ana Mountains were assigned to the Abo Formation by Seager et al. (1976). These strata are considered correlative (Wahlman and King 2002) and are referred to in this study as the Abo member for ease of discussion. Although fusulinids are absent in the Abo member in both mountain ranges (Wahlman and King 2002), a marine invertebrate fauna in the Robledo Mountains suggests a late Wolfcampian (Sakmarian) age (LeMone et al. 1975; Kues 1995; Kietzke and Lucas 1995). In addition, conodonts collected from the lower part of the Abo member in the Robledo Mountains were assigned to the Sterlitamakian, the upper substage of the Sakmarian Stage

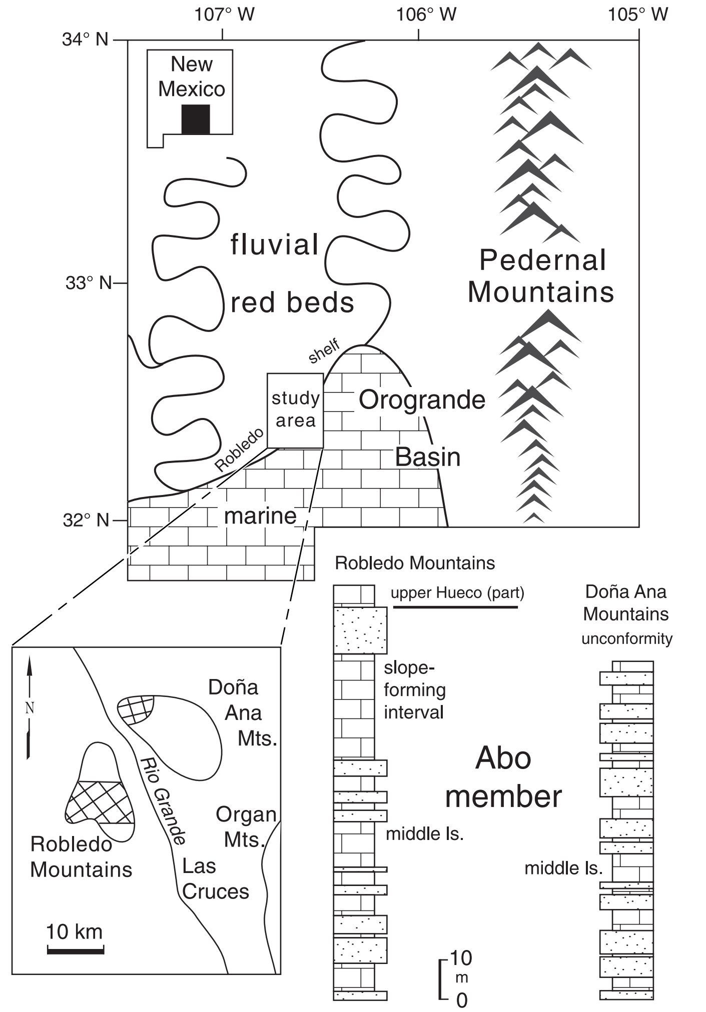


FIGURE 2—Paleogeographic map of south-central New Mexico in Early Permian (late Wolfcampian/Sakmarian) time; study area includes modern Robledo and Doña Ana Mountains, where the Abo member is exposed (cross-hatched area); stratigraphic sections show red-brown sandstone-siltstone intervals, as well as some other key horizons used for correlation. Note that the "slope-forming interval" in the Robledo Mountains has no equivalent in the Doña Ana Mountains.

peritidal dolomicrite	light tan to yellow, desiccation cracks, fenestral fabric, stromatolites, breccia, root traces, calcite pseudomorphs after swallowtail gypsum; 0.2–0.5 m
shallow-marine limestone	micrite matrix; divisible into lagoonal, restricted-marine, and normal-marine by increasing diversity of invertebrate benthic fossils; 0.2–2 m
lower shoreface shale-siltstone	gray shale with ostracodes and bivalves; tan siltstone with hummocky strata and oscillation ripples; 0.1–2 m
estuarine	gray and green, very fine sandstone, siltstone, mudstone; hummocky strata, oscillation ripples, root traces; plant fossils, ripple cross-laminae, horizontal laminae, ostracodes, bivalves; 0.5–8 m
fluvial	red, brown, thin-bedded crevasse-splay siltstones with ripple cross-laminae, horizontal laminae, root traces, desiccation cracks, plant fossils, burrows, tracks; red overbank mudstone, desiccation cracks; 1–7 m red-brown, channel very fine sandstone and siltstone with lateral accretion sets, ripple cross-laminae, horizontal laminae, crossbeds, desiccation cracks, plant fossils, burrows, tracks; 2–11.5 m

FIGURE 3—Summary of the depositional environments of the Abo member in the Robledo and Doña Ana Mountains. Colors are the same as in Figures 4–7.

(Kozur and LeMone 1995). Based on these biostratigraphic data and current estimates of the age of the boundaries between the stages of the Lower Permian (Gradstein et al. 2004), the duration of deposition of the Abo member is between 2 and 10 m.y. The longer duration assumes the Abo member spans the entire Sakmarian Stage, whereas the shorter duration applies if the Abo member is restricted to the Sterlitamakian substage, as estimated by Ross et al. (1994).

The Abo member in the Robledo Mountains contains nine to 11 red or brown sandstone-siltstone intervals (Fig. 2), which were correlated throughout the range by their stratigraphic relationship to four yellow dolomicrites, a "middle limestone" approximately 5 m thick, and the overlying cliff-forming upper Hueco member. The Abo member in the Robledo Mountains also has a distinctive "slope-forming interval" of very fossiliferous limestone and shale approximately 25 m thick that is not present in the Doña Ana Mountains. Although the Abo member in the Robledo Mountains is exposed over an area of approximately 25 km^2, structural complications hamper correlation, resulting in logged sections spaced 0.1–2.3 km apart.

In the Doña Ana Mountains, the Abo member consists of 12 intervals of red or brown sandstone or siltstone (Fig. 2), which can be correlated over an area of approximately 5 km^2 by tracing individual sandstone-siltstone intervals along canyon walls or by their stratigraphic relationship to five yellow dolomicrites, two *Rhizocorallium*-bearing beds, and a "middle limestone" that is comparable in stratigraphic position and thickness to that in the Robledo Mountains. The contact between the Abo member and overlying upper Hueco member is not exposed in the Doña Ana Mountains due to Laramide (latest Cretaceous–Eocene) erosion (Seager et al. 1976). Because of excellent exposures in Lucero Arroyo and adjacent canyons, logged sections in the Doña Ana Mountains are spaced only a few hundred meters apart.

