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Zebra rock and other Ediacaran paleosols from Western Australia

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ABSTRACT
Zebra rock is an ornamental stone from the early Ediacaran, Ranford Formation, around and in Lake Argyle, south of Kununurra, Western Australia. It has been regarded as a marine clay, liquid crystal, groundwater alteration, unconformity paleosol or product of acid sulfate weathering. This study supports the latter hypothesis and finds modern analogues for its distinctive red banding in mottling of gleyed soils. Other acid sulfate paleosols of desert playas (Gypsids) are also found in the Ranford Formation, as well as calcareous desert paleosols (Calcids). The megafossil *Palaeopascichnus* also found in associated grey shales may have been a chambered protozoan, but *Yangtziramulus* in calcic paleosols is most like a microbial earth lichen. Soil climofunctions are evidence of an arid, cool temperate climate during the early Ediacaran.

KEY POINTS
1. Ornamental stones known as ‘zebra rock’ are interpreted as Ediacaran gleyed soils.
2. Gypsum desert-rose beds are interpreted as Ediacaran acid sulfate soils (Gypsids).
3. Calcareous nodular loess beds are interpreted as Ediacaran desert soils (Gypsids).

Introduction

Zebra rock is an ornamental stone unique to northern Western Australia, widely used in jewellery and sculpture. The origin of zebra rock has been regarded as hydrothermal alteration (Kelka *et al.*, 2017; Wallace & Hood, 2018), liesegang banding induced by groundwater (Hobson, 1930), partial leaching of an unconformity paleosol (Trainer, 1931), interbedded marine silt and clay (Geidans, 1981; Larcombe, 1925, 1927), ferronematic banding of a colloidal bed (Mattievich *et al.*, 2003), or a product of acid sulfate soil weathering (Loughnan & Roberts, 1990). The current study aims to adjudicate these alternative origins with additional petrographic and geochemical evidence on zebra rock and other beds in the Ediacaran Ranford Formation and underlying ‘cap carbonate’ of the Moonlight Valley Tillite near Kununurra, Western Australia (Corkeron, 2007, 2008).

A secondary objective is to understand Ediacaran paleoenvironments and life. Of special interest are problematic Ediacaran megafossils associated with zebra rock, such as *Palaeopascichnus* (Lan & Chen, 2012), *Yangtziramulus* (Shen *et al.*, 2009), and a variety of microbially induced sedimentary structures (Lan & Chen, 2013). Biological affinities and paleoenvironments of Ediacaran megafossils remain puzzling (Retallack & Broz, 2020).

Geological background

Zebra rock forms discontinuous lenses within the shaley Johnny Cake Member of the Ranford Formation in localities from northeast to central Lake Argyle (Figure 1). The underlying sandy, Jarrad Member of the Ranford Formation conformably overlies the ‘cap carbonate’ of the Moonlight Valley Tillite in Midnight Valley east of Warnum, and at Palm Springs further south (Bao *et al.*, 2012; Corkeron, 2007). The ‘cap carbonate’ sandstones have been identified as the top 8–10 m of the Midnight Valley Tillite (Kennedy, 1996; Corkeron, 2007), but are very similar to the lowest dolomitic sandstones of the Jarrad Member separated by a poorly exposed interval of shale.

The age of the Ranford Formation is Ediacaran as indicated by fossils of *Palaeopascichnus* (Lan & Chen, 2012). *Yangtziramulus* reported here is another indication of Ediacaran age (Shen *et al.*, 2009). Lithological correlation with other Ediacaran successions in Australia is evidence of early Ediacaran age, below limestones with the stromatolite *Tungussia julia* such as the Égan Formation and Boonall Dolomite of Western Australia (Corkeron, 2007; Grey & Corkeron, 1998), the Julie Limestone of central Australia, and the Wonoka Formation of South Australia (Figure 2). The Ranford Formation also conformably overlies Moonlight Valley Tillite, widely correlated with the terminal Cryogenian, Elatina Glaciation (Corkeron, 2008), commonly...
misnamed Marinoan (Williams et al., 2008). A negative $\delta^{13}C$ isotopic excursion has been used to justify this correlation (Kennedy, 1996), and this correlation is now supported by a distinctive $\Delta^{17}O$ excursion (Bao et al., 2012). These anomalies define the base of the Ediacaran Period in the type section in South Australia (Knoll et al., 2006), so that the base of the Ediacaran is at the base of the ‘cap carbonate’ of the Moonlight Valley Tillite rather than 8 m higher.

Figure 1. Zebra rock localities around Lake Argyle (1–5) and caprock locality east of Warnum (6), Western Australia (geology after Dow & Gemuts, 1969; Sweet, Mendum, et al., 1974; Sweet, Pontifex, et al., 1974).
at the contact with the overlying Ranford Formation (Corkeron, 2007; Kennedy, 1996). An age model using 635 Ma for the base of the cap carbonate, and 567 Ma for top Boonall Dolomite (base Nuccaleena and upper Wonoka of Retallack et al., 2014) gives 635 to 623 Ma for the section measured at Moonlight Valley and 600–599 Ma for short sections measured at Donkey Creek and Duncan Road.

Directions of ice movement in the Moonlight Valley Tillite are from the north (Corkeron, 2008), where tillite forms a thin cap on the Mesoproterozoic (ca 1128 Ma) Osmond Range succession (Figure 2). A local Mesoproterozoic source is also indicated by detrital zircons in the Ranford Formation (Lan et al., 2020). The Ranford Formation is disconformably overlain by early Cambrian
Kinevans Sandstone to the north (Sweet, Mendum, et al., 1974; Sweet, Pontifex, et al., 1974), but is overlain by the Ediacaran, Edward Albert Group to the south (Corkeron, 2007; Plumb et al., 1981). The early Cambrian (505 ± 2 Ma) Antrim Plateau Basalt overlies Kinevans Sandstone and Randfontein Formation (Dow & Gemuts, 1969; Evins et al., 2009; Jourdan et al., 2014; Marshall et al., 2016), and is in turn overlain by Cambrian marine rocks dated by trilobites (Kruse et al., 2004; Opik, 1970), and poorly fossiliferous sandstones of presumed Devonian age (Mory & Beere, 1985).

Materials and methods

Zebra rock was obtained from four localities around Lake Argyle: (1) Duncan Road at S16.12519° E129.04961°, (2) Donkey Creek at S16.0622° E128.96213°, (3) Remote Island at S16.30433° E128.755815°, and (4) Snappy Gum Ridge at S16.359677° E128.718314° (Figure 1). Strata at Duncan Road and Donkey Creek are flat lying (Figure 4c), but dip is near vertical at Remote Island and Snappy Gum Ridge (Figure 3b). Another locality (#3 in Figure 1) near Mt Misere near Argyle Downs Station recorded by Loughnan and Roberts (1990) is now submerged under Lake Argyle. Whole beds of zebra rock were cut and polished, and thin-sections prepared (Figure 3). Measured sections and rock collections were made in quarries at Duncan Road and Donkey Creek (Figure 4), as well as in Moonlight Valley near Warnum (Figures 5 and 6).

