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Towards a glacial subdivision of the Ediacaran Period, with an example of the Boston Bay Group, Massachusetts

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ABSTRACT

After the Elatina glaciation of Snowball Earth, at least four distinct glacial advances and sea-level retreats punctuated Ediacaran time: Gaskiers glaciation (580 Ma), Fauquier glaciation (571 Ma), Bou-Azzer glaciation (566 Ma) and Hankalchough glaciation (551 Ma). Tillites or diamicrites are commonly controversial, but periglacial paleosols with distinctive physical structure and degree of chemical weathering offer supporting evidence of glaciation and sea-level change useful for stratigraphic correlation. This paper reviews glacial advances of the Ediacaran stratotype and other sequences, and also reveals the value of paleosols and chemical index of alteration to understand the upper Squantum and Brookline members of the Roxbury Conglomerate near Boston, Massachusetts. The Boston Bay ice wedges are periglacial paleosols, and evidence of maritime glacial climate like that of modern coastal Greenland and Arctic Canada. Simple discoidal vendobiont fossils (Aspidella terranoivica) in the Dorchester Member of the Roxbury Conglomerate and in the Cambridge Argillite are in heterolithic shale-siltstone facies that are interpreted as intertidal to shallow marine environments. Local marine transgressions and other paleosols showing significant chemical weathering represent temperate interglacial paleoclimates. Short glacial advances affecting climate and sea-level enable subdivision of the Ediacaran Period.

KEY POINTS

1. Four distinct glacial advances and sea-level retreats punctuated Ediacaran time: Gaskiers (580 Ma), Fauquier (571 Ma), Bou-Azzer (566 Ma), Hankalchough (551 Ma).
2. Paleosols with distinctive structures such as ice wedges were periglacial.
3. Squantum Member diamicrites near Boston, Massachusetts are Gaskiers age.

Introduction

The Ediacaran is the longest geological period (Xiao & Narbonne, 2020), and efforts to subdivide it have included glacial advances (Xiao et al., 2016), such as the Gaskiers glaciation of Newfoundland (Eyles & Eyles, 1989; Myrow & Kaufman, 1999; Pu et al., 2016; Retallack, 2013a). The rationale for using glacial advances as temporal markers is that such times in Earth history, like the Pleistocene, Permian, and Cryogenian, are not only revealed by glacial tillites, but by a suite of other changes in paleosols, climatic belts, global sea-level, and ecosystems (Chumakov, 2011a; French, 1996; Hoffmann & Li, 2009; Retallack, 1999, 2011; Williams et al., 2008; Xiao & Narbonne, 2020). These other lines of evidence are reviewed here and detailed with an example from the Boston Bay Group of Massachusetts. Interpretation of coarse-grained rocks as tillites or conglomerates has a long and controversial history, and few units have been more controversial than the tillites and conglomerates of the Squantum Member of the Boston Bay Group (Carto & Eyles, 2011; Dott, 1961; Lahee, 1914a, 1914b; Lindsay et al., 1970; Mansfield, 1906; Rehmer, 1981; Rehmer & Hepburn, 1974; Sayles, 1914; Smith & Soci, 1990; Soci & Smith, 1987, 1990; Stuart et al., 1975).

This study attempts to widen the identification of Ediacaran glacial episodes with evidence from paleosols, geochemical indices such as chemical index of alteration (CIA; Nesbitt & Young, 1982), and fossils. Paleosols in Ediacaran sedimentary rocks are subtle and lack many features of post-Silurian paleosols, such as root traces (Driese et al., 1997; Retallack & Huang, 2011), but do preserve soil horizons with chemical weathering trends, and soil structures, such as ice wedges (Williams, 1986; Williams et al., 2016), peds, and nodules (Retallack, 2011, 2012, 2013b, 2016a, 2016b; Retallack et al., 2015). Ediacaran paleosols are now sufficiently abundant to form the basis of paleoclimatic time series useful for correlating glacioeustatic events worldwide (Retallack, 2014a, 2016a, 2016b; Retallack et al., 2014). Periglacial ice wedges and other terrestrial frost-deformation features are now widely recognised as...
parts of a distinctive group of frigid soils dominated by physical rather than chemical weathering (gelsols of Soil Survey Staff, 2014). Bulk chemical analysis and density determinations have been used here to calculate chemical indices of weathering (Nesbitt & Young, 1982; Retallack, 1997), and also geochemical strain and mass transfer (Brimhall et al., 1992). Not only can periglacial paleosol-bearing facies be identified, but also cool temperate intertidal to marine megafossils such as *Aspidella* (Bailey & Bland, 2001) and microfossils such as *Bavinella* (Lenk et al., 1982).

**Toward an Ediacaran glacial stratigraphy**

**Ediacaran stratotype record of glacials in South Australia**

The stratotype section of the Ediacaran Period (Figure 1) is in the Flinders Ranges, just east of the Ediacara Hills of South Australia (Knoll et al., 2006). Four stratigraphic levels with glacial features are recognised in the Ediacaran sequence of South Australia: (1) dropped pebbles in the Bunyeroo Formation (Gostin et al., 2010, 2011), (2) a deep (600 m) palaeovalley and palaeokarst from sea-level drop exposing marine carbonates of the Wonoka Formation (Retallack et al., 2014; von der Borch et al., 1989; Williams & Schmidt, 2018), (3) dropped pebbles in the Blythes Springs Formation (Jenkins, 2011), and (4) periglacial soil involutions and needle ice in the Ediacara Member (Retallack, 2013a; 2016a). The sequence of these glacial advances is clear, but their age is uncertain due to lack of radiometric dating (Retallack et al., 2014). Figure 1 shows likely correlations with better dated glacial advances elsewhere (Figure 2).

Paleosols in the Flinders Ranges are common in red beds of Cryogenian, Ediacaran, and Cambrian age (Figure 1). They were first recognised by ice wedges and sand wedges (Williams, 1986; Williams et al., 2016). Other field characters are sharp top with drab-haloed filament traces, desiccation cracks, diffuse horizons of sulfate evaporate sand crystals, and carbonate nodular horizons (Retallack, 2012). Granulometry of the red siltsstones revealed poorly sorted, highly angular, silt-size grains characteristic of loess (Mawson & Segnit, 1949; Retallack, 2012). Wind-drift loess on fluvial levees also explains distinctive interflag sandstone laminae that preserve Ediacaran fossils (Retallack, 2019; Tarhan, Droser, et al., 2017a). Petrographic study revealed clay formation and oxidation from feldspar as in soils (Retallack, 2012, 2013b). Chemical differentiation of the individual profiles demonstrated bed deflation (negative strain) and cation depletion (negative mass transfer) characteristic of soils (Retallack, 2012, 2013b). Carbonate nodules of soils had a very strong (R² > 0.9) correlation of δ¹⁸O and δ¹³C characteristic of soil, in contrast with marine limestones and dolostones in the sequence, which show no such correlation (Retallack, 2016b). Finally, early diagenetic silica cements in weakly developed paleosols had Ge/Si ratios of up to 9 μmol/mol, much greater than the upper limit for marine silica of 1 μmol/mol (Retallack, 2017; Tarhan, Hood, et al., 2017b; Tarhan et al., 2016). These weakly developed paleosols were previously regarded as marine based on oscillation ripple marks and flaggy sandstones (Tarhan et al., 2015).

Several features of the paleosols are evidence of paleoclimate. Depth to carbonate nodules (BK or calcic horizon) and to sulfate sand crystals (By or gypsic horizon) in Ediacaran paleosols are guides to both paleoprecipitation and paleoproductivity (Retallack, 2012, 2013b), by comparison with modern soils (Breecker & Retallack, 2014; Retallack, 2005; Retallack & Huang, 2010). Compaction-corrected BK depth is evidence for semiarid to arid paleoclimate of 139–579 ± 149 mm mean annual precipitation for the Ediacara Member (Retallack, 2013b), 140–407 ± 149 mm for the Bonney Sandstone, 377 ± 149 mm for the upper Wonoka Formation, 377 ± 149 mm for the upper Bunyeroo Formation (Retallack et al., 2014), 432 ± 149 mm for the lower Brachina Formation, and 342 ± 149 mm for the Nuccaleena Formation (Retallack, 2011). Known Ediacaran glaciations generally correspond with times of local aridity in the Flinders Ranges, although these data are incomplete (Figure 1).

