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Research paper

Controls on Precambrian sea level change and sedimentary cyclicity

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Abstract

Although uniformitarianism applies in a general sense to the controls on relative and global sea level change, some influences thereon were more prominent in the Precambrian. Short-term base level change due to waves and tides may have been enhanced due to possibly more uniform circulation systems on wide, low gradient Precambrian shelves. The lack of evidence for global glacial events in the Precambrian record implies that intraplate stresses and cyclic changes to Earth's geoid were more likely explanations for third-order sea level change than glacio-eustasy. Higher heat flow in the earlier Precambrian may have led to more rapid tectonic plate formation, transport and destruction, along with an increased role for hot spots, aseismic ridges and mantle plumes (superplumes), all of which may have influenced cyclic sedimentation within the ocean basins. A weak cyclicity in the occurrence of plume events has an approximate duration comparable to that of first-order (supercontinental cycle) sea level change. Second-order cyclicity in the Precambrian largely reflects the influences of thermal epeirogeny, changes to mid-ocean ridge volume as well as to ridge growth and decay rates, and cratonic marginal downwarping concomitant with either sediment loading or extensional tectonism. Third-order cycles of sea level change in the Precambrian also reflected cyclic loading/unloading within flexural foreland basin settings, and filling/deflation of magma chambers associated with island arc evolution.

The relatively limited number of studies of Precambrian sequence stratigraphy allows some preliminary conclusions to be drawn on duration of the first three orders of cyclicity. Archaean greenstone basins appear to have had first- and second-order cycle durations analogous to Phanerozoic equivalents, supporting steady state tectonics throughout Earth history. In direct contrast, however, preserved basin-fills from Neoarchaean–Palaeoproterozoic cratonic terranes have first- and second-order cycles of considerably longer duration than Phanerozoic examples, supporting less evolved tectonism affecting cratonic plates. It is possible that oceanic tectonic realms underwent more rapid and dynamic plate movements and arc generation, whereas early continental cratonic plates offered more stable platforms and may have been subject to slower migration rates. The wide range of controls on Precambrian sea level change, allied to their apparent variability (in rates and periodicity) through

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Precambrian time supports the conclusion that each order of cyclicity is relative and must be defined within the stratigraphic context of each individual case study. This underlines the importance of establishing a hierarchical order of cyclicity in sequence stratigraphic interpretations of Precambrian basins based on the relative importance of sequences rather than their temporal duration.

Keywords: Precambrian; Sequence stratigraphy; First- to third-order depositional cyclicity; Cycle duration; Relative sea level change; Eustasy

1. Introduction

Sea level change, one of the most important variables in sedimentological and sequence stratigraphic studies, forms the theme of this special issue dedicated to Pradip K. Bose, one of the prominent scientists who has studied this topic for most of a lifetime. This paper aims to review briefly the controls on sea level change during the Precambrian Era, and to address major influences on higher-order depositional cycles. We here use "higher-order" to denote more important cycles (cf. sequences) which are of larger scale and lower frequency, as opposed to "lower-order" cycles of greater frequency. Although the principle of Lyell (1830) uniformitarianism can generally be applied to the Precambrian sedimentary record (e.g., Donaldson et al., 2002; Eriksson et al., this volume), certain influences on Precambrian eustasy require either qualification or further discussion, which will be attempted here. Although not of great significance in Phanerozoic sequence stratigraphic studies, factors like Earth surface and mantle heat flow, continental crustal growth rates, global rotation rates and celestial mechanics, palaeo-atmospheric composition (and its changes over time), the "faint young Sun" and the concomitant inferred global greenhouse, mantle superplumes and superplume events (SPEs) and divergent views on early Precambrian plate tectonic regimes loom large in Precambrian examples (e.g., Eriksson et al., 2004a and references therein).

The concepts of sequence stratigraphy (see Miall, 1997, for a balanced commentary) owe much of their origin to the early ideas of Sloss (e.g., 1963), which were refined through the concepts of seismic stratigraphy by the "Exxon school" led by Vail (e.g., Vail et al., 1977), to result in the well known seismic stratigraphic volume edited by Payton (1977). The hierarchy of depositional cycles identified by many workers in many areas, both prior to and after this seminal 1977 publication, was ascribed by this school to predominant eustasy, against a background of relatively constant tectonic subsidence along an extensional continental margin (Miall, 1997; Miall and Miall, 2001, 2002).

Depositional cycles were classified into 1st-, 2nd-, 3rd-, 4th- and 5th-order cycles (Sloss, 1963; Vail et al., 1977; Miall, 1990; Duval et al., 1992). First-order (300–225 My in the Vail et al., 1977 model; generally 500–200 My in duration) cycles are ascribed by most workers to changes in ocean basin volume consequent upon the plate tectonic or supercontinent cycle (e.g., Dewey and Burke, 1974; Pitman, 1978; Worsley et al., 1984; Gurnis, 1988; Nance et al., 1988; Hoffman, 1989) and can thus be identified in the Phanerozoic record on formerly adjacent continents. In the Precambrian, there is evidence to support a first supercontinent ("Kenorland") at c. 2.7 Ga (e.g., Condie, 2004), and the Wilson-cycle of breakup–ocean growth–reassembly appears to have been more protracted than Phanerozoic equivalents (Aspler and Chiarenzelli, 1998; Aspler et al., 2001). The widely favoured concept of more rapid plate movements in the earlier Precambrian (e.g., Hargraves, 1986) stands in contrast to this, leading some workers to emphasise variable rather than universally more rapid rates of plate movement for these times (e.g., Catuneanu, 2001; Eriksson and Catuneanu, 2004a). Duration of these first-order cycles may thus have reached c. 300–650 My in the Neoarchaean–Palaeoproterozoic (Aspler and Chiarenzelli, 1998; Catuneanu and Eriksson, 1999), longer than the duration range found for the Phanerozoic rock record (e.g., Miall, 1997, his Table 3.1). Other workers (e.g., Worsley et al., 1984; Krapez, 1993) favour an essentially uniform supercontinental cycle length of c. 330–440 My.

Second-order (10–80 My, Vail et al., 1977; 10–100 My, Miall, 1997) cycles of sea level change reflect major phases in sedimentary basin evolution, consequent upon changes in mid-ocean ridge (MOR) spreading rates and their growth/destruction, extensional downwarping due to crustal flexure and crustal loading, and due to mantle-related thermal processes; they are correlatable across two or three (formerly contiguous) tectonic plates in the Phanerozoic (Hallam, 1963; Pitman, 1978; Krapez, 1993; Emery and Myers, 1996; Miall, 1997). Within the c. 2.7–2.1 Ga Transvaal basin (Kaapvaal craton, South Africa), five second-order sequences have been identified, with an average duration of c. 130 My; less evolved and thus, at least in the case of Kaapvaal, slower plate tectonic regimes are inferred (Catuneanu and Eriksson, 1999). Again, there is an indication that Precambrian (2nd-order) cyclicity may have had differing duration ranges (discussed in more detail later in this paper). On most Archaean cratons, basin formation and sediment deposition occurred within well-delineated regions and during time intervals that were separated by lengthy periods of tectonic quiescence characterized by intermittent deposition, non-deposition or peneplanation (e.g., Pilbara between 3.1 and 2.8 Ga, and Kaapvaal at c. 3.0 Ga). The evolution of many, but not all, Archaean sedimentary basins can be interpreted in terms of modern-day plate tectonic processes, such as orogeny related to plate collision. Sedimentary depositories associated with Archaean greenstone belts may be related to plate rifting or subduction processes, but do not appear to have exact modern-day tectonic analogues. Nevertheless, most greenstone-related depositories were active for intervals of between 10 and 100 My, and so may be considered as analogous to 2nd-order cycles, particularly those strongly influenced by MORs and destructive oceanic plate margins. First-order, supercontinental cyclicity cannot be applied readily to greenstone belts as these belts are inherently part of craton formation.

Third-order (1–10 My) cycles often have preserved thicknesses which are well resolved on seismic logs, and thus form an important cornerstone of “classical (cf. Exxon-school)” sequence stratigraphy (e.g., Emery and Myers, 1996). Vail et al. (1977, 1991) ascribe these cycles to glacio-eustasy; however, a major problem with this interpretation is the uncommon occurrence of glaciation within the known Precambrian record. Additionally, for much of the Phanerozoic, there was no low altitude glaciation, yet third-order (and higher) sea level cyclicity persisted (e.g., Reading and Levell, 1996). Other, tectonic mechanisms are preferred by many other workers as an alternative to glacio-eustasy: e.g., intraplate stresses leading to large scale flexure of tectonic plates (Miall, 1984, 1991, 1997; Cloetingh et al., 1985; Cloetingh, 1988a).

Fourth- and fifth-order (0.1–0.5 My) cycles (cf. parasequence cycles) probably reflect, at least partially, the influence of autocyclic processes within the sedimentary systems (e.g., Vail et al., 1977; Emery and Myers, 1996). The duration of these (and higher frequency) cycles overlaps with those of the Milankovitch cycles: eccentricity of the annual solar orbit (0.1–0.4 My); cyclicity in the tilt of the Earth’s axis (c. 41,000 years); the wobble on Earth’s rotational axis (c. 23,000 years) (e.g., Reading, 1996). Miall (1997), however, stresses that the lengths of these cycles may have varied significantly over the geological past. Implicit in the Milankovitch effects are changing distributions of plates, water, ice and gravity forces, and these can deform Earth’s geoid, thereby changing sea level by 50–250 m over time periods of c. 1 My (Mörner, 1994), analogous to third-order glacio-isostatic cyclic rates (Eriksson, 1999). Thermally induced geoid highs (cf. mantle plumes, Condie, 1998, 2001, 2004) associated with supercontinental breakup and analogous geoid lows where daughter plates reassemble can also result in third-order cylicity (e.g., Miall, 1997).

The resolution of geochronology required to identify lower-order cyclicity cannot be achieved within the Precambrian rock record, but interpretations of cyclicity up to the third-order are possible based upon detailed facies and architectural studies (e.g., Catuneanu, 2002; Embry et al., 2004). Basing a hierarchy of cyclicity essentially on sequence boundary frequency will not be applicable to Earth’s entire rock record, where, especially for the Precambrian, changes in the dynamics of plate tectonics are inferred; instead, Catuneanu and Eriksson (2004) stress the importance of treating each case separately. As will be seen below, different authors assign the same unconformity-bound units within certain early Precambrian basins to different orders of cyclicity; this problem is especially found in assigning second- and third-order ranks to cycles. This paper will thus focus on possible controls on first- to third-order depositional cyclicity within the Neoarchaean–Palaeoproterozoic period.

2. Controls on sea level change, and unique Precambrian variables

Eustatic sea level changes are measured relative to the fixed datum of the centre of the Earth; they reflect variation in either ocean water or ocean basin volume, changes to the hypsometric curve (which summarises areal distribution of global elevations below and above mean sea level) (Fig. 1) or variation of Earth's geoid (an equipotential surface of the globe's gravitational field corresponding to mean sea level) (Schopf, 1980; Jervey, 1988; Mörner, 1994; Reading and Levell, 1996). In contrast to the global nature of eustasy, relative sea level change, reflecting the interplay of sea level change and tectonism, as well as the effect of sedimentation and sediment compaction, is more localised and is difficult to differentiate in the rock record from eustatic sea level change (Hubbard, 1988; Posamentier et al., 1988; Posamentier and James, 1993; Emery and Myers, 1996).

