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Basin architecture of Pliocene-Lower Pleistocene alluvial-fan and axial-fluvial strata adjacent to the Mud Springs and Caballo Mountains, Palomas half graben, southern Rio Grande ritft

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BASIN ARCHITECTURE OF PLIOCENE-LOWER PLEISTOCENE ALLUVIAL-FAN AND AXIAL-FLUVIAL STRATA ADJACENT TO THE MUD SPRINGS AND CABALLO MOUNTAINS, PALOMAS HALF GRABEN, SOUTHERN RIO GRANDE RIFT

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ABSTRACT—In the northern Palomas half graben, adjacent to both the footwall scarp of the Caballo Mountains and the intrabasinal Mud Springs Mountains, the Pliocene-lower Pleistocene Palomas Formation was separated into five mappable facies assemblages, including (1) volcanic-clast conglomerate (VCC), deposited on alluvial fans derived from the hanging wall mountains (Black Range, Salado Hills, Sierra Cuchillo), (2) sedimentary, granitic and metamorphic-clast conglomerate (SGMC), deposited on alluvial fans derived from the footwall scarps of the Caballo and Mud Springs Mountains, (3) sedimentary-clast conglomerate (SCC), deposited on alluvial fans derived from the eastern dip slope of the Mud Springs Mountains, (4) crossbedded sandstone (CS), deposited by the ancestral Rio Grande and (5) fine-grained facies assemblage (FG), representing a combination of deposition on distal alluvial fans derived from the hanging wall mountains, on the floodplain of the ancestral Rio Grande and on distal alluvial fans from the eastern dip slope of the Mud Springs Mountains.

The basin-scale architecture is characterized by (1) narrow (< 1 km) aprons of alluvial-fan deposits adjacent to the Mud Springs (SGMC and SCC) and Caballo (SGMC) Mountains, (2) a narrow (< 2 km) belt of fine-grained sediment (FG) that extended to within a few hundred meters of the eastern dip slope of the Mud Springs Mountains, (3) a narrow (< 5 km), south-trending belt of axial-fluvial deposits (CS) and (4) a broad (15-20 km) belt of alluvial-fan sediment derived from the western hanging wall mountains (VCC). The width of these facies belts and the nature of their interactions are a response to a combination of eastward tilt of the Palomas half graben, the relative size of the catchments and the role of paleoclimate on the erosive power of channels.

INTRODUCTION

The Rio Grande Valley of south-central New Mexico is an excellent natural laboratory for studying basin-scale architecture of alluvial-fan and axial-fluvial sediment deposited in active continental rift basins. This is because Pliocene-lower Pleistocene basin fill is widely exposed as a result of post-early Pleistocene incision by the Rio Grande and its tributaries. Pliocene-lower Pleistocene strata are exposed along the walls of canyons oriented transverse to the basin axes, as well as in badlands occupying the interfluves between the mouths of the canyons. In addition, the constructional top of the Pliocene-lower Pleistocene strata is locally preserved as a gently dipping geomorphic surface that preserves the depositional slope just prior to the onset of regional incision (Gile et al., 1981; Mack et al., 2006). Interpretation of basin architecture is further enhanced by the fact that Pliocene-lower Pleistocene strata were derived from uplifts and deposited in basins that form the modern landscape, eliminating the uncertainties of paleogeographic reconstructions.

The Pliocene-lower Pleistocene Palomas Formation and its associated constructional surfaces are especially well exposed in the Palomas Basin, an eastward-tilted half graben (Fig. 1; Seager et al., 1982). Previous work in the southern and central part of the Palomas Basin documented the sedimentologic characteristics of alluvial-fan and axial-fluvial facies and their distribution within the basin (Seager et al., 1982; Mack and Seager, 1990). Further research demonstrated that the axial river periodically moved toward the footwall by toe-cutting the footwall-derived alluvial fans, a process interpreted to have been driven by basin tilting toward the footwall during intervals of fault activity (Mack and Leeder, 1999; Mack et al., 2002; Seager and Mack, 2003).

The present study involves mapping at the scale of 1:10,000 of facies assemblages of the Pliocene-lower Pleistocene Palomas Formation adjacent to the footwall scarp of the Caballo Mountains and to the Mud Springs Mountains, a small, intrabasinal uplift in the northern part of the Palomas Basin (Fig. 1). The principle goals of the study are to document the basin-scale distribution of and interaction between the facies assemblages and to define the roles of tectonics and paleoclimate on facies architecture.

Supplemental data for this paper can be accessed at:
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METHODS

The Pliocene-lower Pleistocene Palomas Formation along the flanks of the southern Mud Springs Mountains was divided into five facies assemblages that were mapped on parts of two United States Geological Survey (USGS) 7.5 minute quadrangles (Williamsburg and Cuchillo) and corresponding aerial photographs that were scaled to 1:10,000 (Figs. 1 and 2; Foster, 2009). In addition to the facies assemblages, the constructional top of the Palomas Formation (Cuchillo surface) was mapped, as were two widespread geomorphic surfaces inset against the Palomas Formation. A more detailed map of multiple inset surfaces east of the study area in the Elephant Butte quadrangle is presented by Lozinsky (1986). Also shown on the map are the locations of a distinctive white pumiceous bed and two diatomites, as well as

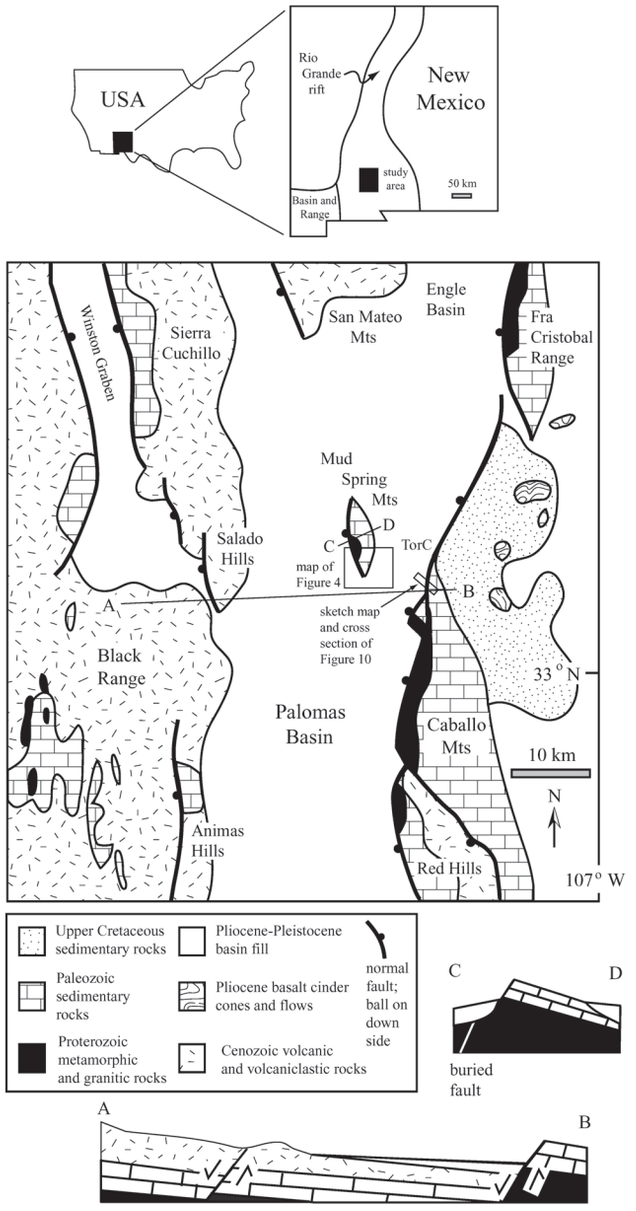


FIGURE 1. Generalized geologic map and cross sections of the northern part of the Palomas Basin and surrounding bedrock uplifts. Adapted from Clemons et al. (1982).

paleocurrent arrows based on 50 measurements of imbrications or the axes of trough crossbeds per site. The bedrock geology of the Mud Springs Mountains is taken from Maxwell and Oakman (1990), although the Palomas-bedrock contact was mapped in more detail in this study. The extreme southeastern part of the Mud Springs Mountains and adjacent basin were previously mapped by Lozinsky (1986), but were remapped in this study with locally different results.

In addition to the map and cross sections of the southern Mud Springs Mountains and adjacent basin, a sketch map and cross section were constructed adjacent to the Caballo Mountains, based on the geologic map of Kelley and Silver (1952) and field work in this study (Figs. 1 and 2). During the latter, two of the facies assemblages recognized adjacent to the Mud Springs Mountains were applied to the Palomas Formation along the flanks of the Caballo Mountains.

GEOLOGIC SETTING

History of the Palomas Basin

The Palomas Basin is an eastward-tilted half graben, whose footwall is the Caballo Mountains and Red Hills and whose hanging wall is the Black Range, Animas Hills, Salado Hills and southern Sierra Cuchillo (Fig. 1; Seager et al., 1982). The footwall scarp of the Caballo Mountains consists of middle Proterozoic metamorphic and granitic rocks that are overlain by Paleozoic sedimentary rocks, mostly carbonates. In contrast, the mountains that comprise the hanging wall are composed primarily of Cenozoic volcanic and volcaniclastic rocks of rhyolitic, andesitic and basaltic composition, although a few exposures of Proterozoic and Paleozoic rocks are present (Fig. 1).

