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Calcareous paleosols of the Upper Triassic Chinle Group, Four Corners region, southwestern United States: Climatic implications

Lawrence H. Tanner[†]
Department of Biological Sciences, Le Moyne College, 1419 Salt Springs Road, Syracuse, New York 13214, USA

Spencer G. Lucas
New Mexico Museum of Natural History, 1801 Mountain Road N.W., Albuquerque, New Mexico 87104, USA

ABSTRACT

Paleosols are prominent features of the Upper Triassic Chinle Group. The oldest (Carnian-age) formations of the Chinle Group (Zuni Mountains and Shinarump Formations) contain kaolinitic paleosols that display gley features but generally lack calcretes. Paleosols of the (Upper Carnian) Blue Mesa Member of the Petrified Forest Formation are mostly mature Alfisols that have distinctive horizonation and commonly host stage II to III calcretes. Mudstones of the Jim Camp Wash Bed of the overlying Sonsela Member host similarly mature paleosols with abundant stage II to stage IV calcretes. The (Lower Norian) Painted Desert Member of the Petrified Forest Formation is characterized by paleosols that lack well-developed A horizons but display thick, red B horizons in which pedogenic slickensides, rhizocretions, and stage II to III calcretes are locally abundant. Immature paleosols hosting stage II to stage III calcretes characterize the lower part of the (Middle Norian) Owl Rock Formation. The upper Owl Rock Formation contains stage III to IV calcretes and laterally persistent limestone ledges that formed as palustrine limestones and groundwater calcretes. The (Norian-Rhaetian) Rock Point Formation generally lacks pedogenic features in most of the study area, but the uppermost strata in some locations host multiple pedogenic horizons that display drab root traces, desiccation cracks, and stage II to III calcretes. Interformational variations in the types of paleosols and the maturity of calcretes in Chinle Group strata reflect gradual aridification across the Colorado Plateau during the Late Triassic. This climatic change overprinted variations in basin sedimentation rate that were potentially controlled by base level and tectonics.

Keywords: Chinle, pedogenic, calcretes, Late Triassic, palustrine.

RESUMEN

La presencia de paleosuelos es uno de los rasgos más característicos del Grupo Chinle del Triásico Superior. Las formaciones más antiguas, Zuni Mountains y Shinarump, son de edad Carniense, no tienen calcretas y los paleosuelos son caoliníticos con rasgos de gley. Los paleosuelos del Miembro Blue Mesa de la Formación Petrified Forest son sobre todo Alfisoles maduros con horizontes bien diferenciados y con

[†]E-mail: tannerlh@lemoyne.edu.

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calcretas de estadios II a III. Las lutitas de las capas Jim Camp Wash del Miembro Sonsela suprayacente también tienen paleosuelos maduros con frecuentes calcretas de estadios II a IV. El miembro Painted Desert (Noriense inferior) de la Formación Petrified Forest se caracteriza por presentar paleosuelos que carecen de horizontes A bien desarrollados, pero que presentan horizontes B rojos y muy potentes con slickensides pedogénicos, rizocreciones, y calcretas de estadios II a III localmente abundantes. Los paleosuelos inmaduros que contienen calcretas de estadio II a III caracterizan la parte inferior de la Formación Owl Rock (Noriense medio). La parte superior de dicha formación contiene calcretas estadio III a IV y lateralmente incluyen lentejones de calizas, que se han interpretado como depósitos palustres y calcretas freáticas. La formación Rock Point (Noriense-Rhetiense) no presenta rasgos pedogénicos en la mayor parte del área estudiada, pero localmente en los estratos superiores hay horizontes pedogénicos múltiples que presentan trazas de raíces, grietas de desecación y calcretas estadios II–III. Las variaciones en el tipo de paleosuelos y en los estadios de madurez de las calcretas en las distintas formaciones del Grupo Chinle reflejan una aridificación gradual a lo largo de la Meseta del Colorado durante el Triásico Superior. Este cambio climático, controló las variaciones en la tasa de sedimentación en la cuenca que también estuvieron potencialmente controladas por cambios en el nivel de base y por la tectónica.

Palabras clave: Chinle, pedogénico, calcretas, Triásico Superior, palustres.

INTRODUCTION

The utility and temporal resolution of paleoclimate modeling has improved dramatically in recent decades, in part through the increasing use of paleosols and pedogenic features as paleoclimate archives. Calcrete, the accumulation of $CaCO_3$ in the subsurface environment as nodules (glaebules) or cemented horizons (see review in Wright and Tucker, 1991), has proven particularly useful because of its high preservation potential, and common association with soil-forming processes in semiarid to arid climates. In addition to climate, however, the morphology and maturity of all paleosols, including those that are calcareous, are controlled by a variety of factors, including sediment accumulation rate, which controls residence time of sediment in the soil-forming environment, vegetative cover, subsurface biotic activity, and host sediment composition (see reviews in Kraus, 1999; Retallack, 2001; Alonso-Zarza, 2003). The controls on the rate of sediment accumulation, which in general is inversely related to paleosol maturity, are particularly complex. Although climate exerts some control over the delivery of sediment to the receiving basin, tectonic activity greatly influences sediment deposition, both through enhancement of sediment production by source area uplift, and through control of basin configuration and accommodation space.

Paleosols have long been recognized as prominent features in Upper Triassic Chinle Group strata, and general features of many of these paleosols have been described previously. Some previous studies have focused on specific local occurrences within an individual formation; for example, Kraus and Middleton (1987) described a history of floodplain incision and aggradation cycles during deposition of the Petrified Forest Formation using, in part, variations in paleosol maturity on the alluvial plain; Therrien and Fastovsky (2000) documented that paleosol hydrology varied with distance from the channel in the same formation. Other studies have cited general characteristics of Chinle Group paleosols, typically in the context of regional changes in climate during the Late Triassic, but generally without detailed documentation of specific pedogenic features (Dubiel, 1987, 1994; Dubiel and Hasiotis, 1994a, 1994b; Hasiotis and Dubiel, 1994; Demko et al., 1998; Hasiotis et al., 1998). The Chinle Group outcrops cover a broad area of the Southwestern United States (Fig. 1) and record deposition during a major portion of the Late Triassic. A thorough and detailed examination of the lateral and temporal variations of the Chinle paleosols is not possible within the scope a single paper. Rather, we survey the spectrum of paleosols present in Chinle Group strata, with particular attention to calcretes, by describing the results of previous studies, as well as providing new data. From these data (old and new), we attempt to interpret the general sedimentologic and climatic conditions that existed during deposition and pedogenesis of Chinle strata.

Calcretes occur in most Chinle formations, yet they have received only limited attention. Most previous studies have focused on individual formations at specific locations; examples include Blodgett's (1988) study of the Norian-Rhaetian Dolores Formation, in the uppermost Chinle of southwest Colorado, the examination of Carnian-Norian age paleosols of the Petrified Forest Formation in the Petrified Forest National Park (PFNP; Figs. 1 and 2) by Therrien and Fastovsky (2000), and the description of pedogenic features of the Norian-aged Owl Rock Formation in the Four Corners region by Tanner (2000). The accumulation of carbonate in the B horizon is not specific to any single soil type or climatic region, however, so the conditions of calcrete

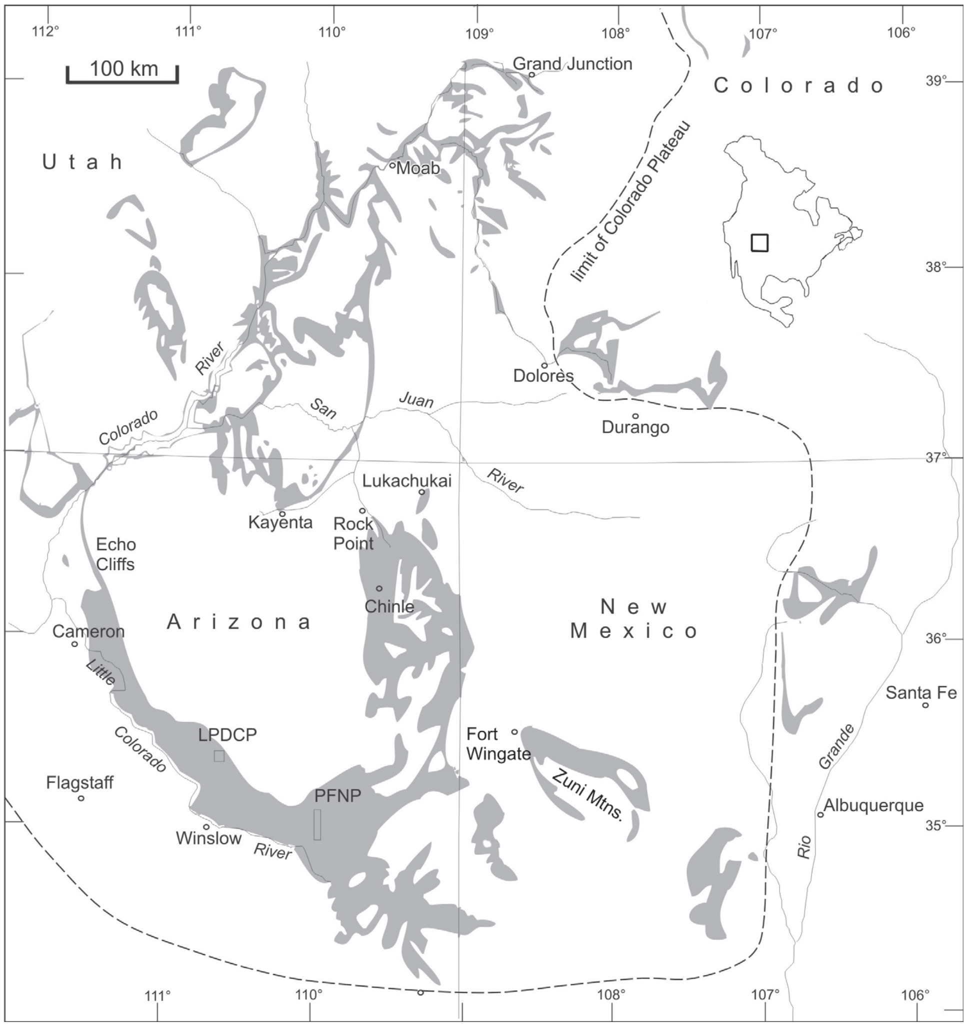

Figure 1. Distribution of outcrops of Chinle Group strata (shaded) in the Four Corners region (after Stewart et al., 1972). Locations mentioned in the text are indicated; PFNP—Petrified Forest National Park, LPDCP—Little Painted Desert County Park.

formation can only be understood within the context of the soils, or paleosols, in which the calcrete occurs.

GEOLOGIC SETTING

The Chinle basin formed as a continental retro-arc basin on the western edge of the North American craton during the initial growth of the Cordilleran magmatic arc in the early Mesozoic (Dickinson, 1981; Lawton, 1994). This basin extended from southwestern Texas to northern Wyoming and was the site of terrestrial sedimentation from the Late Triassic until the beginning of the Early Jurassic (Lucas et al., 1997). Strata of the Chinle Group, ranging in age from Late Carnian to possibly Rhaetian, are exposed across much of the Colorado Plateau (Fig. 1). The Four Corners region (the common border of the states of Arizona, Colorado, New Mexico, and Utah), the focus of this study, was situated within the basin at near-equatorial latitudes (between 5° and 15°N) during Late Triassic time (Scotese, 1994; Molina-Garza et al., 1995; Kent and Olsen, 1997).

Deposition of the Chinle Group sediments was controlled by predominantly west- to northwest-flowing stream systems crossing broad, low-gradient alluvial plains. In the Four Corners region, proximal source areas for these sediments were the Mogollon highlands, located ~500 km to the south and southwest, and to a lesser extent the Uncompahgre highlands located 200–300 km to the east and northeast (Blakey and Gubitosa, 1983; Marzolf, 1994). Syndepositional arc volcanism contributed an appreciable amount of volcaniclastic sediment to the basin. Across most of the Four Corners region, the lowermost Chinle strata were deposited unconformably on Middle Triassic or older strata following an interval of lowered base level and incision.

