Sedimentology and Chemostratigraphy of the Ediacaran Shuram Formation, Nafun Group, Oman
Erwan Le Guerroué

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Sedimentology & Chemostratigraphy of the Ediacaran Shuram Formation, Nafun Group, Oman
Sedimentology and Chemostratigraphy of the Ediacaran Shuram Formation, Nafun Group, Oman

A dissertation submitted to the SWISS FEDERAL INSTITUTE OF TECHNOLOGY ZURICH for the degree of Dr sc. ETH Zürich

presented by
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2006
May peace be with you
Abstract

A large portion of the Neoproterozoic Ediacaran period, extending from the end of the Marinoan glaciation (c. 635 Ma) to the Precambrian-Cambrian boundary (c. 542 Ma), is occupied by large negative carbon isotope excursions that are closely linked in time with the first appearance of animals in the fossil record. These carbon isotopic signals include the Marinoan ‘cap sequence’ excursion, the Shuram-Wonoka and Precambrian-Cambrian boundary excursions (Fig. 0.1).

The Neoproterozoic of Oman displays an essentially complete succession from the Marinoan upward, resting on top of a 822-825 Ma basement. The Huqf Supergroup is well exposed in the core of the Jabal Akhdar of northern Oman. It contains the glacigenic Abu Mahara Group, which has yielded a U-Pb zircon date of 723+16/-10 Ma from a tuffaceous
The Nafun Group above the Marinoan cap carbonate (Hadash Formation) is well exposed in both the Jabal Akhdar and the Huqf region of east-central Oman. It comprises two siliciclastic to carbonate ‘grand cycles’, both initiated by significant transgressions: these cycles comprise the Masirah Bay/Khufai formations and the Shuram/Buah formations. The Khufai Formation displays a pattern of deposition consistent with a carbonate ramp setting and represents a shallowing-upward carbonate cycle (HST), from outer-ramp facies to cross-stratified grainstones and back-shoal mid-ramp deposits, to inner-ramp shallowing upward cycles. The end of Khufai highstand deposition is marked, basinward, by small incised channels, followed by a flooding into transgressive monotonous shales deposited below storm wave base. The Shuram Formation in the Huqf area (proximal part of the basin) records progressive shoaling through a stack of shallowing upward, storm-dominated parasequences. Eventually the siliciclastic Shuram Formation is gradationally overlain by the progradation of the Buah carbonate ramp.

The Shuram (Nafun Group, Huqf Supergroup) excursion of Oman is characterised by an exceptional amplitude (+5‰ to -12‰ δ¹³C; Fig. 0.1) and long stratigraphic record (~800 m). This carbon isotopic trend is reproducible throughout Oman, from outcrops to the subsurface, and irrespective of sedimentary facies. The entire excursion is essentially in phase with longer term relative sea level change, with the nadir in δ¹³C occurring at the level of the maximum flooding zone of the lower Shuram, and the return to positive values occurring within the overlying Buah highstand.

The Shuram Formation is extremely well exposed for over 40 km in a roughly N-S oriented escarpment in the north of the Huqf area, where it displays a stack of shallowing-upward storm-dominated parasequences. At this parasequence scale, carbon isotopic
values are shown to reflect stratigraphic position within the parasequence stack, and each individual parasequence shows a trend in $\delta^{13}$C values in the direction of sediment progradation. These combined stratigraphic-carbon isotopic observations and the fact that the trend is reproducible throughout Oman support a primary, oceanographic origin for the carbon isotopic ratios.

Radiometric ages combined with thermal subsidence modelling constrain the excursion in time and indicate an onset at ~600 Ma, and duration of approximately 50 Myrs.

The excursion is widely recognised in Oman and has potential correlatives in Ediacaran strata elsewhere. It may thus represent a characteristic feature of the middle Ediacaran period. However, these possible correlatives have a limited $\delta^{13}$C data set and are dissected by unconformities, so that the full excursion cannot be recognized. The Doushantuo Formation of China probably records the end of the excursion at around c. 551 Ma.

The Ediacaran period is also marked by the non-global, short-lived Gaskiers glaciation around 580 Ma and possible more loosely dated coeval events. If the proposed chronology is correct, the Gaskiers-aged glaciation is embedded within the large-amplitude, long-term Shuram anomaly and appears to have had no effect on the chemostratigraphic records of Oman and other sections worldwide.

The fact that a carbon isotope excursion of this magnitude can be recognized in marine Ediacaran rocks from several continents indicates that it was a very widespread oceanographic phenomenon, reflecting the composition of seawater from which carbonate minerals were precipitated. However, the Shuram evidence demonstrates that the negative carbon isotopic excursion is unrelated to glaciation per se and that the marine carbon isotopic record cannot be used as a direct recorder of Neoproterozoic ice ages.
Any explanation of the Shuram shift must therefore imply an extreme disruption of biological systems within the oceans and a cessation of photosynthesis/biological productivity for a prolonged time (compared to Phanerozoic examples of perturbations of the carbon cycle; Fig. 0.1), challenging our understanding of the global carbon cycle. The preservation of the mass balance suggests an involvement of a sufficiently large reservoir of $^{13}$C-depleted material (e.g. organic carbon).
Résumé

Une grande partie de la période Néoprotérozoïque Ediacarienne, depuis la fin de la glaciation Marinoan (c. 635 Ma) jusqu’à la limite Précambrienne-Cambrienne (c. 542 Ma), est caractérisée par d’amples excursions négatives de la courbe isotopique du carbone dans l’enregistrement CArbonaté. Ces excursions sont observables dans la ‘cap séquence’ Marinoan, la Shuram-Wonoka et la limite Précambrienne-Cambrienne (Fig. 0.1).

La succession Néoprotérozoïque d’Oman, quasiment complète, repose sur un socle daté à 822-825 Ma. Le SuperGroup de Huqf affleure largement au nord de l’Oman, dans le cœur de l’anticlinal du Jabal Akhdar. Il est constitué des sédiments glaciogéniques du group de Abu Mahara (723+16/-10 Ma). L’événement glaciaire se termine avec la ‘cap carbonate formation de Hadash d’âge supposé Marinoan (c. 635 Ma). Le groupe de Abu
Mahara contiens aussi les volcanoclastiques du Membre de Saqlah caractérisant un épisode d’extension crustal.

Le Groupe de Nafun, situé au-dessus de la Marinoan cap carbonate (Formation de Hadash), représente deux ‘grands cycles’ siliciclastiques à carbonatés, tous deux initiés par de grandes transgressions. Ces cycles comprennent les formations de Masirah Bay/Khufai et les formations de Shuram/Buah. La Formation de Khufai est organisée selon un profil de rampe carbonatée et représente des cycles carbonatés en «shallowing-upward» (HST), allant de facies de rampe externe à des facies à grainstones cross-stratifiés et des dépôts de back-shoal de milieu de rampe, à des cycles «shallowing upward» de rampe interne. En domaine distal, la fin du Khufai highstand est marquée par des petits chenaux incisifs, suivi d’un système transgressif dominé par des argiles monotones, déposées sous la limite d’action des vagues de tempête. Dans le Huqf (domaine proximal du bassin), la Formation de Shuram enregistre une diminution progressive de la tranche d’eau qui se traduit par une série de parasequences «shallowing upward» constituées de dépôts de tempête. Les dépôts siliciclastiques de Shuram passent progressivement aux environnements en rampe carbonatée progradante de Buah.

L’excursion de Shuram (Group de Nafun, Supergroup de Huqf) d’Oman est caractérisée par une amplitude exceptionnelle (+5‰ to -12‰ δ^13C; Fig. 0.1) et un long enregistrement stratigraphique (~800 m). Cette tendance est observée à travers tout l’Oman, de l’affleurement à la sub-surface, et ce indépendamment des faciès sédimentaires. L’excursion est globalement en phase avec les changements relatifs du niveau marin, le paroxysme du δ^13C étant atteint dans la partie inférieure de Shuram, au niveau de la zone d’inondation maximum, alors que le retour aux valeurs positives apparaît avec le prisme de haut niveau de Buah.

La Formation de Shuram affleure parfaitement sur les 40 km d’escarpement N-S du nord de la région du Huqf. Elle montre un empiètement de parasequences «shallowing-
upward» dominées par des dépôts de tempête. À l’échelle de la parasequence, les valeurs isotopiques du carbone reflètent la position stratigraphique dans la pile de parasequences, et chaque parasequence montre une variation des valeurs du δ¹³C avec la direction de progradation. Ces informations stratigraphiques et isotopiques ainsi que la reproductibilité du signal isotopique à travers l'Oman suggèrent une origine primaire et océanographique du rapport isotopique du carbone.

L'excursion est contrainte par des âges radiométriques combinés à une modélisation de la subsidence thermique du bassin. Son initiation est datée à ~600 Ma, pour une durée d’approximative de 50 Ma.

L'excursion, bien documentée en Oman, présente des équivalents probables dans d’autres séries Ediacariennes et pourrait constituer une spécificité de la période Ediacarienne. Cependant, une corrélation globale demeure spéculative en raison d’un enregistrement sédimentaire souvent limité par de nombreuses discordances. En Chine, la Formation de Doushantuo enregistre probablement la fin de l’excursion autour de 551 Ma.

La période Ediacarienne est aussi marquée par la glaciation non-global et de courte durée de Gaskiers, autour de 580 Ma, et ces possibles corrélats moins bien temporellement contraints. Si la chronologie proposée est correcte, la glaciation de Gaskiers est alors comprise dans une excursion négative de large amplitude et de longue durée du carbone isotopique. La glaciation apparaît aussi sans effet direct sur l’enregistrement isotopique et stratigraphique d’Oman et des autres sections du même âge.

Le fait qu’une excursion de l’isotope du carbone de cette amplitude peut être identifiée dans les sédiments Ediacariens de plusieurs continents indique qu’il s’agit d’un phénomène océanographique global, reflétant la composition océanique par laquelle les carbonates sont précipités. Une telle excursion est inhabituelle dans les temps géologiques et son explication représente un vrai défi. Cependant, le cas Shuram démontre que les
excursions négatives des isotopes du carbone ne sont par liées aux glaciations per se et que l’enregistrement du carbone isotopique marin ne peut être utilisé comme témoin direct des glaciations Néoproterozoïques.

Toutes tentative d’interprétation de l’excursion de Shuram doit alors inclure un temps de résidence exceptionnellement long en comparaison avec les perturbations du cycle du carbone Phanérozoïques (Fig. 0.1) et doit engager un réservoir de matériel appauvri en $^{13}$C suffisamment large (e.g. carbone organique dissous).
Acknowledgements

I should start the beginning of this three and half years trip, with the day I arrived in Zurich on the really first time... A friend and I had an interview planned with Philip Allen, following the ‘sparkling’ recommendation of Jean-Pierre Burg… This was some kind of an excitement for two young French fellows... We had the meeting with Philip and Andrea without any optimism, I must admit, to get a position there. I never really know what kind of impression we left this day in Philip’s office but to be honest my friend and I didn’t understand much of the conversation!

I will be forever grateful to both my supervisor Philip Allen and Andrea Cozzi for the time and freedom they gave me and the lead they provided through this work. Thanks a lot for all your advice, support and the fun we had in the desert …

Many thanks as well to Petroleum Development Oman (PDO) and in particular Jan Schreurs and Hisham al Siyabi for their help and constant support during this project. Shukram to the many Mohammads there. Thanks for sorting out my offices problems!

Many thanks as well to all the ones that have contributed in the scientific discussion around this project. Especially to James Etienne (keep practice your darts) and Ruben Rieu (you’re next!), Galen Halverson, Martin Kennedy (what have you done for me today?), John Grotzinger, Adam Maloof, David Fike and Paul Hoffman.

Thanks as well to the field assistant Alex (c’est lourd le gamma ray!), Mat (sheesha?) and Ben (al-hamdoulillah).

Thanks also to all the dudes here in ETH. They are to many but thanks to you Nick, Ruben, James, Darrell, Ansgar, Leo, Chris, Pauline, Alex, Andrew, Marcus... and all the others. Vive le Jura libre!

Grand merci aussi au soutiens inconditionnel des Bretons, et entre autre ceux de Rennes : Ben (merci de montrer la voie, maintenant arrête d’être premier !), Flo (le crétacé sera vert pour toujours, même au pays de Chopin), Nico (la stabilité c’est bon, ça évite de tomber), Alan (tabernacle), Thomas (tu nous réconcilieras presque avec la géophysique), Eva (vive le heavy metal en voiture), Chrys (j’ai pas touché à Plume) et Cat (Madame Grande Bretonne). Merci pour tout le bonheur depuis ces premiers jours de Licence. Merci Ben pour ton aide en Oman, merci aussi de m’avoir tiré en haut du Grand Paradis… Merci au Lorientais de toujours : Joann (celui qui vis ces rêves..) et Fred (Groix forever) évidemment les branleurs du fond de classe à qui tout réussit ! mais aussi Ronan (personal shrink), Brendan (28 ans maintenant non ?), Benoit (dure dure de suivre ton rythme)... Mention spécial pour un gars qui un jour m’a fait confiance (j’avais 16 ans) : merci Gaby. Last but not least, thanks to you Poone.

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Chapter 1

Introduction

Outlines

1.1. Foreword

1.2. Introduction

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   1.2.2. Palaeogeography
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Huoif SuperGroup from Jabal Shams, Jabal Akdhar.
1.1. Foreword

This project started on the 1st of November 2002 under the supervision of Philip Allen and Andrea Cozzi and was based in ETHZ. A summary of the timeline of the research project is as follows:

- A first field season to the Jabal Akhdar of Oman was accomplished during the winter of 2003 (11 weeks) focusing on the Khufai-Shuram boundary.
- A further trip to the Jabal Akhdar in December 2003 (5 weeks) allowed the conclusion the fieldwork in this area.
- Galen Halverson and Paul Hoffman were invited in January 2004 for a two weeks field trip around all Precambrian outcrops. This field trip focused on the Huqf Supergroup in general, in the Jabal Akhdar as well as in the Huqf area.
- Winter 2004 was the first trip to the Huqf area (8 weeks) focusing on the Khufai-Shuram boundary as well as the top Shuram parasequences.
- January to March 2005 was spent on an internship in Petroleum Development Oman (PDO) looking at subsurface data (core, cutting, well data and seismic profiles). This led to an internal PDO report (Le Guerroué, 2005).
- The regional congress of IAS 2005 in Oman was an opportunity to lead an international 3 days field trip through the Neoproterozoic geology of the Jabal Akhdar. The field trip was co-led by John Grotzinger (CalTech), Hisham Al Siyabi (PDO) and Mark Newall (PDO). The excursion is published as a field guide book (Allen et al., 2005).
- In February 2005 a week field trip in the Jabal Akhdar and Huqf was carried out in order to introduce MIT collaborators Adam Maloof and David Fike to the area, looking at Buah Formation and Ara Group in particular. It was also the opportunity to resample the Wadi Aswad top Shuram parasequences.
- February 2005, ETH (Jean Pierre Burg, Andrea Cozzi and Wilfried Winkler) organized a student mapping course in the mountains of northern Oman that focused on mapping the entire Huqf Supergroup in the Jabal Akhdar.

February 2006, a two week field trip in the Jabal Akhdar with James Etienne (ETH) and
Martin Kennedy (Riverside) was carried out, focusing on the Abu Mahara sedimentology and basinal settings.

This work has been structured around submission and publication in scientific journals. Selected publications have been included in this thesis as a chapter, each of which represents a comprehensive sub-project. Chapters are complemented with additional data and links between chapters are provided in order to support the argumentation. A further publication (Le Guerroué et al., 2005) has not been included here as it provides no direct input on the research addressed here.

Although the research findings are my own, I benefited from the following supervision, collaboration and other assistance:

- Chapter 2 is published in Precambrian Research (2006) by Erwan Le Guerroué, Philip Allen and Andrea Cozzi. Field work and writing was led by Le Guerroué with the active supervision of both coauthors. Sedimentary facies analysis and stratigraphic interpretation were carried out in collaboration with Allen. Subsurface data analysis was supervised by Cozzi and numerous staff at PDO (Petroleum Development Oman). Samples for chemical analysis were collected, prepared and measured at the EPFL laboratory of Lausanne (Torsten V.) by Le Guerroué with the assistance of Torsten. The paper published in Precambrian Research benefited from journal reviews by Galen Halverson and an anonymous reviewer.

- Chapter 3 is in press in Basin Research (April 2006 issue) by Erwan Le Guerroué, Philip Allen and Andrea Cozzi. Field work and writing was led by Le Guerroué with the supervision of Allen. Sedimentary facies analysis and stratigraphic interpretation were carried out with the collaboration of Allen. Samples for chemical analysis were collected, prepared and measured at the EPFL laboratory of Lausanne (Torsten V.) by Le Guerroué with the assistance of Torsten. Numerical modeling was conducted by Le Guerroué with supervision from Allen. The paper published in Basin Research benefited from journal reviews by M. Tucker, G. Shields and Paul Wright.
Chapter 4 is published in Terra Nova (2006) by Erwan Le Guerroué, Philip Allen, Andrea Cozzi, James Etienne and Mark Fanning. Writing was led by Le Guerroué with the supervision of Allen and Etienne. Basin subsidence analysis was performed with Allen. Zircon U-Pb dating was carried out in Canberra by Fanning. This paper benefited from a review by Galen Halverson.
1.2. Introduction

1.2.1. Tectono-stratigraphic evolution of Oman

Oman records in its Neoproterozoic Huqf Supergroup three groups overlying the Pan-African basement (800-820 Ma; Husseini, 2000; Leather, 2001; Allen and Leather, 2006; Bowring et al., in review; Fig. 1.1 and 1.2), in ascending order these are: the Abu Mahara, Nafun and Ara Groups (Glennie et al., 1974; Gorin et al., 1982; Rabu, 1988; Rabu et al., 1993).

The Huqf Supergroup records different stages of basin subsidence, probably associated with an evolution of the tectonic settings through time. The Abu Mahara Group represents deposition within a rift basin as illustrated by regional seismic and gravimetric surveys of Oman basement that show horst and graben structures (Loosveld et al., 1996; Romine et al., 2004). Rift basins strike roughly NE-SW, and appear to be cut by NW-SE oriented faults that are commonly attributed to the Najd trend of the Arabian shield (Al-Husseini, 2000; Loosveld et al., 1996). Also, the localized occurrence of pillow basalts and more widespread volcaniclastics of the Saqlah Member (Rabu, 1988; Le Guerroué et al., 2005), suggests the initiation of a rifting phase, which is considered to have continued during deposition of the Fiq Member (the Fiq Formation records sediment supply from both margins of a probable graben structure; Allen et al., 2004). The Fiq Member is glaciogenic and ended with deposition of the characteristic Hadash cap carbonate Formation, considered to be Marinoan in age (c. 635 Ma; Allen et al., 2004; Bowring et al., in review). Glaciations appear to have taken place at times of tectonically generated accommodation, suggesting a link between geodynamics, basin development and climate change. Neoproterozoic glacial strata in Oman are keys to the ongoing Snowball Earth discussion, providing an excellent opportunity to test the hypothesis (Etienne et al., in press).
The Nafun Group is an essentially complete, carbonate-rich Ediacaran-aged (Allen et al., 2004; Allen and Leather, 2006; Le Guerroué et al., 2006b) succession (Fig. 1.2) and overlies the presumed Marinoan, rift-related Fiq Member and ends just below the Precambrian-Cambrian boundary (c. 542 Ma; Amthor et al., 2003). The Nafun Group represents two siliciclastic to carbonate ‘grand cycles’ made of the Masirah Bay/Khufai and Shuram/Buah formations, both initiated by significant transgressions (Fig. 1.2; Allen and Leather, 2006; Le Guerroué et al., 2006a; Le Guerroué et al., 2006b). Tectonic subsidence caused by thermal contraction...
following Fiq-aged rifting allowed the preservation of about 1 km of postrift, laterally extensive stratigraphy, with no major stratigraphic breaks. The Nafun Group therefore is interpreted as the post-rift stage of a failed rift or proximal part of a passive margin (Naylor, 1986; Vroon- ten Hove, 1997; Allen et al., 2004; Allen and Leather, 2006; Le Guerroué et al., 2006a; Le Guerroué et al., 2006b). Some disagreement still exists regarding the interpretation of the tectonic regime during Nafun times and Grotzinger et al., 2002 suggests that the broad, regional subsidence represented by the Nafun Group is related to subsidence of the lithosphere associated with the dynamic topography caused by subduction of cold oceanic lithosphere beneath the Arabian plate.

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Figure 1.2. Simplified sedimentary logs of the Jabal Akhdar and Huqf areas with composite δ¹³C plot and extrapolated ages.
The top of the Huqf Supergroup is calibrated by radiometric dates from the middle of the Fara Formation (Ara Group) at 544.5 ± 3.3 Ma (Brasier et al., 2000; Fig. 1.2). This age is consistent with the occurrence of the skeletal fossils *Cloudina* and *Namacalathus* in correlative subsurface rocks (Amthor et al., 2003). Recent radiometric dates obtained from core material in the South Oman Salt Basin have been used to constrain the age of

Figure 1.3: Paleogeographic model for Ediacaran times around 600-580 Ma. Note the northern passive margin of the Gondwana supercontinent made of Arabia and India. Note also that position is Laurentia is uncertain and that two locations for the Yangtze block of South China are proposed (Jiang et al., 2003; Macouin et al., 2004). Reconstruction adapted from Meert (2004). Paleogeographic keys: Ama: Amazonia; Ara: Arabia; Arm: Armorica; Aus: Australia; Ava: Avalonia; Bal: Baltica; Con: Congo; EAn: East Antarctica; Ind: India; Kal: Kalahari; Lau: Laurentia; SC: South China; Sib: Siberia; Waf: West Africa. Active margins are represented by thick grey lines.
the Cambrian/Precambrian boundary at 542 ± 0.6 Ma, when Cloudina and Namacalathus become extinct (Amthor et al., 2003; Fig. 1.2). This age is now used as the reference for the Cambrian/Precambrian boundary. The Nafun Group-Ara Group transition is marked by a shift from extensive post-rift thermal subsidence to a tectonic style marked by uplift of large basement blocks which segmented the broader basin into several fault-bounded sub-basins around ca. 550-548 Ma (South Oman, Ghaba, and Fahud Salt Basin; Immerz et al., 2000; Fig. 1.1) this tectonic style re-used the tectonic fabric of Abu Mahara rifts, making the distinction between the two problematical (Husseini, 2000). The uplifted blocks of the salt basins accumulated primarily carbonate sediments, whereas downfaulted blocks were covered in black shale and silicilyte. Interlayered evaporites blanketed both basins and uplifted blocks (Amthor et al., 2005). Subsidence of the Ara Group is generally attributed to regional strike slip faulting associated with Najd structures (Husseini, 1990; Husseini, 2000).

1.2.2. Palaeogeography

The paleogeography for the Nafun Group is reconstructed based on outcrops and numerous wells throughout Oman. The general pattern of deposition of the grand cycles is consistent, though the quality of well coverage diminishes downwards in the lower Nafun Group (Masirah Bay and Khufai formations; Vroon-ten Hove, 1997; Cozzi and Al-Siyabi, 2004; Romine et al., 2004; Le Guerroué et al., 2006a; see chapter two Fig. 2.12 and 2.13). This consistent trend during the deposition of the Nafun Group is possibly at least in part inherited from the Fiq rifting event (Loosveld et al., 1996; Allen et al., 2004; Romine et al., 2004). Abu Mahara grabens were striking NE-SW and the margin of the Nafun Group probably striking in the same direction appears to open towards the NW, based on paleo-depocenter location (Jabal Akhdar and salt basins low), relative basement elevation (the Huqf high would represent a wider rift shoulder if compared with the Makarem high) and paleocurrent orientation (see chapter 2). Comprehensive palaeogeographic reconstruction using palaeomagnetic data (Dalziel, 1997; Meert and Torsvik, 2003) indicates a roughly
30°S position for Arabian shield at early Nafun time followed by a drift to equatorial position around 580 Ma (Shuram time; Le Guerroué et al., 2006b; Fig. 1.3) (Meert and Lieberman, 2004). Then eventually Oman moved to around 30°S of the equator during Ara Group time (c. 542 Ma; Dalziel, 1997). However direct constraints for Oman palaeo-position are absent within the exception of a 30°N position on the Mirbat sandstones of south Oman (Kempf et al., 2000). Unfortunately the Mirbat sandstones, assigned to 550 Ma, are poorly aged controlled and could be as old as 750 Ma (Rieu et al., 2006). This continental drifting could provide an explanation for the observed sedimentary cyclicity of the Nafun Group that effectively brings the depositional environment from sub-tropical arid conditions (Khufai carbonates) through tropical humid climate (Shuram siliciclastics) back to sub-tropical again (Buah/Ara carbonates).

Such large-scale stratigraphic cyclicity in the Nafun Group of Oman is remarkably similar in the late Neoproterozoic Windermere Group of Canada (Narbonne and Aitken, 1995) and with the Yunnan Province of south China (Kimura et al., 2005) that also contains two siliciclastic-carbonate cycles stratigraphically correlatable with Oman’s Nafun Group (Allen and Leather, 2006; Le Guerroué et al., 2006a). Considering that during the Ediacaran period the Windermere Group of the Laurentian shield was facing the Arabian/Nubian shield over a large ocean (Dalziel, 1997; Meert and Torsvik, 2003; Meert and Lieberman, 2004; Fig. 1.3) such continental drift through climatic belts would not explain the similarities recorded on two distant independent continents, though the Yunnan section was probably close to the Omani margin (Macouin et al., 2004). The Indian Neoproterozoic succession also record marine deposits within a basin that faced the open ocean to the NW and probably shared its margin with the Yangtze block of China (Jiang et al., 2003; Kaufman et al., in press). However, the sedimentology recorded during the Ediacaran aged Krol Group in India is strongly different to that of the Nafun Group of Oman and no large scale cyclicity is recorded there. Jiang et al. (2003) proposed that the Yangtze block was located between the Nubian and Indian blocks. However, paleomagnetic suggest a location between the Indian and Australian block (Macouin et al., 2004; Fig. 1.3).
The two siliciclastic-carbonate ‘grand cycles’ of the Nafun Group represent sequences lasting around 35 and 60 Myrs for each cycle (Le Guerroué et al., 2006b). The processes driving such long-term cyclicity in a thermally driven post-rift basin may be related to sea floor spreading and subduction rate variations or linked to glacial-ages eustasy (Saylor et al., 1998). The first grand cycle records a major relative sea level rise following the Marinoan glaciation, but no equivalent glaciation in term of amplitude and severity precludes the major transgression forming the base of the second cycle. A tectonic explanation is thus preferred for the second cycle, especially regarding the timing of the onset of this cycle, which corresponds loosely with the switch in tectonic regime at the end of the dispersal of continents associated with the break up of Rodinia and the amalgamation of Gondwana (Dalziel, 1997; Meert and Torsvik, 2003). However, it is important to note that both transgressions marking the base of the grand cycles are accompanied by a major δ¹³C perturbation (see discussion below).

