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Mineralogy, chemistry and biological contingents of an early-middle Miocene Antarctic paleosol and its relevance as a Martian analogue

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ABSTRACT

Fossil mesofauna and bacteria recovered from a paleosol in a moraine situated adjacent to the inland ice, Antarctica, and dating to the earliest glacial event in the Antarctic Dry Valleys opens several questions. The most important relates to understanding of the mineralogy and chemistry of the weathered substrate habitat in which Coleoptera apparently thrived at some point in the Early/Middle Miocene and perhaps earlier. Here, Coleoptera remains are only located in one of six horizons in a paleosol formed in moraine deposited during the alpine glacial event (> 15 Ma). A tendency for quartz to decrease upward in the section may be a detrital effect or a product of dissolution in the early stage of profile morphogenesis when climate was presumably milder and the depositing glacier of temperate type. Discontinuous distributions of smectite, laumontite, and hexahydrite may have provided nutrients and water to mesofauna and bacteria during the early stage of biotic colonization of the profile. Because the mesofauna were members of burrowing Coleoptera species, future work should assess the degree to which the organisms occupied other sites in the Dry Valleys in the past. Whereas there is no reasonable expectations of finding Coleoptera/insect remains on Mars, the chemistry and mineralogy of the paleosol is within a life expectancy window for the presence of microorganisms, principally bacteria and fungi. Thus, parameters discussed here within this Antarctic paleosol could provide an analogue to identifying similar fossil or life-bearing weathered regolith on Mars.

1. Introduction

Recent Paleogene and Neogene climatic reconstructions in Antarctica, particularly work by De Conto and Pollard (2003), Anderson et al. (2010), Young et al. (2011) and Passchier (2011), indicate the initiation of Cenozoic ice began on the Gamburtsev Subglacial Mountains and other high mountains, as the global atmosphere cooled ~34 Ma during the Eocene–Oligocene transition (http://canadiangoldprospector.files.wordpress.com/2009/12/geologic-time-scale1.jpg). Decline of atmospheric CO_2 and strengthening of the Antarctic circumpolar current combined to produce several ice margin fluctuations in synch with orbital rhythms, the latter probably related to emplacement of alpine moraines (Campbell and Claridge, 1987) in the New Mountain area (Fig. 1A and B) of the Dry Valleys as discussed herein. The glacial deposits in Fig. 1B have variously been considered to be debris-covered ice (Shean and Marchant, 2010) and/or rock glaciers (Rignot et al., 2002).

Previous work by Lewis et al. (2007), Ashworth and Kuschel (2003) and Anderson et al. (2010) documents Middle Miocene

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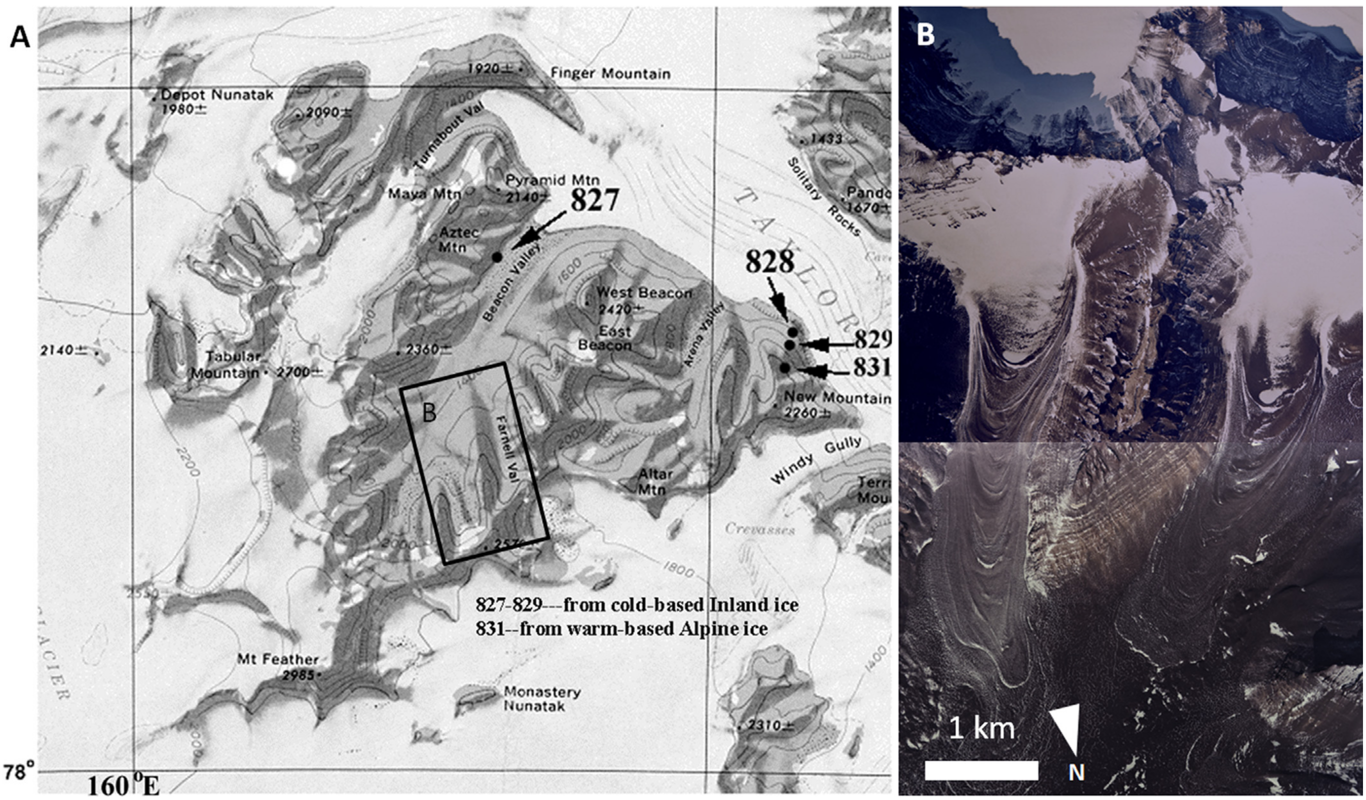

Fig. 1. (A) Location of the 831 profile, New Mountain, Antarctica. The 831 site is 150 m above the embayment where sites 828 and 829 record multiple incursions of ice in the Middle Miocene. Site 827 near Aztec Mountain is close to the Inland Ice Sheet (after Mahaney et al., 2009), the approximate extent of part B is indicated by the black box; (B) aerial images of Beacon Valley from the USGS (CA307900V0296, CA308100V0146), showing two rock glaciers which have similar morphology to glacier-like-forms on Mars (e.g., Fig. 4).

cooling and the demise of biota from East Antarctica, the Dry Valleys, and the Antarctic Peninsula, all part of global climatic change underway in the Late Neogene that began in the latest Eocene/Early Oligocene transition, as stated above. The tundra demise in the Transantarctic Mountains elucidated in these papers focuses on Neogene sediment and fossil biota with little reference to paleosols that are here seen to contain not only microfossils but paleoenvironmental evidence in the enclosing weathered sediment.

Paleosols are long-term recorders of parent material weathering and bioclimatic fluctuations (cf. Beyer et al., 1999; Bockheim, 1979, 1990; Mahaney, 1990; Mahaney et al., 2001, 2013; Birkeland, 1999; Retallack and Krull, 1999; Retallack et al., 2007). They serve as repositories of environmental perturbations that occurred over varying lengths of time and in this case (profile 831; Fig. 2) since at least the Middle Miocene, and possibly earlier in the Early Miocene–Oligocene transition. Adjacent to the inland ice sheet, Antarctica, inordinately slow weathering in paleosols has produced small, but above detection limit concentrations of extractable Fe and Al (Mahaney et al., 2009), as well as exceedingly low concentrations of organic matter derived principally from the decay of microbial populations of bacteria and fungi (Hart et al., 2011). Sluggish paleosol development in the Dry Valleys of Antarctica is not so different from the prevailing cold and dry environmental conditions on Mars, with the latter environment being infrequently interrupted by both transient endogenic and exogenic activities (Baker et al., 1991,2007; Fairén et al., 2003; Fairén, 2010).

As reported by Mahaney et al. (2001), Fell et al. (2006) and Hart et al. (2011) the presence of *Beauveria bassiani,* an insectivorous fungi, restricted to salt-rich horizons in Antarctic pedostratigraphic complexes opens the question as to what insect taxa the fungi might be or have been associated with in this or some past climate. The question has been partly answered by the presence of fossil Coleoptera exoskeletons recovered from a paleosol in a moraine deposited by alpine ice (Mahaney et al., 2012a). Certainly, the presence of fossil Coleoptera argues for a warmer climate in the interior of the Antarctic Dry Valleys at some time during the Middle Miocene or earlier (Marchant et al., 1993). Others (Wilson, 1973; Campbell and Claridge, 1987) argued the region carried alpine ice early in its erosional history, possibly during the Eocene. Evidence of wet-based alpine ice further supports a period of warming of unknown duration, an amelioration presumably accompanied by an invasion of Coleoptera.

Antarctic paleosols are rare and known mainly from the work of Gibson et al. (1983), Campbell and Claridge (1987), Bockheim (1990, 2013), Beyer et al. (1999), Mahaney et al. (2001, 2009) and McLeod et al. (2009), but detailed XRD and SEM analyses of paleosol mineralogy and chemistry are rarer still. Previous work by Retallack and Krull (1999) and Retallack et al. (1998, 2007) documents the composition of soils/paleosols in tills associated with ancient Permian/Triassic forests at Coalsacks Bluff, Antarctica. Recent discovery of exoskeletons of fossil Coleoptera in a paleosol dating to the emplacement of alpine-age moraines during the early growth of temperate ice in the Dry Valleys (Marchant et al., 1993; Mahaney et al., 2012a) makes the mineral and chemical composition of prime importance in understanding the habitat in which the mesofauna and bacteria flourished, possibly in the Early Miocene, albeit for an unknown length of time.

Mineralogical and chemical details of the paleosol horizon and entire moraine profile in which the fossil skeletons were recovered are presented here based on recent XRD and SEM/EDS analyses. Analysis of mineral distributions and chemical trends down profile reveal Early to Middle Miocene habitat constraints on the colonization and evolution of Coleoptera and associated microbial

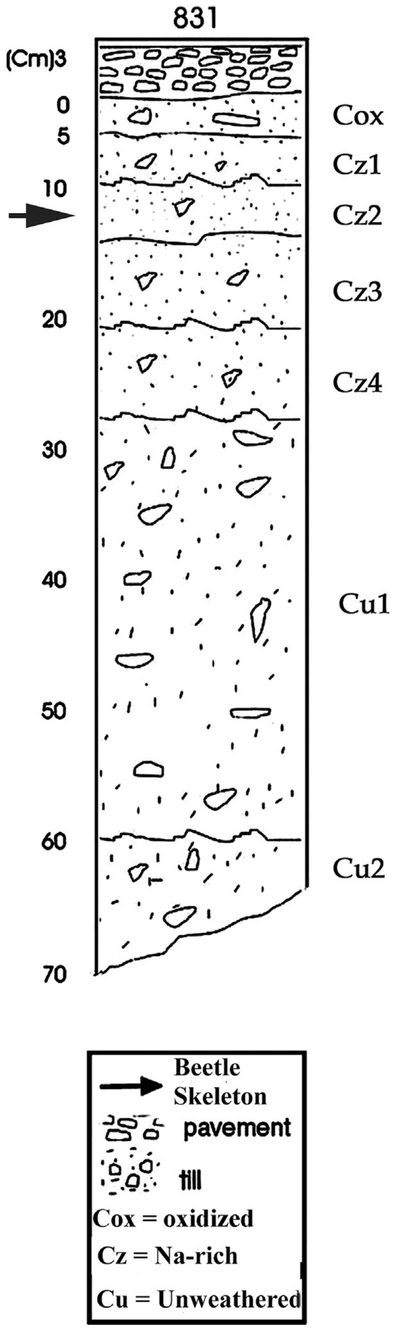

Fig. 2. The 831 section, a single stage profile formed in moraine emplaced by alpine ice during the early stage of glaciation in Antarctica (after Mahaney et al., 2012a,b).

communities. Mycological analysis of nearby profiles, dated to 15 Ma, indicates *B. bassiani* is a key fungal component of local paleosols, prevalent in salt-rich horizons (Claridge and Campbell, 1968; Claridge, 1977; Mahaney et al., 2001). As with the former environmental changes from warm to cold, if global warming continues (Cox et al., 2000), these profiles will warm and colonies of these CO_2 generating fungi will proliferate, enhancing CO_2 loading of the atmosphere. *B. bassiani,* a widespread fungi is, in fact, a prolific CO_2 producer (D. Malloch, personal communication, 2001).

Other recent investigations detail the presence of numerous species of bacteria (Hart et al., 2011; Mahaney et al., 2012a) possibly reducing waste from fungal components and aeolian inputs. To what degree did the lithology of the 831 profile provide a habitat for microbes and mesofauna that colonized the site during the early stage of alpine type (wet based) glaciation in the Dry Valleys? What nutrients were released that allowed various life forms to thrive and proliferate, if even at a slow rate? These are the questions we seek to answer with this investigation. While the study of the Early to Middle Miocene insect fauna in Antarctica is not expected to be found on Mars, it is highly likely the microbe fauna known to exist in the Dry Valleys (Mahaney et al., 2001; Hart et al., 2011) might have correlative species on the Red Planet.

In addition, Antarctica is considered as the nearest Earth analogue to Mars (e.g., Marchant and Head, 2007); hence, we will explore the astrobiological implications of this ancient Antarctic paleosol, which provides a window into the inorganic substrate necessary to support the presence of life in one of the coldest/driest environments on Earth. Our study site has physico-mineral-chemical properties close to those thought to have persisted throughout a large part of the history of Mars, at least since the shutdown of the dynamo or nearly 4.0 Ga (Baker et al., 2007; Fairén et al., 2010a). Future Martian exploration adjacent to the north and south polar ice caps or in regions on Mars thought to have experienced the transgressions and regressions of ice sheets or alpine glaciers (Figs. 3C and D and 4) (Kargel and Strom, 1992; Head and Pratt, 2001; Kargel, 2004), will occur in terrains with a multiplicity of lithologies and with the water necessary for the proliferation of life (Fig. 3). Thus, such environments have the potential to yield evidence of present or former life which exploit similar energy and chemical gradients as the microbial communities uncovered in the Antarctic Dry Valleys (Mahaney et al., 2001, 2009; Hart et al., 2012) (Fig. 2).

2. Regional geology and field area

2.1. Antarctica

The lithology of the Transantarctic Mountains including the Taylor Glacier area (Fig. 1A and B) includes the Beacon Supergroup of quartzitic sandstone and dolerites, an eminently simple stratigraphy (Mahaney et al., 2001). The topographic setting of broad glaciated valleys with alternate benches or steps is similar to the topography of South Africa where the landscape matured under semi-arid climatic conditions. Volcanic vents near McMurdo Station (Armstrong, 1978) remain active and volcanoes such as Mount Terror in the Ross Island area have been only marginally modified by glaciation.

Glaciation in the Antarctic has been the subject of much debate with some arguing for an Early Paleogene onset of ice and others for Late Paleogene/Early Neogene startup (Campbell and Claridge, 1987; Cooper et al., 2008; Florindo and Siegert, 2009). The early phase of glaciation in the Dry Valleys, whenever it began, is seen as a temperate geophysical type with wet-based ice evolving into cold-dry ice which persists to this time (Campbell and Claridge, 1987). Onset of this transition occurred at the termination of the Middle Miocene Climatic Optimum (Warny et al., 2009) and is broadly compatible with the age of the profiles at Aztec Mountain and those situated at lower elevation at New Mountain embayments (Fig. 1A) (Mahaney et al., 2001, 2009). The climate during wet-based glaciation probably resulted in increased summer thaw and melt of both snow and ice and consequently somewhat marginally more water available for weathering and soil genesis. Because pedostratigraphic successions seem to prevail at lower elevations, the periodic invasion of ice from the lower Taylor Valley into New Mountain embayments probably occurred under cold conditions, the substrate frozen and armored and thus self-protected from encroaching ice. While buried paleosols in the New Mountain area appear to have been little disturbed by encroaching cold-based ice, evidence in other areas argues for considerable erosion and modification of the landscape (Cuffey et al., 2000; Davies et al., 2009).

