Terms and Conditions of Use of Digitised Theses from Trinity College Library Dublin

Copyright statement

All material supplied by Trinity College Library is protected by copyright (under the Copyright and Related Rights Act, 2000 as amended) and other relevant Intellectual Property Rights. By accessing and using a Digitised Thesis from Trinity College Library you acknowledge that all Intellectual Property Rights in any Works supplied are the sole and exclusive property of the copyright and/or other IPR holder. Specific copyright holders may not be explicitly identified. Use of materials from other sources within a thesis should not be construed as a claim over them.

A non-exclusive, non-transferable licence is hereby granted to those using or reproducing, in whole or in part, the material for valid purposes, providing the copyright owners are acknowledged using the normal conventions. Where specific permission to use material is required, this is identified and such permission must be sought from the copyright holder or agency cited.

Liability statement

By using a Digitised Thesis, I accept that Trinity College Dublin bears no legal responsibility for the accuracy, legality or comprehensiveness of materials contained within the thesis, and that Trinity College Dublin accepts no liability for indirect, consequential, or incidental, damages or losses arising from use of the thesis for whatever reason. Information located in a thesis may be subject to specific use constraints, details of which may not be explicitly described. It is the responsibility of potential and actual users to be aware of such constraints and to abide by them. By making use of material from a digitised thesis, you accept these copyright and disclaimer provisions. Where it is brought to the attention of Trinity College Library that there may be a breach of copyright or other restraint, it is the policy to withdraw or take down access to a thesis while the issue is being resolved.

Access Agreement

By using a Digitised Thesis from Trinity College Library you are bound by the following Terms & Conditions. Please read them carefully.

I have read and I understand the following statement: All material supplied via a Digitised Thesis from Trinity College Library is protected by copyright and other intellectual property rights, and duplication or sale of all or part of any of a thesis is not permitted, except that material may be duplicated by you for your research use or for educational purposes in electronic or print form providing the copyright owners are acknowledged using the normal conventions. You must obtain permission for any other use. Electronic or print copies may not be offered, whether for sale or otherwise to anyone. This copy has been supplied on the understanding that it is copyright material and that no quotation from the thesis may be published without proper acknowledgement.
SEDIMENTOLOGY, CHEMOSTRATIGRAPHY, AND GEOCHRONOLOGY OF THE LOWER HUQF SUPERGROUP, OMAN

Volume 1 of 2

Jonathan J. Leather

Submitted to the University of Dublin for the degree of Doctor of Philosophy

September 2001
DECLARATION

This thesis is entirely my own work, except where stated. It has not been submitted to this or any other university. All references are duly acknowledged. I agree that the library of Trinity College, Dublin may lend it on request.

[Signature]

[Signature]

[Signature]
SUMMARY

The lower part of the Huqf Supergroup, comprising the Abu Mahara Group and the Hadash and Masirah Bay Formations of the Nafun Group, forms the oldest deposits exposed in Oman. This thesis is based on field study of this lower part of the Huqf Supergroup in the Jebel Akhdar and Huqf areas of Oman. The aim of the thesis is to provide a detailed sedimentological interpretation of this portion of the Oman stratigraphy, as well as analysing the implications of new chemostratigraphic and chronological data from Oman in a regional and global context.

Volcanics and volcaniclastic deposits of the Halfayn Formation in the north of the Huqf area form the oldest deposits of the Huqf Supergroup. Previously, the Halfayn Formation has been correlated with volcanics that occur within the Abu Mahara Group of the Jebel Akhdar. However, U-Pb zircon dating of volcanic horizons suggests that the Halfayn Formation probably pre-dates the oldest deposits of the Jebel Akhdar by 80Myrs. An unconformity occurs at the top of the Halfayn Formation, and no further deposition is recorded in the Huqf area until the base of the Nafun Group.

In the Jebel Akhdar, the Abu Mahara Group preserves a record of a particularly interesting period of Neoproterozoic time when widespread, possibly even global glaciations occurred. An ash bed from within a diamictite unit of the Ghubrah Formation has yielded a U-Pb zircon date of 711.8±1.6Ma. Comparing this to a previous date produced from the same bed of 723±16/-10Ma (Brasier et al., 2000), confirms that the glacial horizons of the Ghubrah Formation are Sturtian equivalents. The date of 711.8±1.6Ma actually comes from within a glacial unit, and consequently provides one of the best estimates yet produced on the age of the Sturtian glaciations.

The Ghubrah Formation is overlain by the deposits of the Ghadir Manqil Formation, which form the upper part of the Abu Mahara Group in the Jebel Akhdar. Chronological and stratigraphic considerations suggest that the Ghadir Manqil Formation is actually a Marinoan equivalent, and that the boundary between the Ghubrah and Ghadir Manqil Formations represents an unconformity spanning c. 100Myrs. At the base of the Ghadir Manqil Formation in the east of the Jebel Akhdar, the basaltic Saqlah Member occurs.
This has been linked to the more widely developed volcanics of the Saih Hatat region further to the east. The geochemistry of these basalts suggests they formed from intracratonic extension, with stretch factors higher than 1.5 indicated in the Saih Hatat region. The upper Ghadir Manqil Formation comprises the well-exposed Fiq Member. Stratigraphic events within the Fiq Member can be recognised across the whole Jebel Akhdar, allowing confident correlations to be made. A number of discrete glacial units separated by non-glacial facies occur within the Fiq Member. These are incompatible with the long-lived, synchronous global glaciations suggested by the snowball Earth model (eg. Hoffman et al., 1998a, b).

The Ghadir Manqil Formation in the Jebel Akhdar is overlain by the Hadash Formation, which is taken as the base of the Nafun Group in this thesis. The Hadash Formation is a typical Neoproterozoic ‘cap carbonate’. As cap carbonates have been recognised globally, and contain a distinctive negative carbon isotopic signature, a detailed isotopic study of the Hadash Formation was conducted. Attempting to correlate the Hadash Formation on a regional scale based on isotopic signatures proved problematic. This suggests that attempting to correlate sections globally based on the presence of cap carbonates may be hazardous, and has implications for the proposed definition of a new ‘Terminal Proterozoic System’.

The Masirah Bay Formation is well-exposed in both the Huqf and Jebel Akhdar areas. In the Huqf, shallow marine deposition dominates and two progradational, coarsening-up cycles occur. These cannot be recognised in the Jebel Akhdar, where deeper-water sedimentation dominates.

The extensive, laterally persistent nature of the deposits of the Hadash and Masirah Bay Formations suggest that the extensional tectonics and basin partitioning that characterised the Ghadir Manqil Formation had finished by the start of Nafun time.
ACKNOWLEDGEMENTS

I would firstly like to thank my two supervisors, Philip Allen and Martin Brasier, who have both been very supportive throughout this project. Philip, thanks for all the advice, both in the field and on the manuscript, and also for some great games of snooker; and Martin, thanks for suggesting the project to me in the first place and for always being so enthusiastic and encouraging ideas. Gretta McCarron provided a great introduction to fieldwork in Oman and has always been more than helpful with the project. Many thanks are also due to Petroleum Development Oman and Forbairt for the generous funding of this project.

I am very much indebted to all four of my field assistants – not only for all the sampling, driving, discussions, and cooking, but also for putting up with me for long periods of time. My time in Oman was always made much more enjoyable by your company, so Tim, Jamie, Malone, and Niamh – thank you all very much – I couldn’t have done it without you.

I also owe many thanks to all the staff in the exploration department of Petroleum Development Oman, who have always been friendly and keen to help, even when they were busy. Especial thanks are due to Salim al Maskery, who showed us how to do field work in style, and went to every effort to help get us equipment and access to military areas that otherwise would have stayed off limits, and also to Joachim Amthor, who put so much time into the project, was always enthusiastic, and who could be persuaded now and again to have a few beers. Many thanks to everyone else who was involved with the project in Oman – Andrea Cozzi, Stefan Schroeder, Albert Matter, John Grotzinger, Sam Bowring, Charlotte Schreiber, and of course Chris Nicholas who I’d like to thank for all the help and the cheeky pints in the boat club.

My time spent in Dublin working on this project has been thoroughly enjoyable and has been made so by all the people I’ve got to know here. I’d like to thank Duncan, Gerry, and John Murray and the rest of his class for making me feel very welcome when I first arrived. Along with Jamie, John Malone, Jackie James (who put up with me in a confined space for so long), Jacqui Connolly, Niamh, Phonsie and Mike who’ve all been around for
the entirety of the project, there’s always been somebody willing to accompany (and very occasionally drag) me to the pub. Many thanks to the rest of the postgrads, Mags, Paulo, Brian, Kay, Jeff, Karl, Beccy, Stuart, Alan, Cristina, Pete, Kerstin, Gemma, Claire, Kate, Ric, Dave, and of course Sarah, some of whom are no longer students, but all of whom have always been good fun and happy to help with any problems.

I’d also like to thank all the rest of the staff in the department who have always been nothing but helpful. Thanks to Neil, Frank, Declan, and Stephen for all the help, and in the past few months especially thanks to Declan for all the help with the printers and photographs. Thanks to Joann for all her help, George Sevastopulo, who has been willing to discuss any aspect of the project over the past few years, John Graham, Chris Stillman, Geoff Clayton, Dave Doff for all his help in the geochem lab, Ian Sanders, Adrian Phillips, Patrick Wyse Jackson, and of course Alex Densmore. I’d also like to thank everyone in the labs in Oxford and Royal Holloway, particularly Julie Cartlidge and Nikki Paige.

Thanks to the rest of the friends I’ve made in Dublin not mentioned above, particularly Declan, Ann, Liadhan, Melissa, Maura, Barbara, Skelton, the badminton club, and all the staff at the Library Shop, who made working there very pleasant.

I would like to give particular thanks to Dad, Cathie, Mum, Amy, and Daisy for all the support you’ve given me and for always being there for me – not just over the past four years. I really can’t thank you enough for everything.

As for Karen, thank you so much for all the support, for all the fun we’ve had, and just for generally being great. You’ve made the last two years in Dublin some of the happiest times of my life and I don’t know how to thank you properly.

Finally, I’d like to thank my undergraduate tutor, Keith Cox, and dedicate this thesis to his memory. He was a great teacher who instilled a real enthusiasm for geology in everyone he taught.
TABLE OF CONTENTS

VOLUME 1

Declaration
Summary
Acknowledgements

Chapter 1. Introduction

1.1. Introduction 1
1.2. Regional setting 1
1.3. Tectono-stratigraphic history 2
   1.3.1. Arabian tectonics in the Neoproterozoic 2
   1.3.2. Tectonic framework for the Neoproterozoic of Oman 3
1.4. Huqf Supergroup nomenclature 6
1.5. Lower Huqf Supergroup stratigraphy and facies associations 7
1.6. Thesis outline 7
1.7. Project development 8

Chapter 2. The Halfayn Formation

2.1. Introduction 10
2.2. Methods 10
2.3. Previous work 11
2.4. Subdivision of the Halfayn Formation 12
2.5. Facies analysis of the Halfayn Formation 12
   2.5.1. Unit H1 – Green rhyolitic ignimbrite 12
   2.5.2. Unit H2 – Unwelded distal pyroclastic flow deposits 12
   2.5.3. Unit H3 – Shallow marine to fluvial pyroclastic sediments 13
2.6. Halfayn Formation synthesis 14
2.7. Age of the Halfayn Formation 14
Chapter 3. The Abu Mahara Group in the Jebel Akhdar: The Ghubrah and Ghadir Manqil Formations

Part I: Introduction

3.1. Introduction
3.2. Methods
3.3. Previous work
3.4. Subdivision and nomenclature of the Abu Mahara Group in the Jebel Akhdar
3.5. Stratigraphy of the Abu Mahara Group in the Jebel Akhdar
  3.5.1. Jabir Formation
  3.5.2. Ghubrah Formation
  3.5.3. Ghadir Manqil Formation: Saqlah Member
  3.5.4. Ghadir Manqil Formation: Fiq Member

Part II. Analysis of the Abu Mahara Group in the Jebel Akhdar

3.6. The Jabir Formation
3.7. Facies analysis of the Ghubrah Formation
  3.7.1. Distal glaciomarine facies association
  3.7.1.1. Massive diamictite facies (Dm)
  3.7.1.2. Massive and graded siltstone facies (Fm and Fg)
  3.7.1.3. Dropstone laminite facies (Fld)
  3.7.1.4. Carbonate facies (L)
  3.7.1.5. Tuffaceous deposits
  3.7.2. Ghubrah Formation synthesis
3.8. The Saqlah Member of the Ghadir Manqil Formation
3.9. Facies analysis of the Fiq Member of the Ghadir Manqil Formation
  3.9.1. Distal glaciomarine facies association
  3.9.1.1. Massive diamictite facies (Dm)
  3.9.1.2. Dropstone laminite facies (Fld)
  3.9.2. Proximal glaciomarine facies association
3.9.2.1. Massive diamictite facies (Dm) 35
3.9.2.2. Stratified diamictite facies (Ds) 37
3.9.2.3. Conglomeratic facies (C) 39
3.9.2.4. Massive/graded sandstone facies (Sm/Sg) 40
3.9.2.5. Rippled sandstone and siltstone facies (Sr and Fr) 41
3.9.2.6. Dropstone laminite facies (Fld) 43
3.9.2.7. Massive/laminated mud/siltstone facies (Fm/Fl) 43
3.9.3. Non-glacial sediment gravity flow facies association 44
3.9.3.1. Conglomeratic facies (C) 44
3.9.3.2. Pebby sandstone facies (Sp) 46
3.9.3.3. Massive/graded sandstone facies (Sm/Sg) 47
3.9.3.4. Rippled sandstone and siltstone facies (Sr and Fr) 50
3.9.3.5. Massive and laminated mud/siltstone facies (Fm and Fl) 52
3.9.3.6. Cross-stratified sandstone facies (Sx) 55
3.9.3.7. Volcaniclastic sandstone facies (Sv) 56
3.9.3.8. Carbonate facies (L) 58
3.9.4. Non-glacial shallow marine facies association 63
3.9.4.1. Conglomeratic facies (C) 63
3.9.4.2. Rippled sandstone facies (Srw) 63
3.9.4.3. Rippled heterolithic (Sr, Srw, Fr, Fl) and massive sandstone facies (Sm) 64

3.10. Fiq Member synthesis 65
3.10.1. Facies evolution of the Fiq Member 66

Part III. Summary

3.11. Summary of the Abu Mahara Group in the Jebel Akhdar 74

Chapter 4. Discussion of the Abu Mahara glacials

4.1. Introduction 76
4.2. Global time-scale for the Neoproterozoic glaciations 77
4.2.1. Age of the Sturtian glacial epoch 78
4.2.2. Age of the older Marinoan glaciations 78
4.2.3. Age of the younger Marinoan glaciations 78

4.3. Age of the Abu Mahara glacial events in Oman 79

4.4. Nature of the widespread Neoproterozoic glaciations 84
   4.4.1. Snowball Earth hypothesis 85
   4.4.2. High obliquity hypothesis 86
   4.4.3. Discussion of models 87
   4.4.4. Discussion of the Abu Mahara Group data 88

4.5. Summary 90

**Chapter 5. The Hadash Formation**

**Part I. Introduction and facies analysis**

5.1. Introduction 91
5.2. Methods 92
5.3. Previous work 93
5.4. Placement of the Hadash Formation at the base of the Nafun Group 94
5.5. Lithological subdivision of the Hadash Formation 95
5.6. Facies analysis of the Hadash Formation 96
   5.6.1. Facies E1: Dolomitic microspar with locally developed siltstone intercalations 96
   5.6.2. Facies E2: Recessive, siltstone dominated deposits 98
   5.6.3. Facies E3: Planar bedded dolomitic microspar to dolospar 98
   5.6.4. Facies E4: Interbedded crystalline limestone and dolospar 99
   5.6.5. Facies E5: Thin beds of dolomite in siltstone and sandstone 100
   5.6.6. Facies W1: Thinly-interbedded dolomite and siltstone 101
   5.6.7. Facies W2: Dolomitic microspar with locally developed siltstone intercalations 102
   5.6.8. Facies W3: Thin beds of dolomite (‘stringers’) within laminated mud/siltstone 103
   5.6.9. Facies H1: Crystalline dolomite 104

5.7. Lithological synthesis of the Hadash Formation 104
Part II. Isotopic and geochemical analysis of the Hadash Formation

5.8. Introduction

5.9. Principles of carbon and oxygen chemostratigraphy
5.9.1. Carbon-isotope fractionation
5.9.2. Post-depositional isotope fractionation in sedimentary and organic carbonates
5.9.3. Oxygen-isotope fractionation
5.9.4. Post-depositional isotope fractionation in sedimentary and microbial carbonates

5.10. Elemental chemistry of the Hadash Formation
5.10.1. Mg/Ca ratios
5.10.2. Strontium (Sr) and Sr/Ca ratios
5.10.3. Iron (Fe)
5.10.4. Mn/Sr as a proxy for diagenetic exchange
5.10.5. Mn and Fe as a proxy for diagenetic exchange
5.10.6. Carbon and oxygen isotopes
5.10.7. Component analysis

5.11. Data correction
5.12. Discussion of stable isotope data from the Hadash Formation
5.13. Use of stable isotopes for intra-basinal high resolution correlations within the Hadash Formation

Part III. Discussion of the Hadash Formation

5.14. Discussion of the Hadash Formation
5.15. Summary of the Hadash Formation

Chapter 6. The Masirah Bay Formation

6.1. Introduction
6.2. Methods
6.3. Previous work
6.4. Subdivision of the Masirah Bay Formation

6.5. Facies analysis of the Masirah Bay Formation in the Huqf area

6.5.1. Facies Association A: Wave-dominated shoreface deposits

6.5.1.1. Facies A1: Wave-rippled sandstone and siltstone

6.5.1.2. Facies A2: Wave-rippled sandstone

6.5.1.3. Facies A3: Thinly-bedded planar-laminated sandstone

6.5.1.4. Facies Association A: Synthesis

6.5.2. Facies Association B: High-energy current dominated deposits

6.5.2.1. Facies B1: Hummocky/swaley cross-stratified sandstone

6.5.2.2. Facies B2: Sigmoidally cross-stratified sandstone

6.5.2.3. Facies B3: Planar cross-stratified sandstone

6.5.2.4. Facies B4: Trough cross-stratified sandstone

6.5.2.5. Facies B5: Large-scale planar cross-stratified sandstone

6.5.2.6. Facies B6: Granular lag deposits

6.5.2.7. Facies Association B: Synthesis

6.5.3. Facies Association C: Offshore storm-influenced deposits

6.5.3.1. Facies C1: Dark grey to red shales

6.5.3.2. Facies C2: Interbedded shales/siltstones and sandstones

6.5.3.3. Facies Association C: Synthesis

6.5.4. Facies Association D: High-energy current-dominated deposits

6.5.4.1. Facies D1: Thinly-bedded undulatory to wave-rippled sandstone

6.5.4.2. Facies D2: Planar cross-stratified sandstone

6.5.4.3. Facies D3: Granular deposits

6.5.4.4. Facies D4: Trough cross-stratified sandstone

6.5.4.5. Facies D5: Bi-directionally cross-stratified sandstone

6.5.4.6. Facies D6: Tidally-bundled sandstone

6.5.4.7. Facies Association D: Synthesis

6.5.5. Facies Association E: Lower shoreface to offshore deposits

6.5.5.1. Facies E1: Very thinly-bedded sandstones with microbial ‘wrinkle’ structures

6.5.5.2. Facies E2: Interbedded carbonate and siltstone beds

6.5.5.3. Facies Association E: Synthesis

6.6. Synthesis of the Masirah Bay Formation in the Huqf area
6.7. The Masirah Bay Formation in the Jebel Akhdar

6.7.1. Facies analysis

6.7.1.1. Facies 1: Argillaceous siltstone

6.7.1.2. Facies 2: Limestone

6.7.1.3. Facies 3: Medium- to coarse-grained cross-stratified sandstone

6.8. Synthesis of the Masirah Bay Formation in the Jebel Akhdar

6.9. Summary of the Masirah Bay Formation

Chapter 7. Discussion and conclusions

7.1. Introduction

7.2. Discussion of the Huqf Supergroup

7.2.1. The Halfayn Formation

7.2.2. The Jabir Formation

7.2.3. The Ghubrah Formation

7.2.4. The Ghadir Manqil Formation

7.2.5. The Hadash Formation

7.2.6. The Masirah Bay Formation

7.2.7. The Khufai, Shuram, and Buah Formations

7.3. Subsidence history of the Huqf Supergroup

References
All tables, figures, and plates are presented by chapter.

**LIST OF TABLES**

**Chapter 1**

Table 1.1. Radiometric dates from the crystalline basement of Oman
Table 1.2. Facies associations of the Abu Mahara and lower Nafun Groups

**Chapter 3**

Table 3.1. Facies associations within the Abu Mahara Group of the Jebel Akhdar
Table 3.2. Summary of facies present in the Ghubrah Formation
Table 3.3. Chemical composition of volcanic samples from Wadi Mistal and Saih Hatat
Table 3.4. Summary of sedimentary facies present within the Fiq Member
Table 3.5. Isotopic and geochemical data from the Abu Mahara Group in the Jebel Akhdar

**Chapter 5**

Table 5.1. Facies present within the Hadash Formation
Table 5.2. Geochemistry data from the Hadash Formation

**Chapter 6**

Table 6.1. Facies associations within the Masirah Bay Formation in the Huqf area
Table 6.2. Summary of sedimentary facies present in the Masirah Bay Formation in the Huqf area
Table 6.3. Stable isotope results from the Masirah Bay Formation
LIST OF FIGURES

Chapter 1

Fig. 1.1. Map of Oman showing the distribution of Precambrian exposure
Fig. 1.2. Simplified tectonic map of Oman
Fig. 1.3. Simplified geological map of the central Oman Mountains
Fig. 1.4. The Precambrian terranes of the Arabian Peninsula with the Neoproterozoic-
Cambrian Najd sinistral strike-slip fault system
Fig. 1.5. Chronostratigraphic chart of the Arabian accretionary terranes
Fig. 1.6. Late Neoproterozoic (545Ma) continental configuration of ‘Pannotia’
Fig. 1.7. Map showing the Late Precambrian-Cambrian geology of Gondwana with
emphasis on Africa and Arabia
Fig. 1.8. Comparison of the Mirbat, Huqf, and Jebel Akhdar areas
Fig. 1.9. Huqf Supergroup nomenclature used within the present study in the Huqf and
Jebel Akhdar areas

Chapter 2

Fig. 2.1. Dubreuilh et al.’s (1992) log of the Halfayn Formation
Fig. 2.2. Sketch map of basement and Halfayn Formation exposures in the Al Jobah area
Fig. 2.3. Log of the Halfayn Formation at Al Jobah
Fig. 2.4. U-Pb concordia plot for single zircons from granodiorite at Al Jobah, Huqf area
(Sample AJD2)
Fig. 2.5. U-Pb concordia plot for single zircons from granodiorite at Al Jobah, Huqf area
(Sample AJD5)
Fig. 2.6. U-Pb concordia plot for single zircons from welded tuff of Unit H1, Halfayn
Formation (Sample AJD3)
Fig. 2.7. U-Pb concordia plot for single zircons from unwelded ignimbrite of Unit H1,
Halfayn Formation (Sample AJD1)
Fig. 2.8. U-Pb concordia plot for single zircons from angular pyroclastic conglomerate of
Unit H3, Halfayn Formation (Sample AJD4)
Chapter 3

Fig. 3.1. Locality map for the Abu Mahara Group in the Jebel Akhdar
Fig. 3.2. Summary logs of the Abu Mahara Group in the Jebel Akhdar
Fig. 3.3. The Ghadir Manqil Formation in Wadi Sahtan
Fig. 3.4. The Ghadir Manqil Formation in Wadi Hajir
Fig. 3.5. The Ghadir Manqil Formation in Wadi Mistal
Fig. 3.6. Clast analysis method
Fig. 3.7. Lower Huqf Supergroup nomenclature used to date and within the present study in the Jebel Akhdar
Fig. 3.8. Cluster diagram showing the relationships between diamicrite units based on clast type
Fig. 3.9. Nb/Y plot for the volcanic rocks of Saih Hatat and the Saqlah Member, Jebel Akhdar
Fig. 3.10. Nomenclature of normal volcanic rocks defined by plot of total alkalis vs silica
Fig. 3.11. Ti/100-Zr-Y*3 for the volcanic rocks of the Hatat Formation, Saih Hatat, and the Saqlah Member, Wadi Mistal
Fig. 3.12. 2Nb-Zr/4-Y plot for the volcanic rocks of the Hatat Formation, Saih Hatat, and the Saqlah Member, Wadi Mistal
Fig. 3.13. Processes acting, and some of the facies deposited in the glaciomarine facies associations envisaged for the Ghadir Manqil Formation.
Fig. 3.14. Palaeocurrent data from the Fiq Member of Wadi Hajir, Wadi Sahtan, and Wadi Mistal
Fig. 3.15. Deposits of postulated channel and associated overbank turbidites in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir during Fiq time
Fig. 3.16. δ13C and δ18O for the carbonate beds at the base of the Fiq Member in Wadi Mistal
Fig. 3.17. Isotopic and elemental plots for the carbonate beds at the base of the Fiq Member in Wadi Mistal
Fig. 3.18. Changes in relative sea-level at the three main logged localities in the Jebel Akhdar
Fig. 3.19. Development of the Abu Mahara Group in the Jebel Akhdar
Chapter 4

Fig. 4.1. Global chronostratigraphic framework for the Neoproterozoic and possible scenarios that could be envisaged for the Omani stratigraphy

Fig. 4.2. U-Pb concordia plot for single zircons from sample Wadi MT, Unit F1a of the Fiq Member

Fig. 4.3. U-Pb concordia plot for single zircons from sample Wadi MT, Unit F1a of the Fiq Member

Fig. 4.4. U-Pb concordia plot for single zircons from an ash bed within diamictite of the Ghubrah Formation

Fig. 4.5. Stratigraphic columns showing carbon-isotope excursions and glacial horizons in the Oman reference section and other correlative sections

Fig. 4.6. Chronological constraints and suggested global correlations of Neoproterozoic glacial intervals

Chapter 5

Fig. 5.1. Locality map for the Hadash Formation in the Jebel Akhdar

Fig. 5.2. Summary logs of the Hadash Formation in the Jebel Akhdar

Fig. 5.3. Possible Hadash Formation equivalent at Al Jobah in the Huqf area

Fig. 5.4. Depositional models for the Hadash Formation in the Jebel Akhdar

Fig. 5.5. Plots of 1000Sr/Ca vs Mg/Ca for the Hadash Formation in the Jebel Akhdar

Fig. 5.6. Plots of 1000Sr/Ca vs Mn for the Hadash Formation in the Jebel Akhdar

Fig. 5.7. Mn vs Sr plot for the Hadash Formation data

Fig. 5.8. Plots of Mn vs Sr for the Hadash Formation by locality and facies

Fig. 5.9. Mn/Sr vs δ13C for the Hadash Formation

Fig. 5.10. Mn/Sr vs δ13C for the Hadash Formation by locality and facies

Fig. 5.11. Mn and Fe plotted against carbon and oxygen isotope values for the Hadash Formation

Fig. 5.12. δ13C vs δ18O for the Hadash Formation in the Jebel Akhdar

Fig. 5.13. Plots of δ13C vs δ18O for the Hadash Formation by locality and facies

Fig. 5.14. Isotopic curves from the Hadash Formation

Fig. 5.15. Summary of carbon isotopic data from the Hadash Formation
Fig. 5.16. Carbon isotopic curve plotted for Hadash 1 when only samples run for elemental geochemistry are included

Fig. 5.17. Stable isotopic curves produced by Kennedy (1996) from Marinoan postglacial cap dolostones in Australia.

Chapter 6

Fig. 6.1. Locality map for the Masirah Bay Formation in the Huqf area
Fig. 6.2. Log of corehole-17, drilled in the centre of the Khufai Dome
Fig. 6.3. Locality map for the Masirah Bay Formation in the Jebel Akhdar
Fig. 6.4. Summary logs of the Masirah Bay Formation in the Huqf area
Fig. 6.5. Summary logs of the Masirah Bay Formation in the Jebel Akhdar
Fig. 6.6. Overview of the Masirah Bay Formation exposed in the centre of the Khufai Dome
Fig. 6.7. The Masirah Bay Formation in Wadi Bani Awf, Jebel Akhdar
Fig. 6.8. Lower Huqf Supergroup nomenclature used to date and within the present study
Fig. 6.9. Depositional model for the Masirah Bay Formation in the Huqf area
Fig. 6.10. Sequence stratigraphic development of the Masirah Bay Formation in the Huqf area
Fig. 6.11. Development of Facies D4 of the Masirah Bay Formation in the Khufai Dome
Fig. 6.12. Lateral variation and extent of the Masirah Bay Formation in the Huqf area inferred from surface outcrops
Fig. 6.13. Well-penetrations of the Masirah Bay Formation in the vicinity of the Huqf area
Fig. 6.14. Postulated palaeocoastline and sediment transport directions during the upper part of Member 2 deposition in Masirah Bay time

Chapter 7

Fig. 7.1. Schematic chronostratigraphic representation of the Huqf Supergroup in the Jebel Akhdar and Huqf areas
Fig. 7.2. Abu Mahara and Nafun Group subsidence in the Jebel Akhdar
LIST OF PLATES

Chapter 2

Plate 2.1. Outcrop of granodioritic basement near the village of Al Jobah, north Huqf area
Plate 2.2. Typical exposure of the Halfayn Formation near Al Jobah, north Huqf area
Plate 2.3. Unit H1 of the Halfayn Formation
Plate 2.4. Photomicrograph of welded tuff from Unit H1 of the Halfayn Formation
Plate 2.5. Exposure of Unit H2 of the Halfayn Formation
Plate 2.6a. Outcrop of swaley cross-stratified dolostone at the base of Unit H3 of the Halfayn Formation
Plate 2.6b. Sample of swaley/hummocky cross-stratified dolostone taken from the base of Unit H3 of the Halfayn Formation
Plate 2.7. Angular conglomerate, Unit H3 of the Halfayn Formation
Plate 2.8. Bed of fine-grained dolomite within conglomerate of Unit H3, Halfayn Formation
Plate 2.9. Sample showing carbonate ‘rip-ups’ at the base of the dolomite bed shown in Plate 2.8
Plate 2.10. Cross-stratification within the upper part of Unit H3 of the Halfayn Formation
Plate 2.11. Chert-filled fissures below the top of Unit H3 of the Halfayn formation

Chapter 3

Plate 3.1. The view looking north into the Ghubrah Bowl, which drains through Wadi Mistal
Plate 3.2. The Ghubrah and Ghadir Manqil Formations in the north of Wadi Mistal
Plate 3.3. The Fiq Member behind the village of Dabu’t, north Wadi Sahtan
Plate 3.4. Well-cleaved diamictite of the Ghubrah Formation
Plate 3.5. Large granite clast within diamictite of the Ghubrah Formation
Plate 3.6. Striated clast within diamictite of the Ghubrah Formation
Plate 3.7. Deformed carbonate bed within diamictite of the Ghubrah Formation
Plate 3.8. Ash bed within diamictite of the Ghubrah Formation
Plate 3.9. Ash bed within the Ghubrah Formation, 6km to the north-west of Plate 3.8
Plate 3.10. Pillowed basalt of the Saqlah Member, Wadi Mu’aydin
Plate 3.11. Vesicled basalt of the Saqlah Member in Wadi Mistal
Plate 3.12. Clast-poor distal glaciomarine diamictite of the Fiq Member, Wadi Sahtan
Plate 3.13. Granite dropstone in laminated siltstone of the Fiq Member, Wadi Hajir
Plate 3.14. Proximal glaciomarine diamictite rich in granite clasts, Fiq Member, Wadi Hajir
Plate 3.15. Typical facetted clast from diamictite of the Fiq member, Wadi Mistal
Plate 3.16a. Striated clast from diamictite of the Fiq Member, Wadi Sahtan
Plate 3.16b. Striated clast from diamictite of the Fiq Member, Wadi Mistal
Plate 3.17. Aggregate of coarse-grained and pebbly material within diamictite of the Fiq Member, Wadi Sahtan
Plate 3.18. Stratified diamictite, Fiq Member, Wadi Hajir
Plate 3.19. Conglomerate within proximal glaciomarine facies association of the Fiq Member
Plate 3.20. Channelised sandstone within proximal glaciomarine facies association, Fiq Member
Plate 3.21a. Rippled sandstone of the proximal glaciomarine facies association, Fiq Member
Plate 3.21b. Rippled sandstone of the proximal glaciomarine facies association, Fiq Member
Plate 3.22. Dropstones in laminated deposits of the Proximal glaciomarine facies association, Fiq Member
Plate 3.23. Conglomerate of non-glacial sediment gravity flow facies association, Fiq Member
Plate 3.24. Erosively based conglomerate of the Fiq Member, Wadi Hajir
Plate 3.25. Pebbly sandstone of the Fiq Member, Wadi Hajir
Plate 3.26. Soft sediment deformation at contact between pebbly sandstone and shale, Fiq Member
Plate 3.27. Massive turbiditic sandstones of the Fiq Member, Wadi Sahtan
Plate 3.28. Flute mark at the base of turbiditic sandstone bed, Fiq Member, Wadi Sahtan
Plate 3.29. Channel-filling graded sandstones of the Fiq Member, Wadi Hajir
Plate 3.30. Sharp, granular base to graded sandstone of the Fiq Member, Wadi Hajir
Plate 3.31. Rippled bedding surface at top of turbiditic sandstone, Fiq Member, Wadi Sahtan
Plate 3.32. Rippled sandstone beds interbedded with siltstones, Fiq Member
Plate 3.33. Sharply based sandstone with rippled top, Fiq Member, Wadi Mistal
Plate 3.34. Coarser rippled horizons within dominantly planar siltstones, Fiq Member
Plate 3.35. Climbing ripples within sandstone, Fiq Member, Wadi Hajir
Plate 3.36. Symmetrical ripple profiles, Fiq Member, Wadi Hajir
Plate 3.37. Siltstone beds, Fiq Member, Wadi Hajir
Plate 3.38. Low-angle scours in laminated siltstone, Fiq Member, Wadi Sahtan
Plate 3.39. Laterally persistent laminations within siltstones, Wadi Sahtan
Plate 3.40. Low-angle cross-stratification in sandstone, Fiq Member, Wadi Bani Awf
Plate 3.41. Erosive base to coarse-grained volcaniclastic sandstone, Fiq Member
Plate 3.42. Chaotic and convoluted bedding in volcaniclastic sandstone facies, Fiq Member
Plate 3.43. Carbonate beds below volcaniclastic sandstone, Fiq Member, Wadi Mistal
Plate 3.44. Rippled carbonate bed overlain by siltstone, Fiq Member, Wadi Mistal
Plate 3.45. Photomicrograph of carbonate bed at the base of the Ghadir Manqil Formation
Plate 3.46a. Photomicrograph of carbonate bed at the base of the Ghadir Manqil Formation
Plate 3.46b. Photomicrograph of spherical structures within carbonate bed at the base of the Ghadir Manqil Formation
Plate 3.47. Carbonate bed at the base of Unit F6b, Fiq Member, Wadi Sahtan
Plate 3.48. Carbonate bed at the base of Unit F6b, Fiq Member, Wadi Hajir
Plate 3.49. Photomicrograph of carbonate bed at the base of Unit F6b, Fiq Member, Wadi Sahtan
Plate 3.50. Photomicrograph of carbonate bed at the base of Unit F6b, Fiq Member, Wadi Hajir
Plate 3.51. Conglomerate of the shallow marine facies association, Fiq Member, Wadi Sahtan
Plate 3.52. Symmetrical ripple profiles, Fiq Member, Wadi Sahtan
Plate 3.53. Straight-crested wave-ripples, Fiq Member, Wadi Sahtan
Plate 3.54. Rippled heterolithic beds interbedded with massive sandstones, Fiq Member
**Chapter 4**

Plate 4.1. Clastic dyke splitting granite clast, Mirbat Sandstone, south Oman

**Chapter 5**

Plate 5.1. The Hadash Formation at locality Hadash 1, Wadi Mistal
Plate 5.2. The Hadash Formation at locality Hajir 1, Wadi Hajir
Plate 5.3. The Hadash Formation in Wadi Bani Jabir
Plate 5.4. The upper part of the Hadash Formation in the west of Wadi Sahtan
Plate 5.5. Facies E1 at the base of the Hadash Formation, Hadash 1
Plate 5.6. Photomicrograph of Facies E1
Plate 5.7. Photomicrograph of locally developed ‘clotted’ texture in Facies E1
Plate 5.8. Facies E2 at locality Hadash 2
Plate 5.9. Facies E3 of the Hadash Formation at locality Hadash 1
Plate 5.10. Pyrite crystals within Facies E3 of the Hadash Formation, Wadi Mu’aydin
Plate 5.11. Photomicrograph of Sample HSH54, Facies E3, Hadash 1
Plate 5.12. Cathodoluminescence photomicrograph of Sample HSH54, Facies E3, Hadash 1
Plate 5.13. Facies E4 of the Hadash Formation at locality Hadash 2
Plate 5.14. Microbial roll-up structure within Facies E4, locality Hadash 2
Plate 5.15. Sample showing microbial roll-up structures within Facies E4
Plate 5.16. Photomicrograph of Sample HSH63, Facies E4, Hadash 1
Plate 5.17. Cathodoluminescence photomicrograph of Sample HSH63, Facies E4, Hadash 1
Plate 5.18. Siltstone and graded sandstone of Facies E5, Hadash Formation, Wadi Bani Jabir
Plate 5.19. Interbedded siltstone and dolomite of Facies W1 at the base of the Hadash Formation, locality Hajir 1
Plate 5.20. Facies W2 of the Hadash Formation at locality Hajir 2
Plate 5.21. Depressions filled with mudstone on the uppermost bedding surface of Facies W2, Hajir 2
Plate 5.22. Carbonate ‘stringers’ interbedded with grey siltstone, Facies W3, Hajir 1
Plate 5.23. Facies W3 of the Hadash Formation at locality Hajir 2
Plate 5.24a. Photomicrograph of Sample HAJ66, Facies W3, Hajir 1
Plate 5.24b. Cathodoluminescence photomicrograph of Sample HAJ66, Facies W3, Hajir 1
Plate 5.25. Outcrop of dolomite unconformably overlying the Halfayn formation near the village of Al Jobah, north Huqf area
Plate 5.26. Base of the dolomite at Al Jobah
Plate 5.27. Photomicrograph of dolomite at Al Jobah

Chapter 6

Plate 6.1. Outcrop of Facies A1, Masirah Bay Formation, in the centre of the Khufai Dome
Plate 6.2. Wave-ripples of Facies A2 in the centre of the Khufai Dome
Plate 6.3. Swaley cross-stratification of Facies B1, Masirah Bay Formation, Khufai Dome
Plate 6.4. Sigmoidally cross-stratified sandstone, Facies B2, Khufai Dome
Plate 6.5. Cross-stratified sandstone beds of Facies B3, Khufai Dome
Plate 6.6. Outcrop of trough cross-stratified sandstone, Facies B4, Khufai Dome
Plate 6.7. Erosively-based trough cross-set of Facies B4, Khufai Dome
Plate 6.8. Trough cross-sets of Facies B4, Masirah Bay Formation, Khufai Dome
Plate 6.9. Large-scale planar cross-stratified sandstone, Facies B5, Al Harf
Plate 6.10. Granular bed of Facies B6, Khufai Dome
Plate 6.11. Shales of Facies C1, Masirah Bay Formation, Al Harf
Plate 6.12. Sandstone lenses within siltstone, Facies C2, Khufai Dome
Plate 6.13. Locally developed cross-stratification within Facies D1, Al Harf
Plate 6.14. Spindle-shaped structures resembling subaqueous shrinkage cracks, Facies D1, Khufai Dome
Plate 6.15. Spindle-shaped structures, Facies D1, Khufai Dome
Plate 6.16. Outcrop of the upper part of the Masirah Bay Formation in the west of the Khufai Dome
Plate 6.17. Trough cross-sets of Facies D4a, Masirah Bay Formation, Khufai Dome
Plate 6.18. Ridge-forming trough cross-sets of Facies D4b, Khufai Dome
Plate 6.19. Bi-directional cross-stratification, Facies D5, Mukhaibah Dome
Plate 6.20. Tidal deposits of Facies D6 at the top of Member 2, Khufai Dome
Plate 6.21. Thinly-bedded sandstones of Facies E1, Khufai Dome
Plate 6.22. Low-angle scour within sandstones of Facies E1, Khufai Dome
Plate 6.23. Series of sub-parallel grooves and ridges on bedding surfaces of sandstones of Facies E1, Khufai Dome
Plate 6.24. Series of sub-parallel grooves and ridges on bedding surfaces of sandstones of Facies E1, Khufai Dome
Plate 6.25. ‘Stepped’ appearance to beds of Facies E2 caused by alternation of well-indurated dolomite and more friable siltstone, Khufai Dome
Plate 6.26. Dolomite and siltstone beds of Facies E2 below the Khufai Formation, Khufai Dome
Plate 6.27. The Masirah Bay Formation near the village of Hadash, south Wadi Mistal, Jebel Akhdar
Plate 6.28. Typical nature of Facies 1 of the Masirah Bay Formation in the Jebel Akhdar
Plate 6.29. Dark grey, organic-rich beds of Facies 1, Masirah Bay Formation, Jebel Akhdar
Plate 6.30. Swaley cross-stratification, Facies 3, Masirah Bay Formation, Wadi Bani Jabir
Plate 6.31. Erosively-based cross-set, Facies 3, Masirah Bay Formation, Wadi Bani Jabir

APPENDICES

Appendix A – U-Pb dating techniques and data

1. U-Pb zircon separation and dating techniques
2. U-Pb isotopic data from basement, the Abu Mahara Group, and the Fara Formation, Oman

Appendix B – Clast analysis data

1. Clast analyses: location and average clast size
2. Clast analyses: lithological data

Appendix C – Isotopic and elemental geochemistry techniques and data

1. Methods of sample evaluation
2. Isotopic analysis methodology
3. Elemental analysis
4. Calculation of correlation coefficient (r)
5. Ghadir Manqil Formation, stable isotope data
6. Hadash Formation, stable isotope data
7. Masirah Bay Formation, stable isotope data
8. Geochemical data from the Ghadir Manqil, Hadash, and Masirah Bay Formations

Appendix D – Logs of the Abu Mahara Group in the Jebel Akhdar
1. Logs made in Wadi Sahtan
2. Logs made in Wadi Hajir
3. Logs made in Wadi Mistal
4. Logs made in Wadi Bani Awf
5. Logs made in Wadi Bani Kharus
6. Logs made in Wadi Mu’aydin
7. Logs made in Wadi Bani Jabir
8. Key

Appendix E – Logs made of the Masirah Bay Formation in the Huqf area
1. Log AH1 (Al Harf)
2. Log MD1 (Mukhaibah Dome)
3. Log MD2 (Mukhaibah Dome)
4. Log KD1 (Khufai Dome)
5. Log KD2 (Khufai Dome)
6. Log KD3 (Khufai Dome)
7. Key
Chapter 1

Introduction
CHAPTER 1. INTRODUCTION

1.1. Introduction

The primary aims of this project were: 1) to provide specific data on the evolution of sedimentary environments and processes in the Late Proterozoic of Oman, derived from sedimentological field and laboratory study of the surface exposures of the lower part of the Huqf Supergroup; 2) to refine the time frame for Huqf Supergroup deposition using chemo- and chronostratigraphic methods; and 3) to integrate these data into a regional context to help define basin and biosphere evolution at this time in Earth history. The impetus for this work resulted from the fact that Huqf Supergroup sediments are components in proven hydrocarbon plays (Lake, 1996; Nederlof and al Ruwehy, 1996). As Huqf source sediments are deeply buried beneath the Oman salt basins, they lie beyond the resolution of seismic and core analysis. Petroleum Development Oman, comprising the Omani Government (as major shareholder) and the Shell Oil Company, therefore funded this project to gain a better understanding of the lower Huqf Supergroup (Abu Mahara Group and lower Nafun Group) from surface exposures that could be integrated into their petroleum exploration program.

