Meteorite water/rock ratios and the significance of sequence and parasequence boundaries in the Gobbler Formation (Middle Pennsylvanian) of south-central New Mexico

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ABSTRACT

The upper carbonate member of the Middle Pennsylvanian Gobbler Formation of south-central New Mexico is composed of a succession of 3–20-m-thick (10–66 ft) parasequences exhibiting petrographic and isotopic evidence of subaerial exposure and early meteorite diagenesis at parasequence tops. Groups of four to six parasequences are arranged in 30–70-m-thick (100–230 ft) sequences that grade upward from nearly basal members into gravely capping members exhibiting brecciation, karst solution features, and laminar and oxidized crusts. About 2%–3% enrichment of micrite δ13C values downward from individual parasequence tops reflects a decrease in average meteorite water/rock ratios from about 800 at parasequence tops to about 200 at a depth of 10 m, reflecting reduced fluid flux and/or evolution of fluid isotopic compositions downstream toward primary marine-carbonate values (+4.0‰ ± 0.5‰ δ13C). Depletion of micrite δ13C values is somewhat stronger with respect to sequence boundaries than to parasequence boundaries in the Mockingbird Gap Hills section, indicating that the former represent surface of longer or more intense meteorite diagenesis than the latter.

INTRODUCTION

Despite extensive research on the internal architecture of sequences and parasequences (e.g., Vail, 1987; Posamentier et al., 1988; Sarg, 1988; Van Wagoner et al., 1988, 1990), the relation of diagenesis to sequence stratigraphic boundaries has received comparatively little attention (e.g., Driese et al., 1992). The Middle Pennsylvanian Gobbler Formation of south-central New Mexico offers an excellent case study for such genetic relations owing to an abundance of sequence boundaries (SBs) and parasequence boundaries (PSBs) that are readily identifiable and that exhibit evidence of strong early meteorite diagenesis associated with subaerial exposure. Stable carbon isotopes are used to quantify meteoritic water/rock (W/R) ratios within this cyclic carbonate succession and to investigate the significance of SBs and PSBs. The goals of this contribution are to (1) describe the sequence stratigraphy of the Gobbler Formation, (2) document patterns of stable carbon isotopic variations with respect to SBs and PSBs, (3) develop a mass-balance model to estimate meteoritic W/R ratios, and (4) evaluate the relative intensity of meteorite diagenesis in relation to SBs and PSBs.

PALEOGEOGRAPHY AND PALEOClimate

The Gobbler Formation of the Sacramento Mountains and stratigraphic equivalents in the San Andres and Hueco mountains of south-central New Mexico and west Texas (hereafter collectively termed "Gobbler Formation" for brevity) were measured and sampled at meter-scale intervals at eight localities in southern New Mexico and west Texas (Fig. 1; Algeo et al., 1991, 1992). The formation constitutes a 250–450-m-thick
(820–1,480 ft) succession of shallow-marine sediments of Derryn–Dresnoeventian age that is divisible into a lower 75–125 m–thick (250–410 ft) siliciclastic member composed of cross-bedded sandstones, gravelly limestones, and shales, and an upper 175–325 m–thick (570–1,070 ft) carbonate member composed of bioturbated micritic limestones and marls ("Bug Scatcll Member") of PhD. 1961). These units, which overlie a Cardishauan–Maltashlian passive-margin succession, represent the first deposits within the Orogodan Basin, an intracratonic trough that actively subsided during the Pennsylvania Period owing to crustal stresses associated with the Ouachita–Marathon orogeny (Rattow, 1963; Goetze and Dickerson, 1985; Ross and Ross, 1985a). The basin was markedly asymmetric, having a steep, fault–bounded eastern margin (Serramento Shelf) and a gently inclined western margin (Robledo Ramp; Algo, 1992). The Orogodan Basin was located at low paleolatitudes, within about 5° of the equator, during the Pennsylvania Period (Ontang and Scotson, 1988). A warm, humid paleoclimate is indicated by the absence of arid–zone features (e.g., dolomite or evaporite, deilaciation fabrics, or caliche structures such as nodules, pisoids, and terreni) and by strong meteoric diagnostica of subaerially exposed surfaces in the Gobbl Formation and in overlying units of Late Pennsylvania age (Goldstein, 1991). Coeval strata in the Paradox, central Colorado, and northern Denver basins to the north, at 10°–20° N paleolatitude, contain extensive evaporite deposits, reflecting their location within an arid climatic belt (Wilson, 1975). However, the Orogodan Basin area was not as humid during the Middle Pennsylvanian as the eastern and mid-continent regions of North America, where moisture-laden trad winds resulted in development of extensive coal swamps (Frikes, 1979).

