Neoproterozoic Snowball Earth extent inferred from paleosols in California

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ABSTRACT

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Gelisol paleosols with sand wedges and sorted stone stripes are reported from the early Cryogenian (717–659 Ma), Surprise Diamictite Member and Sourdough Limestone Member of the Kingston Peak Formation in Redlands Canyon, western Panamint Range, California. The Surprise Diamictite was thus not entirely marine, although glaciomarine sediments and tectonically induced, mass wasting deposits, may be present in other parts of the Kingston Peak Formation. Sand wedge and stone stripe paleosols are evidence of local ice–free land with frigid continental climate at paleolatitude as low as $8 \pm 4^{\circ}$ from paleomagnetic studies of the Surprise Diamictite. The Sturt glaciation was a dramatic global cooling, but not a global snowball. Bare ground of landslides, alluvial fans, till and loess with mineral nutrients, and microtopographic shelter for complex life on land would have been important for survival of life on Earth from glacial destruction.

Key-words-Sand wedge, Sorted stone stripe, Paleosol, Cryogenian, Death Valley.

INTRODUCTION

TNCONFORMITIES and plausible tectonic megabreccias in the Kingston Peak Formation have been documented from the Kingston Range and southern Black Mountains well east of the study area (Fig. 1) as evidence for origin of Kingston Peak diamictites as debris flows and submarine landslide deposits from rift valley faulting (Prave, 1999; Kennedy & Eyles, 2021). In the Panamint Range (Fig. 1), conformable sequences of Kingston Peak Formation diamictites have been interpreted as glacial tillites (Miller, 1985), because of faceted and striated boulders, and lonestones in shaley facies. Isotopic chemostratigraphy and new radiometric dating show that the Kingston Peak diamictites date to the Cryogenian "Snowball Earth", a time of global cooling when even tropical regions were glaciated (Kirschvink, 1991). The extreme or "Hard Snowball" hypothesis of global glacial cover (Hoffman & Schrag, 2002; Pierrehumbert, 2005) would have imperiled life on the planet, contrary to the Cryogenian fossil record (Moczydłowska, 2008). An alternative is a "Soft Snowball" or "Waterbelt" hypothesis allowing a tropical seaway for survival of marine life (Runnegar, 2000; Pierrehumbert et al., 2011). This study examines a third alternative of life surviving on land, as well as the question of glacial versus non-glacial diamictites, from newly recognized periglacial paleosol structures, as constraints on paleoclimate during the Cryogenian in tropical latitudes.

This is not the first time periglacial structures have been reported from Cryogenian paleosols, which are widespread in South Australia (Williams, 1986; Retallack *et al.*, 2014), British Isles (Kilburn *et al.*, 1965; Spencer, 1971, 1985; Johnston, 1993), Sweden (Kumpulainen, 2011), and Mauretania (Deynoux, 1982). Detailed paleoclimatic modelling allows not only limited waterways, but bare rock mountains, sedimentary loess and till plains, and cryoconite dustings of glacial ice during the Cryogenian (Benn *et al.*, 2015; Hoffman *et al.*, 2017). These refugia would have allowed survival of life on land, which was substantial enough to consume CO_2 for biomass and silicate weathering to initiate Snowball Earth in the first place (Retallack, 2021, 2023; Retallack *et al.*, 2021a).

Sand wedges and sorted stone stripes of periglacial patterned ground are but two of many structures indicative of particular frigid paleoclimatic conditions (Williams, 1986). Such large structures are well preserved in paleosols, and obvious guides to past subaerial exposure in the field (Krull, 1999; Retallack, 2022a). Frigid soils of Antarctica (Campbell & Claridge, 1987; Bockheim, 2015) and Arctic regions (Tedrow, 1966; Ugolini, 1986) have limited, but measurable, geochemical and petrographic differentiation, and develop much more slowly than temperate and tropical soils. Petrographic and geochemical data on periglacial paleosols are showing similar trends (Retallack, 2022a; Retallack *et al.*, 2015), as also explored here.

GEOLOGICAL BACKGROUND

The Kingston Peak Formation of diamictites, sandstone and dolostone is highly variable in thickness, up to 450 m thick (Miller, 1985), and well exposed in the Panamint Range west of Death Valley, California (Figs 1-2). The diamictites fill valleys into dolostones of the Tonian (775 Ma), Beck Spring Formation, and Stenian (1087 Ma), Crystal Spring Formation, and gneisses of Paleoproterozoic (1400-1700 Ma) basement (Nelson et al., 2020). These valleys have been interpreted as failed rift valleys or fault-bounded grabens (Williams et al., 1974), but glacially incised valleys with considerable topographic relief may equally explain why 1700 m of Tonian and Stenian rocks were removed so that the Kingston Peak Formation unconformably overlies Paleoproterozoic crystalline basement in the central Panamint Range (Miller, 1985). The valley fill of the Kingston Peak Formation includes sandy Limekiln Spring Member, then Surprise Diamictite Member, then Sourdough Limestone Member overlain by South Park Member, further divided into sandstone of the Middle Park Submember, then conglomerate of Mountain Girl Submember, dolostone of the the Thorndike Submember, and diamictite of the Wildrose Submember (Nelson et al., 2021). This study measured a section through most of the Kingston Peak Formation, but identified candidate periglacial paleosols only in the Surprise Diamictite and lower Sourdough Limestone Members (Figs 3-4).

Geological dating of the Kingston Peak Formation by international correlation of carbon isotope anomalies is insecure (Nelson et al., 2021), because the anomalies represent sea level change with different timing at different locations with different elevations (Retallack et al., 2021a). However, glacioeustatic correlation has been used to date the Surprise Diamictite Member with the Sturt Glaciation of South Australia (Prave, 1999; Kennedy & Eyles, 2021), and diamictite of the Wildrose Submember with the last Cryogenian glacial advance, often called Marinoan (Prave, 1999; Kennedy & Eyles, 2021), but better termed the Elatina Glaciation (Williams et al., 2008). Fortunately, there are U-Pb radiometric dates for detrital zircons in the Kingston Peak Formation: a cluster of grains dated 705.4 ± 0.3 Ma in the middle of the Limekiln Spring Member, and 651.7 ± 0.6 Ma for the upper Thorndike Submember (Nelson et al., 2020). These dates are compatible with onset of the Cryogenian diamictites of the Surprise Member at 717 Ma, the Sourdough Limestone and Middle Park Members from 659-640 Ma, and diamictites of the Wildrose Submember from 640–635 Ma (Halvorsen *et al.*, 2020).

