Neoproterozoic Chuar Group (~800–742 Ma), Grand Canyon: a record of cyclic marine deposition during global cooling and supercontinent rifting

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Abstract

Chuar Group sediments were deposited in a marine cratonic basin synchronously with marine deposition in other basins of western North America and Australia. The top of the Chuar Group is 742 Ma; thus, this succession offers a potentially important record of global climatic and tectonic changes during the mid-Neoproterozoic — a critical time leading to possible global glaciation during supercontinent rifting. Chuar Group cycles and sequences may record glacioeustatic fluctuations during the climatic transition into the Sturtian Ice Age (~750–700 Ma), as well as extension related to the dispersal of Rodinia.

The mid-Neoproterozoic Chuar Group (1600 m thick) is predominantly composed of mudrock with subordinate meter-scale beds of dolomite and sandstone. Facies analysis leads to the interpretation of a wave- and tidal-influenced marine depositional system. Diagnostic marine features include: (1) marine fossils and high local pyrite content in the mudrock facies; (2) mudcracked mud-drapped symmetric ripples and reverse flow indicators in the sandstone facies; (3) facies associations between all facies; and (4) no unequivocal terrestrial deposits. Chuar facies stack into ~320 dolomite- and sandstone-capped meter-scale cycles (1–20 m thick) and non-cyclic intervals of uniform mudrock facies (20–150 m thick). Nearly all cycles have mudrock bases indicating subtidal water depths. Dolomite-capped cycles shallow to peritidal environments (peritidal cycles), some indicating subaerial exposure of varying degrees (exposure cycles). Sandstone-capped cycles shallow to peritidal environments with no evidence of prolonged exposure (peritidal cycles). Non-cyclic intervals represent the deepest water depths (~10s of meters). Peritidal cycles and non-cyclic intervals dominate the lower and middle Chuar Group. In the upper Chuar Group, the percentage of exposure cycles increases, as does the thickness of individual cycles and non-cyclic intervals, and the degree of subaerial exposure in the cycle caps. Correlation of cycles across 10 km shows that many cycles are laterally continuous and some vary in thickness. Some thickness changes are coincident with local and regional extensional structures.

These cycle trends reflect a resolvable mixture of tectonic and climatic controls. Cycles are interpreted to be high frequency and glacioeustatically-controlled based on comparison of cycle character and thickness to Phanerozoic examples, and lateral continuity of many cycles, respectively. Cycles in the lower and middle Chuar Group show characteristics suggestive of low-amplitude sea-level changes similar to meter-scale cycles from global greenhouse climates in the Phanerozoic. Upper Chuar Group cycles show characteristics suggestive of moderate-amplitude sea-level changes similar to Phanerozoic greenhouse–icehouse transition climates, indicating an increase in continental ice volume. Variability in cycle thickness and lateral pinching of some cycle caps is controlled by variable tectonic subsidence, evidenced by thickening trends coincident with

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active rift-related structures, and also by variability in sediment supply. The presence of dolomite caps, in this otherwise siliciclastic succession, is interpreted to be due largely to rapid changes in climate related to short-term glacioeustatically-controlled sea-level changes. Four crude lithostratigraphic sequences (150–775 m thick) are defined based on dolomite-poor to dolomite-rich stratigraphic intervals. Long-term changes in cycle cap-lithology may be driven by long-term changes in climate, similar to short-term glacioeustatic controls on meter-scale cycle cap lithology. Large-scale thickness variability in these sequences is due to local differential tectonic subsidence. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Neoproterozoic; Chuar Group; Grand Canyon; Glaciation; Rodinia; Cyclicity

1. Introduction

The Neoproterozoic era (1000–543 Ma) is an important time in Earth history in terms of tectonic, atmospheric, oceanic and biologic evolution. It included the breakup of the supercontinent Rodinia (Dalziel, 1997), dramatic climatic changes of possibly global scale (Hoffman et al., 1998), and the related biological evolution leading up to the advent of multicellular life (Knoll, 1994). A refined understanding of these phenomena and how they are interrelated requires detailed multidisciplinary studies of Neoproterozoic sedimentary successions.

The Chuar Group provides an important and well-dated reference succession in southwestern North America (Link et al., 1993; Karlstrom et al., 2000) that has potential global importance for understanding the middle Neoproterozoic. A new U–Pb date of 742 ± 6 Ma on an ash at the top of the Chuar Group (Karlstrom et al., 2000), combined with new evidence for a marine origin (this paper), allows us to use Chuar strata to understand mid-Neoproterozoic oceans and to correlate it with other successions in the western US and globally. This is one of several papers that link the Chuar Group to global Neoproterozoic systems (Karlstrom et al., 2000; Porter and Knoll, 2000; Timmons et al., 2001), and will be followed by another paper on the C-isotope stratigraphy of the Chuar Group (see Karlstrom et al., 2000 for preliminary data).

This paper synthesizes new and published data on the sedimentology and stratigraphy of the Chuar Group, and presents new interpretations of depositional environments, depositional models, and controls on sedimentation and stratigraphy. We integrate these results with information from correlative sequences in western North America and Australia in an attempt to view these scattered outcrops of mid-Neoproterozoic marine successions bordering the Laurentian margin as components of a developing miogeoclone.

2. Geologic setting, age control, and stratigraphy

The Chuar Group is exposed exclusively in several right-bank tributaries to the Colorado River in eastern Grand Canyon, Arizona, USA (Fig. 1). This exposure is bounded on the east by the Butte fault zone, and on all other sides by the Great Unconformity and Cambrian Tapeats Sandstone, or locally by the overlying Neoproterozoic Sixtymile Formation or underlying Nankoweap Formation (Figs. 1 and 2). The Chuar Group is gently folded, the main structure being the north-trending Chuar syncline in the eastern part of the exposure belt (Fig. 1). The Chuar Group is also present in the subsurface along the Arizona–Utah border where three wells penetrated as much as 700 m of Chuaria-bearing mudrocks, interbedded with dolomite and sandstone (Rauzzi, 1990; Wiley et al., 1998).

Structural and stratigraphic interpretations of the Chuar Group suggest that it was deposited in a rift basin related to the development of the Cordilleran margin (Sears, 1990; Timmons et al., 2001). Chuar Group sediments were deposited synchronously with movement on a series of N–S striking normal faults as shown by thinning of some strata towards the adjacent Butte fault zone (Fig. 1), subtle thickening of some strata into the hinge region of the Chuar syncline, and thickness changes across subsidiary intraformational faults (Timmons et al., 2001). Although the Chuar Group is preserved today only on the west side of the Butte fault zone, the persistent fine-grained character suggests that the Chuar Group overlapped this fault zone throughout the Chuar deposition, and that the Butte fault did not become emergent until after Chuar deposition.

The age of the basal Chuar Group is constrained by
Fig. 1. Geologic map of the Chuar Group, eastern Grand Canyon (modified from Timmons et al. (2001)) and location of the measured sections in this study.
Table 1
General characteristics of the Chuar Group and overlying sixtymile Formation

<table>
<thead>
<tr>
<th>Member/Formation</th>
<th>Thickness (m)</th>
<th>Main rock types</th>
<th>Fossils(^a)</th>
<th>Lower contact</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sixtymile Fm.</td>
<td>~ 40–60</td>
<td>Silstone, sandstone, intraclastic conglomerate and breccia</td>
<td>None found</td>
<td>Unconformity?</td>
</tr>
<tr>
<td>Walcott Mbr</td>
<td>~ 250</td>
<td>Mudrock and dolomite</td>
<td>Acritarchs, vase-shaped microfossils, bacteria, eukaryotic filaments?</td>
<td>Gradational</td>
</tr>
<tr>
<td>Awatubi Mbr</td>
<td>200–344</td>
<td>Mudrock, sandstone, and dolomite</td>
<td>Acritarchs, vase-shaped microfossils, bacteria, eukaryotic filaments?</td>
<td>Gradational</td>
</tr>
<tr>
<td>Carbon Butte Mbr</td>
<td>45–70</td>
<td>Mudrock and sandstone</td>
<td>None found</td>
<td>Gradational</td>
</tr>
<tr>
<td>Duppa Mbr</td>
<td>30–180(^b)</td>
<td>Mudrock, sandstone, and dolomite</td>
<td>Acritarchs</td>
<td>Gradational</td>
</tr>
<tr>
<td>Carbon Canyon</td>
<td>~ 350</td>
<td>Mudrock, sandstone, and dolomite</td>
<td>Acritarchs, bacteria</td>
<td>Gradational</td>
</tr>
<tr>
<td>Jupiter Mbr</td>
<td>400</td>
<td>Mudrock, sandstone</td>
<td>Acritarchs</td>
<td>Gradational</td>
</tr>
<tr>
<td>Tanner Mbr</td>
<td>185</td>
<td>Mudrock, sandstone, and dolomite</td>
<td>Acritarchs</td>
<td>Angular unconformity</td>
</tr>
</tbody>
</table>

\(^a\) See Table 2 for references.
\(^b\) Specific thickness measured by Ford and Breed (1973).

a Rb–Sr whole-rock date of 1070 ± 70 Ma derived from the underlying Cardenas Basalt and associated diabase dikes and sills (McKee and Noble, 1974; Elston and McKee, 1982). A direct age for the top of the Chuar Group comes from a U–Pb zircon date on an ash bed of 742 ± 6 Ma (Karlstrom et al., 2000). Paleomagnetic correlations of the Grand Canyon Supergroup with other Meso- and Neoproterozoic successions from North America indicate that Chuar deposition occurred between ~850 and 740 Ma, at paleolatitudes of 5°–20° north of the equator (Weil et al., 1999). Stromatolites and microfossil assemblages found throughout the succession also indicate a Neoproterozoic age (Vidal and Knoll, 1983; Vidal and Ford, 1985; Porter and Knoll, 2000).

The Chuar Group is a 1600-m thick, apparently conformable, fossiliferous, unmetamorphosed succession that is composed of dominantly mudrock with subordinate, but recurring, sandstone and carbonate beds. This succession is >85% mudrock, and the interbedded sandstone and carbonate (mostly dolomitic) beds are typically cm-scale to m-scale in thickness (Fig. 2). It is subdivided into the Galeros Formation (lower) and the Kwagunt Formation (upper) at the base of the prominent, thick sandstone unit of the Carbon Butte Member (Ford and Breed, 1973) (Fig. 2). The Galeros Formation is further divided into the Tanner, Jupiter, Carbon Canyon and Duppa Members; the Kwagunt Formation is divided into the Carbon Butte, Awatubi, and Walcott Members (Ford and Breed, 1973). We found these member subdivisions more useful for stratigraphic analysis and summarized these in Table 1.