SEQUENCE STRATIGRAPHY

The carbonate upper member of the Gobbl Formation consists of a succession of 3–20 m–thick (10–66 ft) marl–limestone cycles that are hierarchically arranged in parasequences and sequences (Fig. 1). Individual parasequences are composed of conformable groups of beds and beds (terminology of Sarg, 1988; Van Wagben and others, 1990), each of which is relatively homogeneous internally and is separated from adjacent beds and bedsets of similar lithology by distinct bedding surfaces (Fig. 2). Parasequences exhibit strong asymmetry, being composed of generally thin (<1–2 m [3.3–6.6 ft]) recessive beds and thick 15–20 m [10–66 ft] resistant caps. Recessive basal units mainly consist of nodular marls, locally containing abundant sponge spicules. Resistant caps are largely composed of fossil wackestone and packstone containing an abundant, diverse open–marine biota, including echinoderms, bryozoans, brachiopods, phylloid algae, fusulinids, endothyrs, tubular foraminifers, rugose and tabulate corals. Komia is a small dendroid stromatoporoid, and Chieata (a sponge–like organism of uncertain affinity, Wilson, 1975). In more proximal areas, some parasequences are capped by cross–bedded fossiliferous or calcilutitic grainstone. Each parasequence is interpreted to represent a single transgressive–regressive sea–level cycle, in which the maximum flooding surface (MFS) is located within the recessive basal unit and the overlying resistant beds and bedsets represent deposition under conditions of stable or falling sea level. The tops of individual parasequences are marked by sharp, disconformable surfaces exhibiting brecciated rubble (generally of the same lithology as the underlying rock and of paradocho– nous origin), black pebble conglomerates, karst solution pits and rills (rinkeken), and sandy laminar or ferric oxidized crusts (Fig. 2; Algo et al., 1991). Directly overlying many cycle tops are thin (1–10 cm [0.4–4 in]) layers of coarse–grained, echinoderm–rich fossil debris containing abundant phosphatic grains, glauconite pellets, and quartz sand. Irregular small–scale relief (<1 m [3.3 ft]) is developed along the upper surface of some parasequences. Collectively, these features document subaerial exposure of parasequence tops and deposition of winnowed logs during subsequent transgressions, suggesting that these surfaces are type I unconformities (Sarg, 1988). Sequences consist of groups of four to six parasequences and range in thickness from 30 to 70 m (100–230 ft) (Fig. 3). They exhibit moderate to strong asymmetry owing to the decreased shelf content and increased weathering resistance of component parasequences at stratigraphically higher levels. Within each sequence, parasequences become coarser grained and cleaner upward, developing caps composed of skeletal packstone or grainstone. This asymmetry permits ready field recognition of parasequence–successive hierarchies at most study locales.
Figure 2. Lithologies and diagenetic features of Gobblie Formation parasequences. Parasequence tops exhibit a variety of features associated with subaerial exposure and marine diagenesis. (A) karst solution pit (sp), lined by sandy limestone crust (lc) and infilled by silicified mudstone lag. (B) coarse-grained fossil lag overlying planar oxidation effect (ox) at parasequence boundary; and (C) rillenkarren (r) developed on cycle-top exposure surface (sf) and associated by later silification. Major lithologies include: (D) fossil wackestones with abundant echinoderms (e), fusulinids (f), phylloid algae (a), and gastropods (g); and (E) argillaceous wackestones containing abundant sponge spicules (sp).
Figure 3. Stratigraphic columns and micritic carbon isotope stratigraphy of the Fresnal Canyon and Mockingbird Gap Hills sections. Sequences are numbered 1–5. SB = sequence boundary, PSB = parasequence boundary, c = covered interval. Note the relation between micritic δ¹³C depletion and sequence and parasequence boundaries. The base of the Fresnal Canyon section is incomplete owing to non-exposure.
The carbon isotopic compositions of Gobbler Formation micrites range from +5.0% to +5.8% and exhibit high-frequency variation closely associated with parasequence boundaries (Fig. 3). The uppermost few meters of many parasequences are characterized by large negative δ13C excursions with progressively heavier δ13C values stratigraphically downward within the unit. Generally, a sharp transition exists between 13C-depleted values at the top of one parasequence and 13C-enriched values at the base of the next parasequence. Few 13C-depleted values occur more than 1–2 m (3.3–6.6 ft) below PSBs (Fig. 4).

