The role of siliciclastics in carbonate fabric diversity and preservation: A case study from the Neoproterozoic carbonate–siliciclastic Horse Thief Springs Formation, Death Valley

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ABSTRACT

The sedimentary record of the Pahrump Group in Death Valley comprises well-exposed successions of mixed carbonate and siliciclastic deposits. Despite the abundance of studies focussing on the depositional dynamics of mixed carbonate–siliciclastic deposition in the Phanerozoic, the record of similar Proterozoic examples is comparatively sparse. Using high-resolution stratigraphy and microfacies analyses, this study investigates the Tonian Horse Thief Springs Formation within the Pahrump Group of Death Valley, California, in order to propose first-order constraints on the interplay between carbonate and siliciclastic deposition in the early Neoproterozoic. The mixed successions are unlike many previously studied examples interpreted as being caused by glacio-eustatic sea-level changes, thus arguing for a more relevant tectonic influence on siliciclastic deposition. This interpretation is supported by detailed microfacies analyses of siliciclastic-rich dolostones, which show abundant soft-sediment deformation features suggesting sudden pulses of sandstone deposition onto a shallow marine carbonate shelf. The Tonian carbonate factory recovered quickly after being smothered by siliciclastics, particularly due to abundant stromatolite growth. A new relationship between siliciclastic input, carbonate fabric diversity, and carbonate preservation is established based on microfacies analyses, putting forward that siliciclastic deposition had a significant impact on the formation and preservation of later-stage diagenetic dolomite cements, as well as on stromatolite morphology and carbonate fabric diversity within the microbialites. This study shows how repeated siliciclastic incursions had a significant impact on the Proterozoic carbonate factory of the Horse Thief Springs Formation, as well as on diagenetic modification and preservation of shallow marine carbonates, and establishes previously unexplored relationships between carbonate and siliciclastic strata in the Proterozoic.

Keywords Carbonate factory, carbonate microfacies, Death Valley, mixed carbonate–siliciclastic sequences, Neoproterozoic, stromatolites.
INTRODUCTION
Mixed carbonate – siliciclastic depositional sequences are a mainstay in the research and reconstruction of regional basin-wide depositional settings worldwide (Dolan, 1981; Mack & James, 1985; Dorsey & Kidwell, 1999; Tucker, 2003; McNeill et al., 2004; Chiarella et al., 2012; Zeller et al., 2015). Although sedimentary sequences have traditionally been viewed as being either carbonate-dominated or siliciclastic-dominated, extensive deposits that exhibit compositional and strata mixing (sensu Chiarella et al., 2017) at different scales exist worldwide, and are by no means exceptions to an otherwise normal depositional trend (Mount, 1985; Roberts & Doyle, 1987; Ferro et al., 1999; McNeill et al., 2004; Chiarella & Longhitano, 2012; Chiarella et al., 2017). Mixed carbonate – siliciclastic deposits have mainly been studied in the Phanerozoic with respect to their environmental conditions of deposition, the relationships between the two lithological components, the processes that govern the modes and mechanisms of carbonate versus siliciclastic deposition, as well as their potential for hydrocarbon exploration (McNeill et al., 2004; Chiarella et al., 2017). Mixed carbonate – siliciclastic deposits are sensitive records of environmental change throughout Earth’s history due to the different responses of siliciclastics and carbonates towards global and regional climatic change, as well as tectonic activity (Puigdefábregas et al., 1986; Cosovic et al., 2018; Chiarella et al., 2019). It is therefore crucial to understand the first-order principles of both carbonate and siliciclastic deposition of mixed systems in order to arrive at a comprehensive understanding of their role as archives of global change.

Due to the cyclic nature of many Phanerozoic mixed successions, one of the most widely invoked environmental constraints on the deposition of mixed systems is variations in sea level, as well as subordinate local or regional influence of tectonic activity (Dolan, 1981; Walker et al., 1983; Brett & Baird, 1985; Gillespie & Nelson, 1987; Myrow & Landng, 1992; Sanders & Höfling, 2000; Tucker, 2003). Depositional environments of these Phanerozoic and modern systems are extensive shallow marine continental shelves and epeiric seas, where the interplay between carbonate and siliciclastic deposition is related to phases of sea-level highstands and lowstands (Dorsey & Kidwell, 1999; Breda & Preto, 2011; Bourget et al., 2013; Zeller et al., 2015).

The cyclicity of many mixed carbonate – siliciclastic deposits has led authors to interpret their depositional environment in terms of sequence stratigraphy, by defining endmember sequences ascribed to icehouse and greenhouse conditions of sea-level lowstands and highstands, respectively (Holland, 1993; Rankey et al., 1999; Cicero et al., 2010; Bourget et al., 2013; Carlucci et al., 2014). However, more recent work has stressed the need for more complex depositional models that require the consideration of additional factors besides (glacio) eustasy, such as the type and productivity of the carbonate factory, the mode and mechanisms of siliciclastic sediment transport, the hydraulic regime, the morphology and topography of the basin, and water depth (McNeill et al., 2004; Zeller et al., 2015). Despite numerous examples of mixed carbonate – siliciclastic successions from the Mesoproterozoic (Turner, 2009; Bellefroid et al., 2019), the Tonian (Thomson et al., 2014), the Cryogenian (Giddings et al., 2009) and the Ediacaran (Haines, 1988), the Precambrian is still sparse compared to the Phanerozoic. Precambrian siliciclastic sedimentary lithologies and the formation of their structures have well-defined equivalents in the Phanerozoic, and have been interpreted using similar models due to their likely similar depositional mechanisms (Bradshaw, 1988; Tirsgaard, 1996; Eriksson et al., 1998, 2005; Saylor, 2003). However, rates of physical and chemical processes including erosion, transport, deposition, as well as lithification and diagenesis were possibly different (Donaldson et al., 2002).

Although there is increasing evidence that carbonate proliferation in the face of continuous siliciclastic sedimentation is well-known from numerous Phanerozoic examples (e.g. McNeill et al., 2004; Lokier et al., 2009), comparatively little is known on responses of the non-skeletal carbonate factories in the Proterozoic towards siliciclastic input in mixed systems. Neoproterozoic deposits in the Death Valley region have been studied over the past decades in order to unravel the Precambrian carbon cycle, modes of carbonate precipitation, the waxing and waning of global ice ages, and the evolution of multicellular life on Earth (Prave, 1999; Corsetti et al., 2003, 2006; Hurtgen et al., 2004; Harwood & Sumner, 2011; Petterson et al., 2011; MacDonald et al., 2013; Le Heron et al., 2017). This study presents the first comprehensive case study of macroscale and microscale carbonate – siliciclastic mixing processes from the Proterozoic Horse Thief Springs Formation, deposited during a...
time of incipient supercontinent breakup in the lead up to putative, global snowball Earth events (MacDonald et al., 2013; Mahon et al., 2014a; Smith et al., 2016), and highlights the potential of carbonate – siliciclastic deposits to record changing environmental conditions and address depositional dynamics of mixed systems in the Proterozoic.

THE NEOPROTEROZOIC PAHRUMP GROUP

Tectonic setting

The Pahrump Group in Death Valley is an easily accessible and remarkably complete section of Mesoproterozoic and Neoproterozoic deposits, and comprises the Crystal Spring and Horse Thief Springs Formations, followed by the Beck Spring Dolomite and the Kingston Peak Formation (Fig. 1). The structural and tectonic setting of the Pahrump Group has been discussed since the 1970s, when the aulacogen tectonic model applied to sediment-filled troughs into a cratonic platform was introduced by Hoffmann et al. (1974). This model suggested that Pahrump Group strata were deposited as a continuous sequence within an East – West trending extensional failing rift basin (Wright, 1968; Roberts, 1974, 1982). Evidence for the existence of such a fault-bounded basin was primarily stratigraphic, including facies thickness, clast size and palaeocurrent trends that suggested provenance of sedimentation from palaeohighs to the north and south of the basin (Wright et al., 1976; Roberts, 1982; Maud, 1983). Rocks of the Pahrump Group were subjected to various episodes of tectonic deformation from the time of their deposition to the Cenozoic (Wright, 1976). Roberts (1974) suggested that the Crystal Spring Formation was deposited within an intracratonic trough, in which uplift-controlled transgression and regression, as well as erosion and transport of siliciclastic material from northern and southern uplands and controlled sedimentation in the basin. Maud (1983) applied the aulacogen model to the ‘carbonate – terrigenous member’ of the Crystal Spring Formation, suggesting that fluvio-marine siliciclastics were deposited during graben formation, followed by finer siliciclastics and carbonate strata covering the extinct aulacogen rift arm by Upper Crystal Spring Formation mixed carbonate – siliciclastic sequences during a time of reduced tectonic activity. In a revised tectonic model Mahon et al. (2014a) suggested punctuated intervals of tectonic activity and basin development in a series of intracratonic basins during Pahrump Group time, rather than sedimentation within a single, relatively short-lived basin.

The Crystal Spring Formation

The stratigraphic and sedimentological complexity of the Crystal Spring Formation has led to comprehensive subdivisions based on observable changes in lithofacies and diverse sedimentological features. Primary markers for these subdivisions are diabase intrusions and dykes that intrude the Crystal Spring Formation, and have enabled to designate a lower, middle and upper part (Wright, 1968; Roberts, 1974). The lower and middle parts of Crystal Spring Formation are characterized by various siliciclastic deposits including arkosic and feldspathic sandstones (lower part, Roberts, 1976, 1982), mudstones, algal dolomites and chert members (middle part, Wright, 1968; Roberts, 1976), all of which have been intruded by diabase sills. The upper part is characterized by conspicuous intercalations of mixed carbonate and clastic deposits that differ significantly from the middle and lower units of the Crystal Spring Formation. In two seminal works, Maud (1979, 1983) first described this ‘carbonate – terrigenous member’ in detail and subdivided the cyclic sequences into six units, which he called ‘A’ through ‘F’. Revisiting the Crystal Spring Formation Mahon et al. (2014b) referred to the siliciclastic sandstones and mudstones as ‘lower member’, the middle part consisting of algal, siliciclastic dolomite and chert as the ‘middle member’, and the younger cyclic carbonate – clastic sequence above as ‘upper member’. In addition to local changes in composition and texture of the deposits in different sections, Maud (1979, 1983) also noted significant thickness changes in the upper member from several hundreds of metres in the East to just over one hundred metres in the West, suggesting a linkage between source area and basin, and tectonic activity controlling the deposition of siliciclastic – carbonate cycles.

