Siliciclastic sedimentation in the interlude between two Neoproterozoic glaciations, Mirbat area, southern Oman: A missing link in the Huqf Supergroup?

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ABSTRACT

The Huqf Supergroup in Oman contains an exceptionally well-preserved and complete sedimentary record of the Late Neoproterozoic era, including the oldest components in some of Oman’s hydrocarbon plays. Outcrops of the Huqf Supergroup in northern and central Oman are now well-documented. However, a key succession in the Mirbat area of southern Oman, the Mirbat Group, which includes a stratigraphic interval missing elsewhere in the Arabian Peninsula, remains poorly understood. The <1.5 km-thick Cryogenian (850–635 Ma) Mirbat Group comprises an essentially continuous succession of little-deformed sedimentary rocks containing two glacial intervals separated by c. 1 km of non-glacial marine deposits. The lowermost glacial interval (Ayn Formation) occupies deep paleovalleys incised into crystalline basement. The overlying Arkahawl Formation records at its base a major post-glacial transgression over the previous basin margin and a 300 to 400 m-thick turbidite complex consisting of 1 to 5 km-wide, coarse-grained depositional lobes embedded vertically and laterally in fine-grained distal turbidite fan deposits. Ayn Formation paleovalleys continued to serve as sediment transport routes for the coarse-grained turbidite complexes of Arkahawl times. The turbidite complex deposits gradationally pass up into a c. 500 m-thick unit of distal-marine mudstone and siltstone. The overlying c. 100 m-thick Marsham Formation records highstand deposition and the pulsed progradation of shallow-marine and fluvial deposits over offshore mudstone and siltstone in the approach to a second glaciation, represented by the Shareef Formation. The sedimentary succession described in this paper is believed to largely fill the stratigraphic gap present between the Ghubrah and Fiq formations in the Al Jabal al-Akhdar in northern Oman represented by an unconformity.

INTRODUCTION

The Late Neoproterozoic era (850–542 millions of years ago) was punctuated by several major geological events, some of which are without parallel in the Phanerozoic. These major events include the break-up of the Rodinian Supercontinent and subsequent assembly of eastern Gondwana (Torsvik et al., 1996), some of the most extensive glaciations in Earth history (Harland, 1964; Hoffman and Schrag, 2002), the largest known perturbations of the carbon and sulphur cycles (Halverson et al., 2005; Hurtgen et al., 2005), the rise of atmospheric and oceanic oxygen (Kennedy et al., 2006; Fike et al., 2006), and the priming of complex animal life (Knoll, 2003). The Huqf Supergroup of Oman contains an exceptionally well preserved record of the time interval between the Cryogenian and the Early Cambrian (Brasier et al., 2000; Rieu et al., 2007b; Allen, 2007), including some of the world’s best examples of Neoproterozoic glaciations (Allen et al., 2004; Rieu et al., 2006), the largest known perturbation of the carbon cycle in Earth history (Burns and Matter, 1993; Le Guerroué et al., 2006b), oxygenation of the Ediacaran ocean (Fike et al., 2006) and the Global Stratotype Sections and Points for the Precambrian-Cambrian boundary (Amthor et al., 2003).

The current intense interest in the Neoproterozoic era (Hoffman and Schrag, 2002; Fairchild and Kennedy, 2007) and, as such, in the Huqf Supergroup (Allen, 2007), invites a detailed study of the excellently exposed, but poorly documented Mirbat Group in southern Oman (Figure 1). This siliciclastic-dominated sedimentary succession, formerly known as Mirbat Sandstone Formation (Qidwai et al., 1988), is of particular interest as it comprises a c. 1.5 km-thick, essentially continuous succession framed by the deposits of two Neoproterozoic glaciations (Ayn and Shareef formations, Figure 2). The intervening c. 1 km-thick succession of marine sedimentary rocks (Arkahawl and Marsham formations) represent a period between an older Cryogenian and younger Cryogenian
glaciation that has not been preserved elsewhere in Oman (Rieu et al., 2007b) and that is therefore important in reconstructing the geological history of the Cryogenian in this part of the Arabian Peninsula. The succession described in this paper may also provide an interesting comparison with ‘interglacial’ or ‘non-glacial’ successions between older and younger Cryogenian glaciations found elsewhere around the world, including those of the Windermere Supergroup of the McKenzie Mountains, northwest Canada (Narbonne and Aitken, 1995; Day et al., 2004), the Umbertana Group of the Adelaide Geosycline, South Australia (Preiss and Forbes, 1981; Sohl et al., 1999) and the Polarisbreen Group in Svalbard, Norway (Fairchild and Hambrey, 1995; Halverson et al., 2004).
The Huqf Supergroup sedimentary rocks are furthermore interesting from a petroleum geology point of view, as they are components in proven hydrocarbon plays (Lake, 1996; Vroon-ten-Hove, 1997; Cozzi and Al-Siyabi, 2004). The understanding gained from surface exposures is thus especially valuable.

The sedimentary succession in the Mirbat area has been described by a number of authors (Lees, 1928; Beydoun and Greenwood, 1968; Platel et al., 1987; Qidwai et al., 1988; Platel et al., 1992; Kellerhals, 1993; Kleiber, 1993). In a preliminary study, Kellerhals (1993) interpreted the Arkahawl and Marsham formations as pro-delta and shallow-marine deposits, respectively, and first recognized diamictites at the top of the Mirbat Group (Shareef Formation). Several other comprehensive sedimentological studies have focused mainly on the lowermost glaciogenic deposits of the Ayn Formation (Kellerhals and Matter, 2003; Rieu et al., 2006).

Figure 2: Overview of the Neoproterozoic stratigraphy of Oman showing correlations between the Huqf Supergroup and the Mirbat Group after Rieu et al. (2007b). Previously used nomenclature for the Mirbat Group is shown in italics. Ages from Gass et al. (1990), Leather (2001), Amthor et al. (2003), Worthing (2005), Mercolli et al. (2006), Rieu et al. (2007b) and Bowring et al. (2007). See text for discussion. Correlations of the Mirbat Group based on Rieu et al. (2007b).
In this paper we provide a sedimentological analysis of the largely undocumented marine succession (Arkahawl and Marsham formations) above the Ayn Formation, and document the second glacial interval (Shareef Formation) at the top of the succession. Our more comprehensive study includes a new detailed geological map of the complete Mirbat Group and sedimentological logs of well-exposed sections through the Arkahawl, Marsham and Shareef formations (Figure 3). The results reveal a variety of facies that include mudstone, siltstone, sandstone, conglomerate and diamictite. In total six different units are distinguished comprising distal-marine (units A1 and A3), proximal and distal turbidite (unit A2), shallow marine (unit M1), fluvial (unit M2) and glaciomarine (unit S1) facies associations (Figure 4; Table 1).