1.2.3. Remaining problems

Because of a lack of biostratigraphic and radiometric age constraints, age assignments and global correlation in the Neoproterozoic is mostly based on the carbon isotope signal. Putative snowball Earth event are in this way correlated because of their stratigraphic relationship with negative excursions in their associated carbonate records. A large portion of the Ediacaran period, extending from the end of the Marinoan glaciation (c. 635 Ma) to the Precambrian-Cambrian boundary (c. 542 Ma), is occupied by large negative carbon isotope excursions. These include the Sturtian and Marinoan ‘cap-carbonate sequence’ excursions, the Shuram-Wonoka and Precambrian-Cambrian boundary excursions (Fig. 0.1). The Shuram-Wonoka excursion is particularly large in amplitude and duration. The presence of carbon isotopic excursions of such amplitude and duration are difficult to explain in steady state conditions and implies an extreme disruption of biological systems within the oceans and a cessation of photosynthesis/biological productivity for a prolonged time (Kennedy, 1996). Indeed, Phanerozoic negative excursions are conspicuously short-lived and reach less than -2 ‰ in amplitude (e.g. Hayes et al., 1999; Hesselbo et al., 2000; Galli et al.,
Different working hypotheses have been proposed (among them is Grotzinger and Knoll, 1995; Hoffman et al., 1998; Kennedy et al., 2001). However, detailed mechanisms are poorly constrained and scenarios fail simple stratigraphic and geological tests. Therefore, extremely low values are highly debated regarding their origin and significance as a primary widespread oceanographic signal, although the $\delta^{13}C$ proxy remains the main correlative tool of the Precambrian. Future studies should be targeted at understanding the geochemical origin of the carbon isotopic excursions recorded in Neoproterozoic rocks within a firm sedimentological and stratigraphic context.
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Chapter 2

Chemostatigraphic and sedimentological framework of the largest negative carbon isotopic excursion in Earth history: The Neoproterozoic Shuram Formation (Nafun Group, Oman)

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Precambrian Research, 2006a

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Abstract

Oman records in its Neoproterozoic Nafun Group (Huqf Supergroup) an essentially complete, carbonate-rich Ediacaran succession. The Nafun Group overlies the presumed Marinoan rift-related Fiq Member (c. 635 Ma) and ends just below the Precambrian-Cambrian boundary (542 Ma). Tectonic subsidence caused by thermal contraction following Fiq-aged rifting allowed the preservation of about 1 km of postrift stratigraphy, with no major stratigraphic breaks. The Nafun Group above the Marinoan cap carbonate (Hadash Formation) represents two siliciclastic to carbonate ‘grand cycles’, both initiated by significant transgressions: these cycles comprise the Masirah Bay/Khufai formations and the Shuram/Buah formations. The Khufai-Shuram boundary is associated with the start of a major carbon isotope perturbation. The uppermost ramp carbonates of the Khufai Formation record a smooth decrease in δ¹³C from about +4‰ to values around 0, followed by 2 descending steps across which values plunge to a nadir of -12‰ in the overlying red siltstones and shales interbedded with thin limestones of the Shuram Formation. This fall in isotopic values is temporally rapid and coincident in both shallow and deep water sections in the time span of a single parasequence. The δ¹³C nadir is then followed by 50 million years of monotonic recovery.

The ‘Shuram shift’ represents the largest δ¹³C inorganic carbon negative excursion in Earth history. Although the snowball Earth theory
links periods of depleted carbon isotopic ratios with periods of global glaciation, the non-glaciated context of the Shuram Formation suggests that the causal relationship between global glaciation and negative carbon isotopic excursions is non-unique. Although the precise mechanism driving this major perturbation of the carbon cycle remains enigmatic, the long term remineralization of an isotopically depleted organic carbon reservoir in ocean water is a promising candidate.
2.1. Introduction

Neoproterozoic sedimentary rocks record periods of extreme climatic oscillations, including putative snowball Earth events. However, the Ediacaran period (635-542 Ma; Hoffmann et al., 2004; Knoll et al., 2004; Condon et al., 2005) contains the largest δ^{13}C excursion in inorganic carbon of marine carbonates in Earth history, possibly lasting from c. 600 to c. 550 Ma (Le Guerroué et al., 2006b), a time period for which there is no evidence for major glaciations. The Nafun Group of the Huqf Supergroup of Oman consists of a conformable >1 km-thick mixed carbonate-siliciclastic succession that appears to be unbroken by major unconformities, thereby distinguishing the Nafun Group from other less complete Ediacaran successions. The Nafun Group shows two major siliciclastic to carbonate ‘grand cycles’ above the Marinoan-equivalent Fiq glacial succession and its cap carbonate (Hadash Formation) (Fig. 2.1). The first grand cycle, composed of the Masirah Bay/Khufai formations, is associated with presumed global transgression following the Marinoan glaciation (Allen and Leather, 2006). A significant transgression also defines the base of the second Shuram/Buah grand cycle. The origin of the transgression initiating the second grand cycle remains uncertain since no glacial deposits have been shown to precede the cycle. The Gaskiers glaciation (580 Ma, Krogh et al., 1988; Bowring et al., 2003) is believed to have been localized, and significantly postdates the start of the isotopic excursion according to the chronology of Le Guerroué et al. (2006b).

The post-Marinoan transgressive ‘cap sequence’ (Hoffman et al., 1998) is depleted in ^{13}C with values dropping to -5‰, whereas the transgressive base of the second grand cycle records values down to -12‰ in the Shuram Formation (Burns and Matter, 1993; Le Guerroué et al., 2006b). This very large negative excursion persists through up to 1 km of overlying stratigraphy, with a cross-over to positive values within the ramp carbonates of the Buah Formation. The carbon isotopic excursion is therefore in phase with relative sea level change interpreted from sedimentary facies. A similar excursion in terms of amplitude is found in the Wonoka Formation of Australia (Calver, 2000), Johnnie Formation of SW USA

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1 Details in Chapter 4.
The presence of a carbon isotopic excursion of this amplitude and duration is difficult to explain, since Phanerozoic negative excursions are conspicuously short-lived and represent shifts of generally less than 2 ‰ (e.g., Hesselbo et al., 2000; Galli et al., 2005; Fig. 0.1). A number of theories have been proposed for the isotopic shifts associated with Precambrian glaciations, including the overwhelming of the atmosphere-ocean carbon reservoir with mantle-derived carbon (-5‰; Des Marais and Moore, 1984), accumulated during a long period during which silicate weathering was largely inoperative (snowball Earth theory; Hoffman et al., 1998), the upwelling of stratified, depleted ocean water during periods of deglaciation (Grotzinger and Knoll, 1995), the destabilization of methane clathrates (Kennedy et al., 2001; Jiang et al., 2003), high rates of runoff from large tropical rivers causing efficient burial of organic carbon and subsequent release of large reservoirs of seabed methane (Schrag et al., 2002), and the oxidation of dissolved organic matter in ocean water (Rothman et al., 2003). None of these mechanisms, however, have been applied to the much longer duration and greater amplitude excursion of the Nafun Group, which requires an explanation divorced from global or near-global glaciation.

The aim of this chapter is to provide the sedimentological background for the remarkable changes in the carbon cycle implied by the carbon isotope ratios recovered from marine carbonates in the essentially continuous succession of the Nafun Group of Oman. The Khufai-Shuram boundary and its chemostratigraphic signature are described in deep water facies (Jabal Akhdar, north Oman), shallow water facies (Huqf area, in east-central Oman) and integrated with regional seismic and borehole data provided by Petroleum Development Oman (PDO). In this contribution, the depositional environment and sedimentary processes that were active during deposition of the Khufai-Shuram formations are reconstructed in order to better constrain oceanographic forcing mechanisms for the large Shuram negative δ¹³C anomaly.

More details in Chapter 4.
2.2. The Ediacaran succession in Oman

2.2.1. Stratigraphy

The Neoproterozoic Huqf Supergroup of Oman crops out mainly in north (Jabal Akhdar), east-central (Huqf) and south (Mirbat) Oman and is also penetrated by a large number of boreholes in the salt basins of the Oman interior (Fig. 2.1b). The Huqf Supergroup is subdivided into three groups: the Abu Mahara, Nafun and Ara Groups (Fig. 2.1a, Glennie et al., 1974; Gorin et al., 1982; Hughes-Clarke, 1988; Rabu, 1988; Loosveld et al., 1996; Leather, 2001; Cozzi and Al-Siyabi, 2004; Allen and Leather, 2006).

Figure 2.1: (a) Simplified sedimentary logs of the Jabal Akhdar and Huqf areas. (b) Map of Oman showing Jabal Akhdar, Huqf and Mirbat regions where Neoproterozoic rocks crop out. The Saih Hatat is strongly metamorphosed and deformed and therefore excluded from the present study.
In the Jabal Akhdar, the Abu Mahara Group contains glacigenic intervals in the Ghubrah Formation and Fiq Member of the Ghadir Manqil Formation (Brasier et al., 2000; Leather et al., 2002; Allen et al., 2004), separated by the basaltic volcanics and volcaniclastics of the Saqlah Member (Fig. 2.1a; Rabu, 1988). These two intervals are believed to correlate with two clusters of glacial events labelled Sturtian (700+ Ma) and Marinoan (older than 635 Ma; Condon et al., 2005) respectively (Brasier et al., 2000; Leather et al., 2002; Allen et al., 2004; Halverson et al., 2005). In the Huqf region, the Abu Mahara Group is absent and the Nafun Group rests directly on the felsic volcanics and volcaniclastics of the Halfayn Formation and its granodioritic basement (Plate et al., 1992; Leather, 2001; Allen and Leather, 2006).

The Nafun Group crops out in both the Jabal Akhdar and the Huqf areas. At its base is the Marinoan transgressive cap carbonate of the Hadash Formation (Leather, 2001; Allen et al., 2004; Cozzi and Al-Siyabi, 2004; Allen and Leather, 2006), which directly overlies the Fiq Member (Allen et al., 2004) of the Ghadir Manqil Formation. The Hadash Formation is overlain by two siliciclastic-carbonate depositional cycles of the Masirah Bay-Khufai formations and Shuram-Buah formations (Figs. 2.1a and 2.2). The Nafun Group passes upward into the carbonate-evaporite cycles of the Ara Group, defined in PDO subsurface penetrations in the Oman salt basins (Hughes-Clarke, 1988). Outcrop equivalents of the Ara Group are found in the Huqf area (Fig. 2.1a; Nicholas and Brasier, 2000) and in the Jabal Akhdar, where the Buah Formation is overlain by the volcaniclastics and cherty limestones of the Fara Formation (Brasier et al., 2000).

2.2.2. Geochronology

Both upper and lower limits of the Huqf Supergroup are calibrated by radiometric dates. An ignimbrite in the middle of the Fara Formation (Ara Group) in the Jabal Akhdar yielded a U-Pb age of 544.5±3.3 Ma (Brasier et al., 2000). Recently, core material from the middle part (unit A3 and A4) of the Ara Group in the South Oman Salt Basin yielded an age of 542.0±0.3 Ma, representing the Precambrian-Cambrian boundary (Amthor et al., 2003)
Figure 2.2: Field panoramas of (a) Wadi Bani Awf anticline in the Jabal Akhdar, with Neoproterozoic stratigraphy truncated by the sub-Permian unconformity. Syncline width approximately 3 km. (b) Fiq glacigenic and non-glacial sedimentary rocks overlain by the lower part of the Nafun Group, with Shuram Formation in the core of an asymmetrical syncline, Wadi Sahtan, Jabal Akhdar. Masirah Bay is about 250 m thick. (c) Khufai-Shuram boundary in the Huqf area, showing steeply inclined Khufai Formation, a carbonate-rich upper cycle of the Khufai Formation, and a shale-dominated Shuram Formation, Mukhaibah Dome. (d) Top Khufai Formation dolomitized limestones pass up into bleached siltstones of Shuram Lower Member and then monotonous purple siltstones of the Shuram Middle Member at locality 12, Jabal Akhdar. Telegraph poles are about 5 m high.

which is in good agreement with the Fara Formation age. Therefore, a reasonable estimate for the age of the top of the Nafun Group is c. 550 Ma.

The base of the Huqf Supergroup must be older than the age of the tuffaceous ash interbedded with diamicrites in the Ghubrah Formation in the Jabal Akhdar (723 +16/-10 Ma; Brasier et al., 2000; Bowring et al., in prep). The presence of the volcanic Saqlah Member separating the Ghubrah Formation from the Fiq Member suggests that the latter may be Marinoan in age (ending at 635 Ma\(^3\); Allen et al., 2004; Condon et al., 2005; Bowring et al., in prep). Therefore, the base of the Nafun Group, should this interpretation hold, is regarded to be at around 635 Ma.

\(^3\) See detrital ages of the Lahan well in chapter 4 appendixes
2.2.3. The Nafun Group

The Nafun Group has previously been studied by Gorin et al. (1982), Wright et al. (1990) and McCarron (2000). The Hadash Formation cap carbonate and the siliciclastics of the Masirah Bay Formation strongly overstep basement margins during a major post-Marinoan relative sea-level rise (Allen et al., 2004; Allen and Leather, 2006). The Masirah Bay Formation then passes up gradationally into the prograding carbonate ramp of the Khufai Formation. The Masirah Bay and Khufai formations thus form the first grand cycle of the Nafun Group, sharply transgressive at its base. The overlying Shuram Formation records a significant deepening event, with deposition of siltstones and shales that pass up gradationally into the carbonate ramp of the Buah Formation. The Shuram and Buah formations therefore represent the second grand cycle of the Nafun Group, again with a transgressive base.

Although there are no radiometric dates giving depositional ages within the Nafun Group, a subsidence analysis combined with the ages of detrital zircon populations led Le Guerroué et al. (2006b) to propose that the first grand cycle lasted from c. 635-600 Ma and the second grand cycle from c. 600-542 Ma4.

2.3. The Khufai-Shuram boundary in Oman

2.2.1. Previous work

The Khufai Formation, formerly called the Hajir Formation, was described first by Kapp and Llewellyn (1965), who defined the type section in Wadi Hajir in the Oman mountains (Fig. 2.3). In the Huqf area, the Khufai Formation was first described by Kassler (1965) in the Khufai Dome area (KD; Fig. 2.4). The Khufai Formation attains a thickness of 30-100 m in the Jabal Akhdar and 250-350 m in the Huqf area, and lies conformably on top of the siliciclastic Masirah Bay Formation (Fig. 2.1a). Subsequent studies (Gorin et al., 1982; 4 See details in chapter 4.
Figure 4.3: Detailed geological map of the central part of the Jabal Akhdar between Wadi Bani Kharus and the eastern end of Wadi Sahtan, showing location of logged sections, with paleocurrent indicators. Thickness of the sandstone facies at top Khufai Formation is also indicated. Geological map after Rabu (1988) and Beurrier et al. (1986).
Le Guerroué et al. Figure 4

Figure 2.4: Detailed geological map of the Huqf area showing section locations with paleocurrent indicators. Na1 and Na2: Nafun; BD: Buah Dome; MD1 to 5: Mukhaybah Dome; KD: Khufai Dome; WS: Wadi Shuram, YD: Yidah. Geological map after Dubreuilh et al. (1992) and Platel et al. (1992).
Wright et al., 1990; McCarron, 2000) interpreted the Khufai Formation as a carbonate ramp that passed from inner ramp peritidal facies in the Huqf area to outer ramp organic-rich limestones in the Jabal Akhdar. Khufai carbonate ramps were nucleated on basement highs and passed into deeper water conditions in intervening depressions (see below).

The Shuram Formation, formerly called the Mu’aydin Formation, was described by Kapp and Llewellyn (1965) who defined a stratotype in Wadi Mu’aydin on the south side of the Jabal Akhdar and at Wadi Shuram in the Huqf area (Kassler, 1965; WS, Fig. 2.4). The Shuram Formation is more than 250 m thick in the Huqf area, whereas in the Jabal Akhdar it is approximately 700 m thick, though the thickness is not easily measured due to small and large scale folding. Subsequent studies (Gorin et al., 1982; Wright et al., 1990; McCarron, 2000) interpreted the Shuram Formation as deposited on a storm-dominated siliciclastic shelf in the Huqf area, becoming more distal towards the Jabal Akhdar.

2.3.1. Sedimentary facies

2.3.2.1. Khufai Formation

The Khufai Formation comprises four facies associations (A, B, C and D; Fig. 2.5) distributed along a homoclinal ramp (Fig. 2.6 and Table 2.1) and passes up gradationally, but over a short stratigraphic interval, into the Shuram Formation.

2.3.2.1.1. Facies Association A: fetid carbonates

In the Huqf area facies association A1 consists of dark, fetid dolostones with faint centimetric planar to undulating laminations, rare evidence for cross-stratal truncations, cm-thick graded beds and cm- to m-scale slump structures with vergence generally to the northwest. In the Jabal Akhdar facies association A2 is made of black, fetid, pyritic limestone forming planar, structureless beds. Rare mudstone interbeds have commonly undergone soft sediment deformation (Table 2.1 and Fig. 2.5; McCarron, 2000).
<table>
<thead>
<tr>
<th>Facies association</th>
<th>Sedimentary structures</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Huqf area</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A1 Black fetid dolostones</td>
<td>Centimetric planar to undulating laminations, rare cross-stratal truncations, cm-thick graded beds and cm-m scale slumps</td>
<td>Gently sloping anoxic-suboxic outer ramp</td>
</tr>
<tr>
<td>B1 Grainstones passing into fetid, intraclastic wackestones</td>
<td>Scours, swaley and cross-stratification, soft sediment deformation, stromatolites</td>
<td>Storm dominated shallow mid ramp</td>
</tr>
<tr>
<td>Ca Ooidal/peloidal grainstones with rounded and flattened intraclasts</td>
<td>Cross-stratified beds with grainstone and stromatolite clasts</td>
<td>Subtidal, high energy</td>
</tr>
<tr>
<td>Cb Mudstones with mm-scale siltstone and sandstone beds</td>
<td>Structureless</td>
<td>Protected shallow subtidal, probably lagoon margins</td>
</tr>
<tr>
<td>Cc Microbially laminated packstones, certified evaporite cm-thick lenses</td>
<td>Planar, box-like and conical stromatolites, trochoidal wave ripples</td>
<td>Shallow subtidal (lagoonal), inner ramp</td>
</tr>
<tr>
<td>Cd Sandstones, siltstones</td>
<td>Mini-ripples and bidirectional cross-strata</td>
<td>Wave- and tide-influenced shallow subtidal, inner ramp</td>
</tr>
<tr>
<td>F1 Siltstones with rare calcareous silty interbeds</td>
<td>Small wave ripples, sets of swaley cross-stratification and rare planar beds</td>
<td>Storm influenced shelf</td>
</tr>
<tr>
<td>G1 Siltstones and ooidal grainstones containing intraclasts</td>
<td>Soft sediment deformation, climbing wave ripples, swaley cross-stratification, planar beds and edge wise conglomerates, in progressively thickening parasequences</td>
<td>Stormy proximal shelf to shoreface</td>
</tr>
</tbody>
</table>

| **Jabal Akhdar** |                        |                |
| A2 Black, fetid, pyritic limestone and rare mudstone interbeds | Planar lamination, structureless beds, soft sediment deformation | Distal outer ramp |
| B2 Intraclast grainstones alternating with facies association A | Scours, swaley and cross-stratification, soft sediment deformation | Shallow mid ramp |
| Da Siltstones dispersed in facies association B | Unidirectional current ripples, graded and slumped beds | Storm influenced outer shelf |
| Db Sandstone channels with cm-long intraclasts of reworked mudstones & siltstones, ooids | Rare unidirectional cross-strata, oscillatory and combined flow ripples | Lower shoreface to inner shelf |
| E M-thick dolomitized mudstones alternating with bleached siltstones | Convolute bedding, cm-m scale slump structures, organic-rich | Outer shelf |
| F2 Monotonous shales with local silty carbonate beds | Unidirectional current ripples, rare climbing wave ripples and hummocky cross-stratification | Lower shoreface to inner shelf |
| G2 Shale and mudstone interbeds | Rare edgewise conglomerates, m-scale asymmetric-symmetric parasequences | Inner shelf |

Table 1: Facies associations of Khufai and Shuram Formations in the Jabal Akhdar and Huqf areas.
The presence of pyrite, dark grey colouration and preservation of organic matter suggest deposition in a suboxic environment. The Huqf area sediments are considered to have been deposited in a gently sloping outer ramp setting, whereas the Jabal Akhdar limestones are more typical of distal outer ramp facies (Fig. 2.5; McCarron, 2000).

2.3.2.1.2. Facies Association B: ooidal-peloidal grainstones and stromatolites

Facies association B (Figs. 2.5 and 2.6) is composed of grainstones with a strong facies change from east-central to north Oman. Cross-stratified ooidal/oncolitic grainstones pass into fetid, intraclastic wackestones in the northern part of the Huqf area (Facies association B1; Khufai Dome and especially at Buah Dome), and then to reworked intraclast beds alternating with the fetid carbonates of facies association A1 in the mountains of north Oman (Facies association B2). Cm- to m-scale stromatolites (Mukhaibah Dome) are

Figure 2.5: Sequence stratigraphic conceptual model for the top Khufai and Shuram formations, illustrating lateral variations between the Jabal Akhdar and Huqf area outcrops. Correlation between localities is supported by identical carbon isotopic profiles (heavy lines) between the two areas.
common in the Huqf sections (Table 2.1; McCarron, 2000), but totally absent in those of the Jabal Akhdar.

The presence of stromatolites and associated sedimentary facies in the Huqf area indicates a well-oxygenated, moderate energy environment, possibly in a back-shoal inner ramp setting. The Khufai Dome area represents a moderate to high energy environment in a less protected inner ramp position. The intraclastic/peloidal wackestones of the Buah Dome indicate deposition below fair weather wave base, probably in a storm-dominated mid ramp setting. The Jabal Akhdar reworked intraclast beds and fetid carbonates are more typical of slope ramp facies. Such a lateral variation represents facies changes across a north-deepening ramp (Fig. 2.6; McCarron, 2000).

Figure 2.6: Schematic block diagram showing top Khufai Formation facies distribution along a homoclinal carbonate ramp. A to D refer to facies associations described in text.
Figure 2.7: Detailed sedimentary logs of the top Khufai Formation cycles in the Huqf area. Note that the Khufai-Shuram boundary is placed after the last carbonate-dominated cycle or parasequence. The last carbonate-rich parasequence is equivalent in age to the Lower Member of the Shuram Formation in the Jabal Akhdar.
2.3.2.1.3. Facies Association C: peritidal shallowing-upward m-scale cycles

Facies association C is restricted to outcrops in the Huqf area. At the same stratigraphic position, the Jabal Akhdar sections contain facies association B2 and A2, eventually passing up vertically into D (Table 2.1; Figs. 2.5 and 2.6). The cycles of facies association C in the upper Khufai Formation mark the demise of the carbonate ramp in the Huqf area.
Facies association C is dominated by m-scale cycles (2 to 5 m) recognized by recessive bases and resistant cherty caps (Figs. 2.7 and 2.8). The cycles comprise, from base to top, ooidal/peloidal grainstones containing large (few dm), rounded and flattened intraclasts (Ca; Figs. 2.7 and 2.9c), cross-stratified beds and channel-fills (YD section) with grainstone and stromatolite clasts. The cycles are capped by mudstones with mm-scale siltstone and sandstone beds (Cb) passing up into planar, box-like and conical stromatolites (Facies Cc, Figs. 2.7 and 2.9d), and locally (Mukhaibah Dome) trochoidal wave rippled packstones. Wave ripples have spacings between 5 and 11 cm and steepnesses (vertical form index) of 5 to 6.7, indicating that they are vortex ripples (Cc, Figs. 2.7 and 2.9g). Facies Cc also contains abundant chertified cm-thick lenses (Figs. 2.7 and 2.9b), probably resulting from chertification of lozenge-shaped evaporite mineral precursors. Wright et al. (1990) describe anhydrite inclusions in lutecite rosettes with possibly anhydrite as the precursor mineral.

The m-scale cycles are rare in the Buah Dome, where a calcareous quartz arenite is found (McCarron, 2000). Wright et al. (1990) report dolomite flake breccias in siltstone channels in this facies. This siliciclastic facies (Cd) interfingers with facies association Ca and Cb in the Khufai Dome (WS), the YD section and north Mukhaibah Dome (MD4), where medium siltstones to coarse sandstones in m-scale beds contain mini-ripples and bidirectional cross-strata (Figs. 2.7 and 2.9e,f).

Facies Ca was deposited in subtidal, high energy environments, locally as tidal bars (YD section). The structureless mudstones and wackestones of facies Cb represent

Figure 2.10: Facies typical of the Jabal Akhdar area. (a) Interwoven ripple cross-lamination in grainstone of facies association F2 at location 8; (b) Ripple-laminated and low-angle planar-laminated grainstones, with symmetrically rippled top surface; facies association F2 at location 1; (c) Planar, unidirectional cross-stratified sandstone of facies association Db at location 1; (d) Convolute micro-slumped mudstone of facies association E at locality 12; (e) Khufai-Shuram boundary region in the Jabal Akhdar; brown sandstone-filled channels (S) intercalated with mudstones and limestones (L) of facies association Db at location 1; (f) Unidirectional cross-laminated form-set made of siliciclastic sandstone and grey carbonate mudstone, within the top Khufai Formation of facies association Da at location 4. Coin is about 2,5 cm width.

- 62 -
deposition in protected shallow subtidal environments, probably in lagoon margins with a minor detrital supply. Cc represents evaporitic shallow subtidal- stromatolitic intertidal (lagoonal) environments. Siliciclastic facies Cd requires some continental (possibly aeolian) input, probably from the Buah dome area, where the most siliciclastic sand at this interval occurs.
The presence of bimodal, herringbone cross-stratification demonstrates the influence of tides, and the mini-ripples are interpreted as due to wave action in extremely shallow water depths, which together suggest shallow subtidal to intertidal depositional environments. Analysis of the trochoidal wave ripples found near the top of the Khufai Formation in the Mukhaibah Dome area using the method of Allen (1981, 1984), recently applied to very large ripple structures in Marinoan cap carbonates (Allen and Hoffman, 2005), indicates that maximum periods of formative waves were small (c. 2 secs), suggesting locally generated waves acting in shallow water, most likely in a protected inner ramp setting.

