Condensation origin for Neoproterozoic cap carbonates during deglaciation

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ABSTRACT

Neoproterozoic deglacial stratigraphy is commonly characterized by a sharp contact separating glacial sediments from laminated capping carbonates. This stratigraphic relation is generally assumed to have time significance and to reflect an abrupt shift from icehouse to greenhouse conditions. In contrast to this, sequence stratigraphic field studies of an Ediacaran (ca. 635 Ma) glacial to postglacial transition in the Amadeus Basin of central Australia reveal a complex deglacial stratigraphy, in which more than 175 m of conglomerate, sandstone, marl, and carbonate at the basin margin, and portions of four unconformity-bounded sequences, pass basinward into no more than 3 m of laminated dolomicrite of typical cap carbonate facies. The unconformities, which are characterized by as much as several tens of meters of erosional relief (oblique sections of incised valleys), separate intervals of contrasting sediment provenance, and are confidently mapped on the basis of both criteria. Comparable unconformities are absent in the overlying Neoproterozoic succession, which is >2 km thick and encompasses many tens of millions of years. The Amadeus Basin cap carbonate was thus deposited during a protracted interval of multiphase (cyclical) transgression more similar to Phanerozoic cyclical sea-level rise than to the single catastrophic deglaciation and instantaneous precipitation invoked by popular current models to explain the classic cap carbonate. The superposition of carbonate on glacial facies in distal sections evidently records condensation in the absence of siliciclastic sediment rather than abrupt shifts between glacial and tropical conditions. Facies lithologically similar to cap carbonates may be less obvious in Phanerozoic successions because of a secular change in carbonate composition to reefal and deep-sea pelagic deposits.

INTRODUCTION

Neoproterozoic glacial deposits share a common sedimentary motif, i.e., an abrupt upward transition from glacigenic facies to thinly laminated, fine-grained carbonate deposits typically no more than a few meters to tens of meters thick, and sharply overlain in turn by a thick succession of siltstone (Fairchild and Kennedy, 2007). Sharp, apparently conformable lithological contacts and the lateral continuity of the carbonate at basinal scales, even in successions dominated by siliciclastic deposits, leave the strong impression that cap carbonates record a short-lived chemical oceanographic event punctuating a geologically instantaneous end to one of the most severe ice ages in Earth history (Hoffman et al., 1998). The nature of the geochemical event, however, remains a topic of vigorous debate. Most models link carbonate deposition with the reallocation of carbon between rapidly exchanging reservoirs such as the atmosphere (Hoffman et al., 1998), the deep ocean (Grotzinger and Knoll, 1995), or methane clathrates (Kennedy et al., 2001). Rapid weathering of carbonate shelves (Higgins and Schrag, 2003) and freshwater mixing (Shields, 2005) have also been invoked. The duration of carbonate deposition implied by these models is necessarily short (hundreds to tens of thousands of years), because the non-steady-state processes involved are ultimately constrained by oceanic mixing rates and by the residence time of carbon in the ocean. Such phenomena are not likely to play a significant role beyond ~100 k.y.

An alternative explanation relates cap carbonates to the concentration of hemipelagic and/or pelagic background carbonate sediment through a reduction in siliciclastic dilution during postglacial transgression and landward migration of point sources of sediment (Bjørlykke et al., 1978; Kennedy, 1996). In this model, cap carbonates are not inherently related to geochemical cycling, climate change, or a discrete global event except as a by-product of continental flooding. Delivery of carbonate to the seafloor occurs at time scales that may extend well beyond 100 k.y. Carbonate flux during transgression may be augmented by hypsographically induced concentration increases of alkalinity during the preceding sea-level lowstand (Opdyke and Walker, 1992). This effect, known from the Quaternary as the coral reef hypothesis, would have been much exaggerated in the Precambrian without the buffering of carbonate compensation provided by pelagic calcifiers (Kennedy, 1996; Ridgwell et al., 2003).