Several periglacial features have been documented in paleosols of the Ediacara Member of the Rawnse Quartzite. Polygonal expansion cracks in the sandy surface of paleosols are most like thurf mounds of surficial ice disruption (Retallack, 2012). Soft-sediment slumping in successive episodes of thixotropy can be taken as evidence of periglacial involutions like those in the intermittent permafrost zone (Retallack, 2012). Thurf mounds of the Nuccaleena Formation were embelished by linear shadow dunes (Retallack, 2011). Paleotemperatures of paleosols can be estimated from a chemical index of weathering from major-element chemical composition of paleosols (Oskarsson et al., 2012), and are surprisingly cool at 10.6–11.5 ± 0.4°C mean annual temperature for the Ediacara Member (Retallack, 2013b), 10.3–10.8 ± 0.4°C for the Bonney Sandstone, 7.3 ± 0.4°C for the lower Brachina Formation, and 4.2 ± 0.4°C mm for the upper Elatina Formation (Retallack, 2011). The Ediacara, like the Cryogenian, was evidently a cool period in Earth history, because paleomagnetism of Ediacaran rocks indicates temperate paleolatitude: 30.0 ± 3.5° for the Brachina Formation (Schmidt & Williams, 2010), and 30.0 ± 3.1° for the Wonoka Formation (Schmidt & Williams, 2013).

The principal reason for naming the Ediacaran stratotype in South Australia was the diverse and enigmatic fossil biota of the Ediacara Hills (Fedonkin et al., 2008; Knoll et al., 2006). Also diverse is an associated assemblage of large microfossil acritarchs, although this assemblage was overturned by a major mass extinction above the Wonoka Formation (Grey, 2005; Grey et al., 2003). Opinion remains divided on the biological nature of these fossils, or even whether they are marine or terrestrial (Liu et al., 2015;
Figure 1. Long record of paleosols and carbonate stable isotopic variation in the Flinders Ranges, South Australia. (a) Composite stratigraphic section in Brachina Gorge (Mawson, 1939a) and Ten Mile Creek (Mawson, 1939b). (b) Depth to calcareous nodules in moderately developed paleosols (Retallack, 2008, 2011, 2012). (c, d) Stable isotope compositions of paleosol carbonate (caliche nodules), paleokarst (micritised and ferruginised limestone) and unaltered marine carbonate (data from Calver, 2000; Eickhoff et al., 1988; McKirdy et al., 2001; Retallack, 2008, 2012; Swanson-Hysell et al., 2010; and herein), with named isotopic excursions (after Halverson et al., 2005) and glaciations (Hebert et al., 2010; Jenkins, 2011; Landing & McGabhann, 2010; Williams et al., 2008). Compaction corrected mean annual precipitation is derived from depth to Bk by algorithms of Retallack (2005) and Sheldon and Retallack (2001).
Nevertheless, macrofossils are rare in association with Ediacaran tillites, which include distinctly different microfossils (*Bavilinella faveolate*; Lenk et al., 1982). The Ediacaran stratotype sequence of South Australia with its diverse biota was evidently remote from glaciers.

**Ediacaran glacial advances elsewhere**

The second most fossiliferous and complete sequence of Ediacaran rocks, after the South Australian stratotype, is in Newfoundland (Figure 2d). The Newfoundland sequence also features Ediacaran tillites of the Gaskiers Formation.
(Williams & King, 1979), dated by U–Pb on zircons in volcanic ash at 580.90 ± 0.40 and 579.88 ± 0.44 Ma (Pu et al., 2016). These tillites have been considered submarine (Eyles & Eyles, 1989). More likely, Gaskiers tillites were moraines on land, because they were interbedded with red paleosols as recognised by clay enrichment and depletion of alkali and alkaline earth elements at the expense of feldspar and rock fragments, blocky pedds, argillans and sepic plasmic fabric (Retallack, 2013a). In addition, crystal and lapilli volcanic tuffs overlying fossil beds of the Drook and Mistaken Point formations do not show normal grading required by deposition under water, and show matrix-supported accretionary lapilli, volcanic bombs, and gas escape structures only found in land (Retallack, 2014a, 2016a). These observations falsify past views that Ediacaran megafossils of Newfoundland were deep marine (Liu et al., 2015). A sub-Cambrian (>542 Ma) glacial pavement in nearby Labrador and Quebec was on Laurentia, not Avalonia at that time (Swett, 1981), but the Mystery Lake (531 Ma) glaciation in New Brunswick is evidence of Cambrian glaciation within the same terrane as the Newfoundland Ediacaran sequence (Landing & MacGabhann, 2010). Red beds in the Newfoundland sequence of largely grey marine rocks may record eustatic sea-level retreat at times of glacial advance, including the upper Drook Formation now dated at 570.94 ± 0.33 Ma and the Mistaken Point Formation dated at 566.25 ± 1.3 Ma (Pu et al., 2016). These may correlate with the Fauguier and Bou-Azzer glaciations, respectively (Figure 1).

The Bou-Azzer tillite of the Ouaraazarte Group of Morocco (Vermhet et al., 2012) is below an ignimbrite dated at 561 ± 9.5 Ma (Letsch et al., 2019), and above an ignimbrite dated at 565 ± 6 Ma (Karaoui et al., 2015). The Bou-Azzer tillite is correlated with the Weesenstein-Orellana glaciation of Germany and Spain, ca 565 Ma from detrital zircons (Linnemann et al., 2018). This widespread West African and European glaciation may correspond with 600 m glacioeustatic sea-level drop during deposition of red beds of the Drook and Mistaken Point formations of Newfoundland (Retallack, 2014a, 2016a). Another glaciation more loosely constrained by detrital zircons in Morocco was between 592 and 579 Ma, and may correspond with the Gaskiers glaciation of Newfoundland (Letsch et al., 2018).

The Newfoundland sequence shares exceptional thickness (4–8 km), common red beds, calcalkaline volcanic composition, and terrestrial andesitic volcanic and granitic basement with other Avalonian terranes (Figure 2) of Ediacaran age (Keppie & Dallmeyer, 1989; Nance, 1990; Rast & Skehan, 1990; Thompson et al., 1996). Terrestrial red beds, ash-flow tuffs and unpillowed andesitic flows are characteristic of the Uwharrie Formation of North Carolina (Ingle et al., 2003), Robertson River Volcanics and interbedded Mechum River Formation of Virginia (Hebert et al., 2010; Tollo & Hutson, 1996), Lynn and Mattapan volcanics of Boston (Thompson & Bowring, 2000), Harbour Main Formation of Newfoundland (Williams & King, 1979), and Ragleth Tuffs and other Uriconian volcanic rocks of Longmynd in Shropshire (Compston et al., 2002; Wright, 1968). Basement to the Charnwood sequence of Leicestershire is not exposed and unlike Shropshire, this sequence is just outboard of the Thingstone Fault, which defines the Midlands microcontinent (McIlroy et al., 1998). Nevertheless, bouldery calcalkaline and hyperalkaline volcanics in the Charnwood Forest sequence are evidence of an emergent volcanic source very close at hand (Carney, 1999). These and the Newfoundland sequence were all convergent margin forearc basins to continental volcanic arcs (Retallack, 2013a).

As in Newfoundland, other Avalonian terranes have an incomplete record of Ediacaran glaciations, mainly Hankalchow (551 Ma) and Bou-Azzer (566 Ma) glaciations, but also the Fauguier glaciation (571 Ma). In the Carolina Slate Belt, small (2–3 cm diameter), tuffaceous dropstones in clayey siltstone rhythmites of the middle Tillery Formation (Gibson & Teeter, 1984) are younger than 553 Ma (Ingle et al., 2003), but such small dropstones could be from far-travelled icebergs unreflective of global glacial advance (Weaver et al., 2006). In Virginia, a thick diamictite (Rockfish Conglomerate Member) of the Fauguier Formation is below Catotin Volcanics dated at 571 ± 1 Ma (Hebert et al., 2010). Coarse-grained rocks of the Mechum River Formation are younger than 729 Ma basement plutons, and perhaps also coeval with 705–702 Ma volcanics (Tollo & Hutson, 1996), so may be Cryogenian in age (Chumakov, 2011b; Hoffman & Li, 2009). No glacial facies have been recognised in Ediacaran rocks of England, but there are two suspect coarse-grained facies. On the eastern slopes of Longmynd in Shropshire, these are (1) Helmeth Grits of the basal Stretton Shale locally dated as older than 567 Ma, and (2) Huckster Conglomerate at the base of the Portway Formation locally dated as younger than 556 Ma. The Huckster Conglomerate shows large local thickness variations, and Helmeth Grits include unusually large blocks of Uriconian Volcanics (Wright, 1968). In Charnwood Forest, there are also two coarse-grained horizons: (1) the Sliding Stones Breccia Member of the uppermost Beacon Hill Formation younger than 559 Ma and (2) the South Quarry Breccia Member of the uppermost Ives Head Formation older than 566 Ma (Carney, 1999; Compston et al., 2002). Both have been interpreted as volcanioclastic submarine slump breccias (Boynton & Carney, 2003; Sutherland et al., 1987). Their ages are similar to Hankalchow and Bou-Azzer glaciations.

A thin tillite in the Egan Formation of the Kimberley region of Western Australia underlies stromatolites of Tungussia julia (Corkeron, 2007), also known from the upper Wonoka Formation of South Australia (Retallack et al., 2014). This tillite may be correlative with the Fauguier glaciation (571 Ma).