1.2. Regional setting

Rocks of the Huqf Supergroup are excellently exposed in the core of the Jebel Akhdar in north Oman, in the Huqf area of east-central Oman and near Mirbat in the south of Oman (Fig. 1.1). Further exposures also occur in the Saih Hatat region in the north-east of Oman, where the rocks have undergone complex tectonic deformation (eg. Glennie et al., 1974; Le Metour et al., 1986; Rabu et al., 1993). Although brief excursions were made to both the Mirbat and Saih Hatat regions, the present study has concentrated on the outcrops of the Jebel Akhdar and Huqf areas.

The Jebel Akhdar forms the central part of the Oman Mountains, which occur as an arcuate belt lying parallel to the Gulf of Oman. This belt is 30-130km wide and extends 700km in a NW-SE direction (Figs. 1.2; 1.3). Within the Jebel Akhdar, the Neoproterozoic-Cambrian strata are exposed in a series of erosional windows. From west to east these are
the Sahtan, Kharus, Mistal, Saiq and Jebel Nakhl windows (Fig. 1.3). The pre-Permian rocks of the Jebel Akhdar have been affected by at least two major tectonic events: a late Palaeozoic deformation and a late Cretaceous thrusting which emplaced both the Semail ophiolite and the Tethyan thrust sheets onto the Arabian continental margin (Mann and Hanna, 1990). The latter of these was responsible for the uplift of the present day mountains. The deformation and metamorphism both increase in intensity towards the east and north-east of the Oman Mountains.

The Neoproterozoic rocks exposed in the Huqf area are part of the Huqf-Haushi Uplift. This is an inlier of Neoproterozoic strata which extends approximately 180km north-northeastwards of Duqm, before being covered by the Wahiba Sands in the north-east (Fig. 1.2). Its maximum outcrop width is 40km. The inlier is surrounded by onlapping Cambrian to Cretaceous sequences that rest unconformably on the eroded Neoproterozoic to Cambrian strata. To the west is the subsurface Ghaba Salt Basin, whose subsidence was coeval with the intermittent uplift of the Neoproterozoic-Cambrian strata in the Huqf-Haushi Uplift (Ries and Shackleton, 1990). These surface and subsurface sediments are relatively undeformed.

1.3. Tectono-stratigraphic history

The Sultanate of Oman is currently situated on the south-eastern margin of the Arabian Plate (Fig. 1.4). It is bounded to the south by the Gulf of Aden Spreading Zone, to the east by the Masirah Transform Fault and the Owen Fracture Zone, and to the north by the Zagros-Makram convergent plate margin.

1.3.1. Arabian tectonics in the Neoproterozoic

A number of workers (eg. Stoeser and Camp, 1985; Husseini, 1989; Husseini and Husseini, 1990; Loosveld et al., 1995; 1996) have suggested similar tectonic models for the evolution of the Arabian and adjoining plates in the Neoproterozoic. Initially, a complex collage of oceanic terranes and crustal fragments are thought to have amalgamated through accretionary tectonics to form the Arabian-Nubian shield. Husseini (1989) recognised five distinct terranes in western Saudi Arabia that were amalgamated
during Neoproterozoic time. These are the Asir, Hijaz, Midyan, Afif, and Ar Rayn terranes. The Asir, Hijaz, and Midyan terranes are interpreted as island-arc terranes that fused together along ophiolitic sutures at c. 715 Ma, prior to the accretion of the Afif gneissic terrane along the Nabitah suture between 680 and 640 Ma (Stoeser and Camp, 1985). Finally, the Al Amar island arc and the Ar Rayn gneissic terrane were added along the Al Amar-Idsas suture during the Idsas orogeny (c. 630 Ma, Figs. 1.4, 1.5; eg. Stacey et al., 1984). These accretionary processes closed oceans and merged the terranes of the Middle East with the north-eastern edge of Gondwanaland, forming an extensive area of mountainous stature. The accretion of these terranes was immediately followed by the development of Najd faulting, a major period of dominantly sinistral strike-slip deformation with up to 300km displacement (eg. Stoeser and Camp, 1985; Stern, 1985; Husseini, 1989; Husseini and Husseini, 1990). It has been suggested that Najd shear zones may have extended for long distances under the Rub al Khali towards Oman (Fig. 1.4; eg. Loosveld et al., 1995).

Although the inferred latitude of Arabia at the end of the Neoproterozoic has been taken to be as high as 45°S by some workers (eg. Dalziel, 1997), it seems more likely it was situated close to equatorial latitudes (Beydoun, 1991; McKerrow et al., 1992; Torsvik et al., 1996; Kempf et al., 2000). It is generally accepted that Arabia was one of a number of plates that formed part of the ‘northern’ margin of Gondwana bordering a Palaeo-Tethys ocean (Fig. 1.6; eg. Beydoun, 1991; McKerrow et al., 1992; Torsvik et al., 1996; Dalziel, 1997). Beydoun (1991) suggested that the Arabia, Turkey, central and north-west Iran, Afghan, India and other plates on the northern edge of Gondwana formed a passive margin during the late Neoproterozoic. However, other authors (eg. Shackleton, 1996; Dalziel, 1997; Trompette, 1997) have suggested that at the end of the Neoproterozoic, a collisional mountain range linked to the last stages of collision between East and West Gondwana may have extended for up to 7500 km from Arabia to the Pacific margin of Antarctica (Figs. 1.6; 1.7).

1.3.2. Tectonic framework for the Neoproterozoic of Oman

Crystalline basement is exposed in the south of Oman in the Mirbat region and on the Al Hallaniyah Islands, in east-central Oman in the north of the Huqf area, and near Sur (Jebel
Ja’alan) further to the north. The crystalline basement is composed of metamorphic and igneous rocks that have been intruded by dolerites, granodiorites and granites. U-Pb zircon dates that have previously been obtained from the basement in Oman vary from c. 824 Ma to c. 780 Ma (Table 1.1). Rb/Sr dates have produced some significantly younger ages—most notably a date of 556±11 Ma (Table 1.1; Kramers and Frei, 1992). However, Rb/Sr is prone to resetting (eg. Evans, 2000), and this age may not be reliable.

The Huqf Supergroup contains the Abu Mahara, Nafun and Ara Groups and comprises the oldest exposed sediments in Oman. Rocks of the Mirbat Sandstone unconformably overlie crystalline basement in the south of Oman. These deposits are glacial in origin (Kellerhals, 1998), and are thought to be equivalent to the Abu Mahara Group described from the subsurface of Oman and in the Jebel Akhdar (Fig. 1.8; Bell, 1993a). An ash-bed from within the Abu Mahara Group in the Jebel Akhdar has yielded a U-Pb zircon date of 723 +16/-10 Ma (Brasier et al., 2000; McCarron, 2000). Large lateral thickness variations observed in the Mirbat Sandstone (thickness varying from 600m to 0m over a lateral distance of 8km), and in the Ghadir Manqil Formation in the subsurface (Bell, 1993a) suggests the Abu Mahara Group could represent the infill of a series of half-graben or graben structures. This interpretation is supported by the presence of volcanics and the abundance of volcaniclastic sediments within the Abu Mahara Group of the Jebel Akhdar.

In the Huqf area, crystalline basement is overlain by the volcanics and volcaniclastics of the Halfayn Formation (Chapter 2). These have previously been correlated with the volcanics of the Abu Mahara Group in the Jebel Akhdar (Fig. 1.8; Bell, 1993a). New dates produced as part of this study, however, suggest that the Halfayn Formation pre-dates the volcanics of the Jebel Akhdar (Chapter 2). No Abu Mahara glacial sediments are recorded in the Huqf area, suggesting it may have been a high for much of this time.

The top of the Huqf Supergroup is constrained by a U-Pb zircon date from the Fara Formation overlying the Nafun Group in the Jebel Akhdar. This date of 544.5 ±3.3 Ma (Brasier et al., 2000) suggests that both the Abu Mahara and Nafun Groups are entirely Proterozoic in age.

The tectonic evolution of the Huqf Supergroup has been the focus of debate in recent years, and a number of different tectonic frameworks have been proposed. In the mid-
1990s, a number of models linking Huqf Supergroup basins to the Najd tectonic event to the west were suggested (e.g. Loosveld et al., 1995; 1996). Large lateral variations in thickness and the presence of volcanics within the Abu Mahara Group (Bell, 1993a; Loosveld et al., 1996) at the base of the Huqf Supergroup and within the Ara Group at the top were used to infer rifting at both of these times. This led to the development of two different scenarios. In the first of these, intracratonic rifting during Abu Mahara time was followed by thermal subsidence and tectonic quiescence during the Nafun. Rifting was renewed during Ara time, and this was linked to the Najd tectonic event to the west. In the second model, one rifting event related to sinistral movements along the Najd was envisaged, lasting for the duration of Huqf Supergroup time (Loosveld et al., 1996). This second model requires the Abu Mahara Group to be contemporaneous with the Najd event. However, the date from the Jebel Akhdar of 723 +16/-10 Ma (Brasier et al., 2000) suggests that the Abu Mahara rifting actually represents an earlier event.

Although rifting in Abu Mahara time is generally accepted (e.g. Bell, 1993a; Rabu et al., 1993; Loosveld, 1996), an extensional tectonic setting for the top of the Huqf Supergroup is more questionable. Deformation observed or inferred in wells in the west of Oman that is lacking in more easterly wells has been used to infer the occurrence of compressional tectonics in the late Neoproterozoic of Oman (e.g. Kapellos et al., 1992; Loosveld and Bell, 1994). The boundary between the deformed western area, and the largely undeformed east has been termed the ‘Western Deformation Front’ in Oman (Kapellos et al., 1992). This has been taken as evidence to suggest the arrival of a major thrust belt from the west in Ara time, providing an alternative to rifting to explain the observed compartmentalisation of the basin and voluminous volcanics. It is possible that flexure associated with loading to the west prior to the arrival of such a thrust belt may account for some (or all) of the subsidence that occurred during Nafun time rather than/as well as post-rift sagging. Compressional tectonics in Oman during Ara time are consistent with the collisional belt suggested to extend from Arabia to Antarctica at the end of the Neoproterozoic (Fig. 1.7; eg. Dalziel, 1997).
1.4. Huqf Supergroup nomenclature

Carter (1852) and Lees (1928) published early reports mentioning the ‘micaceous sandstone and argillaceous beds of Dhofar’ and ‘Mirbat sandstone’ overlying the crystalline basement in the south of Oman. The outcrops of the Huqf Supergroup in the Huqf area were first studied by Iraq Petroleum Co. geologists who distinguished four rock units (in Gorin et al., 1982). This fourfold subdivision is mentioned by Henson and Elliot (1958), Morton (1959), and Beydoun (1960, 1964) who used the terminology of First and Second Clastic Groups (Abu Mahara Group and Shuram Formation respectively) and of First and Second Dolomite Groups (Khufai and Buah Formation respectively). Further fieldwork was conducted in the Huqf area by Kassler (1965) who defined the Abu Mahara, Khufai, Shuram, and Buah Formations. In the Jebel Akhdar, Kapp and Llewellyn (1965) defined the Mistal, Hajir, Mi’aidin, and Kharus Formations, which were first correlated with the Huqf succession by Tschopp (1967). Bell (1993a) presented a new stratigraphy, which incorporated the formations of the Huqf and Jebel Akhdar regions into the Huqf Supergroup.

These previous researchers have, along with Petroleum Development Oman geologists, used different nomenclatures and definitions for the formations of the Huqf Supergroup. To avoid confusion, a uniformity of terms will be used in the present study by applying widely used names to all coeval formations. A summary of the terms used in this study is presented in Fig. 1.9. More detailed comparisons between present nomenclature and that used by previous workers are shown in Fig. 3.7 for the Abu Mahara Group and Fig. 6.8 for the Hadash and Masirah Bay Formations.

In this study the Abu Mahara Group is taken to include the Halfayn Formation in the Huqf area, and the Jabir, Ghubrah and Ghadir Manqil Formations in the Jebel Akhdar. The Hadash and Masirah Bay Formations have previously been placed within the Abu Mahara Group (eg. Bell 1993a). However, the base of the Hadash Formation marks a major change in both depositional style and environment, and the Hadash and Masirah Bay Formations are more akin to the carbonate-siliciclastic cycles of the overlying Khufai, Shuram and Buah Formations than the underlying glacial deposits of the Ghadir Manqil Formation. For this reason, the Hadash and Masirah Bay Formations are both placed
within the Nafun Group in this thesis. The boundary between the Ghadir Manqil and Hadash Formations therefore also forms the boundary between the Abu Mahara and Nafun Groups. This slight revision of the nomenclature follows recommendations suggested by McCarron (2000). The nomenclature used for each formation is discussed in more detail within each chapter.

1.5. Lower Huqf Supergroup stratigraphy and facies associations

Published work on the Huqf Supergroup has covered the stratigraphy, structure, and tectonics of both the Jebel Akhdar (eg. Glennie et al., 1974; Beurrier et al., 1986; Rabu et al., 1986; 1993) and Huqf regions (eg. Gorin et al., 1982; Gass et al., 1990; Wright et al., 1990; Dubreuilh et al., 1992). Numerous internal Petroleum Development Oman reports have also been written on the stratigraphy and tectonics of the Huqf Supergroup (eg. Kapp and Llewellyn, 1965; Kassler, 1965; Bell, 1993a; Loosveld et al., 1995; Pilcher and Buckley, 1995).

A recent thesis produced by McCarron (2000) dealt with the sedimentology and chemostratigraphy of the Khufai, Shuram, and Buah Formations of the upper Nafun Group. The present study is concerned with the sedimentology, chronostratigraphy, and chemostratigraphy of the Huqf Supergroup lying below the Khufai Formation. This has included work on the Abu Mahara Group (Halfayn, Ghubrah and Ghadir Manqil Formations) and lower Nafun Group (Hadash and Masirah Bay Formations) in both the Jebel Akhdar and Huqf areas. This is the first study to concentrate on this section of the Oman stratigraphy. Each formation is described in stratigraphical order (Chapters 2, 3, 5 and 6), and where possible, subdivided into facies and facies associations based on similar lithological characteristics and environmental interpretations (Table 1.2).

1.6. Thesis outline

Chapter 2 deals with the volcanics and volcaniclastics of the Halfayn Formation. This is only exposed in the Huqf area and comprises the oldest Abu Mahara Group deposits. These are briefly described, before the significance of new U-Pb age dates are discussed.
The Abu Mahara Group in the Jebel Akhdar is described and discussed in Chapter 3. This includes the glacial Ghubrah and Ghadir Manqil Formations, as well as the possibly older Jabir Formation. The age of these deposits and some of their global implications are discussed in Chapter 4.

Chapters 5 and 6 deal with the Hadash and Masirah Bay Formations respectively, which form the lower part of the Nafun Group. The Hadash Formation is a typical Neoproterozoic cap carbonate, and was studied in detail isotopically as well as lithologically. The Masirah Bay Formation is a siliciclastic formation that records shallow marine deposition in the Huqf area and deeper-water sedimentation in the Jebel Akhdar.

The final chapter (Chapter 7) summarises the findings of this work, before briefly analysing the subsidence/basin history of the Huqf Supergroup.

1.7. Project development

The PhD project undertaken by Gretta McCarron originally encompassed the entire Huqf Supergroup. However, after her first field season, this was deemed too ambitious because of the vast scale and thickness of the exposures and the inhospitable terrain (McCarron, 2000). McCarron (2000) therefore concentrated on the Khufai, Shuram, and Buah Formations of the Nafun Group, and the current study was set up to analyse the lower parts of the Huqf Supergroup.

The first field season in Oman (2/1/98-10/3/98) began with a two week workshop on the Huqf Supergroup, followed by field excursions to the Jebel Akhdar and Huqf areas with Shell personnel, geologists and academics. After this initial introduction to the Huqf Supergroup, the rest of the field season involved reconnaissance studies of the outcrops of the Huqf and Jebel Akhdar regions. Summary logs of the Ghadir Manqil and Masirah Bay Formations were compiled at a number of localities. Samples for radiometric age dating were collected from the granitic basement and the overlying Halfayn Formation at Al Jobah in the Huqf, and from suspected ash-beds in the Jebel Akhdar. The Hadash Formation was also sampled for stable isotopic analysis during this field season.
The first part of the second field season (22/11/98-15/12/98) concentrated on compiling a series of logs through the most complete exposures of the Masirah Bay Formation in the Huqf area, with further reconnaissance of the Ghadir Manqil Formation in the Jebel Akhdar. During the second part of the field season (2/1/99-23/2/99), work was concentrated in the Jebel Akhdar. Detailed logs of the Ghadir Manqil Formation were made in Wadi Mistal and Wadi Sahtan, and further samples from all the formations in the Jebel Akhdar were taken for stable isotope analysis. It was during this field season that two post-doctoral workers were introduced to begin work on the Buah and Fara Formations.

The third field season (8/1/00-25/2/00) began with a short excursion to the south of Oman to compare the sediments of the Mirbat Sandstone to those of the Huqf and Jebel Akhdar regions. After this, work concentrated on detailed logging of the Ghadir Manqil Formation in the Jebel Akhdar, with specific aims to assess lateral variation within it. This involved clast analyses of diamicrite units within the Ghubrah and Ghadir Manqil Formations. A short trip was also made to the Huqf area to investigate the Halfayn Formation at Al Jobah in more detail.

A final trip to Oman was made in January 2001. This was to present at the conference on the geology of Oman, attend wrap-up sessions within Petroleum Development Oman, and to lead field excursions to both the Jebel Akhdar and Huqf areas.
Chapter 2

The Halfayn Formation
CHAPTER 2. THE HALFAYN FORMATION

2.1. Introduction

The Halfayn Formation is exposed in a series of low-lying outcrops in the north-east of the Huqf area, ~15km north-west of the village of Al Jobah. The Halfayn Formation was first identified as a lithostratigraphic unit by Le Metour et al. (1992). It unconformably overlies granodioritic basement at one locality, and is in turn unconformably overlain by a dolomitic unit that passes up into the siliciclastics of the Masirah Bay Formation.

The Halfayn Formation is composed of volcanic and volcaniclastic rocks that form the lowermost part of the Abu Mahara Group in the Huqf area. These have previously been correlated with the volcanics of the Ghadir Manqil Formation (Fig. 1.8; eg. Bell, 1993a; Pilcher and Buckley, 1995). Recent U-Pb dates from zircons within volcanic horizons of the Halfayn Formation produced as part of this study, however, suggest that it represents an earlier phase of volcanism.

The aims of this chapter are: 1) To describe and interpret the lithologies present within the Halfayn Formation; 2) To discuss the age of the Halfayn volcanics.

2.2. Methods

The Halfayn Formation is only exposed near the village of Al Jobah. Exposure here is relatively poor and the nature of the outcrop (Plates 2.1, 2.2) makes accurate measurement of thickness and logging difficult. Short logs were made at a number of localities, however, to try to identify the different lithologies present. Representative samples were taken from the different lithologies for petrographic study. Further samples were taken from the volcanic horizons so that zircons within them could be dated using U-Pb isochrons. For dating methods see Appendix A. The information collected from this work has shown that rhyolitic ignimbrites at the base of the Formation pass up into pyroclastic flow deposits, many of which were reworked in a shallow marine or fluvial environment. U-Pb dating indicates that the Halfayn Formation pre-dates the Ghadir Manqil Formation and that a major unconformity exists at its top.
2.3. Previous work

As the Halfayn Formation was only identified recently (Le Metour et al., 1992), little previous work on it exists. Dubreuilh et al. (1992) first described the Halfayn Formation, finding it to be between 35 and 38 m thick and consisting of deposits of lava and ignimbrite-type rhyolitic tuffs, volcanoclastic breccia and conglomerate with carbonate matrix (Fig. 2.1). At the top of the Halfayn Formation a continuous horizon of brown silcrete was identified. This was thought to be the result of emergence and continental weathering and evidence of a disconformity. Dubreuilh et al. (1992) suggested that the Halfayn Formation was probably deposited in a shallow-marine environment.

Bell (1993a) correlated the volcanics of the Halfayn Formation with the Ghadir Manqil Formation, and this idea was adopted by Pilcher and Buckley (1995) in their report on the outcrops of the Huqf area (Fig. 1.8). Pilcher and Buckley (1995) identified four lithofacies within the Halfayn Formation (red banded welded tuff, carbonate replaced intermediate volcanoclastic agglomerate, lime mudstone, and carbonate/hybrid breccia), and again suggested deposition in a shallow marine environment.

Some previous work has also been done on the dating of possible Halfayn Formation equivalent volcanics. A corehole (CH-17) drilled through the Masirah Bay Formation in the centre of the Khufai Dome bottomed in trachytic volcanics that lay beneath a dolomitic horizon (Fig. 6.2). These possible equivalents of the Halfayn Formation yielded a K-Ar age of 654 ± 12 Ma (Gorine et al., 1982). The validity of this date has, however, since been questioned (Bell, 1993a), due to the susceptibility of K-Ar dates to later resetting. Further geochronological studies were made by Kramers and Frei (1992) on three samples of unweathered rhyolite from the Halfayn Formation near Al Jobah. These produced a Rb/Sr date of 562 ± 42 Ma. A Rb/Sr date of 556 ± 11 Ma from possibly coeval granites and rhyolites in the Mirbat region of south Oman has also been obtained (Kramers and Frei, 1992). A combined Rb/Sr isochron for these two units produces a date of 554 ± 10 Ma (Kramers, 1994).

Dates that have previously been obtained from the crystalline basement of Oman are shown in Table 1.1.
2.4. Subdivision of the Halfayn Formation

There is no complete exposure of the Halfayn Formation in the Al Jobah area. It was therefore logged at three separate localities (AJ1, AJ2, and AJ3), at which different levels of the Halfayn Formation were exposed (Fig. 2.2). From these logs an initial, albeit somewhat crude, subdivision of the Halfayn Formation into three main units (H1, H2, and H3) has been made (Fig. 2.3). These units will be described in chronological order. Note that much of the work in Section 2.5 was done in conjunction with Chris Nicholas.

2.5. Facies analysis of the Halfayn Formation

2.5.1. Unit H1 – Green rhyolitic ignimbrite

The lowest unit described in this study is composed of green rhyolitic tuff (Plate 2.3). This is not observed directly overlying the granodioritic basement at any point, but its base lies at most 3m above the top of the granodiorite. Dubreuilh et al. (1992) reported a red rhyolitic ignimbrite directly overlying basement below the green rhyolite, but this was not observed in this study.

The green rhyolitic ignimbrite is ~ 8-9m thick. Its base is composed of coarse and fine green tuff that probably represents an unwelded ignimbrite. This passes up into vesicular green tuff that has a welded texture (Plate 2.4). The top of this unit is composed of fine flow-laminated green tuff. Both of these upper two lithologies were sampled for zircons and have been dated using U-Pb isochrons (Samples AJD1 and AJD3, Fig. 2.3).

2.5.2. Unit H2 – Unwelded distal pyroclastic flow deposits

This unit is 2.5m thick and lies above Unit H1, although the contact between them was not observed. The basal 90cm of this unit is composed of red and white crystal sand containing lithic clasts including mud/siltstone lithologies (Plate 2.5). These clasts are up to 8cm in size in the basal 20cm, but decrease in size upwards to a maximum of 0.5cm at the top. The basal bed is overlain by red and white crystal sand containing no coarse clasts. This contains finer bands within it and shows a general fining-up trend over its
1.4m thickness. Laminations appear in the uppermost 40cm of this bed. These beds probably represent unwelded distal pyroclastic flow deposits.

2.5.3. Unit H3 – Shallow marine to fluvial pyroclastic sediments

At Locality AJ2, Unit H2 passes up into hummocky cross-stratified dolostone (Plate 2.6), probably representing storm influenced deposits, which forms the basal 50cm of Unit H3. Above this, the lower part of Unit H3 is composed of angular conglomerate of ash/mud lithics and carbonate rip-ups in a carbonate-rich matrix (Plate 2.7). The clasts vary from few mm to 10cm in size and are often concentrated into laterally discontinuous bands. Indeed, the unit as a whole contains a lot of lateral variation, often passing into much finer-grained material and back to coarse again on a metre-scale. Large-scale cross-stratification on a metre scale is locally developed in the lower part of Unit H3. Towards the top of Unit H3 a fine, red-brown dolostone with rip-up fragments resembling fiamme-like structures at its base occurs (Plates 2.8, 2.9). This has locally been injected into the underlying sediments. It is overlain by more of the angular conglomerate containing ash/mud lithics and carbonate rip-ups. This is in general finer-grained than the conglomerate at the base of the unit and contains well-developed cross-stratification (Plate 2.10). Cross-sets are initially on a metre-scale, but quickly pass up into smaller decimetre scale structures. Chert is locally developed within the conglomerate. The cross-stratification present in Unit H3, the coarse-grained, immature nature of the deposits and the lateral variation suggests deposition of subaerial pyroclastic debris in a fluviatile or shallow marine environment. The presence of carbonate clasts within this unit suggests that carbonate had been deposited prior to the formation of this unit.

The uppermost few cm of Unit H3 are much darker in colour than the underlying conglomerate and are siliceous. Chert is more widely developed than below and brecciation of the top of the conglomerate associated with fissures occurs (Plate 2.11). The silicification and reworking of the lower layers indicate a lack of sedimentation and possible emergence. It is at this level that Dubreuilh et al. (1992) postulated a major unconformity.

Sample AJD4 from Unit H3 was taken for U-Pb dating (Fig. 2.3).
2.6. Halfayn Formation synthesis

The lowermost rhyolitic volcanics of the Halfayn Formation unconformably overlie the eroded crystalline basement. They pass up into pyroclastic sediments that were deposited in a shallow marine to fluvial environment. Reworking of some of the lower volcanics into these deposits occurred. At the top of these pyroclastic deposits a major unconformity, recognised by a brown cherty crust and fissures occurs. This probably represents a major time gap before deposition of the dolomitic unit underlying the Masirah Bay Formation commenced.

2.7. Age of the Halfayn Formation

The presence of volcanic horizons containing zircons within the Halfayn Formation allows it to be accurately dated using the U-Pb method. High precision U-Pb dating techniques are generally considered to produce the most reliable ages and have demonstrated inconsistencies with previous Rb-Sr and K-Ar analyses (eg. Evans, 2000). One reason for this is that Rb-Sr and K-Ar have a lower closure temperature in mineral phases than U-Pb (eg. Ghent et al., 1988). This means that any subsequent heating of the rock would be far more likely to reset Rb-Sr and K-Ar dates than U-Pb dates, and lead to a younger age being obtained (eg. Fowler, 1990). It is possible, however, that zircons dated by U-Pb isochrons may be inherited and their crystallisation age may not necessarily be the age of the rock they are now present in. The lithology the zircons occur within therefore has to be considered when an interpretation of the dates is made.

Three horizons within the Halfayn Formation and two samples from the granodioritic basement have been dated using the U-Pb zircon method as part of this study. Their stratigraphic positions are shown in Fig. 2.3.

The crystalline basement yielded U-Pb dates of c. 825 Ma (Sample AJD-2; Fig. 2.4) and c. 822 Ma (Sample AJD-5; Fig. 2.5) (for methods and full data set, see Appendix A). In Unit H1 of the Halfayn Formation, the welded rhyolitic ignimbrite (sample AJD-3) yielded an age of c. 802 Ma (Fig. 2.6) and the overlying unwelded ignimbrite (sample AJD-1) an age
of c. 830 Ma (Fig. 2.7). Zircons from AJD-4, sampled from both the clasts and the matrix of the conglomerate of Unit H3, produced a U-Pb age of c. 825 Ma (Fig. 2.8).

The U-Pb ages from the granodioritic basement at Al Jobah are similar to previous U-Pb and Pb-Pb dates obtained from crystalline basement elsewhere in Oman. These vary from 834 to 780 Ma (Table 1.1). The Rb-Sr dates from basement generally give younger dates than this, but apart from the one date of 556 ± 11 Ma from Mirbat (Kramers and Frei, 1992), they are all still older than 700 Ma.

The U-Pb dates from the Halfayn Formation overlying crystalline basement (802-830 Ma) are, however, all significantly older than any dates previously obtained from it (654 ± 12 Ma, Gorin et al., 1982; 562 ± 42 Ma, Kramers and Frei, 1992). These younger dates come from Rb-Sr or K-Ar dating. This suggests that the younger Rb-Sr and K-Ar ages may have been reset and/or that the zircons dated as part of this study may have been reworked and are inherited. The fact that the zircons taken from higher in the Halfayn Formation produce older dates than those lower down in the stratigraphy (Fig. 2.3) immediately suggests that inheritance must be a factor. This is undoubtedly the case with samples AJD-1 and AJD-4. Both of these samples yield U-Pb basement ages suggesting that the zircons crystallised at the same time as the basement and were later incorporated into the Halfayn Formation. This is not surprising considering that both samples come from lithologies (unwelded ignimbrite (tuff) and angular pyroclastic conglomerate) that would be expected to contain reworked zircons.

The date of c. 802 Ma from sample AJD-3, however, is harder to explain away by inheritance. There is no basement in the area of this age that these zircons could have been reworked from. AJD-3 also comes from the lithology (welded tuff) that would be most expected to contain fresh zircons that had an eruptive igneous origin. It therefore seems likely that this is the true age of the volcanics at this level in the Halfayn Formation. If the Rb-Sr and K-Ar dates were correct, zircons of this age should be present in the Halfayn Formation. Out of 12 zircons used in the isochrons from the Halfayn Formation (Samples AJD 1, 3, and 4) none even approaching 700 Ma, let alone 600 Ma were found. This suggests that the Rb-Sr and K-Ar dates have been reset by a later heating event. This idea is supported by the fact that the K-Ar date of 654 ± 12 Ma (Gorin et al., 1982) has already
been considered to be unreliable (Bell, 1993a), and that the Rb-Sr date of Kramer and Frei (1992) of 562 ± 42 Ma (taking the Halfayn data on its own) has such a large error. I therefore suggest that the Halfayn Formation is significantly older (c. 800 Ma) than previously thought. If this is the case the Halfayn Formation pre-dates the volcanism of the Ghadir Manqil Formation in the Jebel Akhdar, which has been dated at 723 +10/-16 Ma (Brasier et al., 2000) and 711.8 ± 1.6 Ma (this study) using the U-Pb method (Chapters 3 and 4).

The carbonate that unconformably overlies the Halfayn Formation is a possible Hadash Formation equivalent (Chapter 5) and passes up into the clastics of the Masirah Bay Formation. This suggests that no record of Ghadir Manqil deposition is preserved in the Huqf area. The unconformity at the top of the Halfayn Formation may therefore represent a time gap of over 200 Myrs.
Chapter 3

The Abu Mahara Group in the Jebel Akhdar: The Ghubrah and Ghadir Manqil Formations
CHAPTER 3. THE ABU MAHARA GROUP IN THE JEBEL AKHDAR: THE
GHUBRAH AND GHADIR MANQIL FORMATIONS

PART I: INTRODUCTION

3.1. Introduction

The Abu Mahara Group is well-exposed in the Jebel Akhdar (Fig. 3.1), comprising a thick, dominantly siliciclastic succession. It includes the oldest deposits present in the Jebel Akhdar area and its base is not exposed in the region. Where its upper boundary can be seen, it is overlain by the distinctive Hadash Formation of the Nafun Group. This section of the stratigraphy was first described by Kapp and Llewellyn (1965) who included it within the lower part of their Mistal Conglomerate.

In this study, the Abu Mahara Group in the Jebel Akhdar is divided into the Ghubrah and Ghadir Manqil Formations, with a further Jabir Formation being speculatively suggested. The Ghubrah Formation includes the lowermost glacial deposits of the Abu Mahara Group that occur throughout much of the Jebel Akhdar. The Ghadir Manqil Formation lies above these deposits and incorporates the upper part of the Abu Mahara Group below the Hadash Formation. As the Ghadir Manqil Formation is commonly well-exposed and contains distinctive units, confident lithological correlations can be made within it across the Jebel Akhdar. The proposed Jabir Formation is only exposed in Wadi Bani Jabir. In this thesis, it is tentatively suggested to pre-date the Ghubrah Formation.

Volcanic rocks overlying basement in the Huqf area of Oman, termed the Halfayn Formation (Le Metour et al., 1992; Dubreuilh et al., 1992) have been correlated with the Ghadir Manqil Formation in the Jebel Akhdar by some authors (eg. Bell, 1993) (Fig. 2.1. Chapter 2). However, U-Pb dating of zircons as part of this study has suggested that the Halfayn Formation is actually significantly older than the Ghubrah and Ghadir Manqil Formations in the Jebel Akhdar (see discussion in Chapter 2).

The aim of this chapter is to describe the sediments of the Abu Mahara Group in the Jebel Akhdar, establish the facies variation across the region, and to develop a depositional
model. Summary logs from sections across the Jebel Akhdar are shown in Fig. 3.2. These allow an overview of the Abu Mahara Group and show the two formations comprising the group and the main facies associations defined in this study. Detailed lithological logs are in Appendix D.

3.2. Methods

The most complete sections of the Abu Mahara Group that are well-exposed and relatively easily accessible occur in Wadi Sahtan, Wadi Hajir, and Wadi Mistal (Figs 3.3, 3.4, 3.5). Sections at these localities were logged in detail with the aim of characterising lithologies, facies variation and establishing palaeoenvironments. Sedimentary structures were recorded and measured and where possible, beds were traced laterally, though this often proved difficult due to a lack of exposure and, more commonly, inaccessibility. Supplementary sections in Wadi Bani Awf, Wadi Bani Kharus, Wadi Mu’aydin, and Wadi Mistal were also studied to provide further information on facies variation and palaeoenvironments (Fig. 3.2). To supplement these data, representative facies samples were collected for petrographic and microfacies studies and clast analyses of pebbles within diamicite and conglomerate units were conducted in the field. The clast analyses were done by selecting 3 separate lines, each 1m long, and marking them on the outcrop. These were then marked off at 10cm intervals and the lithology, size, and shape of the nearest outsize clast (>3mm in size) to each 10cm mark was recorded (Fig. 3.6). Isotopic and geochemical analysis of the rare carbonate units that occur within the Abu Mahara Group of the Jebel Akhdar was also conducted. The analytical methods are described in Appendix C.

This combined information has enabled reconstruction of the facies relationships and facies evolution through time, and suggests a glaciomarine influence within a rift basin for the setting of the formation.

3.3. Previous Work

The Abu Mahara Group in the Jebel Akhdar was first described by Kapp and Llewellyn (1965) in their survey of the area. It was included within the lower part of the Mistal
Conglomerate, the type section of which was defined in Wadi Hajir. The upper part of Kapp and Llewellyn's (1965) Mistal Conglomerate included the Hadash and Masirah Bay Formations (Chapters 5 and 6) that are now considered to be part of the Nafun Group (Fig. 3.7). Kapp and Llewellyn (1965) split the Abu Mahara Group equivalent lower part of the Mistal Conglomerate into an upper and lower division. Matrix-supported 'conglomerates' of the lower part of the formation were interpreted as being influenced by glacial processes (either terrestrial or glaciomarine), and the overlying sandstones and siltstones of the upper part (still below the Hadash Formation) were thought to be fluvial and/or tidal deposits (Fig. 3.7). Later work conducted by Glennie *et al.* (1974), Beurrier *et al.* (1986), Rabu *et al.* (1986), and Hughes Clarke (1988) generally concurred with this glaciomarine interpretation of the lower part of the formation, the latter suggesting the formation could equate with the youngest Precambrian (Varangerian/Marinoan) glaciation.

Rabu *et al.* (1986) re-named the Mistal Conglomerate the Mistal Formation and divided it into four members (Fig. 3.7). The lower three of these are equivalent to the Abu Mahara Group of this study, and the uppermost (the Amq Member) corresponds to the Hadash and Masirah Bay Formations (Chapters 5 and 6).

Although Rabu *et al.* (1986) still suggested some glacial influence in the Mistal Formation, further work by Rabu (1988), Blendinger and Teyssen (1989), and Rabu *et al.* (1993) began to question the validity of a glacial origin for the diamictites within the formation. Instead it was suggested that they formed as the result of submarine mud flows occurring during tectonically active conditions (horst and graben formation). Much of the work done on subsurface data within Petroleum Development Oman (eg. Kapellos *et al.*, 1992; Bell, 1993a and b; van Marle *et al.*, 1994), continued to evoke a glacial influence however, and suggested that the large thickness variations within the formation (eg. 600m to 0m over 8km (Bell, 1993a)) probably reflected deposition in an extensional graben or half-graben.

Bell (1993a) suggested a new subdivision of the Abu Mahara Group (Fig. 3.7) and first used the term 'Ghadir Manqil Formation' to describe the lowermost sediments in well Ghadir Manqil-1. These are equivalent to the lower three members of Rabu *et al.*'s (1986) Mistal Formation. Bell (1993a) further subdivided his Ghadir Manqil Formation into a clastic, a volcanic, and a diamictite member (Fig. 3.7), as well as proposing divisions for
the overlying Masirah Bay Formation. Note that the Basal Carbonate Member of Bell’s (1993a) Masirah Bay Formation is equivalent to the Hadash Formation.

Recent work by Loosveld et al. (1996) and Kellerhals (1998) on the Mirbat Sandstone Formation in the south of Oman has correlated it with the Abu Mahara Group in the Jebel Akhdar. These workers interpreted the Mirbat Sandstone Formation as having formed in a glacial environment.

3.4. Subdivision and nomenclature of the Abu Mahara Group in the Jebel Akhdar

In this study, the Abu Mahara Group in the Jebel Akhdar has been divided into the Ghubrah and Ghadir Manqil Formations. The Ghubrah Formation is equivalent to the lowermost (Ghubrah) member of Rabu et al.’s (1986) Mistal Formation. The Ghadir Manqil Formation incorporates the Saqlah and Fiq Members that were also originally suggested for the Mistal Formation by Rabu et al. (1986) (Fig 3.7). The Ghadir Manqil Formation referred to in this thesis therefore differs to the Ghadir Manqil Formation originally proposed by Bell (1993a), which also included what is now termed the Ghubrah Formation. This further subdivision has arisen because of the possibility of a large time-gap between the Ghubrah and Ghadir Manqil Formations (see Chapter 4). Note that although the term ‘Ghadir Manqil Formation’ does not originate from the Jebel Akhdar, it has been adopted in this study as it has become generally accepted within Petroleum Development Oman to refer to this section of the stratigraphy.

Deposits in Wadi Bani Jabir are distinctly different to those seen elsewhere in the Jebel Akhdar and may represent an earlier episode of sedimentation (see Section 3.5). These deposits are referred to as the ‘Jabir Formation’.

3.5. Stratigraphy of the Abu Mahara Group in the Jebel Akhdar

3.5.1. Jabir Formation

Deposits in Wadi Bani Jabir in the north-east of the Jebel Akhdar possibly represent lateral equivalents of the Ghubrah or Ghadir Manqil Formations. However, the nature of the
deposits at this level in Wadi Bani Jabir are notably different to those observed in the Abu Mahara Group elsewhere in the Jebel Akhdar. They are composed almost entirely of sandstones and siltstones and there is no evidence for glaciation preserved within them. These deposits have therefore been labelled separately as the Jabir Formation. The base of the Jabir Formation is not exposed, but it is overlain by the Fiq Member. In this study the Jabir Formation is tentatively suggested to pre-date the Ghubrah Formation. The boundary between the Jabir and Ghadir Manqil Formations therefore represents a major unconformity.