Oriented rock samples were collected for laboratory studies of bulk chemical composition (Supplementary Information Table S1), trace element composition (Table S2), and thin-section petrography (Figure 7). Thin-sections were used to quantify grain size (Table S3) and mineral compositions (Table S4) by point counting (500 points) using a Swift automated point counter on a Leitz Orthoplan Pol research microscope. Accuracy of such point counts is ± 2% for common constituents (Murphy, 1983). Major element chemical analysis was used to characterise chemical weathering trends and metamorphic alteration (Figures 8 and 9), determined by X-ray fluorescence at ALS Chemex in Vancouver, Canada, and rare earth elements by inductively coupled plasma atomic emission spectrometry (ICP-AES) in the same laboratory. Ferrous iron was determined by potassium dichromate titration, and total iron as ferric adjusted for that molar proportion. Bulk density was measured by the clod method: from raw weight, then weight in and out of chilled water of paraffin-coated clods.

Metamorphic and diagenetic alteration

Three common early burial alterations were found in the Randfontein Formation and upper Moonlight Valley Tillite, comprising (1) drab-spotting of upper portions of beds owing to burial bleisation of organic matter, (2) dark red (Munsell 10 R) colour from dehydration reddening of ferric hydroxide minerals (Figure 3), and (3) substantial lithostatic compaction (Retallack, 1991a). Burial gleisation is chemical reduction of oxides and hydroxides of iron by anaerobic bacteria on subsidence into anoxic water, and is especially suggested by drab mottles and tubular features radiating down from bed tops (Figure 5b, d), as in Cambrian (Álvarez et al., 2003; Retallack, 2008) and Proterozoic red beds (Driese et al., 1995; Retallack, 2013). Such geologically ancient red beds are also purple to red in colour from burial dehydration of ferric oxyhydroxides (Figure 3), unlike brown to yellow modern soils and late Pleistocene sediments (Retallack, 1991a).

The uppermost Moonlight Valley Tillite 'cap carbonate' and lower Randfontein Formation sandstones have abundant dolomite, both microcrystalline and as silt-sized irregular grains and rhombs. The rhombs are limpid, and not like dolomite replacing other grains, nor forming an interlocking cement, nor baroque with curved cleavage (Mehmood et al., 2018). The microscopic appearance of this 'cap carbonate' is comparable with the Nuccaleena Formation, the basal Ediacaran 'cap carbonate' of South Australia, where micrite and other silt size grains appear to have been loess, with limited burial neomorphism (Retallack, 2011). Correlation of δ13C and δ18O in these dolomites is also highly significant as in soils, unlike weak relationships seen in marine or lacustrine limestones altered by meteoric diagenesis (Figure 10).

One index of lithostatic compaction is the Weaver index of 10 Å/10.5 Å peak height on XRD traces of 3.3 ± 0.2 for the specimens of zebra rock studied by Loughnan and Roberts (1990). This degree of recrystallisation has not resulted in excessive potash metasomatism (Novoselov & de Souza Filho, 2015), because potash enrichment is not seen in tau analysis (Figure 8). A Weaver Index of 3.3 for Ediacaran rocks of the Flinders Ranges of South Australia corresponds to burial depth of 3.6 ± 0.4 km (Retallack, 2013). This is less than 4.4 km of observed overburden in Western Australia (Figure 2), because the Albert Edward Group was not deposited this far north (Mory & Beere, 1985; Plumb et al., 1981). Weaver indices of zebra rock are within the anchizone of diagenesis, but not metamorphic (Frey, 1987), as also suspected from shallow dips, except for the Remote Island and Snappy Gum Ridge localities (Figure 3b). Lack of metamorphism is also indicated by low thermal stability of remnant magnetisation of zebra rock (Abrajевич et al., 2018). Burial compaction (C as a percentage) expected for 3.6 km burial can be calculated as 58%. These calculations use the following formula with depth of burial (B in km) and 0.51, 0.49 and 0.27 as physical constants (of Sheldon & Retallack, 2001) also used to calculate compaction of comparable Ediacaran red beds of the Flinders Ranges, South Australia (Retallack, 2013).

\[
C = \frac{-0.51 \times 100}{\left( \frac{0.49}{B} \right) - 1}
\]
Figure 3. Zebra rock profiles in polished sections and thin-sections from Duncan Road (a–d), Remote Island (e–h) and Lake Argyle (i–l). Yellow epoxy reveals cracks and missing grains.
Sedimentary facies

**TC: trough cross-bedded dolomitic sandstone**

Thick (1–2 m) beds of trough cross-bedded, medium to coarse sandstone in the lower Ranford Formation and upper Moonlight Valley Tillite at Moonlight Valley are prominent cliff-forming units (Figure 5a). From a distance, they can be seen to be lenticular in geometry, with flat tops, but bottoms variably eroding down into underlying beds (Figure 5a). Their bases include angular clasts of redeposited dolomite-cemented sandstone and red claystone breccia, which are intraformational clasts identical to other interbedded layers, including the red clayey Moonlight Valley Tillite.

Trough cross-bedding is created within sand bars confined by channels of rivers (Cant & Walker, 1978) and deltaic distributaries (Gastaldo et al., 2009). A river interpretation is compatible with the diversity of intraformational clasts suggestive of erosional banks rather than constructional deltaic distributaries.

**RG: red-green siltstone**

This is a facies of purple-red siltstone finely interbedded with grey-green siltstone beds. Individual beds are mostly 1–4 mm thick. Some units 15–20 cm thick are more massive with grey-green mottles on the top, but most of the bed is red (Figure 5b, c), and others contain gypsum rosettes (Figure 4c, d). Horizontal planar and wavy bedding are the main sedimentary structures, but there are also straight-crested ripples, shallow hexagonal mud cracks and thin

Figure 4. Field photographs of zebra rock localities in Johnny Cake Member of Ranford Formation: (a) Ruth Duncan with zebra rock from the Duncan Road Quarry; (b) three zebra rock beds tilted vertical at the Snappy Gum Ridge locality; (c) cross-sections (at arrows) of gypsum pseudomorphs in the Thamberalg paleosol at Donkey Creek; (d) gypsum desert roses from Donkey Creek; and (e) aquatic microbially induced sedimentary structure (MISS) Rugalichnus matthewi from Donkey Creek. Scales at base of (d, e) are graduated in millimetres. Photo (b) courtesy of Brian Fennell.
claystone breccias. Mud cracks from RG facies have been illustrated by Lan and Chen (2013).

The fine interlayering of red oxidised and green reduced layers, and claystone breccias with both green and red clasts are evidence that both oxidation and reduction occurred during the Ediacaran (Mawson & Segnit, 1949), rather than oxidation only in outcrop (Gehling, 2000). Lacustrine environments for RG facies are suggested by persistent planar bedding and undulose microbially induced sedimentary structures attributed to aquatic microbial mats (Rugalichnus matthewi of Stimson et al., 2017). These lakes would have been shallow and well aerated to permit alternating oxidation and reduction, and episodic desiccation to produce mud cracks.

IR: interbedded red siltstone and sandstone

Capping beds to facies TC of trough-beded sandstone commonly includes medium to coarse sandy flagstone beds 5–10 cm thick, with interbedded fine-grained sandstones and siltstone, known as interflag sandstone laminae (Figures 5b, c and 6a). Planar bedding, wavy bedding and ripple marks are common, as well as claystone breccias. IR
facies also has strongly textured surfaces informally known as ‘old elephant skin’, and as the microbial trace fossil *Rivularites repertus* (Retallack & Broz, 2020).