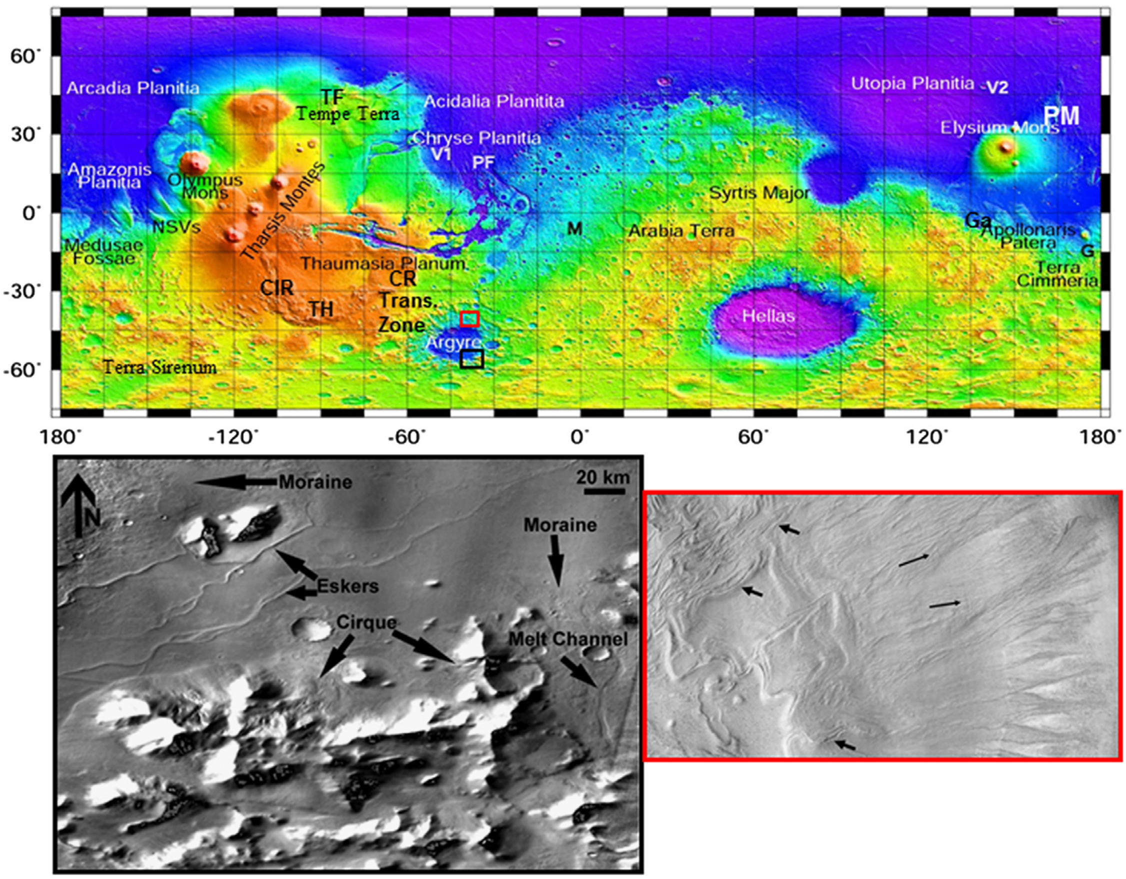

Fig. 3. The top figure is a Mars Orbiter Laser Altimeter (MOLA; credit: NASA) topography map (where white and red are highest elevations and blue and purple lowest elevations). The locations mentioned in the text are labelled and also shown are roving and landing sites of Vikings 1 and 2 (V_1 and V_2, respectively), Pathfinder/Soujourner (PF), Mars Exploration Rovers Opportunity and Spirit at Meridiani (M) and Gusev (G), respectively, and the Mars Science Laboratory ("Curiosity") at Gale crater (Ga). The locations of the lower left and lower right figures are indicated by black and red boxes, respectively. (Lower left) THEMIS IR daytime mosaic (credits: NASA and ASU) highlighting the southeast part of the argyre basin, which includes glacial terrain indicative of the varying environmental condtions through time; features include ridges interpreted to be eskers, cirques, moraines, and melt channels. (Lower right) HiRISE image detailing distinct ridges interpreted to be moraines (black arrows) in an impact crater located along the northern margin of the Argyre basin floor. Data are from the HIRISE image ESP_013507_1395, credit NASA/JPL-Caltech/Univ. of Arizona.

The ice that emplaced the moraine at site 831, higher on the slopes above prominent lower benches at New Mountain (e.g., sites 827–829, Fig. 1A), is considered to have been warm based as evidenced by the greater prevalence of round sands carrying prolific v-shaped percussion cracks considered to result from fluvial transport (compatible grain microtextures in Mahaney, 2002), presumably affected by meltwater, compared with similar clasts in the lower paleosols where angular grains and glacial crushing microfeatures predominate (Mahaney et al., 2009). While it might be possible to argue that round quartz could be inherited from the Beacon sandstone, it is impossible to explain the discrepancy in roundness microtexture frequencies between the 831 profile discussed here, and detailed SEM analysis of nearby tills (sites 827–829) deposited by cold-based inland ice during the Middle Miocene Climatic Optimum or thereafter (Mahaney et al., 1996, 2001). All these paleosols belong to the Cold Desert group of Antarctic soils/paleosols (Campbell and Claridge, 1987) and are classed as subxerous or xerous with moisture regimes at lower moisture levels compared with coastal soil/paleosol systems. Snowfall in the interior mountains near the Inland Ice is infrequent, amounting to about six weeks per year and apparently on the rise in response to present day global warming (Parmesan, 2006; Fountain et al., 2010; Joughin and Alley, 2011). Water from snow melt may stay in liquid form for a few hours but intergranular water films may stay at temperatures well below zero because of the high salt content (Ugolini and Anderson, 1973; Wynn-Williams et al., 2000). Also, because Konishchev (1998) has shown that some mafic minerals, feldspar and quartz are subject to slow chemical weathering (Cryolithogenic Weathering Index) by reaction with micro water films released during periods of freeze thaw in extreme Arctic locales, such processes may have been active in the 831 profile during the warm phase of early Antarctic glaciation.

Lower sites at New Mountain are composed of greater amounts of dolerite relative to quartzitic sandstone; hence, a lower quartz dilution and higher pyroxene, olivine and Ca-plagioclase insure greater Fe content along with other chemical elements.

At Site 831, the upper Cox (ox = oxidized, Birkeland, 1999) horizon is similar with higher Fe and less sandstone, whereas the lower salt-rich horizons are quartz diluted with significantly lower Fe. As a result, the lithologies of the various horizons in 831 are not simple two-phase systems, the differing mineral contents reflecting differences in source rock available to the glacier at different times. Initially, and for some time thereafter, ice depositing the 831 sediment sourced mainly sandstone and granite, reverting later to a mix of dolerite and sandstone as reflected in the composition of the uppermost Cox horizon and pavement.

2.2. Mars

Paleosol systems more highly developed compared to the oldest known Antarctic paleosols are likely to be found in ancient crustal Noachian-age surfaces estimated to be older than 3.7 Ga (Ivanov, 2001; Hartmann and Neukum, 2001), including the rugged rim materials and intervening valleys of the Argyre (Fig. 3C and D) and Hellas impact basins, the mountain ranges of Thaumasia highlands

Fig. 4. A Martian glacier-like form in the eastern part of Argyre (31.07°W, 42.16°S) taken from the Souness et al. (2012) dataset, although larger than the Antarctic counterparts, has longitudinal and terminal ridges, an upland amphitheater-shaped source area and features that indicate deflection of flow around obstacles. Context (CTX) images P14_006624_1385_XI_41S031W and G06_020746_1394_XN_40S031W. Image credit: NASA/JPL-Caltech/MSSS.

and Coprates rise, and the terrains of Terra Cimmeria and Terra Sirenum (Fig. 3A and 3B). Such weathered systems may contain salts, ferric oxides and other water adsorbing compounds many times greater than what is reported here and currently known in Antarctica with a greater diversity of weathered zones (horizons), perhaps even with a similar range of resident clay minerals. More importantly, perhaps Noachian surfaces (>3.7 Ga) probably contained more water, some of which is probably incorporated still in clay minerals and/or as crystalline water in various lithic systems. Hesperian-age surfaces (Fig. 3B), estimated to range from about 3.7 to 3.4 Ga (Ivanov, 2001; Hartmann and Neukum, 2001), may possess paleosols with less weathering, having undergone slower weathering in a cold desiccating environment different than the preceding Noachian period. Paleosols in glacial sediments forming over the Amazonian period (e.g., Conway and Balme, 2014), estimated to be <3.4 Ga (Ivanov, 2001; Hartmann and Neukum, 2001), may reside in relict, buried, or even exhumed positions, such as within some of the Argyre basin deposits (Dohm and Kargel, 2012; Dohm et al., 2013b; El Maarry et al., 2013; Soare et al., 2014a,b) (Fig. 3B). Such glacial sediments (Fig. 4) may include reworked materials sourced from the earlier Noachian and Hesperian periods, described above.

3. Materials and methods

3.1. Antarctica

The sample pit (831) was selected on a moraine ridge deposited by alpine ice above the embayment where sites 828 and 929 are located (Fig. 1A); the site was excavated by hand and cut back to expose fresh material. The soil-horizon descriptions (Fig. 2) used here are genetic and follow guidelines set out by Mahaney et al. (2009), differing somewhat from the system of Campbell and Claridge (1987) which uses an alphabetical enumeration. In the earlier study of the 831 section by Mahaney et al. (2001), horizons were enumerated from A to Z with increasing depth in the profile. In a more recent paper, Mahaney et al. (2009) explored the use of Fe and Al extracts to explore Na-dithionite Fe as an age-control weathering index, and Na-pyrophosphate Al as a proxy for organic carbon. However, the usual downward succession of horizons of Fe-rich material over salt-rich material equates to Cox/Cz horizons in the Canadian and US soil classification systems (CSSC, 1998), Soil Survey Staff (NSSC, 1995; Catt, 1990; Birkeland, 1999). Occasionally, where lower concentrations of oxides exist, a C designation (Birkeland, 1999) is invoked. The “ox” designation simply implies a color stronger than 10YR5/4 (Oyama and Takehara, 1970), a characteristic attributed to more extensive oxidation. The Cu designation refers to unweathered C horizon sediment (Hodgson, 1976). The *z* designation (CSSC, 1998) is preferred for frozen salt-rich horizons over the USDA *n* for unfrozen salt-rich horizons.

Soil color grades are assigned based on the soil chips detailed in Oyama and Takehara (1970). Approximately 500 g samples were collected from each horizon to allow for laboratory work, including particle size analysis following the procedures outlined by Day (1965). Samples were lightly sonicated and wet sieved to separate sands from clay, and silt and particle size follows the Wentworth Scale with the sand/silt boundary at 63 μm, while the clay/silt boundary (2 μm) follows the NSSC (1995). Following particle-size analysis, total dissolved salts were determined through electrical conductivity (Bower and Wilcox, 1965), pH by electrode in a 1:5 ratio (sediment: distilled H_2O ratio) and C, N, and H by Leco apparatus.

For the six samples described here, the <2 mm bulk fraction was ground using a McCrone micronising mill, and the resulting slurry was freeze-dried to precipitate soluble salts before being prepared as a random mount. In addition to analysis of the <2 mm fraction, determination of the clay mineralogy of the <2 μm fraction followed the washing of soluble salts with pure water and removal of organic matter with 30% H_2O_2. As described by Moore and Reynolds (1997), this fraction was extracted by centrifugation, and samples were prepared by drying the resulting suspension onto glass slides. Ethylene-glycol (EG) solvation of the slides was achieved by exposing them to EG vapor at 70 °C for a minimum of 12 h. XRD patterns were recorded with a Bruker D5000 powder diffractometer equipped with a SolX Si(Li) solid state detector from Baltic Scientific Instruments using Cu $K\alpha$ 1+2 radiation. Intensities were recorded at 0.04° 2-theta step intervals from 5 to 80° (6 s counting time per step) and from 2 to 50° (4 s counting time per step) for bulk and clay mineralogy determination, respectively.

Selected sands were subsampled and analyzed under the light microscope. From this grain population, a smaller number of samples were subjected to analysis by JEOL-840-JSM Scanning Electron Microscope (SEM) with Energy Dispersive Spectrometry (EDS) using a PGT System at the Department of Geology, University of Toronto, following methods outlined by Mahaney (2002). Photomicrograph and X-ray microanalyses were obtained at accelerating voltages of 10–20 keV and 10–15 keV, respectively. Some of these micrographs were realized on polished thin-sections of Cz2 horizon sample using a Hitachi S2500 SEM-EDS equipped with a Thermo Noram system at the Isterre laboratory, University of Grenoble. Photomicrograph and X-ray microanalyses were obtained at accelerating voltages of 16 keV.

3.2. Mars

The analysis of ancient crustal rocks on Mars, including basement structures that might contain paleosols, various weathered surfaces, including eolian, alluvial, fluvial, and colluvial covers, are currently ongoing through geologic mapping investigation coupled with analyses of spectrometer data acquired through orbiting, landing, and roving spacecraft (Dohm et al., 2009a, 2009b, 2009c; Murchie et al., 2009b; Morris et al., 2010; Meslin et al., 2013; Arvidson et al., 2014; Fairén et al., 2014).

4. Results

4.1. The Antarctic profile

The 831 profile is a single stage paleosol, as opposed to the pedostratigraphic successions at sites 828–829 (Mahaney et al., 2001), with a 7–8 cm thick pavement, thickest of all pavements in the group of profiles described by Mahaney et al. (2001), which may relate to the inferred ameliorative paleoclimate and length of time the ice had to emplace the pebble blanket. The paleosol consists of seven horizons as shown in Fig. 2; two of the lowermost horizons – Cz4 and Cu1 – are similar in kind to the unweathered parent material (Cu2) below, although with a lower clast count. Horizons are thin, both of the two uppermost being approximately 5 cm thick, spacings within horizons increasing slightly down to the paleosol/parent material contact at just under ~30 cm depth. The Cox horizon under the pavement obtained its slightly stronger (brighter) color because of the higher dolerite content and release of Fe^{+3}, secondary Fe falling to about one-third the concentration in the Cz horizons below, color losing strength accordingly with depth (Table 1). All horizons carry sand textures with the exception of sandy loam in the Cz2 and Cu2. As expected, with little clay content, the soil lacks any structure and the normal soil properties of consistence, plasticity, and stickiness are meaningless to record. What strikes the observer are the color changes, finer textures in the Cz2 and Cu2 horizons, and variations in clast content from soil to parent material. Actual distributions of sand, silt, and clay were previously reported by Mahaney et al. (2001).

Color variations (Table 1) from yellowish brown (10YR5/6) in the Cox horizon to bright yellowish brown (10YR6/6) in the Cz1 and 2 horizons indicate a shift from slightly higher Fe to lower Fe below. The color gradation below in the Cz3 and Cz4 horizons is to paler yellow colors (2.5Y7/4 and 7/3), close to the usual color of near-fresh undifferentiated sediment. Variations in C with depth are near detection limits with the exception of the Cz2 horizon where it 'spikes' to 0.04%. Hydrogen (Table 1) follows a similar trend rising to 0.88%, a fluctuation correlated to the slight reduction in pH. The pH trend is from slight acidity in the Cox to slight alkalinity with depth in the profile. Variations in total salts as determined by electrical conductivity show the highest salt content in the Cz1 and Cz2 horizons, the latter horizon being the weathered zone where the Coleoptera were recovered.

Table 1
Chemical and physical properties of horizons in the 831 Paleosol, Antarctica. Color follows the system of Oyama and Takehara (1970). Color in the pavement is darker by ~2 chroma of colors in the weathered sediment and variable.

