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UPPER TRIASSIC LITHOSTRATIGRAPHY, DEPOSITIONAL SYSTEMS, AND VERTEBRATE PALEONTOLOGY ACROSS SOUTHERN UTAH

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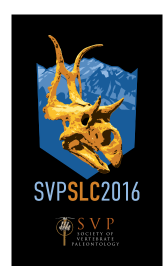

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Cover

The Chinle Formation section above the Shinarump Member near the Circle Cliffs overlook. The Wingate Sandstone forms the jagged cliffs that cap the Chinle.

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Upper Triassic Lithostratigraphy, Depositional Systems, and Vertebrate Paleontology Across Southern Utah

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ABSTRACT

The Chinle Formation and the lower part of the overlying Wingate Sandstone and Moenave Formation were deposited in fluvial, lacustrine, paludal, and eolian environments during the Norian and Rhaetian stages of the Late Triassic (~230 to 201.3 Ma), during which time the climate shifted from subtropical to increasingly arid. In southern Utah, the Shinarump Member was largely confined to pre-Chinle paleovalleys and usually overprinted by mottled strata. From southeastern to southwestern Utah, the lower members of the Chinle Formation (Cameron Member and correlative Monitor Butte Member) thicken dramatically whereas the upper members of the Chinle Formation (the Moss Back, Petrified Forest, Owl Rock, and Church Rock Members) become erosionally truncated; south of Moab, the Kane Springs beds are laterally correlative with the Owl Rock Member and uppermost Petrified Forest Member. Prior to the erosional truncation of the upper members, the Chinle Formation was probably thickest in a southeast to northwest trend between Petrified Forest National Park and the Zion National Park, and thinned to the northeast due to the lower Chinle Formation lensing out against the flanks of the Ancestral Rocky Mountains, where the thickness of the Chinle is largely controlled by syndepositional salt tectonism. The Gartra and Stanaker Members of the Ankareh Formation are poorly understood Chinle Formation correlatives north of the San Rafael Swell. Osteichthyan fish, metoposaurid temnospondyls, phytosaurids, and crocodylomorphs are known throughout the Chinle Formation, although most remains are fragmentary. In the Cameron and Monitor Butte Members, metoposaurids are abundant and non-pseudopalatine phytosaurs are known, as is excellent material of the paracrocodylomorph *Poposaurus*; fragmentary specimens of the aetosaurs *Calyptosuchus*, *Desmatosuchus*, and indeterminate paratypothoracisins were probably also recovered from these beds. Osteichthyans, pseudopalatine phytosaurs, and the aetosaur *Typothorax* are especially abundant in the Kane Springs beds and Church Rock Member of Lisbon Valley, and *Typothorax* is also known from the Petrified Forest Member in Capitol Reef National Park. Procolophonids, doswelliids, and dinosaurs are known but extremely rare in the Chinle Formation of Utah. Body fossils and tracks of osteichthyans, therapsids, crocodylomorphs, and theropods are well known from the lowermost Wingate Sandstone and Moenave Formation, especially from the St. George Dinosaur Discovery Site at Johnson Farm.

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INTRODUCTION

The Triassic Period was one of the most critical intervals in the history of tetrapod evolution—a time when terrestrial ecosystems recovered from the Permian extinction with a spectacular adaptive radiation that included early dinosaurs, crocodylian-like archosaurs, and mammals (e.g., Sues and Fraser, 2010; Nesbitt, 2011). At the end of the Triassic, a globally variable array of terrestrial ecosystems in which different amniote groups enjoyed varying degrees of success (Ezcurra, 2010; Sues and Fraser, 2010) gave way to uniformly dinosaur-dominated ecosystems as a result of complex environmental changes that are only beginning to be understood (e.g., Rowe and others, 2010; Irmis, 2011; Irmis and others, 2011; Olsen and others, 2011; Parker and Martz, 2011). Understanding the nature and timing of these changes requires detailed analysis of lithostratigraphy, depositional systems, and biostratigraphy (e.g., Parker and Martz, 2011, 2017; Martz and Parker, 2017).

The Upper Triassic Chinle Formation, which is exposed across the Four Corners states and southern Nevada (figure 1), provides a remarkable record of environmental and biotic change during the last 25 million years of the Triassic Period, from roughly 230 Ma to sometime before 201.3 Ma (Irmis and others, 2011; Ramezani and others, 2011, 2014; Atchley and others, 2013; Martz and others, 2014). Under current calibrations of the Late Triassic time scale (Furin and others, 2006; Muttoni and others, 2010; Schoene and others, 2010; Hüsing and others 2011), this places the beginning of Chinle deposition very close to the beginning of the Norian, and the end of Chinle deposition during the Rhaetian (e.g., Olsen and others, 2011; Atchley and others, 2013).

The sandstone and mudstone beds of the Chinle Formation were deposited primarily by braided and meandering fluvial systems, associated lacustrine and paludal environments, and in the upper part of the section, rare eolian dunes (e.g., Blakey and Gubitosa, 1983, 1984; Dubiel, 1987a, 1987b, 1989; Demko and others, 1998, 2005; Dubiel and Hasiotis, 2011). Throughout the Norian and Rhaetian, several fluvial aggradational cycles (FACs) were deposited within the Chinle Formation in response to tectonically driven uplift of the early Mesozoic Cordilleran arc to the southwest (figure 1), and associated basin subsidence (Trendell and others, 2012; Atchley and others, 2013; Howell and Blakey, 2013), and/or to variations in discharge and sediment transport related to climate change (Dubiel and Hasiotis, 2011).

The Chinle Formation was deposited in a subtropical climate with distinct seasonality (e.g., Dubiel and others, 1991; Dubiel and Hasiotis, 2011). Paleosols and depositional systems indicate that lakes and swamps were common in western North America early in the Norian, but that the climate became increasingly hot and arid after 215 Ma during the middle-late Norian and Rhaetian. This climate shift may have been due to western North America moving from the tropics into the more arid mid-latitudes, or to the development of a rain shadow resulting from the growing Cordilleran arc accompanied by an associated increase in atmospheric CO_2 (Blakey and Gubitosa, 1983, 1984; Dubiel and others, 1991; Cleveland and others, 2008; Dubiel and Hasiotis, 2011; Atchley and others, 2013; Nordt and others, 2015).

The overlying dune and interdune deposits of the Wingate Sandstone and fluvial deposits of the stratigraphically equivalent Moenave Formation were probably initially deposited during the Rhaetian (Donohoo-Hurley and others, 2010; Kirkland and others, 2014a; Suarez and others, 2017). The Triassic-Jurassic boundary, which is well constrained radioisotopically at 201.3 Ma (Schoene and others, 2010; Blackburn and others, 2013), has long been accepted to lie within the Moenave Formation (Molina-Garza and others, 2003; Cornet and Waanders, 2006; Donohoo-Hurley and others, 2006; Kirkland and Milner, 2006; Tanner and Lucas, 2007; Downs, 2009; Kirkland and others, 2014a). Recent radioisotopic dating places it within the Dinosaur Canyon Member of the Moenave Formation (Suarez and others, 2017). This controversial topic has been discussed in detail in Milner and others (2012), in Kirkland and others (2014a), and in Suarez and others (2017).

The bulk of work on the lithostratigraphy and depositional systems of the Chinle Formation has been done in southern Utah and northern Arizona. The lithostra-

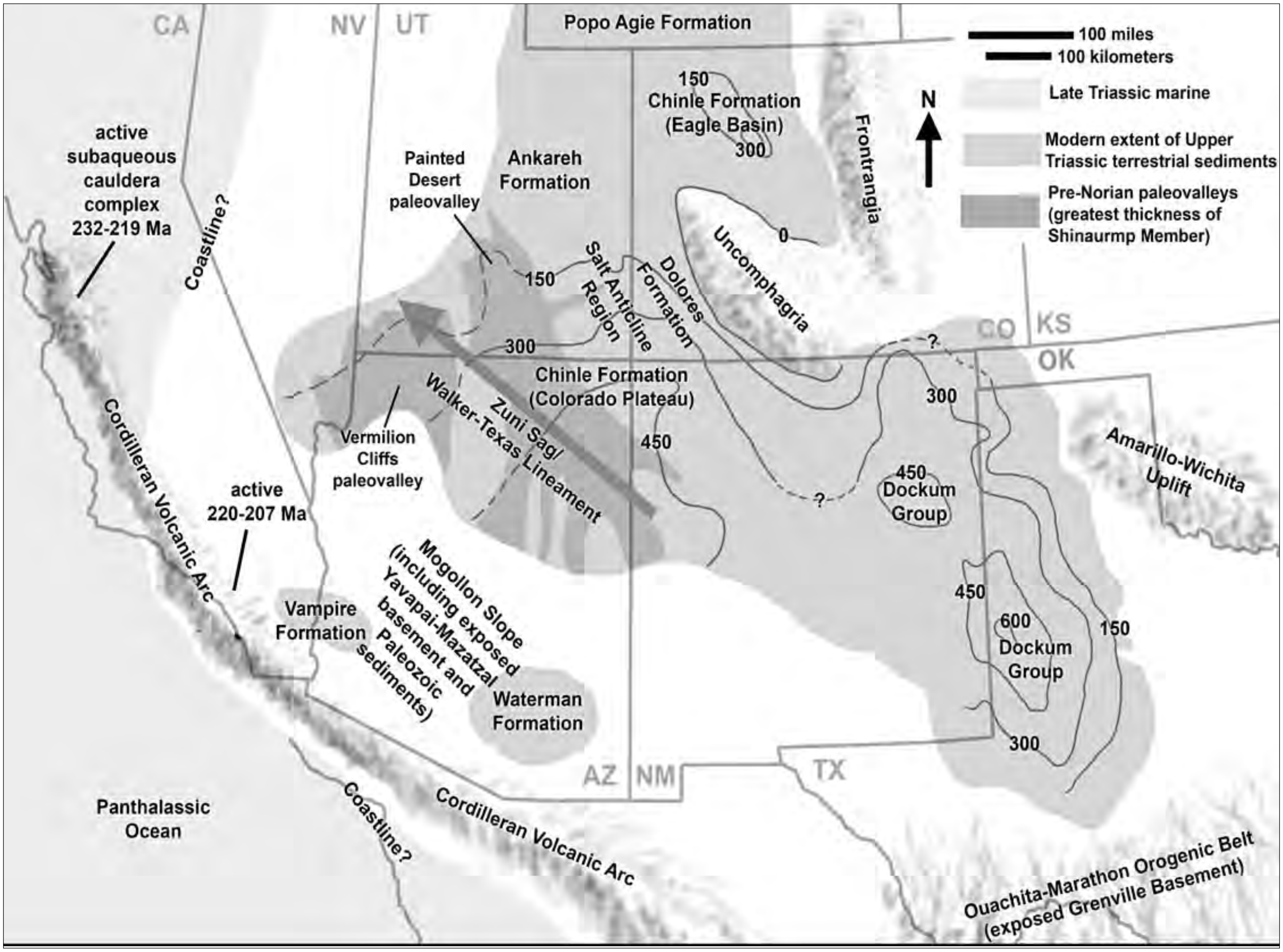

Figure 1. Late Triassic tectonic and geographic controls on Chinle Formation deposition. Extent of Chinle Formation sediment cover, Shinarump paleovalleys, and isopach lines representing thickness (in meters) after Blakey (2008, figures 18 and 21) and Blakey and Gubitosa, 1983, figure 8); extent of Dockum Group sediment cover and isopach lines after Lehman (1994); extent of Vampire and Waterman Formations sediment cover, distribution of Cordilleran volcanic arc, and locations of Late Triassic magmatism after Riggs and others (2013, figure 1; 2016, figure 1); other geographic features after Dickinson and Gehrels (2009, figure 7). Dashed isopach lines indicate region where upper part of Chinle Formation erosionally truncated, indicating that original thickness was probably greater (see figure 2A). Width of modern Basin and Range province compacted to approximate pre-extensional width. Arrow on Zuni Sag trend line shows predominant direction of sediment transport.

tigraphy of the Chinle Formation in southern Utah was described in detail by several studies published between the late 1950s and early 1970s (Stewart, 1956, 1957; Harshbarger and others, 1957; Stewart and others, 1959; Poole and Stewart, 1964; Wilson and Stewart, 1967; Repenning and others, 1969; O'Sullivan, 1970; Stewart and others, 1972a; O'Sullivan and MacLachlan, 1975; O'Sullivan and Green, 1977). These studies established much of the stratigraphic nomenclature that is still in use for the members within the Chinle Formation, and correlated these members across the region. The most notable of these studies is the magisterial work of Stewart and others (1972a), which synthesized previous studies and proposed detailed correlations within the Chinle Formation across the western United States. During the 1980s, Ronald Blakey, Russell Dubiel, and their colleagues published a handful of extremely influential publications presenting lithostratigraphic correlations across southern Utah that largely corroborated previous studies, and gave detailed interpretations of the depositional systems represented by different members (Blakey and Gubitosa 1983, 1984; Blakey and Middleton, 1986; Dubiel, 1987a, 1987b, 1989, 1994).

Lucas (1993; Lucas and others, 1997a) formally raised the Chinle Formation to group rank, presented somewhat different interpretations of correlations within the Chinle Formation across the Colorado Plateau, and also extended the names of several members to areas where they had not previously been used (e.g., the extension of the name Rock Point Member from northern Arizona into southern Utah to replace the name Church Rock Member). Lucas (1993) also proposed several new lithostratigraphic units within the Chinle Group, notably the Cameron Formation and the Blue Mesa and Painted Desert Members of the Petrified Forest Formation. Following most subsequent workers, we differ from Lucas (1993; Lucas and others, 1997a) by using more traditional nomenclature and rank, referring to the Chinle as a formation and its subunits as members (e.g., Dubiel, 1994; Woody, 2006; Blakey, 2008, p. 268; Martz and Parker, 2010; see Zeigler and others, 2008, for revisons in New Mexico).

Since 2000, several studies have developed increasingly detailed lithostratigraphic models for the Chinle Formation that have been used as frameworks for increasingly detailed biostratigraphic studies, permitting, in turn, ever more detailed models relating Late Triassic environmental change to biotic change (e.g., Martz and Parker, 2010; Parker and Martz, 2011; Atchley and others, 2013). These models have become increasingly well-calibrated to the geochronologic time scale by providing a detailed and accurate stratigraphic framework on which to hang the steadily growing body of data on radioisotopic dates and magnetostratigraphy (e.g., Olsen and others, 2011; Irmis and others, 2011; Ramezani and others, 2011, 2014; Atchley and others, 2013; Zeigler and others, 2017). Although the most detailed of these studies have been conducted in Petrified Forest National Park in northern Arizona (e.g., Heckert and Lucas, 2003a; Woody, 2006; Martz and Parker, 2010; Parker and Martz, 2011; Ramezani and others, 2011; Martz and others, 2012; Trendell and others, 2012; Atchley and others, 2013; Howell and Blakey, 2013), there have also been detailed stratigraphic studies of the Chinle Formation in southern Utah (e.g., Beer, 2005; Kirkland and others, 2014b; Martz and others, 2014, 2015). Moreover, our understanding of Chinle depositional and paleoclimatic changes has become enhanced through recent depositional system and paleosol studies (e.g., Therrien and Fastovsky, 2000; Tanner, 2003; Prochnow and others, 2006; Cleveland and others, 2007, 2008; Driese and others, 2010; Dubiel and Hasiotis, 2011; Loughney and others, 2011; Trendell and others, 2012, 2013; Atchley and others, 2013; Howell and Blakey, 2013).

Whereas large vertebrate collections exist from the Chinle Formation of Arizona and New Mexico (e.g., Long and Murry, 1995; Fiorillo and others, 2000; Heckert, 2004; Irmis, 2005; Irmis and others, 2007; Nesbitt and Stocker, 2008; Parker and Martz, 2011), relatively little has been published on the Upper Triassic vertebrate fauna of Utah and collection has been sporadic. Lisbon Valley in southeastern Utah has historically been the most prolific area for collecting vertebrate fossils within the Chinle Formation in Utah (Cope, 1877; Schaeffer, 1967; Morales and Ash, 1993; Lucas and others, 1997a, p. 92, 1997b; Milner, 2006; Milner and others, 2006a, 2008, 2011, 2012; Gibson, 2013a, 2013b; Martz and others, 2014; Hunt-Foster and others, 2016). Comparatively little has been published on Upper Triassic vertebrates from elsewhere in the state and most publications are post 1990 (Heckert and others, 1999; Parrish, 1999; Fraser and others, 2005; Parker and others, 2006; Gauthier and others, 2011; Schachner and others, 2011; Kirkland and others, 2014b; Martz and others, 2015). The quality of much of this material is, however, spectacular (e.g., Schachner and others, 2011; Gibson, 2013a, 2013b). Moreover, paleontological inventories of national parks conducted by the Utah Geological Survey and the National Park Service (Santucci and Koch, 2003; DeBlieux and others, 2006; Kirkland and others, 2010; Santucci and Kirkland, 2010; Madsen and others, 2012; Tweet and others, 2012; Kirkland and others, 2014b; Martz and others, 2015) suggest that the enormous paleontological potential of the Chinle Formation of Utah has barely been tapped.

This paper presents a brief synthesis of what is known of the depositional history, lithostratigraphy, and vertebrate paleontology of the Chinle Formation in southern Utah, with particular regions of interest described in more detail.

ABBREVIATIONS

AMNH: American Museum of Natural History, New York; CANY: Canyonlands National Park; CARE: Capitol Reef National Park; GLCA: Glen Canyon National Recreation Area; GSENM: Grand Staircase-Escalante National Monument; MNA: Museum of Northern Arizona, Flagstaff; PEFO: Petrified Forest National Park; SGDS: St. George Dinosaur Discovery Site at Johnson Farm, St. George; UCM: University of Colorado Museum, Boulder; UMNH: Natural History Museum of Utah, Salt Lake City; USNM: National Museum of Natural History (Smithsonian), Washington D.C.; YPM: Yale Peabody Museum of Natural History, New Haven; ZION: Zion National Park.

GEOGRAPHIC AND TECTONIC CONTROLS ON CHINLE SEDIMENTATION

Deposition of the Chinle Formation across what is now the Colorado Plateau was controlled by a combination of early Mesozoic tectonic activity and inactive remnants of late Paleozoic tectonism (figure 1), many of which followed a striking southeast to northwest trend that parallels or overlies much older Proterozoic basement lineaments with the same orientation (Baars, 1993). The Cordilleran volcanic arc along the western margin of North American continent was well-established by the Triassic (Dickinson, 2004) and was active in southwestern Arizona and eastern California during the Late Triassic (e.g., Barth and Wooden, 2006; Riggs and others, 2013). The arc and the northeasterly sloping region of southwestern Arizona flanking it (the "Mogollon Slope" sensu Bilodeau, 1986) provided important sources of Chinle detrital sediments from plutonic, volcanic, and Paleozoic sedimentary sources, and also formed the southwestern boundary of the Chinle depositional basin (Stewart and others, 1972a; Dickinson and Gehrels, 2008, 2009; Riggs and others, 2012, 2013, 2016). The Chinle basin was also bounded by largely inactive remnant late Paleozoic highlands produced by the accretion of Pangea (figure 1). In central Colorado, the Ancestral Uncomphagre Highlands (also called Uncomphagria or the Ute Highlands; e.g., Nesse, 2008; Blakey, 2008) bounded the Chinle basin to the northeast. The Ouachita-Marathon orogenic belt and Amarillo-Wichita uplift in Texas, as well as Mesoamerican upland sources, lay to the southeast of the Colorado Plateau Chinle Formation depocenter. All of these uplifts were additional sources of Chinle detritus (Stewart and others, 1972a; Riggs and others, 1996; Dickinson and Gehrels, 2008; Dickinson and others, 2010).

The Chinle depositional basin itself was an actively subsiding back-arc basin oriented southeast to northwest across the Four Corners states, roughly parallel to the Cordilleran arc itself (Lawton, 1994; Dickinson and Gehrels, 2008), and was deepest to the southwest, proximal to the arc (see below). A smaller depocenter, the Eagle Basin (Dubiel, 1992), lay between the Ancestral Uncomphagre Highlands and the more easterly located Ancestral Front Range (also called Frontrangia or the Arapahoe Highlands; e.g., Blakey, 2008; Nesse, 2008).

The Chinle Formation overlies a major disconformity wherever it occurs throughout the Colorado Plateau, the Tr-3 unconformity (Pipiringos and O'Sullivan, 1978), usually manifested in the form of southeast to northwest-trending paleovalleys of various depths carved into pre-Chinle strata (e.g., Painted Desert and Vermilion Cliffs paleovalleys in figures 1 and 2A) (Blakey and Gubitosa, 1983, 1984; Blakey, 2008). Most commonly, the Chinle Formation overlies the Early and earliest Middle Triassic Moenkopi Formation (figures 2 and 3) (Stewart and others 1972b; Blakey and others, 1993), but locally it overlies Permian strata such as the Cutler and DeChelly Formations, or even older Precambrian crystalline rocks (e.g., Blakey and Gubitosa, 1983, 1984; Blakey, 2008). As the youngest members of the Moenkopi Formation are early Anisian (early Middle Triassic; Blakey and others, 1993) and Chinle deposition began around the earliest Norian (Ramezani and others, 2011, 2014; Atchley and others, 2013), the Tr-3

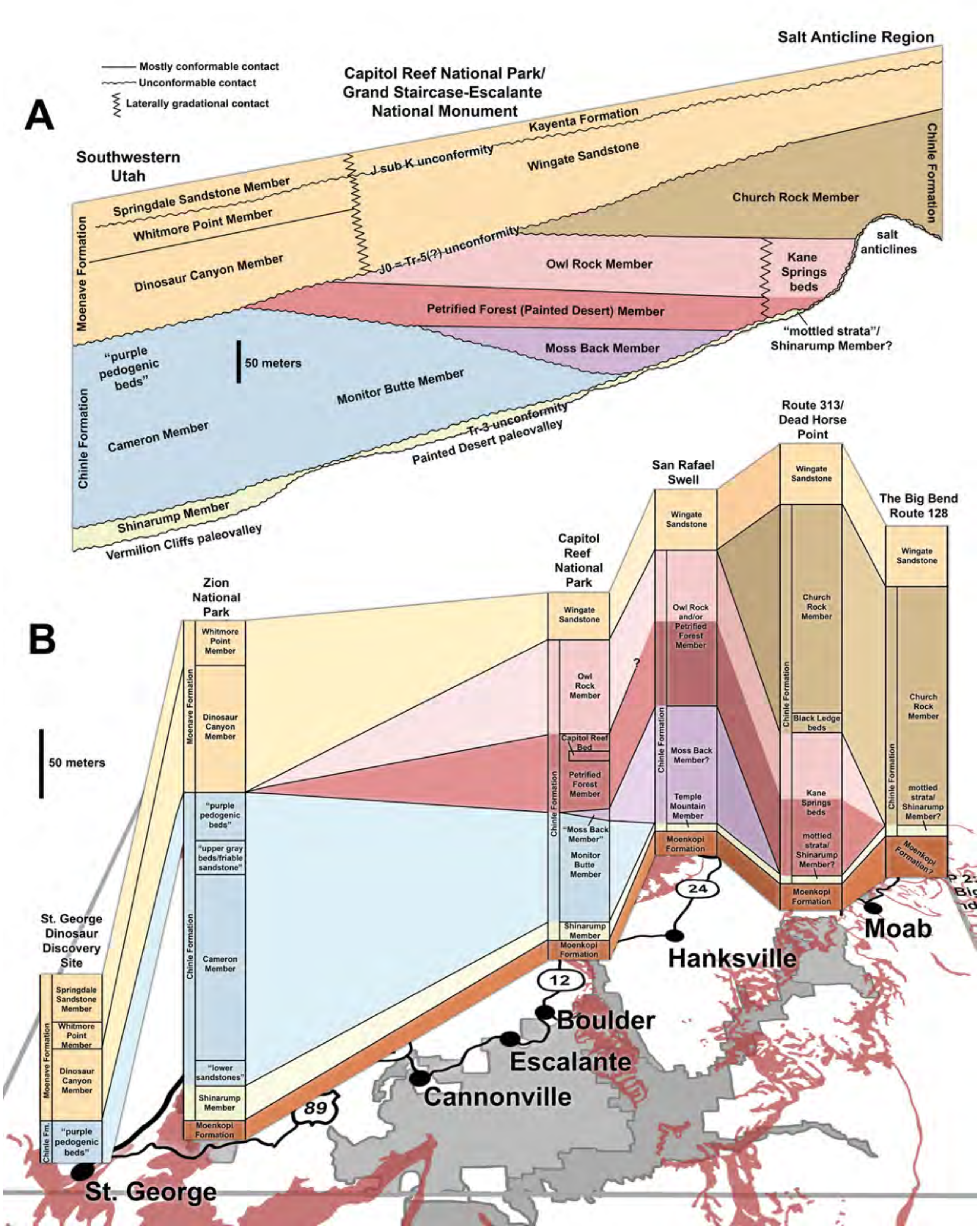

Figure 2 caption on following page.

Figure 2 (figure on previous page). Chinle Formation correlations in southern Utah. (A) Simplified and idealized stratigraphic cross section of Upper Triassic and Lower Jurassic strata from southwestern Utah near St. George and ZION to the salt anticline region near Moab. (B) Simplified stratigraphic sections for areas discussed in detail in the text. In the map beneath, Chinle Formation exposures in dark red, national parks and monuments in gray (see figure 4 for more detail).

unconformity eliminates most of the Anisian, the entire Ladinian (late Middle Triassic), and probably the entire Carnian (early Late Triassic)—a time span of nearly 20 million years (Ogg, 2012).

Paleocurrents indicate that the Chinle fluvial systems flowed predominantly from southeast to northwest, roughly parallel to the long axis of the basin (Stewart and others, 1972a; Blakey and Gubitosa, 1983; Beer, 2005), although tributaries flowed eastwards in the southern part of the state and northwards in northern Arizona (Stewart and others, 1972a; Johns, 1988; Riggs and others, 2013). The Shinarump Member, the basal unit of the Chinle Formation across much of the Colorado Plateau, is thickest in a southeast- to northwest-oriented zone just southwest of the Four Corners region (figure 1) including Canyon de Chelly in Arizona and Capitol Reef National Park (CARE) and the Circle Cliffs area of GSENM, indicating that this zone contained the deepest pre-Norian paleovalleys, the Painted Desert Paleovalley (Blakey and Gubitosa, 1983, figure 8, 1984; Blakey, 2008). The Shinarump river system confined to the Painted Desert paleovalley was fed by tributaries in smaller paleovalleys. A separate Shinarump fluvial system was confined to the northerly oriented Vermilion Cliffs paleovalley extending north from northeastern Arizona to southwestern Utah (figure 1) (Blakey and Gubitosa, 1983, figure 8, 1984; Blakey, 2008).

Several workers (Blakey and Gubitosa, 1983; Demko and others, 1998; Dubiel and Hasiotis, 2011) have suggested that units overlying the Shinarump Member, such as the Monitor Butte and Blue Mesa Members, were also deposited within these paleovalleys. However, the recent recognition that the Mesa Redondo Member is a Shinarump Member correlative (Irmis and others, 2011, supplemental data; Riggs and others, 2016) suggests that, at least in northern Arizona, only the earliest stages of Shinarump deposition were confined to paleovalleys, and that during the Norian these valleys were mostly filled to permit a more laterally continuous depositional system. This is supported by the recognition that the Monitor Butte, Bluewater Creek, Cameron, and lower Blue Mesa Members are probably correlative units overlying the Shinarump/Mesa Redondo fluvial deposits (figures 2 and 3) (Lucas, 1993; Lucas and others, 1997a; Irmis and others, 2011, supplemental data; Kirkland and others, 2014b; Martz and others, 2015). Moreover, Blue Mesa Member deposition in northern Arizona may have been controlled by expansion of a large fluvial fan extending northeasterly from the Cordilleran volcanic arc (Trendell and others, 2013), which would have extended across the areas occupied by the Vermilion Cliffs and Painted Desert paleovalleys (Blakey and Gubitosa, 1983; Blakey, 2008), further supporting the notion that these valleys had filled by the late early Norian so that laterally continuous deposition occurred across the region.

The thickness of the Chinle Formation is greatest in northern Arizona and western New Mexico (Stewart and others, 1972a, plate 1; Blakey and Gubitosa, 1983, figure 3C; Blakey, 2008, figures 16 and 18), indicating that the basin was deepest closest to the Cordilleran arc and shallower to the northeast along the flanks of the Ancestral Uncomphagre Highlands. For example, the stratigraphic thickness of the Chinle Formation in PEFO is about 400 m despite the Rock Point Member being locally absent there (Parker and Martz, 2011, figure 6), and in excess of 500 m in the southern parts of the Navajo Nation and in westernmost New Mexico where the Rock Point Member occurs (Stewart and others, 1972a). Blakey (2008, figure 18) shows the Chinle Formation thinning between northeastern and northwestern Arizona and southwestern Utah (figure 1). However, it is likely that this is a result of the top of the Chinle Formation becoming increasingly truncated by erosion (figure 2A) in the western part of the Colorado Plateau region (this region of truncation is indicated by dashed isopach lines in figure 1 (e.g., Stewart and others, 1972a, plate 2). For example, in Zion National Park (ZION) only the lower part of the Chinle Formation is preserved (figure 2) (Martz and others, 2015), but this

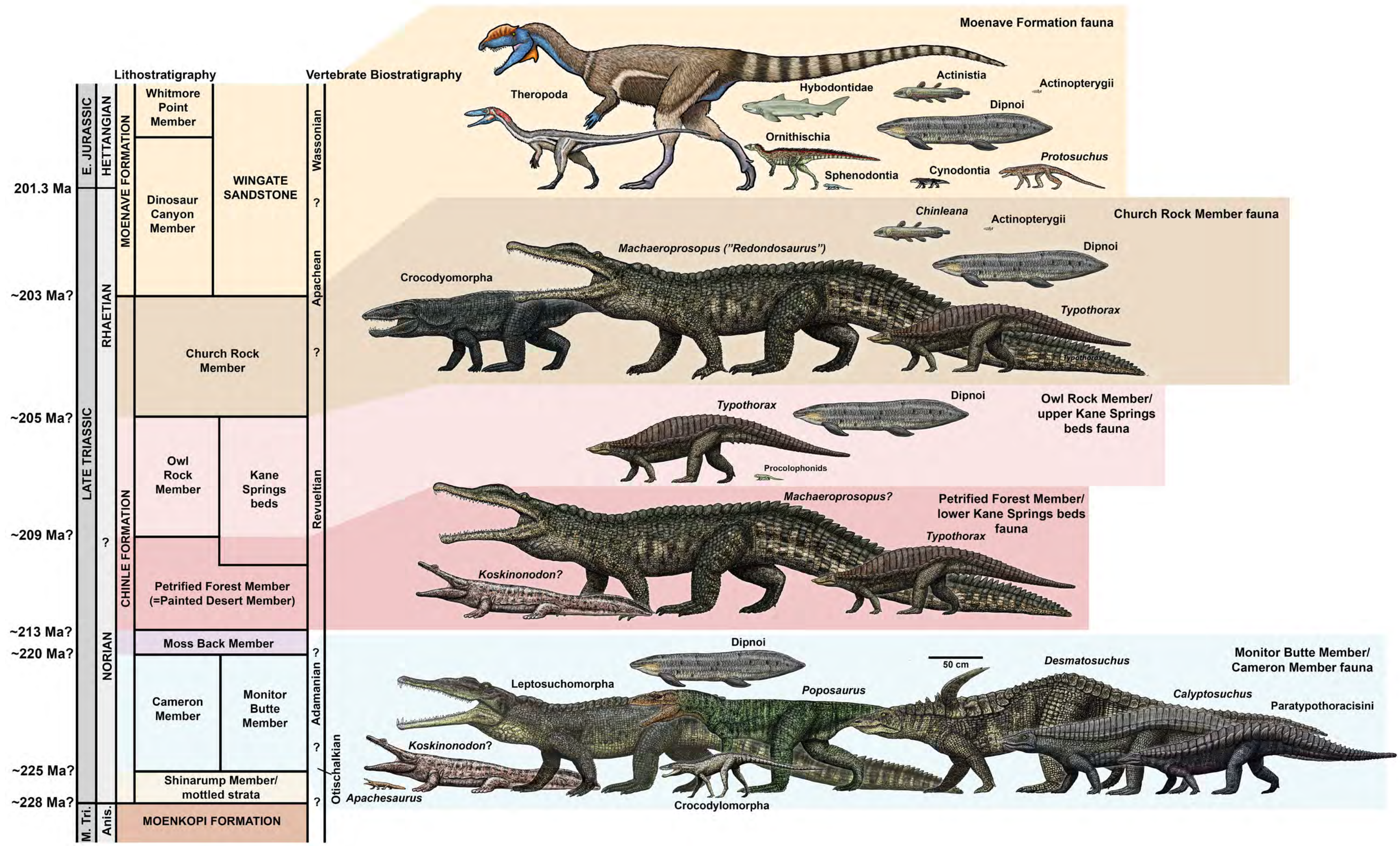

Figure 3. Vertebrate biostratigraphy of the Chinle Formation in southern Utah. Composite stratigraphic column shows correlations of all major lithostratigraphic units in southern part of state. Numerical ages and placement of biozone boundaries are exceedingly tentative and explained in text. Vertebrate taxa known from various members shown on the left; the Moenave Formation taxa are largely known from tracks.

preserved thickness is greater than the nearly complete section at CARE. This suggests that the zone of greatest thickness of the Chinle Formation originally extended along a more southeast to northwest-trending zone that cuts across the divide between the older Vermilion Cliffs and Painted Desert paleovalleys that controlled early Shinarump deposition (figure 1) (Blakey and Gubitosa, 1983).

Interestingly, this southeast- to northwest-trending zone of greatest Chinle thickness closely approximates the Zuni Sag (figure 1), a structural depression considered to have developed in the Early Jurassic due to a regional tectonic event that slightly changed the orientation of the depositional basin (Blakey, 1994). The observation that Chinle deposition occurs across the same regional trend suggests that the Zuni Sag, which overlies the southeast to northwest-trending Walker-Texas Lineament (Baars, 1993), may have actually developed during Norian time (if not before) due to existing weaknesses in the Precambrian basement.

GENERAL LITHOSTRATIGRAPHY OF THE CHINLE FORMATION IN SOUTHERN UTAH

The depositional record of the Chinle Formation in Utah (figures 2 and 4) is most complete in the southeastern part of the state, southwest of the salt anticline region of the Paradox Basin (Stewart and others, 1972a, figures 2 and 10). This area includes Chinle Formation exposures in CARE, GSENM (figure 2A), Red Canyon, and the northern part of the Navajo Nation (e.g., Blakey and Gubitosa, 1983, 1984; Beer, 2005; Kirkland and others, 2014b). Most members of the Chinle Formation in northern Arizona (where the formation is thickest) have correlatives in this region of southern Utah. The section in CARE, the Circle Cliffs region of GSENM, and Lake Powell in Glen Canyon National Recreation Area (GLCA) near the junction of the Colorado and San Juan Rivers varies from approximately 125 m to at least 170 m in thickness, generally thickening to the south (e.g., Kirkland and others, 2014b; Martz and others, 2015; J.W. Martz, unpublished data). At ZION, the preserved section is about 180 m thick (Martz and others, 2015). As discussed above, these thicknesses were probably considerably greater prior to partial erosional truncation of the Chinle Formation across the region (figure 2A).

To the north and northwest, where the basin thins on the flanks of the Ancestral Uncomphagre Highlands, the lower strata pinch out so that Chinle Formation exposures within the San Rafael Swell and salt anticline region are missing lower members such as the Shinarump and Monitor Butte Members (figure 2A) (e.g., Stewart, 1957; Stewart and others, 1972a; Blakey and Gubitosa, 1983, 1984). However, as the upper part of the Chinle Formation in the salt anticline region is generally not strongly erosionally truncated as in CARE, GSENM, and ZION, the preserved section in the salt anticline region is not generally thinner than in sections to the west (figure 2). In Lisbon Valley and the region around Arches National Park, the Chinle Formation comprises mostly the Church Rock Member and locally the Kane Springs beds, and the total formation is generally 110 to 200 m thick (e.g., Martz and others, 2014; J.W. Martz, unpublished data). The region just to the southwest of the Ancestral Uncomphagre Highlands overlies the Pennsylvanian evaporite deposits of the Paradox Basin and was subject to active salt tectonism during the Triassic that formed southeast- to northwest-trending salt anticlines (e.g., Hazel, 1994), as it still does today.