The IR facies with interflag sandstone laminae (Retallack, 2019) and *Rivularites repertus* (Retallack & Broz, 2020) is interpreted as fluvial levee, in part from its close association with paleochannel facies TC. The fine-grained laminae are considered eolian reworking of levees during the interval between floods (Retallack, 2019). *Rivularites repertus* is a texture of microbial earths (Retallack, 2012a) with evidence of multiple episodes of cracking and healing, owing to desiccation alternating with rehydration by rain.

Figure 6. Measured sections east of Warnum (a) and in quarries of Donkey Creek (b) and Duncan Road (c).
and flooding (Figure 11c, d). Aquatic microbial mats by contrast are undulose to stromatolitic (Stimson et al., 2017). RS: red sandstone with dolomitic nodules

Dolomitic nodules of the red sandstone facies are found both in place within thick sandstone beds and redeposited into conglomeratic stringers of paleochannels of facies TC. Thus, they were cemented with dolomite during the Ediacaran and not subsequently.

The micritic cement and rounded form of these nodules within a red matrix (Figure 5d) are similar to soil carbonate, also known as caliche (Retallack, 2015a, 2015b). Furthermore, the nodules are arranged not at the erosional surface of the bed, but a set distance below the top of the bed, as is typical for soil calcic (Bk) horizons (Retallack, 2005) of modern soils (Figure 11f). Carbonate nodules in marine rocks are distinct in generally larger size, enclosing marine fossils, grey to green matrix, and lack evidence of redeposition because they formed during burial (Gariboldi et al., 2015). These nodular profiles thus represent well drained floodplain environments.

### Paleosol recognition

Paleosols are definitively recognised in the field by fossil root traces (Retallack, 1976, 1991b), but vascular land plants did not exist before the Silurian (Retallack, 2015a). Lack of root traces may be why not a single Ediacaran paleosol was recognised until 2011, although hundreds of Ediacaran paleosols have been described since then (Retallack, 2016). Field recognition of Precambrian paleosols relies on soil structures (Figures 3–5) and soil horizons (Figures 6 and 7). Special features of Precambrian paleosols may also include surface textures characteristic of microbial earth soils (Rivularites repertus of Retallack & Broz, 2020) and eolian interbeds (interflag sandstone laminae of Retallack, 2019). The following paragraphs consider each of these criteria in turn.
Crack patterns

Desiccation cracks have a V-shaped profile in clay (Weinberger, 2001), and several examples in shales of facies IR of the Johnny Cake Member of the Ranford Formation have been illustrated by Lan and Chen (2013). Similar tapering cracks were also seen in the dolomitic sandstones of the upper Moonlight Valley Tillite and Jarrad Member of the Ranford Formation (Figure 5b, d). These sandstone cracks emanate from the most hematite-rich tops of sandy beds but are filled with fine-grained white dolomitic sandstone. How sand can crack like clay is explained by Prave (2002) as owing to abundant hydrated microbiota, like that of a thick microbial earth soil rather than thin microbial mat (Retallack, 2012a). This idea is supported by co-occurring microbially influenced sedimentary structures (MISS of Noffke, 2010), such as the microbial trace fossil *Rivularites repertus* (Retallack & Broz, 2020), which has complex, multiple fills of shallow cracks (Figure 5c). A comparable phenomenon is oscillating desiccation cracks of modern supratidal flats (Noffke, 2010), which is the depositional setting envisaged by Lan and Chen (2013).

Sand crystals

Rosettes of gypsum with included detrital grains in the Johnny Cake Member of the Ranford Formation (Figure 4d) have attracted attention as possible Ediacaran discoid fossils (Dunnet, 1965), which have been variously interpreted.
as fossil jellyfish (Sprigg, 1949) or microbial colonies (Grazhdankin et al., 2012). The Ranford rosettes have been debunked as fossils by Grey (1981a, 1981b) and Lan and Chen (2012), primarily because of their radiating crystal structure like desert roses (Almohandis, 2002; Jafarzadeh & Burnham, 1992), although some have a degraded concentric appearance like Ediacaran discoids (Sprigg, 1949). Their occurrence well below the truncated upper surface of beds (Figure 4c) is comparable with gypsic horizons (Figure 11e) of desert soils (Retallack & Huang, 2010). Sand crystals form only in soils, and not in marine or lacustrine evaporites, nor in intertidal sabkhas, where saturated sediment allows clear crystals to grow by displacement of grains (Renaut & Tiecerlin, 1994; Ziegenbalg et al., 2010).

Figure 10. Covariation of carbon and oxygen isotopic composition of carbonate as a characteristic of paleosols, rather than other settings: (a) early Ediacaran cap carbonate of Moonlight Valley Tillite at two localities (Bao et al., 2012), Ediacaran and Cambrian paleosols of South Australia (Retallack et al., 2014), Cambrian Arumera Formation at Ross River (Retallack & Broz, 2020), Ordovician paleosols of Pennsylvania (Retallack, 2015a), and Silurian paleosols of Pennsylvania (Retallack, 2015b); (b) soil nodules (above Woodhouse lava flow, near Flagstaff, Arizona) (Knauth et al., 2003) and in Yuanmou Basin, Yunnan, China (Huang et al., 2005); (c) soil crusts on basalt (Sentinel Volcanic Field, Arizona, from Knauth et al., 2003); (d) Quaternary marine limestone altered diagenetically by meteoric water (Key Largo, Florida, Lohmann, 1988; Clino Island, Bahamas, Melim et al., 2004); (e) Holocene (open circles) and Ordovician (open squares) unweathered marine limestones (Veizer et al., 1999) and lower Cambrian (closed circles), Ajax Limestone, South Australia (Surge et al., 1997); (f) marine methane cold seep carbonate (Miocene, Santa Cruz Formation, Santa Cruz, California (Aiello et al., 2001). Slope of linear regression (m) and coefficients of determination (R²) show that carbon and oxygen isotopic composition is significantly correlated in soils and paleosols.
Interflag sandstone laminae

Thin laminae of fine-grained sandstone and siltstone cap thick cracked beds of dolomitic sandstone in the Moonlight Valley Tillite and Jarrad Member of the Ranford Formation (Figures 5b and 6a), and appear to be interflag sandstone laminae (Retallack, 2019). In the Johnny Cake Member of the Ranford Formation some thin siltstone layers show inverse grading (Figure 3d), characteristic of climbing translatent bedding formed by wind (Hunter, 1977). Traction currents of modern river flooding produces medium-grained beds, and these alternate with fine-grained laminae from wind reworking during the interval of a year or more between floods (Draut & Rubin, 2008; Draut et al., 2008). Wind redeposition is evidence of subaerial exposure, which is also a requirement for paleosols.

