Provenance of a Permian erg on the western margin of Pangea: Depositional system of the Kungurian (late Leonardian) Castle Valley and White Rim sandstones and subjacent Cutler Group, Paradox Basin, Utah, USA

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ABSTRACT

Consideration of petrographic and U-Pb provenance data and paleocurrent analysis of Kungurian (upper Leonardian) Cutler Group strata in the salt anticline province of the Paradox Basin of Utah demonstrates striking contrasts in composition and inferred sources of stratigraphically adjacent eolian and fluvial facies. Eolian strata, termed here the Castle Valley Sandstone, exposed in the Castle Valley northeast of Moab, Utah, and long correlated with the White Rim Sandstone, were deposited on the southwestern flank of a NW-trending diapirc salt wall. The eolian strata, which overlie red fluvial sandstone and conglomerate of the undifferentiated Cutler Formation, are as much as 183 m thick in outcrop and consist of two eolianite members separated by a thin sheet-flood deposit that contains pebbles derived from the salt wall and upturned conglomeratic strata adjacent to it. Both eolian and underlying fluvial deposits thin and onlap eastward onto the now-collapsed salt wall. Fluvial strata at Castle Valley and in exposures to the northeast were transported northward, parallel to the salt wall. Large-scale foresets in the lower eolianite member indicate dominant northeasterly wind directions (present coordinates) and transport directly away from the contemporary Uncompahgre uplift, whereas foresets in the upper member indicate variable northeasterly and northwesterly paleowinds. The eolian strata thus accumulated on the lee side of the salt wall, but sandstone composition and northwestward wind components indicate net transport from the northwest, comparable with dominant southeastward sand transport, away from the Pangean shoreline, documented for the greater White Rim erg to the west and northwest. The NW and NE winds are both predicted by late Paleozoic atmospheric circulation models for western Pangea.

Cutler fluvial sandstones are compositional arkoses (mean Qt56F42L2) containing basement-derived detrital components that include potassium feldspar, plagioclase, biotite, and zircons with a restricted, bimodal age distribution of ~1790–1689 Ma and ~1466–1406 Ma. These grain ages exactly match known basement ages in the nearby Uncompahgre uplift. In contrast, the Castle Valley Sandstone ranges from quartz-rich arkose to subarkose and exhibits a consistent upsection decrease in feldspar content, from Qt90F10L0 in the lower eolianite member to Qt60F40L0 in the upper member. Like the underlying fluvial arkose, the lower eolianite member contains potassium feldspar, plagioclase, and mica derived from the Uncompahgre uplift, but the locally derived zircon age groups constitute only 23%–37% and 13% of the zircon grain ages in the lower and upper eolianite members, respectively; whereas older Archean and Paleoproterozoic grains, including ca. 1.5 Ga grains uncommon in the Laurentian detrital-zircon record, and Grenville, Neoproterozoic, and early Paleozoic grains constitute the bulk of the zircons. Quartzarenite of the greater White Rim erg contains detrital-zircon populations similar to those of the upper eolianite member. The Grenville and younger grains are interpreted as having an eastern Laurentian (Appalachian) source, whereas the ca. 1.5 Ga grains probably had an ultimate source in Baltica. Sediment-transport directions indicate that zircon grains not directly attributable to local basement of the Ancestral Rocky Mountains, including grains with a likely Baltica source, were transported to the western shoreline of Laurentia by transcontinental fluvial systems and then southeastward to their depositional site at the erg margin in salt-withdrawal minibasins.

INTRODUCTION

The nature of late Paleozoic dispersal systems that delivered sediment to the western edge of Pangea and sources of sediment carried by those systems have been topics of speculation and debate since the earliest paleogeographic reconstructions of the supercontinent. Enormous volumes of eolian sediment, which presumably required aerially extensive source areas and possibly transcontinental-sediment-delivery routes, accumulated along the western continental margin during Late Pennsylvanian and Early Permian time (Blakey et al.,
Large erg systems that developed in Early Permian (Wolfcampian and late Leonardian or Sakmarian–Artinskian and Kungurian) time along the NNE-trending shoreline of Pangea (Permian coordinates; Fig. 1) are inferred to have been fed by littoral sand of the western marine margin (e.g., Blakey et al., 1988; Dubiel et al., 1996; Condon, 1997; Dickinson and Gehrels, 2003). Many potential bedrock sources for sediment existed in Pangea due to the wide extent of deformation that took place during the assembly of the supercontinent:

The intracratonic Ancestral Rocky Mountain deformation event accompanied supercontinent assembly and created basement uplifts that provided coarse arkosic sediment to adjacent sedimentary basins in the present region of the Rocky Mountains and Colorado Plateau (Fig. 1; Melton, 1925; Ver Wiebe, 1930; Baker et al., 1933; Mallory, 1972a; Kluth and Coney, 1981; Kluth, 1986; Barbeau, 2003). At the same time, Alleghenian and Ouachita deformation was only recently completed as a result of diachronous collision and terrane accretion along what had been the eastern and southern flanks of Laurentia.

Figure 1. Paleogeographic map of Pennsylvanian–Permian Ancestral Rocky Mountain province. Locations of uplifts and basins adapted from Kluth and Coney (1981), Geslin (1998), Barbeau (2003), Dickinson and Lawton (2003), and Trexler et al. (2004). Location and trend of Kungurian (late Leonardian) shoreline from the 275 Ma paleogeographic map on the Colorado Plateau Geosystems Web site (cpgeosystems.com/nam.html), last accessed January 2015. Predicted wind directions and paleolatitudes for late Paleozoic from Parrish and Peterson (1988) and Peterson (1988). Bold numerals are locations of stratigraphic columns of Figure 2.
(Hatcher, 1989; Viele and Thomas, 1989; Keller and Hatcher, 1999), and accretion of exotic terranes had recently affected the Cordilleran margin of Laurentia (Speed and Sleep, 1982; Wright and Wyld, 2006). Therefore, discrimination of potential sources for late Paleozoic eolian sediment has important implications for Pangean paleogeography and sediment-transport systems and informs general models for eolian sediment supply and accumulation.

This paper describes the structural and depositional settings of continental deposits of the Permian Cutler Group in and near Castle Valley, Utah, located in the proximal part of the Paradox Basin, and compares the petrography and detrital-zircon content of these deposits to correlative eolian strata of the White Rim Sandstone of the more distal basin to the southwest. The data presented here provide insight into sediment sources and sediment-transport systems for voluminous Lower Permian eolian deposits as well as continent-scale paleogeography of Laurentia in its broader context within Pangea.

### GEOLOGIC SETTING

Upper Leonardian (Kungurian) clastic strata of the Paradox Basin, which include the White Rim Sandstone and its correlatives, were deposited late in the history of the Ancestral Rocky Mountains deformational event, an episode of regional crustal deformation that created yoked basement-cored uplifts and basins extending between Oklahoma and Nevada (Fig. 1; Melton, 1925; Ver Wiebe, 1930; Trexler et al., 2004). During the late Paleozoic, the Ancestral Rocky Mountains province lay at tropical latitudes, ~10° north of the equator (Fig. 1; Scotese et al., 1979; Scotese and McKerrow, 1990; Dubiel et al., 2009), and the climatic regime has been inferred to have ranged from semi-arid in Pennsylvanian time to seasonally wet or even peri-glacial in the Early Permian (Soreghan et al., 2002, 2009).

### Paradox Basin

The northwest-trending Paradox Basin contains an asymmetric sedimentary fill that thickens northeastward toward the basement-cored Uncompahgre uplift. The basin is interpreted as the result of flexural subsidence adjacent to the basement load (Barbeau, 2003; Trudgill, 2011), and a number of faults within the uplift and flanking it have demonstrated late Paleozoic sinistral strike-slip or normal movement (Weimer, 1980; Baars and Stevenson, 1981; Stevenson and Baars, 1986; Thomas, 2007). The basin is defined by the distribution of Pennsylvanian and Permian strata included in the Paradox Formation, Honaker Trail Formation, and Cutler Group in Utah and the Hermosa Group and Cutler Formation in Colorado (Figs. 2 and 3). The Paradox Formation is a cyclic succession of shale, dolomite, and basin-central evaporite as much as 4300 m thick (Baker et al., 1933; Hite, 1970; Hite and Buckner, 1981; Doelling, 2002a). Evaporites within the cycles include gypsum, anhydrite, halite, carnallite, and sylvite (Hite, 1970; Hite et al., 1972; Hite and Buckner, 1981). The Paradox Formation is overlain by, and grades southwestward into, mixed carbonate and silicilastic strata of the Honaker Trail Formation, which records glacio-eustatic fluctuations driven by Milankovitch cyclicity (Goldhammer et al., 1994). The Paradox Formation grades to coarse-grained silicilastic strata of the undifferentiated Hermosa Group on the northeast flank of the basin (Dubiel et al., 2009). The Cutler Formation overlies the Honaker Trail Formation and consists of undifferentiated alluvial and fluvial arkose and conglomerate as much as 2450 m thick in the proximal northeastern part of the basin near the Uncompahgre uplift (Condon, 1997; Doelling, 2002a; Venus et al., 2015). Proximal fluvial strata of the Cutler Formation in the northeastern part of the basin have been differentiated on the basis of unpublished, proprietary seismic data (Trudgill, 2011). The upper part of the undifferentiated Cutler Formation has been identified as the Organ Rock Formation (Rasmussen, 2009, 2014), a formal term applied to red continental strata overlying eolian strata of the Cedar Mesa Sandstone and underlying the eolian White Rim Sandstone in the central and southwestern parts of the basin (Fig. 2; Baars, 1962; Stanesco et al., 2000; Dubiel et al., 2009); however, the eolian units, particularly the Cedar Mesa Sandstone, that permit unambiguous identification of the Organ Rock Formation are absent from most outcrops of the proximal part of the basin. Because of persisting uncertainty regarding correlation, the term, undifferentiated Cutler Formation, is retained in this paper for fluvial and alluvial facies of the basin northeast of Moab, Utah. The undifferentiated Cutler Formation onlaps and buries the Uncompahgre uplift near Gateway, Colorado (Figs. 3 and 4; Melton, 1925; Soreghan et al., 2009; Kluth and DuChene, 2009; Rasmussen, 2009), where the formation consists of boulder conglomerate and coarse-grained sandstone generally interpreted as proximal to distal alluvial-fan deposits (Schultz, 1984; Mack and Rasmussen, 1984).

The undifferentiated Cutler Formation fines southwestward across the basin, grading laterally to a succession of formations considered components of the Cutler Group (Fig. 2; Baars, 1962; Condon, 1997). These formations alternate between red-weathering structureless siltstone and fluvial sandstone and siltstone represented by the Halgaito Formation (or “lower Cutler beds”), which overlies the Honaker Trail Formation, and the younger Organ Rock Formation (e.g., Langford and Chan 1988, 1989; Condon, 1997; Stanesco et al., 2000; Soreghan et al., 2002; Mountney and Jagger, 2004) and visually striking, thick-bedded white to pink eolian sandstone intervals that include the Cedar Mesa Sandstone and younger White Rim Sandstone (Loope, 1984; Huntoon and Chan, 1987; Blakey et al., 1988; Blakey, 1996; Langford and Chan, 1988; Chan, 1988; Dubiel et al., 1996, 2009). Arkosic composition and southwest-fining trends of fluvial strata of the undifferentiated Cutler Group and the Organ Rock Formation have led to the generally accepted view that basement rocks of the Uncompahgre uplift were the primary source of the fluvial sediment (Ver Wiebe, 1930; Baars, 1962; Cater and Elston, 1963; Mallory, 1972a, 1972b; Campbell, 1979, 1980; Kluth and Coney, 1981; Kluth, 1986; Condon, 1997; Barbeau, 2003; Blakey, 2008). Sediment in the eolian intervals was largely derived from coastal sand at the edge of the Permian seaway to the west and northwest (Kamola and Chan, 1988; Dubiel et al., 1996, 2009).
On the basis of the large volume of eolian sand, its quartzose composition, and subsequently on the basis of its detrital-zircon content, ultimate sources in the Appalachian orogen of eastern Laurentia have been posited for sediment of the eolian sandstones (Johansen, 1988; Marzolf, 1988; Dickinson and Gehrels, 2003).