Horizon	Depth (cm)	C (%)	H (%)	N (%)	pH (1:5)	E.C.[a] (mS)	Color
Pavement	7–0	–	–	–	–	–	Variable
Cox	0–5	< 0.01	0.17	0.06	6.1	4.90	10YR5/6
Cz1	5–10	< 0.01	0.23	0.23	7.6	13.40	10YR6/6
Cz2	10–14	0.04	0.88	0.37	7.2	7.53	10YR6/4
Cz3	14–22	< 0.01	0.06	0.08	7.4	7.26	2/5Y7/4
Cz4	22–30	0.01	0.07	0.10	7.4	5.50	2.5Y7/4
Cu(1)	30–65	< 0.01	0.06	0.08	7.5	2.43	2.5Y7/2

[a] Electrical conductivity.

4.2. XRD (Antarctic profile)

The bulk (Fig. 5A) and clay (Fig. 5B) fractions were analyzed in all six horizons (Cox, Cz1, Cz2, Cz3, Cz4 and Cu1, see Table 2) with a focus on determining the degree of uniformity/heterogeneity down profile, and especially the lithology of the Cz2 horizon where the meso- and micro-fauna (bacteria) were recovered (Hart et al., 2011; Mahaney et al., 2012a). Presumably with down profile movement of clay and nutrients, including Fe from the adjacent Cz1 horizon, the two uppermost horizons might be considered a tandem source of both moisture and chemical stimulants for metabolic functions. Because the bulk and clay fractions produce somewhat similar results, it is best at first to consider them together, and later to focus on those aspects of the clay fraction that differ from the bulk group of samples.

Through the 831 profile (Table 2), all horizons present a chlorite, quartz, and illite assemblage, with albite/microcline observed only in horizons Cox and Cz2. In the upper part of the profile (Cox, Cz1, Cz2), gypsum was detected. This last phase was associated to the hexahydrite and calcite phase in the Cz2 horizon.

In the lower part of the profile, Cz3, Cz4, and Cu1, laumontite, and pyrophyllite were detected. Smectite was only present in the Cz4 horizon, and halite appeared in the Cu1 horizon only.

The chlorite is of interest because of its Fe content (see below), a necessary requirement as an electron acceptor for some types of microbial respiration (Jaisi et al., 2007). Laumontite, a member of the zeolite group, is a hydrated calc-alumino-silicate, which both acts as a chemical filter and allows ions and water to flow into and out from its crystalline structure, therefore acting as a chemical sieve.

Hexahydrite, a hydrated Mg sulfate mineral (Palache et al., 1951), common in the two uppermost horizons may grow into thin but massive crystals, even ≥ 1 cm in diameter. With six water molecules, it is nearly as heavy with water as with sulfur and magnesium, and may be a source of trapped moisture for the growth of microorganisms in an otherwise highly desiccating environment. Terrestrial extremophiles growing under cryogenic temperatures have been associated with hydrated minerals such as hexahydrite (Dalton et al., 2003) which supports this contention. Laumontite, like most zeolites, often forms inside vesicles of dolerite. With exposure to air and light, laumontite may dehydrate into leonhardite, but despite the porosity and freeze dried nature of the sediment, XRD analysis did not detect leonhardite.

Within the 831 profile (Fig. 5A) concentrations of quartz were noticeably lower in the upper two horizons, possibly a product of dissolution over long time intervals when the climate was warmer and with greater tendency for meltwater production. Alternatively, decreasing quartz might be related to freeze/thaw processes as discussed by Schwamborn et al., (2013) in NE Eurasia.

4.3. Antarctic Coleoptera

Mesofauna were discovered by light microscopic analysis of sands in the 831-Cz2 horizon only. These samples were coated with carbon, and following SEM analysis, EDS signatures were averaged out on other silicate minerals, the difference taken as normal coating concentrations subtracted from high carbon peaks to establish the presence of organic material. Fig. 6A is representative of a relatively clean Coleopteran exoskeleton, the cleanest of all specimens recovered still with appreciable embedded salts, but despite well preserved parts, identification of species proved impossible. The presence of well-preserved fossil Coleoptera from the upper 831 profile ranges from a leg segment and procoxa (Fig. 6A) of a specimen belonging to the family Scarabaeidae, possibly belonging to the genus Aphodius or Ataenius. The specimen demonstrates an articulated tibia, femur, trochanter and attached procoxa seen from

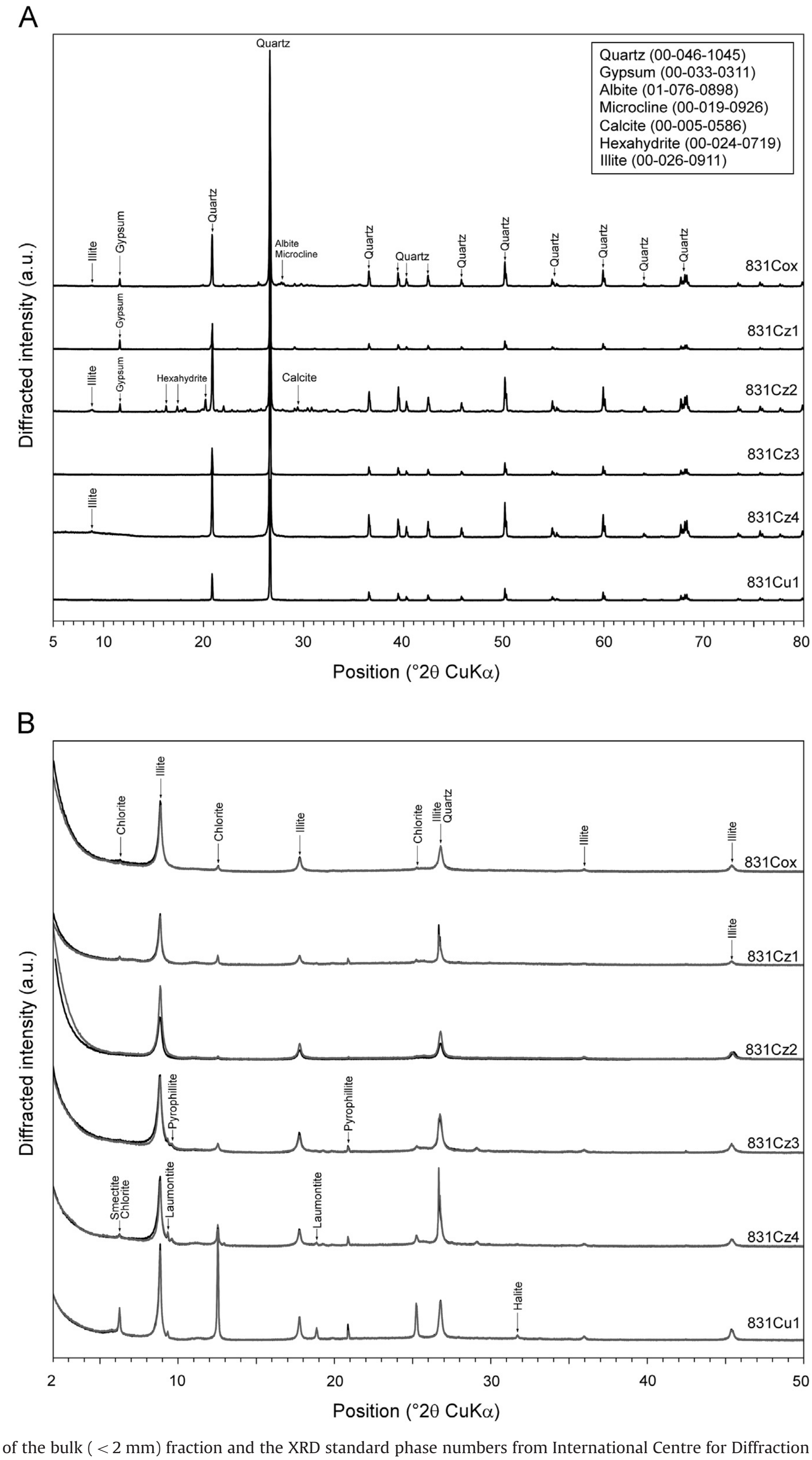

Fig. 5. A, XRD diffractograms of the bulk (< 2 mm) fraction and the XRD standard phase numbers from International Centre for Diffraction Data base (ICDD) are indicated; B, XRD diffractograms of the clay (< 2 μm) fraction in the 831 profile. Patterns recorded under air-dried conditions are shown as solid black lines, whereas those recorded after ethylene-glycol salvation are shown as solid red lines. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

the ventral side. Other descriptive body parts such as the abdomen and head-capsule were not found.

The long extended filamentous-like fiber shown in the SEM frame (Fig. 6A) could possibly be a hyphal mat created by a fungus (Webster and Weber, 2007). Although identification of fungal species was not carried out, previous work showed that the dominant culturable bacterial phylum for the 831-Cz2 horizon is *Actinobacteria* (Hart et al., 2011). *Actinobacteria* are well known as major contributors to the degradation of chitin (Trujillo, 2008) and carry a wide degree of chitinases (Kawase et al., 2004). Chemical spectra of both the

Table 2
Mineral assemblage for the six horizons of the profile 831 obtained by XRD analyses. * Observed in clay fraction $< 2\ \mu m$, *n.d.* not detected. Alb/Mi = Albite/Microcline; Hexa = Hexahydrite; Lau = Laumontite; Pyro = Pyrophillite.

Mineral	Quartz	Alb/Mi	Illite	Chlorite	Gypsum	Hexa	Calcite	Halite	Lau	Pyro	Smectite
Horizon											
Cox	+	+	+	+*	+	*n.d.*	*n.d.*	*n.d.*	*n.d.*	*n.d.*	*n.d.*
Cz1	+	*n.d.*	+*	+*	+	*n.d.*	*n.d.*	*n.d.*	*n.d.*	*n.d.*	*n.d.*
Cz2	+	+	+	+*	+	+	+	*n.d.*	*n.d.*	*n.d.*	*n.d.*
Cz3	+	*n.d.*	+*	+*	*n.d.*	*n.d.*	*n.d.*	*n.d.*	+*	+*	*n.d.*
Cz4	+	*n.d.*	+	+*	*n.d.*	*n.d.*	*n.d.*	*n.d.*	+*	+*	+*
Cu1	+	*n.d.*	+	+*	*n.d.*	*n.d.*	*n.d.*	+*	+*	+*	*n.d.*

filamentous form and the beetle show high concentrations of carbon along with salts enriched with S, Mg, Na and Cl. The image (Fig. 6B) is possibly an encrusted beetle specimen, although with a thickened crust of salts unaffected by light sonication, it is impossible to confirm a biogenic composition. However, EDS spectra on the specimen indicates high C with lesser amounts of K and Cl. Imaged at 15 keV, with penetration of the electron beam limited to about 1 μm, most probably salt encrustations are $\pm 1 \times 10^3$ nm thick. The excessive carbon strongly indicates a beetle skeleton and is similar to the analysis undertaken of like specimens from the same horizon. The filamentous string-like forms attached to the skeleton in Fig. 6B are highly encrusted with salts and minor Si:Al (ratio of 2:1) suggesting, with K present, the presence of illite.

The occurrence of Mg and Fe chlorite, within the sand fraction (Fig. 6C), may act as a source of Fe^{+3} for facultative anaerobic microbe respiration as an electron acceptor (Burdige et al., 1992; Glasauer et al., 2003), which aside from moisture, are important requirements for the growth and preservation of life forms, even in the Antarctic. The clear transformation of Mg chlorite to Fe chlorite is shown in the image and accompanying EDS spectra in Fig. 6C. Most of the white forms on the image are secondary precipitates of P, most probably deposited microbial by-products and very possibly produced over an inordinate length of time, i.e., from the Middle Miocene or longer. The temperature range required for microbe and Coleoptera survival in these paleosols is impossible to determine without detailed site instrumentation and more detailed analysis of micro-moisture and salt content in the sediment mix of felsic and mafic minerals. It is likely the survival temperature of bacteria, fungi and Coleoptera vary within margins, each species escaping freezing and low temperature exposure by adjustment of physiological and biochemical processes that enhance a tolerance to prolonged freezing. It is well known that freeze-tolerant species are able to withstand formation of ice in their intercellular body fluids (Bentzi and Mullins, 1999) and such tolerance is constrained, at least in part, by subtle variations in microhabitat. While super cooling lethal temperatures, at which nucleation of water occurs, varies in each species it must be dependent upon species physiology and various paleosol/sediment properties, all of which are impossible to determine with precision given the data available at present.

4.4. *Antarctic lithology*

A range of grains, mostly quartz of variable grade sizes, is shown in Fig. 7A–D, all of which are representative of sands recovered from the 831 profile. Approximately 80% of quartz sands in the 831 paleosol are subround to round suggesting considerable transport in meltwater. The frequency of adhering particles, considered the product of glacial abrasion (Mahaney, 2002) varies from sample to sample as shown in Fig. 7A and B, with some grains exhibiting a higher frequency of v-shaped percussion cracks considered to be the hallmark of fluvial transport (Krinsley and Doornkamp, 1973; Mahaney, 2002). In contrast, some sands (Fig. 7E and F), from site 829 (location, Fig. 1A) and heavily coated with salts, assume a subangular form typical of glacial crushing (Mahaney, 1995, 2002) and with masked grooves and striae resulting from glacial abrasion and transport.

5. Discussion

5.1. *Antarctica*

5.1.1. *Classification and comparison with nearby sites*

The 831 paleosol described here classifies within the Gelisol order in that it has semi-gelic materials within 100 cm of the soil surface and permafrost within 200 cm of the soil surface (NRCS, http://soils.usda.gov/). The dry frozen sand texture of the horizons consists of primary and secondary mineral gelic materials, with exceedingly low carbon, showing little evidence of cryoturbation. Hence, the 831 profile belongs to the Orthel suborder and keys closely to Anhyorthels, profiles with weathered gravels over salt-rich horizons (Cn or Cz) over frost cemented beds. While little is known about the properties and composition of soils/paleosols adjacent to the Inland Ice Sheet, the 831 profile exhibits similar horizon characteristics to the soil profile at Mt. Fleming Climate Station Site, Upper Wright Glacier, adjacent to margins of the Inland Ice Sheet. The elevation of the Mt. Fleming site is 1698 m, (5568 ft) and the location is 77°32′S; 166°17′E, ~150 m lower than the 831 site. The soil at Mt. Fleming is resident within a polygon consisting of coarse clastic material with a much higher gravel content compared with the pebbly sandy loam matrix of the 831 profile. However, the strongly weathered polygonal gravel of the Mt. Fleming soil, with colors (10YR 5/4), closely akin to profile 831, and at a similar depth, suggest it may have a similar age, correlating with the alpine glacial event. Only weak evidence of cryoturbation is apparent in the 831 profile sketch shown in Fig. 2, the result of the sandy texture and dry frozen state of the weathered sediment from which the Coleoptera were recovered.

Sands in the 831-Cz2 horizon, thickly and thinly coated with salts, display differing degrees of roundness and angularity. For the most part, approximately 80% of all grains analyzed are round or subround, many carrying a wealth of v-shaped percussion cracks from meltwater transport. In contrast, in sites 827–829 (Mahaney et al., 2001), quartz consists of subangular to angular grains carrying varying degrees of glacial crushing, particularly subparallel fractures, striae, grooves and many microfeatures masked with salt encrustations. The degree to which Antarctic paleosols are coated with salts, stands in contrast to tills analyzed in other nearby sections and horizons (Mahaney et al., 2001; Mahaney, 2002) that lie below former active soil forming processes, more or less apart from active pedogenesis above in the profiles.