STRATIGRAPHY

In the Four Corners region, basal Chinle Group strata rest unconformably (the Tr-3 unconformity) on strata of the Moenkopi Formation. Stewart et al. (1972) used the informal designation “mottled strata” to describe alluvial sediments at the base of

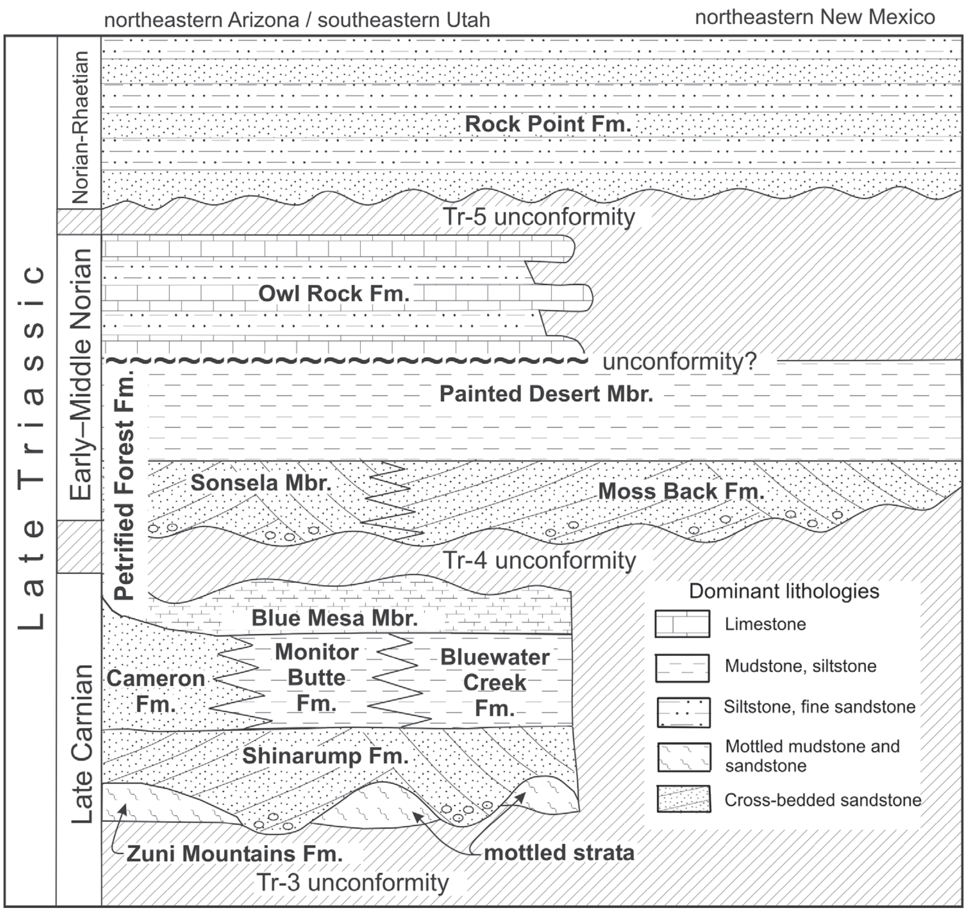

Figure 2. Stratigraphic hierarchy for the Chinle Group used in this paper (after Lucas et al., 1997).

the Chinle Group that exhibit strong pedogenic mottling. These strata underlie or are laterally equivalent to the basal strata of the Shinarump Formation (Lucas et al., 1997). Equivalent strata in the San Rafael Swell were named the Temple Mountain Formation by Robeck (1956), and more recently, Heckert and Lucas (2003) proposed the name Zuni Mountain Formation for these same strata in west-central and north-central New Mexico (Lucas et al., 2003). The Shinarump Formation consists of cross-bedded conglomerates and sandstones of late Carnian age. The Temple Mountain and Shinarump Formations were deposited in paleovalleys incised into the underlying Moenkopi Formation strata (Stewart et al., 1972; Blakey and Gubitosa, 1983; Demko et al., 1998).

Upper Carnian strata immediately above the Shinarump Formation are named regionally the Cameron, Bluewater Creek, and Monitor Butte Formations. Lucas (1993) and Lucas et al. (1997) demonstrated the stratigraphic equivalence of these formations. The lowermost strata of the Petrified Forest Formation in the Four Corners region are designated the Blue Mesa Member (Lucas et al., 1997). These strata, also of late Carnian age, overlie the Cameron–Bluewater Creek–Monitor Butte Formations. The Blue Mesa Member and the underlying Cameron–Bluewater Creek–Monitor Butte and Shinarump Formations were mapped collectively as the lower bentonitic part of the Chinle Formation by Stewart et al. (1972). Thickness of these formations varies from <30 m in southeastern Utah to >450 m in northwestern New Mexico.

An unconformity (Tr-4) separates the Blue Mesa Member from the overlying sandstone-dominated Sonsela Member of the Petrified Forest Formation and the equivalent Moss Back Formation (Lucas, 1993; Heckert and Lucas, 1996; Lucas et al., 1997). Heckert and Lucas (2002a) interpreted the Sonsela Member as filling erosional scours in the underlying Blue Mesa Member, which thins beneath the unconformity to the east. The Painted Desert Member, of early to middle Norian age, overlies Sonsela–Moss Back strata (Lucas et al., 1997). As mapped by Stewart et al. (1972), the thickness of the entire Petrified Forest Formation ranges from just over 30 m at its northeastern limit in eastern Utah, increasing southward to over 400 m at its southeastern extent in northwestern New Mexico. The overlying Owl Rock Formation is composed of up to 150 m of strata that crop out in northern Arizona, northwestern New Mexico, and southern Utah (Stewart et al., 1972; Lucas and Huber, 1994).

Across the Four Corners area, the upper Norian to (possibly) Rhaetian Rock Point Formation is recognized as the youngest stratigraphic unit of the Chinle Group (Lucas, 1993; Lucas et al., 1997). The Rock Point Formation, termed the Rock Point Member of the Wingate Sandstone by Stewart et al. (1972), includes strata formerly assigned to the Church Rock Member of the Chinle Formation (Stewart et al., 1972; Dubiel, 1989; Lucas et al., 1997). The contact between the Rock Point Formation and underlying Owl Rock Formation is unconformable (the Tr-5 unconformity). The Rock Point Formation grades vertically to the eolian-dominated Wingate Formation of Rhaetian to Hettangian age (Harshbarger et al., 1957; Tanner et al., 2002; Molina-Garza et al., 2003).

PALEOSOLS OF THE CHINLE GROUP

Many descriptions of paleosols have suffered from the lack of a single system of paleosol description applied by all researchers. In part, this is because traditional modern soil classification (i.e., Soil Conservation Service, 1999) requires knowledge of such information as vegetative cover and soil moisture that may not preserved in paleosols. Additionally, postdepositional processes (erosion, compaction, diagenesis) may obscure some primary pedogenic features and induce other nonpedogenic features (Retallack, 2001). Although the precise interpretation of paleosols in terms of modern soil classification standards is often problematic, sufficient information is often retained to allow classification at the order level (Kraus, 1999; Retallack, 2001). This study assigns order names to paleosols that are consistent with modern soil usage unless otherwise noted (Table 1). Where such assignments are not possible, we use the alternative classification system specific for paleosols, which abandoned some traditional soil orders and created new paleosol orders (Mack et al., 1993). The term calcrete is used here to refer to the displacive and replacive growth of calcium carbonate in the soil-forming environment, and includes carbonate of both pedogenic and groundwater origins (see reviews in Wright and Tucker, 1991; Tandon and Kumar, 1999; Alonso-Zarza, 2003). The maturity of nodular calcrete is described using the stage concepts of Gile et al. (1966) and Machette (1985).

Zuni Mountains Formation (= Mottled Strata)

Lithostratigraphy

Stewart et al. (1972) used the informal designation "mottled strata" to describe alluvial sediments (mudstones, sandstones, and conglomerates) at the base of the Chinle Group that exhibit strong pedogenic mottling. These strata underlie or are laterally equivalent to the basal strata of the Shinarump Formation (Lucas et al., 1997). Equivalent strata in the San Rafael Swell in central Utah were named the Temple Mountain Formation by Robeck (1956), and more recently, Heckert and Lucas (2003) proposed the name Zuni Mountain Formation for these same strata in west-central and north-central New Mexico. These same authors also noted that, locally, the Shinarump Formation may be absent, in which case the Zuni Mountains Formation is overlain by the Bluewater Creek Formation. At the type location, near Fort

TABLE 1. SUMMARY OF PALEOSOL TYPES AND CALCRETE FEATURES IN STRATA OF THE CHINLE GROUP

Age	Stratigraphic unit	Depositional environment	Paleosols	Calcretes
Late Norian–Rhaetian	Rock Point Formation	Eolian, ephemeral lake, minor alluvial	Aridisol/Inceptisol, Calcisol	Local stage II–III
Middle Norian	Owl Rock Formation	Alluvial channel, floodplain, minor lacustrine and wetland	Calcisol, calcic Alfisol	Common stages II–III, local stage IV, groundwater calcretes
Early to middle Norian	Painted Desert Member, Petrified Forest Formation	Alluvial channel and floodplain	Vertisols, Alfisols	Stage II common, local stage III, common calcrete channel lag deposits
Carnian-Norian boundary	Sonsela Member, Petrified Forest Formation	Alluvial channel and minor floodplain	Calcic Alfisols	Stage II common, local stage III, common calcrete channel lag deposits
Latest Carnian	Blue Mesa Member, Petrified Forest Formation	Alluvial channel and floodplain	Alfisols, Vertisols	Abundant stage II, some local stage III
Late Carnian	Monitor Butte, Mesa Redondo and Cameron Formations	Alluvial channel, floodplain, and minor lacustrine	Vertisols, Alfisols	Stage II common locally
Late Carnian	Shinarump Formation	Alluvial channel and minor interfluve	Bioturbated with gley features	Absent
Late Carnian	Zuni Mountains and Temple Mountain Formations–mottled strata	Interfluve valley fills	Spodosols	Generally absent/rare stage III

Wingate, New Mexico, over 20 m of pedogenically modified strata overlie the Middle Triassic Moenkopi Formation (Heckert and Lucas, 2003).

Pedogenic Features

Strata of the Zuni Mountains Formation display, as do other correlative exposures in the region, extensive pedogenic modification, consisting of some combination of gleying (the presence of low chroma colors) and bioturbation. Typical features of these paleosols are crudely prismatic fabric, bluish to yellowish gray mottling in a dark reddish to orange brown host (Fig. 3), relict bedding, meniscate burrows, sandy horizons that are cemented almost entirely by hematite (spodic or Bs horizons), a clay mineral assemblage dominated by kaolinite, and penetration of the beds by near-vertical sandstone cylinders up to 1.5 m long (Tanner, 2003a). Indeed, the most striking characteristics of the Fort Wingate section are the prominent mottled horizons and penetration of these horizons by the vertical sandstone-filled casts. Originally, these latter structures, present in much of the Chinle Group, were interpreted as lungfish aestivation burrows (Dubiel et al., 1987); more recently, they have been reinterpreted as crayfish burrows (Hasiotis and Dubiel, 1993a) and the casts of deeply penetrating taproots of monopodial vegetation (Lucas and Hayden, 1989; Tanner, 2003a). Both deep taproots and crayfish burrowing would be possible, perhaps even likely, in regions where meter-scale, water table fluctuations occur regularly. These hydrologic conditions would have been conducive to water-logged soils for humid intervals, but periodic, perhaps seasonal, drawdown of the water table would have been sufficient to allow translocation and oxidation of iron and manganese and shrinkage of expandable clays. However, the abundance of kaolinite in the clay mineral assemblage of this formation (Tanner, 2003a; Tabor et al., 2004) suggests substantial humidity.

Figure 3. Typical pedogenic features of the Zuni Mountains Formation in section near Fort Wingate, New Mexico (= mottled strata), include pale greenish-yellow (10 YR 8/2) and light greenish-gray (5 GY 8/1) mottling in moderate brown (5 YR 3/4) to dark reddish-brown (10 R 3/4) matrix. Hammer head for scale is 17 cm long.

Lower Chinle (undifferentiated) paleosols have been described previously as Gleysols (sensu Mack et al., 1993) formed in a humid but seasonal environment (Dubiel and Hasiotis, 1994b; Hasiotis et al., 1998). Demko et al. (1998) suggested that hydromorphism in lower Chinle paleosols was a consequence of stratigraphic proximity to an aquitard in the underlying Moenkopi Formation. Bown and Kraus (1987), Mack et al. (1993), and Retallack (2001) noted, however, that true hydromorphic (gleyed) soils rarely display extensive bioturbation or evidence of desiccation, and the identification of Gleysols cannot be based solely on the presence of mottled horizons. Conversely, pseudogleying, as demonstrated in these lower Chinle paleosols, indicates that hydromorphic conditions existed only periodically. The abundance of root casts in these paleosols demonstrates that the sediment surface was quite well-vegetated. The presence of hematite-cemented sandstone units in the profile at Fort Wingate documents instead the formation of a well-developed spodic (Bs) horizon. The sandy host was cemented, principally by hematite, in a zone in which clays had largely been destroyed by weathering (Birkeland, 1984). Therefore, we suggest that some of these paleosols formed as Spodosols, which typically form beneath forests (Birkeland, 1984; Retallack, 2001). Thus, these profiles may represent composite palosols in the sense that soil layers buried by subsequent increments of sediment remained in an extensively thick, active soil-forming environment (Wright and Marriott, 1996). A climate characterized by abundant but highly seasonal precipitation is consistent with these features. Alternatively, individual paleosol horizons may have formed as time-separated increments within a complex paleosol that was subjected to an overall pedogenic overprint by later conditions; e.g., long-term climate change could have imparted an overprint through downward translocation of clays or oxides formed under more humid conditions than existed during earlier soil formation. This also may account for the presence of gley features in horizons displaying evidence of desiccation.

