ABSTRACT

Coral Pink Sand Dunes State Park, located in southwestern Kane County, Utah, contains a variety of geologic features including one of the largest areas of freely migrating dunes in the Colorado Plateau. The semiarid climate, strong prevailing southerly winds, sparse vegetation, and abundant supply of sand-sized sediment make this area susceptible to eolian processes.

Picturesque exposures of Jurassic rocks are present within the park. The stratigraphic sequence ranges from the Triassic-Jurassic Moenave Formation to the Middle Jurassic Carmel Formation. The most widespread bedrock unit exposed within the park is the Navajo Sandstone (Lower Jurassic). The Navajo Sandstone is also widely exposed across the Mocassin Terrace southwest of the park and is the most likely source for the sand that comprises the dune field. The “coral pink” color of the dune sand is the result of iron-oxide stains on the surface of the sand grains inherited from the source sandstones.

Migrating dunes, whose morphology is primarily a function of wind characteristics, include transverse ridges, barchanoid ridges, and a solitary star dune. Dunes influenced or impeded by topographic obstacles or vegetation include climbing dunes, echo dunes, parabolic dunes, vegetated linear dunes, and nebkhas.

We divide the dune field into major geomorphic units based on the dominant dune type. A largely stabilized (vegetated) sand sheet and partially stabilized, poorly organized dunes are present at the southern (upwind) end of the dune field. The active core of the dune field contains transverse ridges and barchanoid ridges. Barchanoid ridges at the northern (downwind) end of the active core grade into climbing dunes that ramp up the bedrock escarpment associated with the Sevier fault. The climbing dunes in turn grade into large parabolic dunes that dominate the downwind end of the dune field.

Coral Pink Sand Dunes lies within the structural transition zone between the Great Basin section of the Basin and Range province to the west, and the core of the Colorado Plateau to the east. The north-south-trending Sevier fault cuts through the length of the park. The fault trace is marked by a west-facing bedrock escarpment that divides the park into two topographic units (a forested plateau to the east and a relatively low-lying valley floor to the west) and acts as a major control over the accumulation of sand within the dune field.

Important events recorded in the geologic features of the park include the Triassic and Jurassic depositional history of the Glen Canyon Group, the Cretaceous to Cenozoic structural history of the Colorado Plateau, and the late Holocene history of the active dunes. Optically stimulated luminescence dates from the active core of the dune field indicate that Holocene eolian deposition began at least 4,000 years ago. Radiocarbon dating of organic materials from an exhumed soil surface suggests a period of landscape stability approximately 500-200 years ago, coincident with the Little Ice Age. Dendrochronologic data from the ponderosa pines in the park, along with historic photographs, indicate the dune field has experienced alternating wet periods and drought since the end of the Little Ice Age, which have influenced vegetation coverage and dune activity in the area.

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INTRODUCTION

Geographic Setting

Sand dunes occur in two distinct habitats; along the coasts of seas and rivers, on the one hand, and on the barren waterless floors of deserts, on the other....Here [in deserts] instead of finding chaos and disorder, the observer never fails to be amazed at a simplicity of form, an exactitude of repetition and geometric order unknown in nature on a scale larger than that of crystalline structure.

— Ralph A. Bagnold, 1942

The many geologic wonders of southern Utah include a picturesque area of active dunes in southwestern Kane County, located about 27 miles (43 km) northwest of Kanab near the Utah-Arizona border (figure 1). This is one of the largest areas (approximately 5.4 square miles or 14 km²) of freely migrating dunes within the Colorado Plateau Province. This dune field has become known as the Coral Pink Sand Dunes. The term “coral” in this case refers to the deep orange pink color of the sand and has nothing to do with the marine reef-building organisms of the same name. Using a soil-color chart, geologists or soil scientists would formally describe the sand as reddish yellow (5YR 6/8); it is probably a good thing that a geologist did not name these dunes. A wide variety of eolian (an adjective referring to wind activity, derived from the name of the Greek god of the wind, Aeolus; also commonly spelled as aeolian) features and landforms are present within the Coral Pink Sand Dunes, making this an excellent field laboratory in which to learn about the role of wind in shaping the surface of the Earth.

The northern half of the dune field is federal land administered by the Bureau of Land Management (BLM). The southern half of the dune field and adjacent land (3,730 acres or 15 km²) comprises Coral Pink Sand Dunes State Park (CPSDSP), the subject of this paper. The Yellowjacket Canyon 7.5-minute topographic quadrangle covers the park area.

The park was established in 1963 and initially visited primarily by small numbers of off-highway vehicle (OHV) enthusiasts. Over the years visitation has greatly increased and today the majority of park visitors arrive without OHVs to enjoy the scenic beauty of this distinctive landscape. Park facilities include a 22-unit campground, modern restrooms with hot showers, a boardwalk overlook, and a one-half-mile nature trail. In addition, the park serves as an excellent base for exploring other areas of geologic interest in southern Utah, including Zion and Bryce Canyon National Parks, Cedar Breaks and Grand Staircase–Escalante National Monuments, and Kodachrome Basin State Park.

Sand Dune Road, a paved road that runs south from U.S. Highway 89 through Yellowjacket Canyon, and Hancock Road, a paved road that runs southwest from U.S. Highway 89, provide access to the park (figure 1). Together, Sand Dune Road and Hancock Road comprise the state-designated Ponderosa/Coral Pink Sand Dunes Scenic Backway.

Figure 1. Location of Coral Pink Sand Dunes State Park (CPSDSP) in Kane County, Utah.
Topography

Topographically, the park can be divided into two distinct units. The eastern half is a forested upland, part of the Moquith Mountains, bounded by a west-facing escarpment. This escarpment forms the eastern redrock backdrop to the dunes, visible from the dune field and campground to the west (figure 2). This escarpment is fault-controlled (see Structure section) and increases in height from north to south, reaching a maximum of approximately 700 feet (213 m) near the southern end of the park. Although given the name “mountains,” the Moquiths are actually a deeply incised plateau that is tilted toward the northeast. The highest elevation within the park, 6,977 feet (2,126 m), lies on this plateau. The fault-controlled escarpment exposes the same rock units that form the Vermillion Cliffs of the Grand Staircase, which lie generally east and south of the park. For a complete description of the rock units and features that comprise the Grand Staircase of southwestern Utah (see Grand Staircase–Escalante National Monument article in this volume).

The western half of the park is relatively low terrain lying between the Moquith Mountains and the mesa just northwest of the park. Though unnamed on modern topographic maps, this mesa was called Esplin Point by Gregory (1950) in his early regional study, a name we will use in this paper. Esplin Point lies at the eastern end of a line of cliffs and mesas that extend from Elephant Butte to the park, which collectively form the western expression of the White Cliffs of the Grand Staircase. Esplin Point has a maximum elevation of approximately 6,850 feet (2,090 m).

Most of the dune field within the park boundaries lies on this lowland at the base of the scarp bounding the Moquith Mountains. Near the northern end of the park the dunes climb the escarpment and continue to the northeast, forming part of the dune field on land administered by the BLM.

The park area is drained by a number of ephemeral streams whose channels extend from the upland areas, on either side of the dune field, to the margin of the dune field itself. Channels are largely absent within the dunefield, owing to the high infiltration rates within the dune sand. However, the steep ephemeral channels which drain the eastern escarpment act as tributaries to Sand Canyon Wash, lying at the base of the escarpment and along the eastern margin of the dune field. Sand Canyon Wash flows southward to Kanab Creek and hence to the Colorado River. The lowest elevation within the park, approximately 5,620 feet (1,710 m), is in the channel of Sand Canyon Wash where it crosses the park’s southern boundary. Thus, the total topographic relief within the park is approximately 1,357 feet (416 m). The drainage divide between Sand Canyon Wash and Yellowjacket Canyon to the north is located near the intersection of Hancock and Sand Dune roads, approximately 0.9 miles (1.4 km) north of the park. North of this divide, Sand Dune Road descends Yellowjacket Canyon, a north-flowing tributary of the East Fork of the Virgin River.

Climate and Vegetation

Eolian processes are important in areas with strong prevailing winds, sparse vegetation, and an abundant supply of sand-sized sediment. This is why most of the world’s major dune fields are found in association with arid or semiarid climates. The Coral Pink Sand Dunes are no exception, being located within the extensive semiarid climatic zone of Utah known as steppe (Köppen climatic classification: BS). In Utah, the steppe climate is transitional between the true arid/desert climates of the Great Basin and the moister climates of the state’s mountainous regions (Murphy, 1981). The mean annual temperature in Kanab, Utah, is 54.4°F (12.4°C); the mean annual precipitation is 13.3 inches (33.8 cm) (Utah Climate Center, as reported in Pope and Brough, 1996). The distribution of precipitation in this area, as evidenced by the data from Kanab, is distinctly bimodal with winter-summer precipitation peaks and spring-autumn drought. The closest weather station to the park with long-term wind data is Las Vegas, Nevada, located approximately 140 miles (225 km) to the southwest. The mean annual surface wind at this location blows from the southwest at 9.3 mph (15.0 kph). It is interesting to note that in Las Vegas, the two months of May and June have the greatest average wind speed, which coincides with the two driest months of the year in the Kanab area. If this relationship is extrapolated to the Coral Pink Sand Dunes, one would hypothesize that eolian activity would generally peak during this time.

Outside of the distinctive habitats of the dune field itself (see Castle, 1954), the dominant vegetation of the western half of the park is typical of pinyon-juniper woodlands (Pinus edulis, Juniperus spp.) and sagebrush scrub (Artemisia spp.) of the Colorado Plateau—with one notable and scenic exception. Impressive stands of pon-
derosa pine (*Pinus ponderosa*) are present within the more stabilized areas of the dune field. Ponderosa pines also dominate the forested upland of the eastern side of the park.

**Previous Investigations**

Geologic aspects of Coral Pink Sand Dunes have been previously reported in more broad-based investigations, most notably Gregory (1950), Doelling and others (1989), Anderson and Christenson (1989), and Davis (1999). Castle (1954) investigated the soil conditions and vegetation communities within the dune field. We are greatly indebted to these researchers, as much of this paper is a synthesis of their work. However, this is the first work to focus specifically on the geology of Coral Pink Sand Dunes. One of our major contributions to the geologic knowledge of this area lies in the complete characterization of the dune field, including its subdivision into distinct geomorphic units that reflect the relationship between surface forms and dominant surface processes. In addition to our literature review, we conducted an analysis of color aerial photography (scale 1:4,000), augmented by reconnaissance-level field research during the spring and fall of 1999.

Subsequent research by the authors has utilized ground-penetrating radar (GPR) to characterize the internal structure of the dunes (Wilkins and others, 2002; 2003). Wilkins and others (2005) also developed a chronology of Holocene dune activity based on radiocarbon and optically stimulated luminescence (OSL) dating. Wilkins and Ford (2007) used repeat photography and spatial statistics to document recent changes in the spatial organization of the dune field.