All sections show a repeated stack of subtidal facies (Ca) and shallow subtidal facies (Cb/Cc). Base and top facies transitions of each cycle are sharp in most cases, but the transition from grainstones (Ca) to evaporite-bearing mudstones (Cb/Cc) is gradual. Consequently, these m-scale cycles are interpreted as small shallowing upward parasequences (Figs. 2.7 and 2.8).

Although facies association C reflects very shallow water depths at the top of the ramp sequence, no evidence for subaerial exposure in the form of major erosion surfaces, paleosols or karstification has been identified.

2.3.2.1.4. Facies Association D: intermixed siltstones and sandstones

Facies association D is only found in the Jabal Akhdar on top of facies association B (Fig. 2.5). It records input of siliciclastics at the top of the Khufai carbonate ramp. Siliciclastics occur as dispersed quartz grains in a carbonate matrix (forming discontinuous cm-thick layers or graded in the carbonates; Da) whereas at a stratigraphically higher level, coarser clastics (Db) are found in erosive channel-fills eroded into facies association B and the thinly bedded pinkish mudstones of facies association E (see below; Table 2.1). Facies association D marks the demise of the carbonate ramp in the Jabal Akhdar.
Facies Association Da comprises siltstones intercalated in the carbonates of facies association B as cm-thick layers, and as graded basal layers of carbonate beds. They show unidirectional current ripples, shallow erosive and slumped beds (Fig. 2.10f). Regional paleocurrent indicators show a roughly northward directed flow at the time of deposition (Fig. 2.3).

Coarser clastics form shallow, erosive, m-thick channels, with abundant flattened cm-size clasts of reworked pink mudstones and siltstones, ooids and carbonate intraclasts (facies Db; Fig. 2.10e). Sandstone channels contain rare unidirectional cross-strata (Fig. 2.10c). Laterally, the channel-fills comprise clast-supported conglomerates (section 16 for example) with dm-long, subangular, flattened intraclasts. A maximum of 5 channels occur at locality 1, and extend for hundreds of metres laterally, before wedging out (Figs. 2.3 and 2.11).

Some sandstone beds are dm-thick, and form laterally extensive sheets (in section 5 in particular). The tops of beds contain dm-high symmetrical and asymmetrical ripple marks with unidirectional internal foresets. Ripple wavelengths reach up to 75 cm, but more commonly are 40 cm, with heights of ~15 cm and grain size around 0.25 mm. In plan view, they form individual hummocks or less regular 3-dimensional structures. Top profiles of ripple form-sets are typically trochoidal, with thicker drapes of light grey carbonate mud in ripple troughs. Ripple profiles are slightly asymmetrical; the final stage of ripple migration is commonly recognized by a fine interlamination of quartzose sand-silt and carbonate silt, demonstrating that fine carbonate sediment was carried in suspension by the same currents that transported the coarser siliciclastic sand as bedload. Some ripple profiles appear to be modified by tectonic deformation.

The top of the Khufai carbonate ramp in the Jabal Akhdar records the deposition of the siliciclastics of facies association D (Fig. 2.11), which is interpreted as having been deposited mostly below storm wave base (SWB). As in the Huqf area, there was most
likely a shallowing of depositional environments at the top of the Khufai carbonate ramp in the Jabal Akhdar region, indicated by these siliciclastic inputs, locally (as in section 5) reworked by oscillatory and/or combined flows (Dumas et al., 2005). Analysis of the wave ripple structures (method in Allen 1981, 1984) indicates that maximum wave periods were up to 10 secs, suggesting that the Shuram basin was large enough to allow the generation of long period waves. Shallowing of depositional environments relative to the underlying Khufai ramp is also suggested by the derivation of ooids from a shallow ramp setting. However, this shallowing of the ramp was of limited magnitude, as no major erosion surfaces or lowstand deposits have been identified. Indeed, facies association D passes to facies association E gradationally in localities where sandstones and conglomerate channels do not truncate the uppermost carbonate beds. We associate the minor relative sea level fall causing siliciclastic

Figure 2.11: 3-dimensional perspective of the geometry of the top Khufai surface and the position of channelized sandstones, using N-S and E-W profiles in the Jabal Akhdar. Note that the steepest local slopes in the basin are c. 0.5°. Numbers at top and bottom of panels are section locations shown in Figure 2.3.
2.3.2.2. Shuram Formation

The Shuram Formation in this study is divided into three facies associations (E, F and G; Fig. 2.5 and Table 2.1), corresponding to the three members of the Shuram Formation. The boundaries of the Khufai and Shuram formations are defined lithostratigraphically. However, laterally, facies association E of the Shuram Formation in the Jabal Akhdar passes to time equivalent facies association C of the Khufai Formation in the Huqf area (Fig. 2.5). That is, the youngest carbonate-dominated cycle (1 to 5 m-thick) of the Huqf area is equivalent in time to the oldest siliciclastics in the Jabal Akhdar (see chemostratigraphic section below).

2.3.2.2.1. Facies Association E: dolomitic mudstones and bleached siltstones

Facies association E is found only in the Jabal Akhdar, and is made of finely convoluted m-thick dolomitized mudstones (Fig. 2.10e) alternating with multicoloured and bleached siltstones. Slump structures are common at the cm- to m-scale within the mudstone beds and locally involve the whole bed. Organic matter is present within the dolomitized mudstones (up to 4.5% TOC) in the form of thinly laminated black shale interbeds. Facies association E varies in thickness in the Jabal Akhdar area and thickens to 50 m towards the northwest (Fig. 2.11). Rarely, facies association E records minor coarser siliciclastics in the form of cm-thick siltstone and sandstone beds. Sections 5, 6 and 16 contain only a few bleached siltstones of facies association E interbedded with the coarse channelized clastics of facies association D (Fig. 2.11).

The upward change into the fine-grained lithologies of facies association E, and the absence of any storm-generated features, suggests increasing water depths. Transgression is also supported by the presence of interlaminations of organic-rich black shales at the base of the facies association. The convolutions and slumps in thin mudstone beds are the
result of soft sediment deformation due to shaking or gravitational instability.

Facies association E makes up the lower member of the Shuram Formation. Chemostratigraphic correlation (see Figs. 2.5 and 2.11 and below for further chemostratigraphic details) shows this facies association to correlate laterally with facies association D, which contains oscillatory and combined flow ripples. This suggests the existence of a gentle topography in the Jabal Akhdar during the time period of the Lower Member, as supported by the regional distribution of facies association E (Fig. 2.11). The upward transition of carbonate-dominated sediments into the siliciclastic-dominated facies association E indicates a shut-down of the Khufai carbonate factory, possibly related to climate change and/or hydrochemical changes.

2.3.2.2.2. Facies Association F: monotonous siltstones and shales

The sediments of facies association E pass up into purple, monotonous siltstones and shales characteristic of the Shuram Formation Middle Member. The Huqf region contains siltstones with rare calcareous silty interbeds. The siliciclastics locally contain small wave ripples, sets of swaley cross-stratification and rare planar beds (F1; Table 2.1). The Jabal Akhdar records monotonous shales that commonly contain mm-scale, unidirectional current ripples in silty carbonate beds (F2; Table 2.1). Locally (sections 1, 2 and 16), the Jabal Akhdar sections contain brownish dm-thick carbonates with climbing ripples with steep foresets (section 4). Ripples ranging from 5 to 25 cm in wavelength and 10 cm in amplitude have their tops reworked into symmetrical profiles, but rare plan views show linguoidal ripple crestlines. These brown carbonate beds also contain detrital silt-sized quartz grains, rare wave ripples and an undulatory lamination reminiscent of hummocky cross-stratification. On the south west of the Jabal Akhdar, section 8 shows cosets of interwoven wave ripple cross-lamination (Fig. 2.10a).
Rippled silty carbonate beds in the Jabal Akhdar are interpreted as having been reworked from the shallow part of the shelf and resedimented during storms under oscillatory and combined flow. However, they are rare and their distribution is probably linked to local basin highs. Storm influence is also evident in the Huqf region. Monotonous siltstones and shales of facies association F are interpreted as being deposited slightly below the SWB in the Jabal Akhdar and above it in the Huqf region, without recording fair weather structures. However, coarser grain size as well as a greater abundance of oscillatory structures suggests a shallower paleowater depth in the Huqf region. In the Huqf area facies association F of the Shuram Formation represents a significant increase of water depth compared to the peritidal environments of the Khufai Formation. The presence of storm-induced sedimentary structures in this deeper water facies association attests to an increased storm regime. This facies association, comprising the Middle Member, makes up the bulk of the Shuram Formation in the deeper-water Jabal Akhdar, but only a few tens of metres in the shallower-water Huqf area. A maximum flooding zone of the depositional sequence is located towards the base of the Middle Member, where it records minimum storm influence.

2.3.2.2.3. Facies Association G: clastic/carbonate parasequences

In the Huqf area, facies association F gradationally passes upward into planar and swaley cross-stratified coarse siltstones with interbeds made of cross-stratified and wave-ripped ooidal grainstones containing intraformational clasts (G1; Table 2.1 and Fig. 2.9a). Storm-generated sedimentary structures such as climbing wave ripples, swaley cross-stratification, planar beds and edge-wise conglomerates become more common upwards. Eventually this facies association develops into progressively thickening parasequences individually attaining a maximum of 20 m in thickness (Le Guerroué et al., in press)5. The top of the Shuram Formation in the Jabal Akhdar also contains siliciclastic/carbonate m-scale cycles, though these are much finer grained than underlying stratigraphy and contain only rare evidence for storm reworking (edge-wise conglomerates) (G2; Table 2.1). They form m-scale symmetric and asymmetric cycles, before passing into the limestones at the base

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Figure 2.12: Bathymetric facies and decompacted thickness map of the Shuram Formation in Oman. Decompaction was carried out using present-day burial depths for subsurface data, and a maximum burial depth of 4 km for outcrops, using initial porosities and porosity-depth coefficients for shales and carbonates. Map inset gives abbreviations of well names.
of the Buah Formation (Cozzi et al., 2004b).

The Huqf facies association G1 is interpreted as progressively thickening and shallowing upward parasequences of storm-dominated siltstones capped by ooidal grainstones deposited on a shallow shelf to shoreface, with water depths no greater than a few tens of metres (Le Guerroué et al., in press). Facies association G2 in the Jabal Akhdar was deposited in an outer shelf environment with accumulation of thin-bedded siltstones below SWB.

Facies association G makes up the Upper Member of the Shuram Formation and gradationally passes upwards into the carbonate ramp succession of the Buah Formation (Cozzi et al., 2004b). In the Huqf area, parasequences gradually deepen upward, reaching a second order maximum flooding zone recognized by the presence of 10 m of monotonous shales marking the Shuram-Buah transition (Cozzi and Al-Siyabi, 2004; Cozzi et al., 2004b; Le Guerroué et al., in press).

2.3.3. Regional depositional patterns

32 wells and regional seismic lines covering all of Oman (data provided by Petroleum Development Oman) have been interpreted in this study and integrated with field observations in order to document the regional depositional pattern across Oman. Wireline logs (gamma ray, neutron-bulk density, resistivity, etc.) as well as cuttings and core material have been used to reconstruct the regional distribution of Shuram depositional facies. Based on this enlarged Shuram data set, individual wells have been classified into 5 different facies from deep to shallow water, with the Jabal Akhdar and Huqf as deep and shallow water end members, respectively (Fig. 2.12). Seismic profiles have been used to verify the lateral consistency of the well ties as well as to check depositional geometry, lateral thickness variations and the nature of the formation boundaries.

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The Shuram Formation facies deepen from the Huqf area northward to the Jabal Akhdar and westward into the Ghaba Salt Basin. Farther to the west, the Makarem high marks another area of shallow water deposition. To the south of the Huqf outcrops, deep water facies are encountered, before reaching shallow water facies again in the vicinity of Salalah in the Dhofar area (Fig. 2.12). This general pattern of deposition is consistent with those of the Masirah Bay and Khufai formations, though the quality of well coverage diminishes downwards in the lower part of the Nafun Group (Vroon-ten Hove, 1997; Romine et al., 2004; Fig. 2.13). This consistent trend during the deposition of the Nafun Group is at least in part inherited from the Fijr rift event (Loosveld et al., 1996; Allen et al., 2004; Romine et al., 2004) since regional seismic and gravimetric surveys of basement
show horst and graben structures (Loosveld et al., 1996; Romine et al., 2004). Abu Mahara (Fiq) rift basins strike roughly NE-SW, and appear to be cut by NW-SE oriented faults that are commonly attributed to the Najd trend of the Arabian shield (Fig. 2.13; Loosveld et al., 1996; Al-Husseini, 2000). The Nafun Group stratigraphy is thought to have been deposited during the thermal relaxation phase following Saqlah/Fiq-aged Abu Mahara rifting, its depocentres reflecting the inheritance of long-lived horsts and grabens in the Panafican basement (Fig. 2.13).

### 2.3.4. Khufai-Shuram depositional model and sequence stratigraphic interpretation

The Khufai Formation displays a pattern of deposition consistent with a carbonate ramp setting (Fig. 2.6). A homoclinal ramp depositional model is preferred here to a distally steepened ramp (Read, 1985), as no seismic lines showing a sharp slope break and slope breccias, suggesting steep slopes, have been found in the deep-water Jabal Akhdar

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Table 2.2: Reference composite δ¹³C and δ¹⁸O isotope data of section 12 in the Jabal Akhdar and sections Na1, Na2 and MD5 in the Huqf region. Other isotopic data are found in Figures 15 and 16.
sections. The Khufai represents a shallowing-upward carbonate cycle (HST), from outer-ramp facies to cross-stratified grainstones and back-shoal mid-ramp deposits, to inner-ramp shallowing upward cycles (McCarron, 2000). The end of Khufai highstand deposition is marked, in the Jabal Akhdar, by small incised channels, followed by a flooding into transgressive monotonous shales deposited below storm wave base (Fig. 2.11). However, in the Huqf region, the flooding is associated with the uppermost carbonate cycle or parasequence, followed by further deepening into storm-dominated siltstones deposited above storm wave base (Fig. 2.5). The Jabal Akhdar remained in deep water until progradation of the Buah carbonate ramp (Cozzi and Al-Siyabi, 2004; Cozzi et al., 2004b).
whereas the middle and upper members of the Shuram Formation in the Huqf area record progressive shoaling through a stack of shallowing upward storm-dominated parasequences (Fig. 2.5).

2.4. Khufai-Shuram Chemostratigraphy

2.4.1. Materials and methods

16 outcrop sections in the Jabal Akhdar and 16 sections in the Huqf area spanning the Khufai-Shuram boundary have been sampled, in the course of measuring stratigraphic sections, for inorganic carbon stable isotope analyses (about 700 measurements in total; two composite sections for the Jabal Akhdar and the Huqf area are shown in Table 2.2). Samples were drilled with 1 - 5 mm dental drill bits from freshly cut rock slabs avoiding sparry cement and vein material. The C and O isotope composition of powder from the carbonate samples were measured with a GasBench II connected to a Finnigan MAT DeltaPlus XL mass spectrometer, using a He-carrier gas system according to methods adapted after Spoetl and Vennemann (2003). Samples are normalized using an in-house standard calibrated against δ^{13}C and δ^{18}O values of NBS-19 (+1.95 and –2.20 ‰, relative to VPDB). External reproducibility for the analyses estimated from replicate analyses of the in-house standard (n=6) is ±0.07‰ for δ^{13}C and ±0.08‰ for δ^{18}O.

2.4.2. Alteration

Some bulk carbonate samples were geochemically screened (Burns and Matter, 1993; Burns et al., 1994; McCarron, 2000; Leather, 2001). Elemental Sr, Mn and Fe were measured by McCarron (2002) on samples from Khufai, Shuram and Buah formations (Fig. 2.14A and B). The Mn/Sr ratio is a common proxy for diagenetic exchange. During meteoric diagenesis Sr is expelled from marine carbonate whereas Mn is incorporated. However, absolute concentration of Sr, Mn and Fe reflects availability of reduced Sr, Mn and Fe in the

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7 See full data table in Appendixes Page 70
original sea water or diagenetic fluid. The Mn-Sr plot for the carbonate of the Nafun Group shows no clear trend (Fig. 2.14B).

Samples have Mn/Sr ratios <10, with many samples <3, suggesting that most of the δ\(^{13}\)C and δ\(^{18}\)O values are not significantly diagenetically altered (Fig. 2.14C; Kaufman and Knoll, 1995). Additional isotopic data, which are consistent with the values in the samples screened for diagenetic alteration, were therefore used without further elemental screening.

### Figure 2.15: Carbon isotope data plotted against simplified sedimentary logs around the Khufai-Shuram boundary in the Jabal Akhdar. East-west and north-south profiles. Note the slow decrease of δ\(^{13}\)C in the uppermost Khufai as well as the two drops at the base and top of the bleached siltstones of the Shuram Lower Member.
When a plot of $\delta^{13}$C versus $\delta^{18}$O (Fig. 2.14D) produces a straight line of positive slope, the covariance is thought to be due to meteoric alteration, which reduces both the carbon and oxygen isotopic ratios (Fairchild et al. 1990). In the Nafun Group excursion (samples ranging from top Khufai to Buah; i.e. from $+5\%$o down to $-12\%$o then back to $+2\%$o), covariance $R^2$ is 0.461 for top Khufai plus Shuram samples in the Huqf area and 0.0010 for Jabal Akhdar samples. $R^2$ for Buah samples is 0.098 in the Huqf and 0.660 in the Jabal Akhdar. There is therefore no clear covariance between oxygen and carbon isotopic ratios, and the Shuram excursion is assumed to be negligibly affected by meteoric alteration.

Regional consistency between outcrop sections and well data, as well as geochemical screening for diagenetic alteration, confirm the authenticity of the chemostratigraphic trends (Burns and Matter, 1993; McCarron, 2000; Cozzi and Al-Siyabi, 2004; Cozzi et al., 2004a,b; Le Guerroué et al., 2006b).

2.4.3. Results

The pattern of variation in $\delta^{13}$C quickly rises from the negative (-5 to $-3\%$o) Hadash cap carbonate values to positive values through the Masirah Bay Formation ($+3\%$o, see data in Allen et al. 2004 and Leather 2001) and Khufai Formation ($+4$ to $+5\%$o). Variation from the Khufai to the Buah Formations shows a positive-negative-positive cycle. Values around $+4\%$o characterize the top of the Khufai limestones, followed by a smooth decrease to zero within the last carbonate cycle (Facies association C; Figs. 2.15 and 2.16). This smooth $\delta^{13}$C isotopic decline is associated with a general minor shallowing at the top of the carbonate ramp from facies association B1 to C in the Huqf (Fig. 2.16) and from A2 to D in the Jabal Akhdar (Fig. 2.15). The decline to zero is followed by step-wise negative falls, at first to values of zero to $-5\%$o, and then to values as negative as $-12\%$o. The Jabal Akhdar section shows the two descending steps at the top and bottom boundaries of the Lower Member of the Shuram Formation (Facies association E) (Table 2.2; Fig. 2.15), whereas the section in the Huqf shows this double stepped excursion at the boundaries of the uppermost
carbonate cycle of the top Khufai ramp (Facies association C) (Table 2.2; Fig. 2.16). Based on this double chemostratigraphic step, the Khufai-Shuram boundary in the Jabal Akhdar is placed directly above the facies Db sandstone channels (or A2 facies when sandstones channels are missing; Fig. 2.11) and below the bleached siltstones of facies association E (Fig. 2.15). The uppermost Khufai carbonate parasequence in the Huqf area (Facies association C) therefore correlates with the bleached siltstones (Facies association E) of the Jabal Akhdar (Figs. 2.15 and 2.16). Although the Khufai-Shuram boundary marks the beginning of transgression, the top of the Lower Member of the Shuram Formation marks a more important transgressive event, which floods both the Jabal Akhdar and the Huqf areas to establish a relatively deep water sea (Facies association F). Negative isotopic values persist throughout the Shuram Formation and into the base of the carbonates of the Buah Formation (-6‰), and then gradually climb back to positive values (+2 ‰; Cozzi et al., 2004a,b; Cozzi and Al-Siyabi, 2004) at the top of the Buah Formation. The Precambrian-Cambrian boundary in Oman is marked by a smaller negative excursion in the middle of the Ara Group (-3 to -5‰; Amthor et al., 2003).

It is critical to appreciate that the positive-negative-positive isotopic trend is regionally consistent and unbroken by offsets. This confirms the results of several region-wide sedimentological studies (Cozzi et al., 2004a; Cozzi and Al-Siyabi, 2004; McCarron, 2000) that failed to identify any major unconformities within the Nafun Group. Furthermore, carbon isotopic ratios show a systematic variation within a stack of parasequences near the top of the Shuram Formation in the Huqf area (facies G1), each cycle showing a trend in δ¹³C in the direction of sedimentary progradation (Le Guerroué et al., in press). This strongly suggests a long-term secular variation in δ¹³C rather than a post-depositional overprint. These combined stratigraphic-carbon isotopic observations support a primary, oceanographic origin for the carbon isotopic ratios.
2.5. Implications for Ediacaran oceanography, paleoclimatology and correlation

2.5.1. Timing, duration and severity of the carbon cycle perturbation

Using a time transformation of the stratigraphic sections in Oman, based on the modelling of subsidence due to postrift thermal contraction (Le Guerroué et al., 2006b), the major negative carbon isotopic excursion is interpreted to rapidly begin around c. 600 Ma. The negative excursion continues with a steady rising limb crossing over to positive values at c. 550 Ma. The negative excursion therefore has a duration of about 50 million years, starting some 35 My after the end of the Marinoan glacial epoch, dated in Namibia at 635.5 ± 0.6 Ma (Hoffmann et al., 2004) and at 635.2 ± 0.6 Ma in south China (Condon et al., 2005), and ending less than 10 My before the Precambrian-Cambrian boundary (542 Ma; Amthor et al., 2003). A δ¹³C_carb excursion of this amplitude and duration is unique in Earth history.

2.5.2. Worldwide correlation of the δ¹³C excursion

Ediacaran strata recording negative δ¹³C_carb values similar to those of the Khufai-Shuram boundary in Oman are found in the Wonoka Formation of the Adelaide Rift Complex of South Australia (Gostin and Jenkins, 1983; Preiss, 1987; Calver, 2000), the Johnnie Formation of the Death Valley region of western U.S.A. (Christie-Blick and Levy, 1989; Corsetti and Hagadorn, 2000; Corsetti and Kaufman, 2003), the Doushantuo Formation of south China (Condon et al., 2005), the Chenchinskaya, Nikolskaya and Torginskaya Formations of Siberia (Melezhik et al., 2005) and the Krol Group of northern India (Chen et al., 2004; Jiang et al., 2002). However, these possible correlatives have a limited δ¹³C dataset and are dissected by unconformities, so that the full excursion cannot be recognized⁸. The Doushantuo probably records the end of the excursion at around c. 551 Ma (Condon et al., 2005; Le Guerroué et al., 2006b).

⁸ See correlation in chapter 4
Although the Windermere Supergroup of Canada contains two grand cycles that appear to correlate with the Nafun Group of Oman, and strontium isotopes (Narbonne et al., 1994) support a correlation of the Blueflower and Shuram Formations (Le Guerroué et al., 2006b), the carbon isotopic data from the Windermere section (Narbonne et al., 1994) do not replicate the Shuram shift. Similarly, the Jiucheng member of south China (Kimura et al., 2005), which is a possible Shuram equivalent, does not bear a carbon isotopic excursion in carbonate of comparable magnitude to the Shuram shift.

### 2.5.3. ‘Gaskiers’ and equivalent glaciations

Glaciation during Ediacaran times is commonly accepted and recognized as the Varangerian/Gaskiers event (Halverson et al., 2005). Age constraints on glaciation remain a highly debated issue since only sparse absolute ages are available. Nevertheless, the best constraint on a post-Marinoan event is given by the Gaskiers glaciation of Newfoundland (Krogh et al., 1988), which is precisely dated at 580 Ma and lasted less than 1 My (Bowring et al., 2003). Coeval glacial deposits are also found in the Boston basin, where the Squantum Formation is bracketed to between 590 and 575 Ma (Thompson and Bowring, 2000), though its glacial origin is debated (Eyles and Januszczak, 2004). Similar ages obtained on the Loch na Cille Boulder Bed in Scotland (Elles, 1934) and equivalent beds in Ireland (Condon and Prave, 2000), constrained to be younger than 601 ± 4 Ma (Dempster et al., 2002), are in accord with the age of the Gaskiers glaciation. Other constraints are given by the Cottons Breccia and equivalent Croles Hill Diamictite to 582±4 Ma and 574.7±3 Ma respectively (Calver et al., 2004) and the Moelv tillites constrained to be younger than 620 Ma (Bingen et al., 2005). Finally, the Olympic and Inishowen ice-rafted debris are both constrained to be <592±14 Ma (Re-Os age; Schaefer and Burgess, 2003) and ca. 590-570 Ma (Condon and Prave, 2000) respectively. However, the Olympic is considered to be Marinoan aged (ca. 635; Halverson et al., 2005, Kendall et al., 2004). Ediacaran glaciation is therefore thought to have occurred around 580 Ma and to have lasted a short time (Halverson et al., 2005), unlike postulated snowball Earth events (Hoffman et al., 1998).
The short-lived Gaskiers glaciation is therefore embedded within a much longer time period occupied by the Shuram Formation $\delta^{13}C$ excursion. The isotopic excursion appears to have begun and ended independently of glaciation. Consequently, any link between the Shuram excursion and the negative $\delta^{13}C$ values of carbonates bracketing Gaskiers-aged glacial deposits in Newfoundland (Myrow and Kaufman, 1999), the Quruqtagh Formation in northwest China (Xiao et al., 2004), and the Mortensnes Formation in Svalbard (Edwards, 1984) could be coincidental. If our chronology is incorrect and the sharp fall in carbon isotopic values at the Khufai-Shuram boundary correlates in time with the shortlived and localized 580 Ma Gaskiers glaciation, the amplitude of the excursion and subsequent recovery through 1 km of stratigraphy, crossing over to positive values at c. 550 Ma, would still be unexplained. There is consequently no justification for linking the Shuram carbon isotopic excursion to the Gaskiers event.