A belt of salt walls cored by Paradox evaporite, termed the salt anticline province, occupies the thick proximal part of the Paradox Basin and forms a broad region of transition between coarse-grained undifferentiated Cutler Formation and finer-grained differentiated stratigraphic entities of the Cutler Group (Figs. 3 and 4; Baker et al., 1933; Shoemaker et al., 1958; Jones, 1959; Doelling, 1988; Trudgill, 2011). The salt walls developed by rapid diapiric movement of Paradox Formation evaporite during Late Pennsylvanian, Permian, and Triassic time; salt migration continued at slower rates in the Jurassic and Cretaceous (Doelling, 1988, 2002a; Trudgill, 2011). Evidence for syndepositional salt rise includes thinning of strata and angular unconformities on the flanks of diapiric structures (Shoemaker et al., 1958; Jones, 1959; Elston et al.,...
Lawton et al. | White Rim–Castle Valley erg, Paradox Basin, Utah

1962; Cater and Elston, 1963; Trudgill et al., 2004; Matthews et al., 2004; Lawton and Buck, 2006; Trudgill, 2011) and fluvial sediment transport parallel to the axes of salt-withdrawal “minibasins” formed by the migration of evaporite into the diapiric structures (Matthews et al., 2004; Banham and Mountney, 2013, 2014). Rising Paradox evaporite entrained blocks of dolostone and shale interbeds to shallow levels in the diapirs; thus, local diapir exposure is indicated by conglomerate beds near several of the salt walls containing clasts of Paradox carbonate and less common gypsum eroded from the exposed salt walls and deposited in Permian and Triassic strata (Shoemaker et al., 1958; Elston et al., 1962; Lawton and Buck, 2006).
Figure 4. Map showing thickness of White Rim Sandstone and correlative strata in Salt anticline region (isopachs in feet), paleocurrent data from eolian strata and fluvial strata roughly correlative with Organ Rock Formation, and distribution of Proterozoic rocks in Uncompahgre uplift. Labels as in Figure 3 and CC—Cane Creek anticline; SD—Shafer Dome. Thickness data, adapted from Baars and Seager (1970), Condon (1997), Trudgill (2011), and Parr (2012), indicate strong influence of salt-withdrawal minibasins southwest of Salt Valley (SA), Castle Valley (CV), and Moab Valley (MV) salt walls on thickness and orientation of Kungurian erg margin. Paleocurrent data from Baars and Seager (1970), Huntoon and Chan (1987), Buller (2009), Venus et al. (2015), and this study. Basement rock units from Trudgill (1979) and Doelling (2002a). Dashed rectangle at NW end of Castle Valley salt wall indicates location of Figure 5. Oil well symbols indicate selected wells that indicate subcrop relations near structural front of Uncompahgre uplift and key thickness localities of eolian strata at top of Cutler Group, explained in text.
Uncompahgre Uplift

The Uncompahgre uplift, which lay along the northeastern margin of the Paradox Basin, was a doubly vergent basement-involved uplift as much as 160 km wide with marked, but as-yet undetermined, topographic relief (Figs. 1 and 3; DeVoto, 1980; Hoy, 2000; Hoy and Ridgway, 2002). Seismic data and drilling indicate that basement rocks are thrust southwestward over Pennsylvanian and Permian strata, generally synorogenic clastic deposits, near the mountain front, in Utah (Fig. 1; Mobil #1 McCormick well; Frahme and Vaughn, 1983) and possibly over salt of the Paradox Formation in southwestern Colorado (Kluth and DuChene, 2009). Frontal structures on the southwestern side of the Uncompahgre uplift are extensively buried beneath Permian and Triassic clastic rocks, but exposed reverse faults of Pennsylvanian–Permian age encompass basement rocks over Pennsylvanian and Permian conglomerate of the Central Colorado trough on the northeast flank of the uplift (Fig. 3; Hoy, 2000; Hoy and Ridgway, 2002). Alluvial-fan facies of the Permian Maroon Formation directly west of Redstone, Colorado, on the northeast flank of the uplift, indicate a nearby boundary between the uplift and the Eagle basin (Fig. 3). Basement rocks on the southwest flank of the uplift are overlain by the upper part of the Cutler Formation near Gateway, Colorado, and along the frontal part of the uplift (Soreghan et al., 2009), by Triassic beds of the Chinle Formation along the Colorado River and southeastward along the uplift (Tweito, 1979; Doelling, 2002a), and Jurassic strata in the vicinity of the Black Canyon of the Gunnison River to the northeast (Fig. 3; Tweito, 1979). The distribution of strata deposited on Proterozoic rocks therefore demonstrates progressive northeastward onlap of basement perpendicular to the mountain front well into Mesozoic time and indicates that Uncompahgre basement was extensively exposed during the Permian. Farther southeast, on the San Juan dome of Laramide age, Proterozoic rocks are overlain by a thin veneer of Cambrian–Mississippian strata and a thick succession of Pennsylvanian rocks (Tweito, 1979; Weimer, 1980; Thomas, 2007) and thus were not exposed during Pennsylvanian and Permian time. South of Ouray, Colorado, where the structural front of the Uncompahgre uplift appears to be offset by a major basement fault (Fig. 3; Weimer, 1980), west-trending faults with demonstrated Pennsylvanian and Permian displacement form a system of horsts and grabens south of the main uplift (Baars, 1966; Weimer, 1980; Thomas, 2007).

Pre-Pennsylvanian rocks of the Uncompahgre uplift consist of metamorphosed volcanic, plutonic, and sedimentary rocks intruded by posttectonic granites and overlain by a thin cratonic succession of Cambrian–Mississippian strata. Paleoproterozoic rocks include foliated granites and metavolcanic rocks with U-Pb ages ranging 1.78–1.69 Ga (Silver and Barker, 1968; Bickford et al., 1989; Gonzales and Van Schmus, 2007), which are overlain by quartzite and phyllite with maximum depositional ages near 1.67–1.65 Ga (Jessup et al., 2006; Jones et al., 2009). The metasedimentary succession is folded and intruded by porphyritic Mesoproterozoic granitoids with U-Pb ages near 1.44 Ga (Silver and Barker, 1968; Bickford and Cudzilo, 1975; Gonzales and Van Schmus, 2007). Basement rocks of these ages are presently exposed in the southwestern part of the Uncompahgre uplift near Ouray, Colorado; to the south in the San Juan dome north of Durango, Colorado; in the northwestern part of the uplift in Unaweep and Black canyons; in the vicinity of Almont, Colorado, on the northeastern side of the uplift; and in the Colorado River canyon west of Grand Junction, Colorado (Fig. 3).

Following late Paleozoic uplift, the Uncompahgre uplift was reactivated and its structures overprinted by Laramide shortening, which resulted in the development of reverse faults and monoclinal folds (Lindsey et al., 1983); it was then buried by Paleogene volcanic and volcaniclastic rocks and cut by normal faults and sediment-filled grabens of the Neogene Rio Grande rift (Tweito, 1979; Hoy and Ridgway, 2002). These younger events and features have created uncertainty as to the stratigraphic and structural relations unique to the Paleozoic history of the uplift, and even as to its Paleozoic extent (compare Baars, 1966; Condon, 1997; Hoy and Ridgway, 2002; Barbeau, 2003; Thomas, 2007).

Stratigraphic Relations and Age of the White Rim Sandstone

Throughout its geographic distribution, the White Rim Sandstone lies at the top of the Cutler Group. It overlies fluvial strata of the Organ Rock Formation on a sharp contact interpreted as a sequence boundary (Blakey, 1996) and is unconformably overlain by the Moenkopi Formation. The White Rim Sandstone interfingers westward with the fossiliferous Toroweap Formation, which establishes its age as late Leonardian (late Kungurian; Fig. 2; Irwin, 1971; Blakey et al., 1988; Blakey, 1996). On the eastern flank of the Circle Cliffs uplift and the San Rafael Swell (Fig. 3), the White Rim also interfingers with the lower part of the Kaibab Limestone, the age of which is poorly known, and on the San Rafael Swell, the White Rim is overlain by the upper part of the Kaibab, which is of Wordian age (Fig. 2; Irwin, 1971; Huntoon and Chan, 1987; Kamola and Chan, 1988). The Kaibab Limestone is not present along the Green and Colorado Rivers, and the upper surface of the White Rim Sandstone displays irregular topography (Baars and Seager, 1970; Orgill, 1971; Huntoon and Chan, 1987; Kamola and Chan, 1988). Where the top of the White Rim Sandstone is exposed between the Green and Colorado rivers and west of their confluence, the upper surface of the formation displays numerous linear ridges 3–5 m high that trend NNW and are asymmetric with steeper west flanks; the most prominent of these ridges is 75 m high and trends NE (Baars and Seager, 1970; Huntoon and Chan, 1987). These ridges were initially interpreted as elongate marine bars oriented parallel with a dominant SSE transport direction indicated by large-scale foreset dips (Baars and Seager, 1970), an interpretation contravened by subsequent eolian interpretation of the White Rim Sandstone (Huntoon and Chan, 1987; Chan, 1989; Dubiel et al., 1996, 2009). The topographic features were subsequently attributed to both erosional sculpting of uppermost White Rim strata and preservation of relic dune topography during peak transgression of the Kaibab seaway (Huntoon and Chan, 1987; Chan, 1989), which suggests that the White Rim Sandstone is entirely older than the upper part of the Kaibab Limestone in the vicinity of the Colorado and
Supplemental Table 1. U-Pb geochronologic analyses of detrital zircons in Permian strata of Paradox Basin. Please visit http://dx.doi.org/10.1130/GEOS1174.S1 or the full-text article on www.gsapubs.org to view Supplemental Table 1.

Data collection included geologic mapping of the NW end of Castle Valley at a scale of 1:10,000, measurement of stratigraphic sections at accessible localities (details in Parr, 2012), and measurement of eolian foresets. Single measurements were taken for each dune bed set between bounding surfaces on a reference section along Castle Creek, but ten measurements were made per location elsewhere in the study area (Parr, 2012). Trough axes were measured in fluvial strata of the Cutler fluvial beds along Onion Creek northeast of Castle Valley and at locations on the flanks of Castle Valley (Buller, 2009). Petrographic and detrital-zircon samples were collected within the map area of Figure 5 and elsewhere in the basin, as described later.

Standard petrographic thin sections stained for potassium feldspar were counted using the Gazi-Dickinson technique to minimize compositional dependency on grain size (Ingersoll et al., 1984). Four hundred framework grains were counted per sample to achieve a 2σ confidence of ±5% (Van der Plas and Tobi, 1965). Sandstone modal compositions are listed in stratigraphic order in Table 1.

