MAP I-2555

U.S. DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY

CHART SHOWING LITHOLOGY, MINERALOGY, AND PALEONTOLOGY OF THE PENNSYLVANIAN HERMOSA GROUP AT HERMOSA MOUNTAIN, LA PLATA COUNTY, COLORADO

By Karen J. Franczyk, Germaine Clark, Douglas C. Brew, and Janet K. Pitman

MISCELLANEOUS INVESTIGATIONS SERIES
Published by the U.S. Geological Survey, 1995

CHART SHOWING LITHOLOGY, MINERALOGY, AND PALEONTOLOGY OF THE PENNSYLVANIAN HERMOSA GROUP AT HERMOSA MOUNTAIN, LA PLATA COUNTY, COLORADO

By
Karen J. Franczyk, Germaine Clark, Douglas C. Brew, and Janet K. Pitman

INTRODUCTION

A detailed study of the Pennsylvanian Hermosa Group was undertaken at Hermosa mountain (about 19 km (12 mi) north of Durango, Colo.) on the west side of the Animas River Valley as part of an investigation of the Pennsylvanian System in the Paradox Basin. This investigation was part of the Paradox Basin Project of the U.S. Geological Survey. The objectives of the Hermosa Group study were to (1) establish better correlations between Pennsylvanian strata along the eastern basin margin and the central basin area, and between the Hermosa mountain outcrops and those to the north in the San Juan Mountains area and (2) achieve a greater understanding of the depositional processes operative during the Pennsylvanian along the active margin of the Paradox Basin. To accomplish these objectives, a section of the Pennsylvanian Hermosa Group at Hermosa mountain (fig. 1) was measured in detail and the sedimentology, clastic and carbonate petrology, and macrofossil assemblages were interpreted. Franczyk measured the section and interpreted the clastic sedimentology; Clark analyzed the carbonate samples and interpreted the carbonate sedimentology; Brew collected the macrofossils and analyzed the assemblages; and Pitman analyzed the clastic samples.

The exposures at Hermosa mountain are the southernmost exposures of Pennsylvanian strata on the eastern side of the Paradox Basin. To the west, Pennsylvanian strata dip into the subsurface and only crop out again in the canyon of the San Juan River near Mexican Hat, Utah, about 160 km (100 mi) west of Hermosa mountain. The Hermosa Group undergoes abrupt and dramatic facies changes across the Paradox Basin. Along the eastern margin of the basin, near the Uncompahgre uplift, interbedded clastic and carbonate lithologies compose the Hermosa Group. These lithologies grade rapidly westward into interbedded evaporite, carbonate, and minor clastic units that characterize the Hermosa of the central part of the basin. From the central to the southwestern part of the basin, the Hermosa Group grades into shelf-carbonate deposits that have locally abundant algal-mound buildups and minor clastic units.

The Pennsylvanian outcrops from Hermosa mountain north to Silverton, Colo. show lithofacies characteristic of the eastern basin margin; whereas age-equivalent outcrops in the canyons of the San Juan River show lithofacies characteristic of the southwestern shelf area. By using the abundant subsurface geophysical data obtained from numerous oil fields in the southern part of the Paradox Basin, good correlations of Pennsylvanian strata between the southwestern shelf lithofacies and the evaporite-bearing basin-center lithofacies have been established (Wengerd and Matheny, 1958; Hite, 1960; Irwin, 1976; Hite and Buckner, 1981). Correlation of the eastern basin-margin lithofacies to the basin center lithofacies is less well established because of the paucity of subsurface data in the eastern part of the basin. This detailed study of the Hermosa mountain section shows how the predominantly glacioeustatic-controlled cycles that are recognized in Pennsylvanian strata throughout the rest of the Paradox Basin are manifested along the southeastern margin of the basin.

ACKNOWLEDGMENTS

Tracy Whitesell assisted in measuring this section. Her capable and good-natured assistance in difficult terrain is greatly appreciated. Steve Condon and Gary Gianniny carefully and thoughtfully reviewed this document.

METHODS

The Hermosa mountain measured section is comprised of six segments (A-F) located in secs. 24, 25, 26, 35, and 36, T. 37 N., R. 9 W. (La Plata County, Colo.), in the Hermosa 7.5 minute quadrangle (fig. 2). Hermosa mountain (not designated on the quadrangle map) is the local name of the cliff-forming highlands due north of the small town of Hermosa, Colo. Vegetated to densely vegetated slopes characterize most of Hermosa mountain and surrounding areas. The location of each segment of the measured section was chosen on the basis of the quality of exposure and the presence of a bed which could be traced to the next segment. With the exception of segment B, all segments are located in gullies, which generally have the most complete exposures. The abundance of vegetation and slope cover prohibits tracing most units laterally.

Most of the section was measured with a Jacobs staff and Abney level; where the units are thick and cliff forming a tape was dropped from the top of the cliff. Carbonate units and coarser grained sandstone units thicker than 1 m (3 ft) generally form resistant outcrops, whereas claystone, siltstone, and fine-grained sandstone units generally weather to a slope. These slopes were trenched to expose the underlying units. Field descriptions were recorded at a scale of 2.5 cm to 3 m (1 in. to 10 ft). These descriptions were simplified to the 2.5 cm to 30 cm (1 in. to 100 ft) scale in this report. This measured section, at a scale of 2.5 cm to 6 m (1 in. to 20 ft), is included in Franczyk (1992), which shows more detail on lithologic variations, contact types, sedimentary structures, and accessories than is shown in this version.

Samples of the sandstone units were collected at regularly spaced intervals for thin-section analysis to document changes in mineralogy. Thin sections were impregnated with blue epoxy to identify pore space and stained to facilitate mineral identification. Staining with Alizarin Red-S and potassium ferricyanide permitted iron-free carbonate minerals to be distinguished optically from iron-bearing carbonate minerals. Sodium cobaltinitrite stain aided in the identification of potassium feldspar. Mineral abundances were estimated visually using percentage charts of Terry and Chilingar (1955). Sorting and roundness values were determined using visual comparison charts of Pettijohn and others (1972) and Powers (1953), respectively. Grain size was determined by measuring the long dimension of 10 randomly selected quartz grains in each sample.

A sample of almost every carbonate unit in the Hermosa mountain section was collected for slabbing and thin section analysis. Each thin section was stained with Alizarin Red-S and potassium ferricyanide to distinguish calcite, dolomite, ferroan calcite, and ferroan dolomite. The thin section analysis consisted of two parts: (1) a visual estimate of the mineralogy in which the volume estimate equals 100 percent, and (2) a visual estimate of abundance and type of allochem, matrix, and cement in which the total of all catagories equals 100 percent. The descriptive rock name, fabric, and inferred depositional environment for each carbonate sample are also shown on the chart.

Samples of dark-gray claystone, mudstone, and shale were collected throughout the section for vitrinite and palynomorph analysis. Vitrinite reflectance of five Hermosa Group samples gave R_0 values of 1.48 to 1.62, indicating the organic material was over mature with respect to hydrocarbon generation (V.F. Nuccio, written commun., 1993). Palynomorphs were sparse from these outcrop samples, and only long-ranging Pennsylvanian species were recovered (R.M. Kosanke, oral commun., 1992).

The profile on the left side of the lithology column shows the grain size for clastic units. The lithologic profile for carbonate units is arbitrarily extended to the medium-grained interval; the fabric for each carbonate unit is listed to the left of the lithology column. The interpreted depositional environments for clastic and carbonate units is listed to the left of the lithology column. A depositional environment interpretation for each carbonate sample is also listed in the carbonate section of the chart. On the basis of the interpreted environments, a relative sea level curve was constructed. The open-marine carbonate units and deep-water deltaic clastic units are considered to represent a high relative sea-level position and alluvial units a low position. Shallower deltaic and delta-plain deposits and tidal and supratidal carbonate deposits are transitional between the two end members. Abrupt changes from high or intermediate to low relative sea level may correspond to a period of base-level drop and subaerial exposure.

STRATIGRAPHY

For nearly a century, Pennsylvanian outcrops along the Animas River Valley have been studied. Spencer (1900, p. 48) first applied the name Hermosa Formation to these rocks. Although no type section was designated, Spencer derived the name from Hermosa Creek in the area where it enters the Animas River. Roth (1934) measured a section of the Hermosa, which he designated the type section, in secs. 26 and 35, T. 37 N., R. 9 W. (just north of the confluence of Hermosa Creek and the Animas River). Later, Bass (1944) illustrated a Hermosa measured section from the same area and correlated this section to other Pennsylvanian outcrop and subsurface sections in the Four Corners area. Wengerd and Strickland (1954) presented data from a section they measured at Hermosa mountain. Wengerd and Matheny (1958) presented a diagram of this section, and they proposed raising the Hermosa Formation to group rank and subdivided the group into, in ascending order, the Pinkerton Trail Formation, the Paradox Formation, and the Honaker Trail Formation. Because of the complex lithofacies changes of Middle Pennsylvanian and Lower Permian rocks across the Paradox Basin, this nomenclature has been hotly debated ever since. Workers such as Hite and Buckner (1981) and Loope and others (1990) have advocated the return to the older terminology of Bass (1944) who subdivided the Hermosa Formation into a lower member, the Paradox Member, and an upper member. The following section briefly discusses these problems and explains why in this report we have chosen to use Hermosa Group subdivided into the Pinkerton Trail Formation at the base and undifferentiated part of the Hermosa Group above for Middle Pennsylvanian rocks in the study area.

The Middle Pennsylvanian Pinkerton Trail Formation, which lies above the Early Pennsylvanian Molas Formation and below the Middle Pennsylvanian Paradox Formation, is nearly ubiquitous throughout the Paradox Basin. The Pinkerton Trail is a limestone-dominated unit that in parts of the basin exceeds 61 m (200 ft) (Wengerd and Strickland, 1954); depending on where the upper contact is placed, the thickness may exceed 122 m (400 ft) (Fetzner, 1960). Wengerd and Strickland (1954) first named and described the Pinkerton Trail from surface exposures in the general locality of segment A of our Hermosa mountain measured section (fig. 2); later Wengerd and Matheny (1958) included it in the Hermosa Group. At segment A, an 8 m (25 ft) thick covered section immediately overlies the Mississippian Leadville Limestone. Throughout most of the Paradox Basin the red, fine-grained Molas Formation of highly variable thickness lies between the Leadville and the Pinkerton Trail. At segment A there is no evidence of red material in the slope-covered interval between the Leadville and the lowest exposed carbonate bed of the Pinkerton Trial. This suggests that the Molas is probably thin to absent, or not red colored, in the Hermosa mountain area. The Pinkerton Trail in the Hermosa mountain area is 26 m (85 ft) thick (extending from 8 to 34 m (25–110 ft) in the section) and comprises fossiliferous limestone, rare dolostone and clastic units, and covered intervals that may be fine-grained clastic units. Locally carbonaceous shale and scattered coaly material is between the carbonate units, but there is no definitive evidence at this locality of subaerial exposure of beds in the Pinkerton Trail. The upper contact of the Pinkerton Trail is not exposed in the Hermosa mountain area. A few kilometers to the north, a road cut through the upper part of the Pinkerton Trail shows a sandstone-dominated interval sharply overlying the carbonate unit at the top of the Pinkerton Trail. We have chosen to apply the name Pinkerton Trail in this section because of its long-term usage along the outcrop belt on the southeastern side of the basin (Spoelhof, 1974).