Calcretes are generally absent in lowermost Chinle Group strata (i.e., Zuni Mountains and Shinarump Formations) in New Mexico and Arizona, although Tabor et al. (2004) reported a significant accumulation of pedogenic carbonate at the base of a Chinle profile in eastern Utah. These authors described the occurrence of a 1.5-m-thick coalesced nodular (Bkm) horizon (stage III) at the base of a 3.5 m profile, overlain by noncalcareous, mottled kaolinitic mudstone with a blocky fabric and containing goethite nodules. Thick Bk horizons, as described by Tabor et al. (2004), are not typically associated with paleosols containing gley colors, but as noted by Bown and Kraus (1987), calcareous nodules may form in hydromorphic paleosols. Indeed, the formation of a clay pan in the lower B horizon is likely to enhance carbonate accumulation and nodule formation. Tabor et al. (2004) suggested, however, that the profile they described records multiple episodes of pedogenesis, potentially under varying climatic conditions. Therefore, this Bkm horizon may be a relict paleosol formed in Moenkopi strata and unrelated to the later pedogenesis concomitant with initial Chinle deposition.

Shinarump Formation

Lithostratigraphy

The Shinarump Formation consists of cross-bedded conglomerates and quartz arenite sandstones of late Carnian age deposited in paleovalleys incised into the underlying Moenkopi Formation strata (Stewart et al., 1972; Blakey and Gubitosa, 1983; Demko et al., 1998). The quartz arenite sandstones and extrabasinal conglomerates of the Shinarump Formation attain a maximum thickness of 76 m in the Four Corners region, although in many locations, this unit is generally thinner or absent (Fig. 2). The formation records infill of incised paleovalleys by northwest-flowing streams that generally carried a high bedload and were of low sinuosity (Stewart et al., 1972; Blakey and Gubitosa, 1984).

Pedogenic Features

Paleosols are rare in these sandstone-dominated fluvial channel deposits, but interfluve mudstones display extensive pedogenic mottling and bioturbation, as described by Dubiel (1994) and Tanner (2003a). The nature of this pedogenic alteration appears as purple, orange, and gray mottles in a sandy mudstone exhibiting a prismatic fabric, desiccation cracks, and relict ripple lamination. Bedding-plane exposures reveal centimeter-scale concentric banding of purple and yellow zones, potentially a consequence of iron translocated downward along taproots, the casts of which are abundant in Shinarump Formation paleosols. The character of these paleosols is generally similar to that exhibited by the older Zuni Mountains Formation, although the lack of a spodic horizon makes classification more ambiguous. The gley features present here are likewise considered pseudogleying and not indicative of true soil hydromorphism; rather, these features record substantial fluctuations in the position of the water table.

Cameron–Bluewater Creek–Monitor Butte Formations

Lithostratigraphy

The laterally equivalent Cameron, Bluewater Creek, and Monitor Butte Formations, of Upper Carnian age, consist mainly of gray bentonitic to red mudstones, and laminated to cross-bedded fine-grained sandstones. At Petrified Forest National Park, the Bluewater Creek Formation is made up of mainly interbedded sandstone, siltstone, and reddish-purple to grayish-red mudstone (Heckert and Lucas, 2002a). In the Zuni Mountains, however, the Bluewater Creek Formation consists of three distinct lithofacies assemblages: reddish-brown, bluish-gray, and grayish-purple mudstones; ripple-laminated to plane-bedded sandstones; and interbedded bentonitic mudstone and dark shale (Heckert and Lucas, 2002b). The mudstones locally contain abundant plant debris (Ash, 1987, 1989), and thin micritic limestone occurs near the base of the formation (Heckert and Lucas, 2002b).

Pedogenic Features

Tanner (2003a) noted pedogenic features in the basal strata of the Cameron Formation near Cameron, Arizona, that are similar to those observed in the Zuni Mountains and Shinarump Formations, e.g., mottling with gley colors, a high kaolinite content, and pedogenic slickensides. Dubiel (1987) described a similarly mottled unit in the Monitor Butte Formation of southeastern Utah. Dubiel and Hasiotis (1994a, 1994b) interpreted such profiles in (undifferentiated) lower Chinle strata as Gleysols (sensu Mack et al., 1993), although as noted previously, this classification is not consistent with the strict definition of this paleosol order and the evidence for a greatly fluctuating water table. Following the usage of Mack et al. (1993), these paleosols might best be considered gleyed Vertisols.

Paleosols that are stratigraphically higher in the Cameron and Monitor Butte Formations differ in that they typically consist of simple profiles with decimeter-scale light-colored horizons overlying thick (up to 8 m) reddened argillic (Bt) horizons (Dubiel and Hasiotis, 1994b; Hasiotis et al., 1998). Bluewater Creek mudstones commonly display pedogenic slickensides and scattered (stage II) centimeter-scale calcrete nodules with alpha fabrics (Heckert and Lucas, 2002b). Dubiel and Hasiotis (1994b) and Hasiotis et al. (1998) labeled these argillic paleosols with albic horizons as Alfisols.

Petrified Forest Formation: Blue Mesa Member

Lithostratigraphy

The lowermost strata of the Petrified Forest Formation (PFF), designated the Blue Mesa Member (of late Carnian age), conformably overlie the Cameron–Bluewater Creek–Monitor Butte Formations in northeastern Arizona and southeastern Utah; this stratigraphic unit is not present in Colorado (Lucas et al., 1997). This interval consists of bentonitic mudstones with variegated hues of blue, gray, purple, and red, and interbedded thin coarse-grained to very fine-grained sandstones, and it attains a thickness of 100 m or more (Lucas, 1993; Lucas et al., 1997). Most of the sandstones are compositionally and texturally immature, with the exception of the Newspaper Rock Bed, a prominent ledge-forming quartz arenite sandstone unit within the Blue Mesa Member that may be local to the Petrified Forest National Park (Heckert and Lucas, 2002a). The alluvial architecture of the Blue Mesa Member consists of thick muddy floodplain deposits, deeply incised by meter-scale channels. The channels are filled by very fine-grained sand and mud, commonly displaying inclined heterolithic strata (Fig. 4A; lateral accretion surfaces), and are surrounded by well-developed levee complexes with splay deposits (Kraus and Middleton, 1987; Therrien and Fastovsky, 2000). Intraformational disconformity surfaces record cycles of valley incision and fill (Kraus and Middleton, 1987). Kraus and Middleton (1987) proposed that base-level changes during deposition of lower Chinle strata resulted in part from episodes of thermotectonically controlled uplift and subsidence in the Mogollon highlands. The presence of volcanic detritus in the

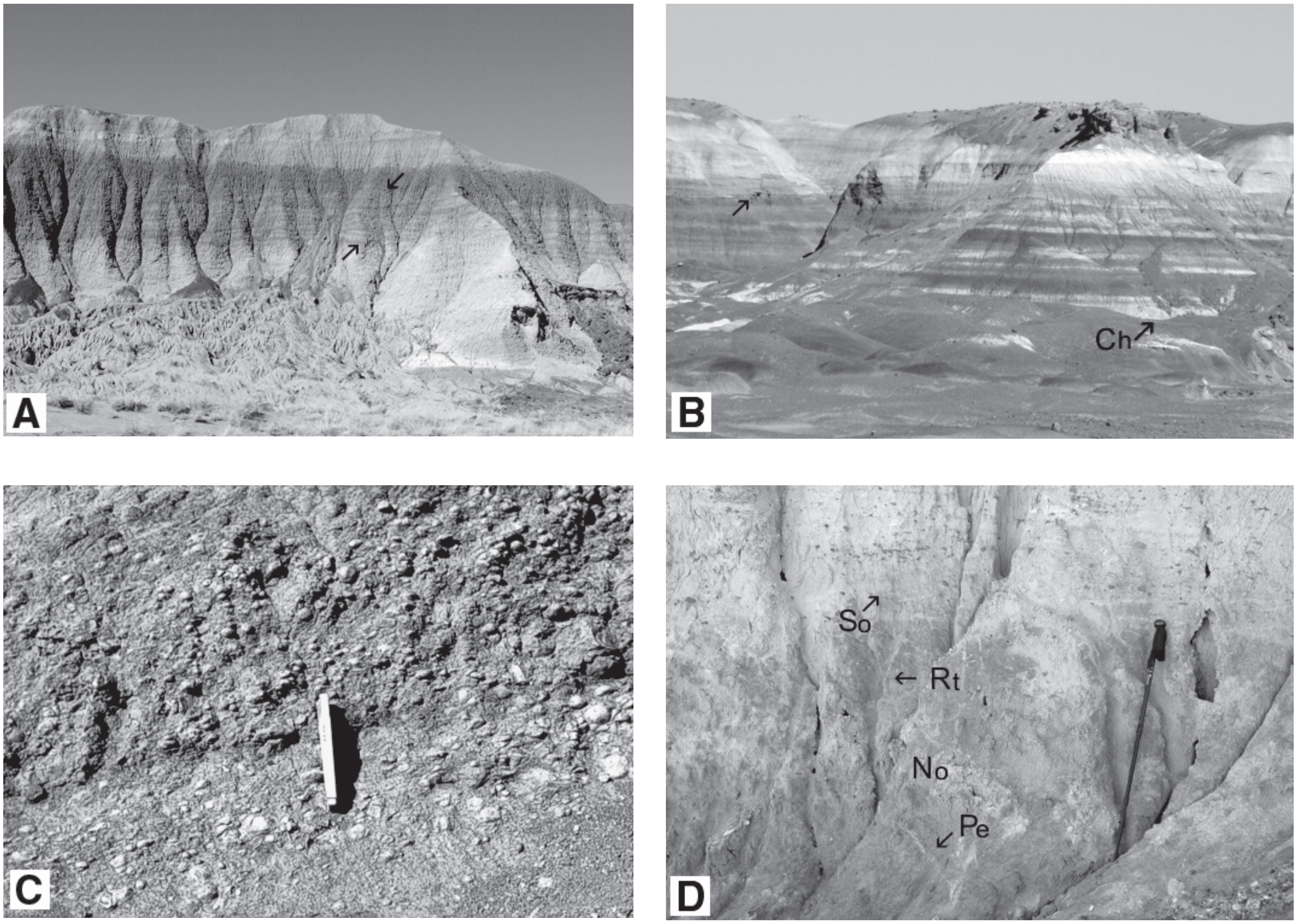

Figure 4. Features of the Blue Mesa Member, Petrified Forest Formation. (A) Channel fill sequence exposed in southern Petrified Forest National Park (the Tepees) displays inclined heterolithic strata set (arrows) ~6 m thick. (B) View of Blue Mesa, in southern Petrified Forest National Park, illustrates contact (at arrow) between Blue Mesa strata and overlying Sonsela Member. Laterally continuous banding in the Blue Mesa strata results from pedogenesis (translocation of oxides and clays) in individual soil profiles formed on an alluvial plain. Ch = lenticular channel-fill deposit. (C) Bk horizon containing abundant (stage II) calcrete nodules in upper Blue Mesa strata, exposed near Moab, Utah. The rule (for scale) is 17 cm. (D) Pedogenic features of the uppermost Blue Mesa Member at the contact with the overlying Sonsela Member (So; Rainbow Forest Bed) at Blue Mesa, Petrified Forest National Park. Rt—drab root traces; No—thin calcrete nodule horizon, Pe—pedogenic slickensides. Hiking staff (for scale) = 120 cm. The mudstone host darkens downward from pale purple (5 P 6/2) to grayish-purple (5 P 4/2).

Chinle, particularly in the Petrified Forest Formation, provides compelling evidence for arc-related magmatism at this time.