Plate tectonic reconstructions show that southeastern California was equatorial during the Cryogenian, on the northern margin of an east–west oriented Laurentian craton (Trindade & Macouin, 2007; Scotese, 2021). The paleolatitude



Fig. 1—Cryogenian Kingston Peak Formation sample sites in Redlands Canyon on a simplified geological map emphasizing the distribution of Ediacaran and Cryogenian rocks (data from Wright *et al.*, 1976).



Fig. 2—Kingston Peak Formation outcrops in Redlands Canyon, Panamint Range, California: (A) Overview of the measured section from the southeast (Fig. 3); (B) Thin Nawogan pedotype on Wepaxaku pedotype at 30 m; (C) Wepaxaku pedotype at 172 m with sand wedge at arrow; (D) Masonih pedotype at 80 m, with vertical pebble in sorted stone stripe at arrow; (E) Large clast with weathering rind in diamictite at 255 m. of the Cryogenian Kingston Peak Formation has been estimated at $8 \pm 4^{\circ}$ (Evans & Raub, 2011).

Evidence against pervasive late diagenetic potash metasomatism (Novoselov & de Sousa Filho, 2015) includes low and highly variable K₂O, at 0.29–3.47 wt% (Table 1). Observed overburden above the Kingston Peak Formation is at least 6 km (Corsetti & Kaufman, 2003), and an additional 1.5 km of Permian to Jurassic strata are shown in regional cross sections by Wernicke et al. (1988). Burial compaction expected for 6.0-7.5 km can be calculated as 53-54 %, using a formula from Sheldon and Retallack (2001) with 0.51 solidity, 0.49 initial porosity, and 0.27 fitting constant. The overlying Noonday Formation dolostones were locally affected by Mississippi Valley style lead-zinc mineralization from early Paleozoic migration of hydrothermal brines (Carlilse et al., 1954; Church et al., 2005). The Queen of Sheba Mine, 18 km east of the Redlands Canyon outcrops examined for this study, is a replacement zone mainly of galena and chalcopyrite, only 12 m thick and 123 m long (Morton, 1965). This region also has local polymetallic skarn mineralization associated with Mesozoic granitic intrusions (Newberry, 1987; Newberry et al., 1991). A contact metamorphic aureole around the Jurassic (151 Ma), Manly Peak quartz monzonite pluton 6 km to the southeast of the measured section (Andrew, 2022) is similar to other highly metamorphosed rocks in the Panamint Range (Labotka et al., 1985, 2000), but the line of section documented here is no more metamorphosed than greenschist facies, dipping gently, and sparsely faulted (Prave, 2000; Nelson et al., 2021). This degree of alteration is well below that observed to create carbon isotope exchange in polymetamorphic rocks elsewhere in the Panamint Range (Bergfeld et al., 1996). The measured section is from an area of coherent Cryogenian stratigraphy between the Goldberg Thrust to the west and the Manly Peak Rim Fold and metamorphic aureole to the east (Andrew, 2022).

MATERIAL AND METHODS

This study examined and sampled a geological section in the north wall of Redlands Canyon, 1 km west of the junction of Redlands and Wood Canyons, in the southwestern Panamint Range (N35.94126° W117.13336°) under U.S. National Park permit DEVA–2008–SCI–0034. A long geological section was measured (Fig. 3), including detailed measurements of representative beds (Fig. 5).

Distinctive, repeated bed types in the Surprise and lowest Sourdough Limestone Members of the Kingston Peak Formation were given nongenetic field names, Wepaxaku, Masonih, and Nawogan, using words for "cut up", "fold", and "striped" in the Shoshoni Native American language (Shoshoni Language Project, 2019). Wepaxaku profiles are 60 cm of breccia, with wide tapering sandstone dikes extending down from the upper surface and nodules of carbonate in the subsurface (Fig. 2B–C). Masonih profiles are 40 cm of conglomerate with distinct inverted chevrons of the long axes of grains (Fig. 2D). Nawogan are thin (10 cm) cracked and laminated profiles of siltstone and sandstone. Examples of each kind of profile were selected for detailed petrographic (Fig. 4) and geochemical study, including point–counted proportions of sand, silt and clay, and of different mineral components, as well as molar weathering ratios (Fig. 5).

Samples were collected for major element geochemical analysis x-ray fluorescence spectroscopy with British Columbia granodiorite as a standard (Table 1) by ALS Chemex, of Vancouver, Canada, who also measured ferrous iron by Pratt titration. Petrographic thin sections were point-counted to determine mineral composition and grain size using a Swift automated stage and Hacker counting box (Tables 2, 3). Both composition and size counts were 500 points, which has accuracy of 2% for common components (Murphy, 1983).

WAS THE SURPRISE MEMBER A MARINE OR TERRESTRIAL DIAMICTITE?

Both marine and terrestrial diamictites have chaotically deposited breccia and conglomerates (Miller, 1985; Kennedy & Eyles, 2021). One way to distinguish them is by included marine fossils, but these are sparse and microscopic in Cryogenian deposits, such as the Kingston Peak Formation (Corsetti et al., 2003, 2006). Another distinction is interbedded paleosols, especially an array of periglacial structures, such as ice and sand wedges (Krull, 1999; Retallack, 1999a, b). Periglacial paleosols are evidence of terrestrial exposure, but their parent diamictites may still have been marine due to large sea level changes or isostatic rebound commonly associated with ice ages (Boulton, 1990). The South Park Member between the two thick diamictites of the Kingston Peak Formation may be marine and include turbidites, as proposed by Nelson et al. (2021). Furthermore, there are lacustrine or marine pillow basalts within upper the Kingston Peak Formation (Labotka et al., 1980; Miller 1985). Precambrian paleosols lack fossil roots and other features of Phanerozoic paleosols, but periglacial soil features are well preserved in them (Williams, 1986; Williams et al., 2008; Retallack et al., 2015; Retallack, 2022a). Nevertheless, subtle features of grain size variation and geochemical differentiation are worth examining in Precambrian periglacial paleosols as well (Retallack, et al., 2015; Retallack, 2022a). The following paragraphs explore each criterion for distinguishing marine from non-marine tillites, first with a description and then an interpretation, for the Surprise Diamictite Member of the Kingston Peak Formation in the Panamint Range.