The Paradox Formation was named for exposures in Paradox Valley, Montrose County, Colo. by Baker and others (1933) for the Pennsylvanian evaporite-bearing interval that occurs throughout most of the Paradox Basin and is exceptionally thick (1524–2134 m (5000–7000 ft)) in the central part of the basin (Wengerd and Strickland, 1954; Hite, 1960). The base and top of the Paradox Formation are defined by the lowest and highest occurrence of evaporite beds. However, the lateral extent of each evaporite bed in the formation differs greatly due to the original depositional distribution and post-depositional evaporite erosion, solution, or movement. Thus, the thickness and areal extent of the Paradox is highly variable if a strict lithologic criterion is used to define the unit. Calcareous black shale beds in the Paradox extend across the basin outside of the limits of evaporite deposition. Wengerd and Matheny (1958) and many subsequent workers correlated these shale beds throughout much of the basin to extend the name Paradox Formation to basin-margin, nonevaporite-bearing, shelf-carbonate lithofacies. In areas of dense subsurface control, these correlations are generally accurate. However, miscorrelations in some basin-margin areas have resulted in placement of the top of the Paradox stratigraphically lower than the uppermost salt beds in the basin center (Hite and Buckner, 1981).

Hite (1960) recognized 29 salt cycles, numbered 1 to 29 in descending order, in the Paradox Formation in the northern part of the basin where evaporite deposition was most extensive. Above these cycles, there are four additional evaporite cycles (numbered 0 to 0000 in ascending order) that are restricted to the northern part of the basin (Huffman and Condon, 1993; D.L. Rasmussen, 1993, oral commun.). Wengerd and Matheny (1958) subdivided the Paradox into "zones" or intervals based on correlation of oil-producing intervals in the Paradox Basin. This interval nomenclature was expanded and has been used extensively throughout the hydrocarbon-producing part of the basin. These intervals are, in assending order the Alkali Gulch interval (that includes salt cycles 29–21), the Barker Creek interval (salt cycles 20–10), the Akah interval (salt cycles 9–6), the Desert Creek interval (salt cycles 5–4), and the Ismay interval (salt cycles 3–2) (Hite and others, 1984). The Barker Creek and Akah intervals contain the most widespread evaporite beds (Hite and Buckner, 1981). Applying this interval nomenclature and the name Paradox Formation nomenclature to basin-margin areas is speculative because of rapid lithofacies changes and sparse subsurface control in the basin-margin areas.

In the outcrops along the southeastern part of the basin, evaporite beds are only observed in the Hermosa mountain area. There are about 30 m (100 ft) of gypsum interbedded with a minor amount of shale starting at about 402 m (1320 ft) in the measured section. The gypsum pinches out to the north, but the exact location of the pinchout is unknown because of extensive cover in this part of the Hermosa along the length of the Hermosa Cliffs. Evaporite beds are absent in almost all of this outcrop area and subsurface control to the west is sparse to absent, therefore, the name Paradox Formation is not applied along the outcrop belt. We use the term undifferentiated part of the Hermosa Group for that part of the Hermosa Group above the Pinkerton Trail Formation. Correlating the Hermosa mountain section to subsurface wells, the closest of which is 16 km (10 mi) to the southeast, the interval from 33.5 to at least 599 m (110 to 1965 ft) in the measured section appears to be the approximate stratigraphic equivalent of the Alkali Gulch through Ismay intervals. Other workers (G.M. Stevenson, written commun., 1993) consider the stratigraphic equivalent of these intervals at the Hermosa mountain section to extend from 33.5 to 667.5 m (100 to 2190 ft) on the basis of the fusulinid data of Wengerd and Matheny (1958). A major assumption in our interpretation was the correlation of the gypsum interval in the Hermosa mountain section with Hite's (1960) cycle 6, which was done on the basis of the thickness and stratigraphic position of this interval. Cycle 6 contains the most regionally extensive evaporite unit in the Paradox Basin.

Preliminary results of conodont studies indicate that a sample of calcareous shale from 552 m (1810 ft) contains assemblages found in the lower part of the Ft Scott Limestone of the Marmaton Group (Desmoinesian) of the midcontinent. This is one cycle higher than the Verdigris Limestone Member of the Senora Formation of the Cabaniss Group (Desmoinesian) that has a faunal assemblage similar to that contained in the Chimney Rock shale unit, which is at the base of the Desert Creek interval (B.R. Wardlaw, oral commun., 1994). A sample of calcareous shale from 587 m (1925 ft) contains conodont assemblages found in the Pawnee Limestone of the Marmaton Group or in a slightly higher cycle of the midcontinent. The Gothic shale unit, which forms the base of the Ismay interval, contains assemblages found in the Pawnee Limestone (S.M. Ritter, BYU, oral commun., 1994). Further study of fusulinid and conodont assemblages from the Hermosa locality will better constrain correlations with other areas of the basin.

The Honaker Trail Formation that overlies the Paradox Formation was defined by Wengerd and Matheny (1958) for exposures along the Honaker Trail in the canyon of the San Juan River in the southwestern part of the Paradox Basin. At this locality, the Paradox Formation does not contain evaporite beds, thus the conventional application of the highest evaporite bed as the top of the Paradox could not be used. Wengerd and Matheny (1958) chose a bed, the top of which they called the 'P datum', that they believed was traceable and time synchronous throughout the basin in the subsurface. They interpreted the 'P datum' to be equivalent to the top of the Ismay interval of the Paradox Formation. Using more recent and abundant subsurface control, Hite and Buckner (1981) showed that the P datum of Wengerd and Matheny (1958) was locally miscorrelated and therefore the top of the Paradox Formation at the Honaker Trail locality lies stratigraphically below the highest evaporite bed in the subsurface. At Honaker Trail, there is no lithologic difference between the Paradox and Honaker Trail Formations. Outside the area of evaporite deposits, most workers placed the contact between the Paradox and Honaker Trail Formations at the top of the bed that they believed to be equivalent to the top of the Ismay interval. At the Hermosa mountain locality, we did not assign the name Honaker Trail Formation to part of the Pennsylvanian section because of the absence of a lithologically definitive Paradox Formation and the uncertainties in correlations with the more basinward evaporite cycles. On the basis of our correlations of the Hermosa mountain section to the subsurface, the stratigraphic equivalent to the Honaker Trail Formation in the measured section extends from 599 m (1965 ft) or slightly higher to the top of the Hermosa Group.

The Hermosa Group at Hermosa mountain has a gradational contact with the overlying alluvial red beds of the Cutler Formation (Lower Permian). The top of the Hermosa is placed at the top of the highest carbonate bed of probable marine origin. Carbonate nodules observed in the Cutler are of pedogenic origin.

We have not applied the name Rico Formation, a name commonly applied in the Paradox Basin, to the section at Hermosa mountain. Spencer (1900) first applied the name Rico to approximately 300 ft of interbedded clastic and carbonate beds below the interval of red, clastic beds he assigned to the Triassic Dolores Formation. The Dolores of Spencer (1900) was subsequently subdivided into the Permian Cutler Formation and the Triassic Dolores Formation. Spencer (1900, p. 60) defined the Rico as composed of "sandstone and conglomerate with intercalated shales and sandy fossiliferous limestones" and having a faunal assemblage that has a much higher ratio of molluscs to brachiopods compared to the assemblage in the underlying Hermosa. The Rico was also dominantly red in contrast to the dominantly green Hermosa. Spencer interpreted the lower contact of the Rico as conformable with the Hermosa and the upper contact as the top of the bed containing "the highest known occurrence of the Rico fossils" (Spencer, 1900, p. 60). The upper contact of the Rico also appeared to be conformable with the overlying formation that Spencer believed to be of Triassic age. Spencer (1900, p. 65), citing personal communication from G.H. Girty, reported fossils from the Rico Formation having "both Permian and Carboniferous affinities", but "if it were necessary to assign the fauna to either the Coal Measures or the Permian, the line between the two should be drawn at the base of the Rico formation rather than at its top". If, as Spencer (1900) proposed, the upper contact was conformable and the overlying unit was of Triassic age, it was highly plausible that the Rico could contain Permian fauna. Later, Herman and Sharps (1956, p. 81) interpreted a Desmoinesian age for the Rico in the Rico area. Blodgett (1984, p. 68) stated that upper Desmoinesian fusulinids were found in one of the uppermost limestone beds exposed on Sandstone Mountain, which has the most complete exposures in the Rico area.

Since the Rico was defined, it has been a source of confusion and controversy among geologists. Many workers applied the name Rico to the uppermost part of the carbonate-bearing Pennsylvanian or Permian rocks (depending on location in the basin) that contained nonmarine clastic rocks, commonly red, interbedded with marine carbonate rocks (Baker, 1936; McKnight, 1940). The Rico was commonly considered to represent the transition from the entirely marine Hermosa to the entirely nonmarine Cutler (Wengerd, and Matheny, 1958). At Rico, nonmarine deposits are common in the Hermosa below beds assigned to the Rico Formation (K.J. Franczyk, 1993, unpubl. data). Nonmarine deposits also occur throughout the Hermosa Group in the outcrop belt along the southeastern margin of the basin (Spoelhof, 1974; Franczyk, 1992, 1993). Baars (1962) summarized the many problems involved with the Rico terminology and suggested that the name be abandoned. Because of the greatly varying criteria used in applying the name Rico Formation and the difficulty of mapping a unit based solely on the character of faunal assemblages, we have chosen not to apply this name to the Hermosa mountain section and also recommend that it not be used in the Paradox Basin.

Spoelhof (1974), in a study of the Pennsylvanian outcrops in the San Juan Mountain area, used a slightly different nomenclature than that used in this report. Spoelhof (1974) applied the name Pinkerton Trail Formation to the interval of dominantly fossiliferous limestone above the Molas Formation. Because there are no Pennsylvanian evaporite beds in the San Juan Mountain area, he applied the name Honaker Trail Formation to the rest of the interbedded clastic and carbonate sequence above the Pinkerton Trail. On the basis of lithologic characteristics, he subdivided the Honaker Trail into lower, middle, and upper members. Because of the problems associated with definition and application of the name Honaker Trail, we chose not to extend this Honaker Trail terminology into the Hermosa mountain area. However, there is excellent correspondence in both the total thickness and the gross lithologic subdivisions of the Hermosa Group between Hermosa mountain and the exposures in the San Juan Mountain area. The limestone-dominated middle member of the Honaker Trail of Spoelhof (1974) is equivalent to the interval in the Hermosa mountain section from 588 to 677 m (1930 to 2220 ft) that contains the most abundant and thickest limestone units in the Hermosa Group exclusive of those in the Pinkerton Trail.

Wengerd and Matheny (1958) applied the Paradox and Honaker Trail Formations and Rico "facies" nomenclature to the interval we call the undifferentiated part of the Hermosa Group at Hermosa mountain. In our section, 552 m (1810 ft) corresponds to the top of their Paradox Formation, 744 m (2440 ft) corresponds to the top of their Honaker Trail Formation, and 805 m (2640 ft) corresponds to the top of their Rico facies. With one notable exception, the thickness measured in our Hermosa mountain section is generally comparable to Wengerd and Matheny's (1958) section. Wengerd and Matheny (1958) measured 643 m (2110 ft) for the Hermosa Group; we measured 843 m (2765 ft). The significant difference in thickness occurs primarily in the interval between the top of the Pinkerton Trail Formation and the base of the gypsum bed in the undifferentiated part of the Hermosa Group. For this interval, Wengerd and Matheny (1958) showed 255 m (835 ft) compared to our 369 m (1210 ft). Abundant cover immediately above and lateral to the Pinkerton Trail measured section locality, laterally discontinuous exposures, and few distinct marker beds make accurate measurement of this interval difficult. In our measurement, we offset northward along the top of the Leadville Limestone to the base of the slope containing the first good continuous exposures. At the location of segment B above the projected top of the Pinkerton Trail, there is 220 ft (67 m) of cover that is overlain by 510 ft (155 m) of nearly continuous exposures. All subsequent offsets were made on traceable marker beds.