The Upper Triassic in southern Utah includes the Chinle Formation, lower part of the Wingate Sandstone, and lower part of the Moenave Formation (figure 3). The Chinle consists of several correlative members (figure 2).

Shinarump Member of the Chinle Formation

Above the Tr-3 unconformity, the base of the Chinle Formation in Utah and Arizona is usually a unit of sandstone and conglomerate called the Shinarump Member (Gilbert, 1875), which is composed almost exclusively of extrabasinal quartz, quartzite, chert, and volcanic clasts characterized by diagenetic quartz overgrowths (Stewart, 1957; Stewart and others, 1972a; Blakey and Gubitosa, 1983, 1984; Dubiel, 1987a, 1987b). In some parts of northern Arizona, there is a reddish mudstone-dominated unit with interbedded siliceous sandstone and conglomerate, the Mesa Redondo Member (Cooley, 1958), which is probably correlative to the Shi-

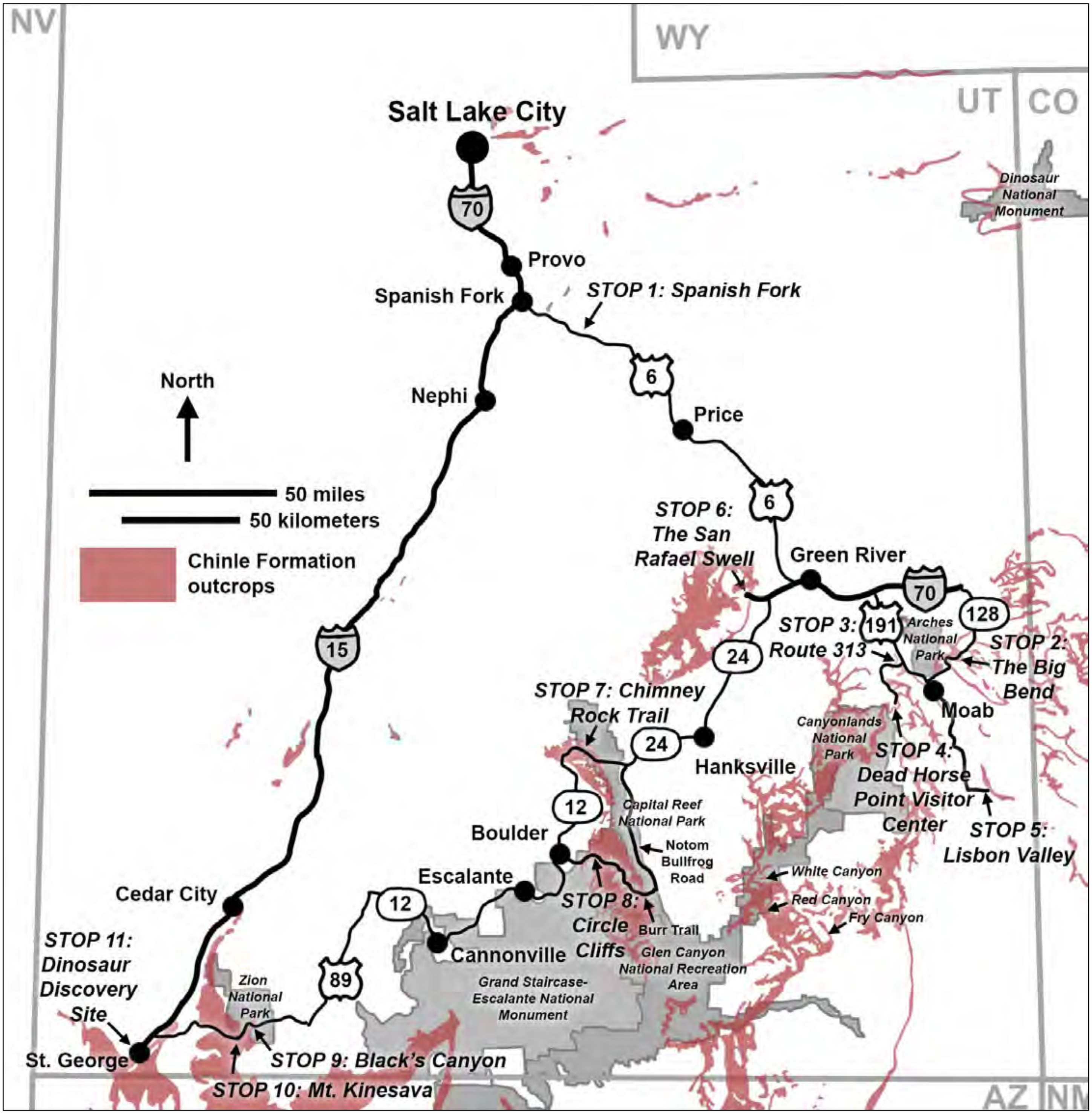

Figure 4. Field trip route and major stops. Chinle Formation outcrops shown in red, national parks and monuments shaded in gray.

narump Member (Irmis and others, 2011, supplemental data). The Mesa Redondo Member has been radioisotopically dated at 228 to 225 Ma, or earliest Norian (early Lacian) (Ramezani and others, 2014), which is therefore presumably the age of the correlative Shinarump Member. The Shinarump Member was deposited by northwest-flowing braided rivers that filled pre-Norian paleovalleys incised into the Moenkopi Formation and pre-Triassic units (Blakey and Gubitosa, 1983, 1984; Blakey, 2008).

Mottled Strata of the Chinle Formation

In most places, both the Shinarump and Mesa Redondo Members are overprinted by a distinctive mottled paleosol that sometimes extends into the uppermost Moenkopi Formation, and is variously referred to as the mottled member, mottled strata, purple mottled unit, Temple Mountain Member, or Zuni Mountain Formation (Robeck, 1956; Poole and Stewart, 1964; Stewart and others, 1972a; Dubiel, 1987a, 1987b, 1992, 1994; Heckert and Lucas, 2003b; Dubiel and Hasiotis, 2011). This mottling marks a gleyed paleosol formed by fluctuating water tables, and is comprised of patches of multi-colored iron oxide interspersed with iron-depleted patches, and often densely packed with crayfish burrows (Tanner and Lucas, 2006; Dubiel and Hasiotis, 2011). A slightly different pattern occurs in CARE and the Circle Cliffs area of GSENM, where the top of the Shinarump Member is eroded and sometimes entirely excised (Stewart and others, 1972a; Dubiel, 1987a, 1987b; Demko, 2003; Beer, 2005; Kirkland and others, 2014b), and the mottled strata seem to be restricted to below the base of the Shinarump Member (Kirkland and others, 2014b; J.W. Martz, unpublished data). In regions where the mottled strata overprint the Shinarump Member (Dubiel, 1987a, 1987b; Dubiel and Hasiotis, 2011; Irmis and others, 2011, supplemental data; Atchley and others, 2013; Martz and others, 2014), formation of this paleosol likely occurred during a depositional hiatus that followed Shinarump deposition. As the uppermost Mesa Redondo Member has been dated radioisotopically at 225.185 Ma or early Norian (Lacian) (Ramezani and others, 2011), the mottled strata overprinting both the Shinarump and Mesa Redondo Members may have formed about 225 million years ago. The mottled strata should not be considered a distinct formation or member (e.g., Zuni Mountains Formation) because they represent an episode of paleosol formation overprinting pre-existing stratigraphic units, not distinct stratigraphic units in their own right.

Monitor Butte and Cameron Members of the Chinle Formation

In southeastern Utah, the Shinarump Member/mottled strata are overlain by a drab-colored, mudstone-dominated unit with interbedded sandstone and conglomerate—the Monitor Butte Member (Kiersch, 1956). The Bluewater Creek Member (Lucas and Hayden, 1989), Blue Mesa Member, and Cameron Member (Lucas, 1993) in southwestern Utah, northern Arizona, and western New Mexico may be correlative to the Monitor Butte Member (Lucas, 1993; Lucas and others, 1997a; Kirkland and others, 2014b). The Blue Mesa Member type section (Lucas, 1993) has been radioisotopically dated between about 225 to 220 Ma (Ramezani and others, 2011, 2014; Atchley and others, 2013); these dates are a reasonable estimate for the correlative Cameron and Monitor Butte Members.

The lower part of the Monitor Butte Member in CARE contains distinctly contorted strata (Dubiel, 1987a, 1991; Kirkland and others, 2014b), similar to contorted strata that occur in the lower Bluewater Creek Member in parts of northeastern Arizona and the Zuni Mountains in western New Mexico (Green, 1956; Repenning and others, 1969; Stewart and others, 1972a, p. 25−26; Ash, 1978; Blakey and Gubitosa, 1983, p. 66–68; Dubiel and others, 1993; Kirkland and others, 2014b). These contorted strata seem to have been deposited along a southeast to northwest depositional trend paralleling the Zuni Sag.

The Monitor Butte Member was deposited by meandering rivers and associated overbank lacustrine and paludal environments (Blakey and Gubitosa, 1983, 1984; Dubiel, 1987a, 1987b; Beer, 2005) indicating relatively high-precipitation subtropic climatic conditions in western North America during the Lacian (Dubiel and Hasiotis, 2011; Atchley and others, 2013; Trendell and others, 2013). The contorted strata in the lower part of the Monitor Butte Member are the result of syndepositional slumping of overloaded lacustrine or paludal deltas (e.g., Blakey and Gubitosa, 1983, 1984; Dubiel and others, 1993).

At ZION in southwestern Utah, the Cameron Member rather than the Monitor Butte Member makes up the lower part of the Chinle Formation that overlies the

Shinarump Member. The Cameron Member (including the sandstone and mudstone member of Phoenix, 1963) is considerably thicker than the Monitor Butte Member to the northeast (figure 2) (e.g., Kirkland and others, 2014b; Martz and others, 2015) and consists of drab-colored mudstone with abundant interbedded pale-colored sandstone, which laterally pinch out to form striking reddish paleosols. As with the contorted strata of the Monitor Butte Member, deposition of the Cameron Member seems to fall within the southeast to northwest trend paralleling the Zuni Sag; this trend includes ZION, Lees Ferry in GLCA, the southwestern part of the Navajo Nation near Cameron, and PEFO. The lower part of the Blue Mesa Member type section in PEFO (Lucas, 1993) is nearly identical to the ZION section (Lucas, 1993; Martz and others, 2015) and should probably be referred to the Cameron Member.

Moss Back Member of the Chinle Formation

The Moss Back Member is a multi-storied, conglomeratic sandstone unit deposited by northwest-flowing braided and meandering rivers that overlies the Monitor Butte Member east of CARE (Stewart and others,1972a; Blakey and Gubitosa, 1983; Blakey, 2008). The Moss Back, like most of the conglomerates in the Chinle Formation above the Shinarump Member, is at least locally dominated by intrabasinal clasts of reworked pedogenic carbonate and clastic sedimentary rock (Stewart and others, 1972a). However, a massive ledge-forming conglomeratic sandstone that is composed almost entirely of extrabasinal siliceous clasts occurs low in the Chinle Formation in the San Rafael Swell area, and was identified as the Moss Back Member by Stewart and others (1972a). A sandstone unit potentially correlative with the Moss Back Member is present at the same stratigraphic position in CARE, the Circle Cliffs, and the surrounding region (Dubiel, 1991; Kirkland and others, 2014b; J.W. Martz, unpublished data). It is likely that the Moss Back Member is at least partially correlative to the Sonsela Member of northern Arizona (Lucas, 1993; Howell and Blakey, 2013), which was deposited between the early and middle Norian (latest Lacian and Alaunian) from 220 to 213 Ma (Ramezani and others, 2011, 2014; Atchley and others, 2013). The Sonsela Member however is a much thicker and more complex sandstone- and conglomerate-dominated unit compared to the Moss Back Member (e.g., Martz and Parker, 2010), and it is not clear which part of the Sonsela Member that the Moss Back Member correlates with.

Petrified Forest Member of the Chinle Formation

The nomenclature of the upper reddish and pastel-colored beds of the Chinle Formation is broadly shared between Arizona and Utah, with some regional variations (e.g., Stewart and others, 1972a; Blakey and Gubitosa, 1983, 1984; Lucas, 1993). Above the Moss Back Member is a reddish unit of interbedded mudstone, sandstone, and conglomerate containing abundant pedogenic carbonate nodules, the Petrified Forest Member (Gregory, 1950 sensu Woody, 2006; Parker, 2006; Martz and Parker, 2010; = Painted Desert Member of the Petrified Forest Formation sensu Lucas, 1993; see Martz and others, 2015, for a discussion of the complex nomenclatural history of this unit). Following the same thickness trends observed with the Chinle Formation as a whole, the Petrified Forest Member is 230 m thick at Monument Valley (Stewart and others, 1972a) and 130 m thick at PEFO (Martz and others, 2012), but less than 60 m thick at Circle Cliffs in GSENM and CARE in southeastern Utah (Kirkland and others, 2014b; J.W. Martz, unpublished data). The Petrified Forest Member was deposited by meandering rivers that underwent frequent avulsions due to large input of volcanic detritus (Blakey and Gubitosa, 1983; Johns, 1988) as western North America was beginning to undergo an interval of aridification (Tanner, 2003; Dubiel and Hasiotis, 2011; Atchley and others, 2013). The Petrified Forest Member was deposited between about 213 to 209 Ma during the middle to late Norian (latest Alaunian to Sevatian), and possibly earliest Rhaetian (e.g., Atchley and others, 2013).

The Owl Rock Member and Kane Springs Beds of the Chinle Formation

The Owl Rock Member (Kiersch, 1956) is a pastel-colored unit dominated by non-bentonitic mudstone and sandstone, with ledge-forming carbonate beds that have a generally gradational lower contact with the Petrified Forest Member (Stewart and others, 1972a; Blakey and Gubistosa, 1983, 1984; Dubiel, 1993; Martz and others, 2012). The Owl Rock Member was deposited by lakes, swamps, and associated river systems during a phase of relatively low volcanic input compared to underlying Chinle deposits (Blakey and Gubitosa, 1983; Dubiel, 1987a, 1987b; Dubiel and Hasiotis, 2011) during the Rhaetian, about 209 to 205 Ma (Atchley and others, 2013).

In the salt anticline region of the Paradox Basin (notably Lisbon Valley, Indian Creek, Canyonlands National Park, and Dead Horse Point State Park), where the lower units of the Chinle Formation are absent, the base of the formation consists of drab-colored sandstone and conglomerate with interbedded mudstone beds. These beds are called the Kane Springs beds (Blakey and Gubitosa, 1983, 1984), and are probably correlative with the Owl Rock Member and possibly part of the Petrified Forest Member (figure 2A) (Blakey and Gubitosa, 1983, 1984; Hazel, 1994; Martz and others, 2014). The Kane Springs beds were deposited by meandering and braided rivers originating off Uncompahgria and flowing through poorly drained floodplains confined within contemporaneous, northwesterly oriented salt anticlines (Hazel, 1994). To the west and southwest, these rivers probably joined with the Petrified Forest and/or Owl Rock Member depositional systems (Blakey, 1978; Huber, 1980, 1981; Blakey and Gubitosa, 1983, 1984; Dubiel and Brown, 1993; Hazel, 1994).

Church Rock Member of the Chinle Formation

The uppermost unit of the Chinle Formation consists of reddish-brown siltstone, sandstone, and conglomerate referred to as the Church Rock Member (Stewart, 1957) in Utah, the Rock Point Member (Harshbarger and others, 1957) in northern Arizona, the Dolores Formation in southwestern Colorado (Cross and Howe, 1905) and the siltstone member in northern New Mexico (Stewart and others, 1972a). Although Lucas (1993) assigned all of these strata to the Rock Point Member, facies variations within these uppermost beds are complex enough that regional variations in nomenclature may be justified (J.W. Martz and W.G. Parker, unpublished data), and it is moreover possible that these strata are not all precisely correlative. O'Sullivan (1970) claimed that the type section of the Church Rock Member in northern Arizona (Witkind and Thaden, 1963) was only correlative with the Hite Bed at the very top of the strata assigned to the Church Rock Member in southern Utah. He referred to the younger southern Utah strata as the "reddish-orange siltstone member." As O'Sullivan's (1970) hypothesis has never been investigated in detail by subsequent authors, we retain the traditional nomenclature, referring to all of the uppermost Chinle Formation strata in southern Utah as the Church Rock Member.

In southern Utah, the Church Rock Member contains prominent ledge-forming conglomeratic sandstone. The black ledge beds (sensu Martz and others, 2014; = Black Ledge sensu Stewart and others, 1959) occur at the base of the member, whereas the Hite Bed (Stewart and others, 1972a) and red ledge beds (Martz and others, 2014) both occur higher in the member.

The lower Owl Rock Member has been dated radioisotopically at about 207 Ma (Atchley and others, 2013), indicating that the overlying Church Rock was probably deposited during the Rhaetian (e.g., Martz and others, 2014). The Church Rock Member was deposited by northwest-flowing, coarse-grained braided streams outcropping in a belt extending from southwestern Colorado to the San Rafael Swell (Stewart and Wilson, 1960). Rare eolian sandstone beds also occur within the Church Rock Member (Stewart and others, 1972a; Dubiel, 1989; Hazel, 1994).

Lower Wingate Sandstone and Moenave Formation

The J-0 (Tr-5) unconformity, which erosionally caps the Chinle Formation across most of the Colorado Plateau, separates the Chinle Formation from the lowermost formations of the Glen Canyon Group: the Wingate Sandstone and Moenave Formation (Pipiringos and O'Sullivan, 1978; Blakey, 1994; Kirkland and others, 2014a).

The Wingate Sandstone is an extremely well-sorted, siliceous, cliff-forming sandstone that occurs northeast of a line extending through Tuba City, Arizona, and Paria, Utah, including the San Rafael Swell, State Road (SR) 128, SR 313, Dead Horse Point State Park, and CARE. Locally, discontinuous lenses of sandstone and conglomerate sometimes referred to informally as the Big Indian Rock beds occur at the base of the Wingate Sandstone (Martz and others, 2014; Irmis and others, 2015).

The Dinosaur Canyon Member of the Moenave Formation, which is laterally equivalent to the Wingate Sandstone (figures 2 and 3), occurs southwest of the line running through Tuba City and Paria, including at ZION and St. George Dinosaur Discovery Site at Johnson Farm (SGDS). The Dinosaur Canyon Member is a reddish-brown unit of interbedded sandstone and mudstone (Clemmensen and others, 1989). The overlying Whitmore Point Member consists of a series of thin-bedded shale, limestone, and sandstone that separates the Dinosaur Canyon Member from the overlying Springdale Sandstone Member of the Kayenta Formation along the Arizona Strip and in southwestern Utah west of Kanab (Wilson, 1967). The contact between the Dinosaur Canyon and the overlying Whitmore Point Members is usually taken as a limestone bed partially replaced by red chert (Wilson, 1967; Kirkland and others, 2014a).

The Big Indian Rock beds represent localized channels that flowed within paleovalleys incised into the uppermost Church Rock Member that were subsequently buried by the Wingate erg before the end of the Rhaetian (Lucas and others, 1997a, 1997b; Martz and others, 2014). The lower part of the Wingate Sandstone is probably Rhaetian, as at least part of the Dinosaur Canyon Member of the Moenave Formation, and possibly part of the Whitmore Point Member (Donohoo-Hurley and others, 2010; Kozur and Weems, 2010; Lucas and others, 2011; Kirkland and others, 2014a). The Triassic-Jurassic boundary, which probably lies high in the fluvially deposited Dinosaur Canyon Member or fluvial-lacustrine lower Whitmore Point Member, is well constrained radioisotopically at 201.3 Ma (Schoene and others, 2010; Blackburn and others, 2013).

STRATIGRAPHIC NOTES ON UPPER TRIASSIC UNITS IN NORTHERN UTAH

Chinle Formation exposures in north-central and northeastern Utah (figure 4) have generally received much less study than the Chinle Formation south of Interstate Highway 70 (I-70). The stratigraphic relationships between the Chinle Formation in northeastern Utah to the rest of the Colorado Plateau are poorly understood. Patchy exposures of Upper Triassic strata across much of north-central Utah have been assigned to the Gartra and Stanaker Members of the Ankareh Formation (Thomas and Krueger, 1946; Kummel, 1954; Rigby, 1968; Brandley, 1988, 1990) and/or Bell Springs Formation (Lucas, 1993; May, 2015), but little work has been done correlating these with the southern Utah members. Informally named units of the Chinle Formation in the Dinosaur National Monument area of northeastern Utah (e.g., Dubiel, 1992; Jensen and Kowallis, 2005; Irmis and others, 2015) also have poorly understood stratigraphic relationships with the Chinle Formation members of the Colorado Plateau with the exception of the Gartra Member, which is generally considered to be correlative with the Shinarump Member (Kinney, 1955; Stewart and others, 1972a; Brandley, 1990; Lucas, 1993). Lucas (1993) correlated the Bell Springs "Formation," the uppermost putative Upper Triassic unit below the Navajo Sandstone, with the Rock Point Member.

VERTEBRATE PALEONTOLOGY OF THE CHINLE FORMATION IN SOUTHERN UTAH

During the Late Triassic, the vertebrate fauna of western North America consisted of a combination of surviving Permian faunal elements and relatively new vertebrate groups that radiated during the Triassic (e.g., Sues and Fraser, 2010; Nesbitt, 2011). The most successful tetrapod clade in western North America were the archosauromorphs, particularly phytosaurs, pseudosuchians (crocodylian-line archosaurs), and dinosauromorphs (e.g., Long and Murry, 1995; Nesbitt and others, 2007; Nesbitt, 2011; Stocker and Butler, 2013; Desojo and others, 2013; Irmis and others, 2013; Langer and others, 2013; Nesbitt and others, 2013). Additionally, metoposaurid temnospondyls, dicynodonts, and other non-mammalian synapsids represent surviving Permian lineages that continued to enjoy a measure of success during the Late Triassic in North America (e.g., Fröbisch, 2009; Sues and Fraser, 2010; Brusatte and others, 2015). Freshwater sharks, coelacanths, lungfishes, and a wide variety of "ray-finned fish" were important components of the freshwater vertebrate fauna (e.g., Schaeffer, 1967; Murry, 1989; Gibson, 2013b). Many of these faunal elements are known from southern Utah (figures 3 and 5).

Osteichthyes

Elasmobranchs are well known from the Chinle Formation and Dockum Group (e.g., Murry, 1989) but have not been described from the Chinle Formation in Utah. Bony fishes (Osteichthyes) are, however, represented, including both cranial material and teeth of lobe-finned fishes (Sarcopterygii) and a variety of spectacular ray-finned fishes (Actinopterygii) (figure 5).

The Church Rock Member of Lisbon Valley has produced many complete coelacanth skeletons, including the holotype specimen of *Chinlea sorenseni* Schaeffer 1967 (figure 5A) (AMNH 5652) and other referred specimens (AMNH 5656, AMNH 5660, AMNH 5704; Schaeffer, 1967), including specimens collected for the Natural History Museum of Utah (Milner and others, 2006a). Lungfish tooth plates have been recovered from the Stanaker Member of the Ankareh Formation near Spanish Fork (Brandley, 1990), two tooth plates of *Arganodus* have been collected from the lower Church Rock Member of the Chinle Formation in Lisbon Valley (A.R.C. Milner, unpublished data), and two specimens are known from the Monitor Butte Member at the Circle Cliffs in GSENM (figure 5B) (YPM 57071) and an un-numbered specimen collected in 2016 (W.G. Parker and J.W. Martz, unpublished data). Another lungfish tooth plate has been collected from the Owl Rock Member of CARE (figure 5C) (un-numbered specimen from locality Wn0183; Kirkland and others, 2014b). Parrish and Good (1987) reported lungfish tooth plates from the Petrified Forest Member at the Four Acres mine in White Canyon.

The Church Rock Member of southern Utah, especially Lisbon Valley has yielded by far the richest and most diverse collection of freshwater ray-finned fishes from the Chinle Formation (Schaeffer, 1967; Milner and others, 2006a; Gibson, 2013a, 2013b, 2015), rivaled only by specimens from the correlative Dolores Formation in nearby southwestern Colorado (Schaeffer, 1967; Dubiel and others, 1989). Schaeffer (1967) described most of these specimens following intensive collecting in 1953 by the Atomic Energy Commission, Utah Geological Survey, and the Museum of Comparative Anatomy at Harvard University. Although Schaeffer (1967, p. 292–293) gave locality information only detailed enough to locate the quarries approximately, more recent excavations by Andrew Milner (Milner and others, 2006a; Martz and others, 2014) have produced numerous new specimens from Walt's Fish Quarry, occurring in fluvial deposits in the upper part of the Church Rock Member. Specimens recovered from Lisbon Valley include the non-teleostean ganoid *Hemicalypterus weiri* Schaeffer, 1967 (figure 5D), the redfieldiiforms *Cionichthys dunklei* (figure 5E) (e.g., AMNH 5615), *Lasalichthys hillsi* (e.g., AMNH 5636), and *Synorichtys stewarti* (e.g., AMNH 5646), the palaeonisciform *Turseodus* sp., and the semionotiforms *Lophionotus sanjuanensis* Gibson, 2013a (figure 5F) (e.g., AMNH 5680) and *Lophionotus chinleana* (e.g., UMNH VP 19417) (Schaeffer, 1967; Milner and others, 2006a; Gibson, 2013a, 2013b, 2015). The type specimens of most species above occur at Lisbon Valley (Schaeffer, 1967; Gibson, 2013a, 2013b). These fish range in length from about 4.5 to 35 cm (Schaeffer, 1967; Gibson, 2013b). Whereas most were probably carnivorous, the teeth of *Hemicalypterus* are multidenticulate, and it may be the earliest known herbivorous fish (Gibson, 2015).

Metoposaurids

Metoposaurids were temnospondyl amphibians that were common large aquatic carnivores in North America, Europe, Morocco, and India (e.g., Hunt, 1993; Sulej, 2002; Brusatte and others, 2015). In western North America, Upper Triassic metoposaurid amphibians are usually assigned to three genera (Colbert and Imbrie, 1956; Hunt, 1993; Sulej, 2002). Large specimens with prominent parietal flanges are usually

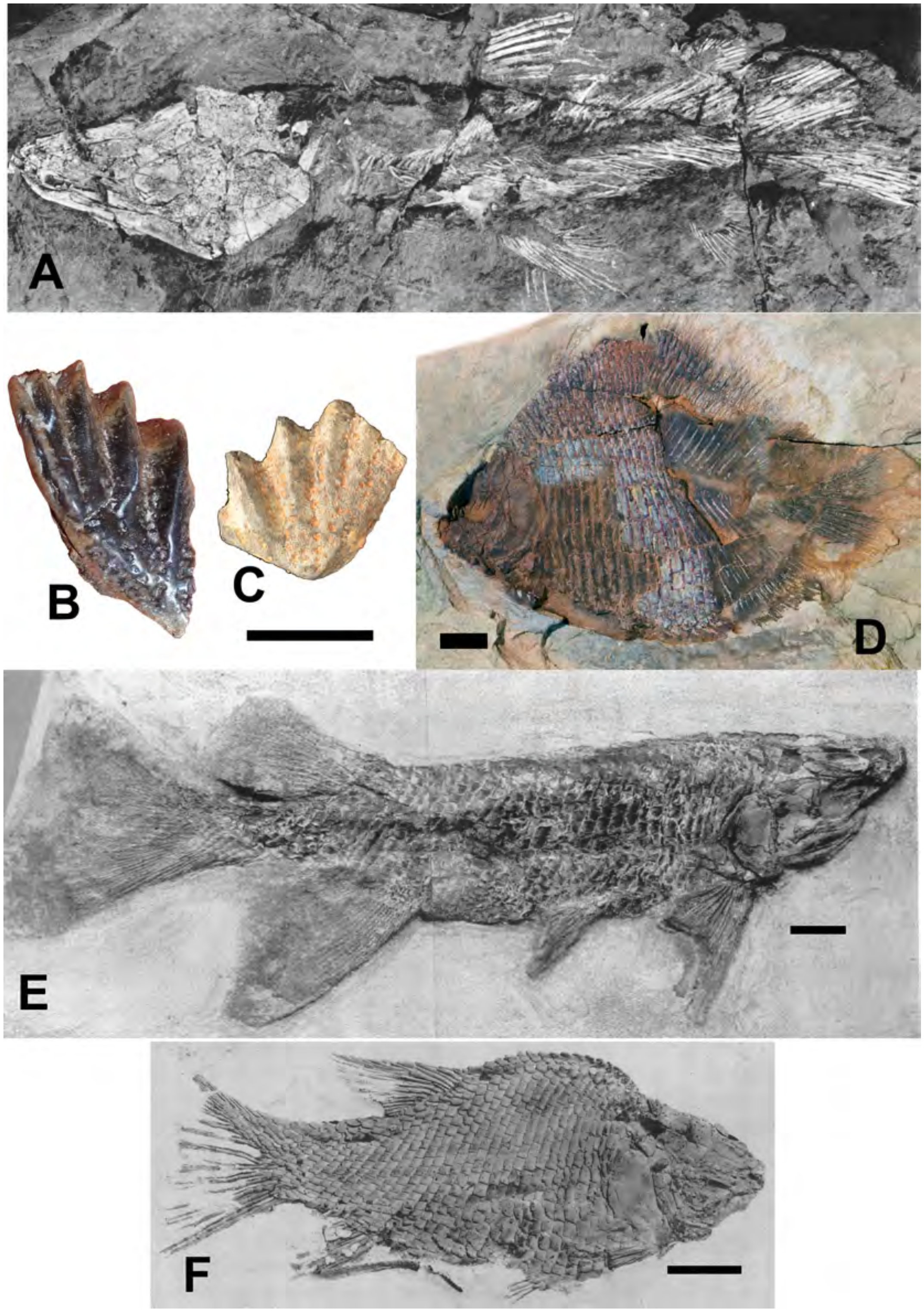

Figure 5. Fish fossils from the Chinle Formation of Utah. (A) Coelocanth *Chinlea sorenseni* AMNH 5652 (Schaeffer, 1967, pl. 26) from the Church Rock Member of Lisbon Valley. (B) Un-numbered lungfish tooth from the Monitor Butte Member of GSENM. (C) Un-numbered lungfish tooth form the Owl Rock Member of CARE (Kirkland and others, 2014a, figure 22). Actinopterygiian fish from the Church Rock Member of Lisbon Valley: (D) *Hemicalypteris weiri* USNM 23425 (Gibson, 2015, figure 1a). (E) *Cionichthys dunklei* AMNH 5615 (Schaeffer, 1967, pl. 11). (F) *Semionotus* sp. AMNH 5680 (Schaeffer, 1967, pl. 22). Scale bars = 1 cm.

placed in the species *Metoposaurus* or *Koskinonodon* (= *Buettneria*; see Mueller, 2007) *bakeri* or *Koskinonodon perfectus*, whereas small specimens with reduced parietal flanges are placed in the genus *Apachesaurus* (e.g., Hunt, 1993; Long and Murry, 1995; Sulej, 2002). In recent treatments *Metoposaurus* and *Koskinonodon* are distinguished almost exclusively based on whether the lacrimal contacts the orbit and the size of the area on the interclavicle covered with rounded and hexagonal pits (Hunt, 1993; Long and Murry, 1995; Sulej, 2002), although more detailed descriptions of metoposaurid variation are currently underway (Brusatte and others, 2015).

Temnospondyl specimens tentatively identifiable as metoposaurid are known in the Chinle Formation of Utah (figures 6A to 6E), including both large and rounded centra with notochordal depressions (figure 6E) and fragments of heavily sculptured cranial and pectoral bones (figures 6A and 6B). Almost none of this material however, can be assigned with certainty to an alpha taxon, and indeed its tentative referral to Metoposauridae is based only on the absence of any other known large temnospondyl taxa from the Chinle Formation. A partial metoposaurid interclavicle was reported from the Shinarump Member in Monument Valley (Lewis and Trimble, 1960), a unit in which tetrapod remains are generally rare; we do not know where this specimen is reposited. Large metoposaurid fragments occur in the Monitor Butte Member of CARE and the Circle Cliffs (figures 6A and 6B) (e.g., YPM 57154, 57126; Kirkland and others, 2014b; W.G. Parker and J.W. Martz, unpublished data) and the Vermilion Cliffs (Parker and others, 2006), as well as in the Cameron Member of ZION (figure 6E) (Martz and others, 2015). Six additional sites are known outside of ZION in the Lower Cameron and Shinarump Members. Additionally, large metoposaurid fragments, including a mandible fragment, have been recovered from the Capitol Reef Bed in the upper Petrified Forest Member at CARE (figures 6C and 6D); these are unusually high occurrences stratigraphically (Kirkland and others, 2014b). Finally, there are two metoposaur sites in the lower Chinle Formation in the Las Vegas, Nevada area (Joshua Bonde, University of Nevada, Las Vegas, verbal communication, 2015).

The large size of the specimens, and the fact that the vertebral centra are shorter than wide (figure 6E) indicate only that they are probably referable to *Metoposaurus* or *Koskinonodon* (= *Buettneria*) rather than *Apachesaurus* (Hunt, 1993). Smaller metoposaurid vertebrae that may be referable to *Apachesaurus* (YPM 57075, YPM 57106) because of having elongate vertebral centra (Hunt, 1993) are known from the Circle Cliffs (W.G. Parker, unpublished data), although their precise stratigraphic position has not been determined. Fragmentary metoposaurid material is also known from the Church Rock Member of Lisbon Valley (Milner and others, 2011; Martz and others, 2014; A.R.C. Milner, unpublished data), indicating that metoposaurids persisted at least until the final stages of Chinle deposition.

Procolophonids

Procolophonid parareptiles are not commonly encountered in the Chinle Formation. *Chinleogomphius jacobsi* (Murry, 1987) from the Blue Mesa Member of Arizona has been identified as procolophonid (Sues and Olsen, 1993), and several undescribed specimens similar to *Hypsognathus* from the Owl Rock Member of Arizona await description (Sues and others, 2000, p. 283). The only known specimen from Utah is a mostly complete skull (figure 6F) (MNA V9953) recovered from the Owl Rock Member in the Manti-La Sal National Forest southwest of Lisbon Valley (Fraser and others, 2005). This specimen is probably a leptopleurine procolophonid closely related to *Hypsognathus*, though distinct from the specimens from the Owl Rock Member of Arizona (Fraser and others, 2005).