Depositional environments and paleosols

The Abo member in the Doña Ana and Robledo Mountains was deposited in fluvial, estuarine, shallow-marine, and peritidal environments. These facies are described by LeMone et al. (1975), Mack and James (1986), Krainer and Lucas (1995), and Mack et al. (2003a, b), and are summarized in Figure 3. In addition, mature, polygenetic paleosols are present in the upper parts of some estuarine and shallow-marine beds (Mack et al. 2003a). Pedogenic features that formed during the initial stage of development include vertical root traces, stage II morphology calcic nodules and tubules, and vertic structures, such as wedge-shaped peds, slickensides, and deep (0.5–1 m) desiccation cracks. Superimposed on these earlier features is gley color mottling

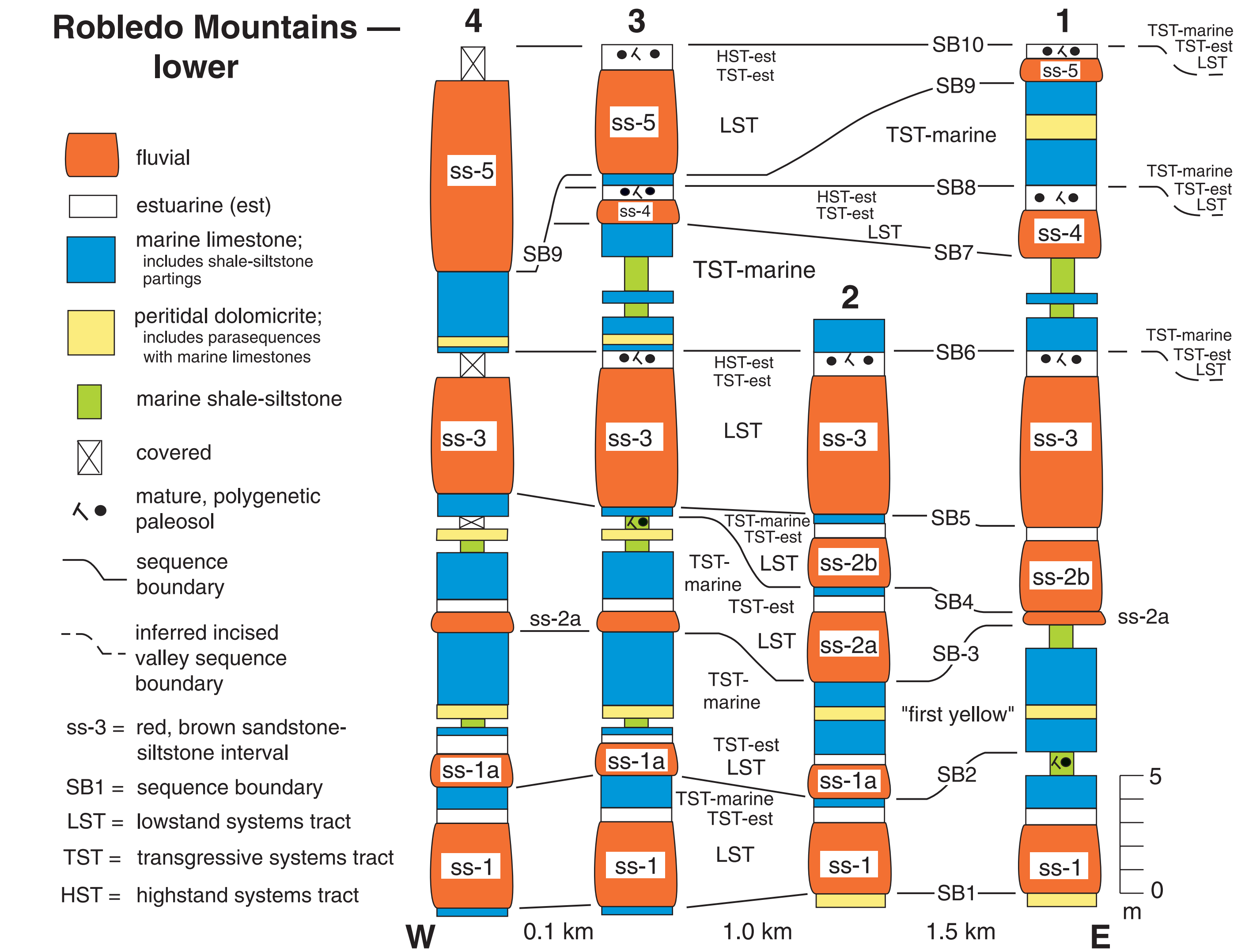


FIGURE 4—Logged sections of the lower Abo member in the Robledo Mountains, projected onto an east-west line. Section 1 = NE¼ SW¼ sec. 30; 2 = NW¼ NE¼ sec. 25; 3 = NE¼ NE¼ sec. 23; 4 = SE¼ SE¼ sec. 23; all sections are in T22S R1W, except section 1, which is in R1E.

and goethite as small nodules, as coatings on peds, and as replacement of calcic nodules.

Sequence stratigraphy

Fourth-order sequences

Interbedding of fluvial, estuarine, and marine/peritidal facies in the Abo member defines 16 sequences in the Robledo Mountains (Figs. 4, 5) and 18 sequences in the Doña Ana Mountains (Figs. 6, 7). The sequences range from 3 to 15 m thick, and each sequence begins at the base of one of the numbered sandstone-siltstone intervals and ends either at the erosional base of the next overlying sandstone-siltstone interval or with a mature, polygenetic paleosol that caps estuarine or marine shale-siltstone deposits. Using the estimates for the age of the Abo member presented above, the average duration of each sequence can be bracketed between about 100 and 625 ka, placing them within the range of fourth-order cycles of Van Wagoner et al. (1990).

The base of each fluvial sandstone-siltstone interval defines a sequence boundary, because it is erosional and because it constitutes a non-Waltherian seaward shift in facies (cf. Van Wagoner et al. 1988, 1990). The existence of incised valleys is inferred by the presence of estuarine facies overlying fluvial deposits, as well as by the ability to correlate some sequence boundaries to adjacent interfluves. In the latter case, mature, polygenetic paleosols cap the interfluves and indicate a long period of landscape stability characterized by the formation of vertic and calcic horizons in a well-drained soil, followed by gleying and precipitation of goethite as a result of a rising and/or perched water table (Fig. 4, SB2, SB4; Fig. 5, SB13; Fig. 6, SB1, SB8).