Tapering sandstone clastic dikes

Clastic dikes filled with sand 50-60 cm deep taper sharply from 30 cm at the surface to only a few mm at their base. The dikes are spaced at intervals of 1-1.5 m along the

Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	CaO	MgO	Na ₂ O	K ₂ O	P ₂ O ₅	SrO	BaO	LOI	TOTAL	g.cm ⁻³
R5861	38.4	0.34	5.23	4.81	0.26	0.1	23.1	3.08	0.03	1.12	0.09	0.01	0.02	23.00	99.34	2.5128
R5862	49.7	0.49	7.43	3.31	0.45	0.09	15.1	4.37	0.03	1.78	0.12	0.01	0.04	18.05	100.53	2.2828
R5863	40.2	0.08	0.56	3.29	0.5	0.21	28.4	1.12	0.02	0.16	0.05	0.01	0.01	25.00	99.05	2.4768
R5864	69.7	0.33	5.21	2.75	0.51	0.06	10.45	0.87	0.03	1.24	0.08	<0.1	0.02	10.3	101.05	2.3882
R5865	47.9	0.09	1.20	3.62	0.32	0.1	25.2	0.52	0.02	0.26	0.03	0.01	0.01	21.1	100.06	2.3767
R5866	61.4	0.08	10.95	3.18	1.68	0.11	7.25	3.46	0.19	3.42	0.13	0.01	0.06	8.49	99.29	2.6495
R5874	79.7	0.13	2.25	2.85	0.71	0.06	7.64	0.34	0.02	0.62	0.04	<0.1	0.01	6.12	99.79	2.4222
R5875	82.7	0.04	0.77	1.18	0.65	0.03	4.54	2.52	0.01	0.22	0.03	<0.1	< 0.1	6.43	98.48	2.3403
R5876	24.8	0.04	3.26	5.20	0.84	0.08	28.7	5.81	0.03	1.06	0.17	0.01	0.02	29.6	99.12	2.4513
R5877	43.8	0.64	10.00	4.93	3.66	0.11	11.95	6.45	0.04	3.3	0.14	0.03	0.05	18.50	99.95	2.5782
R5878	65.4	0.87	9.07	5.23	0.64	0.06	7.25	1.52	0.05	2.32	0.22	0.01	0.06	9.11	101.18	2.4751
R5879	90.7	0.64	0.99	1.4	1.03	0.03	2.85	1.17	0.02	0.29	0.14	<0.1	<0.1	3.74	101.38	2.5568
R5880	49.4	0.6	9.17	4.22	1.23	0.08	12.05	4.66	0.04	2.9	0.16	0.01	0.05	16.20	99.55	2.5404
R5881	56.4	0.6	11.05	4.64	0.84	0.08	10.5	1.57	0.06	2.46	0.14	0.02	0.06	12.70	100.45	2.5204
errors	2.705	0.06	0.825	0.395		0.22	0.18	0.11	0.13	0.025	0.030				0.35	0.003

Table 1-Chemical composition (wt %) from XRF of the Kingston Peak Formation.

Note: Errors are from 10 replicate analyses of the standard, CANMET SDMS2 (British Columbia granodioritic sand) and ten bulk density measurements.

Table 2-Grain-size from point counting thin sections (500 points) of Kingston Peak Formation.

Pedotype	Hoz	No.	% sand	% silt	% clay	Textural class	Grain fabric	Plasmic fabric
Nawogan	above	R5860	30.2	49.2	20.6	loam	agglomeroplasmic	argillasepic
Nawogan	А	R5861	39.8	29.3	31.0	clay loam	porphyroskelic	mosepic
Wepaxaku	Af	R5862	37.8	37.6	24.6	loam	agglomeroplasmic	bimasepic
Wepaxaku	Bf	R5863	35.4	34.9	29.4	clay loam	agglomeroplasmic	mosepic
Wepaxaku	Bf	R5864	34.2	41.2	24.6	loam	agglomeroplasmic	calciasepic
Wepaxaku	Bk	R5865	36.6	43.0	20.4	loam	agglomeroplasmic	calciasepic
Wepaxaku	С	R5866	30.6	55.6	13.8	silt loam	agglomeroplasmic	insepic
Mahonih	above	R5867	31.2	49.6	19.2	loam	agglomeroplasmic	insepic
Mahonih	A	R5868	34.8	39.2	26.0	loam	agglomeroplasmic	mosepic
Mahonih	Bf	R5870	37.0	43.0	20.0	loam	porphyroskelic	mosepic
Mahonih	Ck	R5871	23.8	57.6	18.6	silt loam	agglomeroplasmic	calciasepic
Mahonih	C	R5872	24.2	64.4	11.2	silt loam	agglomeroplasmic	insepic
Mahonih	above	R5873	30.6	54.2	15.2	silt loam	unistrial	silasepic
Mahonih	A	R5874	35.8	42.6	21.6	loam	agglomeroplasmic	insepic
Mahonih	Bf	R5875	43.6	37.0	19.4	loam	agglomeroplasmic	insepic
Mahonih	Ck	R5876	56.2	29.8	14.0	sandy loam	unistrial	calciasepic
Wepaxaku	above	R5877	20.8	57.0	22.2	silt loam	porphyroseklic	mosepic
Wepaxaku	Af	R5878	23.4	53.2	23.4	silt loam	agglomeroplasmic	insepic
Wepaxaku	Bf	R5879	22.0	56.0	22.0	silt loam	porphyroskelic	insepic
Wepaxaku	Ck	R5880	39.6	42.0	18.4	loam	granular	calciasepic
Wepaxaku	С	R5881	35.2	50.6	14.2	silt loam	agglomeroplasmic	insepic
Wepaxaku	С	R5882	35.3	52.2	12.6	silt loam	agglomeroplasmic	insepic



Fig. 3—Measured section of the Kingston Peak Formation in Redlands Canyon, showing identified paleosols and their development, calcareousness, and Munsell color hue. Paleosols were named using words for fold (Masonih), striped (Nawogan) and cut up (Wepaxaku) in the Shoshoni Native American language (Shoshoni Language Project, 2019).



Fig. 4—Petrography of Kingston Peak Formation in thin section scans, all in plane light and oriented vertical to bedding: (A) Contact between Nawogan (above) and Wepaxaku pedotype A horizon (below); (B) A horizon of Wepaxaku pedotype and overlying sandstone; (C) A horizon of Masonih pedotype and overlying sandstone; (D) Desiccation cracking of Masonih K horizon; (E) Gypsum in By horizon of Wepaxaku pedotype; (F) Displacive fabric in micrite of Ck horizon Wepaxaku pedotype; (G) Etched clast in A horizon of Masonih pedotype; (H) Clast with weathering rind (at arrow) in C horizon of Wepaxaku pedotype. Specimen numbers in the Condon Collection of the Museum of Natural and Cultural History, University of Oregon are R5861 (A), R5877 (B), R5868 (C), R5876 (D), R5865 (E), R5875(F), R5870 (G), R5881(H).

outcrop (Fig. 2B–C). They do not extend into the overlying diamictite, but are truncated at that erosional surface. Clastic dikes were only seen in profile, and there were no platform outcrops to reveal their likely polygonal form in plan within this steep canyon wall (Fig. 2A). They were not filled passively by sediment in horizontal layers, but instead show massive, and vertically oriented layers of sand and granules, as if open and closed multiple times (Fig. 2C). Their fill is complex, including suspended, vertically oriented, soft–sediment clasts of their walls.

