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DEPOSITIONAL ENVIRONMENTS OF THE PERMO-PENNSYLVANIAN SANGRE DE CRISTO FORMATION, COYOTE CREEK DISTRICT, MORA COUNTY, NEW MEXICO

C. ELMO BROWN
Placid Oil Company, 410 17th Street, Suite 2000, Denver, Colorado 80202

INTRODUCTION

The Sangre de Cristo Formation is a Pennsylvanian—Permian elastic unit which was deposited in the Paleozoic Rowe—Mora Basin of northern New Mexico and south-central Colorado. Bachman and Read (1952), Zeller and Baltz (1954), Tschanz and others (1958), and May and others (1977) studied the Sangre de Cristo Formation in the Coyote Creek district, an inactive copper-mining area located approximately 8 mi northeast of Mora and 2 mi southeast of Guadalupita in Mora County, north-central New Mexico (Fig. 1). These earlier studies focused on the uranium-rich zones within the formation. A detailed description of the entire Sangre de Cristo Formation within the district, however, was lacking.

The purpose of this study is to describe and classify the various depositional units of the Sangre de Cristo Formation found within the Coyote Creek district. Classification is based on lateral and vertical variations in geometry, lithology, texture, and sedimentary structure. Variations are documented by measured sections, sketches, and sediment-transport direction measurements from better outcrop exposures. From this information, a depositional history can be formulated.

Present Structural Setting

The Coyote Creek district lies within a Laramide imbricate-thrust zone which marks the eastern edge of the Sangre de Cristo Mountains (May and others, 1977). A major fault, with displacement of several thousand feet, separates Precambrian metasedimentary rocks from Pennsylvanian sedimentary rocks (Fig. 2). The structural strike of the sedimentary rocks and minor thrust faults found within them is generally to the north. Dip of beds in the vicinity of the thrust zone varies from nearly vertical to slightly overturned (Zeller and Baltz, 1954). Near the center of Coyote Creek valley, dips abruptly flatten to 7-14° east. Zeller and Baltz (1954) mapped a thrust fault separating the steep dips on the west from the flatter dips to the east; however, no evidence for this fault was found on measured section SDC-1. It appears that only folding of the rocks occurred in this part of the valley.

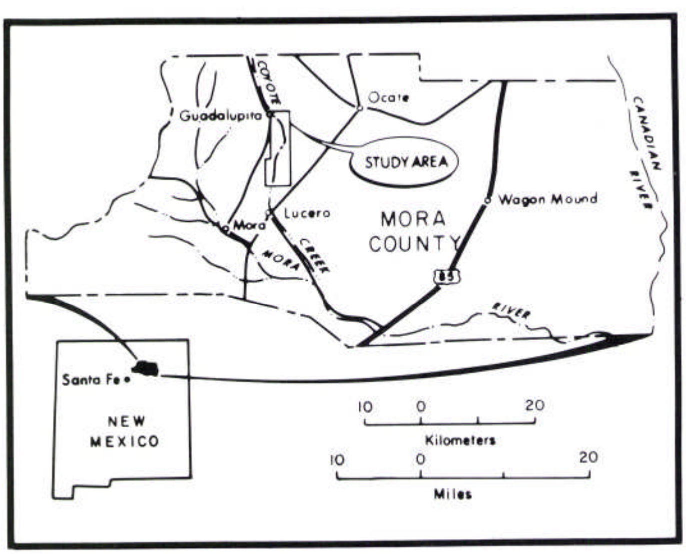

FIGURE 1. Location of study area (modified from Tschanz and others, 1958).

DEPOSITIONAL SYSTEMS OF THE SANGRE DE CRISTO FORMATION

Tschanz and others (1958) divided the Sangre de Cristo Formation in the Coyote Creek district into six lithologic units (Fig. 3). Each unit is characterized by many abrupt facies changes and by interfingering unit boundaries which are difficult to delineate. It is uncertain how persistent these units are due to the lack of detailed mapping.

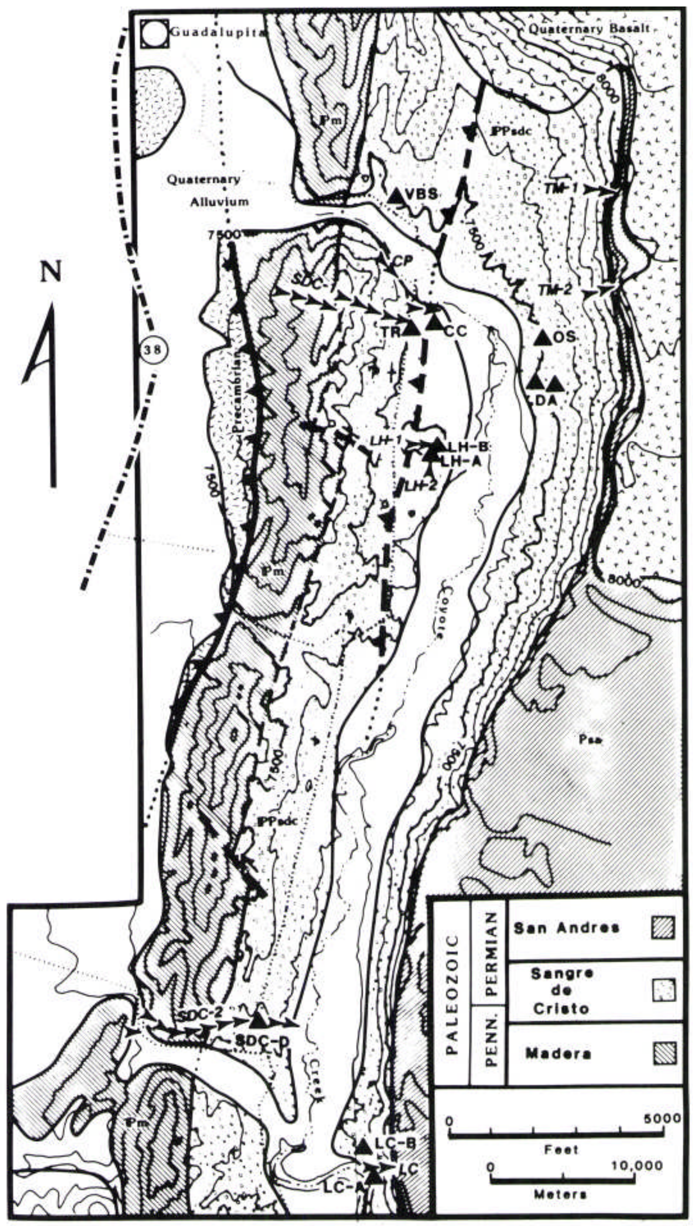

FIGURE 2. Topographic and geologic map of Coyote Creek district showing measured section and sketch locations (modified from Zeller and Baltz, 1954).