3.5.2. Ghubrah Formation

The Ghubrah Formation outcrops in Wadi Mistal and Wadi Mu’aydin in the east of the Jebel Akhdar, and in Wadi Sahtan in the west. The term 'Ghubrah' was first used to describe these deposits by Rabu et al. (1986) after the Ghubrah Bowl, which drains through Wadi Mistal, and in which they are well exposed (Plate 3.1). Note that Wadi Mistal is used synonymously with the Ghubrah Bowl in this thesis. The Ghubrah Formation forms the oldest exposures of the Abu Mahara Group across almost all of the Jebel Akhdar (the only possible exception being Wadi Bani Jabir) and its base is not exposed anywhere within this area. In Wadi Mistal the Ghubrah Formation is at least 200m thick and is overlain by the laterally discontinuous volcanic Saqlah Member or the lower parts of the Fiq Member of the Ghadir Manqil Formation. In Wadi Sahtan, the thickness of the Ghubrah Formation is at least 400m. It is overlain directly by the Fiq Member of the Ghadir Manqil Formation in Wadi Sahtan, the volcanic Saqlah Member being restricted to the more eastern part of the Jebel Akhdar. The boundary between the Ghubrah and Ghadir Manqil Formations is poorly exposed.

3.5.3. Ghadir Manqil Formation: Saqlah Member

The Saqlah Member of the Ghadir Manqil Formation comprises discontinuous flows of mafic lavas up to 60m thick that are well exposed in Wadi Mu’aydin and Wadi Mistal (Plate 3.2). It was named by Rabu et al. (1986) after Wadi Saqlah which lies in the western part of Wadi Mistal, and in which it is well exposed. The mafic flows of the Saqlah Member overlie the top of the Ghubrah Formation and are in turn overlain
conformably by the Fiq Member. They are not observed to the west of the village of Fiq in Wadi Mistal (Figs. 3.1 and 3.2). Intrusive basalt bodies within the Ghubrah Formation of Wadi Mistal are also included within the Saqlah Member here (following the nomenclature of Rabu et al. (1986)).

3.5.4. Ghadir Manqil Formation: Fiq Member

The Fiq Member forms the upper part of the Ghadir Manqil Formation and is well exposed in Wadi Sahtan (Plate 3.3), Wadi Hajir, and Wadi Mistal. Further outcrops also occur in Wadi Bani Awf, Wadi Bani Kharus, Wadi Mu’aydin, Wadi Hedak, and Wadi Bani Jabir. It was named by Rabu et al. (1986) after the village of Fiq in Wadi Mistal. It overlies the Saqlah Member, or where this is absent, the Ghubrah Formation, and is overlain by the Hadash Formation at all the localities where its upper boundary is exposed. In Wadi Sahtan the Fiq Member is approximately 1500m thick. This compares to ~1000m thickness in Wadi Mistal 30km further to the east. In Wadi Bani Jabir a further 30km to the north-east from Wadi Mistal, it is only represented by a thickness of 5m (Fig 3.2).
PART II: ANALYSIS OF THE ABU MAHARA GROUP IN THE JEBEL AKHDAR

3.6. The Jabir Formation

In Wadi Bani Jabir, hundreds of metres of green to red immature sandstones and siltstones containing no outsize clasts and showing no evidence for glaciation occur below the 5m thick Fiq Member of the Ghadir Manqil Formation. The presence of the Hadash Formation overlying these deposits (thought to have been deposited rapidly after the end of Ghadir Manqil deposition – see Chapter 5) suggests that little post-Fiq erosion of them could have occurred. It therefore seems that the most likely explanation for the thin (5m) Fiq Member is that Wadi Bani Jabir was situated on a high and experienced no deposition for much of Ghadir Manqil time.

Many of the sedimentary clasts found in the Ghubrah Formation are of very similar lithologies to the lowermost sediments exposed in Wadi Bani Jabir. This, coupled with the lack of evidence for glaciation within the Jabir Formation suggests that Wadi Bani Jabir may also have been a high during Ghubrah deposition, and was the source for at least some of the material within the Ghubrah Formation. If this is the case the sediments of the Jabir Formation in Wadi Bani Jabir pre-date the Ghubrah Formation. This is, however, only tentatively postulated here as it is also possible that the Jabir Formation may be a lateral equivalent of the Ghubrah Formation. Further work, possibly looking to date horizons within Wadi Bani Jabir, or using heavy minerals to evaluate source areas would need to be done before any further conclusions could be made. However, data from the subsurface suggests the Abu Mahara Group has a very laterally variable thickness, reaching >6km in places (Loosveld et al., 1996). There have been very few well penetrations and no core recovery of these thick, Abu Mahara Group rocks. New U-Pb dates (Chapter 2) suggest over 80Myrs of time occurred between the volcanic deposits of the Abu Mahara Group in the Huqf (Halfayn Formation) and those in the Jebel Akhdar (Ghubrah Formation). Significant deposition of Abu Mahara Group sediments pre-dating the Ghubrah Formation in the Jebel Akhdar could therefore quite easily be envisaged.
3.7. Facies analysis of the Ghubrah Formation

The lithofacies that occur within the Ghubrah and Ghadir Manqil Formations are notably different, and the two formations are therefore dealt with separately here. A summary of the facies associations and lithofacies types of the Ghubrah and Ghadir Manqil Formations is shown in Table 3.1. The facies associations used have been adapted in part from Eyles et al. (1985), Moncrieff and Hambrey (1990), Brodzikowski and van Loon (1991) and Miller (1996). The lithofacies names and codes have been partly adapted from Eyles et al. (1983), Moncrieff (1989), and Moncrieff and Hambrey (1990).

The lack of any bedding within much of the Ghubrah Formation makes it difficult to account for structure and gauge thickness accurately, as well as to correlate it across the Jebel Akhdar. This is compounded by the fact that most of the Ghubrah Formation is very strongly cleaved. Thicknesses given here are therefore estimates. Correlation (Fig. 3.2) is based on lithological considerations and the Ghubrah Formation’s occurrence as the oldest diamictite unit in the area.

3.7.1 Distal glaciomarine facies association

The facies within the Ghubrah Formation are all intimately associated and can be included within a distal glaciomarine facies association. This association is dominated by non-glacial marine and ice-rafting processes (Eyles et al., 1985; Moncrieff and Hambrey, 1990). Four different sedimentary facies types and one volcanic lithology are present (Table 3.2).

3.7.1.1. Massive Diamictite facies (Dm)

The name ‘diamictite’ has been widely accepted as a ‘non-genetic term including terrigenous sedimentary rocks that contain a wide range of particle sizes’ (Flint et al., 1960a, b; Harland et al., 1966). In this study the term is used to describe matrix supported rocks containing >1% outsize clasts larger than 2mm in size, with a significant proportion of mud within the matrix. This definition has been modified from Moncrieff and Hambrey (1990).
Massive diamictites occur within the Ghubrah Formation in Wadi Mistal and Wadi Sahtan. These diamictites are light grey to brown in colour and form bodies that are up to hundreds of metres thick. Upper and lower boundaries of these bodies tend to be transitional and often pass gradationally into massive siltstone facies. The diamictites are structureless and contain no internal sedimentary fabric, although this may often be masked by the strong pervasive cleavage (Plate 3.4). Clasts are dispersed randomly, and where they are unaffected by the cleavage show no preferred orientation. The proportion of outsize clasts in this facies varies from 5% to 20%. Clast size varies from <1cm to >1m, but clasts are mostly a few cm in size, with better indurated lithologies, such as granite or granodiorite, tending to form the larger clasts (Plate 3.5). The clasts are mainly tuffaceous in Wadi Mistal, with sedimentary clasts more common in Wadi Sahtan (Fig. 3.8). Rarely, the clasts within the massive diamictites of the Ghubrah Formation exhibit striated surfaces (Plate 3.6). Facetted clasts displaying bullet- or flat-iron shapes also occur. The matrix of the diamictites is composed of 5-40% mud/silt grade material, with the rest being made up of sand grade grains. Approximately 35% of these larger sand-sized grains are composed of quartz, with the remainder being made up mainly of fine-grained felsic volcanics. Some preliminary work has been done on the clay mineralogy of the matrices of these diamictites. This suggests that illite is the main clay mineral present with smectite also occurring in some of the samples.

A realistic interpretation of diamictite genesis cannot be made until data pertaining to vertical and lateral lithofacies relationships and sequence context together with sediment body geometry are available from detailed sedimentological logging (Eyles et al., 1985). Although the massive diamictite (Dm) facies was not logged in detail due to its lack of bedding and highly cleaved nature, some interpretation, in part based on its association with other facies, can still be made. Deposits in which large clasts are embedded in a muddy sandstone matrix with little consistent preferred fabric, similar to those described within the massive diamictite facies here, can be formed as deposits of cohesive debris flows (eg. Nardin et al., 1979; Lowe, 1982; Stow et al., 1996). This interpretation for the diamictites of the Ghadir Manqil Formation was adopted by Rabu (1988), Blendinger and Teyssen (1989), and Rabu et al. (1993), who suggested that they formed as the result of submarine flows occurring during periods of active extensional tectonics. The lack of any
grading within the diamictites of the Ghubrah Formation, its association with the dropstone laminité (Fld) facies, and the presence of rare striated and facetted clasts however, suggest that an ice-influenced origin may be more likely (eg. Harland et al., 1966; Miller, 1996; Kellerhals and Matter, 2000). Striations can also be formed through other non-glacial processes (eg. Schermerhorn, 1974), but it is worth noting that a lack of striated and/or facetted clasts does not preclude a glacial influence. Material transported englacially, for example, is unlikely to have these features (eg. Eyles and Eyles, 1983; Brodzikowski and van Loon, 1991; Eyles, 1988), nor is material transported by polar glaciers with no well-developed traction layer (Eyles et al., 1983). Overall, however, the transitional upper and lower contacts passing into massive siltstones, the large, relatively homogenous thickness, and the inferred lateral extent (from Wadi Mistal to Wadi Sahtan – a distance of >40km) suggest that the massive diamictites of the Ghubrah Formation formed in a subaqueous environment away from ice-margins as a result of settling of suspended sediment and ice-rafted material, possibly in an outer shelf/slope setting (eg. Anderson et al., 1984; Deynoux, 1985; Eyles, 1988; Eyles and Lagoe, 1990; Moncrieff and Hambrey, 1990; Brodzikowski and van Loon, 1991).

3.7.1.2. Massive and graded siltstone facies (Fm and Fg)

This facies is intimately associated with the Dm diamictites within the Ghubrah Formation. It is commonly light-coloured and is similar in composition to the matrix of the Dm diamictites. Its boundaries are transitional into the diamictite and are defined by a gradual decrease/increase of clasts. These siltstones form bands up to 10m thick within the diamictite and tend to be laterally discontinuous over <100m (although this is possibly due to deformation within the Ghubrah Formation). Within these bands, both massive (Fm) and graded (Fg) siltstones occur. Grading, where it occurs, defines beds and consists of a fining-up from coarse to muddy siltstone on a few cm scale.

The intimate association of this facies with the ice-rafted (Dm) diamictites described above and the gradational boundaries between the two facies, suggests that they represent quiet water marine deposition when the influence of ice-rafting was reduced. This is possibly due to ice retreat and a reduced glacial influence (Eyles and Lagoe, 1990), or a move to
deposition away from lanes of ice-berg transport, either distally (Drewry, 1986; Moncrieff and Hambrey, 1990) or laterally (Miller, 1996).

3.7.1.3. Dropstone laminite facies (Fld)

This facies occurs within the Ghubrah Formation of Wadi Mu’aydin, where it forms a unit ~30m thick below the volcanic Saqlah Member. It is composed of dark muddy siltstones containing few mm-thick planar laminae that are laterally continuous for tens of metres. Rare dropstones up to few cm in size deform these laminae (in a similar manner to dropstones observed in Pleistocene glaciolacustrine deposits by Thomas and Connell (1985)) and are often much larger than the lamination separation, precluding emplacement by traction currents (Miller 1996). It therefore seems likely that these outsize clasts were deposited by ice-rafting processes in which sediment, including larger clasts, held within icebergs was released as the ice melted and was ‘dropped’ into the deposits accumulating on the sea floor (eg. Powell and Domack, 1995). The background sea-bed deposits reflect relatively quiet distal marine sedimentation, with mm-scale lamination forming as a result of varying sediment input, possibly from distal dilute turbiditic flows (Eyles, 1988; Moncrieff and Hambrey, 1990).

The dropstone laminite facies (Fld) is thus interpreted as forming in a distal glaciomarine environment, dominated by marine processes with some influence from ice-rafting. It is worth noting that the dropstones recorded here represent some of the most convincing evidence for glaciation within the Ghubrah Formation, and their occurrence at a similar level in the stratigraphy to the massive diamictites (Dm) lends further weight to the glaciomarine interpretation of the Dm diamictite deposits.

3.7.1.4. Carbonate facies (L)

Carbonate facies form a very minor part of the Ghubrah Formation and are only present at 1 or 2 localities within Wadi Mistal near the village of Fiq. Carbonate occurs as ~10cm thick beds that are better indurated than the Dm diamictites within which they occur (Plate 3.7). The carbonate beds commonly show deformation and boudinage structures, and appear laterally discontinuous because of this. The carbonate beds contain clasts similar to
those in the surrounding Dm diamictite. In thin-section, the carbonate can be seen to be composed of <50µm sized sub- to euhedral dolomite rhombs, commonly exhibiting dark rims. As well as the larger lithic clasts, quartz grains up to 0.2mm in size occur within this dolomitic matrix.

The intimate association of carbonate beds with the massive diamictite (Dm) facies and the presence of outsize clasts within them suggest that they may have formed under similar conditions when ice-rafting was prevalent. The fact that carbonate beds are so sparingly developed is difficult to explain, but their presence possibly reflects local conditions (such as raised alkalinity) that were conducive to carbonate precipitation. Such conditions may have been produced locally by the microbial reduction of sulphate or possibly by subaqueous springs rich in HCO₃⁻ (eg. Knoll et al., 1986; Grotzinger and Knoll, 1995; Wright, 1997). Alternatively, the carbonate precipitation may reflect short-lived changes in oceanic chemistry, as could be caused, for example, by renewed oceanic circulation after a period of stagnation (eg. Kennedy, 1996).

3.7.1.5. Tuffaceous deposits

A tuffaceous bed occurs in Wadi Mistal within massive diamictites approximately 50m below the top of the Ghubrah Formation (Plate 3.8). This has been described in McCarron (2000) and Brasier et al. (2000) and has been dated by the U-Pb isochron method at 723+16/-10Ma. The ash is a pale grey colour, is up to 2m thick and is laterally discontinuous over ~100m. At the same level 7-8km to the north, however, an ash bed is again found. This is formed of a series of few cm thick ash-rich beds that have been deformed and pinch out laterally on a metre scale (Plate 3.9). It is possible that the lateral impersistence is also a function of this deformation, and that originally the ash-bed would have been laterally more persistent. The unit described, sampled and dated by McCarron (2000) and Brasier et al. (2000), was resampled and dated as part of this study. Zircons from within it have yielded a new U-Pb zircon date of 711.8±1.6Ma (Fig. 4.4), the significance of which will be discussed in Chapter 4. The unit is composed of 9 fining-up beds each 10-30cm thick accompanied by minor beds of poorly sorted agglomerates. The bases of the agglomerates comprise volcanioclastic breccia with angular, fine-grained rhyolitic clasts up to 2-3cm in size, which fine upward to medium-grained tuff. The upper
parts of some of the tuffaceous beds, and indeed most of the beds at the locality further to the north, are fine-grained and planar laminated. In thin-section, the rock is highly sericitised with a strong tectonic fabric and without evidence for a glassy volcanic texture.

Zircons taken from these tuffaceous beds in the Ghubrah Formation of Wadi Mistal have euhedral shapes. This suggests that they have not been reworked and represent an eruptive volcanic episode that occurred during the deposition of the Ghubrah Dm diamictites (Brasier et al., 2000; McCarron, 2000). The fine-grained upper parts of some of the beds could represent fallout from suspension and the planar laminated divisions may represent upper-stage plane beds developed during flow of a thin undercurrent (McCarron, 2000).

3.7.2 Ghubrah Formation synthesis

The Ghubrah Formation is dominated by massive diamictites formed by ice-rafting in a distal glaciomarine environment (Fig. 3.19a). During a time of overall glaciation, periods of reduced ice-rafting occurred, reflecting either a move away from sediment entry points or a retreat of the ice margin. Ice-rafting processes dominated in both Wadi Sahtan and Wadi Mistal. The higher proportion of sedimentary clasts present in Wadi Sahtan (Fig. 3.8) suggests that the two areas may have been influenced by different sources and possibly different ice lobes. Ice-shelf rafted debris may, however, show a change in provenance with distance from the grounding line (Powell and Domack, 1995), which could explain the observed change in clast types. To the south, in Wadi Mu'aydin, marine non-glacigenic processes dominated, suggesting it was further away from the ice margin at the time of its deposition. The influence of ice-rafting can still be seen, however, in the rare dropstones within laminated sediments.

Locally developed carbonate precipitation occurred during deposition of the Dm diamictites. This may reflect localised microbial activity or short-lived changes in the ocean chemistry.

The presence of a tuffaceous bed within the diamictites of Wadi Mistal indicates some limited volcanic activity at the time of their deposition. The fact that this bed cannot be found in Wadi Sahtan further to the west suggests that the eruptions may have been centred.
to the east. This would be consistent with the volcanic centre postulated for the Saqlah Member of the Ghadir Manqil Formation (see Section 3.8). The U-Pb zircon ages of 723±16/-10Ma (Brasier et al., 2000) and 711.8±1.6Ma (this study) from the Ghubrah Formation place it as a Sturtian glacial equivalent (eg. Brasier et al., 2000; Young 1995). The implications of this are discussed further in Chapter 4.

3.8. The Saqlah Member of the Ghadir Manqil Formation

The Saqlah Member is a volcanic member composed of the discontinuous lava flows that lie at the base of the Ghadir Manqil Formation overlying the Ghubrah Formation. The discontinuous intrusive basaltic bodies that occur within the Ghubrah Formation of Wadi Mistal have also been included within the Saqlah Member in this study (after Rabu et al., 1986). Rabu (Rabu et al. (1986), Rabu (1988), and Rabu et al. (1993)) conducted a detailed description and interpretation of the Saqlah Member of the Ghadir Manqil Formation. Her work forms the basis for many of the descriptions and interpretations that follow.

The intrusive bodies within the Ghubrah Formation are <10m thick and are discontinuous over 20-30m. Both conformable and unconformable types occur. Rabu (1988) described these as spilitised diorites with a doleritic or microgranular texture. Plagioclase forms 80% of the rock with more minor interstitial chlorite as a biotite replacement. Accessory minerals include apatite, opaques and carbonate.

The discontinuous flows of mafic lavas that occur at the base of the Ghadir Manqil Formation and form the majority of the Saqlah Member are well exposed in Wadi Mu’aydin and Wadi Mistal. Poorly exposed outcrops also occur in Wadi Hedak (Fig. 3.1). The boundary between these flows and the underlying Ghubrah Formation is poorly exposed, but they are overlain conformably by the Fiq Member of the Ghadir Manqil Formation. The volcanics of the Saqlah Member are not observed to the west of the village of Fiq in Wadi Mistal.

Pillowed basalt of the Saqlah member occurs in the south of Wadi Mistal near Wakan and also further to the south in Wadi Mu’aydin (Plate 3.10). In Wadi Mistal, pillowed basaltic
flows of the Saqlah Member can be up to 30m thick, with pillows occurring up to 1.5m in size. In Wadi Mu’aydin, partly brecciated pillowed basalt up to 60m thick occurs. Vesicles up to 0.5cm size, commonly infilled with carbonate, also occur at both localities. Rabu (1988) described the pillowed volcanics as being composed of microlithic lava with a spilitised andesitic-basalt composition. Plagioclase occurs as laths up to 1mm long and often forms an interlocking framework.

Unpillowed basalt of the Saqlah Member also occurs within Wadi Mistal below the Fiq Member. Here, the basalt forms bodies up to a few metres in thickness that are laterally discontinuous over tens to hundreds of metres. Close to the upper and lower surfaces of these bodies numerous vesicles up to 1cm in size occur (Plate 3.11) and locally comprise 20% of the rock. Cryptocrystalline chlorite associated with carbonate fills these vesicles forming amygdales.

Geochemically, the Saqlah Member of the Ghadir Manqil Formation represents a volcanic episode of composition between alkali basalts and trachy-andesites (Table 3.3; Fig. 3.9), (Rabu et al., 1993; Winchester and Floyd, 1977), with one sample plotting in the benmoreite field of Cox et al., (1979) (Fig. 3.10). The principal flows within the Jebel Akhdar occurred in Wadi Mu’aydin, with further flows in Wadi Mistal. These were principally effusive and occasionally hypovolcanic. The pillows suggest subaqueous eruptions. The geochemistry and regional geology suggest that the basalts belong to a differentiated alkali suite emplaced during extension in an intraplate continental setting (Figs. 3.11, 3.12) (Pearce and Cann, 1973; Meschede, 1986).

The volcanics of the Saqlah Member have previously been linked to the more extensive volcanics of the Hatat Formation in the Saih Hatat region (Figs. 1.1, 1.3; eg. Rabu et al., 1986; Le Metour et al., 1986; Villey et al., 1986). The <120m thick Jahlut Member of the Hatat Formation is represented by submarine volcanic and hyaloclastic rocks (Rabu et al., 1993). These represent subalkaline transitional to alkaline volcanism (Fig. 3.9). The subalkaline basalts resemble little differentiated tholeiites, poor in TiO₂, Zr, K₂O, and SiO₂. Their high nickel and chromium content makes them reminiscent of oceanic basalts (Rabu et al., 1993), and they can be compared with the basalts of oceanic ridges, or of ridges becoming oceanised. The transitional to alkaline basalts could correspond to intra-plate
basalts. Recent work (Hill et al., 2000) has questioned the use of Yttrium in discrimination diagrams, suggesting that it is mobile even at the earliest stages of alteration and weathering. Interpretations based on such diagrams should therefore be cautious. However, the association of alkaline, transitional and tholeiitic lavas in a restricted geographical area like Saih Hatat (or even extended as far as the Saqlah Member of the Jebel Akhdar) can still be quite confidently taken to correspond to magmatism accompanying rifting of continental crust (Wilson, 1989). Whether this was active rifting triggered by plume activity or was passive and caused by differential stresses in the lithosphere is hard to say with the available data. For example, when the data are plotted on a 2Nb-Zr/4-Y diagram (Fig. 3.12) (after Meschede, 1986), points fall in all the major categories, including plume-related mid-ocean ridge basalts. Even if the volcanism was related to plume activity and a raised mantle potential temperature, a stretch factor approaching 1.5 would still have been required to produce mantle melting and tholeiitic lavas (McKenzie and Bickle, 1988; Latin and White, 1990). With purely passive rifting, a stretch factor approaching 3 would be needed to produce the same effect (McKenzie and Bickle, 1988; Latin and White, 1990). It seems likely that the volcanism of the Saqlah Member was related to this rifting event, and that the Jebel Akhdar was situated further from the centre of volcanism than the Saih Hatat region (Fig. 3.19b).

3.9. Facies analysis of the Fiq Member of the Ghadir Manqil Formation

Many of the units of the Fiq Member of the Ghadir Manqil Formation are well-bedded and it is in general much less pervasively cleaved than the Ghubrah Formation. This has allowed a number of detailed sedimentological logs to be made through it (Appendix D). These logs are summarised in Fig. 3.2. Lateral variability can often be traced out or inferred from section to section, and events affecting the whole basin can be identified. This has not only allowed confident lithological correlations within the Fiq Member to be made across the Jebel Akhdar, but has also facilitated a detailed interpretation of the Fiq Member.

The facies present within the Fiq Member can be ascribed to four different facies associations (Table 3.1). A more detailed summary of these facies associations is shown in Table 3.4.
3.9.1. Distal glaciomarine facies association

As in the Ghubrah Formation, some facies within the Fiq Member of the Ghadir Manqil Formation represent distal glaciomarine environments where non-glacial marine processes and ice-rafting dominated. The distal glaciomarine facies association is represented in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir and is composed of two facies types (Table 3.4)

3.9.1.1. Massive diamictite facies (Dm)

Massive diamictite deposited in a distal glaciomarine setting forms bodies up to 30m thick within the Fiq Member; they are similar in their overall nature to those described from the Ghubrah Formation. Good examples of this facies can be seen within the Fiq Member in Wadi Sahtan and Wadi Hajir, where individual units can be traced laterally for several kms and maintain a relatively constant thickness over this distance. The facies is either passed into via a gradual increase in the number and size of clasts (eg. 1040m log WS1 (Wadi Sahtan)), or occurs positioned between deep-water deposits (mud/siltstones) with more clearly defined boundaries (eg. 800m log WS1 (Wadi Sahtan)). Clasts form up to 20% of the diamictites and are up to 20cm in size. More commonly however, the clasts form less than 5% of the rock and rarely occur larger than few cm in size, making these particularly clast-poor varieties of diamictite (Plate 3.12). Granite and sandstone clasts are common, as are rhyolite, shale, carbonate and tuffaceous types (Fig. 3.8, analyses Dabut 3, 4 and 5). Many of the clasts have a facetted shape and rare striations can also be observed on some clast surfaces suggesting that they were affected by glacial processes during transportation (eg. Wentworth, 1936; Boulton, 1978; Dowdeswell et al., 1985). As with the Ghubrah Formation, an ice-rafted origin for these diamictites is inferred on the basis of these clasts, the transitional lower boundaries some units display (over which the number of clasts increases gradually), and the lack of any significant internal stratification within them (Anderson et al., 1984; Eyles, 1988; Kellerhals and Matter, 2000). The clast-poor nature of many of the diamictites, their sheet-like geometry, and their position in the stratigraphy between deep-water deposits suggests that they formed in a subaqueous environment away from ice-margins. This was probably as a result of settling of suspended sediment and ice-rafting, possibly in an outer shelf/slope setting (Deynoux, 1985; Eyles and Lagoe, 1990;
Moncrieff and Hambrey, 1990; Brodzikowski and van Loon, 1991). Where the number of clasts gradually increases, an equivalent increase in the amount of ice-rafting is postulated, possibly reflecting increased glacial activity and advance, or a move into lanes of more frequent ice-berg transport (Drewry, 1986; Moncrieff and Hambrey, 1990). The clast-poor diamicite at 1040m (log WS1) in Wadi Sahtan passes laterally eastwards over a few kms into deposits free from outsize clasts and suggests that ice-rafting decreased progressively in this direction.

3.9.1.2. Dropstone laminite facies (Fld)

This facies is present within the Fiq Member in Wadi Hajir (290-340m log WH1) where it forms a unit ~50m thick. At the same level a few hundred metres to the north in Wadi Hajir clast-poor Dm diamicite occurs, illustrating the close association of these two facies within the distal glaciomarine facies association. The laminated siltstone with dropstones facies (Fld) consists of dark grey shales/grey siltstones exhibiting laterally continuous (for 10s of ms) planar laminations on a few mm scale and containing few mm to 10cm thick sandstone lenses/beds. The sandstone beds are sharp-based with more gradational tops often showing rippling, fining-up into siltstone. These sandstones may be laterally continuous for at least tens of metres. Within these deposits outsize clasts up to 10cm in size occur, forming <1% of the rock, and deforming the underlying laminae (Plate 3.13). As discussed in the section on the Ghubrah Formation (Section 3.7), these provide very good evidence for ice-rafting in which sediment, including larger clasts, held within icebergs is released to the seafloor as the ice melts and is ‘dropped’ into the deposits accumulating there (eg. Powell and Domack, 1995). In this case, the dropstones occur in deposits that probably represent relatively quiet distal marine sedimentation: the sandstones represent distal turbidite event beds, and the laminated mud/siltstone formed as a result of variations in the suspended sediment flux, perhaps related to seasonal changes in river/subglacial output or turbiditic plumes (Eyles, 1988; Moncrieff and Hambrey, 1990).

3.9.2. Proximal glaciomarine facies association

The proximal glaciomarine facies association includes deposits that formed close to the grounded ice margin (the line at which ice entering a water body comes afloat (Miller,
1996)) (Fig. 3.13). It is composed of a varied assemblage of facies types, summarised in Table 3.4. Processes acting include rain-out of ice-rafted debris and sediment gravity flows (Fig. 3.13). Although the deposits of sediment gravity flows are likely to extend into distal glaciomarine environments, to avoid repetition they have been dealt with here within the proximal glaciomarine facies association. Sediment gravity flow processes are also not restricted to glaciomarine environments and indeed dominate in the non-glacial facies associations of the Fiq Member. In this study non-glacial sediment gravity flow deposits have been distinguished from those forming in a glaciomarine environment and have been dealt with separately (Section 3.9.3).

3.9.2.1. Massive diamictite facies (Dm)

This facies differs from the distal glaciomarine massive diamictite facies in a number of significant ways, even though there are many similarities between them. Massive diamictites formed in the proximal glaciomarine system occur in every wadi, often at a number of different levels, and vary in thickness from a few metres to over 100m. They may be laterally persistent over kms, but thickness variations and more laterally discontinuous units are more common than in the distal diamictite deposits. In contrast to the distal glaciomarine diamictites, they commonly occur at the top of shallowing-up sequences, often passing up directly (or locally via a metre or so of coarse-grained to pebbly sandstone) into significantly finer-grained deposits. Massive diamictites of the proximal glaciomarine facies association also commonly occur within coarse-grained sections of the stratigraphy where sharp bases, often overlying conglomerates are common. They are also intimately associated with stratified diamictites (Ds) and conglomerates (C), which often form bodies within them. Clasts within the proximal Dm diamictites may form more than 30% of the rock, but more commonly constitute 10-20% (Plate 3.14). These clasts are mainly few cm in size, but larger examples up to a metre or more in size occur; they are commonly larger than the majority of the clasts observed in the more distal deposits. Facetted clasts displaying flat-iron or bullet shapes are common (Plate 3.15). Rarer striated clasts also occur (Plate 3.16). The clast types present vary widely (Fig 3.8), this variation tending to be linked to geographical location. For example, in Wadi Hajir, granite clasts are more common than elsewhere, whereas in Wadi Mistal, tuffaceous clasts tend to dominate. No obvious relationship between clast types present and the relative age
of the Dm diamictite units is apparent, although a general higher proportion of volcanic clasts occurs within the Ghubrah Formation Dm diamictites compared to those of the Ghadir Manqil Formation. Preliminary work on the clay mineralogy of the matrices of these Dm diamictites indicates that illite and chlorite are the main clay minerals present.

Patches up to 40cm in size of coarse-grained material with sharp, irregular boundaries were recorded within this facies in Wadi Sahtan (Plate 3.17). They bear some resemblance to dump structures described by Thomas and Connell (1984) which form through the break up and overturning of dirt-laden, floating icebergs and the consequent release of large quantities of debris to the bed. This phenomenon has been described from contemporary environments by Ovenshine (1970) and Anderson et al. (1980). The shape of the patches observed in Wadi Sahtan and their sharp and irregular boundaries, however, suggests that they must have had some coherence when deposited. The most likely explanation for this is that the material was frozen together at the time of iceberg overturn and melting did not occur until it was incorporated into the sediment. Frozen aggregates of beach sand reported by Gilbert (1990) from northern Baffin Island have a similar form and have been interpreted as forming in a similar manner.

The ‘dump structures’ described above provide very good evidence for ice-rafting processes, and along with the facetted and striated clasts make a glacial origin for these deposits hard to dispute. The position of these Dm units at the top of shallowing-up sequences suggests a proximal location, as does their association with stratified diamictites (Ds) and conglomerates (C), which probably represent winnowed/remobilised diamictite (Section 3.9.2.3). Further evidence for deposition in a proximal environment comes from the large clasts present and the variable nature of the deposits, which would be expected to occur closer to the grounding line zone. The massive diamictites described above are thus interpreted as having formed from continuous rain-out of glacial debris from icebergs or a rafted ice sheet just seaward of the grounding line (Moncrieff and Hambrey, 1990) (Fig 3.13).

The fact that some clast types within the Dm diamictites are more common in certain wadis than in others may be indicative of distance from the grounding line during deposition (Powell and Domack, 1995). However, as these Dm diamictites are all within
the proximal glaciomarine facies association, it is more likely that the variation reflects differing sources for the clasts in the different areas. For example, the high proportion of tuffaceous clasts present in Wadi Mistal suggests that this material was being derived from a volcanic source, whereas the high proportion of granite clasts in Wadi Hajir suggests that the material here was being derived mainly from basement. The fact that different areas were being influenced by different sources explains the lack of any relationship between the relative age of the diamictite and the clast types present. Clast types present cannot therefore be used as a tool for correlation.

Thinner (1-4m) Dm diamictites also occur and are often associated with conglomeratic (C), stratified diamictite (Ds), and sandstone (Sm, Sg) beds. It is possible these deposits could have formed from ice-rafting processes (as described above for the thicker diamictite bodies), but their lower thickness, lateral extent of only tens of metres, and association with other sediment gravity flows makes a debris flow origin equally as likely. If this is the case, the finer material has not been sheared away and a matrix-supported deposit remains.

3.9.2.2. Stratified diamictite facies (Ds)

This facies is similar to the massive diamictite (Dm) facies, but differs in that some stratification, defined either by the clasts or the matrix, occurs within it. Although stratified diamictites (Ds) occur within the Fiq Member of the Ghadir Manqil Formation in most of the sections logged across the Jebel Akhdar, they are strongly subordinate to the Dm diamictite facies. Ds diamictites are intimately associated with the massive diamictites (Dm) and commonly form interbeds (varying from few cm to few metre thickness) that are laterally discontinuous over a few metres, and contain clasts similar to those in the Dm diamictite facies.

The stratification can be defined in a number of ways. Firstly, by clast-rich beds, in which pebbles or granular material are concentrated into bands, usually only few to 10cm thick (Plate 3.18). These are often laterally discontinuous on a metre scale and form close to the base of massive deposits. Sharp, but often slightly irregular lower and upper contacts are common, although more gradational boundaries between clast-poor and clast-rich areas.
also occur. Secondly, stratification can be recognised by variations in the proportion of mud and sand within the matrix. Relatively sand-rich layers are interbedded with relatively sand-poor layers on a cm to few cm-scale. This interbedding is laterally persistent for at least tens of metres (eg. 825m Wadi Sahtan). The boundaries between such layers may be sharp or, equally as likely, diffuse and more gradational.

The fact that the Ds diamictite facies is intimately associated with the massive Dm diamictite facies, and is only present within these massive units suggests that these deposits were also dominated by ice-raifting processes. However, the stratification indicates that other processes were also taking place.

The stratification may have been formed in a number of ways. One possible explanation for variations in the grain size of the matrix and in the concentration of clasts within these stratified diamictites is that winnowing processes were acting on ice-rafted material (Moncrieff and Hambrey, 1990; Eyles et al., 1985). Sandy laminations within the diamictites may have formed as traction currents winnowed out the mud fraction of ice-rafted material during settling. The alternation of these sand-rich laminae with more mud-rich layers suggests that the current intensity may have varied or been discontinuous. With more intense currents, only gravel grade and coarser material would be left in the residual sediment, which could account for the clast-rich beds described above.

A second possibility that could account for the presence of clast-rich beds is overturning of icebergs. In this process, coarse sediment builds up on the surface of an iceberg as it melts until instability is achieved. At this stage the unbalanced iceberg may suddenly tip or overturn and the coarse sediment that has accumulated will then be deposited in one event (Ovenshine, 1970). This would produce a relatively clast-rich bed within otherwise homogenous diamictite.

A further possibility is that small-scale slumping or debris flow processes may have been active. Redeposition of the massive diamictite facies in this manner, possibly triggered by oversteepening of the ice-rafted deposits, by the push of the ice itself, or by water turbulence generated during iceberg calving, could lead to the development of clast-rich layers within the deposit (eg. Miall, 1983).
It is probable that all of the processes described above were acting to some extent to form the stratified diamictites (D) observed within the Fiq Member of the Ghadir Manqil Formation. Although other processes may have been responsible for the stratification, the gross depositional environment was one dominated by ice-rafting just seaward of the grounding line.

3.9.2.3. Conglomeratic facies (C)

The original definition of diamictite (Flint et al., 1960a,b) states that the sediment is matrix-supported, although the term has since been used to describe clast-supported facies (eg. Frakes, 1978; Eyles et al., 1983; Miall, 1985). In this study the original definition has been adopted, and as such, the conglomeratic facies described here encompasses all the clast-supported lithologies present within the Ghadir Manqil Formation. (Note that this is slightly different to the definition of Moncrieff (1989) (adapted from Folk (1954)) who defined a conglomerate as having >80% clasts.) Conglomerates occur in the proximal glaciomarine environment, are commonly associated with the diamictite facies Dm and Ds, and indeed provide a clast-rich end-member for both of these deposits.

Within the proximal glaciomarine facies association, conglomerates (C) occur as laterally discontinuous (over 0.5m to 10m) bands less than a few cm thick (Plate 3.19), as well as occurring as well-bedded deposits generally less than 2m thick, but up to 10m thickness and more in some localities. These conglomerates commonly occur at or near the upper and lower surfaces of massive diamictite deposits. Their basal contacts are usually sharp and sometimes erosive. Grading may be apparent, and in places conglomerates pass gradationally up into sandstones. At most localities, however, the tops of conglomeratic beds are also well-defined. The clasts within conglomerates are commonly very similar in size, shape and lithology to those within Dm and Ds diamictites. Facetted clasts with a distinctive flat-iron or bullet shape have also been found. The matrix within these deposits is commonly composed of coarse-grained sandstone, with little finer-grained or muddy material present.
The thinner, more laterally discontinuous conglomerates probably represent either local concentrations of clasts derived from iceberg overturn (as outlined above for the stratified diamictite facies (Ovenshine, 1970)), or lag deposits. If they are lag deposits, intensive traction current winnowing may have been responsible for producing them (Moncrieff and Hambrey, 1990; Eyles et al., 1985). Alternatively, lags could have been formed by wave and tidal current reworking of ice-rafted diamictite deposits in relatively shallow marine conditions (Eyles, 1988).

The thicker conglomerate beds with erosive bases are unlikely to have formed from such processes however. As well as occurring associated with Dm and Ds facies, thicker conglomerates are also commonly found within sandstone (Sm, Sg) dominated units of the proximal glaciomarine facies association. In fact, such conglomeratic beds often pass gradationally into sandstone facies, and in their upper parts may contain rafted shale clasts. As these conglomerates contain glacially-facetted clasts and are associated with ice-rafted Dm and Ds facies, it is probable that they represent the deposits of debris flows that have remobilised ice-rafted material. Although debris flows are commonly matrix supported (Nardin et al., 1979; Lowe, 1982), fine sands and muds may be sheared from a transitional debris flow at the debris-water interface, and may be expelled into an overlying turbulent suspension (Wright and Anderson, 1982). Eventually the transitional debris flow freezes, leaving a conglomeratic deposit that has been removed of its fine material. The overlying sandstones probably represent deposits of the overlying turbulent flow that are then deposited on top of the ‘frozen’ conglomerate. In a proximal glaciomarine environment debris flows and their deposits have been reported by many authors (eg. Mackiewicz et al., 1984; Miall, 1983; Powell, 1988; Brodzikowski and van Loon, 1990; Miller, 1996). They may be triggered by a number of factors including oversteepening of the ice-raftered deposits, the push of the ice itself, shocks induced by earthquakes, or water turbulence generated during iceberg calving.

3.9.2.4. Massive/graded sandstone facies (Sm/Sg)

Within the proximal glaciomarine facies association, massive and graded sandstone facies are dealt with together as it seems likely that they formed from similar processes and are often interbedded with each other. Sm facies comprises homogeneous sandstones with no
internal stratification or structure. Both fine- and coarse-grained varieties occur and these are usually well-sorted. Graded sandstone (Sg) facies includes sandstones that exhibit a systematic change in grain size to either finer or coarser material throughout their thickness. Most commonly, Sg sandstones fine upwards. Sm and Sg sandstones form beds up to a few metres thick that locally gradationally overlie conglomerate, but which more usually have sharp bases. The Sm and Sg sandstones themselves are commonly overlain by rippled sandstone (Sr) that in turn passes up into siltstone facies. Dropstones can occasionally be seen at boundaries between beds. Laterally, Sm and Sg beds can be traced for hundreds of metres, but as they cannot be correlated precisely from wadi to wadi it seems unlikely that they extend much beyond this. Some massive and graded sandstone beds fill channel-like structures, up to 8m thick and pinching out laterally over 30-200m (Plate 3.20). These sandstones commonly overlie basal conglomeratic horizons.

The Sm and Sg sandstones described above were probably deposited by turbidity currents (Bouma, 1962). It is possible they represent more distal members of debris flows, where the finer-grained material has been stripped from the flow into a turbulent suspension (eg. Wright and Anderson, 1982; Fisher, 1983; Stow et al., 1996). This suspension may then flow downslope and deposit the sand/silt in turbidite deposits, with the coarser fraction being left further upslope. They may alternatively represent the deposits of lower energy flows distinct from debris flows that were supported by fluid turbulence throughout their existence (Stow et al., 1996). In either case, evidence that these deposits formed in a glaciomarine environment comes from their association with more diagnostic deposits such as diamicrites, and also from the presence of rare dropstones on some bedding surfaces. They could well have been triggered by mechanisms similar to those outlined above for the debris flow deposits (Section 3.9.2.3). It is likely that these deposits extended some distance beyond the grounding line zone and are probably not restricted purely to the proximal glaciomarine environment.