2.1. *Short-term and localised changes in sea level: waves, tides, tsunamis, slumps, salinity, temperature and isolated dessicated basins*

Short-term and localised changes in sea level occur due to waves, tides, storms and storm surges, as well as hurricanes, tsunamis and catastrophic sediment slumps; effects on relative sea level will be up to about 20–27 m on open coastlines (Reading and Levell, 1996). Relative sea level fluctuations up to about 10 m occur due to changes in sea water density, resulting from salinity and temperature (themselves mutually dependent): increasing salinity and decreasing temperature lower sea level (Donovan and Jones, 1979; Gross, 1990). In the early Precambrian, when wide and gently sloping shelves were possibly characterised by more uniform circulation systems (Eriksson et al., 1998, 2004b, this volume), waves and tides may have played a more important role in short-term and localised sea level change. The early Precambrian greenhouse palaeo-atmosphere, inferred to have compensated for the "faint young Sun",

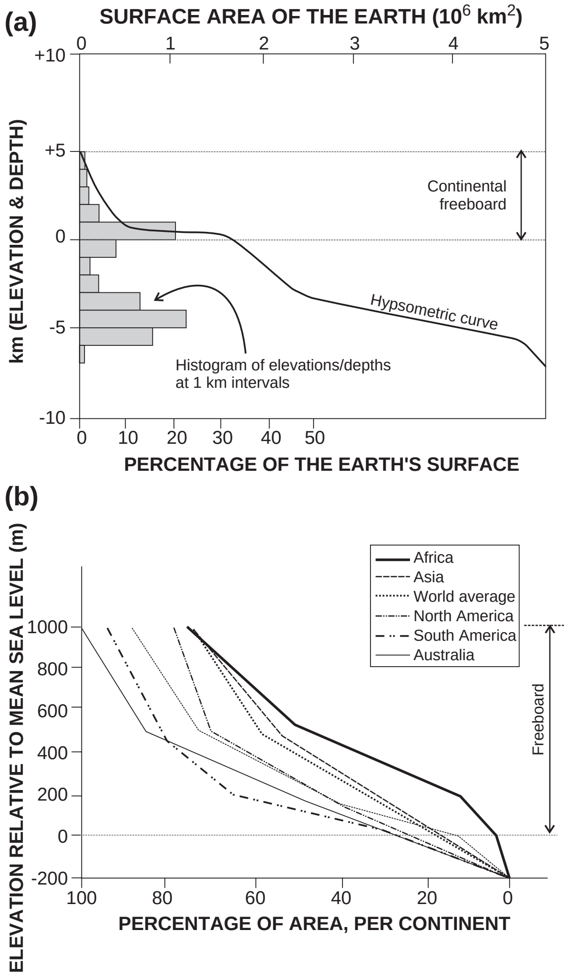


Fig. 1. (a) The hypsometric curve, encompassing the cumulative percentage of the Earth's solid surface at different elevations above/depths below mean sea level, is constructed from a histogram of such elevations/depths (with a 1 km interval in this example). (b) Examples of hypsometric curves for the six present-day continents, assuming the edge of the continental shelf to be at a depth of 200 m, a maximum land elevation of 1000 m above mean sea level, and normalising this interval to 100%. Note also the continental freeboard, defined as the relative elevation of a continent with respect to mean sea level (cf. Wise, 1972, 1974). Modified after Schopf (1980) and Eriksson (1999).

combined with mantle heat flow values 2–3 times those at present (e.g., Eriksson et al., 2004a) would have led to higher oceanic temperatures: possibly up to 100 °C at 3.8 Ga, 70 °C at 3.4 Ga and 22 °C at 2.0 Ga (compared to average surface temperature today of 17.5 °C; Hoyle, 1972; Perry et al., 1978; Knauth and Lowe, 1978; Gross, 1990). However, enhanced evaporation and concomitant increased ocean water salinity would have offset these higher water temperatures (Eriksson, 1999). Longer term eustatic fluctuations up to 10–15 m due to flooding and desiccation of restricted ocean basins (e.g., the Mediterranean; Reading and Levell, 1996), although possible in the Precambrian, are not easily amenable to discrimination in the rock record. However, Christie-Blick et al. (1990) interpret the Wonoka canyons at the base of the Neoproterozoic Wonoka Formation (Flinders Ranges, Adelaide geosyncline, Australia) as reflecting Messinian-type evaporative drawdown.

2.2. Glacio-eustasy and the Precambrian glacial record; intra-plate stresses and Milankovitch effects on Earth's geoid

Glacio-eustasy, responsible for third-order cyclicity within the Vail et al. (1977, 1991) school of thought, affects sea levels by the combination of water removal (or addition upon melting of ice) that is enhanced by increased ocean water density (decreases upon melting). The resultant fluctuations will be reduced by the combined effects of isostatic loading of continents with ice (reduces relative sea level drop, or even induces a relative rise close to the centre of loading) and hydroisostatic unloading (loading upon melting of the ice) of shelves with reduced water depths, with an overall maximum eustatic change of about 150 m (Reading and Levell, 1996). Rapid isostatic rebound of continents (up to about 250 m over <10 ka) upon deglaciation will result in relative sea level drop locally easily outpacing eustatic rise (Reading and Levell, 1996). However, application of glacio-eustasy as an explanation for third-order depositional cycles remains problematic, due to the relative lack of glacial epochs in the Precambrian (e.g., Hambrey and Harland, 1981) and large parts of the Phanerozoic also lack such deposits (e.g., Miller, 1996). For the Precambrian, apart from the well known three Neoproterozoic glacial events (e.g., Young, 1995), glacial periods appear very limited in their occurrence, and their genesis remains subject to debate (e.g., Young, 1991, 2002, 2004). Only one early Precambrian global glacial event is known, that from c. 2.4–2.2 Ga (e.g., Young, 1970, 2004; Hambrey and Harland, 1985; Eyles and Young, 1994; Young et al., 2001; Williams, 2004). Only short-lived and very localised (possibly mountain ice sheets) glaciation is inferred for the Archaean (Page, 1981; Von Brunn and Gold, 1993; Young et al., 1998), and for the almost 1500 My period separating global Palaeoproterozoic and Neoproterozoic events, only one unequivocal glacial deposit has been identified, at about 1.8 Ga in Australia (Williams, 2005).

As third-order and higher cyclicity is well known from almost the entire rock record (Precambrian examples will be discussed below), particularly for the Mesozoic and younger parts thereof (Reading and Levell, 1996), other models are also proposed for these cycles. Cloetingh (1986, 1988a,b) and Cloetingh et al. (1985) infer large scale flexure of tectonic plates due to intraplate stresses, resulting in sea level variation up to about 100 m over c. 10 My periods at the flanks of passive margins. They note that these changes in (relative) sea level can be correlated within neighbouring basins and the scale is thus cratonic or even supercontinental rather than global (eustatic). It should be noted that the action of intraplate stresses is not restricted to passive margins, but also modifies vertical movements within intracratonic basins, foreland basins and flexural moats flanking intra-oceanic volcanic complexes (Cloetingh, 1988b). Moving distributions of tectonic plates, ice, water and gravity forces are implicit within the Milankovitch effects, and can thus result in cyclical sea level changes up to 50–250 m over c. 1 My periods, due to changes in Earth's geoid (Mörner, 1976, 1994).

2.3. Eustasy and changing ocean basin volume; tectonic and magmatic–thermal influences; crustal growth models and the constant freeboard model

Significant eustatic changes also result from changes to ocean basin volume, reflecting essentially plate tectonic–magmatic thermal genesis: enhanced mid-ocean ridge (MOR) spreading rates, creation of new ridges and subduction of old ridges raise sea levels, whereas supercontinent assembly will lower them (Donovan and Jones, 1979; Holser, 1984; Harrison et al., 1981; Pitman and Golovchenko, 1983; Windley, 1995). These processes can affect sea level by up to 350 m over periods of c. 70 My or less (Reading and Levell, 1996). Sea level drop during supercontinent collisions will be mitigated by increased sedimentation along continental margins, this in turn being decreased by isostatic depression of those shelfal margins (Reading and Levell, 1996). Supercontinental rifting, attenuation and dispersal, analogously, raise eustatic sea level and favour carbonate platform development, which further raises water levels (again, subject to hydroisostatic and isostatic depressive amelioration) (Eriksson, 1999).

Significant changes to ocean volume also result from intra-plate hot spots, aseismic ridges (subsiding volcanic island chains) and oceanic plateaus (Reading, 1978; Schlanger et al., 1981; Sykes and Kidd, 1994); presently, plateaus cover 10% of ocean floors and aseismic ridges c. 25% and maximum eustatic fluctuations are estimated at c. 100 m (Reading and Levell, 1996) over time scales of c. 10^6–10^7 My. The thermal character of the Archaean mantle and models for Archaean continental crustal growth strongly suggest an even more important role for magmatic–thermal controls on eustasy in the early Precambrian (Campbell and Griffiths, 1992; De Wit et al., 1992; De Wit and Hynes, 1995; Windley, 1995). In addition, the exponential elevation–age relationship demonstrated for Phanerozoic oceanic lithosphere (e.g., Worsley et al., 1984) was probably more significant in the Precambrian (Eriksson, 1999). The interplay of plate tectonics and mantle (super)plumes (Condie, 1998; 2001) and their products (large igneous provinces or LIPs; e.g., Ernst et al., 2004) was significant during the entire Precambrian (and beyond). A relatively constant occurrence of LIPs is inferred for the Precambrian, with one occurring about every 20 My from 2.5 Ga; decreased frequency prior to 2.8 Ga may be an artifact of data analysis (e.g., Ernst and Buchan, 2001, 2002a,b). These authors find evidence for weak cyclicity in the plume/LIP record, at frequencies of c. 170, 330 and 730–600 My; there are possible gaps in this record at 3.3–3.0 Ga, 2.4–2.22 Ga and 720–615 Ma (Ernst et al., 2004). Mantle plumes (superplumes and superplume events, SPEs) have a direct relationship to the supercontinent cycle (e.g., Peltier et al., 1997; Condie, 1998) and occur close to supercontinent cycle terminations, leading to globally elevated sea levels (thus reinforcing sea level rise due to supercontinent attenuation and breakup) along with significant changes in ocean chemistry and stromatolite occurrence (e.g., Condie, 2004).

Another important factor affecting early Precambrian eustatic variation is the continental freeboard concept (relative elevation of a continent with respect to mean sea level; Wise, 1972, 1974) (Fig. 1), directly related to crustal growth processes and models (e.g., Eriksson, 1999, for detailed discussion). Isotopic and geochemical data strongly support the hypothesis of a near-constant thickness for continental crust since the Neoarchaean (e.g., McLennan and Taylor, 1983); this, in turn, is used as the basis for the constant freeboard model (Wise, 1972, 1974) since c. 2.5 Ga, which allows for the supercontinental cycle controlling first-order eustatic variation (Nance et al., 1986; Windley, 1995). The constant freeboard model also assumes that continental crustal growth largely occurred early in the Precambrian era (McLennan and Taylor, 1982), although there is still much debate about hypotheses of early Precambrian (e.g., Armstrong, 1981), episodic (e.g., Veizer and Jansen, 1979) and constant (e.g., Nelson, 1991) crustal growth models. It is further inferred that constant freeboard pertained prior to c. 2.5 Ga as well, due to the balance between crustal growth (raising eustatic levels) and declining mantle heat flow (lowering sea levels) (Reymer and Schubert, 1984; see also, discussion in Schubert, 1988; Windley, 1995). Obviously, an approximately uniform crustal growth model negates the constant freeboard concept. Linear growth models see crustal recycling as balancing output and input (e.g., O'Nions et al., 1979). A majority of researchers supports the episodic model of continental crustal growth, with c. 80% of crust having formed by about 2.5 Ga (e.g., Eriksson, 1995). It must be emphasised here that continental crustal growth, the freeboard concept and sea level changes are interdependent variables in Precambrian geology (Eriksson, 1999).

3. Controls on Neoarchaean–Palaeoproterozoic depositional cyclicity: a discussion

Although research on the application of sequence stratigraphy to Precambrian examples is still rather limited (e.g., Beukes and Cairncross, 1991; Krapez, 1993, 1996, 1997; Christie-Blick et al., 1988; Catuneanu and Eriksson, 1999, 2002; Ramaekers and Catuneanu, 2004; Eriksson and Catuneanu, 2004b) compared to Phanerozoic studies, it is generally possible to define first- and second-order depositional cycles relatively easily, and, in a number of cases, third-order cycles have been recognized. Here we will briefly examine several Archaean–Palaeoproterozoic studies and discuss the controls on these three orders of cyclicity, with emphasis on differences with younger sedimentary rock records.

3.1. First-order cyclicity

Evidence in support of the supercontinental cycle being responsible for first-order sedimentary cyclicity is well founded (e.g., Worsley et al., 1984, 1986), based primarily on the two supercontinent events known from the Phanerozoic. For the early Precambrian, although there are a number of papers discussing the concept, evidence for supercontinents remains inconclusive and is intrinsically tied to divergent models of continental crustal growth. Assuming validity for the early Archaean growth model of Armstrong (1981) will automatically prejudice the researcher in favour of postulated supercontinents such as "Ur" (centred on India and stable by c. 3.0 Ga; Rogers, 1996), and various configurations suggested for the Neoarchaean (e.g., Button, 1976; Piper, 1983; Williams et al., 1991—"Kenorland"; Gaál, 1992; Stanistreet, 1993—"Zimvaalbara"; Cheney, 1996—"Vaalbara"; Rogers, 1996—"Arctica"). Of these, perhaps the best constrained is Kenorland, which originally comprised the Archaean provinces of North America, but has since been expanded to include the Baltic and Siberian shields. The concept is supported by a large geochronological data base (Aspler and Chiarenzelli, 1998). These authors also postulated that another, "southern" (present-day frame of reference) supercontinent may have existed in the Neoarchaean–Palaeoproterozoic, including at least Kaapvaal and the Pilbara cratons, the two oldest known. However, despite a number of "Vaalbara" reconstructions (see Cheney, 1996 for overview) and a remarkable similarity in lithostratigraphy of their Neoarchaean–Palaeoproterozoic supracrustal successions, precise zircon age data and palaeomagnetic studies do not support this proposal, indicating that widespread "events" do not necessarily imply contiguity (Wingate, 1998; Nelson et al., 1999). Bleeker (2003) recently challenged the widely supported idea of a single (or perhaps a very limited number thereof) Neoarchaean supercontinent, and provides a good case for there possibly having been a relatively large number of "supercratons" which spawned the c. 35 known large Archaean cratonic fragments known today. He suggests that the Slave, Superior and Kaapvaal cratons bear evidence for derivation from independent supercratons, but that the Indian Dharwar, the Wyoming and the Zimbabwe cratons may have been part of an ancestral "Sclavia" supercraton which may also have incorporated the Slave.