### 2.5.4. Climate change and oceanography

Neoproterozoic carbonates record highly negative and positive $\delta^{13}C$ values of strongly debated origin (Christie-Blick et al., 1995; Kaufman and Knoll, 1995; Kaufman et al., 1997; Hayes et al., 1999; Walter et al., 2000; Hoffman and Schrag, 2002; Schrag et al., 2002; Halverson et al., 2005).

The precipitation of isotopically depleted cap carbonates is explained in the snowball Earth theory by the prior build-up of mantle-derived carbon in the atmosphere-ocean reservoir during a period of prolonged hydrological shutdown during which silicate weathering was inoperative (Hoffman et al. 1998). This model is not applicable to the Shuram anomaly since the latter is unrelated to glaciation, and, furthermore, expected $\delta^{13}C$ values in co-precipitated carbonates cannot reach values lower than the -5 to -6‰ $\delta^{13}C$ values typical of mantle sources (Des Marais and Moore, 1984).

The depleted values of Neoproterozoic carbonates may also be the consequence of a biological pumping in a stratified ocean (Grotzinger and Knoll, 1995; Kaufman et al.,
1997). Turnover of the deep-water, δ¹³C-depleted reservoir during deglaciation would cause the precipitation of carbonates with light δ¹³C values in shallow water. Interestingly, Shen et al. (2005) reported a possible carbon isotopic gradient of ~3‰ along a paleoenvironmental transect from shelf to deep basinal sedimentary facies in the Marinoan Nantuo cap carbonate. However, with regard to the Shuram anomaly, no lateral gradient is decipherable between the δ¹³C data sets from shallow and deep water facies (compare Figs. 2.15 and 2.16). In addition, this model cannot explain values as negative as -12‰.

It has also been proposed that negative carbon isotopic ratios in carbonate might reflect flows from a light carbon reservoir such as methane hydrates/clathrates, either in the form of seabed hydrates or in terrestrial permafrost (Kvenvolden, 1998; Kennedy et al., 2001; Jiang et al., 2003; Kasemann et al., 2005). This mechanism has been invoked to explain the highly variable and very negative values of the Doushantuo ‘cap carbonate’ in south China (averaging -5‰ with spikes down to -5‰; Jiang et al., 2003). Schrag et al. (2002) also believed that a large oceanic methane reservoir was slowly released, having built up through high amounts of organic carbon burial due to high river discharges from continental areas assembled at low latitude. In the case of the Shuram anomaly, the severity of the δ¹³C perturbation and its protracted steady recovery towards positive values make it unlikely that these mechanisms acted as the main driver of the Shuram anomaly.

Rothman et al. (2003) used mathematical model to suggest that fluctuation in the size of the reservoir of dissolved organic carbon in ocean water or changes in the fractionation associated with organic production would lead to major variations in the isotopic record. Pulses in the remineralization (enhanced by glaciation) or bioremineralization (enhanced by early metazoans) of a large oceanic organic carbon pool could potentially explain the negative excursions recorded by Neoproterozoic carbonates. However, such a carbon pool remains to be physically identified. The Shuram anomaly conclusively shows that mantle input is insufficient to generate -12‰ values and probably oxidation of organic matter (around -25 to -30‰; Hayes et al., 1999) is required. Rothman et al.’s (2003) non steady-state model demonstrates that long and severe negative perturbations are possible and
most likely linked with evolution of the biosphere during the Neoproterozoic (see also Grey et al., 2003). Slow remineralisation might be reflected in the steady recovery of the negative anomaly.

All of these hypotheses have elements that may go some way to explaining the Shuram shift. However, with the possible exception of Rothman et al.’s (2003) hypothesis of remineralization of dissolved organic carbon, they have difficulty in explaining the extremely long duration of the negative excursion. Others cannot explain the extremely negative isotopic values. There are a number of linked elements in the models so far proposed, including high rates of organic carbon burial, anoxic ocean waters, release of large methane reservoirs, and remineralization of dissolved organic carbon.

In summary, the cause, or causes, of the Shuram shift remain enigmatic, but any explanation must be guided by a number of pertinent observations: (1) the carbon isotopic excursion is in phase with relative sea level, effectively occupying the second grand cycle of the Nafun group; (2) the onset of the fall in δ¹³C corresponds with the demise of the Khufai carbonate factory and the influx of siliciclastics onto the carbonate ramp; (3) the nadir of the carbon isotopic excursion corresponds with a period of maximum flooding, so the period of fall of carbon isotopic values roughly approximates the period of transgression; (4) the sedimentology of the Shuram Formation in the Huqf area demonstrates that the Shuram marine basin was typified by high energy storm conditions; and (5) the change from Khufai to Shuram most likely corresponds to a change from arid to pluvial-humid conditions.
2.6. Conclusions

The Nafun Group of the Huqf Supergroup of Oman records an essentially continuous period of deposition from the post-Marinoan cap carbonate of the Hadash Formation to close to the Precambrian-Cambrian boundary. It comprises two grand cycles of marine siliciclastics to ramp carbonates. Each cycle is initiated by a major transgression, the formation of a relatively deep siliciclastic marine basin, followed by gradual shallowing up into progradational ramp carbonates.

The boundary between the two grand cycles is at the Khufai-Shuram boundary. Both the Khufai and Shuram formations are represented by shallow water facies in the Huqf area of east-central Oman, and by deeper water facies in the Jabal Akhdar of north Oman. In the Huqf area, the Khufai Formation is composed of fetid carbonates, which pass up into m-scale peritidal cycles, indicating progradation of a carbonate ramp. In the Jabal Akhdar however, the Khufai Formation is dominated by black, pyritic limestones deposited in deeper ramp conditions. The Shuram Formation is composed of storm-influenced sandstones, siltstones and limestones in the relatively shallow water Huqf area. In the Jabal Akhdar, however, the Shuram Formation is dominated by deep water, organic-rich dolomitic mudstones and bleached siltstones at the base, which pass up into very thick, purple, monotonous siltstones and shales with thin subordinate carbonates. Shallow water conditions were established over basement highs, which may have been partially inherited from a phase of important rifting during the deposition of the underlying Abu Mahara Group.

The carbonates of the upper grand cycle, comprising the Shuram and Buah Formations, contain a remarkable negative excursion in $\delta^{13}$C values, which start to fall precipitously from the uppermost Khufai Formation, reach a nadir close to the maximum flooding zone of the Shuram, and then recover monotonically through nearly 1 km of stratigraphy, with a crossing point within the Buah Formation. The amplitude of the excursion is from +5 ‰ in the Khufai carbonates, to -12 ‰ in the lower Shuram Formation, and the duration is believed to be c. 50
My. This appears to be the greatest $\delta^{13}C$ negative excursion in inorganic carbon of marine carbonates of Earth history.

The explanation for the Shuram shift is enigmatic. However, there are a number of pointers derived from the sedimentology and stratigraphy of the Nafun Group. Importantly, the carbon isotopic excursion is roughly in phase with relative sea level. The start of the fall in carbon isotopic values commences with the influx of siliciclastics and the demise of the Khufai carbonate factory, and the nadir occurs at the level of the maximum flooding zone of the lower Shuram, so the falling segment of the excursion coincides with a period of transgression. We also note that the transition from the Khufai to the Shuram Formation most likely records a climate change from arid to humid and stormy.

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### Appendixes

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Figure 2.16: Organic versus inorganic Carbon content. Note that analytical precision is about ± 1.

Table 2.3: δ¹³C, δ¹⁸O isotope and carbon yield (average error is about ± 10) data of sections in the Jabal Akhdar and Huqf region. Sample are numbered from stratigraphic base of the log. See plot in figures 2.15 and 2.16.
Figure 2.17: Gamma ray logging through the entire Shuram Formation. Resolution is about one data point every 5 m.
Chapter 3

Parasequence development in the Ediacaran Shuram Formation (Nafun Group, Oman): High resolution stratigraphic test for primary origin of negative carbon isotopic ratios

Erwan Le Guerroué, Philip Allen and Andrea Cozzi

*Basin Research, in press June 2006*

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Abstract

Neoproterozoic carbonates are known to show exceptional variations in their carbon isotopic ratios, and in the absence of biostratigraphy and a firm geochronological framework, these variations are used as a correlation tool. However, it is controversial whether the carbon isotope record reveals a primary oceanographic signal or secondary effects such as diagenesis. The Shuram Formation of the Nafun Group of Oman allows a stratigraphic test of this problem.

The Nafun Group (Huqf Supergroup, Oman) in the Huqf area of east-central Oman consists of inner carbonate ramp facies of the Khufai Formation overlain by marine, storm-generated, red and brown siltstones of the Shuram Formation. Toward its top, the Shuram Formation is composed of distinctive shallowing-upward, 4-17 m-thick parasequences cropping out continuously over 35 km, which show recessive swaley cross-stratified siltstones capped by ledges comprising wave-rippled, intraclast-rich ooidal carbonate. These storm-dominated facies show a regional deepening in palaeobathymetry towards the south.

The carbonates of the Shuram Formation are marked by an extreme depletion in $^{13}$C in bulk rock. $\delta^{13}$C values quickly reach a nadir of -12‰ just above the Khufai-Shuram boundary and steadily return to positive values in the overlying mainly dolomitic Buah Formation. The Shuram excursion is thought to be c. 50 Myr in duration and extends over 600 m of stratigraphy. Carbon isotopic values show a systematic
variation in the parasequence stack, with values varying both vertically through the stratigraphy (~2‰ per 45 m) and laterally in the progradation distance (~1‰ over 35 km). This supports a primary, oceanographic origin for these extremely negative carbon isotopic values and independently argues strongly against diagenetic resetting.
Neoproterozoic sedimentary rocks record periods of extreme swings in carbon isotopic ratios. These swings have commonly been associated with putative snowball Earth events (Hoffman et al., 1998; Hoffman and Schrag, 2002), but long-lived (tens of Myr) major negative excursions are also found where there is no evidence for widespread glaciation (Saylor et al., 1998; McKirdy et al., 2001; Zhang et al., 2005; Le Guerroué et al., 2006). Such excursions show extreme variation from +10 to -12‰ and appear to be up to tens of millions of years in duration (Kaufman et al., 1993; Halverson et al., 2005; Le Guerroué et al., 2006b; Bowring et al., in review). The fact that a carbon isotope excursion of this magnitude can be detected in marine sedimentary rocks from several continents suggests that it represents a global oceanographic phenomenon (Saylor et al., 1998; Condon et al., 2005; Halverson et al., 2005; Le Guerroué et al., 2006b), reflecting the composition of seawater from which carbonate minerals were precipitated. However, the presence of a carbon isotopic excursion of this amplitude and duration is difficult to explain in terms of mass balance. Such negative values (-12‰) are also well below those associated with a complete cessation of biological productivity (Broecker and Peng, 1982). Phanerozoic negative anomalies are usually short-lived and represent excursions of less than -2‰ (e.g., Hayes et al., 1999; Hesselbo et al., 2000; Galli et al., 2005). There is therefore continuing debate as to the primary origin of these long-lived excursions in steady state conditions and their legitimate use as a basinal and global correlative tool in the Precambrian.

The Shuram Formation (Nafun Group, Huqf Supergroup) crops out extensively in the Huqf area of east-central Oman. It is extremely well exposed for over 35 km in a roughly N-S oriented escarpment in the north of the Huqf area, where it displays a stack of shallowing-upward cycles or parasequences (sensu Vail, 1987; Mitchum Jr and Van Wagoner, 1991). Parasequences are common in both siliciclastic- and carbonate-dominated Phanerozoic stratigraphy, but sequence stratigraphic studies dealing with Precambrian examples are scarce (Christie-Blick et al., 1995; Gehling, 2000). The Shuram parasequences are of mixed
lithology, with deeper-water siliciclastics passing up into shallower-water carbonate deposits. Siliciclastic input and carbonate production were most likely controlled by a subtle balance of environmental factors driven by relative sea-level change (Tucker, 2003; see modelling section below).

Carbon isotope stratigraphy assists in shedding light on correlation problems in Precambrian strata where biostratigraphy and radiometric age data are lacking (Knoll et al., 1986; Kaufman et al., 1993; Cozzi et al., 2004a). The particular benefit of an investigation of the parasequences of the Shuram Formation is that their carbonate components yield carbon isotopic values embedded within a long-term secular trend that is reproduced throughout the outcrop areas of Oman and in the subsurface borehole records of the Oman salt basins (Burns and Matter, 1993; Amthor et al., 2003; Cozzi and Al-Siyabi, 2004; Allen and Leather, 2006; Le Guerroué et al., 2006b). At this high resolution scale, bulk rock inorganic carbon isotopic values are shown to reflect stratigraphic position within the parasequence stack, and each individual parasequence shows a predictable trend in δ¹³C values in the direction of sediment progradation. This study therefore provides the sedimentological and stratigraphic context that independent of diagenetic arguments validates the primary origin of the large negative carbon isotopic excursion in Oman as an oceanographic phenomenon. By doing so, it supports the use of the Precambrian highly negative δ¹³C values as a valid correlation tool, but of course does not imply that all Neoproterozoic carbon isotopic excursions can be used in such a manner.

3.2. Geological setting

The Neoproterozoic Huqf Supergroup of Oman crops out mainly in northern (Jabal Akhdar) and central (Huqf) areas of Oman, the latter representing a palaeohigh during Neoproterozoic time, cored by granodioritic crystalline basement dated at 820+Ma (Allen and Leather, 2006). The Huqf Supergroup is subdivided into three groups: the Abu Mahara, Nafun and Ara groups (Glennie et al., 1974; Gorin et al., 1982; Wright et al., 1990; Le
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The Nafun Group in the Huqf area is consistently represented by shallower-water facies relative to those in the Jabal Akhdar (Cozzi and Al-Siyabi, 2004; Le Guerroué et al., 2006a; Le Guerroué et al., 2006b). The Nafun Group overlies the presumed Marinoan (ending at c. 635 Ma; Allen et al., 2004; Le Guerroué et al., 2006b; Bowring et al., in review) glacigenic Fiq Member of the Ghadir Manqil Formation (Abu Mahara Group),

Figure 3.1: General geological setting of the Nafun Group of the Huqf area of east-central Oman. A: Map of Oman showing Neoproterozoic outcrops. B: Summary stratigraphic column of the Nafun Group in the Huqf area with associated δ¹³C record. C: Geological map of the Huqf area showing studied localities. D: Composite sedimentary log of the Shuram Formation and detail of the parasequences in the Wadi Aswad sections.
and is bracketed upward by the Ara Group dated in its middle at 542.0±0.3 Ma (Amthor et al., 2003). The Nafun Group, which is thought to have been deposited during a period of thermal relaxation after the rifting phase of the Ghadir Manqil Formation, comprises at its base the strongly transgressive cap carbonate of the Hadash Formation. Above this, it consists of two major siliciclastic to carbonate cycles (Fig. 3.1; Allen and Leather, 2006; Le Guerroué et al., 2006a). The first cycle consists of marine shales, siltstones and sandstones of the Masirah Bay Formation (150 to 300 m thick; Allen and Leather, 2006), which pass gradationally upwards into the carbonate ramp of the Khufai Formation (50 to 300 m thick; Gorin et al., 1982; Wright et al., 1990; McCarron, 2000). The second cycle comprises thick marine siltstones, shales and subordinate carbonates of the Shuram Formation (250 to 700 m thick; Le Guerroué et al., 2006a), which pass up gradationally into the carbonate ramp of the Buah Formation (250 to 350 m thick; Gorin et al., 1982; Wright et al., 1990; Cozzi and Al-Siyabi, 2004). This large-scale couplet (Shuram-Buah formations) is associated with a major δ13C perturbation in its carbonate record (Fig. 3.1; Burns and Matter, 1993; Cozzi and Al-Siyabi, 2004; Le Guerroué et al., 2006a; Le Guerroué et al., 2006b). The pattern of variation in δ13C from the Khufai to the Buah Formations shows a major positive-negative-positive pattern. Values as positive as +5‰ characterize the top of the Khufai limestones, followed by a precipitous fall in the basal Shuram Formation to values as negative as -12‰ a few tens of metres above the formation boundary. A long, slow, sub-linear trend to less negative δ13C values occurs through the overlying Shuram and Buah Formations, eventually climbing back to positive values (+2 ‰) at the top of the Buah Formation (Fig. 3.1; Burns and Matter, 1993; Cozzi et al., 2004a; Cozzi et al., 2004b; Le Guerroué et al., 2006b). Based on subsidence modelling and the age distribution of detrital zircons, this peculiar perturbation of the carbon cycle is thought to have persisted for approximately 50 million years, representing most of the Ediacaran period (see details in Le Guerroué et al., 2006b).
3.3. Parasequences of the Shuram Formation

The Shuram Formation, comprising the lower part of the second of the Nafun ‘grand cycles’, lies on top of the shallowing-upward carbonate ramp of the Khufai Formation (Fig. 3.1; McCarron, 2000; Le Guerroué et al., 2006a). The Shuram Formation, which reaches 250 m in thickness in the Huqf area, marks an abrupt deepening of the shelf (Fig. 3.1; Le Guerroué et al., 2006a). It is divided into three members. The Lower Member comprises siltstones and subordinate siliciclastic sandstones in the Jabal Akhdar, but transgressive carbonates and subordinate siliciclastic sandstones in the Wadi Aswad N

Figure 3.2: Parasequence correlation panel. A: Southern sections (MD3 and Na1) showing δ¹³C record. B: Correlated sections of Wadi Aswad with primary current lineation (PCL) palaeocurrent indicators compiled per parasequence (rose diagrams). Distances between locations are given in kilometres. The main cross-stratification palaeocurrent mode is to the SW (see text). See figure 1 for location of logs. C: Field panorama around locality WA1 showing the parasequences of Wadi Aswad. Carbonate facies form prominent ledges in the landscape. Parasequence 4 and 5 are not exposed all the way along Wadi Aswad. Top parasequence is about 10 m thick.
The Middle Member in the Huqf area is characterized by hummocky cross-stratified and wave-rippled red siltstones with numerous storm-reworked fine-grained limestone beds. The Upper Member, particularly well developed in the northern extremity of the Huqf area (Wadi Aswad), shows shallowing-upward storm-dominated parasequences (from 4 to 17 m thick; Gorin et al., 1982; Wright et al., 1990; McCarron, 2000; Cozzi and Al-Siyabi, 2004; Fig. 3.1). Facies are arranged into gradually shoaling packages (Figs. 3.1 and 3.2) that crop out continuously along the 35 km extent of Wadi Aswad. More sections are available farther south of Wadi Aswad (Nafun area of Mukhaibah Dome) but the Upper Member occurs here in slightly deeper water facies (Figs. 3.1 and 3.2) without well-developed parasequences. Five distinctive parasequences have been identified in the field in Wadi Aswad, overlain by a package of condensed, amalgamated, m-scale cycles, which in turn pass up into 10 m of shales marking a maximum flooding zone (Cozzi and Al-Siyabi, 2004; Le Guerroué et al., 2006a; Fig. 3.1 and 3.2). The Upper Member of the Shuram Formation then passes into dolomites with abundant edgewise conglomerates of the basal Buah Formation (Cozzi and Al-Siyabi, 2004; Cozzi et al., 2004b), the boundary being defined by the disappearance of siltstone intercalations (Gorin et al., 1982).

Figure 3.3: Idealized sedimentological section of parasequence 1 showing all the characteristic structures within one mixed-lithology cycle. See discussion and interpretation in the text.
Figure 3.4: Siltstone facies. a) Primary current lineation. Note (arrows) two different orientations on different laminae. b) Swaley cross-stratification. Book is about 15cm long. c) Bedding plane covered with trochoidal, irregular wave ripples. d) Low-angle stratification grading to swaley cross-stratification overlain by climbing symmetrical wave ripples. Pencil is 15 cm. e) Water-escape pipe within liquefied laminated siltstones. f) Foundered balls and pseudonodules of laminated siltstone within fluidized (homogenized) bed truncated by swaley cross-stratified siltstones. g) Swaley cross-stratified coset showing one set with pervasive soft sediment deformation. Coin for scale. h) Wavy laminated and draped surface eroding into planar laminae. On all pictures coin is about 2.5 cm diameter.
3.3.1. Storm-Generated Siltstone Facies

The siltstone facies is volumetrically dominant within the Shuram parasequences, occurring in the lower part of each cycle (Figs. 3.1; 3.2 and 3.3). This facies comprises pinkish to yellowish siltstones grading up to subordinate very fine to fine sandstones (Fig. 3.4). In terms of primary sedimentary structures, the siltstone facies is dominated by swaley cross-stratification (SCS; Fig. 3.4b) (Leckie and Walker, 1982). Long wavelength swales, up to 5 m, are symmetrically to slightly asymmetrically filled above a sharp, erosional, and locally highly irregular, lower bounding surface. Locally, undulating erosional surfaces, cut into dark brown argillaceous siltstones, have overhanging steps; overlying swaley cross-strata downlap onto this undulating surface and fill the positive step. Flat and very low-angle laminae commonly occur below or pass gradually upwards into SCS packages, in laterally continuous planar-laminated, locally gently undulating units (Fig. 3.4i). Small wave ripples locally ornament amalgamation surfaces between swaley cross-sets. Parting and primary current lineation (PCL) are abundant on surfaces of planar laminae as well as on very low-angle swaley cross-strata (Figs. 3.3 and 3.4a). Wave ripples are 5-10 cm in spacing, and are commonly trochoidal in profile and linear in crestline (Fig. 3.4c,d and j; Allen, 1997), locally reworking tops of beds. Some examples are ladder-back, with primary and secondary crestline patterns preserved. Aggradational climbing ripples are found in cosets with grain size and wavelength diminishing upward, locally draped by shales (Fig. 3.4j).

Ubiquitous soft-sediment deformation occurs in the form of complexly folded laminae, load balls and fluidization pipes (Fig. 3.4h,i and j), locally homogenized with vestiges of deformed laminae. Asymmetrically deformed (sheared), slumped and convoluted layers are commonly erosionally truncated above. In some places, arrays of small micro-normal faults cut the laminae. Flat to low-angle detachment surfaces are also present, showing limited extensional adjustment (slip of 4 cm) of the sediment pile, together with thin rafts of red siltstone deformed into large irregular folds (30-70 cm).
The abundance of low amplitude, large wavelength swales and the rare preservation of hummocky antiforms suggest proximal storm-influenced water depths as shallow as several metres (Leckie and Walker, 1982; Datta et al., 1999). Their association with flat laminae and primary current lineation supports the activity of high-energy currents, typical of the shoreface during storms (Dott and Bourgeois, 1982). Trochoidal wave ripples are also diagnostic of wave action in shallow water depths (Allen, 1997). Wave ripple crestlines are approximately perpendicular to the direction of primary current lineation (SW-NE), showing that high energy currents acted in the same direction as wave propagation, indicating that water depths were shallow enough for friction to dominate over geostrophic turning (Allen, 1997). Sea-bed currents were clearly powerful and erosive, cutting steep notches in the more cohesive, argillaceous sea-bed sediments and broad undulating scours in the less cohesive coarse silts to very fine sands. Such currents involved a net mass transport to the SW, as indicated by the downlap of swaley cross-strata over lower bounding surfaces. Currents reduced in intensity to allow wave rippling of the sediment surface before burial under new episodes of bedform migration. Currents also reduced in intensity for longer time periods, allowing the accretion of climbing wave-ripple cosets, when the oscillatory component of the flow field dominated over the steady component (Scott, 1992; Cheel and Leckie, 1993).

The overall picture is therefore of an alternation of relatively quiet water deposition of finely laminated argillaceous siltstones, with strong, initially erosive, pulsating storm currents. Storm currents, which caused a net south-westward sediment transport, reduced in intensity to allow residual wave action to ripple the rapidly aggrading sea-bed. Post-storm water depths were very shallow, as indicated by the ladder-back ripples, which were formed by wave propagation under local winds rather than the approach of open marine swells. The resulting m-thick stacked units point to the high amount of deposition associated with major storms taking place in water depths typical of the upper shoreface.

The range of soft-sediment deformation features indicate water escape and gravitational collapse, typical of rapidly deposited sediments such as storm deposits (Mills,
1983). Shortly after deposition, disturbance and dewatering of the bed may have been caused by cyclic wave loading under storm conditions (Staroszczyk, 1996), or by seismic shaking (Allen, 1986). It is certain that the original sediment was a well-laminated siltstone, since the lamination is intricately folded and locally destroyed. In places, fluid escape velocities were high enough to fluidize the bed, leading to homogenization. In general, however, fluid escape velocities were lower, allowing folding by liquefaction. Arrays of small brittle fractures (slip of < 1 cm) and low-angle detachments demonstrate local extension. However, the pervasive and ubiquitous folding shows that most of the deformation involved gravitational collapse and ductile shortening.

The consistently strained load balls erosionally truncated by swaley cross-strata indicate that soft-sediment deformation took place in near-surface sediment, and that the upper portion of the sea-bed underwent deformation under the powerful shear of overlying currents. An origin of the deformation by fluid-sediment interaction during powerful storms therefore appears more likely than an origin by seismic shaking.

### 3.3.2. Intraclast-rich carbonate facies

The vertical transition between the siltstone and carbonate facies is commonly gradual over a few dm. The thickness of the carbonate facies varies from a few centimetres in the older parasequences to up to 7 m in the youngest (Figs. 3.1, 3.2 and 3.3). The

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**Figure 3.5: Carbonate facies.**
- a) Typical parasequence carbonate facies with a cross-stratified siltstone to very fine sandstone interbed. Note the low-angle, planar and swaley cross-stratification in the light grey intraclast-rich carbonate bed.
- b) Bidirectional cross-stratification. Main direction is to the SW.
- c) Ooidal grainstones. Note the presence of ooids in both matrix and carbonate intraclasts.
- d) Large-scale swaley cross-stratification. Swiss army knife for scale.
- e) Surface view of edge-wise conglomerate.
- f) Cross-stratified carbonate bed with a trochoidal wave rippled top. Low-angle large-scale cross-stratified siltstones occur below and above, where they drape the underlying ripple topography.
- g) Edge-wise conglomerates in a nest in cross-section.
- h) Carbonate with orange silt filling pockets and crevices between clasts. Lens cap is 4 cm.
- i) Swaley cross-stratified siltstones truncated by carbonate bed (below compass) passing from intact bed to breccia of intraclasts towards the left. Compass is 8cm width. On all pictures coin is about 2.5 cm.
facies is made mainly of recrystallised and partially dolomitized ooidal-pisoidal grainstones containing numerous cm- to dm-size rounded to angular intraclasts arranged along the main flow direction (locally imbricated) or more commonly organized into edgewise nests (Fig. 3.5).

