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Calcic paleosols of the Plio–Pleistocene Camp Rice and Palomas Formations, southern Rio Grande rift, USA

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ABSTRACT

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Seventy-four separate paleosols are recognized in the Plio–Pleistocene Camp Rice and Palomas Formations of the southern Rio Grande rift. Seven types of profiles are present: A–Bw–Bk–C, A–Bt–Bk–C, Bw–Bk–C, Bt–Bk–C, Bt–K–C, Bk–C, and Ak. Calcic and petrocalcic horizons (Bk and K) commonly overlain by argillic horizons (Bt) dominate most paleosol profiles. Greater pore size and enhanced permeability, typical of sandy and silty parent material, favored the development of argillic horizons. However, micrite is the dominant form of calcium carbonate in Bk and K horizons regardless of parent material grain size. Non-calcified A horizons are typically lighter in color than adjacent subsurface pedogenic intervals due to their oxidized nature or lower clay content. They are only rarely preserved, probably due to erosion prior to final burial. Calcified root mats (Ak) are not directly associated with other pedogenic horizons. Abundant horizontal roots suggest root mats formed under high water table conditions. Restriction of Ak horizons to the distal piedmont lithofacies supports their development as spring or pond deposits near the toes of alluvial fans.

Individual pedogenic horizons can often be traced hundreds of meters to kilometers normal and parallel to a basin's axis. Paleosol maturity and preservation potential are greatest for symmetrical basins, especially where pedogenesis was far removed from active channels within the axial fluvial system or within distal piedmont settings of asymmetrical basins.

Introduction

Authigenic carbonate in the form of nodules or laterally continuous beds is a common feature in nonmarine siliciclastic strata. These calcareous horizons are commonly considered to have developed as the subsurface calcic or petrocalcic horizon of a soil (e.g. Steel, 1974; Hubert, 1978; Hay and Reeder, 1978; McPherson, 1979; Wright, 1982; Retallack, 1983; Monger et al., 1991). However, carbonate can also precipitate in sediment in the zone of capillary rise just above the water table, or as a result of diagenetic processes well below the water table (Semeniuk and Meagher, 1981; Tandon and Narayan, 1981; Arakel and McConchie, 1982). It is not always easy to distinguish between pedogenic and non-pedogenic carbonate, particularly if the observations and analyses are restricted to the carbonate horizon. A pedogenic origin for the carbonate is less equivocal if it can be placed in the context of a soil profile. This involves not only examining the strata above and below the carbonate bed, but also recognizing down-profile changes in texture, structure, and composition that are commensurate with pedogenic processes. This approach, unfortunately, is often overlooked in studies of authigenic carbonate.

The Plio–Pleistocene Camp Rice and Palomas Formations in the southern Rio Grande rift of south-central New Mexico contain numerous horizons of authigenic carbonate. A pedogenic

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origin for a majority of the carbonate intervals can be established by one or more of the following criteria: (1) presence of distinct associated pedogenic horizons, (2) calcareous roots and/or clay-filled root traces, (3) blocky and prismatic peds, (4) illuvial clay, and (5) slickensides. Good exposures and shallow burial ($\leqslant$ 150 m) provide an excellent opportunity to document pedogenic features that are present in paleosols but are little modified from their original state. The goals of this study are to describe the field and petrographic characteristics of these paleosols, to document lateral variations in the paleosols, and to assess the role of sediment texture and depositional environment on paleosol development.

Stratigraphy

The Camp Rice and Palomas Formations were deposited during the most recent stage of extension in the southern Rio Grande rift (Kottlowski, 1953; Seager, 1975; Seager et al., 1984). The Camp Rice Formation was originally defined by Strain (1966) in west Texas and was subsequently used as a map unit in the area between Las Cruces and Hatch, New Mexico (Fig. 1; Seager et al., 1971, 1976, 1982, 1987; Seager and Hawley, 1973; Seager and Clemons, 1975). The term "Palomas" had been informally used since the early part of this century to describe late-rift, basin-fill sediment exposed between Truth or Consequences and Hatch, but was only recently designated a formation by Lozinsky and Hawley (1986). The age of the Camp Rice and Palomas Formations is bracketed between 4.5 and 0.7 Ma (Early Pliocene–Middle Pleistocene) by a combination of radiometric dates of underlying, overlying, and interbedded volcanic rocks (Reynolds and Larsen, 1972; Bachman and Mehnert, 1978; Hawley, 1981; Kortemeir, 1982; Seager et al., 1984; Hawley and Machette, 1987), by the presence of a Blancan vertebrate fauna (Strain, 1966; Tedford, 1981; Lucas and Oakes, 1986; Repenning and May, 1986), and by magnetostratigraphy (Mack et al., 1991).

The Camp Rice and Palomas Formations have a maximum thickness of about 150 m and consist of piedmont and axial-fluvial lithofacies (Fig. 1).

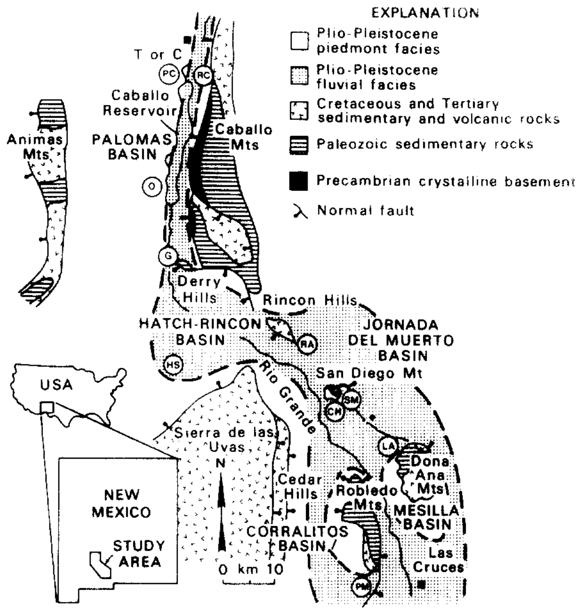


Fig. 1. Index and general geologic map of the study area, southern Rio Grande rift. Palomas Creek (*PC*), Red Canyon (*RC*), Oasis (*O*), Garfield (*G*), Hatch Siphon (*HS*), Rincon Arroyo (*RA*), San Diego Mountain (*SM*), Cedar Hill (*CH*), Lucero Arroyo (*LA*), and Picacho Mountain (*PM*).

The piedmont lithofacies is composed of conglomerate, sand and sandstone, and mudstone that were derived from local uplifts and deposited on alluvial fans and alluvial flats (Mack and Seager, 1990). Proximal piedmont strata, which were deposited within a kilometer or less of the complementary uplift, are composed primarily of conglomerate and a minor amount of sand/sandstone and gravelly, sandy mudstone. Proximal piedmont deposits are especially well exposed along the eastern margin of the Palomas basin, where they interfinger with fluvial sand/sandstone (Fig. 1). Distal piedmont strata, which were deposited kilometers to tens of kilometers from the mountain front, consist of subequal amounts of conglomerate, sand/sandstone, and mudstone (Fig. 1). The axial-fluvial lithofacies was deposited by the Ancestral Rio Grande, which flowed southward and emptied into lakes in west Texas and northern Chihuahua, Mexico (Kottlowski, 1953, 1960; Reeves, 1965, 1969; Strain, 1966; Hawley et al., 1969). Fluvial channel deposits are composed of crossbedded pebbly medium to coarse sand/sandstone, whereas floodplain detritus consists of mudstone and silt or very fine to fine sand (Mack and Seager, 1990; Mack and James, 1992).

Camp Rice and Palomas deposition ended about 0.7 Ma when the Rio Grande began to entrench its basins, as a result of capture by the lower Rio Grande, tectonic uplift, climatic change, or a combination of these features (Hawley et al., 1969; Hawley, 1981). The Rio Grande and its tributaries locally have exposed the entire Camp Rice Formation and the upper 50 m of the Palomas Formation. Thick calcic paleosols, which developed on the constructional top of the formation following entrenchment, are shown in Fig. 2 but are not considered in this study.