Doswellids

Doswellids were armored basal archosauriforms that may have had a semi-aquatic predatory lifestyle as in proterosuchids, proterochampsids, and phytosaurs (Sues and others, 2013). Parrish and Good (1987) and Parrish (1999) noted that fragmentary osteoderms from the Blue Lizard mine locality in Red Canyon, occurring very low in the Monitor Butte Member, have similarities to the North American doswellid taxon *Doswellia* (figure 6G). Specifically, these osteoderms possess fine sub-circular pits and rugose articular

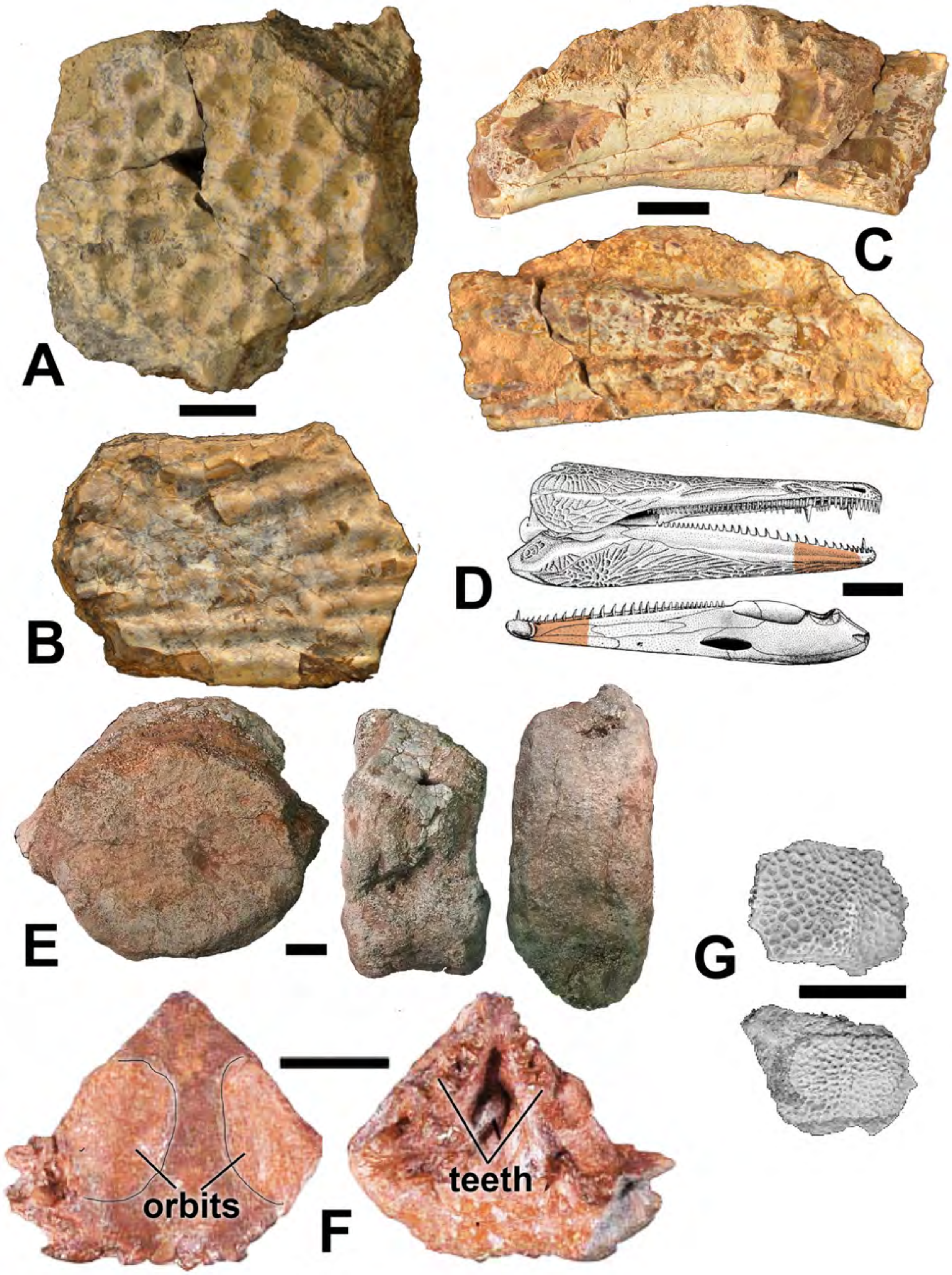

Figure 6 caption on following page.

Figure 6 (figure on previous page). Non-archosauriform tetrapods from the Chinle Formation of southern Utah. (A and B) Metoposaurid fragments. Un-numbered cranial or pectoral fragments from the Monitor Butte Member of CARE (Kirkland and others, 2014a, figures 16B and 16C). (C) Un-numbered right dentary fragment from the Petrified Forest Member of CARE (Kirkland and others, 2014a, figure 20F-H) in (top to bottom) medial and lateral view. (D) Interpretive drawing showing placement of same specimen (modified from Sulej, 2002, figure 3). (E) Metoposaurid centrum (ZION 15679) from the Cameron Member of ZION in (left to right) anterior, lateral, and ventral views. (F) Procolophonid skull (MNA V9953) in (left to right) dorsal and ventral views (Fraser and others, 2005, figure 3.1-2). (G) *Doswellia* osteoderms (UCM 76194) from the Blue Lizard mine in the lowest Monitor Butte Member of Red Canyon (Parrish, 1999, figure 4). Scale bars = 1 cm for 6A to 6C and 6E to 6F; scale bar = 5 cm for 6D.

edges (Parrish and Good, 1987), which also occur in *Doswellia* (Weems, 1980; Sues and others, 2013).

Phytosaurs

Phytosaurs were large predatory archosauriforms that superficially resembled modern crocodylians in form and probably in lifestyle (e.g., Hunt, 1989; Stocker and Butler, 2013), and are among the most commonly encountered vertebrates in the Chinle Formation. Throughout the stratigraphic thickness of the Chinle Formation and the correlative Dockum Group of New Mexico and Texas, the massive, long-snouted skulls of phytosaurs show phylogenetic changes in the posterior region of the skull that correspond well to their stratigraphic position (e.g., Lucas and Hunt, 1993; Martz and Parker, 2017); specifically, the external nares become posteriorly displaced above the antorbital fenestra, the supratemporal fenestra tend to become increasingly reduced and depressed below the skull roof, and the postorbitosquamosal bars become increasingly broadened (Ballew, 1989; Stocker, 2010, 2012). As a result, the utility of phytosaurs as biostratigraphic index fossils has been recognized for decades (e.g., Colbert and Gregory, 1957; Gregory, 1972; Lucas and Hunt, 1993; Lucas, 1998; Parker and Martz, 2011; Martz and Parker, 2017).

The oldest appearances of phytosaur taxa are used to define and bound four Late Triassic land vertebrate "faunachrons" or biozones in western North America; from lowest to highest, these are the Otischalkian, Adamanian, Revueltian, and Apachean (e.g., Lucas, 1998; Parker and Martz, 2011; Desojo and others, 2013). The phytosaur taxa used to define all four biozones are known from the Dockum Group in Texas (e.g., Lucas, 1998; Martz, 2008). In the Chinle Formation, however, the fossil record of phytosaurs is less complete; this is particularly true of Utah. Fragmentary phytosaur material, including teeth, partial skulls and mandibles, and postcranial elements are frequently encountered throughout the Chinle Formation of Utah (e.g., Stewart and others, 1972a; Parrish and Good, 1987; DeBlieux and others, 2005, 2006; Kirkland and others, 2014b; Martz and others, 2014, 2015). Indeed, a phytosaur mandible from Clay Hill in southern Utah identified as *Heterodontosuchus* (Lucas, 1898) is one of the oldest Triassic vertebrate fossils known from the state, although the taxon is a *nomen dubium* (Stocker and Butler, 2013). Phytosaurs are, however, referred to alpha taxa almost exclusively using cranial characters, especially from the posterior region of the skull (Ballew, 1989; Stocker, 2010, 2012, 2013; Hungerbühler and others, 2013), and specimens preserving this region are extremely rare in Utah, with the notable exception of Lisbon Valley (Martz and others, 2014). As a result, biostratigraphic subdivision of the Chinle Formation in southern Utah is generally problematic, with the exception of the Apachean biozone (see below).

Basal phytosaurs (e.g., *Parasuchus* and *Wannia*; Stocker, 2013; Kammerer and others, 2015), which define the Otischalkian biozone (Lucas, 1998; Martz and Parker, 2017), are unknown from the Chinle Formation. Basal members of the clade Leptosuchomorpha (e.g., *Smilosuchus* and *Leptosuchus* sensu Stocker, 2010), which define the base of the Adamanian biozone (Lucas, 1998; Parker and Martz, 2011), are abundant in Arizona (Stocker, 2010), but only two possible specimens are known from Utah. One is a nearly complete phytosaur skull (figure 7A) (UMNH VP.21917) from the lower Cameron Member near ZION (Martz and others, 2015, figure 15E; see below), which has a deep squamosal and partially depressed supratemporal fenestra, char-

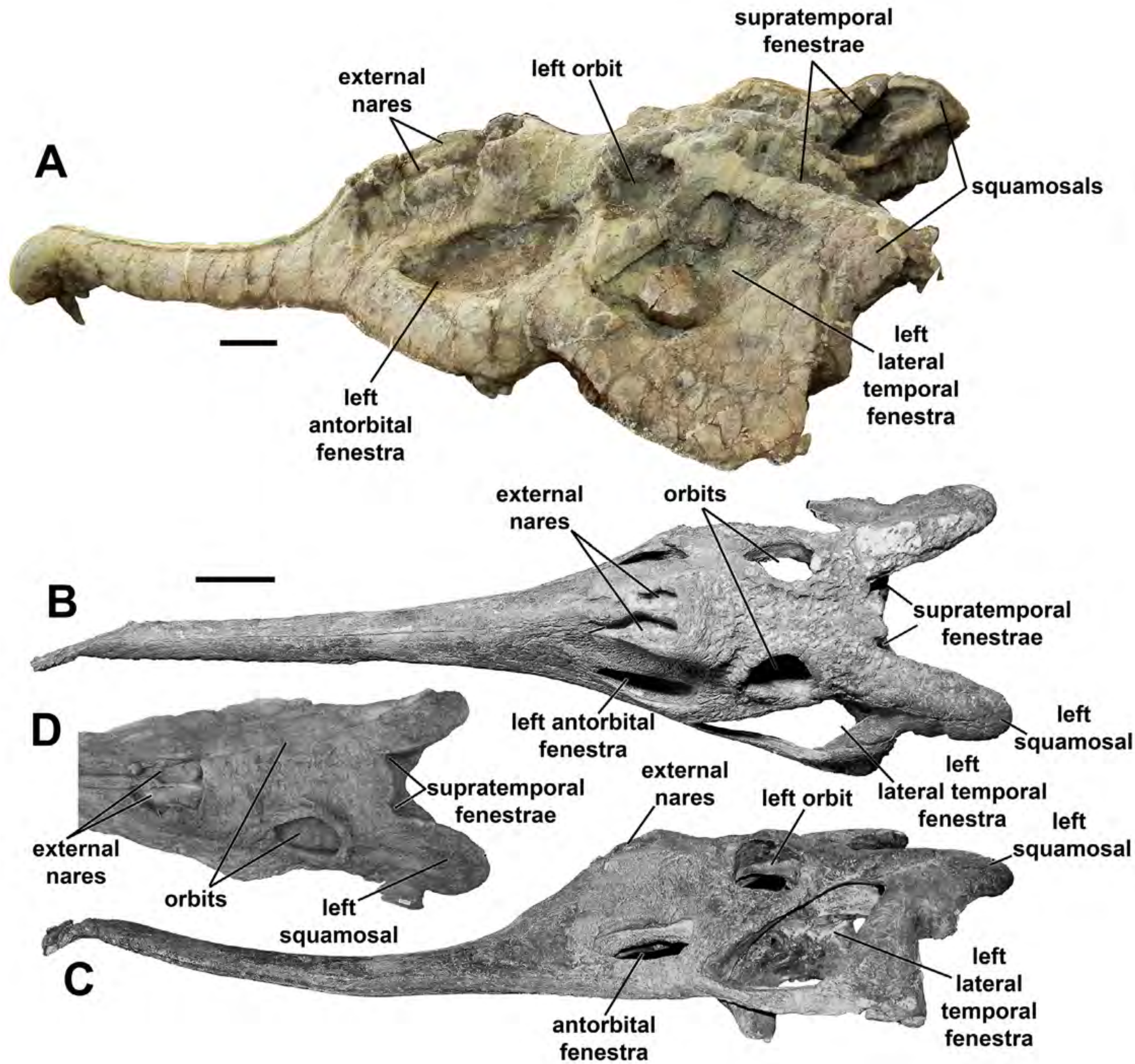

Figure 7. Phytosaur skulls from the Chinle Formation of southern Utah. (A) Leptosuchomorph or non-leptosuchomorph phytosaurid UMNH.VP.21917 in left lateral view (Martz and others, 2015, figure 15E). (B) *Machaeroprosopus* sp. UMNH VP 24304 in dorsal view. (C) Same specimen in left lateral view (Martz and others, 2014, figures 14B and 14C). (D) Latex peel of the "last phytosaur" steinkern, "*Redondasaurus*" (*Machaeroprosopus*) sp. (MNA V3498/UMNH VP 22354) in dorsal view (Martz and others, 2014, figure 14A). Scale bars = 10 cm.

acteristics of basal leptosuchomorphs (Stocker, 2010), although Martz and others (2015) suggested it might be a more basal taxon. Another undescribed, nearly complete skull collected from the Monitor Butte Member of Fry Canyon by Phil Bircheff is a non-pseudopalatine leptosuchomorph (W.G. Parker, unpublished data) that may be referable to the pseudopalatine sister-taxon *Pravusuchus hortus* Stocker 2010 (McCormack and Parker, 2017). All specimens suggest the base of the Adamanian biozone occurs somewhere in the Monitor Butte and Cameron Members.

Three partial phytosaur skulls collected from Fry Canyon by the Atomic Energy Commission in 1953 were assigned to *Rutiodon tenuis* (= *Machaeroprosopus pristinus*; Parker and Irmis, 2006) by Parrish and Good (1987); however, these specimens are probably non-pseudopalatine leptosuchomorphs (J.W. Martz, unpublished data), as they possess deep plate-like squamosals rather than the shallower and knob-like squamosals found in the pseudopalatine *Machaeroprosopus pristinus* (e.g., Long and Murry, 1995; Parker and Irmis, 2006; Stocker, 2010). Parrish and Good (1987) stated that these specimens were recovered from the Petrified Forest Member; we have not, however, had the opportunity to confirm this by revisiting the locality.

Pseudopalatinae (sensu Parker and Irmis, 2006, = Mystriosuchini sensu Kammerer and others, 2015) is a derived phytosaur clade within Leptosuchomorpha very well represented in western North America by the genus *Machaeroprosopus* (sensu Parker and others, 2013; Hungerbühler and others, 2013; = *Pseudopalatus*), the first appearance of which defines the Revueltian biozone (Lucas, 1998; Parker and Martz, 2011). *Machaeroprosopus* can be distinguished from non-pseudopalatine phytosaurs in having broadened postorbitosquamosal bars and greatly reduced and depressed supratemporal fenestrae (e.g., Stocker, 2010; Hungerbühler and others, 2013). These characters are further exaggerated in specimens of *Machaeroprosopus* that are traditionally assigned to the genus "*Redondasaurus*" in which the supratemporal fenestrae are virtually eliminated and the postorbitosquamosal bar is greatly broadened (e.g., Hunt and Lucas, 1993; Spielmann and Lucas, 2012; Hungerbühler and others, 2013; Martz and others, 2014). Whereas the basal species of *Machaeroprosopus* (e.g., *M. pristinus* and *M. buceros*) that define the base of the Revueltian biozone are well known from the Chinle Formation in Arizona and New Mexico (e.g., Long and Murry, 1995), they have not been reported in Utah.

More derived specimens however, at least some of which can be assigned to the "*Redondasaurus*" morphotype, are known from the Church Rock Member of Lisbon Valley (figures 7B to 7D) (Martz and others, 2014). Two of these specimens (UMNH VP 24304 and UMNH VP 24236) are represented by nearly complete skulls (figures 7B and 7C) (Martz and others, 2014). The stratigraphically highest known specimen of *Machaeroprosopus* from anywhere in the western United States is from the Big Indian Rock beds (Martz and others, 2014) at the base of the Wingate Sandstone; this specimen (figure 7D) (UMNH VP 22354) is a steinkern of a skull roof referred to informally as the "last phytosaur" (Morales and Ash, 1993; Lucas and others, 1997b; Spielmann and Lucas, 2012; Martz and others, 2014).

Aetosaurs

Aetosaurs are a clade of armored pseudosuchian archosaurs that were common worldwide during the Late Triassic, with the most diverse fossil record known from the Dockum Group of Texas and New Mexico and the Chinle Formation of New Mexico and Arizona (e.g., Desojo and others, 2012, 2013; Parker, 2016a). Clades and alpha taxa are diagnosed primarily using the osteoderm characters, specifically the shape and ornamentation patterns on the paramedian osteoderms covering the back and the lateral osteoderms covering the sides (e.g., Long and Ballew, 1985; Long and Murry, 1995; Heckert and Lucas, 2000; Parker, 2007; Desojo and others, 2012, 2013). Two major clades have been consistently identified in recent phylogenetic analyses (e.g., Desojo and others, 2012, 2013; Heckert and others, 2015; Parker, 2016a): Typothoracinae, which generally have extremely mediolaterally broad paramedian osteoderms and lateral osteoderms with tapering dorsal flanges, and Desmatosuchinae, which generally have narrower and thicker paramedian osteoderms and spikey lateral osteoderms with more subrectangular dorsal flanges. Other taxa basal to these clades have paramedian osteoderms with radiating surface ornamentation and subrectangular lateral osteoderms without spikes, but these taxa probably do not form a monophylum (Desojo and others, 2013; Parker 2016a). Like phytosaurs, aetosaur taxa have biostratigraphic utility (e.g., Long and Ballew, 1985; Lucas, 1998; Heckert and Lucas, 2000; Parker and Martz, 2011; Martz and others, 2013), and their occurrences may therefore help delineate the Late Triassic land vertebrate biozones in Utah.

Heckert and others (1999) described a badly worn aetosaur paramedian osteoderm (NMMNH P-26938) with an associated tooth and undescribed postcranial material from Canyonlands National Park, which was recovered about 11.5 m above the base of what they identified as the Blue Mesa Member. Examination of their measured section (Heckert and others, 1999, figure 1) suggests that this is actually the middle mudstone-dominated unit of the Kane Springs beds. Based on its narrow width, transverse arching, presence of an anterior bar, and radiating ornamentation pattern, Heckert and others (1999) assigned the osteoderm to *Stagonolepis*. However, North American material often referred to this taxon (e.g., Heckert and Lucas, 2000) is distinct enough from the type material of *Stagonolepis* from Scotland to warrant assigning it to two distinct genera: *Calyptosuchus* and *Scutarx* (Long and Ballew, 1985; Parker, 2016b). We have not had the opportunity to examine the Canyonlands specimen, and cannot comment on its identification. However, the presence of either *Calyptosuchus* or *Scutarx* in the Kane Springs beds would be anomalously high, as these are Adamanian taxa (Heckert and Lucas, 2000; Martz and others, 2013; Parker, 2016b) and the Kane Springs beds probably fall within the Revueltian and Apachean biozones.

In CARE and the Circle Cliffs area of GSENM, more diagnostic material has come to light. A partial paramedian osteoderm with the distinctive pitted ornamentation and pronounced ventral strut of *Typothorax* (e.g., Heckert and Lucas, 2000; Martz, 2002) is known from the Petrified Forest Member of CARE (figure 8A) (un-numbered specimen from locality Wn0147), although the exact stratigraphic level is unclear (Kirkland and others, 2014b). Parrish and Good (1987) reported a fragment of a *Typothorax* osteoderm from the Petrified Forest Member at the Four Acres mine in White Canyon, although the specimen is a fragment and we have not had the opportunity to examine it. *Typothorax* occurs as a rare occurrence in the Adamanian biozone (e.g., Parker and Martz, 2011; Martz and others, 2013) but is far more abundant in the Revueltian and Apachean biozones (e.g., Heckert and Lucas, 2000; Parker and Martz, 2011; Desojo and others, 2013; Martz and others, 2014), which is consistent with the stratigraphic occurrence of the specimen in the Petrified Forest Member at CARE (see below).

Several osteoderms from the Circle Cliffs area collected by Yale University are referable to the aetosaur clade Paratypothoracini (YPM 57096, YPM 57118, YPM 57124, YPM 57113; W.G. Parker and J.W. Martz, unpublished data), which lies within Typothoracinae. One specimen from the same area can be assigned to *Desmatosuchus* (YPM 57132; W.G. Parker, unpublished data), and at least one specimen can be assigned to the genus *Calyptosuchus* (YPM 57152; W.G. Parker, unpublished data; Parker and others [2006] had previously assigned this specimen to *Stagonolepis*). Although paratypothoracins are known from both the Adamanian and Revueltian biozones in western North America (e.g., Parker and Martz, 2011; Desojo and others, 2013), *Calyptosuchus* is restricted to the lower Adamanian biozone (Parker, 2016b) and *Desmatosuchus* to the Adamanian and lower Revueltian (Parker and Martz, 2011; Martz and others, 2013). The stratigraphic level at which these specimens occur is therefore significant, and is currently being investigated using field data collected by Yale University.

Several osteoderms referable to *Typothorax* are known from the Kane Springs beds (figure 8B) (e.g., UMNH VP 24247) and Church Rock Member (figures 8C and 8D) (e.g., UMNH VP 21880; UMNH VP 24232) in Lisbon Valley, making at least some of them Apachean biozone occurrences (Martz and others, 2014). These include both fragmentary paramedian osteoderms with pitted ornamentation and lateral osteoderms with triangular dorsal flanges; both characters diagnose the genus *Typothorax* (e.g., Martz, 2002). A fragmentary osteoderm with pitted ornamentation that may be referable to *Typothorax* is also known from the lower Church Rock Member of ARCH (J.W. Martz and J.I. Kirkland, unpublished data).

Paracrocodylomorphs

Paracrocodylomorpha is the clade that includes mostly carnivorous pseudosuchians traditionally referred to as “rauisuchians,” as well as crocodylomorphs (Nesbitt, 2011; Irmis and others, 2013; Nesbitt and oth-

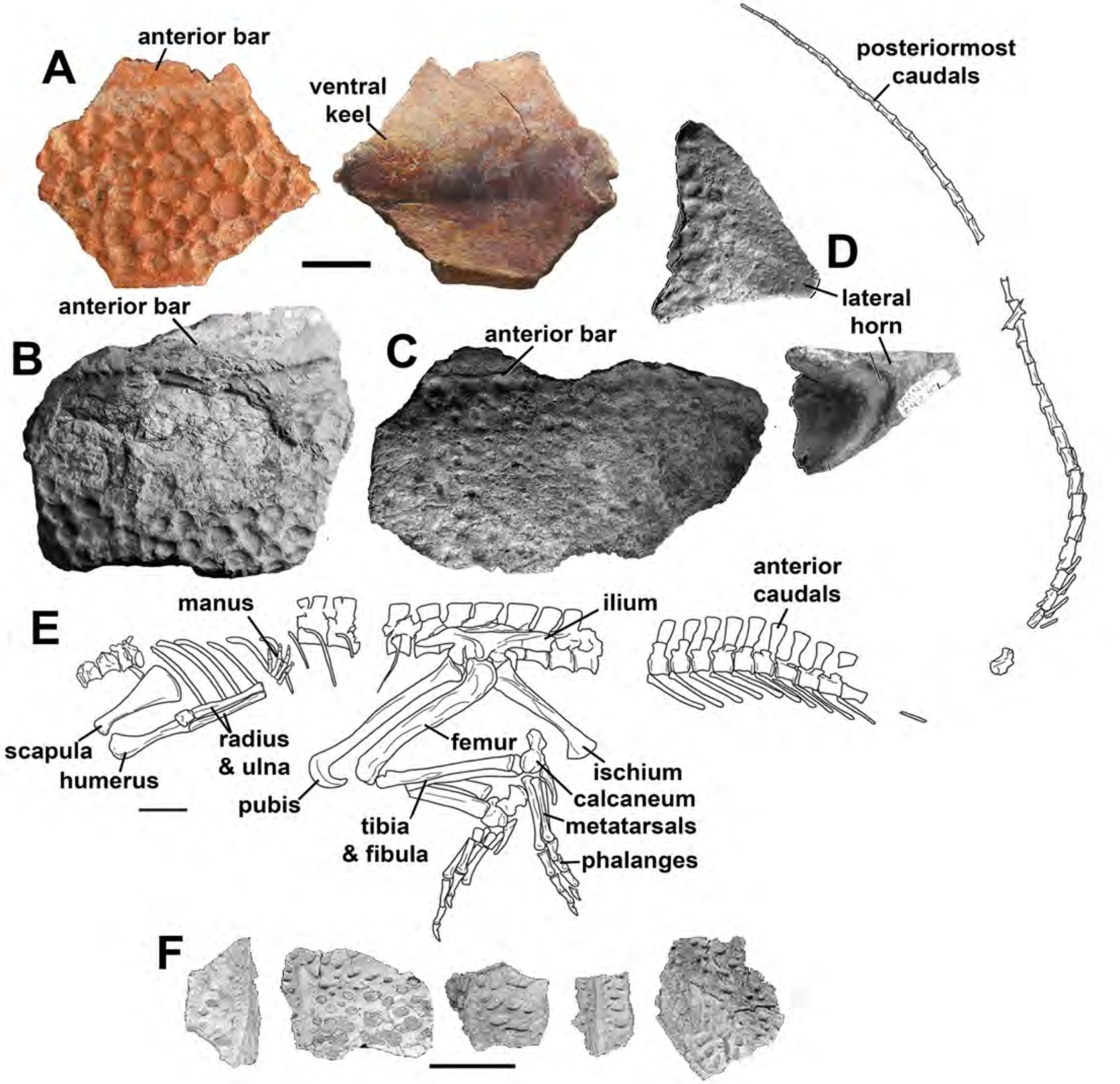

Figure 8. Pseudosuchian archosaurs from the Chinle Formation of southern Utah. (A) Un-numbered partial paramedian osteoderm of the aetosaur *Typothorax coccinarum* from the Petrified Forest Member of CARE in (left) dorsal and (right) ventral views (Kirkland and others, 2014a; figures 20J to 20K). (B) Partial *Typothorax* paramedian UMNH VP 22971 from the Kane Springs beds of Lisbon Valley. (C) Partial *Typothorax* paramedian UMNH VP 24641 from the Church Rock Member of Lisbon Valley (Martz and others, 2014, figures 15F and 15J). (D) Partial *Typothorax* lateral osteoderm UMNH VP 24232 from the Church Rock Member of Lisbon Valley (Martz and others, 2014, figures 16A and 16D). Scale bar for A to D = 2 cm. (E) Nearly complete skeleton of *Poposaurus gracilis* YPM VP 057100 from the Monitor Butte Member of GSENM (Gauthier and others, 2011, figure 1A), scale bar = 10 cm. (F) Possible crocodylomorph osteoderms from the Blue Lizard mine in the lowermost Monitor Butte Member of Red Canyon (Parrish, 1999, figure 1), scale bar = 10 cm.

ers, 2013), although a closer relationship between aetosaurs and crocodylomorphs has been proposed that places the latter clade outside of Paracrocodylomorpha (Brusatte and others, 2010; Butler and others, 2011). Within Paracrocodylomorpha, the clades Poposauroidea and Crocodylomorpha are represented in the Chinle Formation of Utah; Rauisuchidae, a possibly paraphyletic group that consists entirely of large-bodied carnivores such as *Postosuchus* (Weinbaum, 2011, 2013; Nesbitt and others, 2013) and *Vivaron* (Lessner and others, 2016) has not yet been documented in southern Utah.

Poposauroidea is a clade within Paracrocodylomorpha that includes taxa with a wide array of locomotor and dietary adaptations (Nesbitt and others, 2013). In North America, poposauroids are represented by *Poposaurus*, a bipedal carnivore with remarkable convergent resemblance to theropods (Weinbaum and Hungerbühler, 2007; Gauthier and others, 2011; Schachner and others, 2011), and the shuvosaurids (= chatterjeeids), a clade of generally smaller bipedal poposauroids with edentulous beaks that were probably herbivorous (e.g., Nesbitt and others, 2013). One of the most spectacular tetrapod specimens from the Chinle Formation of Utah is a nearly complete and fully articulated skeleton lacking only the skull (figure 8E) (YPM 57100) referable to *Poposaurus gracilis* (Gauthier and others, 2011; Schachner and others, 2011). An isolated ischium (YPM 57129), probably from the Monitor Butte Member, may also belong to *Poposaurus* (W.G. Parker, unpublished data). Shuvosaurid material has not yet been identified from southern Utah.

Several basal crocodylomorph specimens have been recovered in southern Utah. Parrish and Good (1987) and Parrish (1999) described fragmentary material from the Blue Lizard mine locality in Red Canyon (figure 8F) (e.g., UCM 76195). By far the best crocodylomorph specimen from the Chinle Formation is a complete articulated skeleton found in association with the *Poposaurus* skeleton from Circle Cliffs, nicknamed the "popo-buddy." Parrish and Good (1987) tentatively identified a distal tibia from the Moss Back Member at Red Canyon as a crocodylomorph, although we have not examined this specimen and are uncertain where it is reposited. The Church Rock Member in Lisbon Valley has produced an articulated hind limb and partial tail of a crocodylomorph (Milner and others, 2011).

All of these are modest-sized specimens (probably no more than 1 to 2 m long), but larger crocodylomorph material, including a partial skull, have been recovered from the Church Rock Member in Lisbon Valley (A.R.C. Milner, unpublished data). Large crocodylomorph specimens have also been recovered from the Apachean biozone in New Mexico (Nesbitt, 2011, p. 235–236; Nesbitt and others, 2005; Axel Hungerbühler, Mesalands Comminity College, and Jonathan Weinbaum, Southern Connecticut University, verbal communication, 2015), suggesting that crocodylomorphs may have been important large predators during the Apachean (Irmis and others, 2013).

Ornithodira

The ornithodiran record from the Chinle Formation of southern Utah is poor. The proximal end of a tibia referable only to Ornithodira (YPM 57231) is known from the Chinle Formation in Circle Cliffs, but its stratigraphic position has not yet been relocated. Tracks of *Grallator* have been recovered from the Big Indian Rock beds immediately below the Wingate Sandstone (Milner and others, 2006a; Hunt-Foster and others, 2016; A.R.C. Milner, unpublished data), confirming the presence of theropods in the Apachean biozone. Parrish (1999) identified unguals (UCM 76197) from the Blue Lizard mine locality from Red Canyon as theropod, although the assignment was not justified, and a caudal vertebra (UCM 76198) was also assigned from the same locality as theropod due to having a "distinctive, subhexagonal cross section;" it is not at all clear if this character has any phylogenetic significance (Nesbitt, 2011).

LOCAL EXPOSURES OF THE CHINLE FORMATION IN SOUTHERN UTAH

Chinle Formation exposures are readily accessible throughout southern Utah (figure 4), particularly in its many beautiful national parks and monuments. Several key exposures of the Chinle Formation in southern Utah are discussed below, presented as one might visit them in a loop starting at Salt Lake City, moving through southeastern Utah to southwestern Utah and back north again (figure 4). Note, in keeping with odometers used in the United States distances in miles are in parentheses.

Stop 1
U.S. Highway 6 Road Cut Southeast of Spanish Fork

The first Triassic outcrops traversed on this trip occur just southeast of Spanish Fork. After taking I-15 from Salt Lake City south through Provo to Spanish Fork, take U.S. Highway 6 southeast. Red bed outcrops occur about 20 km (12 mi) southeast of the junction of I-15 and U.S. Highway 6, with the best exposures being about 1.2 km (0.8 mi) past the turnoff for Diamond Fork Road (Forest Road 029), in a road cut just northeast of the ghost town of Thistle (figures 4 and 9A) (Rigby, 1968; Brandley, 1988, p. 90–176, 1990; May 2015, p. 26–31).

Stratigraphy and Sedimentology

These red beds have been assigned by various authors (Rigby, 1968; Brandley, 1988, p. 90–176, 1990; May 2015, p. 26–31) to the Ankareh Formation (figures 9B to 9D), a name originally applied to exposures in Park City about 70 km (44 mi) to the north (Boutwell, 1907). The Ankareh Formation consists mostly of Lower and Upper Triassic fluvial deposits and has been divided into three members: the Mahogany, Gartra, and Stanaker (figure 9B) (Thomas and Krueger, 1946; Kummel, 1954; Brandley, 1990).

Unlike the authors cited above, Lucas (1993) followed Stewart and others (1972a) in restricting the term "Ankareh Formation" to Lower Triassic rocks correlative with the Moenkopi Formation, and extended Upper Triassic stratigraphic nomenclature of southern Wyoming into the region by assigning the strata overlying the Garta Member in northern Utah to the Popo Agie and Bell Springs Formations. Although Lucas (1993) did not specifically discuss the Spanish Fork road cut, his application of the names Popo Agie and Bell Springs likely refer to the lower and upper parts of what Brandley (1988, 1990) called the Stanaker Member of the Ankareh Formation.

The lowermost 120 m of the Mahogany Member, including the contact with the underlying Thaynes Limestone, is not exposed in the U.S. Highway 6 road cut, but the Mahogany Member (figure 9B) nonetheless makes up the majority of the roughly 300 m of Triassic exposures in the road cut (Brandley, 1988, 1990). The Mahogany Member consists of primarily dark reddish-brown interbedded siltstone and sandstone with common ripple cross-laminations and rare mud cracks and raindrop marks, as well as some conglomerate composed of re-worked mudstone clasts (Rigby, 1968; Brandley, 1990). The Mahogany Member is Spathian (Late Early Triassic) in age and partially syndepositional with the more extensive Moenkopi Formation of southern Utah and northern Arizona (Blakey and others, 1993). Both the Mahogany Member and correlative members of the Moenkopi Formation were deposited by northwesterly flowing meandering rivers and associated tidal and sabkha environments bordering the marine environments covering much of Utah (Brandley, 1990; Blakey and others, 1993).

Above the Mahogany Member lie the much thinner Gartra and Stanaker Members (sensu Brandley, 1990), which are probably Upper Triassic in age and correlative with the Chinle Formation (Stewart and others, 1972a; Brandley, 1990). These Upper Triassic members are confined to the southeastern part of the U.S. Highway 6 road cut. The Gartra Member, which was deposited by braided rivers flowing across northern Utah and Colorado and is considered to be correlative with the Shinarump Member (Stewart and others, 1972a; Brandley, 1990), was identified as unit 192 in Brandley's (1988, 1990) measured section. This member is a 10-m-thick unit lithologically distinct from both the underlying and overlying strata in consisting of pinkish, resistant, and extremely quartzose sandstone and conglomerate with abundant trough cross-bedding, deposited as multiple lenticular channel fills.

About 40 m below the unit identified by Brandley (1990) as the Gartra Member is an intensely mottled paleosol (figures 9B and 9C) that resembles the mottled strata that typically occur at the base of the Upper Triassic section throughout the western United States. The mottled strata form a resistant ledge (Brandley, 1990,

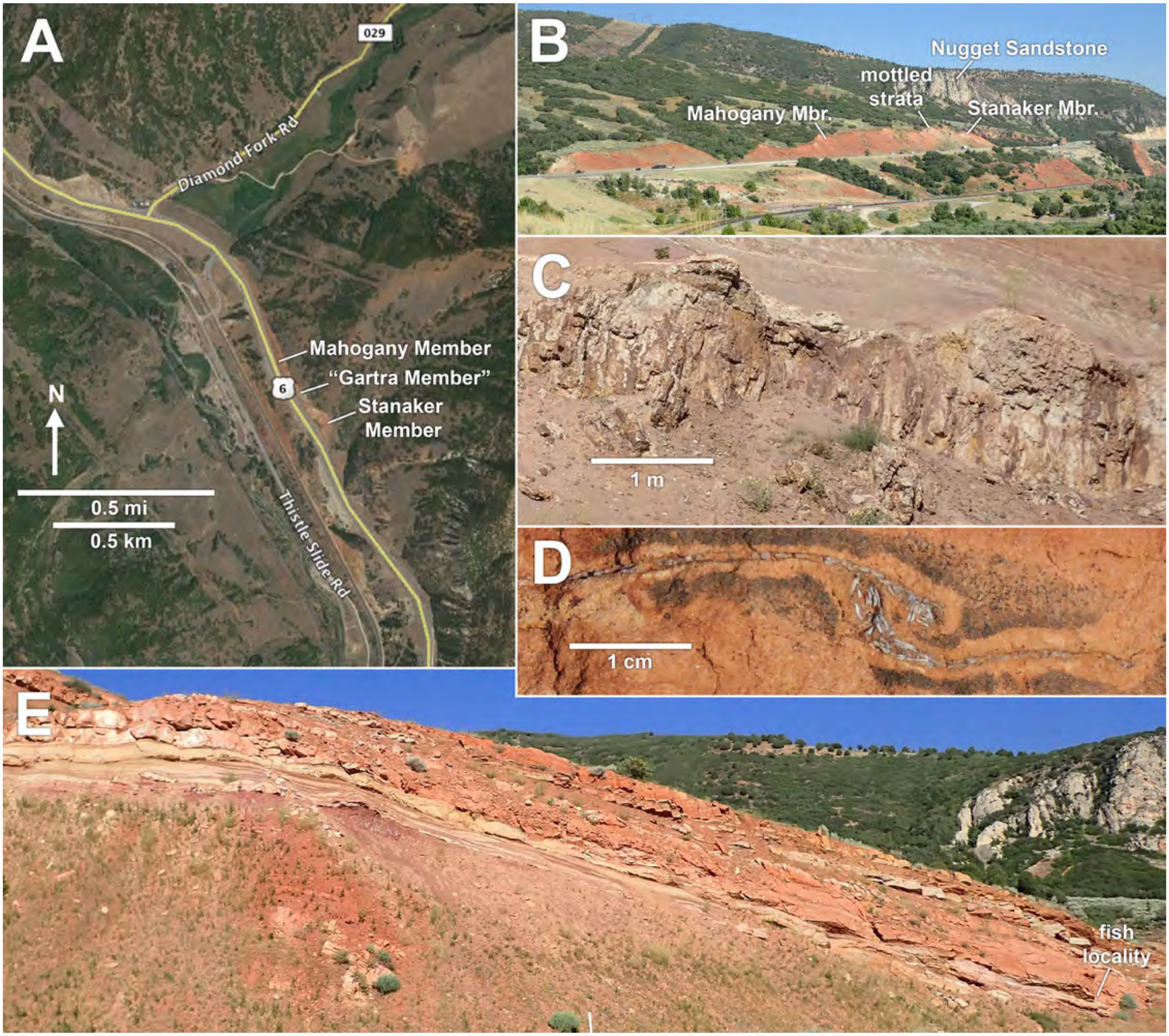

Figure 9. Ankareh Formation exposures near Spanish Fork. (A) Map of outcrop location. (B) Road cut exposures, view southeast along U.S. Highway 6; the Mahogany and Stanaker Members comprise all reddish exposures to the left and right of the mottled strata, respectively. (C) Ledge of well-lithified mottled paleosol that may represent base of Chinle Formation. (D) Fish scales in ledge-forming unit of Stanaker Member (location shown in figure 9E). (E) Stratigraphically highest exposures in road cut of Stanaker Member. The yellow and pinkish ledge-forming unit was interpreted by Brandley (1988, 1990) as meandering channel and levee deposits. The deep red mudstone beneath were interpreted as floodplain deposits.

units 166–170) composed of well-cemented quartz sandstone with "four color" mottling of iron oxide, as is often seen in the mottled strata at the base of the Chinle Formation. It is possible that this ledge is a better candidate for the Gartra Member than the sandstone identified as that unit by Brandley (1990).