The incised valleys cut from 2 to 8 m into underlying strata, locally truncating older sequences and sequence boundaries (Fig. 4, SB9; Fig. 6, SB4). In addition, a fluvial terrace is recognizable at one location (Fig. 6, section 2, ss-2). There, an outcrop of fluvial sediment capped by a mature, polygenetic paleosol can be traced several hundred meters to the northeast, where it is truncated by an erosional surface produced during a younger episode of valley incision. Fluvial strata on either side of the truncation surface have different channel architectures, consistent with deposition by different channels at different times. As expected, the inferred fluvial terrace and its paleosol are overlain by a thinner succession of estuarine sediment than exists in the younger, deeper part of the incised valley. Using this example as a guide, the mature, polygenetic paleosol separating fluvial and estuarine sediment in ss-10 in the Doña Ana Mountains (Fig. 7, section 9) is interpreted to have developed on an inset terrace, even though the margins of the terrace are not exposed.

Following the convention of Allen and Posamentier (1993, 1994), fluvial sediment deposited within an incised valley is designated as the lowstand system tract, despite the fact that some or all of it may have been

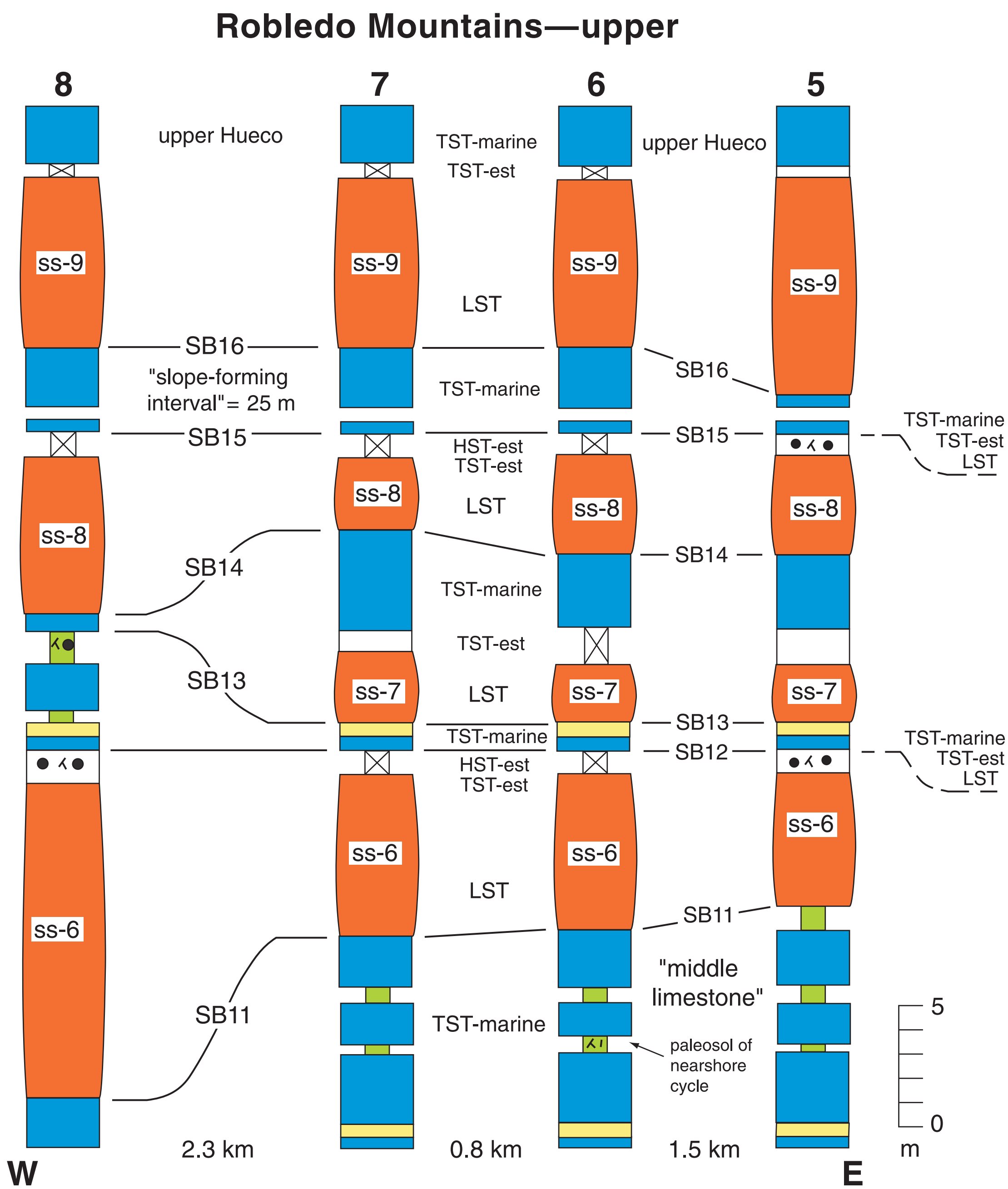


FIGURE 5—Logged sections of the upper Abo member in the Robledo Mountains, projected onto an east-west line. See Figure 4 for key to symbols. In order to save space, the thickness of the slope-forming interval is not to scale. Section 5 = NE¼ SW¼ sec. 30; 6 = NE¼ NW¼ sec. 25; 7 = SE¼ NE¼ sec. 26; 8 = NW¼ SE¼ sec. 22; all sections are in T22S R1W, except section 5, which is in R1E.

deposited as a result of base-level rise during transgression (cf. Van Wagoner et al. 1990). Designating fluvial sediment as the lowstand systems tract is preferred in this study, because it is not possible to accurately correlate regional sea-level turnarounds with the logged sections.

During sea-level rise, the incised valleys and their lowstand fluvial fills were buried by estuarine sediment and/or were partially eroded by one or two marine ravinement surfaces (Mack et al. 2003a). The marine ravinement surfaces are locally mantled by thin (<0.5 m) granule and pebble lags composed of marine fossils and rip-up clasts of marine limestone and/or fluvial or estuarine sediment. The transgressive surface, defined as the boundary between the lowstand and transgressive systems tracts, corresponds to either the lowest marine ravinement surface or to the contact between fluvial and overlying estuarine deposits (Mack et al. 2003a).

Incised valleys of the Abo member may be described as either "estuarine-filled" or "transgression-filled" (Fig. 8). Estuarine-filled incised valleys were filled to the level of the adjacent interfluves with lowstand fluvial sediment and overlying transgressive and highstand estuarine deposits (Fig. 8A). In this case, the fill of the incised valley constitutes a complete sequence and is immediately overlain by the next sequence boundary. In contrast, transgression-filled incised valleys were filled to the level of the interfluves with lowstand fluvial and transgressive estuarine deposits, but the transgression continued, such that marine sediment was deposited above the incised valley fill and onto the adjacent interfluves. Thus, in a transgression-filled incised valley, the transgressive systems tract consists of transgressive estuarine deposits and their associated marine ravinement surfaces, as well as overlying marine limestone and shale-siltstone (Fig. 8B).