Three glacioeustatic disconformities marked by paleokarst are known from the Doushantou Formation of the
Yangtze Gorges area of Hubei, China: (1) above ash dated at 632.5 ± 0.5 Ma, (2) above an ash dated at 576 ± 14 Ma and below the ECAP acritarch extinction, and (3) below an ash dated at 551.1 ± 0.7 Ma (Condon et al., 2005; Xiao et al., 2012). Chinese paleokarst 3 is the same age as the Hankalchough glaciation, paleokarst 2 coeval with the Bou-Azzer glaciation, and paleokarst 1 may be combination of the Fuauquier and Gaskiers glaciations (Figure 1).

In the White Sea region of Russia, a low CIA (<67) is found in the upper Verkhov to Erga formations, which include a U–Pb age of 555.3 ± 0.3 Ma, and before that in the lower Lyamitsa Formation below a U–Pb date 558 ± 1 Ma (Grazhdankin et al., 2005). A bed in the lower Erga Formation also shows permafrost frost boils (Retallack, 2016b). In the north and central Urals of Russia are several diamictites: one in the Starye Pechi Formation between 561 ± 36 Ma from U–Pb on zircons and 559 ± 16 Ma by Rb–Sr on trachyandesite pyroxene, above two horizons in the Koiva Formation and another two horizons in the Tany Formation all younger than 598.1 ± 6 Ma from U–Pb of zircons (Chumakov, 2011b; Maslov et al., 2013). These may represent the Hankalchough (551) and Bou-Azzer (566 Ma) glaciations.

Central Asia and West Africa have extensive evidence of the latest Ediacaran advance widely named Hankalchough glaciation. Glacial tillites are known from the successive Bayisi, Tereeken, and Hankalchough formations of the Quruqtagh region of the eastern Chinese Tianshan (Chumakov, 2009). Tereeken and Bayisi formations are older than 615 ± 6 Ma and are likely Cryogenian (Xu et al., 2009). The Hankalchough Formation postdates a deep carbon isotope anomaly considered equivalent to the global Shuram-Wonoka anomaly and late Ediacaran fossils of Vendobonata and is also below fossils of the basal Cambrian (Chumakov, 2009; Xiao, 2004). The Hongtegou Formation of the Chinese Tianshan is dated at 558 Ma (Ren et al., 2010). The Baykonur Formation of Kazakhstan and Kyrgyzstan includes diamictites directly underlying marine basal Cambrian (Chumakov, 2009). Glacial pavements, dropped pebbles and diamictites of the Zhengmuguan and Luquan diamictites in the Qinglefing range of Henan, North China, are associated with acritarchs of late Ediacaran age and underlie early Cambrian carbonates (Baode et al., 1986). Diamictites of the Fersiga Group of Algeria are above basement granites dated by Rb–Sr at 556 ± 12 Ma, and below volcanics dated by the same method at 519 ± 11 Ma (Bertrand-Sarfati et al., 1995).

The Mortensnes Formation in northern Norway has diamictites filling a paleovalley 400 m thick (Halverston et al., 2005). The Mortensnes tillite is correlated by sequence stratigraphy with the Moelv Tillite of southern Norway, which underlies Ekre shale poorly dated by Rb–Sr isochron at 612 ± 18 Ma (Vidal & Moczydlowska, 1995), and overlies Rendalen Formation with detrital zircons dated by U–Pb at 620 ± 14 Ma (Bingen et al., 2005). These are of uncertain Ediacaran or Cryogenian age.

**Controversies concerning the Boston Bay Group of Massachusetts**

**Stratigraphy and age**

The city and bay of underlain by the Boston Bay Group, a 5.7-km-thick sequence of volcanioclastic sedimentary rocks (Billings, 1929, 1976; Billings et al., 1939; Thompson et al., 2014) unconformably overlying Lynn and Mattapan volcanics, and Dedham Granite (Figures 3–5). The Roxbury Conglomerate at the base of the Boston Bay Group includes andesitic subaerial Brighton Volcanics at several stratigraphic levels and has been divided, in ascending order, into Franklin Park (conglomerate), Brookline (conglomerate), Dorchester (slate), and Squantum (diamictite) members (Thompson et al., 2007, 2014). The age of the Squantum Member was widely considered Permio-Carboniferous (Billings, 1976; Dott, 1961), as suggested by putative fossil plants including ‘trunk casts’ (Burr & Burke, 1900) and ‘megasporangia’ (Pollard, 1965), which appear to have been water-escape structures and vesicles, respectively (Bailey & Newman, 1978; Billings, 1976; Cameron & Jeanne, 1976). An Ediacaran age of the Cambridge Argillite was eventually established by the discovery of partly pyritised cyanobacteria (Bavlinella sp. cf. B. favoleata; Lenk et al., 1982) and the discoid fossil Aspidella terranovica (Bailey & Bland, 2001; Clarke, 1923).

Two outcrops of Brighton Volcanics separated by 1 km of Brookline Member have returned nearly identical 206Pb/238U CA-TIMS ages of 584 ± 0.7 Ma and 585.4 ± 0.7 Ma. The lower collection of dated zircons is from brecciated andesite, and may have been a feeder for the amygdaloidal andesitic basalt of the upper dated rock (Thompson et al., 2014). This new dating contradicts traditional mapping of facies interdigation of fluvial Roxbury Conglomerate south of the Charles River Syncline with grey marine Cambridge Argillite to the north (Bailey, 1987; Bailey & Bland, 2001; Billings, 1976; Billings & Tierney, 1964; Socci & Smith, 1987). Zircons from volcanic ash in the Cambridge Argillite to the north in the Mystic River quarry have an array of ages ranging from 697 ± 5 Ma to as young as 568 ± 3 Ma by 207Pb/206Pb (Thompson & Bowring, 2000). The youngest tuff clasts and zircons from the Squantum Member and Dorchester Member are indistinguishable in age from 600 to 595 Ma Lynn-Mattapan Volcanic basement source terrains (Thompson et al., 2007, 2014; Thompson & Bowring, 2000). All these ages are Ediacaran (635–541 Ma; Xiao & Narbonne, 2020).

**Sedimentary environments**

The Squantum Member of the Roxbury Formation was first interpreted as a tillite because of associated dropped
pebbles, contorted lamination, varves, poorly sorted diamictites with oversized clasts, and faceted and striated pebbles (Lahee, 1914a, 1914b; Mansfield, 1906; Sayles, 1914). Lack of a striated pavement and rounded edges of clasts were emphasised by Dott (1961), who considered the Squantum Member a non-glacial, deep marine, conglomerate. An intermediate view of submarine till was proposed by Carto and Eyles (2011), Smith and Socci (1990) and Socci and Smith (1987, 1990). Additional evidence for glacial origin came from scanning electron microscope studies of quartz grain surface textures, including conchoidal fractures, semi-parallel steps, parallel striation, and intricate breakup blocks (Rehmer, 1981; Rehmer & Hepburn, 1974). However, Stuart et al. (1975) judged these features inherited from geologically older sources. Lindsay et al. (1970) measured the long axis orientation of grains in the Squantum Member, and found that they formed a clear girdle dipping steeper than bedding on a stereonet, unlike the random orientation of genuine tills. Robertson et al. (1971) disputed the approach of Lindsay et al. (1970) as an artefact of tight folding (Socci & Smith, 1987). Glacial interpretation has returned more recently, with discovery of dropped pebbles (McMenamin, 2018; McMenamin & Beuthin, 2008; Williams, 2008).

In contrast, the Roxbury Conglomerate has always been interpreted as alluvial fan and fluvial gravels (Billings, 1976). In addition, paleosols have been recognised in the Boston Bay Group for some time: a 1.5 m-thick, red, grus paleosol has been reported below the fluvial Roxbury Conglomerate, where it overlies Dedham Granite at World’s End Reservation (Bailey, 1987; Bailey & Bland, 2001), and near Hingham (Nellis & Hellier, 1976). Periglacial paleosols represented by ice wedges were recognised in the Brookline Member (Cameron, 1979a, 1979b; Cameron & Jeanne, 1976).

