Secular changes in sedimentation systems and sequence stratigraphy

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ABSTRACT

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The ephemeral nature of most sedimentation processes and the fragmentary character of the sedimentary record are of first-order importance. Despite a basic uniformity of external controls on sedimentation resulting in markedly similar lithologies, facies, facies associations and depositional elements within the rock record across time, there are a number of secular changes, particularly in rates and intensities of processes that resulted in contrasts between preserved Precambrian and Phanerozoic successions. Secular change encompassed (1) variations in mantle heat, rates of plate drift and of continental crustal growth, the gravitational effects of the Moon, and in rates of weathering, erosion, transport, deposition and diagenesis; (2) a decreasing planetary rotation rate over time; (3) no vegetation in the Precambrian, but prolific microbial mats, with the opposite pertaining to the Phanerozoic; (4) the long-term evolution of the hydrosphere-atmosphere-biosphere system. A relatively abrupt and sharp turning point was reached in the Neoarchaean, with spikes in mantle plume flux and tectonothermal activity and possibly concomitant onset of the supercontinent cycle. Substantial and irreversible change occurred subsequently in the Palaeoproterozoic, whereby the dramatic change from reducing to oxidising volcanic gases ushered in change to an oxic environment, to be followed at ca. 2.4-2.3 Ga by the "Great Oxidation Event" (GOE); rise in atmospheric oxygen was accompanied by expansion of oxygenic photosynthesis in the cyanobacteria. A possible global tectono-thermal "slowdown" from ca. 2.45–2.2 Ga may have separated a preceding plate regime which interacted with a higher energy mantle from a ca. 2.2–2.0 Ga Phanerozoic-style plate tectonic regime; the "slowdown" period also encompassed the first known global-scale glaciation and overlapped with the GOE. While large palaeodeserts emerged from ca. 2.0 – 1.8 Ga, possibly associated with the evolution of the supercontinent cycle, widespread euxinia by ca. 1.85 Ga ushered in the "boring billion" year period. A second time of significant and irreversible change, in the Neoproterozoic, saw a second major oxidation event and several low palaeolatitude Cryogenian (740-630 Ma) glaciations. With the veracity of the "Snowball Earth" model for Neoproterozoic glaciation being

under dispute, genesis of Pre-Ediacaran low-palaeolatitude glaciation remains enigmatic. Ediacaran (635–542 Ma) glaciation with a wide palaeolatitudinal range contrasts with the circum-polar nature of Phanerozoic glaciation. The observed change from low latitude to circum-polar glaciation parallels advent and diversification of the Metazoa and the Neoproterozoic oxygenation (ca. 580 Ma), and was succeeded by the Ediacaran-Cambrian transition which ushered in biomineralization, with all its implications for the chemical sedimentary record.

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Contents

1. Introduction

The Precambrian sedimentary record, in contrast to the Phanerozoic, while much longer also offers much greater challenges in studying secular change, not the least being the reality that error margins of at least ca. 1 Myr in radiometric dating techniques effectively negate any possible time control for sequence stratigraphy at a chronological resolution of $<10^0 - 10^1$ Myr (Catuneanu, 2006). Almost all rock types and sedimentary facies identified in the Precambrian record have Phanerozoic-Modern counterparts, with the significant exception of iron formations. The interplay between "first-order" controls of interacting mantle thermal and plate tectonic influences and "second-order" interaction of oceanic-atmospheric-biological influences (both in themselves subject to secular change) on the sedimentary record apply equally to basins of all antiquities (e.g., Eriksson et al., 2001, 2007; Bose et al., 2012). Few would challenge the dictum that it is the tectonic settings with their implicit effects on eustasy, topography, climate, denudation rates and basin formation that are the principal primary influence on global sedimentation patterns.

While a rather uniform chronological framework of changing first-order elements is often perceived (e.g., supercontinentality, global glaciations, whether of "Snowball-Earth" affinity, cf. Kirschvink, 1992, or not), it is possible that diachroneity of influences and products was a more likely scenario (e.g., Eriksson et al., 2011b). Some unique events may also have intervened in the Earth's secular history, such as the possible "magmatic slowdown" (ca. 2.45-2.2 Ga) espoused by Condie et al. (2009a).

Non-uniformitarian influences on sedimentation patterns include decreasing Earth-rotation rates over time (e.g., Williams, 2000, 2004), concomitant changes in the Earth's climatic and thermal zonation (e.g., Rautenbach, 2001), and the influence of microbial mats on sediment stacking patterns and even sequence architecture (Sarkar et al., 2005; Catuneanu, 2006). This paper will attempt a succinct analysis of the complex array of fixed and changeable influences on global sedimentation patterns and their secular variation.

2. The nature of the sedimentary record and sequence stratigraphy

It has long been understood that the stratigraphic record is fragmentary (cf., Blackwelder's [1909] unconformity bounded successions of North America, see also Levorsen (1943) and Sloss's [1963] "sequences"). Barrell (1917) in a paper that was many years ahead of its time, was the first to clearly understand (1) the importance of what we now term *accommodation*, the space available for sediments to accumulate, and (2) the very episodic way in which accommodation is created and removed by geological processes. He demonstrated that under typical conditions of base-level rise and fall only a fraction of geological time may actually be represented by accumulated sediment. This was emphasized by Ager (1973) who remarked that "the stratigraphic record is more gap than record", and in a later study of major unconformities at the Grand Canyon, he stressed that every bedding plane "is, in effect, an unconformity" and that it "was during the breaks that most events probably occurred" (Ager, 1993, p. 14).

The description and interpretation of bedding planes and bounding surfaces has become part of the standard practice of *facies analysis*. These surfaces constitute a hierarchy of importance reflecting their duration and extent, and there have been several attempts to develop hierarchical classifications of these surfaces and the units they enclose (e.g., van Wagoner et al., 1990; Nio and Yang, 1991; Miall, 1996). At the larger scale, Vail et al. (1977) erected a hierarchical classification for stratigraphic sequences.

Sequence stratigraphy has become the standard framework for the description and interpretation of the stratigraphic record at the regional scale (Catuneanu, 2006; Miall, 2010), and one of the central elements of sequence geology is the cyclicity of the stratigraphic record. Sequences are generated by a range of causative mechanisms operating over time scales ranging from 10^4 to 10^8 years (Table 1).

An important, yet neglected discovery about the nature of the sedimentary record is the correlation between the duration of a sedimentary unit and its sedimentation rate (Sadler, 1981; Fig. 1). With sedimentation rates ranging over more than eleven orders of magnitude, it is implied that, at every scale, from the individual bedset to the scale of a basin-fill, the time that can be accounted for by the accumulation of a given thickness of sediment, when measured at the appropriate time scale, accounts for a very small fraction of the total of elapsed time. To extend Ager's famous thought: there are gaps within the gaps, and the record is permeated with them, at every scale. Bailey and Smith (2010, p. 57-58) pointed out the ephemeral nature of most sedimentary processes: "…stratigraphic records are better viewed as the outcome of temporary cessation of the erosion and redistribution of sediment: 'frozen accidents' of accumulation."

It is now widely recognised that not only the durations of stratigraphic gaps, but also the distribution of layer thicknesses and sedimentation rates in stratigraphic successions are fractal in nature (Plotnick, 1986; Schlager, 2004; Bailey and Smith, 2005). The fractal model provides an elegant basis for integrating our knowledge of the processes of accommodation generation with the data on varying sedimentation rates and the varying scales of hiatuses and the processes that operate over these various time scales (Fig. 2). Miall (in press) explores the various autogenic and allogenic (respectively, equate to intra-basinal and extra-basinal) processes by which the "frozen accidents" achieve preservation.

Fragmentary the stratigraphic record might be, but the fractal nature of the record means that it consists of intervals of succession fragments separated by larger gaps that developed at higher time scales. These larger gaps constitute the boundaries between stratigraphic sequences. Several decades of analysis have now indicated that there is a limited number of sequence types, which develop because of the occurrence of particular allogenic processes that are characterized by particular time scales (Miall, 1995, 2010; Table 1). These natural time scales, because of their predominance, tend to lead to enhanced preservability, and it is for this reason that sequence stratigraphy "works." Miall (in press) proposes the definition of a suite of *Sedimentation Rate Scales* to encompass the range of time scales and processes that can now be recognized from modern studies of the stratigraphic record (Fig. 2).

In the absence of any techniques for the accurate dating of short-term events in the Precambrian it is difficult, if not impossible, to determine how the processes identified in Fig. 2 and Table 1 might have varied in duration and importance over the course of the

Earth's history. It seems unlikely that there have been any changes in the physics of clastic sedimentation, so that, at the smaller, outcrop scale, few differences in clastic deposits may be expected through the Earth's history. Indeed, close-up images from Mars are indicative that at this scale the clastic record there is quite similar in facies.

3.   External (allogenic) controls at planetary or larger scale

3.1. Solar system mechanics and the Earth's rotation rate

The length of the year implicit in geochronology is that of the tropical year (or solar year, from equinox to equinox) of 365.242 mean solar days (Allen, 1973; Gradstein et al., 2004). However, a changing gravitational constant G would alter the scale and periods of the orbits of the planets and satellites, raising the question whether radiometric ages in years equate to the Earth's orbits around the Sun. J.G. Williams et al. (2004, p. 4) found that there was no discernible variation of $G,$ and that cosmological expansion is not shared at the solar system scale; radiometric ages do thus equal Earth orbital revolutions.

Information on the Earth's palaeorotation and the past lunar orbit best comes from the analysis of sedimentary tidal rhythmites. Internally self-consistent palaeotidal data recorded by tidal rhythmites from the late Cryogenian (ca. 635 Ma) Elatina Formation in South Australia include 13.1 ± 0.1 synodic (lunar) months/yr, 400 ± 7 solar days/yr, a length of day of 21.9 ± 0.4 hours, and an Earth–Moon distance of 0.965 ± 0.005 of the present distance (Williams, 2000, 2004). The data indicate a mean rate of lunar recession since the late Cryogenian of 2.17 ± 0.31 cm/yr caused by tidal friction, little more than

half the present rate of lunar recession of 3.82 ± 0.07 cm/yr obtained by lunar laser ranging (Dickey et al., 1994). An even lower rate of lunar recession in the early Precambrian is implied to avoid a catastrophic close approach of the Moon prior to 3 Ga (Fig. 3), for which no geological evidence exists.

Secular change in the Earth–Moon system has influenced terrestrial sedimentary systems and has implications for global tectonics: (1) A decreasing frequency of powerful tides with time (ca. 30 spring tides/annum in the early Palaeoproterozoic, 26.2 in the late Cryogenian and 24.7 today; Williams, 2000, 2004). (2) A faster rotating Proterozoic Earth led to less efficient poleward transport of heat, with a slightly warmer equator and cooler poles (Kuhn et al., 1989; Jenkins, 1993), compounding the enigma of Proterozoic low-palaeolatitude glaciation. (3) The hypothesis that the orbital periods for precession and obliquity have increased with time due to secular increase in the Earth–Moon distance (Berger et al., 1989) is confirmed by the precession period of 19.6 ± 1.1 kyr and the obliquity period of 31.3 ± 3.0 kyr recorded by Late Ordovician–Early Silurian evaporites in Western Australia (Williams, 1991), compared to respective periods of 23 kyr and 41 kyr today. (4) Resonances of the Earth's fluid core with the semi-annual and annual forced nutations for the despinning Earth, estimated to have occurred more strongly in the Neoarchaean and around 530 Ma, respectively, would have amplified core motions and possibly increased the flux of mantle plumes and tectonothermal activity at those times (Williams, 1994, 2004). (5) The Elatina palaeotidal data indicate no overall change in the Earth's moment of inertia since the late Cryogenian, ruling out secular change in the Earth's radius at least since that time (Williams, 2000, 2004).

3.2. Change from mantle-domination to plate-domination in the Earth's geodynamics

While few would dispute there was a change from an earlier mantle-dominated Earth to one where plate tectonics became paramount, the timing of this gradation remains contentious. Some argue for a very early onset of plate regimes (e.g., de Wit, 1998; de Wit and Hynes, 1995), thereby making the debate on timing peripheral to the preserved continental crustal and sedimentary record, whereas others see this transition as being significantly later (at ca. 2.0 Ga; e.g., Hamilton, 1998). Both short-period and longer-term secular changes and significant variations in geological processes argue against a simple transition of the Earth's geodynamic evolution along a continuous trend. Problems with strict interpretation of uniformitarian plate-tectonic models when applied to the Neoarchaean and Proterozoic record have been noted (e.g., Hoffman, 1989; Bickford and Hill, 2007; Stern, 2008), and Mints (this volume, and references therein) discusses specific examples of apparent complexity in the evolution of the Earth's geodynamics. The latter include a possible sudden turning-point in crust-forming processes in the Neoarchaean, as well as an alternative model for intracontinental formation of granulite-gneiss belts with protoliths of metasedimentary and metavolcanic granulites filling depressions above mantle plumes, rapidly followed by high-grade metamorphism (e.g., Mints, 2007; Mints and Konilov, 2004; Mints et al., 2007).