Zircon separates from Permian strata were analyzed using a laser-ablation, inductively-coupled plasma mass spectrometer (LA-ICP MS) at the University of Arizona LaserChron Center. Zircons were separated using standard mineral separation techniques. Approximately 100 individual zircon analyses were conducted per detrital sample. Analytical errors and procedures are described elsewhere (Gehrels et al., 2008; Gehrels, 2012). A 90%–110% concordance filter was applied to the zircon-grain analyses; the filter resulted in rejection of 2% of the grains from the detrital-grain suite. Rejected grain ages are indicated by strike-through text in Supplemental Table 1. In this paper, we employ the 2012 GSA Time Scale (Walker et al., 2012).

### STRATIGRAPHY AND STRUCTURAL GEOLOGY OF CASTLE VALLEY

The northwestern end of Castle Valley contains superb exposures of Cutler Group strata and geometric relations that are the keys to understanding the interaction of the strata with the developing salt wall, controls on facies distribution, and sources of sediment for the Cutler Group (Fig. 5). The Castle Valley Sandstone and undifferentiated Cutler Formation are described briefly with other local stratigraphic units in the following section, and in more detail in the section on sedimentology.

#### General Stratigraphy

**Paradox Formation**

Pennsylvanian Paradox evaporitic strata are mostly covered beneath surficial deposits of the valley floor in the northern extent of the valley, but limited exposures are present in the northwesternmost part of the valley along the base of cliffs directly south of the Castle Creek gorge and north of the creek in a wedge-shaped exposure that separates fluvial Cutler strata on the west from lower members of the Moenkopi Formation on the east (Figs. 5 and 6). The Paradox Formation forms low mounds mantled by gypsum crusts south of Castle Creek, but gypsum and shale are well exposed in a stream bank on the north side of Castle Creek.
Figure 5. Geologic map of northwestern end of Castle Valley. Lines of cross sections of Figure 6 indicated. Labels 7a, 7b, 7d, and 7e indicate locations of photo sites in Figure 7. Oblique aerial view of Figure 7C spans south edge of map.
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(continued)
Undifferentiated Cutler Formation

Permian red sandstone and conglomerate are exposed along cliff bases directly south of Castle Creek, where they abut Paradox exposures or surficial deposits, and in an amphitheater-like bowl north of Castle Creek, where Cutler beds form a wedge of strata in which dip values decrease upsection but are everywhere steeper than those of overlying eolian strata. In the amphitheater, Cutler strata strike north and northeast; individual beds steepen and thin toward Paradox and Moenkopi strata on the northeast flank of the salt wall. Near-vertical conglomeratic beds are truncated beneath eolianite strata high on the northern slope of the amphitheater (Fig. 6, section C–C′), where conglomeratic Cutler beds are bleached white below the contact. Cutler strata thicken into northwest-elongate, salt-withdrawal minibasins on both flanks of the Castle Valley salt wall (Doelling and Ross, 1998; Doelling, 2002a, 2002b; Kluth and DuChene, 2009; Trudgill, 2011).

Castle Valley Sandstone

The Castle Valley Sandstone overlies undifferentiated Cutler strata on a discordant, sharp contact that ranges from only slightly angular on Castle Creek (Fig. 7A) to strongly discordant at northernmost exposures. Bed sets of cross-bedded sandstone within the eolianite onlap Cutler strata along Castle Creek in the direction of the former salt wall (Fig. 7B). The eolianite crops out as cliffs on the southwestern flank of Castle Valley (Fig. 5) and thins to a pinch out beneath the Moenkopi Formation southeastward along the valley wall (Fig. 7C; Doelling and Ross, 1998; Doelling, 2002a, 2002b; Kluth and DuChene, 2009; Trudgill, 2011). The Castle Valley Sandstone is 158 m thick along Castle Creek (Fig. 8). Eolian strata are present in the subsurface of the salt-withdrawal minibasin, termed the Big Bend minibasin (Matthews et al., 2004, 2007; Banham and Mountney, 2013), between the Castle Valley and Moab Valley salt walls, where 140 m of massive sandstone was penetrated beneath the Moenkopi Formation in the Burkholder #1 well, 7 km southwest of the outcrop (Fig. 4; Trudgill, 2011; Parr, 2012). The Castle Valley Sandstone does not crop out on the northeastern flank of the salt wall but is present in the subsurface north of the plunging nose of the structure, where the eolian sandstone is ~50 m thick in the Grand River O and G #1 State well (Fig. 4; Trudgill, 2011; Parr, 2012). The Castle Valley Sandstone is not exposed along the northeastern flank of the valley (Doelling and Ross, 1998; Doelling, 2002b), both because it was originally depositionally thin east of the nose of the salt wall (Fig. 4) and because it is truncated in the subsurface beneath the Moenkopi along the trend of the salt wall, as it is in outcrop on the southwest side of the salt wall (Fig. 7C).

Moenkopi Formation

The Lower Triassic Moenkopi Formation unconformably overlies the Castle Valley Sandstone on a discordant contact (Figs. 5, 6, and 7C). The top of the Castle Valley Sandstone is an irregular surface marked by extensive desert varnish, pitting, and small sculpted ridges with amplitudes <1 m in the surface of the sandstone. The Moenkopi Formation is locally composed of four members defined in nearby eastern Colorado and at Castle Valley (Shoemaker and Newman, 1959).

TABLE 1. RECALCULATED MODAL POINT-COUNT DATA FOR CUTLER GROUP SANDSTONES, PARADOX BASIN (continued)

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Note: Chert is included in Qt, Lt, and Qp grain categories.
2Locations: Castle Valley (CV)—west side of valley 700 m south of Castle Creek; Castle Valley W—west side of valley 1.7 km south of Castle Creek; Castle Rock—northeast side of valley south of Castle Rock; CV anticline—crest of anticline north of Castle Creek; South of Castle Creek—west side of valley 300 m south of Castle Creek.
3Approximate detrital-zircon locality 11CT01 at Castle Rock.
Figure 6. Geologic cross sections of northwestern end of Castle Valley salt wall. Map symbols as in Figure 5, plus Pcc, undifferentiated Castle Valley Sandstone. (A) Cross section A–A′. (B) Cross section B–B′. (C) Cross section C–C′. (D) Cross-section D–D′.
Figure 7. Field photos and photomicrograph of Castle Valley Sandstone. (A) Contact of Castle Valley Sandstone (Pccl) with underlying red arkosic strata of undifferentiated Cutler Formation (Pca) directly north of Castle Creek. Scale bar is 1.5 m high. (B) Stratigraphic relations in Castle Valley Sandstone directly north of Castle Creek. View north. Pccl—upper bleached part of lower eolianite member; Pcci—interdune member; Pccu—upper eolianite member. Onlap of interdune member by upper member is demonstrated by thinning of lowermost set of upper member cross-beds, which pinches out updip of normal fault indicated by white arrow. Small foresets in upper left dip counter (black arrow) to larger foresets in lower part of upper member. Interdune member is 7.5 m thick at right edge of photo. (C) Erosional truncation of Castle Valley Sandstone beneath Moenkopi Formation along Porcupine Rim on southwestern flank of Castle Valley. Base of Castle Valley Sandstone indicated by white arrow. Pca—undifferentiated Cutler Formation arkose; Pccl—lower eolianite member of Castle Valley sandstone; Pcci—interdune member; Pccu—upper eolianite member; Trm—Moenkopi Formation; Trc—Chinle Formation; Jw—Wingate Sandstone. The inferred paleofluid contact in the Castle Valley Sandstone, marked by contact between pink and white strata of lower eolianite unit, is truncated beneath the unconformity (orange arrow). Upper cliff of Porcupine Rim is 60 m high. (D) Fault in upper eolianite unit (Pccu) draped by basal Moenkopi beds (Trm). View north. Fault scarp is 3 m high. (E) Tar-saturated medium- to coarse-grained sandstone of grain-flow laminae alternate with fine-grained grain-fall and wind-ripple laminae, upper member of Castle Valley Sandstone. Hammer handle is 42 cm long. (F) Photomicrograph of lower eolianite member of Castle Valley Sandstone (sample CVSM23JA) near crest of Castle Valley anticline. Most grains are monocrystalline quartz; rounded grain in center is partially dissolved plagioclase; black interstitial material is tar; blue material is epoxy (pore space). Bimodal texture is evident.
Figure 8. Measured section of Castle Valley Sandstone along Castle Creek. Dip tadpoles adjacent to section indicate corrected foreset dip azimuths from measured section. Stratigraphic variation in QtFL%Qt (open dots) and QtFL%F (black dots) indicated by curves to right of dip tadpoles. Stereonets indicate all dip directions and dip values (black dots), and resultant rose diagrams for the lower and upper eolianite units measured throughout the study area (see Fig. 10 for off-section localities).
These members, which vary in thickness on the flanks of the Castle Valley salt wall (Fig. 6; Lawton and Buck, 2006; Banham and Mountney, 2013), include the Tenderfoot, Ali Baba, Sewemup, and Parriot members (spelling of Parriot Member by Shoemaker and Newman [1959] differs from the modern spelling of Parriott Mesa, adjacent to Castle Valley). The Moenkopi Formation ranges from 135 to 375 m thick in exposures adjacent to Castle Valley (Lawton and Buck, 2006) but probably exceeds 400 m along the Colorado River northeast of Castle Valley where the basal contact lies in the subsurface (Shoemaker and Newman, 1959).

**Upper Triassic and Jurassic strata.** The Upper Triassic Chinle Formation overlies the Moenkopi throughout the Castle Valley region. The contact lies at the base of a coarse-grained angular sandstone on two hilltops north of the amphitheater, where an angular unconformity at the base of the Chinle truncates 50 m of the Parriot Member of the Moenkopi Formation. The Chinle is overlain by thick-bedded eolian strata of the Jurassic Wingate Sandstone, which forms a steep escarpment, termed the Porcupine Rim, along the southwestern flank of Castle Valley (Fig. 7C). Ledgy outcrops of the Jurassic Kayenta Formation form the rim of the escarpment (Figs. 5 and 6).

**Structural Geology**

The major structural feature of the NW end of the Castle Valley is an open northwest-plunging anticline that forms the continuation of the salt wall beyond the inferred surface extent of the Paradox Formation. The anticline, best seen on cross section A–A′ (Fig. 6), is expressed in the Castle Valley Sandstone and Moenkopi Formation north of Castle Creek (Fig. 5) and plunges northward to terminate abruptly at the Colorado River; it has little structural expression in the Chinle Formation, which thins southeastward onto the anticlinal nose by onlap onto the Moenkopi Formation (e.g., Shoemaker and Newman, 1959; Matthews et al., 2004, 2007). North of Castle Creek, a monocline with ~150 m of structural relief in the middle part of the Moenkopi Formation trends northward, obliquely to the crest of the anticline, and decreases in amplitude along its northward plunge to a termination at the Colorado River. The monocline creates only 30 m of structural relief on the unconformity at the base of the Chinle Formation, and 49 m of the Parriot Member is removed beneath the unconformity across the monocline (Fig. 6, D–D′), relations which indicate that most structural development of the monocline took place prior to Chinle deposition.

North of Castle Creek, the Paradox Formation occupies a northward-tapering exposure that separates fluvial Cutler strata on the west from the Tenderfoot Member of the Moenkopi Formation on the east. The Paradox Formation thins and pinches out northward, merging into a fault-like surface that separates Cutler from Moenkopi strata, all of which face away from the surface. This surface represents a salt weld (e.g., Jackson and Cramez, 1989) formerly occupied by diapiric evaporite between the Cutler and Moenkopi beds (Lawton and Buck, 2006). The weld continues upslope and is truncated at the base of the Moenkopi Formation in the anticlinal hinge of the monocline (Fig. 6). Cutler Formation beds west of the Paradox Formation strike north to northeast, steepening and thinning progressively to onlap the Paradox Formation and the weld. All Moenkopi strata thin over the anticline and thicken eastward across the monoclinal flexure above the weld (Fig. 6).