**STRATIGRAPHY AND GEOMORPHOLOGY**

The generalized geologic map of the park area (figure 3) and diagrammatic cross section (figure 4) provide an idea of the geology. The general geology of the park can be characterized as an area of cliff-and-bench topography, developed on Mesozoic sedimentary rocks, with a Quaternary dune field at the surface. The oldest Mesozoic unit exposed at CPSDSP is the Moenave Formation (Triassic-Jurassic). If one were to drill below the present-day dune field, the Chine (Upper Triassic) and Moenkopi (Lower Triassic) Formations would be successively encountered below the Moenave Formation (figure 4). The youngest bedrock unit exposed within the park boundary is the lowest member of the Carmel Formation (Middle Jurassic); the Co-op Creek Limestone Member of Doelling and others (1989). The various unconsolidated eolian deposits and active dunes, for which the park is named, are the youngest of all the geologic units in the area.

**Mesozoic Rocks**

The picturesque Moenave/Wingate, Kayenta, and Navajo Formations make up the Glen Canyon Group. These units have historically been regarded to be Triassic and Jurassic in age; however, subsequent stratigraphic work has shown that the Glen Canyon Group is entirely Early Jurassic in age (Pipiringos and O’Sullivan, 1978; Imlay, 1980; and Doelling and others; 1989). The general stratigraphy and depositional environments of the various Jurassic-age bedrock formations exposed in the park area are discussed below. A generalized stratigraphic column is presented in figure 5.

**Moenave Formation (Jmo)**

In western Kane County, the Moenave Formation (Triassic-Jurassic) overlies the Chinle Formation and the boundary between them was thought to represent a period of erosion called the J-0 unconformity (Pipiringos and O’Sullivan, 1978), but new evidence suggests the contact is conformable and the "J-0" unconformity is within lower Moenave strata (Lucas and others, 2005). The Moenave Formation is now divided into three members (in ascending order, the Dinosaur Canyon and Whitmore Point), the Springdale Sandstone was formerly the upper member but is now reassigned to the basal Kayenta Formation (Lucas and Tanner, 2006). Only the upper part of the Moenave Formation is exposed in fault blocks of the Sevier fault zone south of the park (figure 3). Thin bedded sandstone and siltstone with bedding parallel lamina and low-angle cross-beds in the Dinosaur Canyon Member and siltstone, fine-grained sandstone, thin limestone, and mudstone with freshwater-fish fossils in the Whitmore Point Member suggest a terrestrial depositional environment for the Moenave Formation. Specific environments probably included lakes, mudflats, and fluvial channels as part of a broad floodplain (Doelling and others, 1989).

**Kayenta Formation (Jk)**

The Kayenta Formation (Lower Jurassic) unconformably overlies the Moenave Formation (figure 5). The Kayenta Formation is exposed within the park in the Sevier fault zone along the base of the escarpment that forms the western edge of the Moquith Mountains (figure 3). The Springdale Sandstone Member is also exposed in fault slices along the base of the Moquith Mountains. The unit, 160 to 185 feet (49–56 m) thick, consists of very fine grained, pale-reddish-brown sandstone with subordinate amounts of cliff-forming conglomerate (Doelling and others, 1989). The Kayenta Formation above the Springdale consists primarily of deep red siltstones and mudstones. Smaller amounts of lenticular, medium-grained, trough-cross-bedded sandstone are also present. In Kane County, the thickness of the Kayenta Formation (not including the Springdale Sandstone Member) varies from 190 to 340 feet (58–104 m) (Doelling and others, 1989).

Although the Kayenta Formation is mostly fluvial in origin (formed by shifting braided streams), some eolian sandstone and lacustrine limestone beds, deposited in interfluve areas, are locally present (Doelling and others,
Figure 3. Geologic map of the Coral Pink Sand Dunes area in Kane County, Utah (after Sargent and Philpott, 1987). The map legend is included with the cross section shown in figure 4 and identifies the geologic units and symbols used. Note the trace of the Sevier fault running north-south the length of the park.
These streams originated in highlands to the east, probably the Uncompahgre uplift (Barnes, 1993).

A thin, though mappable, tongue of sandstone, known as the Tenney Canyon Tongue, extends eastward from the main body of the Kayenta Formation into the lower portion of the overlying Navajo Sandstone (figure 5). The portion of the Navajo Sandstone lying between the main body of the Kayenta Formation and the Tenney Canyon Tongue is called the Lamb Point Tongue of the Navajo Sandstone (Doelling and others, 1989). These two bedrock units are also exposed within the park along the trace of the Sevier fault (figure 3).

**Lamb Point Tongue of the Navajo Sandstone (Jnl)**

The Lamb Point Tongue (Lower Jurassic) is named for a location on the Vermilion Cliffs south of the park. The unit consists of white, tan, or gray fine-grained sandstone with thick sets of eolian cross-beds, locally greater than 25 feet (7.6 m). The lower and upper contacts with the Kayenta Formation are sharp. This unit is 90 to 410 feet (27–125 m) thick and pinches out to the west of the park. The Lamb Point Tongue contains distinctive contorted cross-beds in the upper 10 to 15 feet (3–4.6 m) of the unit. These contorted beds have been interpreted to be the result of slumping on the steep downwind side of dunes or possibly the result of earthquake-induced liquefaction (Doelling and others, 1989). The Lamb Point Tongue unit is an intertongue of the Navajo Sandstone resulting from widespread eolian deposition in Early Jurassic time.

**Tenney Canyon Tongue of the Kayenta Formation (Jkt)**

The Tenney Canyon Tongue (Lower Jurassic) of the Kayenta Formation is pale-reddish brown in color and comprised of siltstone, mudstone, and very fine thin-beded to laminated sandstone, representing floodplain and channel deposition. It is 90 to 170 feet (27–52 m) thick and typically displays low-angle cross-bedding. This unit has sharp contacts with both the underlying Lamb Point Tongue and the overlying main body of the Navajo Sandstone (Doelling and others, 1989). The tongue thins from west to east and pinches out east of Kanab, Utah.

**Navajo Sandstone (Jn)**

The Navajo Sandstone (Lower Jurassic) stands out as
the premier formation of the remarkable scenery of the Colorado Plateau. In the vicinity of CPSDSP, the Navajo Sandstone is a thick unit, 1,800 to 2,300 feet (549–701 m), of well-sorted, fine-grained quartz sandstones of varying colors and degrees of cementation. In addition to the cross-bedded sandstone, the Navajo contains a few lenticular beds of dense limestone or dolostone. These carbonate rocks are thought to have accumulated in interdune lakes or playas. Being more resistant than the sandstones, the limestones and dolostones commonly form benches or shelves (Doelling and others, 1989).

The Navajo Sandstone is world-renowned for its elaborate array of high-angle cross-beds and stark erosional forms, including cliffs, domes, and monuments. The light colors of the Navajo have been described Various as white, tan, buff, salmon, pink, vermillion, brown, red, yellow, cream, orange and gray. The sandstone is weakly cemented by calcite or dolomite and varying amounts of iron oxides. Hematite and goethite cements, produce the red, orange, and yellow colors; ferrous-iron minerals contribute to the brown and occasional green colors (Doelling and others, 1989). The diagenetic coloration of the Navajo Sandstone and related units tells an interesting story of iron cycling, documented in the recent studies of Chan and others (2000), Chan and Perry (2002), and Beitler and others (2003 and 2005).

The Navajo Sandstone is the most extensive bedrock unit within the park (figure 3). The lower and upper parts of the formation are more strongly cemented than the middle third and thus are cliff-forming units. The lower cliff-forming unit is typically reddish in color and, along with the underlying Kayenta Formation, form the Vermillion Cliffs of southwestern Utah. The upper cliff-forming sandstones typically lack the reddish hues characteristic of lower section and thus form the White Cliffs of the Grand Staircase. The lower reddish cliff-forming portion of the

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**Figure 5. Stratigraphic column of the Mesozoic bedrock units present within Coral Pink Sand Dunes State Park (modified from Doelling and others, 1989).**

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Navajo is exposed in the bedrock escarpment east of the active dunefield, whereas the upper white cliff-forming portion of the Navajo is exposed in the face of Esplin Point (figure 6), the prominent mesa located due north of the park’s campground. In active gullies and washes along the margin of the Coral Pink Sand Dunes, the Navajo Sandstone—presumably the middle to upper third of the formation—directly underlies the modern dunes.

Deposition of the Navajo Sandstone occurred in a large coastal to inland desert, possibly similar to the modern Sahara. Orientation of the cross-bedding, essentially the lithified slip faces of the Jurassic dunes, indicates that the Jurassic winds blew primarily from the north and northwest (Stokes, 1986; Parrish and Peterson, 1988; Peterson, 1988). The mineralogy of the Navajo Sandstone, 90 to 98 percent clear quartz with minor amounts of feldspar, magnetite, tourmaline, staurolite, zircon, garnet, and mica, suggests the source area of the sand was dominated by metamorphic rocks (Doelling and others, 1989). Recent research on the detrital zircons found within the Navajo Sandstone suggests that the sediments were eroded from the Appalachian Mountains, Canadian Shield, and the Ancestral Rocky Mountains (Dickinson and Gehrels, 2003). The sediments were probably transported westward by large rivers and deposited on a coastal plain and shallow-water shelf that lay up-wind of the Navajo erg. Geologists use the Arabic term “erg” to refer to large (≥ 48 square miles or 125 km²) desert areas dominated by wind-deposited sand and complex dune forms (smaller areas of dunes, such as the Coral Pink Sand Dunes, are defined as “dune fields” [Pye and Tsoar, 1990]).

Some early workers (e.g., Stanley and others, 1971) favored a marine origin for the Navajo Sandstone as large offshore sandbars cross-bedded by ocean currents. This controversy was essentially put to rest by the work of Hunter (1976, 1977), who established the criteria for distinguishing eolian and water-laid cross-stratification. The discovery of local occurrences of petrified trees within the sandstone (presumably spring-fed desert oases) and animal tracks (from vertebrates, including dinosaurs, and invertebrates) preserved within the interdune limestone added further evidence of terrestrial deposition (Barnes, 1993; Wilkens and Farmer, 2005; Parrish and Falcon-Long, 2007). Recently BLM geologists confirmed the existence of a substantial dinosaur trackway, known to OHV enthusiasts for some time, located on BLM land just southwest of Coral Pink Sand Dunes State Park (Mathews and others, 2010). The exposure in Navajo Sandstone, known as the Moccasin Mountain Tracksite, contains tracks from at least six different types of dinosaurs.