Pedogenic Features

The Blue Mesa Member contains thick, well-developed paleosols with distinctive horizonation (Fig. 4B); these are strikingly well-exposed in the strata in the southern end of the Petrified Forest National Park. Typical are composite profiles consisting of stacked, repetitive sequences of thin, light-colored, cross-bedded to ripple-laminated sandstones and mudstones in beds up to 8 m thick that are greenish-gray to dark reddish-brown and mottled gray, purple, and red. Individual profiles within composite profiles commonly contain a thin, sandy ochric epipedon (A horizon), typically overlying a well-defined pale albic (E) horizon. Thick clay-rich B (Bt) horizons (up to 8 m) are reddish gray to (more commonly) grayish purple, and host pedogenic slickensides (wedge-shaped peds) and pseudoanticlines, downward-tapering sandstone-filled fissures (desiccation fractures), sandstone-filled root casts, a variety of arthropod burrow structures, drab root traces up to 0.1 m long, centimeter-scale reduction spheroids, and calcrete nodule horizons (Kraus and Middleton, 1987; Hasiotis and Dubiel, 1993b; Therrien and Fastovsky, 2000). Calcrete consists most typically of scattered (stage II) centimeter-scale nodules, which commonly display vertical stacking (rhizocretions, sensu Blodgett, 1988), mainly in the uppermost 0.5 m of the horizon. Dubiel and Hasiotis (1994b) and Hasiotis et al. (1998) noted these features and labeled paleosols of the Blue Mesa Member as Vertisols, where vertic features predominate, and Alfisols, where pale A and/or E horizons overlie reddened or purple argillic B horizons. Bown and Kraus (1987) noted, however, that the A horizon may be thin or absent in some Alfisols; we suggest, therefore, that the primary features of these paleosols is the thick B horizon, even where vertic features are common, and that they also should be considered Alfisols. The purple color that is characteristic of many Blue Mesa paleosols probably results from the coarse crystal size of the hematite in the B horizon, and indicates a high maturity of the profiles (Bown and Kraus, 1987). Kraus and Middleton (1987) noted a systematic variation in pedogenic development correlating with position on the alluvial plain; the paleosol maturity is significantly lower in the incised valley infill deposits than on the surrounding floodplain. Therrien and Fastovsky (2000) noted localized gleying in Blue Mesa paleosols, and interpreted it as poor drainage on low areas of the alluvial plain. Consistent with this observation, Heckert and Lucas (2002a) noted the local occurrence of sideritic nodules in Blue Mesa paleosols, demonstrating reducing conditions in the soil-forming environment. Additionally, we note the presence of thin (centimeter-scale) localized organic-rich layers in the epipedons at the top of Blue Mesa paleosols in Little Painted Desert County Park, north of Winslow (Fig. 1). These horizons are too thin to meet the accepted definition of histic epipedons (O horizon), but they suggest the local formation of Histosols on areas of the floodplain with impeded drainage. Kraus and Middleton (1987) described a catenary relationship in which the paleosol maturity correlates with distance from the channel (Platt and Keller, 1992; Mack and Madoff, 2005).

Calcretes are common in Blue Mesa paleosols. Therrien and Fastovsky (2000) described the presence of centimeter-scale (up to 5 cm in diameter) nodules in most Blue Mesa paleosols. These calcretes are mainly limited to horizons in which isolated nodules are abundant (stage II), although more mature (stage III) calcretes occur locally (Fig. 4C; Heckert and Lucas, 2002a). The nodules typically display alpha fabrics; they range from 2 to 8 cm in diameter, have distinct boundaries, are subspherical to irregularly shaped, and commonly display septarian cracking. These nodules occur widely scattered (stage II), but they are locally concentrated in discrete horizons (stage II to incipient stage III) in the middle to upper part of the B horizon of profiles, where they may be associated with pedogenic slickensides; they are particularly prominent immediately below the Newspaper Rock Bed and at the top of the Blue Mesa Member beneath the Rainbow Forest Bed of the Sonsela Member. Numerous locations occur in southern Petrified Forest National Park where the uppermost Blue Mesa strata consist of dark bluish-gray mudstone displaying vertic fractures and centimeter-scale calcareous nodules, in some locations with diffuse boundaries (Fig. 4D). Drab root traces are locally abundant in the mudstone host. These observations are consistent with the interpretation that the paleosols represent mainly Alfisols, which typically form on well-vegetated (forested) surfaces (Bown and Kraus, 1987; Retallack, 2001).

Petrified Forest Formation: Sonsela Member and Moss Back Formation

Lithostratigraphy

The overlying sandstone-dominated Sonsela Member of the Petrified Forest Formation and the laterally equivalent Moss Back Formation consist of up to 50 m of ledge-forming litharenite sandstone and conglomerate, including both intrabasinal and extrabasinal clasts (Stewart et al., 1972; Lucas et al., 1997). Heckert and Lucas (2002a) interpreted the Sonsela Member as filling erosional scours in the underlying Blue Mesa Member, which thins beneath the unconformity to the east. The lower contact with the Blue Mesa Member is clearly erosional and has been interpreted as a regional unconformity (the Tr-4; Lucas et al., 1997). Arc-related tectonism, suggested as the cause of incision-infill cycles in the underlying Blue Mesa Member (Kraus and Middleton, 1987), may explain this unconformity; source area uplift and an increase in the local depositional gradient may have caused incision and reworking of the Blue Mesa strata prior to Sonsela–Moss Back deposition. In particular, the uppermost unit of the Sonsela, the Agate Bridge Bed, contains a significant extrabasinal component, along with a high load of reworked calcrete (type 3 deposit of Gómez-Gras and Alonso-Zarza, 2003). Alternatively, eustasy might have caused a significant base-level drop; a regional unconformity coincident with sea-level fall occurs in the middle Keuper at about the Carnian-Norian boundary (Aigner and Bachman, 1992), approximately correlative with the Tr-4 unconformity (Lucas et al., 1997; Heckert and Lucas, 2002a).

Heckert and Lucas (2002a) examined in detail the stratigraphy of the Sonsela Member in the Petrified Forest National Park and proposed that the Sonsela is composed of three subunits of mappable extent. The lowermost sandstone-dominated unit, which they designated the Rainbow Forest Bed, consists of up to 6 m of quartzarenite sandstone and conglomerate deposited by north-northeasterly flowing streams (Deacon, 1990), and locally contains abundant silicified logs of *Araucarioxylon*. The gradationally overlying Jim Camp Wash Bed consists of up to 30 m of grayish-purple to pale red bentonitic mudstone and interbedded sandstone. The uppermost unit, the Agate Bridge Bed, consists of up to almost 7 m of cross-bedded quartzarenite and sublitharenite sandstone and conglomerate. The conglomerate contains a significant proportion of intraformational clasts, including mudstone rip-ups and calcrete. Heckert and Lucas (2002a) and Lucas et al. (2003) noted that the Sonsela Member fills scours on the Blue Mesa erosional surface.

Pedogenic Features

The Jim Camp Wash Bed displays an abundance of pedogenic features including distinct horizonation, pedogenic slickensides, decimeter-scale sandstone-filled desiccation cracks, and abundant calcrete nodules and coalesced calcrete nodule layers (stage II to III). The sedimentology of the Jim Camp Wash Bed is similar to that of the underlying Blue Mesa Member in

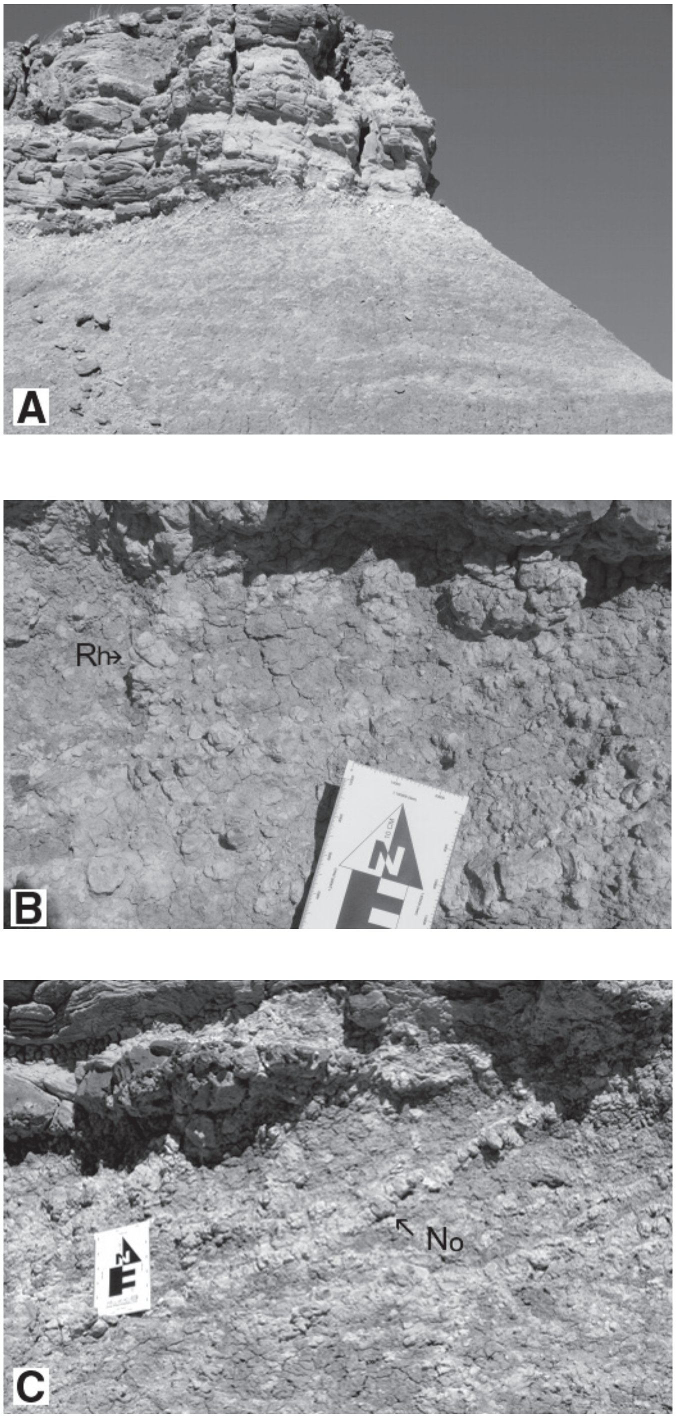

Figure 5. Jim Camp Wash Bed section. (A) The top of the section is the lower portion of the Agate Bridge Bed and here consists of ~2 m of mainly planar cross-bedded sandstone with intraformational conglomerate (calcrete nodule) lag. The calcrete described in the text occurs immediately below the base of the sandstone. (B) Calcrete morphologies include rhizocretions (vertically stacked calcrete nodules; Rh) with downward-tapering arrangement. (C) Coalescing nodule horizons (No; stage III) along arcuate surfaces are typically one to two nodules thick and up to 1 m long.

that meter-scale channels are incised into muddy floodplain deposits on which mature paleosols formed. Jim Camp Wash paleosols differ from those of the Blue Mesa Member, however, in their generally redder color and the typically greater maturity of calcrete.

Excellent Jim Camp Wash Bed paleosol outcrops occur in southern Petrified Forest National Park. Near the Rainbow Forest Museum, the uppermost Jim Camp Wash Bed consists of grayish-purple mudstone that hosts abundant pale nodules ranging in diameter from 0.5 to 3 cm (Fig. 5A). In the uppermost 40 cm of the mudstone, the nodules form coalesced horizons that are laterally discontinuous, extending a maximum distance of 50 cm, and botryoidal masses up to 30 cm long. Near the top of this zone, nodules form vertically stacked bodies, or rhizocretions, up to 20 cm long (Fig. 5B). Smaller nodules commonly exhibit crosscutting burrows. Notably, nodules are concentrated along arcuate curviplanar surfaces, presumably pedogenic slickensides, that extend laterally up to 1 m (Fig. 5C). These features have relatively smooth upper surfaces and are bounded below by a single layer of coalesced nodules projecting downward. The surfaces dip varying directions and form pseudoanticlinal intersections. Discrete fragments of charcoal, recognizable by a silky, fibrous luster, also occur in the uppermost 40 cm of the mudstone. The size and abundance of nodules decrease markedly in the reddened mudstone below the grayish-purple horizon. Nodules occur to a depth of 50 cm within this zone. The entire mudstone section is overlain by sandstone displaying planar cross-beds and conglomerate lags (Fig. 6). The lag deposits are composed of mudstone rip-ups and micritic nodules.

The Jim Camp Wash Bed consists of thick floodplain mudstone sequences overlain by a high bedload stream deposit containing a lag of reworked calcrete nodules. The previously described mudstone represents a floodplain paleosol in which the uppermost horizon (epipedon) has been partially removed by erosion. The shallow soil was subject to pronounced biotic (rooting and burrowing) and vertic activity, the latter facilitated by the smectitic nature of the soil material. A distinct Bk horizon forms the uppermost preserved horizon, containing discrete and coalesced nodular masses of micritic mudstone. The vertical orientation of some nodular masses (rhizocretions) indicates a profound influence of plant roots in the formation of at least some nodules. Vertic fractures, which can form rapidly (Birkeland, 1984), comprise arcuate surfaces with downward-shallowing dips; these served as pathways for the downward translocation of calcium carbonate, as evidenced by the formation of laterally continuous horizons of coalesced nodules along these surfaces. Although this paleosol profile is truncated, we designate it a calcic Alfisol. We note, however, that the morphology of the calcrete displays significantly greater maturity than is generally present in the Blue Mesa paleosols. This disparity suggests that conditions for carbonate accumulation in the Bk horizon were enhanced during this depositional interval.