BRACHIOPOD BIOSTRATIGRAPHY

The faunas of the Hermosa Group (Pennsylvanian) at its type locality on Hermosa mountain are dominated both in numbers and diversity by brachiopods. Whereas other fossil groups such as fusulinid foraminiferans, conodonts, or ammonoid cephalopods are clearly the fossils of choice for constructing zonations of Pennsylvanian-age marine rocks, if those forms are not present or occur sporadically throughout the stratigraphic column, the brachiopods can provide biostratigraphic information to supplement or replace that normally sought from these other taxa.

ZONATION OF THE PENNSYLVANIAN

The Fusulinid Zones

Fusulinid zones have been delineated in Pennsylvanian rocks in several areas in the southwest. Ross and Sabins (1965) and Ross (1973) described the fusulinid zones in southeastern Arizona. Waddell (*in* Sutherland and Harlow, 1973) recognized seven fusulinid zones in the southern Sangre de Cristo Mountains of New Mexico. Sutherland and Harlow (1973), in their work on the taxonomy and biostratigraphy of Pennsylvanian brachiopods of the southern Sangre de Cristo Mountains, described the stratigraphic distribution of brachiopods in the overall framework of the fusulinid zones delineated by Waddell (*in* Sutherland and Harlow, 1973). Brew (1979) did the same for the brachiopods of the Naco Formation (Pennsylvanian) in central Arizona, while employing the fusulinid zonations of Ross and Sabins (1965) and Ross (1973). In addition, Brew (1979) outlined the Pennsylvanian fusulinid zonation in central Arizona. In both regions, the biostratigraphic usefulness of brachiopods was clearly demonstrated. Study of the brachiopods of the Hermosa Group suggests that they too can be useful in dating the Hermosa in its type locality.

The fusulinid zones recognized by Waddell, and their correspondence to zonations elsewhere, are as follows:

Zone I (upper Atokan)—Characterized by species of *Fusulinella.*

Zone II (lowermost lower Desmoinesian)—Species of *Fusulinella* in association with early forms of *Beedeina* (= *Fusulina*; see Ross, 1969; note that Waddell refers to all forms herein termed *Beedeina* as *Fusulina*) as well as *Wedekindellina*; this appears to correspond to the *Fusulinella dosensis* subzone of Ross and Sabins (1965) and Ross (1973) in Arizona.

Zone III (lower Desmoinesian)—Species of *Beedeina* associated with *Wedekindellina*; this appears to correspond to the *Wedekindellina euthysepta* subzone (Ross and Sabins, 1965; Ross, 1973).

Zone IV (middle Desmoinesian)—Various species of *Beedeina* including *B. haworthi* and others recognized by Waddell in the Ardmore Basin (Waddell, 1966); appears to correspond to the *Beedeina girtyi* subzone of Ross and Sabins (1965) and Ross (1973).

Zone V (upper Desmoinesian)—Characterized by *Beedeina cappensis* in northern New Mexico; appears to match the *Beedeina eximia* subzone of Arizona (Ross and Sabins, 1965; Ross, 1973).

Zone VI (Missourian)—Diagnostic forms included *Triticites ohiocensis* and *T. irregularis*.

Zone VII (Virgilian)—More advanced species of *Triticites*, including *T. cf. T. cullomensis*.

Stratigraphic Distribution of Brachiopods

Most of the brachiopods studied by Sutherland and Harlow (1973) in New Mexico are of Morrowan, Atokan, and Desmoinesian age. Their work revealed what they described as a "major change" between faunas of Morrowan and Atokan age; Atokan faunas were more similar to those of the overlying Desmoinesian than to underlying Morrowan forms. An increase in the coarseness of costation was noted in several genera and generic pairs that span the Morrowan through Desmoinesian interval; the taxa that was defined clearly could serve as the basis for an intrabasinal zonation, if not one of wider application.

The associations that Sutherland and Harlow (1973) noted between various species of brachiopods and the recognized fusulinid zones are outlined below (note that species marked by an asterisk occur in the Hermosa mountain section in Colorado):

Zone I (upper Atokan)—*Antiquatonia coloradoensis, Neochonetes henryi, N. whitei, Desmoinesia ingrata* (principal diagnostic species).

Zones I and II (upper Atokan and lower Desmoinesian)—*Neospirifer cameratus**, *Mesolobus striatus, Anthracospirifer curvilateralis chavezae**, *Kozlowskia montgomeryi**.

Zones II and III (lower Desmoinesian)—*Antiquatonia hermosana**, *Kozlowskia haydenesis**, *Hustedia "mormoni"*, *Anthracospirifer rockymontanus, Derbyia crassa** (principal diagnostic species).

Zones IV and V (middle and upper Desmoinesian)—*Desmoinesia muricatina, Mesolobus euampygus, Antiquatonia portlockiana**.

Zone VI (Missourian)—*Echinaria cf. E. semipunctata, Linoproductus cf. L. platyumbonus, Neospirifer alatus*.

Zone VII (Virgilian)—*Wellerella immatura, Leiorhynchoidea? rockymontana, Beecheria millipunctata, Neospirifer dunbari*.

AGE OF THE HERMOSA GROUP

Fusulinids collected from the Pinkerton Trail Formation as part of this study were not identifiable because of extensive recrystalization (R.S. Nail, Texas Tech. Univ., written commun., 1994). Although published information contains some inconsistencies, it typically indicates that *Fusulinella* and *Fusulina (=Beedeina)* occur together near the base of the Pinkerton Trail Formation of the Hermosa (Wengerd and Strickland, 1954; Wengerd and Matheny, 1958). This portion of the Pinkerton Trail has been commonly identified as Atokan in age, however, that age cannot be sustained if the two fusulinid genera named above always occur together; if *Fusulinella* occurs by itself, however, stratigraphically below beds containing both genera, an Atokan age would be valid for the lowermost beds of the Pinkerton Trail.

The brachiopods identified in the lowermost beds of the Pinkerton Trail suggest an age of early Desmoinesian; there are no brachiopods in these beds that were identified by Sutherland and Harlow (1973) as confined to rocks of Atokan age. The species that are present are longer ranging forms found in zones I through III, Atokan through lower Desmoinesian.

Preliminary analysis of fusulinids from 656 m (2153 ft) provides only a Desmoinesian age that is not well constrained (R.S. Nail, Texas Tech. Univ., written commun., 1994). Published information on fusulinids from about 200 ft below the top of the Hermosa Group at Hermosa mountain (Wengerd and Matheny, 1958) suggests that the youngest beds at the type locality are no younger than Desmoinesian in age. Our brachiopod data from this locality appear to confirm this. In New Mexico, Sutherland and Harlow (1973) considered *Antiquatonia portlockiana* to range from uppermost zone IV through lower zone V (upper middle through lower upper Desmoinesian). This is generally consistent with the stratigraphic distribution of this species in Arizona (Brew, 1979); however, *A. portlockiana* is slightly longer ranging in Arizona, extending into the lowermost Missourian.

In the Hermosa mountain section, *A. portlockiana* occurs approximately 150 feet (46 m) below the Hermosa-Cutler contact. Another form, the chonetid *Mesolobus mesolobus* which is confined to rocks of Desmoinesian age wherever it has been found, occurs approximately 320 feet (98 m) below the Hermosa-Cutler contact. Thus, the uppermost beds of the Hermosa are probably no younger than Desmoinesian, although the possibility of an early Missourian age cannot be eliminated.

In contrast to the above findings, Petuch (1988), citing evidence from bivalves, assigned an early Permian (Wolfcampian) age to Hermosa Group beds exposed along East Animas Road (County Road 250) approximately 2 mi (3 km) south of Hermosa mountain. No diagnostic zonation for Pennsylvanian and Permian rocks based on bivalves exists, and bivalves have not been studied in any detail in this region. Thus Petuch's revision of the age of these rocks remains unsubstantiated, and his assignment of these beds to the Cutler Formation based on their purported Permian age remains unjustified.

In summary, on the basis of the brachiopod faunas, the entire Hermosa Group in its type locality at Hermosa mountain is primarily of Desmoinesian age, ranging from the earliest Desmoinesian through at least early late Desmoinesian. The possibility does exist that the youngest beds may be of early Missourian age.

SEDIMENTOLOGY

DEPOSITIONAL CYCLES

The undifferentiated part of the Hermosa Group above the Pinkerton Trail Formation at Hermosa mountain consists of stacked depositional cycles that usually contain a carbonate unit at the base overlain by a upward coarsening clastic interval. There are at least 28 and, by interpolation in the covered intervals, possibly 40 depositional cycles that range in thickness from 3 to 45 m (10–148 ft). Similar cycles characterize the Hermosa Group in outcrops to the north in the San Juan Mountains area (Spoelhof, 1974) and in the Ouray, Colo. area (Franczyk, 1993). Within the Pinkerton Trail, extensive cover of either the entire formation or the interbeds between carbonate units make it difficult to discern or define a cyclical character, if present. The general character of a cycle is described in the following paragraphs. Details of both the carbonate and clastic components of the cycles are described in the following sections "Carbonate Sedimentology and Petrography" and "Clastic Sedimentology".

The base of a cycle is placed at the base of the carbonate unit because of the lithologic contrast and sharp contact between clastic and carbonate units. In the Hermosa mountain area, the lithologic and sedimentologic features of the clastic parts of a cycle remain relatively consistent throughout the entire Hermosa Group. However, the carbonate units show notable differences in thickness, lithology, faunal assemblages, and depositional environment depending on stratigraphic position in the Hermosa Group (discussed below). Regardless of stratigraphic position, the carbonate unit almost always has a sharp, nearly flat contact with the underlying clastic unit. Throughout the entire Hermosa Group, the contact between the carbonate and overlying clastic unit is either sharp or slightly gradational over less than a 0.5 m-thick interval. Calcareous, fine-grained sandstone or siltstone characterize this gradational zone. The carbonate units range in thickness from 0.3 to 21 m (1–70 ft) and are thin to thick bedded. Locally, interbeds of clastic mudstone to siltstone are in carbonate units. Depending on stratigraphic position, the carbonate unit may be a limestone, dolostone, or slightly dolomitized limestone. Lime mudstones, wackestones, and less commonly packstones, compose most of the limestone units; grainstones are rare. The dolostones are recrystallized lime mudstones having rare bioclastic material.

The clastic part of the cycle overlying the carbonate unit shows an overall coarsening upward trend and is highly variable in thickness. The contact between the carbonate unit and overlying clastic rocks is commonly covered to partly covered. At the base of the clastic interval, mudstone or shale, if present, grades upward into thinly interbedded siltstone, mudstone, and very fine grained sandstone. Gradationally overlying this interbedded interval is parallel- to ripple-laminated, locally cross-stratified, very fine to fine-grained sandstone. This sandstone interval may constitute the uppermost unit of a cycle or alternatively be gradationally to sharply overlain by a series of interbedded siltstone, mudstone, and tabular to lenticular sandstone units, or be sharply and irregularly overlain by a medium- to coarse-grained to pebbly sandstone interval that has a fining-upward grain-size trend. Locally, the uppermost few inches (centimeters) of the clastic part of a cycle may be a coarse-grained to granule sandstone.