3.9.2.5. Rippled sandstone and siltstone facies (Sr and Fr)

Rippled siltstones (Fr) and sandstones (Sr) form a minor facies within the proximal glaciomarine facies association, but can occur in a number of forms.
Firstly, Sm and Sg facies occasionally display asymmetrically rippled tops in fine-grained sandstone or siltstone. These ripples have vertical form ripple indices (ripple wavelength/ripple height) of between 8 and 14 and rarely have internal structure preserved. They probably represent current ripples linked to the turbiditic flow that formed the Sm or Sg sandstone (Stow et al., 1996).

Rippled siltstones and sandstones also form thicker units within the proximal glaciomarine facies association. For example, within a part of the stratigraphy dominated by Dm diamictites in Wadi Hajir (605m log WH1), a rippled unit ~8m thick occurs. At this locality, over a vertical distance of a few metres, siltstones containing planar laminated to undulatory structures pass up into increasingly well-rippled siltstones (wavelengths of 8-12cm and heights of <2cm, giving vertical form ripple indices of between 6 and 8). These in turn pass up into siltstones and sandstones displaying climbing-ripple forms that gradually become increasingly steep and in-phase (Plate 3.21) (wavelengths typically ~14cm and heights of 2-8cm, giving ripple indices as low as 2). Symmetrical and trochoidally-shaped profiles were also observed.

The planar laminations in the siltstones at the base of this unit pinch out laterally over a few metres in low angle scours. This suggests that they formed in an upper flow regime, where flow velocities were either very high or the flow itself was very shallow (eg. Allen, 1997). The upward increase in grain size and formation of undulations and then ripples suggests lower flow regime conditions becoming established. The out-of-phase climbing ripples above these passing up into in-phase climbing-ripples indicate high rates of sediment fallout relative to downstream migration (Collinson and Thompson, 1989). No internal structure was observed in the steep in-phase climbing ripples with very low ripple indices however, and it is probable that a superimposed tectonic fabric was partly responsible for their formation and that they were originally similar to the out-of-phase climbing ripples observed directly below them. A subglacial stream outwash is postulated here to explain all of these observed features. Such a stream could maintain high velocity flows, and would also be capable of transporting the high concentrations of sediment required to form the climbing ripples.
3.9.2.6. Dropstone laminitite facies (Fld)

Laminated siltstones with dropstones (Fld) occur associated with diamicrites of the proximal glaciomarine environment as well as with the more distal diamicrite units (Plate 3.22). In the proximal environment, the siltstone is commonly coarser and contains a higher proportion of sand-sized material. The dropstone laminitite facies forms a relatively minor facies type in the proximal glaciomarine environment, comprising thinner units (usually no more than 5-10m thick) than those observed in the distal glaciomarine environment. They were probably deposited by similar processes, although the presence of coarser-grained material suggests higher energy conditions. This Fld facies again provides good evidence for the occurrence of ice-rafting within background marine sedimentation.

3.9.2.7. Massive/laminated mud/siltstone facies (Fm/Fl)

Mud/siltstone facies (Fm/Fl) free from dropstones also occur in the proximal glaciomarine facies association. They form a relatively minor part of the facies association, but do occur in a number of different forms:

i) Thinly-bedded mud/siltstone facies (Fm and Fl) occur interbedded with the massive (Sm) and graded (Sg) sandstones described above. The Fm and Fl facies probably represent the fine-grained fall-out from suspension from the tail of the turbidity current responsible for the deposition of the sandstones (Stow et al., 1996).

ii) Thicker (<few m) units of commonly laminated mud- and siltstone (Fl) also occur within the proximal glaciomarine environment (eg. 70m log WH1). The absence of dropstones in this Fl facies could be explained by invoking a glacial retreat during deposition. However, the lack of dropstones can also be explained if high-rates of deposition are postulated. High rates of deposition of fine-grained sediment (several metres per year) can be achieved from fall-out from overflow plumes associated with turbidity currents (eg. Mackiewicz et al., 1984; Powell and Molnia, 1989). If rates were high enough, the effect of ice-rafting would be masked, and laminated fine-grained deposits free from dropstones could accumulate in a glaciomarine environment.
Mackiewicz et al. (1984) suggested that such deposits may form at distances of between 0.5 and 2km from the grounding line, putting them as transitional between proximal and distal glaciomarine sedimentation.

iii) Planar laminated (Fl) to slightly undulatory siltstones are also developed locally within the proximal glaciomarine environment. The planar laminations in these siltstones are usually on a mm to few mm scale and pinch out over a few ms laterally in low-angle scours. These represent upper stage plane beds caused by either high flow velocities or thin flow thicknesses (eg. Allen, 1997).

3.9.3. Non-glacial sediment gravity flow facies association

The Fiq Member also contains sediments that formed in a non-glacial environment. As can be seen in Fig. 3.2, the Fiq Member is composed of distinct horizons of glacially influenced sediments that are separated by non-glacial facies in units up to 250m thick. Many of the processes that formed these sediments are linked to sediment gravity flows similar to those acting within the glaciomarine environment (Section 3.9.2). A number of facies occur within this non-glacial sediment gravity flow facies association. These are summarised in Table 3.4.

3.9.3.1 Conglomeratic facies (C)

Conglomeratic facies (C) of the non-glacial sediment gravity flow facies association occur mainly within sandstone (Sm, Sg) units (Plate 3.23), but also within mud/siltstone (Fm, Fl, Fr) dominated sections. They vary in thickness from few cm to 2m, and may pinch out laterally over less than a few metres, or maintain a relatively constant thickness for tens of metres. Conglomerates (C) are often interbedded with pebbly sandstones (Sp); these two facies commonly dominate the stratigraphy for up to 70m (eg. 50-120m Wadi Hajir, log WH1). Lower boundaries of the conglomerates are almost always sharp and erosive (Plate 3.24) and may show loading structures. Upper bedding surfaces are more variable, and although they are locally sharp, often fine-up gradationally into sandstone (Sm, Sg) or pebbly sandstone (Sp) facies. Indeed, the lower few cm of many of the graded sandstones is commonly conglomeratic. Clast size within different conglomerate beds varies from
few mm to 30cm, with most clasts falling in the size range of 1 to 5 cm. Clast shape is also variable, but rounded forms are more common than in the glaciomarine diamictites (Dm, Ds). Facetted clasts are extremely rare. Lithologies of clasts are also similar to those described from the diamictites, although different clast types may be preferentially concentrated in different beds. The matrix is variable and ranges from mud to fine-grained sandstone. The matrix of the conglomerates is generally fairly structureless.

The erosive bases of the conglomerates (which may cut down more than a metre) and their often lenticular shape preclude an origin from ice-rafting and current winnowing processes (Miall, 1985). The coarse-grained, clast-supported nature of these conglomeratic deposits combined with the presence of ripped-up and rafted clasts up to 80 cm in length suggest instead that a debris flow or high density turbidite origin is more likely (Nardin et al., 1979). Debris flows are plastic flows in which sediment and water are fully mixed and where any original bedding and lamination is largely destroyed (Stow et al., 1996). Although debris flows are commonly matrix supported (Nardin et al., 1979; Lowe, 1982), fine sands and muds may be sheared from a transitional debris flow at the debris-water interface, and may be expelled into an overlying turbulent suspension (Wright and Anderson, 1982). Such flow-stripping processes acting on debris flows could account for many of the structureless clast-supported conglomerates. However, where the upper boundary of the conglomerate gradationally fines up into Sm or Sp sandstone facies, an origin from a turbiditic flow seems more likely. It is quite possible that transitional flows or a combination of both debris and high density turbiditic flows could have been responsible for depositing this conglomeratic facies.

Debris or turbiditic flows may be triggered by a number of mechanisms including earthquakes, storm waves and depositional oversteepening. The presence of large granite clasts and the occasional facetted clast within some of the conglomeratic beds (e.g. 110m log WH1) suggests that the precursor to some of these C deposits may have been ice-rafted diamictite. However, the paucity of facetted clasts and the association of these C beds with other non-glacial facies suggests that no active glacial influence was occurring during their formation.
3.9.3.2. Pebby sandstone facies (Sp)

This facies is commonly associated with the conglomeratic facies, and indeed, is transitional between conglomerates (C) and sandstone (Sm/Sg). It is distinguished from the diamicite facies by the lack of any significant proportion of mud within its matrix and by the presence of grading and rafted mud clasts.

The facies is not particularly widespread but typically forms units up to 10m thick (most commonly 1-2m thickness) (Plate 3.25). Basal contacts are commonly erosional, cutting into underlying sandstone or mudstone. Pebby sandstones may also gradationally overlie erosively-based conglomerate, however. Where the basal contact overlies mudstone, the pebbly sandstone (Sp) is loaded, and the lower part of the bed is locally disrupted by soft sediment deformation in the form of convoluted beds up to 1m high (Plate 3.26). Pebby sandstone (Sp) beds may pass gradationally upwards into sandstone (Sm, Sg) free from pebbles, or more commonly are erosively overlain by conglomerates (C) that in turn pass up into pebbly sandstones (Sp). Clasts from few mm to 10cm in size (and rarely up to 70cm) occur within a finer sandstone matrix and usually form 5-20% of the rock. The clasts range from being randomly distributed throughout the sandstone to being concentrated into matrix-supported layers 1 or 2 clasts thick. Clasts are usually subrounded and include granite, rhyolite, tuff, shale, and sandstone. Wispy-edged, elongated laminated shale clasts (<80cm long) occur and lie approximately parallel to bedding. Their shape suggests they were ripped-up when only partially lithified. Slight fining-upwards of the matrix often occurs, as does a slight upwards reduction in the size and number of the larger clasts (coarse-tail grading).

The erosive base to these pebbly sandstone (Sp) deposits, and the ripped-up and rafted clasts suggest that they originated from a debris or high-density turbiditic flow, similar to that described above for the conglomerates (Section 3.9.3.1). The fact that they are matrix supported rather than clast-supported may indicate a low proportion of pebble-sized material in the precursor, or that the majority of the coarser material was deposited earlier. An alternative possibility is that the flow-stripping of debris flows envisaged to explain the clast-supported nature of the conglomerates described above did not occur to the same extent (Wright and Anderson, 1982). The pebbly sandstone (Sp) deposits would then
simply record a matrix-supported debris flow. Where grading occurs, however, it seems likely that turbiditic flow processes were dominant.

3.9.3.3. Massive/graded sandstone facies (Sm/Sg)

Massive and graded sandstones (Sm and Sg) form a major part of the non-glacial stratigraphy of the Fiq Member and are represented in most of the logged sections. Particularly good examples occur in Wadi Sahtan (680m log WS1), Wadi Bani Awf (0-40m log WBA1), Wadi Hajir (360m log WH1), and Wadi Mistal (390m log WM1). Sm and Sg facies were probably formed by similar processes and are interpreted together here. As their lithological characteristics differ, however, they will firstly be described separately.

i) Massive sandstone facies (Sm)

Massive sandstone (Sm) forms a relatively common facies within the non-glacial sediment gravity flow facies association of the Fiq Member, occurring at a number of different levels in the stratigraphy. It comprises homogeneous sandstones with no internal stratification or structure. Massive sandstones (Sm) form beds varying from few cm thickness to >5m with commonly well-defined, sharp bases. These beds are locally amalgamated, forming units up to 40m thick with minor intercalated siltstones <1cm thick (Plate 3.27). Individual beds are thinner and are often associated with mud/siltstone (Fm, Fl, Fr) dominated units. Occasionally, massive sandstone (Sm) beds have rippled tops.

Some sandstone (Sm) beds can be traced laterally for hundreds of metres to kms and maintain very constant thickness over these distances, suggesting that they form sheet-like deposits. The same sandstone bodies can be found in different sections several kms away, where they are often associated with conglomeratic beds (C) or with finer siltstones (Fl, Fm, Fr). In Wadi Sahtan (eg. 670m log WS1), where the lower bedding surfaces of these sheet-like sandstones are exposed, they occasionally display flute marks up to 15cm long and few cm deep (Plate 3.28). These indicate palaeocurrents towards the east.
In terms of grain-size, the sandstones (Sm) vary from fine-grained to coarse-grained, with the finer sandstones usually occurring within siltstone dominated units. The sandstones (Sm) are generally well-sorted with subangular to subrounded grains mainly composed of quartz (~80%) with less common microcrystalline lithics, feldspars and mica. Rare, larger lithic clasts and quartz grains up to 0.5cm in size may also occur within the sandstones.

ii) Graded Sandstone Facies (Sg)

Graded sandstone (Sg) is again a fairly common facies within the non-glacial sediment gravity flow facies association of the Fiq Member, occurring at a number of different levels. Note that although some of the pebbly sandstones (Sp) may show some fining-up they are not included in this facies, as due to their pebbly nature described earlier, they are easily distinguishable from the purer graded sandstones (Sg) referred to here.

These sandstones (Sg) form beds varying in thickness from few cms to few ms, and locally dominate the stratigraphy for up to 30m. Alternatively, they may occur interbedded with other more prominent facies. In places, graded sandstones occur in channel-like bodies up to 30m and more across (Plate 3.29). The lateral shape of each bed, is however, commonly hard to follow beyond a few tens of metres, and many of the beds may be more laterally continuous than this. This facies is commonly associated with other sandstones (Sm and Sp), conglomerates (C), and siltstones (Fm, Fl, Fr) – within which it often forms a much more minor component.

Graded sandstones (Sg) have sharp to erosive bases that may be rich in granular or even pebbly material (Plate 3.30); indeed, the facies commonly gradationally overlies erosively-based conglomerates. Scour and loading structures are common on lower bedding surfaces and the coarse fraction at the base is often discontinuous on a dm scale, filling shallow scours/loads. The coarse basal material tends to fine up throughout the graded bed, but is commonly most pronounced over the basal part of the bed, especially when the base is rich in granular or pebbly material. In the basal cm of some of the sandstones, reverse grading can be seen before fining-up is established.
The upper bedding contacts tend to be more transitional, often fining-up gradationally from convoluted sandstone into siltstone (Fm, Fr). This siltstone often contains asymmetrical rippled structures and is in turn overlain by another erosively-based fining-up sandstone.

Grains within the graded sandstones are mainly subrounded, but range from subangular to subrounded. Quartz forms ~80% of these grains and may show undulose extinction. The rest of the grains are composed of polycrystalline quartz, very fine-grained lithics (felsic volcanics) and rare feldspars. Rafted mudflakes up to few cm long and few mm wide commonly occur within the graded sandstones, as do irregular bands <1mm thick of clay-rich material that often marks the tops of smaller scale fining-up cycles within each sandstone bed.

**Interpretation**

The sharp bases and lateral extent of both the massive (Sm) and graded (Sg) sandstones described above suggest they represent the deposits of turbidity currents. Turbidites are suspensions of sediment that are sustained by fluid turbulence (eg. Bouma, 1962; Stow et al., 1996). Within an active turbidity current, the upwards components of turbulent fluid motion provide the main grain support mechanism and this behaviour can be sustained over long distances (tens of kms (Stow, 1994)) through a feedback loop known as autosuspension, which is most effective in fine grain sizes (Bagnold, 1962; Southard and Mackintosh, 1981). True autosuspension may be rare in nature (Clark and Pickering, 1996), but turbidity currents are able to operate on very low gradients and may transport sediment of a wide range of grain sizes from gravel to the finest clays and at a wide range of grain concentrations. Sm and Sg sandstones represent the lower ‘A’ division of the Bouma classification (Bouma, 1962). They probably formed from flows in the upper part of the upper flow regime where sediment arrived at the bed too quickly for it to be reworked into any bedform or lamination. Rapid deposition would also account for the common occurrence of water escape structures and convoluted bedding observed in the Sg sandstones (Stow et al., 1996). Another possibility is that the sandstones formed from the occurrence of sustained flow through which sand accumulated continuously though a basal layer of hindered settling (Kneller and Branney, 1995). Grading occurs where grains of differing sizes were available. Pebbly bases to some Sg beds and the presence of rafted
clasts within beds suggests high-density turbidity currents or hyperconcentrated flow. Reverse grading in the basal cm of some of the sandstone beds demonstrates localised development of kinetic sieving or the action of dispersive pressure in a highly concentrated traction carpet. Where Sm and Sg sandstones occur as channel-like bodies they probably represent the deposits of turbiditic flows constrained within subaqueous channels.

3.9.3.4. Rippled sandstone and siltstone facies (Sr and Fr)

Rippled facies (Sr/Fr) often occur at the top of Sm or Sg sandstone beds (Plate 3.31), but also occur interbedded with siltstones (Fm, Fl) where they may form ripple lenses (Plate 3.32) or as slightly thicker beds 1-10cm thick with sharp bases (Plate 3.33) that are laterally continuous over tens of metres. Rippled siltstone (Fr) dominated units containing lenses or thin beds of rippled sandstone (Sr) also occur and may form thicknesses >100m. These thick Fr siltstone units are dominated by ripples with heights <0.5cm and wavelengths <10cm. Fr siltstone facies also occur within thick, laminated siltstone (Fl) dominated units (Plate 3.34). These commonly contain fluid escape and loading structures. Thick Fr/Fl siltstone units commonly coarsen up gradually into Sr facies containing climbing and symmetrical ripple forms and cross-lamination showing bi-directionality (eg. 540-600m log WS1; 420m log WH1; Plates 3.35, 3.36). In general, however, ripples within Sr facies are asymmetrical and are usually <1cm in height with a wavelength of <9cm (ripple indices of between 8 and 14). Internal structure is rarely preserved. On some ripple crests, off-shooting laminations or washover structures can be seen (Plate 3.33). The sandstones (Sr) are composed of well-sorted, dominantly subrounded to rounded grains that are made up of 95% quartz with minor feldspar and mica. They have sharp bases that may exhibit small (~0.5mm) load structures into the underlying siltstones, as well as well-defined upper surfaces. Carbonate is locally developed between some of the sandstone grains.

Palaeocurrent directions taken from these rippled Sr and Fr facies are variable (Fig. 3.14). Throughout most of the stratigraphy logged in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir asymmetrical ripples suggest currents generally moving towards the east. Water escape structures within rippled facies in these wadis also indicate either palaeoslopes or
overlying shear towards the east. In Wadi Mistal, however, palaeocurrents (although less common) are generally indicated as flowing towards the west.

Similarly to the Sm and Sg sandstones, these rippled Sr and Fr facies are the products of turbidity currents. Their rippled nature and common occurrence at the top of Sm and Sg facies suggests that they represent the ‘C’ interval of the Bouma classification (Bouma, 1962). The Sr and Fr facies described above therefore probably represent fallout of sand or silt from suspension while lower flow regime current ripples were moving on the bed. Where fallout was rapid, climbing ripple cross-lamination occurred. This interval may also include convolute lamination attesting to short-lived liquefaction (Stow et al., 1996).

Sections of the stratigraphy in which silt-sized material dominates for thicknesses of >100m probably represent deposits derived from fallout from turbiditic tails/plumes. The lack of any coarse-grained material within them suggests either that only fine-grained material was being derived (eg. Piper, 1978; Stow, 1979; Kelts and Arthur, 1981), or that flow-stripping was occurring and the coarser fraction was deposited earlier or elsewhere. The latter possibility seems more likely given the large stratigraphic thickness these siltstones comprise, and suggests that they were either distal deposits, or that they were the products of overbank turbidites, with the coarser fraction remaining in submarine channels not observed at this level in outcrop. In many of the localities where such turbiditic siltstones can be observed, they exhibit a gradual coarsening up (eg. 560-600m Wadi Sahtan; 200-240m Wadi Bani Awf; 410-425m Wadi Hajir). The fact that this upward coarsening is usually accompanied by an increasing frequency of ripple forms, a reduction in the amount of background sedimentation, and also by the development of climbing ripples suggests that an increase in energy and a relative shallowing was occurring. The climbing ripples indicate high concentrations of sediment coming from rapid fallout and attest to continuing high rates of deposition (Collinson and Thompson, 1989). Where the upward coarsening culminates in coarse siltstone containing ripples with symmetrical profiles, undulatory bases and bi-directional cross-lamination, the action of waves and probable shoreface deposition is indicated (De Raaf et al., 1977). In instances where such shallowing events are overlain by proximal glaciomarine deposits (eg. 600m log WS1), it is possible that they were caused by a lowering of sea-level (forced regression) associated
with colder temperatures and an increased volume of ice on land. This is conjectural however, as no independent evidence was observed to verify this suggestion.

### 3.9.3.5. Massive and laminated mud/siltstone facies (Fm and Fl)

Mud/siltstone facies (Fm/Fl) also occur in the non-glacial sediment gravity flow facies association. They occur in a number of different forms:

i) Massive dark grey shales (Fm) occur at a number of horizons and commonly abruptly overlie coarser-grained deposits before passing up into laminated mud/siltstones (Fl). The shale units are up to 10m thick, but may be interspersed with thin sandstone beds or laminated siltstones. They are very fine-grained in character and composed dominantly of muddy material. The overlying laminated shales/siltstones (Fl) may form units up to 100m thick. The lamination is on a mm to few mm scale and is picked out by colour differences (dark grey to lighter grey). These laminations are often laterally persistent for tens of metres despite their thin nature. Few further bedding structures occur. Such deposits are often overlain transitionally by slightly coarser, laminated siltstones containing a slightly undulatory lamination. These in turn often pass up gradationally into rippled siltstones (Fr) in an overall coarsening upwards sequence.

The very fine-grained nature of the shale deposits (Fm) and the lateral continuity of the laminae in the Fl facies suggest that these sediments are the products of settling from suspension, possibly from fall-out from turbiditic plumes/tails or river plumes/underflows (Stow et al, 1996).

ii) Massive siltstones (Fm) and laminated siltstones (Fl), commonly dark grey in colour, also occur (Plates 3.37). They are more frequently associated with interbedded sandstones (Sm, Sg) than the more shale-rich deposits described above and form beds <50cm thick with reasonably well defined bases and tops within packages dominated by sandstone (Sm, Sg) facies (eg. 160-200m log WH1). Sandstone grade material forms up to 30% of these siltstone facies. Soft sediment deformation forming convoluted bedding is common.
These Fm and Fl facies probably represent the upper parts of turbidite deposits that were either fully or partly homogenised by fluidisation and dewatering (Nardin et al., 1979; Stow, 1994).

iii) Coarse laminated siltstones (Fl), often associated with thinly-bedded sandstones occur and contain laminations that pinch out under very low-angle scours on a m to few m-scale (Plate 3.38). More laterally persistent laminations also occur (Plate 3.39) and are locally disrupted by soft sediment deformation in the form of convolutions and flame structures.

These coarser Fl siltstone facies probably represent fine-grained fall-out from suspension from the tail of the turbidity current responsible for the deposition of the thinly-bedded sandstones associated with them (Stow et al., 1996). Where planar laminations pinch out, the Fl facies represents upper stage plane beds due to either high flow velocities or shallow flow depths (eg. Allen, 1997).

The descriptions and interpretations above illustrate that a number of different facies within the non-glacial sediment gravity flow facies association can be attributed to the passage of turbidity currents. For example, in places pebbly-based massive sandstones dominate (eg. 0m Wadi Bani Awf) whereas in others siltstones are the prevalent facies type (eg. 780m Wadi Mistal). This suggests that both low and high density currents were acting. At some levels of the stratigraphy these varying facies can actually be shown to be lateral equivalents. In Wadi Bani Awf and the western part of Wadi Sahtan in Unit F4, for example, sheet-like massive medium-grained sandstones occur interbedded with 0.1 to 1m thick conglomeratic beds (0-40m log WBA 1). The conglomerates probably represent the deposits of debris flows or high-density turbidity currents, with the massive sandstones representing the A division of Bouma sequences. At the same level further to the north (670-770m log WS 1), massive sandstones again occur, but here do not contain any material coarser than gravel grade and are interbedded with planar laminated siltstones. Bouma intervals A-E are all developed at least locally in some beds at this level, and all the sandstone beds can be seen to have sharp bases and persistent thickness laterally for well over a hundred metres. Ripple structures and tool marks also occur and indicate palaeocurrents towards the east and south-east. These generally finer grained deposits in the north of Wadi Sahtan at the same level as the coarser material to the south can be
explained if a submarine channel is postulated running through Wadi Bani Awf (Fig 3.15). In this scenario, most of the coarse-grained and pebbly material would be concentrated in the channel as flows moved through it towards the east, resulting in the deposition of conglomerates and sandstones. When flow rates were high, material would be likely to flow over the channel sides and lead to the deposition of overbank turbidites. These overbank turbidites would be composed of flow-stripped finer material (the coarser fraction remaining within the confines of the channel), which is reflected in the preservation of more of the Bouma divisions and the presence of siltstone and mudstone interbedded with the sandstones (log WS 1).

At the same level (Figs. 3.2, 3.15), but further to the east in Wadi Hajir, further evidence for this channel can be seen. Here, in the north of the wadi, a series of fining-up gravelly-based sandstones occur in which individual beds are generally 5-20cm thick. Coarse-grained deposits dominate for the lower 15m, but these pass up into siltstones and few cm-thick rippled sandstone beds that dominate for approximately 10m, before coarser-grained deposition is again established. The upper parts of the fining-up sandstone beds commonly contain rafted mudflakes and suggest that they were deposited by hyperconcentrated flows, either as high-density turbidity currents or as distal members of debris flows. Ripples and fluid escape structures are also common and indicate palaeocurrents to the north-east. The presence of the gravelly material suggests that these deposits formed in a submarine channel or lobe. It seems likely that these sediments were deposited further downslope within the same channel that was postulated in Wadi Sahtan and Wadi Bani Awf; the thinner beds, generally finer grain size and palaeocurrent directions suggest more distal deposition.

Further to the south in Wadi Hajir, low-density turbidites dominated by siltstone and fine-grained sandstone are preserved. These probably represent overbank turbidite deposits linked to the same channel. Their position further downslope to those recorded in Dabu’t in Wadi Sahtan is reflected by their finer-grained nature.

In Wadi Mistal, lateral variation of turbidite deposits can again be observed, reflecting a proximal-distal relationship. In the east of Wadi Mistal near Wijmah (370m log WM 1), massive sandstone beds up to 10m thick containing rafted mudflakes occur. These are
interbedded with minor siltstones, but the stratigraphy is dominated by Bouma A divisions suggesting deposition by hyperconcentrated/high-density turbidity currents. Further to the west at the same level, the stratigraphy is dominated by planar laminated siltstones and thin fine-grained sandstone beds. As these sediments are mostly composed of Bouma D and E divisions, they were probably deposited by fallout from suspension in low-density turbidity currents. What palaeocurrent indicators there are in Wadi Mistal at this level suggest palaeocurrents to the west. The deposits logged near F iq are therefore quite possibly distal equivalents of those logged near Wijmah.

Palaeocurrent and palaeoslope indicators in Wadi Mistal occurring in the opposite direction to those in Wadi Sahtan and Wadi Hajir are not restricted solely to this level. The consistent occurrence of these differing directions suggests that during F iq deposition sediment was being derived from both the west and the east. This will be discussed further in Section 3.10.

3.9.3.6. Cross-stratified sandstone facies (Sx)

Cross-stratified sandstones (Sx) constitute a very minor facies only developed locally within the F iq Member in Wadi Hajir, Wadi Sahtan, and Wadi Bani Awf. In many instances it is only poorly developed and its presence is equivocal.

In the massive sandstone (Sm) unit at ~1030m in Wadi Sahtan (log W S l) trough cross-sets 20cm high and 60cm wide, overlain by oversteepened cross-sets 15cm high with some top-set preservation are apparent at one locality. The cross-bedding is picked out by local weathering of the rock and is not seen anywhere else within this unit.

Lower down in the same section at 850m (log W S l), two sharp-, pebbly-based, fining-up sandstones (Sg) occur. Within the coarse-grained portions of these sandstones immediately above the pebble-rich base, planar cross-sets 30cm high occur. These are overlain by finer-grained rippled sandstones. At a similar level in Wadi Bani Awf (136m log WBA1), low-angle to trough-like cross-beds occur up to 10cm high (Plate 3.40).
Similarly, in Wadi Hajir the upper parts of graded sandstone beds rarely exhibit hints of poorly-preserved cross-bedding. This is often planar with steeply dipping (30°) cross-strata. Also in Wadi Hajir, within the sandstone at ~270m (log WH1), cross-sets up to 50cm high occur within medium-grained to coarse-grained sandstone. These are overlain by coarse-grained to pebbly sandstone that is locally erosive and loads into these underlying sediments.

As already mentioned, most of the structures just described are only poorly and locally developed making further analysis difficult. A detailed interpretation is therefore not really possible. Many of the cross-bedded sandstones occur associated with sediment gravity flow deposits. As these were being deposited offshore, it seems likely that the cross-bedding was also formed in deep water. One possibility is that the cross-bedding was formed during storm events, when strong currents could cause dunes to form and migrate in deeper water offshore.

3.9.3.7. Volcaniclastic sandstone facies (Sv)

The volcaniclastic sandstone facies (Sv) is distinguished from other sandstones (Sm, Sg, Sr) on the basis of its immature, volcaniclastic nature. It occurs near the base of the Fiq Member and has been recorded in Wadi Mistal (eg. 25-80m log WM1), Wadi Sahtan (20-200m log WS1) and Wadi Hedak (not logged due to its poorly exposed and deformed nature) (Figs. 3.1 and 3.2). Individual beds from few cm to 10m thick occur, although bedding on a dm to m scale is most common. Sv beds may be conglomeratic, massive or graded and commonly contain pebbles and rafted mudflakes within them. Rare ripple marks also occur in some of the finer-grained beds.

Volcaniclastic sandstone (Sv) beds maintain relatively constant thickness laterally for tens of metres and possibly more, although this is often hard to trace due to poor exposure of the facies (Plate 3.41). Basal contacts are sharp to erosional and often overlie siltstones into which loading can occur. Rare flame structures are also present in the lower few cm of some beds in Wadi Mistal; where present they suggest palaeoslopes towards the west. Soft sediment deformation, especially in the fine-grained upper parts of graded beds is
common in the form of convolutions in fine-grained sandstones, siltstones and mudstones (Plate 3.42).

The more coarsely-grained members of this Sv facies are poorly sorted with subangular to subrounded grains varying from 0.2mm to >1mm in size comprising most of the rock. These grains are composed of 30-80% mostly tuffaceous, lithic material. Pebbles commonly form the lower few cm of fining-up beds. Pebbles are usually 0.5cm to few cm size, but within some of the thicker (>2m) conglomerates associated with this facies, occur up to 30cm in size. The pebbles are mostly subrounded to rounded and are composed of tuffaceous material. Within the more pebble-rich graded beds, large (up to 1m long) elongated mudstone clasts occur. These occur in the middle or towards the top of beds and have irregular wispy edges suggesting they were only semi-lithified when they were incorporated into the bed.

Finer-grained Sv sandstones composed of similar tuffaceous material occur and are commonly rippled and associated with siltstones. These sandstone beds are usually only a few cm thick and have asymmetrically rippled tops. In Wadi Mistal the ripples indicate palaeocurrents flowing towards the west.

Although these Sv sandstones are rich in volcanic material, it seems unlikely that they were deposited by transport processes resulting directly from volcanic activity. This is indicated by the rounded nature of the clasts and grains indicating significant transport and also by the relatively high proportion (35-70%) of non-volcanic grains (Orton, 1996). Further evidence that they are reworked volcanic deposits comes from preliminary U-Pb dates that have been obtained from zircons within them as part of this study. These give dates (c. 755-765Ma from 25m log WM1; Figs. 4.2, 4.3, Chapter 4) that are significantly older than the date of 711.8±1.6Ma obtained from the underlying Ghubrah Formation. This suggests that the zircons dated from the volcaniclastic sandstone (Sv) of the Fiq Member have been reworked and the deposits they are in did not form directly from volcanic activity. Many features of these deposits, such as their lateral persistence, and the presence of rafted mudflakes and Bouma divisions, suggest that they were the products of sediment gravity flows (both debris flows and turbidites), with the finer ash-like material representing fall-out from turbiditic plumes. In many respects, pyroclastic gravity flows
are governed by similar processes and produce broadly similar deposits to sediment gravity flows (Fisher and Schminke, 1994; Orton, 1996). However, as the term ‘pyroclastic’ (and related terms such as ‘tuff’) is restricted to deposits ‘formed by explosive volcanic activity and deposited by transport processes resulting directly from this activity’ (Cas and Wright, 1987), these sediments cannot be referred to as pyroclastic flow deposits. They are simply sediment gravity flow deposits that contain a high proportion of volcaniclastic and tuffaceous material. The presence of so much volcanic material suggests that a volcanic succession must have been exposed as a source area. Palaeocurrents in Wadi Mistal, and a general fining and thinning of these deposits to the west implies that the source area (at least for Wadi Mistal) was to the east, and may well have been linked to the volcanism recorded in the Saih Hatat region (Le Metour et al., 1986; Villey et al., 1986). Palaeocurrents are lacking from the section logged through the volcaniclastic-rich deposits in Wadi Sahtan, and so although it is possible that these were derived from a different source, this cannot be unequivocally shown.

3.9.3.8. Carbonate facies (L)

Carbonates form a very minor part of the non-glacial sediment gravity flow facies association of the Fiq Member, but are particularly interesting as they form a distinctive lithology in the dominantly siliciclastic succession. Carbonate is restricted to two main horizons within the Fiq Member: a lower horizon in Unit F1 and F2, and an upper horizon in Unit F6 (Fig. 3.2). These are dealt with separately below.

i) Lower carbonate horizons in Units F1 and F2

In the lower 25m of the Fiq Member in Wadi Mistal and even locally developed below the Saqlah Member (20-45m log WM1), well-bedded carbonate units up to 1m thick occur and are separated by volcaniclastic deposits (Plate 3.43). In Wadi Mu’aydān, further to the south, similar carbonates are found within planar laminated tuffaceous siltstones (120-160m log MU1). The carbonates are orange-brown to tawny in colour and are bedded on a cm to few cm scale, often with interbedded siltstones. The carbonates in Wadi Mistal become increasingly impure and silty upwards in the stratigraphy. Planar laminations occur in the lower beds, but further up, rippled laminations and rippled tops to beds are
more common (Plate 3.44). In thin-section, detrital clastic grains can be seen forming up to 50% of the rock. These are commonly quartz grains with embayed edges (Plate 3.45). Most of the beds have been degraded and are coated with brown clay minerals that are almost opaque. Some of the carbonate has also been replaced with silica. Where the degradation has not proceeded too far, small clusters of 5-10µm sized microspheres can be seen (Plate 3.46). Other structures that can be seen within these carbonates include slight fining-up cycles on a cm scale and poorly developed thrombolitic-like textures.

The small clusters of 5-10µm sized microspheres that can be seen in the lower carbonate beds (20-25m log WM1) possibly represent the remains of colonial coccoid cyanobacterial unicells and/or capsules (Kellerhals and Matter, 2000; David Wright, pers. comm., 1999). The presence of detrital quartz that has been embayed suggests trapping and binding in alkaline conditions (David Wright, pers. comm. 1999). This suggests that the lower beds probably represent cyanobacterial stromatolites. The rippling and grading present in the increasingly silt-rich overlying carbonate beds indicate that they were probably deposited by density currents and suggest higher energy events during which stromatolitic limestones were partly eroded and redeposited by turbidity currents. The ripples indicate that these currents were flowing from the east to the west.

This lower carbonate horizon from near the base of the exposed Fiq Member was sampled for preliminary stable isotope analysis in both Wadi Mistal and Wadi Mu’aydin (analytical methods are described in Appendix C). A total of 43 samples from Units F1 and F2 of the Fiq Member were analysed for stable carbon and oxygen isotopes (6 from Wadi Mu’aydin and 37 from Wadi Mistal) (Table 3.5). Each of the samples was selected as ‘least altered’ based on location away from areas of veining, good fabric preservation, and petrographical study. 13 of the samples from Wadi Mistal were also analysed for major and minor elements (Table 3.5). A fuller discussion of the significance of isotopic trends and diagenetic indicators is presented in Chapter 5, Part II.

The δ13C results from Wadi Mu’aydin show a general trend towards increasingly positive values over the 32m of section they span, varying from −3.24‰ at the base to 0.11‰ at the top (Table 3.5). Although an elemental analysis was not carried out on any of these samples, all but one of the samples have δ18O values >-5‰. This suggests that the isotopic
signature has not been significantly altered by diagenesis (Kaufman and Knoll, 1995 – see Chapter 5).

In Wadi Mistal, the δ¹³C signature shows an initial upwards move to more positive values, followed by a negative shift. More positive values are then re-established within the uppermost carbonate unit (Table, 3.5; Fig. 3.16). There is significant variation within the δ¹³C signal from Wadi Mistal, both between carbonate units and within them. For example, within the 1.2m thickness of the second carbonate unit, δ¹³C values vary by 7‰ (Fig. 3.16). To try and investigate whether this large variation reflects the primary isotopic signal or whether it has been caused by later alteration, some preliminary diagenetic screening of the samples was undertaken.

The δ¹⁸O signature from Wadi Mistal is variable but values >-5‰ are recorded in the vast majority of the samples from the lower three carbonate units, suggesting little diagenetic alteration (Kaufman and Knoll, 1995). δ¹⁸O values <-10‰ occur in the uppermost carbonate unit, however (Fig. 3.16). Although the δ¹³C values within this unit are all positive, covariation between δ¹⁸O and δ¹³C (correlation coefficient, r = 0.857, Fig. 3.17a) suggests the lower values of δ¹³C may have been diagenetically reset in a negative direction. There is no overall covariation between δ¹⁸O and δ¹³C within the samples from Wadi Mistal, however, suggesting no consistent alteration of the isotopic signatures.

Plots of Mn vs Sr, δ¹³C vs Mn/Sr, δ¹³C vs Mn, and δ¹³C vs Fe (Figs. 3.17b-e) were also made as further diagenetic screens (for explanation of elemental plots see Chapter 5). The majority of the Mn/Sr ratios are >10 (Table 3.5, Fig. 3.17b), suggesting that diagenetic alteration may have occurred (Kaufman and Knoll, 1995). No consistent covariation occurs between any of the elements, however, again indicating no consistent shift in values due to diagenesis.

What the isotopic curve from this portion of the stratigraphy actually signifies is debatable. If the shifts in the isotopic curves represent basinal changes, they may reflect basin-wide changes in organic productivity and/or organic deposition (eg. Marshall, 1992 – see Chapter 5). However, the widely variable signal and high Mn/Sr values from Wadi Mistal suggest alteration of the isotopic signature there. The fact that carbonates are not found
elsewhere at this level in the stratigraphy means that the signature cannot be compared to other localities within the Jebel Akhdar. This makes it hard to distinguish between trends reflecting basinal changes in the isotopic signature of seawater and local trends that are more likely to be linked to diagenesis. No further conclusions have therefore been drawn from this preliminary study of the isotopic signatures from the carbonates of the lower Fiq Member.

**ii) Upper carbonate horizon in Unit F6**

Well-bedded carbonate also occurs above the conglomerate at 1056m in Wadi Sahtan (log WS1; Plate 3.47) and at the same level above the conglomerates at 425m in Wadi Hajir (log WH1; Plate 3.48) and at 251m in Wadi Bani Awf (log WBA1). This carbonate is brown-coloured, has well-defined upper and lower boundaries and occurs either within or at the base of very fine-grained deposits. In Wadi Sahtan, two beds 10-30cm thick are present, separated by ~20cm of laminated grey siltstone. These beds can be traced for over 100m laterally, but beyond this accessibility becomes a problem. The lower parts of each of these beds contain slightly crinkly laminations that pass up into less well-laminated material. In thin-section, it can be seen that these are fairly pure carbonates containing both calcite and dolomite with detrital grains forming <2% of the rock (Plate 3.49a). Subhedral dolomite crystals 0.1 to 0.2mm in size are set in an anhedral calcite cement. The calcite cement is formed of similar sized crystals, but these are often optically aligned and go into extinction at the same time (Plate 3.49b). Some ferroan calcite is also present.

A few samples from these carbonate beds in Wadi Sahtan were run for exploratory stable isotope analysis. The results are shown in Table 3.5. Highly negative values for both $\delta^{13}C$ and $\delta^{18}O$ were obtained (as low as $-13\%o$ for the carbon and $-15\%o$ for the oxygen). This suggests that the results have been significantly affected by later diagenesis and do not represent the original isotopic signal (eg. Kaufman and Knoll, 1995).

The carbonate bed that overlies the conglomerate at the same level in Wadi Hajir and Wadi Bani Awf differs from that described above from Wadi Sahtan in a number of ways. It occurs as a single bed up to 10cm thick that is only locally developed and may pinch out over 10m laterally. In thin-section a series of cycles can be seen on a cm scale. Dark, silt-
rich bands up to 3mm thick are overlain by bands of similar thickness containing siliciclastic grains held in a carbonate cement. These siliciclastic grains form ~40% of the band, are 0.2mm to 1mm in size and are dominantly composed of quartz, with minor feldspars and lithics (Plate 3.50). The grain-size decreases upwards and the band passes into purer carbonate that dominates for 1cm or so before it is sharply overlain by the silt-rich layer again and the cycle is repeated. This pure carbonate layer contains crystalline carbonate that forms an interlocking mosaic with each crystal <0.5mm in size and with an irregular boundary.