If the mottled strata do indeed represent the base of the Upper Triassic section, then the Stanaker Member in the U.S. Highway 6 road cut comprises units 176–211 in Brandley's (1988, 1990) section, and is about 76 m thick, rather than 10 m thick as it would be if the Stanaker Formation only comprises Brandley's (1990) units 203–211. Like the Mahogany Member, the Stanaker Member consists of dark reddish-brown mudstone and sandstone deposited primarily by meandering rivers (Brandley, 1990). The lowermost 10 m of the Stanaker Member consists predominantly reddish-brown mudstone (figure 9D), with abundant root traces (including rhizoconcretions) and siltstone lenses with erosional bases (Brandley, 1988, p. 90–98), which have been interpreted as floodplain deposits and channel fills deposited by meandering streams. The uppermost beds of the Stanaker Member in the Highway 6 road cut consist mostly of ledge-forming sandstone, conglomerate, and some interbedded mudstone (figure 9D). The lower part of this sequence is a conglomeratic channel lag with a distinctly scoured base, with the overlying sandstone and interbedded mudstone having been interpreted as splay deposits by Brandley (1990).

The uppermost Stanaker Member, which may be equivalent to the strata that Lucas (1993) assigned to the Bell Springs Formation in northern Utah, is not exposed in the road cut. These beds contain evaporites and reddish-brown and grayish-orange quartz sandstone that have been interpreted as eolian deposits, recording a climatic shift to arid conditions that culminate with the Early Jurassic eolian deposition of the Nugget Sandstone (Brandley, 1990). The Nugget Sandstone forms prominent cuestas just to the southeast of the road cut (figure 9B), and may be correlative with the Wingate Sandstone or Navajo Sandstone farther south (Kinney, 1955; Poole and Stewart, 1964; Peterson, 1988).

Paleontology

Brandley (1990) reported small pieces of silicified and carbonized permineralized wood up to 10 cm in diameter, as well as a lungfish tooth plate, from the conglomeratic channel unit (unit 202). The overlying sandstone identified as splay deposits by Brandley (1990; Brandley's unit 203) contain in situ fish (figure 9D). None of these plant or fish fossils have been assigned to an alpha taxon.

Stop 2
Big Bend along SR 128 along the Colorado River, northeast of Moab

From the Spanish Fork road cut, continue to follow U.S. Highway 6 southeast for 185 km (115 mi) until it joins I-70 just west of Green River, then proceed east along I-70 for 45 km (28 mi) to the junction with U.S. Highway 191. Proceed south on U.S. Highway 191 south for 47 km (29 mi), turning northeast onto SR 128 just north of Moab, and follow it along the Colorado River for about 19 km (12 mi) to the parking area on the north side of the road near Big Bend, just west of where the canyon opens into Castle Valley (figure 10A).

Stratigraphy and Sedimentology

SR 128 is one of the most spectacular scenic drives in southern Utah. Here, thick exposures of the Moenkopi Formation are capped by the Chinle Formation, which is in turn capped by the tremendous vertical cliffs of Upper Triassic-Lower Jurassic Wingate Sandstone (figure 10B). In Castle Valley and Richardson Amphitheater to the north, the Moenkopi is well exposed; farther southwest, heading towards Moab the entire section dips so that the Moenkopi enters the subsurface and the Chinle Formation is exposed closer to the road, particularly between westernmost Castle Valley and the Big Bend along the Colorado River. Approaching Moab along SR 128, the Chinle Formation also disappears into the subsurface and the Wingate lines the road. At the pull-out stop, the Chinle Formation is fully exposed.

SR 128 occurs near the edge of the Chinle basin, in the salt anticline region. This geographic placement has two major impacts on the lithostratigraphy of the Chinle Formation:

1. Only the uppermost Chinle Formation is preserved, and the bulk of the section is usually assigned to the Church Rock Member (e.g., Stewart and others, 1972a, section U-18; Martz and others, 2014). Indeed, one problematic aspect of studying the Chinle Formation along SR 128 is that the contact between the Moenkopi Formation and the Church Rock Member of the Chinle For-

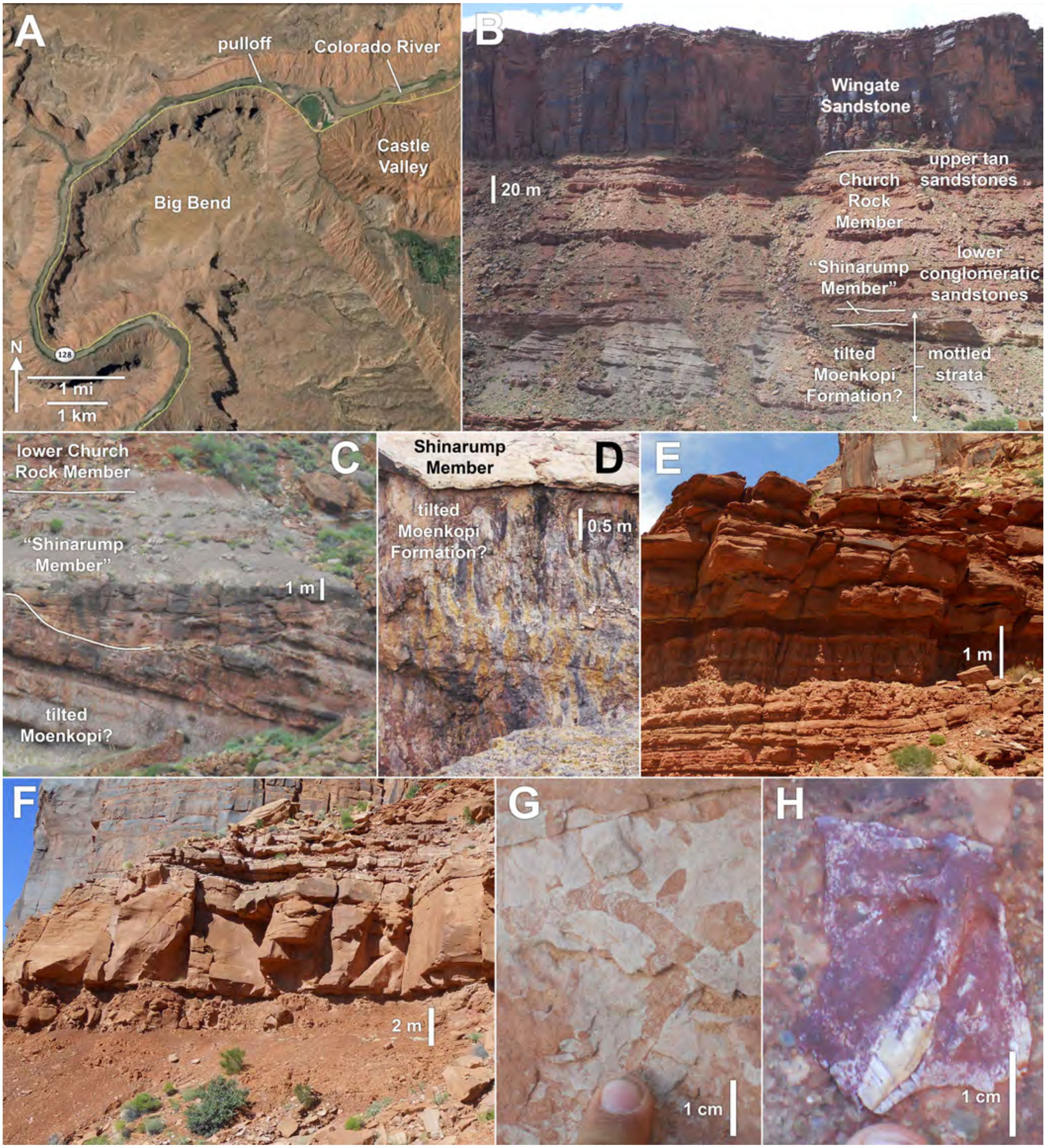

Figure 10. Big Bend along SR 128 near Moab. (A) Map of Big Bend showing approximate pulloff location. (B) Outcrops on the northern side of the Colorado River, easily visible from the road. (C) Same outcrops, showing incision of the sub-horizontal "Shinarump Member" into the tilted strata of the Moenkopi Formation; both units are intensely mottled. (D) Mottled strata with vertical streaks probably representing crayfish burrows in the tilted Moenkopi (?) Formation. (E) One of the lower ledge-forming conglomeratic sandstone beds in the Church Rock Member. (F) Upper tan-colored sandstone in the Church Rock Member with little discernible bedding. (G) Small burrows in the upper tan sandstone. (H) Phytosaur or crocodylomorph osteoderm in the lower conglomeratic beds of the Church Rock Member.

mation is locally difficult to discern (Baker, 1933), as here both formations are predominantly reddish-brown units of interbedded sandstone and mudstone.

2. The thickness of the Chinle Formation is highly variable, as it was controlled largely by syndepositional salt diapirism that generated salt anticlines during Late Triassic time. Along SR 128, the thickness of the Chinle Formation is about 140 m (Stewart and others, 1972a; J.W. Martz and others, unpublished data), notably thicker than the sections in nearby ARCH (J.W. Martz, unpublished notes), and was probably deposited in an area of subsidence that Prochnow and others (2006) and Prochnow (2007) called the Big Bend Minibasin, resulting from subsurface salt evacuation. Despite variations in thickness, however, the Chinle Formation sections along SR 128 and ARCH are essentially identical (J.W. Martz, unpublished data).

Along SR 128, there are striking variations in the nature of the Chinle-Moenkopi contact and the ease with which it is identified. In Richardson Amphitheater and Castle Valley, the contact can be very difficult to discern. There is, however, a thin layer of light-colored siliceous sandstone at the contact that can be faintly discerned in this area that Stewart and others (1972a, section U-18) referred to as the "basal sandstone unit." Within Castle Valley and at the Big Bend, where there is a pronounced angular unconformity at the base of the Chinle Formation (figures 10B and 10C), this siliceous unit is a thick, ledge-forming conglomeratic sandstone composed almost exclusively of extrabasinal quartz. The unit is intensely mottled with shades of red, purple, yellow, and gray taking the form of vertical streaks probably representing crayfish burrows; both mottling and burrowing extend into the tilted strata below the contact (figures 10C and 10D), which the siliceous unit incises to varying depths (figure 10C). From a distance the unit is distinctly paler than the overlying and underlying units, making its identification easy. This unit corresponds to FAC set 3 of Prochnow and others (2006).

Although Baker (1933) identified this siliceous mottled unit as the Shinarump Member, Stewart and others (1972a, figure 10) assigned it to Moss Back Member, and Ronald Blakey (Northern Arizona University, written communication, 2016) also favors this latter correlation. However, we have not observed the development of a four-color gleyed paleosol on top of the Moss Back Member elsewhere (including the San Rafael Swell, see below), whereas this mottling occurs above the Shinarump Member in the majority of the areas of southern Utah where that unit is deposited. For this reason, identifying this basal sandstone as either the Shinarump Member or a correlative is our favored interpretation. If the siliceous sandstone and mottled strata at the base of the Chinle Formation along SR 128 are correlative with the Shinarump Member, the unconformity between these beds and the Church Rock Member is at least 12 Myr, and probably greater (Martz and others, 2014).

Above the mottled siliceous sandstone, the remainder of the Chinle Formation can be assigned to the Church Rock Member (figures 10B, 10E, and 10F) corresponding to FAC sets 4-6 of Prochnow and others (2006). Along the Big Bend, the siliceous mottled sandstone is less than 10 m thick and the overlying 125 m of the Chinle Formation are assigned to the Church Rock Member and shows a broadly fining-upward sequence (figure 10B); reddish ledge-forming sandstone and conglomerate, composed mostly of reworked intrabasinal clasts, occur low in the sequence (figure 10E), and generally become less conglomeratic and dominated by climbing ripple cross-stratification higher in the section. The uppermost Church Rock Member consists of dull tan sandstone that often lack any discernable bedding (figure 10F), and resemble eolian sandstone of the Wingate Sandstone. The Hite Bed of the Church Rock Member, which occurs in this part of Utah (Stewart and others, 1972a), may be reworked eolian sandstone (Hazel, 1994), and possibly, these uppermost beds may represent similar depositional conditions.

The angular unconformity developed along SR 128 at the base of the mottled siliceous sandstone between the Big Bend and Castle Valley to the east is easily seen by looking across the Colorado River to the northwestern side of the canyon (figure 10B). Interpretations of this unconformity have varied. Popular opinion is that the tilted strata below the unconformity are part of the Chinle Formation, with the angular unconformity representing deformation due to salt tectonism that occurred during Late Triassic time (e.g., Prochnow and others, 2006; Martz and others, 2014).

However, based on recent investigations, it seems likely that the beds below the unconformity are pre-Chinle strata, and probably part of the Moenkopi Formation; this conclusion follows the interpretations of Dane (1935) and O'Sullivan and MacLachlan (1975, p. 135) in Richardson Amphitheater. The overall sequence of strata is far more consistent with Chinle Formation sections throughout the salt anticline region if strata below the angular unconformity are excluded. In ARCH, where no angular unconformity occurs, the Chinle Formation section is identical to the SR 128 sequence aside from being slightly thinner. In ARCH, an intensely mottled, siliceous, conglomeratic sandstone is overlain by reddish-brown conglomerate with gravel-sized clasts mostly composed of reworked intrabasinal sediments that generally fine upwards into sandstone with abundant ripple cross-laminations, and then tan sandstone with little discernible bedding. Elsewhere in the salt anticline region, such as Dead Horse Point and along SR 131 (see below), the Kane Springs beds underlie the Church Rock Member, but are virtually identical to the Church Rock lithologically other than being drab-colored (Martz and others, 2014). Moreover, along SR 313, mottled siliceous sandstone beds locally occur immediately below the Kane Springs beds (see below), forming the base of the Chinle Formation as they do along SR 128. If our interpretation is incorrect and the beds below the mottled siliceous sandstone and angular unconformity along SR 128 are actually part of the Chinle Formation (e.g., Prochnow and others, 2006; Martz and others, 2014), they represent a unit present nowhere else in the entire salt anticline region.

Moreover, including the tilted strata below the mottled siliceous sandstone would add considerably to the thickness of the Chinle Formation along SR 128. The thickness of the Chinle Formation throughout the salt anticline region varies from less than 100 to more than 200 m (Stewart and others, 1972a; Blakey and Gubitosa, 1983; Prochnow and others, 2006; Martz and others, 2014; J.W. Martz, unpublished data); the section along SR 128 above the angular unconformity is ~140 m thick (Stewart and others, 1972a; Martz and others, unpublished data). Whereas we have not measured the section below the angular unconformity, it is easily in excess of 60 m, indicating that the total thickness of the Chinle Formation would be greater than we have observed anywhere else in the salt anticline region. It is far more parsimonious to conclude that the strata below the angular unconformity do not belong to the Chinle Formation.

Paleontology

Burrowing is common throughout the Chinle Formation along SR 128. The probable crayfish burrows in the mottled strata are generally 20 to 30 cm wide and vertically oriented, but smaller burrows less than 1 cm wide and in more variable orientations are often seen in the finer grained sandstone beds higher in the section (figure 10G).

Fragmentary vertebrate bones and teeth have been observed in the conglomeratic sandstone in the lower part of the Church Rock Member at the Big Bend, including a probable phytosaur or crocodylomorph osteoderm (figure 10H). Within ARCH, vertebrate bone is also common in the lower conglomeratic sandstone of the Church Rock Member, although little of it is identifiable. Aetosaur osteoderm fragments that may be referable to *Typothorax* have been recovered from this interval within ARCH, as have natural molds of the enigmatic broad leaf plant *Sanmiguelia* (J.I. Kirkland and J.W. Martz, unpublished data), which occurs at several stratigraphic levels in the Chinle Formation (Ash, 1976; Ash and Hasiotis, 2013).

Stop 3
SR 313 Near Junction with U.S. Highway191

From the Big Bend Road cut, take SR 128 back west about 19 km (12 mi) to the junction with U.S. Highway 191, then proceed north 13.6 km (8.5 mi) to the junction with SR 313. Proceed west about 2.1 km (1.3 mi) and pull off to examine the outcrops north of the road (figures 11A to 11C).

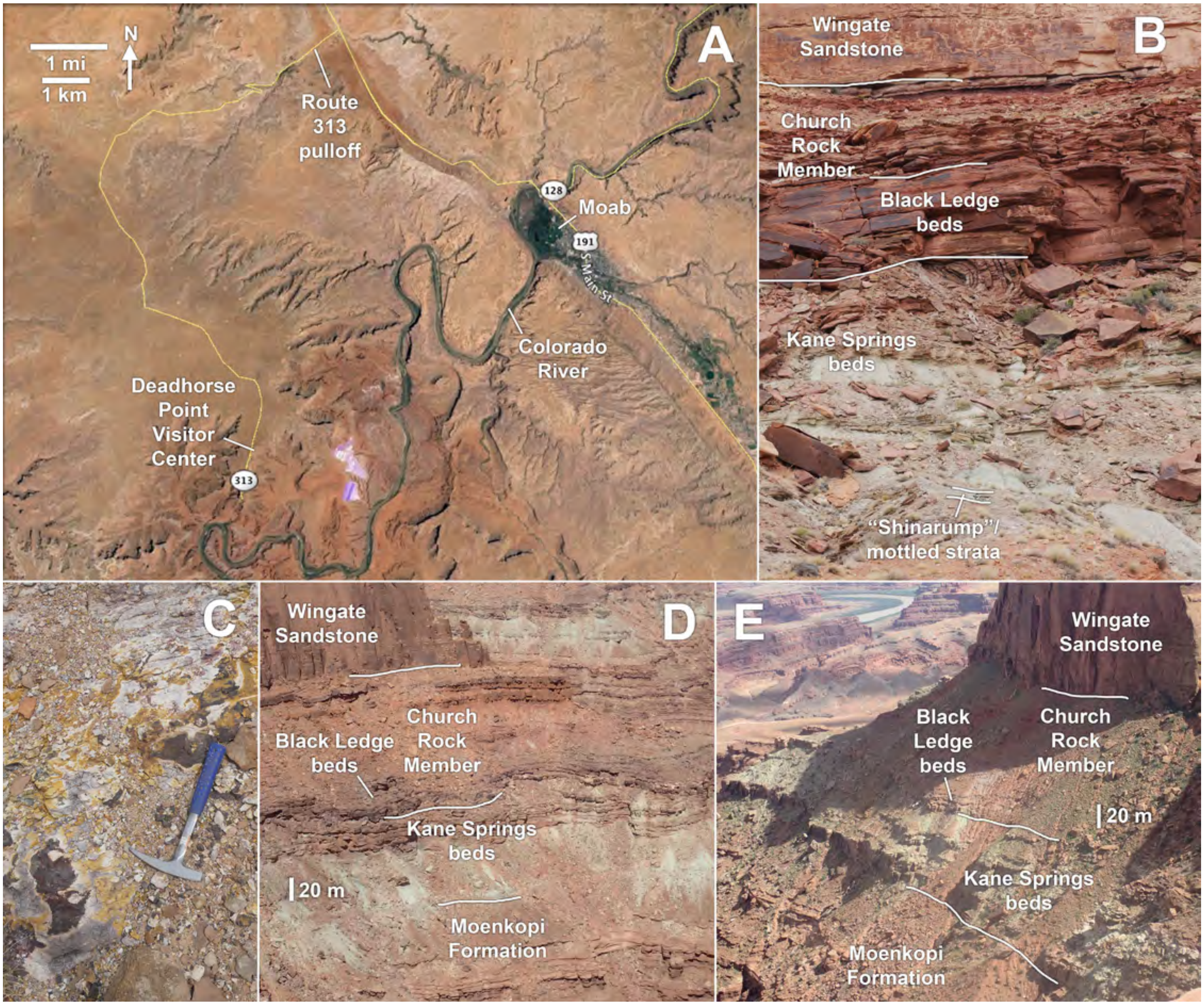

Figure 11. SR 313 and Dead Horse Point State Park. (A) Map of SR 313 showing stops for exposures. (B) Exposures along SR 313 near the junction with Highway 191. (C) Mottled strata in same area. (D) Exposures visible from Dead Horse Point State Park visitor center overlook on the northern side of the canyon. (E) Exposures visible from Dead Horse Point State Park visitor center overlook on the southern side of the canyon.

Sedimentology and Stratigraphy

As with ARCH and along SR 128, the area that includes Canyonlands National Park and Dead Horse Point State Park, Lisbon Valley, and Indian Creek were situated near the edge of the Chinle depositional basin in the salt anticline region (Blakey and Gubitosa, 1983, 1984; Hazel, 1994; Martz and others, 2014); however, unlike ARCH and SR 128, in this region the Kane Springs beds underlie the Church Rock Member (figures 2 and 3). The Chinle Formation section at both Canyonlands and Dead Horse Point cannot be easily examined close up by the casual visitor, requiring a lengthy drive or hike from the top of the Wingate Sandstone and overlying units capping the cliffs to the valley floor. However, the section is more accessible along SR 313 near the junction with U.S. Highway 191, just north of Moab (figure 11A). The section is similar to that exposed elsewhere in the region in being composed primarily of two units: the lower greenish-gray Kane Springs beds and the upper reddish-brown Church Rock Member (figure 11B) (e.g., Blakey and Gubitosa, 1983, 1984; Hazel, 1994; Martz and others, 2014). As noted by previous workers (Stewart, 1956, 1957; Stewart and others, 1959, 1972a; Martz and others, 2014), the Petrified Forest and Owl Rock Members are virtually absent in this region of southeastern Utah except for locally occurring tongues of the Owl Rock Member (Blakey and Gubitosa, 1983, figure 10; Martz and others, 2014, p. 426–427).

Along SR 313, a siliceous mottled sandstone that may be equivalent to the one seen at ARCH and along SR 128 occurs at the base of the section below the Kane Springs strata (figures 11B and 11C) (Martz and others, 2014). It is only a few meters thick here, and is locally completely truncated farther south (including Canyonlands, Lisbon Valley, and Indian Creek) wherever the base of the Kane Springs beds is a thick ledge-forming sandstone (Martz and others, 2014). Russ Dubiel (U.S. Geological Survey, verbal communication, 2016) has also observed remnants of the siliceous mottled strata below the Kane Springs beds near Six Shooter Peaks and west of Lisbon Valley.

The Kane Springs beds have a bimodal grain size composition of predominantly siltstone to fine-grained sandstone, often with horizontal planar bedding and climbing ripple cross-lamination, and conglomerate composed of reworked intrabasinal sedimentary clasts with some extrabasinal siliceous clasts (Hazel, 1994; Martz and others, 2014). Although the Kane Springs beds in Lisbon Valley and some other outcrops such as Bridger Jack Mesa in Indian Creek Valley to the west of SR 211 are divisible into lower and upper ledge-forming conglomeratic sandstone separated by a finer grained middle unit (Huber, 1980, 1981; Blakey and Gubitosa, 1983, 1984; Hazel, 1994; Martz and others, 2014; A.R.C. Milner, unpublished data), there is considerable variation throughout the surrounding region. In the outcrop north of SR 313 (figure 11B), these upper and lower ledge-forming sandstone beds are absent, and the overall unit resembles the middle Kane Springs beds in Lisbon Valley in being predominantly slope-forming mudstone with interbedded sandstone which rest unconformably on the siliceous sandstone and mottled strata (Martz and others, 2014).

Contrary to many workers (e.g., Stewart, 1956, 1957; Stewart and others, 1959; Huber, 1980, 1981; Lucas, 1993; Lucas and others, 1997a; Baars, 2010), the Kane Springs beds are not correlative to the Moss Back Member (Blakey, 1978; Blakey and Gubitosa, 1983, 1984; Martz and others, 2014). In Lisbon Valley, the Kane Springs beds have a clear sedimentological similarity to the overlying Church Rock Member, and there is a gradational contact between the two units (Blakey, 1978; Martz and others, 2014). Moreover, the Kane Springs beds are at least partially correlative with the Owl Rock Member and possibly the Petrified Forest Member (Stewart and others, 1972a; Blakey and Gubitosa, 1983, 1984; Martz and others, 2014), which occur stratigraphically much higher in the section than the Moss Back Member.

The Church Rock Member in this part of Utah is dominated by siltstone and very fine to fine-grained sandstone with abundant climbing ripple cross-lamination and horizontal planar bedding, and some interbedded conglomerate dominated by intrabasinal clasts of reworked siltstone, sandstone, and pedogenic carbonate (e.g., Martz and others, 2014). Along SR 313, a massive ledge-forming sandstone at the base of the Church Rock Member caps the Kane Springs beds unconformably (figure 11B), as it does in Lisbon Valley (Martz and others, 2014), although along SR 313 this ledge makes up most of the member. Martz and others (2014) assigned this ledge to the black ledge beds, although all the ledge-forming units assigned this name may not form a continuous outcrop belt.

Paleontology

Although fossils have not been reported along SR 313 or in Dead Horse Point State Park, the section is similar to that exposed in Canyonlands National Park and especially Lisbon Valley, where fossils are abundant (e.g., Schaeffer, 1967; Hasiotis, 1993; 1995; Heckert and others, 1999; Milner, 2006; Milner and others, 2006a, 2011; Santucci and Kirkland, 2010; Gibson, 2013a, 2013b, 2015; Martz and others, 2014).

Stop 4
Dead Horse Point State Park Visitors Center

From the SR 313 stop, proceed west along SR 313 for about 34 km (21 mi) to the visitor center in Dead Horse Point State Park (figure 11A). A short path leads from the parking lot and visitor center to the overlook into the canyon below, where the Chinle Formation section can be observed (figures 11D and 11E).

Although we have not had the opportunity to examine the Chinle Formation in Dead Horse Point State Park close-up, it seems to have the same members seen in Canyonlands National Park and Lisbon Valley (Martz and others, 2014). These units are easily visible from the Dead Horse Point State Park visitor center overlook, about 2 km (1.2 mi) north of Dead Horse Point (figures 11D and 11E). The Chinle Formation section visible from Dead Horse Point is estimated to be about 200 m thick, with the lower 60 m comprising the Kane Springs beds and the remaining 140 m comprising the Church Rock Member. Here the Chinle Formation is considerably thicker than the ~130-m-thick section exposed along SR 128 near Moab, and not quite twice as thick as the Chinle Formation sections in ARCH and Lisbon Valley.

At the visitor center overlook, lithologic variation in the Kane Springs beds can be observed. On the northern, south-facing exposures (the left-hand exposures seen from the overlook; figure 11D), the lower part of the Kane Springs beds seems to be predominantly fine grained and slope-forming as along SR 313 and in the middle unit at Lisbon Valley, but with an upper ledge-forming sandstone immediately below the Black Ledge beds of the Church Rock Member, similar to the upper unit of the Kane Springs beds in Lisbon Valley. However, on the southern, north-facing exposures (the right-hand exposures seen from the overlook; figure 11E), there seems to be a lower ledge-forming sandstone at the base of the Kane Spring beds, but no upper ledge, as is also the case in parts of Canyonlands National Park (Martz and others, 2014). Ron Blakey (Northern Arizona University, written communication, 2016) believes that the Black Ledge beds may have locally truncated the upper ledge of the Kane Springs beds. Blakey and Gubitosa (1983, figure 10) indicated that the Kane Springs beds at Dead Horse Point contain a tongue of the Owl Rock Member, but this is difficult to evaluate without examining the outcrop up close. It is also currently unknown if the siliceous mottled unit occurring at the base of the Kane Springs beds along SR 313 also occurs at Dead Horse Point State Park.

On both sides of the valley (figures 11D and 11E), a thick ledge where the section changes from greenish-gray to reddish-brown, is interpreted as the transition between the Kane Springs beds and the black ledge beds at the base of the Church Rock Member. Other ledge-forming units occur higher in the section; one occurs approximately in the lower third of the Church Rock Member and two superimposed ledges are visible near the top of the Church Rock Member on the south-facing exposures (figure 11D); on the north-facing exposures these ledges seem to be less prominent (figure 11E). Whether these ledges correlate to the red ledge beds in Lisbon Valley (Martz and others, 2014) or the Hite Bed (Stewart and others, 1972a) is unclear.

Stop 5
Lisbon Valley

From the Dead Horse Point State Park visitors center, proceed back north along SR 313 for 36 km (22.4 mi) to U.S. Highway 191. Turn right and head south on U.S. Highway 191 for 73 km (45.6 mi), passing through Moab, until reaching the junction with Big Indian Road (figure 12A). Proceed about 12 km (7.5 mi) east along Big Indian Road, through the west-dipping cuestas (dipping mesas) capped by the Upper Triassic-Lower Jurassic Glen Canyon Group, until reaching the Big Indian Rock, on the south side of the road. Park to examine the reddish outcrops of Chinle Formation north of the road (figures 12C and 12D).

Sedimentology and Stratigraphy

The valley before you, bounded on the west by the cliff-forming sandstone of the Wingate Sandstone and other overlying units of the Glen Canyon Group, is Lisbon Valley. Lisbon Valley represents the exposed por-

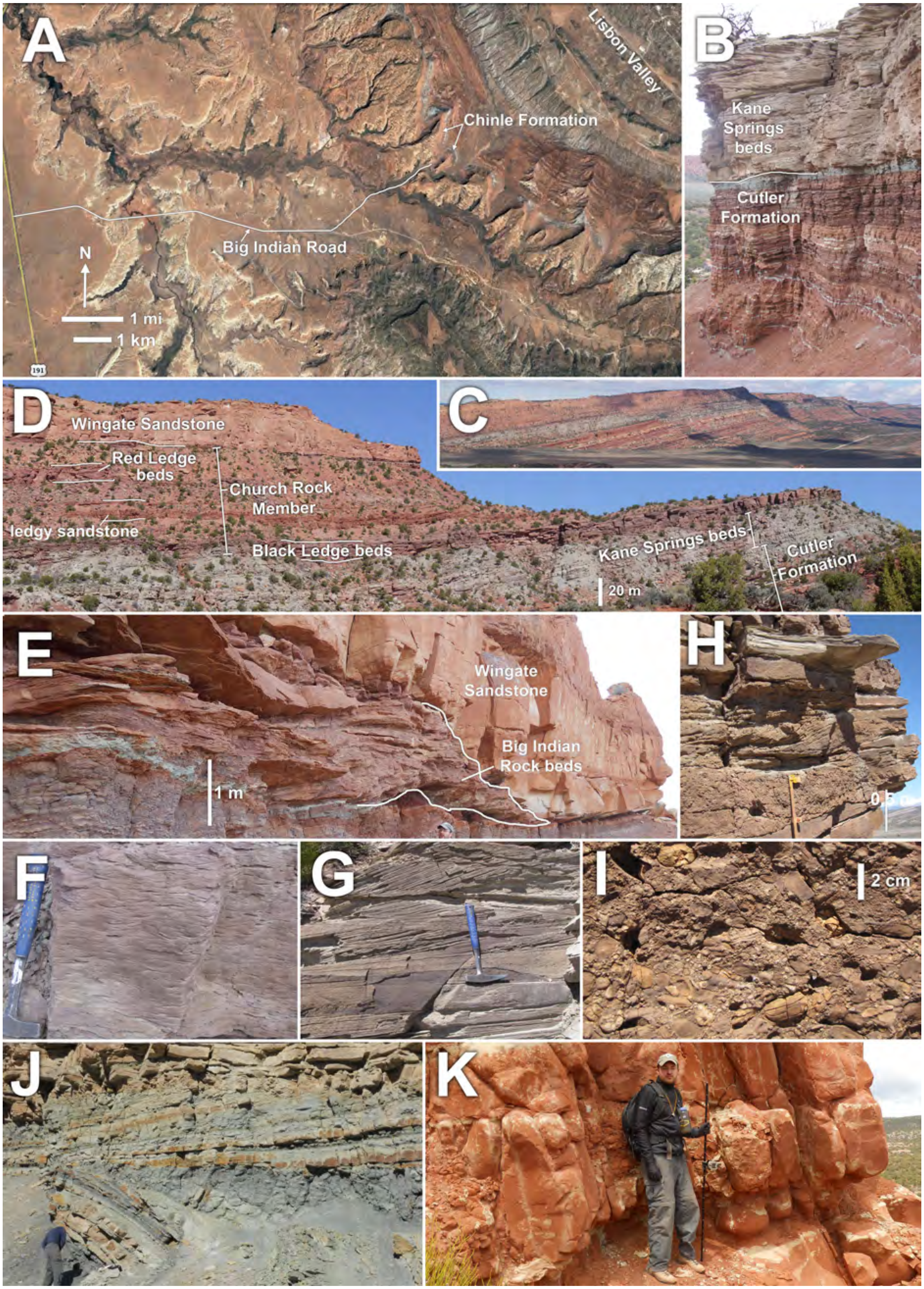

Figure 12 caption on following page.