The Horse Thief Springs Formation

Mesoproterozoic and Neoproterozoic Death Valley strata feature several unconformities produced by a series of Neoproterozoic...
Syndepositional tectonic events, which have enabled the definition of several tectonostratigraphic units (Summa, 1993; Prave, 1999; MacDonald et al., 2013). The middle and upper members of the Crystal Spring Formation are separated by an unconformity (Fig. 1B), which was first recognized locally by Maud (1979, 1983), and later mapped regionally by Mbuyi & Prave (1993). The extent of missing time was quantified by Mahon et al. (2014a) using detrital zircons from sandstones directly overlying the unconformity, suggesting a depositional hiatus of at least 300 Ma. The significant time gap and the lithological and stratigraphic differences above and below the unconformity encouraged Mahon et al. (2014b) to separate the ‘carbonate–terrigenous member’ from the Crystal Spring Formation, thus defining the Horse Thief Springs Formation. This Formation comprises the siliciclastic–carbonate units defined by Maud (1979, 1983) from the top of the unconformity to the base of the Beck Spring Dolomite. Mahon et al. (2014b) hypothesized that the unconformity separating the Crystal Spring from the Horse Thief Springs Formation records a period of tectonic quiescence in the western Laurentian margin between 1070 Ma and 780 Ma, and that Horse Thief Springs strata represent extensional tectonism during the onset of the breakup of Rodinia between ca 780 Ma and 740 Ma (Mahon et al., 2014a, b). The Horse Thief Springs Formation has a mostly conformable contact to the Beck Spring Dolomite, which consists predominantly of dolomite and has been interpreted as being deposited in a shallow subtidal to peritidal marine environment (Harwood & Sumner, 2011).

MATERIALS AND METHODS

Stratigraphy, sample collection and petrography

Two sections with 260 m of continuous exposure were selected for logging the Horse Thief Springs Formation. The first section is located in Beck Canyon of the Kingston Range (35°47.478′N, 115°55.574′W), the second section is located in the Alexander Hills (35°46.198′N, 116°07.278′W, Fig. 1A). The two sections are well-exposed in gullies that run from the bottom of the Beck Spring Dolomite to the Crystal Spring Formation in both localities. Sedimentary
logging was conducted from the top downward, starting at the lowermost member of the Beck Spring Dolomite through the underlying Horse Thief Springs Formation until non-exposure precluded further surveying. Beds were logged at a decimetre scale, and carbonate samples were taken in situ at metre to several metre intervals along the entire section. Rock samples were cut with a rock saw and thin sections were prepared in standard (7.4 x 9.7 cm) and large (up to 15 x 10 cm) size for petrographic investigations at the Institute for Geology, University of Vienna. Thin section microscopy was conducted using a Leica DM4500 P polarization microscope (Leica Microsystems, Wetzlar, Germany). Thin section micrographs were taken with a Leica DFC 450 C camera using the Leica Application Suite 4.4.0 software. To discriminate different carbonate minerals, some thin sections were partially stained with a mixture of potassium ferricyanide and Alizarin red solution, dissolved in 0.1% HCl (cf. Dickson, 1966).

**Mineralogy and stable isotopes**

Rock samples were milled to fine powder using a Retsch RS 200 disc vibrating mill (Retsch, Haan, Germany), and powdered samples were prepared for X-ray diffraction. The powdered samples were analysed using a Panalytical PW 3040/60 X’Pert PRO (CuKa radiation, 40 kV, 40 mA, step size 0.0167.5 s per step; Malvern Panalytical, Malvern, UK) diffractometer. The Panalytical software ‘X’Pert High score plus’ was used to interpret diffraction patterns and to estimate relative percentages of carbonate, quartz and siliciclastic components in the samples. Sample powders for stable carbon and oxygen isotope analysis were extracted from sliced and polished rock slabs using a hand-held microdrill at low rotational speed. Care was taken to avoid recrystallized cavity cements, sparry cements and detrital clasts by comparing rock slabs with thin section observations of the rock slab counterparts. Sample powders were reacted with 103% orthophosphoric acid at 100°C with a rock saw and thin sections were prepared in standard (7.4 x 9.7 cm) and large (up to 15 x 10 cm) size for petrographic investigations at the Institute for Geology, University of Vienna. Thin section microscopy was conducted using a Leica DM4500 P polarization microscope (Leica Microsystems, Wetzlar, Germany). Thin section micrographs were taken with a Leica DFC 450 C camera using the Leica Application Suite 4.4.0 software. To discriminate different carbonate minerals, some thin sections were partially stained with a mixture of potassium ferricyanide and Alizarin red solution, dissolved in 0.1% HCl (cf. Dickson, 1966).

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**RESULTS**

**High-resolution stratigraphy of the Horse Thief Springs Formation**

**Lithostratigraphy: Beck Canyon, Kingston Range**

Approximately 260 m of almost continuous exposure is present in the Kingston Range from the base of the Beck Spring Dolomite to the lower part of the Horse Thief Springs Formation, which is terminated abruptly by non-exposure below (Fig. 2). This non-exposure did not allow logging through to the top of the Crystal Spring Formation, and we therefore set the base of the present log at 35°47.392’N, 115°55.584’W. In this study the Horse Thief Springs Formation was divided from bottom to top into four units. Unit 1 is represented by a 20 m package of siliciclastics composed almost entirely of sandstone and siltstone. Sandstones are poorly sorted from fine-grained to very coarse-grained, and comprise sedimentary structures including sigmoidal bedforms, current ripples, possible mud drapes, subtle parallel laminations, fine ripple cross-laminations, scour surfaces and rip-up clasts (Figs 2 and 3A). Individual measurements of cross-bed dips and foresets indicate a flow in a northerly direction. The upper part of the sandstone unit is capped by a fining-upward bed grading from light grey poorly sorted pebble sandstone into very fine-grained maroon sandstone – siltstone interbeds with ripple cross-stratification and bright orange laminated dolostone (Fig. 3B).

This fining-upward graded bed marks the onset of Unit 2a of the Horse Thief Springs Formation at Beck Canyon (Figs 2 and 3B). Unit 2a is capped by a sharp contact by a sequence of granular dolostone with sandstone intervals topped by sets of fining-upward sequences. These sequences feature 20 to 30 cm thick tabular sets of trough cross-laminae, complex geometries including sigmoidal bedforms and current ripples with parallel bedding at the base (Fig. 4A). Siliciclastic lithologies of Unit 2a comprise coarse-grained maroon to lilac
Fig. 2. Lithostratigraphy and mineralogical compositions of dolostones from the studied sections at the Alexander Hills and the Kingston Range, units of Maud (1983) are shown for comparison.
Fig. 3. Units 1 and 2a in the Kingston Range. (A) Massive, cross-bedded sandstone and large rip-up clasts from section 1. (B) Boundary between sections 1 and 2 with a sharp contact between sandstone below and dolostone above.
sandstones forming well-expressed fining-upward cycles into blueish structureless sandstone beds (Fig. 4B). The orange dolostone beds are micritic, crinkly laminated to poorly stratified and structureless, and feature lenses and thin trains of fine-grained sandstone that produce channel-like infill and scours (Fig. 4C), as well as either crinkly or parallel laminations (Fig. 4C and D). Transitions between dolostone and sandstone beds are mostly abrupt (Fig. 4A–C). Dolostones in the middle and upper part of Unit 2a show first occurrences of scattered centimetre-sized chert nodules and layers that follow the wavy and planar lamination. Unit 2a transitions into Unit 2b, which is a 20 m thick sequence of orange sandy dolostone beds that are occasionally interrupted by decimetre to metre-scale fine-grained sandstones and laminated siltstone (Figs 2 and 5). There is a clear trend towards less frequent sandstone/siltstone intervals from Unit 2a to 2b, although the siliciclastic content in these dolostones varies significantly between 10% and 80% (Fig. 2). Domal stromatolites occur either directly on top of the sandstone beds or are overlain by sandstones with an abrupt contact (Fig. 5A). These dolostones exhibit partially sand-rich domains with wavy lamination. Lenses and nodules of whitish, unidentified material occur between crinkly, convoluted and parallel oriented laminations within centimetre-sized, sand-poor sub-beds (Fig. 5B).

Unit 2 follows conformably to Unit 3, which is dominated by metres to tens of metres-thick dolostone beds with very fine-grained, parallel to wavy and crinkly lamination (Fig. 6A). The more massive dolostone beds are interrupted by 20 to 30 cm thick beds of poorly sorted fine-
grained sandstone with coarser grains and occasional granules quickly passing into bedding-parallel trains. Some of these sandstone beds are only 1 to 2 cm thick, and 1 to 20 cm wide, well-developed, delicate domal stromatolites occur directly on top of these sandstone beds (Fig. 6B). The overall siliciclastic content in Unit 3 dolostones decreases from 30 to 10% from bottom to top, whereby non-quartz minerals including feldspars and clays disappear completely towards the top (Fig. 2). In the middle of Unit 3, roughly 100 m from the bottom of Unit 1, a distinct bed of black chert featuring abundant concentric coated grains, most likely silicified...
After 15 m of non-exposure above dark green, feldspar, Unit 1, which is not in the Kingston Range section could be identified from the Kingston Range, is also present here. The top of Unit 3 is capped by a lithological change represented by Unit 4. The bottom part of Unit 4 comprises siliciclastic deposits, followed by an interval of mixed carbonate – siliciclastic intercalated successions that lead up to the base of the Beck Spring Dolomite. The lower part of Unit 4 in the Alexander Hills comprises 10 m of reddish, coarse-grained sandstones with millimetre-sized quartz grains followed by medium-coarse sandstone –

carbonate ooids, can be observed (Figs 2 and 6C). The authors considered this a marker bed within the Horse Thief Springs Formation at Beck Canyon, as well as in the Alexander Hills section. This bed has a sharp contact to finely laminated, light orange to grey dolostone below, and is capped by an oncoidal sandy dolostone bed directly above, which then transitions gradually into a sequence of light grey to buff coloured, stratified and locally wavy laminated, almost 20 m thick dolostone (Fig. 2).