Ooidal grainstones also occur as planar, tabular and cross-stratified <10 cm-thick beds. Beds show evidence of in situ brecciation, with a transition from allochthonous clast-rich units passing laterally into undisturbed cross-stratified beds. The broken limestone clasts show a progression from angular pieces close to the intact parent bed, to rounded where they are far separated (Fig. 3.5). This range of states of break-up into intraformational breccias is especially noticeable in the first carbonate beds marking the 1-2 m-thick transition between the two facies. Ooids vary considerably in size from bed to bed, ranging from 0.1 mm to 5 mm in diameter, and are generally well sorted, the younger parasequences containing the larger examples (Fig. 3.5c). Radial fabrics of ooids are formed of very thin micritic lamellae (Strasser, 1986). Orange to brown siliciclastic silt mixed with the ooids is commonly preferentially preserved underneath <5 cm-sized limestone intraclasts and in interstices between intraclasts.

The facies shows abundant small-scale planar (<10 cm) cross-sets with angular or tangential cross-strata, as well as undulatory styles of cross-stratification (SCS) and trains of trochoidal wave ripples (Fig. 3.5a,d). Some ripple cross-sets evolve from undulatory lamination to a bidirectional build-up with a symmetrical upper surface. In general, the tops of beds are either rippled, or erosive and irregularly covered with limestone clasts derived locally from the underlying bed. Palaeocurrents from the thin limestones intercalated with laminated siltstones are bidirectional, with a dominant mode to the SW. Planar sets locally contain rippled cross-strata lined with a fine sediment drape, indicating a pause in bedform migration.
The association of undulatory (swaley) cross-stratification, wave ripple cross-lamination and planar, tabular limestone beds are indicative of storm and fair-weather processes in shoreface water depths. Intense current or wave activity is shown by the high-energy structures, bidirectional palaeocurrents, local break-up of the partially lithified sea-bed, and formation of edgewise conglomerates (Mount and Kidder, 1993). The interbedding of carbonate and siliciclastic beds in the transition zone below the carbonate ledge, and the mixing of ooids with siliciclastic silt and very fine sand, demonstrate that the carbonate factory operated at the same time as siliciclastic deposition, in laterally time-equivalent locations. The preservation of orange-brown quartzose silt in pockets and crevices between limestone clasts suggests late infiltration into a clast-supported framework. This might have taken place on the sea-bed during quiet periods, or by aeolian transport onto an emergent, rubbly, sediment surface. These surfaces may be diagnostic of sediment bypass, but no clear evidence for surface exposure such as karstification features has been recognized.

The lack of evidence for soft-sediment deformation in the carbonate facies indicates a lithological control on high-pore fluid pressure in the siltstone facies.

Although depositional water depths were shallow, facies remain uniform throughout the parasequence stack at Wadi Aswad. No mudstone, stromatolite or peritidal facies have been recognised, and there is little evidence for erosional surfaces, soil formation or karstification in the upper parts of shallowing-up parasequences.

### 3.3.3. Paleocurrent and water depth indicators

The orientation of primary current lineation indicates currents oriented NE-SW. This orientation is regionally consistent among all parasequences (Fig. 3.2). The main cross-stratification direction is to the SW with a subordinate mode to the NE (Fig. 3.5b, f). The wave propagation direction is also SW-NE based on crestlines of vortex ripples on the tops of beds. The filling direction of troughs by siltstones and very fine sandstones above the
wave-rippled top surface of carbonate beds is also to the SW. These different palaeocurrent indicators are mutually consistent: assuming that the net sediment transport was offshore-directed, the direction of storm attack was from the SW onto a NW-SE oriented coastline.

The preservation of wave ripples with steep trochoidal profiles allows the estimation of the wave period of formative waves and a range of depositional water depths (see discussion of technique in Allen, 1981 and Allen, 1984, and a recent application in Allen and Hoffman, 2005b). Taking a steep, trochoidal wave ripple with spacing 250 mm and height 45 mm (vertical form index of 5.5) preserved above a limestone bed made of medium-coarse ooids, a maximum wave period of 5.5 seconds has been calculated, assuming the wave ripple was formed at the threshold condition. Formation at the threshold is supported by the preservation of the trochoidal top profile prior to cessation of sediment movement. However, if the dimensionless wave shear stress was in excess of the threshold value (Jerolmack and Mohrig, 2005, and reply by Allen and Hoffman, 2005a), the formative wave period reduces to c. 4 secs. Using the 5.5 secs estimate of the wave period, formative waves must have acted in water depths of less than 20 m, and most likely of less than 10 m. This supports the view that the parasequences were formed on a storm-dominated shoreface. By extending this argument, the swaley cross-stratified siltstones were also formed in similar or slightly greater water depths, and the carbonate components of parasequences were deposited in water depths shallower than 10 m. However, the period of the waves that generated the primary sedimentary structures in the siltstones and their subsequent soft sediment deformation, or that assisted the in situ break-up of the limestone beds on the sea floor, is not known. Since the trochoidal wave ripples signify the end of deposition following bedform migration and break-up of the sea-bed, it is highly likely that storm conditions were more intense in the preceding phase.
Table 3.1. Carbon and oxygen isotope data (from locality 1 to 9) for each parasequence, and average parasequence thickness and carbonate (Carb.) to siltstone (Silt.) facies ratio.
3.4. The space-time evolution of the parasequences

The strike of the outcrop belt is oriented parallel to the palaeocurrent directions (Fig. 3.1 and 3.2) and extensive exposures provide a rare opportunity to trace parasequence architecture in the down/up-dip direction. Based on the 5 most complete parasequences of Wadi Aswad, the water depth trend suggested by lateral thickness and facies variations shows a primary deepening toward the south, which is consistent with the palaeocurrent direction (Fig. 3.2). This is supported by sections in the Mukhaibah Dome and Nafun village areas (MD3 and Na1; Fig. 3.1), which show deeper water facies (Fig. 3.2). Local variation in parasequence thickness is present along Wadi Aswad and is interpreted as due to basinal relief. In addition, section 10 located north of Wadi Aswad records the northward deepening of depositional environments at the northern limit of the Huqf palaeohigh described in other regional studies (Le Guerroué et al., 2006a). The increasing parasequence thickness upwards in the stack, combined with the increased carbonate preservation in the same direction (Fig. 3.2 and Table 3.1) suggest a trend in time of increased accommodation, which was easily filled by a plentiful supply of both siliciclastic and carbonate sediment. This accommodation was most likely generated by eustasy superimposed on background basin subsidence (Vail, 1987; Adams and Grotzinger, 1996; see also modelling section below).

3.5. Progradation and the carbon isotope record

3.5.1. Methodology

Bulk-rock inorganic carbon and oxygen isotope measurements have been carried out on each of the carbonate ledges within the Wadi Aswad parasequences (measurements are shown in Table 3.1) as well as on the carbonate beds of the southern sections at Mukhaibah Dome and Nafun (MD3 and Na1; Figs. 3.2 and 3.3). Hand specimens were drilled with 1 - 5 mm dental drill bits from freshly cut rock slabs avoiding sparry cement and vein material. Ooid rich samples were preferentially analysed, and where possible ooids have been directly
micro-drilled for isotopic measurements. The C and O isotope composition of powder from
the carbonate samples was measured with a GasBench II connected to a Finnigan MAT
DeltaPlus XL mass spectrometer, using a He-carrier gas system (methods adapted after
Spoetl and Vennemann, 2003). Samples were normalized using an in-house standard
calibrated against δ\textsuperscript{13}C and δ\textsuperscript{18}O values of NBS-19 (+1.95 und –2.20 ‰, relative to PDB).
External reproducibility for the analyses estimated from replicate analyses of the in-house
standard (n=6) is ±0.07‰ for δ\textsuperscript{13}C and ±0.08‰ for δ\textsuperscript{18}O.

3.5.1. Diagenesis

In thin section, carbonates show partial dolomitization in a texture that is mainly
recrystallised, all samples showing similar microfacies. Some carbonates of the Shuram
Formation were screened under a cathodoluminescence microscope. Good preservation of
low-Mg calcitic carbonates in the Shuram Formation is indicated by the radial fabrics of the
oolids (McCarron, 2000). Lack of CL zoning in the cements suggests that they did not re-
equilibrates under anoxic conditions in the sulphate-reducing zone (McCarron, 2000). Some
bulk carbonate samples were geochemically screened (Burns and Matter, 1993; Burns et
al., 1994; McCarron, 2000; Leather, 2001; see data in Le Guerroué et al., 2006a). Mn/Sr
ratios in screened samples were <10, with many samples <3, which has been taken to
imply that negligible diagenetic alteration has occurred (Kaufman and Knoll, 1995, but see
McKirdy et al., 2001). Additional isotopic data, which are consistent with the values in the
samples screened for diagenetic alteration, were therefore used without further screening.

On the basis of geochemical screening, therefore, carbon isotopic values may be
interpreted as primary, whereas oxygen values show values randomly ranging from -6 to
-9‰ (Table 3.1) indicating probable mixing with meteoric fluid (Fairchild et al., 1990). The
rarely preserved positive δ\textsuperscript{18}O values occur in a different part of the Shuram δ\textsuperscript{13}C excursion
(at the base of the excursion; see data in Le Guerroué et al., 2006a). During early diagenesis,
alteration of the oxygen isotopes is common, leaving the carbon isotopic values intact
The strongest indication that the δ\(^{13}\)C values are primary is the regional reproducibility of the signal, with little variation regardless of facies changes between outcrop sections and well data (Burns and Matter, 1993; McCarron, 2000; Cozzi and Al-Siyabi, 2004; Cozzi \textit{et al.}, 2004a; Le Guerroué \textit{et al.}, 2006a; Le Guerroué \textit{et al.}, 2006b). The analysis of carbon isotopic trends within the parasequence stack of the Upper Member of the Shuram Formation strongly supports this view.

**3.5.2. Results**

Systematic measurements of δ\(^{13}\)C along individual parasequences in Wadi Aswad show a persistent trend of less negative values when moving laterally to the south (Fig. 3.6). Some imprecision occurs due to analytical errors (instrumental precision is ±0.07‰ for δ\(^{13}\)C measurement), possible local minor alteration of the primary δ\(^{13}\)C signal, and sedimentary reworking of older material along the parasequence. Given the southward progradation direction, time lines cut across facies packages, with lateral facies changes between co-existing shallow-water carbonates and relatively deeper-water siliciclastics (Fig. 3.7). The carbon isotopic ratio is isochronous along individual time lines, making its variations predictable as the result of the development of the prograding parasequence stack (Fig. 3.7).

If we assume a 50 Myr duration of the principal Nafun Group δ\(^{13}\)C excursion (Le Guerroué \textit{et al.}, 2006a), the linear rate of carbon isotope recovery over the Shuram to Buah stratigraphic interval is about 0.3‰/Myr. Assuming a δ\(^{13}\)C variation from -10‰ in the oldest Shuram parasequence to -7.5‰ in the youngest (Fig. 3.6), this implies a total duration of about 8 Myr for the 5 parasequences and an average of c.1.6 Myr per parasequence. This average duration is difficult to attribute to orbital forcing, but cycles in the upper Ara Group (Amthor \textit{et al.}, 2003; Amthor \textit{et al.}, 2005) are radiometrically constrained to last 1.3 Myr.
per cycle (3.5 cycles present) (Bowring et al., in review). The greater thickness of the Ara cycles (50-200 m) would require considerably higher tectonic subsidence and sediment accumulation rates during Ara times compared to the Shuram. Lateral variation in δ¹³C is of the order of <1‰ over a 25 km N-S distance along each parasequence. This gives an average variation of 0.04‰ per km and a progradation rate of ~7.5 km/Myr. The time-averaged vertical sediment accumulation rate, corrected for compaction, was of the order

![Figure 3.6: Composite isotope plot for the carbonate components of 5 parasequences at Wadi Aswad. Colour key shows location of section in a north to south profile. For each parasequence, isotopic ratios become less negative towards the south, which is the progradation direction.](image-url)
of ~12 m/Myr, which is typical of the late stages of failed rifts or passive-margins (Allen and Allen, 2005).

Figure 3.7: Conceptual model of the Shuram parasequence stack with isotope variation following a linear secular trend defined in figures 1 and 6. Four parasequences are drawn bounded by sequence boundaries (SB) and labelled by circled number 1 to 4. A: Chronostratigraphic depositional model represented in dip section architecture, showing less negative carbon isotope ratios in the direction of sedimentary progradation and vertically through the stack of parasequences. B: Same model represented as a Wheeler diagram showing hiatuses due to non-deposition or sediment bypass, as predicted by numerical modelling (Fig. 3.8). $\delta^{13}C$ values are identical along individual time lines (numbered 0 to 10) and follow a secular trend. Moving along one carbonate facies crosses different time lines resulting in an expected variation of the $\delta^{13}C$ record through time.
3.6. Modelling

3.6.1. Methodology

The stack of parasequences in the Shuram Formation at Wadi Aswad has been investigated using a 1-D model based on simple algorithms for accommodation and sediment supply (see Allen and Allen, 2005, p. 269-275, and code in supported online material). The model allows prediction of the thickness and water depth variation through a stack of parasequences based on input of the eustatic variation, maximum sediment supply rate and tectonic subsidence rate (Fig. 3.8A). No distinction between siliciclastic and carbonate sediment is made, and no sediment dynamics are implied. Furthermore, cyclicity is assumed to be driven by a eustatic control rather than by unforced internal dynamics (cf. Burgess et al., 2001). When water depth is very low, and accommodation tends to zero, further autogenic sediment production or alloogenic influx is assumed to be transported into deeper water and not preserved in the 1D profile.

Long-term tectonic subsidence, balanced roughly by sediment supply expressed as a vertical velocity, is assumed to be linear (11 m/Myr) over the time span of the deposition of the parasequence stack. To replicate the cyclicity of the parasequence stack, two eustatic sinusoids were combined, one with amplitude and duration of 30 m and 1.6 Myr, and the other with 10 m and 16 Myr (Fig. 3.8A). Sediment supply is coupled to the eustatic variation, with a maximum (14 m/Myr; i.e. 6 and 8 m per cycle) at times of relative sea-level fall and a minimum at times of relative sea-level rise (Fig. 3.8A). The water depth variation through time plotted against the potential sediment accumulation gives an illustration of

Figure 3.8: Numerical modelling parameters and results. A: Various parameters used in the models plotted against time. PSA: Potential sediment accumulation. B: Potential sediment accumulation plotted against water depth. Bypass threshold depth and facies transition limit (x-3 m) are chosen in order to obtain the best fit with field observations. Note that two models are run with different parameter values. C: Best fit obtained with parameters defined in A, compared with field observations (see average thickness in Table 1). More detailed explanations in text. See computer code in online supported material.
cycle development through the stack (Fig. 3.8B). Facies distributions are influenced by the choice of bypass depth \(x\) and the palaeowater depth of the carbonate-siliciclastic transition (Fig. 3.8C).

### 3.6.2. Model results

The model has been run as a simple simulation exercise in order to identify the most likely range of values of the forcing mechanisms. These parameter ranges can then be directly compared with the estimates derived from analysis of sedimentary facies. Two superimposed eustatic signals are required to produce a thickening-upward parasequence stack assuming a constant subsidence rate. The longer scale eustatic variation allows progressive minor deepening, less bypass and thicker carbonate deposition through time (Fig. 3.8). The shorter scale eustatic variation drives the primary stratigraphic cyclicity.

Simulations have been run with different values of the bypass threshold depth (probably between 1 and 2 m water depth) and palaeowater depth of the facies transition from carbonate to siliciclastics. Best results are obtained when carbonate deposition is constrained in a zone 3 m below the bypass threshold depth \((x-3); \text{ Fig. } 3.8B\). Siltstone deposition is constrained within a zone between 3 m and \(\sim 20\) m below the bypass threshold depth. The bypass threshold could be as shallow as 0 m water depth. These results are in harmony with the palaeowater depths interpreted from facies and wave-ripple analysis.

The m-scale package of cycles overlying the parasequence stack (Figs. 3.2 and 3.8) represents a sharp change in depositional setting. Field observations show that the m-scale package of cycles slowly deepens upwards, eventually reaching 10 m of shales marking a maximum flooding zone. To simulate these relatively deeper water cycles, two scenarios were tested: (a) reduced sediment supply causes a decrease in cycle thickness and carbonate deposition to be replaced entirely by siltstone deposition, whereas an increase in sediment supply causes more bypass and preservation of shallow water carbonates at
the expense of deeper water siltstones; (b) reduction of the amplitude and period of the shorter eustatic component allows the preservation of thin cycles comprising both carbonate and siltstone (two eustatic sinusoids are combined with amplitude, duration and coupled sediment supply of 10 m, 0.32 Myr and 5 m/Myr, and 30 m, 16 Myr and 5 m/Myr; Fig. 3.8). The gradual upward transition into the maximum flooding zone may be due to a longer-term rise in absolute sea level. The change from a thickening-upward parasequence stack into condensed cycles followed by a maximum flooding zone therefore suggests a fundamental change in the eustatic forcing of cycle development.

3.7. Discussion

The simulation results described above are clearly non-unique. Nevertheless, the presence of a clearly decipherable eustatic signal and the observation that mixed carbonate-siliciclastic sequences are more typical of icehouse rather than greenhouse periods (Tucker, 2003), raises the possibility that the Upper Member of the Shuram Formation was deposited during a period of glaciation. Glaciation is known during Ediacaran time in the form of the ‘Gaskiers’ event. The Gaskiers Formation is dated at c. 580 Ma (Krogh et al., 1988; Bowring et al., 2003), but the ages of possible correlatives are loosely constrained (see Halverson et al., 2005; Le Guerroué et al., 2006a for a review). Subsidence analysis of the Nafun Group suggests deposition of the top Shuram parasequences at around 580 Ma, making them possibly synchronous with the ‘Gaskiers’ glacial period (see Le Guerroué et al., 2006b). As Le Guerroué et al. (2006b) have shown, the Gaskiers glacial event is unnoticed in the Huqf Supergroup of Oman in terms of direct sedimentological evidence, such as a significant sea-level lowering, or in terms of its impact on the carbon cycle, but it might be recognized in terms of an enhanced eustatic signal in the stratigraphic architecture of the Upper Member of the Shuram Formation. Speculatively, the end of glacioeustasy may coincide with the sharp transition between the stack of thick parasequences and the package of m-scale cycles that leads eventually to the maximum flooding zone marking the Shuram-Buah boundary (Fig. 3.8). If so, the Khufai-Shuram-Buah succession of the Nafun Group may represent an
alternation of greenhouse/arid and icehouse/pluvial periods. However, as a caveat, it must be emphasized that there are other mechanisms possible for the relative sea-level changes seen in the upper Shuram stratigraphy, especially bearing in mind the long duration of the Shuram cycles (~1.6 Myr).

Depleted values of Neoproterozoic carbonates have been tentatively explained by a biological pumping in a stratified ocean (Grotzinger and Knoll, 1995; Kaufman et al., 1997). Turnover of the deep-water, $^{13}$C-depleted reservoir during deglaciation would cause the precipitation of carbonates with light $\delta^{13}$C values in shallow waters of the continental shelf. Interestingly, Shen et al., 2005) reported a carbon isotopic variation of ~3‰ along a palaeoenvironmental transect from shelf to deep basinal sedimentary facies in the Marinoan Nantuo cap carbonate at a given stratigraphic time. However, with respect to the Shuram excursion, no lateral gradient is decipherable between the $\delta^{13}$C data sets from shallow and deep water facies (separated by 500 km; Le Guerroué et al., 2006a). At the parasequence scale, the limited depth range and lateral extent of facies preserved in the stack in Wadi Aswad do not allow the existence of a lateral gradient in carbon isotopic values to be tested.

The entire excursion occupying the stratigraphic interval from top Khufai Formation to Buah Formation is essentially in phase with longer term relative sea-level change (Le Guerroué et al., 2006a), with a nadir in $\delta^{13}$C occurring at the level of the maximum flooding zone of the lower Shuram, and the cross-over to positive values occurring within the Buah highstand. This feature is also found in other post-Marinoan and post-Sturtian transgressive 'cap sequences' (see compilation in Halverson et al., 2005). At the parasequence scale, however, the carbon isotopic ratios are unaffected by the c.1.6Myr frequency cycles of relative sea-level change, which were most likely driven by eustasy. Any explanation of the Shuram shift must involve an exceptionally long residence time or lag compared to Phanerozoic examples of perturbations of the carbon cycle and should involve a sufficiently large reservoir of $^{13}$C-depleted material (e.g. organic carbon). Although such an explanation is hard to demonstrate (but see non steady state mathematical model from Rothman et al.,
2003), any linkage between Ediacaran-aged (Gaskiers) glaciation and the Shuram carbon isotopic excursion can be ruled out, since they have different ages and durations.

The fact that the carbon isotopic record varies systematically and predictably through the stack of prograding parasequences is a powerful justification for regarding the carbon isotopic ratios as reflecting a primary variation in the chemistry of ocean water (Fig. 3.7). It is inconceivable that post-depositional processes could mimic this, since the carbonate facies comprising the upper parts of parasequences are lithologically identical (ooloidal grainstones).
3.8. Conclusions

Neoproterozoic extremely negative $\delta^{13}$C values have been investigated at the scale of a stack of thickening-upward parasequences in the Upper Member of the Shuram Formation of the Nafun Group of Oman. Although a number of hypotheses exist for the origin of extremely negative carbon isotopic values in marine carbonate, there is currently no satisfactory explanation for such long-lived and strongly negative $\delta^{13}$C values available. Consequently, there is considerable debate as to whether such values are primary (and so reflect the carbon isotopic composition of the ocean water from which carbonates were precipitated) or the result of diagenetic alteration.

The parasequence stack consists of shallow-water, storm-dominated, mixed-lithology sedimentary cycles bounded by flooding surfaces. Siltstones typified by large-scale swaley cross-stratification and ubiquitous soft-sediment deformation pass up into wave-rippled, calcareous ooidal grainstones with abundant intraclasts, reflecting various stages of break-up of carbonate beds and dispersal by vigorous currents. Five such parasequences show a systematic trend of progradation away from a palaeohigh situated in the northern Huqf region towards deeper palaeowater depths in the south.

The occurrence of extremely well-developed parasequences at this level of the stratigraphy suggests the superimposition of eustatic change on a long-term regional basin subsidence, although unforced internal cyclicity cannot be ruled out. The approximate age of the Upper Member of the Shuram Formation is essentially unconstrained by radiometric dates from within the Nafun Group, but a subsidence analysis bracketed by dates of the underlying Marinoan-aged Fiq Member and overlying Ara Group suggests that it is in the region of 580 Ma. This date is similar, perhaps coincidentally, to the reported age of the Gaskiers glaciation recorded in Laurentia. The mixed carbonate-siliciclastic parasequences may therefore reflect increased icehouse conditions at this time in the Ediacaran period.
The carbonate-dominated tops of the parasequences show a systematic trend in carbon isotopic values, becoming less negative from the oldest parasequence to the youngest, and $\delta^{13}$C values also become less negative in the direction of sedimentary progradation. These combined stratigraphic-carbon isotopic observations validate the genetic stratigraphic model, but in addition corroborate a primary, oceanographic origin for the carbon isotopic ratios.

The progradational architecture of the parasequence stack results in a 0.04 per mil per kilometre lateral variation along an individual parasequence. This lateral variation is simply caused by facies boundaries crossing inclined time lines within the parasequence stack. This coeval isotope variation and stratigraphic evolution demonstrate a primary origin of the extremely negative carbon isotopic values in the Shuram sea. Parasequence-scale variations in accommodation generation, freshwater influx, bioproductivity and upwelling had negligible effects on the carbon isotopic composition of seawater.

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Le Guerroué Erwan  

References


Appendixes

3.9.1 Parasequence code

% the code is been written to work under MathLab. You can cut and past it in the program.
% the code after been fill with parameters will draw Water Depth (WD) vs Potential Sedimentation Accumulation (PSA). Parameters used now are the one within the text of the paper.
% then you’ll be ask to define a position in the graph that represent the bypass sedimentation threshold. It can be done automatically (has it is set now).
% therefore a theoretical parasequence stack is drawn, with in green your deeper facies and blue the shallower one.
% the pink line represents a time gap or bypassed sedimentation.
% then you’ll be ask to define a position in a 2nd graph to plot a real
% sequence that you can change (it’s now correspond to the Shuram parasequences).
% finally the code returns the matrix of the theoretical thicknesses called ‘results’ (even bypassed ‘missing thicknesses is given that roughly gives a time indication more than a thickness indication!).

clear all; clf;

%-----------Parameters-----------
threshold=0;% define manually here your threshold
(threshold would be the depth at which your log would be draw) or leave 0 to get a graphical one. The model in the paper has been run at 60.
secondfacies=3; % represent the depth at which you facies is changing.
t = (0:100:32000000)';%duration of simulation in years form 0: increment: until what ever.