The oldest supracrustal successions to which sequence stratigraphy has been applied are those of the granite-greenstone terranes of the Pilbara craton, Australia, where Krapez (1993; see also, 1996, 1997) has identified two first-order (Fig. 2) and seventeen second-order cycles from c. 3.5 to 2.7 Ga. He equates the former with the "classical" supercontinental cycle length of 440 My (Worsley et al., 1984; Nance et al., 1988), but notes that assembly and breakup phases of the cycle may have overlapped with the c. 120 My long stasis period after formation of the new landmass; first-order cycle length was thus inferred as 380–340 My. Krapez (1993) further refers to the analogous duration of the 2770–~2400 Ma Mount Bruce (succession) cycle of the overlying Hamersley basin succession of the Pilbara craton (e.g., Blake and Barley, 1992) and therefore suggests that steady state plate tectonic processes have operated throughout Earth history. Krapez (1993, 1996, 1997) supports the Armstrong (1981) hypothesis of early Precambrian crustal growth in his essentially "uniformitarian" Archaean sequence stratigraphic models.

The Fortescue and Hamersley Groups (forming the lower, c. 2.7–2.45 Ga portion of the Mount Bruce succession; Nelson et al., 1999) of the Pilbara craton show a very close lithostratigraphic relationship with equivalent, broadly coeval rocks of the Ventersdorp Supergroup and overlying Ghaap–Chuniespoort Groups of Kaapvaal (lower Transvaal Supergroup; e.g., Altermann and Nelson, 1998). Episodic volcanism, rather than sea level fluctuations, probably played a dominant role in the control of second-order depositional cycles within both the Fortescue and Ventersdorp successions. Cycles (second-order according to Nelson et al., 1999; more likely, third-order) with a high degree of similarity occur in both cratons within the Hamersley–lower Transvaal succession, and will be discussed below. Catuneanu and Eriksson (1999) have interpreted the sequence strat-

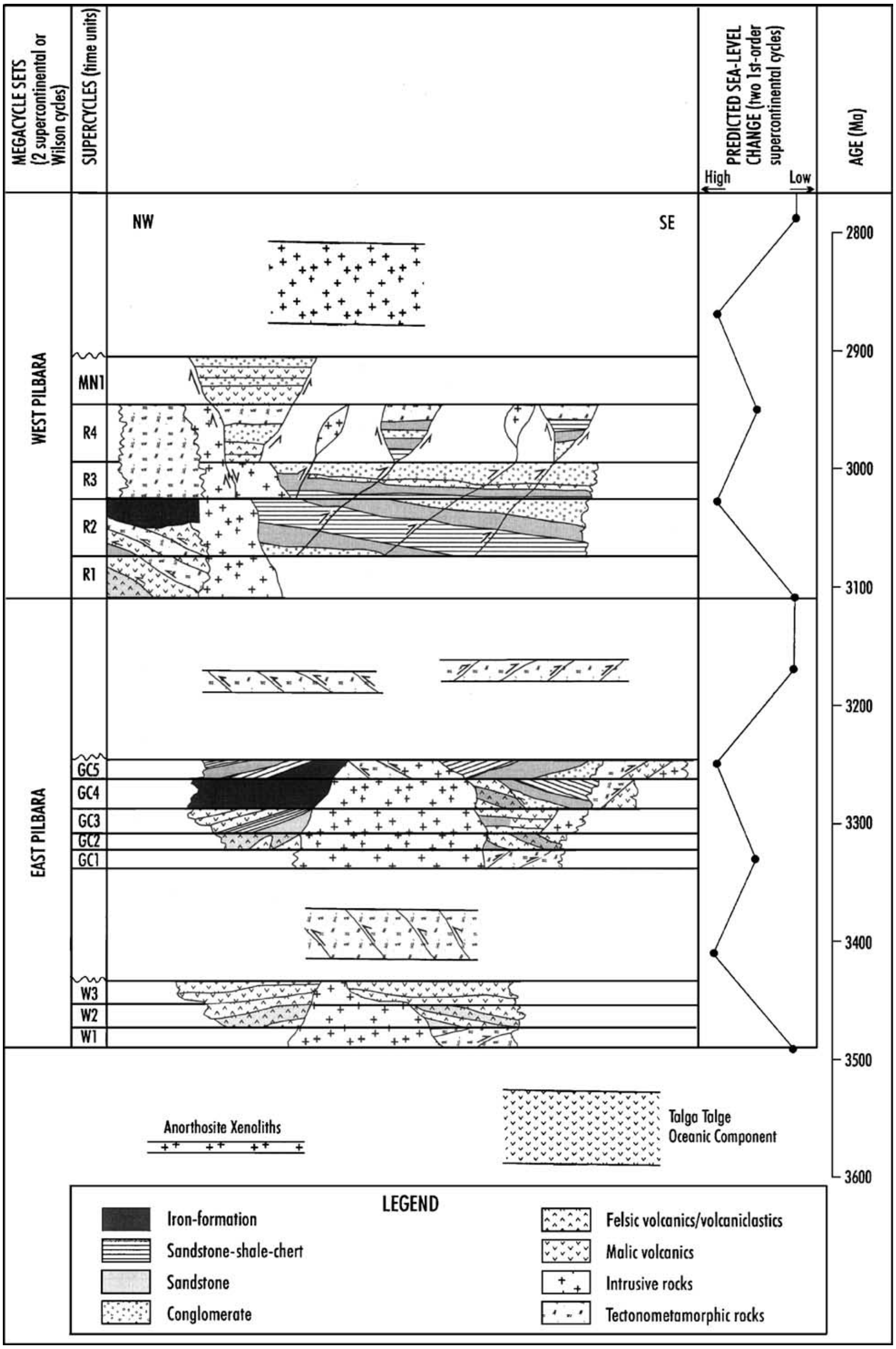


Fig. 2. Sequence stratigraphy interpreted for the Archaean Pilbara granitoid-greenstone terrane by Krapez (1993). The vertical and lateral associations of rock units and interpreted geotectonic settings are shown in the centre, inferred sea level change and a time scale at right, with the two first-order cycles (West Pilbara and East Pilbara) shown at left. Modified after Krapez (1993).

igraphy of the c. 2.7–2.1 Ga Transvaal Supergroup as comprising a first-order cycle of c. 650 My duration, with five second-order cycles (Fig. 3), discussed in the next section. They ascribe the first-order cycle length to a less evolved plate tectonic regime in contrast to the faster plate movements of the Phanerozoic. The sequence stratigraphic models of Krapez (1993) for Pilbara and Catuneanu and Eriksson (1999) for Kaapvaal for the c. 2.7–2.4 period are thus in direct contrast to each other, and both do not accord with the paradigm of faster plate movements in the Archaean when mantle heat flow was 2–3 times that at present (e.g., Nisbet, 1987).

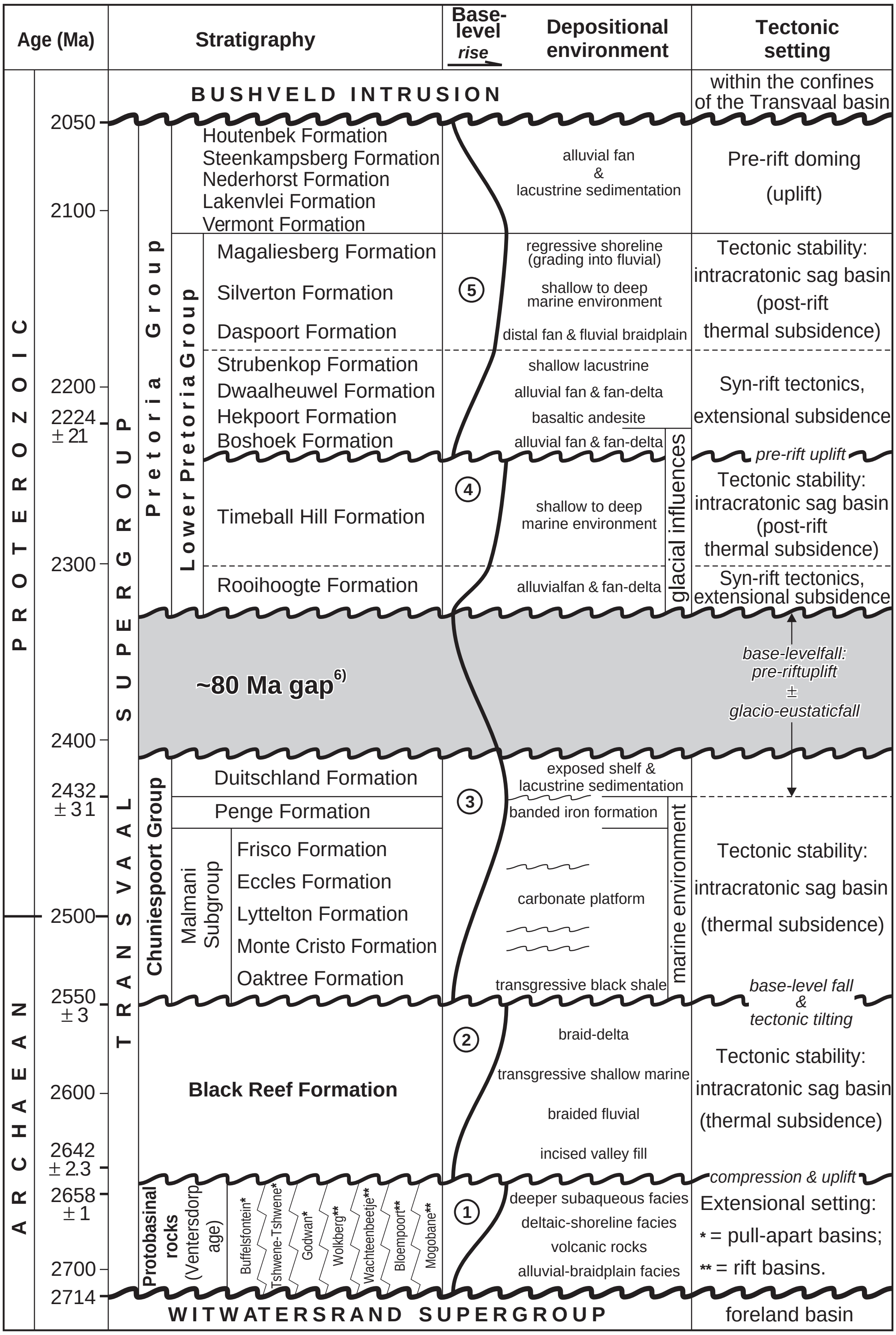
An essentially slower plate tectonic regime is also interpreted by Catuneanu (2001) for the c. 3.1–2.8 Ga Witwatersrand basin of Kaapvaal, where there is direct evidence of insufficient dynamic loading of the foreland system. Analogously, protracted attenuation and breakup of the Kenorland supercontinent is proposed by many researchers working in different segments of this dispersed possible supercontinent. Young et al. (2001) suggest a full Wilson cycle and a first-order cycle duration of c. 550 My for the c. 2.4–1.85 Ga Huronian Supergroup and Sudbury basin of the Superior Province, Canada. Aspler et al. (2001) propose an analogous supercontinental first-order cycle from c. 2.45–1.9 Ga for the succession preserved within the Hurwitz basin, Hearn domain, northern Canada. Similarly, for the essentially coeval Karelain Supergroup (Baltic shield), Ojakangas et al. (2001a) suggests a 550 My first-order Kenorland cycle.

The Kenorland cycles support the viewpoints on slower tectonic regimes and concomitant longer first-order cycles gained from the Transvaal and Witwatersrand basin-fills. Although these combined views may suggest that the work of Krapez (1993) on the 3.5–2.7 Ga Pilbara succession errs in its adherence to Phanerozoic-type first-order cyclicity and a steady-state plate tectonic system throughout Earth history, another scenario is also possible. The assumption that Meso-Neoarchaean plate tectonics was more rapid than that inferred for the c. 3.1–2.1 Ga Kaapvaal basins and comparable (at least in its plate movement rates) with Phanerozoic supercontinental cycles, as suggested by Krapez (1993), is in accord with the opinions of most workers studying Archaean greenstone belts, who find a strong similarity with Phanerozoic plate tectonic processes and island arc systems (e.g., Mueller and Corcoran, 2001; Daigneault et al., 2004). The entire question of early Precambrian plate tectonics and its nature remains a highly contentious issue (see Eriksson and Catuneanu, 2004a, for a recent relevant discussion); however, Trendall's (2002) "plughole" model provides a means of providing a transition from a whole mantle convecting Hadean Earth to a layered mantle–plate tectonic Neoarchaean globe. The latter model also encompasses a gradual transition from mantle-dominated early crustal evolution to a fully plate tectonic regime and explains the formation of deep continental roots beneath early Precambrian cratons.

It is possible that enhanced heat flow in the Neoarchaean–Palaeoproterozoic led to rapid rates of island arc accretion and intra-oceanic obduction tectonics (cf. De Wit et al., 1992), combining to result in rapid crustal growth and an apparent dominance of Phanerozoic-style plate tectonics within greenstone (cf. arc) systems, whereas more stable and rigid early cratons like Kaapvaal and Pilbara, with significant roots, were less amenable to rapid breakup during onset of Wilson cycles and were subject to smaller dynamic loading within foreland systems. The small size of these early cratons possibly limited the thermal blanketing effect and, additionally, the contrast in rheological behaviour and thermal character between such early small cratonic plates and the large, dynamic oceanic crustal realms may have been greater than that known from the Phanerozoic. These factors could, perhaps, explain more rapid plate tectonic regimes and shorter first-order sequence stratigraphic cyclicity in Archaean greenstone belts, with both becoming reduced when applied to emerging cratonic realms in the Palaeoproterozoic.