The Mud Springs Mountains are a northeast-tilted fault block that is ~10 km long and ~2 km wide. It occupies the northeastern part of the Palomas Basin, just south of the right-stepping junction between the Palomas and Engle Basins (Fig. 1). Like the Caballo Mountains, the west-facing footwall scarp of the Mud Springs Mountains consists of Proterozoic metamorphic and granitic rocks and Paleozoic sedimentary rocks, while the eastern dip slope of the range consists almost exclusively of Pennsylvanian limestones and shales with a few thin andesitic sills (Maxwell and Oakman, 1990).

The Palomas Basin first developed ~27 Ma ago, resulting in deposition of ~2 km of uppermost Oligocene and Miocene alluvial-fan and lacustrine strata (Seager et al., 1984; Mack et al., 1994). The basin narrowed near the Miocene-Pliocene boundary by uplift of the Red Hills, Animas Hills and Salado Hills, followed by deposition of 100-150 m of the Palomas Formation on alluvial fans and by the axial ancestral Rio Grande (Lozinsky and Hawley, 1986a, b; Seager et al., 1982; Seager and Mack, 2003). A Pliocene to early Pleistocene age (~5 to 0.8 Ma) for the Palomas Formation has been established by a combination of radioisotopic and tephrochonologic dates of underlying, overlying and interbedded volcanic rocks (Bachman and Mehnert, 1978; Seager et al., 1984; Mack et al., 2009), vertebrate biostratigraphy (Tedford, 1981; Lucas and Oakes, 1986; Repenning and May, 1986) and reversal magnetostratigraphy (Mack et al., 1993, 1998, 2002).

At ~0.8 Ma ago, the ancestral Rio Grande and its tributaries began to alternately incise and partially backfill the Palomas basin, producing a series of basinward-stepping geomorphic surfaces that are inset against the Palomas Formation (Gile et al., 1981; Lozinsky, 1986). This period of net incision, which continues to the present day, was probably driven by glacial-interglacial climatic cycles and ultimately lowered the river by ~100 m, exposing all or the upper part of the previously deposited Palomas Formation (Mack et al., 2006). The constructional top of the Palomas Formation (Cuchillo surface) is locally preserved as

a gently sloping geomorphic surface capped by a petrocalcic soil (Lozinsky, 1986; Lozinsky and Hawley, 1986a, b).

Major Faults Bordering the Northern Palomas Basin

The study area is located at or near the junction of four faults: Caballo, Hot Springs, Williamsburg and Mud Springs (Fig. 2; Kelley and Silver, 1952; Seager and Mack, 2003). The Caballo and Hot Springs faults constitute the border faults of the Caballo Mountains and intersect in what is informally referred to by Seager and Mack (2003) as the "northern Red Hills". Both the Caballo and Hot Springs faults bring middle Proterozoic metamorphic and granitic rocks to the surface and display up to 30 m of offset of the constructional top of the Palomas Formation (Cuchillo surface), indicating post-0.8 Ma fault activity. Extending north-northwest from the northern Red Hills, the Williamsburg fault is manifested as a series of scarps that offset middle to late Pleistocene geomorphic surfaces. Trenching of the northern scarp by Foley et al. (1988) established late Holocene offset on the Williamsburg fault. Near the northern Red Hills, the relationship between the Williamsburg and Hot Springs fault is not clear, because of poor exposure. Initiation of the Hot Springs fault as a major Laramide (latest Cretaceous-early Paleogene) strike-slip fault has been suggested by Cather and Harrison (2002) and Harrison and Cather (2004). Mason (1976) and Seager and Mack (2003) also recognized that parts of the Hot Springs fault could represent reactivation of a Laramide fault, but in their interpretation the fault was a reverse fault related to local Laramide deformation.

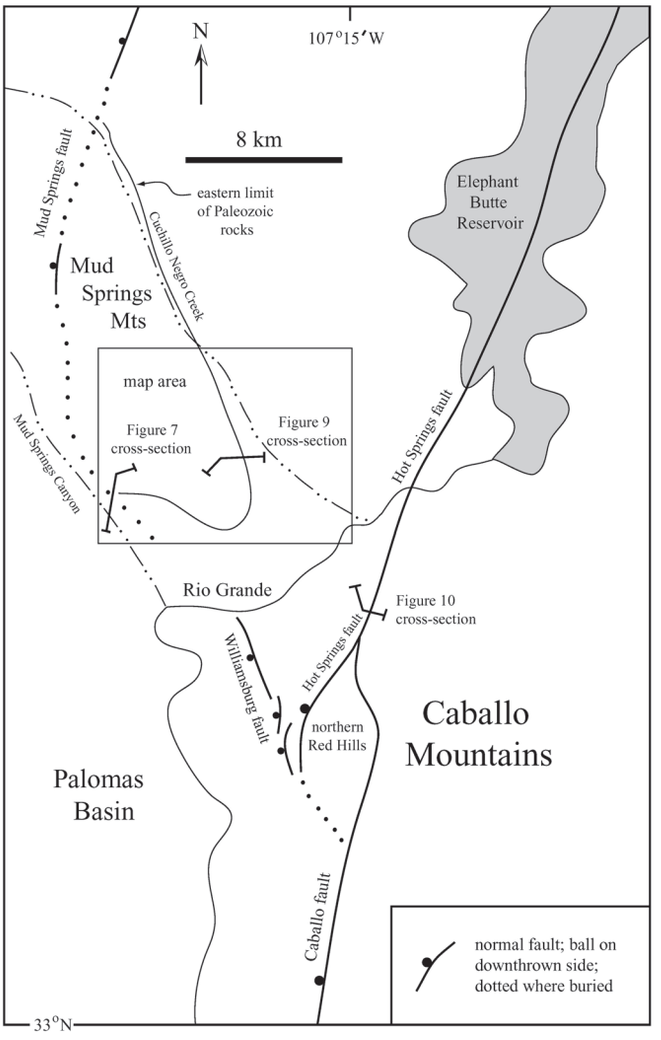


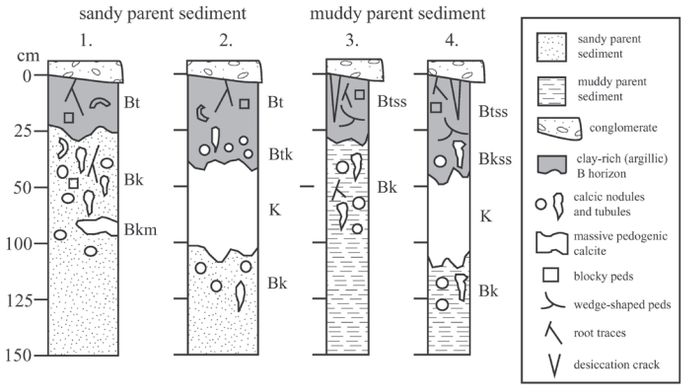
FIGURE 2. Major faults in the vicinity of the northern Palomas Basin. Adapted from Kelley and Silver (1952), Clemons et al. (1982), and Maxwell and Oakman (1990).

The Mud Springs Mountains are bordered on the west-southwest by the Mud Springs fault (Fig. 2; Maxwell and Oakman, 1990), which is buried over most of its length. Based on an apatite fission track age of 8.3 ±3.8 Ma from Precambrian granitic rocks at the base of the footwall scarp, Kelley (1997) suggested that most of the offset on the Mud Springs fault occurred during Miocene time. The fault is interpreted to have been inactive during deposition of the Pliocene-early Pleistocene Palomas Formation, because the fault is everywhere buried beneath Palomas sediment and the footwall scarp of the Mud Springs Mountains was progressively onlapped by the Palomas Formation (Maxwell and Oakman, 1990). Following deposition of the Palomas Formation the Mud Springs fault was reactivated, offsetting the Cuchillo geomorphic surface in the northern part of the range (Machette, 1978; Maxwell and Oakman, 1990). Other faults with a few meters of displacement also cut the Palomas Formation (Hawley and Seager, 1978; Maxwell and Oakman, 1990). Neither the Mud Springs fault nor the Williamsburg fault can be traced across the modern Rio Grande Valley, although Foley et al. (1988) and Seager and Mack (2003) have suggested that the faults may connect. Finally, based on well-log data, Lozinsky (1987) suggested that the Mud Springs Mountains fault block extends at least 6 km north of the northern topographic edge of the range and was bounded on the northeastern side by a northeast-dipping normal fault. However, a fault on the northeastern side of the range in the study area is not evident, because the Palomas Formation onlaps northeast-dipping Paleozoic bedrock (Maxwell and Oakman, 1990).

Paleosols as Indicators of Paleoclimate and Landscape Stability

Interpretation of paleoclimate in south-central New Mexico during Pliocene and early Pleistocene time is primarily based on the physical characteristics of mature calcic and vertic paleosols in the Palomas Formation and coeval Camp Rice Formation to the south. These paleosols have been described in detail by Mack and James (1992) and Mack et al. (2000) and are summarized in Figure 3.