The simple tripartite lithostratigraphic architecture of the vast majority of cap carbonates globally (i.e., glacial deposits, thin carbonate, shale) is consistent with both non-steady-state models (e.g., snowball type) and steady-state models (condensation). However, since the range of time scales appropriate for each is different, evidence for the duration of deposition provides a criterion for discrimination. Owing to limitations in both sampling and geochronological resolution, it has not yet been possible to bracket the ages of cap carbonates and to test the models in this way. Several proxies for depositional duration have been suggested. Rapid deposition (<10 k.y.) has been inferred on the basis of rates of aragonite crystal fan growth (Hoffman and Schrag, 2002), rates of carbonate deposition in slope settings (Shields, 2005), diurnal time scales for millimetric laminae (Hoffman et al., 1998), and Quaternary rates of glacioeustatic rise (Higgins and Schrag, 2003). By contrast, evidence for hiatal surfaces, karst (James et al., 2001), or condensation intervals (slow sedimentation engendered by a marked decrease in siliciclastic supply; e.g., Loutit et al., 1988) within cap carbonates (Dyson and von der Borch, 1994; Kennedy, 1996), diachronous stratigraphic patterns (Hoffman et al., 2007), features common to condensation including metal enrichments, barite crystal growth, and Mn-rich microstromatolites (Kennedy, 1996), along with multiple magnetopolarity reversals suggest deposition of the cap dolostone over considerably longer time scales, >100 k.y. (Trindade et al., 2003; Kilner et al., 2005; Schmidt et al., 2009).

Here we describe the stratigraphic relations of a cap carbonate unit in structurally continuous exposures at the margin of the Amadeus Basin of central Australia. While basinal sections show the typical tripartite motif, in proximal locations the cap interval expands greatly in thickness and passes laterally into coarse-grained siliciclastic facies punctuated by mappable unconformities. We discuss the significance of stratigraphic discontinuities within the cap carbonate in the context of the rate of deglaciation and origin of the cap carbonate.

CAP CARBONATES IN THE AMADEUS BASIN

Glacial deposits at two discrete levels in the Neoproterozoic stratigraphic record of the Amadeus Basin (Areyonga and Olympic

Formations; Figs. 1A, 1B, and 2) have been correlated on a lithologic basis to the better known Sturtian and Marinoan glacial intervals of South Australia (Preiss, 1987; Kennedy et al., 1998). Over much of the Amadeus Basin, the upper cap interval shows the typical tripartite relationship, consisting of thinly laminated dolomite, underlain by heterolithic glacial deposits of the Olympic Formation, and overlain by siltstone of the Pertatataka Formation (Fig. 1A). At the basin's northeastern margin, however, a more varied succession of conglomerate, sandstone, marl, and carbonate has been exposed in two panels as a result of Devonian folding. Within these panels, the simple tripartite relation expands as it is traced landward toward the basin margin. Here an ~175 m heterolithic succession of fluvial and marine sediments, including intervals of cap-like carbonate, is exposed. This succession, which unconformably overlies glacigenic sediments of the Olympic Formation and underlies basin-wide siltstone of the Pertatataka Formation, is called the Gaylad Sandstone (Fig. 1B; Freeman et al., 1991).

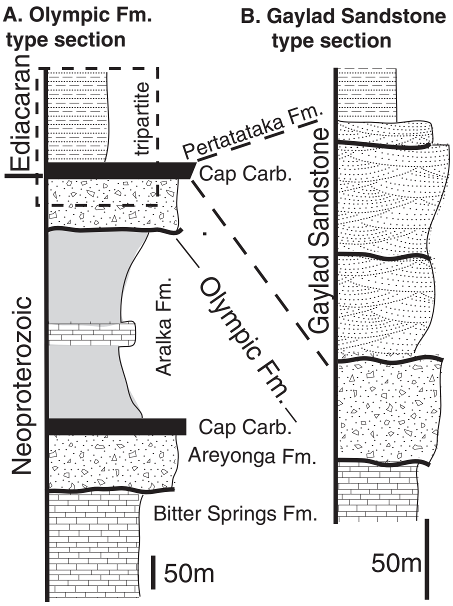

Figure 1. Generalized stratigraphic sections of Neoproterozoic succession in northeastern Amadeus Basin. Type section of Olympic Formation is representative of basinal sections and of Neoproterozoic successions in general. Fm.—formation; carb.—carbonate.

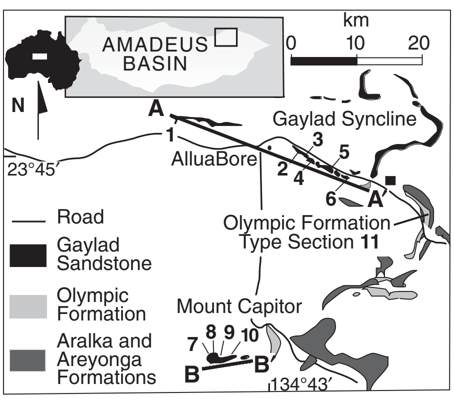

Figure 2. Location map for study area. Transects and measured section numbers correspond to Figure 3.