THE HUQF SUPERGROUP OF OMAN

Neoproterozoic rocks of the Huqf Supergroup crop out extensively in northern (Al Jabal al-Akhdar and Saih Hatat), central (Al Huqf area) and southern (Mirbat area) Oman and are also penetrated by many boreholes (Figures 1 and 2). The Abu Mahara Group of the Huqf Supergroup in northern Oman is thought to record two Cryogenian glacial epochs represented by the older Cryogenian Ghubrah Formation diamictites (723 ± 16/-10 Ma, Brasier et al., 2000; 712 ± 1.6 Ma, Leather, 2001; 714.2 ± 0.6 Ma, Bowring et al., 2007) and the younger Cryogenian Fiq Formation glaciogenic deposits (Leather et al., 2002; Allen et al., 2004; Rieu et al., 2007a), which are in part younger than 645 Ma on the basis of detrital zircon ages (Allen, 2007; Bowring et al., 2007). The Fiq Formation is restricted to the depocenter of a rift basin and is overlain by the Nafun Group, which at the base contains the transgressive Hadash Formation ‘cap’ carbonate (Allen et al., 2004).

The Nafun Group (McCarron, 2000), thought to be Ediacaran (c. 635–542 Ma) in age (Allen and Leather, 2006; Le Guerroué et al., 2006b), strongly oversteps basement-cored basin margins demonstrating a major sea-level rise at its base. Above the Hadash Formation, the Nafun Group contains two

| Table 1 |
| --- |
| Overview of units and lithofacies in the Arkahawl, Marsham and Shareef formations of the Mirbat Group. |

| Unit | Lithofacies (LF) | Occurrence |
| --- | --- | --- |
| **Arkahawl Formation** | | |
| A1: Distal-marine deposits | interbedded mud, silt and sandstone | very common |
| A2: Turbidite complex (sandstone bodies) | A2-1: thick-bedded graded sandstone | very common |
| | A2-2: planar cross-stratified sandstone | rare |
| | A2-3: conglomerate | rare |
| | A2-4: diamictite | very rare |
| | A2-5: thin-bedded sandstone | common |
| | A2-6: siltstone | rare |
| | A2-7: interbedded mudstone, siltstone and thin sandstone | very common |
| A3: Distal-marine deposits | A3-1: laminated mudstone and siltstone | very common |
| | A3-2: graded sandstone | very rare |
| **Marsham Formation** | | |
| M1: Shallow-marine deposits | M1-1: mudstone and siltstone | very common |
| | M1-2: lenticular, wavy and flaser bedded sandstone | very common |
| | M1-3: planar low-angle stratified sandstone | common |
| | M1-4: deformed massive sandstone | rare |
| | M1-5: coarse-grained lag-deposits | very rare |
| M2: Fluvial deposits | M2-1: massive sandstone | very common |
| | M2-2: horizontally and trough cross-stratified sandstone | common |
| | M2-3: conglomerate | common |
| **Shareef Formation** | | |
| S1: Glaciomarine deposits | S1-1: diamictite | very common |
siliciclastic-to-carbonate cycles, starting with the marine shales and sandstones of the Masirah Bay Formation (Allen and Leather, 2006) that pass gradationally up into the prograding carbonate ramp of the Khufai Formation. A major basin deepening is recorded by the siltstone- and mudstone-dominated Shuram Formation (Le Guerroué et al., 2006a), which passes up gradationally into the carbonate ramp of the Buah Formation (Cozzi et al., 2004). The Nafun Group is overlain by the Ara Group carbonate-evaporite succession (Schröder et al., 2003; Schröder et al., 2004) in the Al Huqf area and subsurface salt basins in southern Oman, within which is the Ediacaran-Cambrian boundary (542 ± 0.6 Ma, Amthor et al., 2003), or by the time-equivalent Fara Formation in the Al Jabal al-Akhdar.

THE MIRBAT GROUP, HUQF SUPERGROUP

The Neoproterozoic sedimentary succession in the Mirbat area was formerly referred to as the Mirbat Sandstone Formation, comprising Lower, Middle and Upper members (Ayn, Marsham and Arkahawl members) (Qidwai et al., 1988; Platel et al., 1987, 1992; Kellerhals and Matter, 2003). Based on distinct lithological differences and the large-scale sequence stratigraphic framework, it has been proposed by Rieu et al. (2007b) to informally revise this nomenclature. The former Mirbat Sandstone Formation will be referred to as the Mirbat Group and its former members are raised to formation status and are named the Ayn, Arkahawl and Marsham formations (Figure 2). A new stratigraphic unit at the top of the Mirbat Group has been introduced as the Shareef Formation.

Outcrop area

The Mirbat Group crops out along a 20-km-long, NE-striking escarpment 80 km east of the town of Salalah near the village of Mirbat (Figures 1 and 3). It comprises <1.5 km of well-preserved, weakly deformed, mainly siliciclastic sedimentary rocks that generally dip c. 10° to the northwest. The Mirbat Group unconformably overlies a complex crystalline and metamorphic basement comprising metasediments and highly deformed, banded gneisses intruded by tonalites, gabbros and various kinds of dykes (Platel et al., 1992; Hauser and Zurbriggen, 1994; Worthing, 2005; Mercolli et al., 2006). At the top, the Mirbat Group is truncated by a sub-Cretaceous unconformity, overlain by Cretaceous to Tertiary carbonates. Towards the northeast this unconformity cuts progressively deeper down in the stratigraphy. A post-depositional sub-vertical fault in the northeastern part of the outcrop belt has caused juxtaposition of the lower part of the Mirbat Group and basement with the sub-Cretaceous unconformity (Figure 3).

Stratigraphy

The Mirbat Group is an essentially continuous succession of sedimentary rocks, containing two glacial intervals (Ayn and Shareef formations) separated by c. 1 km of non-glacial marine deposits (Arynahwals and Marsham formations). The Ayn Formation, at the base of the Mirbat Group, comprises <400-m-thick glacially influenced basin-margin deposits, comprising intervals of glacimarine rain-out diamicite and gravity-flow sediments, alternating with intervals of fluvial and deltaic sandstone and conglomerates (Kellerhals and Matter, 2003; Rieu et al., 2006). The Ayn Formation is preserved within several paleovalleys eroded into the basement, which is reflected by its irregular along-strike distribution pattern (Figure 3). The overlying ‘cap’ carbonate transgresses over the basement highs that bounded the previous depocenter (Rieu et al., 2006).