Zebra rock

Zebra rock is a clayey siltstone bed, 5–15 cm thick, with alternating diffuse red and grey bands, 4 to 25 mm thick, varying from subhorizontal to angles up to 50° from regional bedding (Figures 3 and 4a, b). These distinctively
banded rocks can form successive beds (Figure 4b), or occur singly (Figure 4a), within the Johnny Cake Member of the Ranford Formation. The Duncan Road quarry exposes zebra rock 90 m along strike and it varies in thickness considerably, pinching out in places and elsewhere swelling to thickness of 15 cm. Less marked lateral variation in thickness is also seen in the Snappy Gum Ridge quarry (Figure 4b). Tops of the zebra rock bed are red and over- lain abruptly by coarse-grained grey sandstone (Figure 3a, e, i). The bottoms of the zebra rock bed are grey sandstone (Figure 3d, h, l), although that grey colour can be obscured by pits remaining from pyrite ferruginised in the modern outcrop (Figure 3e). Gleisation as indicated by ferrous/ferric ratios >1 is found at the base, not at the top of the bed (Figure 7). Two of the zebra rocks analysed had very high proportion of mica, but this was less abundant in a third example, and all three showed strong depletion of feldspar toward the top of the bed (Figure 7). Both molecular weathering ratios (Figure 7) and tau analysis (Figure 8) of major elements show that zebra rock is intensely weathered from the surface down, as also indicated by identification of the main components as quartz, kaolinite and dickite, with minor sericite and alunite (Loughnan & Roberts, 1990). Ferrihydrite has been detected in zebra rock by electron paramagnetic resonance in both red and white regions (Boas et al., 2005). Intense leaching also is supported by REE enrichment toward the top, without fractionation of light vs heavy REE (Figure 9). Red bands have 30 wt% iron as Fe₂O₃, but drab bands have less than 1 wt% iron largely as FeO (Table S1). Loughnan and Roberts (1990) found a disparity of 8–9 wt% total iron in red bands, but only 0.5 and 0.07 wt% in the white bands. Some remnant paleomagnetic directions in zebra rock and associated sediments can be identified as Paleozoic overprints, but zebra rock has a very distinctive paleomagnetic orientation regarded as primary Ediacaran by Abrajevitch et al. (2018).

Several past hypotheses for the origin of zebra rock can now be dismissed. Low metamorphic grade and primary paleomagnetic signal (Abrajevitch et al., 2018; Loughnan & Roberts, 1990) is evidence against formation by hydrothermal mineralising solutions (Kelka et al., 2017). Multiple horizons of zebra rock tipped on end (Figure 4b) are evidence against liesegang banding induced by weathering of bedrock (Hobson, 1930), or within a single unconformity paleosol (Trainor, 1931). Liesegang bands are asymmetric with intensity increasing gradually then ceasing abruptly and follow the outlines of joint blocks or clasts (Sadek et al., 2010), unlike zebra rock (Figure 3); nor do the bands correspond to variations in grainsize (Figure 3) to support a primary depositional origin as cross-bedding (Geidans, 1981; Larcombe, 1925, 1927).

In contrast, zebra rock beds are thicker and more massive than associated shaley laminae (Figures 4b and 6b) as if each bed were deposited in one episode as an airfall crystal tuff, and in two cases as a biotite tuff (Figure 7). Once in place, profound chemical weathering (Figure 7) and paleomagnetic characters (Abrajevitch et al., 2018) are evidence that they were soils comparable with modern gleyed soils near Union, South Carolina (Figure 11a, b). This distinctive grey-red banding in soils within diffuse margins comparable with zebra rock (Figure 3) is considered the result of gleisation of a red soil by pockets of anaerobic bacteria active during waterlogging (Chen et al., 2019), like other examples of soil gleisation (Schulz et al., 2016). It is the opposite of conservation of total iron in red and grey areas of paleosols with closed-system burial gleisation (Retallack, 1991a). Loss of iron in white bands of zebra rock by open system reduction down to reduced lower horizons is compatible with groundwater rather than surface-water gleisation (Vepraskas & Sprecher, 1997). Soil lamellae are another comparable feature, but these ferruginised seams within soils are much more sandy than zebra rock (Bockheim & Hartemink, 2013; Rawling, 2000). An alternative view of Mattievich et al. (2003) is that zebra rock formed as a liquid crystal in which hematite grains in the tens of nanometre size range had sufficient magnetic moment to segregate within a colloidal bed under a normal intensity of Earth’s magnetic field. Evidence against this is the preservation of sand and silt grains in zebra rock (Figure 3), which were far from colloidal, with no more than 50% grains of clay or finer grainsize (Figure 7; Table S3). Among these grains are oxidised pyrite, supporting the idea of acid sulfate soil weathering to account for the localised intensity of weathering of thin zebra rock beds (Loughnan & Roberts, 1990).

**Mineral weathering trends**

Point counting of individual beds shows surface depletion of rock fragments and feldspar in proportion to degree of destruction of bedding, especially marked within the thin zebra rock beds (Figure 7). Clay enrichment is commonly abrupt below sharp grainsize discontinuities with overlying sandstone and confined to intervals only 5–15 cm thick, unlike alteration by symmetrical hydrothermal or diffuse metamorphic alteration (Kelka et al., 2017; Wallace & Hood, 2018). This is best explained as the result of hydrolytic weathering to clay of feldspar within soil profiles (Retallack, 2013, 2019). These abrupt contacts are well-developed soils overlain sharply by fluvial sandstones.

These trends are also seen in sandstones of the upper Moonlight Valley Tillite and Jarrad Member of the Ranford Formation, which contain large amounts of silt size dolomite (Figure 7). Dolomite is depleted near the top of the profiles but is very abundant at the base of beds as well. Some leaching of dolomite during soil formation is likely, but such substantial amounts of dolomite are evidence that this mineral was also detrital, either eolian or fluvial.
Chemical weathering trends

Molar weathering ratios do not show the erratic depth functions of chemically distinct sedimentary beds, but relatively smooth trends compatible with soil formation of two divergent kinds (Figure 7). Uniformly low soda/potash ratios are evidence against salinisation in these beds. Alkaline earths/alumina spike in dolomitic nodules of two Galadil profiles so are evidence for calcification, but the zebra rocks (Wajing and Thamberalg) are non-calcareous. Hydrolysis of feldspar to clay is revealed by increases in alumina/silica and alumina/bases at the tops of profiles, especially marked in the zebra rocks (Wajing and Thamberalg profiles), but not in the dolomitic profiles (Galadil). Ba/Sr molar ratios as evidence of chemical leaching are low in dolomitic profiles (Galadil), but very low in the zebra rocks (Wajing and Thamberalg). Low ferrous/ferric iron confirms red colour from hematite is evidence of oxidising conditions, especially in zebra rocks (Wajing and Thamberalg), but the dolomitic (Galadil) profiles show modest chemical reduction (gleyisation; Retallack, 1991a). Oxidation during the Ediacaran, rather than in modern outcrop, is indicated by red claystone clasts redeposited within grey sandstones (Figure 8).

Rare earth elements show modest enrichment toward the surface with only slight rare earth enrichment in a Wajing profile (Figure 9), as in many other Precambrian arid land paleosols (Retallack, 2018; Retallack & Mao, 2019). Comparable enrichment slightly favouring light rare earth elements is also seen in modern aridland loessial soils (Compton et al., 2003; Ramakrishnan & Tiwari, 1999), unlike depletion seen in humid granitic soils (de Sá Paye et al., 2016; Kurtz et al., 2001). There is a slight negative europium anomaly, as in sediments from granitic (Foden et al., 1984), rather than basaltic rocks (Bavinton & Taylor, 1980). There is no marked enrichment in europium as in hydrothermally altered rocks (Bolhar et al., 2005; Sugahara et al., 2010). Nor are rare earth elements unchanged within the bed as in marine deposition and hydromorphic alteration of marine beds (Claucer et al., 1990; Setti et al., 2004).