Unit 4 of the Horse Thief Springs Formation comprises almost entirely siliciclastics, starting at the bottom with very coarse-grained sandstones (Unit 4a, Fig. 2) that gradually pass to fine-grained sandstones and siltstone (Fig. 2). This unit is the least complete of the four units in Beck Canyon due to non-exposure (Fig. 2). Upward, the succession consists of a 20 m thick interval of frequently interbedded claystones, siltstones and fine-grained sandstones (Unit 4b). They gradually transition into monotonous, dark green, well-bedded sandstone intercalated with thin, decimetre-thick intervals of fissile reddish-brownish coloured shales followed by a first interval of non-exposure. The upper part of Unit 4b comprises thick siliciclastic deposits with very thin, poorly exposed lenticular dolostones and fine sandstones interbedded with grey-blueish fine sandstones at the base (Fig. 2). The transition of Unit 4 into the Beck Spring Dolomite is not exposed in Beck Canyon. The lowermost exposure of the Beck Spring Dolomite in this section is located at 35°47.478’N, 115°55.574’W. A transitional section from the Horse Thief Springs Formation and the Beck Spring Dolomite was located at 35°47’28.26”N, 115°55’30.78”W, approximately 90 m laterally due NNW from the topmost part of the Beck Canyon log.

Lithostratigraphy: Alexander Hills

The studied outcrop section in the Alexander Hills measures approximately 260 m, and is exposed continuously in a gully from the base of the Beck Spring Dolomite. The same units as in the Kingston Range section could be identified with the exception of Unit 1, which is not exposed in the Alexander Hills. Exposure begins after 15 m of non-exposure above dark green, fine-grained diorite (located at 35°46.254’N, 116°07.426’W), and sediments are composed of 50 m of high-frequency carbonate – siliciclastic interbedded intervals and are designated as part of Unit 2a (Figs 2 and 7A). This unit is dominated by siliciclastic deposits with frequent intercalations of dolostone beds (Fig. 2). The top of Unit 2a comprises two sets of fining-upward sequence packages ranging from several decimetres to 1 to 2 m in thickness that are separated by a thin dolostone bank. The first set features coarse-grained, stratified parallel-bedded clast-supported sandstone at the base that grades gradually into purple, laminated siltstone. Sandstones occasionally feature cross-lamination, as well as trough cross-beds. The second set of fining-upward sequences overlies the dolostone bed by basal coarse-grained, clast-supported greenish to blueish sandstone grading into reddish laminated siltstones. The top of these siliciclastic intervals is capped by sand-rich, olive grey, massive dolostones, which represent the beginning of Unit 2b. Here, there is a gradual transition from higher-frequency (Unit 2a) to lower-frequency (Unit 2b) intercalations of sandstones with sandy dolostones, bearing resemblance to the lithological evolution observed in the Kingston Range section (Fig. 2). Siliciclastic content among sandy dolostones shows little variation among these two units with a homogeneous distribution of around 25 to 30%, whereby 5 to 10% comprise clay minerals and feldspars (Fig. 2). Unit 2a features very coarse, clast-supported brownish-grey sandstones intercalated with orange dolostone beds that comprise dark brown to reddish silt to sandy layers and lenses. Carbonate – siliciclastic sequences in Unit 2b are devoid of fining or coarsening-upward grading sequences, and transitions between carbonates and siliciclastics are almost exclusively sharp. In the carbonate-dominated part in Unit 2b the first occurrence of stromatolites in the Alexander Hills section is observed, as they are absent in Unit 2a below.

Unit 3 in the Alexander Hills comprises mainly dolostones with few siliciclastic intercalations. Dolostones mostly feature parallel laminations or are featureless and massive. The chert marker bed featuring silicified coated grains (Fig. 7B), identified from the Kingston Range, is also present here. The top of Unit 3 is capped by a lithological change represented by Unit 4. The bottom part of Unit 4 comprises siliciclastic deposits, followed by an interval of mixed carbonate – siliciclastic intercalated successions that lead up to the base of the Beck Spring Dolomite. The lower part of Unit 4 in the Alexander Hills comprises 10 m of reddish, coarse-grained sandstones with millimetre-sized quartz grains followed by medium-coarse sandstone –
siltstone intercalations. This package is designated as Unit 4a (Fig. 2), and grades upward into Unit 4b, consisting of 30 m of monotonous finely laminated and bedded grey siltstone that gradually changes colour to orange towards the top. The top of Unit 4b represents a transition to the Beck Spring Dolomite of roughly 40 m of dolostone beds and individual dolostone breccias, intercalated and interrupted by siliciclastic deposits, most of which are siltstones. Transitions between carbonate and siliciclastic lithologies are abrupt with no transitions, except at the

Fig. 7. Outcrops in the Alexander Hills. (A) Unit 1 high-frequency carbonate–clastic sequences, person for scale (ca 1.8 m tall) highlighted by the red circle. (B) Black chert bed from Unit 3, coin for scale. (C) Large domal stromatolite enveloped by chertified laminated sandy dolostone from stromatolite hill, hammer (28 cm) for scale. (D) Horizontally intergrown domal stromatolites from stromatolite hill, hammer for scale. (E) Silicification of fine stromatolite laminae from stromatolite hill, hammer for scale. (F) Silicified coated grain-bearing dolostone from Unit 3, coin (1.5 cm) for scale.
topmost part of Unit 4b, where coarse sandstone grades into grey, well-stratified shales, grading into finely laminated orange dolostone with chert nodules. The topmost part of the Horse Thief Springs Formation comprises a 5 m thick well-stratified siltstone featuring lenses of sandy dolostone, capped by a sharp transition of very dark, almost black dolostone transitioning into sandy dolostones with well-defined lamination. These two latter dolostones comprise the lowermost interval of the Beck Spring Dolomite in the Alexander Hills section.

Stromatolite Hill (35°46'12.27"N, 116°7'25.49"W) is an area where stromatolites crop out extensively, located 80 m to the south-west of the gully that represents a lateral extension of the logged section shown in Fig. 2. The black chert marker bed is also exposed here in situ, as are numerous large domal stromatolites, which are the largest observed in the study area (Fig. 7C and D). The stromatolite domes are several decimetres to half a metre in diameter and composed of alternating centimetre-wide orange and light brownish laminae. Some occur in isolation, yet most of the stromatolite assemblage is arranged in several laterally joined domes (Fig. 7D). Stromatolites are for the most part unaffected by chert (Fig. 7C), yet laminae are locally silicified (Fig. 7E). The position of the stromatolites in the stratigraphic level can be determined due to the presence of the exposed in situ black coated grain-bearing chert bed (Fig. 7F) that was correlated to the exposed section in the gully (Fig. 7B).

**Thin section petrography**

**Sandstones and sandy dolostones**

The typical fabric of sandstones from Unit 1 in the Kingston Range is characterized by irregular grain-size variation and parallel lamination with alternating layers of matrix-supported and grain-supported material (Fig. 8A). Sand grains are mostly monocrystalline quartz, largely subangular to subrounded (Fig. 8B). In the matrix-supported layers the sand grains are not cemented, and held together by alternating light to dark-brown and black fine-grained material, whereby the latter shows fine laminae that run parallel to bedding. Sand grains are partially enclosed by haloes of and unidentified brownish dark material, possibly interstitial clay minerals and/or organic material forming a pseudomatrix around the sand grains (Fig. 8B). Grains in both matrix-supported and grain-supported layers show a chaotic fabric with no preferred orientation. Some of the thin and very dark brown to black layers are devoid of quartz grains. The quartz grains within sandstones and sandy dolostones are devoid of deformational features such as undulose extinction, subgrain development, or intracrystalline microstructures, grain flattening, or pressure solution seams.

Sandy dolostones in Unit 2 have varying contents of quartz sand either dispersedly distributed as grains among the finely crystalline matrix of dolomicrite, or as cohesive, tightly packed clasts and layers that follow the internal lamination (Fig. 9A). Quartz grains within these layers are usually submillimetre in size, yet also some larger chaotically organized angular grains are present within these layers, specifically in Unit 2a (Fig. 9B). In Unit 2b convoluted, clast-supported sand layers within a dolomicrite matrix are present (Fig. 9C), as well as distinct
millimetre-scale fining-upward sequences that are separated by sharp erosive boundaries (Fig. 9D). Above these fining-upward sequences are isolated ripples, growing towards the left and dying out gradually (Fig. 9D). Further observable features are millimetre-sized mud
clasts enclosed by convoluted and distorted layers of sand, and abundant irregular, deformed lamination resembling small-scale load structures (Fig. 9E). Dolomites with less sand content are composed of a dolomicrite mud matrix with randomly distributed millimetre-sized angular quartz grains, some of which feature brecciated and disrupted clasts of lighter and darker lithified mud fragments (Fig. 9F).

Dolostones in Unit 3 are less sandy with only a few samples showing alternating sand-rich and sand-poor layers. Sand grains are chaotically organized, lacking orientation, and are angular to subangular. Small-scale sand lenses and layers form recumbent and wavy dispersed features within the partially disrupted dolomicrite matrix (Fig. 10A). Sand lenses in these dolostones are less thick and less well-packed than those observed in Unit 2. Dolostones devoid of sand exhibit various putative microbially-related features, such as micritic peloids and fragmented laminae (Fig. 10B), diffusive, darker and irregular microcrystalline laminae (Fig. 10C), clayed and peloidal micritic fabric, as well as dark laminae composed of diffuse smaller and larger clots (Fig. 10D). Dolostones within Unit 4 are only exposed in the Alexander Hills and contain little sand, with some exceptions at the top. These dolostones are volumetrically dominated by microcrystalline dolomite matrix dissected by millimetre-thick bands of sub-rounded to well-rounded sand grains (Fig. 10E). In the Alexander Hills, the topmost dolomites just below the Beck Spring Dolomite are brecciated and partially silicified. It is characterized by a matrix of whitish, sparry dolomite containing chips of darker microcrystalline dolomite, as well as submillimetre-sized, angular to rounded dolomite grains (Fig. 10F).