**Temple Cap Sandstone (Jtc)**

A major unconformity (J-1) marks the top of the Navajo Sandstone (figure 5). Above it is the Temple Cap Sandstone of Middle Jurassic age, named for the East and West Temple features of Zion National Park. This formation contains, at its base, a prominent reddish, slope-forming siltstone and sandy siltstone, that is 40 to 50 feet (12–15 m) thick (Sinawava Member). This member is overlain by up to 150 feet (46 m) of light-gray to tan, cross-bedded, cliff-forming sandstone (White Throne Member) (Doelling and others, 1989). The sandstones and siltstones of the lower Sinawava Member are marine to marginal marine (Sprinkel and others, 2009); the cross-bedded sandstones of the White Throne Member represent aeolian deposition. However, the two members are difficult to separate owing to the extensive interfingering between the two (Doelling and others, 1989).

The Temple Cap Sandstone is exposed in the central portion of the face of Esplin Point north of the campground (figure 6). This outcrop belt extends to the western edge of the park (figure 3).

**Co-op Creek Limestone Member of the Carmel Formation (Jc)**

The Carmel Formation (Middle Jurassic), named for Mt. Carmel, Utah, is a lithologically complex unit of limestones and sandstones that is subdivided into numerous members (Hintze, 1988; Doelling and others, 1989). The lowermost Co-op Creek Limestone Member caps Esplin Point (figure 6). Co-op Creek beds also occur just west of the road near the northern boundary of the park. The Co-op Creek Member consists of a basal siltstone, 10 to 20 feet (3–6 m) thick; however, this unit is similar to the Sinawava Member of the Temple Cap and is considered the upper part of the Temple Cap (Sprinkel and others, 2009). The remaining Co-op Creek Members is more than 200 feet (61+ m) of thin- to medium-bedded light-gray limestone and shaly limestone (Doelling and others, 1989).

The Co-op Creek Limestone Member of the Carmel Formation conformably overlies the Temple Cap Sandstone (figure 5). The Carmel Formation in turn is locally truncated by the unconformity at the base of the Cretaceous System. Most importantly, the Carmel Formation...
marks the end of terrestrial deposition and the return of regional shallow marine conditions to this part of Utah during Middle Jurassic time. Subsequent geologic events uplifted the sediments of the Carmel Formation from below sea level and raised them to their present elevation of over 6,600 feet (2,011 m) in the vicinity of the park.

Quaternary Deposits and Geomorphology of the Dune Field

Although the local Jurassic bedrock units are impressive, the foremost geologic features of this park are the various unconsolidated eolian deposits and landforms that make up the dune field. Whereas the Jurassic formations provide information about ancient depositional environments, the modern dunes are the result of recent and on-going geological processes during the Quaternary Period (the last 2.6 million years of Earth’s history).

For a dune to form, a small patch of sand must first accumulate where the wind speed has been reduced, generally by an increase in surface roughness. Once formed the patch of sand grows by trapping bouncing (saltating) grains that are unable to rebound upon impact as easily as those that impact harder non-sandy surfaces (an example of a positive feedback). As a result of the separation and deceleration of the airflow on the lee (downwind) side of the growing sand body, sand accumulates more rapidly than it is removed by eolian erosion; thus the dune increases in size (Summerfield, 1991). A mature dune is thus shaped by the airflow over it and in turn modifies the airflow. As a result, many dunes attain a characteristic equilibrium profile of three components: (1) facing upwind is a gently inclined (typically 10–15 degrees) backslope, (2) facing downwind is a steep slipface, standing at the angle of repose for sand (generally 30–34 degrees), and (3) separating the two slopes, a dune crest (Ritter and others, 1995).

This asymmetric equilibrium profile can be seen in a variety of dune types with differing plan forms (figure 7).

Wind direction and velocity, sand supply, and the presence of topographic obstacles or stabilizing vegetation are the most important factors influencing dune morphology or type. Considering these factors, it is useful to classify the various dune types as either free dunes, whose morphology is primarily a function of wind direction and sediment supply, or impeded dunes whose form is greatly influenced by the effects of vegetation, topographic obstacles, or highly localized sediment sources (Summerfield, 1991). Many of the classic dune types are present within Coral Pink Sand Dunes State Park (table 1).

Sargent and Philpott (1987) divide the dune field into two units: (1) a larger core of active sand dunes and (2) a smaller area of inactive eolian sand along the southern margin of the dune field. Doelling and others (1989) map the dune field as a single unit. Neither report identifies the various types of dunes present in the area. Based on field observations and an analysis of aerial photography, we subdivide the active core of the dune field into five (5) map units that reflect the dominant dune type present (figure 8). In addition, we retain the stabilized-sand unit (Qes) of Sargent and Philpott (1987). These units are described below in order of their occurrence, starting at the south end of the dune field.

Stabilized Eolian Sand Sheet (Qes)

In the southwestern corner of the park, and extending beyond the park boundaries, is an area of wind-blown sand largely stabilized by sagebrush scrub (figure 8). Although slipfaces are not preserved on the dune forms, this unit does retain some original depositional relief in the form of rolling hills of sand. The western and southern boundary of this unit is gradational with the mixed eolian

Figure 7. Morphology of the major types of dunes. Arrows indicate the formative prevailing wind direction (modified from McKee, 1979).
and alluvial sediments of the valley floor (Qae), which support a pinyon-juniper woodland. The interior boundary with the partially stabilized transverse dunes (Qeps) is sharper, being marked by a distinct change from sagebrush scrub to a more sparsely vegetated area. This low-relief accumulation of eolian sand is best termed a sand sheet. Sandsheets develop under conditions unfavorable to dune formation, including a high water table, periodic surface flooding, and a vegetative cover (Lancaster, 1995). Each of these conditions, which act to limit the amount of sand available for dune formation, may be significant in this area, especially the cover of vegetation.

**Partially Stabilized Dunes (Qeps)**

Located at the southern (upwind) end of the dune field and interior of the stabilized sand sheet is an area of partially stabilized dunes (figure 8). Dune morphology within this map unit is poorly developed and somewhat chaotic, but includes small transverse and parabolic-like forms (discussed in detail in following sections). Many of the low rolling dunes in this area lack obvious slipfaces. Topographically, this map unit is transitional, in terms of dune activity, relief and vegetative cover, between the stabilized sand sheet (Qes) to the south and the core of active dunes (Qetd) to the north.

The most distinctive characteristic of this geomorphic unit is the presence of scattered ponderosa pines (*Pinus ponderosa*) amongst the dunes (figure 9). Ponderosa-pine forests are best developed on the Colorado Plateau on mesa tops and mountain slopes at elevations between 6,900 and 8,200 feet (2,100–2,500 m) (Betancourt, 1990). Their presence at this relatively low elevation, approximately 5,680 to 5,800 feet (1,731–1,768 m), on a basin floor otherwise dominated by a pinyon-juniper woodland, is curious. Ponderosa pines can tolerate a variety of soil conditions, but trees on thin, dry soils are usually dwarfed (Harlow and Harrar, 1968). These trees are not dwarfed and thus indicate a moisture regime more typical of cooler habitats at higher elevations. We speculate that the southward slope of the dune field, combined with the high hydraulic conductivity of dune sand, produces a localized southward flow of shallow groundwater. The water table and ground surface may converge as one moves south to a point that is within the rooting depth of ponderosa pines. Four-year-old trees may have taproots 4 to 5 feet (1.2–1.5 m) long (Harlow and Harrar, 1968).

The ponderosa pines in Qeps are found in various situations, ranging from being rooted in shallow (less than one meter deep over bedrock) cover sheet sand to having partially buried trunks (up to an estimated 4 meters) in the dunes. The duff from the trees (e.g., needles, branches, and cones) appears to add stability to the dune and swale surfaces under and around the trees, and as a result very little surface dune activity is observed in the stands.

In the summers of 2002 and 2003, 15 ponderosa in this section of CPSDSP were cored. Using standard dendrochronological methods, the cores were processed, dated, and the rings visually cross-dated (Stokes and Smi-
The range of variability in the trees’ environments, that is, either partially buried or rooted in a swale, results in different growth-ring responses to the same annual rainfall variability. Rings in cores collected from trees partially buried in the dunes show wider rings than trees in the swales show in the same years; it is thought that the dunes allow greater retention of soil moisture during drought or dry years, and also enhance the growth in wetter years. These differences, though, prevent statistical cross-dating so these cores are best used to provide minimum ages of tree and stand establishment (Wilkins and others, 2005).

Because sampling was stratified towards the oldest trees, the majority of those cored are greater than 150 years old. Minimum dates of establishment range from the oldest at 1562 CE to the youngest tree in the last decade of the 20th Century, indicating the stand may have first become established around the beginning of the Little Ice Age, sometime around the early 15th century (Bradley and Jones, 1993) but has continued to survive through the present.

The ages of the trees in CPSDSP were compared to a published ponderosa tree-ring chronology from Navajo Mountain (Dean and Bowden, 1971), 150 miles to the east at roughly the same elevation. Establishment dates of the CPSDSP trees correspond broadly with periods of wide ring widths in the Navajo Mountain trees, indicating they became established during periods of more effective moisture. A similar comparison against the reconstructed Palmer Drought Severity Index for the dunes (Cook and others, 2004) also supports tree-establishment dates corresponding with higher levels of moisture, such as would have existed during the Little Ice Age.

A model of stand establishment and dune activity in this part of the dune field can be reconstructed using these observations. Assuming trees need moist conditions and a stable surface to germinate, ages for the oldest trees of 440 years suggests that activity in this portion of the dune field has been climatically limited, at least intermittently, since the beginning of the Little Ice Age. Those same climatic conditions would result in sediment-limited conditions, as

![Figure 8. Geomorphic map of the Coral Pink Sand Dunes. Trace of the Sevier fault from Sargent and Philpott (1987). See text for detailed descriptions of the Quaternary-age map units.](image-url)
sand does not travel well when conditions are damp (Kocurek and Lancaster, 1999). Additional limitations on the sand available for transport are imposed by the needle duff observed on the surfaces of the dunes. Finally, as trees have increased in height, the resulting feedback is increased friction and reduced wind velocity at the surface, further limiting eolian transport. Thus, dune activity in this portion of the dune field may have been waning since the establishment of the first trees over 450 years ago.

**Active Core of Transverse Dunes (Qetd)**

The most active portion of the dune field is characterized by the presence of large, well-developed transverse dunes, amid very sparse vegetation. These are the signature dunes of the park, and this area is the most popular with the OHV recreationists. This portion of the dune field lies on the valley floor immediately below the west-facing escarpment of the Moquith Mountains (figure 8).