CARBONATE SEDIMENTOLOGY AND PETROGRAPHY

Four distinct carbonate lithofacies are identified in the Hermosa Group at Hermosa mountain on the basis of mineralogy, texture, biotic assemblages, and inferred depositional environment. This division is unusual in the Pennsylvanian outcrop belt in the eastern Paradox Basin. In other areas along this outcrop belt, carbonate rocks show fairly uniform characteristics throughout the Hermosa section. The four lithofacies at Hermosa mountain only partly coincide with previously discussed formal stratigraphic subdivisions. Lithofacies no. 1 is stratigraphically lowest and corresponds to the Pinkerton Trail Formation; it extends from 8 to 34 m (25–110 ft). Lithofacies nos. 2, 3, and 4 are in the undifferentiated part of the Hermosa Group: no. 2 extends from 119 to 399 m (390–1310 ft), no. 3 extends from 485 to 599 m (1590–1965 ft), and no. 4 extends from 599 to 843 m (1965–2765 ft). Carbonate units constitute only about 20 percent of the entire Hermosa Group.

Lithofacies No. 1

Carbonate, in medium to thick beds, comprises 60 percent of lithofacies no. 1; the section between the carbonate units is dominantly covered, but probably contains fine-grained clastic units. Wackestone is the dominant carbonate texture with less common packstone and rare grainstone. The biotic assemblages in these carbonate units include crinoid stems; brachiopod shells and spines; fusulinids; *Climacammina* sp.; monaxon sponge spicules; bryozoans such as *Prismopora triangulata*, fenestelloid or string bryozoans, and encrusting *Fustulipora* sp.; and thin-shelled brachiopod debris. The packstone textures generally occur in the units rich in sponge spicules (sample no. PH91KF7).

Vertical changes in biotic assemblages and textures through the Pinkerton Trail Formation indicate rapid changes in depositional facies. The lowest carbonate unit (sample nos. PH91KF1, PH91KF2) contains crinoid sand grains mixed with angular, medium-grained quartz sand grains indicating a very shallow subtidal, high-energy, shoreline environment. Packstone in the overlying unit (sample no. PH91KF3) contains an assemblage of bryozoans, forams, brachiopods, crinoids, and sponges strongly suggesting that deposition occurred on an open, irregular shelf floor at or near wave base. The next two stratigraphically higher carbonate units (sample nos. PH91KF5, PH91KF7) are wackestones to packstones that contain a faunal assemblage dominated by monaxon sponges with local thin-shelled brachiopod debris. This change in faunal assemblages and the carbonate texture suggests below wave-base conditions on an irregular, sloping, shelf floor onto which sponge spicules drifted and accumulated. Faunal abundance and diversity increases in the overlying carbonate units in the Pinkerton Trail indicating a return to shallower water, moderate-energy conditions.

To the north in the outcrops of the San Juan Mountains area, faunal abundance and diversity is greater in the carbonate beds in the Pinkerton Trail Formation. Spoelhof (1974) documented assemblages of small gastropods, *Chaetetes*, bryozoans, phylloid algae, *Komia*, and *Girvanella* in wackestones and packstones. These assemblages and textures indicate deposition in shoaling, shallow subtidal conditions—shallower, higher energy environments than those present in the Hermosa mountain area. In addition to this change in depositional environment, the Pinkerton Trail thickens northward from the Hermosa mountain area. Spoelhof (1974) measured as much as 84 m (275 ft) of Pinkerton Trail in the San Juan Mountain area outcrops; however, the Pinkerton Trail is also locally absent over fault blocks.

Lithofacies No. 2

The base of lithofacies no. 2 begins at 119 m (390 ft), the base of the lowest carbonate exposed above the Pinkerton Trial, and ends at 399 m (1310 ft), the top of the carbonate below the gypsum unit; sample nos. PH91KF15–PH91KF49 are in this stratigraphic interval. Carbonate units that are thin to thick bedded and dominantly dolostone and minor dolomitic limestone comprise about 15 percent of this stratigraphic interval. Diagenesis and pressure solution have obliterated much of the original texture and structure in the carbonate beds; however, reconstructed precursor textures suggest mud-supported sediments. Biotic constituents are absent in most beds. Where rare fossils were observed, the assemblages consist of crinoid stems, echinoderm plates, and brachiopod shells; faint algal laminations and stromatolitic festooned structures are visible in some beds. Other characteristic features in these carbonate units are platey desiccation breccias that show little or no reworking, desiccation cracks, fenestral porosity, small evaporite crystals and pseudomorphs in some reworked intraclasts, probable evaporite collapse structures, dolomitized calcretes, and traces of very fine, angular, windblown sand grains visible only under cathodoluminescence.

Lithologies, type and distribution of biotic components, and syndepositional structures in the carbonate beds of lithofacies no. 2 indicate deposition under arid, hypersaline conditions in the intertidal to supratidal zone. Much of the deposition occurred on tidal flats that were subjected to periodic subaerial exposure and received very fine, windblown sand. Carbonate beds in this lithofacies most commonly lie directly on alluvial- and delta-plain deposits, indicating deposition during the transgression that accompanied the initial phase of a relative sea-level rise when clastic sediment was trapped in a more landward setting. Prodelta to distal delta-front clastic beds overlying carbonate beds indicate that water depth continued to deepen but carbonate production was halted. Clastic influx or a rapid deepening may have terminated carbonate production. Presence of a condensed section at the top of the carbonate or in the immediately overlying fine-grained clastic unit could not by definitively determined. If a condensed section developed above the carbonate during transgression, clastic influx would probably not have been great enough to halt carbonate production.

To the north, in the Hermosa Group outcrop belt on the western edge of the San Juan Mountains, there is no evidence of restricted, hypersaline, peritidal carbonate deposition in the stratigraphic interval equivalent to carbonate lithofacies no. 2. In this area, detrital intervals I, II, and part or all of III in Spoelhof's (1974) lower member of the Honaker Trail Formation are the stratigraphic equivalents of carbonate lithofacies no. 2. The ratio of clastic to carbonate beds is similar to that in the Hermosa mountain area. However, the character of carbonate beds in the San Juan Mountain area is notably different. The beds in the San Juan Mountain area are wackestones and packstones that are locally oolitic; dolostones are localized and grade laterally into oolitic packstones and wackestones. A diverse faunal assemblage of corals, brachiopods, molluscs, crinoid stems, bryozoans, algae, fusulinids and other forams, *Chaetetes*, and *Komia*, which indicate normal marine waters, characterizes many of these carbonate beds. There is no evidence of evaporative conditions. Spoelhof (1974) interpreted the carbonate environments as high-energy, shallow, subtidal marine, quiet, shallow marine, supratidal, and very shallow subtidal, depending on locality. Geographic and temporal variations were due in part to subtle, local tectonic activity creating localized topographic highs (Spoelhof, 1974).

Lithofacies No. 3

At Hermosa mountain, the stratigraphic interval from 421 to 485 m (1380–1590 ft) is poorly exposed; only the more resistant sandstone units crop out. Therefore, the character of the carbonate beds that might occur in this interval is unknown. Lithofacies no. 3 includes the interval from 485 to 599 m (1590–1965 ft), which includes sample nos. PH91KF51–PH91KF63. A diverse carbonate suite is present in this lithofacies: a partly covered calcareous shale unit that may extend into the overlying covered interval is from 485 to 491 m (1590–1610 ft); a dolomitic siltstone unit (sample no. PH91KF52); and two well exposed calcareous shale units (sample nos. PH91KF53, PH91KF58) that are each capped by cliff-forming, fossiliferous limestone units (sample nos. PH91KF54, PH91KF60–PH91KF63).

The shale units are very dark gray to black and fissile with nearly equal parts silt-sized calcite (and less commonly dolomite) grains and clay to silt-sized (and locally medium sand-sized) quartz grains. These beds are devoid of fossils with the exception of sponge spicules. The fine grain size and general lack of fauna indicate a low-energy, restricted marine environment, possibly a deep water, lower slope or basin-floor setting.

Sample nos. PH91KF53 and PH91KF58 show two different modes of origin of the laminae in the shale. In sample no. PH91KF53, clay and or silt-sized quartz-grain laminae alternate with laminae composed of silt-sized calcite and less abundant dolomite grains. These carbonate grains are neomorphic, show broken grain contacts, and are interpreted to have originated on the carbonate platform where they were neomorphosed shortly after deposition. Storms eroded the platform and transported the carbonate grains basinward, where they were deposited as sheets on basinal muds and silts. Although the carbonate now occurs as laminae, the original depositional thickness is unknown as pressure solution during burial may have removed significant volumes of carbonate.

In sample no. PH91KF58, the carbonate fraction appears to be authochthonous. The sample is composed of detrital, medium-sized, angular quartz grains in a matrix of micritic peloidal muds. The quartz was transported from a nearby source, probably during storm events, into a low-energy, carbonate environment. Lack of additional samples makes it impossible to know whether a single process dominated the deposition of each calcareous shale unit, or whether both processes were operative.

The dolomitic siltstone unit (sample no. PH91KF52) in this lithofacies is also devoid of fauna. It may reflect some variation in water depth from that in which the overlying and underlying shales formed, but lack of diagnostic features make the depositional environment difficult to determine.

The sharp contact between the two cliff-forming, fossiliferous limestone units (sample nos. PH91KF54 and PH91KF60–PH91KF63) and the underlying calcareous shale units indicate an abrupt depositional transition from restricted, quiet water to carbonate-platform environments. The lower limestone unit is 5 m (18 ft) thick; and the upper limestone unit, including thin clastic interbeds in its upper part, is 10 m (34 ft) thick. Both units are mud- to grain-supported bioclastic wackestone to packstone. Their biotic components, in decreasing order, are brachiopod shells and spines, bryozoans such as *Fistulipora* sp., echinoderm plate sands, fusulinids, small forams, and the first appearances of phylloid algae and *Komia*. The faunal material appears to be transported into place rather than forming in place as a biostrome. The assemblages and fabrics in these units suggest sedimentation under normal, shallow-water, subtidal, marine conditions on the carbonate platform.

In both the Hermosa mountain and San Juan Mountains outcrop areas, there is an increased abundance of carbonate units in lithofacies no. 3 and its stratigraphic equivalents have increased in abundance compared with the underlying part of the Hermosa Group. In the San Juan Mountains area, the upper part of detrital-carbonate interval IV of the lower member of the Honaker Trail Formation of Spoelhof (1974) and the lowest carbonate unit in the middle member of the Honaker Trail Formation of Spoelhof (1974) are the stratigraphic equivalents of carbonate lithofacies no. 3. At Engineer Mountain, in the southern part of the San Juan outcrop belt, the carbonate units are mudstone and wackestone having a sparse faunal assemblage of crinoid stems, brachiopods, forams, and molluscs; Spoelhof (1974) interpreted a low-energy, restricted marine environment for these units. An increase in the abundance and diversity of fauna to the north indicate a change to a high- to moderate-energy, open, normal-marine environment. Spoelhof believed the low-energy environment occurred over a structural platform.