Tau analysis

A definitive method to disentangle soil formation from sedimentation is tau analysis (Brimhall et al., 1992). Tau analysis deconstructs two separate aspects of weathering: mole fraction mass transport ($\dot{e}_{j,w}$) of a mobile element and mole fraction strain ($e_{i,w}$) of an immobile element (Ti used here), using the following formula including bulk density ($\rho$ in g.cm$^{-3}$) and oxide assay (C in weight %) for successive samples (subscripts $i, j$) of weathered material (subscript w) and parent material (subscript p).

$$e_{i,w} = \frac{\rho_p C_{j,p}}{\rho_w C_{j,w}} - 1$$

Soils and paleosols lose mass with weathering and so have negative strain ($e_{i,w} < 0$), and also lose nutrient cations and silica, so have negative mass transfer ($\dot{e}_{j,w} < 0$). In contrast, sediment accumulation and diagenetic alteration other than weathering adds elements and mass so has positive strain and mass transfer. Tau analysis has been widely used for Precambrian paleosols (Liivamägi et al., 2014; Retallack & Mindszenty, 1994), as well as Cenozoic paleosols (Bestland et al., 1996; Sheldon, 2003), and modern soils (Chadwick et al., 1990; Hayes et al., 2019). On this basis, most analyses of beds within the Moonlight Valley Tillite and Ranford Formation are within the collapse and loss quadrant, with the exception of iron enrichment of red, but not drab parts of zebra rocks (Wajing of Figure 8a). Zebra rock beds are thin but show profound chemical differentiation. In contrast, deep depletion from the surface is well demonstrated by dolomitic profiles (Galadil of Figure 8b).

Stable isotopic correlation

Analyses of dolomitic nodules in the Ediacaran ‘cap carbonate’ of the upper Moonlight Valley Tillite at both Moonlight Valley and Palm Springs by Bao et al. (2012) show significant correlation of $\delta^{13}C$ and $\delta^{18}O$ values like other Ediacaran and Paleozoic paleosols (Figure 10a), and Holocene soils (Figure 10b) in China (Huang et al., 2005) and Arizona (Knauth et al., 2003). Less statistically significant correlations (Figure 10c, d) are found in soil carbonate crusts (Knauth et al., 2003), and marine limestone altered by deep circulation of meteoric water (Lohmann, 1988; Melim et al., 2004). Although similar relationships in seasonally dry lake carbonates have been attributed to mixing of distinct fluid sources (Talbot, 1990), these may also be cases of alteration by meteoric weathering during lake low stands. In contrast, unaltered marine limestones and sea shells (Figure 10e) and perennial lake carbonates show no hint of correlation (Surge et al., 1997; Talbot, 1990; Veizer et al., 1999), and near constant $\delta^{13}C$ values but highly varied $\delta^{18}O$ values but highly varied $\delta^{13}C$ values (Figure 10f) is created by microbial methanogenesis in carbonate of marine methane seeps (Aiello et al., 2001; Peckmann et al., 2002), and siderite of wetland paleosols (Ludvigson et al., 1998, 2013). The position of data within this cross-plot is determined by a variety of paleoenvironmental factors, especially temperature and CO$_2$ partial pressures, but it is the tightness of the correlation that distinguishes paleosols.

Such tight correlations of $\delta^{13}C$ and $\delta^{18}O$ values are not accidental, but owing to selection for light isotopologues of CO$_2$ (Retallack, 2016). This may be due to kinetic evaporative effects in narrow spaces of soils (Ufnar et al., 2008), but is more likely related to stomatal conductance and fractionation during photosynthesis by the enzymes.
rubisco and carbonic anhydrase because $\delta^{13}$C and $\delta^{18}$O covariance is seen in respired soil CO$_2$ (Ehleringer & Cook, 1998; Ehleringer et al., 2000), and in plant cellulose (Barbour & Farquhar, 2000; Barbour et al., 2002). A role for stomates is undermined by strong correlation of $\delta^{13}$C and $\delta^{18}$O values in early Paleozoic and Ediacaran paleosols, because they predate the evolution of stomates (Figure 10). Similar covariance of $\delta^{13}$C and $\delta^{18}$O values is found in Ediacaran carbonate in the Fauquier Formation of Virginia (Hebert et al., 2010), and the Shibantan Member of the Dengying Formation of China (Duda et al., 2014). The covariance is not destroyed by metamorphism to green- chis facies in paleosols of the Gunjiata and Bloomsburg formations (Retallack, 2015). Values of $\Delta^{17}$O found in carbonate associated sulfate of the Moonlight Valley Tillite by Bao et al. (2012) are so low that they would not have been preserved in either marine or freshwater. Such mass independent fractionations are only preserved in terrestrial tuffs (Bindeman et al., 2007; Martin & Bindeman, 2009). This is evidence that the Moonlight Valley Tillite ‘cap carbonate’ in its type section was a terrestrial deposit.

**Paleosol interpretation**

The preceding paragraphs are evidence for a variety of paleosols in the Upper Moonlight Valley Tillite and Ranford Formation. The remainder of this paper develops paleoenvironmental conclusions based on these paleosols. The various kinds of beds analysed as putative paleosols have been named (Table 1) using the Mirriwoong (for Lake Argyle) and Gija (for Warnum) aboriginal languages (Jarragigirem, 2017; Olawsky & Kofod, 2019). Despite burial illitisation, these pedotypes can now be interpreted in terms of soil taxonomy and various soil-forming factors to build a detailed model of their paleoenvironment (Table 2; Figure 12).

**Paleosol identification**

Thamberalg profiles with cracked surface (A horizon) over a diffuse horizon with mottles and sand crystals (Bk or gyp- sic) are most like Gypsids (Soil Survey Staff, 2014). Galadil profiles, on the other hand, have cracked surfaces (A horizon) over a deep horizon (Bk or calcic) of pedogenic carbonate nodules, as in Calcds (Soil Survey Staff, 2014). Comparable considerations can be used to classify these paleosols in other classifications (Table 1) of Australia (Isbell, 1996; Stace et al., 1968), and of the Food and Agriculture Organization (1978). Other profiles are less well developed Entisols and Inceptisols (of Soil Survey Staff, 2014), and would have been restricted to disturbed parts of the landscape (Table 2).

In the Food and Agriculture Organisation map classification (Food and Agriculture Organization, 1978) the best-developed soil type of the Jarrad Member of the Ranford Formation was Orthic Solonchak, this soil assemblage would have a map code of Zo + Jd,Gd (Table 1). The closest modern match is map unit Zo36–2a + Rd,Rc,Xk,Yh in hot subtropical desert around Lake Mackay and other allu- vial and playa plains on the border of Northern Territory and Western Australia (Food and Agriculture Organization, 1978). At nearby Walungurru Airport, mean annual temperature is 26.2°C and mean annual precipitation is 280 mm (Bureau of Meteorology, 2020). The calcareous upper Moonlight Valley Tillite on the other hand has mainly Calcic Xerosols, in a map code Xk + Rc, most like Xk45–21 + Rc similar to the coastal plain inland of Eighty Mile Beach in the northwest coast of Australia (Food and Agriculture Organization, 1978). At nearby Mandora, mean annual temperature is 26.8°C and mean annual precipitation is 378 mm (Bureau of Meteorology, 2020). Both these places are in Western Australia within 800 km of Lake Argyle.