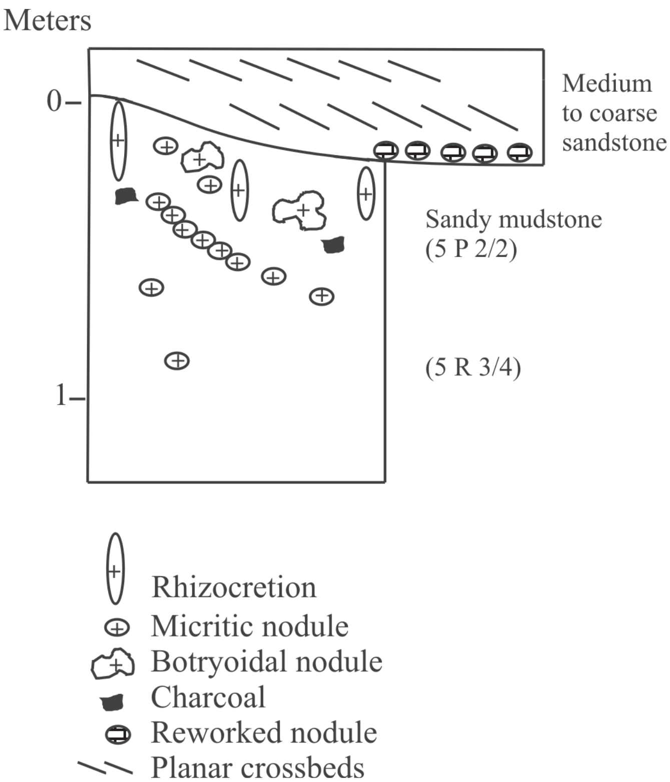

Figure 6. Representative measured section of the truncated paleosol in the Jim Camp Wash Bed described in the text and shown in Figure 5A. The section is located in southern Petrified Forest National Park along the park road (near the Flattops).

Figure 7. Features of Painted Desert Member deposition. (A) Channel-fill sequence between the arrows consists of 3 m of (mostly) red mudstone with (minor) interbedded sandstone displaying inclined heterolithic strata. This section is located in the northern Petrified Forest National Park (near Lacey Point). (B) Locally, the Black Forest Bed consists of up to 3 m of intraformational (primarily calcrete nodules) conglomerate. This section is located in the northern Petrified Forest National Park (at Lacey Point). Scale (hiking staff) is 120 cm.

Petrified Forest Formation: Painted Desert Member

Lithostratigraphy

The Painted Desert Member, of early to middle Norian age, overlies the Sonsela–Moss Back strata and consists of grayish-red and reddish-brown mudstones and thin interbedded sandstones (Lucas et al., 1997). As mapped by Stewart et al. (1972), thickness of the entire Petrified Forest Formation ranges from just over 30 m at its northeastern limit in eastern Utah, increasing southward to over 400 m at its southeastern extent in northwestern New Mexico. Like the Blue Mesa Member, the Painted Desert Member consists of thick mudstone intervals incised by channels with a fine-grained fill, locally displaying prominent inclined heterolithic strata (Fig. 7A) and levee complexes. Alternating with these dominantly suspended-load channel fills are sandstones that are predominantly multistoried and characteristically display tabular to lenticular sets of trough and planar cross-beds (Espegren, 1985). At Petrified Forest National Park, Heckert and Lucas (2002a) recognized correlatable sandstone units within the Painted Desert Member. These are, in ascending order, the Flattops, Lithodendron Wash, and Black Forest Beds. The last of these has a considerable volcaniclastic content. Riggs et al. (1994) reported a U-Pb age of 207 ± 2 Ma for zircon from the Black Forest Beds, and Riggs et al. (2003) later obtained a maximum age of 213 ± 1.5 Ma, although the authors conceded a possible age as young as 209 Ma. The Painted Desert Member sandstones generally lack extrabasinal clasts, but locally the Black Forest Bed, and to a lesser extent, the Lithodendron Wash Bed, contain thick lag deposits of calcrete-dominated intraformational conglomerate (Fig. 7B; Heckert and Lucas, 2002a).

Pedogenic Features

Painted Desert paleosol profiles display A horizons that are thin to absent and B horizons that are typically several meters thick, brick red, and locally display pedogenic slickensides, burrows, drab root traces, and rhizoliths. Calcrete nodules with alpha fabrics are commonly scattered in these thick B horizons (stage

II), but more mature calcretes (stage III and rare stage IV) occur immediately below the high bedload stream deposits. Notably, the Black Forest Bed contains a gravel conglomerate load that is up to 2 m thick in places and consists entirely of calcrete nodules (Fig. 7B).

Excellent exposures of Painted Desert paleosols occur in northern Petrified Forest National Park, particularly in exposures along Lithodendron Wash. Below the Lithodendron Wash Bed of Heckert and Lucas (2002a), the reddish-brown mudstone hosts stage II to stage III calcrete horizons and lenticular bodies of conglomerate (Fig. 8A), which consist of centimeter-scale calcrete nodules and mud chips and pedogenic slickensides. Calcrete nodules in stage II horizons are up to 5 cm in diameter; the nodules commonly exhibit a reduced (drab) interior cut by sparry calcite veins, or circumgranular cracking (crystallaria; Fig. 8B), and some nodules are penetrated by thin (1 mm diameter) burrows. Vertical stacking of nodules (rhizocretions) occurs locally (Fig. 8C). Gray reduction spheres and small drab root traces are also common in the mudstone host. Similar features are present in numerous arroyos that cut Painted Desert strata north of Cameron, Arizona, where rare laminar (stage IV, or K) horizons occur beneath sandstone beds (Fig. 8D).

Dubiel and Hasiotis (1994b) and Hasiotis et al. (1998) described Painted Desert paleosols as Vertisols, largely on the basis of abundant pedogenic slickensides and the presence of illuviated clay on ped surfaces. This designation may be appropriate in instances where no other significant pedogenic features occur, but weak horizonation is present in much of the mudstone-dominated section. Bown and Kraus (1987) noted that Alfisols may display profiles in which the A horizon may be thin or absent,

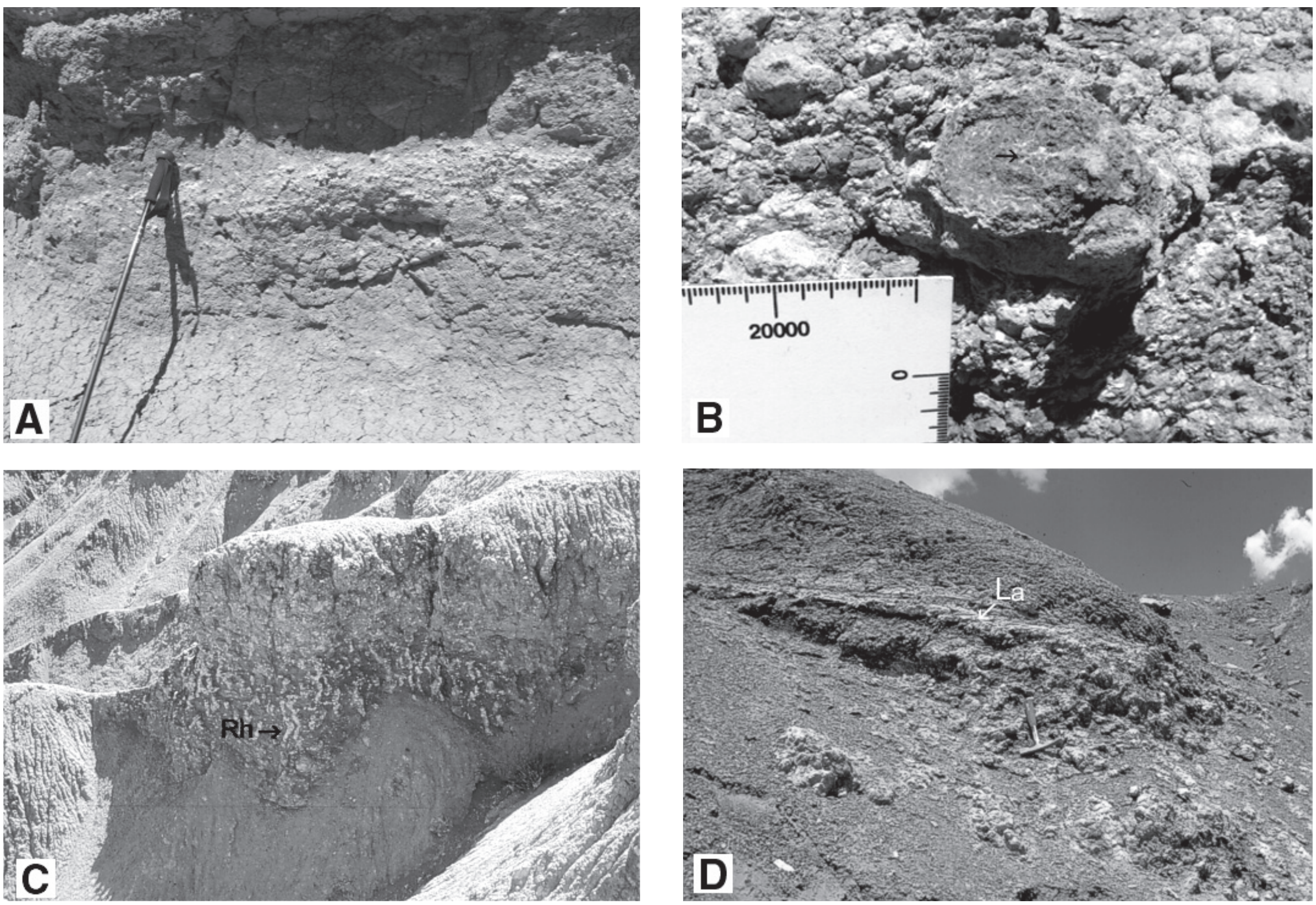

Figure 8. Painted Desert Member pedogenic features. (A) Lens of calcrete nodule and mud-chip conglomerate occurs at the level of the hiking staff handle (staff is 120 cm). The sandstone bed just above the staff handle is the Lithodendron Wash Bed. The mudstone below the calcrete lens has a wedge-shaped ped structure formed by intersecting pedogenic slickensides. (B) Detail of calcrete nodule from lens in (A) illustrating crosscutting crystallaria (sparry calcite veins; arrow). (C) Bk horizon in uppermost Painted Desert strata consists almost entirely of rhizocretions (Rh). The rhizocretion indicated by the arrow is 10 cm long. This section is located is located in the northernmost Petrified Forest National Park (Chinle Mesa). (D) Rare laminar (La; stage IV) calcrete horizon in Painted Desert. Irregular and vertically elongate nodule masses are up to 50 cm long. The base of the Bk horizon is gradational and extends to a depth of 1.5 m below the laminar horizon (hammer is 26 cm long). Location is north of Cameron, Arizona (arroyo near RR 6731).

and the B horizon, which may be thick and brick red in color, also may be calcareous and display vertic features. Therefore, we identify those Painted Desert paleosols that are not dominated by vertic features, and which display weak horizonation, as immature Alfisols.

Owl Rock Formation

Lithostratigraphy

The overlying Owl Rock Formation consists of up to 150 m of interbedded mudstones, sandstones, and limestones of approximately middle Norian age. These strata crop out in northern Arizona, northwestern New Mexico, and southern Utah (Stewart et al., 1972; Lucas and Huber, 1994; Lucas et al., 1997). Dubiel and Good (1991) noted that the contact between the Owl Rock Formation and the underlying Painted Desert Member of the Petrified Forest Formation appears disconformable, and is marked in many places by the presence of a thick intrabasinal conglomerate composed mainly of reworked calcrete clasts and locally abundant unionid bivalves. The upper part of the formation is characterized by distinctive submeter scale beds of ledge-forming limestone. Earlier workers (Blakey and Gubitosa, 1983; Dubiel, 1989, 1993) described these as lacustrine limestones and interpreted them as deposits of a large lacustrine system centered on the Four Corners region. Other workers, however, recognized pervasive pedogenic fabrics in these beds and suggested that they represent mature (stage III and IV) calcretes and palustrine carbonates (Lucas and Anderson, 1993; Lucas et al., 1997; Tanner, 2000).