Carbonate is retained in modern soils with non-calcareous parent material in regions that experience between ~25 and 80 cm of annual precipitation (Retallack, 2005), corresponding primarily to semi-arid climatic zones on the Earth today (Strahler and Strahler, 1983). Although a relationship between depth to the top of the calcic B horizon and mean annual precipitation has been



horizon	characteristics
Bt	red-brown, vertical root traces, burrows, blocky and/or prismatic peds, clay coats on sand grains and/or peds.
Btss	same as Bt, plus wedge-shaped peds, slickensides and deep desiccation cracks
Bk	vertical root traces, burrows, blocky and/or prismatic peds, white microcrystalline calcic nodules and tubules
Bkss	same as Bk, but parent material has wedge-shaped peds, slickensides and deep desiccation cracks
Btk	same as Bt, plus scattered calcic nodules and tubules
Bkm	laterally discontinuous patches of well-indurated, massive white microcrystalline calcite < 30 cm thick
K	laterally continuous, well-indurated, massive white microcrystalline calcite > 30 cm thick; may have calcified root traces and burrows, and a few small patches of unaltered parent material

FIGURE 3. Common examples of calcic paleosols in the Pliocene-lower Pleistocene Palomas and Camp Rice Formations, south-central New Mexico, based on Mack and James (1992) and Mack et al. (2000). Profiles show common variations in pedogenic features and horizons as a function of parent material (sandy 1 and 2; muddy 3 and 4) and increasing soil maturity (more immature 1 and 3; more mature 2 and 4).

suggested by Retallack (2005), it is difficult to apply to paleosols of the Palomas and Camp Rice Formations, because they commonly lack recognizable A horizons and have erosive upper contacts. However, the relative thickness (50-100 cm) and diffuse nature of the Bk horizons of many of the Palomas and Camp Rice paleosols are similar to those of modern soils in regions experiencing a tropical, monsoonal climate (Retallack, 2005). Strongly seasonal precipitation is also suggested by vertic features, such as wedge-shaped peds, deep desiccation cracks and slickensides. Vertic features result from the shrinking and swelling of expandable clays and require a dry season at least four months long to develop (Ahmad, 1983). However, there is no evidence in the Camp Rice and Palomas Formations to indicate whether the paleoclimate was winter-wet or summer-wet as it is today.

In addition to their utility as paleoclimate proxies, the morphology of calcic horizons provides insight into the duration of landscape stability and paleosol development (Gile et al., 1966; 1981; Machette, 1985). Stage II Bk horizons, which consist in muddy and sandy parent material of scattered calcic nodules and tubules, are common in the Palomas and Camp Rice Formations and require thousands of years to develop in non-calcareous parent sediment. Also present in the Palomas Formation in the study area are stage III K horizons composed of massive carbonate 50 to 70 cm thick. In non-calcareous parent sediment, stage III morphology calcic horizons require tens to hundreds of thousands of years of landscape stability to form.

FACIES ASSEMBLAGES AND SEDIMENT DISPERSAL

Five facies assemblages were recognized in the study area (Fig. 4; Table 1, supplemental data). Three of the facies assemblages (VCC, SGMC, SCC) are predominantly conglomeratic and are distinguished from each other by the composition of their clasts, giving them aspects of both lithofacies and petrofacies. The other two facies assemblages are finer grained and are noted for a predominance of crossbedded sand/sandstone (CS) or an abundance of red mudstone and very fine sand (FG).

Volcanic-Clast Conglomerate Facies Assemblage (VCC)

Description

The most common rock type in the VCC facies assemblage is weakly cemented, purplish-gray, grain-supported granule, pebble and cobble conglomerate (Fig. 5A). It is characterized by rhyolitic, andesitic and basaltic clasts, although limestone and granitic clasts locally constitute less than one percent of the total. Beds of conglomerate range from 0.5 to 7 m thick and are lenticular or tabular over distances of 30 m. Horizontal layering exhibiting imbrication is common, while planar crossbeds 0.3 to 2 m thick are locally present. Rarely, the conglomerate contains inclined bedsets dipping 5 to 15 degrees.

Interbedded with the grain-supported conglomerates are beds 0.2 to 2 m thick of granular to pebbly sand (Fig. 5A). Many of the sand beds are internally structureless and have widely dispersed granules and small pebbles. In a few cases, the gravel fraction exists as thin (<5 cm), laterally discontinuous lenses that can either coarsen or fine upward. The least common rock type in the VCC consists of beds of red-brown mudstone 0.3 to 1 m thick, which may be interbedded with both conglomerate and sand (Fig. 5A). Some beds of sand and mudstone display calcic and vertic paleosols, similar to those shown in Figure 3.

Interpretation

The VCC facies assemblage is interpreted to have been deposited in mid to distal parts of alluvial fans derived from the Black Range, Salado Hills and Sierra Cuchillo, which constitute the hanging wall of the Palomas Basin. This is supported by east- and southeast-oriented imbrication paleocurrent data (Fig. 4), as well as clast compositions and relative abundances that coincide with the rocks exposed in the hanging wall mountain ranges. The paleofans extended 15 km or more from their sources and had slopes of 0.5° to 1°, based on the slopes of the locally preserved Cuchillo surface. The grain-supported conglomerates within

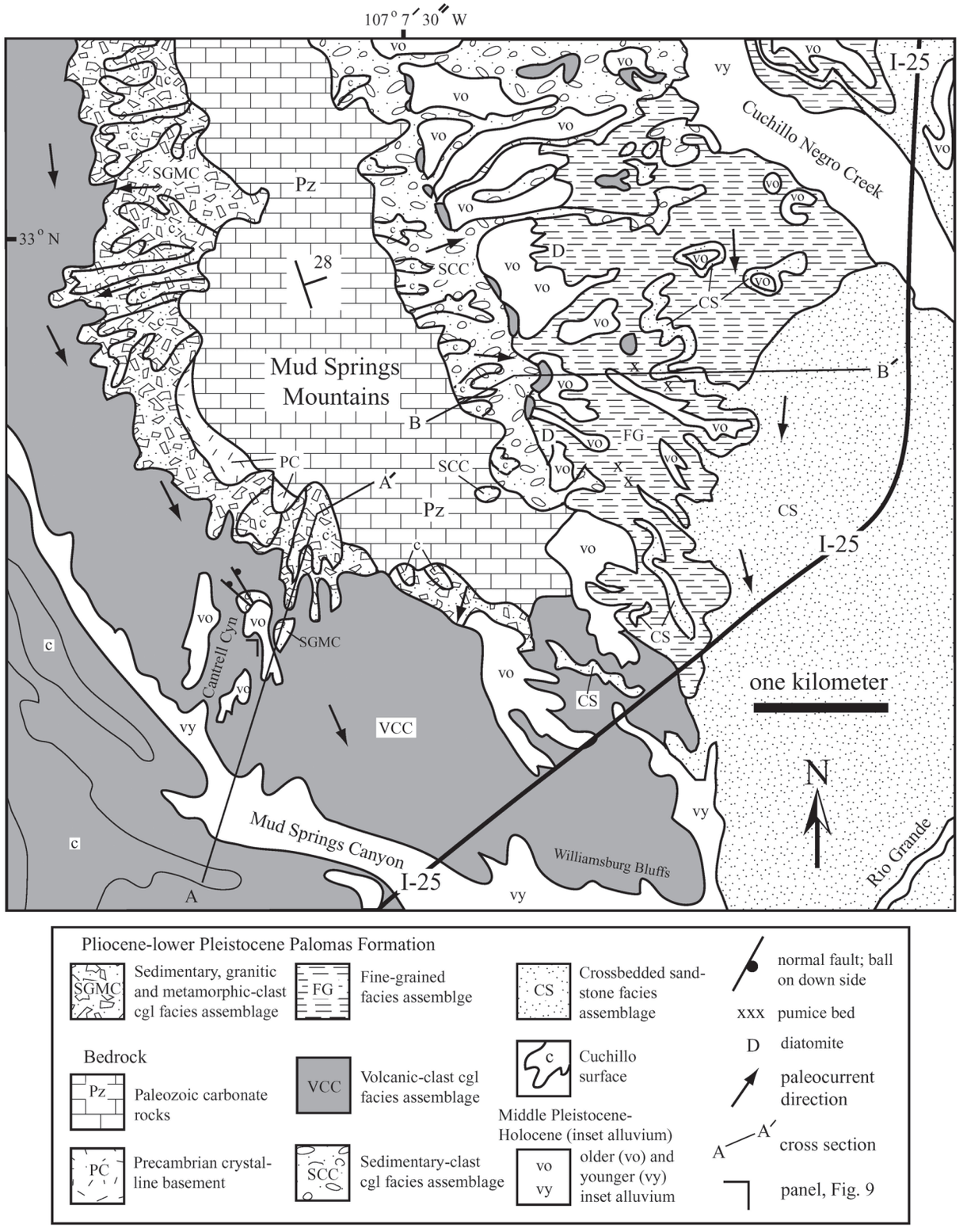


FIGURE 4. Map of the distribution of facies assemblages of the Pliocene-lower Pleistocene Palomas Formation and major Quaternary inset surfaces along the southern tip of the Mud Springs Mountains and adjacent northern Palomas Basin, adapted from Foster (2009). Simplified bedrock geology of the Mud Springs Mountains is from Maxwell and Oakman (1990). Cross sections A-A' and B-B' correspond to Figures 7 and 9, respectively.

the VCC are interpreted to have been deposited in alluvial-fan channels, based on lenticular bed geometry and abundance of current-generated sedimentary structures. Horizontally layered conglomerates with imbrication may have formed as longitudinal bars, while planar crossbeds were either transvere bars or down-stream-migrating bars with slip faces (Hein and Walker, 1977).