Figure 12 (figure on previous page). Lisbon Valley. (A) Map of route between U.S. Highway 191 and Lisbon Valley along Big Indian Road. (B) Contact between Permian Cutler Formation and Upper Triassic Kane Spring beds at base of Chinle Formation. (C) Panoramic view across Lisbon Valley from the south of Big Indian Road to the north showing the dipping cuestas; the gray Kane Springs are distinctly visible. (D) Skull Ridge, with different units of the Chinle Formation labeled. (E) The Chinle-Wingate contact, showing the Big Indian Rock beds lensing in at the contact. (F) Climbing ripple cross-stratification in the middle unit of the Kane Springs beds. (G) Horizontal planar bedding and gently dipping cross-bedding in the upper unit of the Kane Springs beds. (H) Lower Kane Springs beds sandstone with distinct conglomeratic beds. (I) Close-up of intrabasinal conglomerate in the lower Kane Springs beds; the orange clasts are mostly reworked siltstone. (J) Fine-grained fissile sandstone and mudstone of the middle Kane Springs beds; the dipping block on the left appears to represent syndepositional slumping similar to that seen elsewhere in the contorted beds of the Monitor Butte Member. (K) Reddish-brown siltstone of the Church Rock Member with some greenish-gray mottling, but no clear signs of paleosol formation.

tion of the doubly plunging and faulted Lisbon Valley anticline, one of several northwest- to southeast-trending salt anticlines within the salt anticline region of the Paradox Basin of southeastern Utah and southwestern Colorado (e.g., Stewart, 1956; Stewart and Wilson, 1960; Hazel, 1994). The salt is produced in the Paradox Formation of the Pennsylvanian Hermosa Group, which occurs at a depth of about 500 m in Lisbon Valley (Byerly and Joesting, 1959; Stewart and Wilson, 1960). Southwest-dipping exposures (figures 12C and 12D) of the Middle Pennsylvanian Hermosa Group, Upper Pennsylvanian-Lower Permian Cutler Group, Upper Triassic Chinle Formation, and the Upper Triassic-Lower Jurassic Wingate, Kayenta, and Navajo Formations occur on the southwestern side of the fault in Lisbon Valley (e.g., Doelling, 2004). Unlike most of southeastern Utah, the Lower to Middle Triassic Moenkopi Formation is locally absent, pinching out near the southwestern flank of Big Indian Wash (Stewart, 1957), so that the Chinle rests on the Cutler Formation (figure 12B). The hogbacks on the west side of Lisbon Valley are bisected by drainages into a series of cuestas (figure 12C) exposing the Wingate Sandstone, Chinle Formation, and Cutler Group. Skull Ridge (figure 12D) is the cuesta lying immediately northwest of Big Indian Road.

The Chinle Formation in southern Lisbon Valley varies from about 112.1 to 123.4 m in thickness (Martz and others, 2014), and the stratigraphy is broadly similar to that seen along SR 313 and in Dead Horse Point State Park (figure 12C). The Kane Springs beds are approximately 30 to 45 m in total thickness, whereas the Church Rock Member in the study area varies in thickness from 59.6 to 112.1 m, with the top of the member being truncated to varying degrees below the Wingate Sandstone (figure 12E) (Martz and others, 2014). Siliceous mottled strata are not seen below the Kane Springs beds here, possibly because they have been erosionally truncated by the lower Kane Springs sandstone (Martz and others, 2014).

Overall, the Kane Springs beds and Church Rock Member in Lisbon Valley seem to be fairly similar lithologically, with no major unconformities and only slight erosional contacts at the bases of coarser grained subunits that tend to fine upwards. The Kane Springs beds and Church Rock Member both have strongly bimodal grain-size distributions, which are most pronounced in the Church Rock Member, whereas the Kane Springs beds are slightly coarser and compositionally immature (e.g., Huber, 1980; Blakey and Gubitosa, 1983; Chenoweth, 1990; Martz and others, 2014). Both the Kane Springs beds and Church Rock Member consist predominantly of very fine to fine-grained sandstone and siltstone (figures 12F and 12G). However, both the Kane Springs beds and Church Rock Member also contain interbedded conglomerates with granule to cobble-sized clasts predominantly composed of reworked intrabasinal micrite, sandstone, and siltstone, and a matrix of medium- to coarse-grained sandstone (figures 12H and 12I) (Blakey, 1978; Dubiel and Brown, 1993). Extrabasinal clasts composed of chert, quartz, and quartzite occur locally in conglomerates of both the Kane Springs beds (Huber, 1980; Matrz and others, 2014) and the lower part of the Church Rock Member, although these do not exceed a few centimeters in diameter (pebble-size) and are never in greater abundance than the reworked sedimentary clasts. Conglomeratic beds and many sandstone beds are well-cemented and resistant ledge-forming units.

Sedimentary structures in sandstone beds of both the Kane Springs beds and Church Rock Member are dominated by horizontal planar bedding (figure 12G) (Blakey, 1978; Martz and others, 2014), although festoon stratification and climbing ripple cross-stratification are also extremely common (figure 12F). Scour surfaces, trough cross-bedding, planar cross-bedding, and mud cracks occur more rarely. Lateral accretion beds occur locally (Blakey, 1978; Martz and others, 2014).

In Lisbon Valley, the Kane Springs beds consist of lower and upper ledge-forming multistoried conglomeratic sandstone separated by a more mudstone-dominated middle unit as is seen elsewhere (Blakey and Gubitosa, 1984; Hazel, 1994; Martz and others, 2014). The lower ledge-former occupies scours incised into the Cutler Formation, whereas the upper unit is more laterally continuous. The more mudstone-dominated middle Kane Springs beds locally contain fissile mudstone possibly representing lacustrine or paludal deposition (figure 12J) (Martz and others, 2014), similar to that seen in the Monitor Butte Member (Dubiel and Hasiotis, 2011). Unoxidized uranium, vanadium, and sulfide minerals are locally concentrated in sandstone in the Kane Springs beds, both in grain interstices and as clast replacements; this mineralization is especially pronounced in the lowest ledge-forming sandstone, particularly north and southeast of the study area (e.g., Chenoweth, 1990). Blakey (1978) reported limestone beds from the Kane Springs beds, although we have not observed these.

The upper contact between the Kane Springs beds and Church Rock Member is gradational south of Big Indian Rock, but near Big Indian Rock Road a prominent dark reddish-brown conglomeratic sandstone is lithologically similar to the ledge-forming upper and lower Kane Springs beds. Martz and others (2014) assigned this unit to the black ledge beds (figure 12D), a modification of the term Black Ledge applied to the dark conglomeratic sandstone locally occurring at this level in the salt anticline region (Stewart and others, 1972a; Blakey and Gubitosa, 1983; Hazel, 1994).

Above the black ledge beds, the bulk of the lower Church Rock Member is reddish-brown, locally massive, ripple cross-laminated (and sometimes horizontal planar bedded) siltstone and very fine grained sandstone with some greenish-gray mottling. These strata are interpreted here as overbank deposits. However, slightly more complex and coarse-grained ledge-forming units, generally less than 15 m thick, commonly occur 20 to 25 m below the red ledge beds, including at Skull Ridge (figure 12D). These sandstone beds tend to be very fine to fine-grained, interbedded with minor amounts of conglomerate, dominated by horizontal planar-bedded and climbing ripple cross-lamination with some planar cross-bedding and scour surfaces, and usually have a somewhat lighter orange hue than the red ledge beds. These are interpreted as being deposited by channel systems prior to deposition of the red ledge beds (Martz and others, 2014).

The red ledge beds (figure 12D) are the most prominent ledge-forming sandstone beds in the Church Rock Member in Lisbon Valley other than the black ledge beds (Martz and others, 2014). The red ledge beds are generally single-storied and about 8 to 12 m thick. Although Huber (1980) suggested that the red ledge beds represent point bar deposits, Martz and others (2014) did not observe any lateral accretion sets, and the red ledge beds are generally similar to the Kane Springs beds in architecture and sedimentary structures. Martz and others (2014) therefore interpreted the red ledge beds as bedload-dominated braided channel deposits. Bimodal grain-size distribution, mud cracks, and vertebrate tracks (e.g., Milner and others, 2006a, figure 1), suggest highly variable (and even ephemeral) flow and occasional subaerial exposure (Martz and others, 2014).

The thickness between the top of the red ledge beds and the base of the Wingate Sandstone is about 30 m at Skull Ridge (figure 12D). The upper part of the Church Rock Member contains abundant ripple cross-laminated sandstone, often directly overlying the red ledge beds that are probably proximal overbank deposits. The upper part of the Church Rock Member also contains slightly more gravel than the lower part of the Church Rock Member, and these conglomerates are interpreted as small bedload-dominated channel systems.

At Skull Ridge, a distinctive sandstone unit about 1.5 m thick locally occurs about 3 m below the Wingate Sandstone, below the variegated sandstone. This bed is tabular, almost structureless, well sorted, very fine grained, and weathers with conchoidal fracture. Similar sandstone in the Owl Rock Member have been interpreted as loessites by Dubiel and Hasiotis (2011, p. 403), and may indicate more arid conditions than observed lower in the Church Rock Member.

Paleosols are only occasionally encountered in the Church Rock Member in Lisbon Valley but occur locally in finer grained sediments (usually siltstone and very fine grained sandstone) interpreted as overbank deposits. The Lisbon Valley paleosols are generally poorly developed or weakly calcified (figure 12K) (Martz and others, 2014), consistent with paleosols observed elsewhere in the non-bentonitic mudstone beds of the Dolores Formation, and Rock Point and Church Rock Members of the Chinle Formation (Blodgett, 1988; Prochnow and others, 2006; Tanner and Lucas, 2006; Cleveland and others, 2008; Dubiel and Hasiotis, 2011). Greenish-gray mottling in reddish-brown overbank siltstone and very fine grained sandstone are extremely common. In some cases, these mottles are nearly spherical haloes varying from a few millimeters to centimeters in diameter, suggesting that they nucleated around tiny particles of organic matter that generated local reducing conditions. More irregular and often elongate greenish-gray mottles most likely formed around burrows and roots that generated localized reducing conditions (e.g., Cleveland and others, 2008).

However, clear evidence of distinct A or B horizons (including well-calcified Bk horizons) within the Church Rock Member is extremely rare. As a result, Martz and others (2014) considered most of the overbank deposits to be entisols, inceptisols, or protosols (figure 12K) (sensu Mack and others, 1993). Some intervals can be recognized in which overbank deposits weather into spheroidal shapes suggesting nucleated calcification, sometimes associated with greenish-gray mottling, desiccation cracks, rhizoliths, and/or burrows. These were tentatively classified by Martz and others (2014) as aridisols or calcisols (e.g., Dubiel and Hasiotis, 2011). At least locally in the uppermost Church Rock Member, immediately below the Wingate Sandstone, blocky peds and arcuate slickensides are developed into overbank siltstone beds, and were interpreted by Martz and others (2014) as vertisols. Vertisols are generally rare in the uppermost part of the Chinle Formation compared to lower bentonitic units such as the Petrified Forest Member (e.g., Therrien and Fastovsky, 2000; Dubiel and Hasiotis, 2011).

A light greenish-gray band, possibly indicating reducing conditions, frequently occurs at the base of the Wingate Sandstone, and mud cracks also occur locally at the contact, leading Martz and others (2014) to conclude that the Wingate erg was initially deposited on a ground surface that was at least locally saturated. In Lisbon Valley, the sharp erosive contact is most discernible at the base of discontinuous multistoried lenses of interbedded sandstone and conglomerate, referred to by Martz and others (2014) as the "Big Indian Rock sandstones" (figure 12E), which occur locally at the Chinle-Wingate contact (Lucas and others, 1997a, 1997b). These sandstone beds are similar in color and superficial appearance to the eolian deposits of the Wingate Sandstone, but distinct from them in containing conglomeratic layers and more variable bedding.

At Skull Ridge, the Big Indian Rock beds are mostly very fine to fine-grained and horizontally planar bedded, although climbing ripple cross-stratification and highly convoluted bedding produced by syndepositional slumping also occurs. The Big Indian Rock beds at this locality comprise several superimposed sandstone sequences separated by basal scours; at the base of each sequence is an intrabasinal conglomerate that contains clasts of reworked sandstone and siltstone up to 10 cm in diameter. These conglomerates locally contain isolated vertebrate bones.

Martz and others (2014) agree with the interpretation of Lucas and others (1997a, 1997b) that the Big Indian Rock beds are probably fluvial deposits given the lenticular nature of the unit and the combined characteristics of conglomerates, trough cross-bedding, and climbing ripple cross-lamination. The thinness of individual tiers, combined with the conglomeratic bases and absence of distinct lateral accretion beds, suggests that they were deposited by small bedload-dominated channels that preceded deposition of the Wingate erg system. These lenses were probably deposited in localized scours incised into the top of the Church Rock Member. The presence of conglomeratic layers and lack of trough cross-stratification differentiates them from ephemeral stream deposits found within the Wingate Sandstone (Clemmensen and others, 1989), although they are similar to these other Wingate fluvial deposits in being dominated by very fine to fine-grained sand and possessing abundant climbing ripples.

The Big Indian Rock beds (figure12E) generally have a conformable relationship with the overlying Wingate Sandstone (Martz and others, 2014). Although Lucas and others (1997a, 1997b) claimed that the contact between these lenticular sandstone beds and the Wingate Sandstone is gradational, there is usually a discernible sharp contact between them. However, the contact appears to be conformable (i.e., there is not a clear erosional scour incised into the Big Indian Rock beds), and probably does not represent a great deal of missing time.

Paleontology

At Lisbon Valley, abundant petrified logs occur in the lower Kane Springs beds (Martz and others, 2014), and flattened permineralized wood occurs near the top of the Church Rock Member, as do foliage compressions of the giant horsetail *Neocalamites*, the fern *Cynepteris*, the bennetitalean *Zamites*, and the enigmatic seed plants *Pelourdea* and *Sanmiguelia* (Ash, 1987; Milner, 2006). Invertebrates include ostracods, conchostrachans, and rare gastropods from the Kane Springs beds; invertebrate trace fossils are also present (A.R.C. Milner, unpublished data; Martz and others, 2014).

Coelacanth material, much of it referable to *Chinlea*, is known from the Kane Spring beds (figure 5A) (Schaeffer, 1967; Milner and others, 2006a). Lisbon Valley is well known for its spectacular collection of ray-finned fishes (figures 5A and 5D to 5F) (redfieldiid palaeonisciforms, perleidiforms, and semionotiforms) including several holotypes from ledge-forming sandstone beds in the Church Rock Member (figures 5A and 5D to 5F) (Schaeffer, 1967; Milner and others, 2006a; Gibson, 2013a, 2013b, 2015). Excellent ray-finned fish fossils, including the lobe-finned *Chinlea*, are also known from the correlative Dolores Formation in southwestern Colorado (Schaeffer, 1967; Elliott, 1987). Fragmentary metoposaurid material has been recovered from the Kane Springs beds and Church Rock Member in Lisbon Valley (Martz and others, 2014; A.R.C. Milner, unpublished data).

Phytosaur postcranial material has been recovered from the Kane Springs beds in Lisbon Valley, although it cannot be referred to an alpha taxon (Milner and others, 2006a; Martz and others, 2014). Several specimens of the derived phytosaur *Machaeroprosopus*, including some specimens referable to "*Redondasaurus*," occur in the Church Rock Member and overlying Big Indian Rock beds at the base of the Wingate Sandstone (figures 7B to 7D) (Morales and Ash, 1993; Lucas and others, 1997a, 1997b; Martz and others, 2014). The so-called "last phytosaur" (figure 7D) from the Big Indian Rock beds at the base of the Wingate Sandstone in Lisbon Valley is the stratigraphically highest known phytosaur in the western United States (Morales and Ash, 1993; Lucas and others, 1997a, 1997b; Martz and others, 2014). Putative phytosaur tracks and swim traces have also been recovered from the Church Rock Member at Lisbon Valley (Milner and others, 2006a Milner and Lockley, 2016; Hunt-Foster and others, 2016; A.R.C. Milner, unpublished data).

Multiple specimens referable to the aetosaur *Typothorax* were recovered from the upper Kane Springs beds in Lisbon Valley (figure 8B) (Martz and others, 2014), and the poorly preserved aetosaur osteoderm from Canyonlands National Park described by Heckert and others (1999) was probably recovered from the middle Kane Springs beds (see above). Specimens of *Typothorax*, including at least one referable to *T. coccinarum*, also occur within the Kane Springs beds and Church Rock Member at Lisbon Valley (figures 8C and 8D) (Martz and others, 2014). An excellent *Brachycheirotherium* trackway has been recovered from the upper Church Rock Member (Milner and others, 2006a); this ichnotaxon may have been made by *Typothorax* (Heckert and others, 2010). Undescribed crocodylomorph material has also been recovered from the Church Rock Member at Lisbon Valley (A.R.C. Milner, unpublished data), as have *Grallator* tracks indicating the presence of theropod dinosaurs (Milner and others, 2006a).

Stop 6
The San Rafael Swell, Black Dragon Overlook

From Lisbon Valley, return west along Big Indian Road to U.S. Highway 191, and proceed north through

Moab to I-70. Take I-70 west. Approximately 65 km (40 mi) after joining I-70, the highway approaches the steeply rising slopes of the San Rafael Swell, a large double-plugging anticline with a 70 km (44 mi) long axis oriented approximately southwest to northeast (figure 4). The cliff-forming Middle Jurassic Entrada Sandstone caps the reddish-brown Carmel Formation on the lower east-tilting flanks of the Swell, whereas the pale-colored sandstone of the Early Jurassic Navajo Sandstone form most of the sloping surface (figure 13A). After passing west through the road cut lined by sheer cliffs of Navajo Sandstone on either side, the cliffs above are now capped by the Early Jurassic Kayenta Formation and cliff-forming Wingate Sandstone, with the Chinle Formation occurring immediately below, and the reddish and orangish rocks of the Lower Triassic Moenkopi Formation forming the slope below (figure 13B). Exit at the Black Dragon Canyon overlook, about 82 km (51 mi) after exiting U.S. Highway 191 to join I-70. The east-facing exposures along the west side of the San Rafael Swell (figure 13B) represent the Spotted Wolf Canyon section of Beer (2005); the Buckhorn Wash (U-3) section of Stewart and others (1972a) is about 26 km (16 mi) to the north.

About 19 km (12 mi) farther west of the Black Canyon overlook on I-70, exit north on County Road (CR) 332 (Buckhorn Draw Road). This exit affords a view of a mesa capped by the Chinle Formation just north of the I-70 (figures 13C to 13F) just south of the small hill called the Wickiup, and just west of Stewart and others' (1972a) Cane Wash (U-4) section. Although we have not had an opportunity to closely examine the Chinle Formation on the eastern flank of the San Rafael Swell, we were able to briefly examine the section south of the Wickiup.

Sedimentology and Stratigraphy

The Chinle Formation in this part of the San Rafael Swell is about 106 m thick (Stewart and others, 1972a; sections U-3 and U-4). The formation is dominated by a massive cliff-forming sandstone that could easily be mistaken at a distance for part of the Wingate Sandstone (the "Moss Back Member" in figures 13B and 13D), locally with well-developed paleosols at the base (figures 13D and 13E), and overlain by a sequence of interbedded conglomeratic sandstone and mudstone. The stratigraphic identity of these units is open to debate.

The paleosol at the base of the Chinle Formation is the Temple Mountain Member (Robeck, 1956), a sequence of mottled strata that can locally be as much as 30 m thick but is usually 6 to 9 m thick (Robeck, 1956; Stewart and others, 1972a). As indicated by Beer (2005), there are two distinct paleosol zones within the Temple Mountain Member (figure 13E). The lower paleosol is a quartz sandstone with "four-color" mottling that was identified by Beer (2005) as a gleyed oxisol, with vertical streaks interpreted as crayfish burrows that seem to be identical to those often occurring in the mottled strata above, within, or below the Shinarump Member in other parts of southern Utah and northern Arizona (e.g., figure 10D). Above the lower mottled paleosol is a mudstone-dominated drab-colored paleosol identified by Beer (2005) as a gleyed vertisol.

The overlying massive ledge-forming sandstone varies in thickness in the San Rafael Swell from about 18 to 51 m (Stewart and others, 1972a), although Beer (2005) indicates the unit can be up to 100 m thick. This ledge locally truncates the Temple Mountain Member, and is overwhelmingly dominated by pale-yellow siliceous sandstone with interbedded mudstone and conglomerates composed mostly of quartz and quartzite with some chert and reworked sedimentary clasts (Stewart and others, 1972a; sections U-3 to U-6). Dominated by siliceous sand and conglomerate, the unit is lithologically similar to the Shinarump Member. This, combined with the position of the unit at the base of the section, would suggest the unit belongs to the Shinarump Member as indicated by Lucas (1991). However, this unit was identified as an unusually siliceous representation of the Moss Back Member by Stewart and others (1972a), Blakey and Gubitosa (1983), and Beer (2005), who explain its basal position within the Chinle Formation in the San Rafael Swell as due to the thinning of the formation along the flanks of the ancestral Uncompahgre highlands, where the lower units of the Chinle Formation have mostly pinched out (figure 2A). Ultimately, the question cannot be answered without more careful tracing of the unit southwards to where unquestioned outcrops of both the Shinarump and Moss Back Mem-

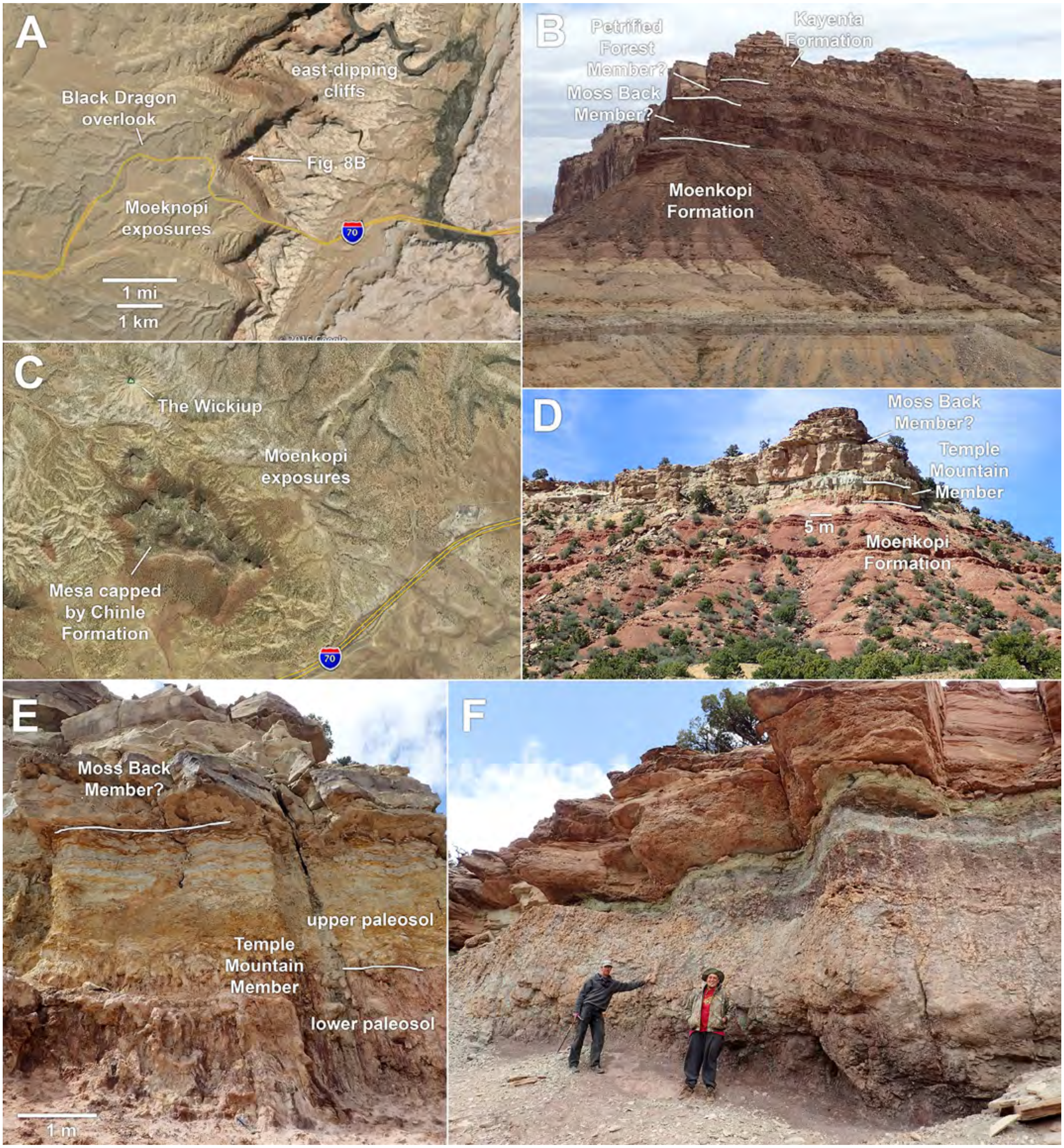

Figure 13. The northern San Rafael Swell. (A) Map showing location of Black Dragon Canyon overlook at the eastern edge of the San Rafael Swell. (B) Cliffs to the north of I-70 visible from Black Dragon Canyon overlook. (C) Map showing location of The Wickiup and the mesa to the south near the western side of the San Rafael Swell. (D) Mesa visible from pulloff by I-70 south of The Wickiup. (E) The Temple Mountain Member at same mesa, showing upper and lower paleosols. (F) The upper Chinle exposures on top of the mesa probably represent the Petrified Forest or Owl Rock Member. The well-developed calcareous paleosol below the ledge-forming conglomeratic sandstone is very similar to the paleosol developed underneath the Capitol Reef Bed in CARE.

bers occur. The presence of a siliceous mottled unit below the massive ledge-former supports the identification of this unit as the Moss Back Member, although it is worth noting that "four color" mottled strata occur below the Shinarump Member in the Circle Cliffs region to the southwest (J.W. Martz, unpublished data).

The relatively thin section of interbedded sandstone and mudstone overlying the cliff-forming sandstone (figures 13B and 13E) has been assigned to the Church Rock Member by Stewart and others (1972a), but Lucas (1991, 1993) correlated these beds to the Cameron, Moss Back and Rock Point Members. Although we have difficulty assigning any of these upper strata in the San Rafael Swell to any particular member, we at least agree with Lucas (1991) that these strata probably do not all belong in the Church Rock (or correlative Rock Point) Member. The Church Rock Member tends to be somewhat unimodal and dominated by reddish-brown siltstone to fine-grained sandstone, or bimodal with those fine-grain sizes interbedded with pebble-cobble conglomerate (Martz and others, 2014). Whereas, the sequence above the massive ledge-forming sandstone in the San Rafael Swell consists largely of purplish- and greenish-gray mudstone with well-developed carbonate nodule horizons interbedded with ledge-forming conglomeratic sandstone dominated by intrabasinal clasts of reworked siltstone and carbonate (Stewart and others, 1972a). These beds are more reminiscent of the Petrified Forest Member or even Owl Rock Member than of the Church Rock Member.

Paleontology

Little information exists on the paleontology of the Chinle Formation in the San Rafael Swell. In addition to crayfish burrows in the Temple Mountain Member (Beer, 2005), carbonized permineralized wood occurs in the massive ledge-forming "Moss Back Member" (Stewart and others, 1972a). In the beds above the "Moss Back Member," the intrabasinal conglomerates contain vertebrate bone fragments, including one tiny jaw with exposed alveoli that was observed on the mesa south of The Wickiup (J.I. Kirkland and J.W. Martz, unpublished data).

Stop 7
Chimney Rock Trail, Capitol Reef National Park

Return east along I-70, leaving the San Rafael Swell, and drive for about 11.4 km (7.1 mi) east of the Black Dragon Canyon overlook until reaching SR 24. Proceed south on SR 24 about 70.8 km (44 mi) to Hanksville, and from there continue west about 64 km (40 mi). This route passes through the Waterpocket Fold, a monoclinal fold that extends for approximately 160 km (100 mi) southeast from the edge of the San Rafael Swell to Lake Powell in GLCA. Passing through the Waterpocket Fold, SR 24 follows the Fremont River through a canyon lined by the sheer, east-dipping cliffs formed by the Navajo Sandstone to eventually enter the wider valley in the western part of the Waterpocket Fold at the town of Fruita. Here, as in the other localities visited on this tour, the cliffs are capped by the Wingate Sandstone, which lies stratigraphically beneath the Kayenta Formation and Navajo Sandstone. The slope below is formed by the Moenkopi and Chinle Formations (figures 14B and 14C). The Chimney Rock trailhead is encountered about 5.3 km (3.3 mi) west of the CARE visitor center on the north (right-hand side driving east) side of the road (figure 14A).

A dramatic fault is visible from the Chimney Rock Trail parking lot. To the right of the fault (including the Chimney Rock pinnacle), reddish-brown Moenkopi Formation is capped by the Shinarump Member of the Chinle Formation, whereas to the left of the fault section the Shinarump Member is exposed close to the level of the parking lot, and is overlain by the stratigraphically higher grayish-colored Monitor Butte Member. The trail leads up through the entire Chinle Formation section, passing through the Shinarump Member and up the steep slope formed by the drab-colored Monitor Butte Member. On top of the mesa, the trail continues past exposures on the northern (left) side of the reddish-colored Petrified Forest Member capped by the Capitol Reef Bed. Leaving the trail and climbing on top of the Capitol Reef Bed allows access to the Owl Rock Member and the base of the Wingate Sandstone.

Sedimentology and Stratigraphy

CARE (figures 14 and 15) and the adjacent Circle Cliffs region of GSENM contains one of the most

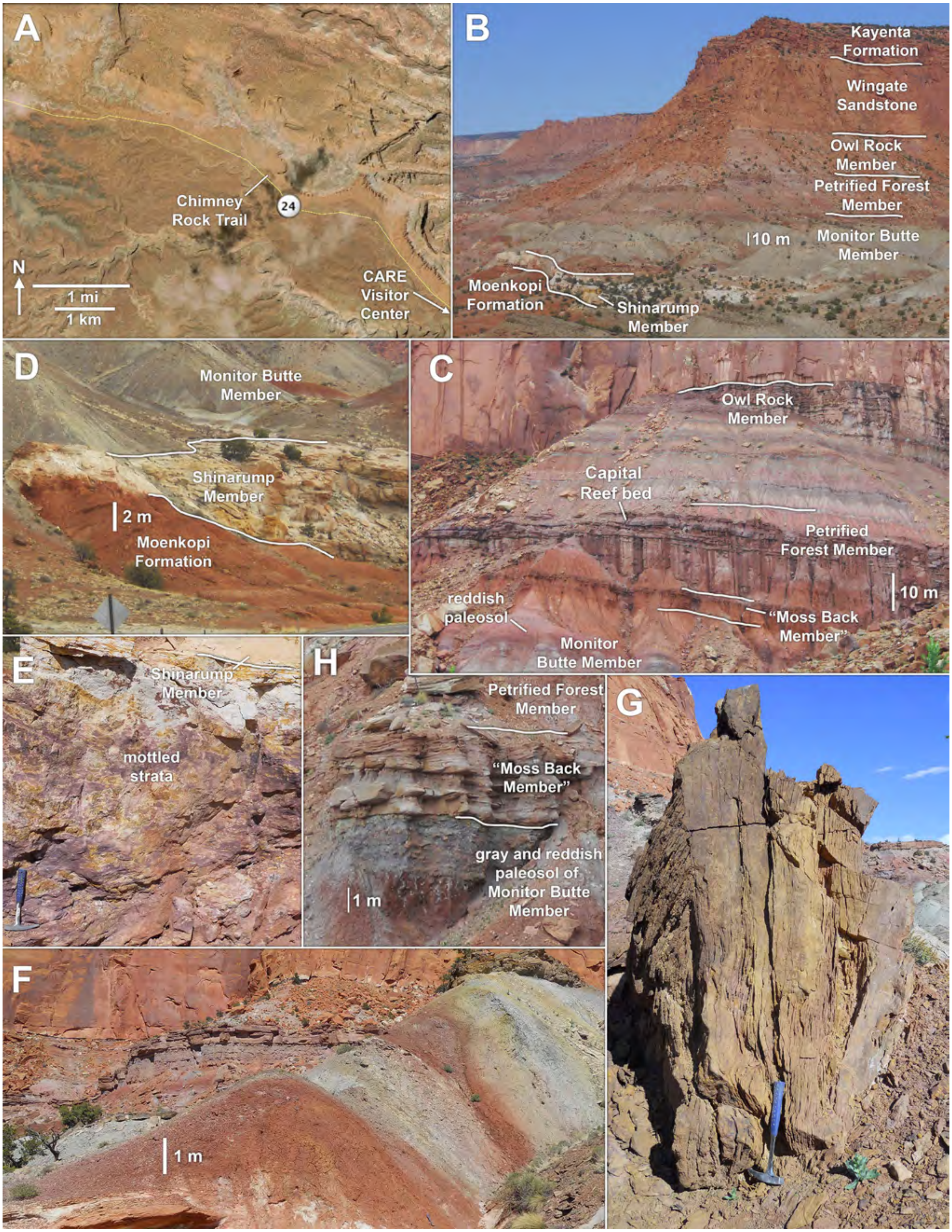

Figure 14 caption on following page.

Figure 14 (figure on previous page). CARE. (A) Map showing location of Chimney Rock Trailhead. (B) Cliffs just to the west of the Chimney Rock Trailhead showing the complete Chinle Formation section. (C) Exposures of the Chinle Formation along the scenic drive south of the visitor center; the color difference between the reddish-colored Petrified Forest Member and pastel-colored Owl Rock Member is especially striking; note the difference in appearance between the fresh outcrop of Owl Rock Member immediately below the Wingate Sandstone and the eroded slope of the same unit below. (D) The Shinarump Member near the western boundary of the park along SR 24, showing the variability in thickness of the unit due to erosion. (E) Mottled strata exposed below the Shinarump Member west of Chimney Rock Trail. (F) Contorted strata in the lower Monitor Butte Member; the strongly variegated beds dip to the right at almost 90°, although beds below and above are roughly horizontal. (G) Ripple cross-laminated sandstone bed with a dip of 90° in the contorted beds. (H) The carbonate nodule-rich paleosol at the top of the Monitor Butte Member, truncated by the "Moss Back Member."

complete Chinle Formation sections in southern Utah, and is located far enough from the flanks of the ancestral Uncomphagre highlands that the lower part of the section is well preserved (figure 2A). To the west (including in ZION; see below), the upper part of the section becomes increasingly erosionally truncated at the base of the Glen Canyon Group (figure 2A), but in CARE only the Church Rock Member is absent. The Shinarump, Monitor Butte, Petrified Forest, and Owl Rock Members are all well exposed in CARE (figures 14B and 14C), and a conglomeratic sandstone occurs at the base of the Petrified Forest Member that may be correlative with the Moss Back Member (Kirkland and others, 2014b).

In CARE, the Shinarump Member overlies the Tr-3 unconformity truncating the Moenkopi Formation. The Shinarump Member in CARE is a well-cemented white and yellow conglomeratic sandstone with abundant clasts of quartz, quartzite, and chert (Kirkland and others, 2014b). The Shinarump Member forms a distinct ledge-forming unit (figure 14D) that varies from 0 to 25 m in thickness (Dubiel, 1987a). At CARE the contact with the overlying Monitor Butte Member is disconformable (Stewart and others, 1972a), and the thickness of the Shinarump Member is more controlled by the degree to which it is incised below the Monitor Butte than the degree to which it incises the underlying Moenkopi Formation (Beer, 2005; Kirkland and others, 2014b). In places, the Shinarump Member is almost completely eroded away, that resulting in deep scours on the surface of the unit into which the overlying Monitor Butte Member was deposited.

Unlike most areas in southern Utah and northern Arizona, in CARE and at the Circle Cliffs, the "four-color" mottled strata that are usually developed on top of (and sometimes within) the Shinarump Member occurs below it (figure 14E) or not at all. Mottled strata have been observed beneath un-mottled Shinarump Member, and at the Moenkopi-Chinle contact in places where the Shinarump is absent (Kirkland and others, 2014b). As with the mottled strata in ZION and elsewhere (Martz and others, 2015), reddish-colored silicified layers locally occur in the mottled strata at CARE (Kirkland and others, 2014b).