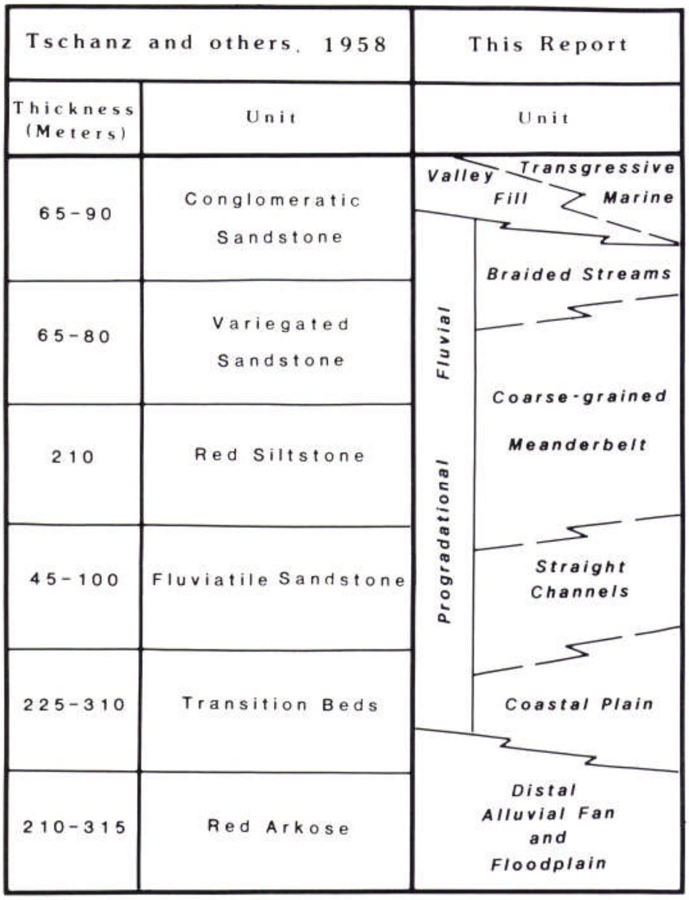

FIGURE 3. Classification of Sangre de Cristo units within Coyote Creek district.

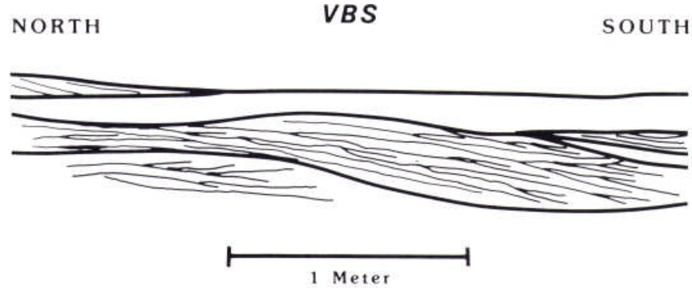

FIGURE 4. Sketch of distal alluvial-fan, sheetflood sandstone. In this and all following figures, blacked-in boxes denote clay clasts and open circles denote gravel.

For the purpose of this study, the Sangre de Cristo Formation is divided into three genetically different depositional systems. In ascending order they are: (1) distal alluvial fan with associated floodplain. (2) progradational fluvial system, and (3) shallow-marine bar and valley fill.

DISTAL ALLUVIAL-FAN AND ASSOCIATED FLOODPLAIN SYSTEM

Deposits of the distal alluvial-fan and associated floodplain depositional system occupy the lower 400-650 m of the formation where outcrop attitude is nearly vertical to slightly overturned. Approximately 30% of the system consists of red to light-brown, arkosic-sandstone beds which form distinct parallel ridges. Separating these ridges are wide, covered areas which represent finer-grained facies.

Distal Alluvial-fan Facies

Ridge-forming sandstones in the lower part of the Sangre de Cristo Formation are coarse- to very coarse-grained, submature, subrounded to angular arkoses. Some of the beds have pebbles of granite and feldspar, either scattered throughout or concentrated along scour surfaces. Angular, unweathered potassium-feldspar grains indicate short transport distances.

The most common sandstone type is represented by sheetlike arkoses which are dominated by small-scale, low-angle, trough crossbeds with very little basal scour (Fig. 4). Pebbles are often scattered throughout with occasional heavy minerals or pebbles lining trough scours. Large grain size and lack of textural variability make internal stratification difficult to see. Paleocurrent direction is interpreted to be to the south, parallel to trough axes.

The small scale and nonerosional nature of the crossbedding suggest that these numerous sandstone packages were deposited in fairly shallow water with very low turbulence and high sedimentation rates, indicating sheetflood deposition (Collinson, 1978). These sheetflood sediments are interpreted to represent deposition in a southward-prograding distal alluvial-fan environment similar to the terminal-fan model defined by Friend (1978). The terminal-fan model is characterized by a distributary channel pattern (Allen, 1965) which eventually disappears down-fan. This loss of channelization is produced by a decrease in discharge caused by several factors including water loss into permeable alluvium and a high evapotranspiration rate. A high evapotranspiration rate would also inhibit bank stabilizing plant growth in the distal fan while increasing the rate of runoff and sediment transport in the source area.

It is believed that when sediment-loaded floodwaters emerged from the channelized system to the unconfined distal alluvial fan, water depth and velocity were reduced, thus decreasing the capacity of the water to carry sediment. This resulted in rapid sheetflood sedimentation.

Floodplain Facies

Approximately 70% of the distal alluvial-fan and associated floodplain system is fine-grained sediment. Even though exposures are limited, lithologies appear to be primarily orange to red, micaceous, silt and sandy silt with some thin beds of limestone, maroon and brown shale, and red, coarse-grained sandstone. The maroon and brown shales occur commonly near the contact with the overlying fluvial system and are often lenticular. Limestone beds are light gray to reddish gray, often silty, unfossiliferous, very hard, nodular micrites. These beds increase in number and thickness to the south and are interpreted as having formed in "K"-horizon caliche zones (tile and others, 1965) or in highly alkaline playa lakes similar to those described by Goudie (1973). In either case, micrite precipitation requires an environment with infrequent elastic sedimentation.