**Stromatolites and coated grains**

Stromatolites from the Kingston Range and the Alexander Hills are both domal with distinct convex-upward laminae of darker and lighter brown microcrystalline dolomite (Fig. 11). Kingston Range stromatolite laminae are almost exclusively micritic with very little to no detrital material or non-carbonate minerals (Fig. 11A). The micritic laminae feature cavernous pockets filled with later-stage dolomite cement. Dolomite cement precipitated between the laminae along the entire width of the domal structure parallel to the micritic laminae, where it occurs as bladed cement growing inward from the micritic laminae into the void space (Fig. 11A). The microcrystalline dolomite matrix of the laminae, as well as at the outer rims of the domal stromatolite are penetrated by darker brown micritic mud forming a cloudy and convoluted texture. Cavity-filling dolomite cement is present in two generations. The first generation shows larger, bladed crystals and flat, rhombic crystal terminations (Fig. 11B). These crystals measure several hundreds of microns in length and feature irregular extinction patterns under crossed-polarized light. Crystal fabric of the first-generation cement is coarse-grained with larger crystals, and the blades are composed of numerous small equant crystals. Second-generation cement features smaller, thin crystals with lower interference colours under crossed-polarized light.

Stromatolites from the Alexander Hills are domal, yet wider and less convex than Kingston Range stromatolites, and consist of alternations of laminae measuring several millimetres in thickness separated by thin, darker brown laminae (Fig. 11C). The thick laminae consist almost entirely of peloidal micritic dolomite within a light brownish to grey sparry dolomite matrix (Fig. 11C and D). The laminae are partly separated and displaced, and occasionally show dark brown to reddish streaks possibly produced by pressure solution (Fig. 11D). Some of the dark laminae are enriched in well-rounded to sub-rounded quartz grains (Fig. 11E). These grains are enveloped by the laminae, whereby the grain boundaries have been replaced by dolomicrite to different degrees (Fig. 11E). Some of these sand grains resemble coated grains (cf. Fig. 12) and have a dissolved or replaced nucleus with well-defined, partly laminated cortices (Fig. 11F). Stromatolites from Alexander Hills are devoid of cavity-filling bladed dolomite cement observed in Kingston Range stromatolites. They have a much higher sand content and occurrence of detrital grains concentrated within specific laminae (Fig. 11E and F). Some of the grains have nuclei composed of opaque black minerals, possibly iron oxides.

Coated grains occur in two horizons in Units 2a and 3 in the Horse Thief Springs Formation, in sandy dolostones, as well as within and in the vicinity of the black chert marker bed (Fig. 2). Coated grains from Unit 3 in the Alexander Hills occur within a dolomicrite matrix and show clearly identifiable crystals with sharp grain boundaries (Fig. 12A). Coated grains form horizontal packages with sharp
boundaries forming hardgrounds (Fig. 12A). The intraclasts or nuclei are replaced by silica, and may have originally been either siliciclastic sand fragments and grains, or carbonaceous clasts. In some instances, the material in the nuclei has been replaced by dolomite cement. The grains...
have faint concentric cortices that are bound around the nucleus, from which a radiating fabric fills the grains towards the outer cortex rim (Fig. 12B). Most of the smaller grains feature one prominent concentric lamina either around the nucleus or located outward towards the rim. In
some coated grains the nucleus has been completely replaced by secondary precipitates, possibly silica, and is not recognizable. In contrast to coated grains from Unit 3, coated grains in dolostones from the top of Unit 2a in the Alexander Hills are much larger, measuring up to 1 mm in diameter (Fig. 12C). These coated grains have clearly identifiable laminae surrounding their nuclei of mostly recrystallized material, probably either quartz grains or dolomite. The large grains float within a matrix of dolomitic spar with partially euhedral crystal boundaries (Fig. 12C).

Mineralogy and stable isotopes

The mineralogical trend in dolostones and sandy dolostones from the Kingston Range and the Alexander Hills from Units 2 and 3 is shown in Fig. 2. X-ray diffraction patterns reveal that the dominant carbonate mineral in dolostones and sandy dolostones is dolomite. Samples from the lowermost Beck Spring Dolomite at both localities, as well as stromatolites and coated grains from the Kingston Range, contain subordinate calcite. Siliciclastic content in dolostones from the Kingston Range varies between 10% and 75%, whereby quartz is the main non-carbonate mineral (Fig. 2). Subordinate silicate minerals are muscovite, chlorite, potassium feldspar, and in fewer cases plagioclase. Beck Spring Dolomite from the Alexander Hills is devoid of calcite, yet contains at least 10% quartz. Dolostone samples transitioning into Unit 4b contain significant siliciclastic components (over 50%), whereby quartz is the dominant mineral with subordinate muscovite, chlorite and potassium feldspar. Unit 3 dolostones from the Alexander Hills contain 10% non-carbonate components on average. Unit 2 dolostones contain 20% siliciclastic material on average.

Stable carbon and oxygen isotope values of dolostone and sandy dolostones from the Horse Thief Springs Formation in the Kingston Range are between $-0.37^{\%}_{oo}$ and $5.72^{\%}_{oo}$ and $-10.8^{\%}_{oo}$ and $0.23^{\%}_{oo}$, respectively. The $\delta^{13}C$ values of Horse Thief Springs dolostones from the Alexander Hills range between $-3.70^{\%}_{oo}$ and $5.08^{\%}_{oo}$, $\delta^{18}O$ values are between $-3.83^{\%}_{oo}$ and $1.78^{\%}_{oo}$. The lowermost unit of the Beck Spring Dolomite in the Kingston Range reveals $\delta^{13}C$ values between $-2.56^{\%}_{oo}$ and $5.01^{\%}_{oo}$ and $\delta^{18}O$ values between $-11.78$ and $0.36^{\%}_{oo}$. In the Alexander Hills, the range of $\delta^{13}C$ values in the lowermost Beck Spring Dolomite is similar to those in the Kingston Range, ranging from $-3.04$ to $5.04^{\%}_{oo}$. All isotope values are given in Table S1.
**A PROTEROZOIC CASE STUDY FOR TECTONICALLY-INDUCED MIXED DEPOSITIONAL SYSTEMS**

Mixed carbonate – siliciclastic deposits are common on shallow water carbonate platforms and shelves, where siliciclastics are introduced by storms or tidal currents, river discharge, gravity flows, or by erosion of hinterland and/or shore-line deposits (Gillespie & Nelson, 1987; Yose & Heller, 1989; Harwood & Sumner, 1987; Dorsey & Kidwell, 1999; McNeill et al., 2004; Chiarella & Longhitano, 2012; Chiarella et al., 2012). These processes are mostly short-term events that may operate continuously on a local scale, and are responsible for facies mixing in nearshore sediments. On a regional scale, mixed successions have been interpreted in terms of either greenhouse versus icehouse conditions producing transgressions and regressions, respectively, tectonic basin uplift or subsidence, or a combination of both (Dolan, 1981; Mack & James, 1985; Gillespie & Nelson, 1987; Kamp & Nelson, 1987; Tucker, 2003; McNeill et al., 2004; Tănăsvuu-Miikeviiene et al., 2009; Cosovic et al., 2018; Chiarella et al., 2019; Uhlein et al., 2019). The Horse Thief Springs Formation records facies that not only pre-date the first of at least two Sturtian glaciations, but also a period of supercontinental rifting leading up to the first putative global glaciation, represented by the Beck Spring Dolomite, interpreted as being deposited in peritidal to subtidal, largely euxinic shallow marine conditions (Harwood & Sumner, 2011; Shuster et al., 2018). Both the Horse Thief Springs Formation and the Beck Spring Dolomite were deposited in a basin that experienced pulses of tectonic activity (Smith et al., 2016), producing an extraordinary example of a mixed carbonate – siliciclastic succession. Basin development and rifting affected most Pahrump Group strata during the Tonian associated with the incipient breakup of Rodinia (Timmons et al., 2005; Li et al., 2008). In the light of these palaeoenvironmental and tectonic reconstructions, lithostratigraphic evidence for possible mechanisms of carbonate – siliciclastic deposition in the Horse Thief Springs Formation will be examined.

