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Tectonic control on facies distribution of the Camp Rice and Palomas Formations (Pliocene-Pleistocene) in the southern Rio Grande rift

GREG H. MACK
WILLIAM R. SEAGER } *Department of Earth Sciences, New Mexico State University, Las Cruces, New Mexico 88003*

ABSTRACT

The Pliocene-Pleistocene Camp Rice and Palomas Formations in the Rio Grande rift of southern New Mexico provide an excellent test of the role of basin symmetry in the distribution of piedmont and axial-fluvial facies. In the asymmetrical Palomas and northern Mesilla basins, the axial-fluvial facies is characterized by multistory channel sands and sandstones and is concentrated near the locus of maximum subsidence within a few kilometers of the footwall scarp. Fanglomerate derived from the footwall uplift extends only a few kilometers or less from the footwall scarp, whereas alluvial-fan and alluvial-flat conglomerate, sand, and mudstone deposited on the hanging-wall dip slope occupy a much wider outcrop belt. In the symmetrical Hatch-Rincon basin, the axial-fluvial facies extends to within a few kilometers of both the northern and southern basin margins and has a much higher percentage of fine-grained overbank deposits, some of which contain calcareous paleosols. In all three basins, a tongue of fanglomerate as much as 30 m thick prograded over the axial-fluvial facies near the end of Camp Rice and Palomas deposition.

The distribution of facies of the Camp Rice and Palomas Formations supports previously published models that predict a two-stage history of asymmetrically subsiding basins. During tectonically active periods, axial-river channels preferentially avulse into the topographically lowest area of the alluvial plain, which is directly above the axis of maximum subsidence. During the postorogenic stage, when erosion rate exceeds subsidence rate, coarse, transverse-dispersed sediment progrades toward the center of the basin.

INTRODUCTION

Tectonism exerts a significant influence on sedimentation, not only on the type and thickness of sediment, but also on the areal distribution of facies within the basin. The role of tectonism in determining facies distribution is especially important in asymmetrically subsiding basins, such as half-grabens, foreland basins, and some strike-slip basins (Bridge and Leeder, 1979; Leeder and Gawthorpe, 1987; Alexander and Leeder, 1987; Blair and Bilodeau, 1988; Heller and others, 1988). Recent depositional models, based largely on theoretical grounds, suggest that periods of rapid subsidence in asymmetrical basins result in deposition of fine-grained sediment, such as lacustrine or axial-fluvial facies, directly above the locus of maximum subsidence and close to the margin of the uplifted terrane (Fig. 1; Bridge and Leeder, 1979; Leeder and Gawthorpe, 1987; Alexander and Leeder, 1987; Blair and Bilodeau, 1988; Heller and others, 1988). Because the rate of denudation is significantly slower than the rates of uplift and subsidence, coarse-grained alluvial-fan or braid-plain sediment is restricted to a narrow zone directly adjacent to the uplifted terrane. As the rate of subsidence and uplift diminishes, however, erosion rate eventually surpasses subsidence rate and a sheet of coarse sediment spreads across the basin (Fig. 1). Thus, near the locus of maximum subsidence, fine-grained facies are synorogenic and coarser-grained facies are postorogenic (Fig. 1).

Although intuitively appealing, this model has not yet been fully tested in the rock record. Presented in support of this model are tectonically active extensional basins, such as the Death Valley region of Califor-

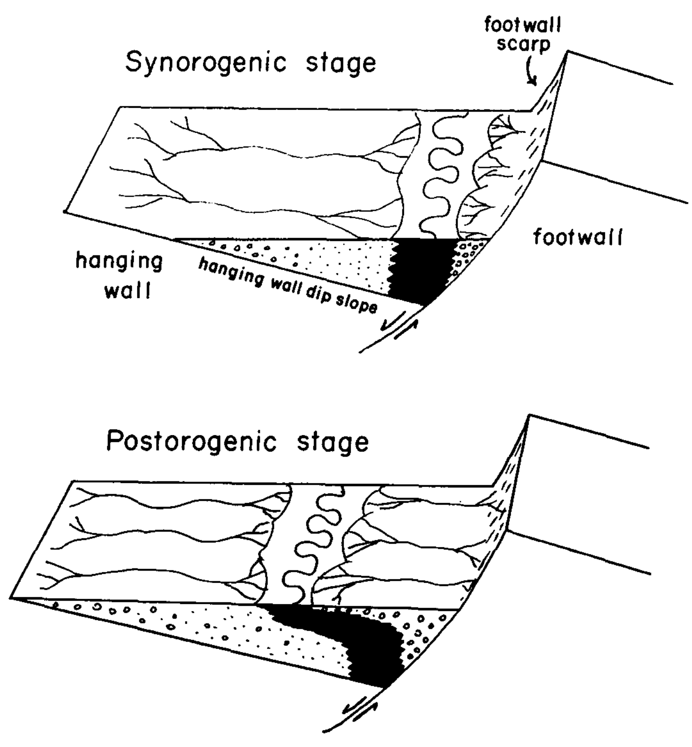

Figure 1. Two-stage depositional model for asymmetrically subsiding basins, adapted from Leeder and Gawthorpe (1987), Blair and Bilodeau (1988), and Heller and others (1988).