The strongly tapering clastic dikes are not tension gashes, nor landslides, nor fault cracks, because they lack slickensides or other evidence of shear, and are truncated by overlying beds (Fig. 2B–C). Nor is there any streaming, soft sediment convolution, fluidization of the fill, or lateral spread as would be expected if these were water release structures, clastic dikes, or sand blows created by upward injection (Cox *et al.*, 2007; Maurer *et al.*, 2019). There is no isoclinal deformation as in locally highly metamorphosed rocks (Labotka *et al.*, 1985; Newberry, 1987; Newberry *et al.*, 1991), nor is there



Fig. 5—Petrographic and geochemical data on the Kingston Peak Formation in Redlands Canyon, California. Data displayed include Munsell hue, calcareousness determined by acid application, grain size and mineral composition determined by point counting petrographic sections, and molecular weathering ratios of major and trace elements determined by XRF analysis. Individual sampled levels are shown in Fig. 3.

mineralization as in local hydrothermal veins (Carlilse *et al.*, 1954; Morton, 1965; Church *et al.*, 2005).

These distinctive clastic dikes match modern sand wedges in their upward flare, regular spacing, and vertically oriented fill (Black, 1973, 1976a, b, 1982). Furthermore, these sand wedges fulfill six other criteria outlined by Black (1976a). First, there is other evidence of limited weathering (alumina/ bases and Ba/Sr molar ratios of Fig. 5). Second, erratic FeO/ Fe_2O_2 (Fig. 5) is independent evidence of local waterlogging. Third, the spacing of wedges is wider for deeper wedges (Fig. 2B-C). Fourth, significant alumina enrichment at the surface (Fig. 5) is evidence of weathering by rain on soil with little or no summer snow cover. Fifth, edges are locally upturned by pressure (Fig. 2C). Sixth, slumped blocks are included in the fill (Fig. 2C). Sand wedges can be distinguished from ice wedges, which have passive, sometimes horizontally bedded fill, and flare more widely toward the top (Leffingwell, 1915; Kokelj et al., 2007; Raffi & Stenni, 2011).

Inverted chevron deformation and jacked pebbles

Some conglomerate beds 30 cm thick are characterized by inverted chevron alignment of pebbles with wavelength of 50 cm (Fig. 2D). Pebbles lean into the peak of the chevron on the flanks, and some pebbles are vertical in the core of the chevron. The chevrons are entirely within the conglomerate bed, and are elongate structures best seen when outcrop intersects them at a right angle.

This distinctive deformation is similar to sorted stone stripes and stone jacking of periglacial soils (Kessler & Werner, 2003; van Vliet-Lanoë, 2010). The high pebble content and elongate form distinguishes these forms from earth hummocks (thufur of van Vliet-Lanoë, 1991; Grab, 2005). The angularity of chevron form is distinct from curved convolute lamination produced by earthquakes or foundering of quicksand beds (Wheeler, 2002; Owen, 2003), and earthquake induced sand blows and quicksand are finer in grain size. The chevron form is similar to folding of some metamorphic rocks (Ramsay, 1974; Bastida et al., 2007), but in this case is entirely within thin beds, rather than distributed throughout the outcrop. Wedged opening of carbonate layers (Fig. 4D), may also represent frost disruption (van Vliet-Lanoë, 1991, 2010). Although Fig. 4D may have the superficial appearance of desiccation or synaeresis cracks (Plummer & Gostin, 1981; Weinberger, 2001), it differs in its complex lateral system of branching fractures including breccias indicative of brittle failure.

Granulometry

Point counting of the non-gravel matrix of selected beds showed very little clay, but much silt: usually over 50%, ranging from 29–64% by volume (Table 2). The silt grains are angular (Fig. 4A–C), and include many dolomite grains, with lesser feldspar and rock fragments (Fig. 5).

Silt-rich beds of the Surprise Diamictite may have been loess, deposited on land by wind, because they are similar in grain size, angularity, and texture to Quaternary Peoria Silt (Swineford & Frye, 1951; Pye & Sherwin, 1999), which can have 31-42% limestone and dolomite clasts (Fisk, 1951; Grimley et al., 1998). Chinese loess of the Central Loess Plateau is also similarly calcareous (Nugteren et al., 2004; Sun et al., 2004). Carbonate grains within Pleistocene Peoria Loess were abraded by glaciers from Paleozoic limestones and dolostones in Illinois and Wisconsin, but the source of carbonate silt-size clasts for the Surprise Diamictite would have been glaciated Mesoproterozoic Beck Spring and Crystal Spring marine dolostones with stromatolites (Wright et al., 1976; Corsetti & Kaufman, 2003). Much eolian silt may have fallen into the sea before landscapes were stabilized by plants, but the Surprise Diamictite examples are grain supported (Fig. 4B) and poor in clay (Fig. 5). Cambrian marine siltstones in contrast are clay supported (Dalrymple et al., 1985).

Within-bed petrographic and chemical composition

Point counting of thin sections of individual beds shows clay enrichment toward the top of beds, usually at the expense of dolomite and rock fragments (Fig. 5). In addition to clay there also are significant amounts (from 8 to 50 % by volume) of other silicate minerals, mainly quartz and feldspar. Some chemical weathering is revealed by etched clasts (Fig. 4G) and clasts with weathering rinds (Fig. 4H). Carbonate is depleted toward the surface of the beds, and this is reflected in molar ratios of CaO+MgO/Al₂O₃. Nevertheless, the profiles have calcite (Fig. 4F) and gypsum (Fig. 4E) at depth. Leaching is modest as indicated by Ba/Sr molar ratios. Chemical reduction (gleization) from FeO/Fe₂O₃ ratios is modest in all profiles, and two of the three profiles are brown with hematite (Fig. 5).