%-----------Tectonic-----------
A = 10 ;'meter/My';% subsidence rate in case of linear subsidence
A=A/1000000 ;%scaling of parameter
T=-A*t;
%-----------Cycles-----------
% you’ve got up to 3 orders of cycles you can add (put values or leave ‘0’):
% 1st cycle; H = ½ amplitude (m); L = cycle duration (year); S = max sedimentation associated (meter/My)
Ho = 4 ; 'm'; Lo = 16000000 ; 'year'; So=6 ; 'meter/My';
H1 = 14 ; 'm'; L1 = 8000000/5 ; 'year'; S1=6

%-----------Eustasy-----------
% 2) sinusoidal Eustasy function :
E = Ho*(sin ((2*pi*t)/Lo))+H1*(sin ((2*pi*t)/L1)) ;
%-----------Relative Sea Level-----------
RSL = E+T ;
%-----------Potential sedimentation accumulation-----------
% 2) sinusoidal Potential sedimentation accumulation function:
PSA = ((So/2)*((Lo/(2*pi)))*(((2*pi*t)/Lo)-cos((2*pi*t)/Lo)+1) + (S1/2) * (L1/(2*pi)))*(((2*pi*t)/L1)-cos((2*pi*t)/L1)+1);

%-----------Water Depth-----------
WD = RSL + PSA ;
%-----------Real sequence-----------
%plot real sequence?
plotreal=1; % '1' is for yes
% the sequence set up here is an average of the wadi Aswad measured section.
resultre(1,1)=2.25;% thickness of the unit 1
resultre(1,2)=2;% represent a deeper facies.
resultre(2,1)=.63;% thickness of the unit 2
resultre(2,2)=1;% represent a shallower facies.
resultre(3,1)=3.13;% thickness of the unit 3
resultre(3,2)=2;% represent a deeper facies.
resultre(4,1)=.21;% thickness of the unit 4
resultre(4,2)=1;% represent a shallower facies.
resultre(5,1)=7.10;% thickness of the unit 5
resultre(5,2)=2;% represent a deeper facies.
resultre(6,1)=.53;% thickness of the unit 6
resultre(6,2)=1;% represent a shallower facies.
resultre(7,1)=11.29;% thickness of the unit 7
resultre(7,2)=2;% represent a deeper facies.
resultre(8,1)=1.57;% thickness of the unit 8
resultre(8,2)=1;% represent a shallower facies.
resultre(9,1)=11.40;% thickness of the unit 9
resultre(9,2)=2;% represent a deeper facies.
resultre(10,1)=5.36;% thickness of the unit 10
resultre(10,2)=1;% represent a shallower facies.


%\%resultre(11,1)=5;\% thickness of the unit 11
%\%resultre(11,2)=2;\% represent a shallower facies.

%etc..... you can add some more.

\%---end of parameters and equations---

clear all; clf;
figure(1); plot( WD , PSA, 'r') ; hold on;
threshold0=threshold; FE(1,1)=threshold;
if threshold=0;
    figure(1); title 'Clic y<o to return or >o to define a Threshold';
    FE=ginput(1);FE=round(FE);
    if FE(1,2)<0;return;end;
    threshold0=FE(1,1);
end;

\% calcul ligne 0
figure(1); plot(threshold0*ones(length(PSA),1),PSA);
flag = ones(length(PSA),1) ;
flag(find(WD>threshold0))=0;

indices0 = find(abs(diff(flag))) ;
intersects0 = PSA(indices0) ;
distances0 = diff(intersects0) ;
flag_0 = ones(length(PSA),1) ;
flag_0(find(WD>threshold0))=-1 ;

\% calcul ligne 1
threshold1= threshold0-secondfacies;
figure(1); plot(threshold1*ones(length(PSA),1),PSA);

\%---end of parameters and equations---

calc line 0
figure(1); plot(WD,PSA, 'r'); hold on;
threshold0=threshold; FE(1,1)=threshold;
if threshold0=0:
    figure(1); title 'Clic y<o to return or >o to define a Threshold';
    FE=ginput(1);FE=round(FE);
    if FE(1,2)<0;return;end;
    threshold0=FE(1,1);
end;

% calcul ligne 0
figure(1); plot(threshold0*ones(length(PSA),1),PSA);
flag = ones(length(PSA),1) ;
flag(find(WD>threshold0))=0;

indices0 = find(abs(diff(flag))) ;
intersects0 = PSA(indices0) ;
distances0 = diff(intersects0) ;
flag_0 = ones(length(PSA),1) ;
flag_0(find(WD>threshold0))=-1 ;

% calcul ligne 1
threshold1= threshold0-secondfacies;
figure(1); plot(threshold1*ones(length(PSA),1),PSA);

3.9.2 M-scale packaging code

for i=1:length(resultint);
    if resultint((i,2)==0; resultint((i,1)=0; end;
end;

\% plot lithologies
y=0;
for e=1:length(resultint);
    if resultint((e,2)==1; C='b' ; D='b'; L=.3; end;
    if resultint((e,2)==2; C='g' ; D='b'; L=2; end;
    if resultint((e,1)==0; C='m' ; D='m'; L=1;
    resultint((e)=0.01; end;
    if resultint((e,2)==0; C='m' ; D='m'; L=1;
    resultint((e)=0.01; end;

    figure(2); rectangle('Position', [0,y,L resultint(e,1)], 'Curvature', [0,0], 'FaceColor', C, 'EdgeColor', D);
    y=y+resultint(e,1);
end;
Q=sum(resultint)+1;
text(0.01,sum(Q(1))+1,'Threshold = ',num2str(threshold0), 'FontSize', 10);
results=
% draw real sequences.
if plotreal==1;
   figure (2); title 'plot real sequences';
   FE=ginput(1);
y2=sum(resultre)-resultre(10,1);
y2=y2(1,1);y=FE(1,2)-y2;
for e=1:length(resultre);
    if resultre(e,2)==1; C='k' ;inc=resultre(e)/15; L=.3; end;
    if resultre(e,2)==2; C='y' ;inc=resultre(e)/25; L=.2; end;
    figure (2);rectangle('Position', [0.5,y,L resultre(e,1)], 'Curvature', [0,0], 'FaceColor', C, 'EdgeColor', 'k');
    hold off;
    y=resultre(e,1)+y ;
end;
figure (2); title 'Cycles';
text(0.5,FE(1,2)+1+resultre(10,1),'Real Sqs','FontSize', 8);
end;
hold off;

3.9.2 M-scale packaging code

clear all; clf;
%--------Parameters---------
threshold=0;\% define manually here your threshold (threshold would be the depth at which your log would be draw)or leave 0 to get a graphical one. The model in the paper has been run at 60.
secondfacies=3; \% represent the depth at which you facies is changing.
t = (0:100:32000000); \%duration of simulation in years form 0: increment: until what ever.

\%--------Tectonic---------
A = 10 ; \%'meter/My'; \% subsidence rate in case of linear subsidence
A=A/1000000 ; \%scaling of parameter
\[
T = -A*t;
\]

%------------Cycles------------
% you've got up to 3 orders of cycles you can add
% (put values or leave '0'):
% 1st cycle; H = \(\frac{H}{2}\) amplitude (m); L = cycle duration
% (year); S = max sedimentation associated (meter/My)
\[
H_o = 4; \quad m; \quad L_o = 16000000; \quad 'year'; \quad S_o = 6;
\]
% meter/My;
\[
H_1 = 14; \quad m; \quad L_1 = 8000000/5; \quad 'year'; \quad S_1 = 6;
\]
% meter/My;
\[
S_o = S_o/1000000; S_1 = S_1/1000000; S_2 = S_2/1000000;
\]
% scaling of parameters

%------------Eustasy------------
% 2) sinusoidal Eustasy function:
\[
E = H_o*(\sin ((2*\pi*t)/L_o))+H_1*(\sin ((2*\pi*t)/L_1));
\]

%-------------Relative Sea Level-------------RSL = E+T ;

% Potential sedimentation accumulation
% 2) sinusoidal Potential sedimentation accumulation function:
\[
PSA = ((So/2)*(Lo/(2*\pi)))*((2*\pi*t)/Lo) - (cos ((2*\pi*t)/Lo)+1) + (S_1/2)*(L_1/(2*\pi)) - (cos ((2*\pi*t)/L_1)+1); \]

%------------Water Depth------------
WD = RSL + PSA ;

%plot real sequence?
plotreal=0; % '1' is for yes

%--end of parameters and equations--

clf;
figure(1) ; plot( WD , PSA, 'r') ; hold on;
threshold0=threshold; FE(1,1)=threshold;
if threshold==0;
    figure(1) ; title 'Clic y<o to return or >o to define a
Threshold';
    FE=ginput(1);FE=round(FE);
    if FE(1,2)<0;return;end;
    threshold0=FE(1,1);
end;

% calcul ligne 0
figure(1) ; plot(threshold0*ones(length(PSA),1),PSA);
flag = ones(length(PSA),1);
flag(find(WD>threshold0))=0;
indices0 = find(abs(diff(flag)));
intersects0 = PSA(indices0);
distances0 = diff(intersects0);
flag_0 = ones(length(PSA),1);
flag_0(find(WD>threshold0))=-1 ;

% calcul ligne 1
threshold1= threshold0-secondfacies;
figure(1) ; plot(threshold1*ones(length(PSA),1),PSA);
text(FE(1,1)+5,50,\['Threshold = ',num2str(threshold0)\],'FontSize',10);
hold off;
flag1 = ones(length(PSA),1);
flag1(find(WD>threshold1))=0;
indices1 = find(abs(diff(flag1)));
intersects1 = PSA(indices1);
distances1 = diff(intersects1);
flag_1 = 2*ones(length(PSA),1);
flag_1(find(WD>threshold1))=-2;
[intersects ic] = sort([intersects1\ intersects0\]) ;
indices = [indices1\ indices0\] ;
titles = [WD , PSA] ;
if i=1:length(indices);
    index = indices(i) ;
    if flag_1(index-1)>0 % deeper facies
        if flag_0(index-1)==2 ;% by pass
            else flag_0(index-1)==1 % shallower facies
    end;
    end;
    thicknesses = diff(intersects) ;
    results= [thicknesses\ titles\] ;
    end;
    % resetting by pass thicknesses for plotting purposes
    resultint= results;
    for i=1:length(resultint);
        resultint(i,2)==0; resultint(i,1)=0; end;
    end;

% plot lithologies
y=0;
for e=1:length(resultint);
    if resultint(e,2)==1 ; C='b' ; D='b'; L=.3; end;
    if resultint(e,2)==2 ; C='g' ; D='b'; L=.2; end;
    if resultint(e,1)==0 ; C='m' ; D='m'; L=1;
    if resultint(e,1)==.001; end;
    if resultint(e,2)==0 ; C='m' ; D='m'; L=1;
    if resultint(e,1)==.001; end;
figure (2); rectangle(['Position',[0,y,L,resultint(e,1)]],\'Curvature',[0,0],\'FaceColor',C,'EdgeColor',D)
y=y+resultint(e,1);
end;
Q=sum(resultint)+1;
text(0.01,sum(Q(1))+1,\'Theoretical Sqs\',\'color\',b',\'FontSize',8);
results

%draw real sequences.
if plotreal==1;
    figure (2); title 'plot real sequences';
    FE=ginput(1);
y2=sum(resultre)-resultre(10,1);
y2=y2(1,1);y=FE(1,2)-y2;
for e=1:length(resultre);
    if resultre(e,2)==1 ; C='k' ; inc=resultre(e)/15; L=.3; end;
    if resultre(e,2)==2 ; C='y' ; inc=resultre(e)/25; L=.2;end;
    end;
}

figure (2); rectangle('Position',[0.5,y,L,resultre(e,1)],'Curvature',[0,0],'FaceColor','C','EdgeColor','k')
hold off;
y=resultre(e,1)+y;
end;
figure (2); title 'Cycles';
text(0.5,FE(1,2)+1+resultre(10,1),'Real Sqs','FontSize',8);
end;
hold off;
Chapter 4

50 Myr recovery from the largest negative δ$^{13}$C excursion in the Ediacaran ocean

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Terra Nova, 2006b

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Abstract

Sedimentary rocks deposited during the Ediacaran period (635-542 Ma) contain carbonates whose carbon isotopic ratios show a marked negative excursion consisting of a precipitous drop from +5‰ to -12‰, followed by a sub-linear recovery to positive δ¹³C values. Radiometric ages (U/Pb) and thermal subsidence modelling are combined to constrain the excursion in time and indicate an onset at ~600 Ma, and duration of recovery of approximately 50 Myrs. The excursion is widely recognised in Oman and has potential correlatives in Ediacaran strata elsewhere, and may thus represent a characteristic feature of the Ediacaran period. The duration and amplitude of this carbon isotope excursion far exceed those of other Neoproterozoic anomalies. The isotopic trend of negative excursion and long-term recovery spanned at least one short-lived glacial episode (at 580 Ma), but appears unrelated to glaciation, which indicates that negative anomalies in the Neoproterozoic marine carbon isotope record are not directly or uniquely linked to ice ages.
4.1. Introduction

A large portion of the Ediacaran period (Knoll et al., 2004), extending from the end of the Marinoan glaciation (c. 635 Ma; Hoffmann et al., 2004; Condon et al., 2005) to the Precambrian-Cambrian boundary (542.6±0.3 Ma; Amthor et al., 2003), is characterized by negative δ¹³C values in marine carbonate. These negative values comprise the Marinoan ‘cap sequence’, the Shuram-Wonoka and the Precambrian-Cambrian boundary excursions (Fig. 0.1; examples in Hoffman and Schrag, 2002; Amthor et al., 2003; Halverson et al., 2005). The Shuram negative δ¹³C excursion of Oman is characterised by an exceptional amplitude (+5‰ to -12‰) and long stratigraphic recovery (~800 m). Such an excursion and recovery is highly unusual in the geological record and is challenging to explain. Potential correlatives also occur in Ediacaran strata elsewhere (Halverson et al., 2005).

The fact that a carbon isotope trend of this magnitude can be recognized in marine Ediacaran rocks from several continents indicates that it was a very widespread oceanographic phenomenon, reflecting the composition of seawater from which carbonate minerals were precipitated, or the remineralization of a large organic oceanic carbon reservoir (Rothman et al., 2003). Since calibrated carbon isotope records provide a framework for the chronology of the Neoproterozoic (Halverson et al., 2005), it is critical that this major isotopic trend of the Ediacaran period is recognized and its timing and duration constrained. The Ediacaran period also contains a record of glaciation in the Gaskiers Formation of Newfoundland, dated 580 Ma (Bowring et al., 2003) and its possible correlatives (Halverson et al., 2005), and witnessed the early evolutionary changes in the biosphere that culminated in the proliferation of animals life forms in the Cambrian radiation. The relationship between climate change, global oceanography and biological evolution based on available geological proxies thus represents a challenging puzzle in this critical period of Earth history. The solution to the puzzle requires geological proxies to be placed within a sound geochronological framework. The exceptionally preserved and continuous Ediacaran record of the Sultanate of Oman allows a time frame to be established through a subsidence reconstruction of 15 stratigraphic
sections combined with correlation of the carbon isotope record. Although lacking the precision of a geochronologically constrained time frame, the chronology proposed is the best currently available for the Ediacaran period in Oman.

### 4.2. The Nafun Group of Oman

The Neoproterozoic Huqf Supergroup crops out in the Jabal Akhdar and Saih Hatat erosional windows of north Oman, in the Huqf region of east-central Oman, and in the Mirbat area of south Oman (Fig. 4.1). It is also penetrated by many boreholes in the salt basins of the Oman interior. The Huqf Supergroup is currently subdivided into the Abu Mahara, Nafun and Ara groups (Fig. 4.2; Loosveld et al., 1996). It overlies a c. 820 Ma igneous basement in the Huqf area (Allen and Leather 2006), and the Precambrian-Cambrian boundary is contained within the middle part of the Ara Group (Fig. 4.2; Amthor et al., 2003). Two glacial successions, are recognized in the Jabal Akhdar within the Abu Mahara Group, the older one being Sturtian in age (c. 711 Ma; Leather, 2001; Bowring et al., in prep) and the younger being attributed to the Marinoan glacial epoch (the Marinoan assignment is supported by the age of an ash bed interbedded with diamictites a few meters below the top of the glacial succession recovered from a subsurface core in the South Oman salt basin; ~635 Ma, Allen et al., 2005; ; Bowring et al., in prep’), which has been dated elsewhere at 635.5±1.2 Ma (Hoffmann et al., 2004) and terminating at 635.2±0.6 Ma (Condon et al., 2005).

The Nafun Group represents an essentially continuous 1 km-thick succession, commencing with the Hadash Formation cap

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1 See details in Appendixes section.

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carbonate of the younger glacial succession (Fig. 4.2; Leather et al., 2002; Allen et al., 2004). The overlying part of the Nafun Group comprises two major siliciclastic-to-carbonate cycles (Allen and Leather 2006; Le Guerroué et al., in 2006a²).

The lower cycle consists of marine shales, siltstones and sandstones of the Masirah Bay Formation, which pass gradationally upwards into the prograding homoclinal carbonate ramp of the Khufai Formation. The upper cycle is made of thick marine siltstones, shales and subordinate carbonates of the Shuram Formation, which pass up gradationally into the prograding distally steepened carbonate ramp of the Buah Formation (Cozzi et al., 2004; Cozzi and Al-Siyabi, 2004; Le Guerroué et al., in press³). The

2 This publication is Chapter 2.
3 This is chapter 3

Figure 4.2. Simplified stratigraphic logs of the Jabal Akhdar and Huqf outcrops and composite South Oman Salt Basin (SOSB) with essential facies description and interpretation. Maximum flooding surfaces (mfs) are indicated. Compiled after Allen, et al. (2004); Cozzi and Al-Siyabi (2004); Allen and Leather (2006).
Figure 4.3. Carbon isotope profiles from the Nafun Group, Huqf Supergroup, Oman. Data come from sections in the Jabal Akhdar and Huqf regions (see Chapter 2), and from 2 boreholes in the Oman salt basins. All profiles are transformed onto a time axis. Radiometric ages (U-Pb) from Brasier et al. (2000) and Amthor et al. (2003). Previous study carbon isotopic data from Burns and Matter (1993); McCarron (2000); Leather (2001); Cozzi et al. (2004); Cozzi and Al-Siyabi (2004). See Fig. 4.6.1 for sections and well locations.

overlying Ara Group has yielded U-Pb ages from ignimbrites of the Fara Formation in the Jabal Akhdar (Brasier et al., 2000) and from ashes penetrated in subsurface wells (Fig. 4.2; Amthor et al., 2003). The latter indicate the disappearance of Namacalathus and Cloudina at 542.6±0.3 Ma, coincident with the Precambrian-Cambrian boundary and its associated δ¹³C negative excursion (Fig. 4.3; Amthor et al., 2003).

4.3. The carbon isotope stratigraphy of the Nafun Group

Carbon isotopic data from the upper Khufai Formation and the overlying Shuram Formation have been collected from 16 outcrop sections in the Jabal Akhdar and 16 outcrop sections in the Huqf region, making a total of 700 measurements. These data are combined with previously published outcrop and borehole data (Fig. 4.3; Burns and Matter, 1993; McCarron, 2000; Leather, 2001; Cozzi et al., 2004; Cozzi and Al-Siyabi, 2004). This large dataset shows that carbon isotopic trends are reproducible throughout Oman, from outcrops to the subsurface, and irrespective of sedimentary facies (see also Le Guerroué et al., in 2006a). Furthermore, carbon isotopic ratios show a systematic variation within a stack of parasequences near the top of the Shuram Formation in the Huqf area, each cycle showing a trend in δ¹³C in the direction of sedimentary progradation⁴ (Le Guerroué et al. in press). This strongly suggests a long-term secular variation in δ¹³C rather than a post-depositional overprint. These combined stratigraphic-carbon isotopic observations support a primary, oceanographic origin for the carbon isotopic ratios. Bulk carbonate samples were previously geochemically screened (Burns and Matter, 1993; Burns et al., 1994; McCarron, 2000; Leather, 2001). These samples have Mn/Sr ratios <10, with many samples <3, and indicate that most of the δ¹³C and δ¹⁸O values have not been significantly altered by diagenesis (see

⁴ See details in chapter 3.
δ¹³C values quickly rise from the negative (-5 to -3‰) Hadash Formation cap carbonate values to positive values (~+3‰) throughout the Masirah Bay and Khufai formations (+4 to +5‰). Variation between the Khufai and Buah formations shows a major positive-negative-positive cycle. δ¹³C values as high as +5‰ characterize the top of the Khufai limestones, followed by a large negative decline in the basal Shuram Formation, to δ¹³C values as low as -12‰ a few tens of meters above the boundary. Negative values persist throughout the Shuram Formation and into the base of the overlying Buah Formation carbonates (-

Figure 4.4. Interpolation of formation boundary ages for Ediacaran outcrop sections and boreholes (see Fig. 4.6.1 for location). The black line shows a sediment-loaded thermal subsidence curve, with paleobathymetric corrections, that fits the total subsidence for the interval 635-542 Ma and the decompacted thickness (one example is drawn here; see text for explanation). The range of values of interpolated ages for each formation boundary is shown with black lozenges and plotted against time at their corresponding stratigraphic level. See Table 4.1 for exact values and associated calculated stretch factor (β). The mean ages of the distribution of interpolated ages are: top Masirah Bay 609±9 Ma, top Khufai 601±8 Ma and top Shuram 574±7 Ma.
6‰), and then gradually climb back to positive values (+2 ‰) at the top of the Buah Formation. There is a steeper climb in δ¹³C values in the upper part of the Buah Formation. Eventually the Precambrian-Cambrian boundary is marked by a smaller negative excursion (-3‰; Amthor et al., 2003). The overall pattern is therefore a precipitous decline in δ¹³C values, reaching a nadir in the lower Shuram Formation, and a long, slow, sub-linear rise throughout the overlying c. 800 m of stratigraphy (Fig. 4.3).

It is vital to appreciate that the positive-negative-positive isotopic pattern is unbroken by offsets, and that the falling curve through the Khufai-Shuram boundary (Fig. 4.3) is regionally consistent. The continuous pattern in δ¹³C corroborates the Table 4.1. Decompacted stratigraphic thicknesses for formations of the Nafun Group from surface outcrops and boreholes, with the best-fitting stretch factor β, and the interpolated ages of formation boundaries. The decompacted thicknesses for the Jabal Akhdar and Huqf outcrops have been calculated with additional overburdens as described in the text.
results of several region-wide sedimentological studies that failed to identify any major unconformities within the Nafun Group (Fig. 4.2; Cozzi et al., 2004; Cozzi and Al-Siyabi, 2004; Allen and Leather, 2006; Le Guerroué et al., in 2006a).

Figure 2.5. Isochron diagrams for detrital zircon samples from the Jabal Akhdar and the Huqf region.

Figure 4.6. The δ¹³C and strontium isotope composite profile from Oman, plotted against time, shows a major negative excursion and long recovery in the time period c. 600-550 Ma. The Oman section is compared with other successions plotted on a thickness scale: The post-Marinoan Windermere Supergroup of NW Canada (Narbonne et al., 1994; James et al., 2001), the Johnnie Formation of Death Valley (Bänninger, 2003; Corsetti and Kaufman, 2003, 2005), the Doushantuo Formation of the Yangtze Platform (Yang et al., 1999; Jiang et al., 2003; Condon et al., 2005), the Wonoka Formation of the Adelaide rift complex (Calver, 2000; Walter et al., 2000; Foden et al., 2001) and the Nama and Tsumeb groups of Namibia (Kaufman et al., 1993; Grotzinger et al., 1995; Saylor et al., 1998). Isotopic age references are ¹Preiss (1987); ²Schaefer and Burgess (2003); ³Hoffmann et al. 2(004); ⁴Grotzinger et al. (1995); ⁵Condon et al. (2005).
4.4. Transformation into time

It is assumed that the Nafun Group represents postrift subsidence driven by thermal contraction of the previously stretched continental lithosphere. Continental extension during deposition of the underlying Abu Mahara Group has been inferred based on 1) the geochemical signature of pillowed basalt (alkali basalts and trachy-andesites) of the Saqlah Member marking the beginning of the synrift record (Rabu, 1988); 2) regional seismic and gravity mapping of grabens in the Oman subsurface (Loosveld et al., 1996) and 3) sedimentologic and stratigraphic studies of the synrift strata (Fiq Member; Leather et al., 2002; Allen et al., 2004). Three operations are carried out to transform stratigraphic thicknesses of the Nafun Group into time. First, the stratigraphic sections are decompacted according to their burial history (see Methods section in the appendixes). Second, a stretch factor ($\beta$) is calculated to fit the total decompacted thickness in the given time frame (635-542 Ma; Table 4.1). Third, time lines are interpolated using model thermal subsidence curves corrected for paleobathymetric variations estimated from sedimentary facies (Fig. 4.4).

The ages of formation boundaries calculated using this method (Fig. 4.4 and Table 4.1) vary, but the variation is acceptable given the uncertainties in paleobathymetry, maximum depth of burial, eustatic and facies variations (compare shallow and deep water sections in Fig. 4.2 and 4.3). The average formation boundary age is based on data from 14 boreholes and 2 outcrop sections, which reduces the effects of local variations.

To date, no primary ashes have been recovered from the Nafun Group with which to independently check the ages of formation boundaries presented in Table 4.1. However, zircons, thought to be detrital, obtained from lithic siltstones 8 m below the top of the Khufai Formation in the Huqf region, and within 10 m above the base of the Shuram Formation in the Jabal Akhdar, have U-Pb SHRIMP II ages ranging from approximately 900 Ma to 600 Ma (Fig. 4.5). The youngest sub-population, found in the Huqf region, has a mean age of 609±7
Ma (2σ error; see appendixes section), and the youngest grain has an age of approximately 600 Ma. Therefore, the U-Pb zircon ages indicate that the Khufai-Shuram boundary cannot be much older than 600 Ma, which is very close to the age of the boundary derived from subsidence analysis.

Using the time transformation, the major negative carbon isotopic pattern is interpreted to begin at c. 600 Ma, with a rising limb crossing over to positive values at c. 550 Ma, a negative isotopic trend with a duration of 50 million years. This upper age assignment is consistent with the 551.1±0.7 Ma date obtained in the Doushantuo Formation of China that also records the crossing over to positive values (Condon et al., 2005) and a 548.8±1 Ma age date in the Nama basin of Namibia within already positive δ\textsuperscript{13}C carbonates (Grotzinger et al., 1995).

### 4.5. Correlation with Ediacaran sections worldwide

Similar Ediacaran δ\textsuperscript{13}C depleted sections may correlate with the Shuram negative excursion of Oman (Fig. 4.6). Very negative values are recorded in post-Marinoan Ediacaran stratigraphy in the Death Valley region of USA (Johnnie Formation; Corsetti and Kaufman, 2003, 2005), the Adelaide rift complex of southern Australia (Wonoka Formation; Calver, 2000; Walter, et al., 2000), the Yangtze Platform of South China (Doushantuo Formation; Yang, et al., 1999; Jiang, et al., 2003; Condon, et al., 2005) and the same anomaly may be recorded in the upper Tsumeb Subgroup and Nama Group of Namibia (Kaufman, et al., 1993; Grotzinger, et al., 1995; Saylor, et al., 1998). Each of these successions potentially contains significant time gaps, given the recognition of unconformities that affect the completeness of the chemostratigraphic record. The post-Marinoan stratigraphy of the Windermere Supergroup of NW Canada is thought to be free of major unconformities (James et al., 2001) and very negative δ\textsuperscript{13}C values have not been reported (Fig. 4.6). However, the succession is siliciclastic-rich and only sporadic data points are available. Other sections such as the Krol Group in India (Jiang et al., 2002) show repetitive low-amplitude negative excursions,
with no apparent shift comparable to that preserved in the Shuram Formation. The recently documented Zhuinska Group of Central Siberia (Pokrovsky and Melezhik, 2005), on the other hand, tends to support a similar long-lived and very negative $\delta^{13}C$ trend.