Cool paleoclimate during deposition of the Cambridge Argillite are indicated by the Ediacaran marine cyanobacterial microfossil Bavlinella, which is associated with glacial facies in Greenland, Norway, Svalbard and Utah (Lenk et al., 1982). Furthermore, a paleolatitude of 55° (Evans & Raub, 2011) and at least one magnetic reversal was found by paleomagnetic studies of the Squantum Member, whereas unconformable basement of Mattapan Volcanics formed at a paleolatitude of 38 ± 8° (Thompson et al., 2007). Finally, chemical alteration indices of the Squantum Member and
Cambridge Argillite by Passchier and Erukanure (2010) were interpreted as evidence of limited alpine glaciation, rather than extreme continental glaciation of the sort proposed by the Snowball Earth hypothesis (Hoffman & Li, 2009; Hoffman & Schrag, 2002). Diamictites of the Squantum Member of the Roxbury Conglomerate near Boston, Massachusetts, U.S.A., have also been considered evidence of Ediacaran glaciation (Carto & Eyles, 2011; Thompson & Bowring, 2000). Neither Gaskiers nor Squantum tillites were as extensive as Cryogenian Elatina and Sturt tillites that inspired the Snowball Earth hypothesis (Hoffman & Li, 2009; Hoffman & Schrag, 2002). Furthermore, the Squantum Member diamictites remain controversial, with three divergent interpretations: (1) terrestrial tillites (Bailey et al., 1976; Dodge, 1875; Lahee, 1914a, 1914b; Mansfield, 1906; Rehmer, 1981; Rehmer & Hepburn, 1974; Rehmer & Roy, 1976; Sayles, 1914; Wolfe, 1976), (2) glaciomarine tillites (Carto & Eyles, 2011; Passchier & Erukanure, 2010; Smith & Socci, 1990; Socci & Smith, 1987, 1990) and (3) deep marine non-glacial gravity-flow deposits (Bailey, 1987; Caldwell, 1981; Dott, 1961; Lindsay et al., 1970; Robertson et al., 1971; Stuart et al., 1975).

**New approaches using paleosols**

Stratigraphic sections were measured using tape and the level of a Brunton compass, with degree of development and of dilute acid reaction estimated in the field using the scales of Retallack (1997), and colour from a Munsell chart (Munsell Color, 1975) with additional tropical and gley pages (Figure 6). Samples of paleosols were collected in the field for preparation of sawn slabs and petrographic thin-sections, and also analysed for major oxides using X-ray fluorescence by ALS Chemex of Vancouver, British
Columbia, who also determined ferrous iron using the Pratt titration, and weight percent organic carbon using a LECO carbon analyser (Table 1). Bulk density was determined by weighing paraffin-coated clods some 2 cm² in size both in and out of water at 6°C (Table 1). Errors were calculated from 10 replicates for bulk density, and from 89 laboratory trials for other analyses. Thin-sections were point-counted using a Swift Automated stage and collator and an eyepiece micrometre to determine proportions of sand–silt–clay components of the paleosols (Figure 7; Table 2). Past trials have shown that the error associated with such 500-point counts is 2% (Murphy, 1983). Molar ratios were also calculated from bulk chemical analyses to give products over reactants of common soil-forming chemical processes (Retallack, 1997), such as salinisation, calcification, clayeyness, base loss and gleisation (Figure 7). A more detailed accounting of geochemical change following Brimhall et al. (1992) is mass transfer of elements in a soil at a given horizon ($s_j,w$), calculated from the bulk density of the soil ($q_w$ in g cm⁻³) and parent material ($q_p$ in g cm⁻³) and from the chemical concentration of the element in soils ($C_{j,w}$ in weight %) and parent material ($C_{j,p}$ in weight %).

Changes in volume of soil during weathering are called strain by Brimhall et al. (1992), and estimated from an immobile element in soil (such as Ti used here) compared with parent material ($e_i,w$ as a fraction). The relevant Equations 1 and 2 are the basis for calculating divergence from parent material composition (origin in various panels of Figure 8).

$$s_j,w = \left( \frac{p_w \cdot C_{j,w}}{p_p \cdot C_{j,p}} \right) \left[ e_{i,w} + 1 \right] - 1 \quad (1)$$

$$e_{i,w} = \left( \frac{p_p \cdot C_{j,p}}{p_w \cdot C_{j,w}} \right) - 1 \quad (2)$$

Also calculated were two weathering indices: (1) a simple molar ratio of Ba/Sr (Retallack, 1997), and (2) CIA of Nesbitt and Young (1982; Bahlburg & Dobrzeński, 2011). CIA ($I$ in mole per mole) is calculated from molar proportions ($m$) of alumina, lime, potash and soda according to Equation 3. The lime is non-carbonate lime, but no correction for carbonates was made because no carbonate was seen in thin-section (although two samples with >5% CaO were excluded; Table 1).

$$I = \frac{100 \cdot (m\text{Al}_2\text{O}_3)}{(m\text{Al}_2\text{O}_3 + m\text{CaO} + m\text{Na}_2\text{O} + m\text{K}_2\text{O})} \quad (3)$$

**New observations of the Boston Bay Group**

**Strongly tapering clastic dykes**

The most striking discovery was strongly tapering clastic dykes in the intertidal rock platform exposures of upper Squantum Member at Squantum (Figure 9b, c and 10), like those previously recorded from the Boston Bay Group by Cameron (1979a, 1979b) and Cameron and Jeanne (1976).
The cliff faces at a confusing angle to dip (22° east on strike 113° grid) and cleavage (51° west on 102° grid) show few unequivocal dykes (Figure 6), but the rock platform reveals bed cross-sections orthogonal to fracture cleavage, synsedimentary deformation, and regional folding of the Squantum Anticline (Figure 9b, c). In all cases the dykes are strongly tapering fissures down into purple, laminated siltstones, and show massive gravelly fill, similar to gravelly layers that extend laterally above them. Only three such strongly tapering dykes were seen below the conglomerates of the upper Brookline Member at Newton (Figure 9a; Cameron, 1979a, 1979b; Cameron & Jeanne, 1976). Near Squantum, strongly tapering clastic dykes have a restricted stratigraphic range (Figure 6) within the upper diamicite and overlying bedded siltstones. In some intervals at Squantum, clastic dykes are strongly overlapping (Figure 6), and range in size from millimetres (Figure 11f) to metres in depth (Figure 9b, c). Microscopic tapering clastic dykes have small grains at the narrow tip (Figure 11f). Above the stratigraphic level of common clastic dykes, clastic dykes are less common, but subhorizontal clastic dykes cut across several levels of bedding (Figure 6).

Lonestones
At Squantum and Newton, but not at other localities, are oversize clasts of volcanics and granite, which are either isolated within purple siltstone, or arrayed in trains of a single pebble thickness. These stones range in size from granules (Figure 11g) to pebbles (Figure 9f), and cobbles. As also noted by Cameron (1979a, 1979b), Cameron and Jeanne (1976), McMenamin (2018) and Williams (2008) oversize pebbles compress underlying silty laminae and are onlapped by overlying laminae without evident disconformity as in dropped pebbles and pebble trains. Pebble trains are laterally impersistent and there is no detectable disconformity to indicate that these were clast pavements. Small granules are commonly tipped at unusually high angles to bedding and have very fine-grained rims on their lower side (Figure 11g).

Faceted and striated clasts
Bedrock clasts with rounded as well as flattened and striated surfaces are uncommon in the Squantum Member, but not rare (Figure 9e, right), as apparent from many examples photographed by Sayles (1914). These hard subangular clasts contrast with common rounded clasts, like those of other members of the Roxbury Conglomerate. Some clasts of granite in the Squantum Member were weathered before transport (McMenamin, 2018) and are so soft that they fractured through the centre (Figure 9e, left), rather than around the margins like other nearby clasts of hard volcanics.

Graded beds
Laminae in the upper Brookline and Squantum members are graded by grainsize (Cameron, 1979a, 1979b; Cameron & Jeanne, 1976). The silty bases to these 1–3 mm thick laminae are sharp and erosional, and pass upward gradationally into clayey layers, which are commonly wrinkled by carbonaceous layers (Figure 11b, c, e). Not all laminae are clearly graded: some thin silty laminae have diffuse contacts both above and below (Figure 11f, g). No thick graded beds were seen in the upper red beds of the Squantum Member chosen for detailed study (Figure 6), but the overlying grey Cambridge Argillite has beds 10–20 cm thick and graded from sand up to shale.

Soft-sediment clasts
Large (>1 m) rafts of sandstone are common within diamicites of the Squantum Member at varied angles, commonly with near-perpendicular bedding. Evidence of soft-sediment transportation includes marginal sinuous
deformation yet bedding within the rest of these large clasts remains undisturbed (Figure 9d). The blocks may have been frozen but thawed marginally. The blocks are not cavity fills, because they lack bedding accommodation to walls or parallel to the regional bedding.

**Microfaults**

Small faults in bedded sequences of the upper Squantum Member viewed on the scale of thin-sections show brittle failure as normal faults yet were growth faults with thicker sediment on one side than the other (Figure 11e). The

Table 1. New chemical analyses of the Boston Bay Group, Massachusetts.