**A relative sea-level rise**

New lithostratigraphic and chemostratigraphic data, complemented by microfacies analyses provide a revised classification scheme of four distinct units recognizable among the two studied sections. Unit 1 is only exposed in the Kingston Range, lying above the regional unconformity separating the Crystal Spring Formation below from the Horse Thief Springs Formation above. Because the transition to the Crystal Spring Formation is masked by non-exposure in this locality, it is likely that not the complete Horse Thief Spring Formation is exposed. Lithostratigraphic data suggest that Unit 1 and the lower parts of Unit 2a represent a siliciclastic-dominated, terrestrial environment that experienced short-lived episodes of flooding and carbonate precipitation, whereas Units 2b and 3 indicate a carbonate-dominated marine environment perturbed by repeated pulses of siliciclastic deposition. Sandstones in Unit 1 in the Kingston Range are mostly coarse-grained with subangular to angular clasts arranged in parallel bedded millimetre to centimetre-sized beds (Fig. 8). Most exhibit planar bedding, recurring trough-cross and wavy laminations, high grain-size variations, as well as rare occurrences of fining-upward sequences (Fig. 2). Sandstones from Unit 1 are most likely fluvial with regional dips of trough-cross-beds and foresets indicating flow in a northerly direction, agreeing with observations by Maud (1979) who based his interpretation on documented cross-bedding and petrographic analysis of grain size and composition. Despite the presence of non-exposure below Unit 1, this unit is most likely the equivalent of the bottom half of unit D as defined by Maud (1979, 1983) being siliciclastic, partly conglomeratic, coarse to medium-grained sandstones interpreted as fluvial facies. These grade into Unit 2, interpreted as a gradual transition towards a greater marine influence on deposition with a displacement of siliciclastic sedimentation in favour of shallow marine sandy dolostones culminating in Unit 3 (Fig. 2), reflecting a gradual sea-level rise. Units 2 and 3 likely equate to the top half of Maud’s (1979, 1983) unit D. Sandy dolostones from Units 2a and 2b are mostly composed of dolomite lime mud or dolomicrite, interrupted by pulses of fine to medium-grained sand. Angular and chaotically arranged clasts within the sand lenses and bands, but also free floating within the lime mud (Fig. 9B) indicate that siliciclastic material was transported over short distances and deposited rapidly onto the shallow, near-shore environment where carbonate precipitation occurred. Unit 2a sandy dolostones are almost entirely composed of a matrix of fine-grained dolomicritic mud, into which pulses of sand were episodically deposited (Fig. 9).
Unit 4 siliciclastics, equating to Maud’s (1979, 1983) Unit F, likely represent tectonically-triggered gravity flows producing fining-upward sequences that mark an abrupt end of the shallow water carbonate factory of Unit 3 in both study sites, although this facies boundary is more transitional in the Alexander Hills (Fig. 2). A relative sea-level rise enabled the deposition of shales and siltstones; greenish, grey and reddish shales and siltstones are finely laminated, arguing for low energy, quiet, and possibly deeper water environments below the storm wave base (e.g. Eriksson & Reczko, 1998; Gehling, 2000; Krassay et al., 2000). Severe deformation or faulting is not observed in the equivalent Beck Spring Dolomites in the Kingston Range, suggesting that not all parts of the basin were affected equally by tectonic activity at the same time. Gradual shallowing occurred in Unit 4b, as suggested by the gradual reappearance of sandy dolostones. The transition to the the Beck Spring Dolomite is similar in both localities, albeit the top of Unit 4b in the Kingston Range section is thinner and more transitional in the Alexander Hills (Fig. 2). The lowermost Beck Spring Dolomite exhibits microbial facies and diagenetic features described elsewhere (Lloyd & Corsetti, 2010), arguing for a return to a subtidal environment (Harwood & Sumner, 2011, 2012).

Tectonically-induced soft-sediment deformation in sandy dolostones

Siliciclastic grains within the dolostones are almost devoid of deformational microstructures, such as undulose extinction under cross-polarized light or grain fracturing, and even indicators for burial diagenesis such as pressure solution seams are absent. The observed angular to subrounded siliciclastic grain morphology, as well as their chaotic organization and lack of sorting suggests that the siliciclastic material was transported over short distances at high velocity, deposited rapidly into viscous, un lithified to partly lithified carbonate mud, and preserved with little post-depositional diagenetic modification. The absence of microdeformation in quartz grains argues for synsedimentary soft-sediment deformation, instead of pervasive deformation of the deposits after lithification and burial (Meere et al., 2016). Soft-sediment deformation occurred episodically, during or shortly after lithification of dolomitic mud. This is exemplified by sheared sand lenses that deformed into diffuse, rotated and compact sand bands (Fig. 9C), fragments of mud clasts floating within sand lenses (Fig. 9E), recumbent folding of partly lithified sand layers within the dolomitic mud (Fig. 10A), as well as undulose, anticlinal cusps resembling load or water escape structures (Fig. 9E). Large mud and rip-up clasts are also observable within the topmost Unit 1 sandstones in the Kingston Range (Fig. 3A). Mud clasts have been recognized in different shallow-water and deep-water depositional environments, and have been interpreted as related to tsunami, storm and high energy turbidity deposits (Fralick, 1989; Chan et al., 1994; Oliveira et al., 2011; Li et al., 2017; Hill & Corcoran, 2018). The Unit 1 clasts are highly angular and occur randomly within the host sandstone, suggesting high energy deposition possibly due to gravity flows, slumping or deposition during a storm event (cf. Li et al., 2017). Rotated sand lenses (Fig. 9C) and recumbent folding of millimetre-sized sand layers within the micritic dolostone (Fig. 10A) both resemble ‘galaxy’ structures described from subglacial and debris-flow deposits (Menzies et al., 1997; van der Meer, 1997), yet have also been associated with turbulent flow adjacent to irregularities that are not diagnostic for a specific depositional environment (Phillips, 2006). In the case of rotated sand lenses (Fig. 9C), shearing triggered soft-sediment deformation of compacted sand lenses within liquefied dolomitic mud. These features may be the result of tectonic faulting, brecciation of sandy micritic dolostones, or by shear produced by energetic, turbulent flow associated with sediment gravity flows (e.g. Plint, 2014). Most soft-sediment deformational structures are not visible in outcrop, yet Unit 2a contains convoluted and crinkly intercalated laminae of sandy dolostone and fine-grained sandstones (Fig. 4C). Due to the lack of vertical offset and faulting between these intercalations, deformation must have occurred early, shortly after deposition when the dolomitic mud was still viscous, prior to, or during, lithification. Sandy dolostones lack protruding, liquefaction dykes or hydraulic shattering structures related to seismic deformation (Zhang et al., 2007; Brandes & Winsemann, 2013), as well as conduits produced by rapid water escape (Moretti & Sabato, 2007). Despite the lack of clear indications for seismic deformation, the convoluted bedding of carbonate – siliciclastic intercalations may have been produced by tectonically-triggered rapid deposition of overlying sandstones, similar to
soft-sediment deformation produced by earthquake-triggered slumping or gravity flows (Moretti & Sabato, 2007; Yong et al., 2013).

**Regional correlation and origin of siliciclastics**

The lithostratigraphic logs presented here show that a bed-by-bed correlation between the Kingston Range and Alexander Hills localities is not possible. There is, however, an observable overall trend of coarse siliciclastic deposition from Unit 1, fining-upward to carbonate-dominated facies in Unit 3 towards an abrupt episode of sandstone deposition, followed by an interval of fine-grained siliciclastics leading up to the Beck Spring Dolomite (Fig. 2). Maud (1979) reported an almost bed-by-bed correlation of the carbonate–terrigenous member, and suggested a lateral facies model in which carbonates grade laterally into sandstones, and sandstones into siltstones in an open marine depositional environment. Observations in our studied sections show that highly dynamic and changing environmental conditions make such correlations challenging, particularly due to the frequent lateral changes in the amount and spatial extent of carbonate and siliciclastic deposition within the same sedimentary basin.

All studies that have been conducted so far on the Horse Thief Springs formation have classified and categorized its carbonate–siliciclastic sequences within the framework of repeating cycles that can be correlated at least on the outcrop scale. Depositional cycles have been documented within the Horse Thief Springs formation as sandstone–siltstone–sandstone–dolostone, fining-upward cycles of siliciclastics capped by dolostones (Maud, 1979, 1983), cycles within the algal member typified by stromatolitic dolostones capped by silt and chert (Roberts, 1974), and metre-scale cycles of conglomerates and breccias fining-upward towards sandstones and dolostones (Mahon et al., 2014b). The new lithostratigraphic and petrographic data show that although cyclicity in the Horse Thief Springs Formation may be present in the tens to hundreds of metres scale, it is absent on a smaller scale regarding mixed carbonate–siliciclastic sedimentary packages investigated in our section. This observation provides clues as to the mechanisms of carbonate–siliciclastic mixing processes. Viewed from the outcrop scale, Units 1 to 3 represent a gradual decrease of siliciclastic deposition towards the top in both localities replaced by an increase of carbonate-dominated strata (Fig. 2). There is no observable decimetre or metre-scale symmetrical or asymmetrical cyclicity within strata of Units 1, 3 or 4. The depositional sequence of Units 2a and 2b shows the highest frequency of alternating carbonate and siliciclastic beds characterized by a plethora of sedimentological structures, colours, grain sizes and textures, including stromatolitic, wavy to parallel laminated dolostones, trough-cross and ripple stratified sandstones, and several small-scale fining-upward sequences of sandstones grading into mostly finely laminated siltstones (Figs 2, 5 and 7A). Thin section observations show that specifically in Unit 2a sand-rich dolostones are characterized by millimetre to centimetre-scale fining-upward cycles (Fig. 9D), and that beds within our investigated sections show a high variability with regard to sand content, grain sorting and size distribution (Figs 9 and 10). Although Unit 2a in the Alexander Hills contains three packages of coarse-grained sandstone grading into finely laminated siltstone, these do not show the same general sequence. One is terminated by a thin bed of laminated dolostone, whereas others are capped by coarse sandstones or siltstone (Fig. 2). In the Kingston Range section such fining-upward sequences of sandstones, siltstones and dolostones are absent. Instead, the top of Unit 2a exhibits several fining-upward sandstone beds capped by stromatolitic and coated grain-bearing dolostones. Although these features occur in roughly the same stratigraphic horizons, they are inherently different in their structure and composition. These observations suggest that, although Units 2a and 2b are cyclic with respect to the alternation of carbonate and siliciclastic deposits, they do not represent vertically or laterally traceable sedimentary packages between our two investigated sections that reveal a cyclic nature, yet these packages represent lateral facies variations of the same relative sea-level rise. It is therefore unlikely that the deposition of carbonate and siliciclastic sequences from Units 2a and 2b in both localities was primarily controlled by recurring Milankovitch-type orbitally forced cycles, as has been previously observed in Precambrian sequences (Eriksson et al., 2005; Miall, 2005).