5.1.2. *Age of profile 831*

Because the Fe-extracts for the 831 sediment suggest an age older than the 15 Ma documented age of nearby cold-based tills (sites 828 and 829; Mahaney et al., 2009), it is likely soil morphogenesis began during the Early Miocene or earlier under a more ameliorative climate

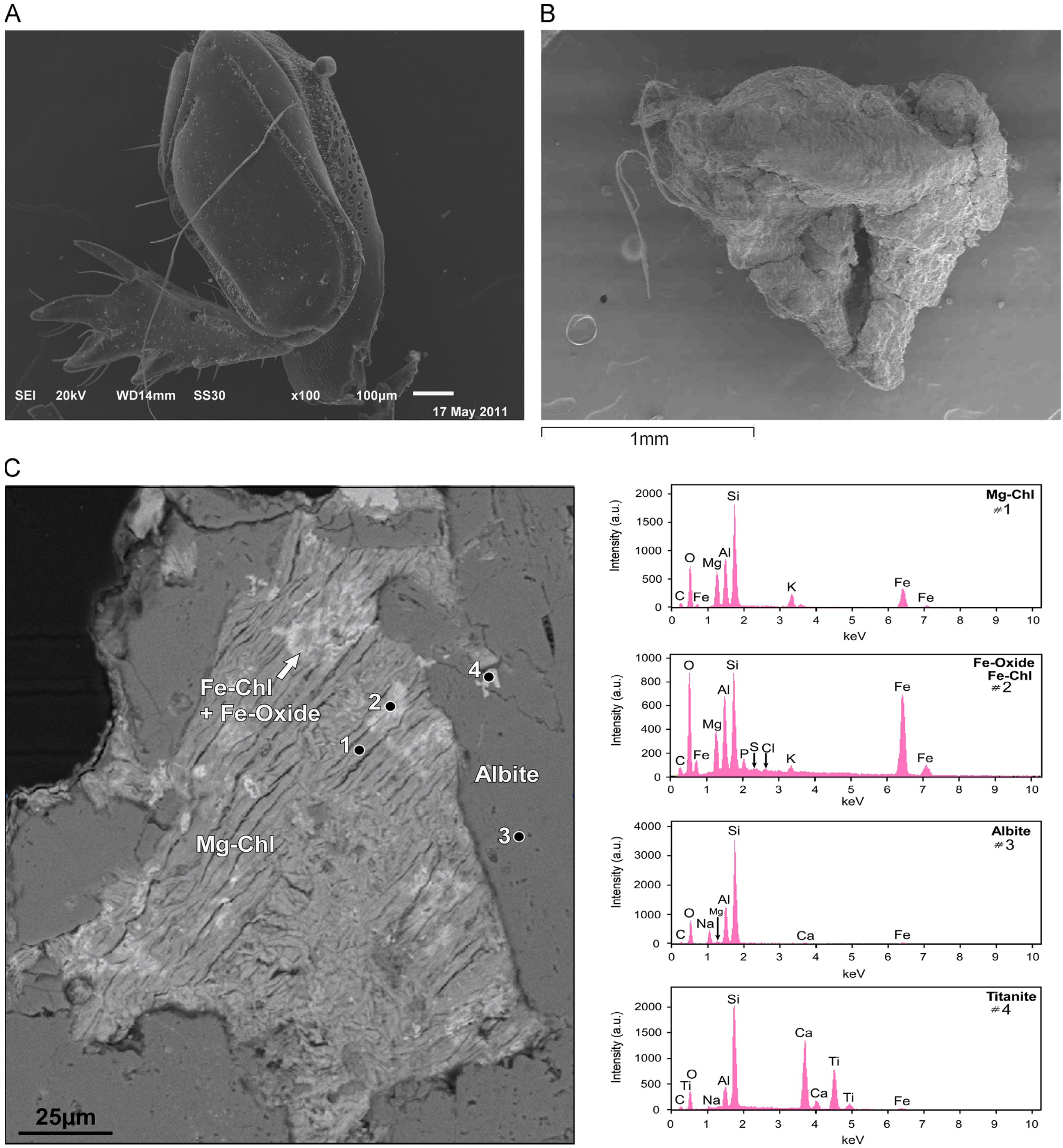

Fig. 6. (A) Coleoptera from Cz2 horizon with a nodule of unknown composition in upper right and a filament encrusted with salts crisscrossing a segment of the leg and procoxa of a Coleoptera specimen. Sensory hairs are intact with slight fractures to exoskeleton. Light sonication may have cleaned salts of this specimen; (B) Grain assemblage that could be a salt-encrusted Coleoptera although the electron beam could only detect Ca, Na, Mg, Cl and S. However, the form suggests wings and dorsal or ventral beetle parts, thickly encrusted with salts; (C), albite chlorite intergrowth illustrating conversion of trioctahedral chlorite to dioctahedral Fe which could result from weathering under more ameliorative soil climate. The presence of P is a secondary precipitate that may have a biogenic origin.

with higher MAAT and MAP. The Fe_d/Fe_t ratio, a measure of age since deposition (Arduino et al., 1986; Torrent and Cabedo, 1986; Mahaney et al., 1999, 2010, 2014) produces a quotient in the 831 profile twice as high as in the 828 and 829 profiles (Mahaney et al., 2009) dated at 15 Ma. While some of the Fe_d increase could be due to the earlier more temperate climate accompanying the growth of alpine ice, the much higher Fe_d/Fe_t ratio surely indicates greater relative age, even if absolute age control is lacking.

While the absolute age of the site is unknown, the evidence strongly suggests the sediment belongs to the alpine phase of Antarctic glaciation which began as warm-based ice in the highlands, eventually spreading to the Antarctic Shield as the climate deteriorated presumably in or before the Middle Miocene. How long the early alpine phase of glaciation lasted is unknown, but could easily range back into the Oligocene. If the 831 section is Oligocene in age, the bacteria and Coleoptera resident there could

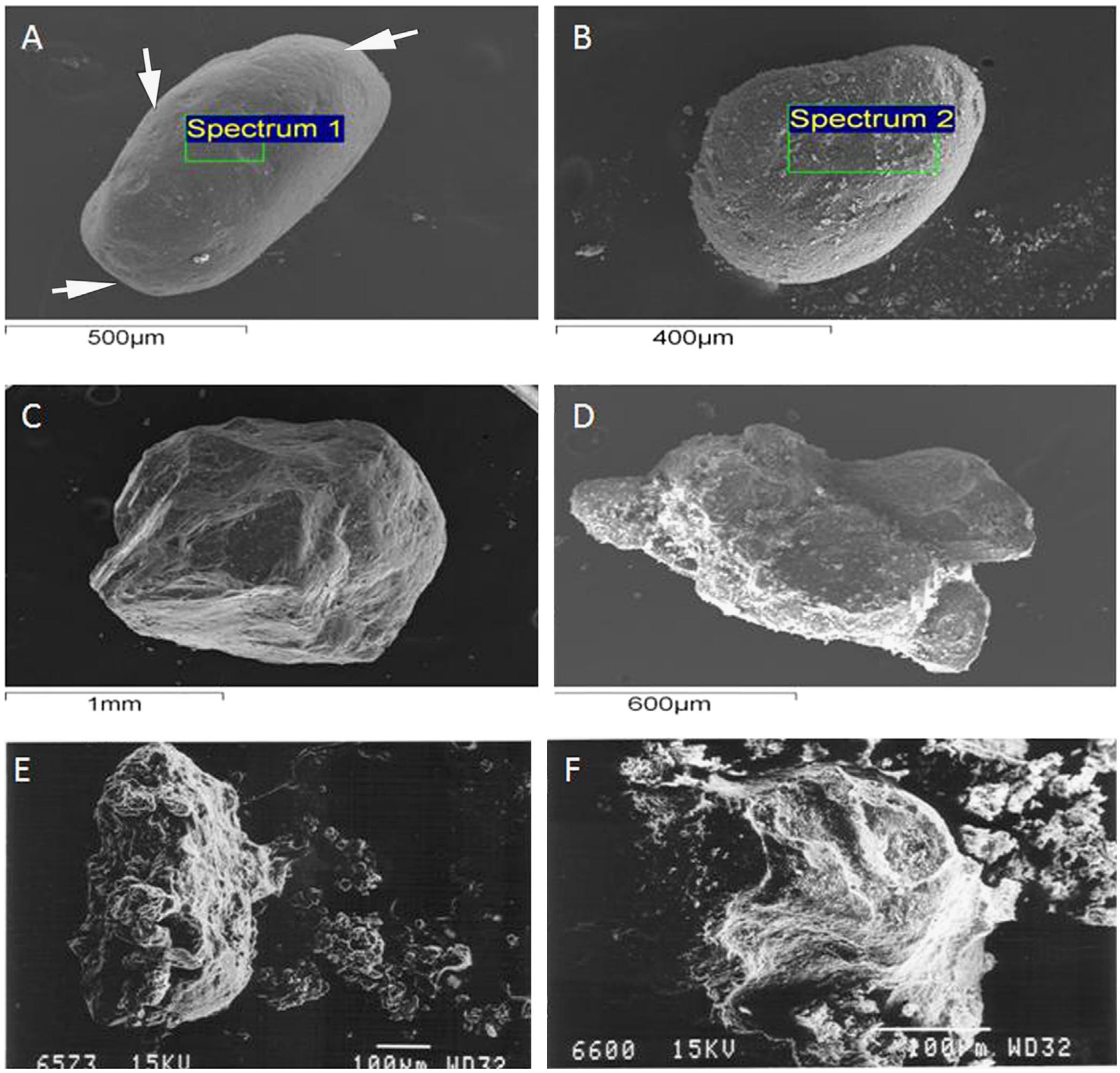

Fig. 7. (A) Subround quartz in the Cz2 horizon with v-shaped percussion cracks (arrows) indicating it had been tumbled in meltwater; (B) round quartz with higher count of adhering particles compared to (A); (C) nearly equidimensional quartz with adhering particles and sharp, blunt edges; (D) salt encrusted subangular grains of quartz and orthoclase confirmed by EDS; (E) complex subangular intergrowth in profile 829 of quartz (left) and orthoclase (right), the entire assemblage covered with $CaSO_4$ and small amounts of Fe, presumably Fe^{+3}; (F) subangular quartz grain (profile 829) thickly coated with salts, mainly $CaSO_4$ and small concentrations of secondary Fe.

easily provide some of the oldest bacteria cells and chitin protein in Antarctica.

5.1.3. *The ecosystem of profile 831*

The Coleoptera recovered from the 831-Cz2 horizon described above are interpreted as burrowing beetles and were not found in overlying horizons, namely, 831-Cox and Cz1, or in the underlying horizons. The presence of beetle mesofauna may relate to the presence of *B. bassiani,* an insectivorous species, which may be related to the former presence of Coleoptera in nearby New Mountain paleosols 828 and 829 (Mahaney et al., 2001). A search for similar Coleoptera fossils in all other profiles – 827–829 – failed to produce positive results. Because fungi were not targeted in the 831 profile, the relationship with *B. bassiani,* while possible, is at the moment only conjecture. However, because bacteria in the horizon have been established (Hart et al., 2011; Mahaney et al., 2012a), it seems plausible to argue that bacterial colonies were metabolizing chitin compounds of the mesofauna and senescent fungi. Another possible link to Coleoptera in the microbial food web is the previous identification of *Micrococcus luteus* in the 831-Cz2 horizon (Hart et al., 2011). This bacteria has been associated as part of the natural flora of Coleoptera (Yilmaz et al., 2006; Cardoza et al., 2006) and is assumed to exist in a symbiotic fashion. Cardoza et al. (2006) isolated *M. luteus* from oral secretions of Coleoptera and argued that they act as primary defense agents against fungal infection. Although only speculative, the presence of these three organisms in the immediate Antarctic vicinity, and because they currently exist in a demonstrated food web at more temperate climates, hints at their interaction prior to the onset of full-polar conditions.

5.1.4. *Coleoptera preservation in profile 831*

While fossil Coleoptera, interpreted to have an age between Pliocene and Middle Miocene, have been found further south in the Meyer Desert Formation near the Beardmore Glacier (Ashworth and Kuschel, 2003), the fossils from New Mountain described here would appear to belong to an earlier age when warm-based ice deposited the alpine moraine. Unlike the site described near the Beardmore Glacier, complete with fossil wood, moss, shells and leaf remains of *Nothofagus,* we recovered only the Coleoptera exoskeletons described herein and in Mahaney et al. (2012a,b). The presence of well-preserved fossil Coleoptera from the upper 831 profile ranges from a leg segment (Fig. 4A) of a specimen belonging to the family Scarabaeidae, possibly belonging to the genus Aphodius or Ataenius, as described above. Given the presumed age of the specimen the amount of preserved detail is remarkable, especially the setae (hairs) still attached to fixed positions on the tibia and femur. The degree of preservation of body parts held together with connective tissue and interwoven with salts, well preserved morphology, and intact chitin lacking appreciable corrosion suggest a number of possible explanations. This specimen, the most highly preserved amongst all recovered samples, is either a recent contamination after sampling or the product of light sonication after salt encrustations were worn off. It is entirely possible the sample was frozen quickly, encrusted (armored) with salt at a later time well prior to lab preparation. Given that the samples were collected in sterile double bags it is unlikely this one sample is a recent contaminant, and if it were, one would expect to find similar specimens in samples from sections 827–829. Potential for the reanimation of dormant coleopterans is possible (Ring and Tesar, 1980; Ring, 1981), considering the more ameliorative climate of the laboratory, and the freeze-dried and stable state of the paleosols between emplacement of the tills and site sampling. Virtually all the Coleoptera recovered maintain some degree of salt encrustation despite the light sonication used in lab preparation of samples for particle size analysis.

The question of salt content in the beetle habitat and its relation to age of the skeletons might be answered several ways. Some specimens analyzed are robustly intact, relatively unencrusted with salt, as shown in Fig. 6A and might be relatively recent invaders of the site. Alternatively, it is possible laboratory pre-treatment with light sonication might have removed salt thus producing a well-articulated specimen. Other specimens (Fig. 6B, for example) are heavily encrusted with salts and could be considered coeval with an early stage of biotic invasion following deposition of the moraine sediment. Variable degrees of encrustation may also derive from position in the horizon itself or from pre-treatment in the laboratory where the entire sample of the <2 mm fraction was subjected to particle size analysis and wet sieving, which may have cleaned up specimens thinly coated with salts. Whatever explanation is preferred, it is possible for some species of Coleoptera to exist in halophilic or hypersaline environments (Velasco et al., 2006). It seems, given the cold desert environment of the last ~ 15 Myr Antarctic environment, attenuated somewhat by the advent of an earlier evolutionary stage of temperate-style glaciation, that it is likely the Coleoptera described here inhabited the site earlier on in its history, dying off with the invasion of more severe dry-frozen conditions. The alternative hypothesis of Coleoptera living a fast-life cycle, invading the site during brief (6–8 week) summer seasons, and expiring with the advent of winter seems less plausible. Because Coleoptera burrows were not detected at the time of collection it would seem the specimens recovered relate to an early phase of pedogenesis following deposition of the moraine. The question of the influence of salt content and habitability can only be answered with a more widespread study of additional sites.

Discordant coatings, with some Coleoptera in near-pristine condition and others coated with thick encrustations of salt and Fe, begs the question as to possible differences in age or simply position within the horizon. If, as suggested by the elevated concentration of secondary Fe (Mahaney et al., 2009), the alpine moraine is indeed considerably older than other moraines at New Mountain (sites 828 and 829), considered to have been emplaced by cold-based ice after the Middle Miocene Climatic Optimum of ~ 15 Ma (Graham et al., 2002 Sites 827, 828, 829; Mahaney et al., 2001), it is likely the 831 profile underwent transition from an ameliorative climatic phase to a colder one. If correct, this transition was likely coincident with the demise of the Coleoptera population, perhaps leaving bacteria either as endospores in a state of freeze-dried stasis, or as psychrophiles.

5.1.5. The role of mineralogy in ecosystem development

The mineralogy of the bulk matrix material of <2 mm fraction and the clay fraction (Fig. 5A and B) shows only minor differences between the two fractions. Within the bulk fraction quartz, calcite, halite and gypsum dominate in variable quantities down section. Within the clay fraction there is an increase in quartz down section that may relate to quartz dissolution in the upper three horizons, possibly the product of silica dissolution over an inordinately long period of time. Otherwise, the illite and illite-chlorite detected in both fractions are considered the product of decomposition of granite and hydrothermal metamorphic rock at some time in the past, probably all preweathered materials, and hence, detrital.

