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The Late Paleozoic Ice Age in western equatorial Pangea: context for complex interactions among aeolian, alluvial, and shoreface sedimentary environments during the Late Pennsylvanian-early Permian

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Abstract

The aims of this study are to analyse the evolution of depositional environments in the Late Pennsylvanian-early Permian of the Paradox Basin in Utah, USA from the lower Cutler beds to White Rim Sandstones, i.e. the Cutler Group. The study combines detailed sedimentological and high-resolution sequence stratigraphic analyses through time and across space, in order to define a model of landscape evolution, to discuss the stratigraphic model, and to evaluate the significance of the cyclicity in the paleoclimatic context.

High-resolution cycles are observed for the first time throughout lower Cutler beds, Cedar Mesa Sandstone, Organ Rock Shale and Rim Sandstone: 40 genetic units within 15 minor stratigraphic cycles. 40 genetic units within 15 minor stratigraphic cycles integrated in two major cycles.

Three steps of the Paradox Basin landscape evolution are recognized with marine influence present in all early Permian formations. The **first step, i.e. lower Cutler beds,** is mainly characterised by a marine environment, with longshore bar, subtidal, tidal deposits and mouth-bar, by some braided rivers with the preservation of trunk of large trees, and by the development of an aeolian dune field in its upper part. In a **second step**, i.e. Cedar Mesa Sandstone, broad erg deposits are present across the entire study area, indicating more arid conditions, even if some wet episodes occur temporarily, allowing the preservation of large and long root traces; longshore bar, subtidal, mouth-bar, and some fluvial deposits are mainly preserved in the northern part. In a **third step**, i.e. Organ Rock Shale and White Rim Sandstone, decreasing aeolian dune field preservation is observed. Within the southern part, the aeolian environments are interbedded with shoreface deposits, whereas in the northern part, fluvial deposits with some mouth-bars are more developed. As indicated by the presence of calcretes, the semi-arid climatic conditions persisted.

In paleogeographic and Late Paleozoic Ice Age contexts, this early Permian succession reflects both relative sea-level fluctuations and the variability of sediment supply resulting from a changing amount of precipitation in the source area. A scenario is proposed to explain the stratigraphic cycles observed. Six stages considering sea-level and sediment supply variations within glacial-interglacial phases allow us to discuss the stratigraphic surfaces, the variation of depositional environments and vegetation taphonomy.

Keywords

High-resolution sequence stratigraphy; Sea-level variation; Paleoenvironment; Paleoclimate

1. Introduction

Since the early 1990's, the concept of high-resolution sequence stratigraphy, initially defined in marine environments (Jervey, 1988, Posamentier & Vail 1988, Posamentier *et al.* 1988, Van Wagoner *et al.* 1988), has been applied to continental environments (e.g. Shanley & McCabe, 1993, 1994; Wright & Marriott, 1993; Leeder & Stewart, 1996; Catuneanu *et al.*, 2009). Studies on the accumulation and preservation of aeolian deposits have focused mainly on the relationship between dry and wet climatic cycles, as reflected by water-table variations. The development of sequence stratigraphy in aeolian systems involves studying the relationship between water-level, subsidence and change in sediment availability (Yang & Nio, 1993; Kocurek & Havholm, 1993; Havholm et al., 1993; Blakey *et al.*, 1996; Carr-Cabaugh & Kocurek, 1998; Veiga *et al.*, 2002; Mountney, 2006). Several models propose either an interaction between aeolian and fluvial (e.g. Clemmensen *et al.*, 1989; Tirsgaard & Oxnevad, 1998; Sweet, 1999; Mountney & Jagger, 2004; Bourquin *et al.*, 2009; Hême de Lacotte & Mountney, 2022) or aeolian and marine environments (Blanchard *et al.*, 2016). Some models discuss the interaction between all three, aeolian, marine and fluvial deposits (e.g. Rankey 1997; Jordan & Mountney, 2010; 2012; Wakefield & Mountney, 2013 and the pioneering work of Terrell, 1972) and their relation through time and space. Other models at reservoir scale study the interaction between water table and sediment supply changes and dune-field events driven by autogenic mechanisms or allogenic forcing (e.g., Gross et al., 2022; Mountney, 2006).

The Pennsylvanian – Permian Paradox Basin, part of the Colorado Plateau located near the Four Corners area, between Utah, Colorado, New Mexico and Arizona (Fig. 1), is characterized by well-exposed early Permian outcrops. The Permian succession (Fig. 2) has been the focus of many studies since the pioneering works of Wengerd & Matheny (1958), Orgill, (1971), Terrell, 1972), Baars (1962), Loope (1981) and Mack (1984). The lower part (lower Cutler beds) is classically considered to comprise shallow marine, aeolian, and fluvial deposits, and overlying deposits are considered to be exclusively continental (e.g. Soreghan *et al.*, 2002a; Jordan & Mountney, 2010; Wakefield & Mountney, 2013). The middle part is considered as dominated by aeolian deposits (Cedar Mesa Sandstone; e.g. Mountney & Jagger, 2004; Mountney, 2006; Jordan & Mountney, 2010), and the upper part by mainly fluvial deposits of the Organ Rock Formation (Fm; e.g. Moore *et al.*, 2008; Soreghan et al., 2009; Keiser, 2015; Venus *et al.*, 2015). Marine deposits reappear at the top of the succession (White Rim Sandstone; e.g. Steele, 1987; Chan, 1989). This early Permian succession was deposited during the Late Paleozoic Ice Age (LPIA; Fielding *et al.*, 2008; Montañez *et al.*, 2016; Griffis et al., 2022) and is located in the central Pangean equatorial region (see Fig. 1 in Soreghan et al., 2020). In consequence, these early Permian series provide an opportunity to study the variation of depositional environments, vegetation preservation and stratigraphic cycles considering sea-level, tectonics and sediment supply variations within glacial-interglacial phases in the equatorial Pangea.

Based on study of early Permian strata of the Canyonlands area in Utah (Fig. 1), the aim of this paper is to reinterpret the palaeoenvironment evolution in space and time and to define a high-resolution sequence stratigraphy in an equatorial context during a glacial-interglacial period. From a detailed sedimentological study, the specific objectives are to: (1) describe the evolution of depositional environments, (2) define the high-resolution stratigraphic cycles and surfaces, (3) propose stratigraphic correlations and a model of landscape evolution of the early Permian succession, and (4) discuss a stratigraphic model and the significance of cyclicity in the LPIA paleoclimatic context.

2. Geological setting

The Pennsylvanian – Permian Paradox Basin was bounded by three main tectonic uplifts that occurred during the Early Pennsylvanian and formed the Uncompahgre Highland, and Defiance-Zuni and Emery uplifts (Wengerd & Matheny, 1958; Baars, 1962). These uplifted areas are onlapped by Permian strata, hence forming a major unconformity within the Colorado Plateau (Wengerd & Matheny, 1958). The Uncompahgre Uplift (Fig. 1) constituted the highest relief during Pennsylvanian and Permian time and therefore the main source of sediments supplying the basin until the late Leonardian, i.e. late Cisuralian, – early Guadalupian (Wengerd & Matheny, 1958).

In the Paradox Basin (Figs. 1, 2), Permian deposits overlie the Hermosa Group (Grp) except in the northeast part of the basin, where they rest on the Proterozoic basement, and in the west where they unconformably overlie the Mississippian formations (Condon, 1997). Since Loope (1984a) questioned the existence of the unconformity defined by Barrs (1962), the Hermosa Grp is considered to be conformably overlain by a transitional unit between underlying predominantly marine deposits and overlying predominantly continental deposits of the Cutler Group. This transitional unit is named the lower Cutler beds, which represents an informal terminology now widely accepted (see discussion in Jordan & Montney, 2010). Lower - Middle Triassic (Moenkopi Fm) or Upper Triassic (Chinle Fm) strata unconformably overlie the Permian (Fig. 2; Condon, 1997).

In the Colorado Plateau, the Permian deposits are characterised by marine and terrestrial sediments grading into each other laterally and vertically (Condon, 1997). The important work of Baker & Reeside (1929) proposed regional stratigraphic correlations within the Colorado Plateau that are still in use today. For the present study, the nomenclature of Baars (1962) is used for the Canyonland section (Fig. 2), integrating the modifications of Loope (1984).

In the Canyonlands National Park, between Moab and Grabens Districts, the Cutler Grp (Fig. 2) is divided into five formations (Wengerd & Matheny, 1958; Baars, 1962; Baars & Molenaar, 1971). The lower Cutler beds, previously named Rico Fm, Wolfcampian Carbonates, or Elephant Canyon Fm, are considered to be of Late Pennsylvanian to early Cisuralian age (i.e. Missourian to lower Wolfcampian, e.g. Soreghan *et al.*, 2002a, DiMichele *et al.*, 2014). This formation is preserved from southeastern Utah to the San Rafael Swell to the northwest (Fig. 1) (Loope, 1984; Sanderson & Verville, 1990). The Cedar Mesa Sandstone (Fig. 2) is composed almost entirely of sandstones. The Organ Rock Fm is laterally equivalent to the Hermit Shale of the Grand Canyon, attributed to late Wolfcampian to early Leonardian, i.e. Cisuralian, on the basis of vertebrate fragments (Vaugh, 1964), plant remains (White, 1929), and conodonts (Lucas & Henderson, 2021). The White Rim Sandstone overlies the Organ Rock Fm and thins to a feather-edge directly below dead Horse point, and is present further northeast in the subsurface (Baars, 1987). The Undivided Cutler Grp south and west of the Uncompahgre Uplift, north-eastward of Moab (Figs. 1, 2), was considered to be deposited from the Upper Pennsylvanian to the Cisuralian (e.g. Moore *et al.*, 2008). To the northeast of Moab, the Organ Rock Fm disappears and the Cedar Mesa Sandstone is directly overlain by White Rim Sandstone (e.g., Baars & Molenaar, 1971; Loope, 1984)

The lower Cutler beds consist of mixed shallow marine fossiliferous limestones and continental red beds that comprise aeolian and fluvial sediments (Jordan & Mountney, 2010). A varied marine faunal assemblage, including brachiopods, bivalves, crinoids and bryozoans, indicates open marine conditions (Terrell, 1972). For Wakefiled & Mountney (2013), laterally extensive and continuous shallow-marine packages that can be correlated over >5000 km^2 represent the maximum of transgression during the deposition of the lower Cutler beds over a coastal plain of very low relief and seaward dip. In the southern Paradox Basin, Soreghan *et al.* (2002a) interpret this formation as composed predominantly of loessite locally reworked by fluvial and marine processes. The top of lower Cutler beds is defined as a prominent, laterally extensive, marine limestone, named Shafer Limestone, deposited during the last major northeastward marine transgression into the basin, above which lie exclusively non-marine sediments of the Cedar Mesa Sandstone (Mountney & Jagger, 2004; Mountney, 2006; Jordan & Mountney, 2010).