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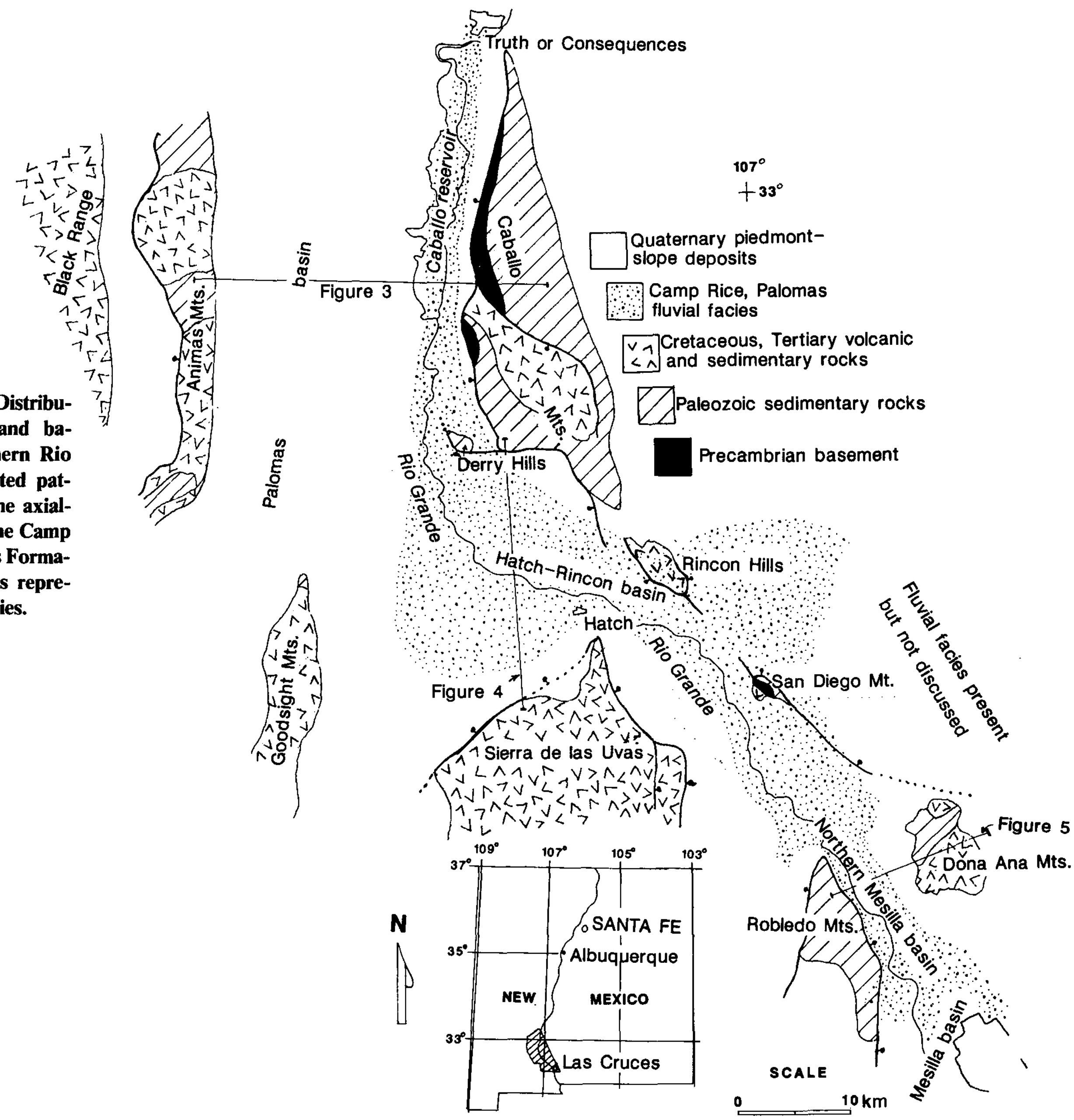

Figure 2. Distribution of uplifts and basins in the southern Rio Grande rift. Dotted pattern represents the axial-fluvial facies of the Camp Rice and Palomas Formations; blank areas represent piedmont facies.

nia (Hunt and Mabey, 1966; Hooke, 1972; Blair and Bilodeau, 1988; Heller and others, 1988), the Hebgen Lake area of Montana (Alexander and Leeder, 1987; Leeder and Alexander, 1987; Leeder and Gawthorpe, 1987), and the margin of the Aegean Sea (Leeder and Gawthorpe, 1987), as well as a variety of Phanerozoic depositional systems (Steel and others, 1977; Gloppen and Steel, 1981; Hayward, 1984; LeTourneau, 1985; Beck and Vondra, 1985a, 1985b; Villien and Kligfield, 1986; Blair, 1987; Heller and others, 1988). The Neogene, strike-slip Ridge basin of southern California (Link and Osborne, 1978; Crowell and Link, 1982; Link, 1987) and Neogene rift basins in East Africa (Frostick and others, 1986; Frostick and Reid, 1987) also are examples of well-documented asymmetrical basins. Modern analogues are compelling because the distribution of tectonic elements is reflected in the modern topography, and historical seismic activity constrains the tectonic regime. Modern depositional systems are less than ideal, however, because they represent only a single instant of time, and it is not always clear for how long the depositional system will persist, or how to translate the surface geomorphology into a three-dimensional unit of sedimentary rocks. In contrast, ancient depositional systems have gone to completion, but in many cases, the basin shape and the chronology of tectonic and/or depositional events are not well constrained. Thus, in many ancient systems, the relationship between tectonics and sedimentation is an interpretation and cannot serve as proof of a particular model.

We present herein a tectono-depositional model for the Pliocene-Pleistocene Camp Rice and Palomas Formations in the southern Rio Grande rift that combines many of the best attributes of modern and ancient analogues. Although an ancient analogue, deposition of the Camp Rice and Palomas Formations ended only a half million years ago (Strain,

TABLE 1. SALIENT STRATIGRAPHIC DATA FROM THE SOUTHERN RIO GRANDE RIFT

Age	Rift stage	Formation name	Thickness (meters)	Rock types	Age indicators: dates
					Basalt flows (post–Camp Rice): 0.55, 0.49 Ma (K-Ar) (Seager and others, 1984)
Pleistocene					Pearlette "O" ash: 0.6 Ma (K-Ar) (Hawley, 1981)
		Camp Rice and Palomas Formations	0–150	Conglomerate, sandstone, shale, basalt	Bishop ash: 0.7 Ma (K-Ar) (Hawley, 1981)
					Las Palomas ash: 1.25–1.45 Ma (K-Ar) (Lozinsky and Hawley, 1986a)
					Basalt flow: 3.1, 2.9 Ma (K-Ar) (Seager and others, 1984; Bachman and Mehnert, 1978)
Late Pleistocene	Late				Medial Blancan fauna (Tedford, 1981; Lucas and Oakes, 1986)
					Very early Blancan fauna (Repenning and May, 1986)
Unconformity					Basalt flow (beneath Palomas Fm.): 4.5 Ma (K-Ar) (Seager and others, 1984)
		Rincon Valley Formation	0–500	Conglomerate, sandstone, shale, basalt	Selden Basalt flow: 9.6 Ma (K-Ar) (Seager and others, 1984)
Miocene					
		Hayner Ranch Formation	0–1,000	Conglomerate, sandstone, shale	
	Early				
		Upper Thurman Formation	100–420	Sandstone, shale, conglomerate	
Oligocene		Uvas Basaltic Andesite	0–300	Basaltic andesite flows, plugs; minor sedimentary rocks	26–27.4 Ma (K-Ar) (Clemons, 1979) 29 Ma (Seager and others, 1984)

1966; Gile and others, 1981; Seager and others, 1984). Consequently, the basins and uplifts produced during the extensional tectonism, which is extant, are largely intact and are responsible for the present-day topography (Seager and others, 1984). Deposition ended when the Ancestral Rio Grande and tributary arroyos began to entrench the basin, probably in response to capture of the upper Rio Grande by the lower Rio Grande (Hawley and Kottlowski, 1969; Hawley, 1981). The Rio Grande and tributaries have downcut as much as 100 m, locally exposing the entire Camp Rice and Palomas Formations. Thus, not only are the basins and uplifts intact, but the basin-fill sediment can be examined in outcrop.