The antiquity of the supercontinent cycle forms the subject of a parallel debate, equally unresolved (e.g., Unrug, 1992; Rogers, 1996; Aspler and Chiarenzelli, 1998). The inherent link (e.g., Zhong et al., 2007; Santosh et al., 2009) between this unresolved question and the mantle plume concept (viz. Condie, 2004 a, b; Condie et al., 2001)

suggests that the supercontinent cycle was generated by the *interaction* of both mantle thermal and plate tectonic regimes. The debate is further complicated by possible global-scale tectonic–thermal "slowdowns" from ca. 2.45–2.2 Ga, and at ca. 1.4 Ga (Condie, 1998; Condie et al., 2009a). It has been suggested that the first (ca. 2.45-2.2 Ga) event even separated an early form of plate regime interacting with a higher energy mantle system, from the onset of truly Phanerozoic-style plate tectonics from ca. 2.2–2.0 Ga. Eriksson and Catuneanu (2004) argued for a gradual transition from the Hadean–Archaean Earth dominated by magmatic processes to one where plate tectonics increasingly interacted with mantle thermal processes, possibly close to the Archaean–Proterozoic boundary; they stressed the possible relevance of Trendall's (2002) "plughole model" for such a gradual transition. However, the issues raised by Mints (this volume) should possibly be superimposed on this simple, gradational model of the Earth's evolution.

While many scientists see supercontinentality becoming almost pervasive by the Neoarchaean (or even earlier) (e.g., Aspler and Chiarenzelli, 1998), others caution against a global-scale onset for this paradigm, arguing instead for the coexistence of one (e.g., "Kenorland" espoused in the Aspler and Chiarenzelli, 1998 view) or more (e.g., Bleeker, 2003) supercontinents with cratons which were not yet part of such amalgamations, with Kaapvaal being one possible such example (e.g., Eriksson et al., 2011 a and b). Mints (this issue) employs data from the Palaeoproterozoic and Meso–Neoproterozoic of Lavroscandia and from other continents, to argue that a long-lived supercontinent (or group of large continents) possibly played the main role in the evolution and growth of the continental crust from ca. 2.80 to ca. 1.0 Ga or maybe even to ca. 0.55 Ga (Fig. 4).

4. Internal (autogenic) sedimentary controls on the planetary scale

4.1. Introduction

Autogenic controls on sedimentation reflect the interplay of geospheric processes within the Earth system (hydrosphere, atmosphere, biosphere). Chemical sedimentary rocks best record such feedbacks because their deposition directly reflects ocean-atmosphere chemistry, climate, weathering, nutrient cycling, and biologic evolution. Thus, variations in feedback rates regulating these factors govern the type and abundance of chemical sedimentary deposits through time (Fig. 5).

The most significant change in chemical sedimentation occurred during the Great Oxidation Event (GOE) between 2.4 and 2.3 Ga (e.g., Bekker et al., 2004; Canfield, 2005; Holland, 2006; Guo et al., 2009; Bekker and Holland, 2012; Pufahl and Hiatt, 2012). The GOE recorded the global expansion of oxygenic photosynthesis in cyanobacteria (Fig. 5; Cloud, 1973; Canfield, 2005; Saito, 2009), the disappearance of reduced detrital minerals such as pyrite and uraninite (Cloud, 1968; Fleet, 1998; England et al., 2002; Hazen et al., 2008) and the concomitant appearance of the large continental margin iron formations, phosphorites, and red beds (Holland, 2002; Bekker et al., 2010; Papineau, 2010; Pufahl, 2010; Pufahl and Hiatt, 2012). Although a number of processes likely preconditioned the atmosphere to accumulate oxygen (Kasting et al., 1979; Des Marais et al., 1992; Catling et al., 2001; Godderis and Veizer, 2004; Zahnle et al., 2006; Konhauser et al., 2009; Papineau, 2010), the most important was probably the switch from reducing to oxidizing volcanic gases (Kump et al., 2001; Holland, 2002). The GOE

irreversibly changed the nature of the Earth system and paved the way for all subsequent keystone changes in chemical sedimentation (Fig. 5).

4.2. Volcanic outgassing and volcanic record

Since the formation of the Earth a steady cooling of the core and mantle is assumed (e.g., Vlaar et al., 1994; Lenardic et al., 2005; Korenaga, 2006, 2008; Herzberg and Gazel, 2009; Herzberg et al., 2010), reflected in the decreasing Mg content of high temperature, ultramafic magmas with time (Skulski et al., 2004): komatiites (>18% MgO; commonly thought to have originated from deep-seated mantle plumes; Campbell et al., 1989; Arndt et al., 2008) were common in the Archaean, komatiitic basalts (13-18% MgO) in the Proterozoic, and picrites (10-13% MgO) in the Phanerozoic.

The secular compositional changes in volcanic rocks can also be found in their erosional products (shales and sandstones; e.g. Schwab, 1978) with trends for sedimentary and igneous rocks (komatiites, basalts, andesites, felsic extrusive rocks, and granites) showing that they changed sympathetically (Migdisov et al., 2003). The abundance of Mg-bearing mineral phases decreased from the Early Archaean to the Phanerozoic and those of K-bearing minerals and, in part, quartz increased with time.

Closely related to the changes in magma composition is also the change in magmatic volatile composition and volcanic outgassing that closely affected seawater chemistry as well as the composition of the atmosphere (Fig. 5). The most abundant gases that are released by volcanic systems and are being added to the atmosphere-ocean system today, are H_2O, CO_2 and SO_2, followed by smaller amounts of H_2S, H_2, CO, HCL, HF, and He

(Symonds et al., 1994; Giggenbach, 1996). These gases are removed from the system by the deposition of organic matter and carbonate minerals (CO_2 and CO), sulphide and sulphate minerals (H_2S and SO_2), and by the reduction of CO_2 and CO to organic matter, the reduction of SO_2 to sulphides, and by loss into interplanetary space (H_2; cf., Holland, 2002).

Probably the most dramatic change in gas chemistry took place between 2470 and 2450 Ma (Kump et al., 2001), when intense magmatic plume activity culminated in a large igneous province recognized on several continents (Barley et al., 1997; Heaman, 1997). During this time, an initially more reduced upper mantle became progressively more oxidized due to the release of reduced volcanic gases (probably mainly H_2 and CO) and subduction of hydrated, oxidized seafloor. This mantle overturn may have changed the redox state and therefore the oxygen fugacity (fO_2) of volcanic gases, taking the system through the crossover point to an oxic environment and eventually the GOE (Fig. 5; Kump et al., 2001; Holland, 2002). In the Phanerozoic, high rates of seafloor spreading and their related increased atmospheric pCO2 through volcanic outgassing as well as lower oceanic Mg/Ca through intensified hydrothermal alteration of basalt, is hypothesized to have caused the change from aragonite to calcite seas (Mackenzie and Pigott, 1981; Sandberg, 1983; Stanley and Hardie, 1998; Kump, 2008; Hasiuk, 2008).

4.3. Secular changes in carbonate sedimentation

The connection between photosynthesis, increasing oxygen, the delivery of sulphate to the oceans, and resulting change of carbonate mineralogy is the first-order secular change

in carbonate sedimentation. Carbonates in the Archaean and Proterozoic were largely formed by evaporative processes and, like chert, were chemical precipitates strongly influenced by microbial processes (Fig. 6). Precambrian carbonate rocks are almost exclusively dolomite (Fig. 7), probably due to evolving seawater chemistry (Fairbridge, 1957; Tucker, 1982; Given and Wilkinson, 1987; Warren, 2000; Machel, 2004). Low sulphate concentrations led to greater dolomite precipitation (e.g., Zentmyer et al., 2011) as sulphate kinetically inhibits dolomite formation (Baker and Kastner, 1981; Morrow and Rickets, 1988; Morrow and Abercrombie, 1994). The removal of SO_4^{2-} occurs via the precipitation of sulphate minerals or by bacterial sulphate reduction, a relationship supported by association of dolomite in evaporative environments or with sulphide minerals (particularly in peritidal evaporative environments; Fig. 6; Zentmyer et al., 2011). In shallow settings where evaporation concentrated silica, abiogenic chert also precipitated (Maliva et al., 1989, 2005; Zentmyer et al., 2011). Weathering of sulphide minerals in the presence of oxygen produced SO_4^{2-}, which accumulated in the oceans following the GOE (Canfield, 1998; Habicht et al., 2002; Poulton et al., 2004). Although it is widely believed that the Earth's early oceans had low sulphate concentrations, the actual levels are debated (e.g., Walker and Brimblecombe, 1985; Grotzinger and Kasting, 1993; Knauth, 2005).

Proterozoic carbonate sedimentary rocks are also characterized by non-actualistic sedimentary structures, reflecting ocean water that was oversaturated with respect to $CaCO_3$, and include giant ooids (Sweet and Knoll, 1989) and aragonite fans (Sumner and Grotzinger, 2004), which only reappear in the Phanerozoic record following major extinction events (Fig. 5; Grotzinger and Knoll, 1999). The Phanerozoic saw a revolution

in which Eukaryotic biomineralization forever changed the nature of carbonate sedimentation (Fig. 5).

4.4. Biologic evolution and chemical sedimentation

Throughout the history of life, innovations in the metabolic processes essential to the growth, reproduction, survival, and evolution of organisms have been a primary control on the Earth's surface processes, the cycling and sequestration of nutrients, and the precipitation of minerals forming chemical sediments. Arguably the most important of these innovations was oxygenic photosynthesis in cyanobacteria. Although photosynthesis is deeply rooted in the tree of life, stretching back to the early Archaean (e.g., Hohmann-Marriott and Blankenship, 2011), oxygenic photosynthesis evolved much later. Biomarker evidence suggests cyanobacteria may have evolved by 2.9 Ga (Nisbet et al., 2007) and were abundant by 2.7 Ga (Brocks et al., 2003a and b, 2005; Canfield, 2005; Schopf, 2006; Buick, 2008), ca. 400 Myr before the GOE. This lag is interpreted to record a period when oxygen-consuming, inorganic and organic reactions prevented the rise of photosynthetic oxygen. The onset of the GOE resulted in the accumulation of oxygen-dependent lithofacies, Mn-oxides, iron formations, phosphorites and glauconites (Roy, 1997; Nelson et al., 2010; Pufahl, 2010), and later allowed for the evolution and diversification of Eukaryotes (Fig. 5; Knoll, 2011).

By ca. 1.85 Ga the bacterial reduction of seawater SO_4^{2-} is thought to have created widespread euxinia, causing the demise of continental margin iron formation (Canfield, 1998; Poulton et al., 2004; 2010) and phosphorite (Nelson et al., 2010; Pufahl and Hiatt,

2012). Such conditions are interpreted to have lasted for nearly a billion years (the "Boring Billion", with dramatic slowing of Eukaryotic evolution; Anbar and Knoll, 2002) and were likely perpetuated by anoxygenic photoautotrophs that tempered oxygen production by using H_2S as an electron acceptor (Johnston et al., 2009). Iron formation and phosphorite reappeared in the Neoproterozoic during several glaciations between 740 and 630 Ma (Klein, 2005; Bekker et al., 2010; Pufahl, 2010).

A second major increase in photosynthetic oxygen at ca. 580 Ma, the Neoproterozoic Oxygenation Event (Fig. 5; Och and Shields-Zhou, 2012), ventilated the deep oceans creating the necessary conditions for forming the first true phosphorite giants (Pufahl, 2010; Pufahl and Hiatt, 2012) and oxygen concentrations rose to levels that led to the diversification of multicellular animals (Canfield et al., 2007; Och and Shields-Zhou, 2012). The diversification of mobility and feeding modes amongst late Ediacaran-early Cambrian metazoans (McIlroy and Logan, 1999; Erwin et al., 2011) had a dramatic effect on the physical and chemical nature of marine sediments. Thus, the Ediacaran-Cambrian transition was marked by an increase in bioturbation intensity (Droser and Bottjer, 1988), a "substrate revolution" (Bottjer et al., 2000) which drastically reduced the thickness and distribution of previously ubiquitous microbial facies and mats (Garrett, 1970; Hagadorn and Bottjer, 1997). Consequently, Phanerozoic seafloors developed large mixed layers with higher water contents, which produced an increasingly heterogeneous sediment-water interface (Seilacher and Pflüger, 1994; Hagadorn and Bottjer, 1997; Seilacher, 1999; Plotnick et al., 2010). Biomineralization by Eukaryotes was another major innovation of the Ediacaran-Cambrian radiation (Weiner and Dove, 2011).

Although older examples of biomineralizing metazoans are known (Porter and Knoll, 2000; Cohen et al., 2011), the Ediacaran-Cambrian radiation (Erwin et al., 2011) marked a widespread appearance of skeletonized taxa in the form of calcified tubes (e.g., *Cloudina*; Germs, 1972) and an array of bilaterian plates, spines and shells (Bengtson, 1992; Maloof et al., 2010). Diversification in biomineralizing organisms was followed by an increase in overall skeletal contribution to shallow-water carbonate facies in the Ordovician, especially by heavily calcified corals, bryozoans, brachiopods, and echinoderms (Pruss et al., 2010). The subsequent evolution of biomineralizing foraminifera (benthic in the Devonian and planktonic in the Jurassic; Hart et al., 2002) and coccolithophores (Triassic) significantly affected carbonate facies distribution by transferring carbonate deposition offshore (Tucker, 1985; Milliman, 1993; Erba, 2006). The Cenozoic explosion of silicifying diatoms severely limited the abundance of silica in surface oceans (Maliva et al., 1989), and may have led to a decrease in silica available for sponges and radiolarians (Lazarus et al., 2009).

The removal of organisms via mass extinctions also profoundly influenced chemical sedimentation (e.g., Payne et al., 2007) with the largest Phanerozoic mass extinction at the end-Permian (Erwin, 1993, 2001; Shen et al., 2011; Payne and Clapham, 2012) returning many carbonate depositional systems to Proterozoic-style sedimentation regimes (Grotzinger and Knoll, 1995; Knoll et al., 1996; Pruss et al., 2004).