The association of 13C-depleted cements with subsahelian exposure surfaces is common in both ancient and modern carbonates (e.g., Alabany and Matthews, 1982; Brown, 1982; Beier, 1987; Budd and Land, 1990; Goedeke, 1991; Hallet and Moore, 1991). Depleted carbon isotopic signatures result from early metazoic diagenesis of the exposed limestone surfaces. Meteoric fluid laden with 13C derived from soil-zones organics fluxes downward and mixes with isotopically heavy carbonate carbon derived from host-rock dissolution. Owing to a progressive increase in fluid-rock interaction downward, meteoric cements precipitated at depth have a lower proportion of fluid-derived carbon than those precipitated close to an exposure surface. The absence of negative carbon isotopic excursions along some Gobbler Formation parasequences suggests that fluid flow by local rock-matrix heterogeneities (Algeo et al., 1992). Petrographic evidence of subsahelian exposure (e.g., Fig. 2) is generally strongest for those parasequence caps exhibiting large negative δ13C excursions, consistent with variation in fluid flux and in the degree of metazoic alteration from one parasequence to the next.

Gobbler Formation micrites have oxygen isotopic compositions ranging from less than −1.5% to −2.9% (Fig. 4). Each section exhibits a characteristic mean δ18O value, ranging from a low of −10.5% at Hembritto Canyon to a high of +4.3% at Fruenall Canyon (at −6.9%; Mockingbird Gap Hills is intermediate). Strong correlation between section-averaged δ18O values and maximum burial depths at a given locale indicate dominant control of oxygen isotopic compositions by burial temperature (Algeo et al., 1992). Micrite δ18O values exhibit no relation to PSBs or SBs. Despite significant shifts in micrite δ18O values during burial diagenesis, carbon isotopic values were largely unaffected owing to low fluid C concentrations and low W/R ratios in the burial diagenetic environment (Algeo et al., 1992). Burial diagenesis resulted in slight homogenization of carbon isotopes (i.e., minor reduction of δ18O range of micrite δ18O values) with progressive 13C depletion (Fig. 4).

In order to evaluate the degree of meteoric alteration of Gobbler Formation micrites, it is necessary to establish a primary marine-carbonate δ13C baseline. First, because most diagenetic processes result in depletion of δ13C and δ18O in secondary precipitates relative to primary marine carbonates (e.g., Meyers and Lohmann, 1985), the isotopically heaviest micrite compositions may approximate such a baseline. The most-enriched isotopic values exhibited by Gobbler Formation micrites are +5.0% δ13C.
The isotopic composition of any diagenetic phase is a function of initial pore-fluid and host-rock isotopic compositions, the relative molar contribution of each source, and the fractionation factor between a precipitate and coexisting fluid at equilibrium. The relative molar contributions of the fluid and rock sources determine the molar W/R ratio, which can be related to volumetric W/R ratios via the relative concentration of the element of interest in the fluid and rock sources. Water-rock ratios calculated either by either basis record only the amount of pore fluid which has moved through and reacted with the host rock. The carbon isotopic composition of a meteoric diagenetic precipitate can be related to W/R ratio through a simple mass balance model, in which the isotopic composition of the precipitate equals the mole-weight average of the isotopic compositions of the pore fluid and host rock:

$$\delta_p = \left( \frac{M_p}{M_R} \times \delta_{pR} \right) + \left( \frac{M_R}{M_R} \times \delta_{pM} \right)$$

where $\delta_p = 8^{13}C$ of diagenetic precipitate, $\delta_{pR} = 8^{13}C$ of pore fluid, $\delta_{pM} = 8^{13}C$ of host carbonate, $M_p = $ molar contribution of carbon from pore fluid, $M_R = $ molar contribution of carbon from host carbonate, and $C_{CaCO_3}/C_{HCO_3^-}$ = carbon isotopic fractionation between aqueous bicarbonate and meteoric calcite expressed in per-mil variation.

Carbon isotopic values of meteoric precipitates reflect the degree of openness of the diagenetic system through $C_{CaCO_3}/C_{HCO_3^-}$ in an open system (i.e., with unlimited fluid supply), $M_p$ is much larger than $M_R$, and maximum calcite-bicarbonate fractionation occurs ($\pm 0.2%$; Romanek et al., 1992). In a closed system (i.e., with minimal fluid and no replenishment), $M_p$ is large in relation to $M_R$, and calcite-bicarbonate fractionation is reduced owing to Rayleigh fractionation of carbon isotopes within the limited volume of available fluid. Equation 1 may be reorganized to solve directly for molar W/R ratio:

$$M_p/M_R = \left( \frac{\delta_{pR} - \delta_{pM}}{\delta_{pR} - \delta_{pC}} \right)$$

(2)

Molar W/R ratio can be converted to volumetric W/R ratio by a constant $\kappa$:

$$V_p/V_R = [M_p/V_R] / [M_p/V_R]$$

(3)

where $V_p/V_R$ is the volumetric W/R ratio. Reorganizing to solve for $\kappa$ yields:

$$\kappa = [M_p/V_R] / [M_p/V_R]$$

(4)

where $M_p/V_R = $ moles of carbon per unit volume of rock, and $M_p/V_R = $ moles of carbon per unit volume of fluid, and

$$M_p = 106C_{CaCO_3}/C_{HCO_3^-}$$

(5)

where $C_{CaCO_3}$ = concentration of carbon in calcite by weight (12 g C/100 g CaCO_3), $C_{HCO_3^-}$ = density of calcite (2.7 g/cc), $\mu_C$ = molecular weight of carbon (12 g/mol), and $10^6$ is a unit-conversion factor. Given these values, $\kappa = 2.7 \times 10^3$ (equation 4), where $TIC$ = concentration of total inorganic carbon in the fluid in mmol. Rearranging equations 2 and 3 and substituting for $\kappa$ yields:

$$V_p/V_R = 2.7 \times 10^3 [\delta_{pR} - \delta_{pC}] / [\delta_{pR} - \delta_{pM}]$$

(6)

All subsequent discussion of molar W/R ratios will be on a volumetric rather than a molar basis.