The lower Cutler beds and Cedar Mesa Sandstone are in part composed of large cross-bedded sandstone bodies. The depositional environment for these sandstones is debated. Early authors (Gregory, 1938; Kunkel, 1958; Wengerd & Matheny, 1958; Baars, 1962; Baars & Molenaar, 1971; Mack, 1977, 1979) suggested that they were either aeolian dunes or marine longshore bar deposits based on the presence of large–scale cross beds, convolute bedding, foraminifers, rare crinoid ossicles, or occasional glauconite grains. Baars (1962, 1979) and Mack (1977, 1978, 1979) proposed that the Cedar Mesa Sandstone is, at least in part, shallow littoral marine. However, Loope (1981, 1984) interpreted all the sand dunes of the Rico Fm, i.e. lower Cutler beds and Cedar Mesa Sandstone, composed of well-sorted fine-grained quartz arenite characterized by thick cross-stratified beds, as aeolian dunes. He argued convincingly for a predominantly non-marine aeolian erg and erg margin origin for the Cedar Mesa Sandstone, noting the occurrence of rhizolith horizons as strong evidence for a terrestrial depositional setting, with prevailing paleowinds from northwest to southeast (present day coordinates). However, Baars (1987) suggested that Cedar Mesa Sandstone are coastal deposits, corresponding to a complex environment in which near-shore marine conditions intermingled with subaerial environment. Recently, based on the presence of rhizoliths, vertebrate trackways associated with the aeolian deposits, and the abrupt contact between the carbonate and the sand deposits, Mountney & Jagger (2004) and Mountney (2006) interpreted the Cedar Mesa Sandstone in the Neddles District as wet aeolian sediments under water-table control. This erg system is bounded to the west by a shallow-marine seaway (e.g. Blakey, 1988; 2008; Blakey *et al.*, 1988; Peterson, 1988) (Fig. S1A; Supplementary Data) and transitioned to sabkha deposits toward the southeast (Condon, 1997; Blakey, 1988; Langford & Chan, 1993; Taagart *et al.*, 2010). These sabkhas have recently been interpreted as saline lakes lacking marine influence (Pettigrew *et al.*, 2021).

In an earlier palaeogeographic interpretation of the Organ Rock Fm, Baars (1962, 1975) suggested this formation was deposited on a seaward-sloping coastal plain by streams, on floodplains and tidal flats, possibly with minimal marine reworking. Blakey (1979, 1980) suggested that some of these deposits were reworked by marine water over a broad flat profile. Stanesco *et al.* (2000) interpreted the Organ Rock Fm as a mixed fluvial and aeolian sediments dominated by braided channels in the proximal area (northeast), grading to meandering channels on a low-relief coastal plain, becoming dominated by aeolian deposits in the distal western part. Recently, Cain & Mountney (2009) re-interpreted the Organ Rock Fm as a semi-arid terminal fan system with sheet-floods and aeolian deposits dominating in the southwest distal part.

The lower part of the White Rim Sandstone is composed of sand-sheet or sabkha deposits characterised by algal laminations, wind-ripple strata, and small-scale cross-bedded, bioturbated intervals, breccia layers, adhesion ripples, and desiccation polygons (Huntoon & Chan, 1987; Kamola & Chan, 1988; Chan, 1989; Langford & Chan, 1989). Its upper part is composed of aeolian dunes intermittently flooded by marine water, attested by *Thalassinoides*, *Chondrites* burrows and glauconite grains (Steele, 1987). Its top is reworked by marine processes, indicated by characteristics such as symmetrical ripples, fluid escape features, and the presence of *Ophiomorpha* burrows (Orgill, 1971).

The Undivided Cutler Grp deposits are interpreted as alluvial fans, debris flows and proximal braided-stream deposits (Campbell, 1980; Mack & Rasmussen, 1984; Schultz, 1984; Dubiel *et al.*, 1996) with some occurrence of aeolian deposits (Hême de Lacotte & Mountney, 2022). Sedimentation was influenced by contemporaneous growth of anticlinal structures, such as the Cane Creek anticline (Wengerd & Matheny, 1958) caused by movements of the deep Paradox Salt Fm (Venus *et al.*, 2015). The structures constituted barriers to the coarse sediments and explain the dramatic changes in thickness across the basin of the Undivided Culter Fm (Condon, 1997; Venus *et al.*, 2015). However, Moore *et al.* (2008) suggested that at the time of deposition of the Undivided Cutler Grp, movement along the subsurface Uncompahgre fault had ceased. Moreover, the Undivided Cutler Grp is attributed to lacustrine-fluvial processes in a cold proglacial system (Soreghan *et al.*, 2009; Keiser *et al.*, 2015).

3. Methods

Five detailed sedimentological sections were measured and analysed across the Canyonlands study area (Fig. 1). From north to south these sections are referred to as: Hurrah Pass (355 m; Figs. 3A, 4), Potash Road (382 m; Figs. 3B to D, 5), Lockhart Canyon (270 m; Figs. 3E to G, 6), Elephant Hill (Fig. 7, constituted of 2 sections, one in the recreation area, 122 m, Fig. 3H to J, and the other outside the Park, 215 m, Fig. 3K, L), and Cathedral Butte (225 m, Figs. 3M, 8). Thicknesses were measured using a 1.5 m Jacob's staff. The facies analyses include sedimentological structures, paleoflow measurements of hydraulic and aeolian deposits, i.e. azimuths of foresets of ripples, megaripples and sand dunes, and orientations of the axes of troughs seen in plan-view sections (not corrected for paleogeography), determination of the fossil content (marine fossils, paleoflora, and bioturbation), and were completed by 24 petrographic thin section analyses collected from the five measured sections (Figs. 4 to 8).

A terminology modified from Miall (1996) and Cain & Mountney (2009) is used to describe fluvial and aeolian deposits. From each measured section (Fig. 3) a sedimentological study of a vertical depositional environment profile has been drawn that allows the recognition of genetic units. Architectural data compiled from a series of panels from direct field measurements (around 10 for each section), from tracing of photomontage by classical methods or by drone acquisition in the Lockhart canyon area, show the lateral and vertical extent of the facies, the spatial arrangement of sets and strata and their bounding surfaces. From each section, the recognition of major surfaces, from panels and satellite images (Google Earth), as well as stratigraphic cycles, is used to perform high-resolution correlation at the scale of the Canyonlands and Grabens districts. The correlation uses the principle of high-resolution sequence stratigraphy, based on the stacking pattern of the smallest stratigraphic units (Van Wagoner *et al.*, 1988) within a sedimentary succession (parasequences, Van Wagoner *et al.*, 1990; Mitchum & Van Wagoner, 1991, or genetic units, Cross, 1988; Cross *et al.*, 1993). By observing the stacking arrangement of genetic sequences, bounded by maximum flooding intervals, above and below, different scales of stratigraphic cycle can be identified. With increasing scale and duration, these stratigraphic cycles are termed genetic sequence sets, minor stratigraphic cycles, and major stratigraphic cycles.

To correlate the sections, it is necessary to define a reference surface, i.e. a horizon that represents the shortest interval of time that is recognizable in all 5 sections. In the Lockhart Canyon section, the youngest carbonate (86 m in Fig. 6) at the top of the lower Cutler beds contains nautiloids, *Tainoceras sp.,* and other marine fossils, such as *Wilkingia* sp. and *Bellerophon*, corresponding to an early Permian fossil assemblage, and is attributed to the Shafer Limestone. This level is overlain by a clay facies with abundant shells of *Wilkingia* sp. in life position, located on top of the first cliff at Lockhart Canyon (Fig. 3E) and overlain by a layer with abundant wood fossils. This surface is easy to recognize in the landscape and was used as a correlation surface (on Google Earth images) at the scale of the study area (Supplementary data X). It is our reference level noted as D on the correlation charts (Figs. 3A, B, F, H, 4, 5, 6). For the Elephant Hill section outside of the park, we walked on a benchmark surface, characterized in this part of the basin by another carbonate level locally preserved which allowed the correlation between the two Elephant Hill sections (noted 1 Fig. 7). Between Elephant Hill and Cathedral Butte sections another characterized surface has been drawn from Google Earth images (Supplementary data X). The top of the section is defined by the Triassic unconformity. To the south of the studied area, i.e. Elephant Hill and Cathedrall Butte sections (Figs. 7, 8), the level located below the unconformity is characterised by numerous well-preserved large stump fossilized in situ and logs (Supplementary Data Fig. 2E2, E3 and 3F3) documenting the occurrence of large trees growing here. Immediately above the unconformity, there is an abrupt change in facies associations marked by the presence of braided river and floodplain deposits with the notable absence of aeolian and playa deposits, which are usually developed in the Organ Rock Fm in this part of the basin (Baars, 1962, 1975; Stanesco *et al.*, 2000; Cain & Mountney, 2009, 2011).

4. Facies association analysis

Sandstone facies are predominant within the Permian succession of Utah. They are associated with some conglomeratic and fine facies (Supplementary Data Table 1). These facies are more or less bioturbated (Supplementary Data Fig. 1) or display pedogenic features (Supplementary Data Fig. 2). Facies associations are described below and summarized in Supplementary Data Table 2.