The Pliocene-Pleistocene Camp Rice and Palomas Formations provide an excellent test of the role of basin symmetry on the distribution of facies because within a 100-km stretch of the southern Rio Grande rift, the basins change from eastward-tilted half-graben in the north (Palomas basin), to full graben in the central part (Hatch-Rincon basin), to westward-tilted half-graben in the south (northern Mesilla basin) (Fig. 2). The model in Figure 1 predicts that the distribution of the axial-fluvial facies of the Camp Rice and Palomas Formations will be different in each of these three basins in response to differences in basin symmetry. The postorogenic stage of the asymmetrical basin model (Fig. 1) can also be tested in the Camp Rice and Palomas Formations. Regional mapping indicates that tectonic uplift and complementary basin subsidence during the most recent stage of rifting reached a peak between 9.6 and 7.1 Ma and has subsequently diminished in magnitude (Seager and others, 1984). The depositional model of Figure 1 predicts that near the end of Camp Rice and Palomas deposition, a fanglomerate facies should have prograded from the footwall block of the half-grabens over the axial-fluvial facies.

TECTONIC AND STRATIGRAPHIC SETTING

The southern Rio Grande rift evolved in two distinct stages (Table 1; Chapin and Seager, 1975; Seager and others, 1984; Morgan and others, 1986). The earlier stage, which began approximately 29 m.y. ago, was characterized by emplacement of basaltic andesite flows; by formation of broad, northwest-trending basins; and by incipient uplift of some of the fault-block mountains of the region. The second stage of uplift and basin subsidence began between 9.6 and 7.1 m.y. ago and has continued with decreasing intensity to the present day (Seager and others, 1984). The second stage is responsible for the largely north-trending ranges and basins that dominate the present topography.

The exposed sedimentary record of the second stage of rifting, and the subject of this study, includes the Palomas Formation in the Palomas basin (Lozinsky and Hawley, 1986a, 1986b) and the Camp Rice Formation in basins farther south (Fig. 2; Table 1; Strain, 1966; Seager and others, 1982, 1987). The age of the Camp Rice and Palomas Formations is bracketed between 4.5 m.y. (early Pliocene) and 0.5 m.y. (middle Pleistocene) by vertebrate fauna and radiometrically dated volcanic rocks (Table 1; Bachman and Mehnert, 1978; Hawley, 1981; Tedford, 1981; Seager and others, 1984; Lucas and Oakes, 1986; Repenning and May, 1986). Deposition of the Camp Rice and Palomas Formations ended when the Ancestral Rio Grande and its tributary arroyos began to entrench the basins (Hawley and Kottlowski, 1969; Hawley, 1981). Entrenchment was already underway by middle Pleistocene time because two basalt flows, dated at 0.55 ± 0.03 and 0.49 ± 0.02 Ma, erupted onto the constructional top of the Camp Rice Formation near Las Cruces and flowed down the incised valley to within 35 m of the present flood plain (Seager and others, 1984).

The Camp Rice and Palomas Formations have a maximum thickness of 150 m and are generally subdivided for mapping purposes into piedmont and axial-fluvial facies (Seager and others, 1971, 1976, 1982, 1987; Seager and Hawley, 1973; Seager and Clemons, 1975; Lozinsky, 1986; Seager and Mack, 1989a, 1989b). The piedmont facies was deposited by transverse drainage in alluvial-fan and alluvial-flat environments. The axial-fluvial facies was deposited by the Ancestral Rio Grande, which

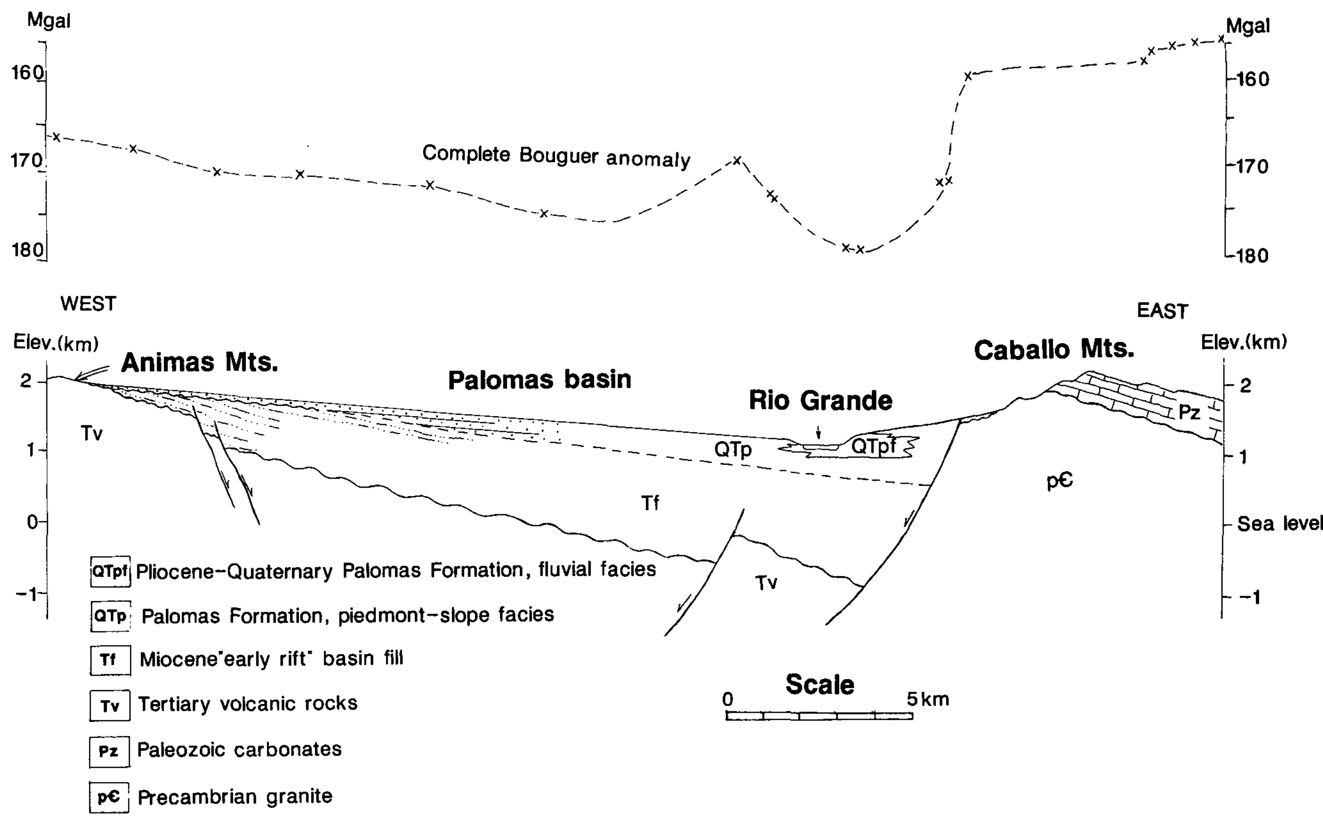

Figure 3. Cross section of the Palomas basin. See Figure 2 for location of section line. Dot pattern for the Palomas Formation and dash-dot pattern for "early" rift deposits in left part of diagram are designed to show angular unconformable relationship between these units.