5. Greenstone belt sedimentation

Pre-rock record sedimentation (>4 Ma) was probably dominated by meteoritic and cometary impact events (cf., Schoenberg et al., 2002) causing very large tsunamis, resulting in very coarse volcaniclastic detritus combined with fine dust settling out of suspension, all reworked by marine current systems and localised turbidites (Eriksson et al., 2007). $^{182}W/^{184}W$ ratios of Isua metasediments suggest derivation of the parent lithologies from meteorite debris (Willbold et al., 2011). Simonson et al. (2004) suggested that at least 6 spherule-layers bear testimony to global-scale impact events in rocks older than 3.2 Ga.

Basement-cover relationships in older and younger greenstone belts (Blenkinsop, 1993; Buick et al., 1995; Bleeker, 2003), epiclastic sedimentary rocks associated with volcanic successions (Mueller and Corcoran, 1998; Corcoran and Mueller, 2007), and abundant, compositionally mature siliciclastic sedimentary rocks (e.g., Krapez and Barley, 1987; Eriksson, et al., 1997; Hessler and Lowe, 2006) provide evidence for exposed land masses and continental weathering. The majority of Mesoarchaean quartz arenite successions are considered to have been deposited on stable continental shelves along rifted margins (Thurston and Chivers, 1990; Eriksson et al., 1994; Mueller et al., 2005), but quartz-rich deposits of fault-controlled, high relief Mesoarchaean (e.g., Krapez, 1996) and Neoarchean (Corcoran et al., 1998) successions are mainly a function of source composition (granodiorite) and intense chemical weathering under reducing atmospheric conditions. These conditions are supported by a paucity of clays and feldspars in Archaean clastic deposits, in addition to which mudstones from the Fig Tree

and Moodies Groups (Barberton belt, Kaapvaal) and from the Gorge Creek and Roebourne Megasequences, Pilbara Block, Australia are enriched in aluminum relative to average shales (Eriksson and Soegaard 1985; Hessler and Lowe, 2006).

Recognizable sedimentary processes in greenstone belts include debris flows on high-gradient alluvial fans, mid-channel bar formation and unconfined sheetflood deposition in low-sinuosity, gravel- and sand-dominated braided rivers, wind ripple and dune migration in aeolian settings, tidal and wave reworking in shallow-marine environments, and deposition from turbulent flows (e.g., Eriksson, 1977, 1979; Eriksson et al., 1994; Heubeck and Lowe; 1994; Corcoran and Mueller, 2004; Mueller et al., 2005; Long, 2011). The predominance of low-sinuosity braided to the exclusion of high-sinuosity meandering fluvial deposits is compatible with high runoff rates linked to steep gradients, related to a lack of bank stabilization in the absence of land plants (Schumm, 1961; Davies and Gibling, 2011; Long, 2011). Unconfined runoff may have been more prominent than stream flow processes as a result of interstratification with erosion-resistant, felsic volcanic flows (Mueller and Corcoran, 2001). Steep gradients and high runoff rates promoted the formation of braid deltas rather than river deltas at the land-ocean interface. Quantitative analysis of tidal sand waves in the Moodies Group, South Africa reveals the presence of semi-diurnal, diurnal and fortnightly signals and implies that tidal current velocities for these deposits were within the range for modern tides (Eriksson and Simpson, 2000).

Maximum water depths of sedimentation are difficult to constrain from the preserved sedimentary record in greenstone belts. The predominance of shallow-water facies does not necessarily imply that Archaean oceans in general were shallow; on the contrary,

estimates of average Archaean ocean depths are circa 2.6 km (Bickle et al., 1994). Siliciclastic turbidites in the Fig Tree Group, South Africa do not necessarily record abyssal depths, as the presence of hummocky cross-stratification in this succession suggests that maximum depths did not exceed tens of metres. It is thought unlikely that secular changes exist for deep water sedimentation, although the prominent controls on deposition during the Archaean may have been tectonic rather than eustatic.

Non-clastic sedimentary rocks in greenstone belts record chemical and biological processes different from those operating in later geological time (discussed in section 4). The significance of Archaean barite is controversial. Lowe (1983) and Buick and Dunlop (1990) consider barite in the Warrawoona Supersequence to be pseudomorphous after gypsum that formed as a normal evaporite from seawater of comparable composition to that of today. Grotzinger and Kasting (1993), in contrast, argue that calcium would have been exhausted before the gypsum field was reached with little or no precipitation of calcium sulphate except near sites of continental runoff from weathering of basalts. In addition to calcium, Archaean seawater also was enriched in Fe^{++}. Upwelling of reducing bottom waters into zones of oxygen productivity resulted in precipitation of iron oxides. Such iron formations are most common in the stable-shelf successions preserved in younger greenstone belts (<3.0 Ga) in the Slave and Superior Provinces, Canada, the Yilgarn Block, Australia and in Zimbabwe (e.g., Blenkinsop, 1993; Bleeker, 2003; Wyche et al., 2004; Mueller et al., 2005). Sulphidic iron formations likely developed below the pycnocline and Eriksson (1983) has likened Archaean iron formations to Holocene pelagic sediment rainout.

6. Earliest cratonic sedimentation

The 4.4 Ga zircons in Australian Meso- or Palaeoarchaean fan-delta metaconglomerates are the first clear evidence for a stable felsic continental crust (Eriksson and Wilde, 2010; Wilde et al., 2001); oxygen isotopes in these zircons provide definitive evidence for a hydrosphere (Cavosie et al., 2005), with lithium isotopes indicating intense weathering of Eoarchaean continental crust in a CO_2-rich environment at elevated temperatures (Ushikubo et al., 2008). The area of continental crust exposed to weathering prior to 3.2 Ga was probably insignificant (Shields, 2007; Bradley, 2011), allied also to high global sea levels (Flament et al., 2011), explaining the absence of extensive terrestrial and shallow marine deposits in Eoarchaean (3.6-4.03 Ga) and Palaeoarchaean (3.2-3.6 Ga) strata. Minor terrestrial strata are present in the Mesoarchaean (2.8-3.2 Ga) of Kenorland (Donaldson and de Kemp, 1998; Breaks et al., 2001), but are more prominent in the Neoarchaean (2.5-2.8 Ga), where alluvial fan and braided river deposits occur in small foreland (Corcoran and Mueller, 2007), piggyback (Kusky and Kidd, 1992; Devaney, 2000), and late transtensional "Timiskaming-type" basins (Mueller and Corcoran, 1998; Bleeker et al., 1999; Eriksson and Wilde, 2010; Corcoran, 2012). Shallow marine quartz-rich strata tend to be preserved in extensional or sag basins (Blight 1985), or where directly associated with komatiites or other mafic lavas, along rifted passive margins (Sakurai et al., 2005; Corcoran, 2012).

The Kaapvaal craton was amongst the earliest to stabilize globally, and hosts the Earth's oldest known large basin, including significant placer gold (Frimmel et al., 2005). Despite a complex geodynamic history (Robb and Meyer, 1995a, b), the craton may not

have been part of a larger supercontinental assemblage during deposition of the Witwatersrand basin, and a sub-Kaapvaal plume may have influenced craton accretion (cf., model of De Wit et al., 1992), formation of a complex flexural foreland basin (Catuneanu, 2001) and enhanced gold provenance (Eriksson et al., 2009, 2011a). Large braided systems, terminating in braid-deltas, debouched directly into a shallow epeiric sea on the distal craton (Els and Mayer, 1992, 1998; Catuneanu, 2001).

Within what became the Kenorland supercontinent (Aspler and Chiarenzelli, 1998), numerous thin successions of quartz-rich sandstones indicate stable shallow water platformal sedimentation (Ojakangas, 1985; Donaldson and de Kemp, 1998). In the Mesoarchaean of the Slave Province, a quartz arenite dominated succession of tide influenced chromite- and fuchsite-bearing sandstones, locally associated with ultramafic flows and sills, and typically overlain by banded iron formations (Bleeker et al., 1999; Corcoran, 2012), are consistent with transgression on a rifted continental margin. Within the Mesoarchaean and Neoarchaean of the Superior Province (Ojakangas, 1985; Donaldson and de Kemp, 1998), and Baltic Shield (Thurston and Kozhevnikov, 2000), there are numerous examples of similar quartz-rich sandstones, mostly of shallow marine origin and constrained to narrow, linear basins. Evidence of extensive carbonate deposition within Kenorland is limited to deposits within the Steep Rock Group that formed on top of a volcanic plateau (Fralick et al., 2008). The major differences in basin and sedimentation style between possible isolated cratons, like the Kaapvaal, and larger continental amalgamations, like Kenorland were probably linked to differences in tectonic control of basin evolution and freeboard, as outlined by Eriksson and Condie (2012).

7. Alluvial sedimentation and secular change

Major secular changes in fluvial style have been driven by variations in atmosphere chemistry, Earth's surface temperature, weathering regime, and most significantly the onset and evolution of terrestrial vegetation (Long, 2011). During the early Hadean (ca. 4.5-4.567 Ga) the Earth's surface was probably molten, with initial terrestrial deposits being dominated by minor talus, loess and aeolian dunes. As surface temperatures dropped to below $100^{o}C$ at approximately 4 Ga (Eriksson et al., 2007), temporary accumulations of water would have permitted development of fluvial- and debris flow-dominated alluvial fans, lakes and fan-deltas, as can be seen on the surface of Mars (Buhler et al., 2011; R.M.E. Williams et al., 2011) and Venus (Jones and Pickering, 2003). All evidence for these early terrestrial facies on the Earth was probably destroyed by the end of the period of heavy bombardment, at approximately 3.9 Ga. Although the paucity of >3.2 Ga lacustrine deposits is a function of poor preservation, the relatively small number of documented Mesoarchaean (2.8-3.2 Ga) and Neoarchaean (2.8-2.5 Ga) examples is probably the result of difficulties in distinguishing between marine and fresh water units in the absence of biological indicators.

Palaeoarchaean (3.2-3.6 Ga) and Mesoarchaean fluvial deposits are poorly represented in the rock record. Most fluvial strata of this age include high gradient alluvial fan deposits with evidence of debris flows (Breaks et al., 2001; Eriksson and Wilde, 2010). Many of the associated sandstone units tend to be planar bedded, implying deposition under upper flow regime conditions. Fralick and Carter (2011) argue that elevated surface temperatures during the Archaean would have significantly reduced the viscosity of

water, generating a thinner viscous sublayer, consequently suppressing ripple formation in sands coarser than fine sand. Dune formation does not appear to have been suppressed, as abundant trough cross-stratification has been recorded in braided fluvial strata at the base of the Mesoarchaean Bababudan Group by Srinivasan and Ojakangas (1986).

By the Neoarchaean (2.5-2.8 Ga) both fan and braided river deposits are better represented, both in transtensional (Corcoran, 2012; Corcoran et al., 1998, 1999; Driese et al., 2011), rift-related, platformal and foreland basin settings (Els, 1998a; Karpeta and Els, 1999; Minter, 1999; Eriksson et al., 2001; Corcoran and Mueller, 2007). Many of the more extensive fluvial systems may have had the form of distributary fan systems (Els, 1998b; Cain and Mountney, 2009; Hartley et al., 2010). As Neoarchaean river banks were not stabilized by plants, width to depth ratios of braided rivers would have been significantly greater, perhaps locally exceeding 1000:1 (Els, 1990). Overbank deposits have not been reported, but lake or pond deposits were present locally, as evidenced by abrupt lithofacies transitions from trough cross-bed - dominated, coarse-grained sandstone and clast-supported, stratified conglomerate to local siltstone-mudstone units containing wave ripples, graded beds and dewatering structures (Krapez, 1984; Mueller et al., 1994; Corcoran et al., 1999; Corcoran and Mueller, 2004). Although most described lacustrine units are laterally restricted, an extensive lake system has been identified in the Tumbiana Formation, Fortescue Group (Buick, 1992; Awramik and Buchheim, 2009). Microbial material may have accumulated in ephemeral ponds in channel thalwegs (Mossman et al., 2008) and in soils (Driese et al., 2011).

Long (2011) provides a comprehensive review of pre-vegetation fluvial systems, with emphasis on the Proterozoic. He noted that due to the absence of rooted vascular plants

most pre-vegetation systems have features that are more closely allied to modern ephemeral and dryland systems. Sand-bed and gravel-bed braided systems are common, and have architecture similar to that of modern systems, except that scour hollows are typically absent or difficult to identify. Sandy braided systems are dominated by composite barforms with predominantly downstream accretionary elements. Sandy ephemeral channelized upper-flow-regime elements, and unconfined sheetflood deposits are common, but lack many of the associated fine-grained components seen in modern systems. Sandy meandering systems can be identified, based on up-section changes in palaeoflow vectors, but typically lack the fining-upward trends seen in modern systems. Overbank deposits are rare, except in marginal marine systems. Point-bars in rare cobble-grade gravel-bed meandering systems were apparently wider and had lower inclination than modern systems. Wandering gravel-bed rivers, low-sinuosity sandy braided systems with alternate bars, and fine-grained meandering and anabranching systems have not been positively identified. Documented lacustrine deposits are more prevalent in the early to middle Proterozoic than the Archaean (Collinson and Terwindt, 1983; Eriksson, 1991; Martins-Neto, 1996).