Some previous workers have interpreted the Chuar Group to represent shallow marine paleoenvironments, and others have interpreted it to represent coastal plain, alluvial plain, and lacustrine depositional settings (Ford and Breed, 1973; Vidal and Knoll, 1983; Reynolds and Elston, 1986; Reynolds et al., 1988; Cook, 1991). Until this study, no facies or stratigraphic analyses had been conducted in terms of basin evolution, nor were depositional models suggested for the majority of the Chuar Group. Attempts to understand Chuar basin evolution are limited by the small areal extent of the Chuar outcrop (exclusively exposed in a 150 km² area in eastern Grand Canyon; Fig. 1), which makes some aspects of traditional basin analysis and reconstruction impossible. This limitation is countered by the excellent vertical exposure of the succession, which provides ample information for interpreting the depositional history of the Chuar basin.

Fig. 2. Stratigraphic column of the Grand Canyon Supergroup and composite measured section of the Chuar Group from this study. The facies distributions throughout the Chuar Group are shown on the right with general environmental interpretations. Facies abbreviations are linked to Table 2.
Table 2  
Summary of sedimentary facies and interpretations, Chuar Group, Grand Canyon

<table>
<thead>
<tr>
<th>Facies</th>
<th>Common sedimentary features and fossils</th>
<th>Facies associations</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dark mudrock facies</td>
<td>Clay-shale/mud-shale/silt-shale, clay is illitic/mixed layer (Dehler, unpublished data). Planar-horizontal laminae to massive, thin-beded lenses of siltstone, sandstone, and silicified ooids and pisoids, mudcracks (≥0.5 cm deep) in heterolithic intervals, abundant local pyrite° (Porter and Knoll, 2000 and references therein), TOC &lt; 8 wt% (Cook, 1991; Karlstrom et al., 2000), meter-scale to 10 meter-scale intervals Vase-shaped microfossils,° algal filaments, acritarchs,° biomarkers (Schopf et al., 1973; Ford and Breed, 1973; Bloeser, 1985; Summons et al., 1988; Porter and Knoll, 2000)</td>
<td>All facies except large-scale crossbedded sandstone facies</td>
<td>Subtidal to intertidal marine, low energy with higher-energy events recorded by coarser grains below fair-weather wavebase. Thicker intervals are deepest water facies in Chuar Group</td>
</tr>
<tr>
<td>Variegated mudrock facies</td>
<td>Mud-shale/mudstone/silt-shale/local clay-shale, clay is illitic/mixed layer (Dehler, unpublished data), silt and sand are composed of quartz. Planar-horizontal and ripple laminae (2-7 mm thick sets) to massive, thinly-beded lenses of siltstone and sandstone, mudcrack casts (cm deep), symmetrical and interference ripples in silt-shale, intraclast impressions common, rare salt pseudomorphs, marcasite nodules in the lower Awatubi Member (Ford and Breed, 1973); meter-scale to 10 meter scale intervals. Acritarchs° (Ford and Breed, 1973 and references therein)</td>
<td>All facies except <em>Izoria</em>—<em>Stratiforma</em> subfacies, coarse dolomite facies, and all Walcott carbonate facies</td>
<td>Subtidal to intertidal marine, low energy wave and tidal influences with rare higher energy events (storms or high tides)</td>
</tr>
<tr>
<td>Laminated and crossbedded sandstone facies</td>
<td>Quartz arenite/subarkose. Commonly massive, parallel to low-angle planar laminations, symmetrical ripplemarks (some with mudcracked mud drapes°), asymmetrical ripple laminae, planar crossbeds and trough crossbeds (≤ 50 cm thick cosets, locally bi-polar°), reactivation surfaces°, rare hummocky crossbeds and associated groove casts, flaser, wavy, and lenticular bedding, rip-up clasts of mudrock and rare dolomite, mudcrack casts common, soft-sediment deformation (load casts, convolute bedding, fluid-escape structures), bed geometry is tabular to lenticular and beds are cm- to meter thick</td>
<td>Laminated and massive dolomite facies and both mudrock facies</td>
<td>Shallow subtidal to intertidal marine, moderate to high-energy (tidal flat to swash) to high-energy (above fair-weather wavebase) to low energy (below fair-weather wavebase)</td>
</tr>
<tr>
<td>Large-scale crossbedded sandstone facies</td>
<td>Quartz arenite/subarkose. Large-scale (1–2 m thick) sets of low-angle tangential crossbeds° and large-scale trough crossbeds. Ripple laminae, soft-sediment deformation, and medium and thinly-beded trough crossbeds found within larger-scale sedimentary structures. Beds tabular to wedge-shaped.</td>
<td>Interbedded with variegated mudrock facies.</td>
<td>Shallow subtidal marine, high-energy shoreface/peritidal (subaqueous tidal and longshore dunes in migrating tidal channels/tidal/longshore sand waves in subtidal area), barrier bar complex?</td>
</tr>
</tbody>
</table>
Table 2 (continued)

<table>
<thead>
<tr>
<th>Facies</th>
<th>Common sedimentary features and fossils</th>
<th>Facies associations&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Laminated dolomite facies</td>
<td>Peloidal dolominitic/dolositite/dolowackestone/rare peloidal boundstone. Smooth to crinkly laminae (1 mm–1 cm thick) with local mm to cm-scale relief, locally contorted laminae, asymmetrical ripple cross- and wavy laminae, massive with wispy laminae, dark mudrock partings common between laminae, symmetrical, asymmetrical and interference ripplemarks, thin beds of flat-pebble conglomerate, mudcracks (cm scale, and 2 horizons with 60 cm-deep cracks), mudcrack casts, rare lenses of ooids and pisoids, locally fetid, rare ‘molar-tooth’ structures and fluid-escape structures; tabular beds 1 cm–1 m thick, rarely lenticular, gypsum, calcite, and silica pseudomorphs after halite(‘?), primary gypsum crystals Filamentous and spheroidal bacteria, vase-shaped microfossils&lt;sup&gt;g&lt;/sup&gt;, acritarchs&lt;sup&gt;g&lt;/sup&gt; (Schopf, 1975; Horodyski, 1993; Porter and Knoll, 2000; this study)</td>
<td>All facies except large-scale crossbedded sandstone facies</td>
<td>Shallow subtidal to supratidal marine, low to moderate energy wave and tidal influences with rare higher energy events (storms or high tides)</td>
</tr>
<tr>
<td>Massive dolomite facies</td>
<td>Doloslitite/dolarenite. Dominantly massive with rare ghost horizontal laminae, vugs, intraclastic breccia, fractures, and dissolution pits</td>
<td>Course dolomite facies, dark mudrock facies, and laminated dolomite facies</td>
<td>Intertidal with modification in the supratidal zone (Cook, 1991; this study)</td>
</tr>
<tr>
<td>Stromatolite buildup facies</td>
<td>Peloidal dolominitic/dolositite boundstone. Crinkly to smooth laminae (0.25 mm ave. thickness) in compound stratiform, domal, and columnar forms, domes range in dia from 1 cm to 2 m wide and 1 cm to 1 m tall, intraclasts abundant (commonly flat pebble and clast-supported), vugs common, calcite pseudomorphs of gypsum crystals near top of unit (Ford and Breed, 1973). Tabular thin to medium beds Wide confluent domes are m-scale domes of <em>Stratiterra</em> and cm-scale confluent columns are <em>Inzeria</em> (Ford and Breed, 1973)</td>
<td>Interbedded with laminated dolomite and dark mudrocks</td>
<td>Shallow subtidal to intertidal marine, dominantly lower energy current with high-energy events. Current was multidirectional based on the equant shape of domes. ‘Littoral to intertidal’ (Ford and Breed, 1973)</td>
</tr>
<tr>
<td>Stromatolite buildup facies</td>
<td>Peloidal doloboundstone. Crinkly to smooth non-isopachous laminae (1–5 mm ave.) in columnar, single and clustered bulbous heads that widen upward (and rarely taper upward of widening) (2–8 cm wide, &lt;15 cm tall), clustered in equant mounds ≤0.6 m in dia; intraclasts of heads common, one to three generations of bioherm, local black chert, one to two tabular beds ≤0.6 cm thick. <em>Baculitosa</em> of Ford and Ford and Breed (1973)</td>
<td>Interbedded with laminated dolomite and variegated to dark mudrocks</td>
<td>Shallow subtidal to intertidal marine, dominantly moderate energy current with high-energy events. ‘Littoral to intertidal’ (Ford and Breed, 1973)</td>
</tr>
<tr>
<td>Stromatolite buildup facies</td>
<td>Peloidal doloboundstone. Crinkly to smooth poorly-defined laminae (0.01–0.03 m thick) in compound stratiform, columnar and domal forms, columns are strait branching (0.04 m dia and up to 0.3 tall), domes are equant to elongate and can be compound (cms to 2 m dia); local mudcracks in top of unit (Ford and Breed, 1973); tabular bed ~2 m thick. <em>Boxonita</em> of Ford and Breed (1973b); <em>Baculitosa</em> of Cloud (1988)</td>
<td>Interbedded with laminated dolomite and variegated to dark mudrocks</td>
<td>Shallow subtidal to intertidal marine, low energy current with higher energy current (wave scour, tidal runoff). ‘Littoral to intertidal’ (Ford and Breed, 1973)</td>
</tr>
</tbody>
</table>
Table 2 (continued)