Pedogenic Features

Previous examination of the Owl Rock Formation, particularly at the type section near Kayenta, Arizona, revealed distinct differences between the upper and lower strata in the types of pedogenic features present (Tanner, 2000). Thick mudstone beds in the lower part of the formation lack distinctive horizonation but host meter-scale stage II to III calcrete (Bk) horizons that display alpha and beta fabrics. The upper Owl Rock Formation hosts limestones that display brecciated to peloidal fabrics, pisoliths, spar-filled circumgranular cracks, root channels, and rare calcite pseudomorphs after gypsum. These beds are laterally gradational with limestones of limited lateral extent that display rare charophyte debris, oscillation ripple lamination, desiccation polygons, and burrowing. Tanner (2000) interpreted the brecciated beds as palustrine limestones, formed by deposition of carbonates in ponds or wetlands on a sediment-starved floodplain that was subjected to intense pedogenesis during base-level fluctuations (Platt, 1992; Platt and Wright, 1992; Armenteros et al., 1997; Alonso-Zarza, 2003). Chert is locally abundant in the brecciated limestones, but lacks the fabrics associated with Magadi-type chert formation and so is interpreted as a secondary replacement feature from groundwater (Schubel and Simonson, 1990; Bustillo, 2001).

Examination of the formation at numerous localities (e.g., the type section near Kayenta, in the Echo Cliffs, at Little Painted Desert County Park, and near Lukachukai, Arizona) has yielded additional details on Owl Rock pedogenic features. At the southern end of the Echo Cliffs, the contact with the Petrified Forest Formation is marked by 5 m of plane-bedded intrabasinal conglomerate and sandstone (Fig. 9A). The conglomerate is composed mainly of reworked calcrete nodules, with a lesser contribution of mudstone and chert pebbles. Lower Owl Rock calcretes (stage II to III Bk to Bkm horizons) are up to 5 m thick, with upper and lower gradational contacts in brown mudstone (Fig. 9B), and they display both alpha and beta fabrics (Fig. 9C). Alpha fabric calcretes comprise micritic nodules that have distinct boundaries and are crosscut by sparry veins. These calcretes are stage II to IV, and they exhibit obvious lateral gradations between stages over distances of hundreds of meters. Paleosols with Bk horizons displaying gradational tops probably represent cumulate paleosols in the sense that continual addition of sediment to the top of the profile gradually caused an upward shift in the depth of carbonate accumulation. Lateral gradations between stages of calcrete development are undoubtedly related to position on the floodplain (i.e., channel proximity), as described previously for the Blue Mesa paleosols. The lower Owl Rock paleosols lack the horizonation and obvious evidence of translocated clays that typifies the paleosols in the underlying formations, making their classification by traditional (i.e., Soil Conservation Service, 1999) soil orders problematic. The nomenclature of Mack et al. (1993), however, allows assignment of these paleosols to the order Calcisol. Some mudstones in the upper Owl Rock, however, exhibit pronounced horizonation, displaying ochric epipedons and pale albic horizons overlying reddened Bt/Bk horizons (Fig. 9B). These paleosols are interpreted as calcic Alfisols. Mudstone beds at various levels in the formation are penetrated by sandstone-filled cylinders that are up 60 cm long and up to 30 cm in diameter (Fig. 9D). These have been interpreted previously as decapod burrows (Dubiel, 1993), but the downward-branching shapes of many of these features leaves little doubt that at least some are instead the casts of deep roots, probably the tap roots of monopodial vegetation.

Many of the ledge-forming calcareous beds in the upper Owl Rock Formation have abrupt contacts and scoured bases with tens of centimeters of relief (Figs. 10 and 11A). These beds commonly overlie mudstone with a platy to prismatic fabric and millimeter- to centimeter-scale calcrete nodules and rhizocretions. These ledge-forming beds are generally greenish-gray to mottled pink-green (on fresh surfaces), and they contain pisoliths, floating siliciclastic grains, root penetration structures, and locally abundant chert (Figs. 11B and 11C). Many of these ledges form multistoried bodies and contain mud-chip lag deposits that are commonly removed by weathering in outcrop. Notably, these beds generally have a massive fabric and lack the distinctly brecciated texture and extensive root penetration that is typical of the limestones at the type section (Figs. 11C and 11D; Tanner, 2000; Alonso-Zarza, 2003). The features we describe here are consistent with an origin as groundwater calcretes; they represent fluvial channel bodies that were pervasively cemented by calcite in

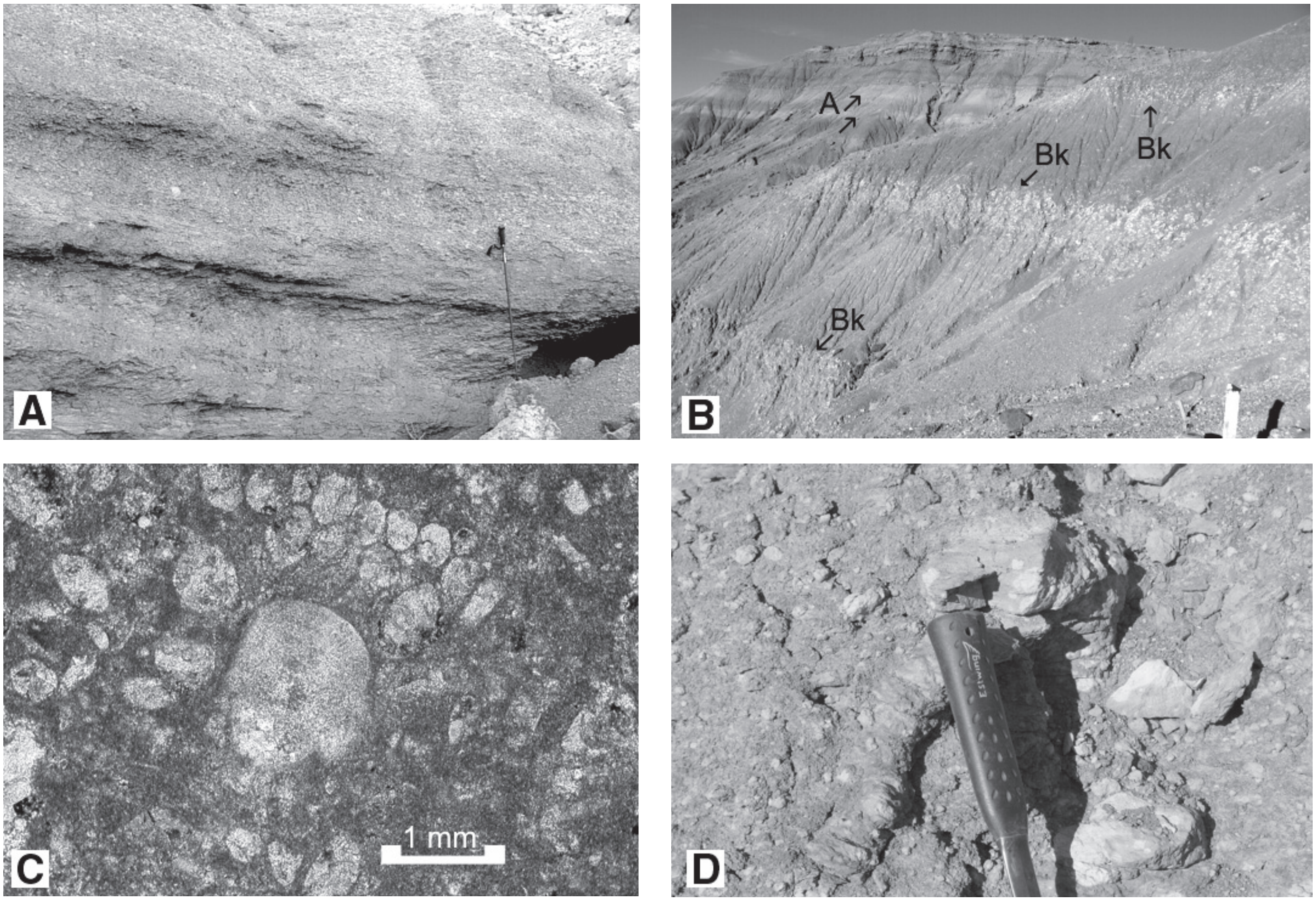

Figure 9. Pedogenic features of the Owl Rock Formation. (A) In the southern Echo Cliffs, the contact between the Owl Rock Formation and the underlying Petrified Forest Formation is marked by a 5 m bed of conglomerate, composed mainly of calcrete nodules and interbedded sandstone lenses. The base and top of the bed are just below and above the field of view in this photograph. The handle of the staff (scale is 120 cm) rests against a conglomerate layer overlying a sandier lens. (B) Overview of the Owl Rock Formation at the south end of the Echo Cliffs. Three Bk horizons with stage II calcrete are displayed. The lowermost has an abrupt top and gradational base, while the others have gradational tops and bases. Stratigraphically higher (to the left), Alfisol (A) profiles with A, E, and B horizons are visible. (C) Rare calcified root-cell structures are visible in thin sections prepared from stage II calcretes in B. (D) Calcareous sandstone cylinders with a twisting and branching morphology are common in the Owl Rock Formation (location at south end of Echo Cliffs). End of hammer handle is 4 cm wide.

the shallow subsurface and lack many of the features of subaerial exposure and desiccation displayed by palustrine limestones (Wright and Tucker, 1991; Alonso-Zarza, 2003).

Rock Point Formation

Lithostratigraphy

Across the Four Corners area, the upper Norian to (possibly) Rhaetian Rock Point Formation is recognized as the youngest stratigraphic unit of the Chinle Group (Lucas, 1993; Lucas et al., 1997). The contact between the Rock Point Formation and underlying Owl Rock Formation is unconformable (the Tr-5 unconformity). The Rock Point Formation, termed the Rock Point Member of the Wingate Sandstone by Stewart et al. (1972), includes strata formerly assigned to the Church Rock Member of the Chinle Formation (Stewart et al., 1972; Dubiel, 1989; Lucas et al., 1997). Strata of this formation consist of up to 300 m of mainly interbedded brown to red, nonbentonitic mudstones and laminated to rippled sandstones (Stewart et al., 1972; Dubiel, 1989; Lucas et al., 1997). The Rock Point Formation grades vertically to the eolian-dominated Wingate Formation (Fig. 12A) of Rhaetian to Hettangian age (Harshbarger et al., 1957; Tanner et al., 2002; Molina-Garza et al., 2003). Much of the formation consists of sandstone and siltstone sheets that display low-amplitude (eolian) ripple lamination. Other lithofacies present include tabular to sheet sandstones with small-scale sets of high-angle trough cross-beds; erosive-based, wedge-shaped sandstones with planar cross-beds and trough cross-beds and ripple translatent strata; and ripple-laminated to massive mudstones. These lithofacies represent deposition on eolian sand sheets (small-scale dunes

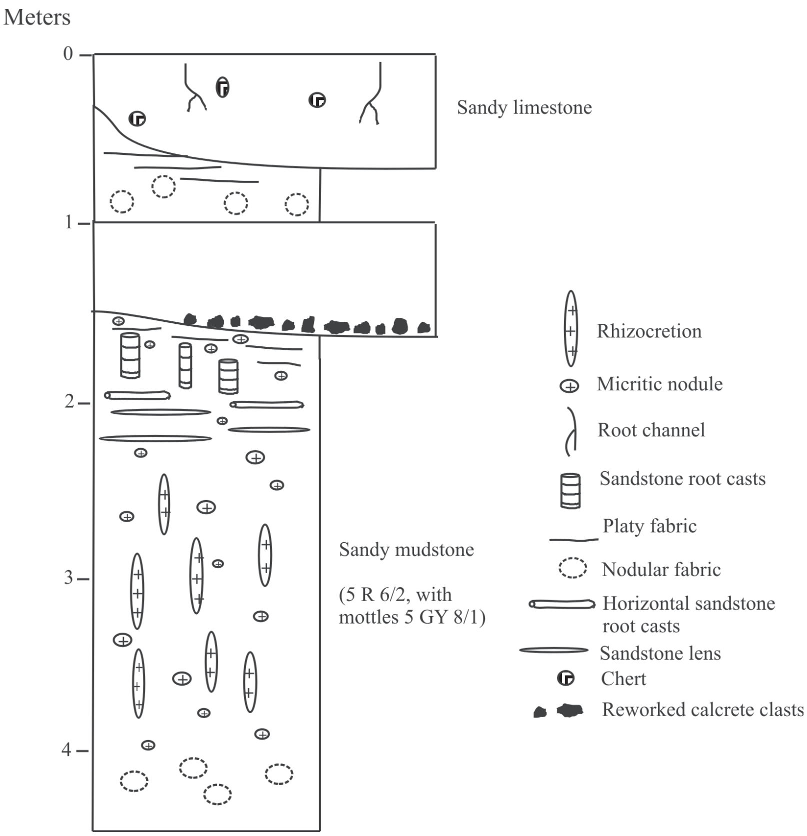

Figure 10. Measured section of paleosol features in upper Owl Rock Formation described from location near Little Painted Desert County Park.

and ripples), on mudflats, in ephemeral lakes, and in ephemeral streams (Stewart et al., 1972; Blakey and Gubitosa, 1984; Dubiel, 1989; Lucas et al., 1997).