Five sequence boundaries in the Robledo Mountains (Fig. 4, SB6, SB8, SB10; Fig. 5,

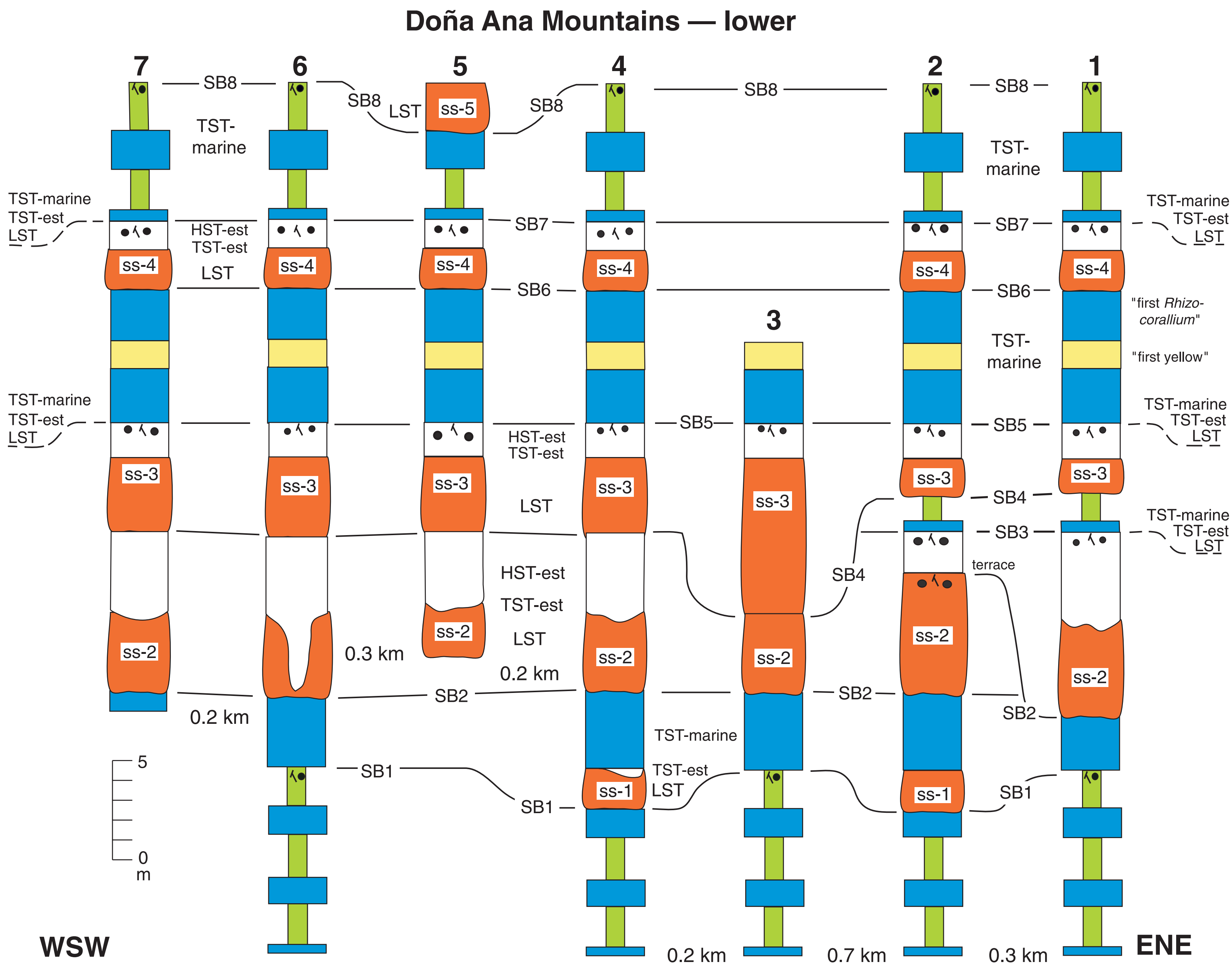


FIGURE 6—Logged sections of the lower Abo member in the Doña Ana Mountains, projected onto an east-west line. See Figure 4 for key to symbols. Section 1 = SW¼ NE¼ sec. 8; 2 = SE¼ NW¼ sec. 8; 3 = NE¼ SE¼ sec. 6; 4 = SE¼ SE¼ sec. 6; 5 = NE¼ SE¼ sec. 7; 6 = NW¼ NE¼ sec. 7; 7 = SE¼ NW¼ sec. 7; all sections are in T21S R1E.