Tectonic uplift and subsidence influences the deposition of mixed sequences in active margins, where abrupt lithological changes result from rapid progression of siliciclastics from an elevated uplift area over the shelf (Dolan, 1981;
Gillespie & Nelson, 1987; Cosovic et al., 2018). The sharp lithological changes observed throughout Units 2 and 3 in the Horse Thief Springs Formation, as well as the lack of periodicity in the mixed carbonate–siliciclastic sequences, argues for a punctuated mixing process on the outcrop scale (cf. Mount, 1984; Chiarella et al., 2017). An important observation regarding a putative tectonically-triggered mixing mechanism is the symmetry of sedimentary sequences. Our observed sections within Horse Thief Springs Strata do not exhibit symmetrical repetitive cycles typical for reciprocal sedimentation in an environment controlled by transgressions and regressions, nor do they show a ‘lowstand carbonate–upper siliciclastic’ architecture that characterizes many Phanerozoic mixed sequences (Catuneanu et al., 2011). The section represents a 200 m thick succession of terrigenous, fluvial siliciclastics followed by continuous deepening from possibly peritidal to subtidal to deep-water sediments, representing an asymmetrical sequence. Such sequences are known from tectonically-controlled mixed carbonate–siliciclastics in which highstand carbonate precipitation coincides with shoreface siliciclastics forming a landward transgression (Saylor, 2003). Rocks of the Pahrump Group including the Horse Thief Springs Formation possibly experienced rift-related extensional tectonics (Mahon et al., 2014b; Smith et al., 2016), similar to foreland basins influenced by uplift and subsidence creating palaeohighs and depocentres that strongly influence the depositional dynamics of mixed systems (e.g. Reis & Suss, 2016). In an active basin that provided accommodation space for sandy dolostones, the major source of siliciclastics was most likely the palaeohigh situated in the south-east to south-west of the basin invoked by Maud (1979). Deposits of Units 1 to 3 represent a more proximal facies with respect to the palaeohigh, which may have acted as an orogenic bulge enabling frequent communication between carbonates and siliciclastics.

A possible alternative mechanism for high-frequency mixing of carbonates and siliciclastics in Units 2a and 2b may have been flexural warping, as suggested by Heller et al. (1988) in two stages: Migration of the flexural bulge forward towards the shoreline promotes the delivery of fluvial and nearshore siliciclastics from streams and rivers onto the shallow carbonate ramp. During periods of tectonic quiescence, erosion of siliciclastics releases the overburden, triggering flexural rebound that creates large unconformities that increase the influx of siliciclastic sediments. Flexural warping may have produced a more inclined topography that initiated siliciclastic sediment deposition from the palaeohigh towards the palaeoshoreline, represented by the massive fluvial sandstones in Unit 1. Subsequent lithospheric rebound would then have flattened the basin gradient, enabling a relative sea-level rise in the basin, progradation of the shoreline, and enhancing accommodation space for carbonate deposition (cf. Heller et al., 1988; Saylor, 2003). The deposition of the high-frequency carbonate–siliciclastic intercalations of Units 2a and 2b would not necessarily have been triggered, but could very well have been enabled by such a tectonic process. However, this model remains at this point speculative for Pahrump Group Strata, because there is no current evidence for deposition within a foreland basin, and future sedimentological and structural geological work may shed additional light on possible mixing processes.

Climate-dependent mixed carbonate–siliciclastic deposition?

The Pahrump Group has been correlated to synchronous strata in the Chuar Group in Arizona, in the Uinta Mountain Group in Utah (USA), and in the Coates Lake Group and Callison Lake dolostone in Canada (Dehler et al., 2005; MacDonald et al., 2013; Strauss et al., 2014). Collectively referred to as ChUMP (Chuar–Uinta–Pahrump) basins, they have been interpreted as being deposited in a large extensional basin within an extensive, climate-sensitive interior seaway producing cyclic deposits in several parts of the basin (Dehler et al., 2010, 2017; Smith et al., 2016). The Horse Thief Springs Formation has been correlated to the Red Pine Shale of the Uinta Mountain Group, and to rock of the Chuar Group using vase-shaped microfossils (Dehler, 2014; Dehler et al., 2017). In the Chuar Group, variability in $\delta^{13}$C values and alternating carbonate-rich and carbonate-poor strata have been interpreted in terms of climate variability, with higher $\delta^{13}$C values in dolostones and increased input of siliciclastics related to wetter climate, and lower $\delta^{13}$C values and decreased occurrence of siliciclastics related to drier climate (Dehler et al., 2005). Hence, a correlation of $\delta^{13}$C values with observed trends in carbonate-rich and carbonate-poor deposits with varying extents of
organic carbon burial during climatic and sea-level fluctuations was established for the Chuar Basin. Similar to the more negative δ13C values obtained from Chuar Group strata, the negative δ13C excursions in the Horse Thief Springs Formation are interpreted as local phenomena related to post-depositional diagenetic modification (cf. Dehler et al., 2005). Such negative 13C excursions are not observed in the Horse Thief Springs Formation due to the different depositional nature between the two localities. Whereas the Chuar Basin strata feature repeating sandstone-capped and dolomite-capped cycles on scales of tens of metres (Dehler et al., 2001, 2005), such cyclicity is absent from our observed sections of the Horse Thief Springs Formation. More importantly, there is no observable carbon isotope variation that can be correlated to lithology throughout Units 1 to 4, suggesting that rates of carbon burial did not vary significantly during deposition.

None of the δ13C values of Horse Thief Springs dolostones investigated in this study are low enough to be interpreted in terms of negative δ13C excursions, which may reach values as low as −6‰ in the Death Valley region (Corsetti & Kaufman, 2003; Bergmann et al., 2011; MacDonald et al., 2013). Negative δ13C excursions within the Horse Thief Springs Formation have been reported previously (Smith et al., 2016), although these values did not exceed −5‰ and could not be associated with local sedimentary processes such as surface exposure or karstification (Kaufman et al., 2006; Knauth & Kennedy, 2009; Smith et al., 2016). The δ13C values of sandy dolostones between +6‰ and −4‰, are in the range of previously reported values from Pahrump Group strata, including the Crystal Spring Formation (Corsetti & Kaufman, 2003) and the Beck Spring Dolomite (Tucker, 1983; Zempolich et al., 1988). Comparing the data from Units 2b and 3, the δ13C curve resembles that of Smith et al. (2016): Our most negative δ13C values from the top of Unit 3 are most likely equivalent to Maud’s (1983) Unit E, and the δ13C plateau represents the top half of Maud’s (1983) Unit D (Smith et al., 2016, Fig. 13). The δ13C composition of seawater has increased steadily from the Late Mesoproterozoic to Neoproterozoic seawater (Bartley & Kah, 2004; Halverson et al., 2005). The low δ13C value of almost −8‰ in the silicified chert horizon is most likely a local phenomenon related to either preserved methane carbon derived from the thermogenic breakdown of organic matter (Bristow et al., 2011), or carbon related to early stages of hydrothermal dolomitization (Hecht et al., 1999).

Given that most samples from Unit 2 and many samples from Unit 3 feature well-preserved primary fabrics that are largely devoid of recrystallization and marine or diagenetic cements, variability in δ18O values in the sandy dolostones may represent primary signals archiving local to regional influx of land-derived meteoric waters to the shallow carbonate ramp. Dolomitization of primary calcite or aragonite in sandy dolostones via marine, meteoric or freshwater would require large volumes of water to be flushed through the sediments during diagenesis, and would likely have homogenized the oxygen isotopic composition across the basin (Banner & Hanson, 1990; Kah, 2000; Lloyd & Corsetti, 2010), instead of producing the bed-to-bed δ18O heterogeneities that vary over several permil (Fig. 13). The δ18O values of dolostones from Units 2 and 3 in the Kingston Range are on average higher than in the Alexander Hills (Fig. 13), which may argue for regional variations in the hydrological cycle within the basin during the time of deposition (Frank & Lyons, 2000), or that Kingston Range shallow water carbonates received more input from meteoric water in a more proximal part of the basin. Overall, the Horse Thief Springs Formation represents an interval of isotopic stasis, despite the occurrence of slight local discrepancies between the two studied localities. The water masses in both localities during Horse Thief Springs time were indistinct, yet the gradual shift towards higher δ13C values across the Horse Thief Springs – Beck Spring Dolomite suggests a water mass evolution within an intercontinental epiric sea on the southern margin of Laurentia approaching the Islay anomaly at the top of the Beck Spring dolomite and the first Cryogenian glaciation (Strauss et al., 2014; Smith et al., 2016). Given the fairly homogeneous carbon isotopic compositions of dolostones through Units 2 and 3, it is likely that the southern margin of Laurentia represented a large homogenized water mass instead of a restricted or isolated basin that would have allowed for larger fluctuations of the geochemical environment (Gilleaudeau & Kah, 2013).
Fig. 13. Chemostratigraphy of the Horse Thief Springs Formation. $\delta^{13}C$ and $\delta^{18}O$ values (permil V-PDB) of sandy dolostones, as well as lowermost dolostones from the Beck Spring Dolomite.
The nature of the carbonate factory and its influence on mixing with siliciclastic deposits is a critical factor in controlling the depositional dynamics of mixed systems (James et al., 1992; Cherchi et al., 2000; Gläser & Betzler, 2002; Halfar et al., 2004). Carbonate production in most Phanerozoic and modern mixed carbonate – siliciclastic systems is governed by biologically-controlled, skeletal carbonates deposited on extensive shallow marine continental shelves and epeiric seas as coral reefs and carbonate platforms (Hallock & Schlager, 1986; Schlager, 2003). Here, the depositional dynamics between carbonate and siliciclastic deposition are governed by sea-level change that controls accommodation space for carbonate production. Many of these mixed systems enact significant environmental stress on carbonate-secreting organisms by influencing ecological parameters including turbidity, light availability, salinity and nutrient availability (Wilson & Lokier, 2002; Pomar & Kendall, 2008; Lokier et al., 2009). Although the large-scale processes enabling carbonate accumulation seem to have changed little over geological time (Grotzinger & James, 2000), the change in style of carbonate precipitation through the Proterozoic to the Phanerozoic may have also affected the depositional controls on Proterozoic mixed clastic–carbonate sequences.