3.2. Second-order cyclicity

Second-order cycles of c. 10–100 My duration can generally be ascribed to a number of possible causes: (1) "dynamic topography", which refers to continental scale thermal uplift, due to the thermal blanketing effect trapping mantle heat beneath a large craton, or which can also result from mantle plumes (large scale) or mantle inhomogeneities (smaller, sub-continental scale)—these cycles will easily be identified within intracratonic basin-fills; (2) changes in volume and


Age (Ma)
Stratigraphy
Base-level rise
Depositional environment
Tectonic setting
BUSHVELD INTRUSION
within the confines of the Transvaal basin
2050
2100
2200
2224 ± 21
2300
2400
2432 ± 31
2500
2550 ± 3
2600
2642 ± 2.3
2658 ± 1
2700
2714
PROTEROZOIC
ARCHAEAN
TRANSVAAL SUPERGROUP
Pretoria Group
Lower Pretoria Group
Chuniespoort Group
Malmani Subgroup
Houtenbek Formation
Steenkampsberg Formation
Nederhorst Formation
Lakenvlei Formation
Vermont Formation
alluvial fan & lacustrine sedimentation
Pre-rift doming (uplift)
Magaliesberg Formation
Silverton Formation
Daspoort Formation
regressive shoreline (grading into fluvial)
shallow to deep marine environment
distal fan & fluvial braidplain
Tectonic stability: intracratonic sag basin (post-rift thermal subsidence)
Strubenkop Formation
Dwaalheuwel Formation
Hekpoort Formation
Boshoek Formation
shallow lacustrine
alluvial fan & fan-delta
basaltic andesite
alluvial fan & fan-delta
Syn-rift tectonics, extensional subsidence
pre-rift uplift
Timeball Hill Formation
shallow to deep marine environment
glacial influences
Tectonic stability: intracratonic sag basin (post-rift thermal subsidence)
Rooihoogte Formation
alluvialfan & fan-delta
Syn-rift tectonics, extensional subsidence
~80 Ma gap6)
base-levelfall: pre-riftuplift ± glacio-eustaticfall
Duitschland Formation
Penge Formation
exposed shelf & lacustrine sedimentation
banded iron formation
Frisco Formation
Eccles Formation
Lyttelton Formation
Monte Cristo Formation
Oaktree Formation
carbonate platform
transgressive black shale
marine environment
Tectonic stability: intracratonic sag basin (thermal subsidence)
base-level fall & tectonic tilting
Black Reef Formation
braid-delta
transgressive shallow marine
braided fluvial
incised valley fill
Tectonic stability: intracratonic sag basin (thermal subsidence)
compression & uplift
Protobasinal rocks (Ventersdorp age)
Buffelsfontein*
Tshwene-Tshwene*
Godwan*
Wolkberg**
Wachteenbeetje**
Bloempoort**
Mogobane**
deeper subaqueous facies
deltaic-shoreline facies
volcanic rocks
alluvial-braidplain facies
Extensional setting:
* = pull-apart basins;
** = rift basins.
WITWATERSRAND SUPERGROUP
foreland basin
5
4
3
2
1

spreading rate at MORs leading to global changes in sea level, which are best observed in successions from extensional craton marginal settings (cf. passive margins); (3) downwarping of cratonic margins due to extensional tectonic regimes and/or crustal loading (such as due to sedimentation) (Miall, 1997). Although the second mechanism will result in correlatable second-order sequences across several continents (e.g., Sloss, 1963; Soares et al., 1978; Hallam, 1984), the other two will tend to be more restricted, with the third having the most restricted distribution.

As pointed out above, mantle plumes and LIPs have a weak cyclicity of greater duration than second-order cycles (c. 170, 330 and 600–730 My; Ernst et al., 2004). They are thus possible contributors to first-order cyclicity or, alternatively may enhance or decrease first-order supercontinent cycles (and plumes have, anyway, a strong connection to such cycles; cf. Condie, 1998, 2001). A possible genetic relationship between LIP cyclicity and second-order sequences would thus appear unlikely, except in the sense that plumes and LIPs may have episodically caused continental epeirogenesis rather than resulting in cyclical changes to the hypsometric curves of individual continents or cratons. Despite a long-held paradigm that cratonic interior epeirogenesis reflected essentially plate marginal far field effects, the work of Gurnis and others (Bond, 1978; Gurnis, 1988, 1990, 1992; Burgess and Gurnis, 1995) has demonstrated clearly that larger and smaller scale convection cells in the mantle can produce cyclical changes such as uplift, subsidence, cratonic tilting and intracratonic basin formation. They also vary in scale from large events related to the supercontinent cycle and global geoid anomalies (supercontinents breaking up above geoid highs and assembling over geoid lows; cf. Condie, 2004) to sub-cratonic scale systems.

Catuneanu and Eriksson (1999) inferred that the first-order c. 650 My cycle of the Transvaal Supergroup, Kaapvaal, could be subdivided into five second-order unconformity-bound depositional sequences (Fig. 3). Each of these latter cycles was characterised by variable preservation of an ideal cyclical alternation of basal lowstand, transgressive, highstand and uppermost falling stage systems tracts. The Transvaal second-order sequences had an average duration of about 130 Ma; as with the first-order cycles discussed above, this is much longer than Phanerozoic equivalents. A less active plate tectonic regime was inferred for the c. 2.7–2.1 Ga period, as for the longer first-order Transvaal cycle (Catuneanu and Eriksson, 1999). These authors postulated that controls on the second-order cyclicity were largely from rifting and thermal subsidence, with a secondary influence from glacio-eustasy during the c. 2.4–2.2 Ga global refrigeration event.

Eriksson et al. (1999) attempted to correlate long-duration depositional cycles from the Neoarchaean–Palaeoproterozoic plates making up the African, Indian and Australian continents, and postulated globally enhanced sea levels at c. 2.6–2.4 Ga, with a second analogous event at c. <2.2 to >2.15 Ga (Fig. 4). These two longer-duration cycles equate with Catuneanu and Eriksson's (1999) second and third, and fourth and fifth second-order cycles (Fig. 3), respectively. Eriksson et al. (1999) interpreted the first, c. 2.6–2.4 Ga cycle as reflecting either "southern" supercontinent breakup or catastrophic global-scale mantle overturn events (cf. Nelson, 1998, 2004) related to a transition from whole to layered mantle convection and the onset of modern-style plate tectonics, at c. 2.7 Ga. The younger, c. 2.2 Ga cycle was inferred to be due to a combination of post-glacial eustatic rise combined with palaeo-atmospheric change and concomitant change in global weathering conditions. The older cycle postulated by Eriksson et al. (1999) would thus be analogous to a first-order cycle, but the second almost certainly belongs to a lower order, second or even third.

This illustrates a complex problem inherent in sequence stratigraphic interpretation of all successions, especially those of early Precambrian age: the difficulty of determining a sequence hierarchy, which leads to incompatibility of various interpretations of cyclic order. This problem is only exacerbated by the

Fig. 3. Summary figure illustrating lithostratigraphy, chronology, inferred tectonic settings and depositional palaeoenvironments for the Transvaal Supergroup in the Transvaal preservational basin. Inferred base-level changes (second column from right) comprise five second-order sequences (numbered 1–5 from the base up), forming the c. 650 My duration Transvaal first-order sequence. For ages and details see Catuneanu and Eriksson (1999, and references therein). Wavy lines represent suggested unconformable contacts.

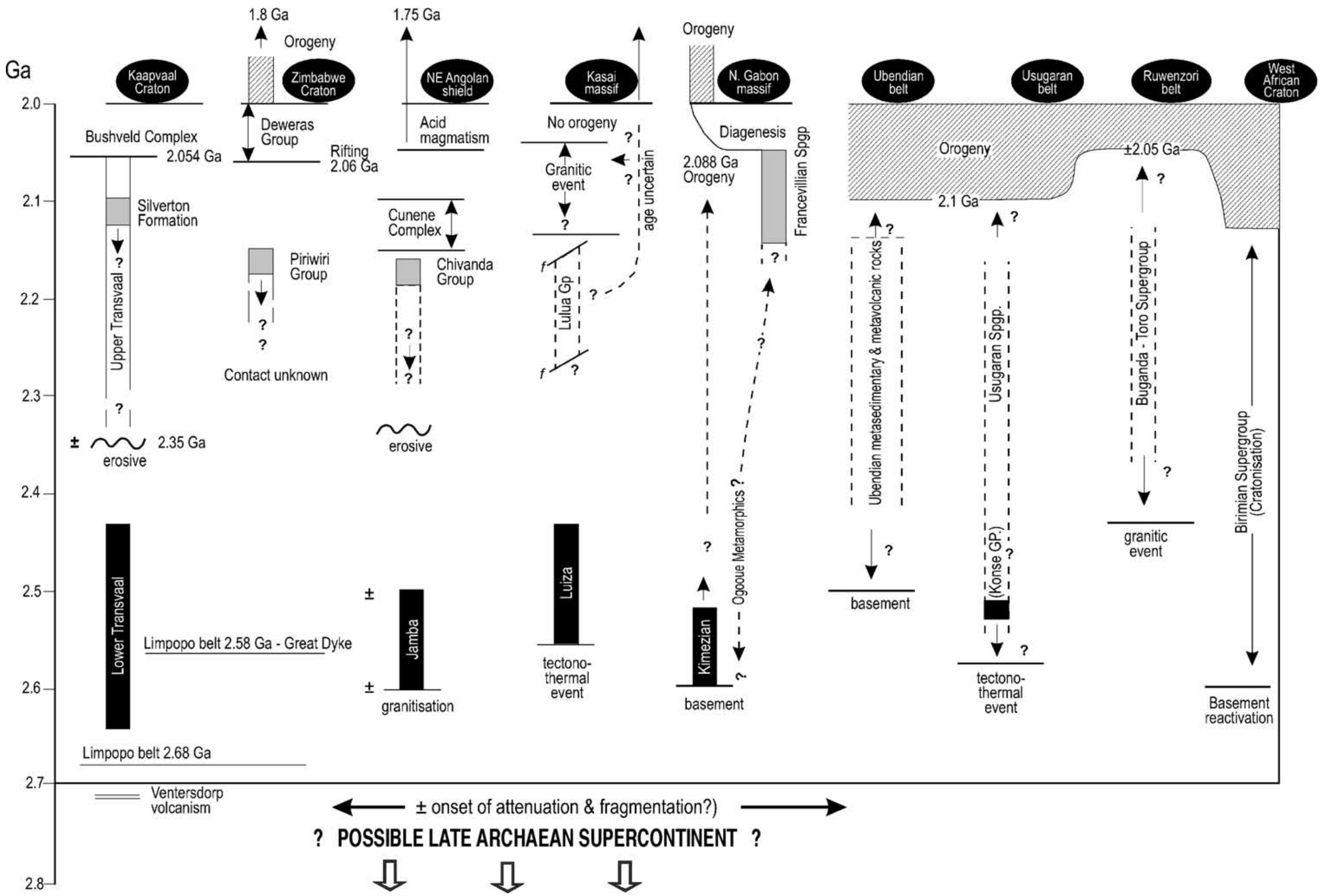


Fig. 4. Geohistory summary diagram for the 2.7–2.0 Ga volcano-sedimentary successions of Africa. Base of each column indicates the approximate age of stabilisation of crustal basement, and at the top of each column the onset of the Eburnean orogeny (cross-hatching) as well as major magmatic/tectonic events are shown. Note lower depositional cycle (transgressive epeiric marine succession, shown in black) at c. 2.6–2.4 Ga, correlatible across much of Africa, and characterised by carbonate-BIF rocks. Upper, analogous, widespread, transgressive depositional cycle, at c. <2.2 to >2.15 Ga (shown by light stippling) was characterised by common black shales. These two long-duration (probably first-order) cycles can also be correlated across large portions of India and Australia (see text). Modified after Eriksson et al. (1999).

almost universally poor chronological control available for most Precambrian successions. Still within the Transvaal Supergroup of Kaapvaal, and considering Catuneanu and Eriksson's (1999) third second-order cycle (the Ghaap–Chuniespoort Groups of carbonate-banded iron-formation affinity) (Fig. 3), it should be noted that Altermann and Nelson (1998; see also, Nelson et al., 1999) define four second-order cycles (duration 10–50 My) within this carbonate platform succession; they attempt to correlate these with analogous cycles from the Pilbara craton (Hamerlsey Group) (Fig. 5), following the ideas of Cheney (1996). They interpret these regressive–transgressive cycles as reflecting craton-marginal downwarping and loading due to sediment and volcanic rock accumulation (cf. (3) above in Miall's, 1997 summary of genesis of second-order cycles). Such downwarping may have been a delayed consequence of the rapid accumulation of the underlying thick Fortescue and Ventersdorp volcanic successions. Within the Catuneanu and Eriksson (1999) model of Transvaal sequence stratigraphy, these same cycles