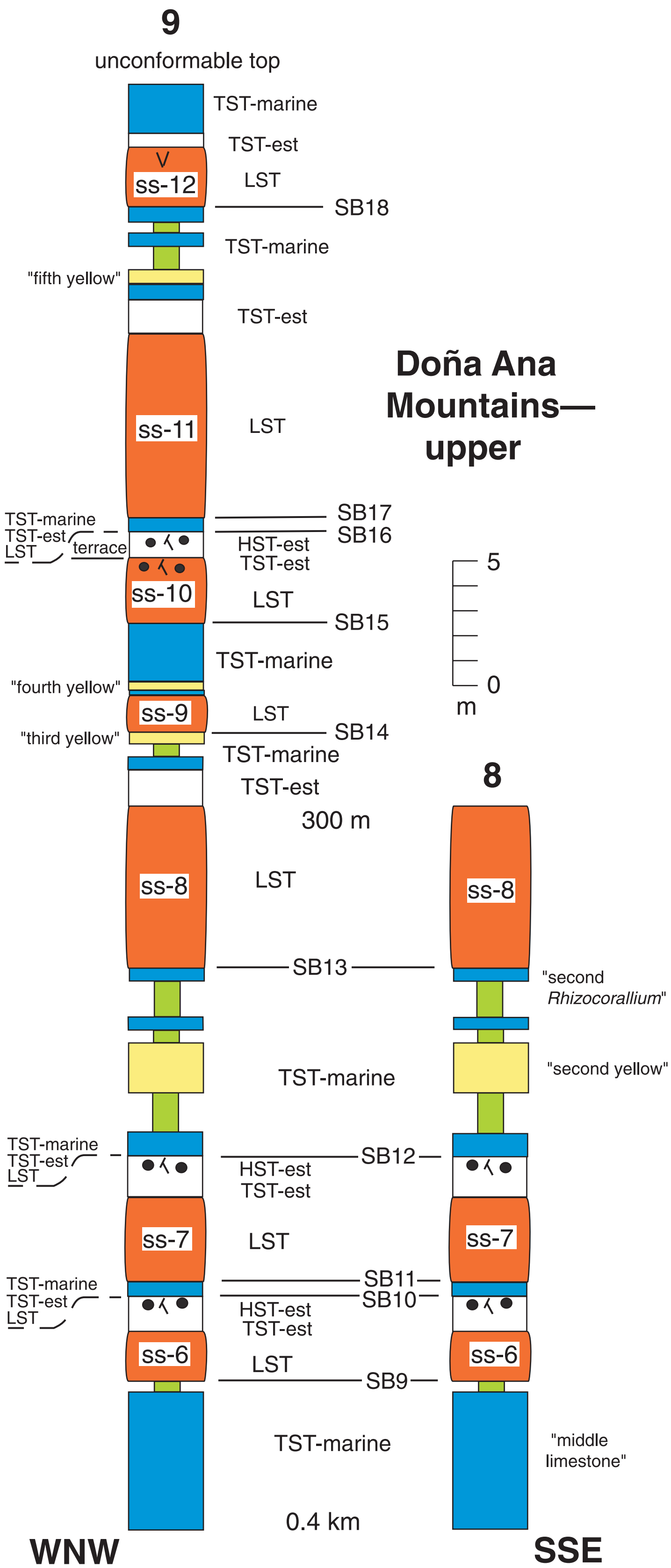


FIGURE 7—Logged sections of the upper Abo member in the Doña Ana Mountains, projected onto an east-west line. See Figure 4 for key to symbols. Section 8 = NW¼ NW¼ sec. 7 T21S R1E; 9 = NE¼ NE¼ sec. 12 T21S R1W.

SB12, SB15) and six in the Doña Ana Mountains (Fig. 6, SB3, SB5, SB7; Fig. 7, SB10, SB12, SB16) were selected based on the presence of mature, polygenetic paleosols that cap the estuarine fill of an incised valley. These paleosols are virtually identical to the polygenetic paleosols discussed previously, and, as such, are interpreted to have formed on interfluves associated with incised valleys that were located beyond the limits of the study areas ("inferred incised valleys" of Figs. 4–7). The interfluve sequence boundaries and their polygenetic paleosols are immediately overlain by marine facies, implying that the inferred incised valleys were of the transgression-filled variety. In several cases, as the interfluve was initially inundated, vertical *Skolithos* and *Diplocraterion* burrows belonging to the *Glossifungites* ichnofacies were emplaced in the firm, upper part of the paleosol before deposition of the overlying marine sediment (Mack et al. 2003a).

At the present time, biostratigraphic resolution is not precise enough to correlate fourth-order sequences and sequence boundaries in the Abo member between the Robledo and Doña Ana Mountains. Thus, identically numbered sequence boundaries in the two ranges do not necessarily represent the same event.

Fifth-order cycles

Embedded within the transgressive systems tracts of the fourth-order sequences are meter- to centimeter-scale parasequences of marine limestone overlain by peritidal dolomicrite (six intervals in the Robledo Mountains; five in the Doña Ana Mountains) and 60 or more cycles of marine limestone overlain by marine shale-siltstone. The latter are comparable to fifth-order cycles described by Olszewski and Patzkowsky (2003) in Lower Permian strata of Kansas and Nebraska (Fig. 9). Because these fifth-order cycles are interpreted to have been deposited between successive inflection points on a sea-level curve and may be separated into systems tracts, Olszewski and Patzkowsky (2003) argue that they represent very thin depositional sequences.

In some cases within the Abo member, the limestone and shale-siltstone beds that constitute the fifth-order cycles are of comparable thickness, from 0.5 to 2 m. In other cases, however, beds of limestone 10–50 cm thick are separated by a shale-siltstone parting only a few centimeters thick. In all cases, contacts between facies are sharp, and locally the basal few centimeters of a limestone may have rip-up clasts of shale or siltstone. Both the base and top of limestone beds commonly display trace fossils of the *Glossifungites* ichnofacies developed on firmgrounds. At the base of the limestone beds are large *Thalassinoides* and *Spongeliomorpha* burrows in convex hyporelief, produced by passive infilling of burrows originally emplaced in the underlying shale-siltstone (Fig. 10A). Similarly, the upper part of limestone beds exhibits passively filled, vertical *Diplocraterion*, *Skolithos*, and *Arenicolites* burrows (Fig. 10B) that extend downward tens of centimeters, locally crosscutting older feeding burrows.

With one exception, interbeds of marine limestone and marine shale-siltstone in the Abo member are comparable to the offshore cycles of Olszewski and Patzkowsky (2003). In their model, offshore cycles are created by the superposition of fifth-order sea-level changes on the steep, transgressive part of a fourth-order sea-level rise (Fig. 9). The resultant composite fifth-order sea-level curves display steep sea-level rise, but very gentle or no sea-level fall. As a result, there is no subaerial exposure nor is there a significant seaward shift in facies, because sea level either rose continuously or fell such a small amount that fluvial incision did not occur. The succeeding rapid rise in sea level on the composite fifth-order curve initially resulted in a condensed section represented by the *Glossifungites* ichnofacies at the base of the limestone beds. The muddy or silty host of the *Glossifungites* burrows may have become firm as a result of early cementation during very slow sediment accumulation or as a result of exhumation by marine erosion of compacted mud and silt, indicated by rip-up clast lags at the base of some limestones. The overlying limestone was deposited during continued sea-level rise and constitutes the transgressive systems tract. The *Glossifungites* ichnofacies at the top of the limestone bed was probably produced by partial cementation associated with slow rate of sedimentation and represents the maximum flooding surface (Fig. 9). As the rate of sea-level rise diminished, marine shale-siltstone was deposited as the highstand systems tract (Fig. 9). The change from carbonate to siliciclastic sedimentation within the fifth-order cycles may also be related to paleoclimate change, which is discussed in more detail below.

In the model of Olszewski and Patzkowsky (2003), fifth-order cycles are also superimposed on fourth-order highstand systems tracts (Fig. 9). Referred to as nearshore cycles, they are distinguished from the offshore cycles described above by the presence of a paleosol at the top of the highstand marine shale-siltstone interval, defining the sequence boundary. However, this has only been observed at one location in the Abo member (Fig. 5, middle limestone). The paucity of highstand, nearshore cycles in the Abo member may be related to truncation of the paleosols during transgression, rendering the cycles impossible to distinguish, and/or to complete removal of the cycles by deep erosion associated with fourth-order incised valleys.