Methods

Seventy-four paleosols were examined at ten stratigraphic sections (Fig. 2). Five sections consist entirely of fluvial strata, two contain interbedded proximal piedmont and fluvial strata, two are composed entirely of distal piedmont lithofacies, and one has interbedded distal piedmont and

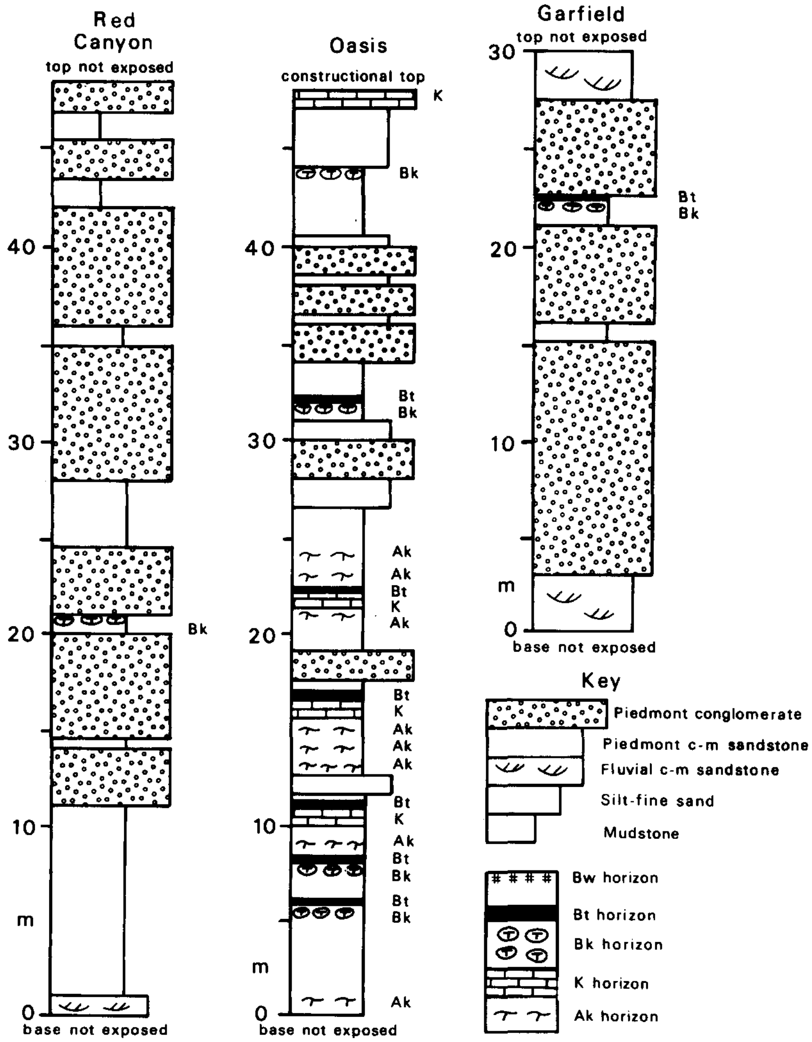


Fig. 2. Measured sections of the Camp Rice and Palomas Formations showing vertical distribution of lithofacies and position and type of paleosols. Sections at Hatch Siphon, Cedar Hill, Rincon Arroyo, Lucero Arroyo, and Picacho Mountain are entirely fluvial, whereas sections at Oasis and Palomos Creek are entirely distal piedmont. Red Canyon and Garfield are mixed proximal piedmont and fluvial, and San Diego Mountain is mixed distal piedmont and fluvial. Because of the scale, C and noncalcified A horizons are not shown.

fluvial strata. Each of the 74 paleosols was described in the field, following standard guidelines and nomenclature used for modern soils (Buol et al., 1980, pp. 21–43; Retallack, 1990, pp. 20–54). Each paleosol was divided into horizons, which were described in terms of thickness, nature of contacts, grain size of parent sediment, color (Munsell color when moist), and structure (root and burrow type and abundance, peds, nodules, and tubules). Thirty-eight paleosols, representative of the range of paleosol types, were sampled for thin- section analysis; one or more samples from each horizon within individual paleosols were collected. A total of 114 thin sections was examined for textural features described by Brewer (1964). Twenty thin sections of argillic B horizons were point-counted, utilizing 300 points per sample and recognizing nine categories: detrital quartz, detrital feldspar, volcanic rock fragments, sedimentary rock fragments, accessory minerals (e.g. biotite, amphibole, opaques), plasma, argillans, authigenic carbonate, and pore space. In addition, point counts were made of eleven carbonate nodules (Bk horizon) or laterally continuous beds (K horizon) and nine calcified root mats (Ak horizon), recognizing micrite, detrital grains, empty or spar-filled vughs, and vugh- and skew-plane argillans. Twenty-eight

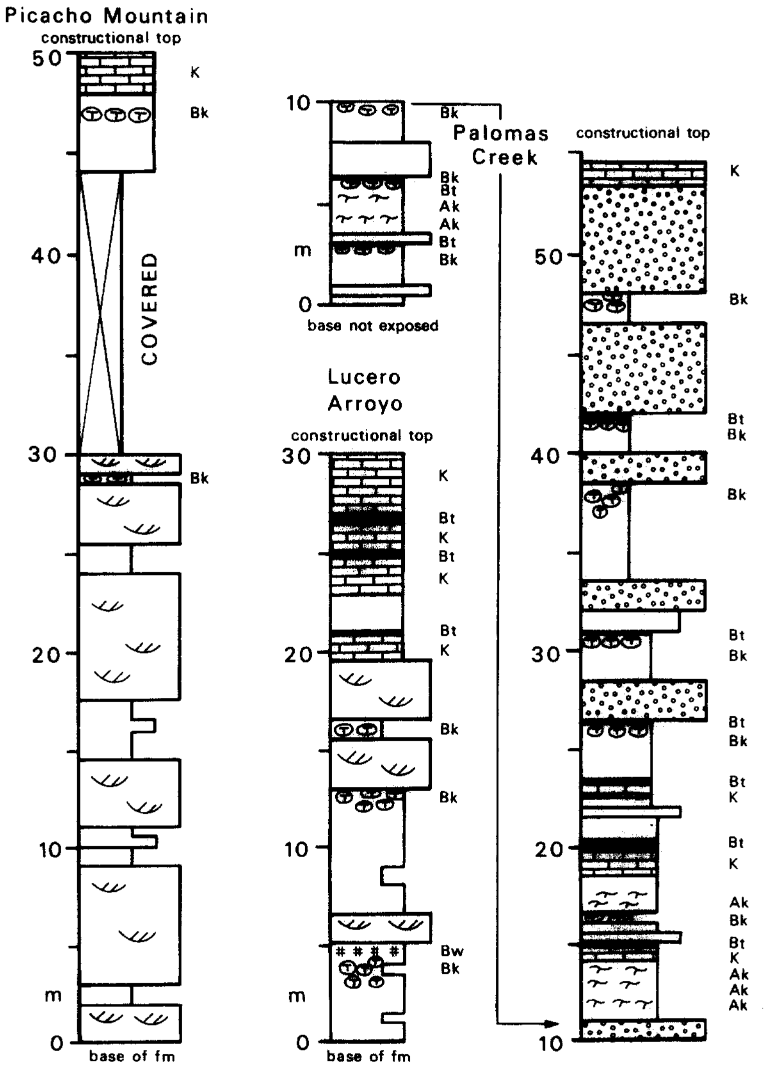


Fig. 2 (continued).

samples were analyzed by scanning electron microscope and EDS for mineral morphology, temporal relations among constituents, and composition.

Standard X-ray diffraction analysis of clay minerals was performed on five samples: three Bk horizons, one Bt horizon, and one Bw horizon. Four analyses per sample were run: air-dried, glycolated, heated to 400°C, and heated to 550°C (Starkey et al., 1984). Clay–mineral cation composition was also estimated for the same five samples by energy dispersive X-ray.

Description of paleosols

Calcic paleosols are present in fluvial and piedmont lithofacies of the Camp Rice and Palomas Formations (Fig. 2). The majority of paleosols in the fluvial lithofacies are in floodplain sediment, including fine sand or silt and mudstone. Rarely, paleosols developed on the top of channel sand bodies. Fine sand, silt, and mud are the parent sediment for the majority of paleosols in the piedmont lithofacies as well. Six types of paleosol horizons exist singly or in combination to