At CARE, the Monitor Butte Member is a complex unit in which generally drab grayish and greenish-colored mudstone and muddy sandstone beds (figure 14B) are interbedded with sandstone and conglomerate that are often resistant and reddish-brown. In some places, including in the vicinity of the Chimney Rock Trail, the lower part of the Monitor Butte Member is strikingly contorted, with some sandstone and conglomerate beds dipping at up to about 90° (figures 14F and 14G). These folds are clearly syndepositional, because they are not only capped by undeformed strata, but overlie undeformed, horizontally oriented strata (Stewart and others, 1972a; Dubiel, 1991; Kirkland and others, 2014b). In contrast to the upper, more uniformly drab-colored part of the Monitor Butte Member, the contorted strata low in the member are strongly variegated. In these variegated beds, dark gray and greenish-gray mudstone beds are interbedded with yellowish-, orangish-, and reddish-colored beds (usually sandstone or conglomerate) (figure 14F). There are abundant anhydrite, calcium carbonate, and limonite concretions, the latter often with botryoidal habit. Gypsum often occurs as slickenside fillings. Sandstone beds in the lower contorted strata are often very fine to fine-grained and dominated by

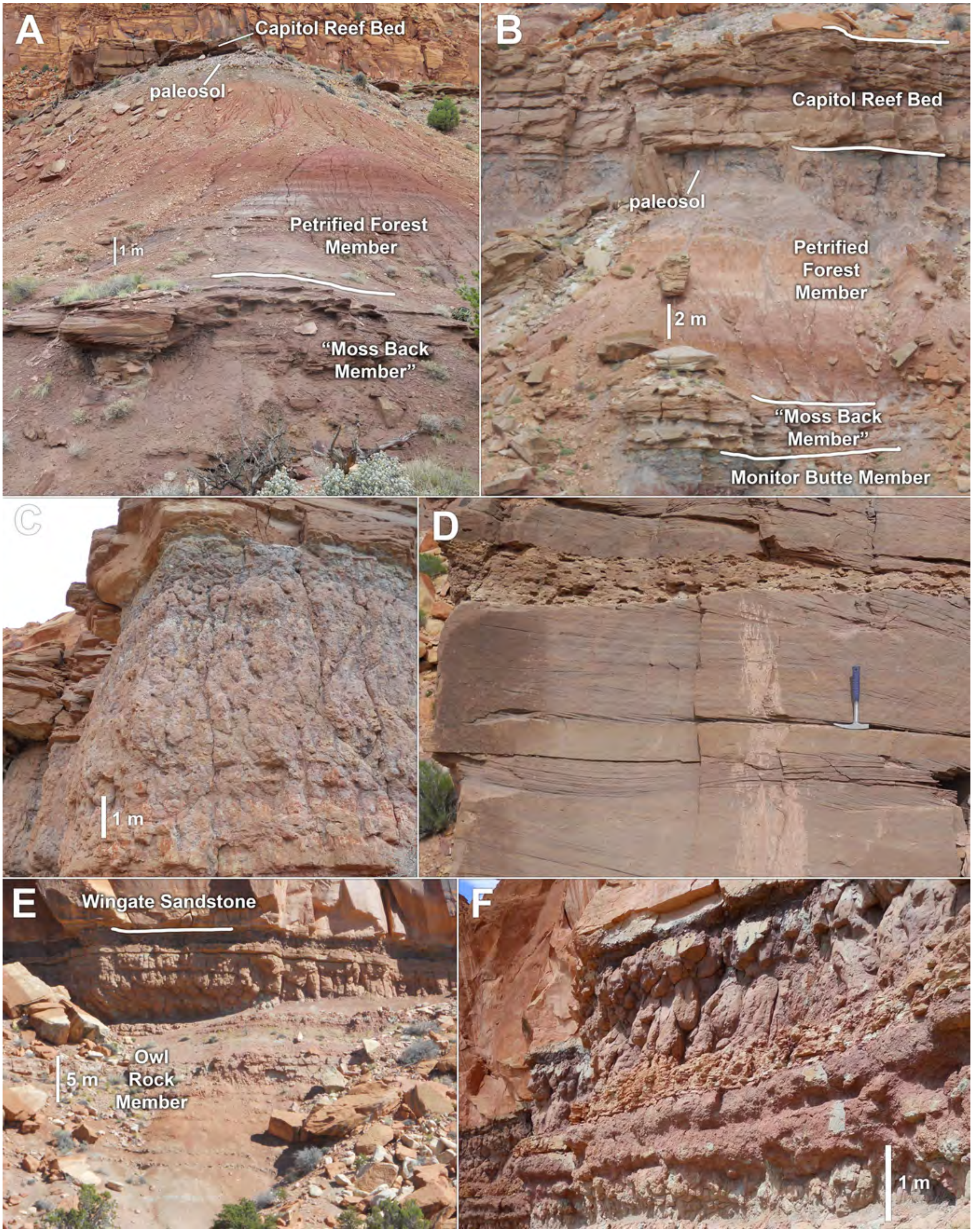

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Figure 15 (figure on previous page). CARE. (A) The lower Petrified Forest Member above the "Moss Back Member" along the Chimney Rock Trail. (B) Nearly the entire Petrified Forest Member near Chimney Rock in CARE; the purple beds immediately below the Capitol Reef Bed are the calcareous paleosol in figure 15C; the outcrop in figure 14C. (C) The calcareous paleosol below the Capitol Reef Bed. (D) Well-expressed bedding in the Capitol Reef Bed, mostly consisting of gently dipping planar cross-beds, although trough cross-bedding is visible just below and to the left of the rock hammer. (E) Owl Rock Member exposures near Chimney Rock Trail, just above the exposures in A. (F) Fresh exposures of the Owl Rock Member (as in figure 14C) showing well-developed calcretes.

climbing ripple cross-lamination (Kirkland and others, 2014b). Above the contorted strata, two horizontal conglomeratic sandstone beds occur in the upper, drab-colored part of the Monitor Butte Member (Kirkland and others, 2014b).

The uppermost few meters of the Monitor Butte Member is a zone of red- and purple-mottled mudstone containing abundant carbonate nodules, sometimes concentrated into horizons (figures 14C and 14H). Fresh exposures of the upper red paleosol contain vertisols that can be identified by the presence of distinct slickensides, and sometimes contain vertical crayfish burrows (Kirkland and others, 2014b).

A medium- to coarse-grained sandstone with some interbedded conglomerate and mudstone dominated by horizontal planar bedding and planar cross-bedding frequently occurs at the base of the Petrified Forest Member at CARE and the Circle Cliffs, capping the upper red paleosol (figures 14C and 14H). Usually the unit is a ledge-former several meters thick, although locally it is friable enough to be a slope-former. The unit varies considerably in color. At CARE, the unit is frequently pinkish- or reddish-colored so that it is difficult to discern from the reddish-colored strata above and below it. We informally refer to this unit as the "Moss Back Member" due to its stratigraphic position at the base of the Petrified Forest (Stewart, 1957; Stewart and others, 1972a), although the true Moss Back Member occurs mainly in southern Utah to the east of CARE (Stewart and others, 1972a; Blakey and Gubitosa, 1983; Dubiel, 1987b).

The reddish-colored mudstone, sandstone, and minor conglomerate overlying the "Moss Back Member" in southern Utah are usually referred to the Petrified Forest Member (e.g., Stewart and others, 1972a; Kirkland and others, 2014b), although they are correlative to what is often called the upper Petrified Forest Member or Painted Desert Member in northern Arizona (e.g., Stewart and others, 1972a; Lucas, 1993; see Martz and others, 2015 for a discussion of the nomenclatural history of the Petrified Forest Member). At CARE and the Circle Cliffs (figures 14C to 14H, 15A, and 15B), the lower part of the Petrified Forest Member is dominated by mudstone with some thinner sandstone and conglomerate beds containing intrabasinal sedimentary clasts. Mudstone beds are commonly vertisols with slickensides, carbonate nodules, and greenish-gray mottling (Dubiel and Hasiotis, 2011).

The Capitol Reef Bed (figures 14C and 15A to 15D) (Stewart and others, 1972a) is a prominent fine- to coarse-grained ledge-forming sandstone with minor interbedded conglomerate that occurs high in the Petrified Forest Member in CARE and the Circle Cliffs. The unit is almost continuous across this region, although it varies considerably in thickness. The Capitol Reef Bed is usually a multistoried channel sandstone dominated by horizontal planar bedding and planar cross-bedding (figure 15D), although locally it contains abundant climbing ripple cross-stratification and has rare trough cross-beds (J.W. Martz, unpublished data). A well-developed, highly resistant calcareous paleosol with abundant crayfish burrows occurs locally below the Capitol Reef Bed in parts of CARE (figures 15B and 15C) (Kirkland and others, 2014b).

The contact between the Petrified Forest Member and the Owl Rock Member is placed at a subtle color change (figures 15B and 15C) from deeper reddish-brown bentonitic mudstone with popcorn weathering to paler pastel colors with harder mudstone generally lacking popcorn weathering (Kirkland and others, 2014b); elsewhere across the Colorado Plateau, Stewart and others (1972a), Dubiel (1993), and Martz and others (2012) also place the contact between the Petrified Forest and Owl Rock Members according to these criteria. The Owl Rock Member at CARE and Circle Cliffs is dominated by slope-forming mudstone, with drab-mottling and large slickensides and desiccation cracks that are often filled with gypsum. A few resistant ledges always occur in the Owl Rock Member, most of which are calcareous paleosols developed into the mudstone units; however, some ledges (especially high in the member) are conglomeratic sandstone. The calcareous paleosol ledges often have a brecciated texture, greenish-gray mottling, and patches that have been replaced by silica (Kirkland and others, 2014b).

Paleontology

CARE and the adjacent Circle Cliffs area of GSENM have produced one of the most important collections of vertebrate fossils in southern Utah. Fossils are known from all stratigraphic levels within the Chinle Formation, giving the CARE-GSENM region the greatest potential within southern Utah for developing a detailed biostratigraphic framework.

Permineralized wood is common throughout the Chinle Formation section, although the nature of preservation varies. The wood is white and well silicified in the Shinarump Member, whereas poorly preserved stumps in both prone and upright growth position occur in the Monitor Butte Member. Ash (1975) described horsetail, fern, conifer, and bennettitalean compressions from mudstone lenses in the Shinarump Member at CARE. Plant remains are also all exceptionally common in the Monitor Butte Member; permineralized logs occur as poorly preserved upright trunks often surrounded by orangish- and yellowish-colored oxides (probably limonite and carnotite), and better preserved yellowish-colored perminerlized logs that occur in the sandstone and conglomerate beds. Giant horsetail fossils occur locally in upright growth position not far the Chimney Rock Trail, and even more impressive specimens referable to *Equisetites* have been documented along South Draw, south of Capitol Gorge (Dubiel, 1987a, 1987b, figure 8; Kirkland and others, 2014b). Within the contorted beds, carbonized plant compressions are preserved, including excellent bennettitalean specimens (Kirkland and others, 2014b).

Permineralized logs also occur in the “Moss Back Member” at the base of the Petrified Forest Member, and from the Capitol Reef Bed near the top of the Petrified Forest in both CARE (Kirkland and others, 2014b) and the Circle Cliffs. Ash (1982) described compressions of the shrubby seed plant *Sanmiguelia* from the Owl Rock Member at CARE.

The drab-colored mudstone beds in the Monitor Butte Member that produce plant compressions have also produced conchostracans in CARE (Kirkland and others, 2014b), just as similar organic-rich mudstone in the Kane Springs beds have produced both plant compressions and small invertebrates (Milner, 2006; Martz and others, 2014). Unionid bivalves occur locally in the Capitol Reef Bed, and occur densely in at least one bed in the upper Owl Rock Member in CARE (Kirkland and others, 2014b). Probable crayfish burrows that may belong to the ichnogenus *Camboygma* (Hasiotis and Mitchell, 1993) occur in the paleosol at the base of the Capitol Reef Bed near Chimney Rock Trail, and other invertebrate traces have been observed in the upper Petrified Forest Member at CARE (Kirkland and others, 2014b).

A lungfish tooth plate has been recovered from the Owl Rock Member at CARE (Kirkland and others, 2014b). Fragmentary metoposaur and phytosaur remains are fairly common in the Monitor Butte Member at CARE. Teeth and fragmentary bones are often found in conglomerates, and one locality in the upper part of the Monitor Butte Member has produced well-preserved phytosaur postcranial elements after only preliminary prospecting (Kirkland and others, 2014b); dull-gray gypsiferous mudstone beds and bone preservation is reminiscent of the *Placerias* Quarry in northern Arizona. So far none of this material can be referred to an alpha taxon. A mandible fragment from a large metoposaurid is known from the top of the Capitol Reef Bed near Chimney Rock Trail, and other indeterminate vertebrate material has been recovered from that level. At least one paramedian osteoderm referable to *Typothorax* is known from the Petrified Forest Member at CARE (figure 8A), although the exact stratigraphic level is unclear (Kirkland and others, 2014b).

Stop 8
Circle Cliffs, Grand Staircase-Escalante National Monument

From the Chimney Rock Trail, there are two possible routes to the Circle Cliffs, both using Burr Trail, which extends from Boulder through Circle Cliffs in GSENM and the Waterpocket Fold in CARE. One can take SR 24 west out of CARE, and then turn south onto SR 12 to Boulder, turning east onto Burr Trail road and drive from there to Circle Cliffs. However, if driving a passenger vehicle without a trailer, we recommended taking the more scenic route from the west (figure 16A). From the Chimney Rock Trailhead, drive east 17 km (10.6 mi) along SR 24 to the junction with Notom-Bullfrog Road, and proceed south. After roughly 30 km (20 mi), this road meets the dipping Mesozoic strata of the Waterpocket Fold, and affords an excellent view of the Mesozoic section. The gray cliffs to the left (east) are Upper Cretaceous Mancos Shale overlain by coastal and floodplain strata of the Muley Canyon Sandstone, Masuk Formation, and Tarantula Mesa Sandstone (Kirkland and others, 2014b; Lawton and others 2014; Titus and others, 2016) and underlain by thin exposures of the Naturita Formation (formerly Dakota Formation; Carpenter, 2014) at the base of the cliffs, with spectacular variegated exposures of the Upper Jurassic Morrison Formation occurring along to the road. On the right (west) side, one sees the upper surfaces of the predominantly reddish-colored Middle Jurassic strata. In decending: the Summerville, Curtis, Entrada, Carmel, and Page Formations; beyond which lie the massive white cliffs of the Lower Jurassic Navajo Sandstone.

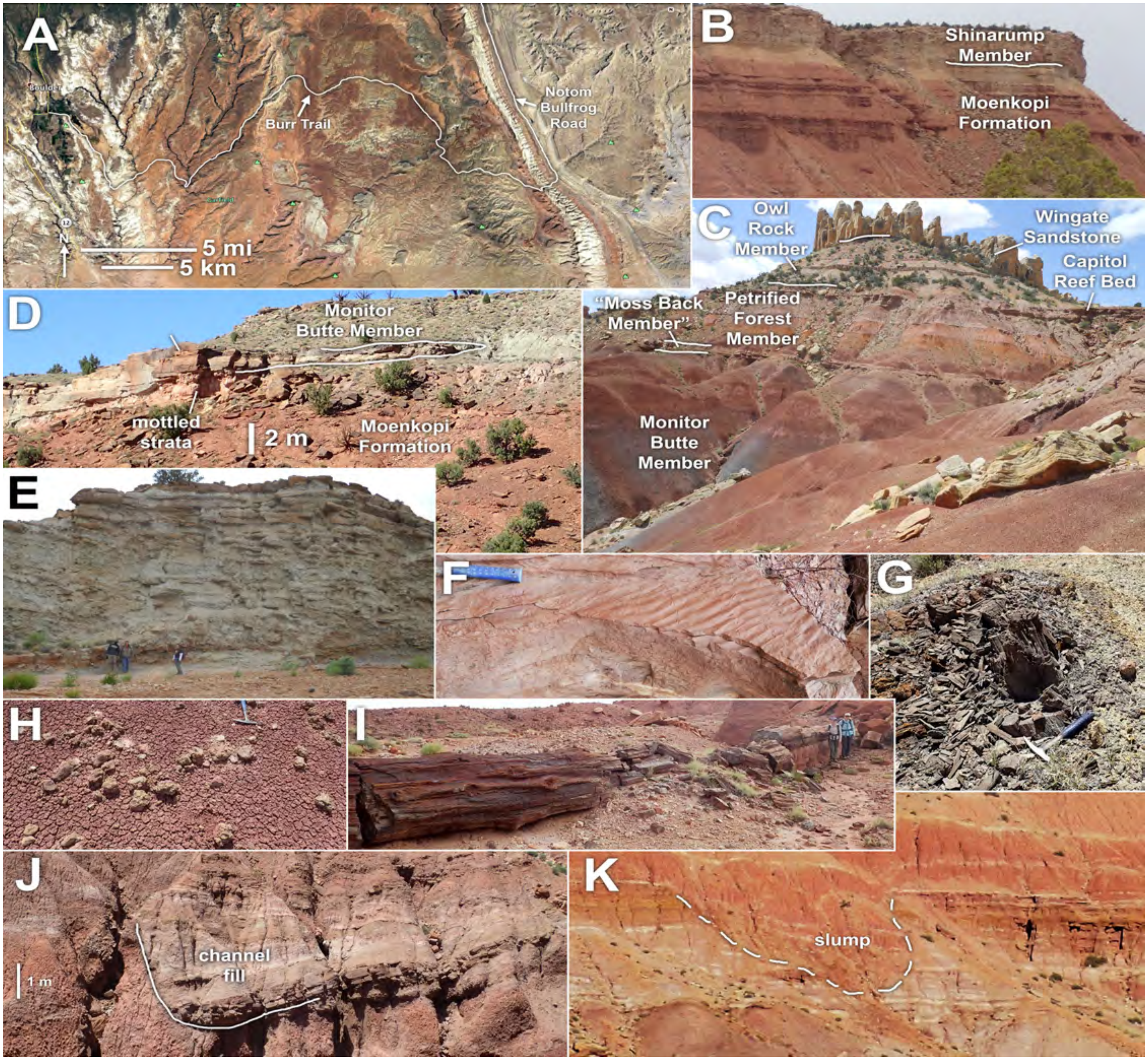
Roughly 50 km (30 mi) after turning onto Notom-Bullfrog Road, the road curves right (west), passing through the Navajo Sandstone and up the Burr Trail switchback (figure 16A). The switchback ascends more or less along the contact between the Navajo Sandstone and the reddish-brown Kayenta Formation. Upon reaching the top, the Burr Trail continues east, passing through the west-facing exposures of the Wingate Sandstone and Chinle Formation (Kirkland and others, 2014b). From here, the Burr Trail passes through exposures of the Moenkopi Formation flooring the central basin of the Circle Cliffs uplift for about another 24 km (15 mi) before reaching the west side. The Circle Cliffs are a topographic result of a domal anticlinal structure for which the Waterpocket Fold in CARE represents the eastern, west-dipping side. The Circle Cliffs expose the Upper Triassic section around their entire circumference. On crossing the central basin of the Circle Cliffs, the Burr Trail climbs upwards through the Shinarump Member (figure 16B), then through most of the Chinle Formation section before reaching the spectacular east-facing overlook (Stop 8) at about the Petrified Forest-Owl Rock Member contact, below cliffs of Wingate Sandstone (figure 16C). The overlook affords an excellent view of the Chinle Formation section.

Proceeding back east a short distance along Burr Trail, one can turn south onto the Wolverine Road, and proceed south about 10 km (6 mi) to the Wolverine Petrified Forest trailhead. A short hike along the trail passes through the Monitor Butte Member, including impressive exposures of lower Monitor Butte ripple cross-laminated sandstone beds, before reaching an area covered with black permineralized logs weathering out of the "Moss Back Member."

Sedimentology and Stratigraphy

The stratigraphy of the Chinle Formation in the Circle Cliffs of GSENM is broadly similar to that in CARE, which is not surprising given its close proximity. As in CARE the stratigraphic sequence consists of the Shinarump (figure 16B), Monitor Butte, "Moss Back," Petrified Forest, and Owl Rock Members (figure 16C), and the overall lithology of these units matches fairly well what is seen in CARE. The entire section is visible from the Burr Trail Circle Cliffs overlook; the Monitor Butte, "Moss Back," Petrified Forest, and Owl Rock Members form a continuous slope capped by the Wingate Sandstone, whereas the Shinarump Member caps a lower tier of mesas overlying the Moenkopi Formation.

As in CARE, the Shinarump Member varies enormously in thickness throughout the Circle Cliffs, and locally pinches out to nothing (figure 16D), although the unit is overall a more continuous resistant ledge former than in CARE (figure 16B). Also as in CARE, the upper contact of the Shinarump Member seems to be fairly sharp and at least locally disconformable, and



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mottling seems to occur locally below the Shinarump rather than above it. Interestingly, as in the San Rafael Swell there seem to be two distinct paleosol horizons below the Shinarump Member (e.g., Beer, 2005)—a lower paleosol with distinct "four-color" mottling, and an upper less mottled zone that in the Circle Cliffs is usually greenish gray (figure 16D) (J.W. Martz, unpublished data).

The Monitor Butte Member is also broadly similar to the unit in CARE, and also shows a degree of lateral variability within the Circle Cliffs. Locally, ripple cross-laminated sandstone beds are abundant in the lower Monitor Butte Member, most notably at a striking outcrop encountered on the Wolverine Petrified Forest trail (figure 16E). These reddish-colored ripple cross-laminated sandstone beds (figure 16F) are some-

Figure 16 (figure on previous page). The Circle Cliffs in GSENM. (A) Map of the Waterpocket Fold and Circle Cliffs, showing the location of Burr Trail and the Circle Cliffs overlook. (B) The Shinarump Member at the base of the Chinle Formation, which is often a ledge-forming sandstone capping the Moenkopi Formation. (C) The Chinle Formation section above the Shinarump Member near the Circle Cliffs overlook. (D) Outcrop where the Shinarump Member locally pinches out (it is not clear if this is due here to erosional truncation or thinning over an interfluve). (E) Outcrop of ripple cross-laminated sandstone in the lowermost Monitor Butte Member along the Wolverine Petrified Forest trail. (F) Ripple cross-laminated sandstone also showing ripples on the bedding plane. (G) Poorly preserved petrified tree stump in the Monitor Butte Member. (H) Carbonate nodules in the reddish-colored paleosol at the top of the Monitor Butte Member. (I) Permineralized log in the "Moss Back Member" at the Circle Cliffs. (J) Cross section through a channel-fill deposit in the lower Petrified Forest Member, showing a probable cut bank and point bar sequence. (K) Slumping at the level of the Capitol Reef Bed in the upper Petrified Forest Member.

times interbedded with interbasinal conglomerate and show a striking degree of variable dip across small areas, although they are not as dramatically contorted as in CARE. However, the senior author has had the opportunity to examine the Chinle Formation section not far from Circle Cliffs along Lake Powell in GLCA with graduate students Susie Wisehart of the University of Washington and Susie Hertfelder of the University of Nevada, Las Vegas. Here, there are beds in the lower Monitor Butte Member that are as strikingly contorted as in CARE; the Lake Powell contorted strata are unusual for the Chinle Formation in containing limestone beds.

As in CARE, the bulk of the Monitor Butte Member is grayish-colored mudstone and interbedded sandstone. Locally, reddish-colored zones occur within the Monitor Butte in the Circle Cliffs that are somewhat reminiscent of the reddish-colored zones seen in the Cameron Member at ZION. Also as in CARE, poorly preserved perminerlized logs (figure 16G) are often found in both upright and horizontal positions within the Monitor Butte Member at Circle Cliffs, and vertebrate fossils occur locally. As elsewhere in southeastern Utah, a well-developed reddish-colored paleosol with abundant large orangish-colored carbonate nodules dominates the uppermost several meters of the unit (figure 16H).

The upper contact of the Monitor Butte Member is strongly unconformable at Circle Cliffs, and has locally been observed to cut down deeply into the Monitor Butte Member, even locally entirely excising the reddish-colored paleosol. In most places, a somewhat resistant ledge-forming conglomeratic sandstone occurs at this contact that is interpreted as being the same "Moss Back Member" seen at CARE. Locally, this unit is a more-or-less uniform dark reddish-brown, but in other places it consists of a lower ledge-forming grayish-colored conglomeratic unit and an upper friable slope-forming orangish-colored unit as is seen in the Wolverine Petrified Forest and the surrounding area, where abundant permineralized logs (figure 16I) and less common vertebrate material occur close to the contact between the two units.

The Petrified Forest and Owl Rock Members are broadly similar to the units in CARE, and the Capitol Reef Bed, or at least a similar ledge-forming sandstone at roughly the same stratigraphic level as in CARE, also occurs in the upper Petrified Forest Member at Circle Cliffs. The lower Petrified Forest Member in Circle Cliffs generally contains a striking zone of bright-orange mudstone not usually observed in CARE, although a similar orangish-colored zone occurs at Lees Ferry in northern Arizona (J.W. Martz, unpublished data). Occasionally, dramatic fluvial cut-and-fill structures can be seen in the Petrified Forest (figure 16J) There is evidence of unusual syndepositional slumping in the Petrified Forest at Circle Cliffs that has not been observed elsewhere (figure 16K); the Capitol Reef Bed and strata immediately above are locally deformed by downward slumping that does not seem to impact strata either lower or higher in the section.

Paleontology

The preservation of plant material in the Circle Cliffs is overall very similar to what is observed at CARE. Upright-growing giant horsetail pith casts have been documented in the lower Monitor Butte Member at the Circle

Cliffs (J.W. Martz, unpublished data). As in CARE, poorly preserved permineralized logs, many in growth position, occur within the Monitor Butte Member (J.W. Martz, unpublished data), and permineralized logs are also common in the "Moss Back Member." At the Wolverine Petrified Forest (Ash, 2003), an unusually dense concentration of permineralized logs in the Circle Cliffs area, most logs occur in the "Moss Back Member," although well-preserved examples are also fairly common in the Capitol Reef Bed. These logs have a preferred southeast to northwest orientation, although many have an orientation that is roughly perpendicular to this (J.W. Martz, unpublished data).

The Circle Cliffs area is slowly yielding one of the most impressive collections of Upper Triassic vertebrate faunas in southern Utah outside of Lisbon Valley, most of it collected by Yale University. Currently, efforts are underway by the senior author to precisely constrain the Yale localities stratigraphically. Yale has collected abundant vertebrate material from the Monitor Butte Member at the Circle Cliffs, including a lungfish tooth plate, metopsaur material, and phytosaur material, although none of the latter can be assigned to an alpha taxon; recently students working with the senior author recovered another lungfish tooth plate from the Monitor Butte Member (figure 5B) (W.G. Parker and J.W. Martz, unpublished data).

Within southern Utah, the Circle Cliffs has produced the largest number of vertebrate specimens referable to an alpha taxon outside of Lisbon Valley (W.G. Parker and J.W. Martz, unpublished data). An astonishing nearly complete and fully articulated specimen of *Poposaurus gracilis* (figure 8E), and an associated crocodylomorph skeleton were recovered by Yale University (Parker and others, 2006; Gauthier and others, 2011; Schachner and others, 2011) from the lower Monitor Butte Member (J.W. Martz, unpublished data), and material referable to the aetosaurs genera *Calyptosuchus* and *Desmatosuchus*, as well as paratypothoracins, have also been recovered, probably from the Monitor Butte Member (W.G. Parker and J.W. Martz, unpublished data).

Stop 9
Black's Canyon, Zion National Park

From the Circle Cliffs overlook, drive west on Burr Trail for 28 km (17.5 mi) to the junction with SR 12 in Boulder. From here, proceed south on SR 12 for 141 km (87 mi) through Escalante, across the northern Kaiparowits Plateau, through Cannonville, and past Bryce Canyon on the Paunsaugunt Plateau to the junction with U.S. Highway 89. Titus and others (2016) described the geology and vertebrate paleontology along this route. Proceed south on U.S. Highway 89 for 70 km (43.2 mi) to the junction with SR 9. Proceed west on SR 9 for 40 km (25.1 mi) through the spectacular exposures of the Navajo Sandstone in the southeastern part of ZION and proceed into the town of Springdale. Turn right (northwest) onto Lion Boulevard and proceed approximately 1.2 km (0.75 mi) to the gate for the O.C. Tanner Amphitheater (figure 17A). Park by the striking purplish-colored beds of the upper Chinle Formation capped by the reddish-brown Moenave Formation (figure 17B).

In ZION, the Chinle Formation section is about 180 m thick, but only the uppermost 25 m or so is exposed at Black's Canyon (figure 17B). The lowermost exposures here are a gray, friable, slope-forming sandstone that may be correlative with the "friable sandstone" in the Mt. Kinesava section (see below). The predominantly purple and gray mudstone and interbedded sandstone beds of the uppermost Chinle Formation in Black's Canyon belong to what Martz and others (2015) refer to as the "purple pedogenic beds" (figure 17B). These beds may be equivalent to the upper Monitor Butte Member or Blue Mesa Member or even the basal Petrified Forest Member, although this is not clear (see below). The "purple pedogenic beds" are purple-weathering mudstone that are a variety of grayish-red, reddish-brown, and yellowish-gray shades in fresh outcrop. These beds are rich in nodules of calcium carbonate and anhydrite as well as non-horizontal veins of evaporite (figure 17C), all of which are absent lower in the section (Martz and others, 2015).

Anhydrite nodules also occur in the uppermost Chinle Formation at St. George (Kirkland and others, 2014a). The nodules are both scattered throughout the unit and concentrated into horizons, often with striking greenish-gray mottling, but there are also zones several meters thick where nodules are totally absent. The evaporite veins occur only in the uppermost 10 m or so of the purple pedogenic beds (figure 17C) and may

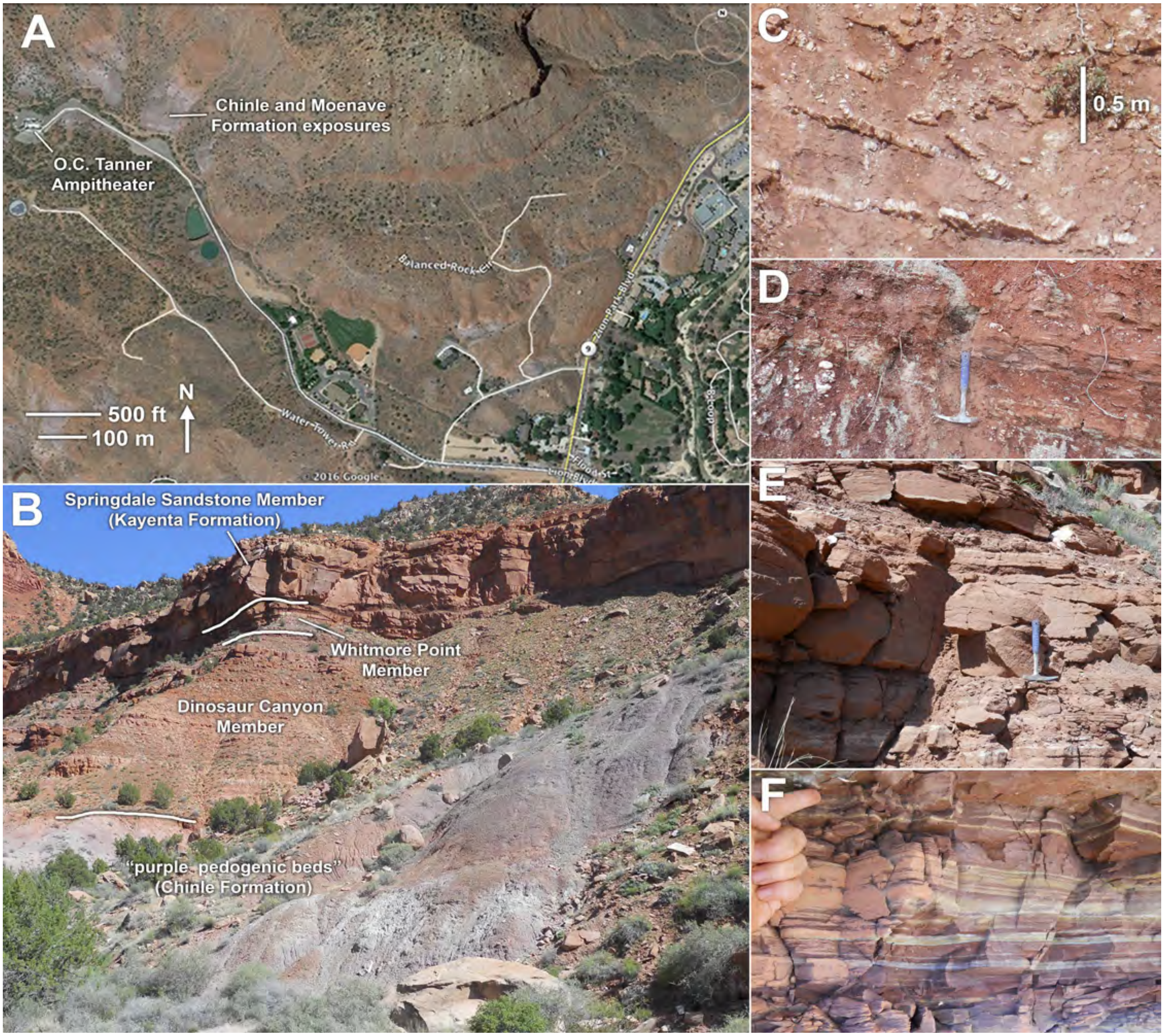

Figure 17. Black's Canyon in ZION. (A) Map showing location of Chinle Formation exposures. (B) Overview of the uppermost Chinle Formation and Moenave Formation exposures near O.C. Tanner Ampitheater. (C) Anhydrite or selenite beds in the uppermost Chinle Formation, just below the base of the Dinosaur Canyon Member. (D) The anhydrite-clast conglomerate marking the base of the Dinosaur Canyon Member. (E) Ledge-forming sandstone in the Dinosaur Canyon Member. (F) Laminated mudstone of the Whitmore Point Member under the overhanging Springdale Sandstone Member.

be composed of anhydrite or selenite. The fact that the veins are generally not perfectly horizontal in orientation suggest they formed as postdepositional diagenetic features when hypersaline groundwater seeped into the subsurface while the Dinosaur Canyon Member was being deposited (see below) and left gypsum in slickensides and desiccation cracks in the purple pedogenic beds. Very thin horizontal beds of gypsum also occur in the purple pedogenic beds, but these may be syndepositional (Martz and others, 2015). The purple pedogenic beds may be equivalent to the red and purple carbonate nodule-rich paleosol at the top of the Monitor Butte Member in CARE; if so, the unconformity between the Chinle and Moenave Formations may span about 18 million years of middle to late Norian and Rhaetian time.

Above the purple pedogenic beds of the Chinle Formation, the section is dominated by reddish-colored subhorizontal sandstone and mudstone of the Moenave Formation. The basal Dinosaur Canyon Member and overlying Whitmore Point Member of the Moenave are exposed here (figure 17B), and are in turn overlain by the Springdale Sandstone Member of the Kayenta Formation (Gregory, 1950; Biek and others, 2010; Milner and others, 2012; Martz and others, 2015). The section is essentially identical to that exposed at St. George (Kirkland and Milner, 2006; Kirkland and others, 2014a).

The base of the Dinosaur Canyon Member is distinctly unconformable, with a conglomerate composed of clasts of reworked white anhydrite and chert pebbles (figure 17D) as it is in other parts of southern Utah (Kirkland and Milner, 2006; Kirkland and others, 2014a). Immediately above is a fluvial sandstone bed also containing conglomerate with clasts of reworked anhydrite. In the lowermost few meters of the Dinosaur Canyon Member, reddish-colored mudstone contain horizontal anhydrite beds probably representing original deposition; as already discussed, the hypersaline conditions that produced these layers are likely also responsible for filling what were then subsurface slickensides in the uppermost Chinle Formation with evaporites. Above these lower mudstone and anhydrate beds, the Dinosaur Canyon Member is about 57 m thick and consists of complexly interbedded, predominantly reddish-brown mudstone and ledge-forming sandstone (figures 17B to 17E); as with the Church Rock Member along SR 128, the upper part of the unit becomes increasingly dominated by climbing ripple cross-lamination. The base of the Whitmore Point Member is a ledge of calcareous sandstone with nodules of reddish-colored chert (Martz and others, 2015) as elsewhere in southwestern Utah (Wilson, 1967; Kirkland and others, 2014a). The Whitmore Point Member is a thin unit only a few meters thick, that is composed of laminated mudstone (figure 17F), and capped by the thick ledge-forming Springdale Sandstone Member of the Kayenta Formation (figure 17B).

Stop 10
Southern Side of Mt. Kinesava, Zion National Park

Return down Lion Boulevard to SR 9 and turn right, proceeding south for 4.8 km (3 mi). Immediately after coming around a right curve, take a hard right onto Anazazi Way (figures 18A and 18B) which proceeds steeply upwards. Parking for the Chinle Trail is just to the east at the start of Anazazi Way near the base of the hill. It is also possible to drop people off farther up the hill. Proceed for approximately 1.8 km (1.1 mi) as Anasazi Way passes north and northwest through a housing complex. Just before the road curves back south, stop at the pull-out by the concrete bridge (figure 18B). In the distance to the north are the Chinle Formation exposures at the base of Mt. Kinesava (figures 18C and 19A). There is a trail that leads into the park from here, but proceeding instead towards the exposures along the bottom of the the wash affords the opportunity to examine outcrops of the Shinarump Member.