The presence of zeolite in the form of laumontite is of special interest in understanding the paleoecology of the site because of its ability to adsorb cations of nutritional importance to the mesofauna. While authigenic chabazite, another zeolite, is known in the Sirius Diamictite and interpreted as a weathering product (Dickinson and Grapes, 1997), laumontite detected in the 831 profile may well represent albitized plagioclase, either authigenic or allogenic in origin. Hexahydrite, as a hydrous molecule similar to laumontite, may well play a role in providing moisture in liquid phase to all life forms that inhabited the site in the past. The question of whether the mesofauna inhabited the site early on during the alpine phase of glaciation and expired as the climate deteriorated or continued to thrive through the later Miocene, Pliocene and Quaternary, is unknown.

The presence of limited foodwebs in the Dry Valleys have previously been demonstrated (Cowan et al., 2010; Pointing et al., 2009) and tend to be governed chiefly by the availability of liquid water (Barrett et al., 2007). As these organisms adhere to strict biochemical requirements, a strong influence is placed over them by the surrounding landscape and geochemistry in regards to nutrient availability. Minerals such as hexahydrite and laumontite release water molecules under dry and exposed conditions, while relatively more complex minerals such as illite can be used as a nutrient resource by microbes after weathering or enzymatic solubilization. The relatively high abundance of N in all of the sampled sites in comparison to the very low to nil determinations of C, indicate very low abundance of live microbial biomass (Simpson et al., 2007) at present. Because liquid water is known to exist, at least at the nanolevel in the Meserve Glacier (Cuffey et al., 2000), and hypothesized to exist in salt-rich horizons of paleosols (Mahaney et al., 2001), sufficient liquid water must have existed in the 831 paleosol after transition to cold-based ice, although microbes might have existed even with little liquid water as described above.

5.1.6. Microbial communities

It is certain that the stratification of soil horizons observed here, both from color change and sediment analyses, is partly linked to microbial activity and to sediment sourcing by ice during the alpine phase of glacier growth as Antarctica cooled. The wide range of microorganisms present in such sediments – often at considerable depth, where anaerobic metabolism most likely dominates (Fredrickson and Balkwill, 2006; Edwards et al., 2012) – has now been demonstrated both through our own research (Hart et al., 2011; Dohm et al., 2011a) and by others (Pointing et al., 2009). In this study we have shown that Fe^{+3} electron acceptors, required for facultative anaerobes involved in both hetero- and auto-trophic processes, are present in plentiful supply. Indeed, it is possible the absence of Fe-chlorite in the Cox horizon is a result of higher levels of microbial activity in the top 5 cm of weathered sediment. The presence of cyanobacteria – probably as endolithic organisms – in this ecosystem is highly likely in the Cox sediment

(Pointing et al., 2009), and this biomass may ultimately serve as a key stimulus for a complex C-cycle, perhaps in the past involving the Coleoptera as well as diverse heterotrophic bacteria and fungi. The presence of chemoautotrophic bacteria in the Cz horizons is also very likely. Such organisms would be involved in Fe leaching processes that could have led to some of the changes observed in the XRD analysis (Niemela et al., 1994).

Despite these possibilities, there is no doubt that the overall level of microbial biomass present is likely to be orders of magnitude lower than what would be expected in a temperate weathered paleosol (Virginia and Wall, 1999). Low microbial biomass may also mean that microbial respiration in the Cz horizons is not actually oxygen limited at all. Further metagenomic analysis of microbial populations in these samples could help to verify this possibility and define the complex microbial community that will certainly be present in the Cz horizons.

5.1.7. *Provenance of materials in profile 831*

The high percentage of round quartz carries equally high percentages of v-shaped percussion cracks, recovered from sands in the 831 section, which opens the question as to origin, specifically whether or not some or all of the grains could be the result of fluvial transport within the early alpine ice or sourced directly from the Beacon Supergroup sandstone. Comparison with quartz sand recovered from tills in nearby sites 828 and 829 at New Mountain and site 827 at Aztec Mountain (Fig. 1A), as well as samples from till at other localities in the Dry Valleys (Mahaney et al., 1996), suggests that grain roundness in 831 is the result of glacial processes, particularly from temperate ice. If otherwise, with rounded grains sourced from the Beacon sandstone, one would expect to find high percentages of round quartz in other Dry Valley till samples. It is impossible to discount some addition of round quartz grains from Beacon sandstone outcrops into the 831 till sediment but it appears the sudden high frequency of such grains in the 831 profile is due to glacial abrasion in meltwater during the early warm/wet glacial phase.

The 831 paleosol carries a single story profile of 30 cm depth, a single stack of Cox/Cz horizons, similar in character to profiles at nearby sites at New Mountain and Aztec Mountain.

5.1.8. *Recent climatic conditions at the 831 site*

The paleosol is within the climate of the polar desert soils of the Dominion Range with a present mean annual air temperature (MAAT) of -39 °C and mean annual precipitation (MAP) of 36 mm (Retallack et al., 2001). While climatic sums on a monthly/daily basis are not available for site 831 ($\sim$1850 m), and/or comparable sites near the inland ice, the only short term climatic data is from McMurdo Station (elevation 24 m a.s.l., http://www.coolantarctica.com, data from Landcare, NZ). At McMurdo the average daily temperature in January is -2.9 °C, mean daily maximum temperature is -0.2 °C and the mean daily minimum is -5.5 °C. The annual mean temperature is -16.9 °C. Applying the normal dry adiabatic lapse rate of 10 °C/km to the McMurdo January mean temperature yields temperature summaries close to the means reported below for the Mt. Fleming site in Upper Wright Valley.

Of the nine stations instrumented and capable of collecting atmospheric and soil climatic data, only Mt. Fleming in Wright Valley (1700 m a.s.l., NRCS, http://soils.usda.gov/) has values close to Site 831 at New Mountain. Mean daily air temperatures collected for 2011 range from -7 °C to -35 °C whereas mean daily soil temperatures at the 2-cm soil level range from $\sim$2 °C in summer to -40 °C in winter. While soil temperatures are for one year only (2011), and considering a longer instrumental record might give a wider spread of temperatures, it is apparent daily temperatures, while close to 0 °C, are mostly well below freezing. While salt content in the Mt. Fleming paleosol is known to be limited to the 2–8 cm level it would appear to be less concentrated than at the 831 site discussed here. Nevertheless, while salt content is sufficient to lower the freezing temperature, it probably does not raise the soil temperature to the 5 °C limit, which is within the lower limit allowable for most Coleoptera to persist at a site. Because the material in which the Coleoptera were recovered was dry frozen at the time of recovery, and barring any climatic perturbations that might have warmed the soil since the Middle Miocene Climatic Maximum, there is little possibility the samples date from less than $\sim$15 Ma with every possibility they are considerably older. With reference to the Fe extracts discussed above the samples are likely Early Miocene or older.

5.2. *Relevance of the Antarctic study to potential development of paleosols on Mars*

5.2.1. *Potential age of Martian paleosols*

Possible paleosol records could date back into a phase of Mars evolution, comparable to the Hadean Earth which has been all but destroyed through plate tectonism. Martian paleosol profiles may not be so distinct compositionally from those described above for the Antarctic paleosol at site 831, New Mountain, Antarctica. This is especially the case for those paleosols in the extremely ancient geologic provinces of Mars, which include Terra Cimmeria and Terra Sirrenum, the mountain ranges of Thaumasia highlands and Coprates rise, and the rugged terrain of Claritas rise (Fig. 3A) (Dohm et al., 2005, 2013a and references therein) as discussed below, even though the age ranges between those of Earth and Mars are vastly different. In comparison with the paleosol at site 831, estimated to be older than 15 Ma, paleosols with source sediment from these extremely ancient geologic provinces of Mars may reach ages greater than 4.0 Ga, or equivalent to Hadean-age rock outcrops. Hadean rocks have been destroyed on Earth through plate tectonism, with only zircons remaining, making such records on Mars valuable not only for informing on its early evolution, but that of the solar system. The age of these extremely ancient geologic provinces is based on stratigraphy (e.g., Scott and Tanaka, 1986; Tanaka et al., 2014), impact crater size-frequency statistics (e.g., Tanaka, 1986; Hartmann and Neukum, 2001), and the existence of remnant magnetic anomalies (e.g., Acuña et al., 1999; Connerney et al., 1999a; Solomon et al., 2005). Such terrains have not been significantly overprinted by volcanism, tectonism, or impact cratering (e.g., Hellas, Argyre, Isidis, and Chryse from the Late Heavy Bombardment), hydrothermalism, erosion, among other processes, since the shutdown of the dynamo nearly 4.0 Ga (Arkani-Hamed, 2004; Dohm et al., 2005, 2013a; Baker et al., 2007; Roberts et al., 2009; Roberts and Arkani-Hamed, 2012).

5.2.2. *Geochemistry and mineralogy of Mars and hypothesized paleosol compositions*

Mars is being revealed to be a more geochemically diverse planetary body through orbiting, landing, and roving robotic spacecraft, since the Mariner and Viking explorations decades ago, and not the mostly basaltic planet that was once popularly reported in the literature.

The paleosol compositions in the extremely ancient terrains of Mars may not be so different from those of Earth. On Mars, in situ analyses have been performed only at landing and roving sites in geologic terrains that are Late Noachian or younger, or not exceeding roughly 3.6 Ga (Fig. 3A and B). Most of these sites have been influenced by geologic and hydrologic processes distinct from those of more ancient times, including those related to the development of the Tharsis superplume until present-day, which includes geochemical signatures of the younger volcanic provinces and their widespread distribution through prevailing long-lived wind activity. Even so, the Pathfinder mission at the debouchment region of Ares Valles revealed rocks of andesitic compositions, which represent highly differentiated crustal materials similar to those found on Earth (Bruckner et al., 2001). These results are in contrast to the typical basaltic signature; elemental concentrations (e.g., Al, Cl, Ca, Fe, Si, S, Mg, Ti) of soils at the Pathfinder site mostly appear similar to those of the Viking 1 and 2 sites, with only potassium being substantially different, with Viking (XRF spectrometer) indicating an upper limit of 0.12% K (Clark et al., 1982), whereas Pathfinder at 0.5% K, the latter of which may be elevated due to weathering of local rocks (Bruckner et al., 2003). The provenance of the Pathfinder soils, includes both Tharsis rocks (basaltic) and rocks associated with a more ancient geologic province of Mars. In addition, the gravel deposits in Gale Crater comprise pebbles that are "unique" from the commonly reported basaltic compositions, including rock clasts with relatively large feldspar crystals (Wiens et al., 2014).

The primary rock compositions of the ancient terrains such as the rugged mountain ranges remain uncertain, though the signatures such as the andesitic composition at the Pathfinder site and the felsic rocks interpreted to be granite associated with Nili Syrtis (Wray et al., 2013; Wiens et al., 2014) are pointing to more felsic parent rock compositions. With increasing orbital reconnaissance, a greater percentage of the ancient crustal rock materials of Mars, including basement, are being revealed through the basaltic mantles, chemical weathering rinds (Mahaney et al., 2012b), and eolian, alluvial, fluvial, and colluvial covers (Dohm et al., 2009a), pointing to a differentiated planetary body not so distinct from Earth. The diversity is revealed by spectrometers onboard the Mars Odyssey, Mars Reconnaissance Orbiter, and Mars Express spacecraft, as well as in situ rover observations. The diversity includes:

- hematite concretions referred to as blueberries (Squyres et al., 2008; Moore, 2004; Ormö et al., 2004; Havics et al., 2009),
- sulfates (Gendrin et al., 2005; Quantin et al., 2005; Mangold et al., 2007a,b; Bishop et al., 2009; Murchie et al., 2009a,b; Flahaut et al., 2010a,b; Fueten et al., 2011),
- clays (Fialips et al., 2005; Poulet et al., 2005; Bibring et al., 2006; Wray et al., 2009; Fairén et al., 2010b, 2011, 2012; Marzo et al., 2010),
- chloride-bearing materials (Osterloo et al., 2008; Davila et al., 2011),
- carbonates (Ehlmann et al., 2008; Morris et al., 2010), serpentine (Dohm et al., 2009b; Ehlmann et al., 2009, 2010),
- pure silica (Squyres et al., 2008), and
- andesitic and possibly granitic rock compositions (e.g., Bandfield et al., 2000; Bruckner et al., 2001; Rieder et al., 1997; Wray et al., 2013).

We expect the paleosols of Mars, particularly those that developed among the more ancient (pre-Tharsis) basement, to be similar to those of the Antarctic paleosol at site 831 described above. This is because such basement rocks of likely felsic compositions, formed at a time when the dynamo and possible plate tectonism was in operation (Baker et al., 2007; Dohm et al., 2009a,b, 2013a), having mixed with younger basaltic rock materials such as those to the average of related to the development of Tharsis and other younger, much less prominent volcanic provinces. Particularly, the felsic rocks and their altered byproduct, as well as mixing with younger basaltic volcanic rock materials, as hypothesized for the extremely ancient provinces of Mars (Dohm et al., 2009, 2013a), may have sourced sediment that formed into paleosols approximating those of New Mountain, Antarctica. As such, with environmental and energy conditions, including expected salts and iron enrichment, the ancient terrains and their associated paleosols are considered to be prime targets for future astrobiologic investigations (Mahaney et al., 2001), among other types of habitable environments on Mars (Dohm et al., 2011a).

Iron enrichment is expected in the Martian paleosols, as with salts and changing pH with associated change in environmental conditions through time (Fairén et al., 2010a). The profiles are also likely to be complex due to impacts, which likely inverted the more ancient felsic crustal rock materials, as well as tapped into olivine-enriched mantle materials (Buczkowski, et al., 2010), with subsequent mixing through eolian (distribution of volcanics such as from Tharsis), fluvial, glacial, and periglacial processes (El Maarry et al., 2013; Soare et al., 2014a,b). The structurally-controlled basins, such as within Terra Sirenum (Anderson et al., 2012; Karasozen et al., 2012), might include (down section), thin veneers of wind-blown basaltic materials, evaporite deposits including salts (Davila et al., 2011), fluvial and alluvial sediments, and basal felsic crustal materials. However, in the ancient mountains of Thaumasia highlands and Coprates rise (Fig. 3A), the rocks are likely to become more felsic with increasing depths toward and within the basement (Dohm et al., 2009a).

Even considering the age differences compared with Antarctica, such paleosols may still contain elevated ferric oxides, salts, and evaporites that allow for slow substrate alteration but over time frames several orders of magnitude greater than in the Antarctic Dry Valleys—one order of magnitude (20 Myr $\times$ 10 = 200 Myr). On a scale such as this, some Amazonian paleosols may well reach 15 orders of magnitude greater thickness/composition than the normal cold desert Antarctic paleosols of 15 cm depth, 1–2% clay content, $\sim$0.5% Fe_d, 0.15% Al_p (proxy for organic carbon derived from microbes-organic carbon measuring.04% in the microbe rich Cz2 horizon of section 831). Adjusting these figures by 15 yields $<$3.4 Ga paleosols in Martian Amazonian terrain to the order of depth = 1.5 m, Fe^{+3} = 7.5%, Al_p = 2.25%, organic carbon = 0.6%, the latter from fossilized microbe membranous tissue if as on Earth. While highly hypothetical, these calculations may well be real given the widespread concentrations of Fe^{+2} known to exist on the Martian surface along with copious water, as confirmed in parts of Mars by the Mars Odyssey (Boynton et al., 2002, 2004; Feldman et al., 2002) and Phoenix (Smith et al., 2009) missions. Considering paleosols to be prime targets to locate extant and/or fossil microbes, principally bacteria and fungi, the presence of salts to maintain liquid water at low temperatures, Fe^{+3} for microbe respiration and as electron receptors, and clay size material for water nutrient adsorption places, these ancient weathering systems should be considered relatively high on the landing site target list.