<table>
<thead>
<tr>
<th>Facies</th>
<th>Common sedimentary features and fossils</th>
<th>Facies associations</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stromatolite build-up facies</td>
<td>Dolomite/dolosiltite. Elongate domal forms (0.3–0.9 m long, 0.3 m wide and 0.15 m tall) with crinkly to smooth laminae (0.8–1.5 mm) of dolomite with alternating laminae of chert, commonly highly contorted, tepee structures and intraclasts common, heads are in tabular beds (0.9 m thick)</td>
<td>Interbedded with laminated dolomite and dark mudrocks</td>
<td>Intertidal marine with supratidal modifications, moderate to high energy (Cook, 1991; this study). Dominantly lower energy current with high-energy events</td>
</tr>
<tr>
<td>Subfacies: flaky dolomite (flaky dol.)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oolitic-pisolitic grainstone facies (dol-oo/pis)</td>
<td>Silica-cemented and -replaced ooids and pisoids. Massive to ghost crossbedded, trough crossbeds, shale intraclasts, 1 cm–1 meter thick tabular to lenticular beds. Filamentous cyanobacteria within grains (Schopf et al., 1973)</td>
<td>Interbedded with laminated dark mudrocks, locally associated with laminated dolomites</td>
<td>Shallow subtidal marine, high-energy beach (thicker beds) and correlative offshore storm deposits and onshore peritidal (thinner lenticular beds) (Cook, 1991; this study)</td>
</tr>
<tr>
<td>Coarse dolomite facies (dol-crs)</td>
<td>Peloidal(?) dolowackestone/grainstone. Planar horizontal and low angle laminae (0.5 cm ave. thickness) and very thin beds to massive, local high-angle planar crossbeds (20 cm thick foreset), hummocky crossbeds in Lava-Chuar Canyon (1.0–3.0 m in dia., 0.5 m thick) and single 1.0 m thick set of low-angle foresets, intraclasts and secondary chert, local dissolution features associated with monomict breccia, medium to thick to massive beds</td>
<td>Overlain by interbedded calcareous dark mudrocks and interbedded laminated and massive dolomite beds</td>
<td>Shallow subtidal marine, moderate to high-energy, near wavebase (subaqueous dunes in migrating tidal channels/tidal sand waves, storm swept, and associated laminites forming in intertidal area)</td>
</tr>
</tbody>
</table>

* See Fig. 2 for stratigraphic distribution of facies.

* Denotes diagnostic marine feature.
3. Facies analysis and paleoenvironmental interpretation

The Chuar Group has been subdivided into nine facies and these are classified using siliciclastic (>50% siliciclastics), carbonate (>50% carbonate) and mudrock nomenclature of McBride (1963), Dunham (1962), and Potter et al. (1980), respectively (Table 2). Salient points, from Ford and Breed’s pioneer work on the Chuar Group (1973 and references therein) and from a detailed study of the Walcott Member by Cook (1991), are integrated into the descriptions and interpretations (Table 2). Many facies appear throughout the succession, and others appear only once. Therefore, facies are listed in decreasing order of abundance between and within each facies category. Details on the distribution of facies are shown in Fig. 2 and Table 2.

3.1. Dark mudrock facies

The dark mudrock facies (~60% of Chuar Group) indicates deposition of organic-rich muds in low-energy, marine subtidal to intertidal environments. This interpretation is supported by mudcrack casts and mudcracks, facies associations, marine fossils and pyrite content. Dark mudrock intervals that do not exhibit mudcracks or mudcrack casts were deposited below fair-weather wave base and represent the deepest water facies in the Chuar Group. The presence of mudcrack casts and mudcracks indicates the periodic submergence, emergence and desiccation of a mudflat. The association of the dark mudrock facies with other facies, interpreted to represent marine environments, strongly suggests related environments of deposition for the dark mudrock facies (Table 2). Fossil types in the dark mudrock facies have been observed in other Proterozoic marine successions (Hofmann, 1977; Vidal and Knoll, 1983; Link et al., 1993; Knoll et al., 1989, 1991) and also suggest a marine environment. The diversity of the fossil assemblage suggests shallow to moderately shallow marine conditions (Vidal and Knoll, 1983). Locally high pyrite abundances have been reported from the Chuar Group (Ford and Breed, 1973; Porter and Knoll, 2000; Shen and Canfield, unpublished data). These pyrite abundances are common in marine and marginal-marine settings and rare in lakes (Berner and Raiswell, 1984; Canfield and Raiswell, 1991).

3.2. Variegated mudrock facies

The variegated mudrock facies (~25% of Chuar Group) is also interpreted to represent deposition in marine subtidal to intertidal environments. These environments are similar to those interpreted for the dark mudrock facies, however, this facies is higher in silt content. This facies also contains greater iron oxide concentrations and less organic matter than the dark mudrock facies (Table 2). The variegation is interpreted to be due to variable post-depositional oxidation processes permitted by the permeability of this silty facies. The environmental interpretation of this facies is also based on mudcracks and mudcrack casts, facies associations, and the presence of marine fossils (Table 2).

3.3. Laminated and crossbedded sandstone facies

The laminated sandstone facies (cm to ~4 m thick) is interpreted to represent deposition by tidal and wave processes in the shoreface and intertidal areas. This interpretation is based on wavy to flaser bedding, an array of bedforms and cross stratification types, associated paleocurrent data, and facies associations.

Wavy to flaser bedding, including mudcracked mud-drapped symmetric ripplemarks, is common in tidal environments and is the result of fluctuating current strengths during tidal or storm deposition on the tidal flat (Fig. 3a; Reineck and Wunderlich, 1968; Klein, 1971; Reineck and Singh, 1980). Paleocurrent analysis of symmetric–ripple–crest trends, dominantly from the Carbon Butte Member (n = 24), and lesser from the Carbon Canyon Member (n = 11) and Awatubi Member (n = 3), shows a broad distribution, yet an overall north-northwest to south-southeast trend (Fig. 4a). This may indicate a shoreline trend that was roughly north-northwest during deposition of the upper Galeros and lower Kwagunt Formations.

The thin- to medium-bedded trough and high-angle crossbed sets indicate high-energy multidirectional, bidirectional, and unidirectional traction currents. Local reactivation surfaces indicate reversing flow, likely in a tidal regime (Fig. 3b). Some of the cross beds could indicate a high-energy shoreface with
dunes and bars, or could represent tidal bedforms. The low-angle and horizontal parallel laminae were likely to be produced in the lower intertidal to subtidal zone by traction flow or suspension settling. The rare hummocky cross stratification indicates that the shoreline was intermittently affected by storms. Paleocurrent analysis of foresets show a radial distribution of maximum dip directions (n = 38), with a dominant flow towards the southeast and bipolar flow towards the northeast and southwest (Fig. 4b).
These data are consistent with the hypothesis of a north-northwest trending shoreline and the sandstone facies recording complex wave and tidal processes. Petrographic comparisons between the Chuar Group laminated sandstone facies and Big Cottonwood Formation sandstones interpreted to represent tidal bundling (Ehlers and Chan, 1999) show striking similarities. Further work needs to be done on Chuar sandstones to demonstrate tidal bundling.

This facies is associated with all other facies and has overlapping sedimentary features with the mudrock facies (wavy to lenticular bedding) and large-scale crossbedded sandstone facies (trough crossbedding). The tabular to lenticular geometry of this sandstone facies indicates that the sand was likely to be distributed in isolated to broad sheets that occupied lower intertidal and shallow subtidal areas.

### 3.4. Large-scale crossbedded sandstone facies

This facies (4–11 m thick) is only present in the basal red sandstone of the Carbon Butte Member (Fig. 2) and is interpreted to represent a combination of tidal- and wave-produced bedforms and tidally produced epsilon crossbeds on a shallow marine shelf. This interpretation is based on the large-scale, low-angle-inclined strata, superimposed smaller-scale sedimentary features within individual cross strata, association with small- to large-scale trough crossbeds, paleocurrent data, and facies associations.

The dominant sedimentary feature in the basal red sandstone is the sharply bounded, thick-bedded, wedge-shaped sets of low-angle cross stratification that are very similar to tidally-produced epsilon crossbedding (Allen, 1963; Clifton, 1983) (Fig. 3c). Individual inclined layers are thinly bedded, commonly undulatory, and show internal ripple laminae. This type of compound stratification is observed in tidally produced epsilon foresets (e.g. Allen, 1963; Clifton, 1983). However, the lack of mud along individual cross-bed surfaces is problematic for a tidal setting. Most descriptions of tidal channel epsilon crossbeds include changes in grain size within the foresets, such as tidal bundles or heterolithic bedding (Allen, 1963; Reineck and Singh, 1980; Clifton, 1983). However, mud-poor tidally produced epsilon crossbeds from the Cambrian Waterfowl Formation are attributed to the predominance of storm-enhanced tidal currents (Cloyd et al., 1990). Mud-poor high-energy tidal deposits also occur in the Lower Proterozoic Mount Guide Quartzite (Simpson and Eriksson, 1991). The lack of mud could also be due to local high tidal current velocity in the lower tidal flat/shallow subtidal area, the presence of an ebb-dominated current, or the scarcity of mud in this part of the system.

The medium-to-large-scale trough cross stratification and its intimate association with the low-angle large-scale crossbeds could indicate the migration of dunes within a tidal channel. Other possibilities are that trough crossbeds reflect dunes within the shoreface of an associated barrier bar, or that all crossbeds are a complex of sandwaves that formed from tidal and/or wave influences in the shallow subtidal and intertidal area.

Limited paleocurrent data of the low-angle large-scale crossbeds and associated trough crossbeds show multi-directional current flow that is interpreted to indicate complex wave and tidal currents. Maximum dip directions of the low-angle large crossbeds trend dominantly to the southeast (n = 16) (Fig. 4c) and indicate either that the edges of migrating channels had trends roughly perpendicular to the flow direction.