The progressive colonization of terrestrial environments by rooted plants in late Silurian to middle Devonian times significantly affected microclimate by modifying albedo and moisture retention in soils, leading to a greater role of organic acids in decomposition of labile components (Jutras et al., 2009). This led to increased production and subsequent trapping of clay minerals in overbank settings. Davies and Gibling (2010a, b) suggest that the initial colonization of land by complex plants (embrophytes) took place between 472 and 436 Ma, but had little effect on fluvial style. The first

advanced vascular plants appear to have evolved in the late Llandovery (Davies and Gibling, 2010a, b). These lacked penetrative roots and were restricted to lowland settings, so had minimal effect on bank stabilization, but may have led to a slight increase in mud retention in overbank settings (Davis and Gibling, 2010b). Davis and Gibling (2010a, 2011) suggest that a transition from predominantly sheet-braided to channel-braided fluvial systems may have occurred at this time. Heterolithic meandering stream deposits, with well-developed lateral accretions sets, are first reported from the late Silurian; they became far more abundant with the development of substantial plant root systems in the Early Devonian.

The onset of advanced terrestrial vegetation systems, including trees, in Middle Devonian tropical to boreal settings marks the first development of extensive forests and extensive stable mud-rich overbank settings. Davies and Gibling (2011) suggest that the continued evolution and expansion of trees and rooted plants in the Carboniferous led to the generation of low-gradient, avulsion prone, organic-rich, fixed multi-channel anastomosed (anabranching) river systems, due to increased root stabilization of river banks and development of extensive riparian and in-channel vegetation. The rapid expansion of grasses into dryland and well drained upland environments during the Cenozoic (Strömberg, 2011) is liable to have had a profound effect on fluvial styles, especially in ephemeral systems as it would baffle overland flows, and possibly enhance rill development.

8. Aeolian sedimentation and secular change

Millimetre- to cm-scale inversely graded strata generated by wind-ripple migration provide an unequivocal means for distinguishing subaerial from subaqueous depositional processes and thereby enable the recognition of aeolian processes in continental settings (Hunter, 1977, 1981; Kocurek and Dott, 1981; Fryberger and Schenk, 1988) or even on Mars (Grotzinger et al., 2005). Based on these criteria, the oldest recognized evidence of aeolian deposition is from the 3.2 Ga Moodies Group of South Africa (Simpson et al., 2012). Other notable Archaean examples of wind processes are ventifacts associated with placer deposits in the 2.9 Ga Witwatersrand Supergroup, South Africa (Minter, 1976; 2006) and sand-sheet deposits composed of aeolian stratification from the 2.6 Ga Minas Supergroup of Brazil (Martins-Neto and Costa, 1985). Reports of Proterozoic and younger aeolianites are more commonplace and probably reflect alternative controls on formation and preservation than those that occurred before ca. 1.8 Ga. (Eriksson and Simpson, 1998; Simpson et al., 2004b).

Early evolved biological soil crust (BSC)-like communities played an important role in the early biological invasion of Precambrian continental landscapes, where they stabilized land surfaces and reduced the flux of sediment into aeolian systems (Campbell, 1979; Belnap et al., 2001; Prave, 2002; Retallack, 2008; Beraldi-Campesi et al., 2009; Finkelstein et al., 2010; Simpson et al., 2010; Beraldi-Campesi and Garcia-Pichel, 2011; Malenda et al., 2012). The oldest recognized BSCs cluster around ca. 1.2 Ga (Prave, 2002; Beraldi-Campesi and Garcia-Pichel, 2011; Beraldi-Campesi et al., 2009); algal mat related roll-up features are reported in 1.8 Ga interdune deposits (Eriksson et al., 2000)

and organic signatures are inferred from $\delta^{13}C$ and bulk rock chemistry in Neoarchaean soils (Watanabe et al., 2000). The paucity of reported examples of Archaean aeolianites suggests the possibility that microbes may have been more important in stabilizing the land surface than previously recognized. The absence of vegetative cover in the Precambrian would generally have enhanced sediment flux (Dott, 2003) which dynamically links fluvial and aeolian systems over different time scales of months to Milankovich periodicities (Clemmenson et al., 1989; Kocurek, 1991, 1996, 1999; Lancaster, 1997; Howell and Mountney, 1997; Chan and Archer, 1999; Loope et al., 2001; Swezey, 2001; Bullard and Livingstone, 2002; Veiga et al., 2002; Scherer and Lavina, 2005; Kocurek and Ewing, 2005).

With higher fluxes of sediment to and from the aeolian systems in the Precambrian, aeolian deposits likely extended across a wider range of climate regimes than today (Eriksson and Simpson, 1998; Simpson et al., 2004b); as a result, we argue that large dune fields and associated facies are the best indicators of deserts in the early rock record. One of the best examples is the Palaeoproterozoic Makgabeng Formation, Waterberg Group, South Africa which consists of dunefield deposits comprising of thick foresets and associated interdune deposits (Eriksson et al., 2000; Simpson et al., 2004a, b). Saline-pan deposits separate the dunefield into two distinct ergs (Eriksson et al., 2000; Simpson et al., 2004a). These saline pan deposits preserve micro-evaporitic features, such as those reflecting solution collapse and that have direct analogues in modern saline pans, and are one of the oldest reported examples (Simpson et al., 2004a). Climate change terminated the youngest erg (Simpson et al., 2002) as evidenced by massive sandstones generated by extreme precipitation events that degraded dune lee faces before truncation by the

overlying Mogalakwena Formation fluvial system (Simpson et al., 2002; Eriksson et al., 2008a).

Eriksson and Simpson (1998) argue that preservation of large-scale Precambrian aeolianites is possibly dependent on suitable tectonic settings of high accommodation during the early breakup of supercontinents and to a lesser degree on the assembly of supercontinents. Numerous semi-quantitative and quantitative predictive models for aeolian subfacies have been developed that examine the roles of subsidence (assumed to be linear within the short-term of the model), sediment availability, water table, and position of the accumulation surface (e.g., Mountney, 2006).

9. Glacigenic sedimentation and secular change

9.1. Archaean and Palaeoproterozoic glaciations

The oldest known glacigenic deposits occur in the Mozaan Group of the ca. 2.9 Ga Pongola Supergroup in southern Africa, with diamictites containing faceted and striated clasts and stratified siltstones with dropstones (Young et al., 1998; Crowell, 1999). Deposition occurred on a stable marine shelf. The apparent mean palaeolatitude of the Pongola Supergroup is ca. 30° but the timing of magnetisation is poorly constrained (Strik et al., 2007).

Early Palaeoproterozoic glaciation is known from four continents, but their temporal relationship is unclear. In Canada, the ca. 2.4–2.2 Ga Huronian Supergroup records three glaciations (Crowell, 1999; Young et al., 2001). Deposits of the 600–1700 m thick

Gowganda Formation are the youngest and most widespread, covering at least 400×200 km, with correlates elsewhere in North America. Facies include massive diamictites, some with faceted and striated clasts, and rhythmites with dropstones and till pellets, deposited in a marine environment on the southern margin of the Canadian Shield (Mustard and Donaldson, 1987; Young et al., 2001). In Fennoscandia, glacial deposition occurred between 2.45 and 2.3 Ga in an area of 300×500 km (Crowell, 1999; Ojakangas et al., 2001). The Urkkavaara Formation in Finland and the Sariolian Group in adjoining Russia have metadiamictite and metasiltstone–argillite with dropstones. In South Africa, the Makganyene Formation (Transvaal Supergroup) is 50–150 m thick and formed after 2436±7 Ma (youngest detrital zircon age; Moore et al., 2012); diamictites with striated clasts were deposited during a regressive, uplift phase. In Western Australia, glaciation between 2.45 and 2.20 Ga is recorded by the 360 m thick Meteorite Bore Member of the Kungarra Formation and the Boolgeeda Iron Formation (Crowell, 1999; Martin, 1999); glacial facies include diamictites with faceted and striated clasts and outsized dropstones in fine-grained beds. A marine environment is envisaged.

Early Palaeoproterozoic glaciations in North America and Fennoscandia occurred in low palaeolatitudes while respective cratons moved across the palaeoequator (Williams and Schmidt, 1997; Schmidt and Williams, 1999; Bindeman et al., 2010). Palaeomagnetic data for 2.2 Ga volcanics in South Africa imply the immediately underlying Makganyene Formation was possibly deposited at a palaeolatitude of $11 \pm 5°$ (Evans et al., 1997). A low palaeolatitude is suggested for early Palaeoproterozoic glaciation in Western Australia (Evans, 2007).

In NW Australia, subglacial channel-fills and glaciofluvial deposits occur at the base of the ca. 1.8 Ga (<1869±7 Ma, youngest detrital zircon, Kirkland et al., 2010; >1790±4 Ma, SHRIMP U-Pb zircon age of intrusive dolerite, Page and Sun, 1994) marine King Leopold Sandstone (Williams, 2005). Glaciation reached close to sea level and coincided with orogeny in Australia and worldwide continental assembly. A low palaeolatitude (8 ± 2°) is indicated (Schmidt and Williams, 2008). Recognition of late Palaeoproterozoic glaciation overturned the view of a 2200–800 Ma nonglacial interval.

9.2. Neoproterozoic glaciations

Two Cryogenian glaciations recognised on most continents are commonly, but incorrectly, referred to as the "Sturtian" and "Marinoan" glaciations based on chronostratigraphic units in South Australia. Allen and Etienne (2008) termed this the "two-epoch paradigm", which carries the unproven assumption of the global synchronism of respective glaciations. They concluded that available data permit numerous Neoproterozoic glaciations from 780 to 580 Ma, with Cryogenian glaciation evidently ending at ca. 635 Ma.

Cryogenian glacial successions attain great thicknesses and cover wide areas. In South Australia, deposits of the middle to late Cryogenian Sturt glaciation are >5 km thick (Preiss et al., 2011) and those of the younger, terminal Cryogenian Elatina glaciation cover 200,000 km^2 (G.E. Williams et al., 2008, 2011). Etienne et al. (2008, p. 343) found that Cryogenian glacial facies "are typical of sedimentary sequences deposited along glaciated continental margins throughout Earth history." Facies of the Elatina glaciation

are particularly varied, ranging from permafrost regolith and periglacial sand sheet, through glaciofluvial, tidal and deltaic sandstones with wave-generated ripple marks, to inner and outer marine-shelf siltstones and diamictites indicating the widespread rainout of ice-rafted debris and several glacial advances and retreats. Cryogenian glaciation occurred in low palaeolatitudes, first firmly established for the Elatina glaciation (Schmidt et al., 1991; Schmidt and Williams, 1995); this datum is supported by positive soft-sediment fold tests and corrected for minor inclination shallowing due to compaction (Schmidt et al., 2009), and remains a benchmark for Cryogenian near-palaeoequatorial (≤10°) glaciation. Subsequent studies have yielded low palaeolatitudes for other Cryogenian glacigenic deposits, with most results between 0–20° and no result >40° palaeolatitude (Evans and Raub, 2011).

Ediacaran (635–542 Ma) glaciation is recognised on most continents (Etienne et al., 2008; Gostin et al., 2010). The Gaskiers glaciation in Newfoundland is dated at ca. 580 Ma and the glacigenic Fauquier Formation in Virginia at ca. 571 Ma (Hebert et al., 2010), but without further radiometric ages their relation to glacial episodes elsewhere remains unclear. Ediacaran glaciation spanned wide palaeolatitudes: ice rafting and glacial deposition in Australia occurred at 10–15° (Gostin et al., 2010) and glaciation elsewhere ranged from ca. 50° to a favoured palaeolatitude of 78 ± 12° for the Fauquier Formation (Evans and Raub, 2011). The palaeolatitude of the Gaskiers glaciation is uncertain (Pisarevsky et al., 2012).

Exceptionally well preserved Cryogenian periglacial polygonal sand-wedges in West Africa (Deynoux, 1982) and South Australia (Williams and Tonkin, 1985; Williams et al., 2008) closely resemble sand wedges in Antarctica. They imply former mean annual

air temperatures of –12 to –20 °C or lower and a seasonal air temperature range of up to 40 C° (Washburn, 1980; Karte, 1983). Hence frigid, strongly seasonal Cryogenian climates existed near sea level in low palaeolatitudes. Wedges interpreted as periglacial occur in other Neoproterozoic and in Huronian glacigenic deposits (references in Williams, 1986).

9.3. Phanerozoic glaciations

Landing and MacGabhann (2010) suggested Early–Middle Cambrian glaciation of the Avalon microcontinent at a high palaeolatitude. Late Ordovician–Early Silurian glaciation from ca. 445–429 Ma, the world's oldest confirmed circum-polar glaciation, was brief but severe (Crowell, 1999). Ice sheets spread from the South Pole in NW Africa to adjoining South America as well as Europe and possibly North America, with grounded ice and periglacial wedges in South Africa at ca. 40° palaeolatitude (Daily and Cooper, 1976; Cocks and Torsvik, 2002). Late Devonian and Early Carboniferous–Late Permian glaciations mark the drift of Gondwana and Pangaea across the South Pole (Caputo and Crowell, 1985; Fielding et al., 2008). Late Palaeozoic ice rafting extended to ca. 32° palaeolatitude (Frakes and Francis, 1988), with Permian mountain glaciers in Colorado evidently descending to relatively low palaeoelevations on the palaeoequator (Soreghan et al., 2008). However, interpreted periglacial wedges near the palaeoequator in the Pennsylvanian–Permian of Colorado (Sweet and Soreghan, 2008) are unlike periglacial forms (see Black, 1983) and are better viewed as wedge-shaped seismites (e.g., Montenat et al., 2007).