Friend (1978) stated that the bulk of distal terminal-fan deposits is usually siltstone and calcareous marlstone. Therefore, fine-grained sediments within this lower interval of the Sangre de Cristo Formation are interpreted as floodplain and playa sediments associated with the distal alluvial-fan sheetflood sands.

PROGRADATIONAL FLUVIAL SYSTEM

The middle 500-700 m of the Sangre de Cristo Formation in the study area is interpreted as having been deposited in a bedload-dominated fluvial system. This system is divided into two facies assemblages. The lowermost is a coastal-plain sequence where straight to slightly meandering fluvial-channel sandstones constitute only a minor percentage of the primarily fine-grained package. Gradationally overlying this facies assemblage is a channel-dominated interval which prograded over the coastal-plain sequence. Only a minor percentage of fine-grained sediment is preserved in the channel-dominated system. It is believed that this upward decrease in fine-grained sediment is a result of an increase in channel-migration rate and a decrease in aggradation rate.

Coastal-plain-fades Assemblage

This 130-m-thick facies assemblage consists of approximately 70% silt and shale, with lesser amounts of limestone, fine- to medium-grained sandstone, and very coarse-grained sandstone. Outcrops are limited to a few small pits dug by Tschanz and his co-workers during their studies of copper mineralization in this part of the Sangre de Cristo Formation (1958).

From these outcrops, three interfingering facies are evident: (1) distal alluvial-fan sheetflood sands, (2) fluvial facies with straight to slightly meandering channels and associated fine-grained overbank sediments, and (3) lacustrine and marsh sediments. Sheetflood sands represent a gradational contact with the underlying distal alluvial-fan system. The upper 60 m of the coastal-plain assemblage contains no sheetflood deposits.

Fluvial Deposits

Over 60% of the coastal-plain system is interpreted as having been deposited by fluvial processes with fine-grained overbank sediments constituting the bulk of the deposits. A very few sandstones exhibit the characteristics needed to classify them as channel-deposited sandstones (Fig. 5). These sandstone beds contain coarse- to very coarse-grained arkosic sand, some clay clasts (3 cm maximum width), and abundant pebbles (10 cm maximum width) which line scours and bedding planes. Large, low-angle, trough crossbeds and large tabular crossbeds are the dominant bedforms. Overlying some of these coarse sand units is a fining-upward sequence consisting of poorly laminated to structureless, occasionally root-mottled, sand and shaly sand, and poorly laminated silt.

Channel sandstones are interpreted as straight to slightly meandering bedload deposits. The low sand/shale ratio can be attributed to the small size of channels (less than 7 m at bankfull depth), the small amount of channel migration, and, most importantly, the apparent high rate of vertical floodplain accretion (Leeder, 1978).

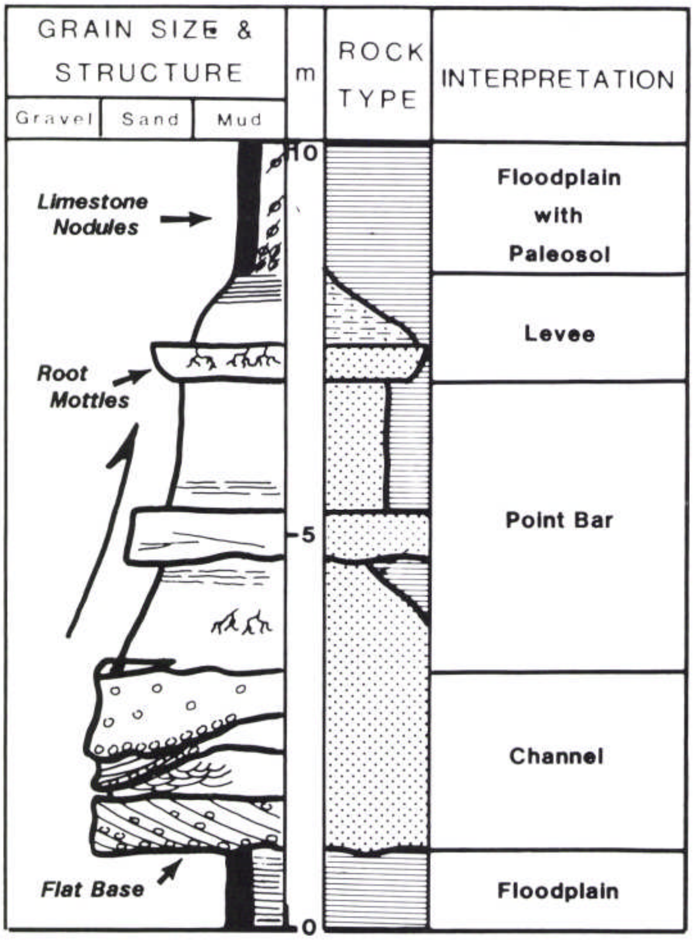

FIGURE 5. Measured section at outcrop TR showing a rare bedload channel sandstone and associated point bar in the coastal-plain-facies assemblage.

Overbank sediments are varied, ranging from laminated to blocky floodplain silts and shales to the crossbedded silts and sands of natural levees and crevasse splays (Fig. 6). Crevasse-splay deposits are the most common sand bodies within the fluvial facies (at least in number, if not in total thickness). These deposits exhibit much the same characteristics as splays described by Reineck and Singh (1975). Even though not one single event can be laterally examined for any distance, lateral variations can be postulated from scattered outcrops.

Lacustrine and Marsh Sediments

Approximately 15% of the coastal-plain-facies assemblage is interpreted as lake or marsh deposits. The lacustrine—marsh facies within the lower 80 m of the assemblage is characterized by copper-rich lenses up to 100 m across, consisting of black to gray-green shales with occasional thin, nodular micrites (Fig. 7). The highest copper percentage is found within the black shales where copper minerals have replaced plant fragments representing *Lebachia, Calamires,* and *Lepidodendron.* As the copper ore seems to be related to syngenetic production of hydrogen sulfide in a highly organic, reducing environment, it is thought that the shales were deposited in small, stagnant ponds. The close areal relationship of limestones with the shales indicates that micrite production may also be lacustrine-related.