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**Fig. 3.** Photos of sedimentary features in selected facies in the Chuar Group: (a) mudcracked, mud-draped symmetric ripples of the variegated mudrock facies. Similar features are found in the dark mudrock facies and the laminated sandstone facies. This suite of features is interpreted to have formed from tidal processes in the intertidal zone; (b) planar tabular crossbeds showing transport to the right (white arrow) and smaller-scale climbing bedform within these foresets showing transport to the left (other white arrow). This suite of features is interpreted to indicate reversing flow direction, likely in a tidal regime; (c) large-scale low angle cross stratification in the basal red sandstone of the Carbon Butte Member. These sedimentary features are interpreted as tidal point bar deposits or possibly tidal dunes or bars; (d) laminated dolomite facies showing crinkly lamination indicating microbial influences on deposition of micrite; (e) large-scale mudcracks in the laminated dolomite facies are an indication of prolonged subaerial exposure and contortion of dolomicrite fill is interpreted as due to compaction and possibly a component of seismic shaking; (f) bedding-plane view of dolomicrite-filled polygonal cracks in the polygonal marker bed of (3c); (g) meter-scale sandstone-and dolomite-capped peritidal cycles in the middle Chuar Group; (h) two meter-scale exposure cycles in the middle Chuar Group. Hammer marks the top of the lower of the two exposure cycles. Meter-scale peritidal cycles stratigraphically above in background.
Fig. 3. (continued)
(N–NE along a shoreline trending N–NW), or that tidally/longshore-drift-produced sandwaves were moving parallel to the shoreline. Maximum dip directions of associated trough-crossbeds (n = 23) show a radial distribution, and imply a complexity of current types that are likely a combination of tidal and wave influences (Fig. 4d). Additionally, cross laminae within symmetric ripple forms in the lower ~2 m of the basal red sandstone show dominantly eastward and lesser westward transport, indicating reverse-flow conditions and flow perpendicular to the interpreted shoreline orientation.

Based on the combined data, the large-scale, low-angle crossbeds are likely epsilon crossbeds, and indicate the migration of tidal channels that were influenced by long-shore drift (e.g. Reinson, 1992). The associated trough crossbeds likely represent dunes, some which were migrating within the interpreted tidal channels, and others that experienced complex currents either in the shoreface or simply within sinuous tidal channels. This combination of wave- and tide-influenced features have been observed in ancient and modern barrier island systems and are most similar to features seen in the barrier inlet area (Reinson, 1992). This facies is intimately associated with the variegated mudrock facies and the laminated sandstone facies, which are interpreted as wave- and tidal-influenced mixed sand–mud tidal flat and offshore equivalents (Table 2).

3.5. Laminated dolomite facies

The laminated dolomite facies (cems to ~3 m-thick) represents marine subtidal and intertidal deposition. The laminated dolomites in the Walcott Member were previously interpreted as shallow subtidal to intertidal marine environments (Cook, 1991). We agree with this interpretation and extend it to apply to all laminated dolomites in the Chuar Group. This interpretation is based on the similarity of a suite of features to those seen in modern and ancient shallow subtidal and intertidal marine environments (e.g. Hardie and Ginsburg, 1977; Esteban and Klappa, 1993; Shinn, 1993; Demmico and Hardie, 1994; Table 2). The most diagnostic features are the diversity and types of lamina, mudcracks, internal upward-shallowing trends, and facies associations.

Laminae (~1–10 mm) range from crinkly to wavy
to planar and form from suspension settling of fine grains, microbial agglutination or baffling (Demmico and Hardie, 1994, and traction flow (Fig. 3d). The crinkly laminations commonly form isolated build-ups (mm to cms tall) and can be recumbent or wildly contorted. Where laminae are associated with ripple marks and within ripple cross-laminated sets, they likely formed from tractional flow produced by tidal or storm processes in the shallow subtidal and intertidal zone.

Within thicker (>60 cm) beds of this facies, an internal upward-shallowing trend is interpreted based on the presence of mudcracks and (or) a massive dolomite layer at the top. The mudcracks are diverse in width (mms to 15 cm) and depth (mms to 60 cm). The larger mudcracks are found in two beds, at least one can be traced across the field area (polygonal marker bed), and are considered to indicate prolonged subaerial exposure (Fig. 3e and f). Facies associations with demonstrably marine facies aid in the environmental interpretation of the laminated dolomite facies (Table 2).

3.6. Massive dolomite facies

This facies (20 cm to ~11 m thick) represents subaerial exposure and modification of other dolomite facies. Birds-eye vugs, intraclastic breccia, and modified sedimentary structures collectively indicate subaerial exposure. Birds-eye vugs are common in the massive dolomite facies and typically form in the upper intertidal to supratidal region (Shinn, 1968). Intraclastic breccia, found irregularly infilling cavities within the massive dolomite facies, are likely due to solution collapse. Associated with the massive dolomite facies are soft-sediment-deformed laminae (wispy laminae) interpreted to represent fluid escape and differential compaction on the upper intertidal and supratidal flat. Massive dolomite facies with a high concentration of exposure features are interpreted to represent zones of prolonged subaerial exposure (Figs. 2 and 6).

3.7. Stromatolite build-up facies

This facies (30 cm–13 m thick) represents marine subtidal and intertidal environments, with some subfacies experiencing subsequent long-term subaerial exposure (Fig. 2, Table 2). Previously,stromatolites in the Chuar Group were interpreted to represent dominantly ‘littoral’ and rare intertidal conditions (Ford and Breed, 1973; Cook, 1991). Our extension of these interpretations is based on the presence of vugs, facies associations, internal upward-shallowing trends, and the occurrence of stromatolite morphotypes identified in the Chuar Group (Table 2) in other marine successions. Inzeria and Stratifera are reported from Precambrian marine successions such as the Uchur-Maya region of Siberia (Serebryakov, 1976), the Beck Springs Dolomite in Death Valley California (Marian and Osborne, 1992), and in the upper Mackenzie Mountain Supergroup and lower Windermere Group in Canada (Narbonne and Aitken, 1995). Baicalia has been reported from Precambrian marine successions such as the Atar Group of Mauritania (Bertrand-Sarfati and Trompette, 1976), the Murky Formation in Canada (Hoffman, 1976), the Uchur-Maya Region of Siberia (Serebryakov, 1976), the Beck Springs Dolomite (Marian and Osborne, 1992), the Australian Supersequence 1 (Hill et al., 2000), the Mackenzie Mountain Supergroup (Narbonne and Aitken, 1995), and Shaler Group (Young, 1981). Boxonia has been reported from marine successions such as the upper Little Dal Group, Canada, the Polarisbreen Group, Spitsbergen, and the Bitter Springs Formation, Australia, as well as from marine Neoproterozoic successions in the Uchur-Maya region, the Olenek Uplift, and the Anabar Massif (Preiss, 1976; Semikhatov, 1976).

3.8. Oolitic/pisolitic grainstone facies

This facies (0.15–2.0 m thick) is interpreted to record shallow subtidal high-energy deposition and coeval offshore and onshore (intertidal) storm and tidal deposits. This facies was previously interpreted as a high-energy environment, specifically shoal and tidal channel deposition (Cook, 1991). Interpretations are based on grain character, rare trough crossbedding, facies architecture, facies associations, and modern and ancient analogues (Cook, 1991; Table 2).

3.9. Coarse dolomite facies

The coarse dolomite facies (6–24 m thick) represents high-energy conditions on a wave-dominated shelf. This interpretation is based on 3-D hummocks,
a diversity of cross stratification types, and a shallow-
ing-upward trend. The hummocks and associated
hummocky cross stratification indicate combined-
flow traction currents generated by storms near fair-
weather wave base (Harms et al., 1975). The associated
high-angle crossbeds are interpreted as the result of
migration of shoreface dunes. The low-angle and hori-
zontal planar laminations were likely produced by
swash or generated by suspension settling and (or)
traction flow.

3.10. Chuar sea vs. Chuar lake

Previous workers have suggested that the Chuar
Group represents lacustrine deposition (Reynolds
and Elston, 1986; Reynolds et al., 1988), although
unequivocal data (diagnostic of a lacustrine setting)
were never reported. The depositional environments
of Precambrian sedimentary successions are difficult
to interpret due to the lack of body fossils and biotur-
bation. We use microfossils, sedimentary structures,
facies associations, and geochemical trends to
discount the lacustrine hypothesis. For example,
shape microfossils and acritarchs similar to
those in the Chuar Group are found in the Draken
Conglomerate Formation, Spitsbergen, which is inter-
preted to be marine based on the dimensions of the
unit (2000 m thick and 650 km wide), sedimentary
facies consistent with a marine platform setting, no
evidence of alluvial deposits, and carbon and sulfur
isotopic values that are suggestive of Proterozoic
seawater (Knoll et al., 1991 and references therein).
Additionally, the Chuar Group microfossil assem-
blage has been found in Proterozoic marine succes-
sions in East Greenland and Svalbard (Vidal and
Knoll, 1983; Knoll et al., 1986), as well as in the
marine Red Pine Shale of Utah (Hofmann, 1977;
Nyberg, 1982a; Link et al., 1993; Crosse et al.,
2000). Vase-shaped microfossils have also been
found in the marine upper Pahrump Group (Link et
al., 1993 and references therein), and the marine lower
Little Dal Group of Canada contains Chuaria circularis
Sedimentary structures and features in the two sand-
stone facies of the Chuar Group are indicative of tidal
processes, and these two facies together are intimately
interbedded with all other facies (Table 2). All Chuar
facies are consistent with a marine setting and there is
no unequivocal evidence of terrestrial deposition in
any facies (e.g. lacustrine and alluvial deposits, paleo-
sols) (Table 2). Carbon isotope data from the Chuar
Group show values consistent with Proterozoic
seawater (Karlstrom et al., 2000 and high local pyrite
concentrations are suggestive of a marine setting (e.g.
Berner and Raiswell, 1984).

The Chuar Group is similar to other Proterozoic
marine successions representing similar depositional
systems. For example, the ‘carbonate-terrigenous’
member of the Crystal Springs Formation and
‘transition beds’ of the Pahrump Group (Link et al.,
1993), the Thule Group of Canada (Hoffman and
Jackson, 1996), and the Burra Group (Walter et al.,
2000) are all thick (1000s of meters), mixed siliciclas-
tic–carbonate shallow-marine successions that are
stromatolite and acritarch bearing. There is no single
modern depositional system that can be used as an
analog for the Chuar Group. However, it contains
carbonate facies similar to those forming today in
the Persian Gulf and Shark Bay, and has features
within the siliciclastic facies that are similar to
many modern tidal flats such as those in the North
Sea and the Sea of Cortez.

3.11. Sediment source

3.11.1. Terrigenous sand and mud
Based on facies analysis, terrigenous sand and mud
were likely brought into the Chuar basin by a combi-
nation of geostrophic, longshore, tidal, and storm
currents, local continental runoff and aeolian trans-
port. There is no evidence for a proximal delta within
the study area, yet it is likely there was one or more,
farther afield, that supplied much of the siliciclastic
detritus to the basin. The overall fine-grained nature of
the siliciclastic sediments in the Chuar Group
suggests low regional relief.