High-angle faults of several trends displace the Cutler Formation and Castle Valley Sandstone. Faults trending west and ENE (~060°) have apparent normal offsets of 15–50 m and displace former fluid contacts, marked by color changes in the Castle Valley Sandstone (e.g., Goren and Chan, 2015) and strata as young as Middle Jurassic on the Porcupine Rim (Fig. 5). These faults have subhorizontal slickensides, are associated with minor anticlinal slip and reverse faults, and are present in the part of the map area where evaporite of the former salt wall pinches out northward, suggesting that they accommodated differential NE-SW shortening where the mechanically weak salt pinches out or thins to the point that it did not exert mechanical control on anticline development. Northwest-, NE-, and north-trending faults with less than 5 m of displacement affect only the undifferentiated Cutler Formation and eolianite beds and terminate in the lowermost Moenkopi Formation. A conspicuous, north-trending normal fault that crosses the crest of the anticline offsets the Castle Valley Sandstone–Moenkopi contact 4 m and is draped by a wedge of strata in the lowermost Moenkopi, indicating fault movement ended near the beginning of Moenkopi deposition (Fig. 7D; Banham and Mountney, 2013).

In addition to local draping of faults by basal Moenkopi strata, thickness trends and stratal geometries of Permian and Early Triassic strata demonstrate syndepositional growth of the folds in the study area. Upturn and onlap of undifferentiated Cutler beds onto the Paradox Formation and weld indicate that the beds were deposited directly on exposed Paradox epeirite and subsequently folded by continued diapirism to form halokinetic sequences adjacent to the diapir (sensu Giles and Lawton, 2002; Rowan et al., 2003; Giles and Rowan, 2012). Thinning of eolianite and Moenkopi beds across the crest of the northwest-trending anticline and thickening of the Moenkopi Formation eastward across the monocline indicate that both the anticline and the monocline grew during deposition of the Castle Valley Sandstone and Moenkopi Formation (Fig. 6; Lawton and Buck, 2006; Banham and Mountney, 2013). The growth of these structures was largely complete by deposition of the lower part of the Chinle Formation. A restoration of cross section D–D′ from the Late Triassic back through deposition of the eolianite indicates that the Paradox Formation was exposed on the crest of the monocline for much of the growth history of the anticline (Fig. 9). Although it is not clear that the Paradox Formation was always exposed during deposition of the Castle Valley Sandstone, sedimentological and compositional features described below suggest that the diapir was exposed during at least some parts of the depositional history of the eolianite and that the diapir had topographic relief that influenced distribution of eolian sand.
**SEDIMENTOLOGY OF PERMIAN UNITS**

The undifferentiated Cutler Formation and Castle Valley Sandstone constitute the exposed components of the Permian depositional system in Castle Valley. The undifferentiated Cutler Formation consists of reddish-brown, angular, medium- to very coarse-grained, poorly sorted sandstone and pebbly sandstone. Sandstone beds consist of upward-fining, broadly channel-form bodies 2–4 m thick with scour fills and pebble-filled troughs near bed bases. Pebbly trough cross-beds decrease in height from 35 to 10 cm upsection through channel bodies. Trough cross-beds are typically overlain by horizontal, discontinuous, normally graded laminae 5–15 mm thick, locally with pebbles dispersed on the laminae. Horizontally laminated sandstone is interpreted as deposits of antidunes by unconfined flow in the shallow parts of channel systems (e.g., Blair and McPherson, 1994). Burrows and rootlet
traces occupy the upper parts of some channel bodies. Pebbles and cobbles are angular to rounded, as much 28 cm in diameter, and consist of granite, biotite schist, gneiss, and quartz. A single paleocurrent site consisting of trough cross-beds on Castle Creek near meter 20 of the measured reference section yielded a NNW (355°) paleocurrent direction (Fig. 8). The resultant direction is similar to consistent northwest-directed paleocurrent indicators reported from approximately correlative Cutler red beds elsewhere in Castle Valley on the SW flank of the salt wall and in the upper part of the undifferentiated Cutler Formation in the Onion Creek drainage, 10 km to the northeast (Fig. 4; Buller, 2009), as well as mean paleocurrent directions reported from the upper part of the Cutler Formation in the same area (Venus et al., 2015). The northwest-oriented sediment-dispersal directions in the upper part of the Cutler section indicate fluvial dispersal parallel to the Castle Valley salt wall. Northwestward sediment transport parallel to the elongate minibasins has also been documented in the overlying Moenkopi Formation (Banham and Mountney, 2013, 2014). Deposition of the undifferentiated Cutler section took place in pebbly braided channel systems in the salt-withdrawal minibasins on both sides of the wall.

The Castle Valley Sandstone consists of two sandstone units with large-scale cross-stratification separated by a laterally continuous, light reddish-brown sandstone interval containing pebble lags and scattered pebbles (Fig. 8). These stratigraphic units, continuous throughout exposures of the Castle Valley Sandstone, are termed the lower eolianite, interdune, and upper eolianite members. Sedimentologic features of the lower and upper eolianite members are similar and are described together. The lower eolianite is 40 m thick along Castle Creek, where it overlies the undifferentiated Cutler Formation on a sharp contact (Fig. 7A) that truncates syndepositional normal faults with centimeter traces of displacement in the underlying arkose. Northward and up dip from the measured section, the contact is gradational and interfingering, with apparent reworking of eolian sandstone into pebbly facies. The lower eolianite consists of moderately sorted, light reddish-brown sandstone with planar cross-beds in tabular sets 2.5–12 m thick in its lower part. The uppermost 5 m of the lower member is composed of trough-form foresets 0.25–0.5 m thick and is white to yellow along the entire outcrop belt, in contrast with the uniform light reddish-brown color of the rest of the member. The upper eolianite is 108 m thick along Castle Creek and consists of well-sorted white to tan sandstone with dominant planar cross-beds in tabular to broadly wedge-shaped sets 1–22 m thick. Subordinate trough cross-beds generally less than 1 m thick are present in the lower part of the member (Fig. 8). At Castle Creek, the lower contact of the upper member is unconformable, indicated by thinning and onlap of the basal dune bed set onto the interdune member (Fig. 7B). The large-scale foresets of both eolianite members are composed of alternating lamina sets of very fine to fine-grained and medium- to coarse-grained sandstone. Coarser laminae are 3–20 mm thick and laterally continuous to broadly lenticular on bedding-parallel surfaces, and have scoured bases. Fine-grained laminae are ~1 mm thick and form co-sets 5–10 cm thick between the coarser laminae to yield a marked cyclicity in the large-scale cross-beds, especially in the upper eolianite (Fig. 7E). Some large-scale bed sets in the upper eolianite are complex and contain small-scale foresets with dip directions opposite to those of the large foresets (Fig. 7B), indicating that smaller bed forms climbed the large-scale dunes. Uncommon low-angle laminae of climbing translatent strata are present in the upper member (Fig. 8). Slumping of slip-face bed sets was not observed in either eolianite member.

The interdune member overlies the lower eolianite on a sharp contact, and consists of reddish-brown medium- to very coarse-grained pebbly sandstone with discontinuous, weakly defined, inversely graded horizontal laminae and shallow pebble-filled scours. The lower contact locally consists of a surface with as much as 50 cm of relief filled locally by a pebble lag and elsewhere by separate pebble lenses in fine-grained sandstone. Pebbles consist of angular to subrounded granules and pebbles of granite, quartz pebbles as much as 2 cm in diameter, and angular clasts of medium-gray dolostone to 6 cm long. The dolostone clasts resemble dolostone of blocks in diapiric Paradox evaporite exposed elsewhere in Castle Valley and also resemble dolostone caprock that flanks the salt wall (Figs. 5 and 6; Shock, 2012). The interdune member is 75 m thick at creek level and thins up dip to a pinch out but persists as an erosion surface with discontinuous pebble lags between the eolianite members to the northern extent of exposure.

Sedimentary structures of the Castle Valley Sandstone indicate deposition by large dunes preserved adjacent to the Castle Valley salt wall. Coarse-grained inversely and normally graded sandstone laminae with scoured bases in large planar foresets are interpreted as sand-flow or grain-flow laminae (e.g., Hunter, 1977; Mountney, 2006), and intervening thin laminae are interpreted as grainfall deposits and wind-ripple laminae (Hunter, 1977; Fryberger et al., 1988), which together constitute slip-face deposits (e.g., Mountney, 2006). Marked cm-scale cyclicity in slip-face deposits containing successions of climbing translatent strata, grain-fall and sand-flow laminae, like those in the Castle Valley Sandstone (Fig. 7E), have been interpreted as a consequence of diurnal onshore sea breezes in coastal dune fields of the Texas Gulf Coast (Hunter, 1977) and thus might record very rapid accumulation of Permian dune sediment, an inference supported by detrital-zircon data described below. Smaller wedge sets represent subordinated transverse or barchanoid dunes superimposed on the larger dunes, with opposing dips possibly resulting from seasonal changes in wind directions. Broadly trough-shaped foresets at the top of the lower eolianite member are interpreted as deposits of small barchanoid dunes deposited prior to cessation of lower member deposition.

The interdune member represents fluvial sheet-flood deposits. Unconfined, supercritical flow created antidune deposits recorded by discontinuous inversely graded sandstone layers with scattered pebbles (e.g., Blair and McPherson, 1994). The flow reworked the upper part of the lower eolianite member and steeply-dipping undifferentiated Cutler beds, the source of many of the pebbles, and also transported pebbles from resistant dolomite blocks and caprock of the diapir, which was exposed during deposition of the interdune member (e.g., Fig. 9F). The interdune member thins up dip toward the diapir as a result of decreased accommodation toward the diapir crest.
Foreset orientations in the Castle Valley Sandstone vary but indicate predominant transport directions to the southwest and southeast (Figs. 8 and 10). The lower eolianite member, in which most of the paleocurrents were measured on the Castle Creek section, is dominated by southwest foreset dip directions, although there are a few large foresets oriented to the northeast and southeast (Fig. 8). The dominant southwest transport direction indicates that the sand was deposited on the lee side of the salt wall. Transport directions in the upper eolianite member were to the southwest and southeast, perpendicular to and parallel to the salt wall, respectively. The variability of wind directions indicated by foreset orientations in both units suggests that the Castle Valley Sandstone records deposits of a draa that consisted of several types of superposed dunes, possibly star, linear, and transverse dunes, in a dune field adjacent to the salt wall. It thus constitutes the exposed part of an irregularly distributed dune field that was preserved against the flanks of salt walls and in intervening minibasins (Fig. 4; e.g., Dubiel et al., 1996; Trudgill, 2011; Parr, 2012; Rasmussen, 2014).

**SANDSTONE PETROGRAPHY**

Sandstone composition and texture change upsection from the Cutler fluvial facies into the eolian facies of the Castle Valley Sandstone and likewise change upsection through the Castle Valley Sandstone itself. The upper member of the Castle Valley Sandstone is compositionally similar to the White Rim Sandstone to the southwest.