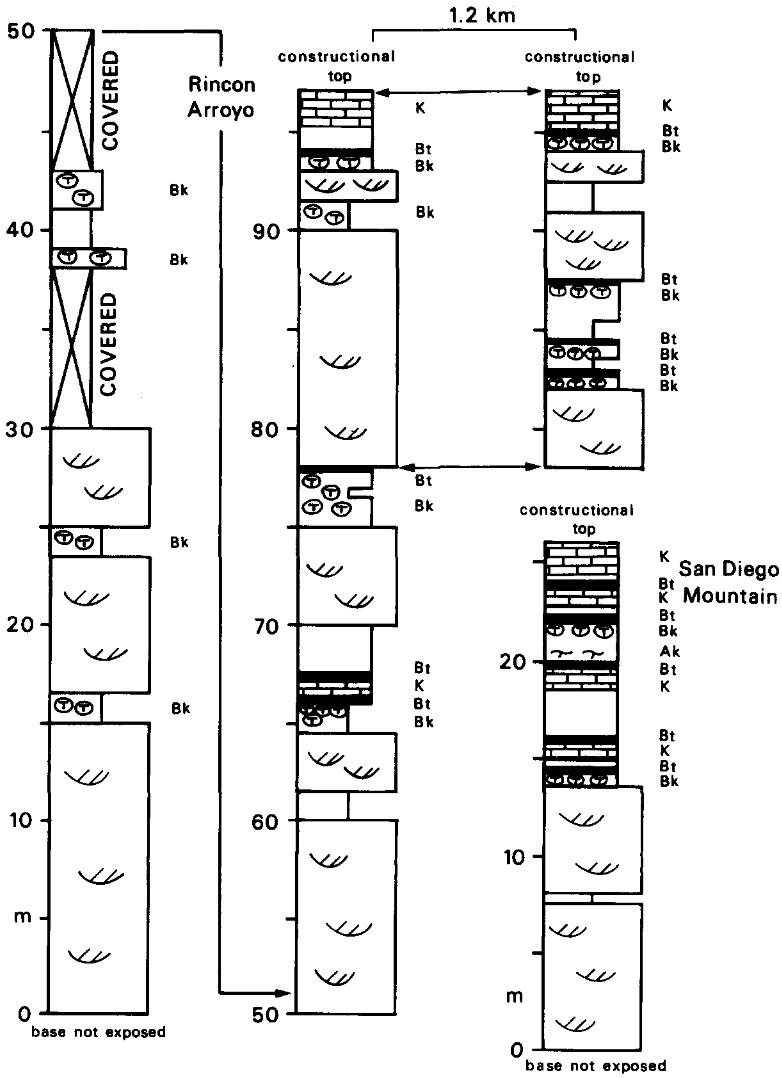
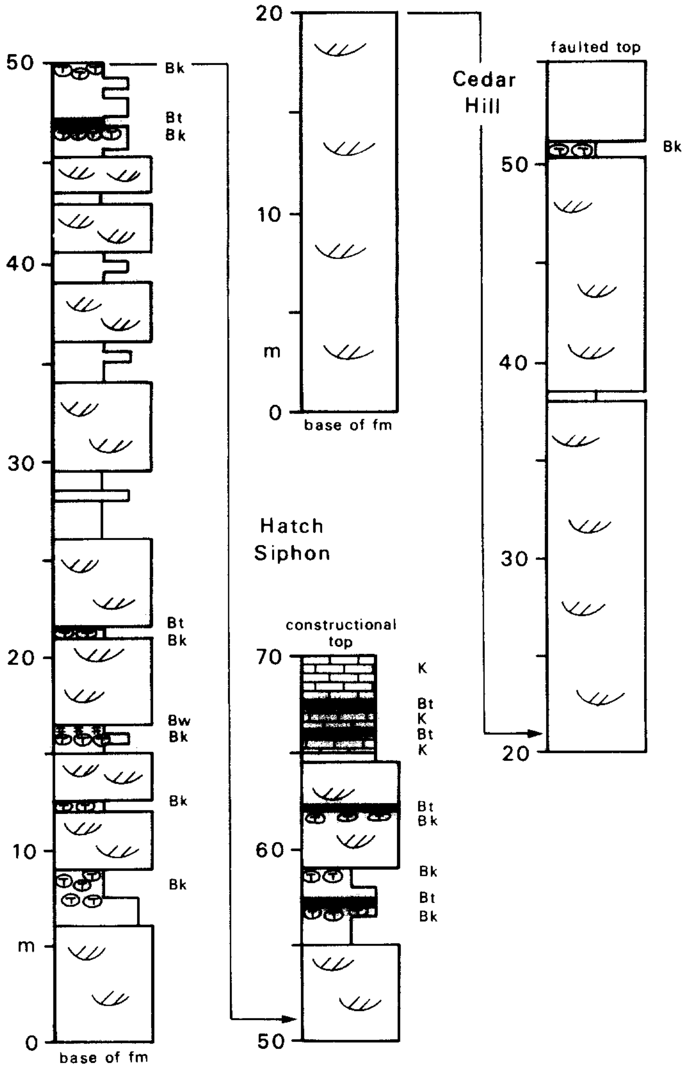


Fig. 2 (continued).

form seven types of paleosol profiles: A–Bw–Bk–C, A–Bt–Bk–C, Bw–Bk–C, Bt–Bk–C, Bt–K–C, Bk–C, and Ak. Representative paleosol profiles are illustrated in Figs. 3 and 4 and the salient characteristics of the horizons are described below.

A horizons

Non-calcified A horizons are present in only 4% of the paleosols examined in this study. A horizons are generally thin ($\leqslant$ 10 cm) and can be distinguished from underlying Bw or Bt horizons by lighter color (light gray, 5YR 6/1) and/or by a greater degree of pedoturbation, which locally resulted in incorporation of darker fragments from the underlying horizon into the base of the A horizon (Fig. 3). The light color probably reflects a lower clay content than the underlying horizon, as well as post-depositional oxidation of the original organic matter. The paucity of non-calcified A horizons is a common feature of calcic paleosols and is probably due to erosion prior to burial (cf. Blodgett, 1988).



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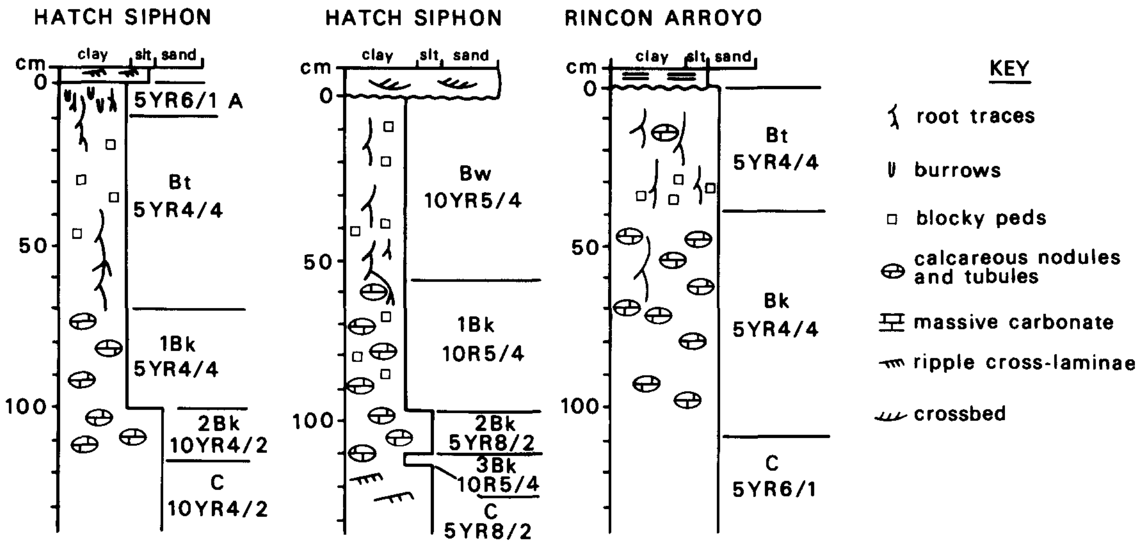


Fig. 3. Columnar sections of typical calcic paleosols showing A, Bt, Bk, Bw and C horizons.

Argillic B horizons (Bt)

Argillic B horizons (Bt) are present in 47% of the paleosols examined in this study. Bt horizons range in thickness from 8 to 100 cm, with an average thickness of about 40 cm, and either gradationally underlie A horizons, or, more commonly have an erosional upper contact (Figs. 3, 4, 5A). Bt horizons are always underlain by either Bk or K horizons and although the contact is

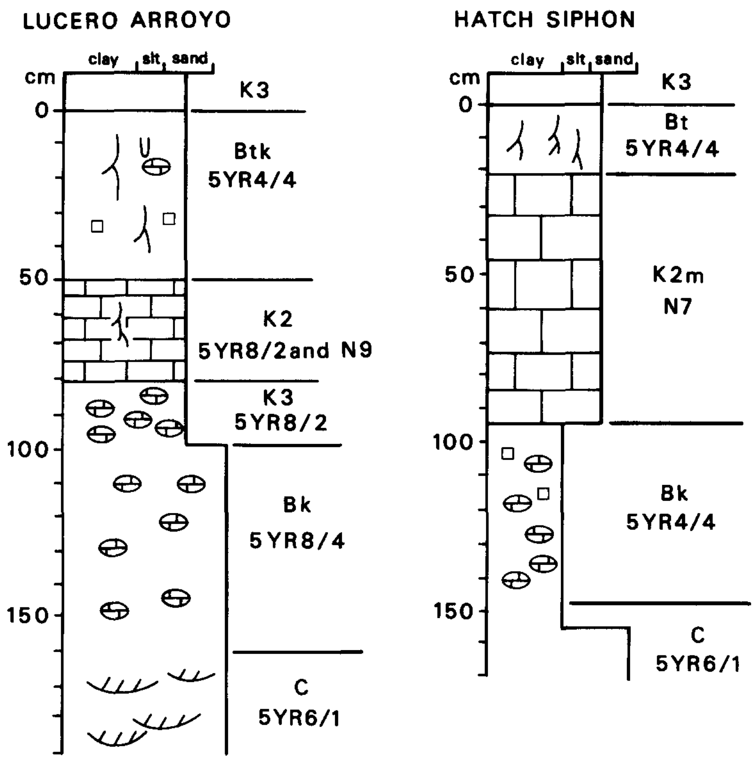


Fig. 4. Columnar sections of typical calcic paleosols showing Btk, Bt, Bk, K, and C horizons. See Fig. 3 for explanation.

gradational, it usually can be accurately placed within a few centimeters. Argillic B horizons most commonly developed in sand or silt parent sediment, although Bt horizons in mudstone parents are present as well (Figs. 3, 4).