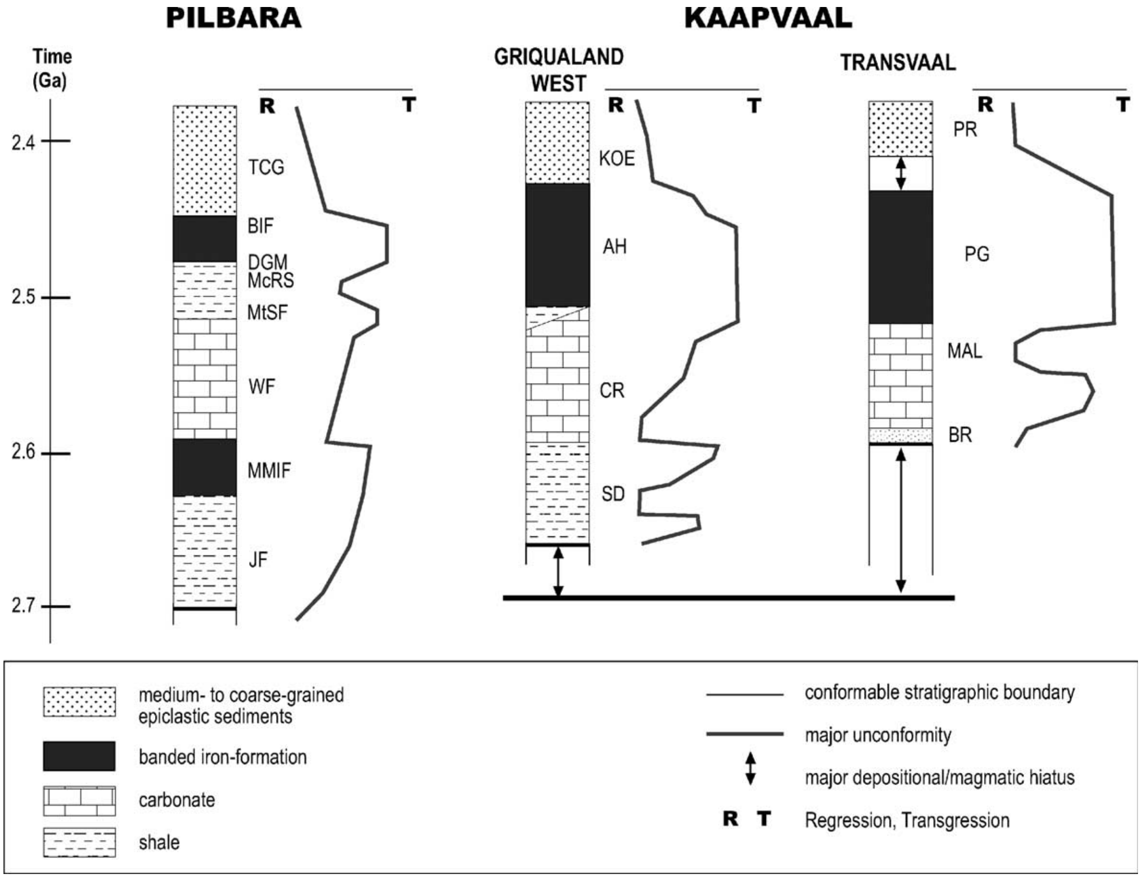


Fig. 5. Lithostratigraphic columns for the Pilbara craton (Hamersley basin) and for the Kaapvaal craton (Griqualand West and Transvaal preservational basins of the Transvaal Supergroup) for the c. 2.7–2.4 Ga supracrustal successions. Note transgression–regression curves for the three successions, defining analogous cyclicity on both cratons; these cycles are inferred to be of second-order rank by Altermann and Nelson (1998; see also Nelson et al., 1999). However, they form part of the third second-order Transvaal cycle shown in Fig. 3, and interpreted by Catuneanu and Eriksson (1999) as third-order cycles (see text for discussion). Modified after Nelson et al. (1999). Abbreviations: *Pilbara*—JF=Jeerinah Formation; MMIF=Marra Mamba Iron Formation; WF=Wittenoom Formation; MtSF=Mount Sylvia Formation; McRS=Mount McRae Shale; DGM=Dales Gorge Member, Brockman Iron Formation; BIF=Boolgeeda Iron Formation; TCG=Turee Creek Group. *Griqualand West*—SD=Schmidtsdrif Subgroup; CR=Campbellrand Subgroup; AH=Asbesheuwels Subgroup; KOE=Koegas Subgroup. *Transvaal*—BR=Black Reef; MAL=Malmani Subgroup; PG=Penge Iron Formation; PR=Pretoria Group.

would be viewed as third-order cycles; Catuneanu and Eriksson (1999) also viewed Transvaal second-order cyclicity as being due to a combination of craton margin downwarping/loading and intracratonic epeirogenesis (similar to (1) and (3) above in the Miall, 1997 scheme).

Within the c. 2.45–1.9 Ga Karelian Supergroup of the Baltic Shield, Ojakangas et al. (2001a) define five second-order cycles, ranging from 30 to 150 My, with an average duration of c. 100 My. They ascribe this cyclicity to changes in ocean basin volume from MOR volume variation, resulting in eustatic movement of sea level. All three major causes of second-order cylicity discussed at the beginning of this section (cf. Miall, 1997) thus appear to have validity. In addition, and as found for the first-order cycles, durations of second-order depositional cycles in the Neoarchaean–Palaeoproterozoic cratonic basinal successions may have been longer than Phanerozoic equivalents; once again, this is taken to reflect a less evolved (i.e., less "Phanerozoic-like") plate tectonic regime globally, or that more rigid continental plate behaviour impeded either more rapid epeirogenesis in craton interiors or faster subsidence along craton margins.

3.3. Third-order cyclicity

These cycles tend to support the Vail et al. (1977) postulate whereby eustasy is the main control on cyclic variation, and the thickness of third-order sequences matches the resolution typically found in seismic logs; in many ways, thus, this order of cylicity is closest to the Exxon school view of depositional cycles and their major controls. Within this paradigm, third-order cycles reflect glacio-eustasy (e.g., Vail et al., 1991). The uncommon preservation of glacial deposits and lack of evidence for global refrigeration events in the Precambrian rock record, already discussed above, do not support this postulate as the main genetic control on third-order cyclicity.

However, despite the general dearth of evidence for glaciation in the Precambrian record, this should not be read as implying that Earth lacked glaciated polar regions, despite a probable greenhouse atmosphere for much if not all of the Neoarchaean–Palaeoproterozoic period (e.g., Holland et al., 1986; Rye et al., 1995; Ohmoto, 2004 and references therein). A major variable in studying the Precambrian Earth is the change in rotation rate of the planet; accurate data are only available for the Neoproterozoic, suggesting a length-of-day (LOD) of c. 18.2 to 20.9 h (Sonett et al., 1996; Williams, 1998, respectively). Rautenbach (2001) used the previous estimate in sophisticated supercomputer models examining climatic change and palaeoclimatic character; even though it is probably the less accurate figure for the c. 900 Ma LOD, the principles resulting from Rautenbach's work remain essentially the same. A more rapid rotation rate will result in an increased number of meridional atmospheric circulation cells (cf. Hadley cells) on Earth compared to those of today, with reduced horizontal winds and increased vertical wind regimes in the atmosphere (Hunt, 1976; Rautenbach, 2001). Together, these differences would have led to less efficient transfer of heat from the warm equatorial regions to polar latitudes (Hunt, 1979; Rautenbach, 2001). Additionally, global climates would have had a greater static and dynamic stability as rotation rate increased (Stone, 1973; G.E. Williams, 2004, personal communication). The warm global greenhouse supported by the majority of researchers into Earth's early Precambrian palaeoclimate (e.g., Ohmoto, 2004) may thus have been pertinent only at relatively low to medium latitudes. Polar latitudes would have been glacial in character. The more extreme palaeoclimatic dichotomy of the Neoarchaean–Palaeoproterozoic allied to its greater latitudinal stability would likely have made a "snowball Earth" (totally frozen globe) model (e.g., Hoffman et al., 1998) more plausible (Molnar and Gutowski, 1995). However, hard geological data negating the snowball Earth model and discussed recently by Young (2004) and Williams (2004) make the possibility of a frozen Earth seem unlikely. G.M. Young (personal communication, 2004) additionally notes that evidence for the inferred Precambrian greenhouse palaeoclimates is based largely on studies of chemical weathering, and that the latter was likely influenced more by the high P_{CO_2} than by elevated temperatures.

Due to the uncommon evidence for global glaciation in the Precambrian rock record and the inferred palaeoclimatic effects of faster rotation during the Neoarchaean–Palaeoproterozoic period, glacio-eustasy cannot generally be accepted as a plausible explanation for third-order depositional cyclicity. Alternative explanations have been detailed above, the most likely being either intra-plate stresses (Cloetingh et al., 1985; Cloetingh, 1988a) or cyclic deformations of Earth's geoid (Mörner, 1994). Tectonism resulting in both relative (i.e., local) and eustatic changes in sea level remains important in generating third-order cyclicity. When applying these concepts to classic plate tectonic scenarios, Exxon-type sea level change curves apply relatively well to extensional continental margins and back-arc basin settings, but within arc-related basins, filling and deflation of magma chambers becomes an additional factor; within foreland basins, flexural loading causes relative sea level changes within the 10^4–10^7 year scale (e.g., Miall, 1997).

Within highly deformed Archaean greenstone belts, the depositional cyclicity usually recorded is that of third-order (e.g., Hofmann et al., 2001 for the Belingwe belt, Zimbabwe craton) or, sometimes, even higher order. The shorter duration of first-order Meso-to Neoarchaean cycles from such terranes, allied to the apparent concordance between many third-order cycle patterns and the Vail et al. (1977)-type eustatic cycles from the Phanerozoic record, may be important reasons for the generally uniformitarian geodynamic models applied by most researchers working on greenstone belts. As noted above, Neoarchaean–Palaeoproterozoic cratonic basin-fills generally appear to suggest slower plate dynamics and concomitantly elongated durations of first- and second-order depositional cycles.

Catuneanu (2001) identified 25 depositional cycles within the c. 3.0–2.8 Ga Witwatersrand Supergroup, Kaapvaal craton; limited age data suggest an average cycle length of c. 8 My. As the bounding surfaces extend across the preserved basin and show no evidence for deformation or for significant changes in sedimentation systems, these are interpreted as third-order cycles (Catuneanu, 2001). Catuneanu and Biddulph (2001) examined four of these cycles in more detail, interpreting a repetition of lowstand, transgressive and highstand systems tracts within the Vaal Reef facies of the Witwatersrand basin (Fig. 6). In view of the flexural foreland basin model applied to this basin (Catuneanu, 2001), the major control for these cycles was probably related to episodic loading and subsidence during advance of the thrust-front, alternating with stages of erosion as unloading and isostatic rebound occurred.

Beukes and Cairncross (1991) examined depositional cyclicity within the Pongola Supergroup, which is generally correlated with the Witwatersrand Supergroup. Their proposed sequence boundaries were based on the relationship between interpreted depositional facies. They inferred a passive margin tectonic setting (on the basis of facies-type sedimentary palaeoenvironmental modeling) for this portion of the greater Witwatersrand basin and thus felt justified in applying Vail et al. (1977) or Exxon school-type sequence stratigraphy, where eustasy within a framework of continuous tectono-thermal subsidence is the main control. However, the flexural foreland model supported by most Witwatersrand researchers (cf. Catuneanu, 2001) would place the Pongola basin within the backbulge part of the depository, where it would have been subjected to cyclical tectonic loading and unloading. Beukes and Cairncross (1991) calculated durations of c. 11 My for the six cycles they identified, and they interpreted these as second-order cycles related either to sea floor spreading rates or due to basin-wide tectonic events. The latter would appear to be in contradiction to their earlier assumption of a passive margin setting, while their support of the Haq et al. (1988) Phanerozoic sea level curves (see also discussion by Miall, this volume) for the Pongola cycles is not in agreement with the tectonic influences implicit in the flexural foreland basin model. We suggest, alternatively, that the Pongola cyclicity of c. 11 My, like the c. 8 My cycles identified within the Witwatersrand Supergroup by Catuneanu (2001) are, in reality, third-order cycles of essentially tectonic origin.