Outsized exotic clasts in mudstones are widespread in the Early Cretaceous of Australia and are known from the mid-Jurassic to mid-Cretaceous in Russia, Alaska, Canada and Spitsbergen (Frakes and Francis, 1988; Frakes et al., 1995). These are interpreted as ice-rafted deposits that formed at palaeolatitudes >60°, implying the presence of high-latitude ice at sea level during the Mesozoic. Diamictite and associated lonestone facies in South Australia indicate Early Cretaceous glaciation at ca. 65° palaeolatitude (Alley and Frakes, 2003).

Ice-rafted debris in early Cenozoic sediments from both southern and northern polar regions indicate glaciers at sea level back to 45 Ma (Dawber and Tripati, 2011) and possibly 58 Ma in Spitsbergen (Frakes and Francis, 1988). Continental-scale glaciation was established in Antarctica by 34 Ma and ice sheets in northern high latitudes by ca. 23 Ma (Tripati et al., 2005; DeConto et al., 2008). The Pleistocene continental ice sheets in North America reached a maximum southern limit of ca. 38°N (Stiff and Hansel, 2004).

9.4. Controls and secular changes

As summarised by Bradley (2011), no consistent relationship is evident between times of glaciation and major tectonism. Palaeozoic ice sheets formed on Gondwana and Pangaea whereas Cenozoic ice sheets formed on dispersed fragments of Pangaea. Some Proterozoic glacigenic deposits are associated with continental rifts and separation (Eyles, 2008), but rifting and subsidence would enhance their preservation thus giving the impression that glaciation is favoured by extensional settings. Moreover, two contrasting global tectonic events in the latter history of the Earth — the 1.3–1.0 Ga Grenville

orogeny and the late Palaeozoic–early Mesozoic breakup of Pangaea — have no known related glaciation.

The outstanding features of the glacial record are the low palaeolatitudes of *pre-Ediacaran* glaciations and the switch to circum-polar glaciations of the Phanerozoic (Fig. 8; Williams, 1993, 2008), with several Palaeoproterozoic glaciations coinciding with drift across the palaeoequator and Palaeozoic glaciations with drift of Gondwana and Pangaea across polar regions. This switch spanned the Ediacaran–early Palaeozoic and the advent of the Metazoa. Palaeomagnetic data accord with the geomagnetic field approximating a geocentric axial dipole during the Proterozoic (Schmidt, 2001; Smirnov and Tarduno, 2004) and probably for all of geological time (McElhinny, 2004), hence palaeolatitudes may be viewed as indicators of geographic latitudes.

Although pre-Ediacaran high-palaeolatitude glacigenic deposits are unknown, the "snowball Earth" hypothesis postulates global glaciation during the Cryogenian with ca. 3 km thick synchronous ice-sheets on all continents and a shut-down of the hydrological cycle (Hoffman and Schrag, 2002; Domack and Hoffman, 2011). However, a large and diverse body of data conflicts with the requirements and predictions of this hypothesis (e.g., Kennedy et al., 2001; Allen and Etienne, 2008; Williams, 2008; Sansjofre et al., 2011). The Elatina glaciation, for example, was marked by extensive and long-lived open seas, unglaciated continental regions and an active hydrological cycle (Williams et al., 2008). Etienne et al. (2008) concluded that Cryogenian glacial facies require neither global synchronism nor necessarily severe climatic excursions.

Alternatively, an obliquity of the ecliptic >54° would render the equator cooler than the poles, on average, and amplify global seasonality, so offering an explanation of all

pre-Ediacaran low-palaeolatitude glaciations and related strong seasonality (Williams, 1993, 2008; Jenkins, 2003). Modelling indicates that a high obliquity is a likely outcome of the Moon-producing single giant impact at 4.5 Ga, but there is no established mechanism to reduce the obliquity during the Ediacaran–early Palaeozoic to accommodate Phanerozoic circum-polar glaciation. Pre-Ediacaran low-palaeolatitude glaciation thus remains a first-order geological and geophysical enigma.

10. Littoral-shelf (epeiric) sedimentation and secular change

Due to the complexity of early plate tectonics and a concomitant long history of deformation and reworking of craton margins, marine successions from Precambrian cratons are strongly skewed towards better preserved epeiric sea deposits on the cratons themselves rather than at their margins (e.g., Eriksson et al., 1998; 2004b) and few modern examples of epeiric seas exist (e.g., Pratt and Holmden, 2008). Clastic successions interpreted as epeiric seaways (essentially shelf-like with strongly directional currents) compare relatively favourably with modern shallow ocean and shelf settings while Precambrian epeiric embayments (lacked shelf breaks and ocean-type currents) were significantly smaller and shallower (Brenner, 1980; Eriksson et al., 2004b, 2008b).

Reduced shelf gradients of many Precambrian epeiric seas promoted predominant tidal energy at the expense of storms/waves (cf., Pratt and James, 1986), as also inferred from extensive peritidal flat deposits identified in the shoaling portions of Phanerozoic-aged epeiric basins (e.g., Swett et al., 1971; Reading, 1978; Pratt and James, 1986; Friedman et al., 1992). Precambrian clastic epeiric sea coastlines were typified by high energy braid-

delta systems (Els, 1998a; Eriksson et al., 2008b) and in wave-dominated coastal segments away from these largely tidally reworked braid-deltas, amalgamation of supralittoral storm beds may have been important (Sarkar et al., 2004). Deposition, erosion and even stratigraphic architecture within Precambrian marine realms are thought to have been heavily dependent on prolific and widespread microbial mat growth (which flourished in the absence of grazers and burrowers), both in siliciclastic and carbonate supralittoral-littoral-shelf settings (e.g., Schieber, 1998; Schopf, 1999; Hagadorn and Bottjer, 1999; Schieber et al., 2007; Noffke, 2010; Sarkar et al., 2008).

Precambrian clastic shelf (cf., shelf-like epeiric) deposits tend to be sandy and to closely mirror Phanerozoic-Modern equivalents (Jackson et al., 1990; Lindsey and Gaylord, 1992; Eriksson et al., 1998), with common hummocky and swaley cross-strata, parallel laminae, trough and planar cross-bedding (Soegaard and Eriksson, 1985; Bose et al., 1988; Jackson et al., 1990; Tirsgaard and Sønderholm, 1997; see, however, some significant exceptions detailed by Sarkar et al., 2005, 2008; Catuneanu and Eriksson, 2007). Inferred open shelf deposits are typically dominated by wave-imprinted nearshore sandstones and massive to laminated mudrocks (± subordinate storm–deposited graded siltstone/fine-grained sandstone beds) and exhibit remarkable preservation of small-scale primary features due to a lack of bioturbation (Schieber, 1989; Chakraborty and Bose, 1992; Sarkar et al., 2002). Very mature shoreface sandstones lack mudstone but show localised pebbles and pebble lags (Harris and Eriksson, 1990; Walker and Plint, 1992; Eriksson et al., 1998). Although literature on Precambrian barrier island, wash-over fan (Eriksson, 1979) and lagoonal deposits is restricted (e.g., Eriksson et al., 1998), Sarkar et al. (2008) have recorded tidal accentuation behind localized shoals. Case studies of

Precambrian epeiric basins are given by Bose et al. (1999, 2001), Chakraborty (2002), Simpson et al. (2004b), Sarkar et al. (2002, 2004, 2005) and Eriksson et al. (2008b).

11. Discussion

11.1. Secular change, sequence stratigraphy and actualism

The principles of sequence stratigraphy are the same irrespective of the age of the strata under analysis. It was generally assumed that conclusions drawn from the earlier study of the Phanerozoic record would be universally applicable to strata of all ages, following the principle of actualism. The more recent applications of sequence stratigraphy to the Precambrian record indicate, however, that the window of time provided by the Phanerozoic is insufficient for meaningful generalizations (e.g., Eriksson et al., 2004a; Catuneanu et al., 2005). Notwithstanding practical limitations imposed by the lesser preservation and poorer time control, the Precambrian stratigraphy enlarges significantly our window into the geological past and affords better insights into some of the first principles of sequence stratigraphy.

The principle of actualism as applied to sequence stratigraphy was placed under scrutiny in recent years (e.g., Eriksson et al., 2004a; Catuneanu et al., 2005, 2012). It is now accepted that both similarities and differences exist between the controls on sequence development and the patterns of sedimentation within Precambrian and Phanerozoic depositional settings.

In terms of similarities, the same set of allogenic controls on sedimentation and sequence development operated throughout geological time and lead to marked similarities with respect to preserved lithologies, sedimentary structures, and the association of sedimentary facies and depositional elements within Precambrian and Phanerozoic sedimentary basins (cf., Eriksson et al., 1998; Altermann and Corcoran, 2002). Contrasts between Precambrian and Phanerozoic sedimentary successions are caused by a variety of factors that changed in terms of nature, rates and intensities during the course of the Earth's history, as have been detailed in preceding sections of this paper.

The combined effect of the secular changes detailed above contributed to differences in the scale and architecture of Precambrian sequences when compared with Phanerozoic counterparts, including the relative contribution of various systems tracts to the makeup of a sequence (e.g., Sarkar et al., 2005; Catuneanu, 2007). These observations provide clues that help in the refining of critical aspects of the sequence stratigraphic method, from the development of a system of sequence hierarchy to understanding the shifts in sequence stratigraphic architecture through time.

In terms of sequence hierarchy, the critical aspect is that the rates and to some extent the nature of basin-forming mechanisms have changed during the Earth's evolution, from a greater measure of competing plume tectonics and plate tectonics in the Precambrian to a more stable plate-tectonic regime in the Phanerozoic (Eriksson and Catuneanu, 2004; Eriksson et al., 2005a, b). The more erratic tectonic regimes, in terms of nature and rates, which controlled the formation and evolution of sedimentary basins in the Precambrian indicate that time is largely irrelevant to the classification of stratigraphic sequences (Catuneanu et al., 2005). Instead, it is rather the record of changes in the tectonic setting

that provides the means for the subdivision of the stratigraphic record into basin-fill successions separated by first-order sequence boundaries. These first-order basin-fill successions are in turn subdivided into second- and lower-rank sequences that result from shifts in the balance between accommodation and sedimentation at various scales of observation, irrespective of the time span between two same-order consecutive events.

In terms of sequence architecture, shifts with time in environmental-energy conditions and in the role of microbial mats on sedimentation may explain observed changes in the relative development of systems tracts within sequences of different ages (Catuneanu, 2007). Notably, the Precambrian sequences lack well-developed transgressive systems tracts, and are dominated by stacked highstand systems tracts that may be separated by thin veneers of transgressive deposits, often reduced to transgressive lags (e.g., Banerjee and Jeevankumar, 2005; Sarkar et al., 2005). The poorer development of transgressive systems tracts in the Precambrian may be attributed to higher wind and wave energy conditions, promoting stronger wave-ravinement erosion during transgression, possibly related to a greater polarization of the Earth's thermal zones (Rautenbach, 2001) combined with enhanced global rotation rates (Williams, 2000, 2004) (section 3.1). The prevalence of highstand systems tracts may be attributed to the prolific growth of microbial mats below the fairweather wave-base in shelfal settings, particularly during the Proterozoic, which prevented the reworking of sediments by the organic binding of particles and promoted aggradation under normal regressive conditions (e.g., Sarkar et al., 2005).

11.2. Significant and irreversible change in the nature of the Earth system: Neoarchaean

Postulated effects of secular change in the Earth's mechanics on global tectonics may have been more extreme at two specific periods in the Earth's history: strong resonances of the Earth's fluid core with the semi-annual and annual forced nutations for the despinning Earth in the Neoarchaean and at ca. 530 Ma might explain the inferred higher flux of mantle plumes and tectonothermal activity at these times (Williams, 1994). It is even conceivable that the significant apparent change in crust-forming processes in the Neoarchaean was related to the semi-annual resonance. Comparably, zircon U-Pb age peaks support episodic rapid crustal growth rates, with the most prominent peak at 2.8-2.7 Ga, thus also suggesting a sharp and sudden turning-point in crust-forming processes in the Neoarchaean (section 3).

A complex interaction between plume and plate tectonic processes for much of the Proterozoic (Mints, this volume) can also be linked inherently to the supercontinent cycle (e.g., Condie et al., 2001; Condie, 2004 a, b; Zhong et al., 2007; Santosh et al., 2009). While some see the onset of this cycle as being of uncertain antiquity, Condie (2001) and Condie et al. (2009b) considered that increased production rates of juvenile crust, starting at ca. 2.7 Ga, correlate with the formation of supercontinents and superplume events. The ongoing and multi-faceted debate is expanded by postulated global-scale tectonic–thermal "slowdowns" from ca. 2.45–2.2 Ga, and at ca. 1.4 Ga (Condie, 1998; Condie et al., 2009a). It can also be hypothesized that the earlier slowdown event may even have separated an early form of plate regime interacting with a higher energy mantle system, from the onset of a more recognisable Phanerozoic-style plate tectonic regime from ca.

2.2–2.0 Ga. While there is a widespread belief in almost pervasive supercontinentality from the Neoarchaean (e.g. Aspler and Chiarenzelli, 1998; Bleeker, 2003), a more diachronous onset may have applied, with coexistence of supercontinents and non-amalgamated cratons (e.g. Eriksson et al., 2011a, b).