flowed southward and emptied into large lakes in west Texas and northern Chihuahua, Mexico (Kottlowski, 1953, 1960; Ruhe, 1962; Reeves, 1965; Strain, 1966; Hawley and others, 1969).

DISTRIBUTION OF FACIES IN THE PALOMAS, HATCH-RINCON, AND NORTHERN MESILLA BASINS

Palomas Basin

Basin Tectonics. The north-trending Palomas basin is one of the major basins in the southern Rio Grande rift. Approximately 65 km long and 25 km wide, the basin is essentially an eastward-tilted half-graben as indicated by both surface and gravity data (Figs. 2 and 3; Keller and Cordell, 1983; Gilmer and others, 1986). Eastward-dipping Tertiary volcanic and sedimentary rocks of the Animas Mountains and Black Range form the hanging-wall uplift, whereas the much deeper, eastern part of the basin is bordered by a major fault zone, the Caballo–Red Hills fault. Structural relief across this fault zone from the estimated deepest part of the Palomas basin to the restored highest part of the adjacent Caballo uplift may approach 6 to 7 km. Maximum thickness of Neogene basin fill in the Palomas basin, estimated from gravity and seismic models, is 1.6 to 2.3 km (Gilmer and others, 1986; Harder and others, 1986).

Although most of the movement on the Caballo–Red Hills fault is probably of Miocene age, movement has continued repeatedly through the Pliocene and Quaternary, concurrently with deposition of the Palomas Formation and younger units. This is indicated not only by the fault truncation of proximal Palomas fan debris but also by composite piedmont scarps in Palomas alluvium (Machette, 1987; Foley and others, 1988). Low scarps are developed in younger parts of Palomas alluvium, as well as in post-Palomas fan gravel, whereas higher scarps are found truncating progressively older deposits. This is a clear indication that higher scarps are composite, the result of multiple faulting events during deposition of the Palomas and younger units. Another indication of active faulting during deposition of Palomas alluvium is the far greater thickness of Palomas fan alluvium on the downthrown side of the Caballo–Red Hills fault compared to the upthrown side.

Sedimentation. The distribution of facies of the Palomas Formation in the Palomas basin is characterized by (1) a narrow belt (≤2 km) of coarse piedmont conglomerate adjacent to and derived from the footwall block (Caballo Mountains), (2) a broad wedge as much as 20 km wide of piedmont conglomerate, sand and sandstone, and shale that was derived from the hanging wall (Animas Mountains and Black Range) and was deposited on the hanging-wall dip slope, and (3) a narrow belt (≤5 km) of axial-fluvial sand and sandstone located a few kilometers from the footwall scarp (Figs. 2 and 3).

The piedmont facies adjacent to the footwall scarp consists of medium- to thin-bedded boulder, cobble, and pebble conglomerate and a few beds or lenses of pebbly coarse sandstone. Most of the conglomerates are grain-supported and imbricated. The thicker beds commonly fill large-scale (tens of meters wide, meters deep) channels, whereas the thin-bedded conglomerates are laterally extensive and display broad (a few meters), shallow (tens of centimeters) cut-and-fill structures. The grain-supported conglomerates were water laid and resulted from channelized stream-flood sedimentation, in the case of the channel-form conglomerates, and unchannelized sheet-flood sedimentation, in the case of the thin-bedded conglomerates (Bull, 1972; Mack and Rasmussen, 1984). A few pebble conglomerates or pebbly sandstones have a single set of foresets as much as 1 m thick, which probably formed by migration of transverse bars (Hein and Walker, 1977; Gustavson, 1978; Massari, 1983). Also present are thick-bedded, poorly sorted conglomerates that lack any organized internal fabric. These disorganized conglomerates were probably deposited by debris flows (Bull, 1972; Shultz, 1984).

Cobble counts (n = 200) near Apache Canyon in the Caballo quadrangle (Seager and Mack, 1989b) indicate that the conglomerates are composed primarily of sedimentary clasts (90%). Limestone and dolomite clasts are much more abundant than sandstone/siltstone or chert clasts, a distribution that reflects the relative abundance of lithologies in the Paleozoic rocks exposed in the Caballo Mountains (Kelley and Silver, 1952; Seager and others, 1982). Also present in the conglomerates are about 10% granitic and gneissic clasts, derived from the Precambrian crystalline basement of the Caballo Mountains, and trace amounts of volcanic clasts.

The hanging-wall dip slope of the Palomas basin is covered by a laterally extensive wedge as much as 20 km wide of alluvial-fan and alluvial-flat detritus that was derived from the Animas Mountains and Black Range (Fig. 3). The most distal portion of this wedge is composed of interbedded conglomerate, sand and sandstone, and red or brown mudstone. The conglomerates make up about 35% of the stratigraphic section. Individual beds of conglomerate range in thickness from 0.5 to 7.5 m. The gravel-sized clasts are generally subrounded to rounded and rarely exceed 10 cm in length (A axis), although locally clasts as long as 17 cm were observed. Clasts in the conglomerates are almost exclusively of volcanic origin. The conglomerates generally are channel-form in cross section and scour as much as 1 m into the underlying sediment. Clast imbrication is especially common and indicates eastward paleoflow. Conglomerate beds commonly contain lenses or interbeds as much as 1.5 m thick of pebbly, granular medium- to coarse-grained sand or sandstone, which display horizontal laminations with parting lineation and, less commonly, trough cross-beds. The percentage of conglomerate increases westward, and within a few kilometers of the Animas Mountains and Black Range, the piedmont facies closely resembles the piedmont facies deposited next to the Caballo Mountains.