Pedogenic Features

In many locations in the Four Corners, the Rock Point Formation displays abundant and various burrows and root traces, but lacks other well-developed pedogenic features. In northeastern Arizona, for example, near the type section for the formation, the red sheet sandstones and coarse mudstones facies that characterize the formation in this area display bedding-parallel burrows and shallow desiccation cracks in some beds, but lack extensive nodular horizons or vertic features. In other areas of the Colorado Plateau, however, much more extensive pedogenesis is evident. Rock Point calcretes are most mature in upland areas; for example, at Colorado National Monument (near Grand Junction), coarse mudstones and very fine-grained sandstones that are age-equivalent to the Rock Point Formation (Lucas et al., 1997; Tanner, 2003a) rest unconformably on granitic basement and are overlain by sandstones of the Wingate Formation. The strata near the top of this section host multiple pedogenic horizons that display drab root traces, desiccation cracks, and stage II to III calcretes (Tanner, 2003a). Pedogenic features in correlative strata north of Durango, Colorado, include desiccation cracks and drab root traces, both of which extend tens of centimeters, crumb and blocky mudstone fabrics, rhizocretions, and stage II to III calcretes in which beta fabrics are common (Fig. 12B; Blodgett, 1988; Tanner, 2003a). Blodgett (1988) interpreted the nodule-bearing horizons in the sheet sandstones of the Dolores Formation as calcareous paleosols of the order Aridisol or Inceptisol, lacking epipedons and argillic horizons. These profiles also could be classified as Calcisols (sensu Mack et al., 1993), an interpretation that can be applied to the paleosols at Colorado National Monument as well. Root traces and rhizocretions are evidence that these soils were vegetated by plants with long monopodial root systems. Calcrete conglomerate lenses in Rock Point mudstones (Fig. 12C) provide evidence of local erosional reworking of the depositional surface.

PALEOCLIMATE SYNTHESIS

Colorado Plateau

Pedogenic processes may be controlled to a large extent by climate, but soil development also depends very much on the rate of sediment accumulation, as paleosol maturity is inversely related to sedimentation rate (Bown and Kraus, 1987). Therefore, any interpretation of the paleoclimatic significance of pedogenic features also must examine changes in depositional rate. This,

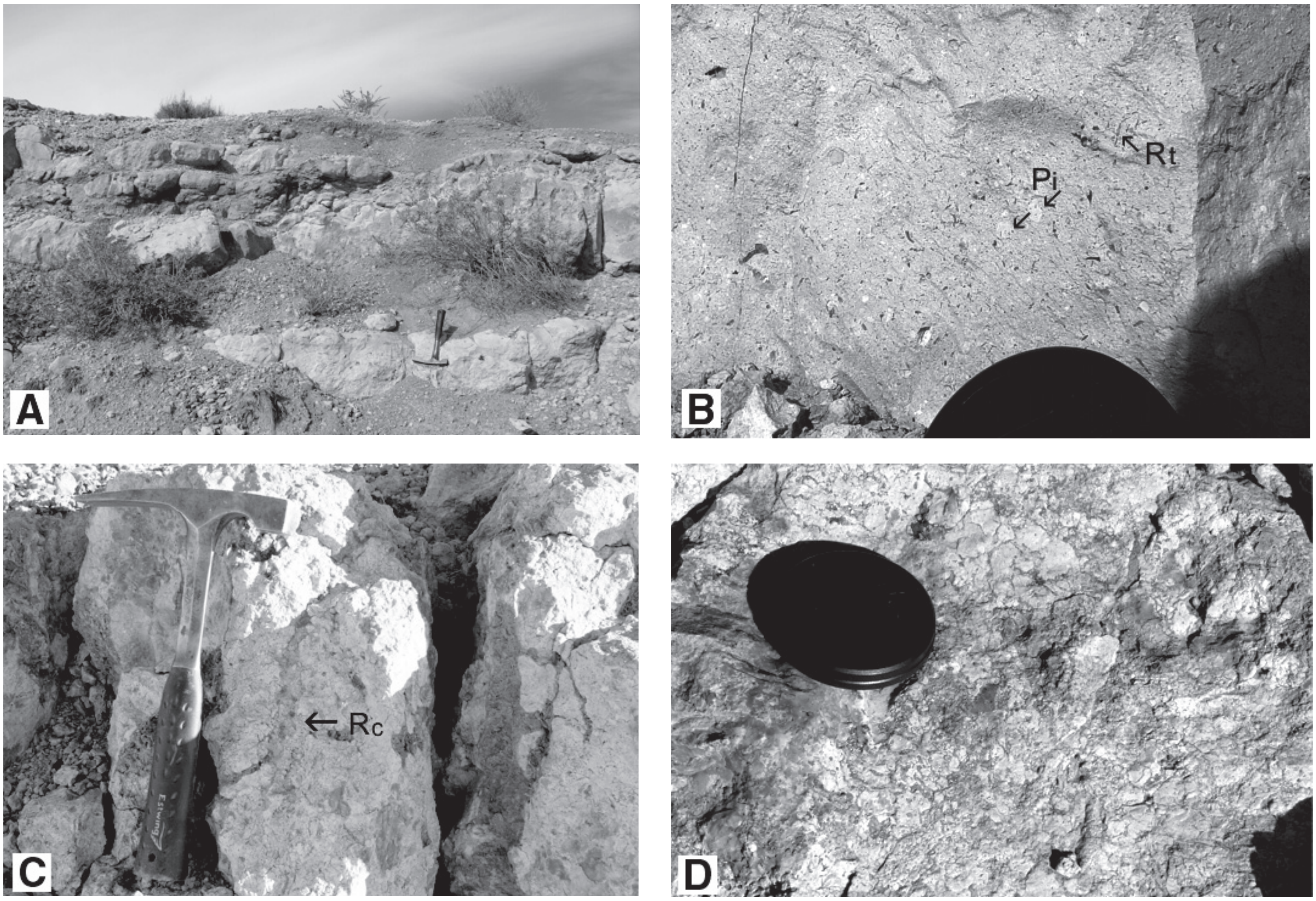

Figure 11. Features of upper Owl Rock Formation ledge-forming beds. (A) Sandy limestone beds have irregular (erosional) bases, a multistory architecture, and exhibit significant lateral thickness variations. Location is north of Little Painted Desert County Park (north of Winslow, Arizona). (B) The fabric of the beds in A is massive, but with numerous coated grains and pisoliths (Pi) and fine sparry veins that may represent root tubules (Rt). Lens cap for scale is 55 mm. (C, D) Limestone beds in section at southern end of the Echo Cliffs display pronounced brecciation fabrics, root channeling (Rc), and extensive chert replacement. Hammer in C is 26 cm; lens cap in D is 55 mm.

in turn can be forced by such extrinsic factors as tectonics and eustasy, both of which may affect base level (Possamentier et al., 1988; Blum and Price, 1998; Possamentier and Allen, 1999). Initial accumulation of Chinle sediment, during the Carnian stage, was limited to paleovalley systems incised in the Moenkopi (Tr-3) surface (Stewart et al., 1972; Blakey and Gubitosa, 1983). Middle Triassic base-level fall and subsequent Late Triassic rise matches the eustatic record of Haq et al. (1987), therefore a eustatic control on alluvial sedimentation is postulated here. The incised paleovalleys and associated tributaries had paleorelief of tens of meters, and so deposition of the Zuni Mountains, Shinarump, and the lowermost strata of the Cameron–Monitor Butte–Bluewater Creek Formations was limited to these topographic lows and was thin to absent between (Stewart et al., 1972; Blakey and Gubitosa, 1983; Demko et al., 1998).

Previous workers (Dubiel and Hasiotis, 1994a, 1994b; Hasiotis et al., 1998) have interpreted a humid but seasonal climate during the late Carnian in the Colorado Plateau region on the basis of the gleyed (or pseudogleyed) and illuviated paleosols in the Zuni Mountains and Shinarump Formations. Demko et al. (1998), however, cautioned that the paleoclimate record of the basal Chinle is biased by deposition within paleovalleys underlain by aquicludes of the Moenkopi Formation, which resulted in artificially high water tables. Indeed, although the prominence of gley features in these paleosols suggests high humidity, the presence of pedogenic slickensides and a prismatic fabric in paleosols in these formations indicates that these soils were allowed to dry completely at times, perhaps seasonally. Numerous authors have commented on the evidence for a strongly seasonal distribution of precipitation during the Late Triassic resulting from a monsoonal effect, both from field studies and from climate models (Robinson, 1973; Parrish and Peterson, 1988; Crowley et al., 1989; Dubiel et al., 1991; Parrish, 1993; Crowley, 1994; Wilson et al., 1994; Pires et al., 2005). This effect presumably was a

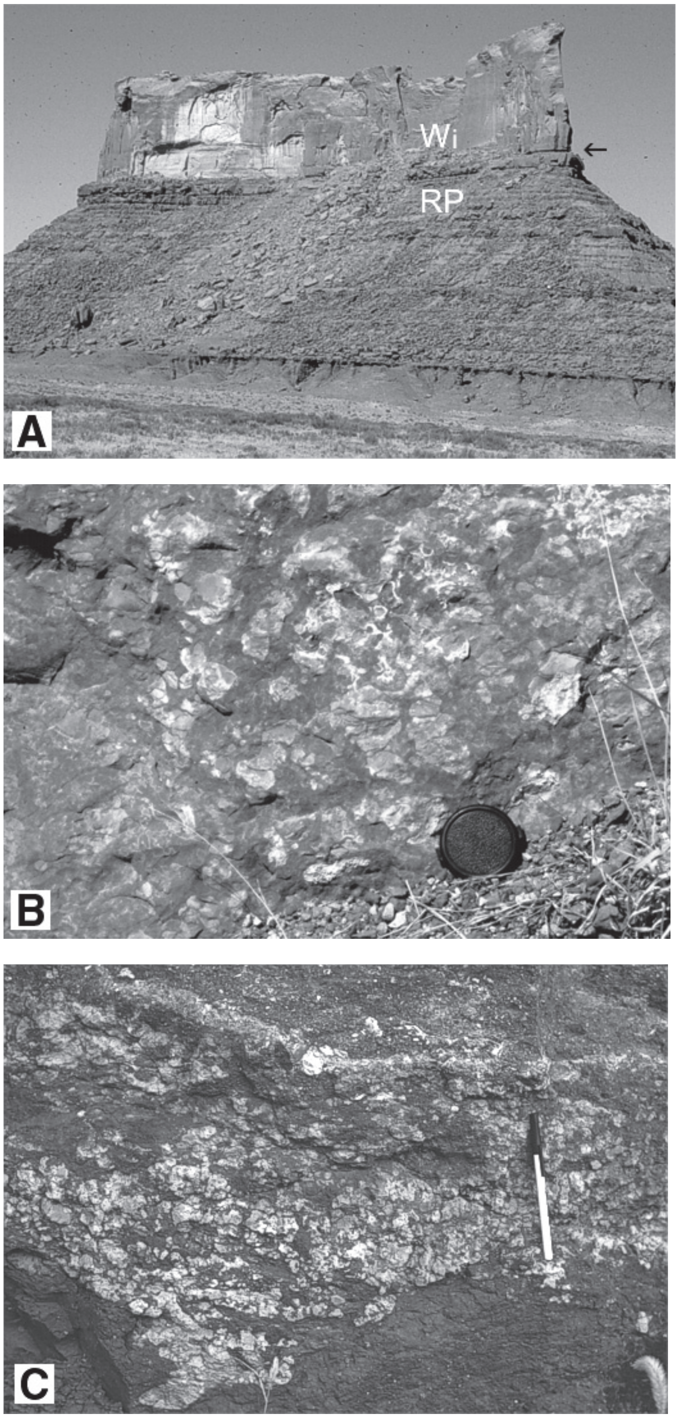

Figure 12. Features of the Rock Point Formation. (A) The section at Little Round Rock, near the type section at Rock Point, Arizona, demonstrates the lithologic transition from interbedded sheet sandstone and mudstone of the Rock Point Formation (RP) to the eolian sandstone–dominated Wingate Formation (Wi). Approximate contact is indicated by the arrow. The entire visible section is ~70 m thick. (B) Rock Point strata near Durango, Colorado, contain stage II calcrete (stage II) consisting primarily of rhizocretions. Lens cap for scale is 55 mm. (C) Calcrete conglomerate lens in mudstone host, in Rock Point strata west of Moab, Utah (San Rafael Swell area).

consequence of the configuration of the Pangean land mass and its position straddling the equator. We speculate that Late Triassic migration and rotation of the continent caused weakening of the monsoon in northern Pangea, resulting in regional drying.