The overlying kilometer-thick succession of marine sandstones and shales of the Arkahawl and Marsham formations form, in contrast to the Ayn Formation, a laterally continuous unit throughout the outcrop area, covering all previous basement relief (Figure 3). Recently, Rieu et al. (2007c) used mineralogical and geochemical proxies to demonstrate that cool and dry conditions prevailed during deposition of the Ayn Formation, and that the Arkahawl and Marsham formations recorded relatively warm and humid conditions during the non-glacial interlude. As demonstrated by earlier work (Kellerhals, 1993; Kellerhals and Matter, 2003; Rieu et al., 2006), and confirmed by this study, the general sediment transport direction in the Mirbat Basin was to the northwest.
Figure 3: (a) Geological map of the outcrop belt of the Mirbat Group showing exposed and interpreted contacts based on field mapping; (b) along-strike profile based on logged sections A-J, with mapped and interpreted contacts.
Correlation

The correlation of the Nafun Group between the Al Jabal al-Akhdar and the Al Huqf regions, and across the Oman subsurface salt basins, is now well-established on the basis of both lithostratigraphy and chemostratigraphy (Burns and Matter, 1993; Cozzi and Al-Siyabi, 2004; Le Guerroué et al., 2006a). The exact relationship between the Abu Mahara Group of northern Oman and the Mirbat Group of southern Oman, which are 1,000 km apart, is less certain (Figure 2). The Mirbat Group unconformably overlies > 700 Ma old crystalline and metamorphic basement (Gass et al., 1990; Worthing, 2005; Mercolli et al., 2006) and is bounded at the top by a sub-Cretaceous unconformity. The most recent study (Rieu et al., 2007b), making use of detrital zircon geochronology, chemostratigraphy and correlations with nearby subsurface data from the South Oman Salt Basin, proposed that the Mirbat Group is time-equivalent to the Abu Mahara Group. Taking into account the uncertainty in correlating glacial events over large distances such as between northern and southern Oman, it should be noted that the Abu Mahara Group in northern Oman, like the Mirbat Group, records two distinct glacial epochs (Le Guerroué et al., 2005).

The Ayn Formation is thought to record an older Cryogenian glaciation that occurred at <722 ± 12 Ma on the basis of detrital zircon geochronology (Allen, 2007; Rieu et al., 2007b) and may therefore be equivalent to the c. 714 Ma-old glaciation represented by the Ghubrah Formation of northern Oman. The younger glaciation, partly

Figure 4: Composite section through the Mirbat Group with a summary of facies associations and depositional environments.
represented by the Shareef Formation, has been correlated with the glacial epoch represented by the Fiq Formation. The age of the Fiq Formation (in part < 645 Ma) is consistent with glaciation taking place near the end of the Cryogenian, as recorded in the Ghaub Formation of northern Namibia and the Nantuo glaciation of South China, both of which terminated close to 635 Ma (Hoffmann et al., 2004; Condon et al., 2005).

Although it is most parsimonious to correlate the Ghubrah-Ayn glacials and the Fiq-Shareef glacials, as proposed by Rieu et al. (2007b) and Allen (2007), it is wise to bear in mind that such correlation is not strictly required by the available geochronological data, and the possibility remains that the base of the Mirbat Group is significantly younger than c. 700 Ma.

As proposed by Rieu et al. (2007b), the Arkahawl and Marsham formations may represent the non-glacial interlude between older and younger Cryogenian glaciations that has not been recognized elsewhere in this part of the Arabian Peninsula, due to the presence of an unconformity between the Ghubrah and basal Saqlah member of the Fiq Formation in northern Oman (Le Guerroué et al., 2005) and to the presence of Nafun Group directly above the 800 Ma-old volcanics and volcaniclastics of the Halfayn Formation in the basement high of the Al Huqf area (Allen and Leather, 2006). The thickness of the Arkahawl and Marsham formations (c. 1 km) suggests that they may record a long period of accumulation, but age constraints are too poor to allow the duration of the non-glacial interlude to be estimated.

**ARKAHAWL FORMATION**

The Arkahawl Formation is approximately 900 m thick and starts with the ‘cap’ carbonate or, where absent, the glaciogenic deposits of the Ayn Formation or basement. The Arkahawl Formation has been subdivided into three units (A1 to A3; Figure 3). At the base is <40 m of distal-marine siltstone and mudstone of unit A1, which is overlain, and in places eroded, by unit A2. Unit A2 is a 300 to 400 m-thick turbidite complex comprising several sandstone bodies ‘embedded’ in fine-grained deposits. The upper part of the Arkahawl Formation, unit A3, is represented by c. 500 m of distal-marine siltstone and mudstone.

**Unit A1: Distal-marine Deposits**

In the largest part of the outcrop belt the base of the Arkahawl Formation is characterized by a < 40 m-thick unit of distal-marine facies. This unit is laterally continuous throughout the study area (Figure 3), although locally deeply incised into, or even completely eroded, by the overlying sandstone of the proximal turbidite facies association (unit A2).

**Lithofacies description**

*Interbedded mudstone, siltstone and sandstone* (LF A1-1) comprises grayish laminated and graded siltstone and mudstone, with white and pink weathering colors (Figure 5a). In the upper part of this unit interbedded thin (10–30 cm) graded sandstone beds similar to those of LF A2-5 occur, becoming increasingly abundant and thicker towards the contact with the overlying sandstone of the turbidite complex (unit A2). Here, soft-sediment deformation is commonly observed, including dm-scale synsedimentary folds and disrupted sandstone beds.

**Interpretation**

The gradual transition of this unit into the overlying turbiditic sandstones of unit A2 suggests that unit A1 was also deposited in a marine environment. The general fine-grained character of this unit indicates that sedimentation was dominated by background hemipelagic sedimentation and/or settling from suspension of distal, low concentration turbidity currents (Pickering et al., 1989; Stow et al., 1996). Graded sandstones at the top of this unit are interpreted to represent episodic mass transport by turbidities (see LF A2-5) and their upward-increasing abundance reflects the gradual transition into the turbidite-dominated middle interval of the Arkahawl Formation (Unit A-2).
Figure 5: Field photographs of representative lithofacies in the Arkahawl Formation:
(a) graded mudstone, siltstone and sandstone (LF A1-1) with folded sandstone layer, ruler for scale;
(b) coarse-grained sandstone channel (facies A2-1) scouring into LF A1-1 (section H), hammer for scale;
(c) two fining-upward sandstone bodies predominantly consisting of proximal turbidite facies LF A2-1, A2-2 (foreground at base of logged section, and middle-ground at top of logged section), alternated with units of outer fan facies (LF A2-7).
See facing page for continuation.
Figure 5 (continued):
(d) graded, gravelly sandstone showing horizontal stratification and large-scale cross-stratification (LF A2-3) in section G, field compass for scale;
(e) graded sandstone with liquefaction structures and rip-up siltstone clasts (LF A2-1) and thin siltstone bed (facies A2-6), hand lens for scale;
(f) graded sandstone with massive (Ta) and parallel-laminated (Tb) divisions (LF A2-1);
(g) conglomeratic channel-fill (LF A2-2), hammer for scale;
(h) thin-bedded turbidites interbedded with siltstone and mudstone of the outer fan facies association (LF A2-7), hammer for scale;
(i) rippled tops (arrow) of graded sandstone interbedded with siltstone (LF A2-5 and LF A2-6), hand lens for scale.
Unit A2: Turbidite Complex

Unit A2 is composed of several discontinuous sandstone bodies (1 to >5 km-wide and 30 to 150 m-thick) comprising mainly thick, ochre-colored, graded turbiditic sandstones and rare conglomerate. They occur throughout the outcrop belt and are vertically alternated and laterally interfingered with fine-grained intervals of mudstone, siltstone and thin turbiditic sandstone beds (Figures 3 and 5c). The sandstone bodies show a typical upward-fining trend and a gradational transition into the overlying finer-grained deposits (Figures 5c and 6b). Unit A2 also shows an overall upward-fining trend, and thickens from the northeast to the southwest. This thickening is accompanied by a change to overall finer-grained facies.