Results of sedimentation and soil formation can be disentangled by tau analysis (Brimhall *et al.*, 1992), which separately resolves mole fraction mass transport $(\tau_{j,w})$ of a mobile element and mole fraction strain $(\varepsilon_{i,w})$ from an immobile element. These are calculated using equations 1 and 2, including bulk density (ρ 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).

$$\epsilon_{i,w} = \left[\frac{\rho_p C_{j,p}}{\rho_w C_{j,w}} \right] - 1 \qquad \qquad -\text{equation 1}$$

$$\tau_{j,w} = \left[\frac{\rho_w C_{j,w}}{\rho_p C_{j,p}}\right] \left[\epsilon_{i,w} + 1\right] - 1$$

-equation 2

Pedotype	Hoz	No.	% clay	% dolo- mite	% gypsum	% rock fragments	% feld- spar	% mica	% quartz	% opaque
Nawogan	above	R5860	22.2	3.6	0	1.8	28.6	4.4	29.6	3.0
Nawogan	А	R5861	30.0	8.6	0	7.6	27.2	0.6	22.8	3.2
Wepaxaku	Af	R5862	27.2	10.0	0	1.4	27.8	3.2	27.2	3.2
Wepaxaku	Bf	R5863	31.8	3.4	0	5.0	30.0	1.0	25.8	3.0
Wepaxaku	Bf	R5864	26.0	10.4	2.0	6.2	24.2	3.0	24.0	3.2
Wepaxaku	Bk	R5865	20.2	12.0	8.6	16.4	22.8	0.2	18.8	1.0
Wepaxaku	С	R5866	16.6	1.2	0	0.6	31.2	17.2	30.0	3.2
Mahonih	above	R5867	21.0	5.2	0	2.6	30.2	10.0	28.2	2.8
Mahonih	А	R5868	25.2	7.0	0	5.2	30.0	0.8	30.4	1.4
Mahonih	Bf	R5870	21.8	5.2	0	4.0	33.8	0.6	32.2	2.4
Mahonih	Ck	R5871	22.0	8.0	0	0.6	39.4	3.0	25.4	1.6
Mahonih	С	R5872	10.6	0.6	0	0.8	34.8	22.6	27.8	2.8
Mahonih	above	R5873	15.4	1.4	0	1.0	34.8	17.2	27.8	2.4
Mahonih	А	R5874	24.8	3.2	0	1.4	33.0	2.2	33.8	1.6
Mahonih	Bf	R5875	22.2	13.8	0	3.2	27.8	2.8	28.2	2.0
Mahonih	Ck	R5876	14.8	21.2	0	2.2	29.2	1.0	28.8	2.8
Wepaxaku	above	R5877	22.8	1.0	0	1.4	33.8	12.0	26.4	2.6
Wepaxaku	Af	R5878	22.2	1.0	0	1.2	31.6	11.8	27.8	4.4
Wepaxaku	Bf	R5879	22.4	0.4	0	1.2	34.8	12.4	27.6	1.2
Wepaxaku	Ck	R5880	17.4	21.2	0	2.0	26.0	10.6	21.4	1.4
Wepaxaku	С	R5881	14.6	0.6	0	1.4	34.6	18.6	28.6	1.4
Wepaxaku	С	R5882	11.6	0.4	0	2.0	33.6	26.8	24.4	1.2

Table 3—Mineral content from point counting thin sections (500 points) of Kingston Peak Formation.

Table 4—Pedotypes and diagnosis for Kingston Peak Formation, California.

Pedotype	Shoshone meaning	Diagnosis	USDA (Soil Survey Staff, 2014)	FAO (1975, 1978)	CLASSIC (Stace <i>et al.</i> , 1968)	AUSTRALIAN (Isbell, 1996)
Masonih	fold	Silty cracked yellow surface (A) over ridged breccia (Bw) and laminar calcrete (K)	Turbel	Gelic Cambisol	Lithosol	Clastic Rudosol
Nawogan	striped	Silty cracked yellow surface (A) over bedded yellow calcareous siltstone (C)	Fluvent	Eutric Flu- visol	Earthy Sand	Stratic Rudosol
Wepaxaku	cut up	Silty cracked yellow surface (A) over sandy fractured breccia (Bw) and wide micritic nodules (Bk)	Turbel	Gelic Cambisol	Lithosol	Clastic Rudosol

Table 5—Interpretation of pedotypes for Kingston Peak Formation, California.

Pedotype	Paleoclimate	Ecosystems	Parent Material	Paleotopography	Time for formation (years)
Masonih	Frigid (MAT–4.4 to–5.6° C) and arid (MAP 226–303 mm) with sorted stone stripes	Microbial earth	Till with capping loess	Till moraines	2,000
Nawogan	Not diagnostic of climate	Unknown	Loess	Loess mantle	100
Wepaxaku	Frigid (MAT–6° C) and arid (MAP 262 mm) with sand wedge patterned ground	Microbial earth	Till with capping loess	Till plain	5,000-7,500



A. mass transfer with strain

Fig. 6—Tau analysis of Kingston Peak Formation beds with strain (A) and with depth (B). Tau is the change in mass or volume with respect to a stable constituent (Ti used here) using equations 1 and 2. These arrays of lost volume and mass are typical for weakly developed soils.

Weathering in soils and paleosols produces negative strain ($\varepsilon_{i,w} < 0$), or loss of mass, as well as negative mass transfer ($\tau_{j,w} < 0$), or loss of nutrient and other elements, both within the lower left quadrant of a crossplot (Fig. 6). Sediment accumulation and diagenetic alteration is usually the opposite, positive strain and mass transfer is in the upper right quadrant of a crossplot (Retallack *et al.*, 2021b). Precambrian paleosols (Retallack & Mindszenty, 1994; Driese *et al.*, 2011) show patterns of tau depletions comparable with Cenozoic paleosols (Bestland *et al.*, 1996; Sheldon & Tabor, 2009), and modern soils (Chadwick *et al.*, 1990; Hayes *et al.*, 2019).

All samples analyzed for this study are within the soil rather than sediment quadrant, and the degree of soil

development in the Wepaxaku loam is greater than for the Wepaxaku silt loam and Masonih loam bed (Fig. 6). In all beds, the depth and intensity of chemical weathering is modest (Fig. 6), even when compared with Ediacaran paleosols (Retallack, 2011, 2013).

Stable isotopic covariation

Micritic calcite nodules low within Wepaxaku silt loam analyzed by Nelson *et al.* (2021) show strong covariance of δ^{13} C and δ^{18} O (Fig. 7A). Furthermore, they show displacive fabric as evidence of low confining pressure when calcite precipitated (Fig. 4F). However, dolostone and altered



Fig. 7—Stable isotopic covariation of Kingston Peak Formation carbonates. (A) lack of correlation in marine–derived clasts of the Kingston Peak Formation, compared with correlated pedogenic carbonate of Kingston Peak Formation pedogenic carbonate, Ediacaran paleosols of South Australia (Retallack *et al.*, 2014), Cambrian Arumbera Formation at Ross River (Retallack & Broz, 2020), Ordovician paleosols of Pennsylvania (Retallack, 2015b), and Silurian paleosols of Pennsylvania (Retallack, 2015a); (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, and Clino Island, Bahamas, Melim *et al.*, 2004); (E) Holocene (open circles) and Ordovician (open squares) unweathered marine limestones (Veizer *et al.*, 1999) and Early Cambrian (closed circles), Ajax Limestone, South Australia (Surge *et al.*, 2001) and Oligocene, Porter, Washington (Peckmann *et al.*, 2002). Slope of linear regression (m) and coefficients of determination (r²) show that carbon and oxygen isotopic composition is significantly correlated in soils and paleosols, but not in other settings.