**Initial parameters for meteoric water/rock ratio model**

In order to calculate model meteoric W/R ratios based on equation 6, values are required for four parameters: (1) meteoric precipitate $8^{13}C$, (2) primary marine carbonate $8^{13}C$, (3) initial meteoric fluid $8^{13}C$, and (4) fluid total inorganic carbon (TIC) concentration. The first two variables can be constrained by direct isotopic analysis of carbonate components within the Gobbler Formation. The carbon isotopic compositions of Gobbler Formation micrites were set within the meteoric environment and altered little within the burial environment (Algeo et al., 1992); thus, meteoric precipitate $8^{13}C$ is measured directly from matrix micrites (Fig. 3). As discussed above, the carbon isotopic composition of primary marine carbonate can be estimated from isotopically heavy micrites and well-preserved aragonitic brachiopods, and is $\pm 4.0\%$ to $0.5\%$ (Fig. 4). The $8^{13}C$ value and TIC concentration of the initial meteoric fluid can be estimated based on the composition of soil solutions in comparable modern climate zones. The immediate source of soil carbon is terrestrial vegetation, and both modern and ancient land plants using the C3 photosynthetic pathway exhibit limited $8^{13}C$ variation (ca. $-25\%$ to $10\%$; Degens, 1969; Doorninck, 1980; Popp et al., 1919). Although plants using the C4 and CAM photosynthetic pathways are isotopically enriched by comparison (ca. $-6\%$ to $19\%$ and $12\%$ to $23\%$, respect
tively), they probably did not exist in the pre-Cenozoic (Cerling et al., 1993; Morgan et al., 1994). Thus, Pennsylvanian soil organic carbon is likely to have been sourced entirely from C3 vegetation.

In addition to vegetation type, the $\delta^{13}$C values of soils of CO$_2$ can be influenced by such factors as atmospheric CO$_2$, soil respiration rate, porosity, and temperature (Cerling, 1991). Although atmospheric pCO$_2$ during the late Paleozoic is not known with any degree of certainty, values were probably relatively low (ca. 0.5-1.0 PAL; Betzer, 1991, 1994) as a result of rapid burial of organic carbon during Late Devonian-Early Carboniferous. At low values of atmospheric pCO$_2$, little, if any, atmospheric CO$_2$ ($\delta^{13}$C = -6.5%) diffuses downward into soils, and soil respiration rate and porosity have little effect on the isotopic composition of soil gas (Cerling, 1991). Temperature is of moderate importance in determining the carbon isotopic composition of soil solutions owing to fractionation between soil gas and dissolved bicarbonate (Cerling, 1991). For temperatures in the range of 10-40°C, HCO$_3$ (-ap-CO$_2$) fractionation is approximately:

$$
\frac{\delta^{13}C_{\text{HCO}_3}}{\delta^{13}C_{\text{CO}_2}} = 1.15(0.36 - 0.12t(0.01)).
$$

(7)

where $T$ is temperature in °C (Romeine et al., 1992). In addition, a 4.4% enrichment of CO$_2$ relative to C$_{HCO_3}$ occurs in soils owing to faster diffusion of $^{12}$C versus $^{13}$C outward to the atmosphere. Thus, the carbon isotopic composition of dissolved bicarbonate in Pennsylvanian soil solutions was approximately:

$$
\delta^{13}C = -20.1 + 4.4(11t(0.5)) - 0.14t(0.01).
$$

(8)

where $\delta^{13}C$ is the $\delta^{13}$C value of dissolved soil bicarbonate. Equitorial temperatures are thought to have been relatively constant throughout the Phanerozoic despite large-scale global climate changes (Frakes, 1979). For modeling purposes, a mean annual temperature of 25 ± 5°C (a range consistent with an equatorial coastal setting) is assumed yielding HCO$_3$ (-ap-CO$_2$) fractionation of +8.0% ± 1.1% (equation 7) and a $\delta^{13}$C value for dissolved bicarbonate in humid, low-latitude Pennsylvanian soils of -13.5% ± 2% (equation 8).

The final model parameter, the initial TIC concentration of meteoric fluids, can be estimated based on the composition of modern soil solutions in climatically comparable environments. In modern temperate- to tropical-zone soils developed on calcareous hor rock ("brown forest soils"), bicarbonate forms in the uppermost (A) horizon through volatization and dissociation of oxidized organic carbon. Dissolved bicarbonate concentrations in such soils are consistently rather high, e.g., 700-1000 mg/l (Matthes, 1982), which is equivalent to a bicarbonate molar concentration of 12 ± 2 mmol/l. Bicarbonate represents by far the largest component of total inorganic carbon in such systems and is a good proxy for TIC.