The source of the sand is hard to determine based
on the scarcity of paleocurrent indicators and sand-
stone maturity (quartz arenite to subarkose; Table 2).
The most likely source for the sand, based on avail-
able paleocurrent data and the patchy distribution of
some sand bodies, is longshore currents from distal
deltaic sources and local derivation of the continent
by wind and local run off. The frosted texture of some
sand grains within the mudrock and sandstone
facies suggests a component of aeolian transport.
The fluctuations in siliciclastic grain size and overall siliciclastic flux throughout Chuar deposition could be due to changes in climate and (or) in the proximity to distant deltas.

Preliminary results from XRD analyses of the Chuar mudrocks indicate predominantly illite-rich clays and quartz silt, with intervals of mixed layer clays within the Tanner, Carbon Canyon and Walcott Members (Table 2; Dehler, unpublished data). The mixed-layer clays likely represent input of volcanic ash into the Chuar basin. This is consistent with reported ash layers from the Chuar Group (Elston, 1979; Link et al., 1993; Karlstrom et al., 2000).

3.11.2. Organic component

Planktonic microbes, such as Chuaria circularis, were the likely source of organic material supplied to the basin throughout Chuar deposition (ave. weight% TOC = 0.83; Dehler, unpublished data, Table 2). Organic carbon concentrations in mudrocks are controlled by a combination of carbon influx from organic productivity, post-depositional preservation potential, and dilution by other sediment types (Pederson and Calvert, 1990). The dark mudrock facies had a greater organic carbon influx, better post-burial preservation, and/or lower siliciclastic sediment supply than the variegated mudrock facies.

3.11.3. Carbonate

Sources for Chuar carbonate grains include precipitation induced by microbial activity, abrasion of pre-existing carbonate rocks (detrital mud and peloids), and direct precipitation from seawater as ooids/pisoloids or carbonate mud. The main source of carbonate was likely precipitation induced by benthic microbial communities living in the peritidal realm. The clotted texture observed in the lamina of some stromatolite complexes (peloidal doloboundstones; Table 2) suggests in situ carbonate production (Fairchild, 1980). Filamentous and spheroidal plant fossils in the outer rinds of coated grains in the ooid/pisoloid grainstone facies indicate that micro-organisms played a role in the precipitation of these grains (Schopf et al., 1973).

4. Chuar Group cycles and stratigraphic trends

The traditional subdivisions of the Chuar Group (members and formations; Ford and Breed, 1973 and references therein) are extremely useful for mapping and descriptive purposes, but do not capture potential scales of genetically related strata. We looked at stacking patterns (cycles and lithostratigraphic sequences) towards understanding the genesis of the stratigraphic framework, and many of these patterns crosscut formation and member divisions. Cyclicity in the Chuar Group is recognized at the meter-scale (<1 to ~20 m), and as lithostratigraphic sequences (100-m-scale) (Figs. 5 and 6), and as non-cyclic intervals (20–150 m-thick intervals). Lateral correlation of cycles and lithostratigraphic sequences was possible in intervals of the Chuar Group where lateral exposure and marker units are present (mostly middle and upper Chuar Group). Describing and interpreting Chuar cyclicity has shed light on the controls on sedimentation and stratal patterns, and will aid in interpretation of large-amplitude C-isotope variability recognized in the Chuar Group (e.g. Karlstrom et al., 2000).

4.1. Meter-scale cycles

Meter-scale cycles (~320 total) are present in all members of the Chuar Group (Fig. 3g and h). Two types are observed: sandstone- and dolomite-capped cycles with mudrock bases. Mudrocks are the dominant rock type in all cycles (80–90% of cycle thickness), and sandstone-capped cycles are the most common cycle type (~75%). Sandstone-capped cycles are dominant in the Tanner Member through the lower Awatubi Member (Figs. 3g and 6). Dolomite-capped cycles are common in the Carbon Canyon and lower Duppa Members and dominate the Walcott Member (Figs. 3h and 6). Many cycle caps can be traced laterally where exposure permits (e.g. Fig. 7a and b).

4.1.1. Sandstone-capped cycles

The sandstone-capped cycles (0.15–19 m-thick, 3.6 m ave.) consist of dominantly variegated or dark mudrock facies (2.9 m-thick ave.), overlain by the laminated sandstone facies (0.7 m-thick ave.). The contacts between facies are gradational or sharp
(Figs. 5 and 6). Commonly, the mudstone grades upward into a siltstone and is capped (sharply or gradationally) by the laminated sandstone facies. The sandstone cap, typically the laminated sandstone facies, is always overlain by one of the two mudrock facies and this upper contact with the overlying cycle is gradational to sharp. The sandstone cap typically has parallel horizontal laminae, and exhibits mud partings and mudcrack casts in ~40% of the cycles.

The sandstone-capped cycles represent shallowing from the relatively deeper water mudrock facies (offshore) to higher-energy, shallow subtidal to intertidal environments (shoreface/peritidal area) and are referred to as peritidal cycles. The gradational or sharp contact between the mudrock and overlying sandstone cap indicates a gradual or abrupt regression, respectively. The gradational or sharp contact between the sandstone cap and overlying mudrock of the next cycle indicates gradual deepening (transgressive-prone cycle) and abrupt deepening (fully regressive cycle), respectively.

4.1.2. Laminated to massive dolomite-capped cycles

The laminated to massive dolomite-capped cycle (0.2–19.1 m-thick, 2.7 m ave.) is present in all dolomite-bearing members. The variegated to dark mudrock facies dominates the cycle (2.1 m-thick ave.) and is capped by laminated to massive dolomite facies (0.6 m-thick ave.). This cycle type doubles its thickness in the Walcott Member (Figs. 5 and 6). The contact between the mudrock and dolomite facies is gradational or sharp. The mudrock facies commonly grades upward into a siltstone and is overlain by the dolomite cap. The contact between the dolomite cap and the mudrock of the overlying cycle is gradational or sharp. The dolomite cap is commonly mudcracked and grades upwards into <10s of cms of massive to wispy-laminated dolomite. It is also common for the cap to be either entirely massive or entirely laminated. Several dolomite caps display evidence of prolonged subaerial exposure, particularly in the Walcott Member. The polygonal marker bed in the upper Carbon Canyon Member contains large-scale and
Fig. 6. Meter-scale cycles and lithostratigraphic sequences in the Chuar Group. Intrabasin correlation shows that many marker bed cycle caps are traceable across the field area. The bundling of dolomite-rich and dolomite-poor intervals define four lithostratigraphic sequences. *Southern localities section is a composite section: Tanner Member measured in Basalt Canyon, Jupiter through Duppia Members measured in Lava Chuar Canyon, Carbon Butte Member measured in Kwagunt Canyon, Awatubi and Walcott Members measured in Sixymile Canyon, as was the Sixymile Formation (Timmons et al., 2001). See Fig. 1 for distances between localities.
deep prism cracks (~0.3–0.5 m deep) that form 30–50 cm in diameter polygons (Fig. 3e and f). Several massive dolomite caps are vuggy or have local solution collapse breccias (Cook, 1991).

This cycle type is interpreted to represent shallowing from relatively deeper water (mudrock facies) to peritidal environments (dolomite facies; i.e. peritidal cycles) with some caps experiencing subaerial exposure (exposure cycles) or prolonged subaerial exposure (exposure cycles E1–E7 in Fig. 6). This cycle type can be transgressive or regressive prone.

4.1.3. Pisolite–oolite-capped cycles

Pisolite–oolite capped cycles (<2% of all cycles) are present only in the Walcott Member and are notably thicker than other cycle types (1.5–10.5 m-thick, 6.4 m-thick ave.). The dark mudrock facies dominates the cycle (5.7 m-thick ave.) and is sharply or gradationally overlain by a cap of silicified oolite–pisolite (0.4 m-thick ave., Fig. 5). No exposure features have been observed in the caps. This cycle type is interpreted to represent shallowing from below wave base (dark mudrock facies) into the shoreface–foreshore/peritidal area (peritidal cycle). This is a fully regressive cycle type.

4.1.4. Stromatolite (build-up)-capped cycles

Four unique stromatolite (build-up)-capped cycles (1.5–12 m-thick, 6.5 m-thick ave.) have been identified (Figs. 2 and 5). The variegated mudrock or dark mudrock facies (1.75 m-thick ave.) is sharply or gradationally overlain by laminated dolomite (5 m-thick ave.) and stromatolite subfacies of variable morphotypes (e.g. Figs. 5 and 7a,b). Thin interbeds of mudrock facies and/or laminated dolomite facies sharply to gradationally overlie the stromatolite cap (Figs. 5 and 7a,b).

The *Inzeral/Stratifera* complex and the ‘flaky dolomite’ cycle caps exhibit features indicating subaerial exposure of significant duration (Fig. 2). These two cycles are interpreted as shallowing from offshore into the peritidal zone, followed by prolonged subaerial exposure (exposure cycles). Both of these cycles are sharply overlain by peritidal facies and are therefore transgressive prone.

The *Baicalia* and *Baicalial/Boxonia* caps show no evidence for long-term exposure and are interpreted to represent shallowing from relatively deeper water into the peritidal zone (peritidal cycles). The upper contact with the *Baicalia* cap is sharply overlain by dark mudrocks of the overlying cycle and is a fully regressive cycle (Fig. 7b). The top of the *Baicalia/Boxonia* cap is gradational with mudstone of the overlying cycle and is transgressive prone.

4.1.5. Non-cyclic intervals

Non-cyclic intervals are recognized in all members (<40% of succession) and are defined by uniform mudrock intervals (20–150 m-thick intervals) that lack meter-scale cycles (Fig. 6). The thickness of uniform mudrock intervals is greater in the upper Chuar Group. The lack of meter-scale cycles may indicate deposition in water that was too deep to feel the effects of sea-level change (subtidal missed beats, e.g. Elrick, 1996). Relatively deeper water conditions were likely the result of either differential tectonic subsidence, which kept local depositional centers in relatively deeper water (mud deposition), or long-term sea-level rise.