Worldwide correlations of Ediacaran strata suffer from a lack of absolute ages and incompleteness of the chemostratigraphic record (Fig. 4.6). Different age brackets for the Shuram and equivalent excursions are suggested, based on U-Pb age data defining the end of the excursion (Condon et al., 2005), subsidence analysis (Saylor et al., 1998) or more comprehensive correlations (Halverson et al., 2005). However, the time lines proposed in this paper combined with a continuous $\delta^{13}C$ record allow the Oman succession to serve as a chemostratigraphic reference for the Ediacaran period (Fig. 4.6).

4.6. Implications for Neoproterozoic ice ages

Improvements in the geochronological constraints on glaciation now suggest at least three glacial epochs for the Neoproterozoic (Hoffmann et al., 2004). The best-constrained glacial event of the Ediacaran is the Gaskiers of Newfoundland, which is dated at c. 580 Ma and is thought to have lasted around 1 Myr (Bowring et al., 2003; Fig. 4.6). Possible coeval glacial deposits exist (Halverson et al., 2005) but suffer from poorer time constraints. However, the Ediacaran (Gaskiers) glacial is thought to be a non-global, localized event (Halverson et al., 2005) and did not leave any recognisable imprint in the Nafun stratigraphic record (Fig. 4.2).

The Oman succession shows that the carbon isotopic record is unaffected by any glaciations taking place on the Ediacaran-age Earth, as opposed to the Sturtian and Marinoan glaciations, which are recorded by both glacigenic deposits and negative carbon isotopic excursions in associated cap carbonates. If the proposed chronology is correct, the local, short-lived Gaskiers glaciation is embedded within a large-amplitude, long-term negative carbon isotopic trend. The effects of such a glaciation, if any, on the carbon cycle certainly
cannot be recognized in the $\delta^{13}$C profile of the Nafun Group (Fig. 4.6; see also Zhang et al., 2005).

The origin of negative $\delta^{13}$C values in the Neoproterozoic remains strongly debated. Different models have been proposed, principally as an explanation of the carbon isotopic composition of 'cap carbonates'. Among these models, negative carbon isotopic values are variously attributed to the prior build-up of mantle-derived carbon in the atmosphere-ocean reservoir during a Snowball period of prolonged hydrological shutdown (Hoffman et al., 1998), to the overturn of a stratified ocean with a strongly functioning biological pump (Grotzinger and Knoll, 1995; Kaufman et al., 1997), to seepage from a light carbon reservoir of methane hydrates/clathrates (Kennedy et al. 2001; Schrag et al. 2002; Jiang et al., 2003) and to remineralization of a large oceanic organic carbon pool (Rothman et al., 2003).

Most of these models require significant deglaciation of a largely glaciated Earth and/or cannot explain the extremely negative and long-lived isotopic trend of the Shuram Formation. For example, carbon isotopic values in the methane clathrate and ocean overturn models would recover in a few mixing times of the ocean, and prolonged shut down in organic productivity cannot explain very negative values. Only Rothman et al.'s (2003) mathematical modelling of remineralization of dissolved organic carbon might theoretically explain such long-lived phenomena, but the large organic carbon pool implied by the model remains to be physically identified. However, a common point between Marinoan ‘cap carbonates’ and the Shuram excursion emerges. Major sea level transgression appears to accompany precipitous declines in $\delta^{13}$C.
4.7. Conclusions

Transformation of the Omani stratigraphic record into time provides a reference chemostratigraphic framework for the Ediacaran period (635-542 Ma). The Shuram carbon isotope record is marked by a perturbation from 600 to 550 Ma, reaching a -12‰ δ¹³C nadir followed by a long sub-linear recovery, and is perhaps the longest-lived and highest amplitude carbon isotope trend to be recognised in the geological record. This trend appears to be reproduced in isotopic records from a number of other Ediacaran reference sections.

The Ediacaran period is also marked by the non-global, short-lived Gaskiers glaciation around 580 Ma (Bowring et al., 2003), but appears to have had no effect on the chemostratigraphic records of Oman and other sections worldwide. The Shuram negative carbon isotopic excursion in the Ediacaran period demonstrates that the marine carbon system was capable of extreme variations unrelated to glaciation per se. This suggests that caution should be used in the interpretation of climatic change from negative carbon isotopic excursions in the Neoproterozoic chemostratigraphic record.

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References


Appendixes

4.8.1 Decompaction and subsidence modelling

Present-day thicknesses of formations were converted to decompacted thicknesses using a program (Allen and Allen, 2005) derived from algorithms in Sclater and Christie (1980). Appropriate initial porosities and porosity-depth coefficients were taken for shales and carbonates. Paleobathymetric corrections were estimated from sedimentary facies. Model subsidence curves were calculated assuming subsidence by thermal contraction of the lithosphere following rift-related heating (McKenzie, 1978; Sclater and Christie, 1980). Outcrop sections are not at their maximum burial depth. An additional overburden of between 1 and 4 km (Huqf area) and between 2 and 5 km (Jabal Akhdar) was therefore added in order to derive realistic decompacted thicknesses. Burial at additional depths of > 5km has little effect on decompacted thicknesses because of the exponential decrease in porosity with depth. Initially, a thermal subsidence curve was fitted to the total subsidence over the interval 635-542 Ma, corrected for estimated water depths at the chronological tie lines of 635 Ma and 542 Ma. Ages of formation boundaries were then interpolated from the thermal subsidence curve with paleobathymetric corrections over the entire interval 635-542 Ma. Eustatic corrections were not applied, since we have no independent estimates of the eustatic curve in the Ediacaran period. This exercise was carried out for 14 boreholes (data obtained from Petroleum Development Oman) and for the outcrop sections of the Huqf region and Jabal Akhdar, yielding a spread of formation ages: 609±9 Ma for the top Masirah Bay Formation, 601±8 Ma for the top Khufai Formation, and 574±7 Ma for the top of the Shuram Formation. Variation in the interpolated ages is most likely due to neglect of the eustatic variation, inaccuracies in the estimation of paleobathymetry, errors in decompaction, and unrecognized time gaps caused by the presence of small unconformities at sequence boundaries. The better-constrained outcrop sections plot close to the mean of the interpolated ages of formation boundaries.
4.8.2. U-Pb SHRIMP geochronology

Zircon grains were hand selected from the heavy mineral concentrates of samples from the Huqf (HA) (c. 60 grains) and Jabal Akhdar (JA) (c. 35 grains) and cast into an epoxy mount together with chips of the FC1 and SL13 reference zircons, sectioned approximately in half and polished. Reflected and transmitted light photomicrographs, and cathodoluminescence (CL) SEM images were prepared for all zircons. The CL images were used to decipher the internal structures of the sectioned grains and to target specific areas within the zircons.

The U-Th-Pb analyses were made using SHRIMP II, each analysis consisting of either 6 or 7 scans through the mass range. The data have been reduced in a manner similar to that described by Williams (1998 and references therein), using the SQUID Excel Macro of Ludwig (2000). The Pb/U ratios for the zircons have been normalised relative to a value of 0.1859 for the $^{206}\text{Pb}/^{238}\text{U}$ ratio of the FC1 reference zircons, equivalent to an age of 1099 Ma (see Paces and Miller, 1993). Uncertainties given for individual analyses (ratios and ages) are at the one sigma level, but the uncertainties in calculated ages are reported as 95% confidence limits.

Many of the zircon grains from both samples have round terminations, and/or pitted surfaces. The grains selected for analysis are the more elongate with simple zoned magmatic internal zoning as determined from CL images. There is clear evidence for surface transport and the majority of the zircons are interpreted to be detrital. Notwithstanding this, it is common to find age groupings within such zircon populations that closely approximate the time of sedimentation; ie the zircon is interpreted to be of a volcanic, or volcanioclastic nature. Initially, 22 grains from sample HA and 21 grains from sample JA were analyzed. A number of the areas analyzed yield Proterozoic and Archaean ages, though most have ages between 600-900 Ma. Several of the grains yield younger ages, interpreted discordant, and thus and are not considered further in this discussion.
Seven grains from sample HA recorded $^{206}\text{Pb} / ^{238}\text{U}$ ages between 600-650 Ma. A second probe session was carried out re-analyzing these grains and others that have similar zoned magmatic internal CL structure. The younger section of the $^{206}\text{Pb} / ^{238}\text{U}$ age spectrum appears to be a mixture of two distinct age populations that can be deconvolved (via ISOPLOT; Ludwig, 1999) using the mixture modeling algorithm of Sambridge and Compston, (1994). This gives $^{206}\text{Pb} / ^{238}\text{U}$ ages of 609±7 Ma and 638±8 Ma. The time of sedimentation is no older than the youngest coherent age population, ie 609±7 Ma. However, it must be noted that this is a restricted dataset and that for sample HA only 43 analyses have been made on 35 zircon grains. Thus, we infer the time of deposition to be no older than 600 Ma, the $^{206}\text{Pb} / ^{238}\text{U}$ age for grains 16, 14 and 3 within this population (see Table 4.2 and 4.3).

4.8.2. Date on the Fiq Formation

The top Fiq Formation is now constrained younger than 645 Ma (Bowring et al., in press) from thought to be ash core material within the Lahan well (south Oman). It is emphasized that the core was taken a thousand km away from the Jabal Akhdar. However, the ash bed is situated within the last meters of glacial diamicites that are capped by -4 to -5‰ $\delta^{13}\text{C}$ carbonates, chemostratigraphically Hadash ‘looking like’ (negative values of -2 to -4.5‰). On top of this carbonates rest siliciclastic then carbonate that would correlate the Masirah Bay and Khufai Formation. This is also supported by $\delta^{13}\text{C}$ values that would match the top Khufai trend. See Allen et al. (2005) and Bowring et al., in press.

Analyses using chemical abrasion method indicate Zircon ages groupings at 715-718 Ma, one at 675 Ma and another at ca. 645 Ma (Bowring et al., in press). Although these thin horizons were sampled as potential volcanic ash beds, the scatter in the dataset and small number of ‘young’ grains (n = 5) make assignment of a depositional age to this interval is equivocal: if they are primary volcanic zircons the ca. 645 population is a maximum depositional age. Alternatively the rock could be a silt/mudstone and the ca. 645 Ma zircons are best interpreted as the youngest detrital component. Bowring et al. consider the ca. 645 Ma age as a maximum age for the Lahan glacial unit at this stratigraphic level.
References


Table 4.2. SHRIMP U-Pb zircon results from the Huqf samples.
### Table 4.3. SHRIMP U-Pb zircon results from the Jabal Akhdar samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>acid used to dissolve sample</th>
<th>Moles Rb</th>
<th>Moles Sr</th>
<th>87/86 Sr (measured)</th>
<th>87/86 Sr (normalized)</th>
<th>(age corrected) 2 σ ext.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Buah 2</td>
<td>5% acetic acid</td>
<td>1.90E-11</td>
<td>6.81E-09</td>
<td>0.708983 ± 5</td>
<td>0.709064</td>
<td>0.708990 ± 17</td>
</tr>
<tr>
<td>Buah 1</td>
<td>5% acetic acid</td>
<td>1.37E-11</td>
<td>5.53E-09</td>
<td>0.708828 ± 9</td>
<td>0.708909</td>
<td>0.708844 ± 17</td>
</tr>
<tr>
<td>Shuram 3</td>
<td>1M HCl</td>
<td>3.34E-11</td>
<td>4.68E-08</td>
<td>0.707938 ± 6</td>
<td>0.708019</td>
<td>0.708000 ± 17</td>
</tr>
<tr>
<td>Shuram 2</td>
<td>ca. 20% acetic acid</td>
<td>1.34E-11</td>
<td>1.81E-07</td>
<td>0.707812 ± 8</td>
<td>0.707892</td>
<td>0.707891 ± 17</td>
</tr>
<tr>
<td>Shuram 1</td>
<td>5% acetic acid</td>
<td>5.50E-12</td>
<td>2.19E-08</td>
<td>0.707839 ± 5</td>
<td>0.707919</td>
<td>0.707913 ± 17</td>
</tr>
</tbody>
</table>

### Notes:

1. Uncertainties given at the one σ level.
2. Error in FC1 Reference zircon calibration was 0.43% for the analytical session. (not included in above errors but required when comparing 206Pb/204U data from different mounts).
3. \( f_{206} \) % denotes the percentage of 206Pb that is common Pb.
4. Correction for common Pb made using the measured 206Pb/204Pb ratio.
5. For % Disc, 0% denotes a concordant analysis.

### Table 4.4. Rb/Sr-Ratios and Sr Isotope Ratios from the Jabal Akhdar samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>acid used to dissolve sample</th>
<th>Moles Rb</th>
<th>Moles Sr</th>
<th>87/86 Sr (measured)</th>
<th>87/86 Sr (normalized)</th>
<th>(age corrected) 2 σ ext.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Buah 2</td>
<td>5% acetic acid</td>
<td>1.90E-11</td>
<td>6.81E-09</td>
<td>0.708983 ± 5</td>
<td>0.709064</td>
<td>0.708990 ± 17</td>
</tr>
<tr>
<td>Buah 1</td>
<td>5% acetic acid</td>
<td>1.37E-11</td>
<td>5.53E-09</td>
<td>0.708828 ± 9</td>
<td>0.708909</td>
<td>0.708844 ± 17</td>
</tr>
<tr>
<td>Shuram 3</td>
<td>1M HCl</td>
<td>3.34E-11</td>
<td>4.68E-08</td>
<td>0.707938 ± 6</td>
<td>0.708019</td>
<td>0.708000 ± 17</td>
</tr>
<tr>
<td>Shuram 2</td>
<td>ca. 20% acetic acid</td>
<td>1.34E-11</td>
<td>1.81E-07</td>
<td>0.707812 ± 8</td>
<td>0.707892</td>
<td>0.707891 ± 17</td>
</tr>
<tr>
<td>Shuram 1</td>
<td>5% acetic acid</td>
<td>5.50E-12</td>
<td>2.19E-08</td>
<td>0.707839 ± 5</td>
<td>0.707919</td>
<td>0.707913 ± 17</td>
</tr>
</tbody>
</table>
Chapter 5

Conclusions

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5.1. Conclusions

The Neoproterozoic Huqf Supergroup of Oman comprises the Abu Mahara, Nafun and Ara groups. The Abu Mahara Group (723±16/-10 Ma) contains a glacial sequence and the volcanoclastic Saqlah Member. The Saqlah-Fiq (Ghadir Manqil Formation) package is interpreted as a rift basin-fill. The initiation of rifting was marked by intraplate basaltic volcanism. The Fiq Member is composed of a mixed assemblage of shallow to relatively deep marine siliciclastics and glacigenic diamicites (Leather et al., 2002; Allen et al., 2004), and most likely correlates with the cluster of glacial events known as Marinoan (c. 635 Ma). Detrital zircon population within the Lahan well of south Oman supports this assignment (~645; Allen et al., 2005; Bowring et al., in preparation). In the case of the Fiq Member, the association of volcanicity, basin development and glaciation invites a geodynamic-climatic linkage of these processes. Ice centres over rift-flank uplifts developed at the same time as basin subsidence accommodated glacigenic diamicites, gravity flows and non-glacial marine sediments.

The Nafun Group of the Huqf Supergroup of Oman records an essentially continuous period of deposition from the post-Marinoan cap carbonate of the Hadash Formation to close to the Precambrian-Cambrian boundary. It comprises two grand cycles of marine siliciclastics to ramp carbonates (Allen and Leather, 2006; Le Guerroué et al., 2006a). Each cycle is initiated by a major transgression, the formation of a relatively deep siliciclastic marine basin, followed by gradual shallowing up into progradational ramp carbonates. Although the basin setting of the Nafun Group is equivocal, the upward evolution from rift-related Fiq Member through the cap carbonate of the Hadash Formation and transition into the siliciclastic-to-carbonate grand cycles of the Nafun Group, and the stratigraphic onlap of Nafun Group over basement-cored rift margins, is strongly suggestive of post-rift subsidence caused by thermal relaxation.
The boundary between the two grand cycles is at the Khufai-Shuram boundary. Both the Khufai and Shuram formations are represented by shallow water facies in the Huqf area of east-central Oman, and by deeper water facies in the Jabal Akhdar of north Oman. In the Huqf area, the Khufai Formation is composed of fetid carbonates, which pass up into m-scale peritidal cycles, indicating progradation of a carbonate ramp. In the Jabal Akhdar however, the Khufai Formation is dominated by black, pyritic limestones deposited in deeper ramp conditions. The Shuram Formation is composed of storm-influenced sandstones, siltstones and limestones in the relatively shallow water Huqf area, eventually developing shallowing upward storm-dominated parasequences. In the Jabal Akhdar, however, the Shuram Formation is dominated by deep water, organic-rich dolomitic mudstones and bleached siltstones at the base, which pass up into very thick, purple, monotonous siltstones and shales with thin subordinate carbonates. Shallow water conditions were established over basement highs, which may have been partially inherited from the phase of important rifting during the deposition of the underlying Ghadir Manqil Formation.

The carbonates of the upper grand cycle, comprising the Shuram and Buah Formations, contain a remarkable negative bulk carbonate excursion in $\delta^{13}C$ values, which start to fall precipitously from the uppermost Khufai Formation, reach a nadir close to the maximum flooding zone of the Shuram, and then recover monotonically through nearly 1 km of stratigraphy, with a crossing point within the Buah Formation. The amplitude of the excursion is from +5 ‰ in the Khufai carbonates, to -12 ‰ in the lower Shuram Formation.

At the scale of the stack of parasequences in the upper Shuram Formation, the carbonate-dominated tops of the parasequences show a systematic rise in carbon isotopic ratio from the oldest parasequence to the youngest, and $\delta^{13}C$ values also become less negative in the direction of sedimentary progradation. This lateral variation can most easily be explained by facies boundaries crossing inclined time lines within the parasequence stack. This pattern of carbon isotope variation associated with high-frequency stratigraphic evolution, as well as the reproducibility of the carbon isotopic pattern throughout both...
the outcrops and subsurface records of Oman, supports a secular trend in isotopic ratios reflecting the isotopic composition of seawater. A strong case can therefore be made for the extremely depleted carbon isotopic values in carbonates of the Shuram Formation as being largely issued from a primary origin.

The stratigraphic record of the Nafun Group is transformed into time, based on thermal subsidence analyses supported by the U-Pb age distribution of detrital zircons. Using this chronology, the Shuram negative carbon isotope excursion and its recovery represents a perturbation of 50 million years duration, from 600 to 550 Ma, quickly reaching a -12‰ δ\(^{13}\)C nadir followed by a long sub-linear recovery. This is perhaps the longest-lived δ\(^{13}\)C negative excursion in inorganic carbon of marine carbonates to be recognised in the geological record.

The Shuram shift provides a reference chemostratigraphic framework for the Ediacaran period (635-542 Ma). The Ediacaran period is also marked by the Gaskiers non-global, short-lived glaciation around 580 Ma (Bowring et al., 2003), but this climatic event appears to have had no effect on the chemostratigraphic records of Oman and other sections worldwide. The Shuram negative carbon isotopic excursion in the Ediacaran period demonstrates that the marine carbon system was capable of extreme long-term variations unrelated to glaciation per se.

The explanation for the Shuram shift is enigmatic. However, there are a number of clues derived from the sedimentology and stratigraphy of the Nafun Group that helps constrain possible causes. Importantly, the carbon isotopic excursion is roughly in phase with relative sea level. The start of the fall in carbon isotopic values commences with the influx of siliciclastics and the demise of the Khufai carbonate factory, and the nadir occurs at the level of the maximum flooding zone of the lower Shuram, so the falling segment of the excursion coincides with a period of transgression. In addition, the transition from the Khufai to the Shuram Formation most likely records a climate change from arid to humid and stormy.
However, at a parasequence scale high frequency variations in accommodation generation, freshwater influx, bioproductivity and upwelling had negligible effects on the carbon isotopic composition of seawater. Although the sedimentological context of the excursion is now well constrained, a satisfactory explanation of the operation of the carbon cycle during this critical period of organic evolution must await further geochemical studies.

The Nafun Group passes upward into the carbonate-evaporite cycles of the Ara Group, defined in PDO subsurface penetrations in the Oman salt basins. The Ara Group spans the Cambrian/Precambrian boundary, dated at 542 ± 0.6 Ma by the disappearance of Cloudina and Namacalathus (Amthor et al., 2003). Outcrop equivalents of the Ara Group are found in the Huqf area and in the Jabal Akhdar, where the Buah Formation is overlain by the volcaniclastics and cherty limestones of the Fara Formation dated at 544.5 ± 3.3 Ma (Brasier et al., 2000).

5.2. Implications for the Snowball Earth hypothesis and climate change in the Ediacaran

Since the ‘popularisation’ of the snowball Earth theory (Hoffman et al., 1998) a lot of attention has been addressed to the Neoproterozoic Era and to critical issues about the functioning and limits of the Earth system. This PhD thesis falls within the broader context of attempting to understand the functioning of the Earth system at a time of extreme climate change, when the carbon cycle appears to have been more strongly perturbed than at any subsequent time in the Phanerozoic.

Neoproterozoic time of Earth (1000 to 542Ma, Fig. 0.1) was characterized by extreme variations in climate indicated by the widespread occurrence of glacial deposits, coupled with extreme swings in the carbon isotopic ratio preserved in mineral carbonate. The ‘snowball Earth’ theory, which advocates a complete and protracted freeze of the Earth on one or more occasions in the Neoproterozoic, is highly debated. Although the paleomagnetic evidence for paleolatitude is open to question, some basins appear to have been located in tropical
to equatorial paleolatitudes at the time of glaciation (see McWilliams and McElhinny, 1980; Schmidt et al., 1991; Meert and Van Der Voo, 1994; Schmidt and Williams, 1995; Sohl et al., 1999; Evans, 2000; Meert and Torsvik, 2003).

There is continuing debate on the exact number, timing and duration of Neoproterozoic glaciations. Inter-regional correlation relies upon an increasing number of radiometric ages (see example in Hoffmann et al., 2004; Condon et al., 2005; Halverson et al., 2005; Etienne et al., in press) as well as the use of carbon isotope proxies (see reviews in Kaufman et al., 1997; Walter et al., 2000; Halverson et al., 2005; Le Guerroué et al., 2006a). On this basis, there appear to have been at least three glaciations in the Neoproterozoic (Fig. 0.1 Sturtian, Marinoan and Varangerian). However resolution of the radiometric constraints probably can resolve global stratal correlation issues if compared with Phanerozoic examples (see review in Etienne et al., in press).

Neoproterozoic times also witnessed the evolutionary changes in the biosphere that led to the radiation of metazoans in the Cambrian (542 My; Amthor et al., 2003, Fig. 0.1). Unfortunately preservation of soft bodied fauna is poor and the link between severe glaciation and evolution of life is controversial (Runnegar, 2000; Schrag et al., 2001; Narbonne and Gehling, 2003). Apart from the small skeletal fossils found in the subsurface Ara Group (Amthor et al., 2003), and unpublished work on acritarchs, no paleontological record has been found in Oman. It is still largely unknown how organic production, decomposition and remineralization affected the carbon cycle in the Neoproterozoic oceans. It remains a possibility that severe climate swings caused evolutionary bottlenecks and stimulated evolutionary developments in their aftermath. The new Ediacaran period (635 to 542 Ma ago; Knoll et al., 2004; Kaufman, 2005; Knoll et al., in press) starts after the last of the great Neoproterozoic glaciations (Marinoan) and encloses the first appearance of Ediacaran fauna (Fig. 0.1).
One of the most striking features about Neoproterozoic time is a strongly disturbed bulk inorganic carbonate carbon isotopic record. Values reach +5 to +10‰ and commonly drop down to -5 and even -12‰ (Fig. 0.1). Among the extreme negative excursions are the Marinoan and Sturtian ‘cap carbonate’ excursions, the Shuram-Wonoka Ediacaran excursion and the Precambrian-Cambrian boundary excursion (see compilation in Saylor et al., 1998; Condon et al., 2005; Halverson et al., 2005; Le Guerroué et al., 2006b, Fig. 0.1). All these excursions are believed to have lasted for millions of years and possibly up to 50 My in the case of the Shuram of Oman (Le Guerroué et al., 2006b, Fig. 0.1) The presence of carbon isotopic excursions of this amplitude and duration are difficult to explain, since Phanerozoic negative excursions are conspicuously short-lived and represent shifts of generally less than 2 ‰ (e.g., Jacobsen and Kaufman, 1999; Hesselbo et al., 2000; Galli et al., 2005, Fig. 0.1).

This work demonstrates through stratigraphic evidence the primary origin, at a basin wide scale, of the Shuram shift. Transformation of this δ¹³C record into time (see chapter 4 and Le Guerroué et al., 2006b) allows a global Ediacaran reference curve to be proposed against which other world wide sections, such as the Windermere Supergroup of NW Canada (James et al., 2001), the Johnnie Formation of Death Valley (Corsetti and Kaufman, 2003), the Doushantuo Formation of the Yangtze Platform of China (Condon et al., 2005), the Wonoka Formation of the Adelaide rift complex (Calver, 2000) and the Kuibis Group of Namibia (Grotzinger et al., 1995; Halverson et al., 2005) can be plotted (Fig. 5.1). Unfortunately such sections suffer from an incomplete sedimentary record through

Figure 5.1. The composite δ¹³C profile from Oman (left) shows a major negative excursion in the time period c. 600-550 Ma. The post-Marinoan Windermere Supergroup of NW Canada James et al., 2001 is also transformed onto the same time axis. Further sections from Death Valley Corsetti and Kaufman, 2003, Yangtze Platform Condon et al., 2005, Adelaide rift complex Calver, 2000 and Namibia Grotzinger et al., 1995; Halverson et al., in press are plotted with interpreted unconformities, which dissect the negative isotopic excursion seen in the complete section of the Nafun Group. References (1 to 15) are: ¹Schaefer and Burgess, 2003; ²Condon et al., 2005; ³Chen et al., 2004; ⁴Barfod et al., 2002; ⁵Preiss, 1987; ⁶Grotzinger et al., 1995; ⁷Hoffmann et al., 2004; ⁸Xiao et al., 2004; ⁹Edwards, 1984; Halverson et al., 2004; ¹⁰Condon and Prave, 2000; ¹¹Calver et al., 2004; ¹²Condon and Prave, 2000; ¹³O’Brien et al., 1992; ¹⁴Myrow and Kaufman, 1999 Bowring et al., 2003; ¹⁵Thompson and Bowring, 2000.
Strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) data plotted on a time axis, with the carbon isotopic record from Oman for reference. Excluding Windermere Supergroup data in the 600-635 Ma interval, strontium isotope ratio data show a well-defined, gradual increase in value through Ediacaran time. In contrast, cap carbonates show very strong variation, including high values indicative of strong pulses in continental weathering at these times. Radiometric age and Sr data references (1 to 9) are: 1Condon et al., 2005; 2Hoffmann et al., 2004; 3Burns et al., 1994; 4Foden et al., 2001; 5Calver, 2000; 6Yang et al., 1999; 7James et al., 2001; 8Kaufman et al., 1993.
the Ediacaran period. However, by dissecting these stratigraphic columns along major unconformities, we obtain a consistent global signal. This signal consists of a sharp drop of $\delta^{13}C$ values down to -12‰ at around 600Ma and slow recovery towards positive values over about 50 Myrs (Fig. 5.1). Using this time frame the Strontium isotopic record is also reconstructed. The integrated values shows a consistent pattern between sections with a slow increase from values around 0.708 after the Marinoan cap to values around 0.709 at the Precambrian Cambrian boundary (Fig. 5.2; as previously established by Kaufman et al., 1993).