The nonfluvial section in the upper 50 m of the coastal-plain-facies assemblage is characterized by cycles of brown to greenish-gray to grayish-green shales grading upward into reddish-brown, silty limestones or limy siltstones. These sediments eventually grade into a very hard, conchoidally fractured, light-gray, slightly silty to pure micritic limestone. Lateral continuity of these units is unknown. The thickness of the limestone beds in the area of measured section SDC-1 averages 1 m. One limestone in measured section SDC-2, however, is approximately 15 m thick, suggesting a possible thickening and coalescing of the limestones to the south.

These unfossiliferous micrites were interpreted by May and others (1977) as having been deposited in a lacustrine environment. Indeed, the limestones seem to fit the lacustrine criteria set forth by Pettijohn (1957) in that they are dense, nodular micrites which are broken by what are interpreted as calcite-tilled shrinkage, or syneresis, cracks. Krumbein and Sloss (1951) indicated that lacustrine deposits may also be powdery marls; a similar example was located in measured section CP.

It is thought that the small, reducing basins represented by the shale lenses in the lower 80 m of the coastal-plain-facies assemblage eventually coalesced into larger lakes which probably existed for long periods. The lack of copper ore in the elastic lacustrine cycles probably reflects more thorough aeration of the water column by increased wind and wave action as the fetch of the lakes increased.

Channel-dominated-facies Assemblage

Finer-grained sediments that dominate the coastal-plain-facies assemblage grade upward into the sandstone deposits of a channel-dominated-facies assemblage. Three channel types are identified as having progressed across the area during deposition of this coarser-grained interval, in ascending order: (1) small, slightly meandering channels, (2) point-bar and chute-bar deposits of coarse-grained meandering streams, and (3) cross-cutting channel deposits of braided streams. Total thickness of the assemblage is estimated at 180 m, approximately half of the thickness postulated by Tschanz and others (1958) for equivalent units.

Small, Slightly Meandering Channels

Medium- to coarse-grained sandstones comprise approximately 50% of the lower 65 m of the channel-dominated-facies assemblage. Sandstones usually occur as single channels less than 2 m thick. Medium-scale, low-angle, trough crossbedding is the major sedimentary structure with lesser amounts of tabular and epsilon crossbedding (Fig. 8). The base of the channels is highly erosional and commonly lined with clay or limestone clasts. Some channels show fining-upwards textural trends grading into root-mottled, silty or shaly sands. Interbedded fine-grained

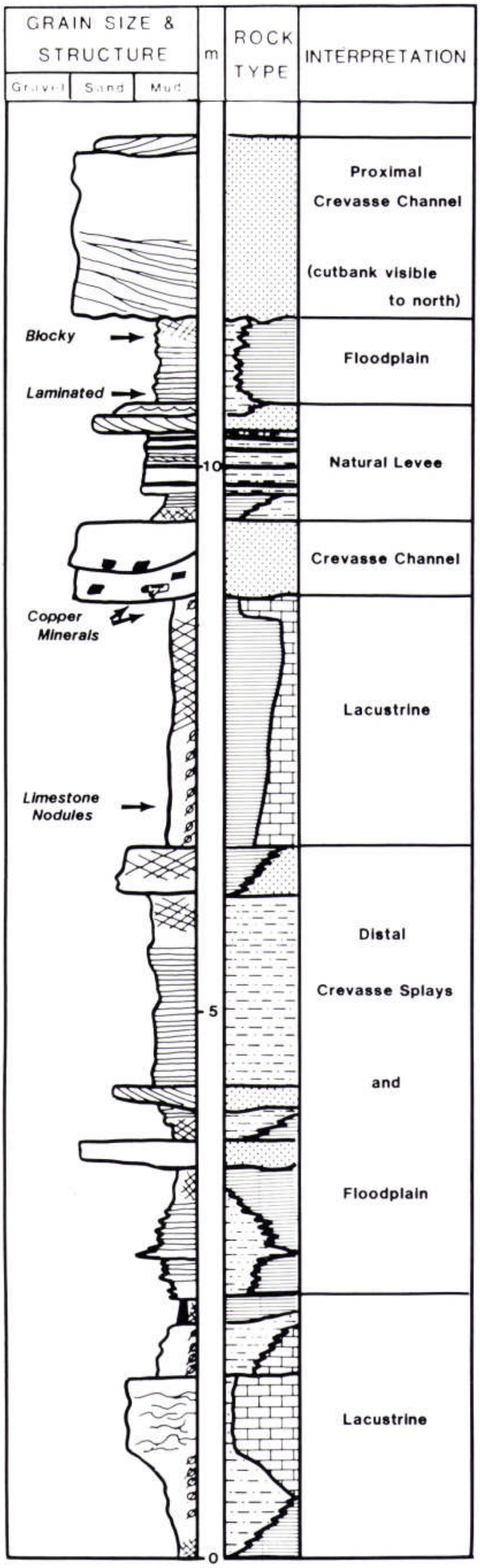

FIGURE 6. Measured section CP showing fine-grained facies of coastal-plain assemblage.

sediments are primarily laminated to blocky, light-green to dark-brown, silty shales.

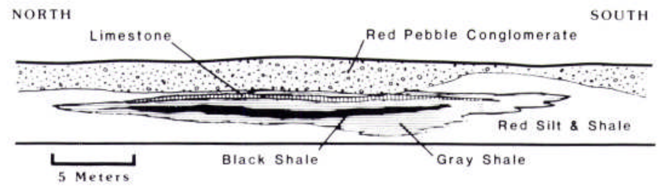

FIGURE 7. Copper-rich shale and limestone lense from the coastal-plain assemblage (from Tschanz and others, 1958).

The origin of these sediments is considered to be similar to the low-sinuosity fluvial deposits in the underlying coastal-plain-facies assemblage. The major differences are the increase of channel sandstones and the decrease of crevasse-splay and lacustrine deposits in the younger, dominately coarse-grained unit.