Argillic B horizons developed in sandy parent materials are easily recognizable because of their reddish brown (5YR 4/4) or red (10R 4/6) color, which is in sharp contrast to the light reddish brown (5YR 6/4), light greenish gray (5GY 8/1), or light gray (5YR 6/1) colors of pedogenically unmodified beds of sand and silt (Figs. 3, 4). In contrast, Bt horizons developed in mudstone display the same range of colors as pedogenically unmodified mudstone, including reddish brown (5YR 4/4), weak red (10R 5/4), and light red (10R 6/6). In a few cases, the argillic B horizon is mottled brown and green, indicating gley conditions and warranting designation as a Btg horizon.

Root traces filled with sediment of different grain size or color than the parent sediment are common in Bt horizons, and vary in size from a few millimeters wide and less than a centimeter long to tens of millimeters wide by 5 to 15 cm long. In many cases, calcareous rhizoliths are present within the argillic B horizon, resulting in designation as a Btk horizon. Peds are rare in sandy Bt horizons, although a few examples of blocky and prismatic types exist. Bt horizons in mudstone, however, commonly display angular blocky peds 1 to 3 cm in diameter and slickensides.

The principal evidence for translocated clay in the Bt horizons is the presence of illuvial argillans. Although best seen in thin section, argillans produce a waxy luster to hand samples, particularly in sandy parent sediment. Argillans are much more difficult to recognize in hand samples of mud-rich parent sediment and in most cases thin

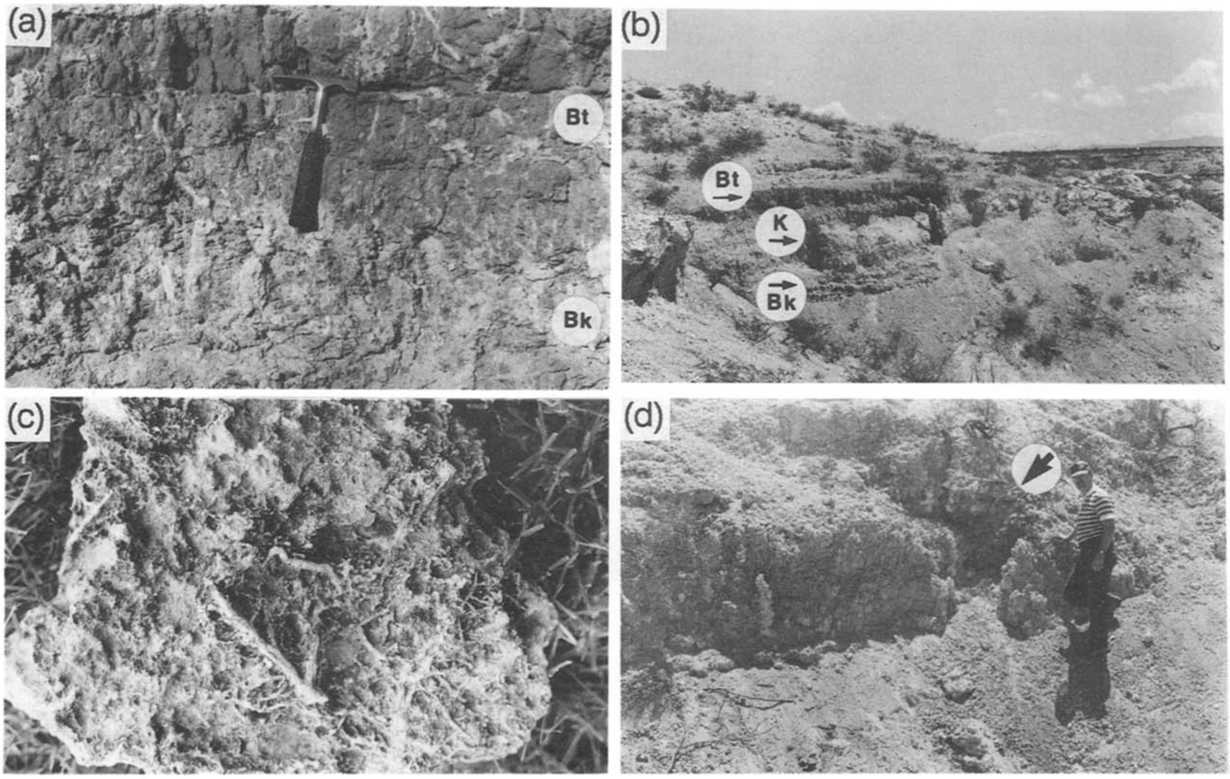


Fig. 5. Outcrop expression of paleosols. A. Dark argillic B horizon (Bt) overlying lighter colored calcic B (Bk) horizon. Calcite tubules especially evident in argillic B horizon. Hammer is 25 cm long. B. Dark argillic B horizon (Bt) overlying K horizon. Man is pointing at the Bt–K contact and standing at the K–Bk contact. Man is 1.75 m tall. C. Bedding plane view of calcified root mat. Field of view is 30 cm wide. D. Pipe structure (arrow) in which dark Bt horizon extends downward into lighter colored Bk horizon. Man is 1.75 m tall.

sections are necessary to demonstrate the presence or absence of argillans. In sandy parent sediment, argillans cover grains that are not separated by plasma (free-grain argillans), as well as grains that are surrounded by plasma (embedded-grain argillans) (Figs. 6A, 6B). The argillans are a few tens of micrometers thick and the basal plane of the clay mineral is aligned parallel to sub-parallel to the substrate (Fig. 7A). Free-grain argillans are more commonly associated with Bt–Bk profiles, whereas embedded-grain argillans are more common in Bt horizons that overlie K horizons. There are several examples of a positive down-profile correlation between the percentage of argillans and the percentage of plasma and a negative correlation between the percentage of argillans and the amount of pore space, supporting the interpretation that the argillans represent translocated clay (Table 1; Gile et al., 1981). Much less common in sandy Bt horizons are ped and skew-plane argillans. Embedded -grain argillans are common in Bt horizons developed in mudstone, as are vugh, skew-plane, and ped argillans (Fig. 6C). Stress argillans, characterized by moderately aligned zones of clay with diffuse boundaries, are common in Bt horizons developed in mud-rich parent sediment and probably resulted from shrinking and swelling of the smectite clay-rich plasma.

The dominant clay minerals in both the soil plasma and as argillans are smectite and mixed-layer smectite–illite; kaolinite is rare. Hematite is also present and probably formed by dehydration of a ferric hydroxide precursor following burial (Van Houten, 1972), or developed by alteration of iron-bearing detrital grains during burial (Walker, 1967). The presence of hematite is responsible for the fact that Camp Rice and Palomas paleosols are redder in color than their Quaternary counterparts in the same region (cf. Gile et al., 1981).

Structural B horizons (Bw)

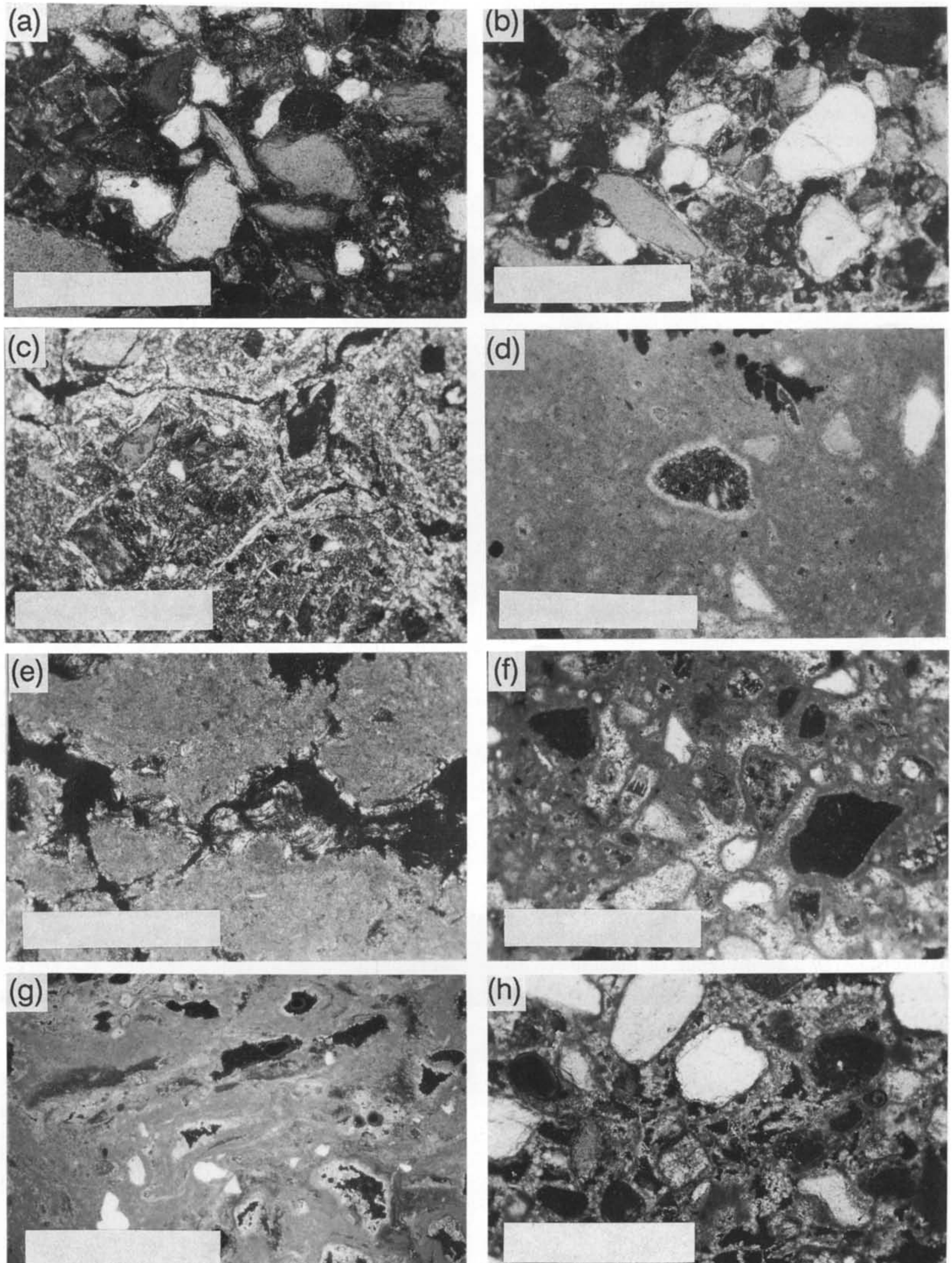
Present in 3% of the paleosols examined in this study, structural B horizons gradationally underlie A horizons or have sharp, erosional upper contacts and gradationally overlie Bk horizons (Fig. 3). Bw horizons range in thickness from 30 to 90 cm and are exclusively developed in mudstone parent sediment. Angular blocky peds and slickensides are common, but free-grain and embedded-grain argillans are rare or absent. Color and root trace abundance and morphology are similar to those of argillic B horizons.