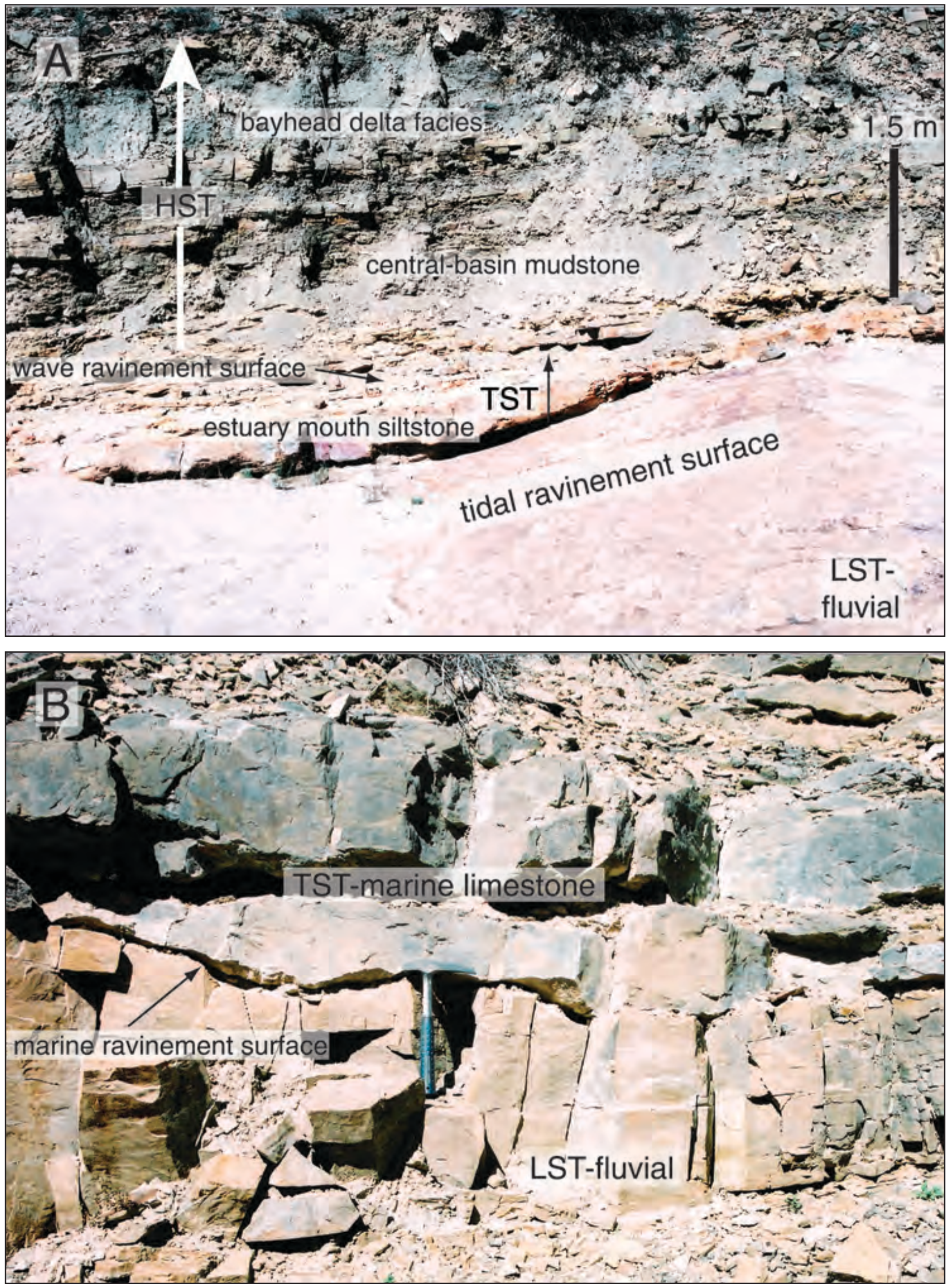


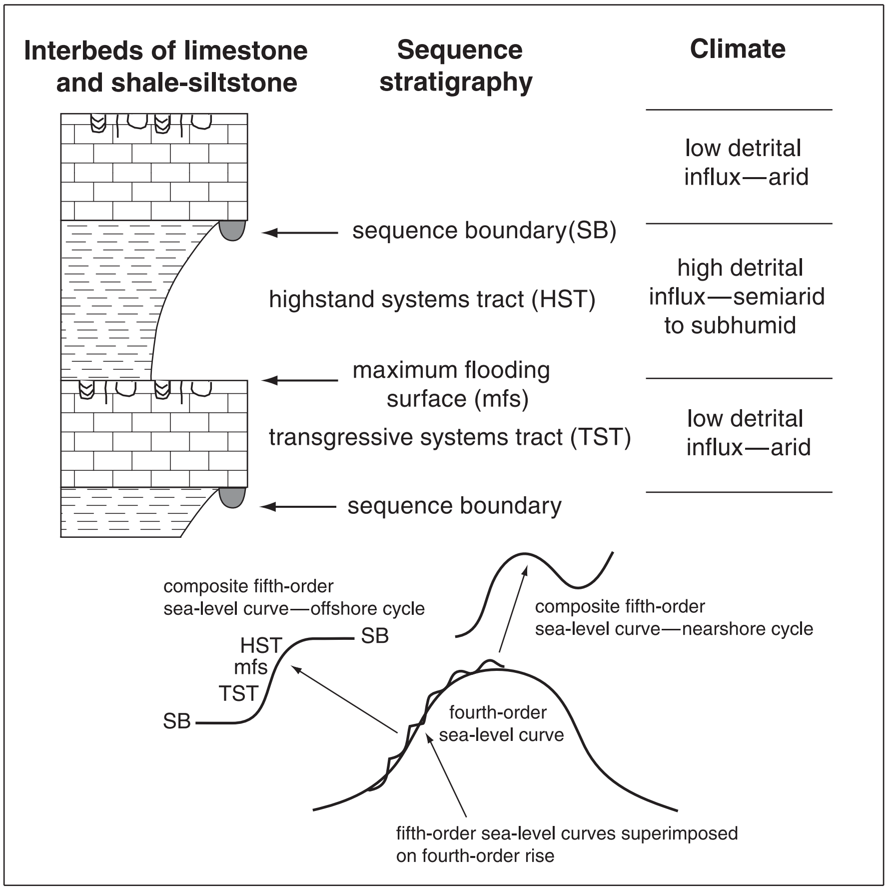
FIGURE 8—**A:** Example of an estuarine-filled incised valley, sandstone-siltstone-2 (ss-2), section 5, Doña Ana Mountains (Fig. 6; Mack et al. 2003a). A tidal ravinement surface and its associated lag cuts into underlying lowstand fluvial sandstones and is overlain by transgressive estuary mouth sandstones that are locally truncated by a wave ravinement surface and its associated lag. The remainder of the outcrop consists of progradational, highstand estuarine sediment. **B:** Example of transgression-filled incised valley, sandstone-siltstone-1 (ss-1), section 4, Doña Ana Mountains (Fig. 6; Mack et al. 2003a). Lowstand fluvial sandstones and transgressive estuarine sandstones (not present in this photo) infilled the incised valley, while continued sea-level rise cut a marine ravinement surface and deposited shallow-marine limestone above the fill of the incised valley and across the adjacent interfluves. Hammer is 25 cm long.