Point- and Chute-bar Deposits

Up-section from the straight to meandering channel facies are at least 60 m of fluvial deposits of which approximately 60% is channel sandstone. Some of these sandstones contain characteristics similar to the point-bar and chute-bar deposit described by McGowen and Garner (1970) from the coarse-grained meanderbelts of the Colorado River in Texas.

Two types of point-bar deposits were recognized in the Sangre de Cristo Formation (Fig. 9). The first type grades upward from medium-grained sandstone with small trough and contorted crossbedding into finer-grained, epsilon-crossbedded sandstone. The top is structureless and is capped by what is interpreted as a crevasse-splay deposit.

The second type of point bar consists entirely of trough cross-stratification. Bedset size decreases upward, as does grain size. A crevasse-splay deposit also caps this particular section.

It is possible that these two types of point bars are actually end points in a depositional spectrum. The trough crossbeds in both types represent lower point-bar deposition associated with more turbulence near the main channel. In the case of outcrop DA (Fig. 9) it appears that sand waves have migrated out of the channel and onto the bar surface. Decrease in bedset thickness can be related to decreasing water depths associated with bar growth and migration. Silt drapes were deposited during low water levels.

Epsilon crossbeds represent interval accretion on an upper point-bar surface. The upper point bar was sufficiently above low-water stand that clay-drape deposition was not possible. Even though the entire bar was crossbedded at one time, subsequent plant growth on the uppermost surface destroyed all sedimentary structures.

A diagnostic feature of Recent coarse-grained meanderbelt systems is the formation of chute-bar complexes. At least one chute bar was recognized in this section (Fig. 10). The chute-channel deposit is coarse- to very coarse-grained sandstone with lenses of gravel and clay clasts up to 10 cm across. The chute bar itself consists of steeply dipping avalanche faces lined with gravel and clay clasts. These large tubular crossbeds are capped by planar topset beds.

Braided-stream Channels

The upper 100 m or more of the prograding fluvial facies is dominated by cross-cutting sandstone channels. Less than 10% of this interval is fine-grained overbank material. The individual sandstone bodies are coarse- to very coarse-grained (Fig. I 1). Most outcrops show no vertical textural trends; however, a few of the sandstone units fine upward.

Most braided-stream sandy facies of Miall (1977, 1978) were recognized in the section with large trough, large planar, and horizontal cross-stratification predominating. Erosional scours and scour fill are also common. Major scour surfaces are often lined with gravel and clay clasts up to 10 cm across. Silt- and clay-draped reactivation surfaces are also present. Siliceous concretions are abundant in the fine-grained

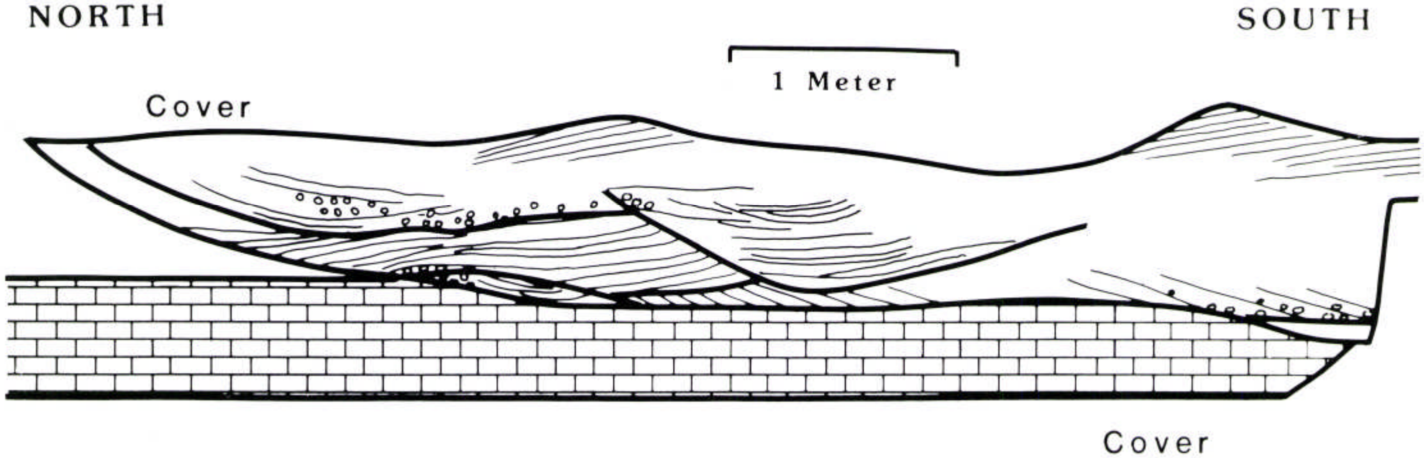

FIGURE 8. Sketch of a singular, medium-grained channel sandstone cutting into a lacustrine limestone at outcrop CC.

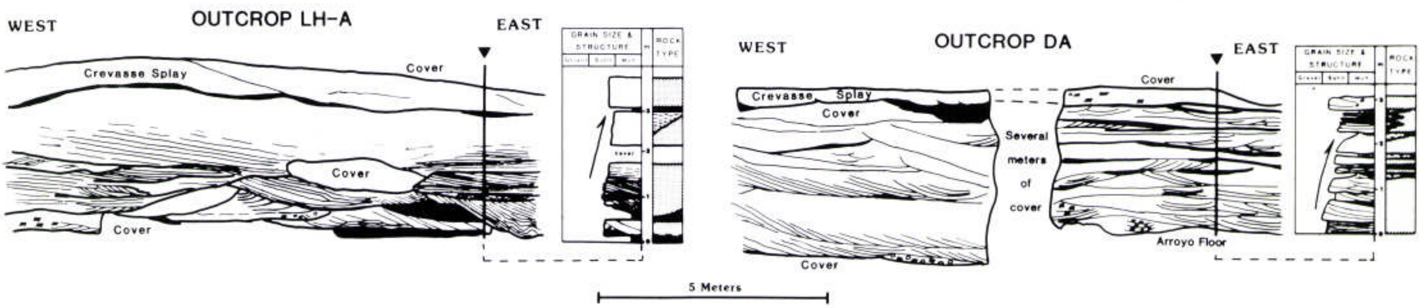

FIGURE 9. Comparison of two point-bar deposits in the progradational fluvial- facies assemblage.