4.1. Facies association E: Erg deposits

4.1.1 Description

This facies association (Fig. 10) is characterised by sandstones facies: AT, Ah, Ar (Table 1). Three facies associations are present (Table 2): E1 (facies AT and PR, Prmo, Fig. 9A, D, F), E2 (facies Ah, Ar; Fig. 9E) and E3 (facies Ar, Aa and PR; Fig. 9F, G). The E1 facies association is mainly composed of AT facies with foreset directions oriented to the southeast (Table 2). Two types of vertical root traces, up to 15 cm in diameter, and several metres long have been observed in the upper portion of some of the dunes: some root traces have preserved wood (named PRmo, Supplementary Data Fig. 2D1), and, in others generally thinner roots, the wood is absent (named PR, Supplementary Data Fig. 2C3). In other cases, the compound coset is capped by a sharp surface that has been colonised by burrowing organisms (Fig. photo à ajouter).

The petrographic analyses of these sandstones (Fig. 13A) show 85 to 90 %, monocrystalline quartz grains, glauconite, fragments of crinoids and foraminifers, and oolites (Fig. 13B). The cement is predominantly calcitic, however in rare instances it is siliceous.

4.1.2. Interpretation

Facies AT, which is the main constituent of the E1 facies association, is interpreted as aeolian dune (Supplementary Data Tables 1, 2) with predominant wind ripples and avalanche features characterising the basal part of the dune (Langford & Chan, 1989; Veiga *et al.*, 2002). Large-scale convolutions are attributed to sand liquefaction during periods with high water table (Allen & Banks, 1972; Doe & Dott, 1980). In consequence, E1 corresponds to aeolian dune migration with paleowinds mostly from the northwest. The vertical root traces with preserved wood (PRmo) observed at the top of aeolian dunes are attributed to plants that have the capacity to produce very large, nearly vertically penetrating roots. At this time in Earth history, the most likely candidates are woody gymnosperms, perhaps some form of cordaitalean, conifer, or dicranophyll. Plants of this type have been reported from xeric, coastal settings of Late Pennsylvanian and early Permian age in New Mexico (Falcon-Lang *et al.*, 2011, 2015, 2016), thus in environments similar to those discussed here. Of course, there are other plant groups reported from water-stressed semi-arid to arid settings of Permian age, in addition to the woody coniferophytes. These include such plants as gigantopterids (e.g. Simon *et al.*, 2018a), supaioids (e.g. White, 1929; DiMichele *et al.*, 2007), comioids (e.g. Mamay *et al.*, 2009), or even certain callipterid peltasperms (e.g. Falcon-Lang *et al.*, 2015). In addition, it must be kept in mind that the parent plants of these roots may have belonged to a yet unknown group. The other root traces (PR) may simply be less mature versions of the larger roots, or could represent some other group of plants capable of tapping into deep water tables. There are no criteria, however, that clearly distinguish these roots, as a different biological taxon from the larger diameter roots. The difference between these two classes of roots may be simply a consequence of sediment built up around a much thinner living root. However, the paleosol above the dunes is rarely preserved and only vertical root-traces indicate that a soil was once present above the sand dunes. The paleosols overlying aeolian dunes attest to a change to more humid conditions that allowed the development of plants such as cordaitaleans, or other drought-tolerant groups.

The aeolian dunes can also be capped by a sharp surface with monospecific burrows attributed to *Diplocraterion*, *Planolites,* or animal bioturbation of indeterminate affinity (Supplementary Data Fig. 1A1, C1). The evidence of colonisation by marine burrowing organisms attests to periodic marine flooding episodes (Supplementary Data Fig. 1A1, C1).

The E2 facies association (Supplementary Data Table 2), consisting mainly of aeolian planar deposits (Ar, Ah facies), characterises either interdunes (Veiga *et al.*, 2002; Kocurek & Nielson, 1986) or aeolian sand-sheets where wind regime conditions and/or sand supply prevent the development of dunes (Kozureck & Nielson, 1986; Trewin, 1993). Facies Ar (Supplementary Data Table 1) is interpreted as migrating wind ripples (Hunter 1977). The aeolian planar deposits of facies Ah (Supplementary Data Table 1) are interpreted as traction deposition by high velocity winds (Hunter, 1977; Clemmensen & Abrahamsen, 1983).

Facies Aa and Ar are interbedded in the E3 facies association (Supplementary Data Table 2). The adhesion structures (facies Aa, Supplementary Data Table 1) are the result of dry, wind-blown sand sticking to a wet or damp surface (Kocurek & Fielder, 1982; Brookfield & Silvestro, 2010). E3 represents alternation of either dry and wet aeolian sand-sheets, or wet aeolian sand-sheets and interdunes (Hunter, 1977; Kocurek & Nielson, 1986). Root traces (PR) are sometimes present at the top of this facies association.

The occurrence of marine fossils together with glauconite within aeolian sandstones attests to reworking of coastal sands by wind as suggested by Loope (1981, 1984). However, for this author, these debris could have been transported over more than 800 km and do not prove an interaction with marine depositional environment (Loope, 1984).

4.2. Facies association F: Fluvial and alluvial plain deposits

4.2.1. Description

This facies association (Fig. 10) is characterised by conglomerates (facies Gm, Gt), sandstones (Sm, St, Sh, Shd, Std, Sr, SF), silty-clay (facies F, Fl) and heterolithic (facies SF, SFl) deposits (Table1) The petrographic analyses of fluvial sandstones show a composition of 75% monocrystalline quartz grains with the remaining 25 % being a combination of biotite, muscovite, plagioclase, microcline, perthitic alkali feldspar, heavy minerals (zircon, tourmaline), and lithic fragments. The cement is predominantly calcitic with some iron oxides (Supplementary Data Fig. XX). Four facies associations are observed: F1, F2, F3 and FE (Supplementary Data Table. 2).

Facies association F1 is characterised by decimetre- to several-metre-thick isolated or stacked channelled sandstones composed of vertical association of facies Gm, Gt, St, Sh, Sm, Sr (Fig. 10C). This vertical association, either complete or truncated, is interbedded with some silt facies (F, SF).

Facies association F2 (Fig. 10C, G) is characterised by tabular beds of conglomerate and sandstone deposited by density flows (Gm, Sm) and tractive currents (Sh, St, Sr) interbedded with heterolithic and silt deposits (SF, SFl, F, Fl). It forms decimetre- to several metre-thick isolated or stacked fining upward sandstone bodies.

Facies association F3 (Fig. 10G) is mainly composed of silt deposits (F, Fl) interbedded with some decimetre sandstones beds (Sm, Sh, Sr, SF, SFl). Within silt facies different pedogenic features are represented either by thin root traces, with or without carbonate concretion (respectively Pr and Prn; Supplementary Data Fig. 2B1, B3; Fig. 10L), or by nodular carbonates (Pn; Supplementary Data Fig. 2A, Fig. 10K). Frequently, the Pn features are located a few decimetres below the Prn root traces. FE facies association (Fig. 10M) is composed of planar bed of Shd, Std, Sh, and Sr sometimes interbedded with aeolian deposits (Ar).

4.2.2. Interpretation

The petrographic analyses indicate basement sources.

F1 suggests deposition in braided rivers, consisting of isolated or stacked channel deposits composed of tractive deposits with mainly 3D megaripple bedding, and unimodal, low dispersive paleocurrent indicators (e.g. Miall, 1996). Vertical evolution into migrating current-ripples and fine facies reflects the end of channel infilling. This facies association comprises mainly southwest-trending paleocurrents (Table 2). Well-exposed trunks of trees, conifers or cordaitaleans (Pw; Supplementary Data Fig. 2F4), are observed interbedded within braided river deposits. The rivers must have been large enough to transport logs without destroying them before fossilization and, due to their large size, a long distance of transport is unlikely (e.g. Wyżga *et al.*, 2017). As a consequence, in accordance with the dynamics of braided systems, the trees must have lived in the braid-plain that could have had interfluves without trees. The occurrence of big trees implies that the water table were at least seasonally shallow. Trees with large trunks were almost certainly growing in areas that experienced wet-dry seasonality with the dry season predominating.

F2 is characteristic of unchannelled streamflows attributed to sheet-floods in the distal braid plain (Miall, 1996), consisting mainly of planar beds deposited by gravity flow and tractive current.

F3 is typical of alluvial plains under semi-arid climatic conditions, with waning floods, overbanks, mud-cracks, and pedogenic carbonate nodules (Hasiotis & Bourke, 2006; Hasiotis et al., 2007).

Within F2 and F3 facies associations, the root traces associated with carbonate nodules (Prn) are interpreted as having formed under arid conditions with intermittent heavy rainfall (e.g. Retallack, 1988; Colson & Cojan, 1996). The irregular beds composed of carbonate nodules, frequently observed below Pr or Prn, result from the mixing of fresh and brine waters, which infiltrates the phreatic zone during periods of high evaporation (Colson & Cojan, 1996).

FE, rarely observed, composed of planar beds, is characterised by deflation lags within stream flood deposits and aeolian deposits (Clemmensen & Abrahamsen, 1983; Langford & Chan, 1989). The lack of both channels and true classical clayey floodplain facies indicates ephemeral floods with numerous channels and a high lateral migration rate within an arid alluvial plain (Bourquin et al., 2009).

4.3. Facies association C: Coastal marine deposits

4.3.1. Description

Facies association C (Fig. 11) is mainly characterised by bioturbated sandstones (Stb, Shb, Sb, Msig, Srw, SFb) or siltstones (Fb, Flb), and comprises some bioclastic carbonates (Cf), sandstones (Stf, Sbf), and conglomerates that occasionally contain marine bioclasts (Gm, Gt). Three facies associations, C1, C2, and C3, are attributed to coastal deposits (Supplementary Data Table 2, Fig. 11N).