The bounding and interfingering lateral stratigraphic relations indicate that the cap carbonate and Gaylad Sandstone are coeval. In the type section of the Olympic Formation, the cap carbonate is <5 m thick and interstratified between the glacimarine diamictite of the underlying Olympic Formation (Kennedy, 1996, his figure 2) and shale of the overlying Pertatataka Formation. The cap carbonate thus constitutes the entire interval between the Olympic and Pertatataka Formations that is also host to the Gaylad Sandstone closer to the basin margin. The laminated dolomicrite facies of the cap carbonate splits, forming marly transitions to siliciclastic sediments within at least four unconformity-bounded sequences. In the most landward areas, where glacimarine deposits are absent, the cap carbonate facies overlies fluvial sediments of the Gaylad Sandstone deposited in local incised valleys in gneissic basement. The overlying Pertatataka siltstone concordantly overlies the cap carbonate.

SEQUENCE STRATIGRAPHY OF THE CAP CARBONATE AND GAYLAD SANDSTONE

The sequence stratigraphic approach adopted here utilizes physical discontinuities such as unconformities (sequence boundaries) and flooding surfaces within sedimentary successions to subdivide the rocks into units with time-stratigraphic significance (Christie-Blick et al., 2007). The surfaces pass laterally through changing facies and are for this reason generally oblique to standard lithostratigraphic map units such as those defined by Freeman et al. (1991). The surfaces record base-level lowering of uncertain origin, but likely reflect a combination of global (eustatic) and regional (tectonic and glacioisostatic) phenomena. Sequence boundaries are most evident in marginal marine deposits, where an abrupt basinward or shoreward shift of facies is most obvious. Deeper water settings are insensitive to changes in base level. Surfaces are traced with difficulty in terrestrial facies, and subject to amalgamation because accommodation is commonly limited.

The Gaylad Sandstone is divisible into four unconformity-bounded sequences (Fig. 3), characterized as follows.

1. An abrupt basinward shift in facies (e.g., from calcareous siltstone and lime mudstone and/or siltstone to conglomerate and sandstone). Each sequence boundary can be traced from fully terrestrial deposits at the basin margin into fully marine strata in thicker basinal successions, where relatively thin coarse-grained deposits are inferred to represent event layers (e.g., sediment gravity flows). In both landward and basinward depositional directions, the sequence boundaries either merge with other surfaces or they become cryptic in unifacial successions.

2. Variations in sediment composition serve to differentiate each sequence and imply a genetic association between lithofacies with similar composition. Carbonate-rich sediments above surface S1 (Fig. 3) are overlain above surface S2 by resistant lithologies derived from a preweathered surface (silicified carbonate, chert, and rounded quartz sand and conglomerate). Above surface S3, the rocks are immature, arkosic, and micaceous, consistent with deposition close to a granitic source. Those above surface S4 are typically quartz rich, and inferred to be marine on the basis of tabular stratification.

3. Marked variations in thickness of individual sequences suggest that although the outcrops at the Gaylad syncline are aligned approximately parallel with depositional dip, oblique sections through incised valleys tens of meters in depth are present.

4. Equivalent sequences in a more basinal setting are well exposed within a 5-km-wide panel in the north face of Mount Capitor (transect B–B′ in Fig. 3). The succession at the most proximal western end of the Mount Capitor exposure (MS-7) is lithologically similar to the most distal setting at the Gaylad syncline (MS-6). Four sequence boundaries overlying diamictite of the Olympic Formation are inferred from abrupt changes in grain size and mineralogy at marine erosional surfaces, and a progressive basinward shift in facies similar to that documented at the Gaylad syncline. The four sequence boundaries present at the western end of this panel can be traced in a continuously exposed face eastward as the succession thins by the loss of the siliciclastic fraction and amalgamates in a basinward direction to single cap carbonate interval.

ORIGIN OF GAYLAD SEQUENCE BOUNDARIES

The sequence boundaries documented at the level of the Gaylad Sandstone are unusual in their frequency (4 in an interval <200 m thick). Fewer comparable surfaces are present in >2 km of overlying Neoproterozoic section. These sequence boundaries were produced by rapid, repeated base-level lowering of many tens of meters, interpreted to be related to Marinoan glaciation. Although evaporites at the level of the Bitter Springs Formation were likely mobile at the time, diapirism alone cannot account adequately for the regional development of the sequence boundaries, because