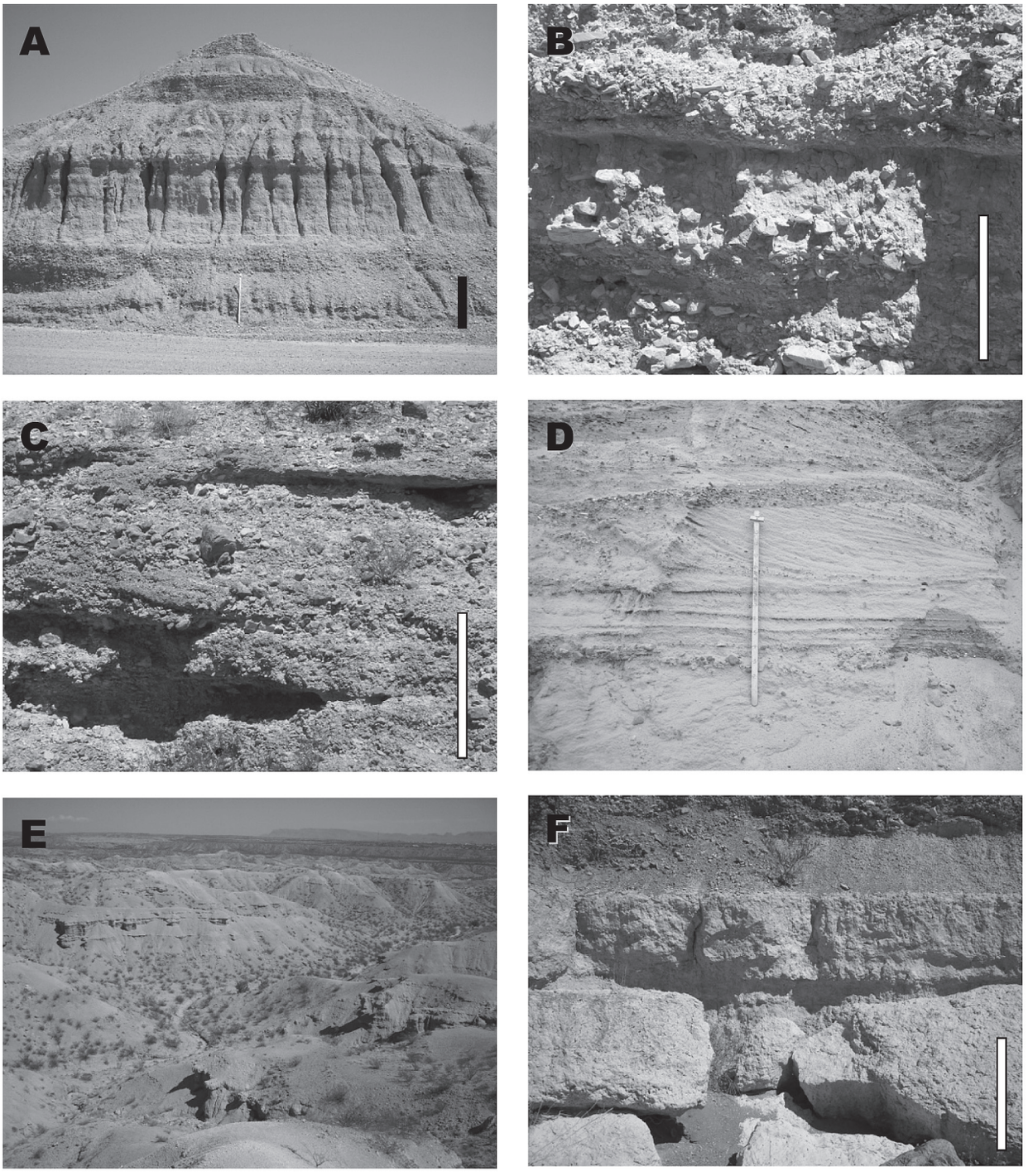


FIGURE 5. Photographs of facies assemblages of the Palomas Formation (Pliocene-lower Pleistocene), northern Palomas Basin. **A**. Interbedded conglomerate, sandstone and red mudstone of the volcanic-clast conglomerate facies assemblage (VCC). Black bar is 1.5 m **B.** Outcrop of the sedimentary, granitic and metamorphic-clast facies assemblage (SGMC), showing a matrix-supported debris flow conglomerate overlain by a grain-supported, waterlaid conglomerate. White bar is 1.5 m. **C.** Grain-supported, well-bedded conglomerates of the sedimentary-clast conglomerate facies assemblage (SCC). White bar is 1.5 m. **D.** Outcrop of the crosbedded sandstone facies assemblage (CS), showing trough crossbeds and horizontal laminae. Jacob's staff is 1.5 m. **E.** View to the northeast of badlands topography developed in the fine-grained facies assemblage (FG). **F.** Light-colored bed of microcrystalline calcite precipitated by shallow groundwater within the fine-grained facies assemblage (FG). White bar is 1.0 m.

Rare inclined bedsets oriented perpendicular to paleoflow likely represent point bars.

Several processes may account for deposition of granular, pebbly sands. Some sand beds interfinger with channel conglomerates and probably represent sandy bars within broad alluvial-fan channels. The most gravel-rich, poorly sorted variety of sand may have been deposited as distal debris flows or hyperconcentrated flows (Smith, 1986; Mack et al., 2002). Mudstones and

thin (0.2 to 0.5 m) beds of sand probably represent overbank deposition and splays from fan channels, respectively.

Sedimentary, Granitic and Metamorphic-Clast Conglomerate Facies Assemblage (SGMC)

Description

The SGMC facies assemblage is characterized by calcite-cemented pebble, cobble and boulder conglomerates composed primarily of limestone, dolostone, sandstone and chert clasts and secondarily of granitic and metamorphic clasts (Fig. 5B). The predominance of carbonate clasts and calcite cement imparts an overall light gray to whitish color to the beds. The SGMG facies assemblage is present directly adjacent to the western margins of the Mud Springs and Caballo Mountains.

Most of the conglomerates are grain-supported and have horizontal layering with imbrication or planar crossbeds up to 2 m thick. Several of the coarsest conglomerates are very poorly sorted and either lack internal layering or have inverse grading (Fig. 5B). Granular, pebbly sand similar to that described in the VCC facies assemblage exists as either thin (<0.5 m) lenses or beds within conglomerates or as laterally continuous beds up to 2 m thick. The latter type of sand is located in distal positions with respect to the adjacent mountain front. Calcic paleosols, like those shown in Figure 3, are present in some of the sand beds.

Interpretation

The SGMC facies assemblage is interpreted to have been deposited on alluvial fans derived from the footwall scarps of the Mud Springs and Caballo Mountains, based on large clast size, paleocurrents directed away from the mountain fronts and a provenance that matches the rocks on the footwall scarps of the two ranges. Where the Cuchillo surface is preserved, paleofan slopes of 7° to 10° are indicated. The coarsest, most poorly sorted conglomerates are the most proximal to the mountain front and probably were deposited as debris flows (Lowe, 1979). The remainder of the conglomerates were waterlaid, as longitudinal bars (horizontal layering with imbrication) and as either transverse bars or downstream-migrating gravel bars with slip faces (planar crossbeds) (Hein and Walker, 1977). Although channel morphology was difficult to recognize in the outcrops, the abundance of gravel beds several meters thick argues for channelized flow.

Granular, pebbly sands interbedded with conglomerates were probably deposited as sandy bars within the fan channels. Because the thicker, more laterally continuous sands are located farthest from the mountain fronts, they may represent distal sheetfloods or, in the case of the most poorly sorted types, distal debris flows or hyperconcentrated flows (Smith, 1986; Mack and Leeder, 1999).

Sedimentary-Clast Conglomerate Facies Assemblage (SCC)

Description

Like the SGMC facies assemblage, the sedimentary-clast conglomerate facies assemblage consists of light-colored, calcite-cemented boulder, cobble and pebble conglomerates that are well bedded and exhibit horizontal layering with imbrication and crossbeds (Fig. 5C). Conglomerates of the SCC facies assemblage are distinguished from those of the SGMC facies assemblage by being composed exclusively of Pennsylvanian limestone and chert clasts and by its location on the eastern flank of the Mud Springs Mountains. Locally interbedded with conglomerate are thin (<0.5 m) beds and lenses of granular, pebbly sand, which may contain calcic paleosols like those in Figure 3.