Calcic B horizons (Bk)

The most common type of paleosol horizon in the Camp Rice and Palomas Formations is the calcic B horizon (Bk), which is present in 55% of the paleosols examined in this study. Bk horizons range in thickness from 20 to 140 cm, averaging 70 cm. Bk horizons are equally as well developed in sand and mud parent sediment, and there are several examples of the Bk horizons crosscutting sand–mud depositional contacts (Fig. 3). Bk horizons gradationally underlie Bt and Bw horizons and may be present above and below K horizons (Figs. 4, 5A, 5B). In 28% of the paleosols, the Bk horizon is the uppermost horizon, as a result of truncation of overlying horizons by channel sand or gravel bodies (Fig. 2).

The diagnostic feature of Bk horizons is the presence of discrete light gray (N7) or pinkish white (5YR 8/2) nodules and tubules of calcite, corresponding to stage II morphology of Gile et al. (1966). Nodules have irregular outlines and range in diameter from 2 to 15 cm. Tubules are comparable in size to nodules, but are elongate perpendicular to bedding and may taper downward, characteristics that suggest an origin as rhizoliths (Klappa, 1980) (Fig. 5A). Nodules and tubules are randomly distributed throughout the horizon, although in some cases there is a marked increase in carbonate near sand–mud contacts, a phenomenon that reflects the fact that downward movement of vadose water is impeded at sharp textural boundaries (Stuart and Dixon, 1973; Aylor and Parlange, 1973; Clothier et al., 1977).

The nodules and tubules consist of euhedral to subhedral crystals of calcite mostly from 2 to 8 μm in diameter (Fig. 7B). Micrite constitutes over 90% of the volume of the nodules and tubules, with the remainder being randomly scattered detrital grains, resulting in a carbonate mudstone


(a)
(b)
(c)
(d)
(e)
(f)
(g)
(h)

texture (Fig. 6D). Some detrital grains have a fringe of sparry calcite, a feature recognized in other Pliocene and Quaternary calcic paleosols (Amiel, 1975; Mount and Cohen, 1984). In some cases the micrite is brecciated, probably as a result of desiccation, and the vughs and planes created by the brecciation are infilled with sparry calcite or are lined with argillans (vugh argillans and skew-plane argillans; Fig. 6E).

Sandy parent sediment located between the nodules and tubules is generally structureless, although blocky and prismatic peds exist. Free-grain and embedded-grain argillans are common in the upper part of Bk horizons and systematically decrease down-profile. Argillans are probably responsible for the color of the parent sand because it is darkest in the upper part of the horizon (reddish brown, 5YR 4/4; weak red, 10R 5/4) and becomes lighter downward (light reddish brown, 5YR 6/4; yellowish red, 5YR 5/6; pale red, 10R 6/2). Mudrock parent sediment is uniformly reddish brown (5YR 4/4) or weak red (10R 5/4), and displays angular blocky peds and slickensides. Embedded-grain, skew-plane, and ped argillans are rare. In a few cases, the parent mud or sand is mottled brown and greyish yellow green (5GY 7/2), suggesting gley (waterlogged) conditions.

K horizons

As proposed by Gile et al. (1965), a K horizon is a soil horizon so strongly impregnated with carbonate that its morphology is determined by the carbonate. Although never formally accepted as a master horizon into Soil Taxonomy (Soil Survey Staff, 1975), nevertheless, the K designation is particularly useful for separation of weaker calcic horizons (Bk) from horizons with moderate to strong carbonate impregnation (Machette, 1985; Retallack, 1990). K horizons are present in 20% of the paleosols examined in this study and consist of laterally persistent beds of light gray (N7) or pinkish white (5YR 8/2) carbonate 50 to 270 cm thick (Figs. 4, 5B). The beds are massive, with a few root traces and small relict patches of brownish sand or mud parent sediment. Following the definition of Gile et al. (1965), a K horizon that consists of more than 90% authigenic carbonate is designated a K2 horizon or a K2m horizon if indurated. A K2 or K2m horizon may be overlain by a K1 horizon and underlain by a K3 horizon, both of which contain between 50 and 90% authigenic carbonate (Fig. 4). In other cases, a K2 or K2m horizon may be underlain or overlain by a Bk horizon, which contains less than 50% authigenic carbonate, or may be directly overlain by a Bt horizon (Fig. 4). All paleosols in the Camp Rice and Palomas Formations that have K horizons also have Bt horizons. All but one of the K horizons corresponds to stage III morphology of Gile et al. (1966). The exception, located at Lucero Arroyo, has a 5-cm-thick laminar and pisolithic cap above the massive carbonate, corresponding to stage IV morphology. K horizons are petrographically similar to calcareous nodules and tubules of the Bk horizon by displaying carbonate mudstone to wackestone textures, sparry calcite fringe cements around some detrital grains, local brecciation, vugh argillans, and skew-plane argillans.

Calcified root mats (Ak horizon)

Approximately 22% of the paleosols examined in this study consist of laterally persistent, relatively thin beds of light gray (N7) or pinkish white (5YR 8/2) carbonate (Figs. 2, 5C). These carbon-

Fig. 6. Photomicrographs of micromorphologic features of calcic paleosols. Bar scales are 0.5 mm. A. Translocated clay coating detrital sand (free-grain argillans); argillic B horizon. B. Translocated clay coating detrital grains that are embedded in soil plasma (embedded-grain argillans); argillic B horizon. C. Translocated clay coating peds developed in silty mudstone (ped argillans); argillic B horizon. D. Sparry calcite fringing detrital grains surrounded by pedogenic micrite; calcite nodule from Bk horizon. E. Translocated clay lining vughs (vugh argillans) developed in brecciated pedogenic micrite; Bk horizon. F. Micrite coatings around detrital grains (calcans); calcified root mat. G. Root molds in pedogenic micrite; calcified root mat. H. Micrite coatings around detrital grains (calcans) and needle-fibre calcite projecting into pore spaces; see Figs. 7C–7F for higher resolution; calcified root mat.