The prominence of deep tap roots and/or crayfish burrows is also consistent with a strongly fluctuating water table. The presence of spodic horizons and the dominantly kaolinitic composition of the clays, however (Tanner, 2003a; Tabor et al., 2004), which are not present in overlying formations, is a clear indication of humidity and strong weathering and translocation of soil materials. Modern Spodosols are generally (but not exclusively) associated with forested regions, typically coniferous, and humid climates. (Birkeland, 1984; Retallack, 2001). Thus, an overall humid to subhumid climate is likely during Carnian deposition of the Zuni Mountains and Shinarump Formations, with high water tables enforced seasonally by the position of the lower Chinle strata in paleovalleys and the locally impermeable nature of the underlying Moenkopi strata. Initial deposition of the Cameron–Monitor Butte–Bluewater Creek Formations took place under similarly humid but seasonal conditions. The observed gleying, or psuedogleying, and desiccation are both consistent with soil development under conditions of strongly fluctuating water tables, suggesting a greatly variable (possibly seasonal) distribution of precipitation under subhumid climate conditions, as described for the paleosols in the underlying Zuni Mountains and Shinarump Formations.

Subsequent accumulation of younger Chinle strata (Cameron–Monitor Butte–Bluewater Creek Formations and the Blue Mesa Member of the Petrified Forest Formation) was widespread and shows no constraint from underlying paleotopography. Paleosols of these strata are primarily vertic Alfisols that are calcic (stage II to III calcretes) in some locations, gleyed, in others, varying by location on the alluvial plain. By analogy to modern soils, this classification implies that the soils formed in woodlands and forests in subhumid to semiarid climates (Birkeland, 1984; Buol et al., 1997; Retallack, 2001). As Bown and Kraus (1987) noted, the presence of gley features does not preclude the formation of calcrete, which can form rapidly in clay-rich paleosols where translocated clays retard the downward movement of meteoric waters. The abundance of pedogenic slickensides and pseudoanticlines in these paleosols further suggests a seasonal, semiarid climate (Therrien and Fastovsky, 2000). The maturity (i.e., horizonation) of the Blue Mesa floodplain paleosols is notable, attesting to a low rate of sediment accumulation. The long residence time of soil materials permitted the effective translocation of oxides and clays within the profiles and the formation of clearly delineated horizons. The general (but not complete) absence of spodic and histic horizons in these strata and the nonkaolinitic composition of the clays signal a decrease in precipitation near the end of the Carnian stage.

Regardless of exact cause, Sonsela–Moss Back deposition marks a significant change in the pattern of alluvial sedimentation in the Chinle basin as the slowly aggrading, high-suspended-load, high-sinuosity stream systems were succeeded by mainly high bedload, low-sinuosity streams. Paleosols in the Jim Camp Wash Bed, however, are mature Alfisols, similar in aspect to those of the Blue Mesa Member, with the exception of greater calcrete maturity (up to stage IV). Therefore, climate was likely more arid (but not greatly so) at the end of the Carnian, during Sonsela deposition, than during Blue Mesa deposition.

Subsequent deposition of the Painted Desert Member during the early to middle Norian took place in a flood basin in which the fluvial style varied markedly; deposition by high-suspended load, high-sinuosity streams was punctuated by episodes of deposition by high-bedload, low-sinuosity streams. Base-level changes may be responsible, but the cause of these changes, i.e., eustasy, climate, or tectonism, is unknown. Paleosols that formed on the floodplains of the high-sinuosity channels are generally vertic Alfisols or Vertisols (sensu Mack et al., 1993) that display only limited translocation of clays and horizonation (Dubiel and Hasiotis, 1994b; Hasiotis et al., 1998). These paleosols display less maturity than do the Blue Mesa or Sonsela paleosols, probably reflecting faster rates of sediment accumulation on the floodplain. Painted Desert calcrete horizons, however, are typically more mature (stage II and III, and rare stage IV) than in the Blue Mesa Member, and so likely reflect more arid conditions, as interpreted for the Jim Camp Wash paleosols. Therrien and Fastovsky (2000) noted that gleying is much less common in the upper (Painted Desert Member) than in the lower (Blue Mesa Member) Petrified Forest Formation, and that Bk horizons are much more prominent. Zuber and Parnell (1989) noted that the clay mineral assemblage in the Painted Desert Member is dominated by mixed-layer illite-smectite, in contrast to the predominantly smectitic mudstones of the Blue Mesa Member. They interpreted this composition as the result of pedogenic illitization of smectitic clays in an alkaline environment in which precipitation was highly seasonal. Retallack (2001), however, viewed claims of illitization in the soil-forming environment with skepticism; the significantly less bentonitic composition of the Painted Desert mudstones may be explained instead by interformational differences in the original clay mineralogy of the sediment load.

At least locally, initial Owl Rock deposition is marked by the infilling of lows incised into the underlying Painted Desert strata by thick sequences of intrabasinal conglomerate, mainly reworked calcrete. Subsequent Owl Rock depositional settings are composed of low- to high-sinuosity streams and floodplain muds, on which calcic Alfisols and Calcisols (sensu Mack et al., 1993) formed. Alluvial deposition alternated with sedimentation in carbonate ponds and marshes that were modified subsequently by pedogenesis. Tanner (2000) interpreted the Owl Rock sequence as consisting of alternating episodes of floodplain aggradation and degradation caused by changes in base level; incision and pedogenesis of highstand mud and carbonate deposits occurred during episodes of base-level fall that may have been climatically induced, similar to the model of climatically forced sequence boundaries of Tandon and Gibling (1997). Although the concept of Owl Rock deposition in large lakes has been dismissed, episodes of high base level are implied by the presence of palustrine and minor lacustrine carbonates (Alonso-Zarza, 2003). Palustrine carbonates may form under climates that range from subhumid to semiarid (Platt and Wright, 1992; Tandon and Andrews, 2001), with drier conditions indicated by the presence of pronounced brecciation fabrics and coated grains, as seen in the Owl Rock Formation. The occurrence of well-developed (stage III and IV) calcrete horizons and Alfisols in the intervening mudstones is consistent with this interpretation of semiaridity.

A pronounced unconformity (Tr-5) separates the Owl Rock and the overlying Rock Point Formations. As noted by Tanner (2003b), Rock Point deposition marked a change in basin configuration that appears to reflect the rise and migration of a forebulge. Rock Point sediments display only weakly developed Aridisols (or Calcisols sensu Mack et al., 1993), probably reflecting a relatively constant influx of sediment. The abundant evidence of eolian deposition and frequent desiccation, however, indicates that deposition took place in a semiarid to arid climatic setting. Dubiel et al. (1991) interpreted the interval of Rock Point deposition as the driest of the Late Triassic. The abundance of faunal bioturbation, however, indicates episodes of significant surface moisture, potentially a consequence of fluctuating water tables, and, locally, the depositional surface was well-vegetated, as indicated by rhizoliths and beta fabrics. This interpretation is consistent with the dominant sedimentary bedform of eolian sand sheets in the Rock Point Formation; sand sheets are an interdunal facies characteristic of wet eolian systems (Lancaster, 1993). Continued aridification during the Rhaetian and Hettangian is clearly indicated by the dominance of eolian and playa sedimentation during deposition of the Moenave and Wingate Formations, as the Wingate erg formed over the Four Corners area.

In sum, evidence from sedimentary facies and paleosols indicates that the climate on the Colorado Plateau was drier during the Norian-Rhaetian than during the Carnian, confirming the interpretation of Blakey and Gubitosa (1984). Dubiel et al. (1991) and Parrish (1993), however, interpreted the same sedimentary evidence as indicating a moist climate until the very end of the Triassic (at least through the Norian). Notably, Parrish (1993) predicted that a strong monsoonal effect would produce abundant moisture in the western equatorial region, which included the Colorado Plateau. Presumably, weakening of the monsoon would have resulted in insufficient strength to draw moisture from the west and aridification of the western equatorial region. Therefore, we must consider the possibility that a weakening monsoon at the start of the Norian caused the observed drying in the Four Corners region.

Global Climate

Overall warm and dry conditions during the Late Triassic are indicated by the abundance of evaporite and carbonate deposits and the restriction of coal formation to high latitudes (Frakes et al., 1992; Lucas, 1999). Indeed, Colbert (1958) first proposed gradual aridification and associated changes in floral patterns during the Late Triassic to explain tetrapod turnover. The configuration of the Pangean continent undoubtedly had a significant effect in controlling this climate (Robinson, 1973). Specifically, the arrangement of land areas likely resulted in a dry climate belt covering a broad region of western and central Pangea at low to mid-paleolatitudes, a consequence of the shrinkage of the humid intertropical convergence zone (ITCZ) and the weakening of zonal circulation. This interpretation received considerable support from early computer modeling exercises (Parrish et al., 1982; Kutzbach and Gallimore, 1989).

A trend of Late Triassic aridification similar to that of the Colorado Plateau is indicated by facies changes, evaporite occurrences, and paleosols in the Upper Triassic to Lower Jurassic formations of the Newark Supergroup (Olsen, 1997; Kent and Olsen, 2000). For example, Norian-age formations contain more mature calcrete paleosols than do Carnian formations in the southern basins, as in the Deep River and Taylorsville basins (Coffey and Textoris, 1996; LeTourneau, 2000; Driese and Mora, 2003). To the north, in the Newark, Hartford, and Fundy basins, the absence of evaporite-bearing or eolian facies in formations of Carnian age and their presence in formations of Norian age demonstrate a similar trend of aridification (see Olsen, 1997, for review). This climate trend in the Newark Supergroup strata, however, has been interpreted as a consequence of the latitudinal drift of eastern North America by 5° to 10°, which carried the basins from a moist subtropical to a more northerly arid climate zone (Olsen, 1997; Kent and Olsen, 2000). Parrish (1993) postulated that aridification on the Colorado Plateau took place during the Early Jurassic as a consequence of the weakening of monsoonal circulation, but we suggest that this aridification took place earlier, during the Norian, as indicated herein. The weakening monsoon, potentially controlled by the northerly drift of Laurasia, resulted in strengthening of zonal circulation and allowed the latitudinal drift of the Colorado Plateau between climate zones (Parrish, 1993).

Similar trends of Late Triassic aridification are seen in facies transitions in other locations globally, as in the succession of the Timezgadouine and Bigoudine Formations in the Argana basin, Morocco (Olsen, 1997; Hofmann et al., 2000), the facies changes in the Upper Triassic Mercia Mudstone Group of England (Talbot et al., 1994; Ruffell and Shelton, 1999), and the Keuper of the Germanic basin (Aigner and Bachmann, 1992). Simms et al. (1994) observed a Late Triassic change in clay mineral assemblages in European sedimentary successions, notably a loss of kaolinite, similar to that observed on the Colorado Plateau. These authors attributed the replacement of the pteridosperm flora by a coniferous, gingko, and fern flora at the Carnian-Norian boundary, to climate change. Not all areas of Pangea became drier during the Late Triassic, however. Extensive coal deposits formed in Australia and China, which became wetter at this time (Fawcett et al., 1994). The growth of large lakes in the Jameson Land basin of eastern Greenland during the Late Triassic is interpreted similarly as a consequence of increasing humidity caused by northward drift of the basin to a humid, temperate climate zone (Clemmensen et al., 1998).

CONCLUSIONS

Paleosols and pedogenic features preserved in the formations of the Chinle Group record a trend of gradual aridification during the Late Triassic. The prominence of gleying in the kaolinitic, bioturbated paleosols of the mottled strata, Shinarump, and basal Cameron formations suggests that climate during the late Carnian was subhumid to humid, and that water tables fluctuated seasonally. High water tables during deposition of these formations may have resulted from their position within paleovalleys incised into the underlying Moenkopi Formation strata. Improved soil drainage during Cameron and Blue Mesa deposition is interpreted from the presence of thick argillic profiles interpreted as Alfisols. This condition may have resulted either from climatic drying or from the position of these soils stratigraphically higher above the Moenkopi Formation. Increasing aridity during early Norian deposition of the Painted Desert Member is clearly suggested by the prominence of vertic features and immature (stage II to III) calcretes. This trend of aridification continued during middle Norian Owl Rock deposition, as indicated by mature (stage III to IV) calcretes. The Norian-Rhaetian Rock Point strata lack mature paleosol profiles, but the predominance of eolian and playa facies in this formation suggests that the trend of increasing aridity continued through onset of the Wingate erg. This trend may have been controlled by the position of the Pangean continent, which led to the weakening of monsoonal flow and the strengthening of zonal circulation.

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