The restricted, possibly deep, marine environment represented by the calcareous shale units in lithofacies no. 3 in the Hermosa mountain area indicate a dramatic change from both the earlier hypersaline, peritidal environments of lithofacies no. 2 and the contemporaneous shallower, restricted- to normal-marine environments to the north. These temporal and geographic changes in carbonate shelf paleotopography may reflect both structural controls along the eastern basin margin and eustatic conditions. A slight structural high in the Hermosa mountain area during deposition of lithofacies no. 2 could account for the persistent, restricted, hypersaline, peritidal environments. A reversal of fault movement may have changed the shelf gradient and configuration so that during deposition of lithofacies no. 3, the Hermosa mountain area was in a deeper-water part of the shelf compared to areas to the north. Alternatively, shelf conditions may have been relatively similar from the Hermosa to the San Juan Mountain area with carbonate production rates varying greatly between these two areas. The interpretation of deeper water environments in the Hermosa mountain area suggests that tectonic processes were in part responsible for variations in depositional facies.

The increase of carbonate deposits across the eastern basin margin during the time of lithofacies no. 3 may indicate a large-scale, possibly third order, eustatic rise. In the Hermosa mountain area, this rise is reflected in the deposition of marine calcareous shales and in the deposition of the first normal-marine carbonate platform deposits above the Pinkerton Trail Formation.

Lithofacies No. 4

Carbonate lithofacies no. 4 extends from 599 to 843 m (1965–2765 ft) and includes sample nos. PH91KF64–PH91KF107. Carbonate beds in lithofacies no. 4 vary in texture from lime mudstones to packstones and local grainstones; the faunal assemblages vary from low to moderate in diversity and abundance. As in all other intervals of the Hermosa Group, the carbonates are interbedded with marine to nonmarine deltaic and alluvial deposits. The depositional environments of the carbonate beds vary from normal to restricted marine. With the exception of the 21 m (70 ft) thick carbonate interval from 655 to 676 m (2150 to 2220 ft) that contains stacked units (each unit composed of carbonate bedsets), the carbonates in lithofacies no. 4 are single units that range from 0.3 to 3 m (1–10 ft) thick.

The basal carbonate unit (sample nos. PH91KF66, PH91KF67) in lithofacies no. 4 is at 620 to 623 m (2035–2043 ft) and is a mudstone to packstone that is locally dolomitic. The faunal abundance is low and restricted to small fauna such as ostracodes and forams in addition to a few brachiopod shell fragments. These features indicate deposition in a low-energy, possibly restricted, environment on muddy floors of the carbonate platform.

The 21 m (70 ft) thick carbonate unit at 655 to 676 m (2150–2220 ft; sample nos. PH91KF73–PH91KF79) is the thickest carbonate unit in the Hermosa Group. Beds in this unit are packstones to wackestones with local pockets or lenses of grainstones; interbeds of claystone, siltstone, and fine-grained sandstone occur in the upper part. The fossil assemblages are enriched with fusulinids, bryozoans, brachiopod spines and shells, and monaxon sponge spicules. Rare thin zones of bryozoan biocoenoses suggest the occurrence of localized biostromes in these beds. However, the mixed and fragmented communities in mud-supported matrices suggest transport before final deposition. Deposition was on an irregular shallow shelf floor in the subtidal zone.

In the remaining part of lithofacies no. 4, carbonate units are thinner and interbedded with thick marine and nonmarine clastic intervals. Most units range from 0.3 to1.5 m (1–5 ft) thick; the thickest unit (sample nos. PH91KF97, PH91KF98) is 3 m (10 ft). Carbonate textures are dominantly mudstone to wackestone and less common packstone and grainstone. The mudstone fabrics are locally fenestral with stromatolites, which strongly suggests sedimentation in the intertidal zone of the platform. Restricted conditions in the intertidal zone may have locally existed based on the very localized occurrence of very fine sized (<0.125 mm) superficial oolites and less common oolites and grapestones that have radial crystaline structures, a texture indicative of formation in hypersaline water (Flugel, 1982). The wackestones contain fossil assemblages of molluscs (gastropods and pelecypods), brachiopod shells, echinoderm plates, crinoid stems, and algal communities of phylloids and Dasyclads; bryozoans and sponges are generally rare to absent. The texture and fossil assemblage indicate deposition in the very shallow subtidal zone. Angular, medium-sized detrital quartz grains are present in many of the carbonate beds; a mixture of detrital lignite and quartz sand grains locally occurs in carbonate mudstones. The common occurrence of clastic detritus suggests that active deltaic areas lay adjacent to or landward of the carbonate zones.

The interval containing the uppermost carbonate unit in lithofacies no. 3 through the thick carbonate unit in lithofacies no. 4 (588–677 m (1,930–2,220 ft)) correlates with Spoelhof's (1974) middle member of the Honaker Trail Formation in the San Juan Mountains area. The middle member and its equivalents are traceable across the eastern basin-margin outcrop belt and mark the time of highest carbonate productivity under normal marine conditions during deposition of the Hermosa Group. There are, however, notable differences in the carbonate units in this zone between the Hermosa mountain and San Juan Mountains area outcrops. The carbonates at the Hermosa mountain locality contain lower faunal abundance and diversity and a higher lime-mud component that do equivalent carbonates in the San Juan Mountain area. There is a scarcity of algal communities and an absence of *Chaetetes* and *Komia* in the carbonate beds at Hermosa mountain. The greater abundance of packstone to grainstone and locally boundstone textures, the more abundant and diverse fauna, and the abundance of fauna requiring photic-zone conditions in the San Juan Mountains area indicate higher energy, shallower water environments than those to the south. A steeper gradient on the carbonate platform in the Hermosa mountain area may account for the deeper water conditions in this area compared to the San Juan Mountains area.

The Hermosa Group above 677 m (2220 ft) in lithofacies no. 4 is time equivalent to Spoelhof's (1974) upper member of the Honaker Trail Formation in the San Juan Mountains area. Carbonate beds in this interval appear to be similar in both areas, but a detailed comparison is complicated by extensive cover and very limited exposure of carbonate beds in the San Juan Mountains area. Where exposed in both the Hermosa mountain and San Juan Mountains areas, these carbonate beds are thinner and have lower faunal diversity than those in the underlying part of the section. Spoelhof (1974) interpreted their depositional environment to be shallow water, low energy, and close to a shoreline. He attributed this change in environment to a decrease in gradient and an increase in width of the carbonate platform due to filling of the Paradox Basin near the end of deposition of the Hermosa Group. This widening of the platform and decrease in shelf topographic relief produced relatively uniform depositional environments along a much greater area of the eastern basin margin than existed previously. These conditions produced carbonate beds of similar characteristics from the Hermosa mountain area northward to the San Juan Mountains area.

CLASTIC SEDIMENTOLOGY

The clastic deposits in cycles formed in marine deltaic and nonmarine deltaic and alluvial environments. Although the clastic parts of cycles show many similarities, the percentage of depositional facies can vary greatly between cycles. The following paragraphs describe the basic components of a clastic part of a cycle in ascending order.

Marine Deltaic Deposits

Mudstone beds occur rarely at the base of clastic sequences. Where present, mudstone is generally less than 0.3 m (1 ft) thick, but a slightly greater thickness may be possible because cover as thick as 1 m (3 ft) thick commonly occurs immediately above carbonate units. As a result of weathering and poor exposure, the internal stratification of mudstone beds is usually obscured, but some very thin lamination is discernable. Mudstone is medium to dark gray and olive gray, generally noncalcareous, and contains scattered organic debris and, locally, impressions of small twigs or plant stems. The fine grain size and presence of organic material indicates deposition in quiet, low-oxygen waters. The environment could be shallow, protected lagoons or deep prodelta open marine conditions. Environmentally diagnostic fauna was not observed in these beds. However, the gradational contact with overlying beds suggests an open-marine environment.

Overlying a mudstone bed is a unit of thinly interbedded siltstone, mudstone, and very fine grained sandstone, which ranges in thickness from less than 1 to 5 m (3–16 ft). In its lower part, this unit is characterized by thin to medium beds of siltstone interbedded with very thin to thin beds of mudstone or sandstone. The siltstone beds are medium to light gray to brownish black; slightly calcareous to noncalcareous; parallel to wavy parallel laminated (where sedimentary structures are discernable); locally burrowed and less commonly bioturbated; and contain fine, disseminated organic debris.

Upward through the interbedded unit, siltstone beds decrease in thickness and abundance and occur as thin to very thin interbeds between very thin to medium beds of very fine to fine-grained sandstone. Sandstone beds are yellow brown to olive gray, slightly calcareous to noncalcareous, have planar upper and lower contacts, are parallel to wavy parallel laminated, have symmetrical to asymmetrical crosslaminations locally in the upper part, and have rare burrows along bedding planes. In some cycles, the highest sandstone beds in these interbedded units have sharp, slightly scoured basal contacts and internally show parallel laminations or possible hummocky to low-angle cross-stratification overlain by symmetrical to asymmetrical crosslaminations.

In many cycles, an interval of stacked sandstone beds overlies the interbedded siltstone, mudstone, and sandstone interval. Within such sandstone intervals medium to thick beds dominate, the basal bedding surfaces are sharp and flat to slightly erosional, fine to medium grain size is dominant with some beds being very fine grained, and the intervals show no perceptible grain-size change or show a slight fining-upward trend. Sedimentary structures in the sandstone beds include planar to wavy laminations, asymmetric to symmetric ripple cross-stratification, and less common trough cross-stratification. In some parts of the section, these intervals are dominated by very thin to medium beds of planar laminations, ripple cross-stratification, and rare hummocky cross-stratification. Both soft sediment deformation and surface weathering commonly obscures the internal structures of many beds. The sandstone beds appear to form tabular units, but the abundant cover prohibits lateral tracing of beds.

One interval of stacked sandstone beds (from 209 to 218 m (687–715 ft)) in the measured section contains large-scale lateral accretion bedding that dips to the northwest. Although not common, units with similar features are observable in the sheer cliff face exposures of Pennsylvanian strata in the Hermosa Cliffs, north of the Hermosa mountain section. In the Hermosa mountain section, the lateral accretion beds have a sharp, but not erosional, contact with the underlying sandstone unit. Medium- to coarse-grained sandstone composes the accretion beds, which are separated from each other by thin splits of micaceous, poorly sorted, dominantly fine grained sandstone. There is only a slight increase in grain size between the lateral accretion beds and the underlying planar bedded sandstone unit that contains alternating fine- and coarse-grained sandstone beds. Internally, the accretion beds are composed dominantly of planar laminations that are locally interbedded with trough cross-stratification. Discrete packages of lateral accretion beds coalesce to form a laterally extensive, tabular sandstone unit; these packages are separated by curved scour surfaces that, from a distance, give the appearance of amalgamated lenticular sandstone units.

Deposits that comprise the interval of interbedded mudstone, siltstone, and sandstone and the overlying interval of stacked sandstone beds are interpreted to have formed in a shoaling marine environment in a deltaic system. In the prodelta and distal delta-front environments, low-energy conditions below wave base were periodically interrupted by flood and storm events that produced the interbedded mudstone, siltstone, and sandstone intervals. Higher energy conditions accompanied progressive shallowing into the delta-front environment and resulted in more abundant and thicker sandstone beds. The large-scale lateral accretion beds that form some sandstone units are interpreted as deltaic foreset beds in Gilbert-type deltas, a type of delta that commonly forms where there is a significant amount of topographic relief along the margins of tectonically active basins. The occurrence of Gilbert-type deltas indicates that locally rivers created steep-sided, wedge-shaped, delta-front complexes where they discharge into the sea.