Transverse dunes, generally associated with unidirectional winds, are asymmetric ridges of sand with a single slipface orientation. The ridge crests are oriented perpendicular or transverse to the prevailing winds (Summerfield, 1991). The crest of the transverse dunes at CPSDSP have a prominent northwest-southeast orientation, indicating a prevailing southwesterly wind, and rise 30 to 50 feet (9–15 m) above their adjacent interdune swales.

Two of the three major types of transverse dunes are present in the Coral Pink Sand Dunes (isolated barchans are not present here). In the southern portion of the map unit, the dunes have relatively straight crests and are thus classified as transverse ridges (figure 7). The crest-to-crest spacing of the ridges is typically 600 to 700 feet (183–213 m). Simple transverse ridges are relatively rare. Apparently, minor surface irregularities commonly create corkscrew-like vortices in the airflow over a transverse ridge. These vortices distort the ridge into a sinuous form called a barchanoid ridge (Summerfield, 1991). The transverse ridges at CPSDSP grade downwind (to the north) into classic barchanoid ridges (figure 10), with an increase in the dune spacing (up to 1,000 feet or 305 m). These barchanoid ridges display the characteristic alternating linguoid (protruding) and barchanoid (recessed) elements (figure 7). Owing to the complex airflow over these sinuous ridges, the barchanoid element of one ridge is commonly aligned with a linguoid element in the adjacent ridge, creating a complex crestal network referred to as aklé or fishscale patterns (Pye and Tsoar, 1990). A weak aklé pattern is present at the northernmost end of this map unit, where the barchanoid ridges impinge upon the bedrock escarpment to the east.

The convergence between the barchanoid ridges and the bedrock escarpment has also produced several echo dunes (figure 11). Echo and climbing dunes (discussed below) are examples of topographically controlled dunes, dunes that owe their existence and morphology to interactions between sand-transporting winds and topographic obstacles (Lancaster, 1995). Echo dunes commonly form in front of cliffs (slope > 50 degrees). A vortex forms immediately upwind of the cliff that acts to “sweep out” a corridor between the cliff and the slipface of the dune (Cooke and others, 1993). The result is the slipface of the dune tends to parallel, or echo, the shape of the cliff face. Echo dunes are best developed where the prevailing wind is near perpendicular to the trend of the cliff or escarpment. Since the sand-transporting winds within the park strike the bedrock escarpment obliquely, only the eastern end of the barchanoid ridges display this echo morphology.

**Small Marginal Parabolic Dunes (Qemp)**

Along the western margin of the dune field, approximately a quarter mile south of the “cattle guard” area (see Classic Geological Sites section) is an area dominated by small parabolic dunes (figure 8). Parabolic dunes are U- or V-shaped in plan with two trailing arms that point upwind (figure 7). Typically, there is a large mound of sand with a
steep slipface at the downwind end of the dune. The outside slopes of the trailing arms are almost always vegetated (Pye and Tsoar, 1990). Parabolic dunes are common in sand accumulations stabilized by vegetation. Localized disturbance of the vegetative cover and subsequent eolian erosion can give rise to circular or elliptical depressions called blowouts. The eroded sand is then deposited downwind, near the edge of the disturbed area, as a parabolic dune (Summerfield, 1991).

The orientation of these parabolic dunes suggest they are controlled by the same wind regime responsible for the large area of transverse dunes. However, these are much smaller features, generally less than 12 feet (3.7 m) in height.

Climbing Dunes (Qecd)

Wind encountering the windward face of a topographic obstacle, such as a hill or scarp, with a slope of 30 to 50 degrees will be decelerated such that a dune is deposited upwind of the obstacle. In this situation, unlike an echo dune, the fixed eddy or vortex is small or absent. Thus, the dune can bank up against the scarp as a sandy ramp, creating a so-called climbing dune. When the ramp attains an equilibrium profile sand can then be carried up and over it (Cooke and others, 1993). This is the process that has taken place at the north end of the park where the active transverse dunes obliquely converge with the faultline scarp of the Sevier fault (figure 8). Figure 12 is a ground-penetrating-radar (GPR) transect from the area that clearly shows a bedrock scarp buried by a ramp of climbing dunes. The climbing dunes have a slope of 5 to 12 degrees and display well preserved internal slip faces. The original slope of the bedrock scarp is estimated to have been 45 to 50 degrees, based on the profile of small outcrops of Navajo Sandstone poking through the sand ramp. The area of climbing dunes is more heavily vegetated than the active core of transverse dunes. The dune morphology is a mixture of poorly developed transverse and parabolic forms.

Large Parabolic Dunes (Qelp)

Where the climbing dunes reach the top of the escarpment they grade into an extensive geomorphic unit dominated by parabolic dunes (figure 8). This geomorphic unit, essentially the northern half of the dune field, lies beyond the park on land administered by the BLM. Slipface orientation in this part of the dune field indicates a more westerly prevailing wind (S. 50˚W.) compared to the area of transverse dunes (S. 40˚W.), probably owing to deflection as the regional wind is funneled through the gap between the eastern escarpment and Esplin Point. The top of the escarpment appears to act as a flow-expansion point, which reduces wind velocity and promotes deposition.

These parabolic dunes are much larger than those along the margin of the dune field farther south. The complex pattern of coalesced and nested dunes in this area, typical of most occurrences of vegetated parabolic dunes, is suggestive of alternating periods of dune migration and stabilization (Cooke and others, 1993). Here the rows of coalesced dunes have an average crest-to-crest spacing of approximately 1,300 feet (396 m). Large circular to elliptical blowouts are present upwind of many of the dunes. A characteristic feature of this part of the dune field is the presence of small groves of ponderosa pines in the interdune swales. Unlike the ponderosa pines present at the extreme southern end of the dune field, these occur at more typical elevations of 6,100 to 6,500 feet (1,859 to 1,981 m).

Most of the eastern and western boundaries of this geomorphic unit are marked by what is best termed a vegetated linear dune. The western feature is quite evident as one travels on Hancock Road. Linear dunes are oriented roughly parallel to the prevailing wind (figure 7). There are many subtypes and their origin is generally poorly understood (Cooke and others, 1993). However, these features are clearly the remnant arms of vegetated parabolic dunes, the noses and other arms having been blown away.

Fluvial Deposits (Qa, Qp, Qae)

Besides the eolian deposits and landforms, a variety of unconsolidated water-laid deposits occur in the park area (figures 3 and 4). The sand and gravel associated with active stream courses are mapped as alluvium (Qa). The largest accumulation of alluvium within the park coincides with the channel of Sand Wash, located along the eastern margin of the dune field. The most widespread group of unconsolidated deposits in the park area is best classified as mixed eolian and alluvial deposits (Qae) (figure 3). These deposits form when accumulations of windblown sand are later reworked by sheetwash or channel flooding during torrential rains (Doelling and others,
These mixed deposits cover much of the area immediately west of the dune field, including the park campground. Unconsolidated gravel deposits lying on bedrock at elevations significantly above the active channels are pediment or terrace deposits (Qp). These remnant deposits indicate that the valley floor once stood at a higher elevation and has since been lowered by erosion. In the park area, pediment gravels are located along both the base of the fault-controlled escarpment to the east and the base of Esplin Point to the west (figure 3). The pediment gravels (Qp) are the oldest of the Quaternary deposits in the area, probably deposited during the Pleistocene Epoch (approximately 2.6 million to 10,000 years ago). All of the other Quaternary units, both aeolian and fluvial, were most likely deposited during the Holocene Epoch, or the last 10,000 years.

Succession of Dune Forms at Coral Pink Sand Dunes

The succession of dune types within this dune field can be generalized as a downwind sequence (south to north) of stabilized and partially stabilized sand sheets and dunes, transverse ridges, barchanoid ridges, climbing dunes, and parabolic dunes. This is a complex sequence that reflects the interplay between topography, sediment supply, vegetation, and possibly age. The progression of forms at Coral Pink Sand Dunes is best understood by considering it to be two subsequences with different primary controls.

The southern half of the sequence represents adjustments in form and process to a topographically controlled increase in sand accumulation with increasing distance from the source area. This subsequence culminates in the deposition of the climbing dunes against the fault-line scarp of the Sevier fault. Other dune fields of the western United States with similar sequences and a pronounced topographic control on dune successions include the Lyndyl Dunes of Utah and Great Sand Dunes National Monument, Colorado (Sack, 1987).

The northern subsequence, climbing dunes to parabolic dunes, reflects a decrease in sand accumulation downwind of the source area, owing to a progressive decrease in the sediment supply and an accompanying increase in vegetation. Similar controls affect the sequence of dunes at the Killpecker Dunes of Wyoming and White Sands National Monument, New Mexico (Sack, 1987).

STRUCTURAL GEOLOGY

Regional Setting

Coral Pink Sand Dunes State Park (CPSDSP) lies within the zone of structural transition between two major geologic or geomorphic provinces of western North America: the Great Basin section of the Basin and Range Province, and the Colorado Plateau province (figure 13). These two provinces are distinguished from one another based on differences in their depositional history and style of structural deformation and related differences in their topography. West of Cedar City and the Hurricane Cliffs (figure 13) lies the Great Basin, an area characterized by north-south-trending fault-block mountain ranges separated by broad intervening basins with internal drainage and filled with Tertiary to Quaternary alluvium and lake deposits. The boundary between the transition zone and the Great Basin coincides with the trend of the Paragonah, Gunlock, and Grand Wash faults (figure 13). The boundary between the transition zone and the more stable Colorado Plateau interior to the east is not as sharply demar-
cated, but generally coincides with the Paunsaugunt fault (Anderson and Christenson, 1989).

The transition zone, designated the High Plateaus section (of the Colorado Plateau) by some workers (for example Davis, 1999), lies between the core areas of the Basin and Range and Colorado Plateau and thus displays characteristics of both major provinces. In simplest terms, the transition zone is that part of the Colorado Plateau that was affected by Basin and Range extensional faulting, more characteristic of the style of deformation further to the west. Basin and Range extension was initiated during Miocene time (approximately 15 million years ago) and is still active today (Davis, 1999). This deformation created east-west-titled structural blocks bounded by widely spaced, north-south-trending, high-angle normal faults; namely the Hurricane, Sevier, and Paunsaugunt faults (Anderson and Christenson, 1989; Davis, 1999). Coral Pink Sand Dunes State Park lies astride the Sevier fault (figures 3 and 13), the central of the three major Basin-and-Range faults of southern Utah.

**Sevier Fault**

The Sevier fault in Utah is the northern portion of a 300-mile-long (500 km) fault zone that includes the Toroweap and Aubray faults in Arizona (Anderson and Christenson, 1989; Davis, 1999; Lund and others, 2008). The fault extends from the Grand Canyon to central Utah, where it loses its discrete character within Miocene volcanic rocks. The north-south trace of the Sevier fault cuts CPSDSP into two roughly equal halves (figure 3).