Lithofacies Description (Sandstone Bodies of Unit A2)

**Thick-bedded, graded sandstone** (LF A2-1) represents about 80% of all beds in the sandstone bodies in unit A2 and is present throughout the outcrop belt. This facies comprises generally 0.3 to 1.5 m-thick (but rarely up to 3 m-thick), medium- to very coarse-grained and gravelly, sandstone beds (Figure 6) that are either laterally continuous over <100 m or, less commonly, occur in channels 10s of meters wide and few meters thick. Channels are most abundant in the north-eastern sections (G and H) (Figure 3b). In addition to normal grading (Figure 5f), typical features include dewatering structures and rip-up siltstone clasts or rafts (Figure 5e), erosive and scoured bases (Figure 5b) and sole marks. The top of these beds may contain fine- to very fine-grained sandstone or even laminated siltstone. Some graded sandstones show horizontal stratification at the top (Figure 5f, bottom). Very coarse-grained, gravelly sandstones may display inverse grading followed by traction carpets (cf. Lowe, 1982). Some sandstone beds at the base of section A contain outsized boulders and cobbles.

**Planar cross-stratified sandstone** (LF A2-2) constitutes a minor lithofacies, comprising coarse-grained sandstone with large-scale (<1-m high) tabular cross-stratification, and is found in association with sandstone channels, commonly overlying graded sandstones (LF A2-1) (Figure 5d).

**Conglomerate** (LF A2-3) and **diamictite** (LF A2-4) are also uncommon and mostly restricted to the easternmost sections (G and H), where they are intimately associated with coarse-grained, commonly channelized, sandstone of LF A2-1. Matrix- and clast-supported pebble- to cobble-conglomerate typically occurs as <30 cm-thick lags at the base of these sandstones. At the base of the second sandstone body in section A, a c. 500 m-wide and 5 m-thick channel-fill contains clast-supported cobbles and boulder conglomerate (Figure 5g). In section H, interbedded with conglomerate, a c. 4 m-thick massive diamictite containing cobbles and boulders is observed. It is important to note that this diamictite, in contrast to those present in the Ayn (Kellerhals and Matter, 2003; Rieu et al., 2006) and Shareef formations, does not contain evidence for a glacial influence such as striated clasts.

**Thin-bedded sandstone** (LF A2-5) and **siltstone** (LF A2-6) occur particularly in the upper part of sandstone bodies (Figure 5i), but may be interbedded with LF A2-1. In particular in the western part of the outcrop area (Wadi Marsham) they comprise a more significant part of unit A2. Rare siltstone beds are massive, laminated or graded and occur in cm- to m-thick units interbedded with fine- to medium-grained, laterally continuous massive or graded sandstone with sole marks (toolmarks, load casts) and, commonly overlain by a horizontally laminated interval and/or a rippled (Figure 5i: arrow) interval.

**Interpretation**

The association in the sandstone bodies of unit A2 of the dominantly thick-bedded sandstones (LF A2-1) with the less common thin-bedded sandstones and siltstones (LF A2-6) and the rare planar cross-stratified sandstones (LF A2-2), conglomerates (LF A2-3) and diamictite (LF A2-4), is interpreted to record a turbidite-depositional system. Figure 7 shows the relationship between the facies described here and facies in a generic turbidite model by Mutti (1992, 1999). The typical upward transition from graded sandstone into horizontally laminated and/or rippled sandstone, observed in LF A2-1, but mainly in LF A2-5, is interpreted as the superimposition of Ta, b and c divisions, formed by low-density turbidity currents (Bouma, 1962; Mutti, 1992, 1999). In addition, the dominant facies (LF A2-1) in unit A2 displays a variety of sedimentary structures, such as normal grading, sole marks,
Figure 6: Representative logs through the turbidite complex (unit A2) of the lower Arkahawl Formation:
(a) interbedded sandstone and siltstone of the proximal turbidite facies association in section B (LF A2-1, A2-2, A2-5 and A2-6);
(b) typical section through a coarse-grained sandstone body, comprising proximal turbidite facies at the base (LF A2-1) gradationally fining up through LF A2-5 and LF A2-6 into;
(c) interbedded mudstone, siltstone and thin sandstone (LF A2-7) of the distal fan facies association.

dewatering structures and rip-up siltstone clasts, which are typically associated with deposition from turbidity currents (e.g. Lowe, 1975; Lowe, 1982; Pickering et al., 1989; Mutti, 1992, 1999; Kneller and Branney, 1995; Stow et al., 1996). The very coarse-grained and massive character, the occurrence of inverse grading and traction carpets, as well as the general lack of Bouma Tb-e divisions, suggests that the sandstones of LF A2-1 were formed mainly by sandy to gravelly high-density turbidity currents (Lowe, 1982; Mutti, 1992, 1999; Kneller and Branney, 1995).

Large-scale planar cross-stratification (LF A2-2) is a relatively common, though poorly understood, feature of turbiditic sandstones in the channel-to-lobe transition zone (Mutti, 1992, 1999; Satur et al., 2004). This feature has been proposed to be related to the change from high-density to low-density flows, related to a hydraulic jump (Mutti, 1992, 1999). The conglomerates and diamictites indicate deposition by hyperconcentrated flows and debris flows respectively (Nemec and Steel, 1988; Mutti, 1992, 1999; Stow et al., 1996). The outsized clasts at the base of section A are, considering the pronounced local paleorelief of the basement (Rieu et al., 2006), interpreted as rock fall or alternatively, represent floats in turbidity currents (Postma et al., 1988). The associated siltstone facies (LF A2-6) are interpreted as hemipelagic background sedimentation or distal, low-density turbidite deposits (Pickering et al., 1989; Stow et al., 1996).
Lithofacies Description (fine-grained intervals of unit A2)
These relatively fine-grained deposits occur in 20 to 150 m-thick intervals, alternating with the sandstone bodies of unit A2. These intervals are laterally continuous throughout the outcrop belt (Figure 3). The contact with the underlying and overlying units (A1 and A3) is gradational.

Interbedded mudstone, siltstone and thin sandstone (LF A2-7) are the main constituent of the fine-grained intervals of unit A2, and are commonly planar laminated (laminae are few mm-thick), graded and locally ripple cross-laminated (Figures 5h and 6c). Fine-grained sandstones are 1–20 cm thick, laterally continuous over few 10s of meters, massive and slightly graded and may have rippled tops. Locally the tops of some sandstones may be undulated or show rare straight-crested and symmetrical wave ripples (< 2 cm-high) (e.g., very base of section A; section F).