Although mixed carbonate – siliciclastic sequences occur throughout Earth history in a variety of depositional settings where they represent a spatial or temporal continuum, significant carbonate production on tropical, eutrophic shelves was for a long time thought to be

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**Fig. 14.** Stable carbon and oxygen isotope cross-plot of sandy dolostones (cf. Fig. 13) from the Kingston Range, the Alexander Hills, as well as data from the lowermost Beck Spring Dolomite, values in permil V-PDB.
drastically hampered or stifled completely by siliciclastic sedimentation (Walker et al., 1983; Hallock & Schlager, 1986; Smosna & Patchen, 1991; Ferro et al., 1999). The Horse Thief Springs Formation is a Proterozoic example of persistent carbonate (re-)proliferation in a favourable environment for carbonate precipitation. Due to the absence of carbonate-secreting metazoans in the Proterozoic, carbonate supply depended less on nutrient availability, turbidity, sea-surface temperature, salinity and light availability (Lees, 1975; Nelson et al., 1983; Lokier et al., 2009). Instead, the carbonate factory in Proterozoic mixed successions would have depended more on physicochemical parameters, such as the saturation state of seawater with respect to carbonate minerals, as well as on physical constraints including basin and shelf morphology, hydrodynamic energy and precursor topography (Testa & Bosence, 1998; McNeill et al., 2004). Mixed carbonate – siliciclastic systems provide highly sensitive records of environmental change because both types of lithologies react differently towards the sea-level change, tectonic events and climatic perturbations (Pigram et al., 1989; Testa & Bosence, 1998; Saylor, 2003; Chiarella et al., 2017, 2019).

In the Horse Thief Springs Formation, the type of mixing is temporal (sensu Budd & Harris, 1990) producing strata mixing (sensu Chiarella et al., 2017), meaning that siliciclastics and carbonate alternate through time as opposed to contemporaneous or contiguous environments producing a compositional mixing (Chiarella et al., 2017). The recovery of carbonate deposition after pulses of siliciclastic sedimentation was very rapid, as suggested by sharp transitions from siliciclastics to carbonates throughout Units 2 and 3 (Figs 2, 3B and 4A to C). Although the carbonate accumulation rate can only be estimated, the Neoproterozoic carbonate factory on the southern Laurentian margin allowed for unrestricted carbonate sedimentation by three main producing agents, stromatolites, coated grains and abiogenic carbonate precipitation from seawater as described below.

**Stromatolites and coated grains**

Stromatolites from the Horse Thief Springs Formation have previously been identified as *Bacalia* and *Conophyton* (Cloud & Semikhhatov, 1969; Howell, 1971; Maud, 1979, 1983). All stromatolites from the Horse Thief Springs Formation form wide domes several centimetres to decimetres in width, resembling Mesoproterozoic to Neoproterozoic *Bacalia Lacera* described from Late Riphean shallow water platform carbonates (Knoll & Semikhhatov, 1998). Variability in stromatolite morphology indicates changing environmental conditions related to water depth, wave energy, flow energy and turbulence, and rates of sediment supply (Hofmann, 1976; Horodyski, 1977; Grotzinger, 1989; Andres & Reid, 2006). Kah et al. (2009) observed variable morphologies of *Conophyton, Jacutopyton* and *Bacialia* in Mesoproterozoic parasequences, interpreting these variations in terms of changing platform geometry, sea-level change and variation in accommodation space. Well-developed stromatolite morphologies in the Horse Thief Springs Formation only include domal to mound-shaped forms besides the recurring stromatolitic parallel, wavy, to crinkly laminations in dolostones throughout Units 2 and 3. This low morphological diversity agrees with the interpreted gradual sea-level change with a concomitant reduction in sediment supply and increasing water depth from Unit 2 to 3. Although the absence of branching, columnar, or bulbous stromatolites species suggests little observable morphological diversity in outcrop, microfacies analyses do provide some evidence for variety among Horse Thief Springs stromatolites. Examples from Unit 3 in the Kingston Range are composed largely of dolomicrite featuring later-stage dolomite cements and lack detrital components (Fig. 11A and B); stromatolites from the Alexander Hills exhibit more variable microfacies. The large, domal, mound-shaped examples feature detritus-rich laminae measuring up to 1 mm in width that are replete in siliciclastic grains, which are partially enveloped by laminae (Fig. 11E and F). Detrital-rich laminae are restricted to flatter, mound-shaped stromatolites, which may indicate an adaptation to a nearshore shallow water environment prone to recurring pulses of siliciclastic sediment. Domes with low convexity provide a large surface area facing upward towards the sea surface that allow for trapping and binding of land-derived input of siliciclastic material providing nutrients for microbial communities. The proliferation of microbial communities such as stromatolites may even have been aided by siliciclastic input by providing templates and substrate for growth aside from delivering nutrients (Andres & Reid, 2006). This is evident from numerous examples of stromatolites that grew directly on top of medium to coarse-grained
sandstones, and incorporated siliciclastic particles and sand grains into their laminae for stability (Fig. 6B).

Although stromatolites re-occur throughout Units 2 and 3 in both localities, they do not form massive metre-scale reefal facies with complex morphological thrombolitic to stromatolitic varieties as observed in other Proterozoic shallow water carbonate platforms (Bekker & Eriksson, 2003; Corkeron & George, 2003; Kah et al., 2009; Winterleitner et al., 2015). The global scale of shallow marine carbonate precipitation has varied little over the course of Earth history (Grotzinger & James, 2000), suggesting a constantly productive shallow water carbonate factory despite dramatic changes in sources for carbonate sediment. Seawater dissolved inorganic carbon is thought to have been significantly higher in the Proterozoic than in the Phanerozoic (Higgins et al., 2009), facilitating the spontaneous precipitation of carbonate minerals from seawater. Dissolved oxygen was low in the Neoproterozoic oceans from which microbial Beck Spring dolomites precipitated, and were likely sulphidic even in the photic zone (Shuster et al., 2018). Reducing, sulphidic, ferruginous and/or organic-rich oceans with higher Mg/Ca ratios than in today’s oceans have been invoked in order to explain the abundance of primary Mg-rich carbonate minerals in Neoproterozoic oceans (Hood & Wallace, 2018).

Analogous to microbial carbonate precipitation from the Phanerozoic to today, microbial metabolisms including sulphate reduction and methanogenesis may have governed the precipitation of dolomite directly from seawater either as a passive organomineralization process, or by forming microbial templates acting as kinetic catalysts for dolomite precipitation (Roberts et al., 2004; Kenward et al., 2009; Krause et al., 2012). The most likely metabolic process supporting the carbonate factory in the Horse Thief Springs Formation is saturation of carbonate minerals in ambient waters via photosynthesis (cf. Dupraz et al., 2009), which is thought to have been carried out by cyanobacteria since the early Archean (Butterfield, 2015; Schirrmieiser et al., 2016). Given the favourable conditions for abiotic carbonate precipitation in Neoproterozoic seawater, the role of microbial photosynthesizing communities in regional-scale carbonate accumulation on shallow shelves and platforms either by photosynthesis or by providing nucleation templates was most likely less relevant than in the Phanerozoic. However, physical processes such as trapping and binding, or providing nucleation templates for mineralization catalysis, may have been more relevant for carbonate accretion in mixed sequences, where the omnipresence of siliciclastics may have stimulated and accelerated microbial carbonate precipitation.

Coated grains are recurring features that occur frequently in the Pahrump Group and younger Death Valley strata up to the Precambrian – Cambrian Boundary, and most have been classified as ooids, giant ooids or oncoids (Marian & Osborne, 1992; Corsetti & Kaufman, 2003; Corsetti et al., 2006; Trower & Grotzinger, 2010; MacDonald et al., 2013). Coated grains from the Horse Thief Springs Formation show little resemblance to the above reported ooids. In contrast to many Proterozoic ooids of several millimetres in size (Sumner & Grotzinger, 1993; Tang et al., 2015; Thorie et al., 2018), coated grains observed here are smaller. Coated grains from the Alexander Hills mostly resemble ooids in shape, size, and the presence of nuclei and cortices (Fig. 12A to C). Grains from Unit 2a are smaller and have the most clearly developed cortices surrounding nuclei of mostly quartz grains. Ooid-like coated grains have been interpreted as being formed in agitated, shallow waters based on their uniform grain shape and size (Corsetti et al., 2006; Trower et al., 2017). In this regard, coated grains in Unit 2a may be interpreted as shallow water, low-energy precipitates forming hardgrounds in the case of irregularly laminated ooids (Fig. 12A and B). Coated grains from Unit 3 (Fig. 12C) are more concentric with visible laminae and dense fabric, suggesting shallow water and more agitated, higher-energy conditions (cf. Perry, 1983; Perry, 1999; Pederson et al., 2015). However, coated grains have been reported from a variety of environments including lakes, shallow water carbonate platforms and hardgrounds, as well as deep water environments (Peryt, 1983), many of which possibly have a microbial origin (Verdrine et al., 2007; Davaud & Girardclos, 2010; Duguid et al., 2010; Pacton et al., 2012; Diaz et al., 2017). Shape, size and fabric of coated grains from Unit 3 may argue for prolonged favourable shallow water, agitated, supersaturated waters producing larger ooid-like coated grains. To summarize, the carbonate factory in the Horse Thief Springs Formation recovered quickly, even after large pulses of rapid clastic sedimentation onto the shallow shelf, most likely by a combination of favourable conditions.
for abiotic carbonate precipitation aided by organomineralization of microorganisms acting as additional catalysts for mineral nucleation.