Interpretation

The SCC facies assemblage represents alluvial-fan deposition adjacent to the hanging wall dip slope of the Mud Springs Mountains, based on overall grain size, paleocurrents and its distinctive provenance. Depositional processes on the fans were similar to those of the previously described VCC and SGMC facies assemblages. Where preserved, the Cuchillo surface indicates paleofan slopes of 7° to 9°.

Crossbedded Sandstone Facies Assemblage (CS)

Description

Gray, crossbedded, granular and pebbly, medium- to coarse-grained sand and sandstone dominate the CS facies assemblage (Fig. 5D). Multistorey channels are common, with storey boundaries defined by erosional scours overlain by gravel and/or mudstone rip-up clasts. In some cases, individual storeys fine upward to fine-grained sand. Trough crossbeds 0.2 to 1.5 m thick are common, although planar crossbeds and horizontal laminae with parting lineation are locally present. Ripple cross-laminae are rare and largely restricted to the fine-grained upper parts of channels. Pebbles within the sandy beds consist of a large variety of volcanic, granitic, metamorphic and sedimentary rocks. Also present in the CS facies assemblage are rare thin (<50 cm) beds of mudstone. Most of the mudstones are red-brown in color and are overprinted by calcic and vertic paleosols, although a few mudstones are green or red and green mottled, and display few pedogenic features.

Interpretation

The CS facies assemblage is interpreted to have been deposited by the ancestral Rio Grande, which flowed southward, parallel to the axis of the Palomas Basin (Mack and Leeder, 1999; Mack et al., 2002). As it does today, the main channel of the ancestral Rio Grande transported a bedload of granular, pebbly sand whose diverse composition is consistent with the variety of

rock types exposed within its large catchment. The abundance of trough crossbeds indicates that the dominant bedforms in the river were three-dimensional dunes (Perez-Arlucea et al., 2000). The rare mudstones represent overbank deposits that were associated with a low water table (red-brown mudstones with pedogenic features), a high water table (green mudstones), or a fluctuating water table (red and green mottled mudstones).

Fine-Grained Facies Assemblage (FG)

Description

The FG facies assemblage contains a variety of rock types, but is dominated by interbeds of reddish-brown mudstone and tan fine- and very fine-grained sand and silt that weather into badlands topography (Fig. 5E). Comprising less than 10 percent by thickness of the facies assemblage are thin beds of conglomerate, green claystone, microcrystalline calcite and three white, low-density, friable sands. The FG facies assemblage is restricted to the east side of the Mud Springs Mountains (Fig. 4).

Beds of red-brown mudstone range in thickness from a few centimeters to 3 m and commonly contain vertic and calcic paleosols. Interbedded with the mudstones are beds of light tan, very well sorted, very fine- or fine-grained sand and silt in beds 0.2 to 1.5 m thick. The sands and silts are commonly internally structureless, although a few beds have faint horizontal laminae. Some of the sand and silt beds display calcic paleosols.

Eight thin (0.2-1.0 m) beds of granule and pebble conglomerate are present within the FG facies assemblage in the study area. These conglomerates have sharp, erosional bases and are laterally discontinuous on a scale of a few hundred meters. Imbrication is common and small-scale crossbeds in sets <30 cm thick are locally present. Most of the conglomerates contain volcanic clasts and are similar to those conglomerates in the VCC facies assemblage. In contrast, two beds are composed exclusively of Pennsylvanian limestone and chert clasts and resemble conglomerates in the SCC facies assemblage.

Ranging from 0.2 to 1.1 m thick, 12 beds of white, microcrystalline calcite are interbedded with mudstone and fine sand or silt in the FG facies assemblage (Fig. 5F). The calcite beds have sharp bases and tops and generally are internally structureless, although a few beds display nodular tops or have horizontal and vertical root traces in the upper 15 cm. Petrographically, the beds consist of ~95 percent microcrystalline calcite that encloses scattered grains of sand. Locally, the white calcite beds are stratigraphically underlain by thin (<0.5 m) beds of finely laminated, green claystone.

Three white, low-density, friable sands are also present locally in the FG facies assemblage. Two of the three beds are ~ 1 m thick, can be traced for about 40 m laterally and consist of numerous diatom frustules. The other bed, 0.1 m thick, is present in two canyons, where it can be traced for a few hundred meters laterally (Figs. 4). It is composed primarily of sand- and granule-sized pumice grains.

Interpretation

The FG facies assemblage is composed of a mixture of sediments that were potentially derived from three different dispersal systems. The volcanic-clast conglomerates were probably derived from the hanging wall mountains, whereas the eastern dip slope of the Mud Springs Mountains was the source of the limestone- and chert-clast conglomerates. In all cases, the conglomerate beds are thin and relatively fine grained, suggesting they were deposited on the distal edge of their respective fans and/or represent small splays of larger fan channels.

The red-brown mudstones could have been derived from overbank flooding of hanging wall-derived alluvial-fan channels, from flooding of the axial Rio Grande and/or were deposited on extremely low-gradient alluvial flats located beyond to the toes of hanging wall-derived alluvial fans (c.f., Mack et al., 1994). The depositional environment of the sand and silt beds is difficult to interpret because of the paucity of sedimentary structures. They may represent deposition by sheetfloods on alluvial flats or by crevasse splays derived from alluvial-fan channels. However, their excellent sorting suggests the possibility of eolian deposition. If eolian in origin, the sand and silt could have blown out of ephemeral alluvial-fan channels or the Rio Grande during periods of low discharge. Many of the mudstones, sands and silts display stage II morphology calcic nodules and tubules, indicating thousands of years of landscape stability and soil formation between depositional events (Gile et al., 1966). The fine-grained floodplain and/or alluvial flat was locally occupied by small ponds, represented by diatomites and laminated green claystones. The pumiceous bed has been correlated geochemically with a 3.1-Ma pumice bed that was first recognized in axial-fluvial facies of the Palomas Formation ~50 km south of the Mud Springs Mountains (Mack et al., 1996; Mack et al., 2009). The pumice is interpreted to have been derived from the Mt. Taylor volcanic field in northwestern New Mexico and was transported to southern New Mexico via the ancestral Rio Puerco and Rio Grande (Mack et al., 2009).

The white, microcrystalline calcite beds are interpreted to be the products of precipitation and replacement of sandy parent material from shallow groundwater (Mack et al., 2000). The nodular top of some beds may have formed by capillary draw just above the water table, while the presence of root traces in upper part of other beds may indicate precipitation near the land surface. The calcite beds thicken and become more numerous toward the Mud Spring Mountains, suggesting that catchments in that range were the source of the laterally flowing groundwater.

SPATIAL AND TEMPORAL DISTRIBUTION OF FACIES ASSEMBLAGES

The facies assemblages are the product of five different sediment dispersal systems (Fig. 6): (1) alluvial fans derived from the hanging wall mountains of the Palomas Basin (VCC facies assemblage), (2) alluvial fans derived from the footwall scarp of the Caballo Mountains (SGMG facies assemblage), (3) allu-

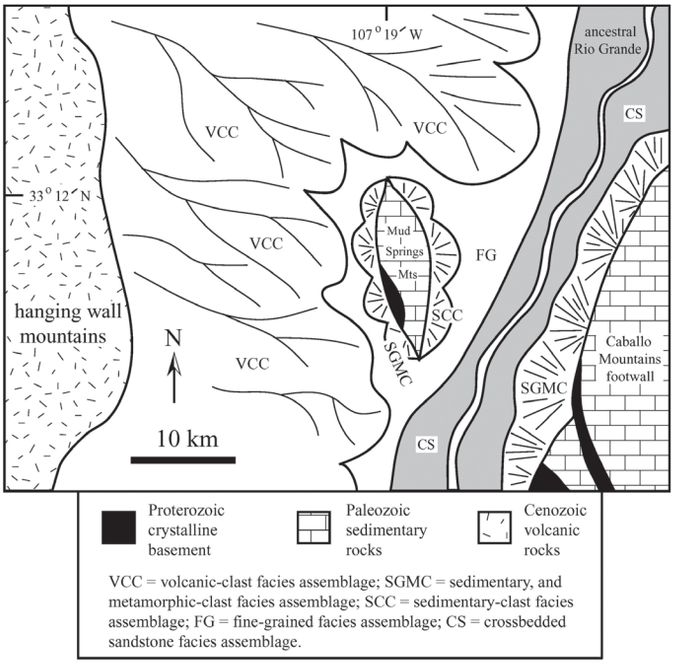


FIGURE 6. Schematic map of the distribution of facies assemblages of the Pliocene-lower Pleistocene Palomas Formation in the northern Palomas Basin, south-central New Mexico.

vial fans derived from the footwall scarp of the intrabasinal Mud Springs Mountains (SGMC facies assemblage), (4) alluvial fans derived from the eastern dip slope of the intrabasinal Mud Springs Mountains (SCC facies assemblage) and (5) axial-fluvial sediment deposited by the south-flowing ancestral Rio Grande (CS facies assemblage). In addition, the fine-grained facies assemblage (FG) covers a large area in the northeastern part of the map area and contains a mixture of sediment derived from the hanging wall mountains and the eastern dip slope of the Mud Springs Mountains, as well as fine sediment deposited by the ancestral Rio Grande (Figs. 4 and 6). The nature of the interaction between the facies assemblages is discussed below.