The presence of graded laminae in this bedded carbonate in Wadi Hajir and Wadi Bani Awf in which siliciclastic-rich bases fine-up into purer carbonate suggests that the carbonate here is the product of density currents. In the carbonate at the same level in Wadi Sahtan, however, these features were not observed. The highly negative $\delta^{13}C$ and $\delta^{18}O$ isotope values from this bed in Wadi Sahtan suggesting diagenetic resetting are consistent with the optically aligned calcite crystals, which further suggest neomorphism/recrystallisation. This hinders interpretation, but as the carbonate beds here are composed of purer carbonate than in Wadi Hajir and Wadi Bani Awf, and as the lower parts of the beds contain ‘crinkly’ (possibly microbial) laminae, they may represent cyanobacterial stromatolites similar to those observed at the base of the Fiq in Wadi Mistal. They were possibly the source of the carbonate present in the beds at the same level further to the west.

Interestingly, the carbonate beds of this horizon are associated with transgressive deposits. Although not developed across the whole Jebel Akhdar, glacial deposits just below the carbonates in Wadi Sahtan suggest that this transgression may have been linked to a postglacial sea-level rise. The occurrence of carbonate above glacial deposits is also recorded by the Hadash Formation, which lies directly on top of the last glacial horizon of the Ghadir Manqil Formation. The reasons why postglacial conditions should promote carbonate precipitation are discussed in more detail in Chapter 5.
3.9.4. Non-glacial shallow marine facies association

The shallow marine shoreface facies association forms a minor part of the non-glacial sediments of the Fiq Member. It contains a number of different facies summarised in Table 3.4.

3.9.4.1. Conglomeratic facies (C)

Conglomerates (C) included in the shallow marine facies association typically occur above wave-rippled facies at the top of coarsening-up sequences and are abruptly overlain by much finer-grained deposits such as dark grey shales (Fm) and locally developed carbonate beds (L) (eg. 423m log WH1). The conglomerates (C) form beds <1m thick that are usually particularly clast-rich (Plate 3.51). They contain commonly well-rounded clasts up to few cm in size. The conglomerates tend to be laterally persistent and maintain consistent thickness for at least several kms. For example, the bed at 423m log WH1 can also be seen at 1056m log WS1 and at 251m log WBA1.

The position of these conglomerates (C) between underlying relatively shallow water deposits and overlying deeper water deposits suggests that they probably represent transgressive lag deposits that have been winnowed of their finer material during flooding. The presence of such conglomerates at the same level in a number of different wadis across the Jebel Akhdar indicates a significant flooding event at the top of Unit F6a (Fig 3.19j).

3.9.4.2. Rippled sandstone facies (SrW)

Rippled sandstone facies (SrW) of the shallow marine facies association occur in Wadi Sahtan overlying dark grey phyllitic siltstones (Fm) (632m log WS1). At this level an extensively rippled sandstone (SrW) unit 6m thick occurs. This is stacked full of ripples commonly exhibiting a symmetrical, trochoidal shape in profile, with straight crests occasionally showing tuning-fork bifurcations (Plates 3.52, 3.53). Typical ripples have ripple indexes of 9-12 (wavelengths of 6-10cm, heights of 0.5 to 1cm), and crests striking between 140 and 165 degrees. Grain size varies from 187.5 to 375μm and tends to get
finer towards the top of the unit where a few of the ripples display flattened or double-crested tops. Within this unit rarer asymmetrical ripple and climbing ripple forms also occur. This rippled sandstone unit is overlain by massive (Sm) sandstone beds 10-15cm thick interbedded with rippled heterolithics (Section 3.9.4.3).

The symmetrical and commonly trochoidal shape of many of the ripples, combined with their straight-crested nature displaying tuning-fork bifurcations indicates that they were formed by the action of waves (eg. Conybeare and Crook, 1968; Allen, 1970; Allen, 1982; Collinson and Thompson, 1989). Storm wave base, even in the most extreme conditions, rarely, if ever, exceeds 200m, while fairweather base is considerably shallower (10-20m) (Johnson and Baldwin, 1996). As many of the sandstone beds in Wadi Sahtan are filled with wave ripples, suggesting they were being continually worked by fairweather waves, it seems likely that they were deposited in depths of less than 20m within a shoreface setting (eg. Boggs, 1987; Reading and Collinson, 1996). On some bedding surfaces towards the top of the wave-rippled sandstones unit, poorly-developed flattened or double-ripple crests can be seen, indicative of shallowing or emergence (Collinson and Thompson, 1989). This suggests even shallower water deposition, probably within the intertidal zone. The crests of the ripples trend very consistently between 140 and 165°, and as wave-ripples commonly lie parallel or sub-parallel with the coastline due to wave refraction (Allen, 1982), this indicates that the palaeocoastline was orientated approximately NW-SE. The asymmetrical ripples indicate the presence of unidirectional currents, and the climbing ripples (where present) indicate high rates of sediment input compared to lateral migration.

3.9.4.3. Rippled heterolithic (Sr, Srw, Fr, Fl) and massive sandstone facies (Sm)

Rippled heterolithic and massive sandstone (Sm) facies occur at 640m log WS1 overlying the wave-rippled sandstone (Srw) described above. Rippled heterolithics comprising rippled sandstone (Sr, Srw), rippled siltstone (Fr) and laminated siltstone (Fl) occur in beds up to 30cm thick (Plate 3.54). The heterolithics contain symmetrical ripples that commonly have undulatory bases and show opposing cross-lamination. This facies is associated with massive sandstone (Sm) that forms beds 10-15cm thick. These have well-defined bases and tops and occur within packages dominated by the heterolithics. The heterolithic and Sm facies pass up into unrippled sediment gravity flow deposits.
The ripples displaying symmetrical forms, often with undulatory bases and opposing cross-lamination, suggest that these deposits were influenced by the action of waves (Collinson and Thompson, 1989). The massive sandstones (Sm) probably represent events, such as storms, when sand was transported offshore, and their occurrence alongside wave-rippled heterolithics suggests slightly deeper water deposition than the wave-rippled sandstone facies (Sr) that occurs directly below them (Allen, 1970), probably in the lower shoreface. These pass up into sediments lacking any evidence of wave action in which density currents start to dominate. This suggests a continuation of the deepening event, and a move from a shoreface depositional environment into offshore conditions.

3.10. Fiq Member synthesis

Before discussing the sequential development and lateral variations that occur within the Fiq Member of the Ghadir Manqil Formation, it is firstly worth looking at how the interpretation of these has been attempted. One of the most striking features of the Fiq Member is the amount of variation that occurs within it. There is not only wide overall variation in the types of facies that occur, but also in the thicknesses of Fiq sediments, and laterally within deposits occurring at the same level in the stratigraphy. In Wadi Sahtan, for example, the Fiq Member is represented by ~1500m of sediment, whereas in Wadi Bani Jabir this thickness is only 5m. This reflects the fact that Wadi Bani Jabir was a high during much of Ghadir Manqil time. Wadi Bani Jabir was therefore experiencing erosion during much of the time that deposition was occurring in Wadi Sahtan, highlighting the differences in depositional rates and environments that were occurring across the Jebel Akhdar during deposition of the Fiq Member. Correlating units within the Fiq Member is therefore problematic. This is worsened by the fact that this is a Precambrian section and is lacking in any fossils that could aid correlation. Most of the correlations that have been made are thus based on lithologies and as the lithologies themselves are often laterally variable, care has to be taken when doing this. However, distinct lithological units and changes in relative sea-level that can be recognised across the whole Jebel Akhdar occur within the Fiq Member (unlike in the Ghubrah Formation). Therefore, so long as lithological correlations are made using sensible inferences as to what would be expected laterally, and the basin-wide sequence stratigraphic events are taken into account, a
relatively robust interpretation of the Fiq Member across the whole Jebel Akhdar can be obtained.

The Hadash Formation that overlies the Fiq Member of the Ghadir Manqil Formation is a widespread and easily recognisable facies that allows a confident correlation to be made at the top of the Fiq Member. Distinct facies types also occur at a number of levels within the stratigraphy of the Fiq Member. Volcaniclastic sandstones lie at its base and allow correlation between Wadi Sahtan and Wadi Mistal. Glacially influenced horizons occur at a number of levels within the Fiq Member, and as glaciation was often widespread and affected the basin as a whole, it provides one means of starting to tie together points from logged sections in different wadis. At the top of most of these glacially influenced horizons, significant flooding events occur. As these events are also basin-wide, they provide points within each logged section that can be tied together confidently. Other basin-wide changes in relative sea-level, not necessarily linked to glaciation, can also be used to tie together points in the stratigraphy between logged sections (Fig. 3.18). Using these tie-points, an accurate reconstruction of laterally linked facies can be attained, and an interpretation of the Fiq Member as a whole is possible (Figs. 3.2; 3.18; 3.19).

3.10.1 Facies evolution of the Fiq Member

Unit FI

The basal part of the Fiq Member is composed of volcaniclastic facies (Fig. 3.19c). These overlie the Saqlah Member, or where this is absent, the Ghubrah Formation. Their volcaniclastic nature allows relatively easy distinction from the massive well-cleaved diamictites of the Ghubrah Formation where the Saqlah Member is missing. These volcaniclastic sediments were deposited by sediment gravity flows in an offshore setting. Palaeocurrents in Wadi Mistal, and a fining and thinning of the beds to the west within Wadi Mistal suggests that, here at least, the volcanic material was being derived from a source lying to the east. Wadi Bani Jabir was a high at this time, but as the volcaniclastic deposits in Wadi Hedak are finer-grained than in the east of Wadi Mistal, it seems unlikely that this was the direct source area for Wadi Mistal. The presence of volcanic material to the east is possibly connected to the powerful eruptions reported from the Saih Hatat
region (Le Metour et al., 1986; Villey et al., 1986) that earlier within this chapter were linked to the volcanism within the Ghubrah and Saqlah Members. However, dates of ~760Ma from the Fiq Member (Bowring, pers. comm.) compared to dates of 723+10/-16Ma (Brasier et al., 2000) and c. 710Ma (Bowring, pers. comm.) from an ash-bed within the Ghubrah Formation suggest the source for the volcanic material in the Fiq Member pre-dates the Ghubrah and Saqlah volcanism. The volcanic material therefore probably represents an earlier phase of volcanic activity that may or may not be linked to Saih Hatat. Stromatolitic carbonates (C) associated with these deposits in Wadi Mistal were possibly able to establish themselves at quiet times between volcaniclastic sediment gravity flow events. Many of these have been reworked by subsequent gravity currents and redeposited. The volcaniclastic deposits in Wadi Sahtan are thicker than those in the east of Wadi Mistal, and are often coarser-grained. Although no palaeocurrents were observed in Wadi Sahtan, it therefore seems likely that the deposits there had a different source to those in Wadi Mistal. Whatever the case, at the onset of Fiq deposition, the Jebel Akhdar was a relatively deep basin, into which material eroded from volcanic terranes was being transported and deposited by sediment gravity flows. The nature of the volcanics of the Saqlah Member, and the large thickness differences observed in the Fiq Member suggest that the basin into which these gravity flows were moving was formed as a graben or half-graben structure.

In both Wadi Mistal and Wadi Mu'aydin, there is evidence for glaciation immediately above the basal volcaniclastic deposits. In the Wijmah section in the east of Wadi Mistal a nearly 40m thick massive diamictite (Dm) containing numerous volcaniclastic clasts occurs. Although this contains no equivocal evidence that it is has a glacial origin, in Wadi Mu'aydin at the same level, a much finer-grained diamictite (Dm) occurs and is overlain by laminated siltstones (Fl) which contain occasional dropstones. These are probably deeper water deposits than those in Wadi Mistal, but they provide good evidence that the eastern part of the Fiq basin was being influenced by glaciation at this time (Fig. 3.19d). One possible explanation why this glacial event is only recorded in the east is that the continental ice-sheets only reached the coast on the eastern landmass. It seems likely from the palaeocurrent indications at this stage that Wadi Sahtan was receiving material derived from the west and Wadi Mistal and Wadi Mu’aydin were influenced more by material coming from the east. It is interesting to note, however, that evidence of
glaciation is not recorded even a few km further to the west within Wadi Mistal, and the glaciomarine deposits are apparently only a locally developed phenomenon. If continental glaciers only reached the coast at certain points on the eastern side of the basin, it seems unlikely that their influence would be seen 30-40km away in Wadi Sahtan. Instead, deposition of volcaniclastic material derived from the west would continue there. Because of this difference the subdivision of Unit F1 in Wadi Mistal and Wadi Mu’aydin has not been made in Wadi Sahtan.

Units F2 and F3 in the west of the Jebel Akhdar

In Wadi Sahtan, quiet, deep-water deposition is established above the sediment gravity flow deposits at the base of the Fiq Member (Fig. 3.18; 3.19e). Increasingly fine-grained volcaniclastic sandstones are overlain by grey siltstones that in turn pass gradationally into dark grey shales (Fm). These dark grey shales contain the maximum flooding surface. They gradually pass up into planar laminated muddy-siltstones (Fl), which continue for the next 200m and demonstrate continuing quiet, deep-water deposition. Above this level the distal members of event beds (probably turbidites) start to make an appearance (suggesting an increase in energy and more rapid deposition), and a gradual shallowing-up suggesting highstand conditions begins. This is represented by an increase in the number of slightly coarser rippled event beds (Fr, Sr), and a reduction in the amount of background sedimentation. Palaeocurrent and palaeoslope indicators at this level in Wadi Sahtan suggest that the landmass from which this material was being derived was situated to the west. This is in contrast to the indicators in Wadi Mistal which suggest the deposits there were derived from the east. Moving upwards through the stratigraphy in Wadi Sahtan, the sediment becomes increasingly influenced by current and wave activity, and climbing and wave ripple forms are developed. The climbing ripples present indicate high concentrations of sediment coming from rapid fallout and relatively rapid deposition. The wave ripples suggest a move from deep-water to lower shoreface deposition, and probable late highstand conditions (Fig. 3.18). At this time then, deposition in Wadi Sahtan moved from a deep-water offshore environment into shallower, possibly shoreface conditions. This shift may have been caused by a number of factors, but one possible explanation is suggested by the overlying deposits of Unit F3. These are composed of ice-rafted diamictite (Dm) and were deposited in a proximal glaciomarine environment (Fig. 3.19f).
The shallowing may thus have been caused by a drawdown of sea-level associated with a deteriorating climate and an increasing volume of ice on land. The culmination of this process occurs when ice-sheets on land have grown so large that they reach the coastline and ice-influenced deposits enter the marine record, producing proximal ice-rafted diamictites.

Units F2 and F3 in the east of the Jebel Akhdar

The F1b diamictite unit shown in log WM1 in Wadi Mistal is overlain by much finer-grained deposits suggesting a flooding event (Fig. 3.18). It is inferred that this is the same flooding event (T1) that occurs at the top of the volcaniclastic sandstones (F1) in Wadi Sahtan. Above this flooding event in Wadi Mistal, coarse-grained sediment gravity flow deposition is re-established (Fig. 3.19e). The coarse-grained gravity flow deposits no longer contain as much volcaniclastic material, but indicate that large amounts of material were still being derived from the east. Deep-water conditions are, however, recorded in Wadi Mu’aydin. The deposits of Unit F2 in Wadi Mistal and Wadi Mu’aydin are overlain by the diamictite of Unit F3, which is correlated here to the F3 diamictite that is present in Wadi Sahtan (Fig. 3.19f). This glaciation is only represented by thin diamictite (Dm) units in the east of Wadi Mistal near Fiq, suggesting again that on the eastern side of the basin, the influence of ice-sheets was concentrated locally. The lowest deposits in Wadi Hajir also record this glacial event. Here, debris flows occur, but the presence of large granite boulders and rare faceted clasts within them suggests that these probably had a glacially influenced precursor. They therefore probably represent ice-rafted glacial material that has been remobilised by later slumping or debris flow events. At this time, then, it seems likely that the whole basin was being influenced by glacial processes.

Unit F4

Above the F3 glacial horizon, a flooding event is recorded across the Jebel Akhdar (T2, Fig. 3.18). This is represented by fine-grained facies (Fm, Fl) in which there is no evidence for continuing glaciation. These deposits signify a period of relatively quiet (but not necessarily deep) water deposition throughout all of the Fiq Member. The flooding is
possibly the result of postglacial transgression and sea-level rise associated with the melting of ice masses on land.

Following this initial postglacial flooding, deep water sediment gravity flow deposition in the form of turbidites is again established in the east of the Jebel Akhdar (Wadi Mistal and Wadi Mu'aydin). Rare palaeocurrent indicators suggest this material is continuing to be derived from the east (Fig. 3.19g). Above the fine-grained deposits associated with the postglacial flooding event in Wadi Sahtan however, shoreface deposits are preserved. These are wave-rippled, shallow-water deposits. Their presence in Wadi Sahtan indicates that the shallow water conditions that existed prior to the F3 glaciation have been re-established. Palaeocurrent indicators in these shoreface deposits are towards the east. Above these shallow water deposits in Wadi Sahtan, a move to increasingly deeper water deposition is recorded until turbiditic sandstones dominate the stratigraphy (Fig. 3.19g). At this stage in the development of the Fiq Member gravity flow deposition dominates right across the Jebel Akhdar. As well as in Wadi Sahtan and Wadi Mistal, similar deposits can also be seen in Wadi Bani Awf and Wadi Hajir. It seems likely however, that the material being deposited was being derived from different sources. The sediments at this level in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir all exhibit structures suggesting palaeocurrents flowing in a general direction from the west towards the east. As has already been discussed (Section 3.9.3.5), these sediments can all be linked if a channel is postulated flowing through Wadi Sahtan and Wadi Bani Awf from the west (Fig. 3.15). The material being deposited would thus have also been derived from this direction. In Wadi Mistal, however a fining trend is observed towards the west and the few palaeocurrent indicators there are suggest flow in this direction. It thus seems likely that at this stage of Fiq deposition, material was being transported into the basin from both the east and the west. The large amount of sediment being deposited and the mainly deep-water gravity flow deposition that dominates the entire basin at this time perhaps indicates renewed uplift of the basin margins coupled with hanging wall subsidence.

Above these mainly turbiditic deposits in Wadi Sahtan and Wadi Mistal a period of quieter water deposition is established. Some influx from distal turbidity currents continues in Wadi Mistal, but in Wadi Sahtan, there is little evidence for any activity and deepwater
conditions are postulated. These fine-grained deposits probably signify another relative flooding event in the basin (T3, Fig. 3.18).

Unit F5

The quiet water conditions sediments at the top of Unit F4 are overlain by sediments indicating another period of renewed glacial activity (Fig. 3.19h). This glaciation affected the whole of the Jebel Akhdar. In Wadi Mistal a proximal glaciomarine environment is recorded in which clast-rich ice-rafted diamictites (Dm, Ds) and sediment gravity flows (perhaps reworking ice-rafted deposits in some cases) are prevalent. Initially in Wadi Hajir and Wadi Bani Awf, similar proximal glaciomarine environments are also indicated by the sediments present. Higher up in these sections, however, the proximal glaciomarine facies association passes into more distal glaciomarine deposition. In Wadi Sahtan, more distal ice-rafted glaciomarine deposition is also recorded. This suggests that at this time these areas were in a deeper water environment than Wadi Mistal further to the east, and possibly indicate more subsidence on the border fault to the west at this time.

Unit F6

The end of the F5 glaciation is again indicated by a flooding event in Wadi Sahtan and Wadi Bani Awf (T4, Fig. 3.18). In the wadis further to the east, however, a series of sediment gravity flow deposits occur above the last glacial deposit before quieter water sedimentation is achieved (Fig. 3.19i). These are possibly the products of remobilisation of sediment deposited by ice-rafting on oversteepened slopes during the glaciation. Alternatively, they may represent a sediment influx caused by the evacuation of moraine and hillslope debris from previously glaciated valleys. In Wadi Mistal, these sediment gravity flow deposits pass up into quieter water sediments, suggesting deepwater deposition. The sediments above this level are cut out by the unconformity with the Permian at WM1 in Wadi Mistal. Further to the south at WM4, the last glaciation of the Ghadir Manqil Formation at the top of the Fiq Member (Unit F7) is recorded. However, the upper part of Unit F6 is not exposed anywhere within Wadi Mistal.
Both lower and upper F6 sediments are well-exposed in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir however. In these wadis to the west, deepwater deposition continues after the postglacial flooding. Current rippled siltstones occur and were probably deposited as distal members of turbidity currents. Palaeocurrent indicators suggest flows moving from the west to the east were continuing. In Wadi Sahtan, these rippled siltstones (Fr, Sr) pass up into massive sandstone (Sm). In Wadi Bani Awf and Wadi Hajir, at the same level as this massive sandstone in Wadi Sahtan, wave-rippled deposits occur. The wave-ripples suggest a shallowing event has been recorded (Fig. 3.18). The sandstone in Wadi Sahtan passes up into very clast-poor diamictite (Dm) that is not observed in Wadi Bani Awf or Wadi Hajir. This diamictite (Dm) possibly represents a glacial period that was not particularly widespread, and in which ice only locally reached the coast (Fig. 3.19i). Overlying this diamictite a significant flooding event occurs that can also be seen in Wadi Bani Awf and Wadi Hajir (T5, Fig. 3.18; Fig. 3.19j). This flooding event is taken as the boundary between Unit F6a and Unit F6b. A clast-rich conglomerate immediately overlies the relatively shallow water deposits described above and probably represents a lag deposit associated with transgression. This conglomerate is in turn overlain by a thin carbonate horizon and then deepwater, massive to planar laminated mudstone. The carbonate in Wadi Bani Awf and Wadi Hajir has been redeposited by density currents, but in Wadi Sahtan, crinkly laminations and the presence of pyrite suggest it may have had a microbial origin. Deepwater conditions persist above this level for some time. In Wadi Sahtan, some rippled and slightly coarser beds suggest the influence of distal density currents as well as fall out processes. Here, the amount of rippling increases upwards and wave and climbing ripple forms can be seen, suggesting a shallowing event (Fig. 3.18). This is also seen in Wadi Hajir, where some rippling is seen and the sediments coarsen up.

**Unit F7**

Overlying the shallowing up sequence at the top of Unit F6b in Wadi Sahtan, coarse-grained sediment gravity flow deposits are preserved before the last glacially influenced interval of the Ghadir Manqil Formation is reached. It is possible that the shallowing (Fig. 3.18) observed below the glacial deposits is also linked to drawdown associated with the onset of glaciation. This final glacial period is recorded in all of the logged sections (Fig. 3.19k). In most of them a proximal glaciomarine facies association is preserved with ice-
rafted diamictites and debris flows. In Wadi Hajir, extensively rippled deposits indicate the proximity of a subglacial stream outwash. The deposits of this final F7 glaciation are overlain by the Hadash Formation, above which no direct evidence for further glaciation is observed in Oman until the Permo-Carboniferous (Gorin et al., 1982; Levell et al., 1988; Wright et al., 1990). The Hadash Formation is a transgressive deposit (Chapter 5). The fact that the uppermost part of the final glaciation of the Fiq Member is present in Wadi Bani Jabir, which was a high for most of Fiq time, suggests that the transgression recorded by the Hadash Formation may have actually begun in the final stages of Fiq deposition (T6, Fig.3.18).
PART III. SUMMARY

3.11. Summary of the Abu Mahara Group in the Jebel Akhdar

This detailed study of the Abu Mahara Group within the Jebel Akhdar has produced a number of interesting conclusions that are briefly summarised below:

i) Deposits in Wadi Bani Jabir, termed the Jabir Formation, are significantly different to any other deposits of the Abu Mahara Group in the Jebel Akhdar. They possibly pre-date the onset of Ghubrah deposition.

ii) The presence of glacial deposits within both the Ghubrah Formation and the Fiq Member of the Ghadir Manqil Formation has been confirmed through the presence of dropstones, striated clasts, and ice-rafted ‘dump’ structures.

iii) The volcanics of the Saqlah Member of the Ghadir Manqil Formation are related to rifting. The fact that the volcanics are absent in the west of the Jebel Akhdar suggest the centre of volcanism was situated closer to Saih Hatat in the east.

iv) The nature of the volcanics of the Saqlah Member, and the large thickness differences observed in the sediments of the Fiq Member suggest that the Ghadir Manqil Formation was deposited in a graben or half-graben structure (Fig. 3.19). This interpretation is consistent with the large amount of material within the Fiq Member derived from sediment gravity flows. Palaeocurrent directions recorded from both the east and the west within the Fiq Member, and the fact that Wadi Bani Jabir was a high for much of Fiq time suggest that the graben or half-graben structure was ~50km across. This is comparable to the size of the graben structures seen from the Abu Mahara Group in the subsurface of Oman (eg. Loosveld et al., 1996).

v) Through lithological and sequence stratigraphic considerations, confident correlations can be made within the Fiq Member of the Ghadir Manqil Formation across the Jebel Akhdar. From these correlations, at least four, and possibly five, discrete glacial intervals separated by non-glacial facies can be seen to occur within the Fiq Member.
vi) New U-Pb zircon dates produced as part of this study have further constrained the timing of the Abu Mahara Group in the Jebel Akhdar. This will be discussed in the next chapter, alongside some of the global implications of the conclusions above.
Chapter 4

Discussion of the Abu Mahara glacials
CHAPTER 4. DISCUSSION OF THE ABU MAHARA GLACIALS

4.1. Introduction

The Abu Mahara Group has long been considered to be of latest Precambrian or Neoproterozoic age (eg. Kapp and Llewellyn, 1965; Glennie et al., 1974; Gorin et al., 1982; Rabu et al., 1986; Loosveld et al., 1996; Brasier et al., 2000; McCarron, 2000). This time period contains a particularly interesting episode in earth history when widespread and possibly even global glaciations occurred. These glaciations are thought to have been so severe that grounded sea-ice occurred in equatorial settings (Schmidt and Williams, 1995; Park, 1997; Sohl et al., 1999). A record of these glaciations is preserved in the Ghubrah and Ghadir Manqil Formations. As these formations are well-exposed and contain volcanic horizons, they provide a good opportunity to further constrain the timing and nature of these possibly global Neoproterozoic glacial events.

Until recently, the placing of the Abu Mahara Group in the Neoproterozoic was based mainly on lithological correlations, its position above pan-African basement, and the presence within it of Vendian acritarchs (Knoll, 1990). Dating of ash-beds within the Huqf Supergroup (Brasier, 1999; Brasier et al., 2000; McCarron, 2000) has now provided tighter constraints on the age of the Abu Mahara Group in the Jebel Akhdar. An ash-bed from within the glacial Ghubrah Formation has produced a U-Pb zircon age of 723±16/-10Ma (Chapter 3, Section 3.7.1.5), and an ash-bed from the Fara Formation, lying above the Nafun Group has yielded a U-Pb zircon age of 544.5±3.3Ma (Fig. 4.1). These dates have been used to suggest that the Abu Mahara Group in the Jebel Akhdar was deposited within the interval 750-590Ma (Brasier, 1999; Brasier et al., 2000). Within this time-span, conventional thinking (eg. Brasier et al., 2000) is that two major glacial epochs occurred (the older Sturtian and younger Marinoan – see below). How the glacial deposits of the Abu Mahara Group fit into a global stratigraphic framework is therefore still questionable. One of the primary aims of this study was therefore to try to further constrain the age of the Abu Mahara Group, both through detailed field-work, and through further U-Pb dating of volcanic and volcanioclastic horizons.
In this chapter, the current age constraints on the widespread Neoproterozoic glaciations will briefly be discussed before the Oman data are assessed. As the glacial deposits of the Abu Mahara Group are well-preserved (particularly in the Fiq Member of the Ghadir Manqil Formation), they also provide an opportunity to discuss current views on the duration and extent of the widely preserved glaciations of the Neoproterozoic.

**4.2. Global time-scale for the Neoproterozoic glaciations**

At least two widespread and broadly synchronous glacial epochs occurred during Neoproterozoic time (eg. Brasier *et al.*, 2000). The older of these epochs is believed to include the Sturtian glaciation of Australia (eg. Christie-Blick, *et al.*, 1995), the Rapitan glaciation of North America (eg. Narbonne and Aitken, 1995; Kaufman *et al.*, 1997), the Chuos/Ghaub glaciations of the Kalahari craton (eg. Kennedy *et al.*, 1998), and possibly the lower ‘Varanger’ glaciation of Svalbard and Scandinavia (eg. Brasier and Shields, 2000; Condon and Prave, 2000). The younger is thought to include the Marinoan of Australia, the Blaubeker/Blasskrans of the Kalahari craton and the Icebrook of North America (eg. Knoll *et al.*, 1986; Narbonne *et al.*, 1994; Hoffman *et al.* 1998a, b; Kaufman *et al.*, 1997; Kennedy *et al.*, 1998; Saylor *et al.*, 1998; Brasier, 1999). Some workers argue on the basis of glacial sediments and carbon isotopic profiles that each of these events was actually composed of more than one glaciation (eg. Kaufman *et al.*, 1997; Saylor *et al.*, 1998; Hoffman *et al.*, 1998b). Many glacial horizons from the Neoproterozoic have poor age constraints on them, and are often assumed to be synchronous with Sturtian-age or Marinoan-age deposits even though there may be no direct evidence for this (cf. Thompson and Bowring, 2000). It is therefore possible that there are numerous glacial events that have been overlooked due to poor chronological constraints. A recent review (Evans, 2000), suggests that if the evidence for the five temporally distinct glaciations reported by Kaufman *et al.* (1997) in northwest Canada, Svalbard, and Namibia is correct, this number should be considered a minimum due to the fragmentary nature of the stratigraphic record. For the sake of simplicity, in this study the terms Sturtian and Marinoan will be used to refer to the two main glacial periods, even though each of these may actually represent multiple events. A chronostratigraphic summary of this time period is shown in Fig. 4.1a.
4.2.1. Age of the Sturtian glacial epoch

The Sturtian glacial epoch is thought to have comprised at least two distinct glaciations and occurred between 760 and 700Ma (eg. Hoffman et al., 1998b; Brasier, 1999; Evans, 2000). An age of 755Ma has been reported for the base of the Windermere Supergroup in Canada (Ross et al., 1995). However, a U-Pb zircon date of 746±2Ma from volcanic beds beneath the Chuos glaciation of the Congo craton, Namibia (Hoffman et al., 1996) may be a more accurate estimate of the onset of Sturtian-age glaciation. In Canada, a U-Pb date of 723+4/-2Ma taken from volcanics overlying lower Sturtian glacial deposits (Heaman et al., 1992) suggests that upper Sturtian glaciations are likely to be younger than this. The upper Sturtian glaciations may even be as young as 700Ma (Young, 1995; Veevers et al., 1997; Brasier, 1999).

4.2.2. Age of the older Marinoan glaciations

Similarly to the Sturtian, at least two distinct glacial events are thought by many workers to have occurred in Marinoan time (eg. Kaufman et al., 1997; Saylor et al., 1998; Brasier, 1999). The older of these is thought to have lasted from c. 600 to c. 590Ma. The Gaskiers tillite in the Conception Group of southeast Newfoundland lies above volcanic rocks U-Pb dated at 606+3.7/-2.9Ma (Krogh et al., 1988) and below tuffs U-Pb dated at 565±3Ma and associated with Ediacara fossils (Benus, 1988). This tillite is thought to correlate with the Marinoan glaciation in Australia (Anderson and King, 1981). A comparable diamictite (the ‘Squantum tillite’) in Massachusetts overlies igneous rocks U-Pb dated at 602±3Ma (Kaye and Zartman, 1980; Grotzinger et al., 1995), and contains clasts that have recently yielded U-Pb dates indicating it has a maximum age of 595Ma (Thompson and Bowring, 2000). The end of the lower Marinoan ice age is estimated to have ended by approximately 590Ma (Brasier, 1999).

4.2.3. Age of the younger Marinoan glaciations

Supposedly younger Marinoan glaciations have also been reported by some workers. These include the Fersiga in Algeria (Bertrand-Sarfati et al., 1995), the Egan in Australia (Grey and Corkeron, 1998) and the Blasskrans or Naukluft in Namibia (Saylor et al., 1998;
The age of these glacial deposits are not well-constrained, but Saylor et al. (1998) estimated a termination date of c. 564-551Ma for Namibia, based on models involving a constant rate of sediment accumulation. The youngest possible age of the Blasskrans is constrained by a U-Pb date of 548.8±1Ma from ash-beds within the overlying Kuibis Subgroup (Grotzinger et al., 1995).

4.3. Age of the Abu Mahara glacial events in Oman

The global time-scale for the Neoproterozoic outlined above indicates at least four distinct glacial periods: an upper and lower Sturtian (760-700Ma), a lower Marinoan (600-590Ma), and possibly an upper Marinoan (c. 575-565Ma). At least four distinct glacial horizons occur within the Abu Mahara Group of Oman (Fig. 4.1c), and how these fit into the Sturtian/Marinoan framework is still debatable. The potential time-frame for the Abu Mahara Group suggested by Brasier (1999) and Brasier et al. (2000) of 750-590Ma suggests that the glacial horizons preserved in the Ghubrah and Ghadir Manqil Formations may span both the Sturtian and lower Marinoan periods. The absolute age date of 544.5±3.3Ma produced from the Fara Formation overlying the top of the Nafun Group does not rule out an upper Marinoan position for the last glacial horizon of the Ghadir Manqil Formation. However, considerations of the stratigraphy and isotopic curves that occur between this glacial horizon and the beds yielding the 544.5Ma date suggest that an upper Marinoan assignment is unlikely (see discussion below). Within this temporal framework a number of different scenarios can still be envisaged for the timing of the Abu Mahara glacial (Fig. 4.1). These will now be considered in turn and analysed in the light of new data produced as part of this study:

Chronostratigraphic scenario 1.

The U-Pb date from within the Ghubrah glaciation of 723+16/-10Ma represents an inherited age, and all the glacial horizons preserved in the Abu Mahara Group of the Jebel Akhdar are younger, probably Marinoan-age equivalents (Fig. 4.1b).

The problems with inheritance that can occur with U-Pb dates were briefly discussed in Chapters 2 and 3. It is possible that the ash horizon in the Ghubrah Formation was
reworked similarly to the volcaniclastic sandstone facies (Sv) of the Fiq Member (Ghadir Manqil Formation, Chapter 3, Section 3.9.3.7). Zircons from a number of samples from this Sv facies were dated using the U-Pb method (Appendix A). The results from two of the samples (Wadi MT and Wadi MT2, log WM1, Appendix D) are shown in Figs. 4.2 and 4.3. Four zircons from sample Wadi MT define a discordia with an upper intercept of 765.5±6Ma and a lower intercept of 63±240Ma at 95% confidence levels (Fig. 4.2). The zircons from sample Wadi MT2 do not define a discordia, and the maximum age is constrained by the youngest analysis (z5) lying close to concordia at c.755Ma (Fig. 4.3). Within both of these samples older zircons occur close to concordia and indicate an inherited component. The date of 765.5±6Ma initially appears relatively robust, but as the dates produced are older than the date from the Ghubrah Formation lower in the stratigraphy, it appears that even the younger zircons have been inherited. The fact that analysis z5 suggests a maximum age for sample Wadi MT2 of 755Ma, and that it comes from approximately the same horizon as Wadi MT (log WM1, Appendix D) indicates that the zircons defining the discordia with an upper intercept of 765.5±6Ma have been reworked, even though they possibly all originated from the same event. A similar reworked origin for the Ghubrah zircons and a Marinoan age for all the glacial horizons in the Abu Mahara Group would, in some ways be a simple explanation of the stratigraphy, giving an essentially unbroken stratigraphic succession with no major gaps (Scenario 1, Fig. 4.1b).

However, a number of arguments work against the supposition that the zircons from the ash-bed in the Ghubrah Formation are reworked. Firstly, the zircons have euhedral shapes and an igneous morphology (Brasier et al., 2000; McCarron, 2000) that would probably not have been retained during abrasion associated with later reworking. The same ash-bed from the Ghubrah Formation that produced the U-Pb date of 723±16/-10Ma was re-sampled as part of this study (Fig. 4.4). Four zircons from this sample define a discordia with an upper intercept of 711.8 ±1.6Ma and a lower intercept of 75.5 ±29.9Ma at a 95% confidence level. Although the age of 711.8 ±1.6Ma is younger than the previously attained date of 723±16/-10Ma, there is an overlap in the errors, and it indicates that the glacial deposits here still fall within the upper Sturtian time bracket. As this new date is still >100Myrs older than any Marinoan deposits, and as no zircons yielding ages younger
than 710Myrs have been found, it seems very unlikely that the Ghubrah glaciations can be of Marinoan age. This first scenario is therefore discounted here.

The date of 711.8 ±1.6Ma produced as part of this study provides a new and useful constraint on the timing of the Sturtian glaciations. As it comes from within glacial deposits instead of bracketing a glacial period with a maximum/minimum age, it provides good evidence of the occurrence of glacial conditions actually at c. 712Ma. In combination with the previous date of 723±16/-10Ma, this date therefore gives one of the best estimates yet produced of the timing of upper Sturtian glaciation.

*Chronostratigraphic scenario 2.*

All of the glacial horizons within the Abu Mahara Group in the Jebel Akhdar are of upper Sturtian age (c. 712Ma). The Nafun Group commences immediately after the end of Sturtian glaciation and extends up to the c. 540Ma Ara Group.

Once the U-Pb dates from the Ghubrah Formation are taken to be accurate and not inherited, this second suggestion is the only way of avoiding having to account for an extra 100Myrs of time within the Abu Mahara Group of the Jebel Akhdar (Fig. 4.1b).

However, a number of problems immediately arise with this conjecture. Firstly, considerations of the stratigraphy lying between the uppermost glacial horizon of the Ghadir Manqil Formation and the U-Pb zircon date of 544.5±3.3Ma from the Fara Formation suggest it is unlikely to represent the 150Myrs required for it to be Sturtian in age. It seems unlikely that deposition would have been slow enough and constant enough to preserve ~1.2km of Nafun stratigraphy over a period of c. 150Myrs. Although the contact between the top of the Ghadir Manqil Formation and the base of the overlying Hadash Formation is sharp, cap carbonate deposition is thought to be intimately linked to postglacial transgression (Chapter 5). No large time gap can therefore be envisaged at this level. The Hadash Formation passes gradationally into the Masirah Bay Formation, which is conformably overlain by the Khufai Formation. A major sequence boundary occurs at the top of the Khufai Formation (Brazier, 1999; McCarron, 2000), but this probably does
not represent a long-lasting break in deposition. No other potential unconformities have been reported from the Nafun Group (McCarron, 2000).

The δ¹³C isotopic curve produced from the sediments overlying the uppermost glacial of the Ghadir Manqil Formation can also be used to investigate the conjecture that the entire Abu Mahara Group is upper Sturtian in age. The negative δ¹³C signature preserved in the Hadash Formation is typical of Neoproterozoic cap carbonates (Chapter 5). This passes up into positive values that persist for the Khufai Formation. These are in turn followed by a large negative excursion that lasts for the duration of the Shuram Formation before more positive values are re-established in the Buah Formation (Burns and Matter, 1993; McCarron, 2000) (Fig. 4.5). It has been suggested that even where glacial deposits are absent in a stratigraphic profile, a negative isotopic excursion may indicate the occurrence of glaciation in other areas (eg. Kaufman et al., 1997; Saylor et al., 1998; Hoffman et al., 1998b). If this is the case, then the negative δ¹³C signature preserved in the Shuram Formation may indicate a glacial event not preserved in the rock record of Oman (McCarron, 2000). One possibility would be that this glacial event corresponds to one of the Marinoan glaciations. Although this provides an explanation for the lack of younger glacial deposits in the Neoproterozoic stratigraphy of Oman that might be expected if the Abu Mahara Group was taken to be entirely Sturtian in age, it still cannot account for the required time interval between the Sturtian and Marinoan periods. For example, even if the Shuram Formation was an equivalent to the lower Marinoan glacial event, if the Abu Mahara Group of the Jebel Akhdar is considered to be entirely Sturtian in age, c. 100Myrs of time still have to be accounted for within the Hadash, Masirah Bay, and Khufai Formations.

Previous correlations have suggested the Shuram Formation is actually an equivalent of the postulated younger Marinoan glaciation (c. 575-565Ma), rather than the older one (600-590Ma) (Fig. 4.5; Saylor et al., 1998; McCarron, 2000). This correlation would suggest that the upper glacial horizons of the Abu Mahara Group represent the older Marinoan glacial period, which is the third suggestion postulated here.
Chronostratigraphic scenario 3.

The Abu Mahara Group in the Jebel Akhdar contains both upper Sturtian and lower Marinoan equivalent glaciations (Fig. 4.1b) and the Nafun Group stratigraphy extends from c. 590-545Ma.

This third suggestion incorporates both the 711.8±1.6Ma date produced from the ash bed within glacial deposits of the Ghubrah Formation and the age date from the Fara Formation of 544.5±3.3Ma. Although this hypothesis requires less time to be accounted for within the Nafun Group, over 100Myrs now has to be incorporated into the Abu Mahara Group of the Jebel Akhdar. A major sedimentary break should therefore be apparent at some point within the Abu Mahara Group, as it is unlikely that sedimentation rates would be low enough for deposition to be continuous throughout this long time period.

The most obvious place for such a break is between the Ghubrah and Ghadir Manqil Formations. Although this boundary is unfortunately poorly exposed, there are a number of features that suggest the two formations may be separated by a significant time gap. The Ghubrah Formation is commonly more pervasively cleaved than the overlying Ghadir Manqil Formation, and in places appears to have undergone higher levels of deformation. Unfortunately, this is often hard to see due to the massive nature of much of the Ghubrah Formation. The overlying deposits of the Ghadir Manqil Formation are also better bedded than the Ghubrah Formation. Beds can often be traced laterally to some extent and reasonably robust correlations can be made across the Jebel Akhdar (Figs. 3.2; 3.18). There were no major unconformities noted in any of the sections logged through the Fiq Member of the Ghadir Manqil Formation (Appendix D).