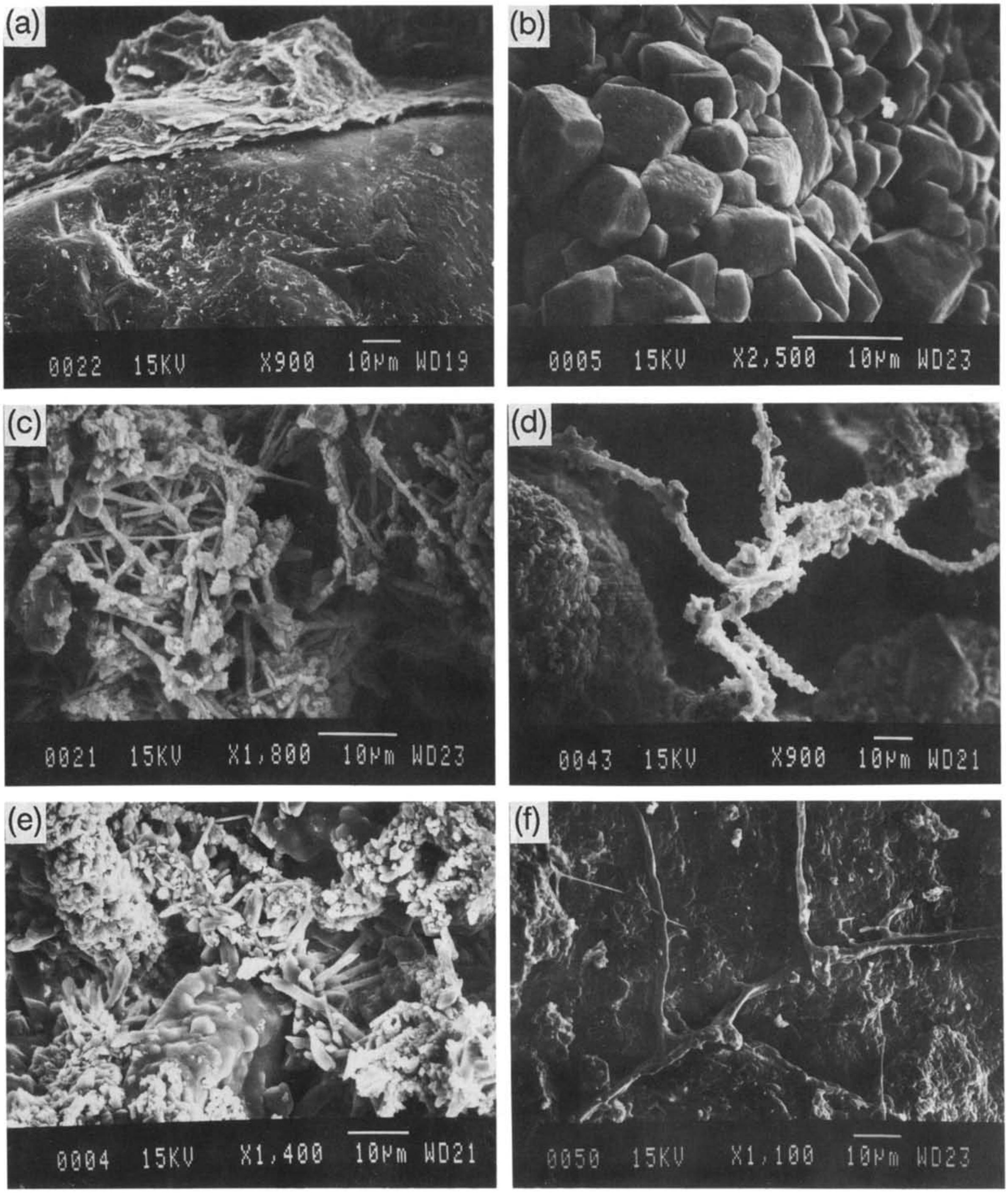


Fig. 7. Scanning electron micrographs of micromorphologic features of calcic paleosols. A. Mixed layer smectite–illite argillan developed parallel–subparallel to detrital quartz grain surface; argillic B horizon. B. Subhedral to euhedral crystals composing calcite nodule; calcic B horizon. C. Needle-fibre calcite; calcified root mat. D. Pore-bridging calcite filaments; calcified root mat. E. Bladed, equant, and smooth calcite filament surface textures of probable fungal origin; calcified root mat. F. Fungal mycelium showing branching filament (hypha) habit and bud structures; calcified root mat.

ate beds superficially resemble K horizons, but closer examination reveals several significant differences: (1) the calcareous beds are composed of numerous calcified roots (rhizoliths) ranging in size from 0.5 mm to 2 cm in diameter; the rhizoliths are parallel to bedding and complexly intergrown, prompting the name calcified root mat; (2) the calcified root mats are generally thinner (10–40 cm) than K horizons (50–270 cm); (3) calcified root mats are not associated with other paleosol horizons, whereas K horizons are always part of a soil profile that at least includes an argillic B horizon; (4) calcified root mats contain a lower proportion of authigenic micrite and higher proportion of detrital grains and empty or spar-filled pore spaces than K horizons or nodules and tubules of Bk horizons (Fig. 8). Detrital grains commonly have micrite coatings (calcans) and root traces display concentric micrite laminae (Figs. 6F, 6G, 6H).

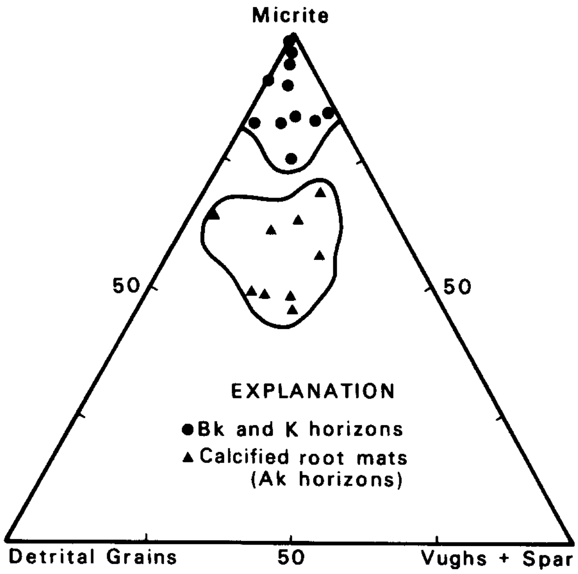


Fig. 8. Triangular diagram plot of micrite, vughs + spar, and detrital grains in Bk, K, and Ak (calcified root mat) horizons.

Calcified root mats develop in the subsurface just above bedrock or hardpan and in areas of very shallow water table (Cohen, 1982; Wright et al., 1988). There is no evidence to suggest that Camp Rice and Palomas root mats developed above a hardpan. However, the abundance of horizontal roots, the absence of vertical roots connecting the horizontal roots to a higher level, and the absence of genetically related paleosol horizons *above* the calcified root mats suggest that the root mats in the Camp Rice and Palomas Formations grew on or close to the land surface in waterlogged soils, perhaps as a result of locally high water table or associated with a spring. Because of their surface or near-surface origin, the calcified root mats are designated Ak horizons.

The calcified root mats contain several textural features, including needle-fibre calcite, alveolar-septal structure, and calcified filaments, that suggest a biochemical origin for at least some of the carbonate. Needle-fibre calcite consists of elongate, 1.6 micrometer diameter (range 0.2–3 μm) and 15 micrometer (range 6–39 μm) long crystals

TABLE 1

Down-profile changes in composition of selected paleosols based on 300 point counts

Type of profile	Horizon	Sample no.	Below top of profile (cm)	$\frac{Q}{F+Lv}$	Plasma (%)	Argillans (%)	Pore space (%)
Bt–Bk–C	Bt	CR-94	25	1.6	0.0	6.0	34.0
		CR-95	83	1.3	15.0	13.0	16.0
		CR-96	125	1.0	22.0	18.0	9.0
Bt–Bk–C	Bt	CR-62	11	1.6	7.0	13.6	20.3
		CR-61	32	1.3	25.3	15.3	14.0
Bt–K–C	Bt	CR-48	7	1.7	9.0	9.6	17.0
		CR-47	25	3.0	15.0	16.6	11.3
		CR-46	42	2.5	15.6	16.3	12.6
Bt–K–C	Bt	CR-91	30	2.8	4.6	23.6	14.6
		CR-92	110	2.0	25.6	23.6	12.3

Q = detrital quartz, F = detrital feldspar, Lv = detrital volcanic rock fragments.

of calcite that are interpreted to have a biogenic origin associated with algae, fungi, or bacteria (Figs. 6H, 7C; Klappa, 1978; Wright, 1986; Phillips and Self, 1987; Phillips et al., 1987). Alveolar-septal structure is composed of a complex network of elongate calcareous tubules and crystals separated by curved partitions that were precipitated by fungi that lived in symbiotic association with root cells (Klappa, 1978; Calvet and Julia, 1983; Wright, 1986; Wright et al., 1988). Also indicative of a biogenic origin are calcareous filaments, which range in diameter from 1 to 14 micrometers (ave. 6.4 μm) with lengths of a few hundred micrometers. Some filaments are hollow while others are filled with calcium carbonate. Filament walls display a wide variety of surface textures (Figs. 7D, 7E, 7F). Similar structures have been described by Klappa (1979), Calvet and Julia (1983) and Phillips et al. (1987), and are interpreted as having a biological genesis. Filament diameter and length suggest they are primarily fungal in origin (Klappa, 1979; Deacon, 1984; Phillips et al., 1987).

C horizons

The C horizon ranges from a few tens of centimeters in thickness to virtually absent. It grades upward into Bk or K horizons and downward into unmodified parent material. The C horizon represents very slightly modified parent material and is often only recognized based on minor color differences compared to underlying sediments.

Lateral variability of paleosols

Calcic paleosols can be traced laterally for hundreds of meters to several kilometers across badlands topography at Picacho Mountain, Lucero Arroyo, Rincon Arroyo, Hatch Siphon, Oasis, and Palomas Creek (Fig. 1). At the latter three locations, the outcrops are oriented approximately parallel to the axis of the basin, whereas at Picacho Mountain, Lucero Arroyo, and Rincon Arroyo the outcrops trend perpendicular to the basin axis. Lateral variability of the paleosols is evident on the scale of a few meters, as well as on the scale of kilometers. Some paleosols, however, display no significant changes in thickness or morphology over distances of hundreds of meters to kilometers.