Fig. 8—Reconstructed Cryogenian landscape and paleosols of Death Valley, California.

limestone of the Thorndike Dolostone (symbol + in Fig. 7A) in the same area shows no such correlation.

Pedogenic carbonate and karst weathering of limestone creates a high correlation of δ^{13} C and δ^{18} O, because weathering is a form of early diagenesis, between deposition and burial (Retallack, 2016). Unaltered lacustrine and marine limestones and sea-shells (Fig. 7E) show no hint of correlation (Surge et al., 1997; Veizer et al., 1999). Significant (P>0.05) covariance is widespread within pedogenic carbonate of Holocene soils (Fig. 7B) in China (Huang et al., 2005) and Arizona (Knauth et al., 2003), and even in metamorphically altered paleosols (Retallack, 2015a, b; Retallack et al., 2014; Retallack & Broz, 2020). Less significant are correlations (Fig. 7C-D) in soil carbonate crusts (Knauth et al., 2003), and in marine limestone altered by meteoric water (Lohmann, 1988; Melim et al., 2004). Isotopic covariance has been recorded from seasonally dry lakes (Talbot, 1990), from weathering during the dry season. Methanogenesis creates a distinctive pattern of constant δ^{18} O but varied δ^{13} C (Fig. 7F) in calcite of marine methane seeps (Aiello et al., 2001; Peckmann et al., 2002), and in siderite of wetland paleosols (Ludvigson et al., 1998, 2013).

Selection for light isotopologues of CO_2 during photosynthesis on land is the best known mechanism for correlation of $\delta^{13}C$ and $\delta^{18}O$ in pedogenic carbonate (Farquhar

& Cernusak, 2012; Broz *et al.*, 2021). However in the ocean or lakes, oxygen of water overwhelms carbon during carbonate precipitation (Retallack, 2016). Kinetic evaporative effects in narrow spaces of soils (Ufnar *et al.*, 2008) involve fractionations much less than stomatal conductance and enzymes such as rubisco (Retallack, 2016), and carbonic anhydrase (Chen *et al.*, 2018). Soil tortuosity is dwarfed as an effect by strong δ^{13} C and δ^{18} O covariance within respired soil CO₂ (Ehleringer *et al.*, 2000; Ehleringer & Cook, 1998), and within plant cellulose (Barbour & Farquhar, 2000; Barbour *et al.*, 2002). Enzymatic control is most likely for Cambrian and Neoproterozoic paleosols which predate evolution of stomates (Fig. 7A). Metamorphism to greenschist facies in paleosols of the Juniata and Bloomsburg formations did not alter covariance (Fig. 7A; Retallack, 2015a, b).

From this perspective, Thorndike Dolostone stable isotope ratios are uncorrelated like marine or lacustrine carbonate (Fig. 7A). In contrast, micrite of the Wepaxaku silt loam has isotopic covariance like pedogenic carbonate, but other dolomite and calcite in the Thorndike Dolostone is uncorrelated like marine or lacustrine carbonates.

Interpretation of paleosols

Soilscape comparisons

As outlined in the previous paragraphs, the named and analyzed beds are distinct kinds or pedotypes of periglacial paleosols supporting interpretation of the Surprise Diamictite in the southern Panamint Range as a terrestrial tillite. Pedotypes of the Surprise Diamictite can be identified within modern soil classifications (Tables 4-5), such as those of the United States (Soil Survey Staff, 2014) and the Food & Agriculture Organization (1974, 1975, 1978), but Australian classifications do not have closely comparable frigid climate soils (Stace et al., 1968; Isbell, 1996). In the Food and Agriculture (1975) system these paleosols represent a map unit code Bx+Je. The most similar Russian soils (units Bx 6-3b of Food & Agriculture Organization, 1978) cover 10,816,000 hectares in floodplain and headwaters of the Olenek River, near Jerbogačon, Siberia, where climate is frigid and vegetation is pine (Pinus sibirica)-larch (Larix dahurica) taiga forest. Comparable Canadian soils (units Bx1-1a and Bx 2-1b of Food & Agriculture Organization, 1975) cover 6,449,000 hectares of marine sediments and glacial outwash west of Hudson Bay and north of Churchill, where climate is frigid and vegetation is mixed spruce (Picea mariana) taiga and tundra. Jerbogačon has a mean annual temperature of -6.7°C and mean annual precipitation of 323 mm, and for Churchill these values are -7.2°C and 435 mm (Müller, 1982). These analogs have mainly ice wedges, and few sand wedges, which are best known from Antarctica, not mapped in the Food and Agriculture Organization system. Sand wedges of Wepaxaku beds are matched by Anhyturbels and sorted stone stripes of Masonih beds are best matched by Haploturbels

in the Dry Valleys of Central Victoria Land, Antarctica (Bockheim, 2013; Bockheim & McCleod, 2015). These soils are hyperarid (45 mm mean annual precipitation) and frigid (temperatures ranging from averages of–37°C in July to +1°C in January, and mean annual temperature of–20°C) at Vanda Weather Station (Campbell & Claridge, 1987). Antarctic Dry Valley soils look barren of vegetation, but host microscopic mosses, lichens, nematodes, collembola, yeasts, and other microorganisms (Chan *et al.*, 2013).

Parent material

Sedimentary parent materials of Surprise Diamictite paleosols are little weathered boulders, loess, and glacial flour derived from dolostones of the Beck Spring and Crystal Spring formations, and granite and gneiss of Paleoproterozoic basement (Miller, 1985; Nelson *et al.*, 2020). Dolostones were more widely exposed than granitic basement at the base of deep glacial valleys (Miller, 1985), and may explain carbonate loess and other layers, but granitic rocks dominate the diamictite material. Some of the clasts have weathering rinds (Fig. 2E, 4H), in part from Cryogenian weathering. These would have been unconsolidated boulder clay at the time of soil formation.