**CARBONATE PRESERVATION AND FABRIC DIVERSITY – THE ROLE OF SILICICLASTICS**

The preservation of primary fabrics in sandy dolostones from Unit 2 is high compared with contemporaneous or younger carbonate deposits in Death Valley and elsewhere. Some examples from Unit 3 in the Kingston Range feature fine to medium-grained, sparry dolomite cement (Fig. 10C and D). Sparry cement probably precipitated during progressive diageneis as a fabric-destructive phase that recrystallized from dolomicroite. Micropar cement has previously been documented from the Beck Spring Dolomite by Harwood & Sumner (2012), who reported similar clotted and laminated microstructures as observed in Unit 3 (Fig. 10C and D). Primary carbonate fabrics and absence of recrystallized or neomorphic fabrics in sandy and cherty dolostones in Unit 2 and parts of Unit 3 is corroborated by little to no correlation between C and O isotopes in these sections in both localities ($r^2 = 0.046$ and $0.301$ for Units 2 and 3 in the Kingston range, and $r^2 = 0.558$ and 0.216 in the Alexander Hills, respectively), suggesting little exchange between isotope pools in dolostones and diageneric pore fluids (Jacobsen & Kaufman, 1999; Bishop et al., 2014). Although Unit 3 dolostones feature a more neomorphic fabric, $\delta^{13}C$ and $\delta^{18}O$ values from this unit also do not correlate. The absence of such correlations in Units 2 and 3 is distinct from data obtained from the younger Beck Spring Dolomite. However, there is a slight covariation in $\delta^{13}C$ and $\delta^{18}O$ values from the lowermost Beck Spring Dolomite and the uppermost intercalated dolostones of Unit 4b in the Alexander Hills section ($r^2 = 0.653$, Fig. 14). Covariance in dolomicroites and early dolospars in the Beck Spring Dolomite has been interpreted as mixing of marine and meteoric waters or karstification (Kenny & Knauth, 2001), that may reflect primary C and O isotopic compositions (Tucker, 1983; Zempolich et al., 1988). However, the base of the Beck Spring Dolomite at Alexander Hills, where the C–O covariation is most well-developed, is deformed, partly brecciated, and most primary micritic components have been recrystallized to microsparitic dolomite, arguing for diageneric alteration in open conditions that affected C and O isotopes in a similar manner (Banner & Hanson, 1990; Frank & Lyons, 2000). Oxygen isotope compositions are largely depleted in $^{18}O$ and more heterogeneous than C isotopes in Units 2 and 3 (Fig. 13). Low $\delta^{18}O$ values are typical for many Proterozoic carbonates (Veizer & Hoefs, 1976), and may argue for an influence of meteoric waters or burial diageneric overprinting during post-depositional recrystallization of primary carbonate (Tucker, 1983; Banner & Hanson, 1990; Gilleaudeau & Kah, 2013). Most $\delta^{18}O$ values (between $-8^{\circ}/_{o}$ and $-2^{\circ}/_{o}$) are too high to represent high-temperature burial diageneric fluids (Brand & Veizer, 1981; Frank & Lyons, 2000; Kah et al., 2012).

With the reduced incursion of sand lenses and beds into dolostones from Units 2a to 3 comes an increased occurrence and diversity of microbial carbonate fabrics. Dolostones from Unit 3 exhibit peloidal micritic dolomite (Fig. 10B) composed of isolated peloids several hundreds of microns in size, as well as partly fractured laminae. This rock fabric of peloidal micritic features floating among whitish, fine-grained sparry cement is similar in fabric and appearance to Phanerozoic microbial carbonates associated with reeval microbialites or mud mounds (Russo et al., 1997; Riding & Tomas, 2006; Zhou & Pratt, 2019). A noteworthy observation is that dolostones from Unit 2 and the base of Unit 3 contain more siliciclastics, and are almost entirely devoid of bladed or banded botryoidal, fabric destructive or replacive dolomite cements, which are common in many dolostones and limestones in Neoproterozoic Death Valley strata (Harwood & Sumner, 2011, 2012; Shuster et al., 2018), as well as in isocho- nous strata elsewhere (e.g. Hood et al., 2011; Zentmeyer et al., 2011; Hood & Wallace, 2012; Gilleaudeau et al., 2018). The well-preserved dolomicrotite in Unit 2 appears to have precipitated directly from seawater similar to non-skeletal automicrite precipitation (sensu Reitner & Neuweiler, 1995; Flügel, 2004). Unit 3 marks a change in facies from orange,stromatolitic, sandy dolostones to cherty, grey dolostones with abundant crinkly lamination and occasional stromatolitic fabrics. Microfacies analyses show a concomitant change with these facies, from well-preserved dominantly micritic dolostones characterized by soft sediment deformation, to dolostones featuring sparry dolomite cements partially replaced by silica with occasionally retained microbial fabrics. In this respect, Unit 3
The relationship between carbonate fabric diversity, contents of siliciclastics within the dolostones, and the degree of carbonate preservation may be due to a protective effect of siliciclastic intercalations in carbonates that shielded them from severe diagenetic alteration and post depositional deformation. Rapid siliciclastic deposition onto shallow-water carbonates increases the rates of carbonate burial and vertical accretion, which would have enhanced the potential for preservation of primary fabrics and shielding from marine or burial diagenetic alteration, as has been documented in alluvial sheetflood systems (Gierlowski-Kordesch, 1998), or by improved fossil preservation in carbonate rocks (Ray & Thomas, 2007; Knaust & Desrochers, 2019). The lack of abundant cavity or fenestral structures also argues for the absence of open voids through which diagenetic fluids could have induced neomorphism of primary fabrics (Lloyd & Corsetti, 2010). Siliciclastic sediments have the additional potential of buffering acid-generating early diagenetic reactions in the pore water – carbonate system, leading to a decrease in carbonate dissolution and, potentially, enhanced preservation of primary fabrics (Best et al., 2007; Wehrmann et al., 2009).

The same relationship between carbonate fabric diversity, abundance of later-stage diagenetic cements and siliciclastic influence in dolostones is observable within the stromatolites themselves. Within some of the stromatolites from the Alexander Hills, clotted micritic dolomite forms up to 1 mm wide laminae (Fig. 11C and D) and is devoid of siliciclastic material. Similarly, the smaller, more convex stromatolites in the Kingston Range contain bladed, later-stage diagenetic cements with very little to no siliciclastics (Fig. 11A and B). Conversely, the wider, domal stromatolites from the same locality replete in siliciclastic material are devoid of any later-stage cements or recrystallized dolomitic spar. These observations may be explained by one of two processes, or a combination of the two: First, stromatolites from Unit 3 in the Kingston Range are smaller, slightly more convex, devoid of detrital material, and also feature bladed, later-stage diagenetic cements (Fig. 11A and B). Clotted micrite represents a primary microbial fabric that was probably produced in response to microbial decomposition of organic matter by calcifying microbial communities dwelling in an environment largely undisturbed by siliciclastic deposition (e.g. Soudry & Weissbrod, 1995; Petrov & Semikhatov, 2001; Riding & Tomas, 2006). Micritic laminae featuring clear, whitish cement resembling fenestral fabrics are widespread in Proterozoic and Phanerozoic thrombolites and stromatolites (Camoin & Montaggioni, 1995; Riding, 2011). Because siliciclastic-replete and siliciclastic-free fabric types are observed in different stromatolites that occur in close proximity, differences in carbonate fabrics could represent a response of different microbial communities to changing environmental conditions (cf. Harwood & Sumner, 2012), such as various degrees of siliciclastic sedimentation. Second, the formation of later-stage diagenetic phases including bladed cements, as well as the recrystallization of primary dolomicrite to dolomitic spar, may have been hindered by the presence of abundant and repeated siliciclastic incursions in a similar fashion as argued for the non-stromatolitic dolostones. Currently, however, more data are required in order to strengthen one of the two hypotheses, and future studies may enable us to arrive at a more solidified conclusion.

CONCLUSIONS AND OPEN QUESTIONS

The Horse Thief Springs Formation in Death Valley represents a world class example of Neoproterozoic mixed carbonate – siliciclastic deposits due to excellent exposure, high degrees of preservation, and the accessibility to multiple sections within the area. Considering the dearth of suitable depositional models for Precambrian mixed carbonate – siliciclastic sequences, the Horse Thief Springs Formation offers multiple insights into alternating depositional processes governing carbonate and siliciclastic deposition. Frequent siliciclastic intervals showing almost exclusively sharp transitions to dolostones and vice versa may have been the result tectonic activity related to the rifting of the Rodinia supercontinent, yet more future work is required to solidify this hypothesis. New chemostratigraphic data further suggest that the Horse Thief Springs Formation, deposited in an epicontinental sea on the southern Laurentian margin, may have been in communication with the open ocean, and did not evolve into an isolated, restricted basin. The absence of carbonate – siliciclastic – carbonate allocycles argues against repeated climatic or glacio-eustatic sea-level changes within the Horse Thief Springs Formation.
Formation. This mixed carbonate–siliciclastic system provides evidence for unhindered carbonate sedimentation in the face of repeated and voluminous siliciclastic input, demonstrating the resilience and effectiveness of the Proterozoic non-skeletal carbonate factory. However, continuous supply of siliciclastics locally hampered the development of diverse, subtidal microbial carbonate fabrics and textures, which only came to fruition during episodes of reduced siliciclastic sedimentation. Concomitant with this inverse correlation of carbonate fabric diversity and siliciclastic content, dolostone preservation is high, and later-stage diagenetic carbonate cements typical for the Proterozoic dolostones are absent in sand-rich intervals, whereas the abundance of bladed cements and sparitic dolomite fabrics increases with decreasing siliciclastic content. These observations on a high-resolution outcrop and microfacies scale imply a potentially important role of siliciclastic deposition on carbonate preservation by shielding primary dolostone fabrics from early and later-stage diagenetic fluid circulation and recrystallization. This may either be a distinct feature of Proterozoic mixed carbonate–siliciclastic systems, or be a recurring phenomenon in mixed systems throughout Earth history. Further systematic studies, however, are currently required in order to test this hypothesis.

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DATA AVAILABILITY STATEMENT

The data that support the findings of this study are openly available in PHAIDRA at https://phaidra.univie.ac.at/.

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**Supporting Information**

Additional information may be found in the online version of this article:

**Table S1.** Stable C and O isotope values of dolostones: AH = Horse Thief Springs Formation, Alexander Hills; BC = Beck Spring Dolomite, Beck Canyon, Kingston Range; TS = Horse Thief Springs Formation (Type Section), Beck Canyon, Kingston Range, all values relative to V-PDB.