C1 is composed of multi-centimetre to several metre-thick, compound or isolated, sandstone and conglomerate bars (Gm, Gt, St, Stb, Sb, Shb) alternating with centimetre- to decimetre-thick heterolithic (Shb, Sb, SFb, SF) and silty (Fb, F) facies. The bars, displaying soft sediment deformation, are either channelled sequences or tabular beds.

A large dispersion of the paleocurrents is observed (mainly towards the southwest and sometimes to the southeast). The facies are bioturbated by animals (Stb, Sb, Shb, SFb, Fb), and vertical root traces (PR, Supplementary Data Fig. 2C2) can occur at the top of the sandstone bars.

Within C2 (Msig, Cf, Gm, Stf, St, Stb, Sbf, Sm, Sb, Sbf, SFl, Flb, Fb) numerous marine macrofossils (Fig. 11K to M) as well as burrowing organism are observed. At the top of sandy limestones root traces are observed (PR, rarely Pr).

C3 is characterised by an alternation of bioturbated sandstones and siltstones mainly with current-ripples and horizontal bedding (Sh, Shb, Sb, Sm, Shb, Srw, Sr, Srb, Stb, SFl, SFb, Flb, Fl, Fb, F). At the top of the sandstones or siltstones, root traces (PR, Pr, Pn, Prn) or desiccation-cracks are observed.

4.3.2. Interpretation

C1 is ascribed to distributary mouth-bars and unconfined flow with paleocurrents in the channelled deposits indicating currents predominantly toward the southwest and occasionally toward the southeast (Supplementary Data Table 2). The bioturbation is attributed to *Diplocraterion* and *Planolites* (Supplementary Data Fig. 1A2, C2) that confirms marine depositional environment. The southeast paleocurrent is interpreted as the result of reworking of the sediment by longshore currents.

C2 characterises sub- to intertidal deposits with storm wash-over deposits. The trough cross-bedding with bipolar paleocurrent directions (or tidal bundles) of the Msig facies is attributed to tidal bars deposited in the subtidal zone or within tidal channels. The occasionally encountered low-angle, horizontally laminated to cross-bedded bioclastic sands and sandy limestones reflect foreshore deposits dominated by storm deposits on a supratidal shelf. The ichnofacies are mainly *Scolocia*-like and *Thalassinoides* (Supplementary Data Fig. 1D, E), rarely *Psilonichnus* (Supplementary Data Fig. 1F).

C3 bioturbated sandstone with an overprint of paleosol development is interpreted to have been deposited at the intersection between the tidal flat and terrestrial environment, corresponding to marsh or coastal plain settings. The indeterminate bioturbation of both Shb facies, with horizontal to subhorizontal lamination and rare fluid escape structures, and Sb facies correspond to sheet-flood deposits formed at relatively shallow depths on the top of mouth-bars (e.g. Pollard *et al.*, 1982; Marshall, 2000; Hinds *et al.*, 2004; Bourquin *et al.*, 2010).

4.4. Facies association S: Subtidal to shoreface/offshore deposits

4.4.1. Description

Facies association S is mainly composed of very fine to medium sand (ST, Sb, Shb, Srb, Srw Sm, Sr; Supplementary Data Table 1), with heterolithic (SFb, SFl; Supplementary Data Table 1), and some fine facies (Fl, Flb, F, Fb; Supplementary Data Table 1). The petrographic analyses of these sandstones (Fig. 13A) show 85 to 90 %, monocrystalline quartz grains. Glauconite, fragments of crinoids and foraminifers, and oolites, also are present. The cement is predominantly calcitic but there are rare samples in which it is dolomitic. Two facies associations are distinguishable, S1 and S2 (Supplementary Data Table 2).

S1 is composed of isolated or compound large-scale dunes frequently intensively bioturbated (ST) grading laterally into bioturbated sands, heterolithic and silty facies

(Fig. 12H). Several metres long root traces are frequently present at the top of compound dunes (PR or PRmo; Supplementary Data Fig. 2D2, D3).

S2, can be up to 8 m in thickness, is composed of fine-grained facies interbedded with bioturbated heterolithic and sandstone facies (Fig. 12H, I), sometimes with root traces (Pr, PR; Supplementary Data Fig. 2B2).

4.4.2. Interpretation

The presence of wave ripples and bioturbation support the interpretation of a marine depositional environment. These marine sandstone facies show more mature material than fluvial deposits with reworked marine-fossils.

S1 (Supplementary Data Table 2) corresponds to offshore bars or detached shoreface dune deposits (Plint, 2010), with a longshore current oriented to the southeast. The organisms responsible for bioturbation are either undetermined or assignable to the ichnogenera *Planolites*, *Palaeophycus,* or rarely crustacean. The vertical root traces of several metres length (PR, PRmo) occasionally observed at the top of longshore dunes, attributable either to cordaitaleans, conifers, or indeterminate vegetation (see above), are indicative of periods of subaerial exposure after the formation of these hydraulic dunes. However, the paleosol above the dunes is rarely preserved and only the vertical root-traces indicate that a soil was once present above the sand dunes. The soil overlying the subaqueous dunes attest to an abrupt transition to terrestrial conditions.

S2 (Supplementary Data Table 2) represents a subtidal environment that evolved vertically to episodic subaerial exposures. It is mainly characterised by current-ripples, oscillatory-ripples, some horizontal bedding, the subaqueous ichnogenera *Planolites* and *Palaeophycus*, and root traces (Pr, PR).

5. Depositional environment and high-resolution stratigraphic cycles

5.1 Early Permian depositional environments

The current study finds that marine influence is a component of all three early Permian formations in the Paradox Basin. Previous studies of these early Permian formations overestimated the occurrence of aeolian dunes and underestimated longshore bar sediments, with obvious and major implications for the reconstruction of the regional landscape. Longshore bars and aeolian dunes were not distinguished because many features are similar in a cursory examination. Additionally, the foreset dip azimuth is similar, both facies being strongly wind-influenced, resulting in a similar southeast dip orientation (Fig. 18). Only detailed facies study allows for the differentiation of these sedimentary bodies, with aeolian dunes having typical grain flow and grain fall strata, and with wind ripples, whereas longshore bars are affected by marine bioturbation in their entire body (not only the top of their preserved foresets) and are characterised by bottomsets interdigitated with silts (Fig. 18). In consequence, a depositional model for the entire Permian succession of the studied area (Fig. 1), i.e. from lower Cutler beds to White Rim Fm, based on the above facies analysis is proposed Figure 14.

5.2 Sequence stratigraphic framework

Using surfaces (supplementary data Google) as anchor horizons, a regional correlation between the five reference sections is possible. The repetition of changing depositional environments through time in the measured section allows to define high resolution stratigraphic cycle, i.e. genetic sequences (Fig. 15) that will be discussed below for each formations. Three types of stratigraphic surfaces can be defined: sand-drift, flooding and maximum flooding surfaces. The criteria for defining allocyclic events are based on the abrupt change of facies across these surfaces and their lateral continuity on a large geographic scale.

5.2.1 Lower Cutler beds

The lower Cutler beds Fm (Figs. 4 to 8) is composed of mixed marine and continental deposits. The shoreface deposits represent the predominant facies association, up to 15 m thick, composed of often well-developed stacked longshore bars. These bars can be intensely bioturbated (Planolites, Palaeophycus, crustacean burrows, or indeterminate bioturbation, Supplementary Data Fig. 1). At the top of these bars, root traces are occasionally observed (Pr, PR). Within the Hurrah Pass section large diameter, long rhizoliths with preserved wood, PRmo, are present, attributable to a tree, likely cordaitaleans or conifers, although, as noted earlier, there certainly are other possibilities (Supplementary Data Fig. 2D2, D3). The coastal facies are characterised by mouth-bars and tidal flats, with various ichnogenre attributed to Psilonichus, Diplocraterion, Planolites, Thalassinoides, Scolocia-like, and *Bichordites*-like (Supplementary Data Fig. 1). The braided rivers have large trunks of trees, probably conifers or cordaitaleans, that were living on the braid-plain where there was at least some kind of wet season. Such trunks are particularly well-exposed in the basal part of the Potash road section (around 6 m, Fig. 4; Supplementary Data Fig. 2F4). In all sections, the aeolian deposits are well-developed in the upper part of lower Cutler beds where 6-m thick stacked aeolian dune or sand-sheet deposits occur. Fossil wood and rhizoliths also occur in the youngest bioclastic deposits that contain marine macrofauna at the top of lower Cutler beds representing the reference level for correlation D (Fig. 3 and Supplementary data). In the Potash road section, it is correlative to the cornaline rhizolith beds (Supplementary Data Fig. 2E1) that overlie bioclastic limestones with macrofauna (at around 163 m, Fig. 5) and are, in turn, overlain by coarse-grained (Gm facies) mouth-bar deposits that contain macrofauna remains. This correlative bed is present at 90 m in the Lockhart Canyon section (Fig. 5) that has abundant silicified wood clasts. This bed is bounded below by a fossiliferous bed containing large bivalves (Wilkingia sp.) and above by mouth-bar deposits. The wood fragments are attributed to coniferophyte branches up to 20 cm in diameter (Supplementary Data Fig. 2F2).

These depositional environments dominated by shoreface and fluvial deposits allow to define the genetic sequence. The maximum flooding episode is located at the base of longshore bar deposits (S1 facies association, Fig. 15A). The progradational phase is characterised by a change from subtidal environments with longshore bar (S1, S2) and tidal flat (C2), coastal plain (C3), and mouth-bar deposits (C1) to fluvial (channelled, F1, or unchanneled streamflows, F2) and alluvial plain (F3) deposits. The paleosol development indicated by root traces is observed within the coastal (C3) and the alluvial (F3) deposits. The retrogradational phase is marked by alluvial plain deposits (F3), dominated by unchanneled deposits (F2) with occasional channelled (F1), overlain by mouth bar (C1) or coastal plain deposits (C3). This sequence is rarely complete, (Fig. 12H) and frequently truncated (Fig. 11N), i.e. the retrogradational phase is rarely recorded. The base of this sequence may also be lacking or truncated, with the maximum flooding episodes having been recorded within mouth-bar and coastal plain deposits. The aeolian deposits (E1, E2 and E3) are superposed on the fluvial deposits (F1, F2, F3 or FE) and terminate the progradational phase (Fig. 15B). From the base of the section to the reference surface D, 14 genetic units have been defined (noted a and 1 to 13, Figs. 1 to 8 and 16). They can be grouped in 5 minor cycles (notes I to V, Figs. 1 to 8 and 16).