The remainder of the sediment in the distal portion of the hanging-wall wedge consists of very fine- to fine-grained sand or silt (45%) and red or brown mudstone (20%) that are interbedded on a scale of 0.2 to 3 m. Both lithologies generally lack primary sedimentary structures, but pedogenic features are common. Paleosols in the fine-grained sediment are characterized by scattered calcareous nodules, tubules, and rhizoliths, which correspond to stage II caliche (Bk horizon), or by massive stage III caliche as much as 0.5 m thick (Gile and others, 1966). The caliche horizons are commonly overlain by dark brown clayey sand/silt or mudstone that contains root traces and in some cases calcareous rhizoliths. Hematitic clay coats around sand grains (ferri-argillans; Brewer, 1964) suggest that the clayey sand/silt is an argillic B horizon (Brewer, 1964; Buol and others, 1980). Twelve individual paleosols were recognized in a 55-m-thick section of distal piedmont facies in the Caballo quadrangle (Seager and Mack, 1989b). The well-developed paleosols indicate that sedimentation was very intermittent or that periodically the alluvial fan or alluvial flat was incised and higher-level surfaces were abandoned (Hooke, 1972; Leeder and Gawthorpe, 1987). Stage III caliches are thought to require as much as 50 ka to develop, whereas stage II caliche may form in as little as 9 to 15 ka (Birkeland, 1974; Gile and others, 1981; Holliday and Gustavson, 1988).

The axial-fluvial facies of the Palomas Formation in the Palomas basin consists of gray, medium-grained sand or sandstone. Scattered grains or thin (≤50 cm) lenses of subrounded to rounded pebbles and small cobbles (≤8 cm length) are present but volumetrically are of minor importance. Clast composition includes volcanic, granitic, metamorphic, and sedimentary, and reflects mixing of detritus from both the adjacent hanging-wall and footwall blocks, as well as from source terranes north of the Palomas basin. Trough cross-beds are particularly abundant and range in set thickness from 20 to 150 cm. Less common are horizontal laminations that are intercalated with the cross-beds. Locally, the sand or sandstone contains a few calcareous nodules or rhizoliths. Green and red shale is quite rare and occurs as thin lenses (10 m wide, 20 cm thick) or as rip-up clasts within the sand or sandstone.

The Ancestral Rio Grande in the Palomas basin is interpreted as a low-sinuosity, sandy bedload stream (Miall, 1985). A narrow flood plain and frequent avulsions resulted in a stratigraphic sequence composed almost entirely of multistory channel sand/sandstone bodies. Channel lag of gravel and/or organic debris commonly defines the base of each channel. Fine-grained overbank sediment was either not deposited in any significant thickness or was eroded by channels shortly after deposition. The presence of shale rip-up clasts in the sand/sandstone supports the latter interpretation.

The axial-fluvial facies is restricted to a narrow belt less than 5 km wide that is within 0.5 to 2 km from the footwall scarp of the Caballo Mountains. The axial-fluvial facies is overlain by about 25 m of coarse fanglomerate. Clast composition and imbrication indicate that the upper fanglomerate prograded westward from the footwall block.

Hatch-Rincon Basin

Basin Tectonics. The Hatch-Rincon basin is a west- to northwest-trending basin that merges into the southern end of the Palomas basin (Fig. 2). As much as 18 km wide, the basin narrows toward the southeast to 9 km, largely as a result of outlying fault blocks (Rincon Hills) of the southern Caballo Mountains extending into the northern flank of the basin. The basin is approximately 25 to 30 km long.

The basic structure of the Hatch-Rincon basin is a symmetrical graben, flanked on both northern and southern margins by normal faults and uplifted rift shoulders (Fig. 4). The northern flank is stair-stepped down toward the basin axis by two normal faults in most places. Symmetry of the graben is indicated by basinward dips of "early rift" basin fill exposed on both flanks of the graben, as well as by gravity profiles across the basin (Fig. 4; Keller and Cordell, 1983). Deposits of the Camp Rice Formation, although extensive across large parts of the basin, have also been eroded over broad areas by the Rio Grande and its tributaries. As in the Palomas basin, Camp Rice units are offset and otherwise deformed adjacent to graben-boundary faults, indicating movement on these faults in late Quaternary time.

Unlike the case in the Palomas basin, "early rift" basin fill is widely exposed beneath Camp Rice strata in the Hatch-Rincon basin. In fact, these "early rift" deposits are folded, block faulted, and generally uplifted relative to similar-age deposits in adjacent basins, and an angular unconformity separates them from overlying, near-horizontal Camp Rice strata. Clearly, the Hatch-Rincon basin is a late-rift (second stage) structure that is superimposed upon an uplifted and deformed segment of an "early rift" basin. The Camp Rice Formation, as well as less voluminous upper Pleistocene and Holocene fan and fluvial deposits, is the only basin-fill unit directly related to the modern Hatch-Rincon basin. The total exposed thickness of these units is approximately 100 m, whereas underlying "early rift" basin fill may be 1,500 m thick or more.