The undifferentiated Cutler Formation consists of poorly to moderately sorted, fine- to very coarse-grained angular to subangular sandstones with pervasive but unevenly distributed hematite grain rims locally as much as 10 µm thick. Coarsely crystalline, granular to rhombohedral calcite forms small patches and locally forms pervasive pore-filling cement. The fluvial strata constitute compositional arkoses with an average composition of Qt56F42L2 (n = 4; Table 1). Potassium feldspar, including abundant microcline, comprises ~60% of the feldspar population, and subordinate plagioclase ranges from unaltered to partly dissolved and replaced by calcite (Fig. 7F). Biotite and muscovite are common, reaching a maximum of 12% of detrital grains; biotite constitutes about three-fourths of the mica. Metamorphic lithic fragments, which include foliated quartz-muscovite schistose fragments and a single observed metarhyolite grain, are uncommon in the samples studied. Granitic rocks dominated the source area of the fluvial sandstones.

The Castle Valley Sandstone ranges from compositional arkose to subarkose; mean quartz content increases upsection (Figs. 8 and 11). Nearly ubiquitous bimodal sandstone textures result from laminae of alternating, coarse, subrounded to rounded grains and fine angular to rounded grains. Grains in coarse-grained laminae range from 0.4 to 2.5 mm in diameter; those in fine laminae are well sorted and consistent at 0.1–0.3 mm in diameter. In the red part of the lower eolianite member, coarse laminae contain grains of rounded monocrystalline quartz, subrounded to rounded potassium feldspar, includ-

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**Figure 10.** Map of wind directions in Castle Valley Sandstone indicated by restored eolian foreset dip directions (dip tadpoles) measured throughout the field area. Locations other than Castle Creek section are averages of ten foreset measurements (Parr, 2012). Data are revised from Parr (2012), except tadpoles along Castle Creek, which are taken from Figure 8 as space permits. A single fluvial paleocurrent locality in the upper part of the undifferentiated Cutler Formation is indicated by arrow north of Castle Creek.
samples. Biotite is typically absent but constitutes 3% of detrital grains in one lithic grain type not observed in the undifferentiated Cutler or lower eolianite intervals are distinguished by the presence of rare chert grains, a sedimentary source, are present and even common in some samples. Both white quartz overgrowths, which indicate recycling of some quartz grains from a sedimentary aspect characteristic of the eolianite. Hematite rims are absent. The fine-grained fraction of both lower and upper eolianite members is somewhat richer in quartz than the coarse-grained fraction; primary quartz overgrowths, developed in situ, are not present in either of the eolianite members.

The interdune member consists of angular to rounded, moderately sorted sandstone with rounded grains to 0.5 mm in diameter, although most grains are subangular to rounded in the range 0.1–0.2 mm in diameter. The grain rounding and range of grain sizes of the unit are thus similar to those of the lower eolianite member, but the sandstones lack the segregated bimodal textural aspect characteristic of the eolianite. Hematite rims range from light to moderate, and are not present on all grains, suggesting recycling of grains from previously oxidized sandstones. The average composition of two samples is Qt80F18L2. Plagioclase constitutes somewhat less than one-third of the feldspar content. Only one sample (CV91; Table 1) contains appreciable metamorphic lithic fragments (3%) consisting of schistose and polygonal quartz-mica fragments. The interdune member is Qt87F12L1 (Table 1), but they differ substantially in feldspar proportions.

White Rim Sandstone samples of the greater erg to the southwest are similar in texture and composition to the white parts of the Castle Valley Sandstone. A sample collected near the base of the Shafer Trail in Canyonlands National Park (Fig. 3; 1WRB) lacks coarse grains, whereas a sample from Hite on Lake Powell (11WRB) contains laminae with scattered coarse grains. As with the Castle Valley Sandstone, the grains are better rounded in the coarser laminae and as much as 0.75 mm in diameter. The two White Rim samples have an average composition of Qt83F16L1 (Table 1), but they differ substantially in feldspar composition: The Hite sample has 16% feldspar and the Shafer Trail sample...
has only 8% feldspar. The alkali feldspar is exclusively orthoclase; microcline is absent. Plagioclase is present but subordinate to orthoclase, and mica is absent. Our counts are somewhat more feldspathic than the mean composition, Qt17F∞L∞, of the White Rim Sandstone between the Green and Colorado rivers in Canyonlands National Park (Fig. 10, Table 1; n = 76; Steele-Mallory, 1982).

In summary, sandstone composition changes upsection through the Castle Valley Sandstone (Table 1). Total quartz decreases from an average of 56% in the fluvial facies to 92% and 90% in the upper white part of the lower eolianite and upper eolianite, respectively. Total feldspar decreases in concert with the upsection increase in quartz, from 42% in the fluvial facies to 8% and 10% in the white eolianite intervals (Fig. 8). Common transported overgrowths on quartz grains and uncommon chert in the white intervals provide evidence for sedimentary rocks in the source area; chert and transported overgrowths were not observed in the fluvial facies or the lower eolianite member. The interdune member is somewhat more feldspathic than the underlying white part of the upper eolianite and so breaks the monotonic trend to higher quartz content with height in the section. It also contains the only detrital carbonate grains observed in the sandstone suite. As noted, the composition and texture of the interdune unit likely resulted from recycling of resistant dolomitic clasts and caprock from the diapir and fluvial Cutler strata upturned adjacent to the diapir, realistic possibilities indicated by the structural restoration (Fig. 9F).

The stratigraphic trends toward increased compositional and textural maturity can be explained by appealing to two possible, but mutually exclusive, mechanisms. Upsection loss of feldspar and concomitant enrichment in quartz could have resulted from feldspar destruction in a single sand population during transport-related abrasion, perhaps aided by chemical weathering of feldspar. The other possibility is that locally derived arkosic sands were mixed with and diluted by a sand population transported to the basin from another source by eolian processes. There is abundant evidence for postdepositional loss of pristine plagioclase, and analogous chemical weathering could have affected the grain populations during transport. On the other hand, the presence of a sedimentary source for grains in the upper part of the Castle Valley Sandstone, indicated by transported quartz overgrowths and chert grains, corroborates the second hypothesis of dilution of local sand populations by eolian sand input. Increased rounding of grains in younger eolianite strata resulting from more extensive transport-related abrasion presumably could have taken place in either scenario.

Detrital zircons from the various units provide a basis for selecting between the two hypotheses.

### DETRITAL ZIRCONS

U-Pb detrital-zircon geochronology of samples from the Cutler Group indicates significant differences in detrital-zircon content among the fluvial facies of the undifferentiated Cutler Formation, the Castle Valley Sandstone, and the White Rim Sandstone of the greater erg in Canyonlands National Park and near Hite, Utah. Samples were collected from the undifferentiated Cutler Formation ~56 m beneath the unconformity with the Moenkopi Formation on the northeast side of Castle Valley (Fig. 4; sample 11CT01) and north of Castle Creek ~30 m stratigraphically beneath the base of the Castle Creek measured section (sample 11CVC01). Castle Valley Sandstone samples were collected on and near the Castle Creek measured section (Fig. 8), and White Rim Sandstone samples were collected at South Fork Wash near the base of the Shafer Trail in Canyonlands National Park and from a road cut on Utah Highway 95 on the north side of the Colorado River (Lake Powell) opposite the site of the former Hite Marina, 1.8 km east of the confluence of the Colorado and Dirty Devil rivers (Fig. 3).

### Zircon Age Populations

Seven zircon grain-age populations were defined on the basis of all U-Pb detrital grain ages measured in the sample set (Fig. 12A; n = 729 individual grain analyses; N = eight sandstone samples; Supplemental Table 1 [see footnote 1]). The discordance and error filters resulted in the rejection of 17 analyses (of 746 total); the rejected grains, of which three are Archean, one is 1550 Ma, five are Grenville, and eight are Paleozoic, are indicated in strikethrough text in Supplemental Table 1 (see footnote 1). Detrital-zircon grain ages range from ca. 3339 Ma to ca. 299 Ma; age populations consist of grain clusters on the probability distribution plots separated by age gaps and, except for Population B, defined below, contain one or more age peaks (Fig. 12A). Table 2 is a compilation of grain-population ages, age peaks, numbers of zircon grains in each population and inferred sources for the grains.

#### Population A (~3539–2548 Ma)

Archean zircons constitute 4% of grains (n = 32), with a wide range of ages with overlapping 1σ age uncertainties. Archean zircons are absent from the fluvial facies of the Cutler Formation, being restricted to the Castle Valley Sandstone (5% of all analyses) and the White Rim Sandstone (9%).

#### Population B (~2456–2008 Ma)

Older Paleoproterozoic zircons are uncommon and represent only 2% of the total grain population (n = 14). This is a population of dispersed grain ages that do not all overlap at 1σ uncertainty; therefore, the population does not contain any significant age peaks. Grains of this population are present in all samples of the Castle Valley Sandstone and the White Rim Sandstone (Figs. 12B, 12D, and 13) but are absent from the Cutler fluvial facies (Figs. 12C and 13). Basement rocks of this age (2.3–1.8 Ga) are present in the Wopmay orogen of northwestern Laurentia and the Trans-Hudson orogen of central Laurentia.
Basement rocks of this approximate age range (2.25–2.05 Ga) are also present in the Maroni-Itacaiunas basement province on the northeastern flank of the Amazonian craton (Cordani et al., 2009; Cardona et al., 2010). Detrital-zircon grains ranging ~2.25–2.00 Ga, of inferred Gondwanan derivation, are present in Paleozoic sedimentary rocks of the Suwannee terrane of Florida, southeastern Alabama, and southern Georgia (Mueller et al., 2014).

Population C (~1991–1599 Ma)

Late Paleoproterozoic zircons of population C constitute abundant grains in the analyzed samples and include 22% of all grain analyses (n = 161) with dominant peaks near 1750 Ma, 1723 Ma, 1686 Ma, and 1650 Ma. This population is of roughly equal abundance in all samples, ranging from 17% to 27% of grains in individual samples; it is somewhat more abundant in White

Figure 12. Probability density plots and grain-age histograms of Cutler Group sandstones. Detrital-zircon populations indicated across top of plots and highlighted by vertical color bars. N is number of samples; n is number of individual grain analyses. All histogram bins are 50 m.y. (A) Age distribution of all zircon grain analyses of this study. (B) Age distribution of Castle Valley Sandstone samples. (C) Age distribution of undifferentiated Cutler Formation samples. (D) Age distribution of White Rim Sandstone samples.
Rim samples (26% and 27%) than in fluvial and eolian Cutler samples. Despite the apparent equal grain abundance in all samples, the range of grain ages in the fluvial Cutler samples is significantly narrower than in the White Rim and Castle Valley Sandstone samples, being restricted to ~1790–1689 Ma, with a mode in the 50 m.y. range of 1750–1700 Ma (Fig. 12C). The distinctive, more restricted age range of ~1790–1689 Ma is designated subpopulation C′.

Grains of population C could have been derived from basement of the Yavapai-Mazatzal province in southwestern Laurentia, and the Trans-Hudson, Central Plains, and Penokean provinces in the interior of Laurentia (see Dickinon and Gehrels, 2009, for a synthesis of basement-age provinces of Laurentia). In the Early Permian, nearby sources of these grains were present in exposed basement rocks of the Uncompahgre uplift. These sources include metavolcanic and granitic rocks now exposed in Unaweep and Black Canyons (Fig. 1) that range in age from 1755 Ma to 1670 Ma (Bickford et al., 1989; Jessup et al., 2006; Gonzales and Van Schmus, 2007; Dickinson and Gehrels, 2009; Jones et al. (2009).