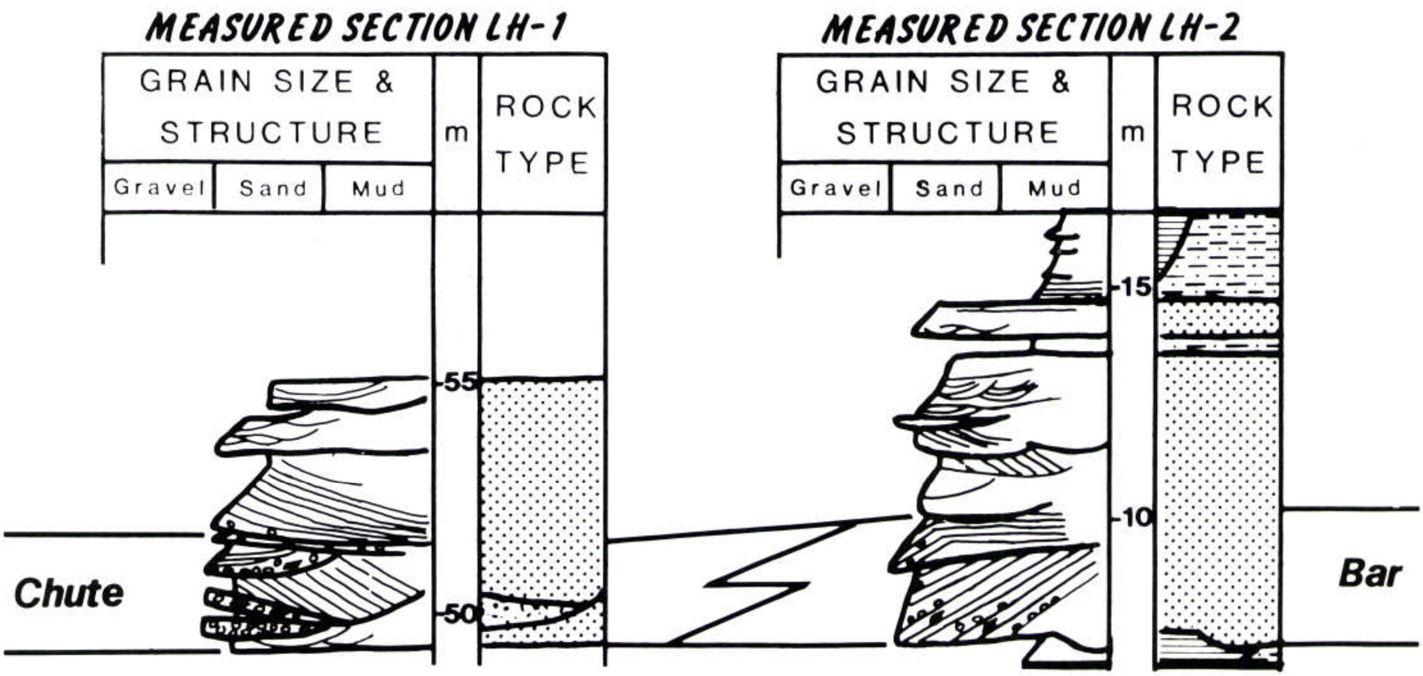

FIGURE 10. Laterally equivalent measured sections interpreted as representing chute-bar deposits.

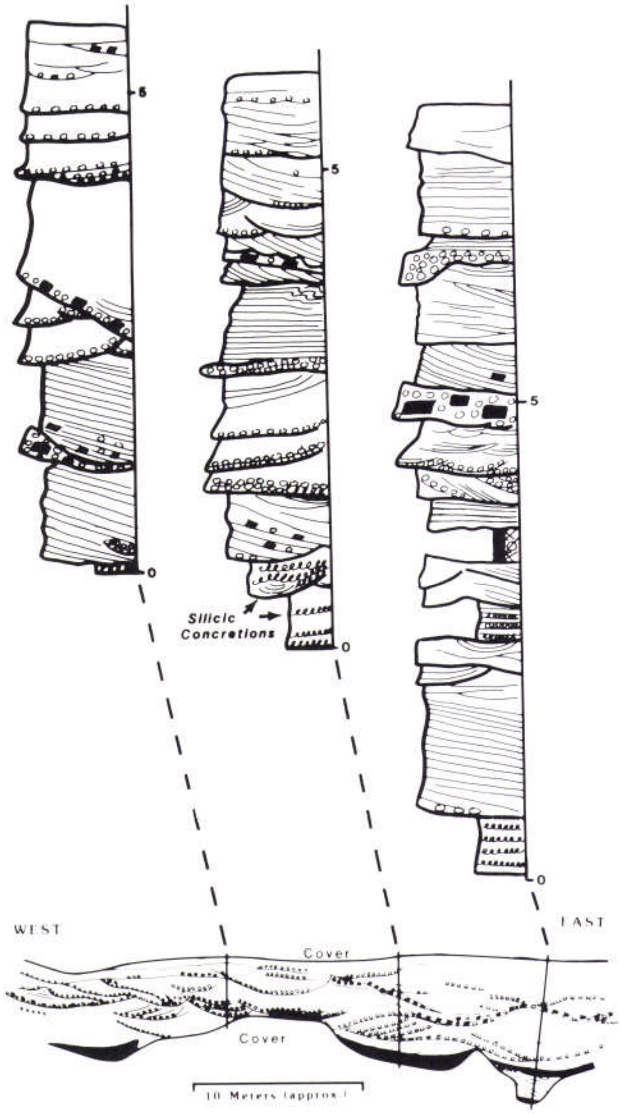

FIGURE 11. Braided-stream deposit at outcrop LC-B.