Delta-Plain Deposits

The unit of stacked sandstone beds may form the top of a cycle where it is overlain by a carbonate unit. In some cycles, this sandstone interval is overlain by a series of intercalated siltstone, mudstone, and tabular to lenticular sandstone units. The thickest intervals of these deposits are from 116 to 146 m (380–480 ft), 168 to 187 m (550–615 ft), and 360 to 393 m (1180–1290 ft) in the measured section; thinner intervals are present throughout much of the section.

Siltstone, mudstone, and claystone, and very fine to fine-grained sandstone commonly occur interbedded with each other in very thin to thin beds. Units composed of these interbedded deposits have gradational upper and lower contacts and may exhibit an upward fining or upward coarsening grain-size trend. Common characteristics and accessory components of such units are abundant mica, small burrows generally parallel to the bedding, fine disseminated organic material, small plant and twig impressions, and local root casts and mud cracks. Ripple cross-stratification interspersed with small scour-and-fill structures are the most common sedimentary structures in the sandstone beds.

Locally, units from 1 to 3 m (3–10 ft) thick are composed solely of claystone or mudstone. These units generally have a sharp contact with underlying sandstone units and grade upward into siltstone beds or units of interbedded mudstone and siltstone or sandstone.

Tabular to lenticular sandstone units form about 50 percent of the intercalated intervals. The tabular sandstone units are dominantly 1 to 2 m (3–6 ft) thick and locally as thick as 6 m (20 ft), have sharp, flat lower contacts, and show a slight upward coarsening or a fining-upward grain-size trend or show no change in grain size. These units are dominantly fine to very fine grained, although medium-grained and rare coarse-grained sandstone occur. Internally, the bedding ranges from very thin to medium with thin bedding most dominant. Most beds are planar laminated and asymmetric ripple cross-stratified; present but less abundant are wavy, continuous to discontinuous laminations, trough cross-stratification, and symmetrical ripple cross-stratification.

The lenticular sandstone units are 1 to 5 m (3–15 ft) thick, have sharp, erosional lower contacts, and show a well developed fining-upward grain-size trend or fine slightly in the uppermost part of the unit. Medium- to fine-grained sandstone dominates, but many units have coarse-grained sandstone to pebbly sandstone at their base. The bedding in each unit commonly shows an upward decrease in size from very thick to thick at the base to medium or thin at the top. Trough cross-stratification is the dominant bedding structure in the lower part of these units; locally it is interbedded with tabular cross-stratification. Planar laminations are locally in the lower part of a unit but more commonly occur interbedded with trough or ripple cross-stratification in the upper part of a unit. Asymmetrical ripple cross-stratification is dominant in the upper part of these units. Rip-up clasts and soft-sediment deformation structures are common.

The intervals of interbedded siltstone, mudstone, and tabular to lenticular sandstone units are interpreted as delta-plain deposits. The conformable position above the shallow marine deltaic deposits, the low energy levels required to form most of these units, and the relatively small size of the channel deposits indicate deposition in low-gradient areas close to the shoreline. Continually shifting channel, overbank, and flood-plain environments resulted in the intercalated character of these deposits. The interbedded claystone, mudstone, and siltstone units were deposited in flood-plain areas on the delta plain. These units commonly coarsen upward into tabular sandstone units of crevasse splay origin indicating the gradual encroachment of active channel environments into parts of the flood plain. Locally flood-plain units are erosionally overlain by lenticular sandstone units of channel origin. Abandonment and avulsion of channels is indicated by claystone or mudstone units sharply overlying channel sandstone units. The geometry, grain size, bedding thickness, and sedimentary structures within the tabular sandstone units indicate deposition in crevasse splays having generally unconfined flow and lower flow regime conditions and less common upper flow regime conditions. The thicker and coarser grained the unit, generally the closer it was to the active channel. The lenticular, erosional-based sandstone units formed as distributary channel fills. Lack of accreting point bar structures indicates low-sinuosity channel systems. Relatively straight channels filled primarily by vertical accretion before avulsion and the abandonment of the old channel.

Large-Scale Channel Deposits

In some cycles, the delta-front sandstone interval is overlain by a sandstone unit that has a sharp or erosional base and a fining-upward grain-size trend. These overlying sandstone units can be divided into two categories: (1) those that are dominantly fine to medium grained with rare pebbles and show evidence of marine influence and (2) those that are medium to coarse grained and commonly pebbly and show no evidence of marine influence.

Sandstone units in the first category have basal contacts that are sharp and flat or erosional. Where erosional, the maximum relief along the surface is difficult to determine due to limited lateral exposure. There is commonly an abrupt change in grain size across the basal contact; mudstone rip-up clasts occur in a matrix of fine to medium and locally coarse grained sandstone. In most units, grain size fines upward. The beds that comprise these units are medium to thick in the lower part and grade upward to thin and very thin. Trough and tabular cross-stratification are dominant in the lower part of the unit. Tabular beds locally show reversal in foreset orientation; reactivation surfaces between beds are also common, especially near the base of the unit. In the upper part of these units, asymmetric and symmetric ripple cross-stratification dominate; flaser bedding, small-scale trough cross-stratification, and planar laminations are locally interbedded with the rippled beds. Horizontal burrows are common to locally abundant. Small, flat, claystone rip-up clasts along bedding and foreset surfaces and laminations of clay draping sandstone beds are common in some units. Fossil fragments are also locally in the basal part of units.

Sandstone units in the first catagory are interpreted as tidal-channel or distributary-channel deposits. The fining-upward grain size trend and decrease in bed thickness indicates shallowing under decreasing energy conditions, which characterize channel deposits. Reversal of cross stratification, alternating deposition of sandstone beds and clay lamination, and erosion of clay laminations as indicated by deposition of small clay clasts along foresets and bedding planes indicate alternating energy levels and current directions that characterize tidal environments. Tidally influenced environments occur most commonly in the uppermost part of the Hermosa Group. Units that do not show tidal features, but have these channel-fill characteristics and are laterally and vertically associated with delta-front and delta-plain deposits, are interpreted to be distributary channel deposits.

The second category of fining-upward sandstone units is distinguished by their much coarser grain size, the common occurrence of pebble- to boulder-sized clasts, and the lack of evidence of marine influence. Exposures on cliff faces and units traced laterally showed as much as 5 to 6 m (15–20 ft) of relief along the basal contacts. These units range from 3 to 16 m (10–52 ft) thick, internally contain 2–5 or more subunits separated by laterally persistent scour surfaces, and exhibit tabular geometries. Claystone rip-up clasts from 0.5 to 30 cm (0.2–12 in.) in length are concentrated near the base; dolostone rip-up clasts as long as 80 cm (2.75 ft) occur in the basal part of the unit that extends from 237 to 255 m (777–838 ft) in the measured section. Pebbles, composed primarily of chert and quartz and less commonly of feldspar, sandstone, and granite, occur as pebbly sandstone at the base of many units and as pebble lenses concentrated at the base of trough cross sets throughout a unit. Average pebble size is 1 to 3 cm (0.4–1 in.); locally clasts as large as 10 cm (4 in.). The dominant grain size is coarse to medium. Grain size is generally consistent throughout a unit or is slightly finer in the uppermost part of a unit. Bedding commonly shows a slight upward decrease in bed thickness from very thick to thick in the lower part and thick to medium in the upper part of a unit. Trough and tabular cross-stratification are the dominant bedding types; in each unit, one type is usually dominant. Planar laminations occur locally in the lower part of a unit or are interbedded with crossbeds in the upper part of a unit. Ripple cross-stratification, where present, is confined to the uppermost part of a unit.

Sandstone units in the second catagory are interpreted as channel-fill deposits that formed on the alluvial plain. The internal bedding structures and grain size profiles suggest deposition in low-sinuosity, bed-load rivers. The locally extensive scour at the base, abrupt change in grain size across the basal contact, and abrupt change in energy conditions and depositional environments from the underlying deposits suggest deposition after a period of erosion caused by a base-level drop. The regional extent of each erosion surface cannot be documented due to the extensive cover; therefore, estimating the scale of base-level change that caused each event is very speculative. The largest magnitude, and possibly most regionally extensive, base level drops occurred prior to deposition of two channel-fill units: one from 237 to 255 m (777–838 ft) and the other from 637 to 652 m (2095–2140 ft). The thickness, coarse grain size, and lateral continuity of these deposits suggest that the long-lived, high-energy braided rivers established subsequent to the erosion events were not localized drainage systems but represent a more regional, integrated drainage system. Marine carbonate beds lie directly on these alluvial channel-fill deposits or are separated from them by about 1 m (3 ft) of covered section. This succession of units indicates fluvial deposition during a base-level rise and the termination of fluvial deposition during transgression.

Clastic Depositional Facies Distribution

The Hermosa Group can be divided into four gross subdivisions based on the percentage of depositional facies that compose the clastic parts of the cycles. From 101 to 395 m (330–1295 ft), the clastic parts of cycles are composed primarily of delta-plain, and in some cycles alluvial-plain, deposits; prodelta and delta-front deposits are thin and not abundant. From 434 to 610 m (1425–2000 ft), the clastic part of cycles are prodelta and delta-front deposits. Because of abundant cover in this interval, it is unknown if other clastic lithofacies are present. However, these prodelta and delta-front deposits are thicker than those found in other parts of the Hermosa Group, suggesting that during this interval conditions were more favorable for accumulation of these deposits. From 610 to 652 m (2000–2140 ft), delta plain and alluvial deposits predominate; delta-front deposits are absent or very thin in cycles. From 677 to 843 m (2220–2765 ft), the clastic part of cycles are composed primarily of tidally influenced marine deposits and delta-plain deposits and common but slightly less abundant alluvial deposits.

Variations in the dominant depositional facies in cycles throughout the Hermosa Group probably reflect changes in magnitude of relative sea level that controlled the cyclicity. These relative changes resulted from a combination of long-term changes in subsidence rates, both local and regional, and changes in the magnitude of eustatic cycles.

CLASTIC PETROGRAPHY

DETRITAL MINERALS

Sandstones in the Hermosa Group are composed of quartz, feldspar, rock fragments, and minor amounts of matrix and miscellaneous minerals. In general, the detrital minerals vary in abundance from sample to sample but show little systematic variation in the group. Most sandstones are poorly to moderately sorted and show slight to moderate mechanical compaction effects. On the basis of the classification scheme of Folk (1974), sandstones in the Hermosa are predominantly arkoses and litharenites. However, a significant number of samples have sufficient quartz to be subarkoses and sublitharenites.

Detrital quartz is a dominant constituent in most sandstones and is present as single and semi-composite to composite (aggregate) grains. The ratio of monocrystalline quartz to composite quartz increases with decreasing grain size. Single quartz grains average 35 percent of the detrital fraction and are angular to subrounded in shape and fine to coarse in size. Grain contacts range from long to suture, but concave-convex contacts are most common. Quartz grains generally are monocrystalline with straight extinction; some grains contain vacuoles and small inclusions. Secondary overgrowths on detrital grains are rare and only occur in coarser-grained sandstones.