Western and Southwestern flanks of the Mud Springs Mountains

Alluvial fans derived from the hanging wall mountains of the Palomas Basin (VCC facies assemblage) spread eastward and southeastward into the basin, occupying most of the western and southwestern part of the map area (Figs. 4 and 6). The VCC facies assemblage is bordered to the southeast by axial-fluvial sediment (CS facies assemblage) and to the east and northeast by alluvial-fan deposits derived from the western footwall scarp of the Mud Springs Mountains (SGMC facies assemblage). Generally extending <1 km from the mountain front, the SGMC facies assemblage is locally capped by the Cuchillo geomorphic surface (Figs. 4 and 7). Late in the history of Palomas deposition, a ~13-m-thick tongue of the VCC facies assemblage was deposited within ~0.5 km of the footwall scarp of the Mud Springs Mountains, while at about the same time at the southern tip of the range, the VCC facies assemblage was deposited directly on bedrock, without any intervening Mud Springs Mountains-derived, alluvial-fan sediment (SGMC facies assemblage) (Figs. 4 and 7). Even in their most distal reaches, the VCC hanging wall-derived fans were characterized by relatively deep (2 to 7 m), broad (10s m) channels that transported pebble- and cobble-sized clasts.

The interaction between alluvial fans derived from the hanging wall mountains (VCC) and from the footwall scarp of the Mud Springs Mountains (SGMC) is especially well exposed ~1 km northeast of the mouth of Cantrell Canyon, where a photographic panel was constructed (Figs. 4 and 8). The panel consists of two parts that are oriented at right angles to each other, with the longer, southern part oriented north-south (Fig. 8). Within each of the parts there are cliff exposures hundreds of meters long and 5 to 10 m high, which form the nucleus of the panel. Key beds were traced between the cliff faces via closely spaced outcrops. The general paleocurrent direction for conglomerates of the SGMC facies assemblage is southwestward, toward the viewer of Figure 8, while that of the VCC conglomerates is generally southeastward, from left to right across the panel.

The majority of the basal 5.5 m of the panel consists of alluvial-fan sediment derived from the hanging wall mountains of the Palomas Basin (VCC facies assemblage). This part of the panel is dominated by granular, pebbly sand, but also has a few thin (<1 m), discontinuous channels filled with pebble conglomerate. Three calcic paleosols are present in the sands, all of which

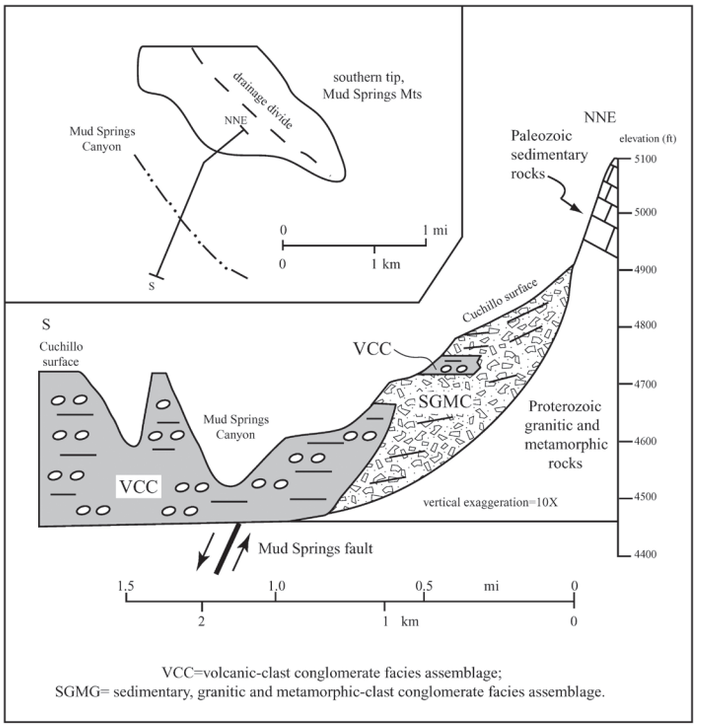


FIGURE 7. Cross section from the southwestern flank of the Mud Springs Mountains into the adjacent Palomas Basin showing the distribution of facies assemblages of the Palomas Formation (Pliocene-lower Pleistocene). Line of cross section is shown on inset map and on Figures 2 and 4.

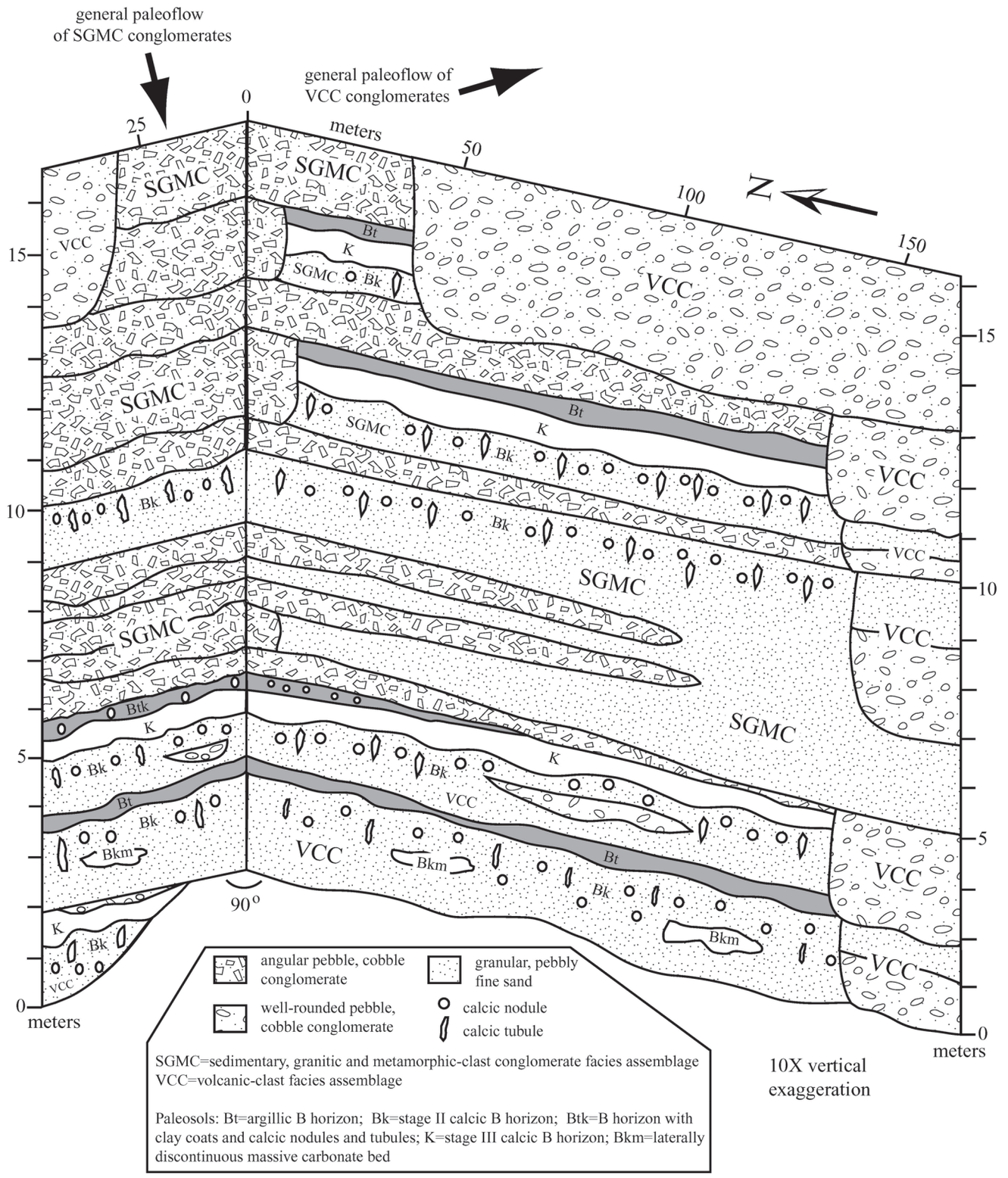


FIGURE 8. The distribution of facies assemblages and paleosols, based on photographic panels of cliff exposures on the north side of Cantrell Canyon about 1 km north of its confluence with Mud Springs Canyon (Fig. 4).

have erosional upper contacts. The middle paleosol has a stage II Bk horizon and is less mature than the upper and lower paleosols that have stage III K horizons. These paleosols indicate long periods of landscape stability (10^3 to 10^4 yrs), when neither Mud Springs-derived nor hanging wall-derived alluvial-fan deposition occurred at this location.

Most of the upper part of the panel is composed of sediment derived from the footwall scarp of the Mud Springs Mountains (SGMC facies assemblage), including both alluvial-fan conglomerates and granular, pebbly sand. The relative abundance of SGMC conglomerate increases northward, resulting in intervals 3 to 7 m thick of stacked conglomerate channels. Five SGMC gravel channels prograded southward across SGMC pebbly sand, while the lowest SGMC conglomerate prograded across and truncated VCC pebbly sand. Progradation of four of these SGMC conglomerates, including the basal one, followed long periods of landscape stability, indicated by the presence of stage II and III calcic paleosols.