Within the mountain ranges, there are relatively large impact basins ($>$50 km in diameter) which would have mixed the more ancient paleosol horizons. Also, the transition zone between the Argyre basin and the Thaumasia highlands mountain range would be a mixture of younger basaltic materials, impact ejecta from the giant Argyre basin-forming event, and felsic materials being shed off the Thaumasia highlands. In the rugged rim materials of both Argyre and Hellas (e.g., Scott and Tanaka, 1986), the profiles are also likely to be complex due to the impacts themselves, which likely inverted the more ancient felsic crustal rock materials, as well as tapped into olivine-enriched mantle materials (Buczkowski, et al., 2010), with subsequent mixing through eolian distribution of volcanics (i.e. from Tharsis), and fluvial, glacial and periglacial processes (Fig. 3C and D and 4) (El Maarry et al., 2013; Soare et al., 2014a,b). The structurally-controlled basins, such as within Terra Sirenum (Anderson et al., 2012; Karasozen et al., 2012), might include (down section), thin veneers of wind-blown basaltic materials, evaporite deposits including salts (Davila et al., 2011), fluvial and alluvial sediments, and basal felsic crustal materials (the primary rocks of such extremely ancient geologic provinces which remain elusive to spectrometers largely due to mantling materials and weathering rinds).

5.2.3. Martian environmental conditions

Since the shutdown of the dynamo and plate tectonism and a thinning of the atmosphere, environmental conditions on Mars approach those of the dominant cold and dry conditions of the Antarctic, with prevailing cold and hyperarid conditions. Though, similar to Earth with ever-changing environmental conditions through time, so too has the prevailing cold and dry conditions of Mars been interrupted by major endogenic releases, as well as exogenic processes such as related to impact cratering and obliquity (Baker et al., 1991; Touma and Wisdom 1993; Laskar and Robutel, 1993; Dohm et al., 1998, 2000, 2007; Fairén et al., 2003; Head et al., 2006; Forget et al., 2006). Though, obliquity-driven resurfacing that has failed to overprint the ancient terrains sculpted by dynamic Tharsis and pre-Tharsis activity (Dohm et al., 2001b), they likely contributed to stacked sedimentary sequences related to the episodic depositional events (e.g., Laskar and Robutel, 1993; Lewis et al., 2008).

As highlighted above, the Antarctic pedon may not be exotic to those of Mars. This is especially true in the extremely ancient terrains, including the Thaumasia highlands and Coprates rise mountain ranges, and the ancient basins of Terra Sirenum and Terra Cimmeria (Fig. 3A), as well as the rugged rim materials of the giant impact structures such as Argyre, there is diverse evidence of environmental conditions distinct from those today, approximating those described above for the Antarctica, including the more ancient wetter phases. Rock glaciers and moraines among other landforms are telltales of such environmental pasts (Mahaney et al., 2007) (Figs. 3C and D, 4). This includes a wide array of glacial and periglacial landforms (Dohm and Tanaka, 1999; Dohm et al., 2001a, 2011b; Rossi et al., 2010; Fairén et al., 2011, 2012, 2014; Dohm and Kargel, 2012). Ancient glacial modification is also observable on the terrains of Gale crater currently traversed by the Curiosity rover (Fairén et al., 2014). As highlighted through this paper, it becomes increasingly apparent that the ancient glaciated Martian terrains merit in situ investigation through future robotic explorers in search for fossil or life-bearing weathered regolith on Mars (Fink et al., 2005; Schulze-Makuch et al., 2007), similar to those characterizing the Antarctic pedon described here.

6. Conclusions

XRD and SEM/EDS analysis of the horizons in the 831 profile reveal a mix of minerals that provide a nutrient base for mesofauna and microbes known to have thrived there at least in the early stage of glacial growth, probably during the Early Miocene, possibly earlier. The 831-Cz2 horizon within the profile, resident with copious exoskeletons of Coleoptera and bacteria, is a case in point as the mesofauna required greater amounts of water and chemical nutrients for survival. Presumably, the presence of illite and chlorite, two Fe-bearing clays may have been sources of Fe necessary for respiration. Laumontite in the 831-Cz3 might have acted to sequester/filter ions and water. Hexahydrite a major source of water as well as salt, combined with halite, may have maintained a liquid supply of moisture at temperatures well below freezing. If quartz dissolution in the upper horizons actually produced silica wreckages these might have adsorbed small amounts of water of benefit to mesofauna and microorganisms.

The 831 profile, with such recorded environmental conditions and geochemical compositions, including records of once living life, is expected to be not so different from paleosols of some ancient terrains of Mars. The possibility of such underscores the necessity of robotic and eventual manned mission, likely to be in tandem, to the extremely ancient terrains of Mars with paleosols of far-reaching archives, including possible life.

The Martian paleosols, including those sourcing from extremely ancient geologic provinces of Mars, have far reaching records. Such paleosol records, some reaching back well into the Hadean-age record of Earth, which has all but been destroyed, make such terrains extremely significant targets not only for informing on the incipient development of Mars, as well as the possible origin and evolution of life, if ever initiated on the Red Planet, but also the early development of the solar system.

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References

Acuña, M.H., Connerney, J.E.P., Ness, N.F., et al., 1999. Global distribution of crustal magnetization discovered by the Mars Global Surveyor MAG/ER experiment. Science 284, 790–793. http://dx.doi.org/10.1126/science.284.5415.790.

Anderson, J.B., Warny, S., Askin, R.A., Wellner, J.S., Bohaty, S.M., Kirshner, A.E., Livsey, D.N., Simms, A.R., Smith, T.R., Ehrmann, W., Lawyer, L.A., Barbeau, D., Wise, S.W., Kulhenek, D.K., Weaver, F.M., Majewski, W, 2010. Progressive Cenozoic cooling and the demise of Antarctica's last refugium. PNASearly edition, 1–5. http://dx.doi.org/10.1073/pnas.1014885108.

Anderson, R.C., Dohm, J.M., Robbins, S., et al., 2012. Terra Sirenum: window into pre-Tharsis phases of Mars evolution. In: Lunar Planetary Science Conference, XXXXIII, Abstract #2803 (CD-ROM).

Arduino, E., Barberis, E., Ajmone Marsan, F., Zanini, E., Franchini, M., 1986. Iron oxides and clay minerals within profiles as indicators of soil age in northern Italy. Geoderma 37, 45–55.

Arkani-Hamed, J., 2004. Timing of the Martian core dynamo. J. Geophys. Res., 109. http://dx.doi.org/10.1029/2003JE002195.

Armstrong, R.L., 1978. K–Ar dating: Late Cenozoic McMurdo volcanics and Dry Valley glacial history. N.Z. J. Geol. Geophys. 21, 685–698.

Arvidson, R.E., Squyres, S.W., Bell, J.F., Catalano, J.G., Clark, B.C., Crumpler, L.S., de Souza, P.A., et al., 2014. Ancient aqueous environments at endeavour crater, Mars. Science 343 (6169), 1248097.

Ashworth, A.C., Kuschel, G., 2003. Fossil weevils (Coleoptera: Curculionidae) from latitude 85°S Antarctica. Palaeo 3 (191), 191–202.

Baker, V.R., Strom, R.G., Gulick, V.C., et al., 1991. Ancient oceans, ice sheets and the hydrological cycle on Mars. Nature 352, 589–594.

Baker, V.R., Maruyama, S., Dohm, J.M., 2007. Tharsis superplume and the geological evo-lution of early Mars. In: Yuen, D.A., Maruyama, S., Karato, S.-I., Windley, B.F. (Eds.), Super-plumes: Beyond Plate Tectonics. Springer, pp. 507–523.

Bandfield, J.L., Hamilton, V.H., Christensen, P.R., 2000. A global view of Martian surface composition from MGS-TES. Science 287, 1626–1630.

Barrett, J.E., Virginia, R.A., Lyons, W.B., McKnight, D.M., Priscu, J.C., Doran, P.T., Fountain, A.G., Wall, D.H., Moorhead, D.L., 2007. Biogeochemical stoichiometry of Antarctic Dry Valley ecosystems. J. Geophys. Res.—Biogeosci. 112 (G1), G01010. http://dx.doi.org/10.1029/2005JG000141.

Bentzi, B.J., Mullins, D.E., 1999. Ecology of Mountain Pine Beetle (Coleoptera: Scolytidae) cold hardening in the intermountain west. Physiol. Chem. Ecol. 28, 577–587.

Beyer, L., Bockheim, J.G., Campbell, I.B., Claridge, G.G.C., 1999. Review, genesis, properties and sensitivity of Antarctic Gelisols. Antarct. Sci. 11, 387–398.

Bibring, J.-P., Langevin, Y., Mustard, J.F., Poulet, F., Arvidson, R., Gendrin, A., Gondet, B., Mangold, N., and the Omega Team, 2006. Global mineralogical and aqueous mars history derived from OMEGA/Mars Express data. Science 312, 400–404.

Birkeland, P.W., 1999. Soils and Gemorphology. Oxford University Press, Oxford, UK p. 430.

Bishop, J.L., et al., 2009. Mineralogy of Juventae Chasma: sulfates in the light-toned mounds, mafics in the sand, and opal in the plains. J. Geophys. Res. 114, E00D09. http://dx.doi.org/10.1029/2009JE003352.

Bockheim, J.G., 1979. Relative age and origin of soil in eastern Wright Valley, Antarctica. Soil Sci. 128, 142–152.

Bockheim, J.G., 1990. Soil development rates in Transantarctic Mountains. Geoderma 47, 59–77.

Bockheim, J.G., 2013. Paleosols in the Transantarctic Mountains: indicators of environmental change. Solid Earth Discuss. , http://dx.doi.org/10.5194/sed-5-1007-2013.

Bower, C.A., Wilcox, L.V., 1965. Soluble salts. In: Black, C.A. (Ed.), Methods of Soil Analysis, Part 2. Am Soc. Agronomy, Madison, WI, pp. 933–951.

Boynton, W.V., Feldman, W.C., Squyres, S.W., et al., 2002. Distribution of hydrogen in the near-surface of Mars: evidence for subsurface ice deposits. Science 297, 81–85.