Clast analyses carried out on glacial diamictites from the two formations suggest that the clast types present were related to locality, and no temporally distinct assemblages of clasts could be recognised (Fig. 3.8). Although no major difference was apparent between the clasts of the potentially much older glacials of the Ghubrah and those of the Ghadir Manqil Formation, a general higher proportion of volcanic clasts within the Ghubrah diamictites can be seen (Fig. 3.8).
Some preliminary work on illite crystallinity was also undertaken to see if there were any major differences between the burial history of the Ghubrah and Ghadir Manqil Formations. However, results were broadly similar for all the samples from both the formations, and the work was not continued further.

Despite the lack of concrete evidence for a major stratigraphic break between the Ghubrah and Ghadir Manqil Formations, it still appears to be the most reasonable level at which to place it within the Abu Mahara Group. As it is more difficult to accommodate the 100Myrs of time within the Nafun Group, a major time break at this level would seem to be the most sensible conjecture to make with the current data available. This would mean that the glacial deposits of the Ghubrah Formation are upper Sturtian equivalents (c. 720-700Ma), and that the five distinct glacial horizons of the overlying Ghadir Manqil Formation are all Marinoan equivalents (c. 600-590Ma). The Abu Mahara Group in the Jebel Akhdar would therefore encompass over 100Myrs of time and two major glacial epochs. Suggested correlations between the Omani section and glacial deposits in Australia, north-west Canada, and Namibia are shown in Fig. 4.6.

It should be noted that the carbonate that occurs at the base of the Ghadir Manqil Formation in Wadi Mistal (Section 3.9.3.8, Chapter 3) has previously been referred to by some workers as a ‘cap carbonate’ (e.g. Brasier et al., 2000). It was suggested that the rapidly rising $\delta^{13}$C values were similar to those reported from carbonates overlying putative Sturtian glacials elsewhere (e.g. Kennedy et al., 1998). Although this carbonate does lie above a probable Sturtian equivalent, it occurs within the Ghadir Manqil Formation. Thus, if the suggested stratigraphic break postulated above is accepted, the carbonate should be separated from the Sturtian-equivalent Ghubrah Formation by ~100Myrs, and could not be linked to this glacial in the manner of a cap carbonate (see Chapter 5). The carbonate at this level also does not appear to have a broad lateral extent. It has therefore not been referred to as a ‘cap carbonate’ in this thesis.

4.4. Nature of the widespread Neoproterozoic glaciations

There has been much debate on the nature of the glaciations recorded in rocks of Neoproterozoic age. Their widespread nature has long been recognised (e.g. Harland,
1964), and their deposits are recorded on all seven present continents, often at more than one stratigraphic level within a sedimentary basin (Evans, 2000). Enigmatically, Neoproterozoic glacial deposits are commonly associated with apparently low-latitude lithological indicators such as carbonate rocks. Recent palaeomagnetic work has confirmed the occurrence of low-latitude, possibly even equatorial sea-level glaciations at this time, suggesting that Phanerozoic-style polar glaciations cannot account for the distribution of Neoproterozoic glacial deposits (e.g. Schmidt and Williams, 1995; Park, 1997; Sohl et al., 1999; Kempf et al., 2000; Evans, 2000). Instead, to explain this occurrence of glaciations in equatorial regions a number of different models have been proposed. These include an equatorial ice ring (Sheldon, 1984), substantial true polar wander (Evans, 2000), and diamictites actually representing impact ejecta (Rampino, 1994). Two of the more widely considered models (the snowball Earth and high obliquity hypotheses) will firstly be outlined and then discussed below, before they are considered in the light of the data from the glacial deposits of the Abu Mahara Group.

4.4.1. Snowball Earth hypothesis

Glaciations of global extent were first envisaged by Harland (1964). This idea of global refrigeration has more recently been dubbed the ‘snowball Earth’ hypothesis (Kirschvink, 1992; Hoffman et al., 1998a, b; Hoffman and Schrag, 2000). In this model, continental break-up brings formerly landlocked areas closer to oceanic sources of moisture. The increased rainfall scrubs more heat-trapping CO₂ out of the atmosphere and increases continental erosion. Global temperatures consequently start to fall and the equator-ward growth of polar ice to c. 30° N-S latitudes is thought to have occurred. Once this point was reached, ice-albedo feedback would have caused the rapid expansion of sea-ice resulting in a globally frozen Earth. During these ‘snowball’ conditions, average global temperatures would have fallen to -50°C, the cold, dry air would have prevented the growth of land glaciers although the oceans would have iced over to a depth >1km, and virtually all marine organisms would have perished (Hoffman and Schrag, 2000). Effectively, the whole hydrological cycle would have been shut down, with no precipitation, evaporation, and no sediment transport or deposition. Thus, only at the onset and end of these snowball conditions would deposition of the primarily subaqueous Neoproterozoic glaciogenic sediments have occurred. During snowball conditions CO₂ from volcanic outgassing
would gradually accumulate in the atmosphere as no rainfall would occur to remove it. Once CO$_2$ levels in the atmosphere had built up to a critical level (~120,000ppm), a rapid change from snowball to greenhouse conditions would occur as the albedo and water vapour feedbacks would enhance the warming with the opening of low-latitude oceans (Hoffman et al., 1998b). During this hothouse aftermath, conditions conducive to carbonate deposition would occur. The snowball Earth model therefore also provides some explanation for the association of carbonates with glacial deposits that is commonly observed in Neoproterozoic strata (Chapter 5).

Estimates for the length of time required for CO$_2$ to build up to the critical level required to begin meltback suggest that snowball conditions must have lasted for ~10 Myrs. The 'Snowball Earth' model therefore predicts synchronous, long-lasting, global glaciations that would have been recorded at all latitudes. It also predicts c. 10 Myr hiatuses in the stratigraphic record, where tectonic subsidence continues, but with no sediment input.

4.4.2. High obliquity hypothesis

The high obliquity model (Williams, 1975) suggests that the preponderance of glacial deposits in equatorial latitudes during the Neoproterozoic can be explained if the present zonation of climate was reversed, and the equator experienced colder conditions than the poles. Any planetary obliquity of >54° leads to the poles receiving more annual sunlight than the equator. In the high obliquity model it is suggested that this has indeed been the case for much of Earth's history (Williams, 1993), and it is only in the past 550 Myrs that the obliquity has approached the current level of 23.5° through obliquity-oblateness feedback (Williams et al., 1998).

The high obliquity model suggests that the poles would have remained ice-free, while equatorial regions would have experienced extreme seasonality and become glaciated. This would explain the paucity of significant evidence for high-latitude glaciations during the Neoproterozoic (Evans, 2000). The oceans would not have been completely iced over and photosynthetic organisms could therefore have survived through the Neoproterozoic glaciations without having to resort to special refuges, as is the case in the snowball model.
In contrast to the snowball model, a high-obliquity scenario would be most compatible with diachronous glaciation as continents moved through tropical latitudes (Evans, 2000).

4.4.3. Discussion of models

The Neoproterozoic glacial paradox and the snowball Earth hypothesis in particular have provoked a lot of recent interest and debate. Some authors have suggested that the long period of time for which marine organisms would have been deprived of sunlight in the Hoffman et al. (1998b) snowball model is in apparent conflict with evidence for the continuous presence of photosynthetic marine organisms in the geological record (eg. Williams et al., 1998; Runnegar, 2000; Hyde et al., 2000). Hyde et al. (2000) therefore proposed a ‘softer’ snowball Earth model based on climatic simulations in which the hydrological cycle was not completely shut down and an equatorial belt of open water may have provided a refugium for multicellular organisms.

Other workers have argued that a number of geological observations refute a snowball Earth scenario (eg. Williams and Schmidt, 2000). Firstly, the presence of thick glaciomarine successions in the Neoproterozoic (eg. Young and Gostin, 1989) suggest that a vigorous hydrological system was maintained during glaciation, rather than ceasing as the snowball model would predict. Secondly, deposits exhibiting wave ripples and tidal rhythmites interbedded with glacial horizons indicate open oceans and unfrozen seas (Williams and Schmidt, 2000). Thirdly, the occurrence of glacial cycles in Neoproterozoic successions suggesting repeated glacial advances and retreats (eg. Spencer, 1971; Condon and Prave, 2000) conflict with the snowball hypothesis in which continuous severe glacial conditions persist for 10Myrs or more. Finally, Williams and Schmidt (2000) provide evidence from periglacial sand wedges for marked seasonality that is also hard to account for during snowball conditions.

As with the snowball Earth model, problems also arise with the high obliquity hypothesis. Although higher obliquity for the Earth would have greatly amplified the seasonal cycle, providing an explanation for the seasonality observed by Schmidt and Williams (2000), this would have resulted in hot biannual summers unfavourable to glaciation (Hoffman et al., 1998a). Evans (2000) also suggests that high obliquity would lead to an inherently
mild climatic regime, further questioning whether conditions capable of producing the widespread glacial deposits in the Neoproterozoic would have occurred with greater obliquity of the Earth. The association of carbonates and glacial deposits commonly observed in the Neoproterozoic is also not fully explained by the high obliquity model (Hoffman et al., 1998a).

Although the high obliquity model (Williams, 1975) was devised as an alternative to Harland’s (1964) idea of global glaciation (snowball Earth model), it is worth noting that the two are not necessarily mutually exclusive. A combination of high obliquity and snowball conditions could produce predominantly low-latitude deposits, but could also allow for the rare occurrence of polar glaciers (Evans, 2000).

4.4.4. Discussion of the Abu Mahara Group data

The sediments of the Abu Mahara Group in the Jebel Akhdar (Chapter 3) offer a good opportunity to assess the applicability of the two models outlined above. The Sturtian-age glacial deposits of the Ghubrah Member are not particularly well-exposed, and do not offer much to the Neoproterozoic glacial debate. The Marinoan-age deposits of the Fiq Member of the Ghadir Manqil Formation, however, are well-preserved and exposed and raise a number of interesting issues.

In all of the stratigraphic profiles logged through the Fiq Member (Appendix D), a number of distinct glacial rain-out intervals occur. Three of these can be recognised across the Jebel Akhdar (Units F3, F5, and F7), with other more localised glacial deposits also occurring (eg. Unit F1b, Log WM1 and the top of Unit F6a, Log WS1) (Fig. 3.2). These rain-out intervals are commonly separated by sediments that contain no evidence for continued glaciation. Indeed, the large (up to 200m) thicknesses of laminated sediments devoid of any dropstones or evidence for ice-rafting that occur between some of the rain-out intervals (eg. Unit F6) suggest that all glacial activity had ceased, at least on a regional scale, in the marine environment as they were being deposited. Other deposits between successive glacial rain-out intervals preserve wave-rippled structures (eg. Facies Srw, Unit F4, Wadi Sahtan), providing further compelling evidence for an unfrozen sea between glacial events. As was discussed above (Section 4.3), all of the glacial deposits of the
The snowball Earth hypothesis (Hoffman et al., 1998b) is supposedly applicable to any Neoproterozoic glacial succession capped by a carbonate unit. The Omani section satisfies these requirements (see Chapter 5), yet does not fit very easily into the model. The discrete glacial rainout intervals provide evidence for significant environmental fluctuations. These would have required areas of unfrozen seas where surface waters could be warmed, circulated, and cause melting of ice sheets and the resulting release of debris (Condon et al., 2001). Thus, in contrast to the extreme conditions envisaged by the snowball Earth model, the Neoproterozoic rock record from Oman indicates that the glacial environment was dynamic and was characterised by contemporaneous open seas and land-based ice sheets. The ‘softer’ snowball Earth suggested by Hyde et al. (2000) could be invoked to explain the open seas. However, the long-lasting (c. 10Myr duration) glacials the snowball Earth model invokes also appear to be absent from the Ghadir Manqil Formation in Oman, which records at least four discrete glaciations, probably within the c. 600-590Ma period.

There is also little evidence from the Ghadir Manqil Formation in Oman (other than the glacial retreats and advances) for the strongly seasonal climate envisaged by the high obliquity model of Williams (1975) and Williams et al. (1998). However, the glacial deposits of the Ghadir Manqil Formation are dominantly marine. Equivalents of the Ghadir Manqil Formation in the south of Oman (the Mirbat Sandstone) preserve shallower-water and continental glaciation, and contain clastic dykes that cross-cut glacial diamictites (Plate 4.1). These may have originated from ice-wedging, which would provide evidence for the marked seasonal changes predicted by the high-obliquity model.

Oman is thought to have been situated in tropical latitudes during Late Neoproterozoic time (Powell et al., 1993; Dalziel, 1997; Kempf et al., 2000). Therefore, if the snowball hypothesis of Hoffman et al. (1998a) is ruled out, either a series of very severe, but not
necessarily global, Phanerozoic-style glacial events reaching down to the tropics occurred, or the high obliquity model has to be seriously considered.

4.5. Summary

From this discussion of the glacial horizons of the Abu Mahara Group in the Jebel Akhdar a number of conclusions can be drawn:

i) A new U-Pb zircon date of 711.8±1.6Ma confirms the presence of upper Sturtian-equivalent glacial deposits in Oman suggested by the previously produced date of 723±16/-10Ma (Brasier et al., 2000). This date comes from within glacial deposits and provides one of the best constraints yet produced on the timing of the upper Sturtian glaciation.

ii) A number of factors including U-Pb zircon dates, the stratigraphy of Oman, and the global Neoproterozoic chronological framework all suggest that the Abu Mahara Group contains glaciations of both Sturtian and Marinoan age. From this study, the Ghubrah Formation is considered to be of upper Sturtian age, and the overlying Ghadir Manqil Formation is taken to be an equivalent to the lower Marinoan glacial period (Fig 4.1). A time gap of ~100Myrs is therefore postulated between the Ghubrah and Ghadir Manqil Formations. If this hypothesis is correct, the negative δ13C signature recorded in the Shuram Formation may well represent the upper Marinoan glacial period (Blasskrans/Wilsonbreen equivalent) that is not otherwise recorded in the rock record of Oman (Fig. 4.5).

iii) Discrete glacial horizons within the Ghadir Manqil Formation suggest repeated glacial cycles and unfrozen, circulating seas that are not compatible with the snowball Earth hypothesis of Hoffman et al. (1998b). It is suggested that a series of severe Phanerozoic-style glaciations may be more characteristic of the Late Neoproterozoic glacial epochs. The high obliquity model for the Earth is also not ruled out. Both of these models could explain the shorter-lived, non-synchronous glacial events within a longer glacial era that the rock record in Oman suggests.
CHAPTER 5. THE HADASH FORMATION

PART I. INTRODUCTION AND FACIES ANALYSIS

5.1. Introduction

The Hadash Formation is well-exposed in the core of the Jebel Akhdar, directly overlying the last glacial diamictite of the Ghadir Manqil Formation. It forms the base of the Nafun Group and passes up gradationally into the Masirah Bay Formation. The Hadash Formation is included as a separate formation here, despite having a maximum thickness of 15m, as it is markedly different in nature to both the Ghadir Manqil and Masirah Bay Formations. It is a carbonate rather than a siliciclastic formation and its base marks a major change in climate and depositional environment at the end of Ghadir Manqil time. Its top is taken as the last carbonate bed before the siliciclastics of the Masirah Bay Formation come to dominate. From outcrop in the Jebel Akhdar, the Hadash Formation can be seen to have a lateral extent of over 70km. Subsurface data (Bell, 1993a), and possible correlations with carbonate horizons of the Mirbat Sandstone in the south of Oman (Loosveld et al., 1996; Kellerhals, 1998) suggest that the Hadash Formation has a lateral extent of hundreds of km right across Oman. Carbonate beds overlying the Halfayn Formation in the Huqf area (Chapter 2) are also possibly a lateral equivalent of the Hadash Formation. These beds have been looked at as part of this study and will be discussed within this chapter.

The Hadash Formation was first described in the Jebel Akhdar by Kapp and Llewellyn (1965), where it was included within the upper part of their Mistal Conglomerate. In the Huqf area, Dubreuilh et al. (1992) were the first to describe the carbonate overlying the Halfayn Formation, including it within the Abu Mahara Formation. The name Hadash Formation has been adopted in this study after excellent exposures which occur close to the village of Hadash in the south of the Ghubrah Bowl (Wadi Mistal, Jebel Akhdar). The name ‘Hadash Formation’ was first used by McCarron (2000).

The occurrence of the carbonate Hadash Formation directly on top of glacial deposits is of particular interest, not only because it marks such a major environmental boundary but also
because, above the base of the Hadash Formation, there is no direct evidence for glaciation in Oman until well into the Phanerozoic. Similar thin carbonate units lying directly on top of Proterozoic glacial successions are well documented from around the world (eg. Williams, 1979; Fairchild and Hambrey, 1984; Tucker, 1986; Narbonne et al., 1994; Kennedy, 1996; Hoffman, et al. 1998; Myrow and Kaufman, 1999). Glacial deposits from this period were particularly widespread and possibly even global in extent (see discussion in Chapter 4). The overlying carbonates ‘capping’ these glacial successions also appear to have had an almost global extent. These ‘cap carbonates’ often have many features in common. They are commonly laterally continuous on a basinal scale, preserve a distinctive negative $\delta^{13}$C signature, and often constitute the sole carbonate horizon in a dominantly siliciclastic succession (Kennedy, 1996). It has been suggested that a cap carbonate overlying Marinoan-age glacial deposits be used to define the proposed new ‘Terminal Proterozoic System’ (eg. Christie-Blick et al., 1995; Knoll, 2000). The well-exposed Hadash Formation preserves a particularly good example of such a cap carbonate. It is therefore not only of importance within Oman, but also has the potential to help fit the Abu Mahara and Nafun Groups into a global context and to shed light on global post-glacial processes in the Neoproterozoic. For this reason the Hadash Formation was studied in detail.

The aims of this chapter are to firstly describe and initially interpret the Hadash Formation lithologically. The isotopic and geochemical data will then be considered. This will be used to further the interpretation of the Hadash Formation and also to investigate the robustness of $\delta^{13}$C stratigraphy at near metre-scale resolution to provide a tool independent of lithology that can be used for both local and global scale correlation.

5.2. Methods

The most complete and least tectonically disturbed exposures of the Hadash Formation occur in Wadi Mistal (Plate 5.1), Wadi Hajir (Plate 5.2), and Wadi Bani Awf in the Jebel Akhdar, with the type section occurring close to the village of Hadash in the south of Wadi Mistal (Figs. 5.1 and 5.2). The Hadash Formation was logged in detail at each of these localities. Samples were taken at 10cm intervals for petrographic and stable isotope analysis. Further logs were made in Wadi Bani Jabir (Plate 5.3), Wadi Mu’aydin and
Wadi Sahtan (Plate 5.4) (Fig. 5.2). Samples for petrographic and stable isotope analysis were also taken at ~30cm intervals from Wadi Bani Jabir and Wadi Mu’aydin, and from supplementary sections in Wadi Hajir and Wadi Mistal. The carbonate unit overlying the Halfayn Formation in the Huqf area was also logged and sampled for stable isotope analysis (Fig. 5.3). Altogether 281 samples from 8 different sections were analysed for carbon and oxygen isotopes. Sampling procedures are outlined in Section 5.8, and analytical methods are described in Appendix C.1. Carbon and oxygen data for all the samples analysed is presented in Appendix C.6.

5.3. Previous work

Carbonates of the Hadash Formation were first referred to in the Jebel Akhdar by Kapp and Llewellyn (1965), where they were included within the upper part of the Mistal Conglomerate along with the Masirah Bay Formation. This laminated carbonate unit within the Mistal Conglomerate was interpreted as probably having formed in a tidal flat environment. Further work in the Jebel Akhdar by Rabu et al. (1986), Beurrier et al. (1986), and Rabu (1988) split the Mistal Conglomerate into four members. The uppermost of these was termed the Amq Member, and spanned the units that are herein separated into the Hadash and Masirah Bay Formations. A possible algal origin for the finely laminated dolomite of the Hadash Formation was postulated by these workers.

The carbonates that outcrop above the Halfayn Formation in the Huqf area were first described by Dubreuilh et al. (1992) where they were taken to represent the base of their Abu Mahara Formation (see Fig. 6.8). A carbonate unit had, however, been noted earlier from Corehole-17 in the centre of the Khufai Dome (Fig. 6.2, Chapter 6) (Gorin et al., 1982). The fact that this carbonate passed up into siliciclastics of the Masirah Bay Formation was used to suggest a tentative correlation between the Huqf and Jebel Akhdar areas at this stratigraphic level. The correlation of these carbonate units was taken further by Bell (1993a). Working mainly on subsurface data he split the Abu Mahara Group into the Ghadir Manqil and Masirah Bay Formations. The Hadash Formation (herein) was then included at the base of the Masirah Bay Formation and termed the ‘Basal Carbonate Member’ (Figs. 3.7, 6.8). Bell (1993a), used this Basal Carbonate Member to correlate surface outcrops in Mirbat with those in the Huqf and Jebel Akhdar, and to tie together the
subsurface sections. The fact that it was the sole carbonate horizon in a dominantly siliciclastic section of the stratigraphy, and that it could be found in outcrop and well-sections so extensively, encouraged these lithological correlations. Bell (1993a, b) suggested that the Basal Carbonate Member was deposited in relatively shallow marine conditions because of domal stromatolites present in the outcrops at Wadi Ercahol (near Salalah in the south of Oman).

The seemingly paradoxical occurrence in the Neoproterozoic of a laterally extensive carbonate lying directly on top of and effectively ‘capping’ a thick glacial succession is not unique to Oman. Similar deposits have been described from a number of localities including Australia (eg. Williams, 1979; von der Borch et al., 1988; Kennedy, 1996), Canada (eg. Narbonne et al., 1994; Narbonne and Aitken, 1994; Myrow and Kaufman, 1999), Africa (eg. Hoffman et al. 1998a, b; Kennedy et al., 1998; Kennedy et al., 2001), California (eg. Tucker, 1986; Prave, 1999(a)), as well as Scotland and the North Atlantic region (eg. Spencer, 1971; Fairchild and Hambrey, 1984; 1995; Fairchild and Spiro, 1987; Brasier and Shields, 2000). Work on these deposits revealed a number of common features including a thin, laterally extensive nature, and a distinctive negative isotopic signature. From this work a number of different models have been put forward to explain the somewhat perplexing association of glacial diamicrites with carbonate as well as accounting for the negative isotopic signature. These models will be discussed in Section 5.14, in a synthesis that includes the lithofacies and isotopic character of the Hadash Formation.

5.4. Placement of the Hadash Formation at the base of the Nafun Group

Bell (1993a) first suggested that the term the ‘Abu Mahara Group’ be used to link together the mainly clastic Ghadir Manqil and Masirah Bay Formations. The Hadash Formation (as used herein) is equivalent to Bell’s (1993a) ‘Basal Carbonate Member’ at the base of the Masirah Bay Formation, and was therefore included within his Abu Mahara Group (Fig. 3.7, Chapter 3). The base of the Hadash Formation, however, marks a major change in depositional style when compared with the underlying Ghadir Manqil Formation and is more closely linked with the Masirah Bay Formation. The Hadash and Masirah Bay Formations are arguably better united with the overlying Nafun Group, which is composed
of siliciclastic-carbonate cycles. For this reason, it is suggested here that the base of the Hadash Formation be taken as the base of an expanded Nafun Group. The Nafun group is therefore taken to include the Hadash, Masirah Bay, Khufai, Shuram, and Buah Formations. The Abu Mahara Group, as used here, therefore ends at the top of the Ghadir Manqil Formation, and includes the Halfayn, Ghubrah, and Ghadir Manqil Formations (Figs. 1.9; 3.7; 6.8).

5.5. Lithological subdivision of the Hadash Formation

Although the Hadash Formation is less than 15m thick, it can be divided into different lithofacies. Exposures in the east of the Jebel Akhdar (ie. Wadi Bani Jabir, Wadi Mistal, Wadi Mu’aydin) contain some notable differences from the exposures in the west (ie. Wadi Hajir, Wadi Bani Awf, Wadi Sahtan). Initially, therefore, the Hadash Formation in the east will be described separately from that in the west. The possible correlative in the Huqf area will also be described separately.

In Wadi Mistal (Hadash sections 1 and 2) and Wadi Mu’aydin, both in the east of the Jebel Akhdar, the Hadash Formation can be subdivided into a lower and upper part (Fig. 5.2). The lower part of the Hadash Formation is typically composed of dolomitic microspar with minor siltstone horizons. This can be further subdivided into a lower, thinly-bedded, fine-grained facies (Facies E1); a middle, siltstone dominated facies (Facies E2); and an upper, more thickly-bedded, planar, and coarser-grained facies (Facies E3). The upper part of the Hadash Formation in the east is coarser-grained than the lower part. It is composed of limestone and dolomite interbedded on a scale of a few centimetres (Facies E4). The Hadash Formation exposed in Wadi Bani Jabir (also in the east of the Jebel Akhdar) differs slightly to the above. Here again, the lower part of the Hadash Formation in Wadi Bani Jabir is composed dominantly of thinly-bedded dolomitic microspar (Facies E1). The upper part, however, is composed of siltstone and medium- to coarse-grained sandstone within which rare beds of dolomite occur (Facies E5). These beds are unlike the dolomite ‘stringers’ found in the upper Hadash Formation in the west of the Jebel Akhdar (Facies W3), being interbedded with much coarser siliciclastic deposits.
The lower part of the Hadash Formation in the west of the Jebel Akhdar is again dominated by dolomitic microspar. The basal part is made up of thin beds of dolomite interbedded with siltstone on a cm scale (Facies W1). Upwards, the amount of siltstone is reduced and thicker beds of dolomite dominate (Facies W2). The upper part of the Hadash Formation in the west of the Jebel Akhdar is dominated by planar laminated mud/siltstone. Within this, 'stringers' of carbonate up to 30cm thick (Facies W3) occur above the main body of the carbonate through a thickness of up to 10m. In the south of Wadi Hajir, the distinction between the east and west of the Jebel Akhdar is less marked, and a basal section composed of Facies E1 passes up into Facies W1.

The carbonate at the base of the Masirah Bay Formation in the Huqf area is composed of relatively pure dolomitic spar (Facies H1). This is here included as a possible correlative of the Hadash Formation based on lithology. This correlation will be discussed in more detail later, alongside the isotopic data (Section 5.12).

5.6. Facies analysis of the Hadash Formation

Each of the lithological facies present within the Hadash Formation will initially be described separately, with a brief interpretation of some of the main features. A synthesis for the whole Hadash Formation will then be put forward on the basis of the lithologies observed. The features noted in each facies will then be used to suggest possible depositional models. A more detailed discussion of the Hadash Formation follows the isotopic results in Sections 5.14 and 5.15.

5.6.1. Facies E1: Dolomitic microspar with locally developed siltstone intercalations

Dolomitic microspar with locally developed siltstone intercalations (Facies E1) forms the basal part of the Hadash Formation in the east of the Jebel Akhdar (eg. 0-2.5m, Wadi Bani Jabir; 0-2.6m, Hadash Cap 1; 0-2.5m, Hadash Cap 2; 0-2.2m, Wadi Mu’aydin; and 0-1.1m, Wadi Hajir 2) (Plate 5.5). The contact with the underlying Ghadir Manqil Formation is sharp, but there is no apparent angular discordance. At every locality visited, Facies E1 (like the base of the Hadash Formation throughout the Jebel Akhdar) lies directly on top of massive diamicrite (Dm) facies. The top of Facies E1 is marked by a series of recessive
beds dominated by mud/siltstone (Facies E2) in Wadi Mistal and Wadi Mu’aydin, and by the influx of siliciclastics that marks the base of Facies E5 in Wadi Bani Jabir.

Like much of the Hadash Formation, the dolomite of Facies E1 has a distinctive tawny colouration on weathered surfaces. Where the dolomite is fresh, a planar, light to dark grey lamination on a few mm- to cm-scale can be seen. At Hadash, thin (cm to few cm thickness) dolomite beds with mm-scale siltstone intercalations make up the base of Facies E1. These beds commonly pinch and swell laterally. They pass up into slightly thicker (up to 20cm), less undulatory beds in which siltstones are no longer present. The light and dark lamination on fresh surfaces is particularly well preserved in these beds. These more planar beds, lacking in siltstone intercalations dominate the whole of Facies E1 in Wadi Mu’aydin. In Wadi Bani Jabir, siltstone intercalations actually become more prevalent upwards within Facies E1.

Although Wadi Hajir 2 is included within the ‘western’ part of the Jebel Akhdar for the purpose of subdividing the Hadash Formation, its basal section is more akin to exposures in the eastern Jebel Akhdar. Unlike the Hadash Formation exposed at Wadi Hajir 1 or further to the west, the lower metre at Wadi Hajir 2 is composed of Facies E1 before passing up into Facies W1.

In thin-section, the dolomite beds of Facies E1 are composed of non-ferroan subhedral dolomite crystals mainly <60μm in size. Most of the crystals are slightly cloudy, with a few appearing brown in colour (Plate 5.6). The dolomite crystals within the darker laminations have an opaque coating. In thin-section, it can be seen that the dark laminations have sharp bases and more wavy, irregular tops that locally develop into a ‘clotted’ fabric (Plate 5.7). A slight fining-up of the dolomite crystals occurs, from ~60μm at the base of Facies E1 to ~40μm at the top. Minor opaque minerals (including pyrite) and quartz also occur within Facies E1. Locally these form clusters up to 100μm in size. Chert nodules up to 2 cm in size are also locally developed within this facies. Both quartz and calcite veining are present. They are concentrated in the lower part of the facies. Under cathodoluminescence (CL), the dolomite rhombs of Facies E1 all show a similar luminescent pattern, with (locally absent) bright centres, non-luminescent inner surrounds, and bright rims (cf. Plate 5.12).
A definitive lithological interpretation of Facies E1 is difficult to achieve (as is the case with many of the facies of the Hadash Formation) because of the lack of primary fabrics or structures. The relatively homogenous, fine-grained crystal size suggests early (and possibly primary) dolomite formation. The cloudiness and colouration of the dolomite crystals is possibly due to the preservation of organic material. If this is the case, the alternately light and dark laminations may well be a product of varying levels of organic matter (Dave Wright pers. comm., 1998). The evidence for a microbial precursor is limited to the preservation of this organic matter, the presence of very minor pyrite, and the equivocal 'clotted' fabric. Degradation of a microbial mat is, however, a possible mechanism for producing the microcrystalline dolomite fabric preserved in Facies E1 (cf. Wright, 1997). The lack of sedimentary structures such as rippling may be due to deposition below storm wave-base.

5.6.2. Facies E2: Recessive, siltstone dominated deposits

Facies E2 occurs above Facies E1 in Wadi Mistal and Wadi Mu’aydin (2.5-2.7m, Hadash Cap 1; 2.6-2.95m, Hadash Cap 2; 2.2-2.5m, Wadi Mu’aydin). It is a recessive facies, and marks the top of a distinct ridge in the lower part of the Hadash Formation at these localities (Plate 5.8). It is composed of light to dark grey friable siliciclastic siltstone within which occur beds of dolomite up to 10cm thick, similar to those described in Facies E1. In Wadi Mu’aydin, rare pyrite crystals up to a few mm in size are found. Facies E2 is overlain by Facies E3 at all localities where it was observed.

The predominance of fine-grained siliciclastics in comparison to carbonate at this level may reflect an increase in the influx of detrital material. This influx may have temporarily hindered carbonate precipitation, or the detritus may simply have accumulated at a greater rate than the carbonate.

5.6.3. Facies E3: Planar bedded dolomitic microspar to dolospar

Planar bedded dolomitic microspar to dolospar (Facies E3) overlies Facies E2 in Wadi Mistal and Wadi Mu’aydin (2.7-5.6m, Hadash Cap 1; 2.95-4.4m, Hadash Cap 2; 2.5-3.9m,
Wadi Mu’aydin), and is in turn overlain by Facies E4. Facies E3 is slightly darker in colour than Facies E1. It is composed of planar beds, 10-20cm in thickness (Plate 5.9), in which a mm-scale planar lamination occurs. The light and dark grey banding observed in Facies E1 is no longer present. On fresh surfaces the dolomite has a beige to light brown colour. In Wadi Mu’aydin, pyrite crystals up to 0.5cm and more in size are common (Plate 5.10). These are particularly concentrated in one band c.10cm thick, some 3m above the base of the formation.

In thin-section, it can be seen that the dolomite is composed of well-developed rhombs that increase slightly in size upwards through the facies from <100µm at the base to ~150µm at the top. The dolomite crystals are similar in nature to those in Facies E1 (Plate 5.11). Although beds are commonly very homogenous, bands rich in quartz grains up to 0.5mm thick occur locally. These define the lamination noted above. Under CL, the dolomite crystals luminesce like those in Facies E1 with (locally absent) bright centres, a dull inner surround, and a bright luminescing rim (Plate 5.12).

Facies E3 probably formed in a similar environment to Facies E1. The large pyrite crystals that are present within this facies in Wadi Mu’aydin may indicate the presence of microbes and anoxic conditions within the sediment. The larger and less ‘dirty’ looking dolomite crystals may therefore simply reflect greater degradation (compared to Facies E1) of an organic-rich, possibly microbial precursor (cf. Slaughter and Hill, 1991; Wright, 1997). The presence of quartz-rich laminae suggests a greater detrital influence.

5.6.4. Facies E4: Interbedded crystalline limestone and dolospar

Interbedded crystalline limestone and dolomite (Facies E4) forms the top part of the Hadash Formation in Wadi Mistal and Wadi Mu’aydin (5.6-7.6m, Hadash Cap 1; 4.4-6.5m, Hadash Cap 2; 3.9-5.6m, Wadi Mu’aydin). Facies E4 overlies the planar bedded dolomite of Facies E3 and is in turn overlain by the siltstones at the base of the Masirah Bay Formation. Facies E4 is composed of grey beds dominated by calcite spar interbedded with beds of tawny-coloured dolomite on a few cm scale (Plate 5.13). Minor interbeds of siltstone a few mm thick occur between some of these carbonate beds. A slightly ‘crinkly’ lamination is widely developed within the carbonate of Facies E4. Locally, this lamination
defines structures (Plates 5.14; 5.15) similar to microbial ‘roll-up’ structures described from Namibia (eg. Hoffman, 1998a). Where these are asymmetrical, they are overturned towards the west. Within this upper part of the Hadash Formation, rare shallow scour structures also occur. Pyrite crystals up to 2mm in size are widely developed within Facies E4.

The dolomite of Facies E4 is composed of almost perfect euhedral rhombic crystals up to 300µm in size. These rhombs are brown in colour and are turbid, often being crowded with inclusions (Plate 5.16). Anhedral crystals of calcite spar occur alongside the dolomite rhombs, forming a minor part of the dolomitic layers and dominating in the grey limestone beds. The layers in which the dolomite rhombs dominate commonly have sharp bases with less well-defined, wispy tops. The calcite crystals are generally of similar size to the dolomite rhombs (up to 300µm), but are colourless and much less turbid in nature. Locally, smaller calcite crystals occur within larger ones, and a neomorphic texture is developed. Very little detrital material is evident in thin-section.

Under CL, the dolomite rhombs show a distinctly different luminescent pattern to that seen in the lower facies of the Hadash Formation (Facies E1 and E3). The majority of the dolomite crystals in Facies E4 have large, uniformly bright orange-red luminescent centres with darker rims (Plate 5.17). The calcite shows dull, yellow-coloured luminescence.

The turbid and brown-coloured nature of the dolomite crystals in Facies E4 suggests the preservation of organic matter. The ‘crinkly’ lamination and roll-up structures suggest cohesion, most likely by mucilaginous microbial membranes (cf. Kennedy, 1996; Hoffman et al. 1998b; Kennedy et al. 2001). The presence of pyrite within the facies further suggests the presence of microbes and indicates anoxic conditions (cf. Kennedy et al. 2001). The fact that the calcite crystals are less cloudy than the dolomite, and locally show neomorphic or poikilotopic textures indicates that they formed at a later stage.

5.6.5. Facies E5: Thin beds of dolomite in siltstone and sandstone

Facies E5 was only seen in the top part of the Hadash Formation in Wadi Bani Jabir. The base of Facies E5 is composed of a white, muddy siltstone that lies on top of Facies E1.
Within this muddy siltstone occur immature sandstone beds up to 60cm thick (Plate 5.18). These are fine- to coarse-grained, commonly contain planar lamina tions, and may show normally graded bedding. Rare carbonate beds up to a few cm in thickness are also present. These occur every few metres, tend to be rich in siliciclastics, and are dolomitic. The last carbonate bed, marking the top of the Hadash Formation here, occurs 6.5m above the top of Facies E1. The Masirah Bay Formation overlying this is dominated by coarse-grained, graded sandstones. Cross-stratification is locally developed within these and suggests palaeocurrents towards the west.

The presence of coarser-grained material within this section of the Hadash Formation at Wadi Bani Jabir suggests that this locality was more proximal to a source of siliciclastic, detrital material than the rest of the exposures in the Jebel Akhdar. This interpretation is supported by the rare palaeocurrent indicators suggesting flow towards the west. Dolomite is restricted to a few, thin beds, suggesting that carbonate precipitation/accumulation was largely swamped by siliciclastic deposition.

5.6.6. Facies W1: Thinly-interbedded dolomite and siltstone

Dolomite and siltstone interbedded on a cm-scale (Facies W1) occur at the base of the Hadash Formation in the west of the Jebel Akhdar (0-1.1m, Wadi Hajir 1; 0-1.2m, Wadi Bani Awf; 0-4.2m, Wadi Sahtan) sharply overlying the uppermost diamictite of the Ghadir Manqil Formation (Plate 5.19). Facies W1 also occurs further south in Wadi Hajir (1.1-2.5m, Wadi Hajir 2), where it overlies the purer dolomite of Facies E1. At all of the localities where it is present, Facies W1 passes up into dolomite beds with a much reduced siltstone component (Facies W2). Throughout Facies W1, the amount of siltstone is reduced upwards, and the dolomite beds gradually thicken. The dolomite beds of Facies W1 are tawny in colour and tend to have undulatory upper and lower surfaces. The interbedded siltstone is friable and light grey in colour.

The dolomite beds are composed of non-ferroan dolomicrospar <40μm in size. The crystals are subhedral and commonly brown in colour. Detrital quartz grains make up <10% of the dolomite. Locally, quartz grains occur in clusters up to 0.1mm in size.
As in Facies E2, the presence of siltstone indicates the influence of a siliciclastic input. The cm-scale interbedding of the siltstone with dolomite suggests alternating periods of deposition during which the siliciclastic input was varied. The fact that dolomite becomes dominant upwards within this facies suggests that either the siliciclastic input became reduced or that rates of dolomite deposition increased. Few structures or fabrics are preserved within the dolomite so that interpretation of the way in which it formed is difficult.

5.6.7. Facies W2: Dolomitic microspar with locally developed siltstone intercalations

The dolomitic microspar with locally developed siltstone intercalations of Facies W2 forms the middle part of the Hadash Formation in the west of the Jebel Akhdar (eg. 2.5-4.1m, Wadi Hajir 2; 1.1-3.6m, Wadi Hajir 1; 1.2-3.5m, Wadi Bani Awf; 4.2-5.2m, Wadi Sahtan). It overlies the interbedded dolomite and siltstone of Facies W1 and passes up into the 'stringers' within siltstone of Facies W3 at all of the localities where it occurs. In many respects, Facies W2 is similar to Facies E1 (Plate 5.20). The dolomite has the same tawny colour, and the amount of siltstone present reduces upwards. Beds are only up to a few cm in thickness, however, and a slightly undulose lamination is preserved on a mm- to few mm-scale. The dolomite is again commonly composed of turbid and brown coloured subhedral crystals but these are finer-grained than in Facies E1, being only up to 40µm in size. Detrital quartz is also present and can form up to 10% of the rock. On the uppermost surface of Facies W2, a series of rounded depressions up to 1cm deep and a few cms in diameter occur (Plate 5.21). These are commonly connected by cracks, which can be seen to radiate out in an almost star-like fashion from some of the hollows. The larger depressions tend to have flattened bases, but the smaller ones can also be rounded. The depressions are commonly filled with the overlying muddy-siltstone of Facies W3, suggesting that they were syndepositional. Rarely, the hollows are infilled with chert.

Similarly to Facies E1, the cloudy nature and brown colour of the dolomite crystals in Facies W2 suggests an organic-rich precursor. The higher percentage of detrital material indicates greater siliciclastic input. The depressions that occur on the top surface of Facies W2 are somewhat enigmatic. They may be the product of dissolution, either prior to the

102
deposition of Facies W3, or by later fluid movement along the boundary between Facies W2 and W3. Alternatively, they may represent concretions that grew, but which were more susceptible than the dolomite to later dissolution. The cracks connecting some of the depressions are probably the product of later jointing exploiting existing weaknesses in the rock.

5.6.8. Facies W3: Thin beds of dolomite (‘stringers’) within laminated mud/siltstone

Facies W3 comprises beds of dolomite up to 30cm thick (‘stringers’) within laminated mud/siltstone (Plates 5.22; 5.23). These occur at the top of the Hadash Formation in the west of the Jebel Akhdar (e.g. 4.1-13.7m, Wadi Hajir 2; 3.6-7.4m, Wadi Hajir 1; 3.5-9.6m, Wadi Bani Awf; 5.1-9.5m, Wadi Sahtan). The last carbonate stringer marks the top of the Hadash Formation and is immediately overlain by siltstone of the Masirah Bay Formation. Each dolomite stringer is separated by up to 1.5m of commonly laminated grey muddy siltstone. The lamination is planar, on a mm scale, and is laterally continuous for at least several metres. The siltstone beds between dolomite stringers tend to thicken towards the top of Facies W3, making the transition from the Hadash Formation into the Masirah Bay Formation gradational in the west of the Jebel Akhdar. The dolomite beds of Facies W3 are typically very weathered and brownish in colour. They are planar, and locally a mm-scale lamination is developed within them. The dolomite is formed of well-developed rhombic crystals up to 100µm in size. The crystals are typically brown in colour and cloudy. Locally, ferroan dolomite and calcite occur. Under CL, the dolomite crystals luminesce a deep red with bright centres, dull inner surrounds and bright rims (Plate 5.24).