In the vicinity of the park, the Sevier fault strikes approximately N. 30° E. and dips about 75° W. It is not a single fault plane but rather a complex zone of braided fault
segments (figures 3 and 4). As a normal fault, the Sevier fault was created by extensional stresses that produced relative motion along the fault zone such that rocks lying above the fault plane and west of the fault’s surface trace moved down relative to the up-thrown block of rocks beneath the dipping fault zone and east of the surface trace (figure 4). Strata within the western down-thrown block dip very gently (<5˚) to the north, as evidenced by the exposures in the face of Esplin Point. The blocks between the fault braids, within the up-thrown eastern block, are locally tilted up to 60˚ (Doelling and others, 1989). Farther east, away from the fault zone, the strata again display a very gentle north to northeast dip.

Total normal displacement along the Sevier fault zone increases from south to north, ranging from approximately 1,560 feet (475 m) at the Utah-Arizona border to 4,920 feet (1,500 m) in southern Sevier County. The stratigraphic displacement within the park is obvious because beds of the upper portion of the Navajo Sandstone (Lower Jurassic), exposed on the western down-thrown side of the fault, are abutted against rocks of the up-thrown block ranging from Moenave Formation (Lower Jurassic) to lower portions of the Navajo Sandstone (figures 3 and 4). Anderson and Christenson (1989) estimated the total displacement in the vicinity of CPSDSP at 1,560 feet (475 m). This fault offset has produced the prominent west-facing bedrock escarpment, a fault-line scarp, east of the campground and main area of active dunes (figure 2).

Movement along the Sevier fault probably began 15 to 12 million years ago, corresponding to the beginning of Basin and Range deformation in this area (Davis, 1999). Assessing the recent history of movement along this fault has proven to be difficult (Lund and others, 2008). Throughout most of its length in southern Utah, the Sevier fault cuts through Mesozoic-age bedrock. This limits the opportunity to document the recent history of movement, which is typically accomplished through the study of fault scarps and stratigraphic offset within unconsolidated surficial deposits (Quaternary age). However, studies of the Sevier fault north and south of the park area suggest movements resulting in surface faulting during the late Tertiary to Pleistocene (approximately 4 to 0.56 million years ago) (Anderson and Christenson, 1989; Lund and others, 2008). Two geomorphic aspects of the fault in the general vicinity of the park provide indirect support for Pleistocene deformation (Anderson and Christenson, 1989). First, the west-facing bedrock scarp displays a distinctive profile, with a moderately sloping rounded upper part and steeper cliff-forming lower part. In several places within the park these two slope elements are separated by a small bench or step in the scarp profile. Anderson and Christenson (1989) suggest that recent movement, rather than differential erosion, is responsible for these profile characteristics. The second feature suggestive of recent movement along the fault is the presence of a small but distinctive closed basin at Clay Flat, approximately 1 mile (1.6 km) north of the park. This area of deposition (depo-center) is located at a major left step in the trace of the Sevier fault. The local stress field has apparently created a small pull-apart basin that interrupts the northward transport of sediment by ephemeral streams in Yellow Jacket and Sethys Canyons. In order to maintain such a local depocenter, active fault movement and related subsidence during the late Quaternary is probably required (Anderson and Christenson, 1989).

Coral Pink Sand Dunes State Park lies within the southern end of the Intermountain Seismic Belt (ISB), the north-south-trending zone of seismicity extending from northwestern Montana to the Las Vegas, Nevada, area. The Wasatch fault of northern Utah is the most infamous feature within this zone of active earthquake activity. Earthquakes in the southern portion of the ISB are generally small to moderate in size, typically less than Richter magnitude (M_R) 4.0, and in many cases are not associated with the mapped Quaternary faults of the area (Anderson and Christenson, 1989). The largest historic earthquake in the vicinity of CPSDSP was the magnitude 6.3 (M_R) Pine Valley earthquake in 1902, located 21 miles (33 km) north of St. George. Estimated magnitude-5.7 and 5.5+ (M_L) earthquakes occurred east of the park near Kanab in 1887 and 1959, respectively. Compared to the Hurricane fault 40 miles (64 km) to the west, which generated the 1992 St. George earthquake (M_w = 5.8), the Sevier fault has had a much lower level of seismicity during historic time. Only two historic (1900-1985) earthquakes have been recorded along the main trace of the Sevier fault in Utah (Davis, 1999), neither of them within the park. Anderson and Christenson (1989) raise the possibility that the Sevier fault may be locally deforming by continuous aseismic fault creep, as opposed to large ground-rupturing earthquakes. Either way, the low-level earthquake activity, coupled with the geomorphic character of the fault-line scarp noted above, suggest that the Sevier fault is still active today.

**Deformation-Band Shear Zones**

The Navajo Sandstone of southwestern Utah displays a variety of outcrop-scale structures that resemble veins or dikes weathering out in positive relief from their host sandstone. These features, called deformation bands, have been recently studied and analyzed in careful detail by Davis (1999) and Fossen and others (2007). Deformation bands are the result of porosity reduction, grain crushing, grain fracturing, and grain flow that occur during the deformation of porous granular materials. Davis recognizes two varieties of deformation bands, cataclastic and non-cataclastic, both of which are present in the Navajo Sandstone within the vicinity of CPSDSP. Cataclastic deformation bands are closely associated with the regional faults and folds of the area, including the Sevier fault. Slip within these features is indicated by slickensides, and a distinctive ladder structure is common. The most distinctive manifestation of cataclastic deformation bands are the formation of “fault fins,” triangular blades of sandstone up
to 26 feet (8 m) tall (see cover photograph of Davis, 1999). Fault fins are produced by differential weathering and erosion; the reduced porosity within the deformation bands increases their resistance. Tilted deformation bands then serve to protect underlying sandstone from erosion (Davis, 1999).

The second or noncataclastic type of deformation band is present in almost every outcrop of the uppermost 33 to 50 feet (10–15 m) of the Navajo Sandstone (Davis, 1999). The noncataclastic deformation bands are vein-like structures that weather in positive relief but show minimal evidence of grain-scale fracturing and no slickensides. Davis (1999) concludes that noncataclastic deformation bands formed in Jurassic time by the loading of undercompacted dune sand of the Navajo Sandstone by sediments that now make up the Carmel Formation. Aided by groundwater infiltration, the Navajo sands failed by vertical compaction. The deformation bands mark the zones where this volume reduction took place. The uppermost part of the Navajo Sandstone outcrops near the northwest corner of the park. Large blocks of sandstone, apparently quarried from that area, have been placed at camp sites and along the loop road. “These blocks are museum pieces of . . . noncataclastic deformation bands” (Davis, 1999, p. 66) (figure 14).

**GEOLOGIC HISTORY**

A complete geologic history of the area around Coral Pink Sand Dunes State Park (CPSDSP) is beyond the scope of this paper. Interested readers should consult Peterson and Turner-Peterson’s (1989) overview of the Colorado Plateau, as well as other papers in this volume. We will focus on those geologic events that have specifically left their mark on the landscape of the park. The most important events are the Triassic-Jurassic deposition of the Glen Canyon Group (Moenave, Kayenta, and Navajo Formations), the Cretaceous to Cenozoic structural deformation of the Colorado Plateau, and the Quaternary formation and evolution of the active dunes. This selective history is primarily abstracted from the work of Baars (1983), Stokes (1986), Harris and Tuttle (1990), Patton and others (1991), Barnes (1993), Morris and Stueben (1994), Elias (1997), Davis (1999), and Blakey and Ranney, 2008).

**Jurassic Environmental Change**

For much of the Paleozoic Era (about 54 to 250 million years ago), Utah was situated on the western edge of the North American continent and dominated by shallow-marine deposition. During Middle Triassic time (245 to 230 million years ago) western Utah was cut off from the ocean to the west by a highland barrier, termed the Mesocordilleran High (Stokes, 1986), which was located in what is now eastern Nevada. To the east, in what is now western Colorado, was another highland area called the Uncompahgre uplift, an eroded remnant of the Ancestral Rocky Mountains. Thus, the subsequent Triassic-Jurassic formations of the Glen Canyon Group were deposited in a sedimentary basin that formed between these two highlands, dominated by terrestrial depositional environments. The Mesocordilleran High blocked moisture-laden air masses and served to enhance arid conditions in Utah by creating a rainshadow. The Mesocordilleran High was also a source area for rivers flowing eastward and carrying sediment into the basin.

Fluvial and lacustrine deposition during the Late-Triassic and Early Jurassic resulted in the sandstones, siltstones, and freshwater limestone of the Moenave Formation. During this time (approximately 200 to 196 million years ago) the North American continent was drifting northward through equatorial latitudes. The rock types and sedimentary structures preserved within the Moenave Formations suggest a landscape of low-gradient stream channels with broad floodplains containing lakes and mudflats. Fossil remains of large freshwater fish, similar to sturgeon, as well as numerous dinosaur footprints and tracks (see Milner and others, 2006), have been found in the Whitmore Point Member.

As North America drifted north away from the equator, the climate in what is now southern Utah became cooler with wet summers and dry winters (Barnes, 1993). Streams deposited silt and sand in channels and on floodplains; dinosaurs left their footprints in the damp sediments that would become the Kayenta Formation (196 to 190 million years ago).

Continued northward drift brought a large portion of North America under the influence of subtropical high pressure, resulting in increasing aridity and the establishment of a huge Jurassic sand sea or erg. One estimate indicates that the erg may have covered 256 million square miles (6.63 x 105 km²) (Marzolf, 1988). However, the intertonguing of the Kayenta and Navajo Formations represented by the Lamb Point and Tenney Canyon indicates an...
initial period of time during which fluvial and eolian systems vied for dominance in southern Utah. A Jurassic desert, represented by the Lamb Point Tongue, appears to have advanced from the east, burying and replacing the fluvial channels and floodplains of the Kayenta Formation. The margins of the sandy desert temporarily receded back to the east, replaced by fluvial environments of the Tenney Canyon Tongue. The desert once again advanced and overwhelmed the fluvial landscape. Eolian deposition, represented by the main body of the Navajo Sandstone, then dominated this area for approximately 15 million years (190 to 175 million years ago).

Paleogeographic reconstructions suggest that the erg was situated near the western edge of the continent between approximately 18° and 25° north latitude, under the influence of onshore north-northwesterly winds associated with the eastern edge of a subtropical high (Marzolf, 1988; Peterson, 1988; Blakely, 2008; Blakely and Ranney, 2008). A shallow marine shelf, estuary, or large lake bordered the lowland desert on the west. Peterson (1988) suggests that the source area for the sand was an uplifted area in central Montana. Rivers carried sediment westward to the coast from this upland, longshore currents moved the sediment southward, and then waves and northwesterly winds delivered the sediment to Utah. Marzolf (1988) suggests a different source for the sand. He suggests that rivers draining northwestward from upland areas in Arizona delivered sediment to the head of the marine embayment to the west, where the silt and clay was winnowed and the sand was returned inland by the onshore winds.