Boynton, W.V., Feldman, W.C., Mitrofanov, I., et al., 2004. The Mars Odyssey gamma-Ray spectrometer instrument suite. Space Sci. Rev. 110, 37–83.
Bruckner, J., Dreibus, G., Rieder, R., Wanke, H., 2001. Revised data of the Mars pathfinder alpha proton X-ray spectrometer: geochemical behavior of major and minor elements. Lunar Planet. Sci. XXXII.
Bruckner, J., Dreibus, G., Rieder, R., Wanke, H., 2003. Refined data of alpha proton X-ray spectrometer analyses of soils and rocks at the Mars pathfinder site: implications for surface chemistry. J. Geophys. Res., 108. http://dx.doi.org/10.1029/2003JE002060.
Buczkowski, D.L., Murchie, S., Clark, R., et al., 2010. Investigations of an Argyre basin ring structure using Mars reconnaissance orbiter/compact reconnaissance imaging spectrometer for Mars. J. Geophys. Res. 115, E12011. http://dx.doi.org/10.1029/2009JE003508.
Burdige, D.J., Dhakar, S.P., Nealson, K.H., 1992. Effects of manganese oxide mineralogy on microbial and chemical manganese reduction. Geomicrobiol. J. 10 (1), 27–48.
Campbell, I.B., Claridge, G.G.C., 1987. Antarctica: Soils, Weathering Processes and Environment. Elsevier, Amsterdam p. 368.
Canadian System Soil Classification, 1998. Soils of Canada, Soil Classification: The Canadian System of Soil Classification. Agriculture and Agri-Food Canada Pub 1646, Ottawa, ON p. 187.
Cardoza, Y.J., Klepzig, K.D., Raffa, K.F., 2006. Bacteria in oral secretions of an endophytic insect inhibit antagonistic fungi. Ecol. Entomol. 31, 636–645.
Palaeopedology Manual. In: Catt, J.A. (Ed.), 6. Quaternary. Int., pp. 1–95.
Clark, B.C., Baird, A.K., Weldon, R.J., et al., 1982. Chemical composition of Martian fines. J. Geophys. Res. 87, 10059–10067. http://dx.doi.org/10.1029/JB087iB12p10059.
Claridge, G.G.C., 1977. The salts in Antarctic soils, their distribution and relationship to soil processes. Soil Sci. 123, 377–384.
Claridge, G.G.C., Campbell, I.B., 1968. Origin of nitrate deposits. Nature 217, 428–430.
Connerney, J.E.P., Acuña, M.H., Wasilewski, P.J., et al., 1999a. The global magnetic field of Mars and implications for crustal evolution. Science 284, 790–793.
Conway, S.J., Balme, M.R., 2014. Decametre-thick remnant glacial ice deposits on Mars. Geophys. Res. Lett 41 (15), 5402–5409. http://dx.doi.org/10.1002/2014GL060314.
(and the 10th ISAES editorial team). Cooper, A.K., Barrett, P.J., Stagg, H., Storey, B., Stump, E., Wise, W. (Eds.), 2008. Antarctica: a keystone in a changing world. In: Proceedings of the 10th International Symposium on Antarctic Earth Sciences. The National Academies Press, Washington, DC.
Cowan, D.A., Khan, N., Pointing, S.B., Cary, S.C., 2010. Diverse hypolithic refuge communities in the McMurdo Dry Valleys. Antarct. Sci. 22 (6), 714–720.
Cox, P.M., Betts, R.A., Jones, C.D., Spall, S.A., Totterdell, I.J., 2000. Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model. Nature 408, 184–187.
Cuffey, K.M., Conway, H., Gades, A.M., Hallet, B., Lorrain, R., Severinghaus, J.P., Steig, E.J., Vaughn, B., White, J.W.C., 2000. Entrainment at cold glacier beds. Geology 28 (4), 351–354.
Dalton, J.B., Mogul, Rakesh, Kagawa, H.K., Chan, S.L., 2003. Methodologies and techniques for detecting extraterrestrial (microbial) life: near infrared detection of potential evidence for microscopic organisms on Europa. Astrobiology 3, 505–529.
Davila, A.F., Gross, C., Marzo, G.A., Fairén, A.G., Kneissl, T., McKay, C.P., Dohm, J.M., 2011. A large sedimentary basin in the Terra Sirenum region of the southern highlands of Mars. Icarus 212, 579–589.
Day, P.E., 1965. Particle fractionation and particle size analysis. In: Black, C.A. (Ed.), Methods of Soil Analysis. American Society of Agronomy, Madison, WI, pp. 545–567.
De Conto, R.M., Pollard, D., 2003. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. Nature 421, 245–249.
Dickinson, W.W., Grapes, R.H., 1997. Authigenic chabazite and implications for weathering in the Sirius Group Diamictite, Table Mountain, Dry Valleys, Antarctica. J. Sediment Res. 67, 815–820.
Dohm, J.M., Tanaka, K.L., 1999. Geology of the Thaumasia region, Mars: plateau development, valley, and magmatic evolution. Planet. Space Sci. 47, 411–431. http://dx.doi.org/10.1016/S0032-0633(98)00141-X.
Dohm, J.M., Kargel, J.S., 2012. Geologic mapping investigation of the Argyre and surrounding regions of Mars. In: Lunar Planetary Science Conference, XXXXIII, Abstract #2468 (CD-ROM).
Dohm, J.M., Anderson, R.C., Tanaka, K.L., 1998. Digital structural mapping of Mars. Astron. Geophys. 39 (3) (20-3.22).
Dohm, J.M., Anderson, R.C., Baker, V.R., et al., 2000. Pulses of magmatic activity through time: potential triggers for climatic variations on Mars. In: Lunar Planetary Science Conference, [CD-ROM], XXXI, abstract 1632 .
Dohm, J.M., Tanaka, K.L., Hare, T.M., 2001a. Geologic Map of the Thaumasia Region of Mars. U.S. Geological Survey Map I-2650 at 1:5,000,000 Scale.
Dohm, J.M., Ferris, J.C., Baker, V.R., Anderson, R.C., Hare, T.M., Strom, R.G., Barlow, N. G., Tanaka, K.L., Klemaszewski, J.E., Scott, D.H., 2001b. Ancient drainage basin of the Tharsis region, Mars: Potential source for outflow channel systems and putative oceans or paleolakes. J. Geophys. Res. 106, 32,942–32,958. http://dx.doi.org/10.1029/2000JE001468.
Dohm, J.M., Kerry, K., Keller, J.M., et al., 2005. Mars geological province designations for the interpretation of GRS data. In: Lunar Planetary Science Conference, XXXVI, #1567 (abstract) [CD-ROM].
Dohm, J.M., Maruyama, S., Baker, V.R., Anderson, R.C., 2007. Traits and evolution of the Tharsis superplume, Mars. In: Yuen, D.A., Maruyama, S., Karato, S.-I., Windley, B.F. (Eds.), Super-plumes: Beyond Plate Tectonics. Springer, pp. 523–537.
Dohm, et al., 2009a. Claritas rise: pre-Tharsis magmatism. J. Volcanol. Geotherm. Res. 185, 139–156.
Dohm, J.M., Williams, J.-P., Anderson, R.C., et al., 2009b. New evidence for a magmatic influence on the origin of Valles Marineris, Mars. J. Volcanol. Geotherm. Res. 185, 12–27. http://dx.doi.org/10.1016/j.jvolgeores.2008.11.029.
Dohm, J.M., Baker, V.R., Boynton, W.V., et al., 2009c. GRS evidence and the possibility of ancient oceans on Mars: special. Planet. Space Sci. 57, 664–684. http://dx.doi.org/10.1016/j.pss.2008.10.008.
Dohm, J.M., Miyamoto, H., Ori, G.G., Fairén, A.G., Davila, A.F., Komatsu, G., Mahaney, W.C., Williams, J.P., Joye, S.B., Di Achille, G., Oehler, D., Marzo, G., Schulze-Makuch, D., Acocella, V., Glamoclija, M., Pondrelli, M., Boston, P., Allen, C.C.R., Anderson, R.C., Baker, V.R., Fink, W., Frazer, A.R., Furfaro, R., Gross, C.H., Hare, T. M., Hart, K.M., Ip, F., Kelleher, B.P., Kim, K.J., Maruyama, S., McGuire, P.C., Netoff, D., Parnell, J., Wendt, L., Wheelock, S., Steele, A., Hancock, R.G.V., Havics, A., Costa, P., Krinsley, D.H., 2011a. An inventory of potentially habitable environments on Mars: geological and biological perspectives. Geol. Soc. Am. Special Paper 483, 317–347.
Dohm, J.M., Ferris, J.C., Baker, V.R., et al., 2011b. Did a large Argyre lake source the Uzboi Vallis drainage system? Post-Viking-era geologic mapping investigation. In: Lunar Planetary Science Conference, XXXIII, Abstract #2255 (CD-ROM).
Dohm, J.M., Miyamoto, H., Maruyama, S., et al., 2013a. Mars evolution. In Mars: evolution, geology, and exploration. In: Fairén, A.G (Ed.), Book Entitled: Mars: Evolution, Geology and Exploration. Nova Science Publishers, Inc., pp. 1–33.
Dohm, J.M., El-Maarry, R., Soare, R.J., et al., 2013b. Geology and astrobiological implications of Argyre, Mars. In: European Planetary Science Congress 2013, EPSC2013-476 .
Edwards, K.J., Becker, K., Colwell, F., 2012. The deep, dark energy biosphere: Intraterrestrial life on Earth. Annu. Rev. Earth Planet. Sci. 40, 551–568.
Ehlmann, B.L., Mustard, J.F., Murchie, S.L., Poulet, F., Bishop, J.L., Brown, A.J., Calvin, W.M., Clark, R.N., et al., 2008. Orbital identification of carbonate-bearing rocks on Mars. Science 322 (5909), 1828–1832.
Ehlmann, B.L., Mustard, J.F., Murchie, S.L., 2009. Detection of serpentine on Mars by MRO-CRISM and possible relationship with olivine and magnesium carbonate in Nili Fossae. In: Lunar and Planetary Institute, Lunar and Planetary Science Conference XL, abstract 1787 (CD-ROM), Houston, Texas.
Ehlmann, B.L., Mustard, J.F., Murchie, S.L., 2010. Geologic setting of serpentine-bearing rocks on Mars. Geophys. Res. Lett. , http://dx.doi.org/10.1029/2010GL042596.
El-Maarry, M.R., Dohm, J.M., Michael, G., et al., 2013. Morphology and evolution of the ejecta of Hale crater in Argyre basin, Mars: results from high resolution mapping 226, 905–922Icarus 226, 905–922.
Fairén, A.G., 2010. A cold and wet Mars. Icarus 208, 165–175.
Fairén, A.G., Dohm, J.M., Baker, V.R., de Pablo, M.A., Ruiz, J., Ferris, J.C., Anderson, R. C., 2003. Episodic flood inundations of the northern plains of Mars. Icarus 165, 53–67. http://dx.doi.org/10.1016/S0019-1035(03)00144-1.
Fairén, A., Davila, A., Lim, D., et al., 2010a. Astrobiology through the ages of Mars: the study of terrestrial analogs to understand the habitability of Mars. Astrobiology 10, 821–843. http://dx.doi.org/10.1089/ast.2009.0440.
Fairén, Alberto G., Chevrier, Vincent, Abramov, Oleg, Marzo, Giuseppe A., Gavin, Patricia, Davila, Alfonso F., Tornabene, Livio L., et al., 2010b. Noachian and more recent phyllosilicates in impact craters on Mars. Proc. Nat. Acad. Sci. U.S.A. 107 (27), 12095–12100.
Fairén, et al., 2011. Cold glacial oceans would have inhibited phyllosilicate sedimentation on early Mars. Nat. Geosci. 4, 667–670.
Fairén, et al., 2012. Glacial paleoenvironments on Mars revealed by the paucity of hydrated silicates in the Noachian crust of the Northern Lowlands. Planet. Space Sci. 70, 126–133.
Fairén, et al., 2014. A cold hydrological system in Gale crater, Mars. Planet. Space Sci. 93-94, 101–118.
Feldman, W.C., Boynton, W.V., Tokar, R.L., et al., 2002. Global distribution of neutrons from Mars: results from Mars Odyssey. Science 297, 75–78.
Fell, J.W., Scorzetti, G., Connell, L., Craig, S., 2006. Biodiversity of micro-eukaryotes in Antarctic Dry Valley soils with $<5\%$ soil moisture. Soil Biol. Biochem. 38, 3107–3119.
Fialips, C.I., Carey, J.W., Vaniman, D.T., Bish, D.L., Feldman, W.C., Mellon, M.T., 2005. Hydration state of zeolites, clays, and hydrated salts under present-day Martian surface conditions: can hydrous minerals account for Mars Odyssey observations of near-equatorial water-equivalent hydrogen? Icarus 178, 74–83.
Fink, W., Dohm, J.M., Tarbell, M.A., et al., 2005. Next-generation robotic planetary reconnaissance missions: a paradigm shift. Planet. Space Sci. 53, 1,419–1,426.
Flahaut, J., Quantin, C., Allemand, P., Thomas, P., Le Deit, L., 2010a. Identification, distribution and possible origins of sulfates in Capri Chasma (Mars), inferred from CRISM data. J. Geophys. Res. 115, E11007. http://dx.doi.org/10.1029/2009JE003566.
Flahaut, J., Quantin, C., Allemand, P., Thomas, P., 2010b. Morphology and geology of the ILD in Capri/Eos Chasma (Mars) from visible and infrared data. Icarus 207, 175–185. http://dx.doi.org/10.1016/j.icarus.2009.11.019.
Antarctic Climate Evolution. Developments in Earth and Environmental Sciences. In: Florindo, F., Siegert, M. (Eds.), 8. Elsevier.
Forget, F., Haberle, R.M., Montmessin, F., et al., 2006. Formation of glaciers on Mars by atmospheric precipitation at high obliquity. Science 311, 368–371.
Fountain, A.G., Nylen, T.H., Monaghan, A., Basagic, H.J., Bromwich, D., 2010. Snow in the McMurdo Dry Valleys, Antarctica. Int. J. Climatol. 30, 633–642.
Fredrickson, J.K., Balkwill, D.L., 2006. Geomicrobial processes and biodiversity in the deep terrestrial subsurface. Geomicrobiol. J. 23 (6), 345–356.
Fueten, F., Flahaut, J., Le Deit, L., Stesky, R., Hauber, E., Gwinner, K., 2011. Interior layered deposits within a perched basin, southern Coprates Chasma, Mars:

evidence for their formation, alteration, and erosion. J. Geophys. Res. 116, E02003. http://dx.doi.org/10.1029/2010JE003695.
Gendrin, A., Mangold, N., Bibring, J.P., Langevin, Y., Gondet, B., Poulet, F., Bonello, G., Quantin, C., Mustard, J., Arvidson, R., LeMouelic, S., 2005. Sulfates in Martian layered terrains: the OMEGA/Mars Express view. Science 307, 1587–1591.
Gibson, E.K., Wentworth, S.J., McKay, D.S., 1983. Chemical weathering and diagenesis of a cold desert soil from Wright Valley, Antarctica: an analog of Martian weathering processes. In: Proc. 13th Lunar and Planetary Science Conf. Part 2. Journal of Geophysical Research 88, A912-A928.
Glasauer, S., Weidler, P.G., Langley, S., Beveridge, T.J., 2003. Controls on Fe reduction and mineral formation by a subsurface bacterium. Geochim. Cosmochim. Acta 67 (7), 1277–1288.
Graham, I.J., Ditchburn, R.G., Claridge, G.G.C., Sheppard, D.S., Whitehead, N.E. Xondervan, A, Sheppard, D.S., 2002. Dating Antarctic soils using atmosphere-derived ^{10}Be and nitrate. In: Gamble, J.A., Skinner, D.N.B., Henry, S., (Eds.). Antarctica at the Close of a Millennium, Proc. Eighth Int. Symposium on Antarctic Earth Sciences, pp. 429–436.
Hart, K.M., Szpak, M.T., Mahaney, W.C., Frazer, A.R., Jordan, S.F., Dohm, J.M., Allen, C.C. R., Kelleher, B.P., 2011. Bacterial enrichment study and overview of the extractable lipids from Antarctic paleosols, Dry Valleys, Antarctica. Astrobiology 11, 303–321.
Hartmann, W.K., Neukum, G., 2001. Cratering chronology and the evolution of Mars. Space Sci. Rev. 96, 165–194. http://dx.doi.org/10.1023/A:1011945222010.
Havics, A., Milner, M.W., Netoff, D., Dohm, J.M., Kalm, V., Krinsley, D., Sodhi, R.N.S., Anderson, R.C., Boccia, S., Malloch, D., 2009. Blueberries on Earth and Mars: Correlations Between Concretions in Navajo Sandstone and Terra Meridiani on Mars. Chicago, McCrone Research Institute, IM-2009, 7 July, 2009.
Head, J.W., Pratt, S., 2001. Extensive Hesperian-aged south polar ice sheet on Mars: evidence for massive melting and retreat, and lateral flow and ponding of meltwater. J. Geophys. Res. 106, 12275–12299.
Head, J.W., Marchant, D.R., Agnew, M.C., Fassett, C.I., Kreslavsky, M.A., 2006. Extensive valley glacier deposits in the northern mid-latitudes of Mars: evidence for Late Amazonian obliquity-driven climate change. Earth Planet. Sci. Lett. 241, 663–671. http://dx.doi.org/10.1016/j.epsl.2005.11.016.
Hodgson, J.M. (Ed.), 1976. Soil Survey Field Handbook. Soil Survey Tech. Monog. No. 5. Rothamsted Experimental Station, Harpenden, Herts. UK, p. 99.
Ivanov, B.A., 2001. Mars/Moon cratering rate ratio estimates. Space Sci. Rev. 96, 87–104.
Jaisi, D.P., Dong, H., Liu, C., 2007. Influence of biogenic Fe(III) on the extent of microbial reduction of Fe(III) in clay minerals nontronite, illite and chlorite. Geochim. Cosmochim. Acta 71, 1145–1158.
Joughin, I.B., Alley, R.B., 2011. Stability of the West Antarctic ice sheet in a warming world. Nat. Geosci. 4, 506–513.
Karasozen, E., Andrews-Hanna, Dohm, J.M., Anderson, R.C., 2012. The formation mechanism of the south Tharsis ridge belt, Mars. In: Lunar Planetary Science Conference, XXXXIII, Abstract #2592 (CD-ROM).
Kargel, J.S., 2004. Mars: A Warmer Wetter Planet. Praxis-Springer p. 557.
Kargel, J.S., Strom, R.G., 1992. Ancient glaciation on Mars. Geology 20, 3–7. http://dx.doi.org/10.1130/0091-7613(1992)020 < 0003:AGOM > 2.3.CO;2.
Kawase, T., Saito, A., Sato, T., Kanai, R., Fujii, T., Nikaidou, N., Miyashita, K., Watanabe, T., 2004. Distribution and phylogenetic analysis of family 19 chitinases in *Actinobacteria*. Appl. Environ. Microbiol. 70, 1135–1144.
Konishchev, V.N., 1998. Relationship between the lithology of active—layer materials and mean annual ground temperature in the former USSR. In: Permafrost-Seventh International Conference Proc., Yellowknife, NT, 55. 591-594.
Krinsley, D.H., Doornkamp, J., 1973. Atlas of Sand Grain Surface Textures. Cambridge Univ. Press, Cambridge.
Laskar, J., Robutel, P., 1993. The chaotic obliquity of the planets. Nature 361, 608–612.
Lewis, A.R., Marchant, D.R., Ashworth, A.C., Hemming, S.R., Machlus, M.I., 2007. Major Middle Miocene global climate change: evidence from East Antarctica and Transantarctic Mountains. Geol. Soc. Am. Bull. 119, 1461.
Lewis, K.W., et al., 2008. Quasi-periodic bedding in the sedimentary rock record of Mars. Science 322, 1532–1535.
Davies, Lloyd, Atkins, M., Meer, C., van der, J.J.M., Barrett, P.J., Hicock, S., 2009. Cold-based glacier advance in the Allan Hills, Antarctica: evidence and preservation potential. Quat. Sci. Rev. 28, 3124–3137.
Mahaney, W.C., 2002. Atlas of Sand Grain Surface Textures and Applications. Oxford University Press, Oxford, UK p. 237.
Mahaney, W.C., 1990. Ice on the Equator. Wm Caxton Press, Ellison Bay, WIi p. 386.
Mahaney, W.C., 1995. Pleistocene and Holocene glacier thicknesses, transport histories and dynamics inferred from SEM microtextures on quartz particles. Boreas 24, 293–304.
Mahaney, W.C., Claridge, G., Campbell, I.A., 1996. Microtextures on quartz grains in tills from Antarctica. Palaeo 3 (121), 89–103.
Mahaney, W.C., Hancock, R.G.V., Sanmugadas, K., 1999. Extractable Fe and Al in soils of the Middle Teton chronosequence, western Wyoming, USA. Z. Geomorphol 43, 393–407.
Mahaney, W.C., Dohm, J.M., Baker, V.R., Newsom, H.E., Malloch, D., Hancock, R.G.V., Campbell, I., Sheppard, D., Milner, M.W., 2001. Morphogenesis of Antarctic paleosols: Martian analogue. Icarus 154, 113–130.
Mahaney, W., Miyamoto, H., Dohm, J.M., et al., 2007. Rock glaciers on Mars: Earth-based clues to Mars' recent paleoclimatic history. Planet. Space Sci. 55, 181–192. http://dx.doi.org/10.1016/j.pss.2006.04.016.
Mahaney, W.C., Dohm, J.M., Kapran, B., Hancock, R.G.V., Milner, M.W., 2009. Secondary Fe and Al in Antarctic paleosols: correlation to Mars with prospect for the presence of life. Icarus 203, 320–330.
Mahaney, W.C., Kalm, V., Hancock, R.G.V., Michel, F., Kapran, B., 2010. Geochemistry and extractable Fe and Al in cold temperature soils of northwestern Siberia. J. Quat. Sci. 25, 178–179.
Mahaney, W.C., Hart, K., Allen, C.C.R., Dohm, J.M., Hancock, R.G.V., Kelleher, B., Milner, M.W., 2012a. Coleoptera and microbe biomass in Antarctic Dry Valley paleosols adjacent to the Inland Ice: implications for Mars. J. Planet. Space Sci. 60, 386–398.
Mahaney, W.C., Dohm, J.M., Fairen, Alberto, 2012b. Weathering rinds on clasts: examples from Earth and Mars as short and long term recorders of paleoenvironment. J. Planet. Space Sci. 73, 243–253.
Mahaney, W.C., Keiser, Leslie, Krinsley, D.H., Pentlavalli, P., Allen, C.C.R., Somelar, P., Schwartz, Stephane, Dohm, James M., Dirzowsky, Randy, West, Allen, Julig, P., Costa, P., 2013. Weathering rinds as mirror images of palaeosols: examples from the Western Alps with correlation to Antarctica and Mars. J. Geol. Soc., Lyell Volume , http://dx.doi.org/10.1144/jgs2012-150.
Mahaney, W.C., Barendregt, R.W., Hamilton, T.S., Hancock, R.G.V., Costa, P.J.M., 2014. Soil mineralogy and chemistry of Late Pliocene-Early Pleistocene paleosols on Mount Kenya: age and paleoclimate reconstruction. Geomorphology 204, 217–228.
Mangold, N., Poulet, F., Mustard, J.F., Bibring, J.P., Gondet, B., Langevin, Y., Ansanm, V., Masson, Ph., Fassett, C., Head III, J.W., Hoffman, H., Nuekum, G., 2007a. Mineralogy of the Nili Fossae region with OMEGA/Mars Express data: 2. Aqueous alteration of the crust. J. Geophys. Res. 112, 1–25. http://dx.doi.org/10.1029/2006JE002835.
Mangold, N., et al., 2007b. An overview of the sulfates detected in the equatorial regions by the OMEGA/MEX spectrometer. In: Seventh International Conference on Mars, July 9–13, 2007, Pasadena CA [CD-ROM], LPI Contrib., 1353, Abstract 3141.
Marchant, D.R., Denton, G.H., Sugden, D.E., Swisher III, C.C., 1993. Miocene glacial stratigraphy and landscape evolution of the western Asgard Range, Antarctica. Geogr. Ann 75A, 303–330.
Marchant, D.R., Head, J.W., 2007. Antarctic dry valleys: microclimate zonation, variable geomorphic processes, and implications for assessing climate change on Mars. Icarus 192 (1), 187–222. http://dx.doi.org/10.1016/j.icarus.2007.06.018.
Marzo, G.A., Davila, A.F., Tornabene, L.L., Dohm, J.M., Fairén, A.G., Gross, C., Kneissl, T., Bishop, J.L., Roush, T.L., McKay, C.P., 2010. Evidence for Hesperian impact-induced hydrothermalism on Mars. Icarus 208, 667–683. http://dx.doi.org/10.1016/j.icarus.2010.03.013.
McLeod, M., Bockheim, J., Balks, M., Aislabie, J., 2009. Soils of western Wright Valley, Antarctica. Antarct. Sci. 21, 355–365.
Meslin, P-Y., Gasnault, O., Forni, O., Schröder, S., Cousin, A., Berger, G., Clegg, S.M., et al., 2013. Soil diversity and hydration as observed by ChemCam at Gale Crater, Mars (2013). Science 341 (6153), 1238670.
Moore, J.M., 2004. Blueberry fields forever. Nature 428, 711–712. http://dx.doi.org/10.1038/428711a.
Morris, R.V., Ruff, S.W., Gellert, R., Ming, D.W., Arvidson, R.E., Clark, B.C., Golden, D. C., Siebach, K., Klingelhöfer, G., Schröder, C., Fleischer, I., Yen, A.S., Squyres, S.W., 2010. Identification of carbonate-rich outcrops on Mars by the spirit rover. Science 329, 421–424. http://dx.doi.org/10.1126/science.1189667.
Moore, D.M., Reynolds Jr., R.C., 1997. X-ray Diffraction and the Identification and Analysis of Clay Minerals. Oxford University Press, Oxford, UK p. 378.
Murchie, S.L., et al., 2009a. The compact reconnaissance imaging spectrometer for Mars investigation and data set from the Mars Reconnaissance Orbiter's primary science phase. J. Geophys. Res. 114, E00D07. http://dx.doi.org/10.1029/2009JE003344.
Murchie, S.L., et al., 2009b. A synthesis of Martian aqueous mineralogy after 1 Mars year of observations from the Mars Reconnaissance Orbiter. J. Geophys. Res. 114, E00D06. http://dx.doi.org/10.1029/2009JE003342.
National Soil Survey Center (NSSC), 1995. Soil Survey Laboratory Information Manual. Soil Survey Investigations Report no. 45, version 1.00, U.S.D.A., pp. 305.
Niemela, S.I., Riekkola-Vanhanen, M., Sivela, C., Viguera, F., Tuovinen, O.H., 1994. Nutrient effect on the biological leaching of a black-schist ore. Appl. Environ. Microbiol. 60, 1287–1291.
NRCS, 〈http://soils.usda.gov/〉.
Ormö, J., Komatsu, G., Chan, M.A., Beitler, B., Parry, W.T., 2004. Geological features indicative of processes related to the hematite formation in Meridiani Planum and Aram Chaos, Mars: a comparison with diagenetic hematite deposits in southern Utah, USA. Icarus 171, 295–316.
Osterloo, M.M., Hamilton, V.E., Bandfield, J.L., Glotch, T.D., Baldridge, A.M., Christensen, P.R., Tornabene, L.L., Anderson, F.S., 2008. Chloride-bearing materials in the southern highlands of Mars Science 319, 1651–1654. http://dx.doi.org/10.1126/science.1150690.
Oyama, M., Takehara, H., 1970. Standard Soil Color Charts, Japan Research Council for Agriculture, Forestry and Fisheries.
Palache C., Berman H. and Fondel C., Dana's System of Mineralogy (7th Edition), John Wiley & Sons, New York, 11, 1951, 494–495.
Parmesan, C., 2006. Ecological and evolutionary responses to recent climatic change. Annual Review. Ecol. Evol. Syst 37, 637–669.
Passchier, S., 2011. Climate change: ancient Antarctic fjords. Nature 474, 46–47.
Pointing, S.B., Chan, Y., Lacap, D.C., Lau, M.C.Y., Jurgens, J.A., Farrell, R.L., 2009. Highly specialized microbial diversity in hyper-arid polar desert. PNAS 106, 19964–19969.