<table>
<thead>
<tr>
<th>Sample</th>
<th>mode level</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>P₂O₅</th>
<th>LOI</th>
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<th>Sr</th>
<th>g/cc</th>
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<tr>
<td>R2740A</td>
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<td>61.83</td>
<td>0.81</td>
<td>18.33</td>
<td>5.58</td>
<td>1.22</td>
<td>0.11</td>
<td>0.25</td>
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<td>0.32</td>
<td>1.67</td>
<td>0.23</td>
<td>2.50</td>
<td>0.18</td>
<td>1.05</td>
<td>4.58</td>
</tr>
<tr>
<td>R2740A</td>
<td>0.6</td>
<td>61.83</td>
<td>0.81</td>
<td>18.33</td>
<td>5.58</td>
<td>1.22</td>
<td>0.11</td>
<td>0.25</td>
<td>0.76</td>
<td>0.32</td>
<td>1.67</td>
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<td>1.05</td>
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<tr>
<td>R2741A</td>
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<td>2.59</td>
<td>1.22</td>
<td>0.13</td>
<td>1.56</td>
<td>0.41</td>
<td>1.34</td>
<td>1.85</td>
<td>0.09</td>
<td>2.09</td>
<td>0.98</td>
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<td>56.49</td>
</tr>
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<td>9.83</td>
<td>2.59</td>
<td>1.22</td>
<td>0.13</td>
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<td>0.98</td>
<td>330</td>
<td>56.49</td>
</tr>
<tr>
<td>R2738A</td>
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<td>0.13</td>
<td>1.56</td>
<td>0.41</td>
<td>1.34</td>
<td>1.85</td>
<td>0.09</td>
<td>2.09</td>
<td>0.98</td>
<td>330</td>
<td>56.49</td>
</tr>
</tbody>
</table>

Note. Major oxides, loss on ignition and totals are in weight %, trace elements in ppm, and bulk density in g cm⁻³. Analyses by XRF, with ferrous iron by Pratt titration, and bulk density by clod method using paraffin. Errors are from 10 replicate analyses of the standard, CANMET SDMS2 (British Columbia granodioritic sand). Samples R3920 (vein in granite) and R3732 (clast in till) were probably caliche nodules and were not plotted. Abbreviation n.d. is not determined.

Figure 7. Grainsize and mineral (from point-counting; Table 2) and chemical composition (from Walkley-Black titration and X-ray fluorescence; Table 1) of Ediacaran paleosols in the upper Squantum Member. Molar weathering ratios are designed to reveal degree of common soil-forming reactions, such as base loss (Retallack, 1997).
faulted beds, however, show sinuous soft-sediment deformation. These microfaults indicate brittle failure at the time of deposition, perhaps of frozen sediment, but soft-sediment deformation at other times.

Filamentous bed disruption

The orientation of rock samples collected for this work was marked in the field so that thin-sections for this study could be cut vertical to bedding, to reveal the degree of post depositional alteration of sedimentary structures. Some grey beds of the upper Squantum Member are un laminated and have near total subvertical filamentous disruption of bedding (Figure 11a and 12e). In purple silty beds of the upper Squantum Member tubular structures are 0.5–1 mm diameter and sparsely distributed across bedding (Figure 11b, c). In each case, the filamentous structures are filled with materials of similar texture and colour to nearby rock layers. Unlike fracture cleavage, which is straight, continuous, and at an angle to bedding, these filamentous disruptions are crooked, sinuous, and intermittent, but generally vertical to bedding.

Paucity of clay

The lower Cambridge Argillite and Roxbury Conglomerate had a large proportion of loess, as revealed by point-counted 34–60 volume percent angular silt grains (Figure 7; Table 2). In contrast, gravelly rocks were nearly equal parts gravel, sand, silt and clay (Table 2). Clay is more common in the uppermost Squantum Member (upper Figure 7) and Cambridge Argillite, where suspension settling may have been a sedimentary mechanism. Even in dark grey argillites at Hewitts Cove and Mystic River quarry, silt is dominant (Figure 11h–j). True shales have only been reported from the upper Cambridge Argillite, which has 37–69 wt% clay minerals (sericite, kaolinite and chlorite).
Figure 9. Field photos of (a) upper Brookline Member at Newton, (j) Dorchester Member at Brighton, (b–i) Squantum Member at Chapel Rocks, and (g, k–m) Cambridge Argillite at Hewitts Cove; (a–c) vertical cross-sections of ice wedges filled with tillite into red siltstone of Chapel pedotype; (d) large block of sandstone transported with mild deformation as if frozen within tillite; (e) hand specimen (R2746) of tillite with cross-section of weathered granite (left) and bev- eled, vertically striated rhyolite clast (to right); (f) string of dropped pebbles in red siltstone; (g) cuspate and linsel bedding of intertidal facies; (h) paleosol of Standish pedotype (recessive red siltstone below tillite) at arrow; (i) thin paleosols of Shepard pedotype; (j–m) bedding lane exposures of vendobiont fossils, Aspidella terranovica, at arrows in Panel (j). Panel (l) shows setuufs (shadow dunes) downwind of Aspidella. Panel (m) also shows four individuals of the marine lichen Verrucaria mucosa, common in the intertidal zone of Boston Bay (Webber, 1975). Panels (k) and (l), are field photos but specimen of (j) is Condon collection F116521A and (m) is Condon Collection F114497.
calculated as modes from chemical analyses (Billings & Tierney, 1964; Rahm, 1962).

**Anhydrite sand crystals**

Red siltstones of the upper Squantum Member at Squantum have common, small (1–2 mm long) rhomboidal crystals scattered at all angles to bedding (Figure 11e), and in one case arrayed on bedding planes as lags (Figure 11c). These crystals are distinctly zoned with outer sharp angles filled with chalcedony, and an inner core of anhydrite (Figure 12f). The central anhydrite core has inclusions of matrix indicating that it formed by replacive incorporation of matrix, rather than displacement of matrix by an optically pure crystal.

**Red colour**

The most striking difference between Cambridge Argillite and Roxbury Conglomerate is colour: purple-red Roxbury Conglomerate contrasts with dark bluish grey Cambridge Argillite. Both diamictites and siltstones of the Squantum Member are purple, but some siltstone beds and thin gravel beds and clastic dykes in the upper Squantum Member are greenish grey. The Dorchester Member of the Roxbury Conglomerate at Brighton has a mix of purple-red siltstone flagstones, and lenticular to wavy bedded bluish grey shale. Purple and red colours were found in freshly excavated deep drainage tunnels (Billings & Tierney, 1964; Rahm, 1962), so are not due to modern weathering. Nor are the colours due to sedimentary alternation of different source materials, because apart from iron redox state, red and green beds are chemically and petrographically similar (Figure 7).

**Chemical differentiation**

Overall chemical composition of the upper Squantum Member at Squantum is varied in degree of weathering, as indicated by big differences in alumina enrichment at the expense of bases, and barium enrichment at the expense of strontium (Figure 7). Also quite varied is organic carbon concentration: highest near the tops of beds and sparse below (Figure 7). These differences are very striking when a stable constituent is used to calculate strain and mass transfer of chemical constituents (Equation 1 and 2). Such analysis (Figure 8) reveals two very distinct groups of beds: (1) poorly weathered beds showing volume dilation and gain of elements, and (2) moderately weathered beds showing loss of volume and most elements. The first group is compatible with cryoturbation and addition of ash and loess, but the second group is best explained by chemical weathering. These distinct kinds of siltstones were also found from calculation of CIA (Equation 3) for samples additional to those analysed by Passchier and Erukanure (2010), who did not find the low values encountered on some horizons analysed here.

An additional molar ratio of Ba/Sr calculated here, also shows predominantly high values, but low values on the same stratigraphic horizons as the low CIA (Figure 4 and 5). There is a risk that Ba/Sr ratios in marine rocks may be compromised by additions of biogenic barium (Reitz et al., 2004), but Ba/Al molar ratios for data compiled here were 0.0019–0.0037, with two outliers of 0.0045 at 1879 m and 0.0079 at 1872 m (Figure 4). All are within the range of this ratio for modern detrital (non-biogenic) Ba/Al of marine rocks (Reitz et al., 2004).