Analogously, the c. 10–50 My, second-order, cycles identified by Altermann and Nelson (1998) within the c. 2.58–2.43 Ga carbonate platform of the Transvaal basin, Kaapvaal (discussed under second-order cyclicity above) (Fig. 5), are possibly of third-order rank. The importance of establishing a hierarchical order of cyclicity (e.g., Catuneanu and Eriksson, 2004) is again emphasised by identification of third-order cycles within the lower part of the Pretoria Group, Transvaal Supergroup (Catuneanu and Eriksson, 2002) (Fig. 7). Here the Rooihoogte and overlying Timeball Hill Formations are bounded by strongly erosional and angular unconformities,

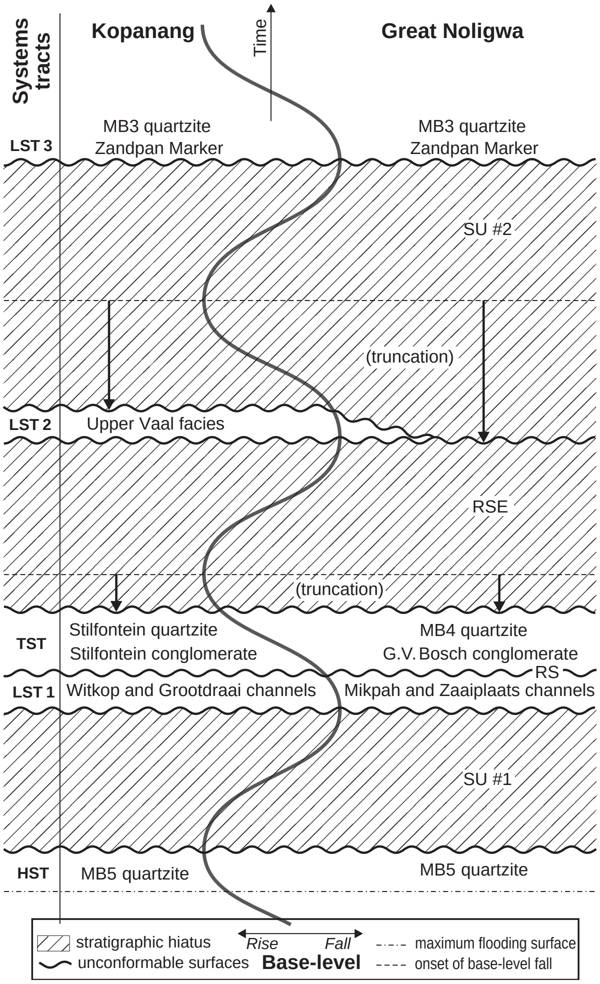


Fig. 6. Sequence stratigraphy of the Vaal Reef facies associations within the Kopanang and Great Noligwa localities, Central Rand Group, Witwatersrand Supergeroup, Kaapvaal craton. The figure illustrates systems tracts inferred for four of the 25 third-order depositional cycles identified by Catuneanu (2001) and Catuneanu and Biddulph (2001). Modified after the later authors. Abbreviations: HST=highstand systems tract; LST=lowstand systems tract; TST=trangressive systems tract; SU=subaerial unconformity; RSE=regressive surface of marine erosion; RS=ravinement surface.

thus indicating major tectonic reorganisation of the Transvaal basin, in accord with the definition of second-order cycles. Additionally, the Rooihoogte–Timeball Hill succession, comprising a conformable package of strata related genetically to a full cycle of rifting (Fig. 3), suggests a second-order character (Catuneanu and Eriksson, 2002). These authors discuss syn-rift extensional and post-rift thermal subsidence that provided the accommodation for sedimentation during the c. 2.4–2.2 Ga period of global glacio-eustatic fall for this second-order cycle.

Having established these controlling factors (rifting with concomitant mechanical and subsequent thermal subsidence) on second-order sea level change, the essentially concordant nature of the basin-wide unconformity at the base of the Klapperkop Sandstone Member, Timeball Hill Formation, can confidently be interpreted as dividing two third-order cycles within the Rooihoogte–Timeball Hill succession (Catuneanu and Eriksson, 2002) (Fig. 7). Thus, although the Rooihoogte–Timeball Hill succession was deposited during active (possibly global) glaciation (with at least

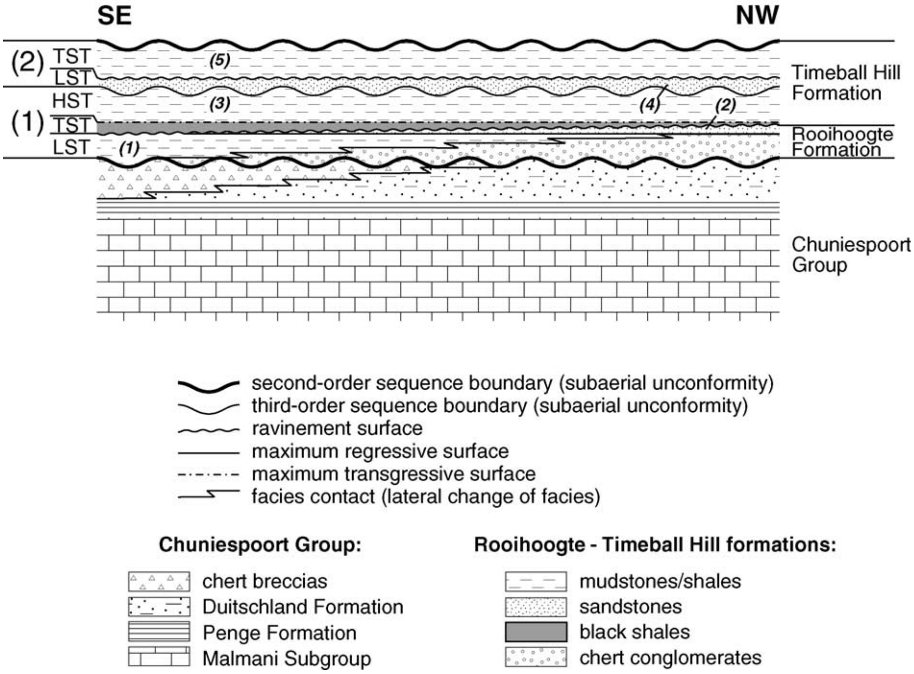


Fig. 7. Sequence stratigraphic interpretation (not to scale) of the Rooihoogte Formation–Timeball Hill Formation (Pretoria Group, Transvaal Supergroup; see second-order cycle number 4 in Fig. 3) succession. Note that this succession is split into two third-order depositional cycles (numbered as (1) and (2) at left of figure) by the basin-wide subaerial unconformity at the base of the Klapperkop Member, Timeball Hill Formation (numbered as 4 above). Abbreviations: LST, TST, HST as for Fig. 6; (1) = lacustrine Mudstone Member, Rooihoogte Formation; (2) = Polo Ground Sandstone Member, Rooihoogte Formation; (3) = Lower Shale member, Timeball Hill Formation, excluding the basal black shales which are shown with stipples; (4) = Klapperkop Member, Timeball Hill Formation; (5) = Upper Shale Member, Timeball Hill Formation. Modified after Catuneanu and Eriksson (2002).

three cyclic advances and retreats of glacial systems, cf. Young, 1991, 2004), the main control on third-order cyclicity within the Pretoria Group basin appears to have been tectono-thermal in character. It should be noted that there is a divergence of views whether the evidence for the c. 2.4–2.2 Ga glaciation observed on several cratons represents a truly global event or more restricted refrigeration; certainly, the lower two diamictite-bearing units in the Canadian Huronian Supergroup are only developed locally within fault-bounded basins, while the upper, Gowganda Formation is much more widespread (e.g., discussion in Young, 2004).

Higher up in the Pretoria Group of the Transvaal basin, the Daspoort Formation is also amenable to third-order sequence stratigraphic studies. As many Precambrian basins are only partially preserved, they should commonly show successions where no shoreline deposits approximately coeval with fluvial deposits are preserved. In such cases, the subdivision of fluvial deposits into lowstand, transgressive and highstand systems tracts cannot realistically be justified (e.g., Shanley and McCabe, 1994; Catuneanu and Eriksson, 2004). The alternative approach, of using changes in fluvial style and architecture, has been applied successfully recently in the Phanerozoic record (e.g., Boyd et al., 1999; Zaitlin et al., 2000). As accommodation is generally limited during early fluvial base level rise, to be followed by creation of accommodation at higher rates, the concept of these authors, namely that of low and high accommodation systems tracts, can be applied (Catuneanu and Eriksson, 2004). The first two applications of this concept to Precambrian fluvial deposits are those for the c. 2.0–1.6 Ga Athabasca basin (Rae and Hearne Provinces, Saskatchewan and Alberta, Canada) (Ramaekers and Catuneanu, 2004) and the c. <2.3 and >2.1 Ga Daspoort Formation (Eriksson and Catuneanu, 2004b). In both cases, third-order cyclicity was controlled essentially by tectonic influences, with tectonic indentation following upon the Trans-Hudson orogen influencing the Athabasca basin deposition (e.g., Ramaekers and Catuneanu, 2004), and post-rift thermal subsidence controlling fluvial sedimentation within the Daspoort Formation (Catuneanu and Eriksson, 1999; Eriksson and Catuneanu, 2004b).

Strand (1993) identified four sequences (mostly comprising transgressive and subsequent highstand systems tracts) of c. 10–20 My duration within the c. 2.5–2.1 Ga supracrustal rocks of the Fennoscandian craton. He related these cycles to variable thermal subsidence of the lithosphere following upon earlier rifting events (Strand and Laajoki, 1999). Within one of these cycles, they found evidence for a lower-order cyclicity (which they related to glaciogenic change in relative sea level) and were able to use these to correlate between two adjacent supracrustal sedimentary belts on the craton. Strand and Laajoki (1999) were uncertain whether the four sequences corresponded to second- or third-order cycles; Ojakangas et al. (2001a) in their study of the Karelian Supergroup, inferred them to be third-order cycles of 10–20 My duration, within the c. 100 My (average duration) second-order cycles they identified. Ojakangas et al. (2001b) also identify c. 1–10 My duration third-order cycles of glacio-eustatic origin within the c. 2.2–1.73 Ga rocks of the Lake Superior region, North America.

4. Conclusions

Generally, the well known controls on sea level change inferred for the Phanerozoic rock record can be applied to the Precambrian record as well, but some qualification is necessary. Possibly more uniform circulation systems on the wide and gently sloping shelves typical of Precambrian continental margins may have enhanced the control on short-term relative sea level resulting from waves and tides. As evidence for glaciation in the Precambrian is even less than that in the Phanerozoic record, glacio-eustasy would appear to have been much less important than intraplate stresses and cyclic changes to Earth's geoid as possible explanations for third-order cyclic changes in sea level. Changes to ocean basin volume due to "hot spots", aseismic ridges and mantle plumes (superplumes and SPEs) would have been more important influences in the early Precambrian due to a higher mantle and surface heat flow. A weak cyclicity identified in LIP events throughout the Precambrian equates approximately with first-order depositional cycle durations, and anyway, the supercontinent cycle is indubitably linked to thermal blanketing and supercontinent breakup due to mantle plumes (e.g., Condie, 1998). The interdependence of continental freeboard, sea level change and continental crustal growth rates was important for the Precambrian era.

Controls on Precambrian second-order cyclicity appear to have been due to thermal epeirogeny, changes to MOR volume and growth/decay rates, and cratonic marginal downwarping due to extensional tectonism and/or sediment loading (cf., Miall, 1997). For third-order cyclicity, additional controls in Precambrian basins included cyclic loading/unloading within flexural foreland settings, as well as filling and deflation within magma chambers related to arc evolution. The duration of third- and lower-order cycles overlaps with the errors normal in precision zircon dating techniques, but there is a strong need for accurate dating of ash beds interbedded within Precambrian sedimentary successions to better define the cyclicity periods of higher-order cycles. The importance of establishing a hierarchical order of cyclicity when interpreting the sequence stratigraphy of Precambrian basins (cf. Catuneanu and Eriksson, 2004) cannot thus be over-emphasised.

Examination of the limited number of studies of early Precambrian sequence stratigraphy suggests that in Archaean greenstone basins, first- and second-order cycles may have been analogous in duration to their Phanerozoic equivalents. This in turn indicates the possibility of steady state plate tectonic processes throughout Earth history (e.g., Krapez, 1993). However, basin-fills from Neoarchaean–Palaeoproterozoic cratonic terranes point to first- and second-order cycle durations which are significantly longer than their Phanerozoic counterparts; this in turn suggests that the supercontinent cycle and plate tectonism affecting these early stable continental plates were less evolved than on the Modern–Phanerozoic dynamic Earth. It is postulated that the contrast between much more dynamic oceanic plates, undergoing fast plate migrations and arc generation, and early stable and deeply-rooted cratons was such that this apparent dichotomy in plate tectonic character became possible.

We conclude that the sedimentary cyclicity observed in Precambrian successions is primarily a function of the interplay of a series of independent tectonic and climatic controls that operate over different time scales. In addition, each control on (relative) sea level shift, such as a cycle of continental rifting for example, may record a change through time in duration and rates of processes during the history of an evolving Earth. This suggests that attempts to classify cycles according to a rigid time framework may prove to be inadequate, and may lead to artificial interpretations of hierarchical orders as sequences are forced to fit pre-conceived classification schemes. Irrespective of time span, the nature of processes that lead to the formation of sequence boundaries, as well as their relative importance within each basin, may provide a more realistic basis for their classification into hierarchical orders. The wide range of controls on Precambrian sea level change, as well as their variability through time in terms of rates and periodicity, supports this conclusion and argues that each order of cyclicity is relative and needs to be defined within the stratigraphic context of individual case studies.

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References

Altermann W, Nelson DR. Sedimentation rates, basin analysis and regional correlations of three Neoarchaean and Palaeoproterozoic sub-basins of the Kaapvaal craton as inferred from precise U–Pb zircon ages from volcaniclastic sediments. Sediment Geol 1998;120:225–56.

Armstrong RL. Radiogenic isotopes: the case for crustal recycling on a near-steady-state no-continental-growth Earth. Philos Trans R Soc Lond, A 1981;301:443–72.

Aspler LB, Chiarenzelli JR. Protracted breakup of Kenorland, a Neoarchean supercontinent? Geochronologic, tectonostratigraphic and sedimentologic evidence from the Paleoproterozoic. Sediment Geol 1998;120/1–4:75–104.

Aspler LB, Wisotzek IE, Chiarenzelli JR, Losonczy MF, Cousens BL, McNicoll VJ, et al. Paleoproterozoic intracratonic basin processes, from breakup of Kenorland to assembly of Laurentia: Hurwitz Basin, Nunavut, Canada. Sediment Geol 2001;141–142:287–318.