Similar soils also form in cooler climates. Hypercalcic Calcarosols similar to Galadil paleosols have formed on an alluvial plain near Clare, South Australia, under mean annual precipitation of 444 mm, and mean annual temperature of 15.2°C (profile CA7 of McKenzie et al., 2004). Red Sodosols similar to Thamberalg paleosols formed on an alluvial plain near Deniliquin, New South Wales, under a mean annual precipitation of 339 mm and mean annual temperature of 15.7°C (profile SO3 of McKenzie et al., 2004).

**Original parent material**

Parent materials to the paleosols were of two distinct kinds: dolomitic sandstones in the upper Moonlight Valley Tillite and basalt Ranford Formation, and reworked felsic tuffs in the middle Ranford Formation zebra rock quarries. Silt size dolomite is as much as 24.4 vol% of parent materials to Jilam and Galadil profiles and was a conspicuous part of their loessic parent materials. A likely source for dolomite is glacial erosion of the Mesozoic reefal limestone (ca 1128 Ma), Bungle Bungle Dolomite, in glacial valleys that were the source of the Moonlight Valley Tillite to the north (Corkeron, 2008). This geologically diverse recycled orogen of the Ranford Formation is also compatible with a relatively flat REE pattern (Figure 10) similar to post- Archean Australian sediment (Nance & Taylor, 1976), and detrital zircons largely of Mesozoic age (2000–1000 Ma; Lan et al., 2020). The basal Ediacaran Nuccaleena Formation of South Australia (Retallack, 2011) is a similar partly eolian deposit of dolomite, quartz and feldspar capping a thick red tillite. A modern analogue is the late Pleistocene Peoria Silt of North America, which has carbonate content of 42% at Vicksburg, Mississippi (Fisk, 1951), 32% at Cumbuck, Indiana (Ruhe & Olson, 1980), and 31% in core G56 in Illinois 80 km north of St Louis (Grimley et al., 1998).
### Table 1. Pedotypes and diagnosis for Ediacaran red beds of Western Australia.

<table>
<thead>
<tr>
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<tbody>
<tr>
<td><strong>Danggang</strong></td>
<td>Red ochre</td>
<td>Red siltstone surface (A) over bedded sandstone or claystone breccia (C)</td>
<td>Fluvent</td>
<td>Dystric Fluvisol</td>
<td>Alluvial Soil</td>
<td>Stratic Rudosol</td>
</tr>
<tr>
<td><strong>Galadil</strong></td>
<td>White paint</td>
<td>Red sandstone with penetrating fractal structures (A) over red sandstone with micritic nodules (Bk)</td>
<td>Calcid</td>
<td>Calcic Xerosol</td>
<td>Red Desert Soil</td>
<td>Supracalciic Calcarosol</td>
</tr>
<tr>
<td><strong>Jilam</strong></td>
<td>ground</td>
<td>Red sandstone with 'old elephant skin' (A) over red bedded sandstone (C)</td>
<td>Psamment</td>
<td>Calcaric Regosol</td>
<td>Coalure Sand</td>
<td>Arenic Rudosol</td>
</tr>
<tr>
<td><strong>Thamberalg</strong></td>
<td>star</td>
<td>Red-grey layered surface (A) over siltstone with gypsum rosettes (By)</td>
<td>Gypsid</td>
<td>Orthic Solonchak</td>
<td>Solonchak</td>
<td>Hypersalicy Hydrosol</td>
</tr>
<tr>
<td><strong>Wajing</strong></td>
<td>rainbow</td>
<td>Red siltstone surface (A) red-grey spotted and banded siltstone 'zebra rock' (Bg)</td>
<td>Aquept</td>
<td>Dystric Gleysol</td>
<td>Non-calciic Brown Soil</td>
<td>Oxyaquic Hydrosol</td>
</tr>
</tbody>
</table>

Indigenous names pedotypes around Lake Argyle are from Mirriwoong language (Olawsky & Kofod, 2019) and for Warnum region from Gija language (Jaraggirrem, 2017).

### Table 2. Interpretation of pedotypes for Ediacaran red beds of Western Australia.

<table>
<thead>
<tr>
<th>Pedotype</th>
<th>Paleoclimate</th>
<th>Ecosystems</th>
<th>Parent materials</th>
<th>Paleo-topography</th>
<th>Time for formation (kyrs)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Danggang</strong></td>
<td>Not diagnostic for climate</td>
<td>Microbial earth</td>
<td>Quartz-feldsparic sand</td>
<td>Lake margin lunette</td>
<td>0.1–1.0</td>
</tr>
<tr>
<td><strong>Galadil</strong></td>
<td>Frigid (-2.5 ± 0.5°C and -1.7 ± 0.5°C mean annual temperature), arid (289 mm ± 147 mm mean annual precipitation), seasonal (44 ± 22 mm mean annual range of precipitation)</td>
<td>Microbial earth with Yangtziramulus zhangi</td>
<td>Quartz-dolomite sand</td>
<td>Well-drained floodplain</td>
<td>11–17 ± 1.8</td>
</tr>
<tr>
<td><strong>Jilam</strong></td>
<td>Not diagnostic for climate</td>
<td>Ruvularites repertus</td>
<td>Quartz-dolomite sand</td>
<td>Near stream levee</td>
<td>1.0–2.0</td>
</tr>
<tr>
<td><strong>Thamberalg</strong></td>
<td>Temperate (11.9 ± 0.5°C mean annual temperature), hyperarid (106 ± 129 mm and 109 ± 129 mm mean annual temperature)</td>
<td>Microbial earth</td>
<td>Quartz-feldsparic silt</td>
<td>Playa lake</td>
<td>89–97 ± 15</td>
</tr>
<tr>
<td><strong>Wajing</strong></td>
<td>Not diagnostic for climate</td>
<td>Microbial earth</td>
<td>Rhyolitic crystal tuff</td>
<td>Seasonally waterlogged floodplain</td>
<td>1.0–2.0</td>
</tr>
</tbody>
</table>
Two of the zebra rock beds studied are mica-rich, but one is mica-poor, and all are relatively massive beds compared with laminated to varved enclosing rocks of the Johnny Cake Member of the Ranford Formation. These zebra rock profiles may have been weathered from biotite tuffs and feldspathic tuffs. The tuffs are andesitic in the alkali/silica discrimination of Le Bas et al. (1986), but that composition is compromised by weathering in place (Loughnan & Roberts, 1990). Immobile trace element ratios of Zr/TiO₂ vs Nb/Y ratios of Winchester and Floyd (1977) are a better indication and are rhyolitic. No local active volcano is known; perhaps they were far-travelled airfall ash from 600 Ma volcanic arcs reconstructed to the west in what is now southern China (Scotese, 2009).

Reconstructed sedimentary setting

Two distinct sedimentary settings are represented by dolomitic sandstones of the upper Moonlight Valley Tillite and lower Jarrad Member, and zebra rock shales of the Johnny Cake Member of the Ranford Formation. Fluvial paleochannels (Facies TC) of dolomitic sandstone are capped by both Jilam and Galadil paleosols (Facies RS), and both paleosols are capped by fine-grained eolian siltstones (Facies IR). Galadil profiles are deeper and have large nodules as indications of greater time for formation (Retallack, 2005) and good drainage in infrequently flooded plains or terraces. In contrast, Jilam profiles are thin and grey at the base, as if waterlogged by groundwater (Vepraskas & Sprecher, 1997). Thus, Jilam profiles formed on alluvial levees.