**Aspidella terranovica**

New collections of these enigmatic fossils from the lower Cambridge Argillite at the well-known localities near Hewitts Cove (Figure 9k, l) were supplemented with new discoveries of *Aspidella* in the Dorchester Member of the Roxbury Conglomerate at Brighton (Figure 9j) and Hingham (Condon Collection, University of Oregon, specimen F116522). In all cases, *Aspidella* was found in grey
Figure 11. Thin-section photoscans of (a–g) Squantum Member south of Chapel Rocks and (h–j) Cambridge Argillite at Hewitts Cove. (a) Surface of Standish silty clay loam paleosol, showing overlying lamination and large grains filling a vertical filament at arrow; (b, c) varved siltstone with crinkly lamination above graded siltstone with dark strata-transgressive tubular features (arrowed); (d) strata-transgressive, irregularly tubular structure from clast in tillite; (e) sand-filled vertical structure (left edge centre) and microfault (at arrow) in varved siltstone; (f) miniature ice wedge, filled with granules from layer above and deforming in varved siltstone; (g) oversized clasts in varved siltstone interpreted as dropped pebbles, because deforming underlying bed and onlapped (at arrow); (h) dark-filled tubular features (left leaning arrows), apparently connected, in varved shale; (i) transverse sections of dark organic matter of Aspidella terranovica (right leaning arrows); (j) horizontal section of Aspidella terranovica, with central sediment fill. Specimens in collections of University of Oregon: (a) R2736, (b) R2735B, (c) R2741, (d) R2745, (e) R2741A, (f) R2730, (g) R2743, (h) F114498 and (i, j) F114496, transverse and horizontal sections, respectively).
clayey rocks of Hewitts Cove exposed only at low tide have large (15 mm diameter) Aspidella with wide central depressions (Figure 9l, m), whereas eastern Hewitts Cove coastal cliffs have small (3 mm diameter) Aspidella with very small central depressions and high relief (Figure 9k). Brighton has medium size (5 mm diameter) Aspidella with low relief (Figure 9j), and Hingham has small (3 mm diameter) Aspidella with low relief. This variation is comparable with that of paratypic material of Aspidella terranovica from Newfoundland (Gehling et al., 2000; Retallack, 2014b).

A wider vision of periglacial facies

Taken together, the above observations support terrestrial tillite and fluvioglacial plain interpretations for the Squantum and upper Brookline members, as developed at length by Cameron (1979a, 1979b), Cameron and Jeanne (1976), Laihe (1914a, 1914b), and Sayles (1914). These authors interpreted tapering clastic dykes as ice wedges, faceted and grooved pebbles as glacially transported, lomestones as pebbles dropped from melting ice, and graded laminae as varves from seasonal snow melt. These authors, as well as Bailey and Bland (1981), Billings (1976) and Billings and Tierney (1964) regard the Squantum and Brookline members as coastal plain deposits formed during successive glacioeustatic drawdowns. The Boston Bay Group was deposited on the basal paleosol in Dedham Granite, and includes several levels of terrestrial Brighton Volcanics, with unpillowed lavas and red interflow paleosols (Figure 4). Marine transgression of interfingering shallow marine Cambridge Argillite, culminated in deeper marine upper Cambridge Argillite (Carto & Eyles, 2011; Passchier & Erkuahne, 2010; Smith & Socci, 1990; Socci & Smith, 1987, 1990).

The strongly tapering, clastic dykes (Figure 9a–c) are not tension gashes, nor landslide, nor desiccation cracks, because their fill is complex, including suspended, vertically oriented, soft-sediment clasts of wall, which appear to have been transported while partly frozen. Nor is there any streaming or vertical orientation of the fill as would be expected if these were water release structures or clastic dykes created by upward injection. These distinctive clastic dykes match modern ground ice and Pleistocene ice wedges in their marked upward flare and regular spacing (Figure 10; Table 3: data from Black, 1976a, 1976b; Davis, 2001; Dylik, 1994; French, 1996; Gell, 1978; Kokelj et al., 2007; Leffingwell, 1915; Owen et al., 1998; Raffi & Stenni, 2011; Wayne, 1991), including microscopic forms of Figure 11f (like those illustrated by Van Vliet-Lanoe, 2010). Furthermore, the Boston fossil ice wedges fulfill the stringent criteria outlined by Black (1976a): (1) other evidence of glacial conditions (low CIA and Ba/Sr of Figure 4 and 5); (2) other evidence of waterlogging (high FeO/Fe2O3 of Figure 5, iron depletion of Figure 7); (3) proportional spacing of polygons (Figure 10); (4) evidence of rain on soil with little or no summer snow cover (significant within-paleosol base depletion of Figures 7 and 8); (5) edges upturned by pressure (Figure 9a–c); and (6) slumping included blocks (Figure 9a). These lines of evidence contradict the view of Dott (1961), Lindsay et al. (1970), Robertson et al. (1971), and Stuart et al. (1975) that these poorly sorted diamictites are non-glacial or submarine landslide deposits. Ice wedges are also evidence against entirely submarine glacial deposition proposed by Carto and Eyles (2011), Passchier and Erkuahne (2010), Smith and Socci...
(1990), and Socci and Smith (1987, 1990), although varved shales and dropstones (McMenamin, 2018; McMenamin & Beuthin, 2008; Williams, 2008) are evidence of at least episodic submergence under a lake or sea for some parts of the sequence, and deeper marine facies may be present in the upper Cambridge Argillite.

Other observations support terrestrial periglacial paleoenvironments for the Squantum and upper Brookline members (Cameron, 1979a, 1979b; Cameron & Jeanne, 1976). Large mildly deformed soft-sediment clasts (Figure 9d) and small brittle growth faults (Figure 11e) may be interpreted as periglacial features, with clasts transported short distances frozen in till, and microfaulting of frozen varved shales. Unograded silty laminae within beds with vertically oriented granules rimmed at the bottom by very fine silt (Figure 11g), are similar to freeze–thaw banding, frost heave, and stone jackding of periglacial soils (Van Vliet-Lanoe, 1998, Vliet-Lanoë, 2010). Paucity of clay in purple-red siltstones with anhydrite sand crystals are complementary lines of evidence that this facies was exposed to weathering, because known Ediacaran paleosols are ferruginised, silty with loess and not so deeply weathered to clay as modern soils (Retallack, 2009, 2011, 2012, 2013b).

In thin-sections cut vertical to bedding, these beds show layered organic matter is strongly deformed, perhaps by needle ice as documented by Van Vliet-Lanoe (1998), in red periglacial siltsstones (Figure 12a, e), but undeformed in non-glacial, grey siltsstones (Figure 12b, d). Finally, the heterolithic, grey, shale–siltstone facies at Brighton and Hewitts Cove includes the problematic fossil Aspidella terranovica. Sediments and fossils of the Boston Bay Group are like shallow marine shales of Newfoundland with this fossil studied by Gehling et al. (2000, p. 432) and Retallack (2014b).

### Table 3. Linear regression statistics for Ediacaran, Pleistocene, and modern ice wedges.

<table>
<thead>
<tr>
<th>Category</th>
<th>Independent variable (x in m)</th>
<th>Dependent variable (y in m)</th>
<th>Slope (y/x)</th>
<th>Intercept (y in m)</th>
<th>Coefficient of variation (r²)</th>
<th>Standard error (±y in m)</th>
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<td>Ediacaran</td>
<td>Width at surface</td>
<td>Depth</td>
<td>+0.7755</td>
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Data from Black (1976a, 1976b); Davis (2001); Dylik (1994); French (1996); Gell (1978); Kokel et al. (2007); Leffingwell (1915); Owen et al. (1998); Raffi and Stenni (2011); Wayne (1991).

### Recognition and classification of paleosols

#### Alteration after burial

Paleosols are a form of alteration before burial and can be distinguished from deep burial and metamorphic alteration. The Boston Bay Group has sericite and slaty cleavage of burial alteration in the greenish facies of regional metamorphism (Billings & Tierney, 1964). In addition, sand crystals of orthohombic anhydrite are interpreted as a deep burial recrystallisation of monoclinic gypsum crystals by dehydration at temperatures in excess of 184–251°C (Jacques et al., 2009). The original form of the gypsum is pseudomorphed by chalcedony flanges to the anhydrite (Figure 12f), which shows that they were from pedogenic gypsum sand crystals, forming horizons of varying density and incorporating matrix like sand crystals of desert soils (Dan et al., 1973; Dan & Yaalon, 1982; Lebron et al., 2009). Other siltstone beds in the upper Squantum Member with abrupt top and gradationally diminishing expression below are green-grey beds within the red beds (Figure 9h).

In thin-sections cut vertical to bedding, these beds show filamentous structure vertical to bedding (Figure 11a), like the microtextures of other Ediacaran and Cambrian microbial earth paleosols (Retallack, 2008, 2009, 2011, 2012, 2013a, 2013b, 2016a, 2016b). Layered organic matter is strongly deformed, perhaps by needle ice as documented by Van Vliet-Lanoe (1998), in red periglacial siltsstones (Figure 12a, e), but undeformed in non-glacial, grey siltsstones (Figure 12b, d).
organic matter (Retallack, 1997). By this explanation, drab layers reflect alteration of soil organic matter very soon after burial.

Field classification of paleosols
At least four distinct kinds, or pedotypes, of paleosols can be recognised and named for nearby geographic features (Table 4 and 5). The ice-wedged horizons are the most conspicuous of these (Chapel pedotype of Figure 7). Above the uppermost lenticular green diamicton of the Squantum Member is a massive paleosol with fine vertical filamentous structures in its surface (Standish pedotype of Figure 7). Other thin horizons of green-mottled siltstone within the purple siltstone show less pedogenic development, with fine mottles and subsurface anhydrite crystals (Nickerson pedotype) or with vertical filaments and mottles (Shepard pedotype). These very weakly developed paleosols were not sampled for geochemical or petrographic analysis.