Stratigraphy and Sedimentology

Along the southern face of Mt. Kinesava near Springdale at the southern end of ZION, the entire Chinle section is preserved between the upper red member of the Moenkopi Formation and the Dinosaur Canyon Member of the Moenave Formation, and the Moenave is capped in turn by the Kayenta Formation and massive cliffs of the Navajo Sandstone (figure 18C). Only the lower part of the Chinle Formation is probably preserved here, and the upper part is truncated by erosion (figure 2A), and is capped unconformably by the Dinosaur Canyon Member (Martz and others, 2015). Nonetheless, the total thickness of the Chinle Formation preserved here is about 180 m (Martz and others, 2015), thicker than the entire Chinle Formation section in CARE (Kirkland and others, 2014a).

The basal unit of the Chinle Formation at ZION is the Shinarump Member (figures 18C and 18D), which is a thick ledge-forming siliceous conglomeratic sand-

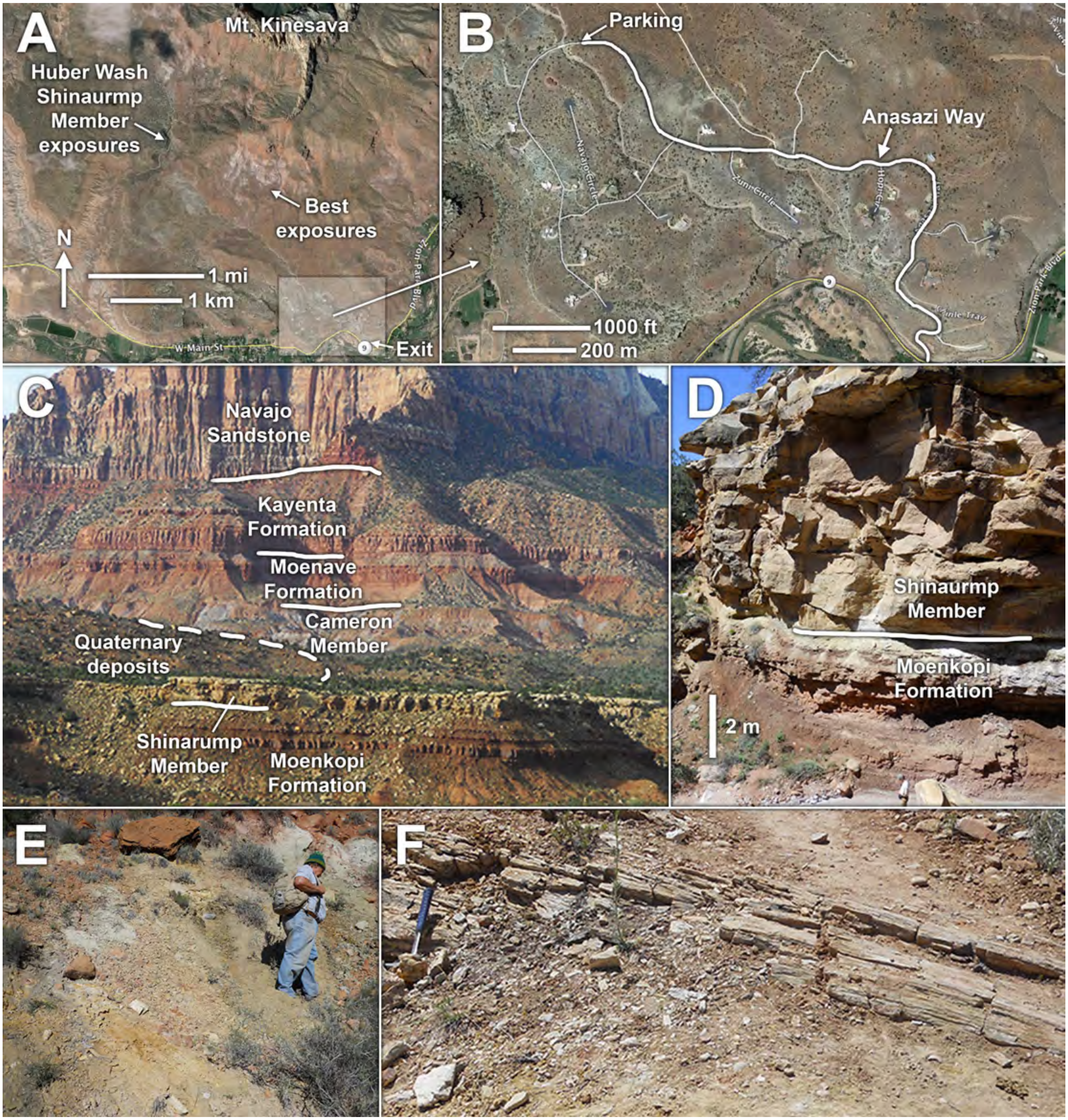

Figure 18. Mt. Kinesava in ZION and the surrounding region. (A) Overview of the area south of Mt. Kinesava, indicating the turnoff for Anasazi Way and the location of exposures of interest; the lightened box shows the area expanded in figure 18B. (B) Close-up of Anasazi Way showing the location of the parking area. (C) Overview of the Lower Mesozoic section at Mt. Kinesava, photographed from across the valley to the south. (D) The Shinarump Member exposed in Huber Wash on the west side of Mt. Kinesava. (E) "Lower sandstones" above the Shinarump Member exposed near Huber Wash showing the unusual orange mottling and abundant permineralized wood. (F) Permineralized log weathering out of the "lower sandstones" along Huber Wash on the Mt. Kinesava trail.

stone as seen elsewhere in southern Utah (e.g., Blakey and Gubitosa, 1983, 1984). In ZION, the Shinarump Member varies in thickness from 18 to 41 m (Biek and others, 2010), but is about 22.6 m thick where measured by Martz and others (2015) near Mt. Kinesava. The Shinarump is well exposed in drainages south of Mt. Kinesava, near the Chinle Trailhead just south of ZION, and even better exposed in Huber Wash on the western side of Mt. Kinesava (figures 18A to 18D). The sandstone and conglomerate beds in the Shinarump Member at ZION are generally poorly sorted, and dominated by siliceous extrabasinal clasts, especially multi-colored quartzite, although there are rare intrabasinal sedimentary clasts. Bedding within the Shinarump Member consists mostly of horizontal planar bedding, planar cross-bedding, and trough cross-bedding (Martz and others, 2015). The uppermost meter of the Moenkopi Formation below the Shinarump Member at ZION is a yellowish-colored zone of alteration that also occurs in CARE. Fossil logs are very common in the Shinarump Member at ZION (DeBlieux and others, 2005; Martz and others, 2015) as they are in the member elsewhere across the Colorado Plateau (e.g., Ash, 1975), and vertebrate remains are present but rare and unidentified (DeBlieux and others, 2005; Martz and others, 2015).

Unlike at CARE, the Shinarump Member does not have an erosionally truncated upper surface, but instead has a gradational contact with the overlying beds as in most areas of the Colorado Plateau. As in ZION, these overlying beds are generally slope-forming sandstone about 14 m thick (Martz and others, 2015) that Gregory (1950) referred to as the "lower sandstones" (figure 18E). These "lower sandstones" are predominantly siliceous as in the Shinarump Member, but differ in containing abundant dark-colored accessory minerals and in being rich in mudstone, organic material, gypsum, anhydrite, carnotite, iron oxide, and concretions (Martz and others, 2015); in these regards, the "lower sandstones" are somewhat similar to parts of the Monitor Butte Member, and have been considered correlative to it (Stewart and others, 1959; Doelling and Davis, 1989).

Gregory (1950) noted the striking similarity between the bulk of the Chinle Formation strata in ZION (figures 19A and 19B) and at Petrified Forest National Park (PEFO) in Arizona, naming the Petrified Forest Member for exposures at both parks, although a type section was not designated. Indeed, the Chinle Formation sediments Gregory (1950) assigned to the Petrified Forest Member also have striking similarities to the lower Chinle Formation seen farther southeast at Lees Ferry in the southern part of GLCA, and in the Navajo Nation between Lees Ferry and Cameron (Lucas, 1993; Martz and others, 2015). At Lees Ferry, Phoenix (1963) referred to these beds as the "sandstone and mudstone member," and Lucas (1993) later proposed the name Cameron Member for the same sediments at Cameron on the Navajo Nation in northern Arizona. Lucas (1993) and Martz and others (2015) also applied the name Cameron Member to the beds in ZION that Gregory (1950) referred to as the Petrified Forest Member, and the overall appearance and lithology of the beds at ZION are indeed strikingly similar to the Cameron Member in northern Arizona. In these areas, drab-colored mudstone (especially light olive-gray) dominate the sequence that are punctuated by reddish-brown zones that locally contain sandstone beds that grade laterally into mudstone (figures 19A and 19B).

These beds are also quite similar, and occupy roughly the same stratigraphic position, as the lower Blue Mesa Member of PEFO; it is very likely that this is what prompted Gregory (1950) to name the beds at ZION the Petrified Forest Member. In PEFO, the Newspaper Rock Bed is a prominent ledge-forming fluvial channel sandstone that grades laterally into reddish-colored, finer grained facies representing a floodplain paleosol (Demko, 1995; Parker, 2006; Trendell and others, 2010), and it is likely that the same is true of these reddish-colored bands seen at Mt. Kinesava.

This southeast to northwest-trending zone of Cameron Member/lower Petrified Forest Member from PEFO to ZION is much thicker than and lithologically distinct from the Monitor Butte Member. The Cameron Member is strikingly different from the Monitor Butte Member in having fewer organic-rich facies with abundant evaporites, oxides, and gypsum (although these do occur, especially low in the Cameron Member). The contorted strata often seen in the lower part of the Monitor Butte and Bluewater Creek Members have not yet been observed in the Cameron Member. We concur with Lucas (1993; Lucas and others, 1997a) and Irmis

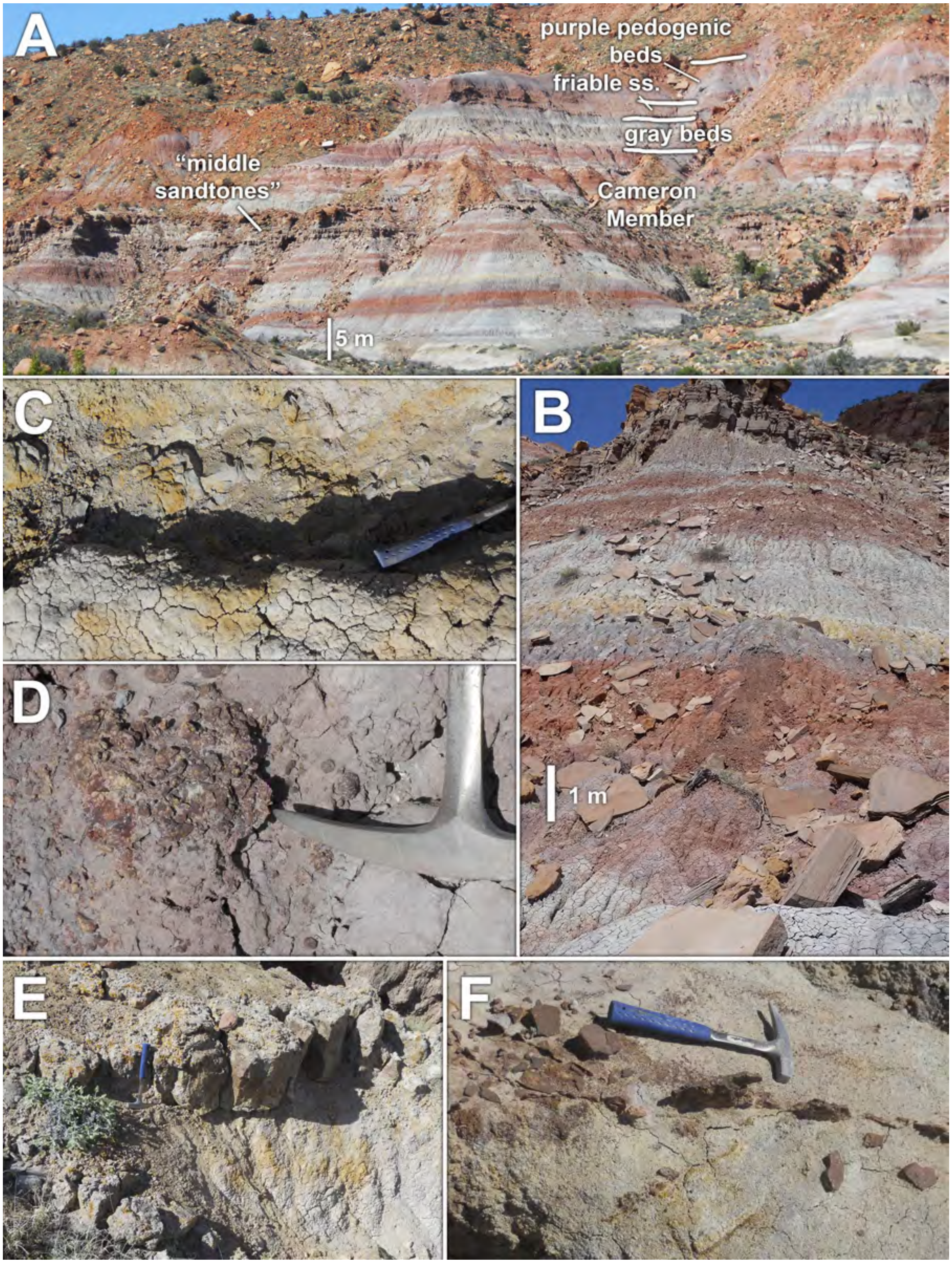

Figure 19. Mt. Kinesava in ZION and the surrounding region (continued). (A) Overview of the Chinle Formation on the south side of Mt. Kinesava. (B) Close-up of the exposures showing the thick reddish-colored zones and unusual yellowish-colored mottled horizons. (C) Close-up of yellowish-colored mottling. (D) Barite concretion at the base of one of the reddish-colored zones. (E) The "four-color" mottled zone, here a well-cemented ledge. (F) The "four-color" mottled zone with thin reddish-colored silicified beds probably filling slickensides.

and others (2011, supplemental data) that the Cameron, Bluewater Creek, and Monitor Butte Members should be considered distinct from each other.

The Cameron Member at Mt. Kinesava is about 110 m thick (Martz and others, 2015), and seems to have a gradational lower contact with the "lower sandstones." The unit consists predominantly of mudstone with interbedded very fine grained sandstone, and less-abundant claystone, coarser sandstone, and conglomerate; the latter two grain sizes mostly occur in ledge-forming channel sandstone beds. Slope-forming mudstone tend to be bluish-gray and greenish-gray, whereas coarser grained beds are often reddish-brown or yellowish-colored (figures 19B and 19C) (Martz and others, 2015). Mudstone beds in both reddish- and drab-colored zones are usually mottled with a variety of shades of yellowish-, reddish-, and greenish-gray, varying from large patches several centimeters across to thin stringers a few millimeters wide. These probably mostly represent root traces and possibly animal bioturbation as well. Although mottling is common in the Chinle Formation, abundant yellowish-colored mottling (figures 19B and 19C) is unusual. Slickensides are common in fresh exposures of mottled mudstone at ZION, indicating that these are probably vertisols. Concretions apparently composed of barite and iron oxides occur at the bases of some of the reddish-colored zones (figure 19D).

Five particularly thick reddish-colored zones are present in the Cameron Member at Mt. Kinesava (figures 19A and 19B); the upper part of the zone is often friable sandstone and the lower part is siltstone. The red bands likely represent paleosols that formed on interfluve sediments, as with the Newspaper Rock paleosol in PEFO (Trendell and others, 2013). Several thick ledge-forming sandstone beds with minor interbedded conglomerate are present that are referred to informally as the "middle sandstones" (figure 19A). These sandstone beds tend to be slightly coarser grained than most slope-forming sands in the Cameron Member. The "middle sandstones" are characterized by mostly horizontal planar bedding and planar cross-bedding, and have small burrows and some greenish copper mineralization.

On the southern side of Mt. Kinesava, the lowermost part of the Cameron Member consists of reddish-colored, largely ripple cross-laminated, very fine to fine-grained micaceous sandstone similar to those seen locally in the lower part of the Monitor Butte Member to the northeast at CARE and the Circle Cliffs, although the sandstone beds at Mt. Kinesava are not contorted into high-angle folds. These sandstone beds often contain burrows consisting of vertical holes and horizontal grooves less than 1 cm in diameter (Martz and others, 2015).

In the lower part of the mudstone-dominant sequence, about 30 m above these lower reddish-colored sandstone beds and just above the second thick reddish-colored zone, is a horizon of "four-color" mottled strata of gray, red, purple, and yellowish-orange colors (figures 19E and 19F). The horizon is usually well lithified, and often contains sheets of reddish-colored chert (figure 19F). In all of these characteristics, the mottled strata are strikingly similar to the mottled paleosols usually developed on top of the Shinarump Member in northern Arizona and parts of Utah (e.g., Dubiel and Hasiotis, 2011; Irmis and others, 2011, supplemental data). However, as this mottling occurs well above the top of the Shinarump Member, it likely does not represent the same episode of paleosol formation, unless the rate of sedimentation following Shinarump deposition was much greater here prior to formation of the paleosol. Given how thick the Cameron Member is here compared to lower Chinle strata in other parts of southern Utah, this possibility cannot be discounted.

Above the Cameron Member are mudstone-dominated strata (figure 17A) that have similarities to both the upper part of the Blue Mesa Member at PEFO (e.g., Martz and Parker, 2010) and the uppermost portion of the Monitor Butte Member elsewhere in southern Utah. The beds immediately above the Cameron Member consist mostly of gray mudstone with one thick red sandstone bed (the "friable sandstone" of Martz and others, 2015). The uppermost 25 m of the Chinle Formation consists of the "purple pedogenic beds"— a purple mudstone with abundant carbonate and sulfate nodules and sulfate veins, as is seen in Black's Canyon and throughout much of the region (Kirkland and Milner, 2006; Milner and others, 2012; Kirkland and others, 2014a). As discussed above, these beds are somewhat similar to the uppermost part of the Monitor Butte

Member at CARE and the Circle Cliffs. The Moenave Formation section has not been examined in detail by several of the authors, but thick horizontal beds of gypsum occur in the lower part of the Dinosaur Canyon Member as seen in Blacks Canyon.

Paleontology

As with the Shinarump Member, the "lower sandstones" contain abundant permineralized wood (figure 18F) and rare vertebrate material, at least some of which has been identified as phytosaurian (DeBlieux and others, 2005; Martz and others, 2015). Preservation of permineralized wood is similar to that seen in the Monitor Butte Member in that trees are poorly preserved, and often associated with carnotite and iron oxides.

Although permineralized wood is common in the Shinarump Member and in the overlying "lower sandstones," it is surprisingly rare in the Cameron Member given that wood is common at the Cameron Member type section in Arizona (J.W. Martz and W.G. Parker, unpublished data). This may be due to the relatively fine-grained nature of the unit at ZION; in the Chinle Formation well-preserved permineralized logs tend to occur in medium- to coarse-grained conglomeratic channel units, which are a relatively minor component of the Cameron Member at Mt. Kinesava. However, poorly preserved permineralized wood with associated orange and yellow oxides (probably carnotite and limonite) has been observed in a drab-colored zone low in the Cameron Member at ZION. This lower zone is unusually rich in organic material with preservation similar to what is seen in the "lower sandstones" at ZION and the Monitor Butte Member elsewhere. Unidentified plant compression fossils have also been recovered from the Cameron Member. Ash and others (2014) reported a well-preserved specimen of *Sanmiguelia lewisii* from the upper part of the Whitmore Point Member of the Moenave Formation in ZION. This is the youngest record of *Sanmiguelia*, and its first documented occurrence in the Early Jurassic.

Small burrow traces including both horizontally and vertically oriented grooves and punctures occur in the fine-grained, ripple cross-laminated sandstone low in the Cameron Member (Martz and others, 2015) as they do in similar sandstone beds in the Monitor Butte Member at CARE. These burrows may be referable to *Scoyenia, Cylindricum*, or the "Type 5" burrows that Hasiotis and Dubiel (1993) identified at PEFO.

Phytosaur and metoposaur material is fairly common in the Cameron Member, although it is usually too fragmentary to identify an alpha taxon. The metoposaurid material includes vertebral centra that are shorter aneteroposteriorly than wide (figure 6E) (Martz and others, 2015), indicating that they are probably referable to *Koskinonodon* or *Metoposaurus* rather than *Apachesaurus* (Hunt, 1993). The phytosaur material from ZION includes abundant teeth and some postcranial material, none of which can be assigned to an alpha taxon. A large, nearly complete phytosaur skull with mandibles has been recovered (DeBlieux and others, 2011a, 2011b) from low in the Cameron Member in the Eagle Crags area south of ZION (figure 6A) (UMNH VP 21917). The skull has fairly deep and plate-like squamosals and partially depressed supratemporal fenestrae (Martz and others, 2015), indicating that it may be a basal leptosuchomorph (Parker and Irmis, 2006; Stocker, 2010) although Martz and others (2015) suggested that it might be a more basal phytosaurid. A possible partial phytosaur humerus has also been recovered from the friable sandstone above the Cameron Member at Mt. Kinesava (Martz and others, 2015).

Stop 11
St. George Dinosaur Discovery Site at Johnson Farm

Although this paper is an overview of the Upper Triassic of southern Utah, it is also a field trip guide. Thus, a stop at the important St. George Dinosaur Discovery Site at Johnson Farm to view the faunas in the overlying Moenave Formation was a logical end to the field trip.

Return down Anasazi Way to SR 9. Proceed 60 km (37 mi) west through Rockville, La Verkin, and Hurricane until reaching the junction with I-15. Proceed south about 8.7 km (5.4 mi) on I-15 towards St. George, exit onto SR 212 and turn left to proceed south. This becomes N 3050 E and then curves southwest to become Riverside Drive. The SGDS is on the left side of the road about 4 km (2.5 mi) after exiting I-15 (figure 20A).

Stratigraphy and Sedimentology

Only the uppermost Chinle Formation (purplish mudstone beds possibly correlating to the purple-colored pedogenic beds at ZION) is exposed around St. George. However, the Moenave Formation (Rhaetian to Hettangian or latest Triassic to earliest Jurassic in age) in the area of the SGDS is 73.97 m thick, and is divided into the Dinosaur Canyon Member (56.41 m thick) and the overlying Whitmore Point Member (17.56 m thick) (Kirkland and Milner, 2006; Milner and others, 2012; Kirkland and others, 2014a).

The Dinosaur Canyon Member was deposited by predominantly northwesterly flowing rivers bordering the southwestern edge of the massive erg that deposited the Wingate Sandstone (Clemmenson and others, 1989; Blakey, 1994). At the SGDS, the predominantly fluvial facies of the Dinosaur Canyon Member is divided into three intervals: (1) a basal conglomerate about 80 cm thick, immediately above the unconformity at the top of the Chinle Formation, (2) a lower mudstone interval about 34.8 m thick, and (3) an upper sandstone interval about 20.46 m thick (Kirkland and Milner, 2006; Kirkland and others, 2014a). Based on a variety of fossil, magnetostratigraphic, and C-isotope chemostratigraphic evidence, the Triassic-Jurassic boundary lies in the Dinosaur Canyon Member (see Milner and others, 2012; Kirkland and others, 2014a; Suarez and others, 2017).

In 1967, Wilson described the Whitmore Point Member as a series of thin-bedded shale, limestone, and sandstone that separated the Dinosaur Canyon Member from the overlying Springdale Sandstone Member along the Arizona Strip and in southwestern Utah west of Kanab. Wilson (1967) also defined the contact between the Dinosaur Canyon and Whitmore Point Members in the Leeds area, Utah, as a limestone bed partially replaced by red chert. This bed defines the contact between these members over much of southwestern Utah (Martz and others, 2015), including at SGDS (Kirkland and others, 2014a).

The Whitmore Point Member at the SGDS was deposited by Lake Dixie, an extensive body of water that formed across a huge area of southwestern Utah and northwestern Arizona during the latest Triassic(?) to earliest Jurassic (Kirkland and Milner, 2006; Kirkland and others, 2014a). The Whitmore Point Member at SGDS has a lower, complex interval (4.48 m thick) composed of shoreline deposits laid down in subaerial and subaqueous environments that display lake-level transgressions and regressions along the western margin of Lake Dixie. This lower interval largely represents deepening and then shallowing of Lake Dixie (Kirkland and Milner, 2006; Kirkland and others, 2014a). Above this are a middle sandstone (7.64 m thick) interval and an upper, shale-dominated (6.55 m thick) interval that represents an upper lake cycle that is best developed in the St. George region, thus two major deepening and shallowing cycles are represented (Kirkland and Milner, 2006; Kirkland and others, 2014a). The upper, shale-dominated interval is unconformably overlain by the Springdale Sandstone Member of the Kayenta Formation (Kirkland and Milner, 2006; Kirkland and others, 2014a).

One of the most striking features of the SGDS is the relationship of trackways, topography, and the local paleogeography. It is not possible to map the paleogeography at every track level because surfaces are exposed only sporadically and the cost of additional excavation would be prohibitive. However, it is possible to map the "top surface layer," on which most of the trackways were made by animals walking on an undulating surface. In contrast to this onshore location, an extensive track-bearing continuation of the "main track layer" surface discovered to the northwest on Washington County School District and Darcy Stewart properties preserves abundant dinosaur swim tracks (*Characichnos*; see below) representing an offshore facies equivalent to the onshore surface marked by well-preserved *Eubrontes* tracks (Milner and others, 2006c; Milner and Lockley, 2016). This same marginal lacustrine onshore-offshore relationship can be seen in the sedimentary structures. Offshore or nearshore, submerged sedimentary structures include a variety of current and symmetrical ripples (figures 21B and 21C), tool marks (figure 21D), flute casts, and scratch circles (figure 21E).

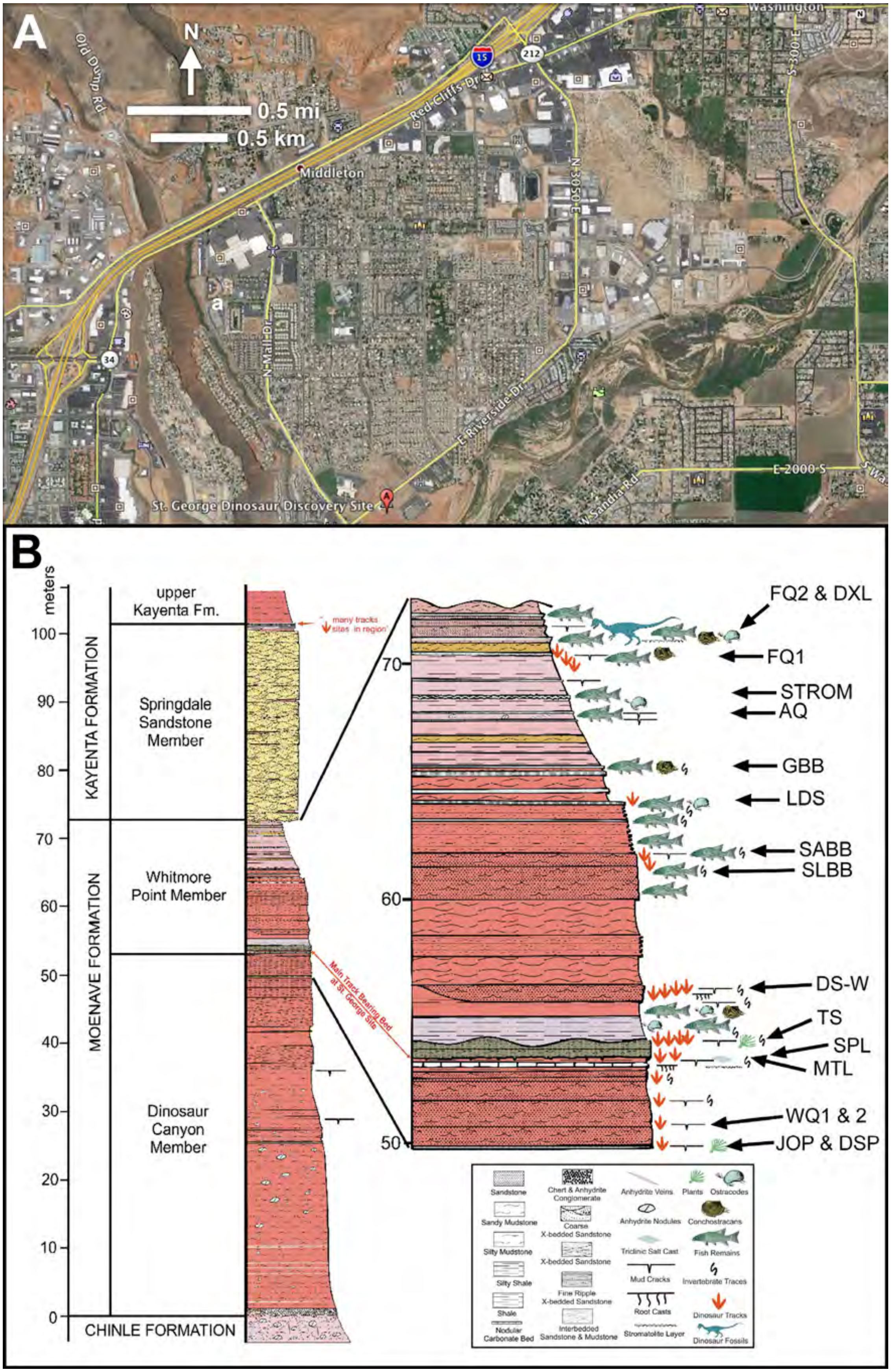

Figure 20. SGDS. (A) Map showing location of SGDS at St. George. (B) Stratigraphic section of the Moenave Formation in St. George. The acronyms refer to paleontology localities discussed in the text.

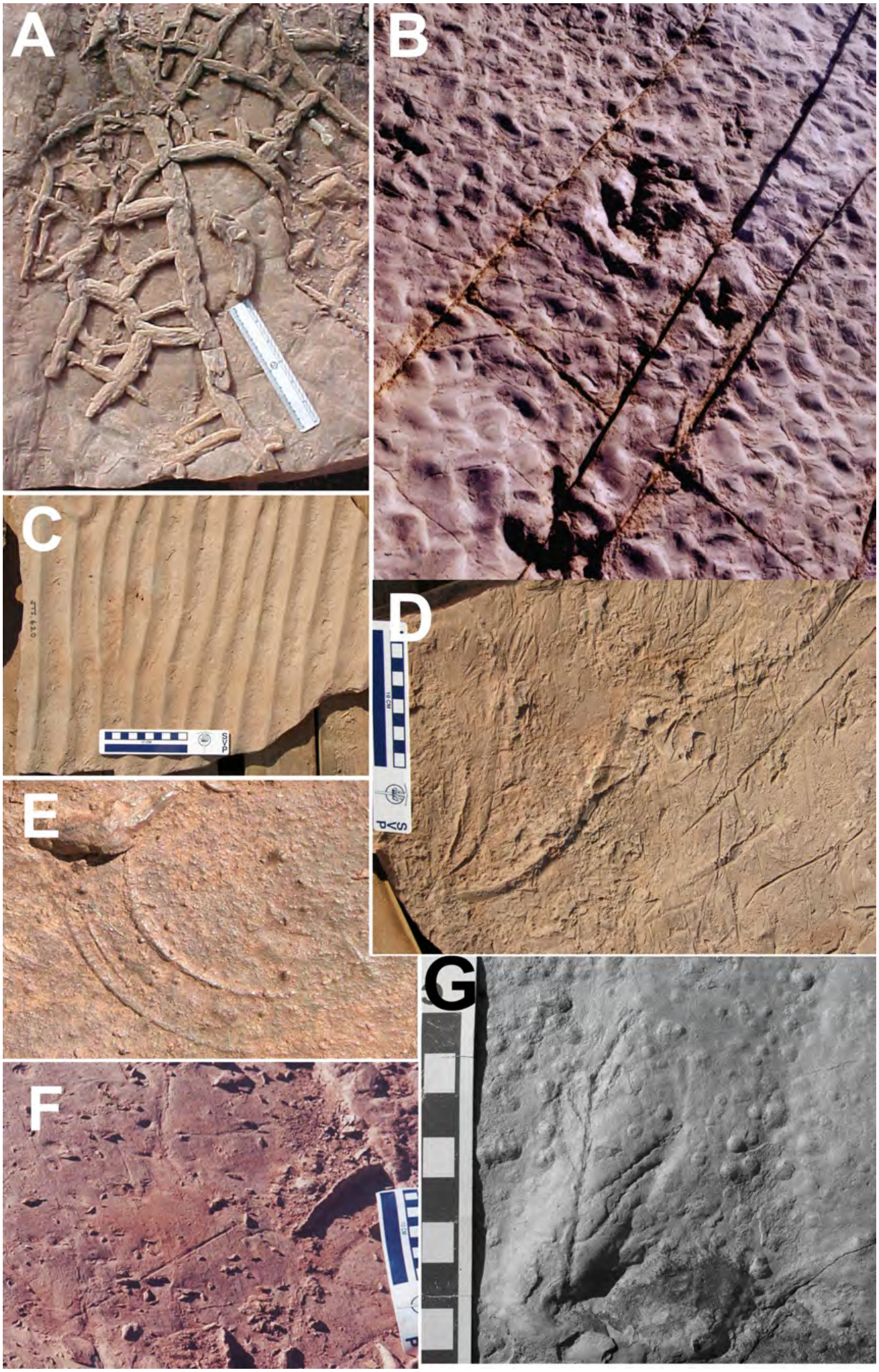

Figure 21. Common sedimentary structures from the SGDS and vicinity. (A) Mud-crack natural casts (SGDS 10). (B) *Eubrontes* and *Grallator* trackways with current ripples and joints (SGDS 18.T1). (C) Symmetrical wave-formed ripples (SGDS 620). (D) Tool mark marks (SGDS 262) and sedimentary structures formed by objects such as mud clasts, rock, plant fragments, etc., bouncing and scraping along a submerged substrate. (E) Scratch circles (specimen number pending). (F) Structures believed to be triclinic sulfate salt-crystal casts, but some researchers suggest they may be the trace fossil *Lockeia* (SGDS 40). (G) Raindrop impressions around a *Pagiophyllum* conifer branch (SGDS 491) (from Milner and others, 2012).

Likewise, onshore, subaerial sedimentary structures include mud cracks (figure 21A), possible evaporative salt-crystal casts (figure 21F), and raindrop impressions (figure 21G).

Many fish remains have been recovered from areas to the north and northwest of the SGDS museum site on former Darcy Stewart, Washington County School District, and LDS properties, especially from higher stratigraphic levels of the Whitmore Point Member (figure 20B) (Milner and others, 2006a, 2006b). This indicates that at the time the "top surface layers" of the Johnson Farm Sandstone Bed were deposited, the lake shoreline ran somewhere between the SGDS and Darcy Stewart sites, probably with a north-northeast–south-southwest trend. Fish scales have been found in mudstone beds that were deposited directly on the "Top Surface" tracksites, suggesting a lacustrine transgression soon after track formation (Milner and Lockley, 2006).

Further support for this interpretation of shoreline trend comes from trackway orientations, symmetrical wave-form and current ripple marks, and other sedimentary structures (figure 21). On the "top surface" are a series of large ridges and swales with a west-northwest–east-southeast trend, and between the ridges, are abundant, unidirectional current ripples that suggest a somewhat consistent flow pattern toward the west-northwest. Other sedimentary structures that support this flow direction include chevron marks, flute casts, and rill marks, all of which formed during high runoff in the direction of the lake (Kirkland and Milner, 2006; Milner and Lockley, 2006; Milner and others, 2012; Kirkland and others, 2014a). Locally, this topography has been eroded and reworked by small-scale water action; however, on a much larger scale, the ridges and swales are erosional megaripples that formed during a deeper lacustrine phase by large wave action moving from the northeast toward the southwest across a partially exposed beach shoal or spit (Kirkland and Milner, 2006; Milner and others, 2012; Kirkland and others, 2014a). Furthermore, symmetrical wave-form ripples (figure 21C), with a N. 30° E. ripple crest orientation, suggest small-scale waves lapping parallel to the paleoshoreline and perpendicular to the ridge-and-swale topography (Kirkland and Milner, 2006; Milner and Lockley, 2006; Milner and others, 2012).