Lateral variation in paleosols over short distances is primarily restricted to changes in thickness of the horizons. Much of this variation is the result of truncation of the top of Bt, Bw, or Bk horizons and probably not related to pedogenic processes. The paleosol horizons are truncated primarily by fluvial and piedmont channels composed of pebble or cobble conglomerate or pebbly medium to coarse sand/sandstone. Calcareous rip-up clasts derived from Bk or K horizons are commonly present within the channel lag. In a few cases, paleosol horizons developed in floodplain mudstone are separated from floodplain silt and fine sand by a sharp contact displaying several centimeters of relief, indicating erosional truncation of the underlying paleosol.

Lateral variation in horizon thickness over short distances is also present in some Bt–Bk, Bt–K, and Bw–Bk profiles. The Bt or Bw horizons change thickness by tens of centimeters over a distance of a few meters laterally at the expense of the underlying Bk or K horizon, resulting in a swaley contact. The swaley contacts are commonly localized in paleosols that otherwise show little variation in horizon thickness. The swaley contacts are similar to the pipes observed by Gile et al. (1981) in Quaternary soils near Las Cruces. Pipes are interpreted to be sites of higher than normal permeability due to textural differences in the parent sediment or to the presence of large tree roots or burrows (Fig. 5D; Gile et al., 1981).

Another variation in paleosols is the presence in Bt–K profiles of localized collapse features. Although about the same scale as pipes, the collapse features are easily recognizable because of: (1) sharp, steep sides, which in many cases are small-scale faults, (2) no change in thickness of the Bt horizon across the structure, and (3) localized brecciation of the K horizon. Collapse features probably developed by dissolution of carbonate of the K horizon when the paleosol dropped below the water table and is, in effect, a dissolution feature.

Lateral variability of paleosol morphology over distances of hundreds of meters or kilometers also exists. At Rincon Arroyo a Bt–Bk–K2m profile can be traced for several kilometers southward where it changes to a Bt–Bk profile. The change is accomplished by gradual thinning of the K horizon and concomitant thickening of the Bk horizon, with little variation in the thickness of the Bt horizon. A similar relationship exists near Picacho Mountain, where a floodplain mudstone lacking pedogenic carbonate changes westward within a kilometer to a Bk horizon. In both of these examples the critical factor controlling paleosol morphology appears to be position within the basin. The morphological stage increases toward the edge of the basin, where sedimentation rates were slower and paleosols had more time to develop and where preservation potential of paleosols was greater due to less frequent erosion by channels (Bown and Kraus, 1987; Kraus, 1987; Kraus and Bown, 1988).

A second type of lateral variation in paleosols over a large area can be illustrated by an example from Rincon Arroyo. A 12-m-thick, multistorey fluvial channel changes northward, away from the basin center, within 1.2 km into two 3-m-thick channels that are separated by 5 m of floodplain mudstone and fine sand displaying three distinct Bt–Bk profiles (Fig. 2). Once again, pedogenesis was more active or paleosols had a higher preservation potential along the margin of the basin, where channel forming processes were less common.

Pedogenic processes

The case for pedogenesis

Calcic paleosols in the Camp Rice and Palomas Formations formed in the vadose zone within a meter or so of the sediment surface. Evidence to support this interpretation includes: (1) paleosol profiles involving two or more horizons with the calcic horizon at the base, (2) translocated clay in the upper part of the paleosol profiles, (3) vertical root traces and calcareous rhizoliths, (4) the presence of slickensides, stress argillans, and carbonate breccia, and (5) red and brown colors.

The presence of argillic B horizons directly above calcic B or K horizons is a common feature of modern soils in relatively dry climates and provides strong evidence that the Bt–Bk or K couplets in the Camp Rice and Palomas Formations are pedogenic (cf. Buol et al., 1980, pp. 246–267). Transported in solution, carbonate moves farther down the soil profile than particulate clay, which is transported through the soil in the solid state (Gile et al., 1981). Fine-grained calcite precipitates in the lower part of the zone of wetting as the soil water is removed by plants and evaporation (evapotranspiration; Birkeland, 1984, pp. 138–146). Alternating wetting and drying of the soil results in slickensides and stress argillans, caused by shrinking and swelling of smectite clays, as well as brecciation of some of the pedogenic carbonate. Red color supports the interpretation that the paleosols developed in a well oxygenated vadose zone, as does the presence of vertical root traces and calcareous rhizoliths.

The characteristics of the calcic paleosols in the Camp Rice and Palomas Formations differ markedly from groundwater carbonate that precipitates below the water table, in the zone of fluctuating water table, or in the zone of capillary draw a few tens of centimeters above the water table (Hay and Reeder, 1978; Semeniuk and Meagher, 1981; Tandon and Narayan, 1981; Arakel and McConchie, 1982). Groundwater carbonate is generally not directly overlain by a horizon of translocated clay and has sharp lower contacts. Zones of groundwater carbonate may also have evidence of reducing conditions in the form of drab hues of gray or green in the parent sediment, colors which are largely absent from parent sediment of Camp Rice and Palomas paleosols. In the zone of fluctuating water table and in the zone of capillary draw, the host sediment of the groundwater carbonate may show mottling of red, yellow, gray or green, as a result of alternating reduced and oxygenated conditions. Color mottling (gley) in Camp Rice and Palomas paleosols is rare.

Calcified root mats are distinctly different than other paleosols in the Camp Rice and Palomas Formations and resemble groundwater carbonate, in that the root mats are not associated with other paleosol horizons, they do not have translocated clay, and they have sharp lower contacts. However, macroscopic and microscopic textures described above suggest that much of the carbonate precipitation was associated with the microchemical environment surrounding roots, rather than with inorganic precipitation at or below the water table. Furthermore, the fact that the rhizoliths are horizontal and there are no genetically related paleosol horizons above the root mats, argue strongly against formation of the calcified root mats in the zone of capillary draw immediately above the water table, which is characterized by long, vertical roots of phreatophytes (Semeniuk and Meagher, 1981). The calcified root mats in the Camp Rice and Palomas Formations appear to have formed at or near the land surface.

Origin of translocated clay and pedogenic carbonate

There are several possible origins for translocated clay in Camp Rice and Palomas paleosols. Some of the clay may have been detrital in origin, that is, it was present in the parent sediment prior to the onset of pedogenesis and was subsequently redistributed through the soil profile. A detrital origin is likely for some of the translocated clay in muddy parent sediment, but is less likely in sandy parent sediment, because pedogenically unmodified sand beds contain little or no clay. A second possibility is that the clay was produced in situ by breakdown of detrital feldspar and rock fragments. The evidence in support of this process is mixed (Table 1). In all but one of the profiles examined there is no consistent correlation between the relative abundance of detrital feldspar or volcanic rock fragments and the percentage of argillans and plasma (Table 1). However, in one Bt–K profile a down-profile increase in the ratio of quartz to feldspar plus volcanic rock fragments corresponds to an increase in argillans and plasma, suggesting that some of the clay may have been derived from breakdown of the feldspars and volcanic rock fragments. It is also likely that some of the translocated clay originated as windblown dust. Gile and Grossman (1979) demonstrated that up to 40% of the windblown dust near Las Cruces consists of clay.

Carbonate in Bk and K horizons was probably also derived from windblown dust, a phenomenon that has been quantitatively demonstrated for Quaternary soils (Gile and Grossman, 1979; Gile et al., 1981; Mayer et al., 1988). A second, although minor, source of carbonate is the dissolution of detrital carbonate rock fragments. This process should have been most important in paleosols developed in piedmont sediment that was derived largely from Paleozoic carbonate rocks, such as along the eastern flank of the Palomas basin (Garfield and Red Canyon sections; Fig. 1). In contrast, fluvial strata contain only a few percent or less of carbonate rock fragments and piedmont strata on the hanging wall dip slope of the Palomas basin (Oasis and Palomas Creek sections) contain no detrital carbonate rock fragments.

A common feature of Bt–Bk and Bt–K profiles in the Camp Rice and Palomas Formations is the presence of translocated clay in the upper part of the Bk and K horizons. Gile et al. (1981) suggest that clay is translocated into the calcic B horizon contemporaneously with carbonate precipitation, suggesting that argillans in the upper part of Bk or K horizons are a natural consequence of pedogenesis of calcic soils. A second possibility is that the zone of carbonate accumulation invaded the base of the argillic horizon in response to soil maturation. For example, as the amount of clay in the Bt horizon increases, so does the water-holding capacity, which inhibits downward movement of water and promotes precipitation of carbonate in the lower part of the Bt horizon (Birkeland, 1984, p. 313). Soil maturation also manifests itself by plugging the calcic horizon with carbonate, ultimately producing a K horizon. Additional carbonate is added to the top of the K horizon, forcing the K horizon to invade the base of the overlying Bt horizon (Gile et al., 1981). It is also possible that argillans in the upper parts of the Bk or K horizons could reflect temporal changes in extrinsic variables, either by addition of sediment to the soil or by a change in climate.