Feldspar is widespread in most sandstones (19 percent average) and occurs as single monomineralic grains or as coarse-crystalline constituents in igneous and metamorphic rock fragments. As a group, feldspars vary in morphology and degree of alteration. Many grains are irregularly shaped, angular to subangular fragments, and are of a size that is similar to quartz grains in the same sample. Two distinct feldspar types, plagioclase and potassium feldspar, were differentiated on the basis of twinning, cleavage, alteration effects and staining by sodium cobaltinitrite. Plagioclase feldspar is unstained and occurs as twinned and untwinned grains. The absence of twinning in some sandstones may be due to grain orientation. In general, grains showing albite twinning are fresher than untwinned grains. A large fraction of plagioclase shows some signs of weathering or chemical alteration. Many grains display a cloudy appearance due to incipient clay mineral replacement whereas others show partial replacement by calcite. Overall, plagioclase content averages about 5 percent by volume although locally it comprises as much as 30 volume percent. Potassium feldspar stains bright yellow and is distributed as discrete twinned and untwinned fragments (orthoclase, perthite, microcline) or as irregular fragments in composite igneous grains. Many single grains appear to have derived from the breakup of granitic rock fragments indicating they are of plutonic origin. Potassium feldspar is most abundant (13 percent average) in coarser-grained alluvial sandstones; the mineral is virtually absent in the lower part of the group.

Igneous, metamorphic, and sedimentary rock fragments constitute a significant proportion (12 percent average) of the detrital mineral fraction in Hermosa sandstones. Rock fragments are best preserved in coarser-grained alluvial sandstones. Metamorphic grains constitute the majority of the lithic fragments and consist of polycrystalline (semicomposite to composite) quartz. Abundances vary from 0 to 60 percent and average 9 percent. Polycrystalline quartz is composed of angular, tightly interlocking crystals and can be difficult to distinguish from metaquartzite fragments. In some sandstones, polycrystalline quartz grains are intergrown with crystals of potassium feldspar; these quartzo-feldspathic grains are of igneous (plutonic) origin. Sedimentary rock fragments are a minor component (less than 5 percent) of the lithic assemblage and are difficult to classify because of their fine particle size. The main sedimentary lithic constituents are argillaceous clasts, clastic carbonate grains, and chert fragments. Argillaceous (mudstone and shale) clasts occur as discrete rounded grains and are in various stages of disaggregation. Mudstone and shale are mechanically and chemically unstable, thus they commonly are deformed between more rigid framework grains. Sparse fine-grained chert is composed of micron-sized subequant crystals. The scarcity of siliceous constituents with gradational crystal sizes allows a clear distinction to be made between chert and quartzite. Chert grains are rounded in size and are fresh to weathered in appearance; relict fossil structures were not observed. Weathered fragments are light to medium brown due to clay-mineral alteration. A few sandstones contain clastic grains of fine-crystalline dolomite; locally, these grains are ferroan. Most dolomite grains show varying degrees of rounding and abrasion, which attests to their detrital origin.

Matrix material can be widespread in prodelta-, delta-front, and delta-plain sandstones. Commonly, detrital matrix is difficult to distinguish from "pseudomatrix" which formed by compaction of labile grains (mudstone and shale). The observed particle size distribution in sandstones thus may bear little resemblance to the initial grain size at the time of deposition. Identification of matrix also may be difficult if authigenic clay minerals (such as, chlorite and illite) are complexly intergrown with this detrital material. As in pseudomatrix, detrital matrix tends to inhibit both cementation and porosity development because of its low permeability.

Miscellaneous detrital constituents in sandstones consist of biotite and muscovite, rare glauconite, and minor unidentified fossil fragments, among others. Of these constituents, mica is most prevalent. In prodelta and delta-front sandstones, biotite tends to concentrate along bedding planes and constitutes as much as 10 volume percent of the detritus. In a few sandstones, reddish brown to dark brown iron oxide (hematite?) is present in sparse amounts. It is unevenly distributed as a coating on detrital grains.

AUTHIGENIC MINERALS

Sandstones in the Hermosa Group have been affected by varying degrees of postdepositional mineral alteration. Major authigenic mineral phases include (1) quartz, (2) multiple generations of carbonate, (3) sulfate minerals, and (4) a clay mineral assemblage.

Authigenic quartz is sparse (< 1 percent) in most sandstones. Where it is present, it forms coarse idiomorphic crystals adjacent to granitic rock fragments and polycrystalline quartz grains. Silica cementation occurred during early diagenesis, before the emplacement of other mineral cements.

A carbonate-mineral assemblage composed of ferroan and nonferroan calcite and dolomite occurs throughout most of the stratigraphic interval. Formation of carbonate was an early- to late-stage diagenetic event. Locally, carbonate minerals comprise as much as 30 volume percent although all phases are not present in every sample. The proportion of noncarbonate cements in sandstones is inversely proportional to the amount of preserved carbonate. In general, sandstones with large amounts of carbonate cement have only trace amounts of noncarbonate cement. Calcite in sandstones forms an intergranular cement and partly replaces detrital framework grains. Locally, calcite is slightly ferroan. Where calcite is widespread, it occurs as large optically continuous poikilotopic crystals in pores. Poikilotopic calcite commonly is pervasive in sandstones that show high initial porosity. In sandstones where authigenic chlorite is abundant, calcite is formed peripheral to chlorite indicating the emplacement of calcite postdated the development of chlorite. Dolomite generally has a more restricted distribution than calcite. Most dolomite forms an anhedral cement that replaces calcite, which implies that the formation of dolomite followed that of calcite. Like calcite, authigenic dolomite may be ferroan. In one sandstone that has abundant ferroan dolomite cement (sample no. PH91KF83) there are numerous clastic ferroan-dolomite grains. These grains appear to have provided nucleation sites for extensive carbonate growth.

Authigenic anhydrite and barite occur sparsely in some sandstones. The mode of occurrence of these minerals ranges from small scattered patches to large poikilotopic crystals that preferentially replace mineral cements and framework grains such as feldspar. Grain to grain relations suggest that sulfate minerals formed relatively late in the diagenetic history.

The authigenic clay minerals illite and chlorite, and interstratified illite-chlorite, occur in various amounts in Hermosa sandstones. They are most abundant in prodelta, delta front, and delta plain sandstones. Authigenic illite is distributed as an alteration product on sedimentary rock fragments and feldspar (predominantly plagioclase), and it occurs as a fibrous mineral in pores. In some sandstones, authigenic illite may be associated with matrix matter, thus the two minerals may be difficult to distinguish from one another. Authigenic chlorite occurs predominantly as pseudohexagonal platelets oriented perpendicular to grain surfaces and pore edges. Where chlorite platelets are well developed, they form isopachous tangential rims that line pores and coat grains except where the grains are in contact with one another. Pores lined with chlorite commonly are open indicating that early chlorite grain coats may have inhibited the formation of mineral cements. *In situ* alteration of ferromagnesian minerals and shales undergoing compaction may have provided the ions necessary for clay mineral development.

Porosity is best developed in coarser grained alluvial sandstones; most porosity probably is of secondary origin. Intergranular porosity is the predominant porosity type although microporosity is common in sandstones that have partially to completely clay filled pores. In some sandstones adjacent to carbonate strata, carbonate cement occludes primary pores. In other sandstones proximal to carbonate beds, residual pore space is preserved. In these beds, pre-existing carbonate cement may have been dissolved thereby enhancing porosity.

DEPOSITIONAL CONTROLS ON SANDSTONE COMPOSITION

Environment of deposition influenced the framework composition of sandstones in the Hermosa Group. The clastic deposits in the Hermosa were deposited in both marine and nonmarine settings in environments ranging from deltaic to alluvial. The samples studied petrographically are representative of several of these environments. In finer grained prodelta and delta-front sandstones, detrital matrix and biotite is moderate to abundant, and metamorphic and igneous rock fragments are virtually absent. In contrast, coarser grained alluvial sandstones have abundant metamorphic and igneous rock fragments are abundant and rare matrix. A similar correlation exists between framework grain size and pore space. Coarser grained alluvial sandstones have more porosity than finer grained sandstones.

The relative abundance of authigenic mineral cements in sandstones in part reflects the lithology of adjacent beds. For example, sandstones adjacent to carbonate strata tend to have substantial calcite and dolomite cement which suggests that precipitation was influenced by fluids sourced from carbonate rocks. Fluids may have migrated intermittently resulting in multiple precipitation events. The timing of sulfate mineral emplacement is uncertain, but the source of fluids most certainly was from gypsum interbeds in the group.

SUMMARY

At Hermosa mountain, we divide the Pennsylvanian Hermosa Group into the Pinkerton Trail Formation and the overlying undifferentiated part of the Hermosa Group. As evaporite beds are absent in nearly all of the outcrop area along the eastern margin of the Paradox Basin and subsurface control to the west is sparse to absent, the name Paradox Formation is not applied at Hermosa mountain or along the outcrop belt. Correlating the Hermosa mountain section to subsurface wells, the closest of which is 16 km (10 mi) to the southeast, the interval from 33.5 to at least 599 m (110–1965 ft) in this section appears to be the approximate stratigraphic equivalent of the Alkali Gulch through Ismay intervals of the Paradox Formation. The thickness and stratigraphic position of the gypsum interval in the Hermosa mountain section suggests that it correlates with Hite's (1960) cycle 6, which is the most regionally extensive evaporite unit in the Paradox Basin. At the Hermosa mountain locality, we did not assign the name Honaker Trail Formation to part of the Pennsylvanian section because of the absence of a lithologically definitive Paradox Formation and the uncertainties in correlations with the more basinward evaporite cycles. On the basis of our correlations of the Hermosa mountain section to the subsurface, the stratigraphic equivalent to the Honaker Trail Formation in the measured section extends from 599 m (1965 ft) or slightly higher to the top of the Hermosa Group. The Hermosa Group at Hermosa mountain has a gradational contact with the overlying alluvial red beds of the Cutler Formation. The top of the Hermosa is placed at the top of the highest carbonate bed of probable marine origin. We have chosen not to apply the name Rico Formation in the Hermosa mountain area because of the greatly varying criteria geologists have used in defining the Rico Formation and the difficulty of mapping a unit based solely on the character of faunal assemblages, which was how Spencer (1900) originally defined the Rico. There is excellent correspondence in both the total thickness and the gross lithologic subdivisions of the Hermosa Group between Hermosa mountain and the exposures to the north in the San Juan Mountain area, although the nomenclature used here differs from Spoelhof's (1974) nomenclature in the San Juan outcrop area. Spoelhof's (1974) limestone-dominated, middle member of the Honaker Trail is equivalent to the interval in the Hermosa mountain section from 588 to 677 m (1930–2220 ft). Along the eastern basin margin, this laterally extensive interval contains the most abundant and thickest limestone units in the Hermosa Group exclusive of those in the Pinkerton Trail.

On the basis of the brachiopod faunas, the entire Hermosa Group at its type locality at Hermosa mountain is primarily of Desmoinesian age, ranging from the earliest Desmoinesian through at least early late Desmoinesian. The brachiopods identified in the lowermost beds of the Pinkerton Trail suggest that these beds are early Desmoinesian in age; no brachiopods noted by Sutherland and Harlow (1973) as confined to rocks of Atokan age occur in them. The brachiopod *Antiquatonia portlockiana* occurs approximately 46 m (150 ft) below the Hermosa-Cutler contact; and the chonetid *Mesolobus mesolobus*, which is confined to rocks of Desmoinesian age wherever it has been found, occurs approximately 98 m (320 ft) below the Hermosa-Cutler contract. Thus, the uppermost beds of the Hermosa are probably no younger than Desmoinesian, although the possibility of an early Missourian age cannot be eliminated as *A. portlockiana* extends into the lowermost Missourian in other regions, such as Arizona.