Paleontology

The SGDS has received enormous attention (Kirkland and Milner, 2006; Lucas and others, 2006; Milner and Kirkland, 2006; Milner and Lockley, 2006; Milner and others, 2006b, 2006c, 2006d, 2012; Milner and Spears, 2007; Kirkland and others, 2014a; Harris and Milner, 2015; Milner and Lockley, 2016) as one of the most important Jurassic dinosaur tracksites in the world. Tracks have been identified on 25 stratigraphic levels in the immediate vicinity of the SGDS (figure 20), and many of these layers have been mapped in situ. Hunt and Lucas (2006) classified the SGDS as a Konzentrat-Ichnolagerstätte, although it could also be considered a Konzentrat-Lagerstätte because of the abundance of associated plants, conchostracans, ostracods, a variety of fishes, and tetrapod material (Milner and Spears, 2007). Collectively, the SGDS provides a window into an Early Jurassic ecosystem associated with the shores of Lake Dixie (e.g., Milner and Spears, 2007; Milner and others, 2012; Harris and Milner, 2015). This is the only tracksite in the western United States that rivals the preservation quality and abundance of the famous Lower Jurassic sites in eastern North America.

Dinosaur tracks and other tetrapod footprints considered to be Early Jurassic in age are well known in the southwestern United States (Lockley and Hunt, 1995). However, until the SGDS discovery, only a small number of tracksites (e.g., the Warner Valley Tracksite; Miller and others, 1989) had been scientifically documented in southwestern Utah, none of which are in the Moenave Formation. The SGDS site, therefore, fills an important stratigraphic gap in the trackway fossils record in the southwestern United States.

Four track-producing layers have been recognized in the uppermost part of the Dinosaur Canyon Member at the SGDS (figure 20B). Two very important localities, Walt's Quarry 1 and Walt's Quarry 2, named in honor of Walter Jessop who discovered both sites, were mapped in situ during careful excavation on former Darcy Stewart property in 2004–2005. Part of the Walt's Quarry 1 site was recovered as a single, 23.59 metric ton block now displayed in the SGDS museum (figure 20A). It was originally mapped and described as having 47 *Grallator* tracks in 11 trackways (Milner and others, 2006d), but remapping in 2011 revealed 58 *Grallator* in 13 trackways, and is one of the most visually spectacular specimens in the SGDS collection. Part of Walt's Quarry 2 was incorporated into a 13.11-m-long by 4.57-m-high wall (SGDS 567) during the first phase of the SGDS museum construction and development; it now occupies much of the rear (west) wall of the museum building and restoration was completed in 2014. About 200 dinosaur tracks (mostly *Grallator*), fish swim traces, and crocodylomorph tracks are on the blocks comprising this wall.

The first tracks discovered at the SGDS are from a horizon called the "main track layer" (figure 20B), situated at the base of the Johnson Farm Sandstone Bed (Kirkland and Milner, 2006; Milner and others, 2012; Kirkland and others, 2014a), a unit that is quite extensive and mapable in the St. George area. Tracks from the "main track layer" (figures 22B and 22C) are preserved as robust, natural sandstone casts (positive hyporelief) and associated with mud cracks, possible sulfate salt-crystal casts, flute casts, and many other sedimentary structures. Internally the bed displays several 15- to 25-cm-thick sets of climbing ripple bedding as in underlying sandstone in the upper Dinosaur Canyon Member. The Johnson Farm Sandstone Bed generally varies in thickness between 30 and 70 cm, although it has been eroded away and infilled by overlying units in some areas at the site. It is well-sorted, fine-grained sandstone about 53 m above the base of the Moenave Formation (figure 20B). The track casts (figures 22B and 22C) have up to 20 cm of relief and can be seen only after blocks of the Johnson Farm Sandstone Bed had been turned over. Flipping the massive blocks required heavy equipment and necessitated removing blocks from their original in situ positions to their current locations in the SGDS building and elsewhere. Several other track layers also occur above the "main track layer," most notably the Stewart-Walker Tracksite, which occurs approximately 1–1.5 m above the Johnson Farm Sandstone Bed upper surface (Kirkland and Milner, 2006; Milner and Lockley, 2006; Milner and others, 2006d; Kirkland and others, 2014a). Together, the mapped areas document approximately 3000 tracks (not including some 3000 theropod swim track claw marks) in their in situ orientations and stratigraphic contexts.

Many of the dinosaur trackways at the SGDS parallel the paleoshoreline, although some are perpendicular to shoreline trends. This same kind of trend for trackways is noted at many other tracksites elsewhere in the world (Lockley and Hunt, 1995). The shore-perpendicular trackways tend to follow the orientations of the ridges and swales on the SGDS top surface. Although many quadruped trackways also tend to parallel the top surface paleoshoreline, a concentration of *Batrachopus* tracks along ridge tops (i.e., shore-perpendicular) indicate track makers preferentially walked across higher terrain (Milner and Lockley, 2006; Milner and others, 2006d, 2012).

Tetrapod tracks from the SGDS have been assigned to the theropod dinosaur (e.g., Olsen and others, 1998) ichnotaxa *Grallator*, *Eubrontes* (figures 22A to 22C), and *Kayentapus,* the ornithischian dinosaur ichnotaxon *Anomoepus* (figure 22D), the crocodylomorph ichnotaxon *Batrachopus* (figure 22E) (e.g., Olsen and Padian, 1986; Olsen and others, 1998; Lockley and others, 2004), and the possibly sphenodontian ichnotaxon *Exocampe* (Milner and others, 2012; Harris and Milner, 2015). This ichnoassemblage is remarkably similar to those described by Hitchcock (1858) and Lull (1953) from the Early Jurassic of New England and strata of similar age from elsewhere around the world (Olsen and others, 2002), including elsewhere in the western United States (Lockley and Hunt, 1995).

Theropod tracks from the SGDS (figures 22A to 22C) have a remarkable size range. The smallest are only about 1.8 to 3 cm long, but several of these tracks have exceptionally long steps between footprints. In fact, one of the smallest track makers, with a footprint length of 2 cm, has step lengths of 23 to 36 cm, suggesting locomotory speeds of 3.2 to 6.4 m/s presuming animals with hip heights equal to five times track length. Another track maker, with a footprint length of 5.2 cm, has a step of 57 cm (Milner and Lockley, 2006; Milner and others, 2006d). Other small tracks (footprint lengths 8 cm, including heel traces) have short steps and exhibit metatarsal and hallux impressions (Milner and Lockley, 2006; Milner and others, 2006d).

Grallator footprints (figures 22A and 22B) are the most common ichnites at the SGDS. *Grallator* tracks at the site range in size from 10 to 25 cm long and 8 to 11

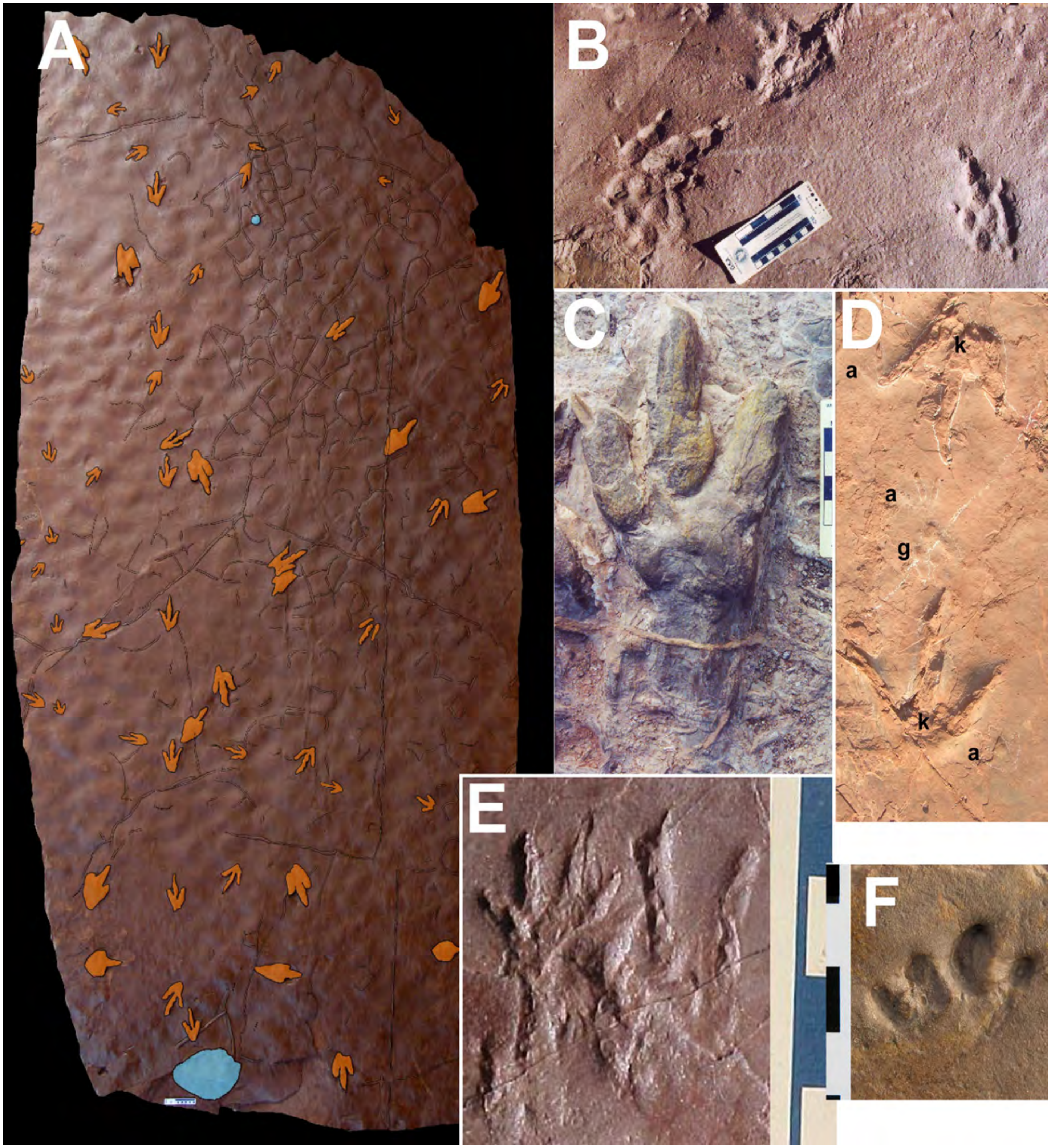

Figure 22. Tracks from SGDS. (A) 23.59 ton-block of sandstone preserving 58 *Grallator* tracks in 13 trackways (SGDS 568) (from Milner and others, 2012). (B) *Grallator* tracks (SGDS 197A). (C) *Eubrontes* natural cast track (SGDS 9). (D) *Anomoepus* tracks in situ from the "Unit 19 Road Cut Site" (site no. 2 on figure 20) a = *Anomoepus*, g = *Grallator*, k = *Kayentapus*. E. Manus and pes set of *Batrachopus* (SGDS 170). F. Possible synapsid pes track (SGDS 190).

cm wide, whereas *Eubrontes* tracks (figure 22C) tend to have foot lengths >32 cm. Very few intermediate track sizes (foot lengths between 25 to 32 cm) are known from the SGDS, and few *Grallator* footprints actually exceed 20 cm in length (Milner and Lockley, 2006; Milner and others, 2006d). With no intermediate footprint sizes, this strongly suggests the likelihood that *Grallator* and *Eubrontes* track makers were different theropod species (figure 3, top), and may not represent an ontogenetic series of tracks produced by the same species as suggested by Olsen and others (1998).

Many small, non-dinosaurian tracks have been found at the SGDS; most resemble the well-known, Early Jurassic ichnogenus *Batrachopus* (figure 22E) (Olsen and Padian, 1986). These tracks typically range from 1 to 5 cm in length, though maximum lengths of about 8 cm are known, although very rare. Pes prints, which are more commonly preserved, are always larger than the manus. In fact, the pes sometimes overprints the manus, occasionally leading to misinterpretations of the trackways and the potential track makers. A classic example is *Selenichnus*, described by Lull (1953), which was produced by animals passing through a soft substrate. Lull (1953) interpreted these tracks as produced by bipedal dinosaurs, but discoveries at the SGDS suggest that *Selenichnus* was produced by a quadrupedal crocodylomorph whose pedes overprinted its manus tracks (Lockley and others, 2004; Milner and others, 2006d). *Batrachopus* trackways are often hard to follow in detail, although several trackways have been recorded at the SGDS, including one trackway that can be traced for 5.8 m paralleling the crest of an erosional megaripple on the "top surface layer."

Various other quadruped tracks suggest the possibly that there are sphenodontian ichnotaxon *Exocampe*, although the true ichnotaxonomic status of *Exocampe* needs to be resolved. *Brasilichnium*-like footprints (figure 22F) that may have been produced by mammalian relatives (probably cynodont synapsids) are also present at SGDS (Milner and others, 2012; Harris and Milner, 2015). "Protomammal" tracks are known from other Triassic-Jurassic boundary sequences in the western United States, though they are most common in eolian deposits, unlike those represented in the Whitmore Point Member of the Moenave Formation. Both *Batrachopus* and the possible "protomammal" tracks are quite common transitioning from walking tracks to swim tracks and vice-versa (Milner and others, 2006c; Milner and others, 2012).

Two very important dinosaur track associations are preserved at the SGDS. These are abundant swim or floundering tracks and a trackway of a large theropod that stops and sits on the substrate.

First, SGDS preserves the largest and best preserved collection of dinosaur swim tracks in the world (e.g., Milner and Lockley, 2006, 2016; Milner and others, 2006c, 2012; Milner and Spears, 2007; Harris and Milner, 2015), which resolved a long-standing controversy among paleontologists as to what these kinds of structures truly represent. Tridactyl swim tracks of dinosaurs are usually arranged in sets of three parallel scrape marks that taper at each end, corresponding to the three functional digits of the theropod dinosaur pes. The longer digit III left a more elongate and deeper scrape mark compared to shorter digits II and IV. Dinosaur swim tracks from the Middle Jurassic of England were described and given the name *Characichnos* by Whyte and Romano (2001). A varying arrangement of theropod swim track morphotypes are represented in the SGDS collection showing: (1) animals swimming against a current, (2) animals swimming with the current that produced claw and digit impressions vertically or clear *Grallator*-type footprints with claw scrape marks directed caudally from the footprints, and (3) animals that were swimming cross-current (Milner and others, 2006c, 2012; Milner and Lockley, 2016). Many of these swim tracks preserve spectacular details such as skin impressions, scale scratch lines, and details on claw cuticles.

The second important association at the SGDS is a 22.3-m-long (73.2-ft-long) *Eubrontes* trackway (SGDS 18.T1) preserved in situ on the Top Surface within the SGDS museum. This trackway displays very rare tail drag marks along much of its length. Near the beginning of this same trackway are crouching traces produced when the track maker sat down on the substrate (on one of the large ridges mentioned previously), then shuffled forward and sat down a second time, creating two overlapping but distinct crouching impressions consisting of pes prints with hallux impressions, ischial callosities, tail traces, and even scale scratch lines (Milner and others, 2009). In addition, this unique trace fossil has clear, associated manus impressions (Milner and Lockley, 2006; Milner and others, 2006d, 2009, 2012; Harris and Milner, 2015). Both tail drag and crouching traces produced by theropod dinosaurs are extremely rare, and the manus impressions are unique among known theropod traces (Weems [2006] reported possible manus traces associated with *Kayentapus* tracks from the Early Jurassic of Virginia, but the manus impressions lack detail). The track maker proceeded through sediments of differing consistencies and gradients, providing an opportunity to better understand the interplay between sediment consistency, substrate morphology, animal behavior, and track morphology (Milner and Lockley, 2006; Milner and others, 2006d, 2009).

The SGDS also has a large number of fish swim trails and coprolites. Fish swimming traces include fine examples of *Undichna*, formed by the caudal fin (and sometimes other fins) of a fish scraping on a submerged lacustrine substrate (Seilacher, 2007), and *Parundichna*, probably made by pectoral and pelvic fins of a coelacanth scraping along a muddy substrate (Simon and others, 2003; Seilacher, 2007). *Parundichna* traces are known elsewhere from the Middle Triassic (Ladinian) Lower Keuper of Rot am See, Baden-Württemberg, Germany (Simon and others, 2003).

A moderately diverse compressional flora is preserved in sandstone beds in the upper part of the Dinosaur Canyon Member a few meters below the "main track-bearing surface" (Kirkland and Milner, 2006; Tidwell and Ash, 2006). The flora is conifer dominated and includes the holotypes of *Araucarites stockeyi* (SGDS.515.A & B), *Saintgeorgeia jensenii* (SGDS.627A & B), and *Milnerites planus* (SDDS.513A).

A low-diversity, invertebrate ichnofauna occurs in close association with tetrapod traces, sedimentary structures, and some body fossils at and around the SGDS (e.g., Lucas and others, 2006; Milner and others, 2012; Harris and Milner, 2015). An unusual feature of the SGDS is the co-occurrence of invertebrate ichnofossils with body fossils in the Whitmore Point Member. Most of the associated body fossils are invertebrates, including the ostracods *Darwinula* sp. and Cypridoidea indet. (Schudack, 2006), and two species of conchostracan *Euestheria brodieana* and *Bulbilimnadia killianorum* (Lucas and Milner, 2006; Kozur and Weems, 2010; Lucas and others, 2011).

Extensive collections of Early Jurassic fishes from the SGDS and surrounding area are currently being prepared and studied (Milner and Kirkland, 2006; Milner and Lockley, 2006; Milner and others, 2006b, 2012; Milner and Spears, 2007; Harris and Milner, 2015). Fishes that have already received some study include two new species described by Milner and Kirkland (2006): the hybodontoid shark *Lissodus johnsonorum* (figures 23A and 23B) and the lungfish *Ceratodus stewarti* (figure 23C). Other fishes, include one palaeoniscoid (figure 23E), abundant semionotids likely belonging to the genus *Lophionotus* (figures 23F and 23G), and a new, large species of *Chinlea*-like coelacanth (figure 23D) represented by many isolated elements including disarticulated and associated skulls and an articulated caudal fin.

Although far rarer than the fishes, tetrapod skeletal remains from the SGDS have also been recovered, all of which pertain to theropod dinosaurs thus far (Kirkland and others, 2005; Milner and Lockley, 2006; Milner and Kirkland, 2007; Milner and others, 2012; Harris and Milner, 2015). At least two types of theropod teeth and a single, well-preserved dorsal vertebra (figures 24A to 24C) have been found. The larger teeth, which are conical and sometimes preserve serrations under the right circumstances (figure 24F), have an overall shape similar to those of spinosaurids from the Early Cretaceous of North Africa and elsewhere (figures 24E and 24F), although they are not spinosaurids. Teeth like these have not been reported in any Early Jurassic theropod and thus likely pertain to a new taxon. Smaller, serrated, blade-like teeth (figure 24G) may belong to *Megapnosaurus/Syntarsus*, fragmentary remains of which have been reported from the Dinosaur Canyon Member of the Moenave Formation (Lucas and Heckert, 2001), or to an unknown coelophysoid theropod. It is likely that theropod dinosaurs were entering the waters of Lake Dixie to catch fish (Kirkland and others, 2005; Milner and Kirkland, 2007).

UPPER TRIASSIC VERTEBRATE BIOSTRATIGRAPHY IN SOUTHERN UTAH

As previously discussed, the Late Triassic Otischalkian, Adamanian, Revueltian, and Apachean land ver-

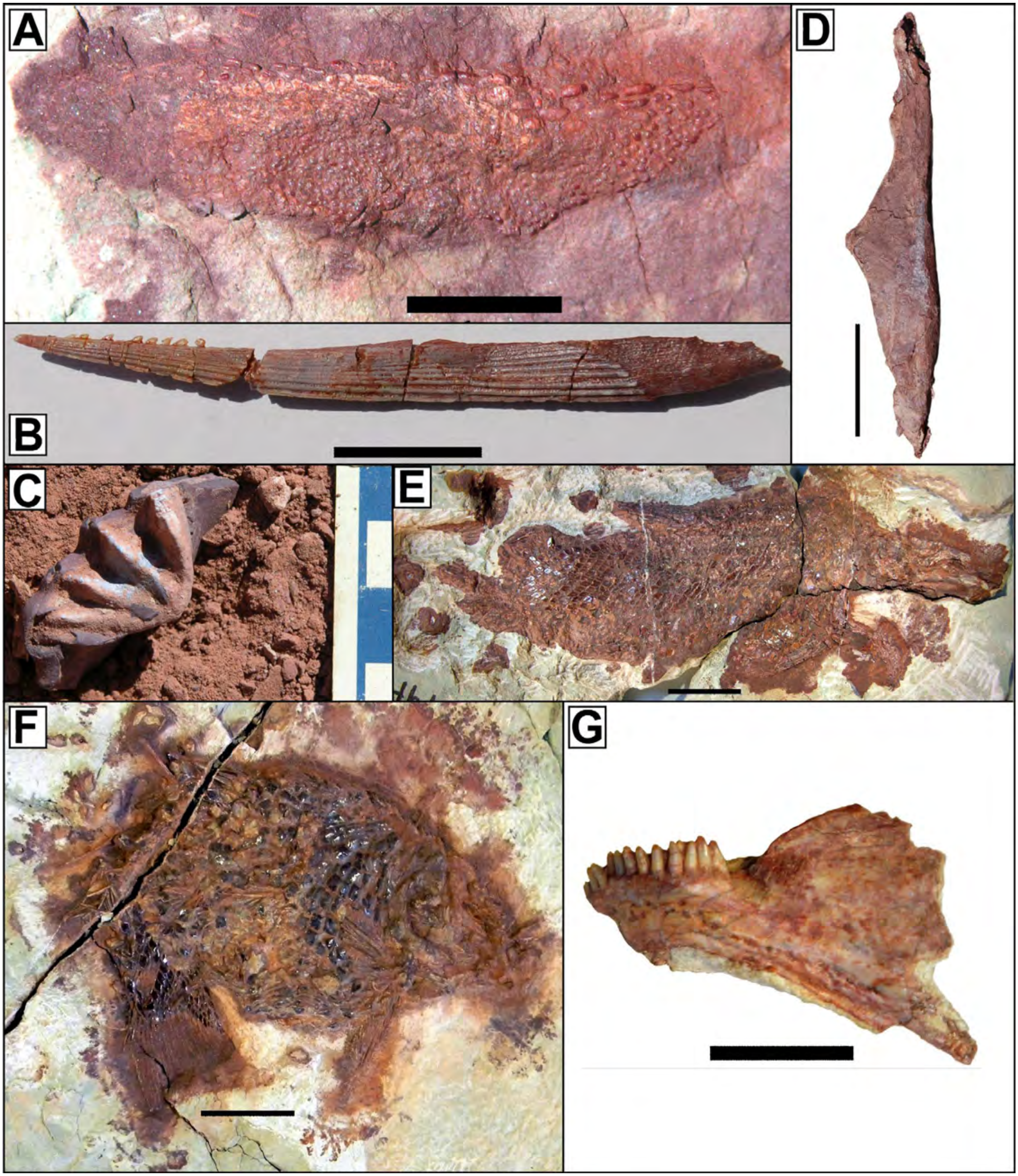

Figure 23 caption on following page.

Figure 23 (figure on previous page). Examples of fishes from the SGDS and surrounding areas. (A) Holotype specimen of *Lissodus johnsonorum* Milner and Kirkland (2006) lower left jaw with complete tooth sets (SGDS 857). Scale bar = 1 cm. (B) *Lissodus johnsonorum* dorsal fin spine in left lateral view (SGDS 828). Scale bar = 2 cm. (C) Holotype *Ceratodus stewarti* Milner and Kirkland (2006) lungfish tooth plate (UMNH-VP 16027). (D) Large left angular from a new species of *Chinlea*-like coelacanth (SGDS 892). Scale bar = 5 cm. (E) Nearly complete, unidentified palaeoniscoid fish (SGDS 1241). Scale bar = 2 cm. (F) Nearly complete semionotid fish (SGDS 1193). Scale bar = 5 cm. G. Lower jaw of a semionotid fish (SGDS 814). Scale bar = 1 cm (from Milner and others, 2012).

tebrate "faunachrons" or biozones of western North America are defined and bounded using phytosaur taxa (e.g., Lucas, 1998; Parker and Martz, 2011; Desojo and others, 2013). The biozones have also long been recognized as containing distinctive faunal associations consisting of various alpha taxa of phytosaurs, aetosaurs, metoposaurids, and other Late Triassic vertebrates (Colbert and Gregory, 1957; Gregory, 1972; Lucas and Hunt, 1993; Lucas, 1998; Parker and Martz, 2011). Although the biozones were recognized in Arizona and Texas, where the Upper Triassic vertebrate fossil record of western North America is known best (Colbert and Gregory, 1957; Gregory, 1972; Lucas and Hunt, 1993; Lucas, 1998; Parker and Martz, 2011; Martz and others, 2013), our knowledge of the vertebrate record from the Chinle Formation of southern Utah is rapidly improving. Although the biostratigraphic significance of many taxa from southern Utah is unclear, a few taxa offer an opportunity for a preliminary assessment of the biozones.

The Shinarump Member of the Chinle Formation in southern Utah and the Santa Rosa Formation of New Mexico and Texas are probably correlative units (Lucas, 1993; Riggs and others, 1996). As the Santa Rosa Formation and overlying beds contain basal phytosaurs that diagnose the Otischalkian biozone (Hunt and Lucas, 1991; Martz and others, 2013), it is likely that the Shinarump Member and beds immediately above, including those containing the Blue Lizard mine in Red Canyon on the east side of GLCA (Parrish and Good, 1987; Parrish, 1999) are also Otischalkian. *Doswellia,* which occurs in the Blue Lizard mine (figure 6G), is known from the Otischalkian of Texas (e.g., Long and Murry, 1995; Lucas, 1998) and from the Adamanian biozone in the Blue Mesa Member in PEFO (Parker and others, 2016), but has not been reported from the Revueltian or Apachean biozones.

The presence of putative basal leptosuchomorph phytosaurs from the Monitor Butte and lower Cameron Members (figure 7A) suggests that these units fall within the Adamanian biozone (figure 3), consistent with the lithostratigraphic correlation of these units with each other and with the Blue Mesa Member in Arizona (Lucas and others, 1997a; Irmis and others, 2011, supplemental data). However, neither the Monitor Butte Member phytosaur specimens nor the Cameron Member specimen have been described in detail, and the precise biostratigraphic position of the Monitor Butte specimens requires further assessment, although the stratigraphic position of the Cameron Formation skull is well constrained as occurring near the base of the unit (Martz and others, 2015).

Stratigraphic reconnaissance of the Circle Cliffs indicates that specimens of the aetosaurs *Calyptosuchus* and *Desmatosuchus* recovered there (figure 3) (W.G. Parker, unpublished data) probably also occur within the Monitor Butte Member (J.W. Martz, unpublished data), although as with the phytosaur specimens the precise stratigraphic horizon of these specimens has not yet been determined. This is also consistent with the near restriction of both aetosaur taxa to the Adamanian biozone of Texas and Arizona (e.g., Lucas, 1998; Parker and Martz, 2011; Martz and others, 2013). The spectacular specimen of the paracrocodylomorph *Poposaurus* from the Monitor Butte Member of Circle Cliffs (figure 8E) (Gauthier and others, 2011; Schachner and others, 2011) is consistent with this, as the *Poposaurus* is probably restricted to the Otischalkian and Adamanian biozones (e.g., Lucas, 1998; Parker and Martz, 2011). Fragmentary metoposaurid remains seem to be more abundant in the Cameron and Monitor Butte Members than in overlying members, again consistent with the relative abundance of metoposaurids in the Otischalkian and Adamanian biozones in Arizona and Texas (e.g., Hunt and Lucas, 1993; Lucas, 1998; Parker and Irmis, 2011).

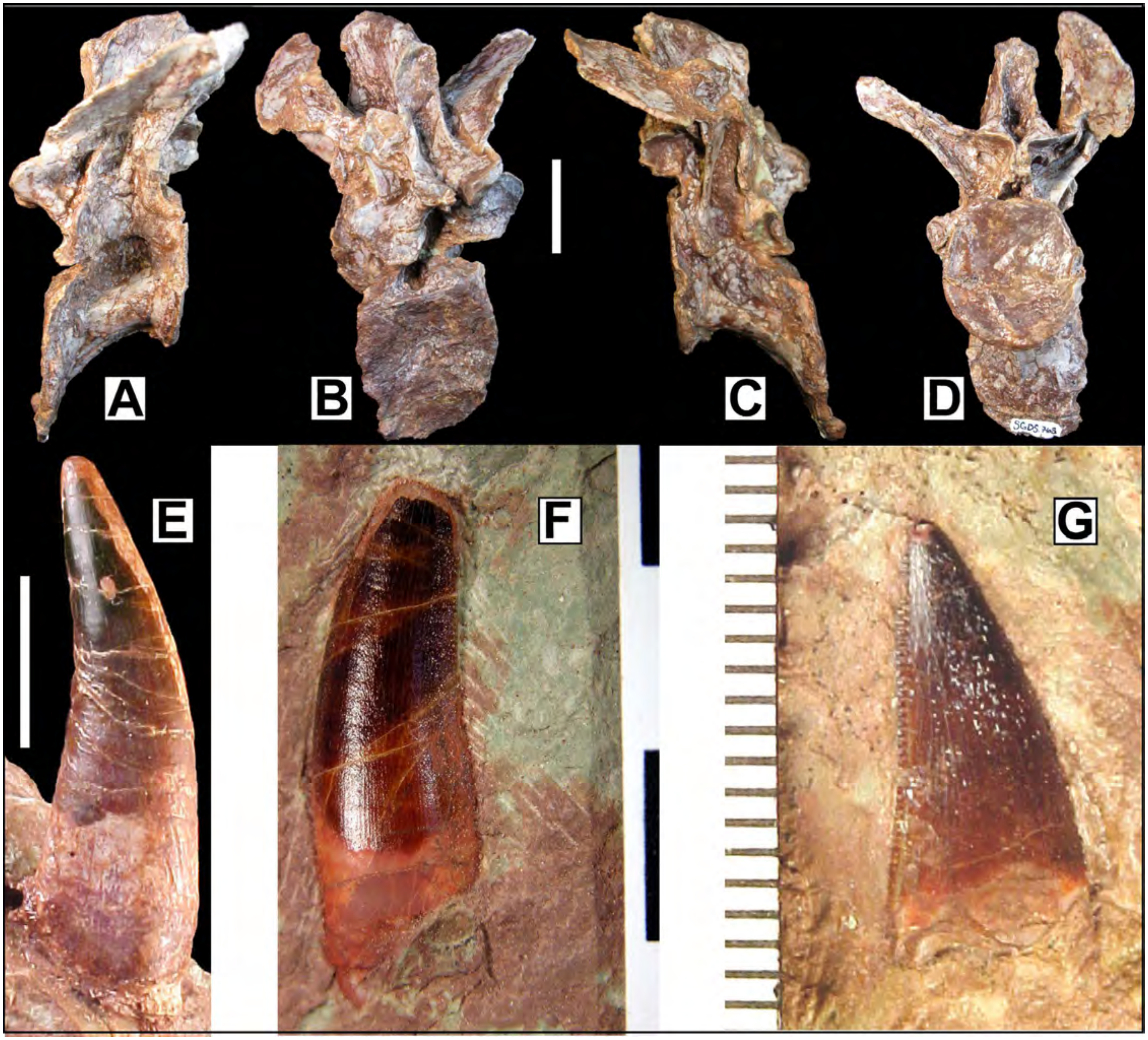

Figure 24. Theropod dinosaur remains from the SGDS. (A to D). Theropod cranial thoracic (anterior dorsal) vertebra (SGDS 768). (A) Left lateral view. (B) Anterior view. (C) Right lateral view. (D) Posterior view. Scale bar = 3 cm. (E) Large theropod tooth (SGDS 852). Scale bar = 2 cm. (F) Large theropod tooth that was broken while still articulated in the jaw. It preserves serrations and wear facets (SGDS 1335). (G) Small serrated tooth, possibly from a coelophysoid theropod (SGDS 851).

The boundary between the Adamanian and Revueltian biozones has been very well constrained in Arizona and Texas (e.g., Parker and Martz, 2011; Martz and others, 2013), but is less clear in southern Utah because this boundary is defined by the lowest stratigraphic occurrence of the phytosaur *Machaeroprosopus*. Species of *Machaeroprosopus* (e.g., *M. pristinus* and *M. buceros*) that occur at the base of the Revueltian in northern Arizona and western Texas (e.g., Parker and Martz, 2011; Martz and others, 2013) are unknown in Utah. This is unfortunate as the Adamanian-Revueltian boundary may indicate a faunal turnover resulting from climatic change or a major bolide impact (Olsen and others, 2011; Parker and Martz, 2011). Although it is tempting to correlate the Moss Back Member with the Sonsela Sandstone Bed of the Petrified Forest Member and Trujillo Formation in Arizona and Texas (e.g., Lucas, 1993) where the Adamanian-Revueltian transition occurs (Parker and Martz, 2011; Martz and others, 2013), this is conjectural until a precise lithostratigraphic correlation between the Moss Back Member and Sonsela Sandstone Bed can be explored in more detail.

The aetosaur *Typothorax* is known from the Petrified Forest Member of CARE (figure 3) (Kirkland and others, 2014b) and from the Kane Springs beds and Church Rock Member of Lisbon Valley (Martz and others, 2014). This taxon is most abundant in the Revueltian and Apachean biozones (e.g., Lucas, 1993; Heckert and others, 2010; Martz and others, 2014), although rare occurrences are known in the Adamanian biozone in Arizona and Texas (Parker and Martz, 2011; Martz and others, 2013). The Utah occurrences support the placement of the Petrified Forest Member in southern Utah into the Revueltian biozone, although as previously noted, the exact position of the lower boundary of the biozone is unclear.

In southern Utah, the base of the Apachean biozone, which is defined by the appearance of derived species of *Machaeroprosopus* traditionally referred to as "*Redondasaurus*" (figures 7B and 7C) (e.g., Lucas, 1998; Martz and others, 2014) is also difficult to precisely place. Parker and others (2011) identified a phytosaur skull from the Owl Rock Member in Arizona as "*Redondasaurus*," but in southern Utah the stratigraphically lowest occurrences are in the Church Rock Member (Martz and others, 2014), which overlies the Owl Rock Member and correlative Kane Springs beds (e.g., Blakey and Gubitosa, 1983, 1984). It is unclear if this is because the lowest occurrence of "*Redondasaurus*" or the beginning of Owl Rock deposition is diachronous between Arizona and Utah, or if the Utah fossil record is merely incomplete (Martz and others, 2014).

The top of the Apachean biozone is difficult to precisely place in southern Utah because it is defined by the lowest occurrence of the basal crocodylomorph *Protosuchus* (Lucas, 1998), which has not been reported from Utah. In Arizona, the lowest occurrence of *Protosuchus* is in the Dinosaur Canyon Member of the Moenave Formation (e.g., Clark and Fastovsky, 1986; Sues and others, 1994; Lucas and others, 2005), which overlies the Chinle Formation. The Moenave Formation is lithostratigraphically correlative with the Wingate Sandstone (e.g., Blakey, 1994), indicating that the top of the Apachean biozone probably occurs within the latter formation as well. The precise placement of the top of the Apachean biozone relative to the Triassic-Jurassic boundary has been difficult, and has relied on various interpretations of fossil and magnetostratigraphic data. However, the Triassic-Jurassic boundary and the end-Triassic mass extinction probably occur slightly after the end of the Apachean biozone (Milner and others, 2012; Kirkland and others, 2014a; Suarez and others, 2017).

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