TABLE 2. U-PB DETRITAL-ZIRCON AGE POPULATIONS IN CASTLE VALLEY EOLIANITE, CUTLER GROUP, AND WHITE RIM SANDSTONE

<table>
<thead>
<tr>
<th>Age population</th>
<th>Age range (Ma)</th>
<th>Age peaks (Ma)</th>
<th>n</th>
<th>% of Total</th>
<th>Possible ultimate source</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>~3539–2548 (Archean)</td>
<td>Ca. 2697</td>
<td>32</td>
<td>4</td>
<td>(1) Wyoming craton; (2) Paleoproterozoic quartzite units of Uncompahgre uplift and San Juan dome</td>
<td>Jessup et al. (2006); Dickinson and Gehrels (2009); Jones et al. (2009)</td>
</tr>
<tr>
<td>B</td>
<td>~2456–2008 (Early Paleoproterozoic)</td>
<td>Ca. 1842, 1800, 1783, 1750, 1723, 1666, 1650</td>
<td>161</td>
<td>22</td>
<td>(1) Yavapai province of SW Laurentia; (2) metamorphosed volcanic and plutonic rocks of Uncompahgre uplift (Gunnison area and Black Canyon) and Needle Mountains of San Juan dome; (3) Paleoproterozoic quartzite units of Uncompahgre uplift and San Juan dome</td>
<td>Bickford et al. (1989); Jessup et al. (2006); Gonzales and Van Schmus (2007); Dickinson and Gehrels (2009); Jones et al. (2009)</td>
</tr>
<tr>
<td>C</td>
<td>~1991–1599 (Late Paleoproterozoic)</td>
<td>Ca. 1575, 1550, 1441</td>
<td>267</td>
<td>37</td>
<td>(1) 1.4 Ga granite-rhyolite suite of Laurentia; (2) 1.44–1.35 Ga plutons in Uncompahgre uplift (Unaweep and Black canyons) and San Juan dome (Needle Mountains); (3) Sveconorwegian orogen; (4) Pinware terrane, SE Labrador</td>
<td>Bickford and Cudzilo (1975); Tucker and Gower (1984); Anderson (1989); Tewksbury (1989); Wasteneys et al. (1997); Jessup et al. (2006); Gonzales and Van Schmus (2007); Bingen and Solli (2009)</td>
</tr>
<tr>
<td>E</td>
<td>~1288–900 (Late Mesoproterozoic–Early Neoproterozoic)</td>
<td>Ca. 643, 624, 608, 569, 536</td>
<td>57</td>
<td>8</td>
<td>Iapetan syntaxis volcanics of eastern and southern Laurentia (~765–530 Ma); Pan-African and peri-Gondwanan terranes of Appalachian orogen and south of Ouachita orogen</td>
<td>Thomas (2011, 2014); Mueller et al. (2014)</td>
</tr>
<tr>
<td>F</td>
<td>~734–499 (Late Neoproterozoic–Cambrian)</td>
<td>Ca. 455, 424, 380, 328</td>
<td>74</td>
<td>10</td>
<td>Greater Appalachian orogen; Taconic orogen (~490–440 Ma); Acadian orogen (~420–350 Ma); Alleghanian orogen (~330–270 Ma)</td>
<td>Dickinson and Gehrels (2003); Thomas (2011)</td>
</tr>
</tbody>
</table>

1Age ranges are cohorts of ages with overlapping 2σ uncertainties separated by age gaps.
2Peak ages picked using Age Pick algorithm of Gehrels (https://docs.google.com/document/d/1MYwm8GcdYFsOvIN62BPLUd-g2r1AS3Vmm4gHMOFxg/preview). Listed peaks (plain text) based on nine or more grains, major peaks (bold text) on 39 or more grains.
3n = number of analyses in age population.
4Total = total number of analyses = 729.
with nearly unimodal age peaks in four samples at 1762 Ma, 1750 Ma, 1746 Ma, and 1740 Ma. Thus, known Proterozoic rock ages in the Uncompahgre uplift east of Castle Valley (Fig. 4) and the San Juan dome to the southeast lie in the range of most zircon ages of subpopulation $C'$ in the fluvial Cutler Formation.

Population D (~1581–1302 Ma)

Early Mesoproterozoic grain ages ($n = 267; 37\%$), with a dominant peak near 1441 Ma, dominate the sample set; nevertheless, this population is unequally distributed between fluvial and eolian facies. The population composes 76% of the two undifferentiated Cutler fluvial samples, in which the age range of grains, ~1466–1406 Ma with one outlier at ca. 1349 Ma, is more restricted than the general population. The age range ~1466–1406 Ma is designated subpopulation $D'$. Population D makes up 27% of the Castle Valley Sandstone and 14% of the White Rim Sandstone samples. The main source for this grain age population in Laurentia is the 1.4 Ga granite-rhyolite suite that extends across the Yavapai and Mazatzal basement provinces (Anderson, 1989; Dickinson and Gehrels, 2009). Potential local sources in the Uncompaghre uplift include the Vernal Mesa Monzogranite in Black Canyon (1434 ± 2 Ma, U-Pb zircon; Jessup et al., 2006), texturally and mineralogically similar quartz monzonite in Unaweep Canyon (1443 ± 22 Ma, U-Pb zircon; Bickford and Cudzilo, 1975), and cross-cutting pegmatite in Black Canyon (1413 ± 2 Ma, U-Pb zircon; Jessup et al., 2006). In the Needle Mountains of the San Juan dome, dated rocks in this age range include the Eolus Granite (1442 ± 3 Ma to 1435 ± 3 Ma, U-Pb zircon upper intercept; Gonzales and Van Schmus, 2007) and Trimble Granite (ca. 1350 Ma; Tewksbury, 1989). The latter age is represented by a single grain in the undifferentiated Cutler samples. Grains older than 1470 Ma in this population ($n = 28$), which are present in the White Rim and Castle Valley sandstones (Supplemental Table 1 [see footnote 1]), are not readily attributable to a local Uncompaghre source.

Population E (~1288–900 Ma)

This population of Late Mesoproterozoic to Early Neoproterozoic grains, referred to as Grenville grains, constitutes 17% of all grains analyzed in the Cutler Group samples. It is common in all samples of the White Rim Sandstone and Castle Valley Sandstone but is rare in the Cutler fluvial facies ($n = 3$). Grains of this age are typically attributed to ultimate derivation from the Grenville orogen of eastern Laurentia (e.g., Dickinson and Gehrels, 2009).

Population F (~734–499 Ma)

Late Neoproterozoic and Cambrian grains constitute 8% ($n = 57$) of grains analyzed. This age range is common in all eolian samples but absent in the Cutler fluvial facies. This age group corresponds in part to the age range of volcanic rocks associated with Iapetan rifting along the eastern and southern margins of Laurentia (~765–530 Ma; Thomas, 2011, 2014), as well as Pan-African basement domains in the Appalachian orogen and south of the Ouachita orogen (Mueller et al., 2014). Grains of this age are commonly associated with Grenville grains inferred to have an Appalachian source (Dickinson and Gehrels, 2003, 2009).
Population G (~491–283 Ma)

Paleozoic grains of this population, with a dominant peak at ca. 424 Ma, constitute 10% (n = 74) of all grains analyzed. They are present in subequal quantities in all eolian samples, but only one grain (ca. 406 Ma) is present in the Cutler fluvial samples. Grains of this age group are commonly attributed to sources in peri-Gondwana assemblages of the greater Appalachian orogen (Dickinson and Gehrels, 2009) including the Taconic orogen (~490–440 Ma), Acadian orogen (~420–350 Ma), and Alleghenian orogen (~330–270 Ma; e.g., Thomas, 2011). A minor source for grains of this population might be present in pre-Permian Paleozoic strata of the San Juan dome, although detrital-zircon data for these strata do not yet exist, and these ages are rare to absent in the Cutler fluvial-facies samples. There are no young grains near the depositional age of the Cutler Group in the sample set.

SANDSTONE PROVENANCE

Detrital-zircon ages, sandstone petrology, and paleocurrent data indicate different sources for Cutler fluvial strata and the eolian samples of the study; whereas the eolian White Rim and Castle Valley sandstones have compositional similarities (Figs. 11–13). Moreover, stratigraphic trends in zircon age populations indicate a lower percentage of locally derived grains in the upper eolianite unit and furthermore is similar to that of the undifferentiated Cutler Formation. The White Rim Sandstone contains a broad range of grain ages that resembles the distribution of grain ages in the upper eolianite unit and furthermore is similar to that of the correlative Diamond Creek Sandstone to the northwest (Figs. 1, 2, and 13). The undifferentiated Cutler Formation consists of first-cycle arkose with a bimodal zircon age population consisting of subpopulations C’ and D’ that can be directly attributed to basement rocks of the Uncompahgre uplift. Thus, petrography and detrital-zircon content indicate that the fluvial Cutler arkoses, at least in the upper part of the undifferentiated Cutler section, constitute an excellent example of locally sourced sediment. The detrital-zircon ages of the Cutler Formation have restricted ranges, ~1790–1689 Ma and ~1466–1406 Ma, subpopulations C’ and D’, respectively, which aggregate 98% of the two combined fluvial samples (Table 3). The other five grains (1349 Ma, 1195 Ma, 1075 Ma, 960 Ma, and 406 Ma) are common components of the eolian strata and were likely blown into the fluvial system. The dominant age ranges correspond to previously reported ages of metavolcanic, plutonic, and metasedimentary rocks in the Unaweep and Black canyons, as discussed above, and thus provide additional insight into the ages and potential abundance of unexposed Uncompahgre basement rocks. Although paleocurrent data in the Cutler Formation indicate that sediment dispersal paralleled the salt walls, there are low-elevation gaps in structural culminations of the diapirs, particularly the one between the Fisher Valley and Sinbad Valley diapirs (Fig. 4), which could have permitted rivers to cross major diapirc trends and deliver sediment to the Castle Valley area. Present data cannot preclude derivation from uplifted basement rocks farther to the southeast, for example in the vicinity of the San Juan dome, and detrital-zircon populations of local Cutler Formation strata in Colorado should provide a basis for retaining or rejecting an exclusive nearby Uncompahgre source.

The lower eolianite member of the Castle Valley Sandstone contains a significant component of the local detritus as indicated by its subarkosic composition and zircon population modes similar to those of the undifferentiated Cutler Formation; nevertheless, the locally derived zircons decrease in abundance upsection in the formation in concert with changing sandstone composition (Table 3). The sample from the base of the eolianite (12CV50) contains 7% and 19% grains in the age range ~1790–1689 Ma and ~1466–1406 Ma, subpopulations C’ and D’, respectively, a 71% decrease relative to the sum of those subpopulations in the fluvial strata (Table 3). In contrast, grain populations scarce or absent in the fluvial samples dominate the basal eolian.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Stratigraphic unit</th>
<th>QtFL (%Qt)</th>
<th>QtFL (%F)</th>
<th>QmPK (%P)</th>
<th>QmPK (%K)</th>
<th>A and B (%)</th>
<th>C and D (%)</th>
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<td>06UT01</td>
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<td>ND</td>
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<td>8</td>
<td>2</td>
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<td>56</td>
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<td>4</td>
<td>13</td>
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<tr>
<td>10CWW</td>
<td>Upper eolianite</td>
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<td>14</td>
<td>32</td>
<td>0 (97)</td>
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<tr>
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<td>61</td>
<td>37</td>
<td>14</td>
<td>21</td>
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Note: Population age ranges: A: 3539–2548 Ma; B: 2546–2008 Ma; C: 1991–1599 Ma; C’: 1790–1689 Ma; D: 1581–1302 Ma; D’: 1466–1406 Ma; E: 1288–900 Ma; F: 734–499 Ma; G: 491–283 Ma. Subpopulations C’ and D’ represent locally derived Uncompahgre grains. ND—not determined; undiff.—undifferentiated.
sample: younger populations E, F, and G constitute 47% of the sample, a 45% increase relative to the fluvial samples; Archean and Paleoproterozoic populations A and B constitute 5% of the zircons analyzed. In higher samples of the lower eolianite, populations E, F, and G similarly constitute 52% (sample 10CVR) and 41% (sample 10CVG) of zircons analyzed, and populations A and B constitute 6% and 8% of the same samples, respectively. These numbers indicate that at least 50% of the zircons in the lower eolianite member were delivered to the site by eolian transport. Compositional and zircon data indicate the White Rim Sandstone and the upper eolianite member contain comparable populations of quartz, feldspar, and non-local zircon populations A, B, E, F, and G, indicating that downwind drift of sediment in the greater White Rim erg could have supplied essentially all detrital components of the upper eolianite member.