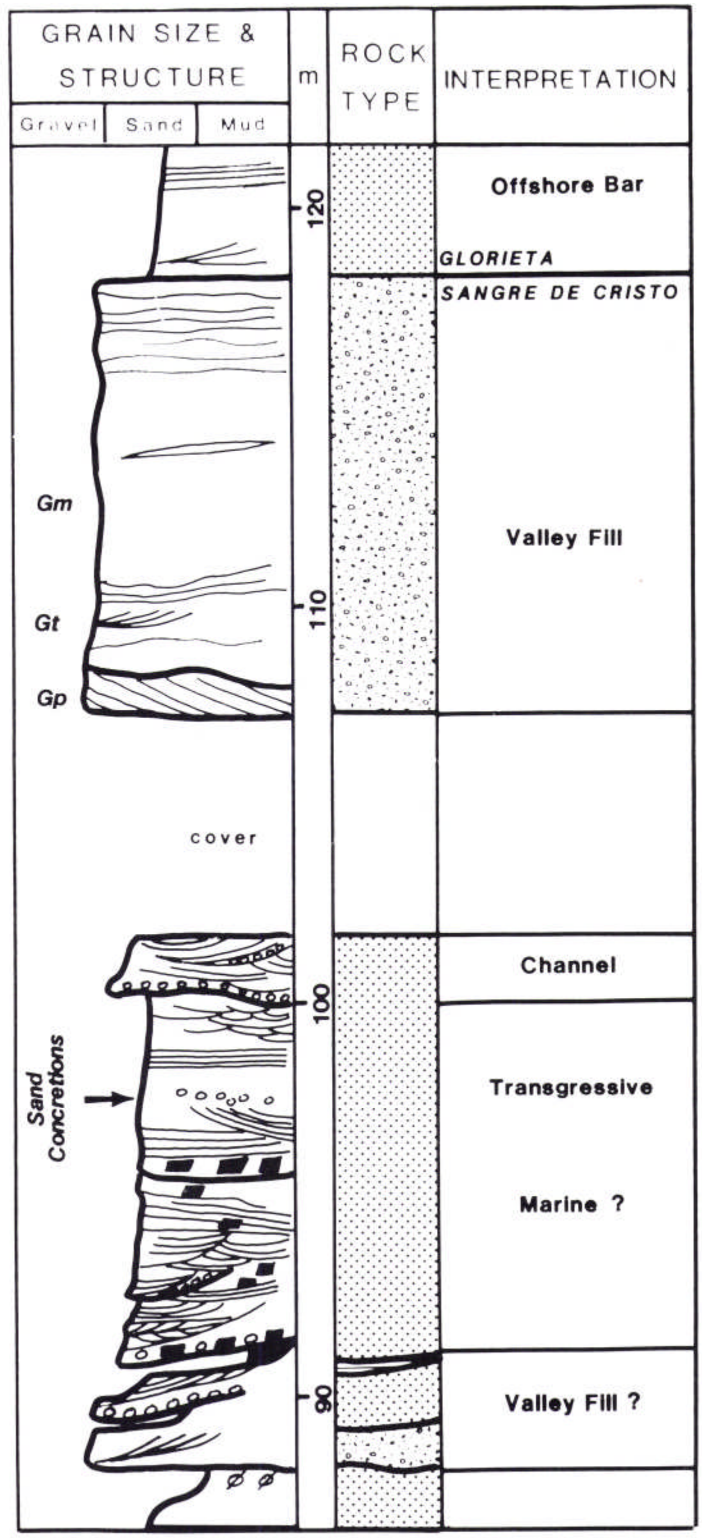

FIGURE 12. Measured section LC showing valley fill and possible transgressive marine units at the top of the Sangre de Cristo Formation. Recognized facies are massive gravel (Gm), trough-crossbedded gravel (Gr), and planar-crossbedded gravel (Gp).

overbank material and are locally present along some scour surfaces. These concretions are interpreted as having been precipitated during post-burial alteration.

The lack of preserved overbank material in this interval is related to a decrease in rate of subsidence. This allowed braided fluvial channels to migrate across the area and erode channel-stabilizing fine-grained sediments.

VALLEY-FILL AND TRANSGRESSIVE MARINE SEQUENCE

A 12-m-thick conglomerate is present at the top of the Sangre de Cristo Formation in measured section LC (Fig. 12). According to Miall's classification, the primary facies is massive gravel with minor occurrences of trough and planar crossbedding. Since equivalent gravels were not recognized in measured sections TM-1 and TM-2 to the north, it is apparent that this accumulation is lenticular. This gravel lens is interpreted as a valley-fill sequence related to a rapid rise in sea level. Sea-level rise can be illustrated by the large offshore-bar complex of the Glorieta Member of the San Andres Formation which directly overlies the valley-fill sequence. The gravel, with individual clasts attaining diameters of 20 cm, was deposited when the capacity of the water to carry sediment was diminished by the rise in base level.

A thin conglomerate, possibly representing another valley-fill deposit, occurs approximately 18 m below the thick gravel lens. The medium-grained, low-angle trough- and planar-bedded sediments overlying the thin gravel deposit may also represent a transgressive marine sand.

DEPOSITIONAL HISTORY

The Rowe—Mora Basin was part of a Paleozoic trough containing both the Maroon and Central Colorado Basins, which extended north-

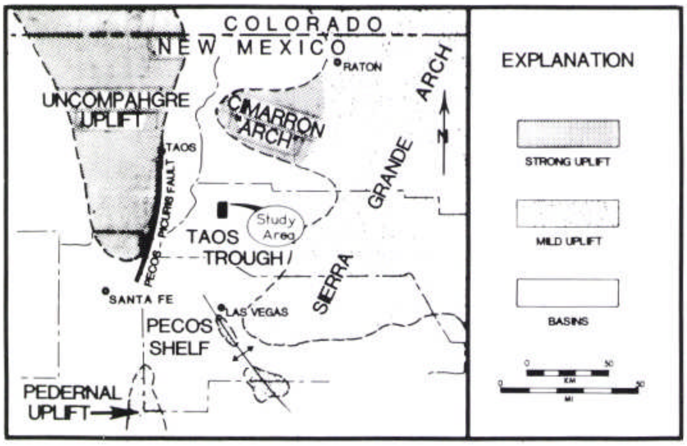

FIGURE 13. Pennsylvanian paleogeography of the Taos trough area (after Casey, 1980b).

ward approximately 600 km into Moffat County, Colorado (Brill, 1952). Baltz (1965) postulated that the Rowe—Mora Basin was separated from the Central Colorado Basin by the northwest-trending Cimarron arch. Sutherland (1963) divided the Rowe—Mora Basin into two distinct regions based on dominant sediment type. To the north was the rapidly subsiding, elastics-filled Taos trough, and to the south the stable, carbonate Pecos shelf (Fig. 13). The Coyote Creek district lies within the Taos trough, about 20 km east of the basinal axis (Bachman, 1953).