11.3. Significant and irreversible change in the nature of the Earth system: Palaeoproterozoic

Probably the most significant change in chemical conditions on the Earth was the Great Oxidation Event (GOE) (e.g., Bekker et al., 2004; Canfield, 2005; Holland, 2006; Guo et al., 2009; Bekker and Holland, 2012; Pufahl and Hiatt, 2012), the rise in atmospheric oxygen between 2.4 and 2.3 Ga which paralleled the global expansion of oxygenic photosynthesis in cyanobacteria (Cloud, 1973; Bekker et al., 2004; Canfield, 2005; Holland, 2006; Guo et al., 2009; Saito, 2009; Bekker and Holland, 2012; Pufahl and Hiatt, 2012). The preceding ca. 2470-2450 Ma interval was marked by intense plume activity terminating in a widespread LIP (large igneous province; Barley et al., 1997; Heaman, 1997) which led to a dramatic switch from reducing to oxidizing volcanic gases (Kump et al., 2001; Holland, 2002). The ca. 2450 Ma age is the same as that inferred for the onset of Condie et al.'s (2009a) global magmatic slowdown.

The onset of the GOE not only led to the disappearance of reduced detrital minerals (e.g., pyrite, uraninite) from the sedimentary rock record (Cloud, 1968; Fleet, 1998; England et al., 2002; Hazen et al., 2008), but also to the accumulation of oxygen-dependent lithofacies like Mn-oxides, phosphorites and glauconites, and red beds (Roy,

1997; Holland, 2002; Bekker et al., 2010; Nelson et al., 2010; Papineau, 2010; Pufahl, 2010; Pufahl and Hiatt, 2012); later, it also enabled the evolution and diversification of eukaryotes (Fig. 5; Knoll, 2011). The GOE thus irreversibly changed the nature of the Earth system and led to subsequent major changes in chemical sedimentation (Fig. 5). Archaean and Proterozoic carbonates were largely of an evaporative origin, with significant influences from microbial processes. Fe-rich dolomite gave way to iron-poor dolomites during the Neoarchaean-Mesoproterozoic period; a change accompanied by a gradual secular increase in limestone (e.g., Bose et al., 2012) (Fig. 7).

A broader range of palaeoclimatic zones is inferred from the Neoarchaean-Palaeoproterozoic transitional period, and warm and humid climates may well have predominated up till at least ca. 2.3 Ga, accompanied by reduction in acid-rain effects and decrease in greenhouse gasses, especially marked after ca. 2.2 Ga. However, evidence for Palaeoproterozoic glaciation is preserved on four continents, but the temporal relationship of the deposits (spread across more than 200 Myr, between ca. 2.45 and 2.2 Ga) remains unclear. Evidence for continent-marginal marine settings for glacial deposition is strong and generally, palaeolatitudes appear to have been low, coeval with host cratons or supercontinents like Kenorland moving across the palaeoequator (e.g., Williams and Schmidt, 1997; Schmidt and Williams, 1999; Bindeman et al., 2010).

Long's (2011) study of pre-vegetational fluvial systems indicates that they most probably closely resembled modern ephemeral and dryland systems with common sand-bed and gravel-bed braided systems; unconfined sheetflood deposits were abundant, with some sandy meandering systems but not fine-grained meandering and anabranching systems, and overbank deposits were rare. Documented lacustrine deposits become more

prevalent in the Palaeo- to Mesoproterozoic (e.g., Collinson and Terwindt, 1983; Eriksson, 1991; Martins-Neto, 1996). The development of advanced terrestrial vegetation systems in the Middle Devonian allowed extensive stable mud-rich overbank settings for the first time on the Earth. The lack of vegetation, and its presence and evolution during the Phanerozoic eon were thus both absolutely critical factors in the development of fluvial systems and their secular changes through the planet's history.

11.4. The "Boring Billion"?

Widespread euxinia (inferred to reflect bacterial reduction of oceanic SO_4^{2-}) by ca. 1.85 Ga is postulated to have ended iron formation and phosphorite deposition (Canfield, 1998; Poulton et al., 2004; 2010; Nelson et al., 2010; Pufahl and Hiatt, 2012) and ushered in the so-called "boring billion" year period which followed. During this time, anoxygenic photoautotrophs that tempered oxygen production by using H_2S as an electron acceptor (Johnston et al., 2009) are thought to have perpetuated euxinia and dramatically curtailed Eukaryotic evolution (Anbar and Knoll, 2002).

An earlier view of a nonglacial interval from 2200-800 Ma, approximately coincident with the "boring billion", was revised by documentation of ca. 1.8 Ga low latitude subglacial-fluvioglacial deposits in Australia, coincident with orogeny and global-scale continental assembly (Williams, 2005; Schmidt and Williams, 2008). On a greater geographical scale within the continental terranes, examples of Proterozoic and younger aeolianites become much more common than prior to the onset of the "boring billion", and may reflect different controls on formation and preservation than occurred prior to

ca. 1.8 Ga (Eriksson and Simpson, 1998; Simpson et al., 2004b) (section 8). Eriksson and Simpson (1998) have argued for an association with supercontinental breakup (and to a lesser extent assembly) to provide high accommodation tectonic settings for extensive desert deposit preservation, with the oldest deposits clustering around 1.8 – 2.0 Ga. Preservation of extensive desert facies probably also requires large land masses at suitable palaeolatitudes (ca. 30°), thus dependent on the scale of continental assemblies. It is also inferred that Precambrian aeolian deposits most probably extended across a wider range of palaeoclimatic conditions than today (e.g., Eriksson and Simpson, 1998; Simpson et al., 2004b).

11.5. Significant and irreversible change in the nature of the Earth system: Neoproterozoic and beyond (Phanerozoic)

Both iron formations and phosphorites reappear in the Earth's sedimentary record in the Neoproterozoic, associated with some Cryogenian glaciations postulated at 740-630 Ma (Klein, 2005; Bekker et al., 2010; Pufahl, 2010). Although two Cryogenian glaciations are recognised on most continents, global synchronism is unproven, and the postulate of numerous Cryogenian to Ediacaran glaciations from 780 to 580 Ma may well be a better interpretation of the Neoproterozoic glacial record (Allen and Etienne, 2008). No palaeomagnetic result >40° palaeolatitude (Evans and Raub, 2011) has been found. Ediacaran (635–542 Ma) glaciation, also recognised on most continents (Etienne et al., 2008; Gostin et al., 2010), again lacks clear evidence of synchronism, but is distinguished by a wide palaeolatitudinal spread (e.g., Gostin et al., 2010; Evans and Raub, 2011). Phanerozoic

glaciation marks the change to circum-polar glaciation, with a localized episode of relatively low-elevation equatorial mountain glaciation during the late Palaeozoic. While the postulate of an obliquity of the ecliptic >54° can account for all pre-Ediacaran low-palaeolatitude glaciation and related strong seasonality (Williams, 1993, 2008; Jenkins, 2003), supported by modelling indicating that a high obliquity is a likely outcome of the Moon-producing single giant impact at 4.5 Ga, no mechanism is known to reduce the obliquity during the Ediacaran–early Palaeozoic, in order to accommodate Phanerozoic circum-polar glaciation.

The Neoproterozoic Oxygenation Event at ca. 580 Ma (Fig. 5) spurred on the Ediacaran diversification of multicellular animals (Canfield et al., 2007; Och and Shields-Zhou, 2012), the succeeding Ediacaran-Cambrian transition, and the concomitant "substrate revolution" (Bottjer et al., 2000) which dramatically affected the physical and chemical nature of marine sediments. As another consequence, the dominance of microbial mats that characterised much of the preceding Precambrian was ended (Garrett, 1970; Hagadorn and Bottjer, 1997). Relatively widespread biomineralization by Eukaryotes at the Ediacaran-Cambrian radiation (Weiner and Dove, 2011) brought not only skeletonized taxa (Germs, 1972) but also the complex array of skeletal structures (Bengtson, 1992; Maloof et al., 2010) that strongly influenced shallow-water carbonate facies in the Ordovician (Pruss et al., 2010) and subsequent siliceous biomineralization (Fig. 5).

12. Conclusions

In all considerations of secular change of sedimentation systems, the fragmentary nature of the sedimentary record must be borne in mind. Almost all sedimentary processes are essentially ephemeral and stratigraphic gaps, distribution of layer thicknesses and sedimentation rates all have an overall fractal character. It was particularly changes in the rates and intensities of controlling factors during the course of the Earth's history, rather than the factors themselves, that were responsible for contrasts between Precambrian and Phanerozoic sedimentary successions.

The Neoarchaean marks a sudden and sharp turning-point in crust-forming processes, with change in the Earth's mechanics and concomitant influences on global tectonics related to inferred peaks in the flux of mantle plumes and in tectonothermal activity, and the onset of the supercontinent cycle starting at ca. 2.7 Ga. In the succeeding Palaeoproterozoic, the combined changes had become both substantial and irreversible. The end of the ca. 2470-2450 Ma interval, a period characterized by intense plume activity, saw a dramatic change from reducing to oxidizing volcanic gases, which brought on the most meaningful change in the Earth's chemical conditions, the Great Oxidation Event (GOE; 2.4-2.3 Ga). Its inception was marked by the disappearance from the sedimentary rock record of reduced detrital minerals and by the onset of the accumulation of oxygen-dependent lithofacies. A global-scale tectono-thermal "slowdown", postulated from ca. 2.45–2.2 Ga possibly marked the divide between an early plate regime interacting with a higher energy mantle system, and a Phanerozoic-style plate tectonic regime which became active from ca. 2.2–2.0 Ga. The "slowdown"

period was also marked by the first global-scale glacial deposits of predominantly shallow marine affinity and inferred low palaeolatitudes.

The lack of vegetation in the Precambrian, and its presence and evolution during the Phanerozoic, were a prime secular control on the nature of, particularly, fluvial systems and their preserved deposits during the Earth's history. Development of large ergs and widespread palaeodeserts globally from ca. 2.0 - 1.8 Ga may reflect different controls on formation and preservation than occurred prior to that time, and an association with the supercontinent cycle is postulated. Precambrian aeolian deposits probably extended across a wider range of palaeoclimatic conditions than their modern equivalents.

A second period of significant and irreversible change occurred in the Neoproterozoic; iron formations and phosphorites reappear in the sedimentary record and several Cryogenian glaciations are inferred at 740-630 Ma. Similar to the global-scale glaciation in the Palaeoproterozoic, global synchronism is not proven, and low palaeolatitudes are determined. The analogously widespread Ediacaran glaciation (635–542 Ma) is marked, in contrast, by a wide palaeolatitudinal range, while Phanerozoic glaciation is circum-polar. It is notable that the change from low latitude to circum-polar glaciation corresponds with the Ediacaran–early Palaeozoic period and the advent of the Metazoa. The Neoproterozoic Oxygenation Event (ca. 580 Ma) and concomitant Ediacaran diversification of multicellular animals were succeeded by the Ediacaran-Cambrian transition, and allied dramatic changes to the physical and chemical nature of marine sedimentation.

Major changes at the global-scale, such as the "Great Oxidation Event" or a widespread low palaeolatitude glaciation event, would have interacted with local-scale

conditions (such as regional tectonic setting, rate of change of relative sea level, palaeomicroclimate perhaps defined by local tectonic setting, unique biological character etc.) and as a result, the imposition of the global event on the nature of the local rock record might have been delayed. The preservation of such global events would thus plausibly have been diachronous at the local, especially outcrop scale. When taken together with the inherently fragmentary nature of the sedimentary rock record of the Earth as a whole, global synchroneity of major events across preserved successions on different cratons will be unlikely. This essentially diachronous character of the rock record when studied at the field-scale should thus not be taken as an indication of an event with broad temporal distribution, but rather as an inherent preservational character of a short-term global event within the sedimentary record as it is implicitly studied from field outcrops.

Acknowledgements

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Figure captions

Figure 1. The relationship between the duration of a sedimentary unit and its sedimentation rate, showing results from some 25,000 data sets; circled Roman numerals denote modes resulting from the most-used means of calculating sedimentation rates (after Sadler, 1981).

Figure 2. Rates and durations of sedimentary processes. Numerals refer to the Sedimentation Rate Scale (Miall, in press).

Figure 3. Change in the mean Earth–Moon distance with time, as suggested by different average rates of tidal energy dissipation. Curve a: present rate of lunar recession of 3.82 cm/yr (Dickey et al., 1994). Curve b: mean rate of lunar recession of 3.16 cm/yr since 500 Ma suggested by palaeontological data (Lambeck, 1980). Curve c: mean rate of lunar recession of 2.17 ± 0.31 cm/yr since the late Cryogenian indicated by the Elatina datum. Curve d: mean rate of lunar recession consistent with the Elatina datum and data for rhythmites from the 2.45 Ga Weeli Wolli Formation. Shaded area shows the error based on the Weeli Wolli datum. Modified from Williams (2000).

Figure 4. Age histograms for the crust and mantle based on U–Pb zircon dates and Re–Os model ages, respectively (modified after Lee et al., 2011). The U–Pb histogram is modified from Hawkesworth and Kemp (2006). Shaded areas refer to clusters of Re–Os model ages (Carlson et al., 2005). T_{MA} = model age.

Figure 5. Factors affecting the temporal abundance of chemical sedimentary rocks. The most significant change in chemical sedimentation occurred during the Great Oxidation Event (orange

bar). OC = hypothesized ocean chemistry varying from ferrous (red) to sulphidic (purple) to oxygenated (blue). P_2O_5 = phosphorites (yellow). Fe = iron formations in Precambrian and ironstones in Phanerozoic (red). Ironstones are fundamentally different from iron formations, having accumulated within an oxygenated ocean. Eva = evaporites (green). CO_3 = dominant calcium carbonate producers including benthic (B) and planktonic (P) sources. SiO_2 = dominant silica sources including sponges and radiolarians (S&R) and diatoms (D). Atmospheric oxygen proxies used to evaluate minimum (dark blue) and maximum (light blue) oxygen concentrations include volcanic outgassing, detrital pyrite occurrence, the distribution of red beds and palaeosols, and black shales. Modified from compilations by Holland (2006), James et al. (2010), Hohmann-Marriott and Blankenship (2011), Pufahl and Hiatt (2012).