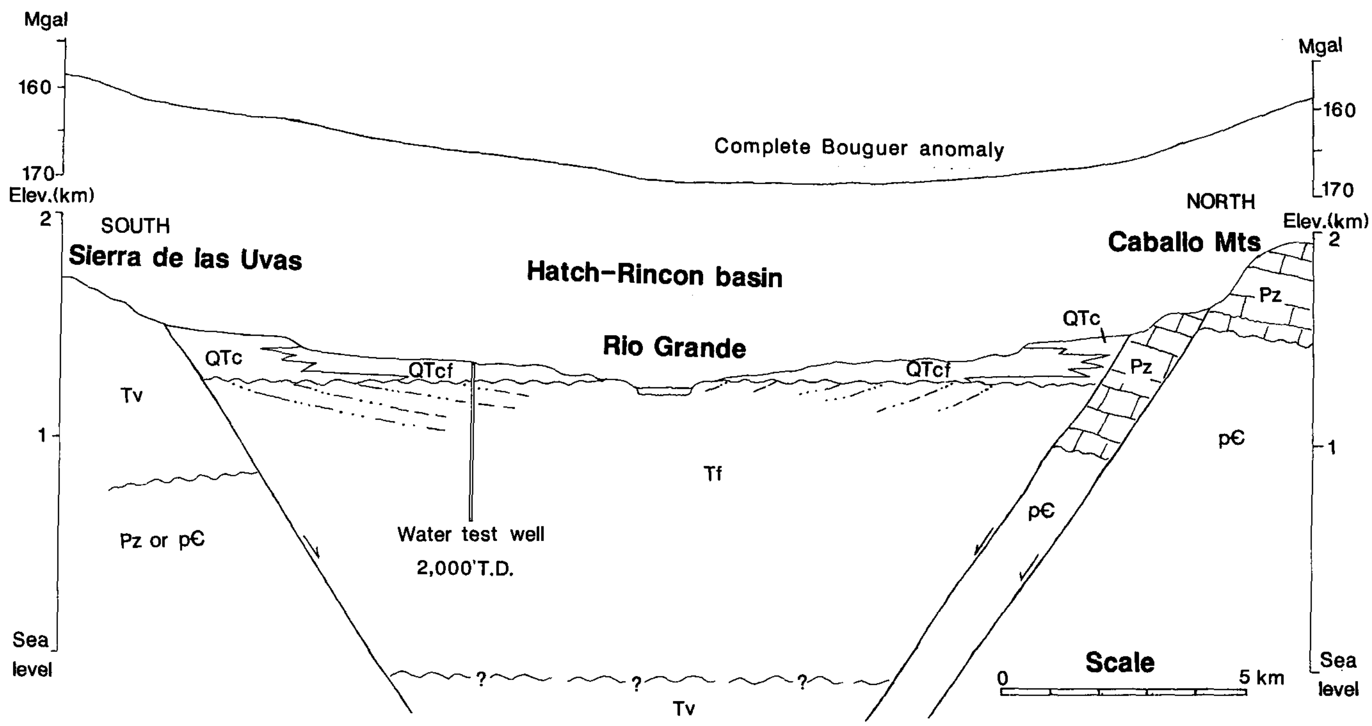

Figure 4. Cross section of the Hatch-Rincon basin. See Figure 2 for location of section line. Symbols the same as in Figure 3 except for the use of Camp Rice Formation (QTcf, QTc) instead of Palomas Formation.

Sedimentation. In the Hatch-Rincon basin, the piedmont facies of the Camp Rice Formation consists of coarse fanglomerate that extends only a few kilometers from the boundary faults of the Sierra de las Uvas, southern Caballo Mountains, and Rincon Hills. These rocks are similar in grain size and sedimentary structures to the piedmont facies deposited along the west flank of the Caballo Mountains in the Palomas basin. Clast composition in the piedmont facies reflects local derivation, a volcanic provenance for the piedmont facies derived from the Sierra de las Uvas and a mixed sedimentary and volcanic provenance adjacent to the Rincon Hills and southern Caballo Mountains.

The axial-fluvial facies in the Hatch-Rincon basin is quite different from the axial-fluvial facies in the Palomas and northern Mesilla basin in terms of rock types and their lateral distribution. The width of the belt of axial-fluvial facies in the Hatch-Rincon basin varies from 8 to 17 km, and axial-fluvial rocks are located within a few kilometers of the basin boundary faults on both the northern and southern sides of the basin (Seager and others, 1982). The increase in width of the belt of axial-fluvial facies is very abrupt and occurs between the Derry Hills and Hatch (Fig. 2). North of the Derry Hills, in the southern end of the Palomas basin, the axial-fluvial facies belt is about 5 km wide. As the axial-fluvial facies enters the Hatch-Rincon basin, a few kilometers south of the Derry Hills (Fig. 2), the belt of axial-fluvial strata increases to about 17 km wide. The presence of axial-fluvial rocks north of the Rincon Hills also indicates that the Ancestral Rio Grande was periodically diverted between the Rincon Hills and southern Caballo Mountains (Seager and Hawley, 1973). As in the Palomas basin, piedmont facies as much as 20 m thick prograded from both the northern and southern margins over the axial-fluvial facies.

The axial-fluvial facies in the Hatch-Rincon basin consists of interbedded channel sandstone or conglomeratic sandstone and overbank mudstone and sandstone. Channel bodies are classified as multistory sheets (Friend and others, 1979), range in thickness from 3 to 18 m, and can be traced laterally for hundreds of meters where the outcrop allows, such as in the badlands west of Hatch or east and south of the Rincon Hills. Clast composition, grain size distribution, and sedimentary structures in the channel bodies are similar to those of the axial-fluvial facies in the Palomas basin. Markedly different from the Palomas basin, however, is the abundance in the Hatch-Rincon basin of overbank sediment, which occupies approximately 40% of the stratigraphic section exposed in the badlands south and east of the Rincon Hills. The overbank sediment is composed of red mudstone and fine to very fine sand or sandstone that are interbedded on a scale of 0.3 to 3.0 m. Paleosols are common in the overbank sediment and consist of stage II caliche (Bk horizon) overlain by reddish-brown, argillic horizon (Bt). In a 53-m-thick section east of the Rincon Hills, 6 individual paleosols were recognized. The relative abundance and thickness of individual beds of overbank facies indicate that flood plains in the Hatch-Rincon basin were stable for long periods of time before they were invaded by fluvial channels. The well-developed paleosols further suggest that flood-plain sedimentation was intermittent, allowing pedogenesis to proceed for thousands or tens of thousands of years at a time (Birkeland, 1974; Gile and others, 1981). These characteristics reflect the increased width of the alluvial plain in the Hatch-Rincon basin.

Northern Mesilla Basin

Basin Tectonics. The Mesilla basin, located in extreme south-central New Mexico near Las Cruces, is a major north-trending basin of the southern Rio Grande rift. Although its southern part is a broad, deep graben, the basin narrows northward, and faulting along the eastern basin margin dies out. Consequently, the northernmost part of the basin is a westward-tilted half-graben in which the hanging-wall dip slope and uplift