Paleosol interpretations
Each pedotype recognised in the Squantum and upper Brookline members of the Roxbury Conglomerate represents a distinct ancient terrestrial environment. Comparable modern soils can be located by identification of the paleosols within soil classifications (Food & Agriculture Organization, 1974; Isbell, 1996; Soil Survey Staff, 2014). Modern comparisons also are used to infer Ediacaran soil-forming factors in the following paragraphs.

Paleoclimate
Paleoclimate can be inferred for ice wedges in the Boston Bay Group from what is known about modern ice wedges, which form under mean annual air temperatures of less than \(-8 \text{ to } -4 ^\circ C\), mean coldest month temperature \(-25 \text{ to } -10 ^\circ C\), mean warmest month temperature of \(10 \text{ to } 20 ^\circ C\), mean annual precipitation \(50 \text{ to } 200 \text{ mm}\), freeze index \(2600 \text{ to } >7000 \text{ C-days a}^{-1}\), thaw index \(100 \text{ to } 1000 \text{ C-days a}^{-1}\), and rapid cooling early winter with thin snow (Williams, 1986). Under colder conditions, sand wedge polygons are formed (Williams et al., 2008), and under warmer conditions a wide array of periglacial structures, including involutions and thufur mounds (Krull, 1999; Retallack, 1999, 2011, 2012, 2013b). Ice wedges are best known from Greenland and Arctic Canada, which are maritime glacial climates, as opposed to the continental glacial climate of modern Antarctica with sand wedges (Black, 1976a, 1976b; Washburn, 1980). Fossil ice wedges and sand wedges are also known from Cryogenian and earlier Proterozoic glacial deposits (Kumpulainen, 2011; Retallack et al., 2015; Williams, 1986; Williams et al., 2008; Young & Long, 1976). Former temperature and precipitation can also be inferred from chemical composition of the paleosols. The paleohyetometer of Sheldon et al. (2002) uses CIA without potash \((R = 1000\text{Al}_2\text{O}_3/\text{mAl}_2\text{O}_3 + \text{mCaO + mNa}_2\text{O})\), in moles, which increases with mean annual precipitation \(P\) in millimetres) in modern soils \((R^2 = 0.72; \ SE = \pm 182 \text{ mm})\), as follows:

\[
P = 221e^{0.0197R}
\]

This formulation is based on the hydrolysis equation of weathering, which enriches alumina at the expense of lime,
magnesia, potash and soda. Magnesia is ignored because it is not significant for most sedimentary rocks, and potash is excluded because it can be enriched during deep burial alteration of clayey sediments (Maynard, 1992; Novoselov & de Souza Filho, 2015). This potential diagenetic compromise is not considered in the otherwise similar CIA (Nesbitt & Young, 1982).

A useful paleothermometer for paleosols predating the evolution of modern vegetation is based on modern soils under tundra vegetation of Iceland (Öskarsson et al., 2012). This linear regression between mean annual temperature (T in °C) and chemical index of weathering \( I = 100(m\text{Al}_2\text{O}_3/m\text{SiO}_2 + m\text{CaO} + m\text{Na}_2\text{O}) \) in molar proportions is given in Equation 5:

\[
    T = 0.21I - 8.93 \quad (5)
\]

where \( R^2 = 0.81 \), and \( SE = \pm 0.4°C \).

Nickerson and Shepard pedotypes are inappropriate for these proxies because they are too poorly developed to reflect paleoclimate, but these equations yield mean annual temperature of \(-8.9 ± 0.4°C\) for the Chapel pedotype and \(9.4 ± 0.4°C\) for the Standish pedotype, and mean annual precipitation of \(662 ± 182\) mm for the Chapel pedotype and \(1255 ± 182\) mm for the Standish pedotype. A colder and drier paleoclimate for Chapel paleosols is also inferred from tillite, anhydrite sand crystals and ice wedges in Chapel paleosols. The absence of these features in the Standish paleosols is also inferred from tillite, anhydrite sand crystals and ice wedges in Chapel paleosols.
differences do not reflect differences in source compositions, which Thompson and Bowring (2000) showed to be uniformly andesitic, from Zr/TiO$_2$ molar ratios of 0.017 ± 0.004 (15 analyses). This molar ratio for 44 analyses obtained for this study and by Passchier and Erkenare (2010) is 0.029 ± 0.012, with a maximum value of 0.047. These values are very uniform, so not compromised by grainsize effects, and also are short of granite–rhyodacite values (Winchester & Floyd, 1977). Glacial sediments have a CIA of 50–55 (Passchier & Krisske, 2008), whereas a CIA > 70 is typical of average shale derived from unglaciated land (Nesbitt & Young, 1982). Glacial conditions are reflected by CIA less than 60 and barium/strontium molar ratios of less than 2 within the upper Brookline and uppermost Dorchester to Squantum members. Within the upper Dorchester and Squantum members are four glacial levels, separated by temperate levels and reflecting advance and retreat of ice during this glacial episode (Figures 2 and 3). Temperate conditions were more common overall during deposition of the Boston Bay Group, as indicated by Passchier and Erkenare (2010), who did not sample the narrow glacial stratigraphic levels.

**Life on land and in the sea**

Traces of organic carbon and the appearance of organic matter in thin-section are very different for the glacial Chapel and Nickerson pedotypes, than for the temperate Standish and Shepard pedotypes. Laminated organic matter comparable with thin microbial mats was seen in periglacial Chapel and Nickerson paleosols (Figure 11b, c, e and 12a, b), but temperate Standish and Shepard paleosols had a vertical filamentous fabric (Figure 11a) more like those of Ediacaran and Cambrian microbial earth ecosystems (Callow & Brasier, 2009; Retallack, 2008, 2011, 2012, 2013b). Both laminated and tubular filamentous structures are comparable with those of cyanobacteria such as *Microcoleus*, which can build both laminated mats and vertical ropes in soils and playa lakes (Garcia-Pichel & Wojciechowski, 2009). Microfossils of *Bavlinella* from the Cambridge Argillite are evidence that cyanobacteria were present in marine to intertidal facies of the Boston Bay Group (Lenk et al., 1982). Microbial earths differ from microbial mats in being thoroughly admixed with their substrates (Belnap & Lange, 2003).

Sparse organic-lined tubes up to 100 μm in diameter at the base of varves in Nickerson paleosols (Figure 9b at arrows) are cut across laminations. Small animals such as worms could explain such structures and are inferred from other Ediacaran trace fossils (Fedonkin et al., 2008), but penetrative burrows are rare and controversial before the Cambrian (Brasier et al., 2013; Dzik, 2005; Rogov et al., 2012). These tiny tubes are more likely the work of the slug phase of migrating cellular slime moulds (Eumycetozoa, Orders Dictyostelia and Protostelia; Bonner, 2009; Brown et al., 2009), or large examples of cyanobacterial or other microbial ropy structures (Garcia-Pichel & Wojciechowski, 2009).

Intertidal to shallow marine, heterolithic, shaley facies yielded abundant fossils identical with recently collected Newfoundland material *Aspidella terranovica* (Retallack, 2014b), a problematic vendobiont considered a cnidarian by Gehling et al. (2000). Thin-sections of these particular examples (Figures 9h–j and 1c, d), however, do not support interpretation as animals. *Aspidella* has a complex microstructure of subhorizontal laminar elements as well as vertical filaments, which ramify down in the matrix, and also protrude for a limited distance into overlying layers. This microscopic structure is more like microbial consortia, such as ministromatolites (Hofmann & Jackson, 1987) or crustose lichens (Brodo et al., 2001). The morphologically similar discoidal marine lichen *Verrucaria mucosa* is common in Boston Bay today (Webber, 1975), and encrusted slabs with *Aspidella* found in the current intertidal zone at Hewitts Cove (Figure 9m).

One thin-section (Figure 11d) shows an *Aspidella*-like carbonaceous body above a diffuse zone of downward deformation of a thin varve, as if the discoid were atop a rooting structure. This is a roughly tubular structure viewed in the slab, not a desiccation or synaeresis crack. It cannot be a water or gas escape structure like those illustrated by Frey et al. (2009) because deformation is downwards not upwards, and it is capped with organic matter where the blowout orifice would have been. It is also unlikely to be a downward penetrating burrow (as figured by Hantzsche, 1975), because it tapers strongly. This specimen is within a clast of grey siltstone from tillite of the Squantum Member, presumably redeposited from comparable beds of the underlying Dorchester Member. Similar deformation below and threadlike extensions from *Aspidella* have been illustrated from the Ediacaran Fermeuse Formation of Newfoundland by Gehling et al. (2000, figures 10 and 14) and Retallack (2014b).