Navajo Sandstone deposition ended, possibly the result of climate change, in the vicinity of CPSDSP when streams, loaded with red mud, flooded the dunes and partially truncated them. For wind-blowsediments to be preserved as eolian sandstones, they must be deposited in an actively subsiding basin. In the case of a coastal erg, if deposition does not keep pace with subsidence, the sea will transgress or flood the lowland. This is what occurred along the western edge of the Navajo erg during Middle Jurassic time, resulting in a major unconformity (J-1) (figure 6). A shallow seaway extended into southwestern Utah from the north.

The mostly marine sediments flooded the region and became the Sinawava Member of the Temple Cap Sandstone (Middle Jurassic). Sea level fluctuated during Temple Cap deposition, which allowed eolian deposition to quickly resume producing the White Thrown Member of the Temple Cap Sandstone. Sea level rose again, but this time the sea flooded farther inland completely inundating the erg field resulting in deposition of limestone beds of the Co-op Creek Member of the Carmel Formation (Middle Jurassic).

Upper Jurassic through Lower Cretaceous beds are not present in the vicinity of the park and were probably removed by a long period of uplift and erosion associated with the Laramide orogeny. The nearest outcrops of Cretaceous rocks are poor exposures of Upper Cretaceous Tropic Shale and Dakota Sandstone located near Mt. Carmel. Thus, the modern sand dunes of the park lie directly on Lower Jurassic formations, most notably the Navajo Sandstone. The boundary between the two aeolian units, one lithified and one still moving, represents a hiatus or gap in the stratigraphic record of approximately 180 million years.

Cretaceous to Cenozoic Tectonics

Towards the end of Mesozoic time, and continuing into the Cenozoic, western North America was subjected to a mountain-building episode known as the Laramide orogeny (approximately 90 to 50 million years ago). This orogeny was the result of compressional stresses generated by plate-tectonic convergence between the North American plate and the oceanic Farallon plate to the west. Subduction of the Farallon plate, and related folding and thrust faulting, gave rise to the Rocky Mountains. The Colorado Plateau was also greatly affected by this deformational event (see Davis and Bump, 2009). Compression of the Colorado Plateau region was accommodated by basement-cored reverse faulting and associated folding which produced the Kaibab uplift, the San Rafael Swell, and the numerous broad anticlines and synclines of the Kaiparowits Plateau, among other major structures. Since the beginning of the Laramide orogeny the nearly circular Colorado Plateau has acted as a coherent structural block and has been uplifted more than 6,600 feet (2,012 m) (Morris and Stubble, 1994).

One hypothesis (Beghoul and Barazangi, 1989) suggests the Farallon plate may have been key to this uplift, becoming attached during subduction to the bottom of the North American plate in the vicinity of the Colorado Plateau. During mid-Cenozoic time the Farallon plate separated from the North American plate and sank into the mantle. As the Farallon plate remnant sank, hot material from the asthenosphere rose to fill the void. This material thermally expanded as it rose and some of it intruded the lower continental crust, thereby increasing its thickness and temperature. The Colorado Plateau is thus seen to have been uplifted by isostatic compensation resulting from a combination of heating and increased crustal thickness.

A second deformational event, Basin and Range extension, also had a pronounced effect on the geology of the park. This episode of normal faulting and crustal extension began approximately 15 million years ago, in mid-Miocene time, and is still active today (Davis, 1999). There is no consensus among geoscientists as to the plate-tectonic explanation for this extension. Some (for example Stewart, 1971) suggest that a slowing of convergence between the Farallon and North American plates coupled with a rollback of the subducting Farallon plate essentially created back-arc spreading within the continental crust. Others (for example Dickinson, 1979) have suggested that the North American plate completely overrode the mid-ocean ridge along the western boundary of the Farallon plate,
thus creating a window in the subducting slab through which heat and magma from the asthenosphere could rise to lift and thin the overlying continental crust. Either way, the beginning of this extensional deformation was coincident with the cessation of subduction along the California coast and the initiation of the transform boundary between the North American and Pacific plates marked by the San Andreas fault.

Whatever its cause, this extensional deformation created the distinctive topography of the Basin and Range Province, including the Great Basin section in Nevada and western Utah, as well as the three major high-angle normal faults located within the southwestern margin of the Colorado Plateau—namely the Hurricane, Sevier, and Paunsaugunt faults (figure 13). Movement along the Sevier fault has had a major influence on the geologic history of the park, creating the fault-line scarp that dominates the present topography. Without this fault scarp, the varied Mesozoic strata would not be exposed within the park. In addition, this scarp is in part responsible for the concentration and accumulation of wind-blown sand at this location.

The uplift of the Colorado Plateau caused a major change in the nature of the dominant geologic processes. Whereas the Mesozoic Era was dominated by deposition, the subsequent Cenozoic geologic history of the Colorado Plateau has been dominated by erosion, producing the world-renowned canyonlands and cliff-and-bench topography. This erosion is very evident in the landscape of CPSDSP, exposing the picturesque formations of the Glen Canyon Group. However, localized eolian deposition during the Holocene Epoch has created the features for which the park is named.

Late Holocene History of the Dune Field

The Quaternary Period is subdivided into the Pleistocene and Holocene Epochs, with the boundary between the two occurring approximately 10,000 years ago. The Pleistocene is noted for its alternating glacial and inter-glacial conditions. The Holocene is essentially the most recent and on-going inter-glacial time period that began when the last ice age (the Wisconsin glaciation) ended. Late Pleistocene full-glacial (22,000 to 18,000 years ago) conditions on the Colorado Plateau probably inhibited major eolian transport and deposition. Paleobotanical evidence suggests that late Wisconsin winters were wetter and summers were cooler and drier than present (Betancourt, 1990). Drier summers probably reflect a decrease in the availability of subtropical moisture via the southwest monsoon, whereas wetter winters were caused by a southwestern shift in the polar jet and an increased occurrence of Pacific storms and associated frontal precipitation over the Plateau. The result was coniferous woodlands, including limber pine, juniper, Colorado blue spruce, Douglas-fir, and sagebrush, were present in areas that now support pinyon-juniper woodlands. The area of the Coral Pink Sand Dunes may have supported a coniferous woodland during the late Pleistocene. Notably, ponderosa pine was apparently absent from the Colorado Plateau during this time, due to a lack of summer moisture and/or a reduction of thunderstorm activity and lightning strikes; ponderosa pine distribution is in part linked to fire (Betancourt, 1990).

The Pleistocene-Holocene transition on the Colorado Plateau is generally marked by gradual changes in vegetation that indicate warming and a progressive decrease in effective moisture. Macrobotanical remains from the Escalante River basin indicate decreased abundances of mesophytic plants and an upslope retreat of Douglas fir and spruce (Withers and Mead, 1993). However, the early to mid-Holocene (10,000 to 6,000 years ago) on the Colorado Plateau saw a significant increase in summer precipitation (and thunderstorm activity?) (Betancourt, 1990). Ponderosa pine expanded during this period beyond even its present distribution. Even during the latter part of this period, commonly associated with hot-dry conditions in other parts of the southwest (the altithermal), the Colorado Plateau appears to have been wetter than today. The most likely explanation is more abundant subtropical moisture from an intensified Bermuda High and southwest monsoon (Betancourt, 1990). Pollen evidence from Posy Lake on the Aquarius Plateau, north of the park, indicates that the period of enhanced summer precipitation persisted until 5,500 years before present (B.P.) (Shafer, 1989). Lake levels were lowest at this site from about 5,000 to 3,500 years B.P., indicating greater aridity than today. The oldest dates that we have obtained from the deposits in the dune field, approximately 4000 yr B.P. (discussed below) coincide with this period of regional aridity.

Late Holocene Geochronology and Dune Activity

Using a combination of dating methods, including radiocarbon dating, optically stimulated luminescence (OSL) dating, and dendrochronology, we have been able to reconstruct the changes in the environment and resulting changes in dune activity over the past several millennia preserved within the Holocene stratigraphy of the dune field (Wilkins and others, 2005; 2007).

Dune migration and erosional lowering of the upwind surfaces of some of the tallest transverse dunes (Qtd) has uncovered the tops of several large ponderosa pine snags, in situ and upright, presumably killed through deep burial by passing dunes. Radiocarbon dating (note: all ages are reported at the 2σ level and calibrated using intercal 09.14c; Stuiver and Reimer, 1993) of the outer rings of a branch near the top of a ten meter-tall snag gives an age of death of 190±50 14C yr BP (Cal. 1643–1950 CE; Beta 181947); ring counting revealed 188 rings, indicating that the stand of trees was established for several hundred years before it was buried.

In a swale (approximately 500 meters north of the snags) between active transverse dunes, a fallen and heavily weathered ponderosa pine rooted in bedrock has been uncovered by dune migration. Radiocarbon dating of the pith, or center rings, at its base yields an age germination
of 250±40 $^{14}$C yr BP (Cal. 1485–1950; Beta 181943); counting the rings from the pith to the outer edge of the stem indicated a minimum age of 75 years before its death.

One hundred meters west of the downed ponderosa, recent dune movement has also exhumed a paleosurface consisting of a muddy soil capping older eolian sediments. The older sediments are two to three meters of friable cross-bedded eolian sand comprised of three units, each one to two meters thick, all with very similar grain mineralogy, color, and sorting. The uppermost eolian unit is unconformably overlain by a more resistant silty mud layer. The mud and soil layer, 0.25 meters thick, exhibits flow features at the base, including rip-up clasts, in the silty matrix characteristic of a sheetflood deposit. The mud layer is topped by an organic layer consisting of what appears to be a duff layer from woodland forest floor, including charcoal and plant macrofossils of juniper berries, cones (Pinus edulis), stems and roots in situ, indicating a pinyon-juniper woodland had once occupied the site.

A method of dating, called optically stimulated luminescence (OSL) dating, is a useful tool in desert environments to date the time since sand was last exposed to sunlight and buried. Samples of sand for OSL dating were collected at the upper contact of each buried eolian unit below the exhumed soil. Dating of the samples record an upper limiting age for the termination of eolian activity and deposition related to that unit. The OSL ages for deposition and burial are in the proper stratigraphic order, with the upper unit terminating 0.51±0.06 ka (1435–1555 CE) at a depth of 0.7 m below the surface. The two lower units had ages of 2.8±0.22 ka (2.0 m, 1015–525 BCE) and 4.1±0.19 ka (2.8 m, 2205–1905 BCE), respectively (Wilkins and others, 2007).