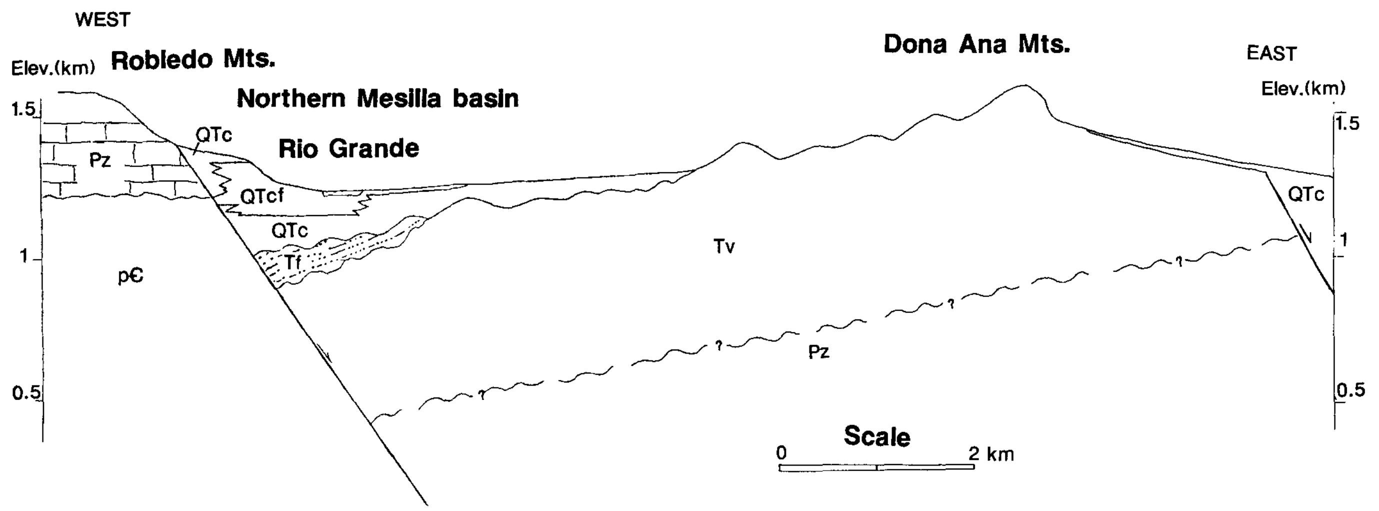

Figure 5. Cross section of the northern Mesilla basin. See Figure 2 for location of section line. Symbols the same as in Figure 3 except for the use of Camp Rice Formation (QTcf, QTc) instead of Palomas Formation.

are composed of Tertiary volcanic rocks of the Dona Ana Mountains and in which the footwall uplift is the Robledo Mountains (Figs. 2 and 5).

The Camp Rice Formation constitutes the bulk of the basin fill in the northern Mesilla basin, but it probably is nowhere thicker than 100 to 150 m, and much of that has been removed by the Rio Grande. Consequently, the gravity anomaly associated with this shallow rift structure is weak. Nevertheless, the asymmetry of the half-graben is well documented by exposures of Tertiary volcanic rocks that floor the structure (Seager and others, 1976, 1982).

The western margin of the half-graben is formed by the East Robledo fault, which exhibits approximately 0.7 km of stratigraphic separation in the vicinity of the cross section (Fig. 5). The fault cuts Camp Rice strata, indicating movement in late Quaternary time. Syndepositional movement with Camp Rice units may be inferred by much greater thicknesses of Camp Rice strata along the downthrown side of the East Robledo fault, compared to equivalent strata on the upthrown side, and by the very coarse-grained tongues of piedmont alluvium that comprise proximal fan facies near the fault.

Sedimentation. Outcrops of the Camp Rice Formation in the northern Mesilla basin are restricted to a narrow belt directly adjacent to the footwall block (Robledo Mountains) and to scattered patches along the flank of the hanging-wall block (Dona Ana Mountains). The bulk of the Camp Rice Formation has been removed from the basin since middle Pleistocene time. Despite the paucity of outcrops, the original distribution of piedmont and axial-fluvial facies can be accurately reconstructed.

Coarse piedmont fanglomerate extends less than half a kilometer away from the footwall scarp of the Robledo Mountains before it interfingers with and is replaced by axial-fluvial facies. The axial-fluvial facies consists almost exclusively of cross-bedded or horizontally laminated pebbly sand or sandstone and is nearly identical to the axial-fluvial facies in the Palomas basin. A tongue of piedmont fanglomerate about 20 to 30 m thick overlies the axial-fluvial facies exposed along the eastern flank of the Robledo Mountains.

Thin relicts of piedmont fanglomerate are also exposed in a belt about 4 km wide on the hanging-wall dip slope along the west flank of the Dona Ana Mountains. The fanglomerate fines rapidly westward and is composed of sedimentary and volcanic clasts that were derived from the Dona Ana Mountains. Also present are thin tongues of fluvial sand and gravel (Seager and others, 1976).

DISCUSSION

A comparison of Camp Rice and Palomas strata in the Palomas, Hatch-Rincon, and northern Mesilla basins suggests a strong tectonic influence on the distribution of piedmont and axial-fluvial facies. In the asymmetrical Palomas and northern Mesilla basins, the axial-fluvial facies is located near the locus of maximum subsidence and within a few kilometers of the footwall scarp (Figs. 3 and 5). Coarse fanglomerate derived from the footwall uplift is restricted to a narrow zone directly adjacent to the footwall scarp, whereas alluvial-fan and alluvial-flat detritus derived from the hanging-wall uplift and deposited on the hanging-wall dip slope occupies a facies belt many times more extensive than the piedmont facies belt on the opposite side of the basin.

The facies relationships in the Palomas and northern Mesilla basins are identical to those shown schematically in Figure 1. According to the models of Leeder and Gawthorpe (1987) and Blair and Bilodeau (1988), the facies distribution largely reflects changes in depositional slope produced by faulting. Periods of fault movement in a half-graben assures that the topographically lowest point of the basin is maintained directly above the zone of maximum subsidence. Consequently, as fluvial channels avulse, seeking the lowest part of the alluvial plain, a sequence of multistory channel sands is constructed above the zone of maximum subsidence (Fig. 1; Bridge and Leeder, 1979; Leeder and Gawthorpe, 1987; Leeder and Alexander, 1987; Alexander and Leeder, 1987). At the same time, fault motion tends to decrease the slope of alluvial fans draining the footwall uplift, causing new fan segments to develop near the fan apex and preventing the fans from prograding too far into the basin (Hooke, 1972; Leeder and Gawthorpe, 1987). In contrast, the hanging-wall dip slope increases during fault movement, which causes channel incision and displacement of coarse detritus basinward (Leeder and Gawthorpe, 1987).

The distribution of facies in a half-graben is affected not only by asymmetrical subsidence but also by the relative size of the drainage basins of the footwall and hanging-wall uplifts. In the Palomas basin, the hanging-wall uplift is much larger than the footwall uplift (Seager and others, 1982). It might be argued that the laterally extensive clastic wedge on the hanging-wall dip slope of the Palomas basin, as well as the narrow axial-fluvial belt near the footwall scarp, is as much a function of the magnitude of sediment derived from the hanging-wall uplift as of asymmetrical subsidence. In effect, the large amount of detritus shed off the hanging-wall block may have prevented the alluvial plain from extending westward. Although this argument may apply to the Palomas basin, it does not apply to the northern Mesilla basin. The size of the drainage basins facing the northern Mesilla basin is roughly equal on the footwall (Robledo Mountains) and hanging wall (Dona Ana Mountains) (Seager and others, 1987), and yet the axial-fluvial facies is concentrated near the footwall scarp, as it is in the Palomas basin. This relationship suggests that asymmetrical subsidence, and not the volume of detritus dispersed across the hanging-wall dip slope, is responsible for the location of the axial-fluvial facies.