Beukes NJ, Cairncross B. A lithostratigraphic–sedimentological reference profile for the Late Archaean Mozaan Group, Pongola Sequence: application to sequence stratigraphy and correlation with the Witwatersrand Supergroup. S Afr J Geol 1991;94(1): 44–69.

Blake TS, Barley ME. Tectonic evolution of the late archaean to early proterozoic Mount Bruce megasequence set, Western Australia. Tectonics 1992;11:1415–25.

Bleeker W. The late Archean record: a puzzle in ca 35 pieces. Lithos 2003;71:99–134.

Bond G. Speculations on real sea-level changes and vertical motions of continents at selected times in the Cretaceous and Tertiary periods. Geology 1978;6:247–50.

Boyd R, Diessel C, Wadsworth J, Chalmers G, Little M, Leckie D, et al. Development of a nonmarine sequence stratigraphic model. American Association of Petroleum Geologists Annual Meeting, San Antonio, Texas, USA Official program; 1999. p. A15.

Burgess PM, Gurnis M. Mechanisms for the formation of cratonic stratigraphic sequences. Earth Planet Sci Lett 1995;136:647–63.

Button A. Transvaal and Hamersley basins - review of basin development and mineral deposits. Miner Sci Eng 1976;8: 262–92.

Campbell IH, Griffiths RW. The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. J Geol 1992;100:497–523.

Catuneanu O. Flexural partitioning of the Late Archaean Witwatersrand foreland system, South Africa. Sediment Geol 2001;141–142:95–112.

Catuneanu O. Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls. J Afr Earth Sci 2002;35/1:1–43.

Catuneanu O, Biddulph MN. Sequence stratigraphy of the Vaal Reef facies associations in the Witwatersrand foredeep, South Africa. Sediment Geol 2001;141–142:113–30.

Catuneanu O, Eriksson PG. The sequence stratigraphic concept and the Precambrian rock record: an example from the 2.7–2.1 Ga Transvaal Supergroup, Kaapvaal craton. Precambrian Res 1999;97:215–51.

Catuneanu O, Eriksson PG. Sequence stratigraphy of the Precambrian Rooihoogte–Timeball Hill rift succession, Transvaal Basin, South Africa. Sediment Geol 2002;147:71–88.

Catuneanu O, Eriksson PG. Commentary. Chap. 8. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 735–8.

Cheney ES. Sequence stratigraphy and plate tectonic significance of the Transvaal succession of southern Africa and its equivalent in Western Australia. Precambrian Res 1996;79:3–24.
Christie-Blick N, Grotzinger JP, Von der Borch CC. Sequence stratigraphy in Proterozoic successions. Geology 1988;16: 100–4.
Christie-Blick N, von der Borch CC, DiBona PA. Working hypotheses for the origin of the Wonoka canyons (Neoproterozoic), South Australia. Am J Sci 1990;290-A:295–332.
Cloetingh S. Intraplate stresses: a new tectonic mechanism for fluctuations of relative sea-level. Geology 1986;14:617–21.
Cloetingh S. Intraplate stresses: a new element in basin analysis. In: Kleinspehn K, Paola C, editors. New Perspectives in Basin Analysis. Berlin: Springer-Verlag; 1988a. p. 205–30.
Cloetingh S. Intraplate stress: a tectonic cause for third order cycles in apparent sea level. In: Wilgus CK, Hastings BS, Kendall CGStC, Posamentier HW, Ross CA, Van Wagoner JC, editors. Sea Level Changes—An Integrated Approach. Spec Publ - Soc Econ Paleontol Mineral, vol. 42; 1988b. p. 19–30. Tulsa, Oklahoma.
Cloetingh S, McQueen H, Lambeck K. On a tectonic mechanism for regional sea level variations. Earth Planet Sci Lett 1985;75: 157–66.
Condie KC. Episodic continental growth and supercontinents: a mantle avalanche connection? Earth Planet Sci Lett 1998;163: 97–108.
Condie KC. Mantle plumes and their record in Earth history. Cambridge, UK: Cambridge University Press; 2001. 305 pp.
Condie KC. Precambrian superplume events. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 163–73.
Daigneault R, Mueller WU, Chown EH. Abitibi greenstone belt plate tectonics: the diachronous history of arc development, accretion and collision. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 88–103.
Dewey JF, Burke KCA. Hot spots and continental break-up: implications for collisional orogeny. Geology 1974;2:57–60.
De Wit MJ, Hynes A. The onset of interaction between the hydrosphere and oceanic crust, and the origin of the first continental lithosphere. In: Coward MP, Ries AC, editors. Early Precambrian Processes. Spec Publ - Geol Soc Lond, vol. 95; 1995. p. 1–9.
De Wit MJ, Roering C, Hart RJ, Armstrong RA, De Ronde REJ, Green RWE, et al. Formation of an Archaean continent. Nature 1992;357:553–62.
Donaldson JA, Eriksson PG, Altermann W. Actualistic versus non-actualistic conditions in the Precambrian: a reappraisal of an enduring discussion. In: Altermann W, Corcoran PL, editors. Precambrian Sedimentary Environments: A Modern Approach to Ancient Depositional Systems. IAS Spec Publ, vol. 33. Oxford: Blackwell; 2002. p. 3–13.
Donovan DT, Jones EJW. Causes of world-wide changes in sea level. J Geol Soc 1979;136:187–92.
Duval B, Cramez C, Vail PR. Types and hierarchy of stratigraphic cycles. In: Centre Nat. Rech. Sci. PR, editor. Mesozoic and Cenozoic sequence stratigraphy of European basins. Abstr Int Symposium, Dijon, France; 1992. p. 44–5.
Embry AF, Catuneanu O, Eriksson PG. Introduction. Chap. 8. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: tempos and events. Amsterdam: Elsevier; 2004. p. 681–4.
Emery D, Myers K. Sequence stratigraphy. Oxford: Blackwell; 1996. 297 pp.
Eriksson KA. Crustal growth, surface processes, and atmospheric evolution on the early Earth. In: Coward MP, Ries AC, editors. Early Precambrian Processes. Spec Publ - Geol Soc Lond, vol. 95; 1995. p. 11–25.
Eriksson PG. Sea level changes and the continental freeboard concept: general principles and application to the Precambrian. Precambrian Res 1999;97:143–54.
Eriksson PG, Catuneanu O. A commentary on Precambrian plate tectonics. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004a. p. 201–13.
Eriksson PG, Catuneanu O. Third-order sequence stratigraphy in the Palaeoproterozoic Daspoort Formation (Pretoria Group, Transvaal Supergroup), Kaapvaal craton. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004b. p. 724–35.
Eriksson PG, Condie KC, Tirsgaard H, Mueller WU, Altermann W, Miall AD, et al. Precambrian clastic sedimentation systems. Sediment Geol 1998;120/1–4:5–53.
Eriksson PG, Mazumder R, Sarkar S, Bose PK, Altermann W, van der Merwe R. The 2.7–2.0 Ga volcano-sedimentary record of Africa, India and Australia: evidence for global and local changes in sea level and continental freeboard. Precambrian Res 1999;97:269–302.
Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004a. 941 pp.
Eriksson PG, Bumby AJ, Popa M. Introduction. Chap. 7. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004b. p. 593–602.
Eriksson PG, Catuneanu O, Sarkar S, Tirsgaard H. Patterns of sedimentation in the Precambrian. Sediment Geol 2005;176:17–37 [this volume].
Ernst RE, Buchan KL. Large mafic magmatic events through time and links to mantle–plume heads. In: Ernst RE, Buchan KL, editors. Mantle plumes: their identification through time. Spec Pap - Geol Soc Am, vol. 352; 2001. p. 483–575. Boulder, Colorado.
Ernst RE, Buchan KL. Maximum size and distribution in time and space of mantle plumes: evidence from large igneous provinces. J Geodyn 2002a;34:309–42.
Ernst RE, Buchan KL. Maximum size and distribution in time and space of mantle plumes: evidence from large igneous provinces. J Geodyn 2002b;34:309–42 [Erratum, 2002, 34: 711–714].

Ernst RE, Buchan KL, Prokoph A. Large igneous province record through time. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: tempos and events. Amsterdam: Elsevier; 2004. p. 173–80.

Eyles N, Young GM. Geodynamic controls on glaciation in Earth history. In: Deynoux M, Miller M, Domack JMG, Eyles N, Fairchild IJ, Young GM, editors. Earth's glacial record. Cambridge Univ. Press.; 1994. p. 1–28.

Gaál G. Global Proterozoic tectonic cycles and early Proterozoic metallogeny. S Afr J Geol 1992;95(3):79–87.

Gross MG. Oceanography. Englewood Cliffs, New Jersey: Prentice Hall; 1990. 441 pp.

Gurnis M. Large-scale mantle convection and the aggregation and dispersal of supercontinents. Nature 1988;332:695–9.

Gurnis M. Ridge spreading, subduction and sea level fluctuations. Science 1990;250:970–2.

Gurnis M. Long-term controls on eustatic and epeirogenic motions by mantle convection. GSA Today 1992;2:141–57.

Hallam A. Major epeirogenic and eustatic changes since the Cretaceous and their possible relationship to crustal structure. Am J Sci 1963;261:397–423.

Hallam A. Pre-Quaternary sea-level changes. Annu Rev Earth Planet Sci 1984;12:205–43.

Hambrey MJ, Harland WB, editors. Earth's Pre-Pleistocene Glacial Record. Cambridge Univ. Press. 1004 pp.

Hambrey MJ, Harland WB. The late Proterozoic glacial era. Palaeogeogr Palaeoclimatol Palaeoecol 1985;51:255–72.

Haq BU, Vail PR, Hardenbol J, Van Wagoner JC. Sea level history. Science 1988;241:596–602.

Hargraves RB. Faster spreading or greater ridge length in the Archaean? Geology 1986;14:750–2.

Harrison CGA, Brass GW, Saltzman E, Sloan II J. Sea level variations, global sedimentation rates and the hypsographic curve. Earth Planet Sci Lett 1981;54:1–16.

Hoffman PF. Precambrian geology and tectonic history of North America. In: Bally AW, Palmer AR, editors. The Geology of North America, vol. A. Boulder, Colorado: Geol. Soc. Amer.; 1989. p. 447–512.

Hoffman PF, Kaufman AJ, Halverson GP, Schrag DP. A Neoproterozoic snowball Earth. Science 1998;281:1342–6.

Hofmann A, Dirks P, Jelsma H. Late Archaean foreland basin evolution, Belingwe greenstone belt, Zimbabwe. Sediment Geol 2001;141–142:131–68.

Holland HD, Lazar B, McCaffrey M. Evolution of the atmosphere and oceans. Nature 1986;320:27–33.

Holser WT. Gradual and abrupt shifts in ocean chemistry during Phanerozoic time. In: Holland HD, Trendall AF, editors. Patterns of Change in Earth Evolution. Berlin: Springer; 1984. p. 123–43.

Hoyle F. The history of the earth. Q J R Astron Soc 1972;13: 328–45.

Hubbard RJ. Age and significance of sequence boundaries on Jurassic and early Cretaceous rifted continental margins. Bull Am Assoc Pet Geol 1988;72:49–72.

Hunt BG. Experiments with a stratospheric general circulation model: Part IV. Inclusion of the hydrologic cycle. Mon Weather Rev 1976;104:333–50.

Hunt BG. The influence of the Earth's rotation rate on the general circulation of the atmosphere. J Atmos Sci 1979;36:1392–408.

Jervey MT. Quantitative geological modelling of siliciclastic rock sequences and their seismic expression. In: Wilgus CK, Hastings BS, Kendall CGStC, Posamentier HW, Ross CA, Van Wagoner JC, editors. Sea-level changes: an integrated approach. Spec Publ - Soc Econ Paleontol Mineral, vol. 42; 1988. p. 47–69. Tulsa, Oklahoma.

Knauth LP, Lowe DR. Oxygen isotopic composition of cherts from the Onverwacht Group (3.4 b. yrs), South Africa, with implications for secular variations in the isotopic composition of cherts. Abstr Programs - Geol Soc Am 1978;10:436.

Krapez B. Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia. Precambrian Res 1993;60:1–45.

Krapez B. Sequence-stratigraphic concepts applied to the identification of basin-filling rhythms in Precambrian successions. Aust J Earth Sci 1996;43:355–80.

Krapez B. Sequence-stratigraphic concepts applied to the identification of depositional basins and global tectonic cycles. Aust J Earth Sci 1997;44:1–36.

Lyell C. Principles of geology, vol. 1. 1st ed. London: John Murray; 1830. 511 pp.

McLennan SM, Taylor SR. Geochemical constraints on the growth of the continental crust. J Geol 1982;90:347–61.

McLennan SM, Taylor SR. Continental freeboard, sedimentation rates and growth of continental crust. Nature 1983;306: 169–72.

Miall AD. Principles of sedimentary basin analysis. New York: Springer; 1984. 490 pp.

Miall AD. Principles of sedimentary basin analysis. 2nd ed. New York: Springer; 1990. 668 pp.

Miall AD. Stratigraphic sequences and their chronostratigraphic correlation. J Sediment Petrol 1991;61:497–505.