Radiocarbon dates were also obtained from this site. A sample of the silty mud was collected from the base of the layer and was dated using the bulk organic carbon component, yielding an age of 470±50 $^{14}$C yr B.P. (Cal. 1400–1490 CE; Beta 190240). A pinecone embedded in the surface soil and duff layer yielded an age of 200±40 $^{14}$C yr B.P. (Cal. 1761–1803 CE; Beta 181944; Wilkins and others, 2005).

The OSL dating of the eolian sand and radiocarbon dating of the exhumed soil, plant matter, and tree snags provide the basis for reconstructing the latest Holocene environments in the Coral Pink Sand Dunes. The luminescence dates indicate that eolian activity has been present in the park area for at least the last 4000 years, though our understanding is still limited as to how episodic or persistent the earlier periods of eolian activity were. The youngest eolian unit, initiating some time prior to its 500 OSL yr BP age, is likely related to one or more of the megadroughts in the western US in the 13th, 14th, and early 15th centuries (Oviatt, 1988; Stahle and others, 2007; Mensing and others, 2008) and also coincides with the regional abandonment of Anasazi farming settlements (Benson and others, 2007). Eolian activity in the park area during that time is broadly similar to activity reported for the Great Sand Dunes (Colorado; Forman and others, 2006) and the Tusayan Dunes (Arizona; Stokes and Breed, 2003).

The termination of eolian activity and apparently abrupt transition to wetter conditions—as recorded in the waterlain muds over the eolian sediments, expansion of the pinyon-juniper woodland into the dunes, and establishment of the ponderosa in the dune field—is dated to the start of the Little Ice Age (LIA). The onset of dune stability corresponds to a period of valley deposition in the surrounding region, as reported by Hereford (2002), as well as an absence of prolonged, multidecadal drought (Cook and others, 2004). Hereford (2002) proposes that the accumulation of valley fill resulted from a general absence of large floods. In the area of the Coral Pink Sand Dunes, the Little Ice Age appears to have been a time of landscape stability and diminished eolian transport, interspersed with periods of drought and resurgence of eolian activity such as that which led to the burial and death of ponderosa pines in the active core of the dune field. The ages of the snags in the dunes (250±50 and 190±50 $^{14}$C yr BP) correspond to a period of eolian activity at the Great Sand Dunes in Colorado (Forman and others, 2006), centered around 1800 CE that is attributed to a period of reduced precipitation in the region.

The last half of the 19th century marked the end of the Little Ice Age in the Colorado Plateau region (Hereford, 2002). Data from Cook and others (2004) and Stahle and others (2007) both report a number of short-term severe droughts during this period; tree rings in cores collected in CPSDSP also record these droughts as very narrow rings, most notably a severe drought in the 1890s that lasted several years.

The end of the Little Ice Age is also marked by a return of frequent large floods in the region (Hereford, 2002; Ely, 1997)—the early 20th century is marked by a prolonged period of increased moisture termed a “pluvial” by Stahle and others (2007), and is recorded in CPSDSP tree cores as very wide rings. The pluvial was followed immediately by the extreme continental-scale drought of the 1930s known as “the Dust Bowl” (Stahle and others, 2007). H.E. Gregory (1950), the first geologist to record a visit to the Coral Pink Sand Dunes, reported a highly active dune field in 1937; this activity was likely a response to the drought conditions still present from the early part of that decade. Aerial photos of the dunes acquired in 1957 and 1960 also show very active dunes, likely a response to the extreme drought in the southwestern U.S. in the 1950s (Stahle and others, 2007).

In the four decades since those photos, 1961–2000, local climate records record a slight increase in mean annual precipitation over the previous three decades, 1931–1960. In response, the vegetation at the margins of the dune field expanded and become more broadly established in the dunes—aerial photos acquired in 1997 show a contraction of active dune areas since 1960 (Wilkins and Ford, 2007).
GEologic uniqUENESS OF THE PARK

Naturalist Freeman Tilden (1970) concluded that sand dunes are nature’s nudes. Indeed, there is something very pleasing, even sensual, about the curves and bare contours of active dunes. Dunes bring to mind romantic adventures in Old World deserts, as well as frontier tales from the American West. It is no wonder then that the Coral Pinks Sand Dunes have been used as locations for several Hollywood films: Arabian Nights (1942)—“corny escapist stuff” with an enslaved Sheherazade and a retired Sinbad the Sailor; Mackenna’s Gold (1969)—“overblown adventure saga about search for lost canyon of gold,” and One Little Indian (1973)—“AWOL cavalry corporal escaping through the desert with young Indian boy and a camel” (Maltin, 1997). Geologic history, OHV recreation, film making, and wildlife habitat (to be discussed below)—all make this is an interesting, important, and valuable landscape.

Source of the Sand

Although the commercial use of the Coral Pink Sand Dunes dates to 1942, the earliest scientific treatment of the dune field is Herbert Gregory’s (1950) description of the dunes in his general survey of the geology and geography of the Zion National Park area:

Likewise the disintegrated products from the bare sandstones of South Mountain are carried eastward across Moccasin Terrace to form the dunes along the cliffs at Sand Canyon and the great sand flats at the head of Three Lakes Canyon. With local accretions en route these deposits are redistributed over Wygaret Terrace and eastward to the base of the Kaiparowits plateau. During sandstorms great quantities of the finest dust rise high into the air and move eastward to distant places or after darkening the skies with whirling clouds return to the place from which they came. . . . Another large area of dunes extends from the head of Cottonwood and Water Canyons northwestward across Sand and Yellow Jacket Valleys to the base of the Block Mesas near Esplin Point. For a distance of about 8 miles northwest of Riggs Spring dunes as much as 200 feet high are nearly stationary; they support growths of pitoon and juniper and in wet seasons hold lakes and give rise to springs. At the head of Sand Canyon they are in motion and in their migration are burying and uncovering trees and converting the original escarpment of the Sevier fault into a slope (Gregory, 1950, p. 188).

Botanist Elias Castle (1954) may have been the first to use the term “coral pink” to describe the color of the dunes. His report, probably the first to focus specifically on the dunes in the park area, summarizes the soil conditions and their influence on the various vegetation communities within the dune field.

Kane County has many aeolian deposits though most are quite small. In addition to the Coral Pink Sand Dunes, the Sand Hills, which extend across U.S. 89 between the Sevier fault and Kanab Creek, are large enough to be portrayed on regional geologic maps (Sargent and Philpott, 1987; Doelling and others, 1989). Of the many Jurassic sandstone units exposed in the vicinity, the Navajo, Page, and Entrada Sandstones are thought to be especially productive in supplying sand for dunes. Doelling and others (1989) suggest that the friable and poorly cemented middle portion of the Navajo Sandstone provided the bulk of the reddish sand for the Coral Pink Sand Dunes. The reddish orange (coral pink) color of the quartz sand is likely inherited from iron oxide grain coatings and cement of the Navajo Sandstone (Chan, 2010, personal communication).

Although Gregory (1950) indicates a western source (South Mountain) for the sand, the prominent southwest-northeast orientation of the dune field and prevailing southwesterly winds suggests to us a source to the southwest. The middle pink portion of the Navajo Sandstone is exposed extensively across the Moccasin Terrace, southwest of the park. In addition, extensive surficial deposits, primarily deposits of fine-grained sand deposited by mixed alluvial and eolian processes (Qae), cover the Moccasin Terrace (Doelling and others, 1989). These unconsolidated deposits are a likely source for the sand that comprises the Coral Pink Sand Dunes.

This suggested source indicates a very interesting path through the rock cycle for these sand grains—ancient eolian sandstones weathering to produce modern aeolian sediments. Recall that recent studies (Dickinson and Gehrels, 2003) suggest that rocks as far away as the Appalachian Mountains were weathered and eroded to produce sediment that was transported westward by rivers to a marine embayment during Jurassic time. This sediment was transported southward by ocean currents, progressively winnowed by wave and tidal action, and eventually transported back to land by onshore winds. These sand grains were then deposited in large dunes of a lowland erg. The dunes were buried by younger sediments and eventually lithified to sandstone, what we now call the Navajo Sandstone. These eolian sandstones remained buried for perhaps 125 million years or more. Uplift of the Colorado Plateau during the Tertiary Period set the stage for the recent and on going weathering and erosion, releasing the sand grains from the ancient rocks. Once again, perhaps 170 million years latter, the sand grains are eroded and again transported by wind. The modern wind regime has concentrated many of these sand grains in the Coral Pink Sand Dunes and continues to slowly transport them to the northeast, to encroach on and bury Jurassic Navajo Sandstone outcrops (figure 11).

Formation of the Dune Field

An obvious question is why have these modern eolian sediments been concentrated here? Although no local wind data are available, a park brochure (Utah State Parks
and Recreation, no date) and interpretative signage describe the existence of several regional winds that converge in the vicinity of the park. A break in the Vermillion Cliffs south of the park, due to offset on the Sevier fault, likely allows winds to accelerate through the gap between the Moquith and Moccasin Mountains and to entrain loose sediment on the Moccasin Terrace. As the gap widens in the vicinity of the park and the winds impinge upon the bedrock scarp of the Sevier fault, velocity may decrease to allow deposition. Named for an Italian physicist, the acceleration of air or liquid as it is forced through a narrow constriction is called the Venturi effect. The southern portion of the dune field has a distinct north-south orientation, possibly reflecting the influence of this southerly wind.

About 2 miles (3.2 km) north of the southern end of the dune field, in the generally vicinity of the park campground, the orientation of the dune field changes to southwest-northeast. This may reflect the influence of prevailing southwesterly winds blowing up from Rosy Canyon, another wind gap in the Vermillion Cliffs to the southwest. These southwesterly winds may be funneled between the Moccasin Mountains to the south and the Block Mesas, including Esplin Point, to the north. The fault-related scarp on the eastern edge of the dune field may act to slow these winds as well, thus initiating deposition at the base of the scarp. The sand deposited at the base of the scarp is the obvious source for the dunes that climb the fault scarp and extend approximately 4 miles (6.4 km) northeast onto BLM land. This secondary sand movement may be facilitated by an acceleration of the wind as it flows through the narrow, 1-mile (1.6 km) gap between Esplin Point and the scarp of the Sevier fault. Thus, the Coral Pink Sand Dunes appear to owe their existence to an abundant supply of fine-grained sand from the Navajo Sandstone and other units, strong prevailing southerly and southwesterly winds, and the decelerating influence of the Sevier fault-line scarp. Specific data about the local wind regime are needed to confirm the highly speculative hypothesis presented above.

**LOCATION AND DESCRIPTION OF CLASSIC GEOLOGICAL SITES WITHIN THE PARK**

**Overview Area and Nature Trail**

A parking lot and overview area (figure 8) is located on the east side of the campground road between the ranger station and the campground. Most non-OHV visitors to the park utilize the boardwalk here to access the dune field. The overview is built atop what is best termed a vegetated linear dune (Tsoar and Moller, 1986) that marks the eastern boundary of this portion of the dune field. Unlike the vegetated linear dunes along the margins of the dune field to the north, the origin of this feature is not obvious. The best interpretation we can provide for this example is that it is the result of complex airflow along the boundary between the dunefield and the vegetated valley floor.