The distribution of facies in the nearly symmetrical Hatch-Rincon basin is markedly different from that in the Palomas and northern Mesilla basins. In the absence of asymmetrical subsidence, the axial-fluvial system was free to occupy almost the entire basin (Fig. 4). As the fluvial channels moved across the wide basin, however, large areas of the basin experienced prolonged periods of overbank sedimentation and subsequent pedogenesis.

In each of the three basins, fanglomerate prograded over the axial-fluvial facies, producing a coarsening-upward sequence. One interpretation of this coarsening-upward sequence is that it represents a period of uplift in the adjacent source terranes. In the models of Leeder and Gawthorpe (1987), Blair and Bilodeau (1988), and Heller and others (1988), however, the upper conglomerate may well be postorogenic, representing a period when the rate of basin subsidence was less than the rate of erosion. In light of the data suggesting that tectonism in the southern Rio Grande rift has decreased in intensity since about 7 Ma, a postorogenic interpretation for the upper conglomerate is favored.

The emphasis in this study is on tectonic control of facies distribution; however, other allocyclic variables, such as eustacy and paleoclimate, may have also been important in determining facies characteristics of the Camp Rice and Palomas Formations and need to be evaluated. Eustatic sea-level changes probably did not affect drainage in the southern Rio Grande rift because the base level for the Ancestral Rio Grande was a system of lakes in west Texas and northern Chihuahua, Mexico, rather than the sea (Kottlowski, 1953, 1960; Ruhe, 1962; Reeves, 1965; Strain, 1966; Hawley and others, 1969). In contrast, paleoclimate probably did have a significant influence on sedimentation in the southern Rio Grande rift, both in terms of the nature of depositional processes and in terms of fluctuating lake levels.

Vertebrate fauna in the Palomas Formation indicate that the southern Rio Grande rift was dominated by open grasslands and that the paleoclimate was frost free (Lucas and Oakes, 1986). These conditions correlate with modern semi-arid prairies or steppes (Strahler and Strahler, 1983). The influence of a relatively dry paleoclimate is evident in the piedmont facies, given the presence of sheet-flood and debris-flow deposits (Bull, 1972; Mack and Rasmussen, 1984). Red, highly oxidized mudstones and calcareous paleosols are also consistent with a semi-arid paleoclimate (Reeves, 1970; Retallack, 1983). Furthermore, a semi-arid paleoclimate, along with coarse bedload, was probably responsible for the broad, low-sinuosity channels of the axial-fluvial facies (Schumm, 1960, 1961; Schumm and Lichty, 1963; Baker and Penteado-Orellana, 1977).

It is also possible that changes in paleoclimate, commensurate with glacial and interglacial periods, resulted in fluctuations in the level of the lakes in west Texas and northern Chihuahua, Mexico. Changing lake levels, assuming they were of sufficient magnitude, would have promoted alternating incision and aggradation by the Ancestral Rio Grande and its tributaries. Periods of channel incision may have been important in the production of paleosols on the relict, higher-level surfaces. There is, however, no quantitative documentation of fluctuations in the level of the lakes in west Texas and northern Mexico, and the effects of this process on deposition of the Camp Rice and Palomas Formations cannot be accurately evaluated.

Finally, it is important to evaluate the possibility that basinward progradation of the upper conglomerate of the Camp Rice and Palomas Formations was climatically, and not tectonically, controlled. The relationship between sediment yield and climate, particularly annual precipitation, is well established (Langbein and Schumm, 1958; Douglas, 1967; Wilson, 1973; Jansen and Painter, 1974; Dendy and Bolton, 1976); however, paleoclimatic changes during Pliocene–middle Pleistocene time in south-central New Mexico have not been well documented. Consequently, paleoclimatic fluctuations for this time period must be deduced from oxygen isotopic ratios of deep-sea marine sediment (for example, Imbrie and others, 1984). Although the isotopic data provide a clear record of glacial and interglacial cycles, the relationship between glacial cycles and precipitation in the southwestern United States is equivocal and controversial (compare with Brakenridge, 1978, and Van Devender and Spaulding, 1979). Thus, at the present time, the data are inadequate to assess the role of paleoclimatic changes on sediment yield during Camp Rice and Palomas deposition. Despite the lack of paleoclimatic data for Pliocene–middle Pleistocene time, several studies of late Pleistocene–Holocene geomorphic relationships may provide a useful analogue to Camp Rice and Palomas sedimentation. These studies indicate that short-term climatic fluctuations result in a complex interplay between deposition, entrenchment, and pedogenesis (Gile and others, 1981; Wells and others, 1987). The stratigraphic record of such complexity is characterized by extreme heterogeneity. Such complexity does not appear to be present in the upper conglomerate of the Camp Rice and Palomas Formations, suggesting that another variable, such as tectonism, was the driving force behind progradation of the upper conglomerate.

CONCLUSIONS

The distribution of piedmont and axial-fluvial facies of the Camp Rice and Palomas Formations (Pliocene-Pleistocene) in the southern Rio Grande rift is controlled primarily by basin symmetry. In the Palomas and northern Mesilla half-grabens, the axial-fluvial facies is located in a narrow belt near the locus of maximum subsidence and is composed predominantly of multistory channel sands and sandstones. The axial-fluival facies is bordered on the footwall side by piedmont fanglomerate that extends only a few kilometers from the footwall scarp, and on the hanging-wall side by a laterally more extensive wedge of piedmont conglomerate, sand and sandstone, mudstone, and calcareous paleosols that was deposited on the hanging-wall dip slope. In the nearly symmetrical Hatch-Rincon graben, the axial-fluvial facies occupies almost the entire basin, except for narrow zones of piedmont fanglomerate. A greater percentage of overbank fine sand and mudstone and calcareous paleosols distinguishes the axial-fluvial facies in the Hatch-Rincon basin from that in the asymmetrical basins. In all three basins, the axial-fluvial facies is overlain by a tongue of piedmont fanglomerate, which results in a coarsening-upward sequence.

Facies distribution of the Camp Rice and Palomas Formations supports a two-stage model of deposition in asymmetrical basins proposed by Leeder and Gawthorpe (1987), Blair and Bilodeau (1988), and Heller and others (1988). During periods of active subsidence, the location of axial-fluvial deposition is maintained near the locus of maximum subsidence by the process of channel avulsion into the topographically lowest area of the basin. As tectonism wanes and erosion rates surpasses subsidence rate, the piedmont facies progrades toward the basin center and over previously deposited axial-fluvial sediment.

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