The subsidence analysis used, coupled with detrital zircon geochronology, is not a precise and fully reliable means to erect an Ediacaran chronology. However, the chronology developed from the essentially continuous succession in Oman represents a useful yardstick for timing events in Ediacaran history.

Neoproterozoic $\delta^{13}C$ values are also highly debated in terms of whether they reflect a primary oceanographic signal or whether they have been reset during diagenesis (see discussion in chapter 2 and 3; Le Guerroué et al., 2006a; Le Guerroué et al., in press). The Snowball Earth hypothesis explains the negative shift in carbon isotopic values recorded in cap carbonates as due to the mixing of a light carbon reservoir built up in the atmosphere during volcanic outgazing during a prolonged hydrological shutdown (Schrag et al., 2002). The ‘cap carbonates’ deposited in the immediate aftermath of glaciation have a negative $\delta^{13}C$ signature (to -5‰) in their basal transgressive phase (Kennedy, 1996; Hoffman and Schrag, 2002; Halverson et al., 2005; Halverson, in press), but pass up into highstand deposits with positive $\delta^{13}C$ values indicative of a resumption of ‘normal’ oceanic biological productivity. The key contrast with the Shuram shift is that the negative excursion is deeper, and the recovery much longer in the latter, suggesting than an alternative mechanism can’t be excluded.
Figure 5.3: High (A) and low (B) latitude model for the 580Ma paleogeographic reconstruction and key Ediacaran successions. See the strong correspondance between passive margin settings and the absence of glacial deposits in the Ediacaran record. Reconstruction adapted from Meert (2004). Ediacaran deposits key: 1Shuram of Oman; 2Kroll of Lesser Himalaya; 3Doushantuo of south China; Location after Macouin et al., 2004; 4Hankalchough of China; Location after Xiao et al., 2004; 5Wonoka of Australia; 6Croles Hill of Tasmania; Location after Calver et al., 2004; 7Johnnie of California; 8Inga of Canada; 9Gaskiers and Squantum of Avalon Terrane; 10Loch Na Cille of Ireland; 11Mortensnes of Norway. Paleogeographic keys: Ama: Amazonia; Ara: Arabia; Arm: Armorica; Aus: Australia; Ava: Avalonia; Bal: Baltica; Con: Congo; EAn: East Antarctica; Ind: India; Kal: Kalahari; Lau: Laurentia; SC: South China; Sib: Siberia; Waf: West Africa. Active margins are represented by thick grey lines. Eyles and Januszczak, 2004.
Glaciation during Ediacaran time is recorded in numerous places. However, poor radiometric constraints exist with the exception of the Gaskiers Formation of Newfoundland, dated around 580 and constrained to have lasted less than 1 My (Fig. 5.1). Clearly, therefore, neither the Marinoan nor the Gaskiers glaciation were responsible for the perturbation of the oceanic carbon cycle recognized in the carbon isotopic ratios of carbonate in the Shuram Formation. A mechanism not directly related to glaciation appears to be required. Based on sedimentological and stratigraphic evidence, the isotopic trend of the Shuram is mirrored by the pattern of relative sea level change, suggesting that a mechanism driven, affected or associated with the transgression of continental margins should be searched for. Glacial Sedimentary record could be linked to paleogeographic settings and/or localised on active margins (Fig. 5.3; see also Eyles and Januszczak, 2004).

In conclusion, although the work contained in this thesis is not specifically targeted at an evaluation of the snowball Earth hypothesis, it highlights a number of relevant issues: (1) the carbon isotopic record of marine carbonate deposited in Neoproterozoic oceans cannot be relied upon as a direct proxy for glaciation, and (2) the largest perturbation of the carbon cycle in the Neoproterozoic took place during the Ediacaran period, commencing at about 600 Ma and lasted for c. 50 Myr. Whatever caused this carbon isotopic perturbation had a significantly larger effect on the carbon cycle than any mechanisms associated with putative snowball Earth events such as the Marinoan.

5.3. Future Outlook

Although the broad sedimentological and stratigraphical history of the Nafun Group of Oman in general, and the Shuram Formation in particular, is now well-established, this is not the case for most of the Shuram global correlatives, which lack detailed chemostratigraphic data and sedimentological context. Improvement in our knowledge on other Ediacaran successions would validate the Shuram $\delta^{13}$C pattern, the global stratigraphic signal and substantiate a reference curve for the Ediacaran period.
In addition, models to explain the highly depleted $\delta^{13}$C signature of the Ediacaran are unsatisfactory at present and many questions remain:

- Can we identify a physical pool of depleted carbon in the Ediacaran oceans and quantify its likely size? Can we formulate and quantify a mechanism capable of mineralizing this depleted carbon pool?
- What are the consequences of such a mechanism in the Neoproterozoic for the evolution of the biosphere, and to what extent was such a mechanism controlled by global tectonics and climatic changes? What triggered what?

Detailed carbonate petrographic and geochemical studies coupled with micro $\delta^{13}$C analysis on Neoproterozoic rocks are rare (cf. Jiang et al.’s 2003 work on the Marinoan Doushantuo cap carbonate of China). Such studies are required to improve resolution on the processes associated with the formation of carbonate grains, cements and replacements during Neoproterozoic time as well as to understand possible alteration of the primary $\delta^{13}$C signal.

Testing the applicability of different processes through geochemical modelling, using all the chemostratigraphic data realised in this and other works, and making use of the fledgling chronology for the Ediacaran proposed here, offers the possibility to better understand the functioning of the Earth’s ocean-atmosphere system at this critical time period in Earth history.
References


Conclusions ...


Thompson, M.D. and Bowring, S.A., 2000. Age of the Squantum 'tillite', Boston basin,


Annexes

5.4 Conferences abstracts

1. First Swiss Geoscience Meeting, Bale, Switzerland. 2003
2. Second Swiss Geoscience Meeting, Lausanne, Switzerland. 2004
3. 24th IAS Meeting, Muscat, Oman. 2005
4. European Union of Geosciences, Vienna, Austria. 2005
5. International Conference on Glacial Sedimentary Products, Aberystwyth, UK. 2005
6. Snowball Earth conference, Ascona, Switzerland. 2006
Stormy climatic phase following last Neoproterozoic glaciation? 
Shuram Formation (Huqf Supergroup) of Oman

Le Guerroué Erwan, Allen Philip and Cozzi Andrea

The late Neoproterozoic period of Earth history was characterized by extreme climatic oscillations, with glaciations that are thought to have reached equatorial latitudes (Hoffman et al., 2002). These possible snowball Earth events produced a distinctive $\delta^{13}$C signature in the carbonate rocks that are associated with such glacial events. Carbonates underlying glacial diamictites are characteristically isotopically heavy, with values falling prior to the deposition of glacial successions, continuing to be isotopically light in the overlying ‘cap’ carbonates.

The Neoproterozoic Nafun Group of Oman is made of two siliciclastic-carbonate depositional cycles immediately following the last Neoproterozoic glacial event (580 Ma) and ending before the Precambrian-Cambrian boundary (543 Ma). The Shuram Formation, marking the base of the upper cycle, is in the Jabal Akhdar (north Oman) deep marine red and brown siltstones comprising very fine storm-influenced sandstones interbedded with fine limestones. It contains at its base local lowstand sandstones and resedimented carbonate breccias. In the Huqf region (central Oman), the Shuram Formation presents storm-dominated oolitic limestones and shales in a shallower paleoenvironment. The Khufai-Shuram boundary shows evidence for relative sealevel fall and contains a major negative shift in $\delta^{13}$C, with values around +6‰ before the boundary, followed by a negative excursion dropping to a maximum of -12‰ a few meters above it, slowly reaching positive values of +2‰ in the overlying Buah Formation. Although these peculiar features fit generally with those predicted by a snowball Earth event, absence of either glacial deposits or cap carbonates questions a causal relationship with a snowball event.

An alternative scenario capable of explaining the observed chemostratigraphic and sedimentological features at the Khufai-Shuram boundary in Oman, is one of a pluvial climatic interlude during the Shuram deposition, causing the anomalous low $\delta^{13}$C values and the Fe- and radiogenic Sr-rich marine red beds of the Shuram Fm.

References
Major negative $\delta^{13}$C excursion at the last Neoproterozoic (Varangerian) glaciation: The Khufai-Shuram boundary of Oman

Le Guerroué Erwan, Allen Philip and Cozzi Andrea

The late Neoproterozoic period of Earth history was characterized by intense climatic oscillations that are commonly interpreted as snowball Earth events. These glacial epochs are coupled with a distinctive $\delta^{13}$C signature in associated carbonate strata. Carbonates deposited before the initiation of a snowball Earth event are characteristically isotopically heavy (+2-6‰ VPDB), whereas the carbonates directly overlying glaciogenic sedimentary rocks are typically isotopically light (-5-6‰), with a trend towards positive values towards the top of the carbonate sequence.

The Neoproterozoic stratigraphy of Oman contains two such events that are recorded in the glacigenic strata of the Ghubrah Formation and the Fiq Member of the Ghadir Manqil Formation (Abu Mahara Group), which seem to correlate well with the two ‘global’ Sturtian and Marinoan snowball Earth-type glacial epochs, respectively. A third younger glaciation (Varangerian) is supposed to have occurred after 600 Ma and possibly being as young as 580 Ma. If so, this glacial epoch should be recorded in the overlying strata of the 1 km-thick Nafun Group of Oman. The Nafun Group is made of two siliciclastic-carbonate depositional cycles, comprising the Masirah Bay-Khufai formations and Shuram-Buah formations, and overlies the Marinoan-aged Fiq Member of the Ghadir Manqil Formation and its cap carbonate (Hadash Formation). The base of the overlying Ara Group is dated c. 543 Ma, and used as the Precambrian-Cambrian boundary.

The Khufai-Shuram boundary, marking the base of the second depositional cycle, is well exposed in the Jabal Akhdar (northern Oman) and in the Huqf region of central Oman. In the former the outer ramp carbonates of the Khufai Formation are overlain by relatively deeper marine red and brown siltstones and storm-generated sandstones interbedded with fine limestones. Conversely, in the Huqf area shallow-water inner ramp carbonates of the Khufai Formation are directly overlain, without evidence for karstification or an erosional unconformity, by the storm-influenced reddish siltstones of the Shuram Formation. The Khufai-Shuram boundary in both Jabal Akhdar and Huqf sections contains a massive negative $\delta^{13}$C excursion, with values dropping from around +5‰ at the top of the Khufai, to values of -12‰ at the base of the Shuram. This negative shift is long lived and persists throughout the Shuram Formation and for most of the overlying Buah Formation, where only at the top positive values are finally reached (+2‰).

The absence of 1) glacial deposits on the top of the Khufai and 2) a conventional ‘cap’ carbonate sequence, together with a $\delta^{13}$C signature not typical of classic cap carbonate sequences, makes a causal relationship with a snowball event improbable.

Moreover, the Khufai-Shuram boundary negative isotopic shift correlates well with other formations worldwide where the same pattern is found (Wonoka Fm of Australia, Johnnie Fm of California, Squantum and Gaskiers Fm of Newfoundland bracketed to 580-595 Ma, and the Hankalchough Fm of China). Some of these contain glacial diamictites but others show only sings for relative sealevel lowering (e.g., incised valleys) associated with the negative shift. A ‘Snowball Earth’ type glaciation thus seems unlikely for the Varangerian. Moreover, the observed negative $\delta^{13}$C excursion is bigger in amplitude than the one predicted by the snowball theory, calling for other mechanisms capable of producing such a shift.
Major negative $\delta^{13}C$ excursion following the last Neoproterozoic (Varangerian) glaciation: the Khufai-Shuram boundary of Oman

Le Guerroué Erwan, Allen Philip and Cozzi Andrea

The late Neoproterozoic period of Earth history was characterized by intense climatic oscillations that are commonly interpreted as snowball Earth events. These glacial epochs are coupled with a distinctive $\delta^{13}C$ signature in associated carbonate strata. Carbonates deposited before the initiation of a snowball Earth event are characteristically isotopically heavy (+2-6‰ VPDB), whereas the carbonates directly overlying glacigenic sedimentary rocks are typically isotopically light (-5-6‰), with a trend towards positive values towards the top of the carbonate sequence.

The Neoproterozoic stratigraphy of Oman contains two such events that are recorded in the glacigenic strata of the Ghubrah Formation and the Fiq Member of the Ghadir Manqil Formation (Abu Mahara Group), which seem to correlate well with the two ‘global’ Sturtian and Marinoan snowball Earth-type glacial epochs (Brasier et al., 2000; Allen et al., 2004), respectively. A third younger glaciation (corresponding to the Varangerian glacial epoch) is supposed to have occurred after 600 Ma and possibly being as young as 580 Ma (Halverson et al., 2005). If so, this glacial epoch should be recorded in the overlying strata of the 1 km-thick Nafun Group of Oman. The Nafun Group is made of two siliciclastic-carbonate depositional cycles, comprising the Masirah Bay-Khufai formations and Shuram-Buah formations, and overlies the Marinoan-aged Fiq Member of the Ghadir Manqil Formation and its cap carbonate (Hadash Formation) (Allen et al., 2004). The top of the Buah Formation, representing the end of Nafun deposition, is dated at c. 550 Ma (Cozzi & Al-Siyabi, 2004), very close to the Precambrian-Cambrian boundary (543 Ma). The Khufai-Shuram boundary, marking the base of the second depositional cycle, is well exposed in the Jabal Akhdar (northern Oman) and in the Huqf region of central Oman. In the former the outer ramp carbonates of the Khufai Formation are overlain by relatively deeper marine red and brown siltstones and storm-generated sandstones interbedded with fine limestones. Conversely, in the Huqf area shallow-water inner ramp carbonates of the Khufai Formation are directly overlain, without evidence for karstification or an erosional unconformity, by the storm-influenced reddish siltstones of the Shuram Formation.

The Khufai-Shuram boundary in both Jabal Akhdar and Huqf sections contains a massive negative $\delta^{13}C$ excursion, with values dropping from around +6‰ at the top of the Khufai, to values of -12‰ at the base of the Shuram. This negative shift is long lived and persists throughout the Shuram Formation and for most of the overlying Buah Formation, where only at the top positive values are finally reached (+2‰) (Cozzi et al., 2004). The absence of 1) glacial deposits on the top of the Khufai and 2) a conventional ‘cap’ carbonate sequence, together with a $\delta^{13}C$ signature not typical of classic cap carbonate sequences, makes a causal relationship with a snowball event improbable. The Khufai-Shuram boundary negative isotopic shift correlates well with other sections worldwide where the same pattern is found (Halverson et al., 2005). Some of these contain glacial diamictites but others show only sings for relative sealevel lowering (e.g., incised valleys) associated with the negative shift. Therefore, although the base Shuram interval may correspond with a young (Varangerian) non-global glaciation, the mechanisms capable of producing such dramatic changes in the C cycle at this time are currently unclear.

References
The late Neoproterozoic period of Earth history was characterized by intense climatic oscillations that are commonly interpreted as snowball Earth events and correlated worldwide by sharp negative $\delta^{13}C$ excursions. On top of a dated 800-825 Ma basement (Leather 2001), the Neoproterozoic stratigraphy of Oman contains two records of two glacial epochs - the Ghubrah (dated 712 Ma; Leather 2001) and the Fiq formations (currently undated), separated by a major unconformity (Le Guerroué et al. 2005b) which seem to correlate well with the two ‘global’ Sturtian (700-750) and Marinoan (630-640) snowball Earth-type glacial epochs (Brasier et al. 2000; Allen et al. 2004; Le Guerroué et al. 2005b). A third younger glaciation (corresponding to the Varangerian glacial epoch; Halverson et al. 2005) is supposed to have occurred after 600 Ma and possibly being as young as 580 Ma (Le Guerroué et al. 2005a). The fingerprint of this last glacial epoch, preserved on every continental fragment (proper equivalent are: Nama Group of Namibia; Wonoka Formation of Australia and Johnnie Formation of California; Le Guerroué et al. 2005a), the largest negative carbon isotopic excursion of earth history.

The Varangerian epoch is recorded in the overlying strata of the 1 km-thick Nafun Group of Oman. The Nafun Group is made of two siliciclastic-carbonate depositional cycles, comprising the Masirah Bay-Khufai formations and Shuram-Buah formations, and overlies the probable Marinoan-aged Fiq Formation and its cap carbonate (Allen et al. 2004; Le Guerroué et al. 2005b). The top of the Buah Formation, representing the end of Nafun deposition, is dated at c. 550 Ma (Cozzi and Al-Siyabi 2004), very close to the Precambrian-Cambrian boundary (542 Ma).

The Khufai-Shuram boundary is well exposed in the core of the Jabal Akhdar (northern Oman) and in the Huqf region of central Oman. In the former the outer ramp carbonates of the Khufai Formation are overlain by relatively deep marine red and brown siltstones and storm-generated sandstones interbedded with fine limestones. In the Huqf area, however, shallow-water inner ramp carbonates of the Khufai Formation are directly overlain by the storm-influenced reddish siltstones of the Shuram Formation.

Although the Khufai-Shuram boundary does not show evidence for glacial depositional settings, in both Jabal Akhdar and Huqf sections it records a large negative $\delta^{13}C$ excursion, with values dropping from around +4‰ at the top of the Khufai, to values of -12‰ at the base of the Shuram. This negative shift observed in the whole Oman, persists throughout the Shuram Formation and for most of the overlying Buah Formation, where only at the top positive values are finally reached (+2‰) (Cozzi and Al-Siyabi 2004; Le Guerroué et al. 2005a). The Khufai-Shuram boundary negative isotopic shift correlates well with other sections worldwide where the same isotopic pattern is found (Le Guerroué et al. 2005a). Most of these sections do not contain glacial diamictites nor cap carbonates associated with the negative shift. Therefore, the isotopically light Shuram Formation interval may correspond with a young (Varangerian) non-global glaciation, discarding any causal relationship between the negative $\delta^{13}C$ excursion and a snowball Earth event. At this time, the
mechanism(s) capable of producing such dramatic changes in the C cycle are still an unresolved problem.

Detrital zircon analysis from north and east-central Oman shows zircon sub-populations characteristic of basement (800-820 Ma), and magmatic events coinciding with the Sturtian (Ghubrah Formation equivalent 710-730 Ma) and Marinoan glaciations (Fiq Formation equivalent 630-640 Ma). The youngest sub-population shows an average age of 609 Ma with grains as young as 590 Ma. This age may be close to the depositional age of the Khufai-Shuram boundary and currently provides the best estimate of a maximum age for the large, possibly Varangerian, isotopic excursion.

References:
The largest $\delta^{13}C$ excursion of Earth History: the late Neoproterozoic Khufai-Shuram boundary of Oman.

Le Guerroué Erwan, Allen Philip and Cozzi Andrea

The Neoproterozoic stratigraphy of Oman contains a record of two glacial epochs represented by the Ghubrah Formation (dated ~712 Ma; Leather 2001) and the Fiq Member of the Ghadir Manqil Formation (currently undated). The Ghubrah and Fiq are separated by a major unconformity, and appear to correlate well with the two 'global' Sturtian (700-750 Ma) and Marinoan (630-640 Ma) Snowball Earth glacial epochs. A third younger glaciation (corresponding to the Varangerian glacial epoch) is thought to have occurred after 600 Ma and possibly as young as 580 Ma. This last period bears, on every continental fragment, the largest negative carbon isotope excursion known of Earth history.

The Khufai-Shuram boundary is well exposed in the core of the Jabal Akhdar (north Oman) and in the Huqf region of east-central Oman. In the former, the outer ramp carbonates of the Khufai Formation are overlain by relatively deep marine red and brown siltstones and storm-generated sandstones interbedded with fine limestones. However, in the Huqf area shallow-water inner ramp carbonates of the Khufai Formation are directly overlain by storm-influenced reddish siltstones of the Shuram Formation.

Although the Khufai-Shuram boundary does not show evidence for glacial depositional settings, in both the Jabal Akhdar and Huqf sections it records a large negative $\delta^{13}C$ excursion, with values dropping from around +4‰ at the top of the Khufai, to values of -12‰ at the base of the Shuram. This negative shift, which is observed through the whole of Oman, persists throughout the Shuram Formation and for most of the overlying Buah Formation, where only at the top positive values are finally reached (+2‰).

Detrital zircon analysis from north and east-central Oman shows zircon sub-populations characteristic of basement (800-820 Ma), and magmatic events coinciding with the Sturtian (Ghubrah Formation equivalent 710-730 Ma) and Marinoan glaciations (Fiq Formation equivalent 630-640 Ma). The youngest sub-population shows an average age of 609 Ma with grains as young as 590 Ma. This age may be close to the depositional age of the Khufai-Shuram boundary and currently provides the best estimate of a maximum age for the large Varangerian isotopic excursion.

The Khufai-Shuram boundary negative isotopic shift correlates well with other sections worldwide where the same pattern is found. Most of these other sections contain neither glacial diamicites nor cap carbonates associated with the negative shift questioning a causal relationship of $\delta^{13}C$ excursion with Snowball Earth events. The Varangerian epoch shows severe glaciation, probably reaching low latitudes but could be restricted to tectonically active margins, with glaciers nucleated on local relief. However, the mechanisms capable of producing such dramatic changes in the C cycle at this time are currently unclear.
The largest $\delta^{13}C$ excursion of Earth History:
Ediacaran Shuram Formation of Oman

Le Guerroué Erwan, Allen Philip and Cozzi Andrea

A large portion of the Ediacaran period (Knoll et al., 2004), extending from the end of the Marinoan glaciation (c. 635 Ma; Hoffmann et al., 2004; Condon et al., 2005) to the Precambrian-Cambrian boundary (542 Ma; Amthor et al., 2003), is occupied by large negative carbon isotope excursions. These include Marinoan ‘cap carbonate’ excursion, the Shuram-Wonoka and the Precambrian-Cambrian boundary excursions.

The negative $\delta^{13}C$ excursion within the Shuram Fm of Oman (Nafun Group, Huqf Supergroup) is characterised by an exceptional amplitude (+5‰ to -12‰ $\delta^{13}C$) and long uninterrupted stratigraphic record (~800 m). This carbon isotopic trend is reproducible throughout Oman, from outcrops to the subsurface, and irrespective of sedimentary facies (Le Guerroué et al., in review-b). The entire excursion is essentially in phase with long term relative sea level changes. The nadir in $\delta^{13}C$ values occurs at the level of the maximum flooding zone of the lower Shuram Fm, while the cross-over to positive values is found within the overlaying Buah Fm highstand systems tract. Radiometric ages and thermal subsidence modelling constrain the excursion in time, and indicate an onset at ~600 Ma, and a duration of approximately 50 Myr (Le Guerroué et al., in review-c).

The Shuram Formation is extremely well exposed for over 40 km in the north of the Huqf area, where it displays a stack of shallowing-upward storm-dominated parasequences (Le Guerroué et al., in review-a). At the parasequence scale, carbon isotopic values are shown to reflect stratigraphic position within the parasequence stack, and each individual parasequence shows a trend in $\delta^{13}C$ values that highlight the direction of lateral facies progradation. These combined stratigraphic-carbon isotopic observations, and the fact that the trend is reproducible throughout Oman, support a primary, oceanographic origin for the carbon isotopic ratios.

The Shuram excursion has potential correlatives in Ediacaran strata elsewhere (Wonoka Formation of the Adelaide Rift Complex of South Australia, Calver, 2000: Johnnie Formation of the Death Valley region of western U.S.A., Corsetti and Kaufman, 2003: Doushantuo Formation of south China, Condon et al., 2005; Krol Group of northern India, Jiang et al., 2002), and may thus represent a characteristic feature of the Ediacaran period. However, these possible correlative sections have a limited $\delta^{13}C$ data set and are dissected by unconformities, so that the full excursion cannot be recognized. The Doushantuo probably records the end of the excursion at around c. 551 Ma (Condon et al., 2005).

The Ediacaran period is also marked by the Gaskiers non-global, short-lived glaciation around 580 Ma (Krogh et al., 1988) and in possible loosely age-constrained correlatives. If the proposed chronology for the Shuram excursion is correct, the Gaskiers-aged glaciation is embedded within a large-amplitude, long-term negative carbon isotopic excursion, and appears to have had no effect on the chemostratigraphic records of Oman and other sections worldwide.

The fact that a carbon isotope excursion of this magnitude can be recognized in marine Ediacaran rocks from several continents indicates that it was a very widespread oceanographic phenomenon, reflecting the composition of seawater from which carbonate minerals were precipitated. Such an excursion is highly unusual in the geological record and is challenging to explain. However, the Shuram excursion demonstrates that this negative carbon isotopic excursion is unrelated to glaciation, and therefore the marine carbon isotopic record cannot be used as a direct proxy for...
Neoproterozoic ice ages.

Any explanation of the Shuram shift must therefore involve an exceptionally long residence time of a sufficiently large reservoir of $^{13}$C-depleted material (e.g., dissolved organic carbon, if compared to Phanerozoic examples of perturbations of the carbon cycle.

References


Curriculum vitae

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