From this vantage point one can look to the south and see the gap between the Moquith Mountains (east, left) and Moccasin Mountain (west, right) through which southerly winds may be accelerated by the Venturi effect. One can also look to the southwest across much of the Moccasin Terrace, the probable source area for much of the sand in the dunes. To the north-northeast is the narrow gap between Esplin Point and the west-facing escarpment of the Sevier fault, which may facilitate acceleration of the wind and the transport of sand over the escarpment to form the large parabolic dunes at the northern end of the dune field (Qelp). Directly east of the overview are the active core of the dune field (Qetd) and the prominent bedrock escarpment of the Sevier fault (figure 2).

**Nebkhas**

A short nature trail begins beyond the boardwalk overview and traverses into the dune field. Interpretative signs along the trail focus on the relationships between vegetation and certain dune forms. Here one can see excellent examples of nebkhas (also called shrub-coppice dunes), mounds or hummocks of sand trapped by clumps of vegetation (figure 15). Although present in each geomorphic unit of the dune field, nebkhas are particullary abundant here along the southwestern margin of the active core of transverse dunes (Qetd).

On flat surfaces wind velocities may decrease as much as 50 percent on the downwind side of a bush or shrub (Francis, 1994). Both the decrease in velocity and the physical obstruction of the stems or branches cause deposition of wind-blown particles within the plant, including sand, silt, organic debris, and nutrients attached to dust nuclei. Such accumulations may locally improve the soil fertility.
A nebkha will develop where the plants grow fast enough to remain on top of the accumulating mound. Nebkas may eventually reach 3 to 15 feet (1–3 m) in height and 6 to 30 feet (2–5 m) in diameter. Typically they are oval or elliptical in plan form. The nebkhas in the Coral Pink Sand Dunes are generally less than 6 feet (2 m) in height. Here, several different plant species are adapted to the stress of progressive burial by wind-blown sand, but *Wye thia scabra* (mule ears) is by far the most common. *Wye thia scabra* nebkhas are particularly abundant along the margins of the dune field. In some areas the relief of the nebkha has been increased by deflation or eolian erosion of the ground adjacent to the nebkha, locally exposing the matted and twisted stems and roots of the mound-forming plant for a considerable depth below the top of the dune (Castle, 1954).

**Star Dune**

The large dune directly east of the boardwalk overview is an excellent example of a star dune (figures 2 and 8). Star dunes commonly have a high central peak and three or four curving arms, with multiple slip faces, that radiate outward from the central mass. Star dunes form where winds from more than two directions have contributed to dune growth. They tend to grow vertically, rather than migrating horizontally, and contain a greater volume of sand than any other dune type (Lancaster, 1995). McKee (1982) noted that most of the star dunes of the Namib Desert, southwestern Africa, developed from linear dunes, although some are associated with transverse dunes, as is the case in the Coral Pink Sand Dunes.

Star dunes typically achieve heights of 45 to 450 feet (14-137 m), though some modern star dunes in the Namib Desert are reported to rise as much as 900 feet (274 m) above their interdune base (McKee, 1982). The large star dune at CPSDSP rises approximately 100 feet (30 m) above its base and has three sinuous arms and several smaller secondary arms.

We were surprised to learn, through the study of aerial photographs, that the star dune did not exist in 1960, but is clearly evident on photographs from 1997 (Wilkins and others, 2003). The photographs suggest that several transverse dunes merged, sometime between 1960 and 1997, to form the lone star dune. We have collected data from several ground-penetrating radar (GPR) transects in the area of the star dune since 2000 (Wilkins and others, 2002; 2003). Data quality is excellent with strong reflectors up to 60 feet (20 m) below the surface being recorded. The presence of unidirectional, high-angle cross-bedding in a GPR transect that crosses an arm of the star dune (figure 16) supports the transverse-dune parentage of this feature. Most star dunes are thought to be composed of low- to moderate-angle ripple-laminated sand (Nielsen and Kocurek, 1987). Thus the internal structure and recent history of this dune suggests a distinctive mode of star dune formation.

**Barchan (Current Crescent)**

Also visible from the boardwalk overview is a large barchan-like dune, approximately 0.2 miles (0.3 km)
south-southwest of the star dune. Barchans are characterized by a distinctive crescentic shape in plan view, a steep concave slipface, and “horns” extending downwind from the central mass of the dune (figure 8). Barchans are the stable dune type in areas where the directional variability of the wind is less than 15°. Isolated barchans generally occur in areas with a limited sand supply; as the sand supply increases barchans coalesce to form barchanoid ridges (Lancaster, 1995). However, it is puzzling that this dune type so indicative of a unidirectional wind regime is in such close proximity to a star dune, indicative of multidirectional winds. A mitigating factor may be the fact that the isolated barchan at CPSDSP is not a freely migrating dune, but instead has formed largely upwind of a small knob of Navajo Sandstone, which can be observed in the dune’s slip face and GPR transects. We believe this dune is not a true barchan, but instead is best termed a current crescent. Current crescents are horseshoe-shaped dunes that accumulate where winds are deflected around and over topographic obstacles such as boulders, escarpments, or hills (Pye and Tsoar, 1990). Sand is deposited both in front of the obstacle and on either side in two tapering wings that resemble the horns of a true barchan. The 1960 and 1997 aerial photography support this interpretation, as the current crescent formed when a migrating transverse dune impinged onto and wrapped around the bedrock knob. We speculate that the formation of the current crescent may have modified the local wind regime, concentrating flow immediately downwind and leading to the formation of the isolated star dune (Wilkins and others, 2003).

Tiger Beetle Conservation Area (“Cattle Guard Turnout”)

Approximately 1.6 miles (2.6 km) north of the ranger station, on the east side of the road, is a turnout and small parking area locally referred to as the “cattle guard” area (figure 8). This site provides easy access to the active core of transverse dunes (Qetd) and the area of small marginal parabolic dunes (Qemp). From this area one can also see the barchanoid ridges (figure 10) impinge upon the bedrock escarpment, transforming into echo dunes and climbing dunes. This is also an excellent location to observe the ecological differences between the largely unvegetated dune ridges and the vegetated interdune swales. Some of the interdune swales are incised, exposing their shallow stratigraphy. This stratigraphy, which includes mudcracks and darker, organic-rich (?) layers, indicates cycles of water ponding and desiccation and periods of weak soil development within these interdune areas.

This general location is also the habitat of the Coral Pink Sand Dunes (CPSD) tiger beetle (Cicindela limbata albissima), which is only known to inhabit this dune field. The biology and ecology of the CPSD tiger beetle has been largely revealed through the research of entomologists Barry Knisley (Randolph-Macon College, Ashland, Virginia) and James Hill. The following account is summarized from their work, primarily Knisley and Hill (2001).

Tiger beetles are predatory insects that prefer open, sparsely vegetated habitats, such as sand dunes. Many species of the genus Cicindela inhabit dune fields of the western United States and several are endemic to specific dune fields, including the CPSD tiger beetle. Surprisingly, the CPSD tiger beetle appears to regularly inhabit only a small portion (approximately 200 acres or 81 ha) of the total dune field, making its geographical range one of the smallest of any organism known to science.

In 1994 the Southern Utah Wilderness Alliance petitioned the U.S. Fish and Wildlife Service (USFWS) to list the CPSD tiger beetle as an endangered species. Currently, the CPSD tiger beetle is recognized as a candidate for listing as an endangered or threatened species. In 1998 the USFWS, BLM, Utah Department of Natural Resources, and Kane County signed an agreement that established a 200-acre (81 ha) conservation area within CPSD State Park. Off-highway vehicles are not permitted in the western portion of the conservation area where most of the tiger beetles are found. The eastern portion of the conservation area serves as a travel corridor between the southern and northern portions of the dune field. An additional 370 acre (121 ha) conservation area has been established on BLM-administered land at the extreme northern end of the dune field where a small number of dispersing tiger beetles have been observed. The CPSD tiger beetle was once considered a subspecies of the Sandy Tiger Beetle (Cicindela limbata). DNA studies published in 2000 (Morgan and others, 2000) show it to be a distinct and separate species and highlight the importance of the conservation measures taken within the park.

Although the range of the CPSD Tiger beetle is quite small, as is its total population, there is no evidence of a progressive decline in CPSD tiger beetles during the time that Knisley and Hill (2001) have monitored their population (1992-present). They believe that the implementation of the conservation plan in 1998, which protects the beetles’ primary habitat, will promote the long-term existence of this species within the Coral Pink Sand Dunes.

A puzzling question, for which there is presently no definitive answer, is why is the distribution of the CPSD tiger beetle so restricted within the overall dune field? Knisley and Hill (2001) point out that there is significant variation in the geomorphology, vegetation, prey insects, and OHV activity throughout the dune field. They note that the percent vegetation cover in the interdune swales of the primary habitat/conservation area is significantly greater than in the swales of other areas of the dune field. Probably related, the numbers and types of prey species are also significantly greater in the conservation area compared to other areas of the dune field. Are these differences related to purely natural factors, such as the geomorphic transition from more active, less vegetated barchanoid dunes in the southern portion of the dune field to the less active, more vegetated climbing and parabolic dunes to...
the north? Or are the habitat differences related to OHV activity? Knisley and Hill (2001) suggest that high OHV use contributes to the highly dynamic nature of the southern portion of the dune field, decreasing the cover of vegetation in interdune swales. These questions point out the distinctly interdisciplinary (biology and geology in this case) nature of many of the resource and land-use issues or concerns now faced by human societies.

Final Comments

This paper has attempted to summarize our current understanding of the particular geomorphic system known as the Coral Pink Sand Dunes. We know of no other dune field in Utah, with the possible exception of the Lynndyl Dunes (BLM Little Sahara Recreation Area) (Sack, 1987), that contains the diversity of eolian landforms present here. Hopefully we have shown that much can be learned at Coral Pink Sand Dunes State Park, both individually and individually by observant visitor-naturalists and formally and cooperatively by professional scientists. Coral Pink Sand Dunes will continue to provide a natural laboratory in which to observe and discern the complex interactions among atmospheric, geomorphic, and ecological processes. The authors plan to continue research here in the hopes of developing a more complete chronology of dune formation and evolution during the Holocene. Of course one doesn’t have to be a scientist to appreciate the importance and beauty of this landscape.

To stand in the midst of a sea of great dunes is to have the sensation of being utterly alone, and vulnerable, in a world of scorching elements and soul-shaking beauty.

— Jan DeBlieu, 1999

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