In summary, the parameters used in calculating model meteoric W/R ratios for the Gobler Formation via equation 6 are (sources in parentheses): $\delta^{13}C$ = measured $\delta^{13}$C in % (Gobler Formation micrites); $\delta^{13}C_{\text{HCO}_3}$ = +4.0% ± 0.5% (Gobler Formation brachiopods); $\delta^{13}C_{\text{HCO}_3} = -13.5% ± 1%$ (calculated assuming C3 vegetation); $\delta^{13}C_{\text{HCO}_3}$ = -4.0% ± 0.2% (experimentally determined; Romeine et al., 1992); TIC = 12 ± 2 mmol/l (estimated based on modern temperate zone soil waters; Matthes, 1982).

Variability of meteoric waterrock ratios within parasequences

Within individual parasequences, micrites generally exhibit 13C enrichment downsection, indicating smaller contributions of isotopically light soil-derived carbon at depth (Fig. 5). Assuming that Gobler Formation micrites initially had an average isotopic composition equal to that of primary marine carbonate ($\delta^{13}$C = +4.0% ± 0.5%), their present degree of $\delta^{13}$C depletion provides a measure of the relative minor contribution of carbon from fluid and rock sources and, thus, of meteoric W/R ratio. Differences in the degree of $\delta^{13}$C depletion in successive parasequences caps probably reflect variation in meteoric fluid flux across these surfaces, although variable development of overlying soil horizons may have played a role. Differences in the rate of micrite $\delta^{13}$C enrichment downsection within parasequences reflect a combination of variable meteoric fluid volume and host-rock susceptibility to diagenetic reaction. Thus, although a number of factors contributed to change in micrite $\delta^{13}$C with depth below parasequence tops, there can be useful characterized as "meteoric W/R ratio," providing a quantitative measure for comparison of meteoric effects among parasequences.

Changes in meteoric W/R ratio downward within individual parasequences can be calculated based on the isotopic mass balance model developed above and documented patterns of downsection enrichment of micrite carbon isotopes. The latter parameter was determined as a weighted running average of micrite $\delta^{13}$C values with respect to depth below PSB’s. At Mockingbird Gap Hills, micrite $\delta^{13}$C values increase on average from -0.4% at parasequence boundaries to +2.4% at a depth of 10 m (33 ft) (Fig. 5). This rate of $\delta^{13}$C enrichment is compatible in magnitude to that reported for several Quaternary limestone units (e.g., Bone, 1987; Budd and Land, 1990). Average meteoric W/R ratios calculated from this trend range from about 800 at parasequence tops to about 200 at 10 m (33 ft) depth (Fig. 5, equation 6). Compounded uncertainties in the various model parameters permit parasequence-cap W/R ratios potentially to be as high as 1500 or as low as 458, and those at 10 m (33 ft) depth to be as high as 500 or as low as 80. All combinations of model values yield relatively high W/R ratios, indicating that meteoric diagenesis proceeded within a largely open system.

Meteoric W/R ratios decrease downsection over a 10 m (33 ft) interval by a factor of three to six (Fig. 5). Such downsection decrease is a result of two processes operating within Gobler Formation meteoric systems. First, lateral diversion of
Figure 5. Model meteoric water-rock ratios with respect to parasequence boundaries at Mockingbird Gap Hills. The model predicts depth-dependent marine δ¹⁸O values based on a weighted mean average (solid line, bottom), marine carbonate δ¹⁸O = 0.0 ± 0.5%, meteoric fluid δ¹⁸O = -13.5 ± 2%, δ¹⁸O_water = +1.0 ± 0.2%, and fluid TIC (total inorganic carbon) concentration = 12 ± 2 mmol/L. Resulting volumetric estimates of meteoric WR ratios range from 800 at a depth of 10 m (solid line, top). Uncertainty limits for meteoric WR ratio estimates (dashed lines, top) were calculated as compounded uncertainty ranges of all model parameters, including ±1 standard mean error for the weighted δ¹⁸O causing average (dashed lines, bottom). A similar range of meteoric WR ratios was calculated for Formal Crayon (not shown).