Poulet, F., Bibring, J.-P., Mustard, J.F., Gendrin, A., Mangold, N., Langevin, Y., Arvidson, R.E., Gondet, B., Gomez, C., 2005. Phyllosilicates on Mars and implications for early Martian climate. Nature 438, 623–627.

Quantin, C., Allemand, P., Mangold, N., Dromat, G., Delacourt, C., 2005. Fluvial and lacustrine activity on layered deposits in Melas Chasma, Valles Marineris, Mars. J. Geophys. Res. 110, E12S19. http://dx.doi.org/10.1029/2005JE002440.

Retallack, G.J., Krull, E.S., 1999. Neogene paleosols of the Sirius Group, Dominion Range, Antarctica. U.S. Antarct. J 30, 10–13.

Retallack, G.J., Seyedolali, A., Krull, E.S., Holser, W.T., Ambers, C.P., Kyte, F.T., 1998. Search for evidence of impact at the Permian–Triassic boundary in Antarctica and Australia. Geology 26, 976–992.

Retallack, G.J., Krull, E.S., Bockheim, J.G., 2001. New grounds for reassessing palaeoclimate of the Sirius Group, Antarctica. J. Geol. Soc. London 158, 925–935.

Retallack, G.J., Greaver, T., Jahren, H., 2007. Return to Coalsack Bluff and the Permian–Triassic boundary in Antarctica. Global Planet. Change 55, 90–108.

Rieder, R., Economou, T., Wänke, H., et al., 1997. The chemical composition of Martian soil and rocks returned by the mobile alpha proton X-ray spectrometer: preliminary results from the X-ray mode. Science 278, 1771–1774. http://dx.doi.org/10.1126/science.278.5344.1771.

Rignot, E., Hallet, B., Fountain, A., 2002. Rock glacier surface motion in Beacon Valley, Antarctica, from synthetic-aperture radar interferometry. Geophys. Res. Lett. 29 (12), http://dx.doi.org/10.1029/2001GL013494.

Ring, R.A., 1981. The physiology and biochemistry of cold tolerance in Arctic insects. J. Therm. Biol. 6, 219–229.

Ring, R.A., Tesar, D., 1980. Cold-hardiness of the Arctic beetle, *Pytho americanus* Kirby coleoptera, Pythidae (Salpingidae). J. Insect Physiol. 26, 763–774.

Roberts, J.H., Arkani-Hamed, J., 2012. Impact-induced mantle dynamics on Mars. Icarus 218, 278–289.

Roberts, J.H., Lillis, R.J., Manga, M., 2009. Giant impacts on erly Mars and the cessation of the Martian dynamo. J. Geophys. Res. 114, E04009. http://dx.doi.org/10.1029/2008JE003287.

Rossi, A.P., van Gasselt, S., Pondrelli, M., et al., 2010. Evolution of periglacial landforms in the ancient mountain range of Thaumasia highlands, Mars. In: Balme, M.R., Bargerym, A.S., Gallagher, C.J., Gupta, S. (Eds.), Geomorphology on Mars and Other Planets, 356. Geological Society of London Special Publication, pp. 69–85.

Schulze-Makuch, D., Houtkooper, J.M., Knoblauch, M., et al., 2007. The Biological Oxidant and Life Detection (BOLD) Mission: An outline for a new mission to Mars. In: Hoover, R.B., Levin, G.V., Rozanov, A.Y., Davies, P.C.W., (Eds.), Proceedings of SPIE—Instruments, Methods, and Missions for Astrobiology X: 6694-23, v. 6694, 10.1117/12.732155.

Schwamborn, G., Schirrmeister, L, Diekman, B., 2013. Onset of intense permafrost conditions in Northern Eurasia at ~2.55 Ma seen in a cryogenic weathering record from Lake El'gygytgyn. Clim. Past 9, 6255–6285.

Scott, D.H., Tanaka, K.L., 1986. Geologic Map of the Western Equatorial Region of Mars. U.S. Geological Survey Miscellaneous Investigations Series Map I-1802-A, scale 1:15,000,000.

Shean, D.E., Marchant, D.R., 2010. Seismic and GPR surveys of Mullins Glacier, McMurdo Dry Valleys, Antarctica: ice thickness, internal structure and implications for surface ridge formation. J. Glaciol. 56, 48–64. http://dx.doi.org/10.3189/002214310791190901.

Simpson, A.J., Simpson, M.J., Smith, E., Kelleher, B.P., 2007. Microbially derived inputs to soil organic matter: are current estimates too low? Environ. Sci. Technol. 41, 8070–8076.

Smith, P.H., Tamppari, L.K., Arvidson, R.E., et al., 2009. H2O at the Phoenix landing site. Science 325, 58–61.

Soare, R.J., Conway, S.J., Dohm, J.M., El-Maarry, M.R., 2014a. Possible ice-wedge polygons and recent landscape modification by "wet" periglacial processes in and around the Argyre impact basin, Mars. Icarus 233, 214–228.

Soare, R.J., Conway, S.J., Dohm, J.M., El-Maarry, M.R., 2014b. Possible open-system (hydraulic) pingos in and around the Argyre impact regions of Mars. Earth Planet. Lett. 398, 25–36.

Solomon, S.C., Aharonson, O., Aurnou, J.M., Banerdt, W.B., Carr, M.H., Dombard, A.J., Herbert, V., Frey, H.V., 2005. New perspectives on ancient Mars. Science 307 (5713), 1214–1220. http://dx.doi.org/10.1126/science.1101812.

Souness, C., Hubbard, B., Milliken, R.E., Quincey, D., 2012. An inventory and population-scale analysis of Martian glacier-like forms. Icarus 217 (1), 243–255. http://dx.doi.org/10.1016/j.icarus.2011.10.020.

Squyres, S.W., Arvidson, R.E., Ruff, S., Gellert, R., Morris, R.V., Ming, D.W., Crumpler, L., Farmer, J.D., Des Marais, D.J., Yen, A., McLennan, S.M., Calvin, W., Bell III, J.F., Clark, B.C., Wang, A., McCoy, T.J., Schmidt, M.E., de Souza, P.A., 2008. Detection of silica-rich deposits on Mars. Science 320, 1063–1067. http://dx.doi.org/10.1126/science.1155429.

Tanaka, K.L., Skinner, J.A., Jr., Dohm, J.M., Irwin, R.P., III, Kolb, E.J., Fortezzo, C.M., Platz, T., Michael, G.G., Hare, T.M., 2014, Geologic map of Mars. U.S. Geological Survey Scientific Investigations Map 3292, scale 1:20,000,000, pamphlet 43 p. ⟨http://dx.doi.org/10.3133/sim3292⟩.

Tanaka, Kenneth L., 1986. The stratigraphy of Mars. J. Geophys. Res. Solid Earth 91 (B13), E139–E158. http://dx.doi.org/10.1029/JB091iB13p0E139.

Torrent, J., Cabedo, A., 1986. Sources of iron oxides in reddish brown soil profiles from calcareinites in southern Spain. Geoderma 37, 57–66.

Touma, J., Wisdom, J., 1993. The chaotic obliquity of Mars. Science 259, 1294–1296.

Trujillo, M.E., 2008. Actinobacteria. Encycl. Life Sci. , http://dx.doi.org/10.1002/9780470015902.a0020366.

Ugolini, F.C., Anderson, D.M., 1973. Ionic migration and weathering in frozen Antarctic soils. Soil Sci. 115, 461–470.

Velasco, J., Millán, A., Hernández, J., Gutiérrez, C., Abellán, P., Sánchez, D., Ruiz, M., 2006. Response of biotic communities to salinity changes in a Mediterranean hypersaline stream. Saline Syst 2, 1–15.

Virginia, R.A., Wall, D.H., 1999. How soils structure communities in the Antarctic Dry Valleys. Bioscience 49 (12), 973–983.

Warny, S., Askin, R.A., Hannah, M.J., Mohr, B.A.R., Raine, J.I., Harwood, D.M., Florindo, F., and the SMS Science Team, 2009. Palynomorphs from a sediment core reveal a sudden remarkably warm Antarctic during the Middle Miocene. Geology 10, 955–958.

Webster, J., Weber, R.W.S, 2007. Introduction to Fungi, third ed. Cambridge University Press, Cambridge, UK p. 841.

Wiens, R.C., Maurice, S., Blaney, D.L., et al., 2014. Geochemistry at Gale from CHEMCAM: implications for Martian igneous and sedimentary processes and for habitability. In: Eighth International Conference on Mars (2014), Abstract #1170.

Wilson, A.T., 1973. The great antiquity of some Antarctic landforms. Evidence for an Eocene temperate glaciation in the McMurdo region. In: van Zinderen Bakker, E,M. (Ed.), Palaeoecology of Africa. Balkema, Capetown, pp. 23–35.

Wray, J.J., Murchie, S.L., Squyres, S.W., et al., 2009. Diverse aqueous environments on ancient Mars revealed in the southern highlands. Geology 37, 1043–1046. http://dx.doi.org/10.1130/G30331A.1.

Wray, J.J., Hansen, S.T., Dufek, J., et al., 2013. Prolonged magmatic activity on Mars inferred from the detection of felsic rocks. Nat. Geosci. 6, 1013–1017. http://dx.doi.org/10.1038/ngeo1994.

Wynn-Williams, D.D., Edwards, H.G.M., Newton, E.M., 2000. The detection of cold desert biomarkers, ancient and modern: Antarctica and Mars. In: Abstract Volume, The Second International Conference on Mars Polar Science and Exploration, Reykjavik, Iceland, August 21–25, 2000, pp. 182–183.

Yilmaz, H., Sezen, K., Kati, H., Demirbağ, Z., 2006. The first study on the bacterial flora of the European spruce bark beetle, *Dendroctonus micans* (Coleoptere: Scolytidae). Biologia 61, 679–686.

Young, D., Wright, A.P., Roberts, J.L., Warner, R.C., Young, N.W., Greenbaum, J.S., Schroeder, D.M., Holt, J.W., Sugden, D.E., Blankenship, D.D., van Ommen, T.D., Siegert, M.J., 2011. A dynamic early East Antarctic Ice Sheet suggested by ice-covered fjord landscapes. Nature 474, 72–75.