4.1.6. Controls on meter-scale cyclicity

Possible controls on the development of meter-scale cycles in carbonate and siliciclastic settings include delta avulsion, tidal-channel migration, carbonate tidal-flat progradation, episodic tectonism, climate-controlled changes in terrigenous sediment supply and glacioeustasy. Several of these controls can be ruled out when considering Chuar cycles. Since there is no evidence for a proximal delta, delta avulsion is not a viable control for the sandstone-capped cycles. Tidal-channel migration (e.g. Cloyd et al., 1990) is a viable control on sandstone-and dolomite-capped cycles, however, there are few tidal channels in the Chuar Group and this mechanism cannot explain the majority of cycles. Ginsburg’s (1971) carbonate tidal-flat-progradation model was developed to explain upward-shallowing cycles on the Bahamas platform. It cannot be used to explain Chuar dolomite-capped cycles because the Chuar Group is a dominantly siliciclastic depositional system that lacks a substantial subtidal carbonate factory and also lacks a mechanism to cause cessation of seaward progradation of the carbonate tidal flat (e.g. edge of steep-sided platform). The remaining potential cycle controls are episodic tectonism, climate-controlled changes in terrigenous sediment
supply, and glacioeustasy, of which the latter is favored (see below). Other factors, including local effects from short-term climate change, variable tectonic subsidence, and sediment supply are considered influences on cycle character (i.e. could control uniformity of the cycle thickness, lateral continuity of the cycle, or cycle cap lithology), yet cannot explain overall cycle development.

Episodic fault-controlled subsidence (Cisne, 1986) is an unlikely driver for generating many of the Chuar cycles. For episodic tectonism to be the driver, we would expect cycles to consistently show abrupt deepening (regressive-prone cycle, e.g. Elrick, 1996). Although ~80% of cycles in the Chuar Group are regressive prone, ~20% are transgressive prone. Therefore, if episodic tectonism was the control on regressive-prone cycles, it would be necessary to call upon another control to produce the transgressive-prone cycles, which are in many cases, otherwise identical to the regressive-prone cycles. It would be easier to call upon a mechanism that could produce both transgressive- and regressive-prone cycles. Additionally, Cisne’s (1986) model predicts the thinning of cycles away from the controlling

Fig. 7. (a) Meter-scale cycle correlation in the marker bed interval of the Carbon Canyon Member. See Fig. 1 for location of measured sections and marker bed interval, and see Fig. 5 for explanation of symbols. Lateral discontinuity of some cycles is attributed to changes in paleobathymetry due to local differential tectonic subsidence and/or sediment supply. Note location of Butte fault relative to measured sections and thinning trend towards fault from site 9b to site 9a. (b) West-to-east correlation of Baicalia marker bed across the Chuar syncline. See Fig. 1 for location of measured sections and see Fig. 5 for explanation of symbols. Note that the Baicalia bed is thickest in the axis of the syncline, suggesting differential tectonic subsidence during deposition. Also note lateral continuity of the Baicalia marker bed.
structure. We observe many laterally continuous cycles that are uniform in thickness, and also observe cycles that thin in different directions relative to one another and relative to active structures (i.e. not all cycles thin in the same direction; Fig. 7a and b).

Climatically controlled changes in terrigenous sediment supply, combined with basin subsidence, have been invoked to explain siliciclastic parasequences (or cycles) in Cretaceous nearshore deposits of the Western Interior seaway (e.g. Elder et al., 1994). Although this is a likely mechanism to explain the origin of purely siliciclastic cycles in the Chuar Group, it cannot explain the origin of dolomite-capped cycles (especially exposure cycles) or the occurrence of interbedded dolomite-capped cycles with sandstone capped cycles.

Evidence supporting glacioeustasy as the control on Chuar cycle development includes the predictability of vertical facies patterns, the nature of cycle types, and the lateral persistence of many cycles. Regardless of the cycle cap rock type, >60% of the Chuar Group consists of stacked meter-scale cycles, which are consistently composed of basal mudrock facies (offshore environment) overlain by peritidal facies (which may be modified by subaerial exposure). This cyclic pattern requires a short-term cyclic control. After ruling out all other potential controls on meter-scale cyclicity, glacioeustasy is the only viable control to produce this consistent pattern for all Chuar cycle types.

The Chuar cycle types (peritidal and exposure cycles) are both transgressive-prone and regressive-prone, which is readily explained by glacioeustasy (e.g. Elrick, 1995). Peritidal cycles show deposition in the offshore area during transgression (mudrock facies) followed by deposition in the shoreface–foreshore/peritidal area (laminated sandstone facies, all dolomite facies) during regression. Exposure cycles require a drop in sea level, because marine carbonate-sediment production cannot fill above sea level.

The lateral persistence of many cycles supports a glacioeustatic control (e.g. Grotzinger, 1986), however, the distance of correlation in this study is limited by the areal extent of Chuar outcrop belt (maximum intrabasinal correlation of 10 km). For example, all stromatolite-capped and many laminated dolomite-capped cycles are traceable across the field area. The sandstone caps are generally not distinctive and therefore make poor marker beds, yet the basal red sandstone and white sandstone marker beds are within cycles that are laterally traceable across the field area (Figs. 2 and 6).

Although there is no time control on meter-scale Chuar cycles, the interpreted corresponding fluctuations
in sea level could be of high frequency (10^4 to 10^5 years). This is likely because the range of thicknesses of Chuar cycles approximates those of meter-scale cycles that are interpreted to be glacioeustatically controlled, and attributed to orbital forcing (e.g. Beach and Ginsburg 1978; Heckel, 1986; Goldhammer et al., 1987). Orbitally forced upward-shallowing cycles typically are generated in ~20, 40, 100 and 400 ky frequencies.

4.1.7. Local and climatic influences on Chuar cycle character

Although glacioeustasy is interpreted to have controlled Chuar cycle development, other factors influenced the variability in thickness and laterally continuity of cycles, as well as the variability in rock type of the cycle cap. These factors are variability in local paleobathymetry (controlled by variation in tectonic subsidence and sediment supply) and short-term climate change.

From Chuar Group facies, cycle analyses and intra-basinal correlation, it is apparent that some Chuar cycles change laterally in thickness and the cycle cap is not always laterally traceable (Figs. 5–7a). These observations can be explained by variations in paleobathymetry, controlled by variability in tectonic subsidence and sediment supply. Differential tectonic subsidence is interpreted as an influence on paleobathymetric variability, and hence, cycle character, because some cycle thickness changes are coincident with intraformational faults, and some cycle caps thicken or thin coincident with tectonic structures. For example, the *Baicalia* cycle cap shows multigenerational bioherm growth at localities that are coincident with the Chuar synclinal axis, which has been demonstrated to be controlled by syndepositional movement on the adjacent north-south-trending Butte fault (Fig. 7b; Timmons et al., 2001). Another example is the three prominent dolomite beds in the uppermost Chuar Group. These beds pinch out laterally to the east (towards the Butte fault), suggesting that deposition during upper Walcott time preceded and/or was concurrent with activity along the Butte fault (Elston, 1979; Cook, 1991). These examples are best explained by the generation of differential tectonic subsidence during development of the Chuar syncline, the Butte fault, and the greater Cordilleran rift basin (Timmons et al., 2001). However, the trends of cycle and cycle-cap thickness variability are not always predictable: all cycle caps do not thin or pinch out towards the Butte fault, nor do cycles consistently thicken into the axis of the Chuar syncline (Fig. 7a). This is likely due to the generation of differential fault-generated subsidence along strike of the Butte fault (e.g. ‘porpoising’ of the Chuar syncline axis, Timmons et al., 2001). Many of the cycle and cycle cap thickness changes can be explained by variations in sediment supply. This factor can cause variations in paleobathymetry that result in laterally discontinuous, lenticular bodies of sediment (e.g. ‘island model’ of James (1984), Pratt and James (1986), Kozar et al. (1990)). This factor is a likely influence on cycles where there is no demonstrable relationship with structural features (Fig. 7a).

The variability in cycle-cap rock type (specifically where dolomite facies intermittently cap terrigenous mudrock facies in a succession otherwise dominated by sandstone-capped cycles) can be explained by short-term climate change associated with glacial-interglacial phases and less so by Walther’s Law. The application of Walther’s Law to the scale of individual cycles implies that carbonate, sandstone, and mudrock facies co-existed on the shelf. If this were the case, we should see significant interfingering and mixing of these rock types on a cm-scale. This is observed locally in some of the more heterolithic intervals in the Chuar Group (basal Jupiter, Carbon Canyon, Duppa, and basal Awatubi Members), where the dolomites are silty to locally sandy or have mudrock partings. However, many Chuar dolomites are low in siliciclastic content, do not have mudrock partings, and exhibit sharp contacts with siliciclastic facies.

4.2. Model for generation of dolomite-capped cycles

From the Quaternary record, it is apparent that it was drier at tropical latitudes during glacial intervals than during interglacial intervals (Clapperton, 1993; Broecker, 1996). Soreghan (1997) recognized how these Quaternary low-latitude wet-to-dry cycles might work in concert with glacioeustatic sea-level changes to generate mixed siliciclastic–carbonate upward-shallowing cycles. She suggested that for the late Paleozoic of the southwestern US, low-latitude climatic conditions during high-frequency
Fig. 8. Depositional model for the generation of (short-term) meter-scale cycles in the Chuar Group. During short-term transgressions, high terrigenous sedimentary input suggests a wet climate phase (linked to short-term interglacial phase in low-latitude regions). During short-term regressions, terrigenous input is diminished due to a change to more arid conditions (linked to glacial phase in low-latitude regions). However, this short-term arid climate phase can be over-run by long-term climate change and (or) tectonism, and can result in high terrigenous input during short-term regressions (e.g. Jupiter Member sandstone-capped cycles). See Fig. 6 for the link to lithostratigraphic sequences.

sea-level rise and highstand were relatively wet and siliciclastic sediment supply increased resulting in siliciclastic-dominated deposition in nearshore and offshore environments. During the following sea-level fall and lowstand, the climate was drier and siliciclastic sediment supply was decreased, permitting an increase in carbonate productivity and carbonate cycle cap deposition. Soreghan’s (1997) model implies that Walther’s Law cannot be applied to successions deposited during rapid climate changes.