Discussion

The specific pattern of sea-level change in a sedimentary basin reflects the interplay between eustasy and local tectonic and climatic effects. Although it is often difficult to distinguish the relative effects of these variables on sea-level change, it is instructive to explore their possible contributions to the sequence stratigraphic record.

Glacial eustasy

Fourth-order sequences and fifth-order cycles in the rock record commonly are assumed to be the result of glacial-eustatic sea-level changes driven by Milankovitch cycles (Van Wagoner et al. 1990). This is a reasonable interpretation for the Abo member, because of the existence of late Paleozoic glaciers on Gondwana (Caputo and Crowell 1985; Veevers and Powell



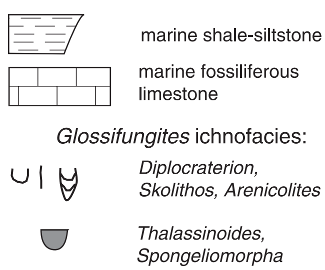


FIGURE 9—Origin of fifth-order cycles represented by marine limestone and overlying marine shale-siltstone, adapted from Olszewski and Patzkowsky (2003). With one exception, the fifth-order cycles in the Abo member correspond to the offshore cycles of Olszewski and Patzkowsky (2003).

1987; Lindsay 1997). Continental glaciation attained peak volumes on Gondwana in Middle to Late Pennsylvanian time, when the amplitude of sea-level change is estimated to have been from 40 to more than 200 m (Heckel 1977; Ross and Ross 1985; Crowley and Baum 1991; Maynard and Leeder 1992; Soreghan and Giles 1999). By Early Permian time, however, the extent of continental glaciation had dramatically decreased, being restricted in Sakmarian/late Wolfcampian time to eastern Australia and, perhaps, to parts of Antarctica (Caputo and Crowell 1985; Veevers and Powell 1987; Lindsay 1997). The fourth-order sequences in the Abo member would most likely correspond to the eccentricity-driven Milankovitch cycles, although Abo chronostratigraphy is not precise enough to distinguish between 100 ka and 400 ka cycles. The fifth-order cycles in the Abo member probably are the result of higher-frequency Milankovitch cycles related to axial tilt and precession. It is also possible that inset fluvial terraces were the product of high-frequency sea-level changes, but they also could have been influenced by climate change or by episodes of downcutting and lateral widening of the valleys that were out of phase with sea-level change (cf. Blum and Valastro 1994; Blum and Törnqvist 2000; Strong and Paola 2006).

Paleoclimate

Paleoclimatic data from the midcontinent of the United States indicate that Permian–Pennsylvanian glacial/interglacial cycles were accompanied by terrestrial paleoclimate change, although disagreement exists concerning the magnitude of the paleoclimate changes and their timing with respect to the sea-level curve (Cecil 1990; Tandon and Gibling 1994; Miller et al. 1996; Rankey 1997; Olszewski and Patzkowsky 2003). In the Abo member, the presence of vertic features and calcic nodules in the polygenetic paleosols that occupied interfluves suggests that a semiarid paleoclimate with seasonal precipitation existed during fourth-order sea-level lowstands (cf. Mack 2003). Unfortunately, there are no reliable paleoclimatic indicators in the estuarine facies of the Abo member to assess paleoclimatic conditions during subsequent transgression and highstand.

A link between glacial/interglacial cycles and paleoclimate has been called on to explain the interbedding of marine limestone and marine shale-siltstone in Permian–Pennsylvanian cyclothems of the midcontinent, and is applicable to the fifth-order cycles in the Abo member (Fig. 9). In the models of Cecil (1990), Soreghan (1997), and Olszewski and Patzkowsky (2003), marine carbonate sediment was deposited during drier periods, whereas deposition of marine shale-siltstone occurred during wetter times, when siliciclastic sediment yields were higher. In the Abo member, a relatively arid paleoclimate existed during deposition of the carbonate part of the fifth-order cycles, based on calcite pseudomorphs after swallowtail gypsum in some lagoonal limestones and peritidal dolomicrites (Mack and James 1986; Mack et al. 2003a). Conversely, a brackish ostracode fauna in the shale-siltstone facies (Kohn 1993) is consistent with greater influx of freshwater to the coast as a result of greater precipitation. However, there are no quantitative data in the Abo member or in other midcontinent cyclothems to determine the absolute difference in precipitation between the "drier" and "wetter" intervals of time.

Tectonics

Tectonics potentially influences sequence stratigraphy by uplift of source areas, which increases siliciclastic sediment yield to complementary basins, and by the rate and symmetry of basin subsidence. By Early Permian time, uplift of the Ancestral Rocky Mountains had largely ceased, and those mountains that fed sediment to the coastal plain of southern New Mexico were being eroded and onlapped by upper Wolfcampian and Leonardian sediment (Mack and Dinterman 2002; Kues and Giles 2004). Evidence exists, however, for differential subsidence of the Orogrande Basin during Wolfcampian time. In south-central New Mexico, red beds of the Abo Formation thicken toward the east-central part of the basin (Kottlowski 1963; Mack et al. 2003b), whereas, more pertinent to this study, Kottlowski (1960, 1963), Seager et al. (1976), and Seager (1981) noted an east-southeastward increase in thickness and relative abundance of shale in uppermost Pennsylvanian and Lower Permian strata between the Robledo, Doña Ana, and Organ Mountains (Fig. 2). They attributed this to local downwarp between their shallow "Robledo shelf" and the deeper floor of the Orogrande Basin, such that the latter subsided at a greater rate and received a greater abundance of prodelta muds (Fig. 2). This idea implies the existence of a break in slope between the Robledo shelf and the deeper part of the basin, which, in the model of Leeder and Stewart (1996), is necessary for rivers to cut incised valleys during sea-level lowstand.