Figure 3. Facies stacking patterns of this type are commonly simulated using a sea-level forcing function modeled by two sine waves of differing periods and amplitudes (e.g., Goldhammer et al., 1991; Read et al., 1991). If formation of sequences and parasequences reflects such a forcing mechanism, this should be manifested in systematic variation in the degree of exposure of PSBs within sequences. In this scenario, high-frequency sea-level falls exhibit reduced amplitude during intervals of rising long-term sea level, but increased amplitude during intervals of falling long-term sea level (e.g., Gibling and Read, 1991, fig. 15). Thus, parasequence tops at the base of a sequence should undergo shorter intervals of exposure than those at the top of a sequence and, hence, should exhibit less intense meteoric alteration.

Significance of sequence versus parasequence boundaries

Within most Goolbler Formation sequences, the lowestmost parasequence has a thick marly base and stratigraphically higher parasequences exhibit increased grausiness and reduced marl content, indicating progressively shallower water conditions.
undergo similar degrees of meteoric alteration and isotopic depletion (e.g., dashed line, Fig. 6A). In this case, micrite δ¹⁸O values should exhibit a strong relation to PSBs (dashed line, Fig. 6B) but only a weak relation to SBSs (dashed line, Fig. 6C). Conversely, if the rate of sea-level change of long-term cycles is large relative to that of short-term cycles (A/ΔA ≥ A/ΔA), then SBSs should undergo more intense meteoric alteration than PSBs (e.g., solid lines, Fig. 6A), and micrite δ¹⁸O values should exhibit a stronger relation to the former (solid line, Fig. 6C) than to the latter (solid line, Fig. 6B).

The strength of these relations can be evaluated based on correlation coefficients (r²), which represent the amount of variance accounted for through linear regression of a diagram. If the r² value for micrite δ¹⁸O versus depth below SBSs is smaller than that versus depth below PSBs (e.g., Fig. 6B), then the former surfaces are equivalent to the latter with regard to intensity of meteoric diagenesis. Conversely, if the r² value for SBSs is larger than that for PSBs (e.g., Fig. 6C), then micrite δ¹⁸O depletion is mainly controlled by diagenesis at SBs, and these surfaces are likely to have undergone stronger or more intense exposure than the "average" PSB. A correlation coefficient approaching 1.0 would indicate strong control of meteoric carbonate δ¹⁸O values by PSBs or SBSs, whereas a value approaching 0 would indicate no control by the respective boundary type. Weak correlation of meteoric carbonate δ¹⁸O values with SBs or PSBs may reflect an incidental result arising from strong controls by the other boundary type. For example, if diagenetic alteration were associated exclusively with SBSs (r² = 1.0), δ¹⁸O values versus depth below PSBs would nonetheless yield r² equal to 1/2 (where n is the number of parasequences per sequence; in Fig. 6, n = 3 and r² = 0.11) for the Gobblert Formation, n = 4-6 and r² = 0.03-0.6. Gobblert Formation micrites exhibit ¹⁴C depletion with respect to both PSBs and SBs (Fig. 7). In the Fresnel Canyon section, micrite ¹⁴C depletion trends are weak in relation to both types of boundaries (r² = 0.07-0.09). Because many exposure surfaces in the Fresnel Canyon section exhibit ¹⁴C depletion (Fig. 3), low r² values appear to reflect a high degree of isotopic scatter about the mean depth trend (Fig. 7, top). This suggests that patterns of meteoric fluid flow in the Fresnel Canyon section were affected by large heterogeneities in matrix permeability, or possibly the effects of multiple diagenetic overprints. In contrast, the depth dependency of micrite δ¹⁸O values is greater for both boundary types in the Mockingbird Gap Hills section, in which micrite ¹⁴C depletion correlates more strongly with SBSs (r² = 0.25) than PSBs (r² = 0.15; Fig. 7, bottom). These results indicate a more intense control of meteoric diagenesis by exposure surfaces and longer or more intense alteration of SBs than PSBs. Differences in ¹⁴C depletion trends between the two study localities may reflect regionally variable isotopic and sedimentation patterns within the Orogrende Basin. The Mockingbird Gap Hills section was located on the northwestern margin of the Orogrende Basin, yielding regular facies stacking patterns (Fig. 3) and control of meteoric diagenesis by quasiperiodic sea-level changes. Conversely, the Fresnel Canyon section was located on the tectonically active eastern basin margin (Sacramento Shelf), resulting in less regular facies stacking patterns and, probably, more heterogeneous matrix permeabilities that influenced vertical and lateral fluid migration within meteoric lenses. Facies control of meteoric fluid flow, rather than vertical motions of the shelf per se, are more likely to have produced the weak relation of micrite δ¹⁸O values to SBs and PSBs in the Fresnel Canyon section (Fig. 7).