These depositional environments have been previously described by Terrell (1972) for the lower Cutler beds, with marine bars and aeolian dunes. This precursory work described nearly cyclic transitional marine to non-marine carbonate and clastic rocks with clastic debris from adjacent Uncompahgre highland and the accompanying retreat of the sea. Jordan & Mountney (2010) defined a depositional model with only shallow marine facies, all the sandstones dunes being described as aeolian dunes. Jordan & Mountney (2012) and Wakefield & Mountney (2013) considered at least 12 marine transgressive-regressive sequences, the fluvial deposits being preserved in incised-valley.

5.2.2. Cedar Mesa Sandstone

The Cedar Mesa Sandstones is classically considered as exclusively continental (Loope, 1981), dominantly of aeolian dune origin, with intervening deposits of dry, damp and wet interdune, restricted lacustrine pond, inland (i.e. continental) sabkha, and fluvial origins (e.g. Pettegrew….). In the studied part of the basin, the Cedar Mesa Sandstone is composed primarily of aeolian dunes, but shoreface deposits were also identified (Figs. 4 to 7). The longshore bars (+ Photo) are less developed than in the lower Cutler beds and do not exceed 9 m in thickness (Figs. 4 to 8). Few fluvial or mouth-bar deposits are observed within the Cedar Mesa Sandstone (+ Photo). Large rhizoliths are preserved on top of the aeolian dunes. In the Hurrah Pass section (Fig. 6), large, long permineralized roots, PRmo, mostly likely attributable to coniferophytes (Supplementary Data Fig. 2D1), are present at the base of this formation. Above the 200-m level (Fig. 6), the rhizoliths lack permineralized wood and are thinner (PR and Pr); there are, therefore, no criteria clearly demonstrating that root size is not simply a consequence of sediment built up around a much thinner root. At Elephant Hill (Figs. 7) and Cathedral Butte (Fig. 8), the root traces are devoid of organics and are simply discolorations of the sediment (Pr, PR).

The genetic sequence type in the Cedar Mesa Sandstone is similar to the one of the upper part of the lower Cutler beds (Fig. 15B). However, the progradational sequence is rarely complete and some facies are missing (Figs. 12I, 15C, D), implying an abrupt transition from longshore bar to aeolian dune facies, occasionally underlain by root traces (Fig. 15E), whereas the retrogradational sequence is rarely preserved (Fig. 15C). In this case, the retrogradational phase is marked by sediments recording aeolian sand-sheet (E3-E2-FE), alluvial (F3, F2, F1, FE), and mouth bar (C1) or coastal environments (C3). The aeolian dune is truncated (MRS) and marked by root traces that indicate the presence of a paleosol that stabilized the dune surface thus also contributing to dune preservation (Fig. 15B). In genetic units lacking retrograde deposits, the retrogradational phase is marked only by an abrupt transition from aeolian dune to longshore bar, either underlain by root traces, especially within the Cedar Mesa Sandstone (Fig. 15E, PHOTO), or, in some cases, by evidence of colonisation by organisms that burrow vertically into the substrate (Fig. 15F, PHOTO). As a consequence, large stacked sandstone bars with the same foreset orientation are preserved, the longshore currents and paleowind having similar orientations towards the southeast. A detailed facies analysis is needed to differentiate these two types of sandstone bars (Fig. 18).

These high-resolution cycles are observed for the first time throughout this formation, and in continuity with the lower Cutler beds. From the surface D, 17 genetic units have been defined (noted a and 14 to 30, Figs. 1 to 8 and 16). They can be grouped in 6 minor cycles (noted VI to XI, Figs. 1 to 8 and 16).

Loope (1984) considered that Cedar Mesa Fm intertongues with marine carbonate only to the northwest and interpreted all the crossbedded sandstone in Canyonland national Park as aeolian in origin. This view contradicted interpretations by earlier authors and especially Mark (1979), who considered longshore and aeolian bars. In this area, Loope (1985) interpreted the thin limestones and mudstones as probably lacustrine in origin, and he considered 40 flat-topped aeolian sand bodies, overlain by rhizoliths and burrows. The further studies proposed high-resolution stratigraphic architecture at reservoir scale (i.e. outcrop) and considered fluvial-aeolian or sabkha-aeolian-fluvial interactions (e.g., Targaart et al, 2010; Mountney 2011; Pettigrew *et al.*, 2021; Hême de Lacotte & Mountney, 2022).

5.2.3. Organ Rock Formation

The Organ Rock Fm is classically considered as mixed fluvial and aeolian and has been described as terminal alluvial fan system (Stanesco et al., 2000). However, our study shows that the Organ Rock Fm (Figs. 4 to 6) is mainly characterised by fluvial and mouth-bar deposits associated with coastal or shoreface environments (+ photos). Aeolian deposits are occasional to the north (i.e. Hurrah Pass, Potasch Road and Lockhart Canyon sections) and more developed to south (Elephant Hill and Cathedral Butte sections). The fluvial deposits consist of southwest oriented braided rivers and unchannelled stream flows (sheet-flood deposits), with a few intercalated floodplain deposits. The mouth-bar paleocurrents indicate a flow direction from the southwest to the southeast. The shoreface deposits are stacked longshore bars, less than 10 m thick (Figs. 4, 6). At Lockhart Canyon (Fig. 5), only the lower part of the Organ Rock Fm is observed, and shoreface deposits are absent. Within this formation, the vertical root traces are similar to those observed in the upper part of the Cedar Mesa Sandstone (long thinner root traces without organic matter preserved).

In the Organ Rock Fm, 8 genetic sequences (noted 31 to 38, Figs. 1 to 8 and 16) are different between the north and the south, where they differ by the absence (15B) or presence (Fig. 15 A) of aeolian deposits, respectively. The main difference with genetic sequences of the lower Cutler beds is in the more developed fluvial deposits (especially at Potash road section), and with those of the Cedar Mesa Fm in the more developed fluvial (Elephant Hill section) or marine deposits (Cathedral Butte section). The genetic units can be grouped in 4 minor cycles (noted XII to XV, Figs. 1 to 8 and 16).

Cain & Mountney (2009) correlated five large-scale finning upward cycles ending by aeolian deposits, below the Triassic unconformity. The upper part of their studied series, with well-developed aeolian facies, could correspond to the White Rim Fm. Moreover, Hême de Lacotte and Mountney (2022) proposed a model of architectural relationships for these fluvial-aeolian deposits.

5.2.4. White Rim Sandstone

The White Rim Sandstone is observed at the top of the sections with around 25 m preserved sediment (Figs. 3D, 5). In this part of the basin, this formation is mainly characterized by fluvial, mouth bar facies and aeolian dunes abruptly and unconformably overlain by Triassic deposits (Fig. 9F). In consequence, only 2 genetic sequences are partially observed in each section (noted 39-40, Figs. 1 to 8 and 16) that characterized the end of the minor cycle XV (Figs. 1 to 8 and 16).

In the northeastern part of the Paradox Basin, NW of Moab, Lowton et al. (2015) also observed two aeolian deposits separated by fluvial environment within White Rim Sandstone. The relatively important thickness of this formation, around 170 m, is attributed to high subsidence rate dues to salt diapirism that occur in this part of the basin (Lowton et al., 2015)

5.2.5 Stratigraphic surfaces

5.2.5.1. Sand-drift surface

The sand-drift surfaces (SDS, Fig. 15B, C, D, E, F) characterise the different types of erosional and depositional contacts between water-lain and wind-driven processes (Clemmensen & Tirsgaard, 1990). SDSs are formed by subaerial removal of previously deposited subaqueous sediments, and form an important bounding surface that represents a significant shift in processes controlling sediment transport. The formation of water-lain deposits predating the accumulation of aeolian deposits, controlled by the water-table position, has previously been documented in other aeolian successions, in coastal marine environments (e.g. Fryberger, 1984; Blakey *et al.*, 1996; Blanchard *et al.*, 2016), or in endoreic basins (e.g. Veiga *et al.*, 2002; Bourquin *et al.*, 2009). After the development of a sand-drift surface, aeolian sand accumulation indicates a period of absence of fluvial sediment supply and is either assumed to express progradation within a terrestrial environment (Bourquin *et al.*, 2009) or the latter part of lowstands system tracts within marine coastal environments (Blanchard *et al.*, 2016). This indicates that the aeolian sediment availability reached a peak during progradation, but the preservation is permitted by the creation of the accommodation space. Therefore, the SDSs are not synchronous surfaces and only preserve a single period during the end of progradation without fluvial sediment supply and the beginning of water table rise, which allows the preservation of aeolian deposits (Fig. 15B to E).

5.2.5.2. Supersurfaces: deflation or flood surfaces

Contact surfaces between aeolian dunes and the overlying lithologies are often sharp. These sharp surfaces are known as supersurfaces (Kocurek, 1988) and result from climatic and relative sea-level changes (e.g. Loope, 1984; Mountney, 2006, Bourquin *et al.*, 2009; Blanchard *et al.*, 2016).

In coastal settings, they represent a period of marine erosion resulting in a deflation surface formed prior to a transgression (Chan & Kocurek, 1988). The deflationary supersurface results from a shutdown of sedimentary source material caused by flooding occurring upwind of the erg margin, or by an increase of vegetation cover, or rise of the water level (Kocurek & Havlom, 1993; Mountney, 2012).