**DISCUSSION**

The Castle Valley Sandstone represents the innermost part of a shoreline-attached erg that extended in a downwind direction at least 250 km from the southeastern flank of the Oquirrh basin to an exposed conspicuous, almost linear, edge along the Colorado River and a buried, intricate edge in the salt anticline province of the Paradox Basin (Fig. 14A), where small salt-withdrawal basins enhanced local accommodation (Fig. 4). Extensively documented paleocurrents from the exposed southeastern edge of the erg (Baars and Seager, 1970; Steele-Mallory, 1982; Huntoon and Chan, 1987) demonstrate that northwesterly winds blew sand toward its current pinch out; data presented here indicate that those winds transported sediment along the axes of the salt-withdrawal basins. Although the present erg margin that parallels the Colorado River has been interpreted as an erosional pinch out beneath the Moenkopi Formation (Dubiel et al., 1996, 2009; Huntoon and Chan, 1987) demonstrate that northwesterly winds blew sand toward its current pinch out; data presented here indicate that those winds transported sediment along the axes of the salt-withdrawal basins. Although the present erg margin that parallels the Colorado River has been interpreted as an erosional pinch out beneath the Moenkopi Formation (Dubiel et al., 1996, 2009; Huntoon and Chan, 1987), the current margin probably does not lie far from its original depositional pinch out for the following reasons: (1) The pinch-out margin trends perpendicular to dominant direction of Permian sand transport, supporting the inference of sediment depletion with transport distance (e.g., Chan, 1989); (2) the White Rim Sandstone thins on the order of 700 m across eastern Utah subparallel to the trend of the pinch out (Fig. 4), which is not likely due solely to pre-Triassic beveling; (3) dominant large-scale trough cross-beds and curved dune crests at our Shafer Trail sample locality (also noted by Dubiel et al., 1996) indicate deposition by barchan dunes and hence diminished sand supply directly upwind of the White Rim pinch out. The downwind erg margin appears to have advanced farther southeast in the foredeep of the Paradox Basin as a result of continued subsidence and the effect of salt withdrawal (Fig. 14A). Sand that occupied the salt-withdrawal basins was reworked extensively during its initial arrival by northeast winds that represent either the general Pangean zonal trade winds (Parrish and Peterson, 1988; Peterson, 1988) or local foehn winds that blew off the high-standing Uncompahgre uplift. The northeasterlies deflated sand from the salt-withdrawal basin northeast of the Castle Valley salt wall (Parriott basin of Banham and Mountney, 2013) and likely resulted in the significant component of Uncompahgre detritus in the lower eolianite member. With time, the northwest winds, more significant contributors to deposition of the upper member of the Castle Valley Sandstone, delivered large volumes of far-traveled sediment to the proximal Paradox Basin. Although no facies or transport data yet exist for the Diamond Creek Sandstone, it likely represents the updip edge of the erg near the marine margin of the Oquirrh basin (e.g., Blakey, 2009). A shoreline source for White Rim sediment is suggested by coarsening grain size westward toward the marine margin and interfingering of eolian and marine strata where they are exposed on the flank of the Circle Cliffs uplift (Fig. 3; Kamola and Chan, 1988), and corroborated by the presence of glauconite grains and a crinoid fragment in the White Rim Sandstone in Canyonlands National Park (Steele-Mallory, 1982). Thus, the primary source for erg sediment was the Kungurian shoreline of western Pangea, as suggested by numerous previous workers (Johansen, 1988; Kamola and Chan, 1988; Marzolf, 1988; Dubiel et al., 1996, 2009; Condon, 1997); nevertheless, the bimodal wind direction recorded in the upper eolianite member of the Castle Valley Sandstone evidently created large linear dunes whose remnant topography is expressed on the upper surface of the White Rim Sandstone (Baars and Seager, 1970; Huntoon and Chan, 1987).

The Namib erg of the West African coast (Fig. 14B) appears to be a reasonable analog for the White Rim erg. The modern African sand sea is affected by a bimodal wind regime consisting of zonal SSE trade winds that blow off the Atlantic Ocean and seasonal easterly orographic winds, referred to as the berg (Bristow et al., 2007), that descend from the Great Escarpment (Fig. 14B; Glennie, 1987). The central part of the Namib erg consists of prominent linear dunes with north-south crests that extend for over 100 km in some examples (Fig. 14B). These complex linear dunes (Bristow et al., 2007) are analogues for the linear features preserved on the surface of the White Rim Sandstone.

Fluvial deposits of the interdune member separate the upper and lower eolianite members of the Castle Valley Sandstone and approximately coincide with a shift to more mature sandstone compositions. Fluvial deposition is indicated by the presence of an irregular scoured base, pebble lags, and horizontal laminae interpreted as sheet-flood deposits. The source of the sediment was the topographically elevated salt wall and its flanking strata. Dolomite clasts were derived from the diapir itself, which was exposed during deposition of the eolian strata and superjacent Moenkopi beds (Fig. 9F; Lawton and Buck, 2008), whereas granite and quartz pebbles were derived from exposed conglomeratic Cutler beds upturned adjacent to the diapir (Fig. 9F). Fluvial deposition may have been triggered by a transient shift to a seasonally wet climate, as has been suggested for progradation of the Organ Rock Formation above the Cedar Mesa Sandstone (Stanesco et al., 2000; Mountney, 2006; Dubiel et al., 2009). Because of the great distances to the time-equivalent marine shoreline, ~250 km upwind and ~100 km directly normal to the shoreline, it is unlikely that bounding surfaces within the eolian succession were generated by water-table changes driven by glacial eustasy, as suggested for Cedar Mesa bounding surfaces (Mountney, 2006). Rather, water table fluctuations...
and shifts in the associated capillary fringe were likely caused by short-term climatic cyclicity in combination with high sediment-accumulation rates in the adjacent salt-withdrawal minibasin.

White intervals of Castle Valley Sandstone are similar petrographically to bleached facies of the White Rim Sandstone in parts of the greater White Rim erg that have been attributed to leaching of previously deposited hematite cement by hydrocarbons (Gorenc and Chan, 2015). The presence of degraded hydrocarbon in the white parts of the Castle Valley Sandstone (Fig. 7F) seems to corroborate similarity of process, suggesting that the persistent color change in the Castle Valley sandstone is also a result of hydrocarbon migration through the sandstone. Moreover, the sandstone composition and detrital-zircon content of the white parts of the sandstone contrast with that of the pink part, suggesting that hematite cementation may have been in part controlled by the presence or absence of detrital iron oxide grains in the arkosic sandstones. Why the color change conforms closely to lithostratigraphy in the tilted sandstone and why it is truncated beneath the Moenkopi Formation remain

Figure 14. (A) Reconstruction of White Rim erg during Kungurian (late Leonardian, ca. 273 Ma) time. Black arrows are average estimated wind directions from White Rim Sandstone foresets discussed in text. Red arrow is average foreset dip direction of Huntoon and Chan (1987). Violet arrows are wind directions estimated from dominant foreset dip in lower member of Castle Valley Sandstone. Dashed black arrow near shoreline is estimated direction of longshore drift driven by NNW winds. Thin arrows are sediment transport directions estimated from fluvial cross-bed data of the undifferentiated Cutler Formation (Buller, 2009; Venus et al., 2015) and Organ Rock Formation (arrow east of confluence of Green and Colorado rivers; Mountney and Jagger, 2004) and provenance data described in this paper. Thin black lines near confluence of Green and Colorado rivers are linear topographic features on upper surface of White Rim Sandstone (Baars and Seager, 1970), here interpreted as remnant topography created by large linear dunes. They appear short on the map because they are only exposed on the surface of the White Rim Sandstone and are buried to the northwest beneath the Moenkopi Formation. Shoreline position from 275 Ma map in Blakey (2009). Position of Diamond Creek Sandstone sample (DC) adjusted 30 km westward to accommodate eastward translation during Late Cretaceous shortening (Kwon and Mitra, 2004). Other locations: CR—Capitol Reef shoreline location (Kamola and Chan, 1988); ND—Nokai Dome, where 7 m of White Rim Sandstone is interpreted to overlie a separate erg deposit, the De Chelly Sandstone (Irwin, 1971). (B) Modern Namib sand sea (note north orientation), illustrating dominant south-southwesterly zonal trade wind direction (black arrows) and orographic berg wind direction (violet arrow). Wind directions from Glennie (1987); base map from GoogleEarth. Thin black lines are prominent linear dunes in central part of erg; thin black arrows indicate fluvial dispersal directions.
open questions, which nevertheless suggest that the hydrocarbons were present in the rocks in the Permian.

The diverse grain ages of the Castle Valley and White Rim Sandstone resemble population assemblages that have been attributed to sources in the Appalachian orogen of eastern Laurentia. In particular these assemblages include Grenville, Neoproterozoic, and early Paleozoic grain ages attributed to 1.1 Ga Grenville granites, pan-African crust (Suwannee terrane), and peri-Gondwanan terranes, respectively, in the Appalachian region and in the subsurface south of the Ouachita orogen (e.g., Viele and Thomas, 1989; Mueller et al., 2014), and present in other Permian eolian sandstones, including the Coconino Sandstone, of southwestern Laurentia (Dickinson and Gehrels, 2003; Gehrels et al., 2011). These grains were likely delivered by transcontinental fluvial systems to the Permian shoreline NW of the Ancestral Rocky Mountain province and transported by eolian processes into the Permian ergs of western Pangea (Johansen, 1988; Marzolf, 1988; Dickinson and Gehrels, 2003; Gehrels et al., 2011). This hypothesis is consistent with the prevailing data that indicate that the greater White Rim erg was deposited by winds that blew from the northwest, away from the shoreline.

Northwesterly winds, rather than zonal northeasterly trade winds (Parrish and Peterson, 1988; Peterson, 1988), thus dominated eolian sand transport in the White Rim erg. Given the long distance of transport from the Kungurian marine margin in north-central Utah, it seems likely that a seasonal, or monsoonal, low-pressure system over Gondwana was the driver of the northwesterly circulation pattern, as predicted by Parrish and Peterson (1988), rather than onshore breezes. Whereas a high-standing Uncompahgre uplift might have created fierce topographic winds in the proximal part of the Paradox Basin, it might have sheltered the more distal part of the basin from zonal trade winds that dominated sand transport in the approximately time-equivalent De Chelly and Coconino ergs (Peterson, 1988). In addition, the northwesterly winds appear to have resulted in large erg deposits on the windward side of the large peninsula that straddled the transcontinental arc (Fig. 1) and lesser sand accumulations on the leeward, southeastern side of the peninsula (e.g., Blakey, 2009).