Using Soreghan’s (1997) model, we suggest the following conditions to explain the deposition of dolomite cycle caps in the otherwise siliciclastic-dominated Chuar depositional system. During short-term transgressions (interglacial phase), the climate in the low-latitude study area was relatively wet, resulting in deposition of up to sand-sized siliciclastic grains in the nearshore and offshore regions (Fig. 8). During the ensuing regression (glacial phase), the climate was dry, resulting in a decrease in the siliciclastic sediment supply (particularly sand-sized fraction), which permitted an increase in carbonate productivity in peritidal environments (Fig. 8). When lowstands were reached, peritidal environments were subaerially exposed. The occurrence of some terrigenous sand and silt in dolomite caps could be due to: (1) reworking of relict grains deposited during
wetter phases; (2) eolian input; and (or) (3) input from
geostrophic or longshore currents.

In this model, we interpret that during all short-term
transgressions and highstands (wet climate), silici-
clastic muds were being deposited offshore (Fig. 8)
with no carbonate production in pertidal environ-
ments. It is only during regressions that siliciclastic
sediment supply decreased enough to permit carbo-
nate production (and during lowstands the carbonates
were subsequently exposed). In the lower and middle
parts of the Chuar Group, many successive cycles are
capped by sandstone. This suggests that over extended
intervals of time, some other mechanism influenced
siliciclastic sediment supply during sea-level fall/
lowstand. Soreghan (1997) suggests that increased
tectonic activity could overprint the effects of glacial–interglacial climatic influences such that the
entire upward-shallowing cycle is siliciclastic in
composition. Tectonic activity could overprint
climate change during Chuar time during generation
of hinterland relief due to faulting. We additionally
suggest long-term climate change as a control on
cycle-cap rock type, as discussed below.

The Soreghan model was developed using icehouse
climate conditions (Pennsylvanian and Quaternary)
when glacioeustatic fluctuations were large (50–
100 m). This model is also relevant during less
extreme climate modes. For example, even during a
greenhouse mode, orbitally induced changes in low-
latitude rainfall patterns has been invoked to explain
meter-scale carbonate-siliciclastic cycles in the
Cretaceous of the Western Interior seaway (Barron
et al., 1985; Elder et al., 1994; Sageman et al., 1997).

4.2.1. Implications of glacioeustasy and rapid climate
change

Almost all of the cycles in the lower and middle
Chuar Group are capped by pertidal facies (pertidal
cycles), less than 2% of the cycle caps display
evidence of prolonged subaerial exposure, and 20%
of all cycles are transgressive prone. In addition, non-
cyclic deeper subtidal mudrock intervals, which are
interpreted as intervals of subtidal ‘missed beats’, are
present throughout the succession. These four key
features are characteristic of cyclic successions of
Cambro-Ordovician, Devonian, Upper Triassic,
and Cretaceous age (e.g. Strasser, 1988; Koerscher
and Read, 1989; Osleger and Read, 1991; Montanez
and Read, 1992; Balog et al., 1997; Elrick, 1995). All
of these Phanerozoic successions were deposited
during greenhouse climate modes when the volume
of continental ice was low to absent (e.g. Fischer,
1982) therefore, the amplitude of the short-term
glacioeustatic sea-level oscillations were low. As a
result of relatively low-amplitude oscillations (less
than 10 m), short-term sea-level fall rates are slow
such that tidal flat progradation rates can keep pace
with or exceed sea-level fall rates generating common
tidal-flat capped cycles. With the slower rise rates,
deeper water conditions are rarely attained such that
cycle basies tend to be composed of relatively shallow
water subtidal facies and deposition during the initial
rise is common (transgressive-prone cycles).

In the upper two members of the Chuar Group, the
character of the cycles changes. Here, the cycles
double their average thickness (from ~3 to ~6 m),
the basal cycle facies are consistently composed of
rocks representing the deepest water facies (dark
mudrock facies interpreted as sub-storm wave-base
deposits), there is a decrease in the number of trans-
gressive prone cycles (<10%), and prolonged subaer-
ial exposure features within exposure cycle caps are
more common (~40% of caps; Fig. 6). These features
are more characteristic of Middle to Upper Ordovi-
cian and Mississippian cycles that have been inter-
preted to form in response to moderate amplitude
glacio-eustatic sea-level oscillations (Elrick and
Read, 1991; Read et al., 1995; Pope and Read,
1998): moderate amplitude oscillations signal an
increase in continental ice volume and a transition
into an icehouse climate mode. This interpretation
of low-amplitude oscillations for the lower and
middle Chuar Group followed by moderate amplitude
oscillations in the upper Chuar Group is consistent
with the age of the uppermost Chuar Group of
742 Ma, which is close to the age of the widespread
Sturtian glaciation (estimated to have occurred
between about 750–700 Ma (e.g. Knoll, 2000; Walter
et al., 2000).

If this interpretation of increasing amplitudes of
short-term glacioeustatic oscillations in the mid-
Neoproterozoic is correct, it has interesting impli-
cations for paleoclimate and paleogeographic
reconstructions of this time interval. Several paleo-
geographic reconstructions for the Meso- and Neopro-
terozoic suggest that the majority of continents lay at
equatorial to low latitudes (Powell et al., 1993; Li et al., 1995; Karlstrom et al., 1999; Evans, 2001), though the lack of paleomagnetic control on the paleolatitude of some continents also permits continental positions at higher latitudes. If the mid-Neoproterozoic world was characterized by dominantly low-latitude continents, then the interpreted low- to moderate-amplitude glacioeustatic oscillations imply not greenhouse/greenhouse–icehouse transition modes, but rather a global climate cold enough to permit small to moderate volumes of continental ice to accumulate on low-latitude continents. The ensuing widespread Sturtian Ice Age then suggests very large continental ice volumes. In fact, the Sturtian Ice Age is the first of at least two widespread glacial events interpreted by Hoffman et al. (1998) to represent a ‘snowball Earth’; a climate mode much more severe than the well known icehouse climate modes of the Phanerozoic Era. Assuming a low-latitude dispersing supercontinent, the Chuar Group records the transition from a very cold Earth (cold enough to accumulate small to moderate volumes of ice on low latitude continents) to a widely glaciated Earth. However, the Chuar Group has no record of tillites or associated incised valleys, and combined with the ~742 Ma date at the top of the succession, this suggests that widespread glaciation post-dates 742 Ma.

Paleomagnetic constraints on continental positions in the Meso- and Neoproterozoic are uncertain enough (Meert, 1999) such that an alternative interpretation can explain the interpreted low- to moderate amplitude glacioeustatic oscillations. If at least one continent lay at relatively high latitudes, then small to moderate amounts of continental ice could accumulate and cold global climatic conditions need not be invoked. In fact, the global climate could be relatively warm and the one or more higher-latitude continent(s) accumulating ice could account for the interpreted low to moderate amplitude glacioeustatic oscillations; more similar to Phanerozoic analogues.

Thus, depending on mid-Neoproterozoic paleogeographic reconstructions, the interpreted low to moderate amplitude glacioeustatic oscillations support either a globally very cold climate (assumes all continents are equatorial to low latitude) leading into the Sturtian Ice Age, or if at least one continent lay at a higher latitude, the global climate could have been warmer. Regardless of continental reconstruction, the interpretation of Chuar Group cycles records a transition from low amplitude to moderate amplitude glacioeustatic sea-level change implying an increase in continental ice volume starting before 742 Ma.

4.3. Lithostratigraphic sequences

Four crude lithostratigraphic sequences (~150–775 m thick) were identified in the Chuar Group and are defined by dolomite-poor to dolomite-rich stratigraphic intervals (Fig. 6). The base of sequence 1 sits above a long-term exposure cycle, E1 (Tanner dolomite, Figs. 2 and 6). Above E1 is ~180 m of sandstone-capped peritidal cycles interbedded with three non-cyclic intervals. The top of this sequence consists of several dolomite-capped peritidal and exposure cycles, including the Inzerial/Stratiforma complex (E2, Figs. 2 and 6). Sequence 2 consists of ~450 m of peritidal cycles (sandstone- and dolomite-capped), and associated non-cyclic intervals. Above this is ~325 m of peritidal (dolomite- and sandstone-capped) and exposure cycles. Sequence 3 is defined by 150 m of stacked peritidal cycles (sandstone caps) and non-cyclic intervals, and capped by ~5 m of the Boxonian/Baicalia complex. Sequence 4 consists of a lower ~200 m dominantly non-cyclic interval overlain by 250 m of peritidal and exposure cycles (all dolomite caps), and non-cyclic intervals. This sequence contains several long-term exposure cycles (E4–E7) (Fig. 6).

These four sequences can be correlated across the field area (Fig. 6) suggesting at least a regional control. The limited age control and limited areal extent of the Chuar Group outcrop belt does not permit further correlation; however, the thickness of the sequences argues for long-term (~10^6 years) controls.

For meter-scale cycles, we interpret the presence of dolomite caps in this siliciclastic-dominated succession as the result of a dry climate associated with lowstands of short-term sea level (glacial phase). If this meter-scale cycle interpretation is correct, then thick stratigraphic intervals that are rich in dolomite caps might suggest long-term dry climate phases, and thick intervals of sandstone-capped cycles might be signaling long-term wet climate phases. These long-term wet-to-dry climate changes might be related to...
long-term changes in continental ice volume (Read et al., 1995; Read, 1998) (Fig. 6).

Intrabasinal correlations between lithostratigraphic sequences show thickness variations across the basin (Fig. 6). The greatest thickness variations are seen in sequences 2 and 3; however, correlation was not possible for the entire succession (Fig. 6). These variations in thickness are likely attributable to differential tectonic subsidence related to regional extensional structures (Timmons et al., 2001; this study).
5. Regional Correlation of Chuar Group

The Chuar Group has been correlated with other successions in North America based on lithostratigraphy, biostratigraphy, and geochronology (Link et al., 1993). The 742 Ma date from the Chuar Group (Karlstrom et al., 2000), when combined with refined paleontologic and stratigraphic information from the Chuar Group (Porter and Knoll, 2000; this study) and new paleomagnetic data from North America and Australia (Karlstrom et al., 1999), allow for an updated synthesis of mid-Neoproterozoic stratigraphic correlations (Fig. 9).