Miall AD. The geology of stratigraphic sequences. Berlin: Springer-Verlag; 1997. 433 pp.

Miall AD. Testing for eustatic sea-level control in the Precambrian sedimentary record. Sediment Geol 2005;176:9–16 [this volume].

Miall AD, Miall CE. Sequence stratigraphy as a scientific enterprise: the evolution and persistence of conflicting paradigms. Earth-Sci Rev 2001;54:321–48.

Miall CE, Miall AD. The Exxon factor: the roles of corporate and academic science in the emergence and legitimation of a new global model of sequence stratigraphy. Sociol Q 2002;43/1:1–25.

Miller JMG. Glacial sediments. In: Reading HG, editor. Sedimentary environments. Oxford: Blackwell; 1996. p. 454–84.

Molnar GI, Gutowski WJ. The faint young Sun paradox: further exploration of the role of dynamical heat-flux feed backs in maintaining global climate stability. J Glaciol 1995;41/137:87–90.

Mörner N-A. Eustasy and geoid changes. J Geol 1976;84:123–51.

Mörner N-A. Internal response to orbital forcing and external cyclic sedimentary sequences. In: de Boer P.L., Smith D.G., editors. Orbital forcing and cyclic sequencesSpec Publ Int Assoc Sedimentol, vol. 19. Oxford: Blackwell; 1994. p. 25–33.

Mueller WU, Corcoran PL. Volcano-sedimentary processes operating on a marginal continental arc: the Archean Raquette Lake Formation, Slave Province, Canada. Sediment Geol 2001;141–142:169–204.

Nance RD, Worsley TR, Moody JB. Post-Archean biogeochemical cycles and long-term episodicity in tectonic processes. Geology 1986;14:514–8.

Nance RD, Worsley TR, Moody JB. The supercontinent cycle. In: Moores EM, editor. Tectonics and shaping the Earth, continents and oceans. Readings from Scientific American. New York: Freeman; 1988. p. 177–87.

Nelson KD. A unified view of craton evolution motivated by recent deep seismic reflection and refraction results. Geophys J Int 1991;105:25–35.

Nelson DR. Granite-greenstone crust formation on the Archaean Earth - a consequence of two superimposed processes. Earth Planet Sci Lett 1998;158:109–19.

Nelson DR. Episodic crustal growth during catastrophic global-scale mantle overturn events. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 180–3.

Nelson DR, Trendall AF, Altermann W. Chronological correlations between the Pilbara and Kaapvaal cratons. Precambrian Res 1999;97:165–89.

Nisbet EG. The young Earth: an introduction to Archean geology. Boston: Allen and Unwin; 1987. 402 pp.

Ohmoto H. The Archaean atmosphere, hydrosphere and biosphere. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 361–88.

Ojakangas RW, Marmo JS, Heiskanen KI. Basin evolution of the Paleoproterozoic Karelian Supergroup of the Fennoscandian (Baltic) Shield. Sediment Geol 2001a;141–142:255–85.

Ojakangas RW, Morey GB, Southwick DL. Paleoproterozoic basin development and sedimentation in the Lake Superior region, North America. Sediment Geol 2001b;141–142:319–41.

O'Nions RK, Evensen NM, Hamilton PJ. Geochemical modeling of mantle differentiation and crustal growth. J Geophys Res 1979;84:6091–101.

Page NJ. The Precambrian diamictite below the base of the Stillwater Complex, Montana. In: Hambrey MJ, Harland WB, editors. Earth's Pre-Pleistocene Glacial Record. Amsterdam: Cambridge Univ. Press. p. 821–3.

Payton CE, editor. Seismic Stratigraphy—Applications to Hydrocarbon Exploration. Mem - Am Assoc Pet Geol, vol. 26; 1977. 516 pp.

Peltier WR, Butler S, Solheim LP. The influence of phase transformations on mantle mixing and plate tectonics. In: Crossley DJ, editor. Earth's Deep Interior. Amsterdam: Gordon and Breach; 1997. p. 405–30.

Perry Jr EC, Ahmad SN, Swulius TM. The oxygen isotope composition of 3800 my old metamorphosed chert and iron formation from Isukasia, West Greenland. J Geol 1978;86: 223–39.

Piper JDA. Proterozoic palaeomagnetism and single continent plate tectonics. Geophys J R Astron Soc 1983;74:163–97.

Pitman III WC. Relationship between eustasy and stratigraphic sequences on passive margins. Geol Soc Amer Bull 1978;89: 1389–403.

Pitman III WC, Golovchenko X. The effect of sea level change on the shelf edge and slope of passive margins. In: Stanley DJ, Moore GT, editors. The Shelfbreak: Critical Interface on Continental Margins. Spec Publ - Soc Econ Paleontol Mineral, vol. 33; 1983. p. 41–58. Tulsa, Oklahoma.

Posamentier HW, James DP. An overview of sequence-stratigraphic concepts: uses and abuses. In: Posamentier HW, Haq BU, Allen GP, editors. Sequence Stratigraphy and Facies Associations. Spec Publ Int Assoc Sedimentol, vol. 18; 1993. p. 3–18. Oxford.

Posamentier HW, Jervey MT, Vail PR. Eustatic controls on clastic deposition I: conceptual framework. In: Wilgus CK, Hastings BS, Kendall CGStC, Posamentier HW, Ross CA, Van Wagoner JC, editors. Sea-level Changes: An Integrated Approach. Spec Publ - Soc Econ Paleontol Mineral, vol. 42; 1988. p. 109–124. Tulsa, Oklahoma.

Ramaekers P, Catuneanu O. Development and sequences of the Athabasca basin, early Proterozoic, Saskatchewan and Alberta, Canada. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 705–23.

Rautenbach CJ de W. A hypothetical approach to determining the effect of palaeorotational rates on Earth's Neoproterozoic palaeoclimate. J Afr Earth Sci 2001;33:463–73.

Reading HG, editor. Sedimentary Environments and Facies. Oxford: Blackwell; 1978. 557 pp.

Reading HG, editor. Sedimentary Environments: Processes, Facies and Stratigraphy, 3rd ed. Oxford: Blackwell; 1996. 688 pp.

Reading HG, Levell BK. Controls on the sedimentary rock record. In: Reading HG, editor. Sedimentary Environments: Processes, Facies and Stratigraphy. Oxford: Blackwell; 1996. p. 5–36.

Reymer A, Schubert G. Phanerozoic addition rates to the continental crust and crustal growth. Tectonics 1984;3:63–77.

Rogers JJW. A history of continents in the past three billion years. J Geol 1996;104:91–107.

Rye RO, Kuo PH, Holland HD. Atmospheric carbon dioxide concentrations before 2.2 billion years ago. Nature 1995;378: 603–5.

Schlanger SO, Jenkyns HC, Premoli Silva I. Volcanism and vertical tectonics in the Pacific basin related to global Cretaceous transgressions. Earth Planet Sci Lett 1981;52:435–49.

Schopf TJM. Paleoceanography. Cambridge, Mass.: Harvard University Press; 1980. 341 pp.

Schubert G. Physical processes in the growth of the continental crust. In: Ashwall LD, editor. Workshop on the Growth of Continental Crust. Houston, Texas: Lunar Planetary Inst. Tech. Rep.; 1988. p. 131–2.

Shanley KW, McCabe PJ. Perspectives on the sequence stratigraphy of continental strata. Am Assoc Petr Geol Bull 1994;78: 544–68.

Sloss LL. Sequence in the cratonic interior of North America. Geol Soc Amer Bull 1963;74:93–114.

Soares PC, Landim PMB, Fulfaro VJ. Tectonic cycles and sedimentary sequences in the Brazilian intracratonic basins. Geol Soc Amer Bull 1978;89:181–91.

Sonett CP, Kvale EP, Zakharian A, Chan MA, Demko TM. Late Proterozoic and Palaeozoic tides, retreat of the moon, and rotation of the Earth. Science 1996;273:100–4.

Stanistreet IG. Ancient and modern examples of tectonic escape basins: the Archaean Witwatersrand Basin compared with the Cenozoic Maracaibo Basin. In: Frostick LE, Steel RJ, editors. Tectonic Controls and Signatures in Sedimentary Successions. Spec Publ Int Assoc Sedimentol, vol. 20. Oxford: Blackwell; 1993. p. 363–76.

Stone PH. The effect of large-scale eddies on climatic change. J Atmos Sci 1973;30/4:521–9.

Strand K. Sedimentation in a Palaeoproterozoic rift to cratonic margin transition, Finland. Acta Univ Ouluensis, A Sci Rerum Nat 1993;A242.

Strand K, Laajoki K. Application of the parasequence concept to the Paleoproterozoic record of the northern Fennoscandian shield. Precambrian Res 1999;97:253–67.

Sykes T, Kidd RB. Volcanogenic sediment distributions in the Indian Ocean through the Cretaceous and Cenozoic, and their paleoenvironmental implications. Mar Geol 1994;116:267–91.

Trendall AF. The significance of iron formation in the Precambrian stratigraphic record. In: Altermann W, Corcoran PL, editors. Precambrian Sedimentary Environments: A Modern Approach to Depositional Systems. IAS Spec Publ, vol. 44; 2002. p. 33–66. Blackwell, Oxford.

Vail PR, Mitchum Jr RM, Thompson III S. Seismic stratigraphy and global changes of sea level. In: Payton CE, editor. Seismic Stratigraphy—Applications to Hydrocarbon Exploration. Mem - Am Assoc Pet Geol, vol. 26, p. 49–212.

Vail PR, Audemart F, Bowman SA, Eisner PN, Perez-Cruz G. The stratigraphic signatures of tectonics, eustasy and sedimentation—an overview. In: Einsele G, Ricken W, Seilacher A, editors. Cyclic Stratigraphy. New York: Springer; 1991. p. 617–59.

Veizer J, Jansen SL. Basement and sedimentary recycling and continental evolution. J Geol 1979;87:341–70.

Von Brunn V, Gold DJD. Diamictite in the Archean Pongola sequence of southern Africa. J Afr Earth Sci 1993;16:367–74.

Williams GE. Precambrian tidal and glacial clastic deposits: implications for Precambrian Earth–Moon dynamics and palaeoclimate. Sediment Geol 1998;120:55–74.

Williams GE. The paradox of Proterozoic glaciomarine deposition, open seas and strong seasonality near the Palaeo-Equator: global implications. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: Tempos and Events. Amsterdam: Elsevier; 2004. p. 448–59.

Williams GE. Subglacial meltwater channels and glaciofluvial deposits in the Kimberley Basin, Western Australia: 1.8 Ga low-latitude glaciation coeval with continental assembly. J Geol Soc Lond 2005;162:111–24.

Williams H, Hoffman PF, Lewry JF, Monger JWH, Rivers T. Anatomy of North America: thematic portrayals of the continent. Tectonophysics 1991;187:117–34.

Windley BF. The evolving continents. Chichester: Wiley; 1995. 526 pp.

Wingate MTD. A palaeomagnetic test of the Kaapvaal–Pilbara (Vaalbara) connection at 2.78 Ga. S Afr J Geol 1998;101: 257–74.

Wise DU. Freeboard of continents through time. Mem Geol Soc Amer 1972;132:87–100.

Wise DU. Continental margins, freeboard and volumes of continents and oceans through time. In: Burke CA, Drake CL, editors. The Geology of Continental Margins. Berlin: Springer; 1974. p. 45–58.

Worsley TR, Nance RD, Moody JB. Global tectonics and eustasy for the past 2 billion years. Mar Geol 1984;58:373–400.

Worsley TR, Nance RD, Moody JB. Tectonic cycles and the history of the Earth's biogeochemical and paleoceanographic record. Paleoceanography 1986;1:233–63.

Young GM. An extensive early Proterozoic glaciation in North America? Palaeogeogr Palaeoclimatol Palaeoecol 1970;7: 85–101.

Young GM. The geologic record of glaciation: relevance to the climatic history of the Earth. Geosci Can 1991;18:100–8.

Young GM. Are Neoproterozoic glacial deposits preserved on the margins of Laurentia related to the fragmentation of two supercontinents? Geology 1995;23:153–6.

Young GM. Stratigraphic and tectonic settings of Proterozoic glaciogenic rocks and banded iron-formations: relevance to the snowball Earth debate. J Afr Earth Sci 2002;35:451–66.

Young GM. Earth's two great Precambrian glaciations: aftermath of the "Snowball Earth" hypothesis. In: Eriksson PG, Altermann W, Nelson DR, Mueller WU, Catuneanu O, editors. The Precambrian Earth: tempos and events. Amsterdam: Elsevier; 2004. p. 440–8.

Young GM, von Brunn V, Gold DJC, Minter WEL. Earth's oldest reported glaciation: physical and chemical evidence from the Archean Mozaan Group (~2.9 Ga) of South Africa. J Geol 1998;106:523–38.

Young GM, Long DGF, Fedo CM, Nesbitt HW. Paleoproterozoic Huronian basin: product of a Wilson cycle punctuated by glaciations and a meteorite impact. Sediment Geol 2001; 141–142:233–54.

Zaitlin BA, Potocki D, Warren MJ, Rosenthal L, Boyd R. Sequence stratigraphy in low accommodation foreland basins: an example from the lower Cretaceous Basal Quartz Formation of southern Albert. Abstracts, GeoCanada 2000 conference. Can Soc Pet Geol, (on CD).