Grains in the age range 1607–1492 Ma (n = 26 or 4% of all analyses), which form a subset of population D and fall within the postulated North American magmatic gap (~1.61–1.49 Ga; Van Schmus et al., 1993; Grove et al., 2008), are difficult to explain by appealing to Laurentian basement sources. Zircon grains in the age range ~1.61–1.49 Ga are present in all Kungurian eolian strata of this study in abundances ranging from 2% to 13% of each sample analyzed (Supplemental Table 1 [see footnote 1]) and 7% of the correlative Diamond Creek Sandstone of north-central Utah (Lawton et al., 2010). Grains in this age range could have been derived directly from continental sources on the western edge of Baltica, present in the early Paleozoic Caledonian suture between Baltica and Laurentia (e.g., Bingen and Soll, 2009) and the Pinware terrane of SE Labrador (Tucker and Gower, 1994; Wasteney et al., 1997), or from western accreted terranes that contain zircon grains and basement fragments possibly derived from the Caledonian orogen by tectonic escape and subsequent long-distance tectonic transport along one or the other margins of Laurentia (Wright and Wyld, 2006; Grove et al., 2008) and now occupying oceanic crustal domains west of the study area in northern California. Rocks in the age range 1.6–1.5 Ga are also present in the Amazonian craton (e.g., Cordani et al., 2009; Cardona et al., 2010), and have been posited as a possible source for grains in Permian sandstones of southwestern Laurentia (Soreghan and Soreghan, 2013).

General paleogeographic arguments against Permian sediment delivery from Cordilleran accreted terranes and from the suture between Gondwana and Laurentia were presented by Dickinson and Gehrels (2003), who favored transcontinental fluvial sediment delivery to the Laurentian marine margin coupled with longshore transport driven by zonal trade winds and deflation of coastal plain sediment into the Early Permian erg (Fig. 1). Those authors objected to sediment sources to the west on the basis of the intervening Pennsylvanian–Permian marine sedimentary basins on the western margin of Laurentia that would have blocked sediment transport from western sources, and sediment sources to the south were rejected on the basis of an assemblage of marine foreland basins that lay north of the Ouachita-Marathon suture belt in Texas and New Mexico. Whereas the eastern components of this foreland basin system ceased to subside in the Early Permian (Ingersoll et al., 1995; Dickinson and Lawton, 2003; Thomas, 2014), remnant topography of the suture zone was likely effective at blocking sediment delivery from the low-lying Amazonian shield.

Zircon grains in Permian eolian strata of the Paradox Basin with likely Pan-African affinity (~765–535 Ma) and sources in eastern and southeastern Laurentia provide additional leverage on ultimate sources of grains in the White Rim erg. Grains in this age range constitute 4%–11% of the six samples of eolian strata sampled in this study and 4% of grains in the Diamond Creek Sandstone (Lawton et al., 2010). In contrast, this age range is absent to rare (generally <1%) in accreted eugeoclinal strata of northern California and related oceanic terranes, which include both lower Paleozoic strata and plutonic rocks, to the north in British Columbia and southeastern Alaska (Gehrels et al., 1996; Grove et al., 2008). Pan-African grains could therefore not have been derived from the Cordilleran terranes on the basis of existing data; nevertheless, because Pan-African grains are likewise uncommon in Baltica-derived sandstones (e.g., Bingen and Soll, 2009), Permian eolian sandstones containing both Pan-African and Baltica-derived grains in any event require a combination of sediment-dispersal systems that transported sediment from different parts of Laurentia. These separate systems could have been: (1) transcontinental drainages with headwaters in the former Caledonian orogen and farther south along the Appalachian orogen and even the Ouachita orogen, to tap northern sources of Baltica and Pinwarian crust and southern sources containing crust of Pan-African affinity, respectively; or (2) transcontinental sources with headwaters in the Pan-African crustal sources of the Appalachian region and shorter sediment-transport systems with headwaters in the west, which somehow bypassed the western marine basins of western Laurentia. The former paleogeographic model is preferred here, consisting of westward-
and southwestward-directed transcontinental fluvial systems that transported sediment from varied northeastern and eastern sources in the Caledonian and Appalachian orogens to a marine mixing zone along the Permian shoreline of Utah, Idaho, and Montana, from which the sands were delivered to the Paradox Basin.

Salt tectonics within the salt anticline province provided critical influence on local eolian sediment transport and accumulation of eolian sand, as well as the facies distribution and composition of the sand, in the Paradox Basin. Rapid subsidence within salt-withdrawal minibasins in the northwestwestern part of the salt anticline province provided accommodation for eolian sand at the same time as topographically expressed salt walls created protected leeward sites for sand accumulation. Indeed, some of the thickest eolian sand accumulations lie in the Big Bend minibasin southwest of the Castle Valley salt wall (Fig. 4). The initial eolian sediment contained a significant component of first-cycle arkose derived directly from the Uncompahgre uplift and deposited on the leeward side of the salt wall as recorded by the lower eolianite member. With time, and deposition of the upper eolianite member, northwest winds evidently became more influential and provided a greater percentage, roughly 100%, of sediment derived from the Laurentian marine margin. At this time, eolian sediment accumulation greatly outpaced supply from local fluvial systems, which could have resulted from decreased uplift rate of the Uncompahgre source (e.g., Soreghan et al., 2009) or high rate of sediment supply to the White Rim erg, an alternative possibility suggested by the general absence of interfingering of fluvial and eolian facies noted by White Rim stratigraphers (e.g., Huntoon and Chan, 1987; Chan, 1989). Eolian sediment transport was impeded to farther southeastern minibasins in the salt anticline province by a topographic obstruction near the intersection of the plunging nose of the Castle Valley salt wall and the west-trending Cache Valley salt wall. Although reasons for the increased importance of northwest winds remain unclear, the corresponding compositional change in the White Rim Sandstone occurred just prior to deposition of the interdune member in the middle of the formation. It seems likely that the change in local depositional style, from eolian to fluvial deposition, followed by a return to eolian deposition with a stronger component of transport from the northwest, might signal an important change in the climate of western Pangea, perhaps due to increasing strength of the southern low-pressure system over Gondwana (e.g., Parrish and Peterson, 1988).

The great thickness of the Castle Valley Sandstone in the Big Bend minibasin (Fig. 4), the difference in dominant wind directions between lower and upper eolianite members, and the compositional difference between the members combine to suggest the possibility that the Castle Valley Sandstone may correlate with both the White Rim Sandstone and the De Chelly Sandstone of the Four Corners region (R.F. Dubiel, 2015, written commun.), as suggested in Figure 2. Formerly considered equivalent to the White Rim Sandstone (Baars, 1982), the De Chelly Sandstone was later interpreted to underlie the White Rim Sandstone in an exploration well at Nokai Dome on the San Juan River (Figs. 3 and 14; Irwin, 1971); accordingly subsequent correlations have generally depicted the De Chelly as somewhat older than the White Rim (Blakey, 1990, 1996; Dubiel et al., 2009). Like the lower eolianite member of the Castle Valley Sandstone, the De Chelly Sandstone has prominent paleocurrent indicators to the SW and SE (Stanesco, 1991; Peterson, 1988), and examination of a single thin section from the Laramide Defiance uplift on the Arizona–New Mexico state line (Pennsylvanian Zuni uplift of Fig.1) indicates that it contains as much as 20% potassium feldspar, including microcline, a compositional characteristic of the lower eolianite member. Future detrital-zircon analysis of the De Chelly Sandstone might provide improved insight into possible correlation with the lower eolianite member of the Castle Valley Sandstone.

CONCLUSIONS

Prominent exposures of Lower Permian (Kungurian and upper Leonardian) eolian sandstone on the flank of the Castle Valley salt wall were deposited at the top of the Cutler Group in the northeastern part of the Paradox Basin in a salt withdrawal sub-basin. Termed here the Castle Valley Sandstone, the eolian deposits are continuous in the subsurface with, and constitute the depositional edge of, an extensive erg recorded by the time-equivalent White Rim Sandstone. The greater White Rim erg extended 125–150 km east from the time-equivalent edge of the west Pangean seaway, the shoreline of which trended approximately north-south, and at least 250 km downwind of littoral sediment sources on the southeastern flank of the Oquirrh basin. The Castle Valley Sandstone, as much as 183 m thick in outcrop, consists of two informal eolianite members separated by a fluvial deposit termed the interdune member. The upper part of the lower member and the upper member are bleached by hydrocarbons that once occupied the eolian sandstone. The Castle Valley Sandstone overlies fluvial red beds of the undifferentiated Cutler Formation, which may be equivalent to the Organ Rock Formation of the southwestern part of the Paradox Basin.

Provenance of fluvial and eolian Cutler Group strata reflects a combination of local and distant erosional sources. Cutler fluvial strata are compositional arkoses with average composition Qt56F42L2 and a narrowly defined bimodal detrital-zircon content with modes at ca. 1724 Ma and ca. 1441 Ma. The arkose was derived entirely from Proterozoic basement of the nearby Uncompahgre uplift, located no more than 40 km from the depositional site. Castle Valley Sandstone compositions change stratigraphically toward increased maturity with height in the section. The unbleached part of the lower member is quartz-rich arkose (Qt75,F30,L5) with a zircon population that includes Archean, Grenville, Neoproterozoic, and Early Proterozoic grains, ages not present in known basement of the Uncompahgre uplift, in addition to grains equivalent in age to the locally derived Uncompahgre zircons. The lower member was transported in part by northeasterly winds that picked up sediment from the Cutler fluvial plain and deposited it on the lee (southwest) side of the salt wall. The upper member of the Castle Valley Sandstone, deposited following an episode of sheet-flood deposition recorded by the interdune member, has an average
composition $Q_{80}F_{10}L_{0}$, a detrital-zircon content that completely lacks locally derived grains but contains abundant Grenville, Neo-Proterozoic (Pan-African), and early Paleozoic grains, as well as some grains in the range $\sim 1.61-1.49$ Ga that are here attributed to an ultimate source in Baltica. The upper member was deposited by a combination of northeasterly and northwesterly winds, the latter similar to the predominant wind direction recorded in the greater White Rim erg. The detrital-zircon content of the White Rim Sandstone, sampled at two localities to the southwest, resembles that of the upper member of the Castle Valley Sandstone.

As suggested by previous authors, sediment was transported via transcontinental rivers to the western marine margin of Laurentian Pangea from the Appalachian region, which provided Grenville, Pan-African, and lower Paleozoic grains to the littoral zone. In addition, northern tributaries to the transcontinental drainage, or a separate fluvial trunk river lying to the north, drained remnant topography of the early Paleozoic Caledonian orogeny, or nearer sources in Labrador, to supply grains of Baltica affinity. Baltica-age grains did not derive from accreted terranes in the Cordillera due to the intervening marine sedimentary basins that lay between the western shoreline of Laurentia and the potential Cordilleran terranes, which moreover lack Neo-Proterozoic Pan-African grains. Abundant sediment was blown southeastward from littoral sources incrementally exposed during Kungurian sea-level drawdown. Rapid transport of voluminous eolian sediment overwhelmed sediment derived from local Uncompaghre sources and resulted in observed compositional changes in eolian relative to fluvial sediment.

A combination of high subsidence rates and topographic expression of salt walls, both induced by salt diapirism in the northeastern part of the Paradox Basin, permitted preservation of the rapidly introduced eolian sediment, which might otherwise have been deeply deformed to more distant sites, on the inboard flank of the White Rim erg. Consideration of basin-scale paleocurrent patterns, sandstone composition from traditional petrographic methods, and detrital-zircon content of the Early Permian depositional system thus demonstrates the importance of regional transport systems in creating large volumes of far-traveled sediment in a local basin setting.

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