Flooding surfaces are defined by the flooding of aeolian deposits, independent of their size and lateral extent (Clemmensen & Tirsgaard, 1990; Fryberger, 1993; Langford and Chan, 1988). In some instances, the preservation of an aeolian dune can be caused only by pure marine flooding, which also induced the generation of that supersurface (Blanchard *et al.*, 2016).

In our study area, these two types of supersurfaces can be observed. The flooding surface (FS, Fig. 15B) marks an increase in fluvially transported sediment but does not represent a stratigraphic surface at the scale of the genetic unit (Bourquin *et al.*, 2009). In fact, the transition from progradational to retrogradation phases that represent the stratigraphic surface is located at the top of the aeolian deposits. Blanchard *et al.* (2016) consider these surfaces equivalent to the maximum regressive surfaces (MRS; Helland-

Hansen & Martinsen, 1986; Catuneanu *et al.*, 2009). The MRS corresponds to the beginning of deflationary episode during which the dune field was colonised by vegetation (Fig. 15B). In other cases, the aeolian deposits are overlain by shoreface facies or coastal deposits, and the top of the aeolian dune can be colonised by burrowing marine organisms (Fig. 15F). If only root traces are observed at the top of aeolian dunes, the presence of a subsequently eroded paleosol is indicated (Fig. 15E). In this case, the retrogradation phase is not preserved; rather, the MRS and the maximum flooding surface (MFS) represent the same surface (Fig. 15E). To conclude, the top of the aeolian dune field is marked by supersurfaces representing either time hiatuses (marked by root traces, Fig. 15E) or direct flooding episodes (burrowed by organisms, Fig. 15F).

5.2.5.3. Maximum flooding surface

The maximum flooding surface (MFS) is interpreted to be marine flooding surface (e.g., Posamentier, 1988; Van Wagoner et al., 1988). However, such surfaces have been recognized in terrestrial environments without any marine influence (e.g. Legarreta *et al.*, 1993; Olsen, 1995; Currie, 1997; Bourquin *et al.*, 1998, 2009). In this case, these surfaces mark the transition from retrogradational to progradational phases, characterised by extensive floodplain or lake sediments.

Within the studied area, the more distal facies are characterised by shoreface deposits, i.e. longshore bar or subtidal deposits. The MFS is located either at the base of the longshore bar or within subtidal facies. Its lateral equivalent corresponds to well-developed coastal plain deposits (Fig. 15).

5.2.5.4. Sequence boundary

In continental settings, the period of progradation is characterised by low sediment preservation. Firstly, the terrestrial surface over which progradation occurs often is marked by well-developed paleosols. Furthermore, as the progradational phase accelerates, it is marked by an increase of sediment supply associated with fluvial incision, and lag deposits or amalgamated fluvial channel beds (e.g., Bourquin *et al.*, 1998, 2009). These specific deposits can be present or absent. When they are present, their base characterises the sequence boundary surface (SB), which is diachronous across the basin. In this study, with mixed marine and terrestrial depositional environments, the increase of sediment supply is greatest at the base of mouth-bars or at the base of fluvial deposits (Fig. 15A). When aeolian deposits are preserved, the SB is clearly delineated at the base of the fluvial or mouth-bar deposits below the aeolian dunes (Fig. 15B). This SB can coincide with the SDS surface (Fig. 15C, D, E).

6. Landscape reconstruction

For the first time, this study shows the general evolution of Permian environments, i.e. the lower Cutler beds, Cedar Mesa Sandstone, Organ Rock Shale and Rim Sandstone, in the Paradox Basin (Fig. 16). The general evolution of this area allows two major cycles to be defined. From the base of the sections to the MRS of the cycle XI, a major progradational phase took place, marked by the evolution from shoreface deposits to a broad aeolian dune field. The retrogradational phase (retrogradational phases of cycle XI and cycle XIV) is mainly characterised by marine, and alluvial plain deposits, up to a widespread marine episode that defines a major MFS. The second major cycle is characterised by a vertical evolution to mouth-bar and fluvial facies in the northern part and to fluvial and aeolian deposits in the southern part (cycles XV). This progradational phase is truncated by the Triassic base unconformity in this part of the basin.

Landscape evolution of the Paradox Basin, based on the correlations, is presented in Figure 17. Three steps of development are recognized (Fig. 17 A to C).

The **first step** (cycle I to V, i.e. basal part of first major cycle or lower Cutler beds; Figs. 16, 17A), is mainly characterised by a shoreface environment, with longshore bar, braided river, and mouth-bar deposits. The occurrence of large root traces and preserved wood (trunk up to 18 m long at the base of the section) and branches of up to 30 cm in diameter at the top of this first step, transported by braided rivers indicates that trees lived in the drainage basin. A variety of vegetation, likely including trees grew on the braid-plains where ground water levels were at least seasonally shallow during a wet season. The development of an aeolian dune field begins in the upper part of this first step (cycles IV and V, Fig. 16).

In a **second step**, (sequences VI to XI, i.e. upper part of first major cycle, Figs. 16, 17B), broad aeolian dune fields are present across the entire study area (i.e. Cedar Mesa Sandstone, Fig. 16). The dune fields overlie rooted shoreface deposits (surface D). Large and long root traces with preserved wood are also present in the top of the aeolian dunes and are tentatively attributed to woody plant, possibly cordaitaleans or conifers, recognizing that there are other possibilities, including plants we know little or nothing of as macrofossils. This feature indicates episodes of wet conditions during which the area was colonised by vegetation, followed by periods with dry conditions accompanied by the development of aeolian features. The sand of the aeolian dunes is slightly finer than that of the subaqueous dunes. The aeolian dunes contain grains attributable to marine fossil remains, such as foraminifers, that demonstrate remobilization of marine sands.

A **third step** (cycle XII to the top, i.e. retrogradational phase of the first major cycle and progradational phase of the second major cycle; Figs. 16, 17C) is characterised by decreasing aeolian dune field deposits, shoreface environment, and mainly fluvial deposits (i.e. Organ Rock and White Rim Fm). Within the southern part, the aeolian environments are interbedded with shoreface deposits, with few longshore bars, whereas in the northern part, fluvial with some mouth-bar deposits are more developed occasionally overlain by aeolian deposits. These features imply a decrease of wind-driven sand mobilisation and an increase of fluvial input in the basin within the second major progradational phase.

7. Discussion

A high-resolution correlation of the entire Permian stratigraphic succession in the Paradox Basin (Utah) allows discussion of landscape evolution in time and space. The proposed scenario documents an interaction between aeolian dune fields, shoreface, mouth-bar and fluvial deposits (Figs. 16, 17). Previous studies have been focused on one formation and therefore missed the significance of a robust regional analysis of sedimentary filling of the basin.

7.1. Significance of cyclicity

Within the early Permian succession studied herein, as preserved in the Paradox Basin, there are 40 genetic units within 15 minor stratigraphic cycles. The sedimentary record reflects both relative sea-level fluctuations and the variability of sediment supply resulting from the changing amount of precipitation in the source area. The Uncompahgre Highlands were the source area (Fig. 1) from which the derived sediment was deposited primarily on fluvial floodplains and in coastal mouth-bar (Lowton et al., 2021). Tidal influences and longshore bars are less dominant vertically and give way to less marine influenced sediments higher in the succession. Aeolian deposits, more prevalent up section, are more common in the southern part of the study area than in the north where they are interdigitated within fluvial deposits.

For the lower Cutler beds, Jordan & Mountney (2012) and Wakefield & Mountney (2013) considered the hypothesis that both climate and eustasy were interdependent and probably responding to a glacio-eustatic driving mechanism as suggested by (Rankey, 1997; Soreghan et al., 2002). For the Cedar Mesa Sandstone, previous workers (e.g., Loope 1984, 1985; Langford & Chan, 1989; Mountney & Jagger, 2004; Mountney, 2006; Taggart et al., 2010; Pettigrew *et al.*, 2021) did not consider marine influence, thus they envisioned accumulation and preservation induced only by slow rises of the relative water-table due to climate variations driven by Milankovitch-type cyclicity. For the Organ Rock Fm, an evolution from progradational alluvial fan and its downstream passage to aeolian-dominated deposits has been put forward (Stanesco et al., 2000; Cain, 2009; Cain & Mountney, 2009, 2011; Hême de Lacotte & Mountney, 2022). The White Rim Sandstone records a coastal aeolian erg system with marine transgression (Huntoon & Chan, 1987; Kamola & Chan, 1988; Chan, 1989; Langford & Chan, 1989).

Wanless & Shepard (1936) had suggested glacio-eustatic control on late Palaeozoic cyclicity, but the magnitude of each sea-level fluctuation is still subject to many uncertainties (e.g. Rygel *et al.*, 2008). Rygel *et al.* (2008) pointed out that large-scale fluctuations (100-120 m) were probably restricted to the very Late Pennsylvanian-Cisuralian times, when ice sheet extent reached a peak during the LPIA. Recently, Blanchard *et al.* (2016) proposed a sequence stratigraphic framework for mixed aeolian, coastal, and marine environments in the Pennsylvanian of western Pangea and suggested that the accumulation of aeolian dunes took place during the latter part of the lowstands system tract and not during the sea-level fall as initially suggested by Carr-Crabaugh & Dunn (1996).

Two types of fluvial deposits are recognized (Fig. 15B). The first, above the longshore dunes in a progradational setting occurs at the end of sea-level fall, and is overlain by aeolian dune fields; the second, which occur during the sea-level rise, is above the aeolian dune fields, i.e. in a retrogradational setting (Fig. 19). Aeolian erg construction occurs during lowstands system tract when sediment availability is at a maximum being preserved during the early stage of transgression (Fig. 19). Aeolian dune fields are sharply truncated marking an MRS. In the southern part of the basin, where no non-marine facies are recognized, the MRSs and MFSs are combined (Fig. 15E). Two episodes of well-developed paleosols occur (Fig. 19): the first during the regression, at the beginning of the sea-level fall, and the second at the end of lowstands, during acceleration of the transgression.