Interpretation
The often close relationship of these fine-grained deposits with the turbiditic deposits of the sandstone bodies suggests that they form part of the same depositional system, as discussed below. Laminated and graded siltstone and mudstone are therefore interpreted to represent suspension settling from the tails of low-density turbidites (Pickering et al., 1989; Stow et al., 1996) and hemipelagic background sedimentation. The interbedded fine-grained sandstones are interpreted to be deposited by low-density turbidity currents (Mutti, 1992, 1999). The overall fine-grained nature of this facies association probably reflects a relatively quiet depositional environment in an outer fan or lobe fringe setting (Pickering et al., 1989; Mutti, 1992, 1999; Stow et al., 1996). The presence of wave ripples indicates wave action in water depths shallow enough for the sea bed to be agitated by storm waves (Allen, 1984).

Depositional Model for Unit A2
Following the model of Mutti (1992, 1999) for the evolution of turbidite flows (Figure 7), conglomerate (LF A2-3) and diamicrite (LF A2-4) facies represent the most proximal deposits and probably formed in an ‘upper fan’ setting by hyperconcentrated flows and debris flows (Figure 8). Such sediment-flows may transform in the downstream direction into sandy and gravelly high-density currents, from which the bulk of the coarse-grained sandstones was deposited (LF A2-2), probably in a ‘middle fan’ setting. Farther down the system, these high-density currents transformed into low-density turbidite currents resulting in the deposition of planar stratified sandstone facies (LF A2-3) and finer-grained turbidites with Bouma divisions in areas otherwise dominated by hemipelagic sedimentation (LF A2-5, A2-6 and A2-7) (‘outer fan’).

Taken as a whole, unit A2 is interpreted as a submarine turbidite complex comprising several small-scale (1–5 km wide), coarse-grained depositional lobes or fans (represented by sandstone bodies; Figure 3) embedded in fine-grained turbidite deposits in an outer-fan environment (Figure 8). Paleocurrent orientations and the overall lateral shift to finer-grained and more distal-lobe deposits from the northeast to the southwest suggest that this depositional system was fed from the northeastern part of the outcrop area (Figure 8) and may have been connected to the entrance point of the large paleovalley that also served as the main sediment transport route during deposition of the Ayn Formation (Figure 3b). The overall fining-up of unit A2 is interpreted as due to simple transgression.

The occurrence of wave-ripples in LF A2-5 indicates that the sea-bed was never completely below storm wave-base (50–100m). We therefore envisage that the turbidite complex described here represents a relatively shallow turbidite system fed by rivers (e.g. Dabrio, 1990; Plink-Bjorklund and Steel, 2004), rather than a deep-sea turbidite fan. A shallow-water interpretation of unit A2 is supported by the presence of wave-ripples in the overlying “distal-marine” unit (A3).

Unit A3: Distal-Marine Deposits
Overlying unit A2 is a second, much thicker (<500 m) interval of distal-marine deposits that forms the upper part of the Arkahawl Formation (unit A3); it is a laterally continuous unit throughout the outcrop belt (Figure 3). The transition between the underlying sandstone-dominated deposits of unit A2 and the mudstone and siltstone of unit A3 is gradational. Some poorly defined upward-coarsening
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Figure 8: Depositional model for unit A2 of the Arkahawl Formation showing part of coarse-grained submarine fan composed of several active small sandy lobes (km scale) that are laterally alternated with finer grained-outer-fan deposits. B, D and G mark the approximate location of the logged sections (Figure 3). In this illustration the fan lobes are fed with coarse-grained debris that is transported through paleovalleys, which also were sediment transport routes during deposition of the underlying glaciogenic Ayn Formation. See text for discussion.

Figure 7: Comparison between generic turbidite facies described by Mutti (1992, 1999) and facies found in the Arkahawl Formation.

This study Lithofacies (LF) Codes

| Flow Type         | Mechanism of Deposition |
|-------------------|-------------------------|
| Flow Transformation (b) | Cohesive freezing (a) followed by cohesive freezing (b) |
| Traction carpets  | Frictional freezing (a) |
| Gravelly High Density Flow | En-masse deposition |
| Sandy High Density Turbidity Current |
| Low Density Turbidity Current |

Boulder, cobble and pebble
Small pebble
Very small pebble, granule and very coarse sand
Medium and fine sand
Very fine sand and coarse silt
Mud

Paleocurrent directions measured from ripple cross-lamination, dune cross-stratification and toolmarks

Distal-marine mudstones (LF A1-1 and A3-1)
‘Outer-fan’ mudstones, siltstones and thin sandstones (LF A2-7)
‘Middle fan’ sandstones and rare siltstones (LF A2-1, A2-2, A2-5 and A2-6)
‘Upper-middle fan’ sandstones and rare conglomerates (LF A2-1 to A2-4)
Basement
cycles can be recognized in this unit (e.g. basal part unit A3 in section F). However, due to the very fine-grained character of the deposits and common vegetation cover, it is difficult to recognize the bounding surfaces of these cycles and to trace them throughout the area.

**Lithofacies Description**

*Laminated mudstone and siltstone* (LF A3-1) is the dominant facies in unit A3 and is represented mainly by greenish, laminated siltstone and mudstone, occasionally showing ripple cross-lamination. Fault scars and slump folds on 10s meter-scale occur in this unit (Figure 9a).

*Graded sandstone* (LF A3-2), occurs as a minor facies in the lower c. 50 m of this unit. It is very fine- to medium-grained and sedimentologically similar to the thin turbidites of the underlying unit (LF A2-5); it represents a transition between the two units. Occasionally these very fine-grained sandstones have symmetrical wave ripples (few cm high, wave-length \( \lambda \) = c. 10 cm). In the western part of the outcrop belt, a 3 to 7 m-thick, coarse-grained, massive sandstone channel-fill occurs isolated within LF A3-1 in the upper part of this unit (sections B and C: c. 900 and 975 m).

**Interpretation**

The dominance of these facies in unit A3 indicates a quiet, distal-marine depositional environment dominated by hemipelagic sedimentation. Ripple cross-lamination suggests reworking by weak bottom currents (Pickering et al., 1989; Stow et al., 1996). The occurrence of large-scale synsedimentary faults and folds in this context points to deposition on an unstable, oversteepened slope. The presence of wave-rippled sandstones implies that, although deposition took place in a quiet distal-marine depositional environment, water depths were shallow enough for the sea bed to be occasionally agitated by storm waves (e.g. Allen, 1984). The massive and graded nature of the isolated sandstones found in sections B and C suggests rapid deposition from a high-density current. In this otherwise quiet depositional environment, this sandstone may reflect a rare, catastrophic event such as a river flooding (Mulder and Syvitski, 1995), transporting coarse-grained sediment into distal parts of the basin.

**MARSHAM FORMATION**

The distal-marine siltstones and mudstones of the Arkahawl Formation gradationally pass up into the Marsham Formation, which can be found throughout the area, except for the northeastern part where it is erosionally truncated beneath overlying Cretaceous deposits. The Marsham Formation is a < 150 m-thick overall upward-coarsening succession of shallow-marine to fluvial deposits, represented by units M1 and M2 respectively.