The presence of the "slope-forming interval" in the Robledo Mountains and its absence in the Doña Ana Mountains may require a tectonic explanation (Fig. 2). Lacking evidence of fourth-order sequence boundaries, the "slope-forming interval" in the Robledo Mountains has the most abundant, diverse invertebrate fauna in the Abo member, reflecting the most favorable conditions for benthic marine life (Kues 1995). The comparable stratigraphic interval in the Doña Ana Mountains has four fourth-order sequences whose marine component is composed of restricted-marine limestones and lagoonal-to-peritidal parasequences, suggesting overall shallower conditions than for the Robledo "slope-forming interval." This difference is difficult to explain if glacial-eustasy was the only force affecting sea level during deposition of the Abo member.

Lawton et al. (2002) recently presented a tectonic model that may be pertinent to the origin of the "slope-forming interval." They documented episodic uplift and erosion along the western margin of the Orogrande Basin during Early Permian time, based on unconformities and associated locally derived conglomerates in the Caballo Mountains, located approximately 50 km northwest of the Doña Ana Mountains. Using these data, along with a nearly symmetrical isopach pattern for Virgilian strata, they argued that the Orogrande Basin may have been a rhomb basin related to a component of right-lateral shear along the eastern margin of the Pedernal uplift. They also implied that the western margin of the basin may have undergone pulses of differential subsidence on a wrench fault or fault system. If this fault/fault system were situated between the modern Robledo and Doña Ana Mountains, then down-to-the-west movement could have produced slightly deeper, more normal marine conditions at the site of the modern Robledo Mountains during deposition of the "slope-forming interval," while the area of the modern Doña Ana Mountains was shallower and more responsive to small changes in sea level. This model remains speculative, however, because an Early Permian fault or fault system between the Robledo and Doña Ana Mountains has not yet been identified. Although the origin of the "slope-forming interval" in the Robledo Mountains may be unresolved, it illustrates the importance of considering variables other than global eustasy in sequence stratigraphy.

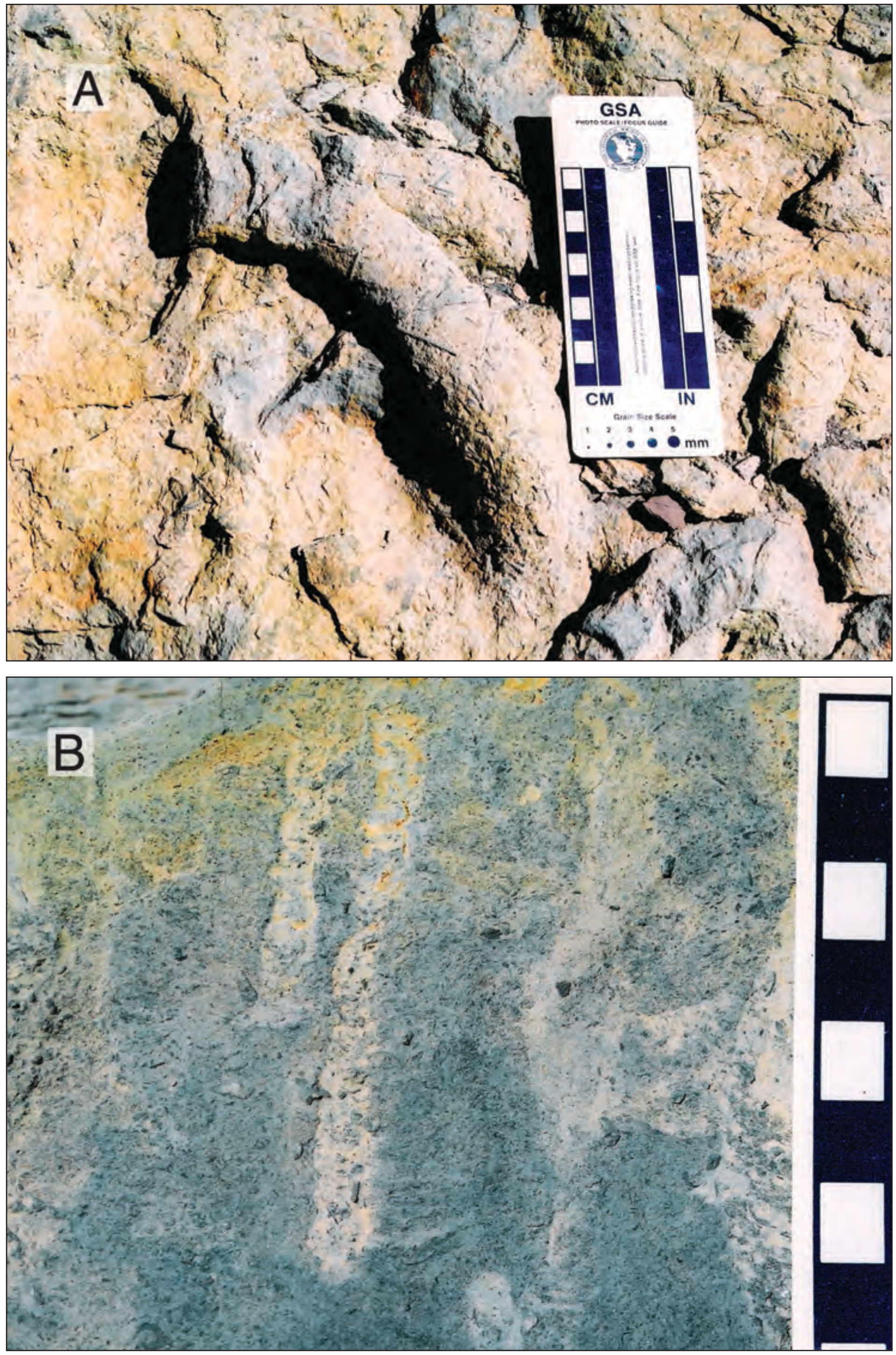


FIGURE 10—*Glossifungites* ichnofacies in marine limestones. **A:** *Thalassinoides* burrows on the underside of limestone bed. **B:** *Diplocraterion* burrows extending downward from the top of limestone bed. Scale is in centimeters.

Acknowledgments

My thoughts on the depositional environments and sequence stratigraphy of the Abo member have evolved over the past 20 years because of discussions in the field with Calvin James, Mike Leeder, Marta Perez-Arlucea, William Seager, Brendon Bailey, Chris Maples, and Neil Tabor. The correlations and interpretations presented here, however, are my own. Maya Elrick and Greg Wahlman read an earlier version of this manuscript and made helpful suggestions for its improvement.

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