Loope (1985) considered tabular genetic units (tabular aeolian sand bodies top by plane bedding with abundant roots and burrows), where diastems present much more time than the rock themselves (around 400,000 years). In the equatorial terrestrial Pangea context, Cecil *et al.* (2014) suggest that a full cycle may have been completed in approximately 100,000 years and these shorter cycles are superimposed on longer, 10^6 yr, intervals of global warming and cooling, and those on a still longer-term trend of increasing equatorial aridity. Moreover, some authors also indicate that the cyclothem record precipitation variability on timescales shorter than the 100–400 kyr periodicities (e.g., Olszewski and Patzkowsky, 2003; Tabor and Poulsen, 2008; Soreghan et al., 2014a, b). As the age of the formations are unconstrained and the top is eroded by an unconformity (Fig. 2), it is very difficult to estimate the full duration of the preserved deposits. Knowing that it is possible that some sequences are absent or stacked, only correlation to the northwest with marine deposits would allow a more detailed discussion of the duration of this series.

Based on our observations, the following scenario is proposed to account for the observed genetic units in the coastal environment of the central Pangean equatorial region during Pennsylvanian-early Permian (Fig. 20). This scenario takes into consideration sea-level variation, the climate variation, which controls sediment supply and vegetation cover will be discussed in the next step.

Stage 1: Highstand system tract, maximum sea-level, end of interglacial phase

In coastal environment, the tidal, subtidal and longshore bars are well preserved. The vegetation cover limits the erosion of the hinterland, also reducing the sediment supply into the basin and favoring fluvial aggradation.

Stage 2: End of highstand, beginning of sea-level fall, beginning of glacial phase

Fluvial sediment supply decreases and vegetation fixes alluvial and coastal deposits. In the coastal environment, tidal and coastal deposits dominate the stratigraphic record with paleosol (root traces or carbonate nodules) development in the terrestrial domain. The presence of vegetation and its accompanying roots strongly limits erosion and thus sediment transport.

Stage 3: Lowstands, minimum sea-level, glacial phase

As sea-level falls, the terrestrial surface emerges with high sediment supply favouring fluvial incisions and amalgamated channel belts, while in marine system the mouth-bar deposits prograde. During lowstands, aeolian dunes start to be present but are very little preserved.

Stage 4: End of lowstands, beginning of sea-level rise, transition to interglacial stage

The fluvial sediment supply decreased, limiting fluvial erosion and allowing maximum migration of aeolian dunes. The beginning of sea-level rise allows aeolian dune preservation. This model is in accordance with those of Loope (1985) that considered aeolian preservation during lowstands. It differs from those of Wakefield & Mountney (2013) where all the large scale crossbedded sandstone are considered as aeolian deposits (Fig. XX de Lockart Canyon) subsequently incised by fluvial processes at the beginning of the falling stage.

Stage 5: Beginning of the transgressive system tract (TST), beginning of the interglacial stage

For the aeolian dunes near the seashore, the beginning of the transgression is recorded at the top of some aeolian dunes where marine animal bioturbation (burrows) is observed. Landward, vegetational growth, as indicated by the rhizoliths, stabilizes the arid alluvial plain and aeolian dunes.

Stage 6: TST - interglacial stage

In the coastal environment, the sediments are rarely preserved and mainly characterised by limited aeolian sand sheets and fluvial deposits. Semi-arid conditions are indicated by the presence of calcretes in paleosols and aeolian deposits.

7.3. Palaeoclimate implications

The aeolian deposits are located at the end of lowstands and in consequence at the end of the glacial phase. Strong linkages exist between global climate, ice volume, sea-level and environmental conditions, with global and regional conditions greatly impacting vegetational change in the late Carboniferous and early Permian glacial world (e.g., Cecil *et al.*, 2014; DiMichele, 2014). However, the interpretation of climate differs according to the authors: either tropical climate is considered during glacial intervals as humid, i.e. wetter and less seasonal precipitation (e.g., Cecil et al., 2003, 2014; Eros et al., 2012) or as arid, i.e. drier and more seasonal precipitation (e.g., Rankey, 1997; Olszewski and Patzkowsky, 2003; Soreghan et al., 2008a; Jordan and Mountney, 2012; Wakefield, & Mountney, 2013), than during interglacial intervals.

In the central Pangean equatorial regions, coal and organic-rich shales occur in cyclic sequences, thus Cecil *et al.* (2014) and DiMichele (2014) conclude that the maximum rainfall and the minimum seasonality occurred during glacial maxima and thus sea-level lowstands. In craton interior, coal preservation occurs when equatorial rainfall exceeds evapotranspiration (Cecil *et al.*, 2014). Initiation of ice-sheet build-up progressively increases precipitation and decreases seasonality in the equatorial regions, which resulted in more vegetation (wetland forest) in all topographic areas surrounding a basin (Cecil *et al.*, 2014). During the early transgression (end of glacial phase), climate began a shift to seasonal, subhumid (Cecil et al., 2003) that reduced vegetation density on the landscape (Cecil *et al.*, 2014). As a consequence, in this equatorial coastal domain, the aeolian deposits reflect a reduction of vegetation that implies greater erosion and is probably also related to the onset of a more extensive seasonal migration of the Intertropical Convergence Zone (ITCZ) with higher wind velocities (e.g., Tabor & Poulsen, 2008).

Some climate simulations of tropical climate supported the hypothesis of dryer climate during glacial interval and consider the effects of glaciating the equatorial mountains (e.g., Heavens *et al.*, 2015). For these authors, this interval corresponds to strong winds caused by glaciation of the Central Pangea Mountains that suppressed precipitation over the Central Pangean Mountains and shifted the ITCZ poleward, like the monsoonal circulation (e.g., Heavens et al., 2015). This model considers the prevailing winds would have been laden with moisture evaporated from Palaeo-Tethys and wind from the east during equatorial glaciations, which is not in agreement with the wind direction measured in Paradox Basin (this study and e.g., Loope, Mountney). In western equatorial Pangea and in the Paradox Basin, Soreghan *et al.* (2008) considered loess-paleosols to record sub-100 kyr fluctuations between the drier, dustier glacial periods, and the wetter interglacial intervals of the late Palaeozoic (Soreghan et al., 2014a, b). The paleowind directions in this region are oriented to the south-southwest, reflective of a megamonsoonal circulation formed during the assembly of the Pangea (e.g. Soreghan *et al.*, 2002b). These authors consider a glacial and thus primarily mechanical weathering as origin of the silt. However, the paleowind measurements in aeolian dunes in our study area have the same orientation of those measured by Soreghan *et al.* (2002b) and consequently, this loess, if time equivalent to the aeolian dunes in the coastal domain, could correspond to aeolian sand reworked from the coastal region. Such an origin would therefore reflect monsoonal conditions fostering strong winds, seasonality, and consequent silt mobilization, instead of the existence of uplands and cold-weathering processes at low latitude (e.g. Soreghan *et al.*, 2008). Moreover, Griffis et al. (2023) also suggest a non-glacial origin for the silt. Overall, at this period of the Earth history, i.e. early Permian, in the western Pangean domain, a large quantity of aeolian dust is available and preserved, whereas in the eastern Pangean domain, coal, lacustrine deposits with Gilbert-type deltas and volcanic ashes are recorded (Schneider *et al.*, 2006; Ducassou *et al.,* 2019; Mercuzot *et al.*, 2021). Consequently, aeolian deposits are equivalent to the end of the glacial phase, and the glacial phase, with peat/coal development in the terrestrial domain, could be equivalent to fluvial and mouth-bar deposits in the coastal domain located just above well-developed paleosols, at the beginning of the glacial phase, as described in American Midcontinent during Late Pennsylvanian (e.g. Cecil *et al.*, 2014).

7.3. Provenance and tectonics

The Paradox Basin display bimodal detrital-zircon ages indicating fluvial sediment derived from local basement of Uncompahgre Uplift but the source of Neoproterozoic grains in marine and aeolian sandstones along the marine shoreline remains uncertain (e.g., Lowton *et al.*, 2021). However, for the White Rim Sandstone, Lowton *et al.* (2015) considered that sediment was transported via transcontinental rivers to the western marine margin of Laurentian Pangea from the Appalachian region and abundant sediment was blown southeastward from littoral sources incrementally exposed during Kungurian sea-level drawdown. Rapid transport of voluminous eolian sediment overwhelmed sediment derived from local Uncompahgre sources and resulted in observed compositional changes in eolian relative to fluvial sediment.

In the northeastern part of the Paradox Basin, the sedimentation of the Undivided Cutler Grp is controlled by high subsidence rate with syn-sedimentary fault (e.g. Wengerd & Matheny, 1958; Condon, 1997; Venus *et al.*, 2015; Lowton et al., 2015). However, in the studied area the subsidence rate is considered as relatively constant as proved by the tabular correlation. If the first order control on the depositional environment, i.e. genetic unit and minor cycles, is glacio-eustasy and thus is linked to the climate condition of the equatorial domain, the general evolution of the series from a major progradation to retrogradation and then back to a major progradation could be a phase reflective of another control (Fig. 16). Indeed, the first major progradational-retrogradational pattern seems to be controlled by large-scale sea-level variations, whereas the progradational phase of the second cycle seems to be controlled by sediment supply variation. In fact, the upper part of the succession is marked by an increase in fluvial sediment supply within the same climate context (indicated by the same association of calcretes, paleosol and aeolian deposits southward), which is therefore probably linked with the syn-sedimentary movements recorded in the northeast part of the study area observed at that time in the Undivided Cutler Gp (Venus *et al.*, 2015), i.e. lateral equivalent of Upper Cedar Mesa and Organ Rock Fm (Figs. 2, 16). However, Moore *et al.* (2008) suggest that movement along the subsurface Uncompahgre fault had ceased by then. Moreover, according to Soreghan *et al.* (2009, 2014a) and Keiser *et al.* (2015), the Undivided Cutler Gp is described as a proglacial to periglacial lacustrine system in proximity of marine and paralic facies, which implies cold temperature at low elevations in the Uncompahgre uplift. If this local glaciation were to have taken place it would have required anomalously cold conditions in an equatorial domain, which raises the question of the contemporaneity of this formation with the other formations in the basin.