Poorly sorted diamicrites, interpreted to be tillites, cap many of the mid-Neoproterozoic successions and are correlated with Sturtian glaciogenic rocks of Australia (~750–700 Ma; Knoll, 2000; Walter et al., 2000). Although precise dates are lacking for many of these glaciogenic rocks, it is often assumed that they may be synchronous and hence the base of these deposits is often used as a datum for correlations (Fig. 9; e.g. Knoll, 2000). Sturtian tillites are not present in the Chuar Group, the Uinta Mountain Group or the Shaler Group. The Sixtymile Formation, which conformably overlies the Chuar Group, does not resemble a tillite, yet does exhibit paleo-channels tens of meters deep that may indicate effects of base-level change caused by a glacially induced marine regression. A 12% negative shift in C-isotope composition in the upper Chuar Group supports this latter interpretation (Dehler et al., 1999; Karlstrom et al., 2000), because this shift is similar in magnitude to C-isotopic shifts worldwide that precede the Sturtian glaciogenic deposits (Kaufman and Knoll, 1995; Kaufman et al., 1997; Hoffman et al., 1998).

Most of these mid-Neoproterozoic successions are dominantly shallow-marine deposits that are stromatolite-bearing and (or) acritarch-bearing and contain a mixture of siliciclastic and carbonate rock types (Fig. 9). Similar marine rock types and fossil assemblages lead to the concept that all of these successions were connected to a common ocean. Emerging carbon isotope data also imply this connection (e.g. Asmerom, 1991; Narbonne and Aitken, 1995; Dehler et al., 1999; Prave, 1999; Karlstrom et al., 2000; Walter et al., 2000). Additionally, all successions in Fig. 9 contain evidence of synextensional deposition, suggesting that basin formation was related to the break up of the Rodinia supercontinent about 750 Ma (Link et al., 1993; Karlstrom et al., 2000).

Many sedimentary successions can be correlated with the Chuar Group based on isotopic age control (Fig. 9). The lower Little Dal Group, Canada is older than the 779 Ma mafic dike that intrudes the lower part (Heaman et al., 1992). The Coats Lake Group unconformably overlies the Little Dal Group and is subsequently overlain by the Rapitan Group which is younger than the 755 ± 18 Ma granite clast in the tillite (G.M. Ross in Klein and Beukes, 1993). The Shaler Group, Victoria Island is older than the 723 Ma Franklin dikes and younger than the 1.2 Ga basalt (Wanless and Loveridge, 1972; Heaman and Rainbird, 1990). In Australia, the upper Callana Group is younger than the 827 Ma Gairdner dikes and contains the 802 Ma on Rook tuff. The overlying Burra Group has a basal rhyolite of 777 Ma and is capped by the Sturtian tillite of Australia (Walter et al., 2000) (Fig. 9).

Despite the long stratigraphic ranges of Precambrian fossils and stromatolites, biostratigraphy can still be a useful correlation tool. The Chuar Group contains numerous stromatolites, the acritarch, Chuaria circularis, and the vase-shaped microfossil Melanocyclium, all of which are found in other Mid-Neoproterozoic deposits (Fig. 9). Perhaps the strongest evidence for correlation of the Chuar Group with the Red Pine Shale of the Uinta Mountain Group lies in the commonality of a Chuaria/vase-shaped microfossil assemblage (Ford and Breed, 1973; Hofmann, 1977; Bloeser, 1985; Vidal and Ford, 1985; Nyberg 1982a,b; Link et al., 1993). The Little Dal Group of Canada and the Burra Group of Australia also contain Chuaria (Narbonne and Aitken, 1995), as do many pre-Sturtian successions in the North Atlantic region (Hofmann, 1977; Vidal and Knoll, 1983; Vidal and Ford, 1985). Vase-shaped microfossils are also known from the Beck Spring Dolomite and the overlying ‘transition beds’ of the Pahrump Group (Horodyski, 1987), the ‘lower Uinta Mountain Group’ (Nyberg, 1982b), and other localities worldwide (Porter and Knoll, 2000).

Formally named stromatolites in the Chuar Group include Baicalia, Inzeria, Stratifera, and Boxonia (Ford and Breed, 1973; Cloud, 1988), and some or all of these stromatolites are present in other mid-Neoproterozoic successions (Fig. 9). Baicalia is the
most cosmopolitan stromatolite, is found in all stromatolite-bearing mid-Neoproterozoic successions included in Fig. 9, and is a potential index fossil for the lower-mid-Neoproterozoic. *Inzeria* is present within all of the North American successions, *S. fera* is present throughout the Canadian successions, and *Boxonia* is present in the upper Little Dal Group (Fig. 9). The Beck Spring Dolomite and ‘transition beds’, the upper Chuar Group, and the Red Pine Shale have the most fossils and (or) stromatolites in common: these successions likely shared the same sub-basin (Fig. 9).

6. Basin interpretation

The Chuar Group is interpreted to represent a cratonic basin (a basin of extensional origin floored by continental crust inboard of the plate margin (Miall, 1984; pp. 505, 596)) that experienced extension related to the rifting of Rodinia. This basin interpretation is based on: the internal stratigraphy of the Chuar Group; N-trending syn-depositional structural data, the position of the Chuar Group relative to the Sr 0.706 line, and the relationship between the Chuar Group and the underlying Unkar Group. One possible shape of the basin (Fig. 9 inset) follows the Rodinia reconstruction of Brookfield (1993), Karlstrom et al. (1999).

The Chuar Group stratigraphy is similar to that in other cratonic basins, which are characterized by successions that are typically kms-thick, can be strongly affected by regional and global cyclicity, and beds can be traced laterally for kms (Miall, 1984, p. 599). These basins are most commonly filled with shallow marine sediments, as is the Chuar Group. The basins are long-lived features, representing 10s–100s of millions of years (e.g. Klein, 1995), which is a reasonable duration for the Chuar basin (e.g. Karlstrom et al., 2000).

North-trending syn-depositional normal faults are present in the Chuar Group and suggest that the active structural grain during Chuar deposition was parallel to the Cordilleran rift margin (Timmons et al., 2001). These data suggest that Chuar Group deposition was concurrent with the rifting of Rodinia and was affected by E–W extension. The position of the Chuar Group relative to the Sr 0.706 line indicates that it was hundreds of kilometers east of the site of eventual rift separation of Laurentia from western continents (Fig. 9 inset).

Cratonic basins commonly overlie pre-existing Precambrian structures (Bally and Snelson, 1980), which is also the case for the Chuar Group. The Chuar Group overlies ~3 km of Mesoproterozoic sedimentary and volcanic rock (Unkar Group, Fig. 2), interpreted to record pre-Chuar deposition (1.2–1.1 Ga) in an extensional basin that formed at high angles to the Grenville collisional orogen during the assembly of Rodinia (Timmons et al., 2001). The Michigan basin is underlain by part of the Keweenawan rift complex (1.1–1.2 Ga) (Fowler and Kuenzi, 1978) and the Williston basin overlies a Precambrian suture zone (Thomas et al., 1987). Likewise, the Adelaidian basins overlie a series of older sedimentary basins (Preiss, 2000). Klein (1995) hypothesizes that the formation of cratonic basins is coeval with the breakup of supercontinents, which we postulate was the fate of Rodinia during Chuar deposition: the Chuar basin likely formed as a downwarped part of the crust corresponding to weaknesses inherited from underlying Precambrian structure (extensional Unkar basin) and responded to extensional stresses applied by the distant rifting of Rodinia.

The mid-Neoproterozoic deposits distributed across western North America and Australia were all likely sub-basins within a larger basin that predated the Cordilleran miogeocline (e.g. Ross, 1991a; Fig. 9). These sub-basins likely formed in response to extension and were drowned by marine waters as Rodinia broke apart. The relative lack of Neoproterozoic mafic intrusive rocks (~770–720 Ma) associated with mid-Neoproterozoic successions in the south and central western US. (Death Valley, northern Arizona, northern Utah), as opposed to the coeval successions in Canada and eastern Australia, indicate that successions were likely deposited in distinctive tectonic sub-basins within the greater intracratonic rift basin (Fig. 9).

7. Conclusions

1. Detailed sedimentologic and stratigraphic analyses indicate that the Chuar Group is marine (not lacustrine). Diagnostic marine features include: (i)
marine fossils and high local pyrite content in the mudrock facies; (ii) mudcracked mud-draped symmetric ripples and reverse flow indicators in the sandstone facies; (iii) facies associations between all facies; and (iv) no unequivocal terrestrial deposits. Fine-grained siliciclastic facies (and lesser dolomite facies) dominate the Chuar Group and imply a wave- and tide-affected marine depositional system. Local basin shape is unknown due to the limited areal extent of the Chuar Group and limited paleocurrent data, yet the shoreline likely had a N-NW trend (at least during middle and late Chuar deposition) and was part of a low-energy, low-relief shelf.

2. Meter-scale cyclicity was likely controlled by high-frequency glacioeustasy. The nature of cycles in lower and middle Chuar Group (dominantly peritidal, thin non-cyclic intervals, lower percentage of exposure cycles showing prolonged subaerial exposure, more transgressive-prone cycles) suggests low-amplitude sea-level changes. The nature of upper Chuar Group cycles (fewer peritidal cycles, thicker non-cyclic intervals, greater percentage of prolonged subaerial exposure cycles, fewer transgressive-prone cycles) suggests moderate-amplitude sea-level changes. This change suggests an increase in continental ice volume between middle and upper Chuar time, and combined with the 742 Ma age of the top of the Chuar Group, suggests that the period ~742 Ma records a global climate change leading to the Sturtian Ice Age. Differential tectonic subsidence, variations in sediment supply, and rapid climate change associated with glacioeustasy controlled the lateral continuity of cycle caps, cycle thickness, and rock type of cycle cap.

3. Lithostratigraphic sequences in the Chuar Group may reflect long-term changes in climate and also tectonic influences. Dolomite-poor to dolomite-rich sequences may be linked to long-term wet and dry phases, respectively. Large-scale thickness variations suggest that differential tectonic subsidence was a control on local accommodation variability.

4. The Chuar basin was a cratonic basin that responded to distributed extension during breakup of Rodinia. It correlates (microfossils, stromatolites, facies interpretations, isotopic dates, and C-isotopes) to mid-Neoproterozoic sequences in western North America and Australia, suggesting the possibility of a greater marine cratonic basin with distinct sub-basins.

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