Laminated and varved shales of the Johnny Cake Member represent a different sedimentary facies (RG) of varved and laminated, red and green, shales, with thin sandstones and weathered tuffs of a lacustrine basin that was intermittently exposed like modern playas and salt lakes (Benison et al., 2007; Benison & Bowen, 2015). Thamberalg profiles with gypsum sand crystals are similar to dry lake Gypsids or Solonchak soils (McKenzie et al., 2004). Zebra rock (Wajing) profiles were rhyolitic feldspar and biotite tuffs with poorly drained lower horizons within the lake basins, and show acid sulfate weathering (Loughnan & Roberts, 1990). Red sandstone Danggang profiles were on local sandy soils, perhaps eolian lunettes which commonly flank playa lakes (Bowler, 1973; Fitzsimmons et al., 2014). Zebra rock’s intimate admixture of oxidised and gleyed areas with open system enrichment and depletion of iron over small distances, but with red at the top and grey at the bottom, is similar to soils within the zone of seasonal rise and fall of water-table (Vepraskas & Sprecher, 1997).

Time for formation

Duration of soil formation for Galadil and Thamberalg pedotypes can be calculated from chronofunctions for modern arid land soils. Diameter of pedogenic carbonate nodules (D in cm) is related to radiocarbon age of nodules (A in kyrs) near Las Cruces, New Mexico, with modest precision ($r^2 = 0.57$, s.e. $= \pm 1.8$ kyr, $p = 0.001$), as follows (Retallack, 2005).

$$A = 3.92D^{0.34} \quad (4)$$

Similarly, abundance of gypsum in a profile (G as percentage surface area using comparison chart of Terry & Chilingar, 1955) is a metric for age (A in kyrs) in the Negev Desert of Israel, with fair precision ($r^2 = 0.95$, s.e. $= \pm 15$, and $p < 0.01$), as follows (Retallack, 2013).

$$A = 3.987G + 5.774 \quad (5)$$

The calcic chronofunction applied to five Galadil paleosols gives durations of $11-17 \pm 1.8$ kyr (13.9 ± 3.4 kyr mean and standard deviation), and the gypsum chronofunction applied to two Thamberalg profiles with 23 and 21% gypsum (Figure 4d) give durations of 97.4 ± 15 and 89.5 ± 15 kyr. These Ediacaran paleosols are thus strongly developed, on a scale in which Holocene (ca 10 ka) soils are moderately developed.

Other paleosols would have been much less developed, with a few millennia likely for Jilam and Wajing paleosols with sedimentary structures obscured to a depth of 15 cm, and only a century or less for Danggang paleosols with clear relict bedding. These are maximal estimates, because

**Figure 12.** Reconstructed early Ediacaran soils of north Western Australia.
they were derived by comparison with homogenisation of bedding in Pleistocene soils of the San Joaquin Valley, California (Harden, 1982), which were more actively rooted and burrowed than likely for Ediacaran soils.

**Paleoclimate**

Gypsic and calcic horizons are today found at depths in soils proportional to mean annual precipitation (Retallack, 2005, 2012b; Retallack & Huang, 2010), and this in turn is related to the balance between atmospheric CO$_2$ and soil CO$_2$ from terrestrial productivity (Breecker & Retallack, 2014). Comparisons of paleoclimatic inferences from Bk and By metrics with independent geochemical proxies do not show serious divergence in Neoproterozoic paleosols (Retallack, 2011, 2013; Retallack & Mindszenty, 1994), but Paleoproterozoic and Archean paleosols are problematic because of aggressive acid sulfate weathering (Retallack & Mao, 2019; Retallack et al., 2016). Weathering by strong sulfuric acid rather than weak carbonic acid also affected Thamberalg and Wajing paleosols of the Ranford Formation, and explains their unusually high degree of chemical weathering (Loughnan & Roberts, 1990). Calccic soils are widespread, but gypsic soils form in extreme deserts such as the Atacama Desert of Chile (Ewing et al., 2008; Navarro-González et al., 2003). For calcic paleosols, mean annual precipitation (P in millimetres) is related to depth in the profile to calcareous nodules (D in cm corrected for burial compaction by equation 1) from a global compilation ($r^2 = 0.52$, s.e. = $\pm 147$, and $p < 0.0001$), as follows (Retallack, 2005).

$$P = 137.24 + 6.45D - 0.0132D^2$$  \hspace{1cm} (6)

For gypsic soils, another global compilation gives mean annual precipitation from depth to gypsum (D again compaction corrected) ($r^2 = 0.63$, s.e. = $\pm 129$, and $p < 0.000000001$), as follows (Retallack & Huang, 2010).

$$P = 87.593e^{0.0209D}$$  \hspace{1cm} (7)

Seasonality of precipitation, defined as wettest minus driest month mean precipitation (M in millimetres) is a function of thickness of the calcic horizon (T in cm), again from a global compilation ($r^2 = 0.58$, s.e. = $\pm 22$, and $p < 0.00001$) as follows (Retallack, 2005).

$$R = 0.79T + 13.7$$  \hspace{1cm} (8)

Application of these Bk metrics to five Galadil paleosols of the upper Moonlight Valley Tillite and Jarrad Member of the Ranford Formation gives mean annual precipitation averaging 289 mm $\pm 147$ (265–325 mm). Mean annual range of precipitation averaged 44 $\pm 22$ (37–48 mm), which is seasonality much less than from modern monsoonal climates (Retallack, 2005).

Application of By metrics to two Thamberalg profiles give mean annual precipitation of 106 $\pm 129$ mm and 109 $\pm 129$ mm, which is hyperarid like the modern Atacama Desert (Ewing et al., 2008; Navarro-González et al., 2003). Pedogenic paleothermometers based on soils of modern woody vegetation (Gallagher & Sheldon, 2013; Nordt & Driese, 2010; Sheldon et al., 2002) are not applicable to likely microbial earths of Precambrian paleosols (Retallack & Mao, 2019). A paleothermometer based on modern soils under lichen-shrub tundra vegetation of Iceland (Öskarsson et al., 2012) is an option, predicting temperature ($T$ in °C) from chemical index of weathering ($W$ in mole fractions) of non-calcareous Bw and By horizons ($r^2 = 0.81$, s.e. = 0.5, $p < 0.001$).

$$T = 0.21W - 8.93$$  \hspace{1cm} (9)

$$W = \frac{100Al_{2}O_{3}}{(Al_{2}O_{3} + CaO + Na_{2}O)}$$  \hspace{1cm} (10)

Another useful climofunction is the molar alkali index ($I$) of Sheldon et al. (2002), which includes desert soils in its training set, and gives temperature ($T$ in °C) following equations 11 ($r^2 = 0.37$, s.e. = 4.4, $p < 0.00001$) and 12.

$$T = 18.5I + 17.3$$  \hspace{1cm} (11)

$$I = \frac{K_{2}O + Na_{2}O}{Al_{2}O_{3}}$$  \hspace{1cm} (12)

This CIW proxy of Öskarsson et al. (2012) gives mean annual temperature of $-2.5 \pm 0.5$°C for the upper Bk horizon of the Galadil silt loam and $-1.7 \pm 0.5$°C for the upper Bk horizon of the Galadil sandy clay loam, which are frigid climates, evidently persisting above deposition of the tillite, as in the basal Ediacaran Nuccaleena Formation in South Australia (Retallack, 2011). These results are not disqualified by high content of lime (3–6 wt% CaO in Galadil profiles), specifically excluded for the similar chemical index of alteration (Nesbitt & Young, 1982, 1989), because the Icelandic training set for this paleothermometer also has high lime (3–11 wt%); however, in Iceland the lime is in volcanic glass rather than dolomite (Öskarsson et al., 2012). This CIW metric also gave 11.9 $\pm 0.5$°C for the Thamberalg profile. The alkali index proxy gave mean annual temperatures of 12.0 $\pm 4.4$°C and 12.4 $\pm 4.4$°C for the Galadil paleosols (respectively) and 16.9 $\pm 4.4$°C for the Thamberalg profile. Paleotemperature was not calculated for Wajing paleosols also analysed because they were waterlogged (see above), and so did not reflect atmospheric temperatures.