In the former case, a cumulate soil, intermittent addition of sediment to the top of the profile would cause the Bt–Bk/K boundary to rise and the Bk/K horizon would invade the former base of the Bt horizon. Similarly, a change to drier climate would cause the top of the zone of carbonate accumulation to rise into the base of the former Bt horizon (Jenny and Leonard, 1934; Arkley, 1963). In contrast, a change to wetter climate would lower the Bt–Bk/K contact and allow the base of the Bt horizon to invade the former top of the Bk or K horizon.

Petrographic data provide evidence to distinguish between the various mechanisms for producing translocated clay in the upper part of the calcic horizon. In the cases of increased water-holding capacity of the Bt horizon, upward growth of K horizon, a cumulate soil, and a shift to drier climate, the zone of carbonate precipitation invades the base of the argillic B horizon. Petrographic evidence of this invasion should exist in the form of pedogenic carbonate crosscutting and replacing argillans. In contrast, the change to wetter climate promotes the invasion of the former top of the calcic horizon by translocated clay, a phenomenon that should be manifested in the presence of argillans lining vughs, planes, and peds developed in pedogenic carbonate. Contemporaneous formation of argillans and pedogenic carbonate should result in a mixture of the petrographic features described above.

For the majority of Camp Rice and Palomas paleosols examined (70%, $n = 16$), petrographic evidence suggests that argillans and pedogenic carbonate formed contemporaneously in the upper part of the Bk and K horizons. Pedogenic carbonate crosscuts argillans, or more commonly, relict detrital grains with argillans are completely surrounded by pedogenic carbonate. There are present in the same horizons, however, argillans lining vughs, skew planes, or peds developed entirely within pedogenic carbonate (Fig. 6E). Only one paleosol exhibits carbonate crosscutting argillans, but with no evidence of argillans coating carbonate structures. In approximately 25% of the paleosols examined, it was not possible to evaluate the temporal relationship between translocated clay and pedogenic carbonate because of the lack or diagnostic crosscutting relationships.

Role of grain size of parent sediment

Grain size of the parent sediment played an important role in the distribution of argillic B horizons, which are more common in sand or silt parent sediment than in mud-rich parent sediment. This relationship probably reflects greater permeability and larger pore spaces in sandy and silty parent sediment compared to finer parent sediment. It is also possible that argillans were originally present in paleosols of muddy parent sediment but were destroyed by subsequent shrinking and swelling of smectite clays (Gile and Grossman, 1968).

The role of grain size of the parent sediment on grain size of pedogenic carbonate was also evaluated for the Camp Rice and Palomas Formations. Wieder and Yaalon (1974) suggested that pedogenic carbonate in predominantly clay-sized parent sediment would more likely be composed of micrite, because of the large number of nucleation sites, than sandy soils, which would consist of microspar and spar. These relationships are not evident in Camp Rice and Palomas paleosols, however. Coarse micrite is the dominant form of pedogenic carbonate regardless of the grain size of the parent sediment. Microspar is rare (< 2% of samples) and is present in both muddy and sandy parent sediment.

Role of depositional environment

Depositional environment exerted significant control on the distribution and morphology of calcic paleosols in the Camp Rice and Palomas Formations. This relationship has been previously demonstrated by Mack and Seager (1990) and Mack and James (1992) and will be summarized here. Four lithofacies described by Mack and Seager (1990) are applicable to this study: (1) proximal piedmont, which is composed primarily of conglomerate and minor amounts of sandstone and sandy mudstone and was deposited adjacent to the footwall block of the Palomas half graben; (2) distal piedmont, consisting of subequal amounts of channel conglomerate and interchannel sand/sandstone and mudstone that were deposited on the hanging wall dip slope of the Palomas half graben; (3) axial-fluvial channel sand/sandstone that was deposited in a narrow belt ($\leqslant 5$ km) near the footwall block of the Palomas half graben, and (4) axial-fluvial sediment deposited in the Hatch–Rincon and Jornada full grabens and consisting of interbedded channel sand/sandstone and overbank mudstone, silt, and very fine sand.

Calcic paleosols are uncommon and restricted to stage II morphology in proximal piedmont lithofacies (Fig. 2, Red Canyon, Garfield). Paleosol development was likely inhibited by rapid sedimentation and small fan size, which did not allow for long periods of non-deposition on the fan surface (Mack and Seager, 1990). Paleosols may also have developed on upper fan surfaces and were subsequently eroded. It is also possible that early cementation by coarse groundwater calcrete obliterated original textures of the pedogenic carbonate. In contrast, calcic paleosols are abundant and attain stage III morphology in fine-grained distal piedmont strata (Fig. 2, Palomas Creek, Oasis), a relationship that is a response to asymmetrical subsidence of the Palomas half graben. During periods of fault movement, the gradient increased on the hanging wall dip slope, causing piedmont channels to incise and shifting the locus of sedimentation basinward (Hooke, 1972; Leeder and Gawthorpe, 1987; Mack and Seager, 1990). During piedmont channel entrenchment, fine-grained interchannel areas became isolated terraces, enhancing the process of soil formation. Calcified root mats (Ak horizons) are restricted to distal piedmont lithofacies (Fig. 2), suggesting that they formed in environments that were unique to broad fans and/or alluvial flats. Several possibilities include the margins of small interchannel ponds, or springs that developed near the toe of the fans.

Asymmetrical subsidence is also largely responsible for the location and width of axial-fluvial lithofacies in the Palomas half graben. The fluvial channel system avulsed into the area of maximum subsidence, which in the Palomas half graben was close to the footwall block (Leeder and Gawthorpe, 1987; Mack and Seager, 1990; Mack and James, 1992). This process resulted in a narrow belt of fluvial lithofacies that interfingered with proximal piedmont conglomerates adjacent to the footwall block. As the channel system moved across the narrow alluvial plain, it eroded overbank fines and previously deposited channel sand, producing multistorey sheets of channel sand with virtually no interbedded overbank fines (Mack and Seager, 1990; Mack and James, 1992). Active channel reworking also inhibited the development of paleosols, resulting in the absence of stage II and III morphology calcic paleosols in axial-fluvial strata of the asymmetrical Palomas basin (Fig. 2, Red Canyon, Garfield).

In full grabens, however, the channel system had the opportunity to migrate across almost the entire basin. Those parts of the basin that were located far from the active channel system experienced overbank sedimentation and pedogenesis. The result was preservation of relatively thick (up to 4 m), laterally continuous beds of overbank fines, many of which contain stage II and III morphology calcic paleosols (Fig. 2, Hatch Siphon, Rincon Arroyo).

Conclusions

Calcic paleosols are common within the Camp Rice and Palomas Formations of the southern Rio Grande rift. Evidence supporting a pedogenic origin for calcic and associated horizons are: (1) vertical profiles of two or more horizons with the calcic interval present toward the base; (2) translocated clay in the upper part of paleosol profiles; (3) presence of calcareous root traces and rhizoliths; (4) slickensides, stress argillans, and carbonate breccia; and (5) red and brown horizon colors. Six types of paleosol horizons are present in combination or singly forming seven types of paleosol profiles: A–Bw–Bk–C, A–Bt–Bk–C, Bw–Bk–C, Bt–Bk–C, Bt–K–C, Bk–C, and Ak. Bt often in combination with Bk and/or K horizons dominate these paleosol profiles. Non-calcified A horizons are only rarely preserved. Calcified root mats (Ak horizons) are common and restricted to distal piedmont lithofacies.

Paleosol horizons are traceable laterally for hundreds of meters to several kilometers both parallel and normal to the axis of basins. Lateral thickness variations over short distances are related to truncation of upper pedogenic horizons by erosion, textural differences in parent material, local presence of large tree roots or burrows, and localized collapse due to partial dissolution of K horizons. Lateral variations over longer distances, such as gradation from a Bk to K horizon, require at least hundreds of meters and resulted from more active pedogenesis away from the locus of fluvial channel activity.

Micrite is the dominant form of pedogenic carbonate in Bk and K horizons regardless of parent material grain size. However, the grain size of parent material does influence the distribution of argillic B horizons in paleosols of the Camp Rice and Palomas. Argillic subsurface accumulations were favored by greater pore size and permeability in sandy and silty sediment.

Paleosol distribution was controlled by depositional facies and ultimately basin symmetry throughout the southern Rio Grande rift. Broad, symmetrical basins favored the development of calcic paleosols (stages II and III morphologies) in the overbank setting of the axial-fluvial system, especially in areas far removed from active channels. There is an absence of stages II and III paleosols in axial-fluvial lithofacies of asymmetrical basins due to rapid channel migration across narrow alluvial belts. In proximal piedmont lithofacies, paleosols are uncommon and develop no higher than stage II morphology. However, piedmont tributary entrenchment, caused by basin boundary faulting, favored development of abundant calcic paleosols with up to stage III morphology in the distal piedmont lithofacies.

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