Figure 6. (A) Polished slab of small domal stromatolites in the Palaeoproterozoic Kona Dolomite, Chocolay Group, Michigan. (B) White dolomite pseudomorphs after gypsum that show the original dolomite formed in association with evaporative conditions, Palaeoproterozoic Kona Dolomite, Michigan. Scale bars are 2 cm long. Samples provided by Dan Damrow.

Figure 7. The relative abundance of non-siliciclastic sediments through time, showing dolomites to be more abundant than limestones during much of the Proterozoic, when microbial ecosystems dominated the biosphere. Modified after Ronov (1964).

Figure 8. Palaeolatitudinal extent of glacigenic deposits. (1–2) Palaeoproterozoic, (3) Cryogenian, (4) Ediacaran, (5) Cambrian (see text for sources of data). (6) Ice-rafted deposits for the Late Ordovician–Early Silurian, (7) Late Devonian, (8) Early Carboniferous–Late Permian, (9–11) Early Jurassic–Late Cretaceous, and (12) Cenozoic (after Frakes and Francis, 1988). Note change of time-scale at 500 Ma.
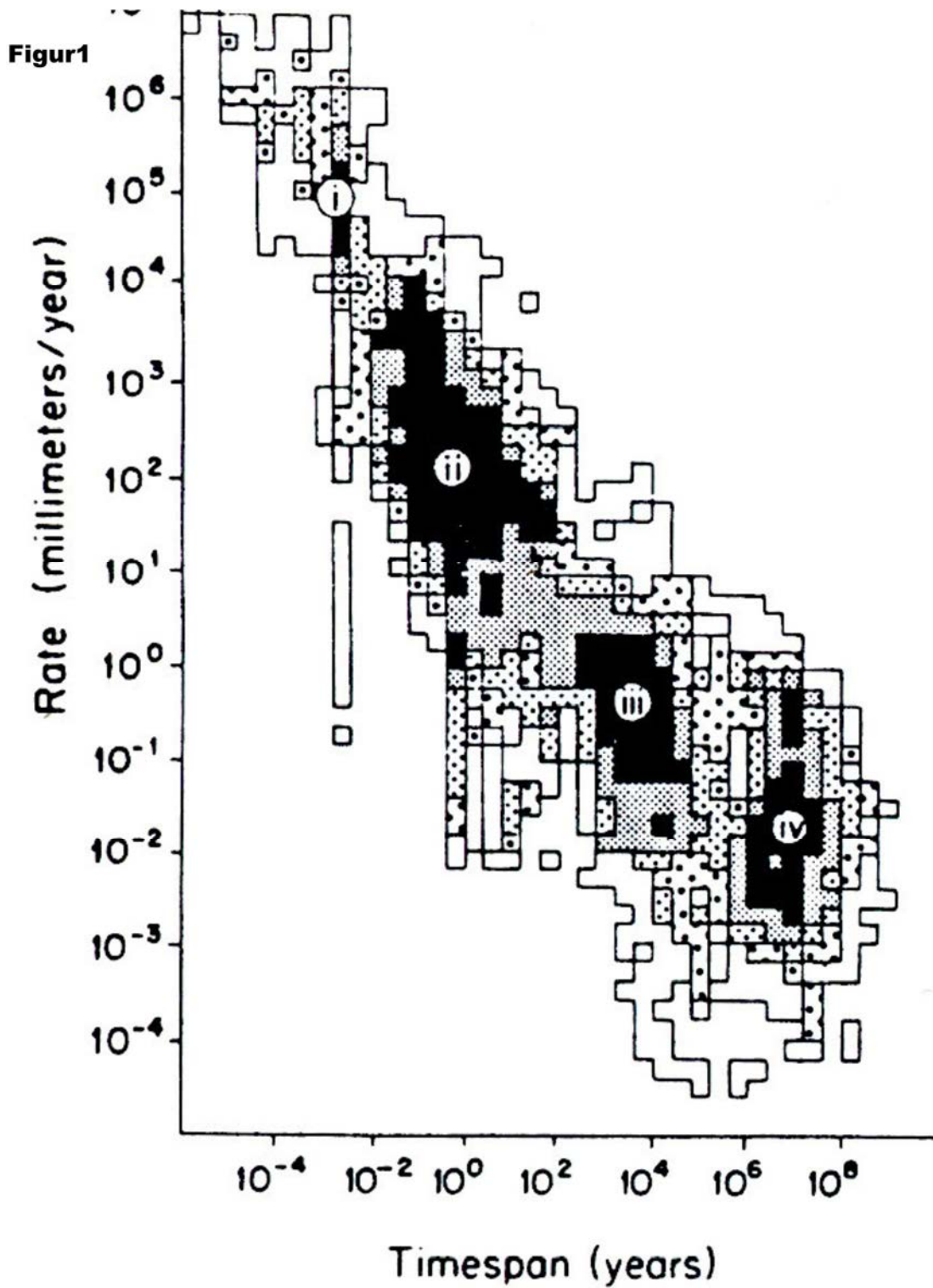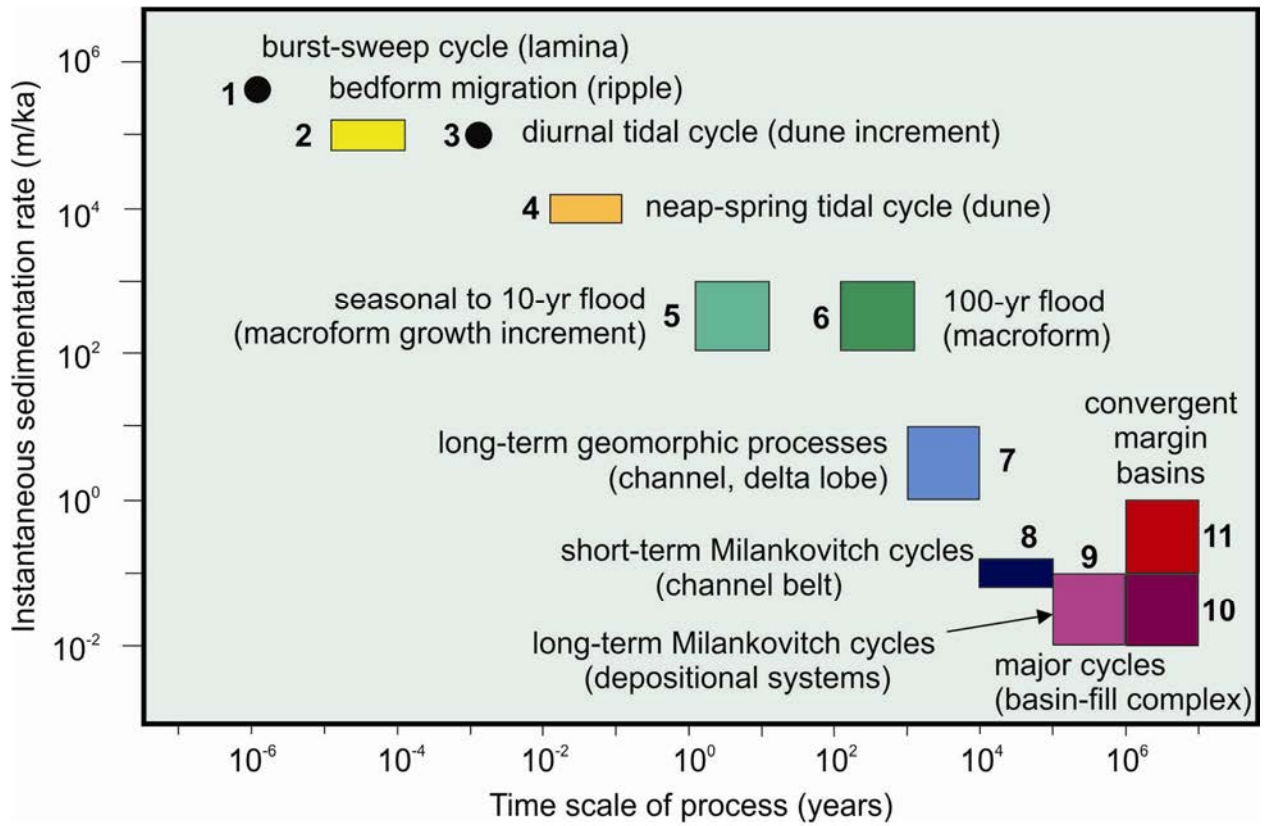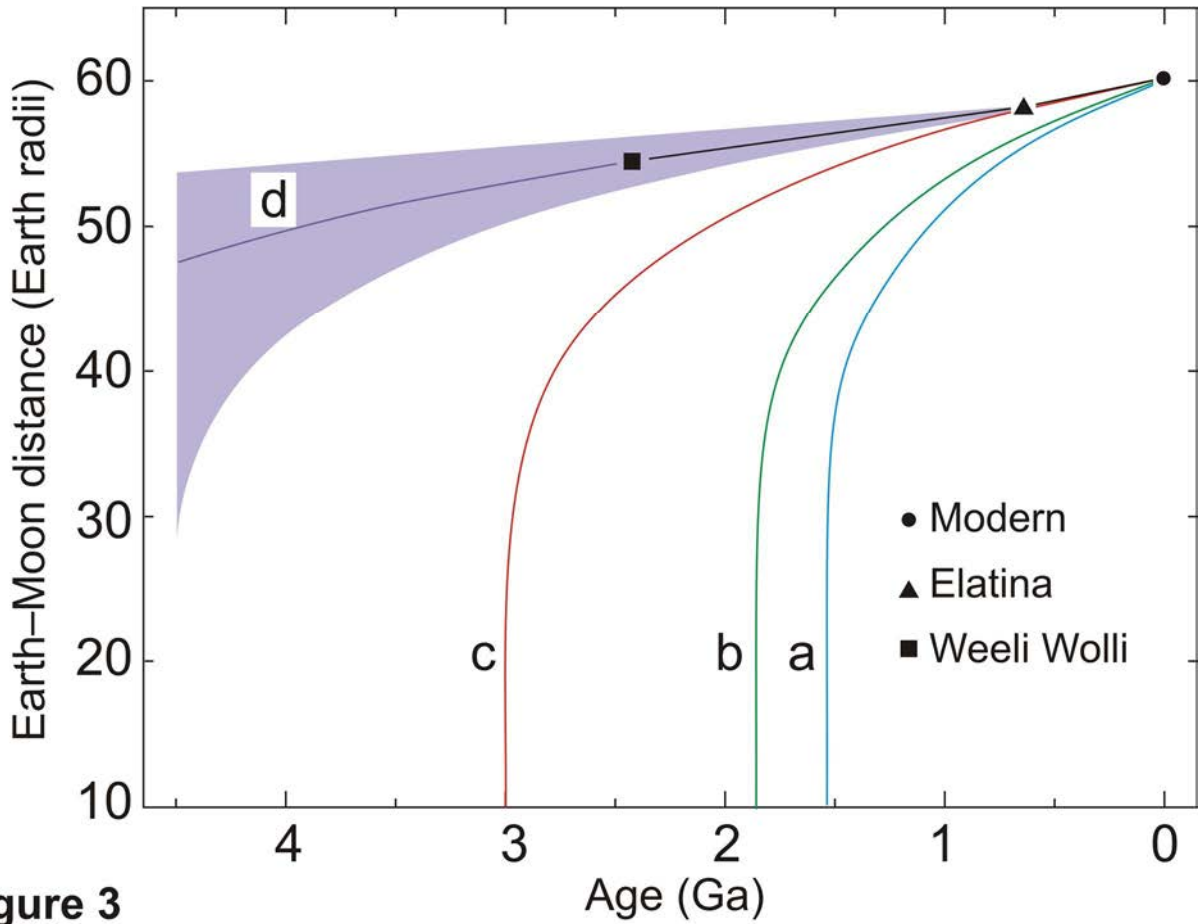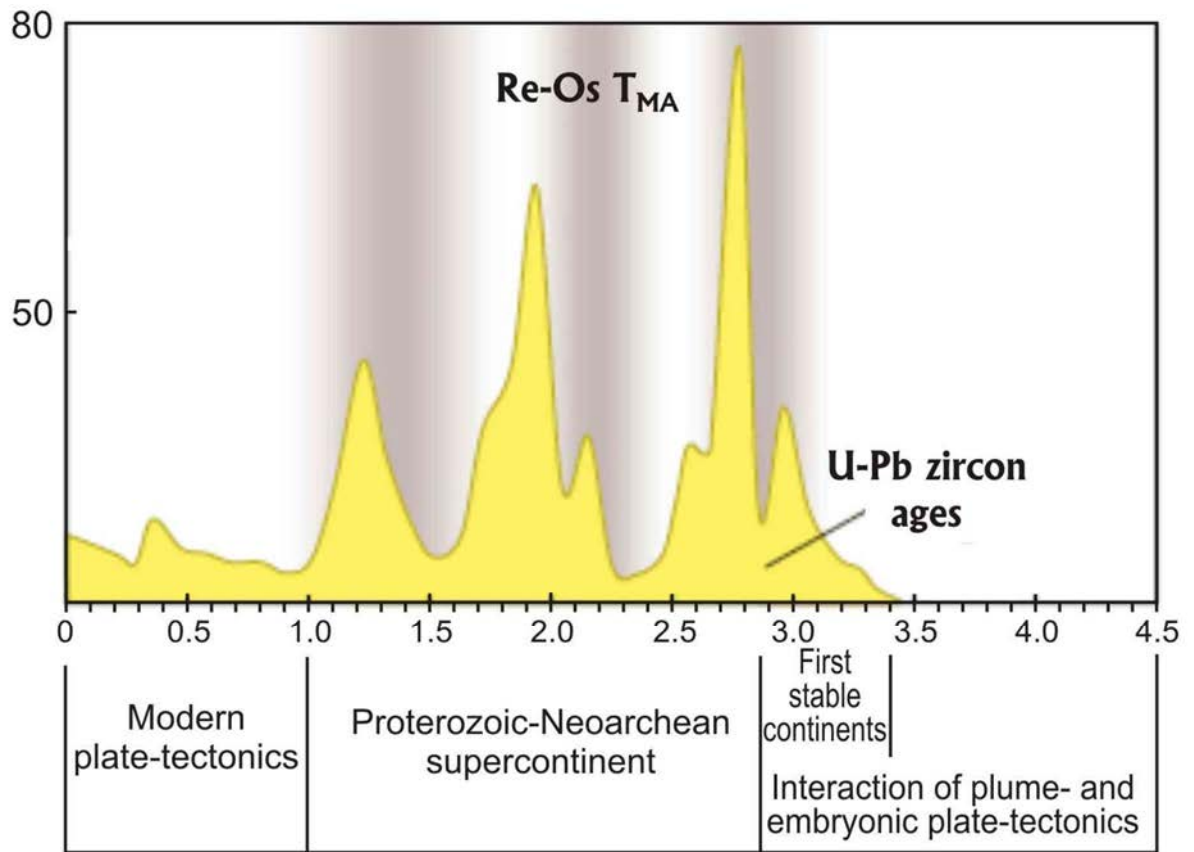
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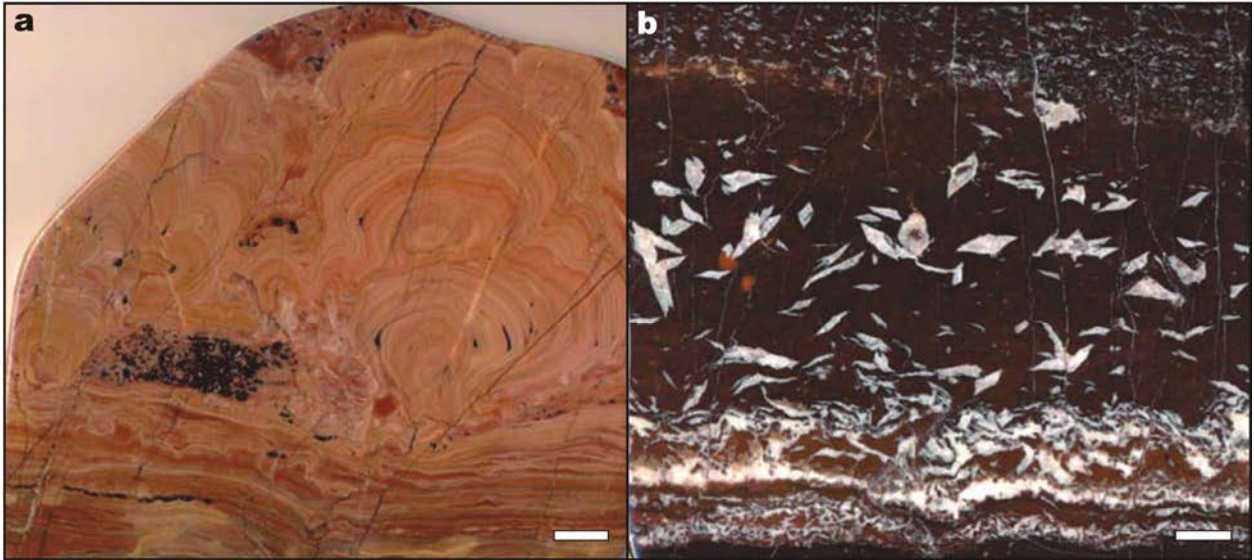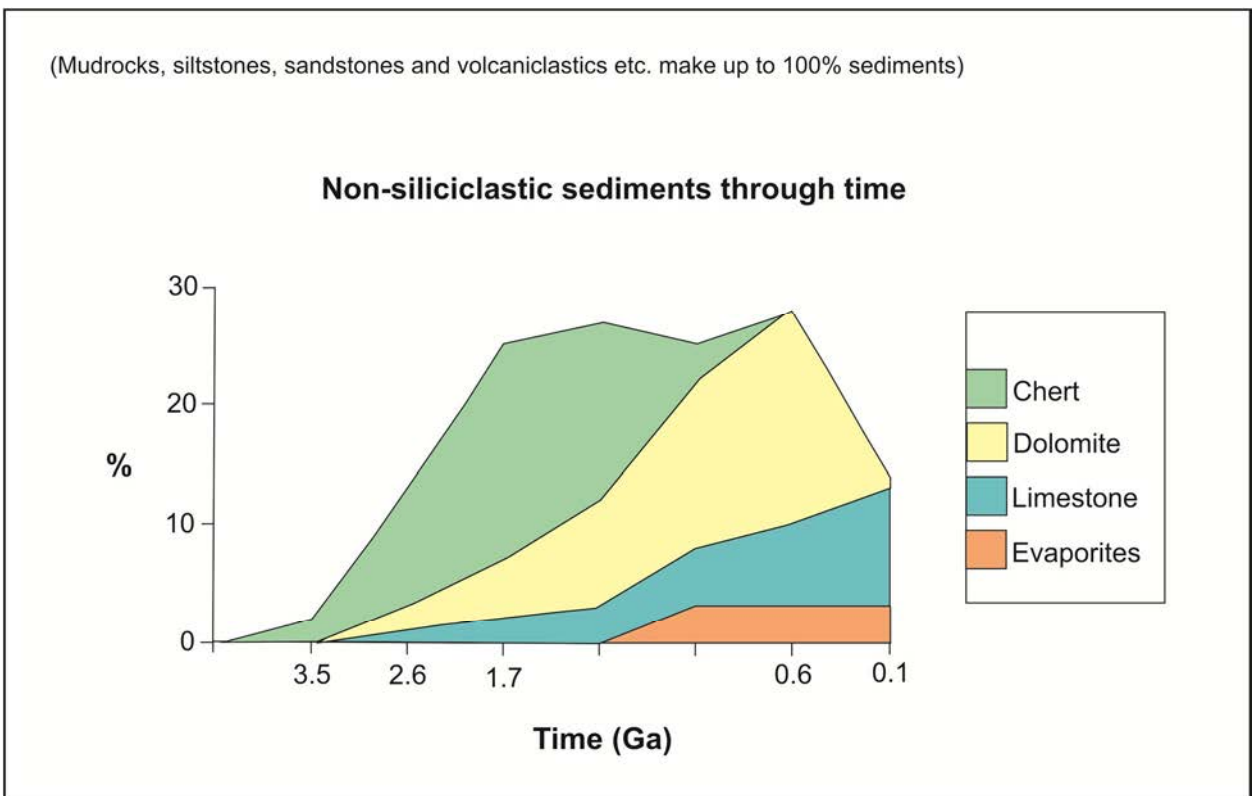**Figure 4**
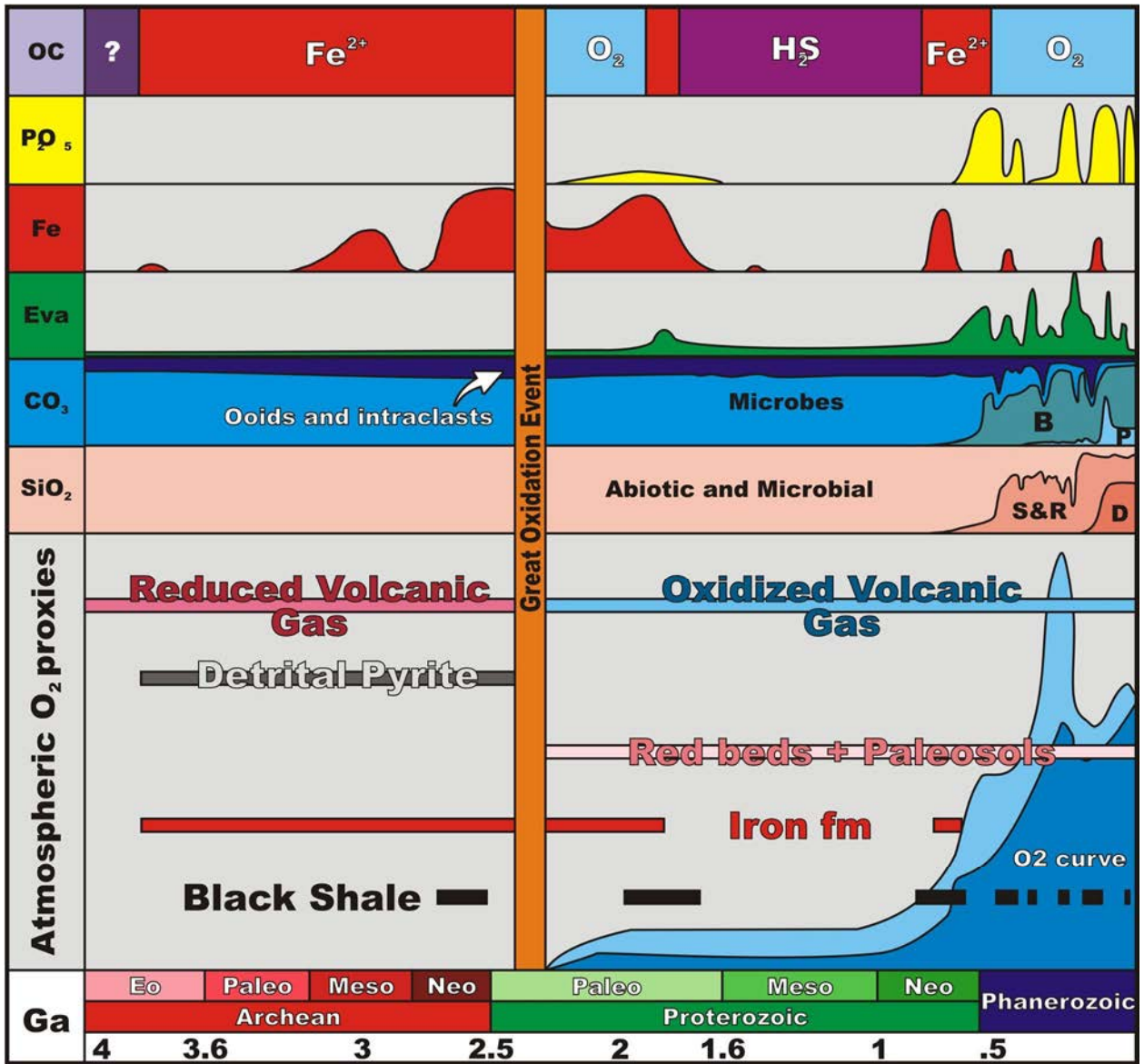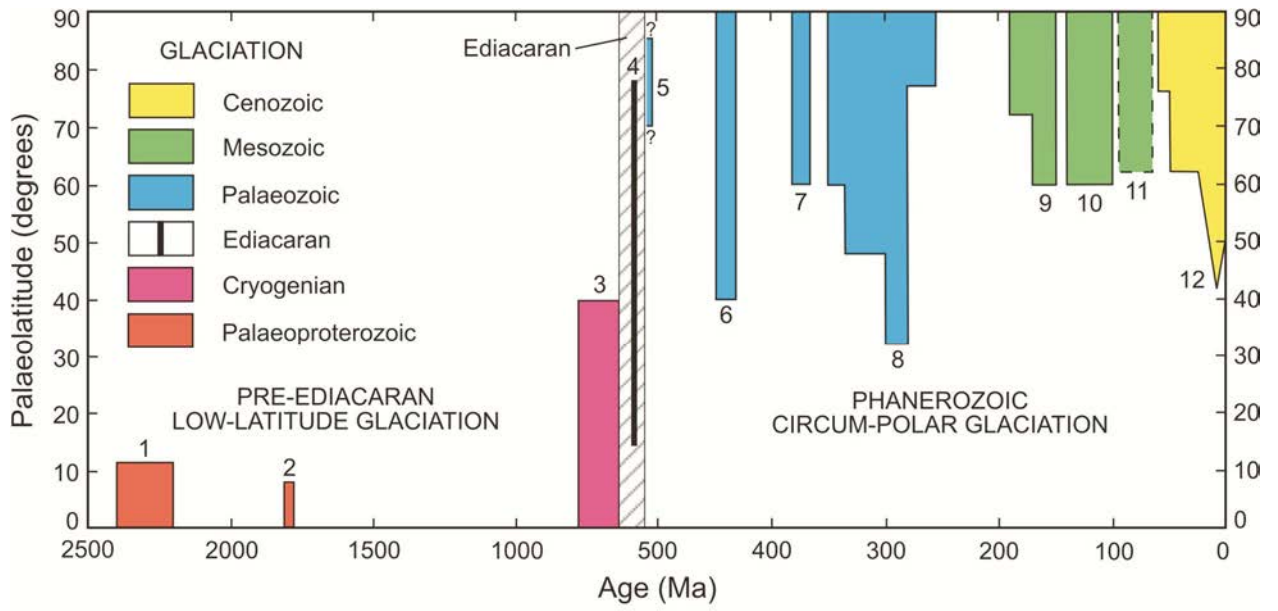
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