The southern part of the panel is dominated by thick channel conglomerates derived from the hanging wall mountains (VCC facies assemblage). These channels truncate finer grained sediment of both the VCC and SGMC facies assemblages. The upper channel, which is present on both parts of the panel, is the thickest and most laterally extensive.

Interbedding of pebbly sand-rich and conglomerate-rich beds and the long periods of landscape stability and paleosol development between depositional events may be explained by a combination of lobe-switching and channel-fan entrenchment on both the hanging wall-derived alluvial fans (VCC facies assemblage) and on the fans derived from the footwall scarp of the Mud Springs Mountains (SGMC facies assemblage). Coarse gravel deposition would have occurred on the active lobe of a fan, while areas directly adjacent to the active lobe would have received pebbly sand deposited in small channels or as splays from the active channels. Farthest from the active lobe, virtually no deposition took place, allowing the development of mature calcic paleosols. As the active lobes switched back and forth across the fans, the features illustrated in the panel could have been created. Fan-channel entrenchment would also leave adjacent areas of a fan above the level of active sediment transport and deposition. Entrenchment could have been caused by climate change or by tectonic tilting of the floor of the basin, the latter of which is discussed in more detail below.

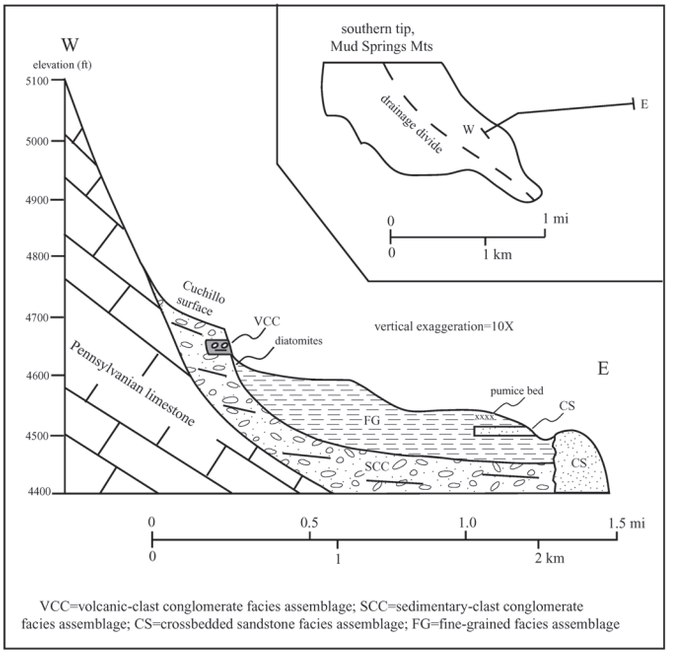


FIGURE 9. Cross section from the southeastern flank of the Mud Springs Mountains into the adjacent Palomas Basin showing the distribution of facies assemblages of the Palomas Formation (Pliocene-lower Pleistocene). Line of cross section is shown on inset map and on Figures 2 and 4.

Eastern Flank of the Mud Springs Mountains

The oldest basin-fill deposits exposed on the eastern flank of the Mud Springs Mountains are conglomerates derived from the eastern dip slope of the Mud Springs Mountains (SCC facies assemblage) (Figs. 4 and 9). Two kilometers north of the map area these basal conglomerates unconformably overlie Paleozoic bedrock (Maxwell and Oakman, 1990).

Most of the eastern part of the map area contains the fine-grained facies assemblage (FG) and axial-fluvial sediment deposited by the south-flowing, ancestral Rio Grande (CS facies assemblage). The fine-grained facies assemblage onlaps the dip slope of the Mud Springs Mountains to within 0.5 km of the mountain front (Figs. 4 and 9). The predominance of mudstone and very fine to fine sand with numerous calcic paleosols suggests a relatively gentle depositional surface that periodically experienced thousands of years of non-deposition and soil formation between depositional events. Only twice during deposition of the fine-grained facies assemblage did thin (<1 m) beds of Mud Springs Mountains-derived, alluvial-fan gravels of the SCC facies assemblage prograde basinward across the muddy or sandy sediment surface.

The fine-grained facies assemblage and the crossbedded sand facies assemblage (CS) deposited by the south-flowing ancestral Rio Grande have a relatively abrupt north to northeast-trending contact (Fig. 4). Although outcrops along this contact are sparse, there are several examples of truncation of mudstones and fine sands by crossbedded channel sands. In addition, one mappable tongue of fluvial sand ~5 m thick extends ~1 km west of the main body of the CS facies assemblage (Figs. 4 and 9). Throughout much of its outcrop area the tongue consists exclusively of crossbedded sand, although several outcrops have two fluvial-channel sands separated by a thin (<1 m) red mudstone with stage II calcic nodules and vertic features. The paleosol-bearing mudstone indicates that there were two fluvial-channel incursions separated in time by overbank deposition and several thousand years of landscape stability and soil formation. Where absent, the bed of overbank mud was presumably eroded by the younger channel.

Exposures of alluvial-fan sediment derived from the hanging wall mountains of the Palomas Basin (VCC facies assemblage) are relatively uncommon and widely scattered in the eastern part of the map area. They exist as thin (<1 m), laterally discontinuous gravel beds in the fine-grained facies assemblage (FG) and as a relatively continuous, northwest-trending belt of gravel-rich beds that extend to within ~0.5 km of the eastern dip slope of the Mud Springs Mountains (Figs. 4 and 9). In a few outcrops there is evidence that the sharp western contact between the northwest-

trending belt of VCC conglomerates and the SCC facies assemblage resulted from toe-cutting of the SCC deposits by VCC alluvial-fan channels. However, during the final stage of deposition of the Palomas Formation, SCC alluvial fans prograded over the interval of VCC conglomerates (Figs. 4 and 9).

Western Flank of the Caballo Mountains

Exposed along the western flank of the Caballo Mountains are alluvial-fan conglomerates derived from the adjacent footwall scarp (SGMC facies assemblage), as well as axial-fluvial sediment deposited by the ancestral Rio Grande (CS facies assemblage) (Fig. 10). Throughout the central and southern parts of the Palomas Basin, and locally exposed west of the map and cross section of Figure 10, Caballo Mountains-derived, alluvial-fan deposits extend <1 km west of the mountain front, beyond which axial-fluvial sediment is exposed (Seager et al., 1982; Mack and Leeder, 1999). On the map of Figure 10, a single tongue of axial-fluvial sediment was mapped between conglomerates of the SGMC facies assemblage, extending to within ~600 m of the Hot Springs fault. However, one cliff exposure reveals that the mapped tongue consists of two fluvial sands separated by a few meters of alluvial-fan conglomerate (Fig. 10).

Late in the history of Palomas deposition, the Hot Springs fault became inactive and the Caballo Mountains footwall scarp underwent erosional retreat. As a result, conglomerates of the SGMC facies assemblage onlapped the fault. Following deposition of the Palomas Formation, the fault was reactivated (Fig. 10). Despite residing on the upthrown side of the Hot Springs fault, local exposures of SGMC conglomerate have survived erosion.

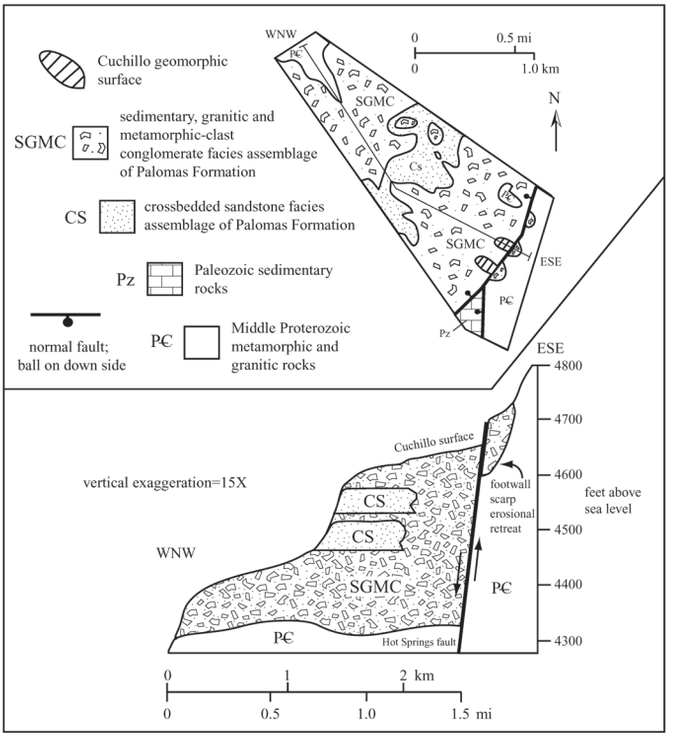


FIGURE 10. Cross section and simplified geologic map from the western flank of the Caballo Mountains into the adjacent Palomas Basin showing the distribution of facies assemblages of the Palomas Formation (Pliocene-lower Pleistocene). Line of cross section is shown on inset map and on Figure 2. Geologic map is adapted from Kelley and Silver (1952) and mapping done in this study.