**Unit M1: Shallow-Marine Deposits**

Unit M1 is composed of shallow-marine deposits and forms a distinct, laterally continuous unit below the overlying Cretaceous limestone cliffs (Figure 9b). In an overall upward-coarsening trend, unit M1 comprises up to four or five, lower-order, 10s of m-thick, upward-coarsening cycles. These cycles are composed of c. 20 m of proximal offshore mudstone and siltstone at the base, passing up into a 10–30 m-thick sandstone-dominated unit of transition zone, shoreface and/or beachface deposits (Figure 10). Each cycle is commonly ‘capped’ by <30 cm-thick, relatively coarse-grained lag deposits.

**Lithofacies description**

*Mudstone and siltstone* (LF M1-1) constitute the lower part of the coarsening-upward cycles. They comprise purple (but greenish in the lower cycles), laminated mudstone and siltstone with rare cross-laminated few cm-thick siltstone lenses. Upward, lenticular bedding and siltstone lenses become more abundant, as do thin (<15 cm-thick) interbedded coarse siltstone with straight-crested wave-rippled tops.

*Lenticular, wavy and flaser-bedded sandstone* (LF M1-2) is a major constituent of the sandstone-dominated intervals that gradationally overlie offshore mudstone and siltstone (LF M1-1). These heterolithic deposits are composed of 1–10 cm-thick, very fine and fine-grained, well-sorted sandstone with
subordinate interbedded mudstone (mm- to few cm-thick). Symmetrical ripple marks and ripple cross-lamination are abundant in this facies (Figures 9c and d). Wave-ripple crests have a roughly consistent NE-orientation and ripple cross-lamination indicates dominantly offshore paleocurrents to the northwest (Figure 10).

*Planar and low-angle stratified sandstone* (LF M1-3) is commonly found in the upper part of the coarsening upward cycles overlying heterolithic transition zone facies (LF M1-2), although locally it also occurs interbedded with them. LF M1-3 is composed of fine- to medium-grained, well-sorted, 0.2-2 m-thick, planar-stratified to low-angle (<20°) cross-stratified sandstone (Figure 9e). These sandstones may also show convolute bedding and mm-thick laminae of heavy minerals.

*Deformed massive sandstone* (LF M1-4) is rare and occurs mainly at the top of coarsening-upward cycles in <10 m-thick pervasively disorganized sandstone units with m-scale load casts.

*Coarse-grained lag deposits* (LF M1-5) commonly occur at the top of the coarsening-upward cycles and are overlain by offshore mudstones and siltstones (LF M1-1) of the overlying cycle. These lag deposits (<30 cm-thick) are continuous throughout unit M1 and range from medium-grained sandstone overlying the lower two cycles, to pebble conglomerate overlying the third cycle.

**Interpretation**

A shallow-marine depositional environment for this facies association is suggested primarily by the ample presence of wave-induced structures particularly in LF M1-2 and also in LF M1-1. The abundance of laminated mudstone and siltstone in facies M1-1 indicates, however, a generally quiet offshore depositional environment. The lenticular and wave-rippled siltstones present in this facies are interpreted to record occasional wave action associated with storms. The upward increase of these storm-generated deposits indicates shallowing above storm wave base.

In LF M1-2, wavy, lenticular and flaser bedding, in addition to symmetrical ripple marks, suggest deposition above storm-wave base, whereas the interbedded mudstone laminae indicate deposition below fair weather wave base, in between storms (Komar, 1976; De Raaf et al., 1977; Hunter et al., 1979; Johnson and Baldwin, 1996). These facies therefore probably formed in the lower shoreface to offshore transition zone (cf. Komar, 1976) dominated by storm-induced underflows, which is consistent with the predominance of offshore directed paleocurrents.

The planar and low-angle stratified sandstone and heavy-mineral concentrations (LF M1-3) indicate upper-stage plane beds formed in high energy conditions, typical for the upper shoreface and beach environments (Hunter et al., 1979; Johnson and Baldwin, 1996). In this context, convolute bedding (LF M1-3) may have been enhanced by air-entrapment in beach sediment, breaking waves and changes in the water table (de Boer, 1979; Collinson and Thompson, 1982). The deformed sandstone facies (LF M1-4) possibly record liquefaction of upper shoreface deposits, as they are found at the top of coarsening-upward cycles. Liquefaction may have resulted from wave and/or sediment loading, or alternatively, from a seismic trigger.

The relatively coarse-grained character of the lag deposits (LF M1-5) compared to the underlying sandstone (LF M1-2 to LF M1-4), and the occurrence of this facies just below transgressive surfaces, suggests winnowing of the sea bed during flooding to produce ravinement deposits (e.g. Bergman and Walker, 1988).

**Depositional model**

Taken as a whole, unit M1 is suggestive of a wave-dominated, shallow-marine and beach environment (Johnson and Baldwin, 1996). The overall upward-coarsening trend suggests simple progradation, with the lower-order coarsening and shallowing-upward cycles indicating shorter-term cyclical progradational pulses. The apparently minor lateral facies changes in this unit are consistent with the NW-directed paleocurrent directions, indicating that the shoreline was oriented roughly parallel to the outcrops, i.e. NE-SW. One exception is that cycles 2 and 3 in sections I and J seem to be thicker, and lag deposits (LF M1-5) in section J, which are correlated with those overlying cycle 3 elsewhere, pass eastward into offshore facies, suggesting faster progradation in the east.
Figure 9: Field photographs of representative lithofacies of the upper Arkahawl Formation (unit A3), the Marsham Formation (units M1 and M2) and the Shareef Formation (unit S1). See facing page for continuation.
Figure 9 (continued):
(a) green distal-marine siltstone and mudstone (LF A3-1) in section C (looking NE at ~550 m), showing disconformable contacts interpreted as slump or slide scars;
(b) coarsening-upward cycles of shallow marine deposits (unit M1; section C), arrows indicate flooding surfaces;
(c) lenticular, wavy and flaser bedded sandstone (LF M1-2), hand lens for scale;
(d) straight crested, trochoidal, wave-rippled top profile in sandstone (LF M1-2), hammer for scale;
(e) horizontally-stratified sandstone with convolute bedding and low-angle cross-stratification of shoreface and beach facies (LF M1-3), hammer for scale;
(f) trough cross-stratified sandstone (LF M2-2) of the fluvial facies association;
(g) clast-poor glaciomarine diamictite with lonestone (LF S1-1);
(h) striated clast from diamictite (LF S1-1), pen for scale.
Unit M2: Fluvial Deposits

A <30 m-thick unit of fluvial deposits overlies, and incises by up to 10 m into, offshore mudstone and siltstone of the shallow-marine facies association in section B (Figures 3 and 10). Owing to vegetation cover and truncation by the sub-Cretaceous unconformity, fluvial deposits cannot be observed elsewhere. There is an overall coarsening-upward trend through this unit from sandstone to gravel containing cobble-size clasts.