8. Conclusions

Interactions between aeolian, fluvial and marine environments within the entire Permian succession of the Canyonlands area in Utah have been demonstrated from a detailed facies analysis and high-resolution sequence stratigraphic correlation. Vertical evolution of depositional environments shows, at the base of the early Permian deposits, mixed-shoreface, mouth-bar and braided-rivers deposits with preservation of large trees. This depositional environment evolves vertically with an increasing preservation of aeolian dunes and decreasing fluvial influences with preservation of root traces. The upper part records an increase in fluvial sediment supply in the north, which caused a decrease of aeolian preservation, always within a marine coastal environment. The longshore bar facies become less developed and the mouth-bar deposits dominate the marine sedimentation.

The detailed exposures of high-frequency cycles of erg construction and marine flooding under some fluvial influence in the Permian of Utah provide a valuable analogue for other records, especially within subsurface data. Moreover, in the palaeogeographic and LPIA contexts, a scenario is proposed to explain the stratigraphic cycles, and the variation of depositional environments and vegetation preservation observed in the early Permian, considering sea-level and sediment supply variations within glacial-interglacial phases in the equatorial Pangea. Two types of fluvial deposits are recognized. The first one, above the longshore dunes, appears at the minimum of sea-level fall or beginning of the lowstands, during the glacial phase, and is in a progradational setting, overlain by aeolian dune fields. The second one, very rarely preserved, is above the aeolian dune fields and occurs during the acceleration of the transgression, at the beginning of the interglacial stage. Aeolian environments occur at the end of lowstands, during the beginning of sea-level rise, at the transition to the interglacial stage. Two stages of paleosol preservation are observed. One occurs during initiation of ice build-up, i.e. during sea-level fall. The second occurs during transgression and the beginning of the interglacial stage, which allows aeolian dune preservation. As a consequence, in this equatorial coastal domain, the aeolian deposits reflect a period of a reduction of vegetation that was accompanied by greater erosion. This occurred during the beginning phases of sea-level rise and is probably also related to the onset of a more extensive seasonal migration of the Intertropical Convergence Zone with higher wind velocities.

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Figure captions

Fig. 1. Location of the studied sections in the Paradox Basin and major paleo-reliefs (modified from Condon, 1997).

Fig. 2. Main lithostratigraphic units in Canyonlands area. Nomenclature of Baars (1962) and Loope (1984). See Fig. 1 for location.

Fig. 3. Outcrops of A) Hurrah Pass section; B to D) Potash Road; E to G) Lockhart Canyon section; H to J) Elephant Hill section inside park; K, L) Elephant Hill section outside park; M) Cathedral Butte section; N) Panorama view of the Six Shooter Peak, near the US road 211, showing the Organ Rock Fm pinched out toward the ENE below the Triassic base unconformity. See Fig. 1 for the location.

Fig. 4. Sedimentological section at Hurrah Pass with interpretations as depositional environments and sequence stratigraphy. See Figs. 1 and 3A for the section location.

Fig. 5. Sedimentological section at Potash Road with interpretations as depositional environments and sequence stratigraphy. See Figs. 1 and 3B to D for the section location.

Fig. 6. Sedimentological section at Lockhart Canyon with interpretations as depositional environments and sequence stratigraphy. See Figs. 1 and 3E to G for the section location.

Fig. 7. Sedimentological section at Elephant Hill with interpretations as depositional environments and sequence stratigraphy. See Figs. 1 and 3H to L for the section location.

Fig. 8. Sedimentological section at Cathedral Butte with interpretations as depositional environments and sequence stratigraphy. See Figs. 1 and 3M for the section location.

Fig. 9. Aeolian facies: A) Overview of the facies association E1 showing stacked AT facies, trough cross-bedded aeolian sandstones, Cathedral Butte (from 0 to 20 m, Fig. 8); B) Detail of contorted bedding in the facies AT; C) Detail of the facies Ar (location on photography F):

subcritical climbing translatent stratification, horizontal to low-angle cross-lamination with inverse grading; D) Facies AT: trough cross-bedded sandstones with grainfall lamination and grainflow (avalanche) strata, Elephant Hill; E) Facies Ah, aeolian sand-sheet; F) E1 and E3 facies association of the White Rim Sandstone at the top of Potash Road (Fig. 5) unconformably overlain by Triassic deposits; G) Facies association Aa, adhesion structure. See Supplementary Data Tables 1, 2 for facies and facies association description.

Fig. 10. Alluvial facies: A) Facies St: trough cross-bedded sandstone; B) Facies St: trough cross-bedded sandstone with mud-clasts (white arrows); C) Overview of alluvial facies association, F1, F2: incised channel deposits with erosive basal boundary, Potash Road; D) Facies Sm, massive sandstone; E) Facies Sh facies, horizontal laminated sandstones; F) Facies Sm and Sh with mud-clasts, respectively massive and horizontal laminated sandstones; G) Facies Fl overlain by facies Sm and Sh; H) Facies Sh and F; I) Mudcracks, J) Facies Sr (ripple cross-laminated sandstones) within facies F; K) Facies Sm and F with pedogenetic nodules (Pn) ; L) facies F, siltstones, with root traces (Pr); M) Facies Std and Shd respectively trough cross and horizontal bedding alternating with deflation lag surface (d); Alluvial facies association F1 (A to C), F2 (D to G), F3 (H to L) and FE (M). See Supplementary Data Tables 1, 2 for facies and facies association description.

Fig. 11. Coastal facies: A) facies Stb: trough cross bedded sandstones bioturbated B) mainly by diplocraterion with frequently C) fluid escape structure; D) Facies Sb, bioturbated massive sandstone with *Planolites*; E) Facies Msig, tidal bar; F) Detail of facies Msig showing clay drapes; G) Facies Msig, tidal bar with hearing bones structures; H) Facies Srw, current and wave ripples; I) Heterolithic biourbated facies SFb; J) Facies Cf, sandy limestone or limestone with bioclasts; K) Brachiopod; L) Gastropods; M) Crinoids; N) Overview of the coastal facies association at Elephant Hill section (from 0 to 20 m, Figs. 3H, 7). See Supplementary Data Tables 1, 2 for facies and facies association description.

Fig. 12. Subaqueous facies: A) Facies ST with contorted bedding; B) Detail of the base of the foreset of the facies ST with C) Facies Sb, bioturbated sandstone and D) Facies Srb, bioturbated current ripples; Facies association S2 with detail of E) facies SFb and FlB, of F) Facies Srb, bioturbated ripple cross-laminated sandstone and of G) Facies Fb and Sm; H) Overview of the marine (S1, S2, C3, C1) and alluvial (F1) facies association, Potash Road section (from 45 to 73 m, Figs. 3E, 5); I) Overview of marine (S2, C1 and C2) and aeolian facies (E1 association), Hurrah Pass section (from 105-118m, Figs. 3A, 4). See Supplementary Data Tables 1, 2 for facies and facies association description.

Fig. 13. Thin section showing of composition of A) marine and B) aeolian sandstones. See explanation in the text.

Fig. 14. Schematic representation of the different depositional environments identified in the early Permian sections of the Paradox Basin.

Fig. 15. Genetic units observed in the early Permian sections of the Paradox Basin. See explanation in the text.

Fig. 16. Correlations at the scale of the study area. See Fig. 1 for the section location and Figs. 4 to 8 for detail sedimentological sections

Fig. 17. Landscape evolution of the Paradox basin during early Permian inferred from the sedimentological analyses and correlations of the studied section with a) First step (cycles I to V, Fig. 16); b) Second step (cycles VI to XI, Fig. 16) and c) Third step (cycles XII to top, i.e. Triassic base unconformity, Fig. 16).

Fig. 18. Comparison between aeolian dunes and longshore dunes.

Fig. 19. Stratigraphic cycles in relation with sea-level variations.

Fig. 20. A model of the genetic unit evolution in the coastal environment of the central Pangean equatorial region during Late Pennsylvanian-early Permian glacial world. See explanation in the text and Fig. 19 for the six stages noted 1 to 6.

Supplementary Data captions

Supplementary Data Table 1. Facies description: code facies, lithology, bioturbation and fossil content, sedimentary structures and their interpretation in terms of depositional process.

Supplementary Data Table 2. Facies Associations: code of the facies association, sedimentary architecture and their interpretation in terms of depositional environments. Paleocurrents measured in each facies association are also represented. See Supplementary Data Tables 1, 2, and Fig. 1, respectively for facies, bioturbation and root traces.

Supplementary Data Fig. 1. Bioturbations: A) *Diplocraterion*: A1 top view, A2 section view; B) *Palaeophycus*: B1 top view, B2 section view; C) *Planolites*: C1: top view, C2 section view; D) *Scolicia*: top view); E) *Thalassinoides*: E1 top view, E2 section view; F) *Psilonichnus*; G) *Taenidium*: G1, Hymenopter trackways, G2 Coleopteran trackways.

Supplementary Data Fig. 2. Root traces and wood: A) Calcrete Pn; B) Thin and short root traces Pr, in siltstones (B1) or sandstones (B2), or associated with carbonate nodule and concretion, Prn (B3); C) Long and large root traces usually observed in sandstones (C3) and sometimes in fine facies with carbonate concretion (C1, C2, lower Cutler beds at Elephant Hill); D) Long and large root traces with preserved wood, PRmo at Hurrah Pass, at top of aeolian dunes (D1) or at the top of subaqueous dunes (D2, D3); E) Stump horizon with cornaline fillings at the top of the lower Cutler beds Fm at Potash Road (E1) and stump horizon at the top of the Cedar Mesa Sandstone at Cathedral Butte (E2, E3); F) Fossil wood: at the base of the lower Cutler beds at Potash Road either interbedded within St facies (F1), or 20 m-long trunk (F4), at the top of the bower Cutler beds isolated within fine facies (F2), or at the top of the Cedar Mesa Sandstone at Cathedral Butte (F3).