CONCLUSIONS
3. The upper carbonate member of the Middle Pennsylvanian Gobblert Formation of south-central New Mexico is composed of 3-20-m-thick (10-66 ft) parasequences capped by
sharp, disconformable surfaces exhibiting petrographic and isotopic evidence of subaerial exposure and meteoric diagenesis. 2. Groups of four to six parasequences comprise individual 30-70-m-thick (100-230 ft) sequences. Within each sequence, individual parasequences become less shaly and coarser-grained upsection, and parasequence caps tend to exhibit stronger development of subaerial exposure features and greater 13C depletion of meteoric carbonates.

3. Gobbler Formation parasequences are wide; correlatable (>100 km [62 mi]) laterally across the Oquirrh Basin, show evidence of early meteoric diagenesis in both shif and basal areas, exhibit strong modal thicknesses at each locale, and represent an average depositional period of ca. 120-180 ka. The most likely mechanism to produce such regular, large-amplitude, high-frequency changes of relative sea level over substantial lateral distances is glacio-eustasy.

4. Well-preserved articulate brachiopods and isotopically heavy micrites indicate a primary marine-carbonate isotopic composition of +4.0‰ ± 0.5‰, δ13C and −3.5‰ ± 0.2‰ δ18O for the Gobbler Formation.

5. Isotopic patterns document two stages of diagenesis within the Gobbler Formation: an early meteoric phase associated with episodic subaerial exposure that largely controlled micrite δ13C values, and a later burial phase that reset micrite δ18O values without substantially altering carbon isotopic ratios.

6. Micrite δ13C values range from +5.0‰ to −5.8‰ and exhibit high-frequency variation closely associated with exposure surfaces, in which the uppermost few meters of many parasequences are characterized by large negative δ13C excursions.

7. Meteoric precipitates record mixing of carbon from isotopically heavy marine carbonate (δ13C = +4.0‰ ± 0.5‰) and isotopically light, soil-derived bicarbonate (δ13C = −13.5‰ ± 2‰) sources.

8. Mass-balance modeling indicates that meteoric diagenesis took place within a largely open system, in which meteoric W/R ratios decreased from about 800 at parasequence tops (uncertainty limits 450-1500) to about 200 at abd (33 ft) depth (uncertainty limits 80-500). A three- to six-fold decrease in W/R ratio at 10 m (33 ft) intervals downstream indicates strong evolution of fluid isotopic compositions and/or substantial lateral flow owing to vertical permeability barriers.

9. In the Mockingbird Gap Hills section, micrite 13C depletion is stronger with respect to sequence boundaries than parasequence boundaries (r2 = 0.25 versus 0.18), implying that the former represent exposure surfaces of greater duration and/or diagenetic intensity than the latter. Sequence and parasequence