Depositional setting

Limestone and dolostone in the Kingston Peak Formation has been assumed to be marine, and thus reflect marine isotopic reservoirs of carbon and oxygen (Miller, 1985; Prave, 1999; Kennedy & Eyles, 2021; Nelson et al., 2021). This study did find uncorrelated oxygen and carbon isotopic values in Thorndike dolostones and limestones, which may be marine or lacustrine, but also highly correlated isotopic values in the Sourdough Limestone suggestive of pedogenic carbonate (Fig.7A). Nor were any pedogenic features seen in the Middle Park Member, which may include marine or lacustrine turbidites from marine transgression between glacial advances, as interpreted by Miller (1985) and Nelson et al. (2021). Pillow basalt in the upper Surprise Diamictite in Pleasant Canyon is evidence that some parts of the Surprise Diamictite were marine or lacustrine (Labotka et al., 1980; Miller, 1985). One fault active during deposition of the upper Kingston Peak Formation has been documented (Prave, 1999), but there is little evidence of normally faulted grabens oriented east-west like paleocurrent directions (Wright et al., 1976): mapped faults are predominantly north-south in orientation (Labotka et al., 1980; Miller, 1985; Andrew, 2022).

Grabens have been inferred not from faults, but from strong thickness variations from one measured section to the next (Kennedy & Eyles, 2021). Mass wasting and fan deposits are best known in the Kingston Range well to the east of the current study (Kennedy & Eyles, 2021). In the Panamint Range there is evidence of glacial deposition from the generally conformable stratigraphic sequence, faceted and striated boulders, and lonestones in shaley facies (Miller, 1985; Nelson et al., 2021). Sand wedges, sorted stone stripes, and other evidence of paleosols (section 4) on the tops of diamictite beds are evidence that this was a frigid plain, presumably within a deep valley with mountainous margins, thus explaining dramatic local thickness variations, fanglomerates, and huge blocks of underlying formations (Miller, 1985; Prave, 1999). Like the modern Dry Valleys and lakes of Victoria Land Antarctica (Black, 1982; Campbell & Claridge, 1987; Bockheim, 2015), the Surprise Diamictite included polygonal patterned ground of sand wedges and lineations of sorted stone stripes (Fig. 8). These different periglacial structures did not necessarily correlate with glacial advance and retreat, because moisture distribution patterns may have maintained frigid ground rimmed by hanging glaciers from nearby peaks, as in the Dry Valleys of Antarctica (Bockheim, 2015).

Within this interpreted till plain, the Wepaxaku silty clay loam and Masonih loam profiles were well drained, with low ferrous/ferric iron ratios near the surface, but the Wepaxaku silt loam has erratic variation of this ratio (Fig. 5), perhaps due to surface waterlogging of frozen ground. Generally good drainage of the profiles is indicated by pedogenic carbonate (Fig. 7A), and evidence of silicate weathering from alkali and alkaline earth depletion and apatite weathering from phosphorus depletion (Fig. 6). Stone stripes rather than polygons develop on ground sloping at 10° to 30° (Kessler & Werner, 2003), such as Alpine moraines in the Dry Valleys of Antarctica (Bockheim, 2015). Also like comparable sand wedges of the Dry Valleys of Antarctica, periglacial patterned ground of the Kingston Peak Formation may have passed laterally into lacustrine and marine basins with ice shelves, thus explaining microfossils, pillow basalts and turbidites in overlying, underlying, and partly interbedded units (Labotka et al., 1980; Miller, 1985; Corsetti et al., 2003, 2006).

Duration of soil formation

Rates of ice and sand wedge growth in Antarctica ranged from 0.79 mm per year in the 1960s (Black, 1973) to 0.04 mm per year in the 1970s (Black, 1982). By this range of growth rates, sand wedges of the Wepaxaku silt loam represent 380–7500 years, and those of the Wepaxaku loam represent 253–5000 years. Such slow rates and long durations are comparable with Canadian and Siberian ages for large ice wedges (Fortier & Allard, 2004; Kokelj *et al.*, 2007; Streletskaya *et al.*, 2011). Long durations are also compatible with measurable chemical weathering (Fig. 6) in a paleoclimate like that of Victoria Land today, where salts present only as small flecks (Fig. 4E) as in weathering stage 1 representing 10,000 to 18,000 years (Bockheim, 1997).

Sorted stone stripes of the Masonih profile (Fig. 2D) are comparable with Glacic Haploturbel with ice cores on the Alpine 1 moraine (mid–Holocene) near Goodspeed Glacier, in the Dry Valleys of Antarctica (Bockheim, 2015). Alpine I surfaces in the Dry Valleys are less than 2100 years old (Campbell & Claridge, 1987).

Other paleosols of the Surprise Member are very weakly developed (Nawogan pedotypes) in the development scale of Retallack (1991). Among modern periglacial and temperate soils such very weakly developed paleosols represent only centuries to decades of soil formation (Birkeland, 1999).

Paleoclimate

Sand wedge polygons form today under the following paleoclimatic conditions: mean annual air temperatures of less than-12 to-20°C, mean coldest month temperature <-35°C, mean warmest month temperature of +4°C, and mean annual precipitation <100 mm, as in Antarctic Dry Valleys today (Williams, 1986). Comparable paleoclimate has been inferred for Cryogenian sand wedges of South Australia: in the Cattle Grid Breccia and lower Whyalla Sandstone of the Mount Gunson Mine on the Stuart Shelf (Williams et al., 2008), and in Reynella Siltstone of the Elatina Formation at Hallett Cove (Retallack et al., 2015). Sand wedge polygons have also been reported from the Cryogenian Port Askaig Tillite of Scotland (Spencer, 1971, 1985), and Lillfjället Formation of Sweden (Kumpulainen, 2011). Under warmer frigid conditions, a variety of structures are formed including ice wedge polygons, periglacial involutions, and thufur mounds (Young & Long, 1976; Krull, 1999; Retallack, 1999a, b, 2011, 2013). Modern ice wedges are best known from Greenland and Arctic Canada, which have maritime glacial climates, as opposed to continental glacial climate of modern Antarctica with sand wedges (Black, 1976a, b). Fossilized ice wedge polygons have been reported from the Ediacaran Jbéliat Group of Mali (Deynoux et al., 1989; Álvaro et al., 2007), and Brookline and Squantum Members of the Roxbury Conglomerate in Massachusetts (Retallack, 2022a), Cryogenian Port Askaig Formation of Ireland (Kilburn et al., 1965; Howarth, 1971; Johnston, 1993), and Bakoye Group of Mauretania (Deynoux, 1982), and Paleoproterozoic Ramsay Lake Formation of Canada (Young & Long, 1976). Slightly less frigid conditions are evident from thufur mounds and shadow dunes in the early Ediacaran Nuccaleena Formation of South Australia (Retallack, 2011).

Sorted stone stripes form in climates with mean annual temperature of 0 to -4° C, mean air temperature of coldest month of -8° C, rapid temperature drops, freezing index of 1000 to >7000°C days/year, and thawing index of 1000 to 2000°C days/year, again as in Antarctica (Williams, 1986).