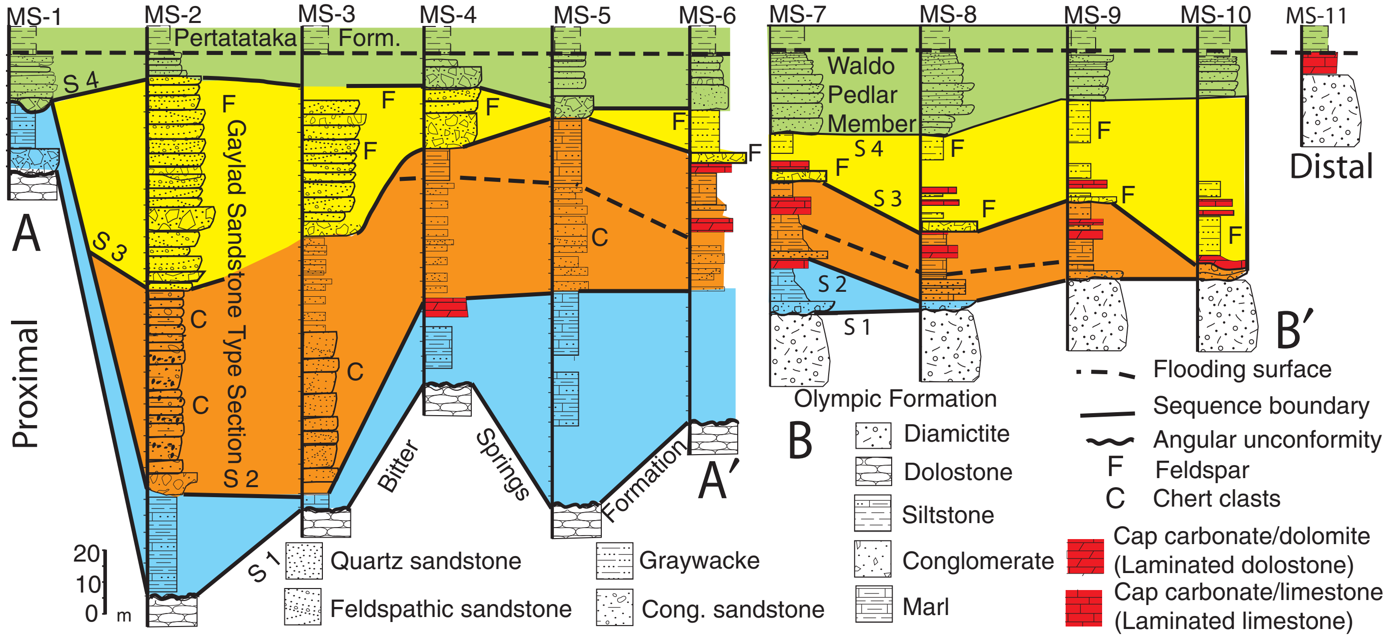

Figure 3. Measured sections from two transects and Olympic Formation type section in northeastern Amadeus Basin (location in Fig. 2). Red units are laminated cap carbonate facies that interfinger within siliciclastic deposits. Cong.—conglomeratic.

the scale of these structures is <20 km; nor can variations in sediment supply. An increase in sediment supply might lead to regression of the shoreline and to gradual upward shoaling, but not to the abrupt basinward shifts in facies that overlie the unconformities along the margin of the Gaylad syncline.

High-frequency (e.g., Milankovitch band) variations of sea level are characteristic of Phanerozoic glacial intervals and may be a feature of Neoproterozoic glacigenic sucessesions as well (Leather et al., 2002). During deglaciation, the focus here, the stratigraphic signal is a product of both eustasy associated with changes in the global ice volume and local isostatic effects that act primarily on length scales of within 200 km of the ice margin. During Cenozoic deglaciation, a complicated stratigraphic signal results from the interplay of these signals and multiple phases of ice advance occurring during overall retreat (Boulton, 1990).

In contrast to the cyclical deglaciation of the Cenozoic, Neoproterozic deglaciation is considered in many current models to have been rapid, verging on globally instantaneous. In this view, in a low-latitude setting (like the Amadeus Basin), where local ice withdrawal led global ice retreat, the 10^5 yr response time of isostatic rebound should have postdated global ice melting superimposing a phase of subaerial erosion on initial glacioeustatic deepening. The existence of multiple sequences within the Gaylad Sandstone suggests considerably more complexity in detail than this simple scenario might imply. Changes in sediment composition between sequences suggest the reorganization of drainage patterns and/or regional uplift between sequences, further supporting a longer term multiphase deglacial process more akin to Phanerozoic stepwise deglaciation. This stratigraphic complexity (and the time implied), evident at the basin margin, is reflected in basinal cap carbonate sections by the interfingering of marl with siliciclastic sediments, and the vertical transition to the typical laminated cap carbonate facies at flooding surfaces.