The undifferentiated part of the Hermosa Group above the Pinkerton Trail Formation at Hermosa mountain consists of stacked depositional cycles that usually contain a carbonate unit at the base overlain by a upward coarsening clastic interval. There are at least 28 and, by interpolation in the covered intervals, possibly 40 depositional cycles that range in thickness from 3 to 45 m (10–148 ft). The base of a cycle is chosen at the base of the carbonate unit because of the lithologic contrast and sharp contact between clastic and carbonate units. The contact between the carbonate and overlying clastic unit is either sharp or slightly gradational over less than a 0.5 m (1.5 ft) thick interval. The clastic part of the cycle overlying the carbonate unit shows an overall coarsening upward trend from mudstone or shale, if present, into thinly interbedded siltstone, mudstone, and very fine grained sandstone that grades upward into fine-grained sandstone. This sandstone may be the uppermost unit of a cycle, or is either gradationally to sharply overlain by a series of interbedded siltstone, mudstone, and tabular to lenticular sandstone units or sharply and irregularly overlain by a medium- to coarse-grained to pebbly sandstone interval that has a fining-upward trend.

Four distinct carbonate lithofacies are identified in the Hermosa Group at Hermosa mountain based on mineralogy, texture, biotic assemblages, and inferred depositional environment. This division is distinct from those documented in other areas the Pennsylvanian outcrop belt in the eastern part of the Paradox Basin. Lithofacies no. 1 is from 8 to 34 m (25–110 ft) and corresponds to the Pinkerton Trail Formation; carbonate deposition occurred in high-energy subtidal areas proximal to the shoreline and in low-energy shelf environments. Lithofacies no. 2 extends from 119 to 399 m (390–1310 ft); carbonate deposition was exclusively in arid, hypersaline conditions in the intertidal to supratidal zone. Lithofacies no. 3 extends from 485 to 599 m (1590–1965 ft); carbonate deposition was in restricted, possibly deep, marine waters and in normal marine waters in the subtidal zone on carbonate platforms. Lithofacies no. 4 extends from 599 to 843 m (1965–2765 ft); carbonate deposition varied from low-energy subtidal carbonate platform environments to very shallow subtidal and intertidal zones to restricted intertidal zones. Except for the uppermost part of the Hermosa Group, carbonate environments in the Hermosa mountain area differed slightly to significantly from synchronous

carbonate environments to the north along the eastern margin of the Paradox Basin.

The clastic deposits in a cycle formed in marine deltaic and nonmarine deltaic and alluvial environments. In some cycles, pebbly braided sandstone units were deposited after a period of erosion caused by a base-level drop. The scale of base-level change that caused each event cannot be accurately determined because the regional extent of each erosion surface cannot be ascertained. The succession of marine carbonate units directly overlying the alluvial units indicate fluvial deposition during a base-level rise and the termination of fluvial deposition during transgression. Although the clastic part of cycles show great similarities in many respects, the percentage of depositional facies composing the clastic part of each cycle can vary greatly. The undifferentiated part of the Hermosa Group can be divided into four gross subdivisions based on the percentage of clastic depositional facies that compose the cycles. From 101 to 395 m (330–1295 ft), cycles are composed primarily of delta-plain, and in some cycles alluvial-plain, deposits; prodelta and delta front deposits are thin and compose a minor percentage of the cycles. From 434 to 610 m (1425–2000 ft), prodelta and delta-front deposits compose the entire part of a cycle; other clastic lithofacies may be present in the covered intervals. From 610 to 652 m (2000–2140 ft) delta-plain and alluvial deposits predominate; delta-front deposits are absent to very thin in cycles. From 677 to 843 m (2220–2765 ft) cycles are composed primarily of tidally influenced marine deposits and delta-plain deposits and common, but slightly less abundant, alluvial deposits. These changes in the dominant depositional facies in cycles through the Hermosa Group probably reflect changes in magnitude of relative sea level change that controlled the cyclicity.

Sandstones in the Hermosa Group are predominantly arkoses and litharenites composed of quartz, feldspar, and rock fragments with minor amounts of matrix and miscellaneous minerals. Potassium feldspar is virtually absent in the lowest part of the formation. Depositional environment influenced the framework composition of sandstones. In finer grained prodelta and delta-front sandstones, detrital matrix and biotite is moderate to abundant, and metamorphic and igneous rock fragments are virtually absent. In coarser grained alluvial sandstones, metamorphic and igneous rock fragments are abundant and matrix is rare. Porosity, of which the intergranular type dominates, is greatest in the coarse-grained alluvial sandstones and is mostly of secondary origin. The relative abundance of authigenic mineral cements in sandstones in part reflects the lithology of adjacent beds. Sandstones adjacent to carbonate strata tend to have substantial calcite and dolomite cement, which suggests that precipitation was influenced by fluids sourced from carbonate rocks. Fluids derived from gypsum interbeds in the formation precipitated sulfate minerals, but the timing of their emplacement is uncertain.

REFERENCES CITED

Baars, D.L., 1962, Permian system of Colorado Plateau: American Association of Petroleum Geologists Bulletin, v. 46, no. 2, p. 149–218.

Baker, A.A., 1936, Geology of the Monument Valley-Navajo Mountain region, San Juan County, Utah: U.S. Geological Survey Bulletin 865, 106 p.

Baker, A.A., Dane, C.H., and Reeside, J.B., Jr., 1933, Paradox formation of eastern Utah and western Colorado: American Association of Petroleum Geologists Bulletin, v. 17, no. 8, p. 963–980.

Bass, N.W., 1944, Correlation of basal Permian and older rocks, southwest Colorado, northwest New Mexico, northeast Arizona, and southeast Utah: U.S. Geological Survey Oil and Gas Investigations Chart 7.

Blodgett, R.H., comp., 1984, Road log from Durango to Telluride, Colorado, via Mancos, Dolores, and Rico, Colorado, *in* Brew, D.C., ed., Field trip guidebook: Four Corners Geological Society, 37th Annual Meeting Rocky Mountain Section Geological Society of America, p. 62–74.

Brew, D.C., 1979, Biostratigraphy of the Naco Formation (Pennsylvanian) in central Arizona: American Geological Institute Selected Guidebook Series 1, p. 119–125.

Fetzner, R.W., 1960, Pennsylvanian paleotectonics of Colorado Plateau: American Association of Petroleum Geologists Bulletin, v. 44, no. 8, p. 1371–1413.

Flugel, E., 1982, Microfacies analysis of limestones: New York, Springer-Verlag, 633 p.

Folk, R.L., 1974, Petrology of sedimentary rocks: Austin, Hemphill, 182 p.

Franczyk, K.J., 1992, Measured section of the Pennsylvanian Hermosa Group Near Hermosa, Colorado: U.S. Geological Survey Open-File Report 92–689, 33 p.

———1993, Measured section of the Pennsylvanian Hermosa Group near Ouray, Colorado: U.S. Geological Survey Open File Report 93–712, 25 p.

Herman, George, and Sharps, S.L., 1956, Pennsylvanian and Permian stratigraphy of the Paradox salt embayment: Intermountain Association of Petroleum Geologists Field Conference, 7th, p. 77–84.

Hite, R.J., 1960, Stratigraphy of the saline facies of the Paradox Member of the Hermosa Formation of southeastern Utah and southwestern Colorado, *in* Geology of the Paradox Basin fold and fault belt: Four Corners Geological Society Field Conference, 3rd, p. 86–89.

Hite, R.J., Anders, D.E., and Ging, T.G., 1984, Organic-rich source rocks of Pennsylvanian age in the Paradox Basin of Utah and Colorado, *in* Woodward, J., Meissner, F.F., and Clayton, J.L., eds,, Hydrocarbon source rocks of the greater Rocky Mountain region: Denver, Rocky Mountain Association of Geologists, p. 255–274.

Hite, R.J., and Buckner, D.H., 1981, Stratigraphic correlations, facies concepts, and cyclicity in Pennsylvanian rocks of the Paradox Basin, *in* Weigand, D.L., ed., Geology of the Paradox Basin: Rocky Mountain Association of Geologists Field Conference, 1981, p. 147–159.

Huffman, A.C., Jr., and Condon, S.M., 1993, Stratigraphy, structure, and paleogeography of Pennsylvanian and Permian rocks, San Juan Basin and adjacent areaas, Utah, Colorado, Arizona, and New Mexico: U.S. Geological Survey Bulletin 1808, Chapter O, 44 p.

Irwin, C.D., 1976, Paleozoic subsurface correlation in the Paradox Basin, Colorado and New Mexico: Denver, Rocky Mountain Association of Geologists Paleozoic Correlation Chart, Sheets 11, 13.

Loope, D.B., Sanderson, G.A., and Verville, G.J., 1990, Abandonment of the name "Elephant Canyon Formation" in southeastern Utah—physical and temporal implications: The Mountain Geologist, v. 27, no. 4, p. 119–130.

McKnight, E.T., 1940, Geology of area between Green and Colorado Rivers, Grand and San Juan Counties, Utah: U.S. Geological Survey Bulletin 908, 147 p.

Pettijohn, F.J., Potter, P.E., and Siever, R., 1972, Sand and sandstone: New York, Springer Verlag, 618 p.

Petuch, E.J., 1988, A new Pholadomyid fauna from the Permian of southwest Colorado: Bulletin of Paleomalocology, v. 1, p.43–48.

Powers, M.C., 1953, A new roundness scale for sedimentary particles: Journal of Sedimentary Petrography, v. 23, p. 117–119.

Ross, C.A., 1969, Middle and Upper Pennsylvanian Fusulinaceans, Gila Mountains, Arizona: Journal of Paleontology, v. 43, p.1405–1422.

Ross, C.A., 1973, Pennsylvanian and Early Permian depositional history, southeastern Arizona: American Association of Petroleum Geologists Bulletin, v. 57, p. 887–912.

Ross, C.A., and Sabins, F.F., Jr., 1965, Early and Middle Pennsylvanian Fusulinids from southeast Arizona: Journal of Paleontology, v. 39, p. 173–209.

Roth, R.I., 1934, Type section of Hermosa Formation, Colorado: American Association of Petroleum Geologists Bulletin, v. 18, no. 7, p. 944–947.

Spencer, A.C., 1900, The sedimentary formations, Chapter 2, *in* Cross, W. and Spencer, A.C., Geology of the Rico Mountains, Colorado: U.S. Geological Survey, Twenty-First Annual Report, part 2, p. 37–88.

Spoelhof, R.W., 1974, Pennsylvanian stratigraphy and tectonics in the Lime Creek-Molas Lake area, San Juan County, Colorado: Golden, Colorado School of Mines, Ph.D. Thesis, 193 p.

Sutherland, P.K., and Harlow, F.H., 1973, Pennsylvanian brachiopods and biostratigraphy in southern Sangre de Cristo Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources Memoir 27, 173 p.

Terry, R.D. and Chilingar, G.V., 1955, Summary of "Concerning some additional aids in studying sedimentary formations" by M.S. Shvetsov: Journal of Sedimentary Petrography, v.25, p. 229–234.

Waddell, D.E., 1966, Pennsylvanian fusulinids in the Ardmore Basin, Love and Carter Counties, Oklahoma: Oklahoma Geological Survey Bulletin 113, 128 p.

Wengerd, S.A., and Matheny, M.L., 1958, Pennsylvanian system of Four Corners region: American Association of Petroleum Geologists Bulletin, v. 42, no. 9, p. 2048–2106.

Wengerd, S.A., and Strickland, J.W., 1954, Pennsylvanian stratigraphy of Paradox salt basin, Four Corners region, Colorado and Utah: American Association of Petroleum Geologists Bulletin, v. 38, no. 10, p. 2157–2199.