Sutherland (1963) believed that the Taos trough was separated from the Uncompahgre uplift to the west by the Pecos—Picuris fault, a reactivated Precambrian fault zone. Large displacement on this fault caused the Uncompahgre uplift to become a major sediment contributor during much of the Pennsylvanian (Foster and others, 1972; Mallory, 1972).

To the east, the Taos trough sloped gently onto the ancestral Sierra Grande uplift. The importance of this positive area, in terms of sediment production, is masked by a lack of subsurface data and the near-coincident position of the Tertiary Sierra Grande arch (Casey, 1980a). Roberts and others (1976) believed that the Sierra Grande uplift was the major source area for the Taos trough during the Late Pennsylvanian. However, this uplift was probably only a mild, northeast-trending feature supplying minor amounts of sediments until it was buried in the Early Permian.

The Sangre de Cristo Formation is a Late Pennsylvanian—Early Permian deposit which asymmetrically filled the Taos trough. Deposits thin southward from a maximum thickness of 1,700 m near the Cimarron arch, 15 km north of the study area (May and others, 1977). The formation represents a north—south progradational unit which for the most part intertongues with, or conformably overlies, the Madera Formation. The basal boundary is considered time-transgressive, ranging from upper Desmoinesian in the north to lower Wolfcampian in the south. In the Coyote Creek district the basal contact is upper Desmoinesian or possibly lower Missourian (Casey, 1980a; Tschanz and others, 1958).

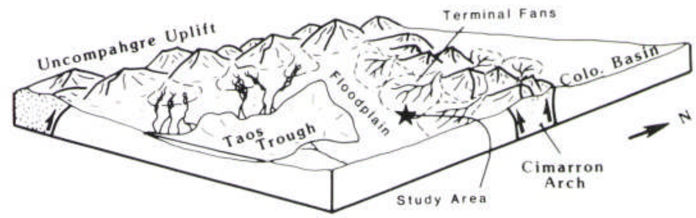

FIGURE 14. Paleogeography of the Coyote Creek region during alluvial-fan deposition of the Sangre de Cristo Formation.

By classifying the separate environments of the Sangre de Cristo Formation and identifying the sedimentary processes controlling those environments, a depositional history of the area can be formulated. The resultant depositional history suggests that Sangre de Cristo sedimentation in the Coyote Creek area was strongly influenced by the rapid uplift, erosion, and eventual burial of the Cimarron arch.

Distal Alluvial-fan Deposition

Early Sangre de Cristo sediments suggest that final uplift of the Cimarron arch began during the upper Desmoinesian, resulting in an initiation of arkosic alluvial-fan deposition (Fig. 14). Grain size decreased within the fans from proximal boulder conglomerates near the arch (May and others, 1977) to distal siltstones and marlstones with thin pebbly sandstone in the study area. Progradation of the fans displaced the fluvial, deltaic, and marine deposits of the time-equivalent Madera Formation to the south. Lack of major variation within the alluvial-fan-facies interval suggests that sedimentation rates were nearly equal to the subsidence.

Coastal-plain Deposition

By the time the coastal-plain-facies assemblage was being deposited, uplift on the Cimmaron arch had essentially halted, thus reducing source potential for the alluvial fans (Fig. 15). Eventually, the arch was eroded to a point where fan deposition ceased in the Coyote Creek area.

The fine-grained coastal-plain sediments, representing minor channels, marsh, and lake deposits, were probably sediments on a large delta plain. It is believed that sediment was fed from the Colorado Basin into the Taos trough through a breach of the still partially positive Cimarron arch. By the end of the coastal-plain depositional sequence, sedimentation was being outpaced by subsidence. At this time, the shoreline of the Taos Sea was probably only a few kilometers to the south.

Progradational Fluvial Sequence

Finally, the Cimarron arch adjacent to the Coyote Creek area was breached by the fluvial systems of the Colorado Basin (Fig. 16). As subsidence rates decreased, bedload fluvial systems prograded across the area into the Taos trough. Fluvial facies consisted of distal, straight

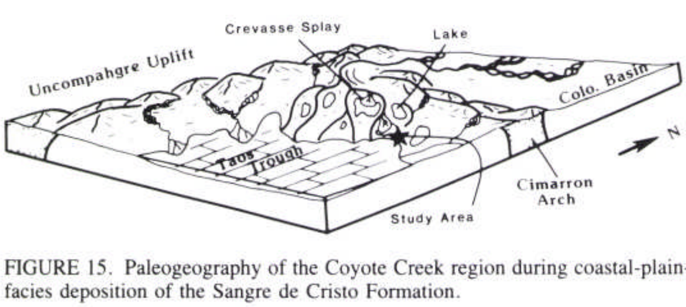

FIGURE 15. Paleogeography of the Coyote Creek region during coastal-plain-facies deposition of the Sangre de Cristo Formation.

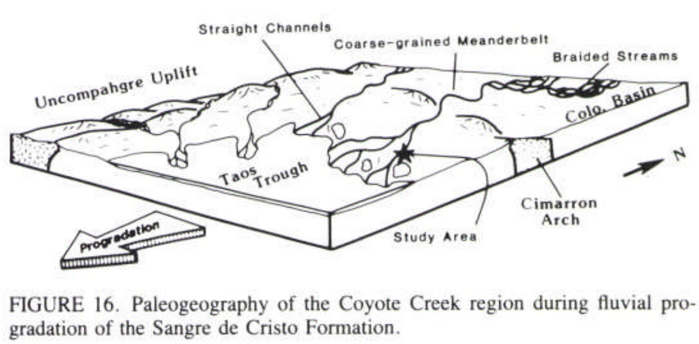

FIGURE 16. Paleogeography of the Coyote Creek region during fluvial progradation of the Sangre de Cristo Formation.

to slightly meandering distributary channels which coalesced both up-dip and up-section into larger, coarse-grained meanderbelt channels. Sinuosity and migration of the channels seem to be directly related to subsidence and aggradation rates. Ultimately, these rates decreased, resulting in the deposition of rapidly migrating braided-stream channels which eroded any bank-stabilizing material. By the end of Sangre de Cristo deposition, the Rowe-Mora Basin had been filled with sediment. Glorieta seas transgressed across the area beveling the top of the Sangre de Cristo Formation to a flat surface, thus closing the sedimentary history of the Paleozoic Rowe-Mora Basin.

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