Geochemical evidence of paleotemperature can be obtained from a pedogenic paleothermometer of Óskarsson *et al.* (2012), based on modern soils under tundra vegetation of Iceland. This linear regression between mean annual temperature (T in °C) and chemical index of weathering (W), in molar proportions is given in equation 4 ($R^2 = 0.81$; S.E. = ± 0.4 °C). This ratio is similar but not identical to the well– known chemical index of alteration (Nesbitt & Young, 1982; Bahlburg & Dobrzinski, 2011).

$$W = \frac{100. \text{ mAl}_2 \text{ O}_3}{\text{mAl}_2 \text{ O}_3 + \text{mCaO} + \text{mNa}_2 \text{O}} - \text{equation 3}$$
$$T = 0.21\text{I}-8.93 - \text{equation 4}$$

Results this calculation for the analyzed type profiles are mean annual temperatures of -5.6 ± 0.4 °C for Wepaxaku silt loam, -4.4 ± 0.4 °C for Wepaxaku loam, and -6.0 ± 0.4 °C for Masonih loam. Other chemical paleothermometers of Gallagher and Sheldon (2013) and of Sheldon *et al.* (2002) have a training set of modern temperate forested soils, and yielded mean annual paleotemperatures in the range 10.5– 12.8 °C, incompatible with evidence of sand wedge polygons (Fig. 2).

Arid paleoclimate for the paleosols of the Surprise Diamictite is supported by a paleohyetometer of Sheldon *et al.* (2002) based on temperate soils of North America, Chemical index of weathering (*W* in equation 4) increases with mean annual precipitation (*P* in mm) in modern soils ($R^2 = 0.72$; S.E. = ± 182 mm), as follows.

$$P = 221 e^{0.0197W} \qquad -- equation 5$$

This formulation is based on hydrolytic chemical weathering, which results in alumina enrichment at the expense of lime, magnesia, potash and soda. Magnesia is not common in most sedimentary rocks, and potash is excluded because confused by deep burial alteration of sediments (Maynard, 1992; Novoselov & de Souza Filho, 2015). Results of these calculations for paleosols of the Surprise Diamictite are mean annual precipitation 303 ± 182 mm for Wepaxaku silt loam, 226 ± 182 mm for Wepaxaku loam, and 262 ± 182 mm for Masonih loam. Also compatible with arid climate is persistent dolomite, calcite and gypsum within the profiles, but these soluble minerals are not organized into hard nodules like those calibrated for proxies of paleoprecipitation in soils (Retallack, 2005; Retallack & Huang, 2010). Arid and frigid continental climate is indicated for the Surprise Diamictite of Death Valley, California, by sand wedges, sorted stones stripes, and a variety of chemical indications of limited weathering.

Life on land

There are no indications of megascopic life in paleosols of the Surprise Diamictite, but reasons to believe that they were alive with microbes. Depletion of phosphorus and iron but limited mobilization of alumina in two of the profiles (Fig. 6) can be taken as evidence of organic ligands in these paleosols (Neaman *et al.*, 2005a, b). Lack of significant depletion of phosphorus in a third paleosol (Fig. 6) may reflect the patchy distribution of life in these profiles, similar to the patchy distribution of algae and lichens observed in soils of Antarctic Dry Valleys (Campbell & Claridge, 1987). Surface textures in thin sections may also reflect thread–like microorganisms creating vertical disruption of silt and clay (Fig. 4B). Carbonate mineralization at the tops of paleosols also shows vertical fabric and laterally variable thickness and zonation (Fig. 4A, C), suggestive of microbial soil crusts or lichen stromatolites (Klappa, 1979).

Life in Cryogenian paleosols is not a surprise, because evidence for life in soils is now known back 3.7 billion years (Retallack, 2014, 2022b). Most of this ancient life was microscopic, but millimetric fungal vesicles are known from 2100 Ma paleosols (Retallack et al., 2013), centimetric slime molds slug traces from 1900 Ma paleosols (Retallack & Mao, 2019), and cyanobacteria, lichens, fungi, and slime mold megafossils from 720 Ma paleosols (Retallack, 2021, 2023). An observed increase in terrestrial biodiversity and weathering in Tonian paleosols at about 720 Ma may have been a carbon consumption trigger for the onset of Neoproterozoic Snowball Earth (Retallack, 2021, 2023; Retallack et al., 2021a). Survival of Snowball Earth by life at sea may have been a challenge (Runnegar, 2000; Hoffman, 2016; Hawes et al., 2018), but survival would have been less problematic on land in conditions similar to biologically diverse soils of the Antarctic Dry Valleys (Campbell & Claridge, 1987). Before the Cryogenian Sturt Glaciation, the only sterane biomarker known was cholesterol from red algae, non-dikaryan fungi, or perhaps animals, but after the Sturt Glaciation cholesterol was joined by substantial amounts of green algal stigmasterane and fungal ergosterane (Gold, 2018; Love & Zumberge, 2021). The Neoproterozoic Oxidation Event and global glaciation, like the Paleoproterozoic Great Oxidation Event, may have been another occasion when terrestrial and freshwater organisms survived to thrive, and even colonize the oceans (Sánchez-Baracaldo et al., 2017; Shinohara & Nishitani, 2021; Žárský et al., 2021).

CONCLUSIONS

Snowball Earth was at one time envisaged as such a total freeze that life on land and at sea would have been threatened with extinction. Survival of marine life in a tropical seaways, demonstrated by a Cryogenian fossil record, has been explained by the "Soft Snowball" or "Waterbelt" hypotheses, including cryoconite ponds on sea ice. Paleoclimatic modelling also now allows limited waterways, bare rock mountains, sedimentary loess and till plains, and cryoconite dustings of glacial ice. This project, and supporting accounts of Cryogenian periglacial paleosols from South Australia, British Isles, and Mauretania, now confirms that some land surfaces suitable for complex life were bare of snow, mantled with fertile loess, and microtopographically diverse habitats of patterned ground, comparable with the Dry Valleys of Antarctica. Frigid, arid, continental climate is indicated by sand wedges and sorted stone stripes in the Surprise Diamictite. This was a much colder world than at present for paleolatitude of the Cryogenian Kingston Peak Formation estimated at $8^{\circ} \pm 4^{\circ}$. Similar frigid Cryogenian paleosols without evidence of glacial tillites are found at paleolatitudes within 12° of the paleoequator in the Reynella Siltstone Member of the Elatina Formation at Hallett Cove, South Australia. Cryogenian Snowball Earth was a major event in paleoclimatic history of our planet, but its severity can now be assessed from Cryogenian paleosols.

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