**Paleotopography**

Red paleosols require elevation above water-table (Retallack, 1997), but facies and thickness changes within the Boston Bay Group are evidence of much more relief in both the depositional basin and source area. Near and east of Hingham, Roxbury Conglomerate filled paleorelief of 260–640 m on granitic basement (Billings et al., 1939). This Dedham Granite is older (612–606 Ma) than overlying Mattapan Volcanics (602–596 Ma; Thompson & Bowring, 2000; Thompson et al., 2007), which would have been deeply dissected calcalkaline volcanic edifices to expose the underlying granite. Normal faulting of 1 km during deposition of the Roxbury Conglomerate has been proposed by Thompson et al. (2014). This rift basin is only 30 km wide from Hingham to Newton, and yet it contains at least 5.7 km of sediment (Billings, 1976), with only one likely volcanic ash (Thompson & Bowring, 2000) and no reported seismites. For these reasons, it may have had
deep glacial valleys incised into granite and dormant calcalkaline volcanics.

Rapid facies changes from fluvial conglomerates and glacial tills of the Roxbury Conglomerate to marine shales and siltstones of the Cambridge Argillite are evidence of a narrow coastal plain with an abrupt transition from high energy fluvial to low energy marine palaeoenvironments (Figure 13a). Lateral mapping of Squantum Member strike ridges has diamicrite in some places and fluvial conglomerates in others (Billings, 1976), which may represent terminal moraines locally breached by braided streams (Figure 13a). Ice wedges and other paleosols (Figures 6–10) are evidence of a well-drained coastal plain, on which tillites would have been arcuate wreeaths of terminal moraines, similar to the glaciated west coast of the South Island of New Zealand (Tovar et al., 2008).

This study confirms that the transition between red Roxbury Conglomerate and grey Cambridge Argillite reflects terrestrial facies overlain by deposits of a lake or sea, as envisaged by Billings and Tierney (1964). Sea-level and water-tables are a redox boundary now and for all known Phanerozoic paleosols (Retallack, 2008; Retallack & Huang, 2011), and also for early Ediacaran landscapes (Retallack, 2011, 2012, 2013a, 2013b). There is thus no evidence from the Boston Bay Group to support the idea, proposed on the basis of Ediacaran rocks from Newfoundland, that biological productivity and organic matter content of the Ediacaran ocean was low enough to permit marine clastic red beds at that time (Canfield et al., 2007). This is not to say that there might not have been Ediacaran equivalents of Phanerozoic red limestones of pelagic shells (ammonitico rosso; Mamet & Preat, 2006), abyssal red clays (Gleason et al., 2002) or oolitic iron ores (Brett et al., 1998).

Unlike the red Squantum Member, these would be chemical sediments with few siliciclastic grains, and perhaps skeletonised fossils, such as Cloudina and thecamoebians, known as far back as the Ediacaran (MacDonald et al., 2010).

Within the coastal plain during deposition of the upper Squantum Member, Chapel and Standish paleosols formed on well-drained terraces, judging from low ferrous/ferric iron ratios and their purple-red colour (Figure 9h). Green-grey colour is more prominent and organic carbon more abundant in the Standish than Chapel pedotype (Figure 9h). This difference is more likely to reflect the process of burial gleisation of organic matter (Retallack, 1997) than original groundwater gleisation, because the lower part of the Standish paleosol remained brown and oxidised, like other Cambrian and Ediacaran paleosols (Retallack, 2008, 2011). By this model, grey-green colour reflects higher biological productivity of the temperate Standish paleosol, compared with organic lean and frigid Chapel pedotype.

**Parent material**

Sedimentary parent materials of Boston Bay Group paleosols are an initial control on their development, and reflect a calcalkaline volcanic source terrain, with minor tuffaceous and granitic material (Bailey et al., 1976; Thompson & Bowring, 2000). Clasts of micritic carbonate were also found in tillite of the upper Squantum Member at both Squantum and Hewitts Cove (Figure 3). The only known source is calcareous fluvial conglomerates of the Brookline Member of the Roxbury Formation near Hingham (Nellis & Hellier, 1976). Such calcareous parent materials were only preserved in Chapel pedotype paleosols, and not found in Standish paleosols. Decalcification and other base depletion of this parent material in temperate pedotypes (Standish and Shepard) are more marked than in the periglacial pedotypes (Chapel and Nickerson).

**Time period for formation**

Ice wedges comparable in size (0.5–1.5 m. wide at top) with those of the upper Brookline and Squantum members took about 3000 years to form by snowmelt water seeping into permafrost and refreezing as veins during the Holocene in the Canadian Arctic (Fortier & Allard, 2004; Kokelj et al., 2007) and Siberia (Streletska et al., 2011). All these modern ice wedges are in the taiga and tundra zones and covered by peats unknown in the Ediacaran, so a better modern analogue may be ice wedges in plant-less soils of Antarctica (Raffi & Stenni, 2011) or Alaska (Black, 1976a, 1976b). Rates of ice and sand wedge growth in Antarctica ranged from 0.79 mm a$^{-1}$ in the 1960s (Black, 1973) to 0.04 mm a$^{-1}$ in the 1970s (Black, 1982). Such wide disparity is problematic for understanding rates and extrapolating age, although comparable with growth of such features in the northern hemisphere (Bockheim, 1995). Nevertheless, 0.5–1.5 m-wide ice wedges like those of the Squantum Member require 63–190 years by the fast rate and 3750–1250 years by the slow rate. This latter slow rate is comparable with Canadian and Siberian ages for large ice wedges (Fortier & Allard, 2004; Kokelj et al., 2007; Streletska et al., 2011).

Other paleosols of the upper Squantum Member are only weakly developed (Standish pedotype) or very weakly developed (Nickerson and Shepard pedotypes) in the development scale of Retallack (1997). Among modern periglacial and temperate soils, such very weakly developed paleosols represent only centuries to decades of soil formation (Birkeland, 1999).

**Conclusions**

Both periglacial and temperate paleosols are now recognised within the Boston Bay Group (Table 2), as well as sediments with a glacial and temperate chemical composition (Figure 2 and 3). The single most compelling evidence of periglacial paleosols are ice wedges recognised here (Figure 9a–c and 11f) and previously (Cameron, 1979a, 1979b; Cameron & Jeanne, 1976), but paleosols were also verified by anhydrite sand crystals (Figure 12f), oxidised...
horizons (Figure 7) and bedding disruption (Figure 12e), and within-bed chemical differentiation (Figure 8). At least two glacial episodes are recognised within the sequence: one episode within the upper Brookline Member of the Roxbury Conglomerate, and a second episode of at least three successive advances and retreats in the upper Dorchester and Squantum members of the Roxbury Conglomerate. Paleolatitude of about 55° has been inferred from paleomagnetism of the Squantum Member (Evans & Raub, 2011), although the possibility of Paleozoic overprint is raised by Thompson et al. (2007), who determined a paleolatitude of 38° for underlying Lynn-Mattapan volcanics. With glaciers down almost to sea-level at such mid to high paleolatitudes, the Roxbury tillites, ice wedges and limited chemical weathering (Figures 2 and 3) are evidence of a world as cool as the Holocene, in which glaciers are within 100 m of sea-level at more than 57° north and more than 38° south (NSIDC, 2012; Pfeffer et al., 2014). The Ediacaran was not a greenhouse world like that of the Mesoproterozoic (Sheldon, 2006) or Ordovician (Berner, 2006), nor was the Ediacaran as severely cold as the Cryogenian, when glaciers extended into tropical paleolatitudes (Retallack, et al., 2015).

The marine fossil Aspidella is evidence of interglacial and postglacial sea-level rise, and also of late Ediacaran age of the Boston Bay Group (Gehling et al., 2000). Discovery during this research of new localities for Aspidella at Brighton and Hingham, additional to those already known around Hewitts Cove (Bailey & Bland, 2001; Clarke, 1923), indicates the potential for further discoveries of Ediacaran fossils in the Boston Bay Group. Worth re-examining from this perspective are localities and specimens of putative plant fossils described by Burr and Burke (1900) and Pollard (1965) from the Roxbury Conglomerate and Mattapan Volcanics but dismissed as pseudofossils by Bailey and Newman (1978), and Cameron (1979a). Glacial-eustatic sea-level variation established by both marine-intertidal fossils and paleosols can be helpful in correlating Ediacaran sequences internationally.

Local radiometric dating (Thompson et al., 2014) is evidence that the Squantum and Dorchester periglacial paleosol levels correlate with the Gaskiers glacialiation in Newfoundland. This is the first of four likely Ediacaran glacial advances (Table 6): (1) Gaskiers at 580 Ma, (2) Fauquier at 571 Ma, (3) Bou-Azzer at 566 Ma and (4) Hankalchough at 551 Ma. Tillites and diamictites have proven so controversial that additional research on paleosols and sequence stratigraphy will be needed for an effective international correlation of Ediacaran glacial advances.

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References