The fact that siltstone is so prevalent at the top of the Hadash Formation in the west of the Jebel Akhdar indicates the increased importance of siliciclastic deposition in the transition to the Masirah Bay Formation. The presence of purely fine-grained (mud/siltstone) siliciclastic deposits within Facies W3 suggests relatively quiet, and possibly deep water deposition. The laterally extensive planar lamination probably indicates that the siltstones formed through background, suspension settling. The cloudy nature of the dolomite crystals again suggests they may have some organic content. Whether the dolomite formed when siliciclastic deposition was reduced or was deposited rapidly as the background sedimentation continued is uncertain. The low proportion of detrital grains
within each dolomite stringer suggests that background siliciclastic deposition may have been reduced during dolomite formation.

5.6.9. Facies H1: Crystalline dolomite

Crystalline dolomite (Facies H1) occurs at the base of the Masirah Bay Formation in the Huqf area and is a possible Hadash Formation equivalent (Fig. 5.3). The dolomite is a few metres thick and outcrops near Al Jobah (Plate 5.25), overlying the unconformity at the top of the Halfayn Formation. It is in turn overlain by siliciclastics that are thought to represent the base of the Masirah Bay Formation (Chapter 6). The dolomite of Facies H1 is pink-to-beige in colour and contains a mm-scale to few mm-scale sedimentary lamination (Plate 5.26; Fig. 5.3). This is best developed in the lower part of the carbonate and is mainly planar, with small, cm-scale scours developed within it. In thin-section, the dolomite can be seen to be composed of interlocking euhedral rhombs up to 250μm in size (Plate 5.27). Zoning of these crystals is locally developed.

The coarser grain size and lack of siliciclastic material make the carbonate in the Huqf area (composed of Facies H1) notably different from the Hadash Formation in the Jebel Akhdar. As in the Jebel Akhdar, though, the crystalline nature of the dolomite makes an interpretation difficult. However, as the overlying deposits of the Masirah Bay Formation in the Huqf area are of shallower origin than those in the Jebel Akhdar, the dolomite here may also be a shallower water correlative of the Hadash Formation in the Jebel Akhdar. The planar lamination with small-scale scours is probably the product of high velocity flows (upper flow regime), consistent with shallow water deposition.

5.7. Lithological synthesis of the Hadash Formation

Within the Jebel Akhdar, the boundary of the Hadash Formation with the underlying Ghadir Manqil Formation is sharp but apparently conformable. The dramatic change to a finer grain-size that occurs at its base, the generally fine-grained nature of the siliciclastics, the overlying deep-water deposits of the Masirah Bay Formation, and the large lateral extent (possibly of many hundreds of kilometres) all suggest deposition of the Hadash Formation in relatively deep-water (below storm wave-base), possibly linked to a
transgression. This interpretation is further supported by the lack of evidence for subaerial exposure, for evaporites, or for the influence of waves within any of the facies of the Hadash Formation. The transgressive nature is clearly suggested in the Huqf area, where the unconformable base of the carbonate preserved there indicates flooding of the high of Ghadir Manqil times and the onset of shallow marine deposition there. The presence of the Hadash Formation lying conformably above the last glacial deposit of the Ghadir Manqil Formation in the Jebel Akhdar suggests this transgression may well have been linked to deglaciation.

Larger amounts of fine-grained siliciclastics in the west of the Jebel Akhdar, the orientation of the roll-over structures at Hadash, and the palaeocurrent directions recorded in Wadi Bani Jabir, all suggest that deposition of the Hadash Formation was proximal to the east and more distal in the west of the Jebel Akhdar. In the Huqf area, the carbonate is inferred to represent shallower water conditions than further to the north in the Jebel Akhdar.

The textures now seen in the carbonate of the Hadash Formation are inferred to be replacive but the fine-grain-size and homogenous nature of many of the beds suggests rapid dolomite precipitation after deposition. Much of the Hadash Formation probably formed as the product of background carbonate precipitation in areas of limited terrigenous input (cf. Kennedy, 1996). However, the turbid nature of much of the dolomite and the roll-over structures within Facies E4 suggest that parts of the Hadash Formation may have formed through the degradation of an organic-rich, probably microbially-influenced precursor (cf. Wright, 1997). Dolomite precipitation may therefore have been promoted by the presence of microbes.

In the Jebel Akhdar, a number of different correlations between the facies of the Hadash Formation can be made, depending on how deposition is modelled. Four possible depositional models for the Hadash Formation in the Jebel Akhdar are shown in Fig. 5.4. In the first of these (Fig 5.4a), deposition of the Hadash Formation is not taken to begin until highstand conditions have become established following the initial post-glacial transgression. Progradation dominates and deposition begins in the more proximal areas. It is not until the deposition of Facies E2 in the east that sedimentation of Facies W1
begins in the more distal west. In this model, deposition also continues for longer in the western parts of the Jebel Akhdar, where the ‘stringers’ of Facies W3 are taken to have formed after carbonate deposition had ceased further to the east. This situation could have arisen as the influence of siliciclastic influxes may have been at a lower level in the west compared to the more proximal east, allowing the intermittent establishment of carbonate deposition there. Alternatively, the carbonate stringers of Facies W3 in the west may represent episodes when carbonate originally deposited in more proximal areas was eroded and transported further offshore, forming event beds within the background sedimentation. The lack of any sedimentary structures within the stringers of Facies W3 apart from a locally developed planar lamination suggests that the latter explanation is unlikely. During deposition of the upper parts of the Hadash Formation (post-Tl), relatively coarse-grained siliciclastics, probably deposited from relatively proximal sediment gravity flows, occur in Wadi Bani Jabir. It may be that this more proximal area was situated close to a point of terrigenous input at the end of Hadash time, preventing significant carbonate precipitation. By T3, the siliciclastic deposition that characterises the Masirah Bay Formation in the Jebel Akhdar is established across the whole Jebel Akhdar. This model is initially attractive as it correlates Facies E2 with Facies W1, both of which record relatively high levels of fine-grained siliciclastic input.

In the second depositional model (Fig 5.4b), similarly to the first, deposition of the Hadash Formation in the Jebel Akhdar does not begin until highstand conditions have become established. In this model, however, aggradation is the dominant process, and the start of deposition is synchronous in both the east and the west. In the upper part of the Hadash Formation, the purer carbonate deposition of Facies E3 and E4 in the east is synchronous with the deposition of the siliciclastic-rich Facies W3 in the west. This could be envisaged if fine-grained detrital material that by-passed the more proximal areas in the east was inhibiting carbonate precipitation in the west, only allowing the thin carbonate stringers to form at certain periods when the terrigenous input was reduced.

Problems with both of these first two models (Figs. 5.4a; 5.4b) arise from the fact that they require the establishment of highstand conditions before the onset of Hadash deposition. Current models proposed for cap carbonate deposition (eg. Kennedy, 1996; Hoffman et al., 1998b; Kennedy et al., 2001) suggest that the carbonate precipitation is linked directly to
postglacial transgression (see later discussion, Section 5.14). Progradation could still be envisaged, however, if there was a pause in sea-level rise during transgression (Fig. 5.4c). Similarly to the first model (Fig. 5.4a), this third depositional model allows the lithologically attractive correlation between the siliciclastic-rich Facies E2 in the east and the thinly-interbedded siltstone and dolomite of Facies W1 in the west (Fig. 5.4c). After the initial progradational deposition, however, at the top of Facies W2 in the west and Facies E4 in the east, renewed flooding occurs. This transgression results in a hiatus in deposition/erosion in the more proximal areas, with onlap of the thin carbonate beds of Facies W3 in the more distal western Jebel Akhdar. The increase in the influx of terrigenous material at the base of the Masirah Bay Formation indicates the re-establishment of highstand conditions.

The fourth and final depositional model suggested here envisages a simpler transgression than the previous model followed by the onset of highstand conditions (Fig. 5.4d). In this model, Hadash deposition begins during transgression when carbonate onlaps onto the shelf. The facies at the base of the Hadash Formation in the west are therefore deposited before those in the more proximal east. Transgression continues until the top of Facies W2 in the west and Facies E3 in the east, when highstand conditions have become established. At this point, thin, probably microbially-bound, carbonate shelf deposits are established in the east (Facies E4), passing distally into the carbonate 'stringers' of Facies W3. As in all the other models, relatively coarse-grained siliciclastic deposition dominates in the most proximal areas (Wadi Bani Jabir) during deposition of the upper Hadash Formation. Carbonate deposition ceases when later highstand conditions develop and siliciclastic input is increased. This marks the base of the Masirah Bay Formation.

The four models outlined above indicate the wide range of possible correlations that can be made between facies within the Hadash Formation based on lithological and sequence stratigraphical considerations. The different depositional models each carry important implications for how closely linked to post-glacial transgression 'cap carbonates' may be. In Part II of this chapter, an attempt will be made to test the correlations within the Hadash Formation suggested by each model (and therefore the applicability of each model) by using carbon isotope stratigraphy.
PART II. ISOTOPIC AND GEOCHEMICAL ANALYSIS OF THE HADASH FORMATION

5.8. Introduction

Cap carbonates reported from Neoproterozoic time from around the world all share a remarkably similar, strongly negative isotopic signature. Any reasons suggesting why deglaciation at this time was so dramatic, and why such deglaciation should produce conditions conducive to carbonate precipitation on a global scale also need to explain the isotopic signature that is so distinctive of cap carbonates.

Unlike the siliciclastics of the underlying Abu Mahara Group and overlying Masirah Bay Formation, the Hadash Formation is dominantly composed of carbonate. The fact that it is well preserved and has been relatively free from tectonic alteration suggests that it has good potential for the acquisition of stable isotopes. Although the dolomite is replacive, some of the features discussed above (Sections 5.6 and 5.7) suggest dolomitisation occurred at an early stage. Early dolomitisation would have occluded pore space, isolating the crystals from later alteration by diagenetic fluids (Tucker, 1983; McCarron 2000). As carbon is a major component of carbonate rocks and a minor component of sedimentary fluids, the $\delta^{13}C$ signature is likely to be relatively insensitive to water-rock alteration. Dolomite in isotopic equilibrium with calcite ought, in theory, to be enriched in $^{13}C$ by approximately 2‰ (Sheppard and Schwarcz, 1970). In practice, however, the difference appears to be minimal (Kaufman and Knoll, 1995). Most studies of Proterozoic limestones and dolostones have demonstrated empirically that they normally retain a C-isotopic composition that must be close to their precursor CaCO$_3$ (eg. Tucker, 1983; Knoll et al., 1986; Derry et al., 1992; Narbonne et al., 1994).

Whole-rock carbon isotopic studies have revealed stratigraphic patterns of $\delta^{13}C$ fluctuations of relatively large magnitude through the late Neoproterozoic (eg. Kaufman and Knoll, 1995; Pelechaty et al., 1996; Brasier et al., 2000; McCarron, 2000). These fluctuations can be used to improve the relatively coarse stratigraphic resolution provided by radiometric age dates and sparse Neoproterozoic fossils (Knoll and Walter, 1992). $\delta^{13}C$ curves produced for the Nafun Group have concentrated on the Khufai, Shuram and Buah
Formations (Burns and Matter, 1993), with the few samples that were run from the Hadash Formation producing $\delta^{13}$C values between $-2.5$ and $-5.3\%$ (McCarron, 2000). The $\delta^{13}$C signature produced above the negative Hadash Formation samples reveals a marked positive excursion in the Khufai Formation, a negative excursion in the Shuram, with a gradual return to positive values in the Buah. This has led to a correlation of the Hadash Formation with carbonates transgressing glacial deposits in Namibia, Western Canada and Australia (McCarron, 2000) (see discussion in Chapter 4).

Carbonates capping glacial horizons in the Neoproterozoic yield negative isotopic values that are similar to those of the Hadash Formation. These consistently show a distinctive signature in which an initial shift towards negative values occurs above the base of the carbonate, followed by a gradual return to more positive values up-section (eg. Kennedy, 1996; Kennedy et al., 1998; Hoffman et al., 1998b; Myrow and Kaufman, 1999). As it is so widely developed, this distinctive negative isotopic excursion has, to some extent, been used to correlate Neoproterozoic sections (eg. Kennedy et al., 1998; Brasier et al., 2000). It has even been suggested that the inflection points of the negative excursion be used to help define a new ‘Terminal Proterozoic System’ (eg. Christie-Blick, 1995; Knoll, 2000). The negative carbon isotopic values have also been incorporated into models explaining the occurrence of Neoproterozoic cap carbonates lying directly above glacial deposits (eg. Knoll et al., 1986; Kaufman et al., 1993, 1997; Grotzinger and Knoll, 1995; Kennedy, 1996; Hoffman et al., 1998a, b; Kennedy et al., 2001).

Although the Hadash Formation is <15m thick it should therefore be possible to use its $\delta^{13}$C signature for chem stratigraphic purposes, as well as using it to help explain its occurrence directly above glacial diamictite.

In this study, 281 samples from 8 stratigraphic profiles were collected for stable isotope analysis (7 in the Jebel Akhdar, and 1 in the Huqf area). This level of sampling for isotopic analysis is at a greater density than any recorded in published studies of cap carbonates (eg. Kennedy, 1996; Myrow and Kaufman, 1999). ‘Least altered’ samples were selected for being well away from areas of veining and for having good fabric preservation, including petrographical and cathodoluminescence data. Lateral samples were taken <1m apart from the same horizon and repeats were run to test the
reproducibility and accuracy of results. Eighty-one samples from five stratigraphic profiles in the Jebel Akhdar area and two samples from the Huqf area were analysed for major and minor elements. Analytical methods are described in Appendix C.1-C.3. Carbon and oxygen data for all the samples analysed is presented in Appendix C.6.

The aim of this study is therefore to characterise the stable isotopic nature of the Hadash Formation in detail, assessing the likely effects of diagenesis, and making corrections where possible. Elemental and isotopic data have been analysed, and where appropriate, cross-plots have been made to screen for diagenesis. This should make it possible to investigate whether the isotopic signatures preserved in the Hadash Formation are likely to represent a primary signal. Where the isotopic signature has been affected, the possibility of ‘backtracking’ to the least-altered isotope signature has been investigated. Screened isotopic data of this kind should then allow higher resolution chem stratigraphic correlations, not only within Oman at this level, but also with published δ¹³C curves from other Neoproterozoic cap carbonates. These data may help constrain models explaining the presence of carbonate directly overlying glacial diamictite at this level.

5.9. Principles of carbon and oxygen isotope chem stratigraphy

In stable-isotope geochemistry, the ratio between two different isotopes of the same element is measured. For stable-isotope geochemistry of carbon, the ratio of ¹³C to ¹²C (expressed as δ¹³C and quoted in ‰) is measured, and for stable-isotope geochemistry of oxygen the ¹⁸O to ¹⁶O ratio (expressed as δ¹⁸O and quoted in ‰) is measured in a sample of calcium carbonate. These ratios are calibrated against the universal PDB standard, enabling the comparison of ratios from different samples and different laboratories (Anderson and Arthur, 1983). The isotopic compositions of elements of low atomic number vary because their isotopes are fractionated by natural chemical and physical processes. Differences that arise because of this fractionation form the basis of stable isotopic investigations in palaeoceanography (Anderson and Arthur, 1983).
5.9.1. Carbon-isotope fractionation

Carbon consists principally of two stable isotopes. These are the light $^{12}$C isotope (98.89% abundance) and the heavy $^{13}$C isotope (1.1% abundance) (Anderson and Arthur, 1983). The short-lived radioactive isotope $^{14}$C occurs only in trace amounts. The isotopic composition of carbon in the biosphere and hydrosphere reflects the complex balance of equilibrium and kinetic fractionation reactions associated with the global carbon cycle. Marked fractionation is associated with photosynthesis and other biological activity, and most organic carbon compounds have negative isotopic signatures compared to the dissolved inorganic carbon in seawater (Anderson and Arthur, 1983; Marshall, 1992). This is due to the fact that photosynthesis discriminates against $^{13}$C with the preferential uptake of $^{12}$C, owing to a kinetic isotope effect inherent in the first irreversible enzymatic CO$_2$-fixing reaction (Schidlowski and Aharon, 1988). Thus, the basic configuration of the global carbon cycle reflects the activity of the Earth’s biosphere.

A number of different factors, including depth, ecology, water energy, proximity to land, and circulation patterns contribute to fluctuations in the ambient seawater $\delta^{13}$C. Today, the world average surface water has a $\delta^{13}$C of +2.0‰ PDB, with the deeper water having lower values, typically with $\delta^{13}$C between −0.5‰ and +1.0‰ PDB (Prothero, 1994). This depth gradient arises because marine organic matter (with very negative $\delta^{13}$C) is dominantly withdrawn from the dissolved inorganic carbon in the surface waters by photosynthesis. In deeper water and on the seafloor, marine organic matter is commonly re-oxidised after the death of the photosynthetic organisms and the light carbon returned to the dissolved inorganic carbon (Marshall, 1992). The difference in the isotopic composition between surface and deep waters accordingly reflects the productivity and circulation within the oceans. Absolute values of $\delta^{13}$C in the photic zone also vary with productivity and surface water mixing, meaning that $\delta^{13}$C can be used as an indication of palaeo-productivity in the oceans (Brasier, 1995).

Changes in climate, caused by or resulting in a change in greenhouse gases (including CO$_2$), can also be reflected in changes in $\delta^{13}$C. This is due to there being an equilibrium between the partial pressure of CO$_2$ in the oceans and that in the atmosphere, and an equilibrium isotopic fractionation between the two reservoirs (Marshall, 1992).
Stratigraphic variations in the carbon isotopic composition of marine carbonates can be interpreted in terms of changes in the balance of the different components of the carbon cycle. For example, positive carbon excursions have been interpreted as reflecting times of increased organic productivity and/or enhanced organic deposition within the oceans (eg. Jenkyns, 1980; Jenkyns and Clayton, 1986). An increase in the rate of withdrawal of isotopically negative carbon in the form of organic matter leaves the residual dissolved inorganic carbon isotopically heavy. Conversely, negative shifts apparently reflect breakdown of oceanic productivity and the large scale return of carbon from the organic biomass to dissolved inorganic carbon (eg. Zachos et al., 1989).

5.9.2. Post-depositional isotope fractionation in sedimentary and organic carbonates

During diagenesis and metamorphism, primary δ13C isotopic signatures are prone to suffering secondary overprints as part of the overall changes experienced by their host rocks during their geological history. For example, carbonate sediments deposited in shallow marine environments may be exposed to highly corrosive meteoric waters charged with CO2. Under such conditions, the unstable primary mineralogical assemblage of high-Mg calcite and aragonite transforms to the more stable low-Mg calcite by sequential dissolution-reprecipitation (Land, 1967; Bathurst, 1975; Tucker and Wright, 1990). However, this is a rock dominated system in which the precursor mineral phases play a dominant role in controlling the chemistry and isotopic composition of waters in the microenvironment. Diagenetically stabilised sedimentary carbonates can therefore preserve the isotopic composition of original carbonate muds within approximately ±1‰ of primary values (Schidlowksi and Aharon, 1988). Thus, carbonates can closely reflect the 13C/12C ratio of the parent bicarbonate pool, making the marine δ13C record a fairly reliable sensor for the isotopic evolution of oceanic HCO3⁻.

The carbon-isotope value of marine organic matter is typically ~25‰ more negative than inorganic carbonate (Marshall, 1992). The degradation of organic matter therefore has the potential to significantly alter the carbon-isotope composition of marine carbonate sediment. Reactions of organic matter degradation include microbially-mediated reduction of iron, manganese, sulphate and nitrate under sub-oxic conditions, and methanogenesis.
under anoxic conditions (Coleman and Raiswell, 1981; Glumac and Walker, 1998). Most of these reactions release organogenic carbon, which decreases the $\delta^{13}C$ value of dissolved bicarbonate and any precipitated carbonate phases (Irwin et al., 1977).

During metamorphism, carbonates with siliceous impurities can undergo decarbonation reactions in which the primary carbonate phases (limestone and dolomite) react with Ca-Mg silicates, with a concomitant release of CO$_2$. Decarbonation processes of this type start at 300-400°C and increase with metamorphic grade. The $\delta^{13}C$ of metamorphosed impure carbonates can therefore be shifted by 2-5‰ in a negative direction (Kaufman and Knoll, 1995).

5.9.3. Oxygen-isotope fractionation

Oxygen has three stable isotopes: $^{16}O$ (forming ~99.63% abundance), $^{17}O$ (~0.0375%), and $^{18}O$ (~0.1995%) (Anderson and Arthur, 1983). Unlike carbon, oxygen-isotope fractionation is strongly temperature dependent. $\delta^{18}O$ is depleted by approximately 1‰ for every 4°C increase at near-surficial temperatures (Marshall, 1992). The oxygen isotopic composition of modern surface waters is also affected by evaporation, condensation, and the mixing of water bodies.

Biogenic fractionation (although not likely to be of significance in Precambrian sequences) can also result in the non-equilibrium fractionation of oxygen isotopes in biominerals.

5.9.4. Post-depositional isotope fractionation in sedimentary and microbial carbonates

Unlike carbon, the proportion of oxygen in the diagenetic microenvironment originating from the precursor carbonate phases rather than the pore waters is negligible. The oxygen isotopes of diagenetically altered carbonates are therefore unlikely to represent their primary isotopic composition. The elevated temperatures that occur when a sediment is buried to depths greater than 400m results in cementation and recrystallisation processes that act to decrease $\delta^{18}O$ values (Marshall, 1992). Sub-aerial exposure also results in meteoric diagenetic alteration of $\delta^{18}O$ values by the precipitation of carbonate depleted in
Evaporation, however, can preferentially remove lighter \(^{16}\)O from the upper parts of the water mass, resulting in carbonates depleted in \(^{16}\)O (Pierre, 1988).

### 5.10. Elemental chemistry of the Hadash Formation

The concentrations of different elements (eg. Mn, Sr, Mg, Ca, and Fe) within a carbonate can be used to help investigate the degree of alteration associated with meteoric and burial diagenesis and dolomitisation (eg. Brand and Veizer, 1980; 1981; Marshall, 1992). In the diagenetic environment, stabilisation of an original metastable carbonate assemblage is achieved through complementary textural, mineralogical and chemical changes involving the dissolution-reprecipitation reaction:

\[
\text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons \text{Ca(HCO}_3)_2
\]

Substitution of Ca\(^{2+}\) by trace elements such as Sr\(^{2+}\), Mn\(^{2+}\), Fe\(^{3+}\), Pb\(^{2+}\), Zn\(^{2+}\), and Na\(^{+}\) in the CaCO\(_3\) lattice can occur to varying degrees due to different partition coefficients and large compositional differences in marine and meteoric water (Brand and Veizer, 1980). Open or partially closed diagenetic systems and/or single or multiple dissolution-reprecipitation events will lead, in general, to a decrease in concentrations of elements in which the partition coefficient \((K_{\text{calcite-water}}) < 1\) (Sr\(^{2+}\), Na\(^{+}\), Mg\(^{2+}\)), and to an increase for those with \(K > 1\) (Mn\(^{2+}\), Fe\(^{3+}\), Zn\(^{2+}\)). The greater the deviation of a particular coefficient from unity, the stronger the depletion or enrichment for a given degree of diagenetic equilibration with meteoric water.

Elemental and stable-isotope data for the 83 selected samples from the Hadash Formation are presented in Table 5.2. The full stable-isotope data set is presented in Appendix C.

#### 5.10.1. Mg/Ca ratios

The oceanic Mg/Ca ratio is largely controlled by hydrothermal weathering of basalts at mid-ocean ridges. As seawater is pumped through the ridge, Mg is preferentially extracted to form clay minerals such as chlorite and epidote. Thus, lower Mg/Ca ratios can be expected during periods of high rates of seafloor spreading, as occur during periods of high
sea-level stand (Tucker and Wright, 1990). It is believed that dolomite, aragonite, and high Mg-calcite are more likely to form when Mg/Ca ratios in seawater are high, and low-Mg calcite when they are low. Dolomite is currently thought to form in evaporitic environments, in meteoric-marine mixing zones, or on burial (Hardie, 1987; Tucker and Wright, 1990). Anaerobic microbial activity may also promote dolomite precipitation (Wright, 1997; Warthmann et al., 2000). Dolomite is often considered to be a problem mineral since it is difficult to study its formation in a laboratory. The reasons why dolomite was far more common than limestone in the Precambrian are also still not fully understood (Tucker and Wright, 1990).

The Mg/Ca ratios of the Hadash Formation in the Jebel Akhdar show variation from locality to locality (Table 5.2), as well as within individual sections. Dolomite samples from across the whole Jebel Akhdar have an average Mg/Ca ratio of 0.617 with a standard deviation of 0.098, whilst the limestones have an average ratio of 0.208 with a standard deviation of 0.126. At both of the localities in Wadi Mistal, there is a distinct drop in the Mg/Ca ratio in the upper part of the Hadash Formation, associated with the change from dolomite to limestone. A slight drop in the Mg/Ca ratio in the upper part of the formation also occurs in Wadi Mu'aydin. This drop is less marked than in Wadi Mistal (an average of 0.640 falling to 0.401 compared to 0.663 falling to 0.163), and is not accompanied by a change of carbonate mineralogy. Wadi Hajir is dominated by dolomite (Mg/Ca ratio of 0.604), but the presence of limestone in Wadi Bani Awf is highlighted by the low average ratio there of 0.343.

The Mg/Ca ratio can be used to characterise the carbonate mineralogy of a sample. Labelling a sample as ‘dolomite’ or ‘calcite’ can often be too simplistic, as often both phases will be present. ‘Ideal calcite’ would have Mg/Ca molar ratios approaching zero, whereas ‘ideal dolomite’ would have Mg/Ca molar ratios close to 1, corresponding to a weight ratio of approximately 0.6 (Veizer, 1983; Nicholas, 1996). Mg/Ca ratios are plotted against 1000Sr/Ca in Fig. 5.5. This shows the dolomites of Hadash Cap 1 and Wadi Hajir plotting well into the ‘ideal dolomite’ field. The limestone samples in general show greater variation in their Mg/Ca ratios. This is particularly apparent in Wadi Bani Awf where there is also considerable spread in the Mg/Ca ratios of the dolomites.
5.10.2. Strontium (Sr) and Sr/Ca ratios

Strontium (Sr) in the ocean comes from two primary sources – the continental crust (by subaerial weathering and river transport), and the oceanic crust (by hydrothermal activity at the mid-ocean ridges and submarine alteration of basalts) (Tucker and Wright, 1990).

Sr values for the Hadash Formation vary from 68 to 564 ppm, with the majority being between 100 and 300 ppm (Table 5.2). These values are slightly lower than reported Sr values from dolomites forming in modern hypersaline and evaporitic environments (600-900 ppm), but are comparable to those forming in modern mixing-zones (100-250 ppm). Similar Sr values have been recorded from other Neoproterozoic cap carbonates (e.g. Kennedy, 1996; Myrow and Kaufman, 1999) and are not uncommon in the Precambrian (e.g. Tucker, 1983; McCarron, 2000). Low Sr concentrations are often considered to indicate diagenetic alteration to some extent (e.g. Brand and Veizer, 1980; Tucker, 1983; Kennedy, 1996). Diagenetic pathways have been modelled as a function of decreasing Sr/Ca and increasing Mn concentrations (Brand and Veizer, 1980; Veizer, 1983). 1000Sr/Ca against Mn ppm is plotted for the Hadash Formation in Fig. 5.6. Mn is notably high in all the samples. Although Mn concentrations above 10,000 ppm have previously been recorded from Precambrian carbonates (e.g. Veizer et al., 1992), the high values within some of the samples possibly indicate that clays as well as carbonate were leached during preparation (Appendix C). Only the samples from Hadash Cap 1 show a relationship involving a decrease in Sr/Ca as Mn increases. This suggests that the samples with low Sr concentrations here may have experienced higher levels of diagenesis.

5.10.3. Iron (Fe)

Due to their low concentrations in seawater, modern carbonates have low concentrations of Fe and Mn (tens of ppm). In contrast, high values of these elements are common in diagenetic porewaters, especially if they have negative oxidation/reduction potential (Eh), which is generally the case in burial environments (Tucker, 1986). Within the Hadash Formation, Fe (as well as Mn) concentrations are high, ranging from 400 to 23,500 ppm, and mainly being above 2000 ppm (Table 5.2). This may simply reflect high levels of
these elements in the diagenetic fluids. Alternatively, the elevated Fe concentrations may be related to seawater chemistry (Grotzinger and Knoll, 1995).

5.10.4. Mn/Sr as a proxy for diagenetic exchange

As has already been mentioned, Mn and Sr have markedly different partition coefficients in calcite, Mn being preferred over Sr to replace Ca in the calcite lattice during dissolution-recrystallisation events. As this should lead to Mn being incorporated into carbonates and Sr being expelled during diagenesis, the ratio of Mn/Sr can act as a useful tool for screening samples (eg. Tucker, 1986; Marshall, 1992; Kaufman and Knoll, 1995). Correlation with isotopic ratios has been taken to suggest that diagenetic recrystallisation of carbonate incurs predictable trace element exchange in an open system (Marshall, 1992). However, absolute concentrations of Mn, Fe, and Sr are dependent on a number of factors including their availability, local redox conditions, and the water/rock ratio (Tucker, 1986). Kaufman et al. (1993) and Kaufman and Knoll (1995) had highest confidence in stable isotope results from carbonates with Mn/Sr ratios of <3, but suggested that carbonates with Mn/Sr <10 also commonly retain near primary $\delta^{13}$C abundances.

The Mn/Sr plot for the Hadash Formation in the Jebel Akhdar is shown in Fig. 5.7. The data is relatively scattered, with only approximately one-third of the samples yielding an Mn/Sr ratio of <10 (27 of 83 samples). This immediately suggests diagenesis may have affected the $\delta^{13}$C values obtained from a number of the samples (Kaufman and Knoll, 1995). With progressive diagenesis, an increase in the concentration of Mn at the expense of Sr would be expected, however, and there is little evidence for this from Fig. 5.7. Correlation coefficients ($r$) for Mn and Sr have been calculated for each lithology the drilled (for calculation of $r$, see Appendix C.4). Note that these lithologies reflect the texture and composition of what was drilled and are not linked to specific lithofacies. The only lithology that shows covariance of any significance between Mn and Sr is calcite microspar (the high negative coefficient obtained from the Dolomite spar being of limited application because of the low number of samples run from that lithology). The calcite microspar exhibits a positive correlation – suggesting that both Mn and Sr increase in value together – the opposite to what would be expected from diagenetic effects. This may
therefore be indicative of an influence from leached clays, rather than dissolution-recrystallisation events.

Mn and Sr data sets from each locality have also been plotted individually, to look at possible local diagenetic effects (Fig. 5.8). The lower part of the Hadash Formation (Facies E1-E3/W1-W2) has been plotted separately to the upper part (Facies E4/W3) in each case, to further screen for small-scale diagenetic patterns. The data from Hadash Cap 1 show some negative covariation \( r = -0.512 \) for the lower part of the formation, suggesting enrichment of Mn at the expense of Sr, and possible meteoric diagenesis at this locality (cf. McCarron, 2000). There is no consistent pattern of covariation shown by the other plots, however.

Although the general lack of significant covariance between Mn and Sr suggests that the samples have not been systematically recrystallised in an open diagenetic system, the overall high Mn/Sr ratios suggest that \( \delta^{13}C \) may have still have been affected by diagenesis. To investigate this, \( \delta^{13}C \) has been plotted against Mn/Sr for covariant analysis (Fig. 5.9). This shows no consistent change in \( \delta^{13}C \) with raised Mn/Sr values through the Hadash Formation. Samples composed of calcite spar show some positive covariation with increased Mn/Sr \( (r = 0.605) \). However, the majority of these samples have Mn/Sr ratios < 10 suggesting that they probably retain near primary \( \delta^{13}C \) abundances. Again, the high negative correlation coefficient for samples composed of dolomite spar is a function of the low number of samples.

Mn/Sr has also been plotted against \( \delta^{13}C \) for the lower and upper parts of the Hadash Formation at each locality (Fig. 5.10). High correlation coefficients obtained for Facies W3 at the top of the Hadash Formation in Wadi Hajir and Wadi Bani Awf \( (-0.945 \text{ and } 0.762 \text{ respectively}) \) are interesting, but may not be significant as they are based on only a few data points. There is some indication that \( \delta^{13}C \) moves to less negative values with increasing Mn/Sr ratios in the upper part of Hadash Cap 1 (Facies E4) and the lower part of Wadi Bani Awf (below Facies W3). In the lower part of Hadash Cap 1, however, \( \delta^{13}C \) becomes increasingly negative with increasing Mn/Sr \( (r = -0.399) \). It is possible that this covariance is a function of diagenetic alteration. However, the low values of \( r \) and the
wide dispersion of the data around the regression line, indicate that any diagenesis has not consistently altered the δ¹³C values.

5.10.5. Mn and Fe as a proxy for diagenetic exchange

Both Mn and Fe are preferentially incorporated into the calcite lattice under reducing conditions. Such conditions rarely occur in the Phanerozoic depositional environment, but are common in the diagenetic environment (Brasier et al., 1992). Hence covariation of either of these elements with δ¹³C or δ¹⁸O should indicate alteration of the isotopic signal through diagenesis. Where such covariation occurs, it can be used to extrapolate back to primary marine values (Marshall, 1992). Within the Hadash Formation, systematic covariance between Mn or Fe and δ¹³C or δ¹⁸O only occurs locally (Fig. 5.11). At Wadi Hajir 1, negative covariance (r = -0.771) occurs between Fe and δ¹⁸O within the Hadash Formation. Similar negative covariance occurs between Fe and δ¹⁸O in the lower part of the Hadash Formation (Facies E1 to E3) at Hadash 1 (r = -0.587). At Hadash 1, there is also correlation between Mn and both δ¹³C and δ¹⁸O within Facies E4 at the top of the Hadash Formation (r = 0.742 and r = 0.590 respectively). Other high correlation coefficients obtained are discarded as they are derived from datasets with 5 or fewer points. No other significant relationships were recorded. The Mn and Fe backstripping method described in Marshall (1992) therefore has only limited potential for the data from the Hadash Formation.

This general lack of correlation between isotopic ratios and Mn and Fe concentrations could be used to suggest that their concentrations are not linked to diagenesis in an open system, but were determined instead by redox conditions, or by their availability in the depositional system (McCarron, 2000). The high Fe and Mn contents could be explained, for example, by deposition in a humid post-glacial climate, when higher continental discharge, hydration, content of organic matter, and lower Eh and pH would occur. This would lead to a higher content of dissolved, colloidal and suspended Fe in surface waters and its precipitation, coagulation and settling upon entering marginal marine basins (Tucker, 1992; Krauskopf and Bird, 1995).
An alternative suggestion to this is that the abundant Fe and Mn, and negative oxygen isotope values indicate the occurrence of recrystallisation in the presence of reduced fluids more typical of local argillaceous material in a mixing or burial environment than marine water at the sea floor (Kennedy, 1996). This interpretation is possibly more consistent with the high concentrations of both Mn and Fe within the Hadash Formation, and the lack of any consistent vertical trends in relative Fe or Mn enrichment through the formation that can be correlated laterally (Table 5.2). Such diagenesis would not affect $\delta^{13}$C values if it occurred where there was a low water/rock ratio (Kennedy, 1996).

As has been mentioned earlier, where Fe and Mn concentrations are very high, it is possible that some leaching of clays occurred alongside carbonate dissolution during preparation. Thus, the high values do not necessarily indicate higher levels of diagenesis.

5.10.6. Carbon and oxygen isotopes

Covariation between $\delta^{13}$C and $\delta^{18}$O can also be used as a screen for potential diagenesis. Where a positive relationship occurs, only the most enriched $\delta^{13}$C and $\delta^{18}$O values are considered to be potentially unaltered (Fairchild et al., 1990). Fig. 5.12 shows a plot of $\delta^{18}$O against $\delta^{13}$C for the whole Hadash Formation split into the different lithologies drilled. Each lithology shows a positive correlation, but this is not particularly strong in any case ($r = 0.524$ for calcite microspar representing the highest correlation coefficient). The consistent positive slopes do suggest that there may have been some re-equilibration during burial diagenesis, however.

$\delta^{18}$O against $\delta^{13}$C has also been plotted for each locality sampled (Fig. 5.13). Positive correlations between $\delta^{18}$O and $\delta^{13}$C are again evident from these plots. The correlation is strongest in the upper part (Facies E4) of Hadash 1 where $r = 0.824$, suggesting some diagenetic re-setting of the isotopic values. High correlation coefficients are also given by the data from Wadi Bani Awf ($r = 0.746$ and $r = 0.644$ for the upper and lower parts of the Hadash Formation respectively) and from Facies E4 at Hadash 2 ($r = 0.731$). The data from these areas is generally more spread about the best-fit line than at Hadash 1, however. This is also the case for Wadi Hajir 1, where the regression line for the lower part of the formation doesn’t describe the data particularly accurately. However, at Wadi Hajir 1
there is a cluster of data with low $\delta^{18}$O values (ringed on Fig. 5.13) that correspond to the upper part of Facies W2. These low $\delta^{18}$O values suggest that all the points in this cluster have undergone significant diagenesis. Two points from this portion of the Hadash Formation at Wadi Hajir 1 have less negative $\delta^{18}$O values and also much less negative $\delta^{13}$C (-2‰ compared to -4‰). This suggests that both $\delta^{18}$O and $\delta^{13}$C may have been diagenetically reset to more negative values here. Interestingly, if just the upper part of Facies W2 is taken at Wadi Bani Awf and Wadi Hajir 1, correlation coefficients between $\delta^{13}$C and $\delta^{18}$O of 0.911 and 0.804 are produced respectively. These further suggest resetting of both $\delta^{18}$O and $\delta^{13}$C to more negative values, and indicate that the large spread of data about the regression line in Figs. 5.13f) and 5.13h) masks some important local relationships.

Oxygen-isotopic compositions can also be used on their own as sensitive indicators of diagenesis, as isotopic exchange with meteoric, burial, or hydrothermal fluids can cause a decrease in $\delta^{18}$O without necessarily affecting $\delta^{13}$C. Kaufman and Knoll (1995) consider that $\delta^{18}$O values of limestones that are <-5‰ indicate a degree of oxygen-isotopic alteration; samples with $\delta^{18}$O <-10‰ are considered unacceptably altered. Although, very few of the samples in the present study have $\delta^{18}$O values <-10‰, the ones that do are dominantly composed of calcite spar (Fig. 5.12). This is consistent with the observation that the calcite spar probably formed later than the dolomite (Section 5.6.4), and indicates it may have formed under the influence of diagenetic fluids. In general, both the dolomite and calcite microspar lithologies maintain less negative $\delta^{18}$O values than the two coarser spar lithologies. This suggests the finer-grained microspar may be more likely to preserve primary isotopic signals.

5.10.7. Component analysis

Component analysis differentiates the amount of alteration between least altered (eg. micrite) and most altered (eg. sparry cement/recrystallised spar) fractions within a single sample. Unfortunately, the Hadash Formation does not lend itself very easily to analysis in this manner, as there is very little micrite or other fractions that would be considered least altered present. Much of the Hadash Formation is composed of relatively homogenous microspar, and it is only in Facies E4 that two different components are present (dolomite
spar and calcite spar). From its cloudier, coloured nature, and from possible neomorphic
textures in the calcite, the dolomite is considered to be less altered. Two samples (HSH 70
and HSH 73) from Facies E4 of Hadash 1 drilled from dolomite-rich areas have markedly
lower δ¹³C and δ¹⁸O values compared to the calcite-dominated areas. However, other
dolomite samples from the same facies (eg. HSH 66 and 67) exhibit no such enrichment in
δ¹³C and δ¹⁸O, and at Hadash 2, calcite samples from Facies E4 do not show the same
depletion apparent at Hadash 1. No consistent trend of enrichment/depletion in isotopic
values between calcite and dolomite spar could therefore be identified within Facies E4.

5.11. Data correction

As has briefly been touched on above, a number of the screens for diagenesis can be used
to estimate and then correct data for diagenetic overprints (Fairchild et al., 1990; Marshall,
1992; McCarron, 2000). These corrections can then be used as a control to assess how
robust isotopic trends are to the effects of diagenesis. Where a consistent trend occurs
between an independent diagenetic variable and δ¹⁸O or δ¹³C, the isotopic value can be
‘backstripped’ to least-altered values (Marshall, 1992). However, as has been
demonstrated above (Section 5.10), few such consistent trends occur within the Hadash
Formation. The Hadash Formation at Hadash 1 may have been reset in a negative
direction during diagenesis. This negative shift is indicated by a negative correlation
between Mn and Sr (Figs. 5.6 and 5.8), a negative correlation for the lower part of the
formation between Mn/Sr and δ¹³C (Fig. 5.10), and good correlations, especially for the
upper part of the formation (Facies E4), between δ¹⁸O and δ¹³C (Fig. 5.13). Backstripping
to primary values is difficult for the lower part of the Hadash Formation at Hadash 1 as the
correlation between Mn/Sr and δ¹³C is not particularly strong. However, the strong
correlation between δ¹⁸O and δ¹³C for Facies E4 at the top of the Hadash Formation allows
some correction to be made. It is worth noting, however, that Marshall (1992) suggested
patterns of covariation in carbon and oxygen values should be used only as a last resort as
they may reflect diagenetic alteration, but may also reflect important primary variation.
The corrected δ¹³C curve should therefore still be treated with some caution.