Cool temperate to frigid paleotemperatures are surprising considering the low paleolatitude of these formations from paleomagnetic data. An average of eight paleolatitudes determined for zebra rock is 10° $\pm$ 6°, although one near-polar paleolatitude of 86° was also found by Abrajevitch et al. (2018) and may be a Paleozoic overprint. Ediacaran periglacial paleosols (Retallack, 2011, 2013) and tillites such as the Gaskiers Formation of Canada (Pu et al., 2016) demonstrate that the Ediacaran Period was not a full recovery from extreme glacial conditions of the Cryogenian Snowball Earth (Hoffman et al., 1998; Retallack, 2011).
Life on land

Two Ediacaran mega-fossils were found in the Upper Moonlight Valley Tillite and Ranford Formation, *Palaeopascichnus linearis* (Hawco et al., 2019; Lan & Chen, 2012), and *Yangtziramulus zhangi* (Shen et al., 2009). *Yangtziramulus* is a large fractally branching structure (Figure 5f, g). *Yangtziramulus* has been considered a ‘diagnostic structure, possibly mediated by microbial activity’ (Conway Morris, 2006), and is similar to the more orthogonal peaked stromatolite *Thyssagites* (Walter, 1976).

In both the Chinese and Australian specimens, the body partitions are encrusted with carbonate that extend deep below the surface. *Yangtziramulus* has irregular fractal architecture also generally similar to less-peaked *Fractofusus* (Gehling & Narbonne, 2007) and *Beothukis* (Taylor et al., 2019). Similarly, it may have been an extinct organism of uncertain affinity within Class Vendobionta, variously interpreted as microbial colonies, xenophyophore foraminifers, fungi, lichens, sponges, cnidarians, bilaterians, or stem metazoan (Retallack & Broz, 2020). In the Ediacaran Shibantan Member of the Denying Formation in China, *Yangtziramulus* was regarded as a shallow subtidal sediment binder (Shen et al., 2009), but intertidal paleosols with oxidised surfaces (A horizon) and gleayed pyritic subsurface (Bg) horizons have been observed in the Shibantan Member (Retallack, 2014). Other evidence for intertidal or supratidal exposure of some horizons of the Shibantan Member comes from strong correlation of $\delta^{18}$O and $\delta^{13}$C values of its carbonates (as in Figure 10a, b), and a divergent mix of local oxic (low Fe and Mn, Ce/Ce*, Y/Ho, V/Cr) and anoxic (V/Sc, Ni/Co–, V/(V + Ni)) chemical indicators (Duda et al., 2014; Wan et al., 2020). In contrast, *Yangtziramulus* in the upper Moonlight Valley Tillite (Figure 5e–g) is on the surface of red Galadil paleosols, suggesting that it may have been a supratidal or terrestrial organism.

In the Donkey Creek outcrop of the Ranford Formation (Figure 6b), Lan and Chen (2012) reported *Palaeopascichnus* sp. indet, which can now be identified as *Palaeopascichnus linearis* (Kolesnikov et al., 2018). At first *Palaeopascichnus* was interpreted as a meandering trail trace fossil (Glaessner, 1969; Parcha & Pandey, 2011), alga (Haines, 2000), faecal string (Jensen, 2003), stratiform stromatolite (Runnegar, 1995), or xenophyophore foraminifer (Kolesnikov et al., 2018; Sellacher et al., 2003), but recent work has largely concluded that it was a chambered protozoan body fossil, distinct from modern xenophyophores (Antcliffe et al., 2011; Dong et al., 2008; Gehling & Droser, 2009; Hawco et al., 2019). *Palaeopascichnus* has been widely assumed to have been marine (Antcliffe et al., 2011; Hawco et al., 2019), and the species *P. delicatus* had a boron content compatible either with marine or terrestrial salt lakes (Retallack, 2020). In the Ranford Formation, *Palaeopascichnus* is in the same sequence as Thamberalg paleosols of hypersaline playa lakes (Figure 6b), but in lacustrine sediments rather than in the paleosols.

A variety of fossils and MISS are known from the Ranford Formation and upper Moonlight Valley Tillite (Lan & Chen, 2012, 2013). Some of these from the Ranford Formation are ridged surfaces with disconnected fragments (Figure 4e), comparable with aquatic microbial mats and the trace fossils *Rugalichnus matthewi* (Stimson et al., 2017). These undulose fabrics contrast with surfaces from the upper Moonlight Valley Tillite showing a patchwork of cracking and healing called ‘old elephant skin’ or *Rivularites repertus* (Retallack, 2013; Retallack & Broz, 2020). ‘Old elephant skin’ refers to a complex pattern of cracks, but there is also evidence of push-up ridges and domes, and thus alternating tension and compression. These textures are identical to surfaces of biological soil crusts (Belnap, 2003) and microbial earths (Figure 11d).

Finally, there is chemical and petrographic evidence for life in these paleosols. Thin-sections show lack of lamination in paleosol surface horizons (Figures 3 and 5f, g), perhaps owing to vertical thread-forming microbial consortia (Retallack, 2012a). All the paleosols also show significant loss of phosphorus (Figure 8), another indication of a microbial earth soil (Neaman et al., 2005a, 2005b).

Conclusions

Ediacaran paleoenvironments and paleosols of the Ranford Formation are reconstructed in Figure 12. This study outlines diagnostic paleosol criteria of pedogenic cracking patterns, sand crystals, interflag sandstone laminae, mineral weathering trends, chemical weathering trends, tau analysis, stable isotopic correlation and biotic microstructures. Once paleosols are recognised implying formation within the critical zone transitional between earth and air, their ancient environments may be interpreted. The enigmatic ornamental zebra rock can now be seen as a lowland seasonally waterlogged, redoximorphic soil (Wajing pedotype). Ediacaran paleosols of north Western Australia are evidence of arid and cool to cold temperate climates. Paleosols also reveal landscapes with moderately developed soils and stable ecosystems, as well as weakly developed and early successional soils and communities. These communities of scattered, ground-hugging, sessile large organisms would have appeared similar to modern polsterlands of alpine, polar and extreme deserts.

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No potential conflict of interest was reported by the author(s).

Data availability statement
Supplementary data that support the findings of this study are openly available from the University of Oregon at https://doi.org/10.7910/DVN/OXAFZI

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