DISCUSSION

The distribution of Pliocene-lower Pleistocene strata in the northern Palomas Basin is characterized by narrow (<1 km wide) aprons of coarse alluvial-fan deposits adjacent to the Mud Springs Mountains and Caballo Mountains, a narrow (<5 km), south-trending belt of axial-fluvial sediment deposited by the ancestral Rio Grande and a broad (15-20 km) belt of alluvial-fan sediment derived from the hanging wall mountain ranges (Fig. 6). Hanging wall-derived alluvial-fan channels extended to the toe of their fans and toe-cut the alluvial fans derived from the Mud Springs Mountains, locally depositing sediment within 0.5 km of the mountain front. Similarly, erosion by the ancestral Rio Grande maintained a relatively sharp boundary with the fine-grained facies assemblage, and the river periodically moved to within ~0.6 km of the Caballo Mountains by toe-cutting the adjacent alluvial fans (c.f., Mack and Leeder, 1999). The extrinsic variables responsible for these processes and the resultant basin architecture are discussed below.

Catchment Size and Tectonic Tilt

Many of the architectural features in the northern Palomas Basin are the result of the size of the catchments in the surrounding mountain ranges and the tilt direction of the basin floor, both of which are inherent to the structural configuration of a half graben. Catchments on the footwall scarp of a half graben have high relief, but are generally small in area. Conversely, catchments on the hanging wall of a half graben occupy greater area than those on the footwall scarp (Leeder and Gawthorpe, 1987). Because the area of an alluvial fan is directly proportional to the area of its catchment (Bull, 1972), variation in the size of alluvial fans during Palomas deposition probably was strongly influenced by differences in the size of the catchments. Although the exact size of the catchments in the mountains surrounding the Palomas Basin during Pliocene and early Pleistocene time is not known, the size of the modern catchments provide a guide to past conditions, because the *relative differences* in catchment size probably have not changed appreciably since early Pleistocene time (Fig. 11). The modern catchments in the hanging wall mountains of the northern Palomas Basin are one to two orders of magnitude larger than those in the Caballo Mountains and Mud Springs Mountains, which, assuming similar conditions in Pliocene and early Pleistocene time, would have contributed to the large size of the hanging wall-derived alluvial fans (VCC) and the small size of the fans derived from the footwall scarps of the Caballo and Mud Springs Mountains (SGMC) during Palomas deposition (Fig.11; Mack and Leeder, 1999; Mack et al., 2002, 2008). Catchments on the eastern dip slope of the Mud Spring Mountains are larger than those on the footwall scarp of the same range, but are sev-

eral orders of magnitude smaller than those in the hanging wall mountains (Black Range, Salado Hills, Sierra Cuchillo), resulting in small size of the SCC alluvial fans.

Flow of the hanging wall-derived, alluvial-fan channels (VCC facies assemblage) far into the basin would have been enhanced by eastward tilt of the Palomas half graben and block rotation on west-dipping faults in the Black Range (Seager et al., 1982). During periods of fault activity, an increase in slope of the basin floor toward the footwall would have caused the hanging wall-derived, alluvial-fan channels to incise, thereby transporting coarse gravel farther into the basin. The regional slope of the Palomas half graben would have been unaffected by the Mud Springs fault, if, as has been suggested, it was inactive during Palomas deposition (Maxwell and Oakman, 1990).

The location and width of the axial-fluvial belt, as well as periodic large-scale (km's) incursions of the river toward the footwall, also may be related to basin tilting and relative catchment size. In the Palomas half graben, the axial river would have been positioned at the topographically lowest area, whose location was dictated by the asymmetrical, eastward tilt of the basin. This relationship, along with the general southward flow of the river, resulted in the relatively narrow, south-trending belt of axial-fluvial sediment (CS) located near the Caballo Mountains footwall (Figs. 4 and 6). Smaller-scale (10's to 100's m), lateral migration of the Rio Grande also could have resulted from avulsion and meander-loop migration.

The relatively sharp boundary between the VCC and CS facies assemblages in the southeastern part of the map area is difficult to interpret because of poor exposures (Fig. 4). It probably reflects a balance between progradation of the large, hanging wall-derived VCC alluvial fans onto the floodplain and periodic toe-cutting of the fans by the axial river.

During periods of active faulting and concomitant tilting of the basin toward the Caballo Mountains footwall, the river moved eastward by toe-cutting the Caballo Mountains-derived alluvial fans (SGMC facies assemblage; Fig. 10). Consistent with this interpretation is the presence of two fluvial tongues adjacent to the Hot Springs fault in the study area, the upper of which is ~10 m below the Cuchillo surface, while 10 km to the south, the northern segment of the Caballo fault has one fluvial tongue whose top is 40 m below the Cuchillo surface (Fig. 1; Mack and Leeder, 1999). This suggests activity on the two fault segments was not synchronous, resulting in a unique number and age of fluvial incursions for each fault segment. If climate had been the driving force behind fluvial incursions toward the footwall, then the number and age of fluvial incursions toward the footwall should have been the same throughout the basin.

Prior to 3.1 Ma, there were two kilometer-scale, westward incursions of the ancestral Rio Grande, which were mapped as a single unit (Figs. 4 and 9). Following the ideas discussed above, these westward fluvial incursions may represent periods of inactivity on the basin-bounding fault, during which time the river infilled the topographic low near the footwall and subsequently spread westward across the basin. Westward movement of the river in the map area also may have been enhanced by relatively easy erosion of mud and fine sand of the fine-grained facies

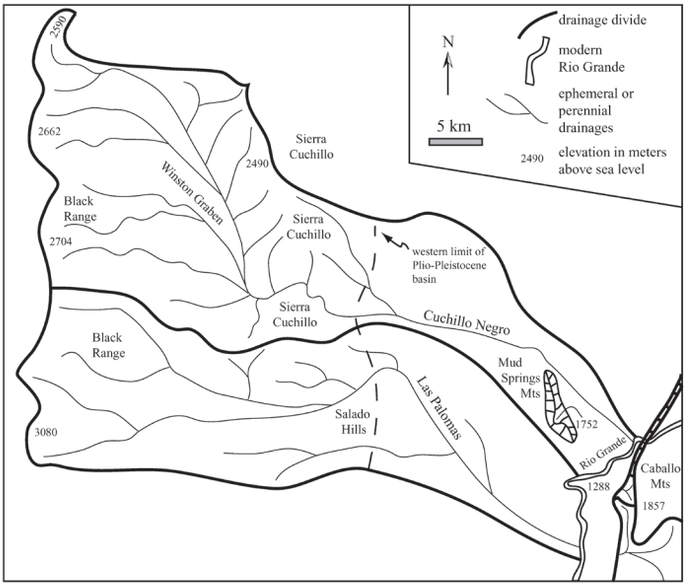


FIGURE 11. The distribution of modern catchments and their elevations in the study area.

assemblage.

The ability of the axial river to toe-cut the footwall-derived alluvial fans reflects, in part, the small size of the fans, which were fed from small catchments on the footwall scarp (Fig. 11). In contrast, the ancestral Rio Grande was fed from numerous mountainous catchments, many over 3000 m in elevation, distributed over a distance of ~500 km upstream of the Palomas Basin (Pazzaglia and Hawley, 2004). This large drainage basin is expected to have provided the discharge and stream power necessary to erode the footwall-derived alluvial fans.

Paleolimatic Influence on Basin Architecture

While large catchments and tectonic tilting may explain why the hanging wall-derived fans (VCC facies assemblage) spread coarse sediment far into the basin, they cannot, by themselves, account for the high discharge and erosive power that the hanging wall-derived fan channels needed to toe-cut the alluvial fans derived from the Mud Springs Mountains. The process of toe-cutting probably was influenced also by local paleoclimatic conditions. As was the case with catchment size, the modern climate in the study area provides insight into the relative differences in paleoclimatic conditions between the Palomas Basin and surrounding mountains during Pliocene-early Pleistocene time.

Maximum elevations in the modern hanging wall catchments range from 2490 to 3080, over 1000 m higher than the floor of the Palomas Basin and the highest points in the Caballo and Mud Springs Mountains within the study area (Fig. 11). High elevation in the modern hanging wall catchments results in cooler annual temperatures and approximately twice the annual precipitation than on the adjacent basin floor (Mueller, 1986). As a result, the modern hanging wall catchments support coniferous forest, while the adjacent basin is desert grassland with scattered shrubs (Dick-Peddie, 1986). Assuming similar differences between the basin-

bounding mountains during Palomas deposition, then the combination of large catchments, high precipitation and heavy vegetative cover may have provided Pliocene-early Pleistocene hanging wall-derived, alluvial-fan channels with high discharge and low sediment/water ratios, resulting in substantial erosive power.

Paleoclimate may also have played a role in the ability of the ancestral Rio Grande to toe-cut alluvial fans and the fine-grained facies assemblage. Prior to damming of the river in the 20th Century, the Rio Grande experienced major flooding in the late Spring and early Summer as the result of snowmelt in its northern catchments (Mack et al., 2008). Similar periods of high runoff may have enhanced the erosive power of the ancestral Rio Grande during deposition of the Palomas Formation.

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Debris flow in Winston New Mexico area, ca 1953. NMBGMR Photo Archive No. p-00329.