Lithofacies Description

Massive sandstone (LF M2-1) facies directly overlie the shallow-marine deposits of unit M1 in a 25 m-thick, upward-coarsening succession of poorly organized, thick and crudely-bedded (<3 m) sandstone. The sandstones are moderately-well sorted, fine- to coarse-grained and contain dispersed gravel and few mm-thick gravel stringers. Individual beds may show subtle upward-fining. At the base of this succession incised channels occur, containing 0.5 to 1 m-thick conglomeratic gravel lags.

Horizontally stratified and trough cross-stratified sandstone (LF M2-2) occurs towards the top of this succession. This facies is composed of medium-grained, well-sorted sandstone with 10–50 cm-high trough cross-sets (Figure 9f) and parallel laminations.

Conglomerates (LF M2-3) comprise the upper 20 m of unit M2. This facies includes massive and stratified gravel and pebble conglomerate, and matrix-supported, poorly-sorted pebble conglomerates with rare cobbles as well as clast-supported conglomerate. The latter is more common towards the top.

Interpretation

The limited exposure of these facies hampers a comprehensive interpretation. However, the association of trough cross-stratified sandstone with channels and conglomerate suggests a fluvial-dominated environment (e.g. Ramos et al., 1986; Collinson, 1996). Trough cross-stratified sandstone (LF M2-2) in this facies association is indicative of bedload transport in migrating dunes, whereas horizontal stratification is interpreted as the product of upper-stage plane bed transport (e.g., Stear, 1985). The massive and graded sandstone facies, in this context, are interpreted to record rapid deposition from heavily sediment-laden flows during river floods (LF M2-1) (Stear, 1985; Collinson, 1996). The gravel stringers present in these facies may record winnowing by strong currents (e.g. McCabe, 1977). The overlying clast-supported, massive and stratified conglomerates indicate bedload deposition during stream floods (e.g. Steel and Thompson, 1983), whereas the massive, matrix-supported conglomerates (LF M2-3) are interpreted as sandy debris flows (Nemec and Steel, 1984).

A fluvial origin of unit M2 is consistent with a continuation of the overall progradational trend observed in the underlying shallow-marine to coastal part of the Marsham Formation. The abrupt change from offshore to fluvial deposits (Figure 10, c. 115 m) may indicate either fluvial erosion of shoreface deposits during progradation, or a rapid shift in facies due to forced regression. It is also worth noting that many coarse-grained coastal systems occur in fault-bounded, steep-sided basins, associated with fan deltas (Nemec and Steel, 1988; Gawthorpe and Colella, 1990).

SHAREEF FORMATION

The Shareef Formation comprises <40-m-thick glaciomarine deposits, which overlie fluvial conglomerates (section B) or offshore mudstone (section B2) of the Marsham Formation (Figures 3 and 10). Outcrops of the Shareef Formation are difficult to access due to their steep nature and vegetation cover. The great variability in thickness (0–40 m) over 100s of meters suggests preferential preservation in incised paleovalleys.

Unit S.1 Glaciomarine Deposits

Lithofacies Description

Glaciomarine deposits (LF S1-1) are the only constituent of the Shareef Formation observed in the study area. These deposits are represented by a 10–40 m-thick unit of diamictite with a dark-
purple siltstone matrix containing c. 5% (locally up to 60%) of pebble to boulder size clasts (LF S1-1). Green, chemically altered 5–20 cm-thick horizons that are oriented parallel to the regional bedding are thought to have formed along primary bedding planes (Figure 9g). The majority of the clasts have striated and/or very well polished surfaces and flat-iron shapes commonly occur (Figure 9h).

**Interpretation**

Diamictite may be formed by different processes including subglacial deformation or meltout, ice-rafting and debris flows and is common in glacial environments (Ovenshine, 1970; Eyles et al., 1985; Syvitski et al., 1996). Although debris-flow diamictites can form in non-glacial conditions (Nemec and Steel, 1988; Arnaud and Eyles, 2002), the abundant striated, polished and flat-iron-shaped clasts indicate subglacial transport and are supportive of an interpretation involving deposition in a glacial environment (Boulton, 1978). The low clast abundance and the muddy nature of the diamictite suggest a relatively distal glaciomarine setting (e.g. Eyles et al., 1985; Syvitski et al., 1996). An alternative interpretation as lodgement or meltout till is also possible.

It is worth mentioning here that lithologically this diamictite is very similar to a diamictite (clast-poor with purple matrix) underlying a thin cap-carbonate and possible Nafun Group stratigraphy in the PDO exploratory well Lahan-1, located about 100 km northwest of the Mirbat area (Bowring et al., 2007).

**STRATIGRAPHIC ARCHITECTURE**

The bulk sediment dispersal direction of the Mirbat Basin is generally towards the WNW as indicated by transport directions between west and north in the Ayn Formation (Kellerhals and Matter, 2003), the N-NW orientation of paleovalleys in the basement (Rieu et al., 2006), and the transport direction to the SW-NW in the turbidites of the Arkahawl Formation.

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**Figure 10:** Representative section through the Marsham Formation at section B, paleocurrent directions measured from ripple-cross lamination (black) and the orientations of wave-ripple crests (gray) in unit M1 (measured at different localities).
Deepening of the basin to the WNW is in agreement with the SW-NE direction of the coastline inferred from the Marsham Formation, and with the interpretation that the Mirbat Group was deposited on the eastern margin of the roughly NE-striking ‘Abu Mahara basins’ based on gravity data (Loosveld et al., 1996). Such a geometry for the Abu Mahara basins is also illustrated by the trend maps of the Abu Mahara Group (Figure 11).

The fundamental sedimentological trend displayed by the Mirbat Group (Figure 12) is of: (1) overall relative sea-level lowstand allowing the cutting and subsequent filling of incised valleys by glacially influenced shallow-water and terrestrial basin-margin deposits (Ayn Formation); (2) a period of marine sedimentation during overall transgression over basin margins and early highstand (units A1 to A3), possibly interrupted by a period of lowstand when turbidite fans formed (unit A2); (3) pulsed progradation of shallow-marine deposits during late highstand (unit M1), (4) fluvial incision and deposition (unit M2) during regression in the prelude to (5) a second period of glacially influenced sedimentation (unit S1).

The major transgression at the base of the Arkahawl Formation is interpreted to represent, at least partly, eustatic sea-level rise during the recovery from a major Neoproterozoic glaciation. The termination of many Neoproterozoic glaciations appears to be associated with a change from syn- to post-rift subsidence (Young, 1995; Eyles and Januszczak, 2004), but there is no field evidence to either support or reject such a scenario for the Mirbat Basin. The abrupt progradation of turbidite complexes of unit A2 over distal-marine mudstone and siltstone (unit A1) is cautiously interpreted as caused by a period of relative sea-level fall and lowstand, but alternatively may reflect the lagged evacuation of large amounts of sediment from the continent during deglaciation (Hinderer, 2001). Vertical alternation between proximal turbidite deposits (formed on the front of a fan-delta) and outer-fan deposits may have been caused by smaller-scale fluctuations in relative sea level and sediment input, or may have resulted from avulsions of the main feeder channels of these lobes on the delta-front unrelated to relative sea-level change. Due to the discontinuity of the fan lobes, physical correlation of stratigraphic surfaces is impossible.