Condensation through a reduced siliciclastic component is consistent with the carbon isotopic profiles from successive sections along the Mount Capitor transect (Fig. 4). These show a decreasing isotopic range, from >8‰ in the thickest, most complete section to <3‰ in the most condensed section. All sections converge on a similar $\delta^{13}C$ minimum ($\delta^{13}C$ ~–5‰) at the top of the section, with none capturing the return to positive values. This suggests that the controlling mechanisms on $\delta^{13}C$ were longer lived than the duration of cap carbonate deposition as a facies. We suggest that the latter was controlled primarily by the time scale of progradation of siliciclastic sediments of the overlying Pertatataka Formation across the shallow Amadeus Basin and the concomitant dilution of carbonate. This is inconsistent with the short time scale and interrelationship of isotopic values and carbonate precipitation called for in the ocean overturn hypothesis (Grotzinger and Knoll, 1995) or the freshwater lens hypothesis (Shields, 2005), or in the still more rapid (~10 k.y.) and more limited (<–3‰ $\delta^{13}C$) isotopic change required by the kinetic and temperature fractionation models of Higgins and Schrag (2003). The data are also inconsistent with a single catastrophic release of methane (Kennedy et al., 2001), but may be consistent with repeated marine clathrate destabilization tracking deglacial cycles. This is supported in the Marinoan type section, where a phase of methane release is thought to have preceded cap carbonate deposition (Kennedy et al., 2008), similar to Mesozoic examples that proceeded in a series of stepwise releases over >200 k.y. (Cohen et al., 2007).

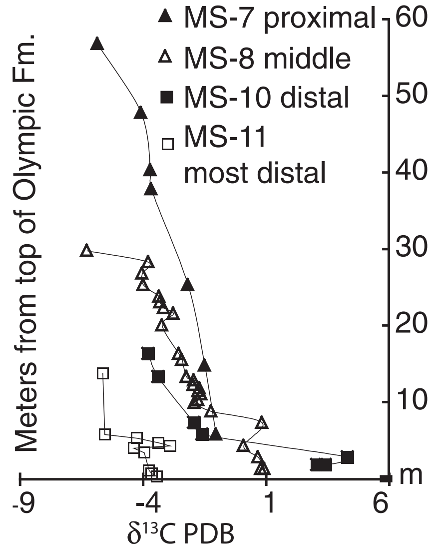

Figure 4. $\delta^{13}C$ values from adjacent sections in Mount Capitor transect from top of Olympic Formation to last carbonate below Pertatataka Formation. Olympic Formation type section is from Kennedy (1996). PDB—Peedee belemnite.

While our data document an exceptionally clear relation between the cap carbonate and basin marginal sediments, they are not unique. An analogous succession was described in the southern Adelaide Foldbelt of South Australia by Dyson and von der Borch (1994), who interpreted tongues of the basal Ediacaran cap carbonate (Nuccaleena Formation) as flooding events within the deglacial Seacliff Sandstone. James et al. (2001) dismissed a single instantaneous depositional event for the Ice Brook cap carbonate of northwestern Canada on the basis of stratigraphic breaks and a karstic unconformity.

CONCLUSIONS

Condensation of carbonate during transgression explains some important properties of cap carbonates, such as their association with glacioeustatic rise, their basin-wide continuity, persistent hemipelagic laminated lithology, their presence in otherwise siliciclastic successions, magnetopolarity reversals, and internal stratigraphic complexity. However, it does not explain a distinctive suite of sedimentary features (Kennedy et al., 2001; Fairchild and Kennedy, 2007) or anomalous isotopic values as low as –40‰ $\delta^{13}C$ (Jiang et al., 2003). Some of these features, like the methane-influenced isotopic values, may be independent of cap carbonate formation and be recorded because the cap carbonate provides a coincidentally timed high-resolution geochemical record. The widespread occurrence of cap carbonates in Neoproterozoic successions may also be a product of the prevalence of broad intracratonic basins in which these deposits are preserved. Thin, laminated cap carbonate-like facies are associated with transgressive shelf successions outside the time frame of snowball events (Bjørlykke et al., 1978). South Australian examples, with negative isotope values, include the Wearing Dolomite and Burr Well Members of the Wonoka Formation (Preiss, 1987). Facies lithologically similar to cap carbonates may be less obvious in Phanerozoic successions because of a greater abundance of skeletal components in the form of reefal and deep-sea pelagic deposits.

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