The overall fining-upward trend displayed by unit A2 is interpreted as the result of gradual transgression. An additional transgressive pulse may be recorded by the change to distal-marine sedimentation of unit A3, although the occurrence of wave-rippled storm deposits in the lower part of unit A3 precludes a major flooding event. The following progradation of shallow-marine deposits of the Marsham Formation (unit M1) reflects highstand deposition. The sequence boundary recorded at the base of the overlying fluvial deposits (unit M2), incising into offshore shelf deposits, is interpreted as the result of a forced regression caused by glacio-eustatic sea-level fall related to the glaciation recorded by the overlying Shareef Formation. In this context, the short-term relative sea-level fluctuation recorded by parasequences of unit M1 could also reflect glacio-eustatic sea-level variations, which would be consistent with the suggestion by Day et al. (2004) that icehouse conditions prevailed prior to the widespread deposition of Cryogenian glacial deposits.

**COMPARISON WITH OTHER NEOPROTEROZOIC ‘INTERGLACIAL’ SUCCESSIONS**

Other ‘interglacial’ or ‘non-glacial’ successions between older and younger Cryogenian glaciations provide interesting comparisons for the Mirbat Group. These include, for example, the Windermere Supergroup of the McKenzie Mountains, northwest Canada (Narbonne and Aitken, 1995; Day et al., 2004), the Umbertana Group of the Adelaide Geosycline, South Australia (Preiss and Forbes, 1981; Sohl et al., 1999) and from Svalbard (Fairchild and Hambre, 1995; Halverson et al., 2004). Glacial deposits in Canada and southern Australia preserved in rift basins (Young, 1995; McKirdy et al., 2001; Hoffman and Schrag, 2002), are, like the Ayn Formation, overlain by transgressive ‘cap’ carbonates and thick turbidite successions of the Twitya (Canada) and Tapley Hill (Australia) formations (Preiss and Forbes, 1981; Eibacher, 1985; Preiss, 1987; Narbonne and Aitken, 1995; Eyles and Januszczak, 2004).

Interestingly, Day et al., 2004 interpreted rapid alternation between deep-water and shallow-water to fluvial facies at the end of the ‘interglacial’ period, underlying the Icebrook Formation (Aitken, 1991; James et al., 2001) and Wilsonbreen Formation glacial deposits in Canada and Svalbard (Fairchild...
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Figure 11: Outlines of Abu Mahara basins in southern Oman (this study) based on seismic interpretations of Petroleum Development Oman (e.g., Vroon-ten-hove, 1997; Romine and Stuart-Smith, 2003) and outcrop data (Mirbat area). Basin flanks are illustrated by dark colors and depocenters by bright colors. Black lines and dots represent seismic lines and wells penetrating the Huqf Supergroup. Note the dominant NE-SW structural trend, also seen on gravity surveys (Loosveld et al., 1996). This trend was also present during much of the deposition of the overlying Nafun Group (e.g., Allen and Leather, 2006; Le Guerroué et al., 2006a).

Figure 12: Conceptual genetic stratigraphic framework for the Mirbat Group in a NW-SE profile, perpendicular to the outcrop belt and the cross section of Figure 3b. Unit A1: distal-marine deposits; Unit A2: turbidite complex on fan-delta front; Unit A3: distal-marine deposits; Unit M1: shallow marine deposits; Unit M2: fluvial deposits; and Unit S1: glaciomarine deposits.
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and Hambrey, 1995; Halverson et al., 2004) as reflecting low-frequency (recorded at a decameter scale) and high-amplitude (c. 100 m) sea-level oscillations, as a hallmark for icehouse conditions (e.g., Read, 1995). The Icebrook glaciation in Canada was furthermore heralded by a change from warm-water carbonate deposition to (cool-water?) siliciclastic deposition (Day et al., 2004). Similarly, the Elatina Formation in Australia was preceded by a change from carbonate deposition to deposition of shallow-water arkosic siltstone and sandstone (Yalpatina Formation and Gumbowie Member; Preis and Forbes, 1981; Sohl et al., 1999). In Svalbard, periglacial deposits are thought to underlie the Wilsonbreen diamictites (Fairchild and Hambrey, 1995).

The Marsham Formation is remarkably similar to these successions in the sense that low-frequency relative sea-level fluctuations in the order of 10s of meters preceded glacially influenced sedimentation recorded in the Shareef Formation. In addition, chemical and mineralogical proxies (Rieu et al., 2007c) indicate a change from relatively humid conditions in the Arkahawl Formation, to more arid conditions in the Marsham Formation. These observations are consistent with the suggestion by Day et al. (2004) that icehouse conditions prevailed prior to the widespread deposition of Cryogenian glacial deposits. If the interpretation of Day et al. (2004) is correct, and the relative sea-level fluctuations observed in the prelude to the glaciations in Canada, Svalbard and Oman were in fact forced by glacio-eustatic mechanisms, this would imply a gradual and pulsed expansion of continental ice-sheets over a time period accounting for >100 m of shallow-marine deposits in the approach to glaciation, rather than (Poulsen, 2003), or prior to, a rapid and catastrophic growth of ice sheets (e.g. Hoffman et al., 1998). Although no sequence boundaries have been documented in the shallow-marine platform successions of the Ombaatje Formation beneath the c. 635 Ma Ghaub glaciation in the Otavi Platform of Namibia, we note that it does record the start of declining carbon-isotopic values in the approach of glaciation (Halverson et al., 2002).

CONCLUSIONS

The siliciclastic Mirbat Group in southern Oman records an apparently continuous succession of two Neoproterozoic glaciations (the Ayn and Shareef formations) separated by a prolonged Cryogenian ‘interglacial’ period (Arkahawl and Marsham formations), which may represent the only record of the period between the older Cryogenian and the end-Cryogenian glaciations on this part of the Arabian Peninsula, possibly filling the gap between the Ghubrah and Fiq glaciations in northern Oman. The Arkahawl Formation records a long-term transgression and overall highstand during the long ‘interglacial’ period. In the aftermath of the first glaciation, a ‘cap’ carbonate and overlying distal-marine shales overstepped the previous basin margins reflecting a major transgression. This was followed by the development of turbidite complexes along the basin margin, consisting of several small-scale, coarse-grained, coalescent fan lobes that were probably connected to a coarse-grained fan delta in the northeast. Gradual deepening followed, giving rise to prolonged deposition of distal-marine shales in the upper part of the Arkahawl Formation. The overlying Marsham Formation records a highstand and the pulsed progradation of shallow-marine and fluvial deposits in advance of a second glaciation represented by the newly introduced Shareef Formation. This pulsed progradation may reflect short-term, glacio-eustatically forced, sea-level oscillations in response to gradual and pulsed expansion of polar ice sheets in the prelude to the Shareef glaciation.

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