δ¹⁸O values of −4‰ have been considered to be typical of least-altered Precambrian
dolomites, with δ¹⁸O values of −5.5‰ to −6.5‰ typical for least-altered limestones (eg.

122
Brand and Veizer, 1981; Veizer et al., 1992; Kaufman and Knoll, 1995). Individual data points for Facies E4 from Hadash 1 cannot be corrected since the regression line is only defined by the data set as a whole (Fairchild et al., 1990). As Facies E4 is dominantly composed of limestone, the covariation apparent in Fig. 5.13b has been used to correct the data set from Facies E4 as a whole to an assumed primary δ¹⁸O value of −5.74‰ (the most enriched δ¹⁸O value recorded from the facies). This corresponds to a δ¹³C value of approximately −1‰ (Fig. 5.14b). This is a relatively crude correction as the δ¹⁸O value at the time of deposition may have been significantly different from that assumed, and any variation that may have naturally occurred in δ¹⁸O or δ¹³C values during the deposition of Facies E4 is not accounted for and is effectively ignored.

A similar procedure could be followed to correct the data from the upper part of Facies W2 in Wadi Bani Awf and at Wadi Hajir 1 (Figs. 5.14f and 5.14g), where significant covariation between δ¹⁸O and δ¹³C also occurs (Fig. 5.13). Such a correction would lead to this portion of the δ¹³C curve being corrected to approximately −1‰ and −2‰ respectively for the two localities. This is a very simplistic correction and the data from Wadi Bani Awf and Wadi Hajir will be discussed in more detail below (Section 5.12).

Further data corrections cannot reasonably be made due to the lack of consistent covariation between δ¹⁸O or δ¹³C and other diagenetic indicators. This could be taken to suggest that the δ¹³C curve for the Hadash Formation should be reasonably robust (McCarron, 2000). However, any interpretations of the stable isotope curves do need to take into account the results from this section, and the potential diagenesis indicated at some localities.

5.12. Discussion of stable isotope data from the Hadash Formation

The aim of this isotopic and geochemical study was not only to use the data to aid in the interpretation of the Hadash Formation and its diagenetic history, but also to investigate whether the δ¹³C signatures can be used at this <15m resolution to provide a correlatory tool that is independent of lithology and that can be used both locally and globally. The carbon- and oxygen-isotope curves from the Hadash Formation are shown in Figs. 5.14a-h and 5.15. These data show that the δ¹³C values from the Hadash Formation in the Jebel
Akhdar are strongly negative (Figs. 5.14a-g), whereas the $\delta^{13}$C values from the carbonate in the Huqf area vary from slightly negative to slightly positive (Fig. 5.14h).

The large difference in the values of $\delta^{13}$C obtained from the carbonate in the Huqf compared to those obtained from the Hadash Formation in the Jebel Akhdar (Fig. 5.15), initially suggests that a direct correlation between these two areas based on lithological similarity at this level is not valid. $\delta^{18}$O values from the two areas are similar, and there is no covariation between $\delta^{13}$C and $\delta^{18}$O from the Huqf carbonate, suggesting that the difference in the $\delta^{13}$C profiles is not linked to diagenetic effects. The $\delta^{13}$C values obtained from the carbonate in the Huqf may represent an intermediate part of the $\delta^{13}$C curve between the negative results obtained from the Hadash Formation and the positive results from the top of the Masirah Bay Formation (Chapter 6). If this is the case, it would suggest that the onset of Masirah Bay Formation deposition occurred later in the Huqf compared to the Jebel Akhdar. Alternatively, the difference in depositional environments between the shallow-water Huqf and the deeper-water Jebel Akhdar provides another possible explanation for the difference between the $\delta^{13}$C signatures. It may be that the carbonates were deposited at the same time, but the $\delta^{13}$C signatures they preserve were affected by local effects (such as upwelling or a $\delta^{13}$C gradient in the water column) and consequently differ. The fact that the carbonate in the Huqf area is also a transgressive deposit, and that models of cap carbonate formation suggest that conditions promoting carbonate deposition should be widely developed during postglacial transgression suggest that it may be more likely that they are quite closely linked in time.

Although the curves in the Jebel Akhdar are noisy, larger-scale trends towards more positive or more negative values can be seen within the $\delta^{13}$C signal from the Hadash Formation. In the east of the Jebel Akhdar (Wadi Bani Jabir, Hadash 1 and 2, and Wadi Mu’aydin), a similar $\delta^{13}$C signature can be seen within the lower part of the Hadash Formation (Facies E1). A negative shift in the $\delta^{13}$C values just above the base of the formation occurs (eg. $-4.1\%o$ to $-5.2\%o$ in Wadi Mu’aydin), followed by a move to more positive values (eg. $-2.9\%o$ in Wadi Mu’aydin). Then, before the top of Facies E1 is reached, another kick towards more negative values occurs. This same pattern can be seen at all four of the more easterly localities even though absolute values of $\delta^{13}$C vary, often to
a large extent (eg. values of $\delta^{13}$C concentrated around $-6\%_o$ (but up to almost $-8\%_o$) at Hadash 1 for the first negative excursion compared to $-4\%_o$ at Hadash 2 and $-3\%_o$ in Wadi Bani Jabir). This variation in absolute $\delta^{13}$C values from locality to locality cannot easily be explained in terms of differing depositional environments as localities Hadash 1 and Hadash 2 are only a few hundred metres apart and share very similar facies throughout their exposure. It is therefore more likely to be a reflection of diagenetic alteration. For example, Hadash 1 contains some of the most negative isotopic values. It also exhibits the most variable or ‘noisy’ $\delta^{13}$C curve, with differences between successive samples in Facies E1 as large as $7\%_o$. As mentioned above, the negative correlation between Mn and Sr and the negative correlation between Mn/Sr and $\delta^{13}$C (Fig. 5.10) that is recorded for the lower part of the Hadash Formation at Hadash 1 suggests a greater degree of diagenetic alteration compared to the other localities. This provides one possible explanation for the negative shift in some of the $\delta^{13}$C values.

Another possible explanation arises from the fact that the Hadash Formation was sampled at a greater density at Hadash 1 than at the other localities. It is possible that the large variation in $\delta^{13}$C at Hadash 1 represents the primary signal, and that some of the details of this may simply have been missed by the other curves. Indeed, if the $\delta^{13}$C curve from Hadash 1 is plotted only for those samples that were run for elemental geochemistry (reducing the number of data points to 29), the shape of it appears much more closely aligned to the curve from Hadash 2 compared to when all the data was plotted (Fig. 5.16).

Moving further up through the Hadash Formation in the east of the Jebel Akhdar, clear shifts in the $\delta^{13}$C trend become less obvious. At Hadash Cap 1, there is a general trend towards less negative $\delta^{13}$C values from approximately $-4\%_o$ to $-2\%_o$. This becomes more apparent (falling to $-1\%_o$) if the correction to the data from Facies E4 at the top of the formation is applied. There is less noise in the $\delta^{13}$C curve than lower in the formation, but there is still variation from point to point of up to $2\%_o$. Within Facies E2 and E3, there is also significantly less variability in the $\delta^{18}$O curve (Figs. 5.14 and 5.15).

Again, although the actual values are lower, the general trend to less negative $\delta^{13}$C values can also be seen at Hadash 2, where they decrease to $-1\%_o$ from $-3\%_o$. However, at Wadi
Mu'aydin, $\delta^{13}C$ values remain relatively constant (at $-4\%$), before becoming increasingly negative at the top of Facies E3 and in Facies E4. Three out of the upper four samples here show signs of having undergone significant diagenesis (Mn/Sr $>10$ or $\delta^{18}O < -10\%$). A negative shift in $\delta^{13}C$ values associated with this diagenesis is one possible explanation for the difference in the curve here compared with Hadash 1 and 2.

Only two data points were collected from the upper part of the Hadash Formation in Wadi Bani Jabir (Facies E5). These record one value of $-1\%$ and one of $-3\%$. As the data set is so poor here (a function of the lack of carbonate), the $\delta^{13}C$ results from the upper part of the Hadash Formation cannot be meaningfully compared to the other localities.

The $\delta^{13}C$ signature obtained from the Hadash Formation in the west of the Jebel Akhdar (Wadi Hajir 1 and 2, and Wadi Bani Awf) is different in many respects from those obtained further to the east. $\delta^{13}C$ values are still negative, but the same excursions that are seen within Facies E1 in the east are not readily apparent. At the base of the Hadash Formation, instead of the negative-positive-negative excursions noted in the east, $\delta^{13}C$ values are reasonably constant. Values stay stable at $-4\%$ in Wadi Bani Awf and at Wadi Hajir 1, and at $-3\%$ at Wadi Hajir 2, where the base of the formation is actually composed of Facies E1 rather than Facies W1.

Although sampled at a much lower resolution, Wadi Hajir 2 exhibits a broadly similar signal to Wadi Hajir 1 for much of its thickness. Above the initially unvarying $\delta^{13}C$ signal, there is a small positive excursion, followed by an overall move to gradually increasingly negative values up-section. This trend ends in Facies W3, where despite some variation, $\delta^{13}C$ essentially remains strongly negative (values of $-6\%$ to $-3\%$).

The $\delta^{13}C$ signal from the Hadash Formation at Wadi Bani Awf has a slightly different character to that seen in Wadi Hajir. The $\delta^{13}C$ curve above the lowermost unvarying portion shows more variation (or noise) between individual data points than occurs in Wadi Hajir. Despite this, a more marked positive excursion can be seen in the $\delta^{13}C$ curve (to values as high as $-1\%$) in Facies W2 in Wadi Bani Awf. Above this, in Facies W3,
more negative values of $\delta^{13}C$ (-5 to $-6\%_o$) are re-established, similar to those recorded in Wadi Hajir.

The reasons for the more marked shift to less negative values in the Hadash Formation of Wadi Bani Awf compared to Wadi Hajir can perhaps be explained by consideration of the $\delta^{18}O$ curves from the different localities. The $\delta^{18}O$ curve from Wadi Bani Awf is highly variable, with values varying between $-2\%_o$ and $-10\%_o$. This suggests some significant resetting of values by diagenesis. In the lower part of the formation, $\delta^{13}C$ values are relatively constant, suggesting they have been largely unaffected by this diagenesis. However, within the upper part of the Hadash Formation (Facies W2 and W3), the $\delta^{13}C$ signature is far more noisy, and strong covariation between $\delta^{18}O$ and $\delta^{13}C$ occurs (Fig. 5.13) ($r = 0.911$ for upper part of Facies W2 in Wadi Bani Awf). As the least negative $\delta^{13}C$ values correspond to the least negative $\delta^{18}O$ values, it appears that the diagenesis that has locally reset some of the $\delta^{18}O$ has also locally reset the $\delta^{13}C$, and the least negative values probably most closely reflect the primary signal. At Wadi Hajir 1, there is less variation in $\delta^{18}O$, but there is a marked negative shift to relatively constant values of $\sim-8\%_o$ within Facies W2 (Fig. 5.14f). This is also reflected in the cluster of data ringed on the $\delta^{13}C$ vs $\delta^{18}O$ diagram (Fig. 5.13f). Below this shift in the $\delta^{18}O$ curve, there is a small but steady rise in $\delta^{13}C$. At the $\delta^{18}O$ change, the rise is halted, and $\delta^{13}C$ values dominantly around $-4\%_o$ persist for the remainder of Facies W2. This suggests that whereas in Wadi Bani Awf only some of the $\delta^{13}C$ values in the upper part of Facies W2 have been shifted in a negative direction by diagenesis, at Wadi Hajir 1, almost all of them have been. The positive excursion recorded in Wadi Bani Awf is therefore much less evident in Wadi Hajir, where the $\delta^{18}O$ indicates that the $\delta^{13}C$ values have been diagenetically reset. If the $\delta^{13}C$ curve at this level in both wadis is corrected (using the covariation between $\delta^{13}C$ and $\delta^{18}O$), a large positive excursion is seen at both Wadi Hajir 1 and in Wadi Bani Awf (Figs. 5.14f and 5.14g).

A possible explanation as to why the upper part of Facies W2 has been diagenetically altered to a greater extent than the rest of the Hadash Formation is provided by the small hollows that occur on its top surface (Section 5.6.7). These possibly represent the effects of dissolution, either before the deposition of Facies W3, or through later fluids moving.
along the boundary between the two facies. The fluids responsible for the dissolution may have had an effect on the upper part of Facies W2 that did not extend to the lower levels of the Hadash Formation. Recrystallisation of the carbonate in the upper part of Facies W2 in the presence of this fluid may then have led to significant alteration of the primary isotopic signal observed there.

5.13. Use of stable isotopes for intra-basinal high resolution correlations within the Hadash Formation

Some similar patterns in the δ¹³C signature from different localities can be identified (especially after diagenetic effects are accounted for), and it would be possible to make a case that these reflect changes in the primary isotopic signal and can be used to tie together the different sections independently of facies. For example, it could be argued that the relatively constant δ¹³C values of −4‰ at the base of the Hadash Formation in the west of the Jebel Akhdar correspond to the δ¹³C signal from Facies E2 and the lower part of Facies E3 in the east. Although there is more noise at this level in the east, at both Wadi Mu'aydin and Hadash 1, values are centred around −4‰. The move to less negative δ¹³C values that occurs towards the top of the Hadash Formation at Hadash 1 and 2 would then be correlatable to the positive excursion within Facies W2 in Wadi Bani Awf. As discussed above, this excursion is probably less defined in Wadi Hajir because of diagenetic effects, and if crude corrections are made, the correlation becomes more apparent. Such correlations would suggest that deposition of the Hadash Formation began later in the west compared to the east. They also suggest carbonate deposition continuing in the west in the form of Facies W3 after it had ceased in the east.

A diachronous base to the Hadash Formation can be envisaged from the lithological argument, stated above, that Facies E2 in the middle of the Hadash Formation in the east can be correlated to Facies W1 at the base of the Hadash Formation in the west by the presence of fine-grained siliciclastics (Figs. 5.4a and 5.4c). However, a number of different scenarios are possible (Figs. 5.4b and 5.4d) and problems arise with this idea if the Hadash Formation is taken to represent a purely transgressive deposit (Fig. 5.4d). Carbonate deposition persisting for longer in the west of the Jebel Akhdar is possibly a more reasonable conjecture. At the top of the Hadash Formation, sporadic carbonate
deposition may well have continued for longer in the more distal western part of the Jebel Akhdar, where the influence of siliciclastic influxes may have been at a lower level compared to the more proximal east (e.g. Fig. 5.4a). It therefore initially appears that the carbon-isotope curve can be used at this level to a limited extent to provide a correlatory tool that is independent of lithology.

Despite possible correlations within the Hadash Formation across the Jebel Akhdar using the stable isotope curves, there are still a number of problems. Even though lithological arguments can be made to support matching the curves in the manner described above, they can also be used to support other possible ties and there is no quantifiable independent measure with which to check the validity of any correlation. For example, it could be argued that the negative shift within Facies W2 in the west of the Jebel Akhdar actually matches the lowermost negative shift in Facies E1 in the east of the Jebel Akhdar. In this scenario, deposition would have begun earlier rather than later in the west. In some regards this is easier to explain considering the transgressive nature of the Hadash Formation, but leads to less obvious correlations of the $\delta^{13}C$ curve higher in the stratigraphic profile. The correlations suggested above are therefore not the only possible way to 'match the wiggles' in the $\delta^{13}C$ curve, and may consequently be incorrect. Models explaining the variation in the Hadash Formation across the Jebel Akhdar cannot rely on the $\delta^{13}C$ signature to provide tie points and are therefore still rather speculative (Fig. 5.4).

The variation that occurs in the values of the $\delta^{13}C$ from locality to locality, even where the same negative and positive shifts can be seen is also a problem. This variation occurs where distance between localities is only a few hundred metres, and even where samples a few cm apart from the same horizon were run. It does not seem likely therefore that this variation is linked to differing environments. Excursions of 2$\%_o$ are used to tie sections together at certain points, yet differences between these tie points of up to 2$\%_o$ in the actual $\delta^{13}C$ values can occur between separate profiles, and cannot always be easily accounted for. If diagenesis is responsible for differences in $\delta^{13}C$ values between sections, then it could also account for the excursions in the curve. These would then no longer reflect changes in the primary $\delta^{13}C$ signal, and would therefore no longer provide a stratigraphic link.
Large variation can also be seen between successive data points at the same locality. This variation often remains after diagenesis has been screened for. Where the Hadash Formation has been less densely sampled and the $\delta^{13}C$ signature is defined by fewer data points, this variation is more likely to distort the shape of the curve – possibly creating the impression of excursions, which denser sampling would reveal to be background noise. This is clearly shown in Fig. 5.16, where the excursions appear more striking when fewer data points are plotted. Excursions that are only defined by a few data points, and any correlations based on them, should therefore be treated with suspicion.

Thus, although correlations can be made within the Hadash Formation using the $\delta^{13}C$ curves, the validity of these correlations, and therefore ultimately their worth, is very questionable. Carbon isotopes have been widely employed at a much larger scale as a useful stratigraphic tool in many Precambrian successions (eg. Knoll et al., 1986; Narbonne and Aitken, 1995; Brasier et al., 1996; 2000; Saylor et al., 1998). Their worth for correlation at this scale, particularly when used alongside radiometric age dates and $^{87}$Sr/$^{86}$Sr isotopic curves, is proven. The negative $\delta^{13}C$ results from the Hadash Formation (varying between $-8\%o$ and $-1\%o$) also allow it to be fitted in to a global context when other constraints (isotopic curves from the Khufai, Shuram, and Buah Formations, and radiometric age dates) are taken into account (Chapter 4). However, from this detailed study of a thin carbonate unit in a dominantly siliciclastic succession, it does not appear that carbon isotopes can be used to correlate Neoproterozoic cap carbonates at the resolution attempted (<15m scale) with any degree of confidence, whether this is on a local, basinal, or more global scale. This conclusion has major implications for the correlation of other Neoproterozoic postglacial cap carbonates. For example, Kennedy (1996), suggested that the consistent trend to more depleted $\delta^{13}C$ values up-section within individual stratigraphic profiles in Australian Marinoan cap dolostones could be used as an independent means of assessing lithostratigraphic correlation between isolated successions. The resolution at which Kennedy (1996) sampled was much lower than in the present study (Fig. 5.17), the trend towards more depleted values up-section was not uniformly developed and the actual values of $\delta^{13}C$ vary from profile to profile. The same arguments against meaningful correlations using $\delta^{13}C$ signatures at this scale put forward above therefore also apply.
Attempts to try and accurately locate the inflection points of the isotopic curve with respect to glacial periods in the Neoproterozoic are therefore likely to meet with difficulties within such thin cap carbonate units. This has major implications for the proposal to define the base of the 'Terminal Proterozoic System' within or at the base of a Marinoan-type cap carbonate, or with reference to their presumed carbon isotopic inflections (eg. Christie-Blick et al., 1995; Knoll, 2000). Where carbonate is more persistent above Neoproterozoic glacial events (eg. Namibia (Hoffman et al., 1998b; Kennedy et al., 2001)), better constraints are possible, but still need to be assessed critically.
PART III. DISCUSSION

5.14. Discussion of the Hadash Formation

Although fine-scale correlations within Oman cannot confidently be achieved using the $\delta^{13}$C profiles from the Hadash Formation, the $\delta^{13}$C values recorded as a whole are typical of the negative values widely reported for postglacial carbonates overlying ‘Marinoan-age’ glacial deposits (Kaufman and Knoll, 1995; Kennedy, 1996; Hoffman et al., 1998b; Myrow and Kaufman, 1999; Kennedy et al., 2001). The enigmatic position of these postglacial cap carbonates, their common occurrence as the only carbonate of any thickness in a dominantly siliciclastic succession, and their distinctive isotopic signatures has provoked a large amount of debate as to how they formed. Thus, although little previous work has been done on the Hadash Formation in Oman, ideas produced from work on similar cap carbonates overlying Neoproterozoic glaciations can be used to help explain some of its features. Many other Neoproterozoic cap carbonates are also interpreted as relatively deep-water, transgressive deposits (e.g. Tucker, 1986; Kennedy, 1996; Myrow and Kaufman, 1999; Kennedy et al., 2001). Their position directly overlying extensive and possibly global (see previous chapter) glacial deposits suggests a fundamental link between their occurrence and postglacial sea-level rise. The major change in depositional style that occurs at the base of the Hadash Formation therefore arguably reflects such a change from a major glacial epoch into immediately post-glacial conditions.

Different classes of models have been proposed as to why post-glacial transgression should lead to carbonate precipitation. One class of models invokes physical stratification of the ocean during glacial conditions. In the poorly ventilated deep oceans, organic matter would be mineralised via sulphate-reduction to $\text{CO}_2$ and $\text{HCO}_3^-$ producing a strong surface to deep carbon isotopic gradient in the ocean (Knoll et al., 1986; Kaufman et al., 1993; 1997; Grotzinger and Knoll, 1995). Postglacial upwelling then delivers both alkalinity and isotopically light carbon to continental shelves and flooded interior basins. This results in massive precipitation of inorganic carbonate depleted in $^{13}\text{C}$ directly on top of glacial deposits. Such models suggest that deposition of cap carbonates is on a time-scale of $\sim 10^3$-
10⁴ years. This is based on analogy with the timing of deglaciation in the Holocene (Kennedy, 1996), and the residence time of carbon in the oceans (Kump, 1991).

Kennedy et al. (2001) suggest that such models have a number of possible weaknesses. These include: 1) the physical difficulty of stratifying the glacial ocean prior to cap carbonate deposition; 2) the positive (counter) effect that enhanced upwelling of nutrient-rich deep water and associated productivity would have engendered on surface-marine carbonate-carbon isotopic values; and 3) the fact that the carbon isotopic signature of cap carbonates would be expected to be more variable, both regionally and globally.

Although, there is some regional variation in the carbon isotopic signature from the Hadash Formation in Oman, on a global scale, with its overall strongly negative character, the Hadash Formation is similar to other reported Neoproterozoic cap carbonates.

In another class of models, a change in the carbon isotopic composition of the entire ocean is envisaged. For example, in the "snowball Earth" hypothesis (Hoffman et al., 1998a, b), this change is produced by the partitioning of the ocean and atmospheric CO₂ reservoirs by a nearly continuous carapace of sea-ice. With photosynthesis shut down and no air-sea gas exchange, after a period of time greatly exceeding the residence time of carbon in the oceans (10⁵ years), both the ocean and the atmosphere would have similar δ¹³C values equivalent to the weathering flux of -5‰ to -7‰ (Kump, 1991). During the "snowball" glacial conditions, a strong albedo feedback and diminished silicate weathering could have permitted atmospheric CO₂ to build up to high levels (350 times greater than at present) before triggering rapid deglaciation and melting of sea ice. Upon this deglaciation or meltback, gas exchange between the surface ocean and the high CO₂ atmosphere would first drive carbonate dissolution and then drive precipitation as cold deep waters with high concentrations of calcium and dissolved inorganic carbon mixed with warm tropical waters. Additional sources of alkalinity would come from intense continental weathering that was driven by warm temperatures, high levels of CO₂, and a strong hydrological cycle. As atmospheric CO₂ levels fell, enough carbon would be available to rapidly form the thin carbonates found globally at this level. Until the biological pump could re-establish itself and drive δ¹³C values upwards, the carbonate precipitated at this time would preserve the negative δ¹³C values of the glacial ocean and atmosphere (Hoffman et al., 1998b).
This model suggests long-lasting negative $\delta^{13}C$ excursions existing for the duration of "snowball" conditions ($\sim 9 \times 10^6$ years according to Hoffman et al. (1998a; b)), rather than the rapid excursions predicted by the first class of models ($10^3$ years (eg. Kennedy, 1996). Both models still invoke rapid precipitation of the cap carbonates.

Difficulties with this second model arise from the implausibility of a frozen tropical ocean under such high levels of atmospheric CO$_2$ (Chandler and Sohl, 2000; Hyde et al., 2000), and from the high rates of silicate weathering required to draw down CO$_2$ in the short interval of cap carbonate deposition (Kennedy et al., 2001). Postglacial weathering rates would have to be 3-4 orders of magnitude greater than those preceding the CO$_2$ build-up, and this is not supported by associated changes in the Sr isotopes during this period (Kennedy et al., 2001).

An alternative model has recently been proposed by Kennedy et al., (2001). Similarly to the snowball model, it involves short-lived changes in the bulk carbon isotopic composition of the ocean, but it avoids some of the obvious difficulties associated with huge changes in atmospheric CO$_2$. The model suggests that the observed negative carbon isotopic excursion in cap carbonates may be due to the destabilisation of gas hydrates in terrestrial permafrost as a result of postglacial warming and marine transgression. Methane would be released as a pulse addition during climatic amelioration over a span of $10^3$-$10^4$ years. The bulk of the methane would have been oxidised by sulphate reduction within anoxic sediment near the seafloor. Sulphate reduction at seeps would be facilitated by a microbial consortium that produced alkalinity enriched in $^{13}C$ as a reaction product. Alkalinity produced by sulphate-reduction of methane enhance carbonate deposition, in contrast to pH lowering and dissolution of carbonate, which occurs when methane is oxidised to CO$_2$ in the ocean and atmosphere. Addition to the ocean of sufficient methane to drive a substantial negative carbon isotopic excursion must be balanced by an equivalent amount of carbon redeposited in a sedimentary reservoir either as biomass or carbonate. Once conditions had stabilised, recovery of isotopic values would have occurred over an interval of $\sim 10^3$ years, as the isotopically depleted stock of carbon was replaced by riverine input over several residence times (Kump, 1991).
This model is attractive, not only because changes in the gas hydrate pool are potentially capable of producing substantial shifts in the carbon isotopic composition of the ocean without requiring huge changes in atmospheric CO₂, but also because of lithological considerations. Kennedy et al. (2001), considered two anomalous and previously enigmatic cap carbonate facies that could be explained by a model invoking gas hydrate destabilisation. The first facies was characterised by abundant bedding expansion and cementation, and the second facies by a microbially laminated nature with abundant tube-like structures.

Each of the models outlined above provides a potential explanation for the occurrence of the Hadash Formation directly overlying glacial deposits and for its negative δ¹³C signature. However, as was discussed in Chapter 4, the underlying Ghadir Manqil Formation suggests that a number of distinct glacial events came and went, rather than a prolonged “snowball” earth. This questions whether the proposed partitioning of the ocean and atmosphere by ice and complete shutdown of photosynthesis, as envisaged by Hoffman et al. (1998b) to explain the negative δ¹³C excursion, could have occurred. Some of the lithological features of the Hadash Formation are similar to the ‘anomalous facies’ that Kennedy et al. (2001) suggested could be explained by gas hydrate destabilisation. Although no sheet cracks, and no domal, tepee-shaped or tube-like structures were observed in the Hadash Formation, features such as the ‘crinkly’ lamination and roll-up structures suggest cohesion, most likely by mucilaginous microbial membranes (Kennedy, 1996; Hoffman et al., 1998b; Kennedy et al., 2001). These features in the Hadash Formation, and the pyrite suggesting anoxic conditions, could be used to support the model that cap carbonate deposition was linked to gas hydrate destabilisation (Kennedy et al., 2001). It is therefore possible that the negative isotopic results from the Hadash Formation are partly related to the degradation of organic matter (for which there was some evidence) and the associated release of light carbon isotopes. However, as the isotopic signatures are so alike from similar deposits globally, it would seem that their negative nature cannot be entirely due to such local effects.

Many questions still remain to be answered about the occurrence of these peculiar cap carbonates. Unfortunately, many of these questions can not be answered form this study of the Hadash Formation. For example, attempts to try and fit the Hadash Formation into a
more tightly constrained depositional framework linking facies in the east of the Jebel Akhdar with those in the west have to be left somewhat open-ended. The fact that the $\delta^{13}$C signature cannot be meaningfully used to correlate stratigraphic profiles independently to facies at this scale, means that a number of different scenarios are still possible. The suggestion that the onset of the Hadash Formation may have begun later in the west of the Jebel Akhdar compared to the east is potentially very interesting as it would indicate (albeit possibly short-lived) highstand conditions and possible deposition after, rather than during, transgression. Other interesting features include the depressions that occur on the upper surface of Facies W2. If these do indicate a period of dissolution on the seafloor, and a lack of deposition for some time, arguments could be made suggesting that deposition of the Hadash Formation was longer-lived than the $10^3$ to $10^4$ years that current models would suggest. However, all of these suggestions must remain rather conjectural at this stage, and no new models of cap carbonate deposition have been proposed here.

So far as global correlation of the Hadash Formation is concerned, although the isotopic signature could not be used for fine-scale correlations, the negative $\delta^{13}$C signature that persists even after data correction indicates that the Hadash Formation can be linked to similar carbonates capping Varangerian-age successions elsewhere in the world (see Chapter 4). However, whether these carbonates and their associated negative $\delta^{13}$C signatures all correspond to exactly the same event is debatable. A significant negative excursion occurs throughout the Shuram Formation, and has been used to suggest that a glacial event not lithologically represented in Oman occurred at this time (Saylor et al., 1998; Brasier et al., 2000). If this is the case it is possible this records the excursion preserved in some cap carbonates that would normally be correlated to the Hadash Formation. Although some authors recognise four Cryogenian glaciations (Kaufman et al., 1997), and others only accept two (Kennedy et al., 1998), within the Fiq Member of the Ghadir Manqil Formation there is evidence for numerous discrete glacial events within the stratigraphy (Chapters 3 and 4). Above one of these glacial events, directly overlying the transgressive lag at the base of Unit F6b, a thin carbonate unit is preserved. Although the isotopic signature from this is strongly negative, it has been diagenetically reset and cannot be taken to represent a primary signal. The features preserved in this carbonate differ from the Hadash Formation. However, the presence of carbonate at this level suggests that conditions conducive to cap carbonate deposition are not unique. This is perhaps not
surprising as the models invoking postglacial overturn of stratified oceans or gas hydrate destabilisation to explain the formation of cap carbonates (discussed above) suggest that conditions conducive to carbonate deposition with negative $\delta^{13}$C are not necessarily limited to the aftermath of one long-lasting glacial event. Thus, if a number of discrete, widespread glacial events within one glacial epoch do occur (as the deposits of the Ghadir Manqil Formation would suggest), global correlations of glacial deposits based on overlying thin carbonates with negative $\delta^{13}$C are likely to be inaccurate. This has serious implications for the reliability of cap carbonates as a marker for the proposed ‘Terminal Proterozoic System’, as without further constraints such as radiometric age dates, detailed $^{87}$Sr/$^{86}$Sr isotopic curves, and well-defined fossil assemblages, it cannot be assumed that cap carbonates at approximately the same lithostratigraphic level globally were deposited synchronously.

5.15. Summary of the Hadash Formation

The Hadash Formation in the Jebel Akhdar represents a typical Neoproterozoic cap carbonate. It directly overlies the last glacial deposits of the Ghadir Manqil Formation, and is a thin, laterally extensive carbonate, possibly linked to postglacial transgression, which preserves a negative $\delta^{13}$C signature. Its lateral extent, the general lack of physical sedimentary structures, and its position above glaciomarine deposits and below deepwater siltstones suggests deposition below storm wave base. Fine-grained siliciclastic deposition accompanied carbonate sedimentation in many of the facies. Greater influxes of terrigenous material as later highstand conditions became established may explain the termination of carbonate deposition at the top of the Hadash Formation. Coarser-grained siliciclastic material and palaeocurrent directions toward the west in Wadi Bani Jabir, and the orientation of roll-over structures at Hadash 1 suggest the west of the Jebel Akhdar was in a more distal position than the east during deposition of the Hadash Formation.

A detailed study of the Hadash Formation in the Jebel Akhdar suggests that some correlations between facies within it are possible. These are somewhat conjectural, and a number of different scenarios could be envisaged. Attempts to use $\delta^{13}$C signatures to tie points together within the Hadash Formation to help with intra-basinal correlation proved unsuccessful. Although some portions of the isotopic curves matched closely between
profiles, other parts did not fit as easily. Diagenetic screening allowed some measure to be made of how reliable the δ¹³C values were likely to be. Little covariation occurred between any independent diagenetic variables, making 'backstripping' to offset any diagenetic effects, and an approximation of original δ¹³C values difficult. Some portions of the δ¹³C signatures could be corrected using covariation between δ¹³C and δ¹⁸O. These were bulk corrections, however, and tended to hide any subtleties in the curve. After these corrections were made close similarities between some profiles were apparent (eg. between Wadi Bani Awf and Wadi Hajir 1, and between Hadash 1 and Hadash 2). Significant differences between isotopic signatures that could not easily be explained by differing degrees of diagenesis occurred across the Jebel Akhdar as a whole, however. Ultimately, the variation in values between sections, and the fact that a number of different ways of matching excursions were possible, meant that the δ¹³C profiles could not be used meaningfully as a correlatory tool at this <15m resolution within the Hadash Formation. The difficulties found in using δ¹³C at this resolution here suggest that isotopic trends used to correlate sections on this near metre-scale scale elsewhere (often where the sampling density is lower) should be treated with caution.

Stable isotopes also failed to resolve the questions posed as to the validity of the lithological correlation between the carbonate unit in the Huqf area and the Hadash Formation in the Jebel Akhdar. Two possibilities still remain. Either the carbonate in the Huqf was deposited later than the Hadash Formation in the Jebel Akhdar when δ¹³C values had returned to more positive values, or the two carbonates are synchronous, but δ¹⁸C values in the shallower-water Huqf area were affected by local factors. The fact that the carbonate in the Huqf area is also a transgressive deposit, and that models of cap carbonate formation suggest that conditions promoting carbonate deposition should be widely developed during postglacial transgression, indicate that they may be quite closely linked in time.

The overall δ¹³C signature from the Hadash Formation in the Jebel Akhdar is very similar to those reported from other Neoproterozoic carbonates capping glacial events (eg. Kennedy, 1996; Hoffman et al., 1998b; Myrow and Kaufman, 1999). A number of models have been suggested to explain the presence of carbonate depleted in δ¹³C directly overlying these Neoproterozoic glaciations. There is some evidence within the Hadash
Formation that carbonate deposition was assisted by a microbial influence, and may have been linked to the degradation of organic-rich material. This supports the gas hydrate destabilisation model of Kennedy et al. (2001) to some extent, but does not necessarily rule out other models of cap carbonate deposition. Consideration of the underlying Ghadir Manqil Formation does suggest, however, that the proposed partitioning of the ocean and atmosphere by ice and complete shutdown of photosynthesis, as envisaged by Hoffman et al. (1998b), did not occur.

The similarity on a global scale of cap carbonates and their distinctive isotopic signatures has led to their use in Neoproterozoic sections as stratigraphic markers (eg. Kennedy et al., 1998), and to the suggestion that a Marinoan-type cap carbonate be used to define the base of a new ‘Terminal Proterozoic System’ (eg. Christie-Blick et al., 1995; Knoll, 2000). However, a thin carbonate overlying glacial deposits within the Ghadir Manqil Formation, well below the Hadash Formation, suggests that conditions promoting cap carbonate formation may not have been unique. There is also evidence within the Ghadir Manqil Formation for a number of distinct glacial events, separated by non-glacial facies. If this is the case, cap carbonates reported globally from Neoproterozoic successions were not necessarily deposited synchronously. Potentially, they could have formed after any one of the glacial events. Therefore, where constraints (such as radiometric dates or isotopic curves) above and below cap carbonates are poor, correlations between sections on a global scale based on their presence should be viewed sceptically. This suggests that placing the base of the ‘Terminal Proterozoic System’ within or at the base of a Marinoan-type cap carbonate may prove to be problematic.
Chapter 6

The Masirah Bay Formation
CHAPTER 6. THE MASIRAH BAY FORMATION

6.1 Introduction

The Masirah Bay Formation is well-exposed in both the Huqf area of east-central Oman and in the core of the Jebel Akhdar (Fig. 1.1). This section of the stratigraphy was first described in outcrop by Kassler (1965) in the Huqf area (where it was termed the 'Abu Mahara Formation') and by Kapp and Llewellyn (1965) in the Jebel Akhdar (where it was included as the upper part of their Mistal Conglomerate). These exposures have since been correlated with each other on the basis of lithology (Tschopp, 1967; Gorine et al., 1982) and from the stable isotopic signatures from the overlying deposits (Burns and Matter, 1993; Burns et al., 1994; McCarron, 2000). In this study the name Masirah Bay Formation is used to include the outcrops in both the Huqf and Jebel Akhdar areas that lie between the Hadash and Khufai Formations. This has been adapted from Bell (1993a) who used the term to describe the upper parts of the Abu Mahara Group above the Ghadir Manqil Formation in the subsurface; the type section being defined from well Masirah Bay-1. The term Masirah Bay Formation is widely used within Petroleum Development Oman, and has also been used in recent publications (eg. Brasier et al., 2000).

In the Huqf area (Fig. 6.1), the base of the Masirah Bay Formation is very poorly exposed. In the Al Jobah area, the carbonate that unconformably overlies the Halfayn Formation can be seen to pass up into siliciclastics. These siliciclastics possibly represent the base of the Masirah Bay Formation, but a thickness of only a few metres is exposed and the outcrops here are well separated (by more than 80km) from the better developed Masirah Bay exposures further to the south. A corehole (CH-17) was drilled by Petroleum Development Oman in the centre of the Khufai Dome directly beneath the lowermost exposures of the Masirah Bay Formation there (Fig. 6.2). Corehole-17 penetrated 250m of siliciclastics (sandstone and shale) before passing through a dolomitic unit and bottoming in volcanics. It seems likely that this is the same carbonate that lies above the Halfayn volcanics at Al Jobah, and is probably equivalent to the Hadash Formation (Chapter 5). This suggests that throughout most of the Huqf only the upper parts of the Masirah Bay Formation are exposed.
In the Jebel Akhdar (Fig. 6.3), the base of the Masirah Bay Formation is well-exposed and gradationally overlies the Hadash Formation. In this study the base of the Masirah Bay Formation is taken as lying directly above the last carbonate bed of the Hadash Formation.

In both the Huqf and Jebel Akhdar areas, the dominantly siliciclastic Masirah Bay Formation is gradationally overlain by carbonates of the Khufai Formation.

The aim of this chapter is to describe the sediments of the Masirah Bay Formation in both the Huqf and Jebel Akhdar areas, with a view to establishing facies and their regional variation. Summary logs from the Huqf area are shown in Fig. 6.4, and summary logs from the Jebel Akhdar are shown in Fig. 6.5. These allow an overview of the Masirah Bay Formation, and illustrate the differing environments that occurred in the Huqf and Jebel Akhdar areas during Masirah Bay time. Detailed lithological logs from the Huqf area are in Appendix E.

6.2 Methods

In the Huqf area well-exposed sections of the upper part of the Masirah Bay Formation occur in the centre of the Khufai Dome (Figs. 6.6a and 6.6b), Mukhaibah Dome, and at Al Harf. In the Jebel Akhdar, complete sections are exposed in Wadi Bani Awf (Fig. 6.7) and Wadi Hajir. Further good exposures occur in Wadi Mistal near the village of Hadash, in Wadi Sahtan, and in Wadi Bani Jabir. All of the above sections were logged in detail with the aim of characterising facies and establishing palaeoenvironments. Sedimentary structures were measured and recorded, and where possible, beds were traced laterally. Supplementary sections in the Buah Dome, in the north of the Huqf area, and at Wadi Bani Kharus in the Jebel Akhdar were also studied to provide further information on facies variations and palaeoenvironments. Representative samples of facies were collected from each locality for petrographic and microfacies analysis. Further samples were collected from the carbonate facies for preliminary stable isotopic analysis.

This combined information has enabled reconstruction of the facies relationships and facies evolution through time. Dominantly shallow marine deposition is indicated in the
Huqf area, whereas further to the north in the Jebel Akhdar, deep, quiet water deposition dominates.

6.3. Previous work

Kassler (1965) first described the sediments of the Masirah Bay Formation in the Huqf area, terming them the Abu Mahara Formation (Fig. 6.8), and defining the type section in the centre of the Khufai Dome. He suggested that these sediments were fluviatile in origin, with the finer clastics representing lagoonal or tidal-flat deposits. A fluviatile origin for the Masirah Bay sandstones was adopted by later workers (Gorin et al., 1982; Hughes Clarke, 1988; Dubreuilh et al., 1992; Platel et al., 1992b), who further reported channels and coarsening-up sequences indicating local progradation of fluviatile outwash deposits. Dubreuilh et al., (1992) kept the nomenclature of Kassler (1965), but also recorded the presence of basement and the volcanic and volcaniclastic Halfayn Formation in the Huqf area (Fig. 6.8). More detailed work by Pilcher and Buckley (1995) on the outcrops of the Huqf area identified five major lithofacies in the Masirah Bay Formation. Conversely to the previous workers, they suggested that a tide-dominated shelf depositional setting was more likely.

In the Jebel Akhdar, Kapp and Llewellyn (1965) first described the Masirah Bay Formation, incorporating it into the upper part of the Mistal Conglomerate. In this work it was briefly described but no interpretation was attempted. It was incorporated by Rabu et al. (1986), Beurrier et al. (1986), and Rabu (1988) into the Amq Member (now Hadash and Masirah Bay Formations) that occurred at the top of the Mistal Formation (Fig. 3.7). The fact that it had been intensely altered, and that thin-section examination at its base revealed mainly silica and iron oxide with varying quantities of carbonate was used to infer a very specific, possibly evaporitic or continental environment.

The Masirah Bay Formation has also been described from the subsurface. Initially, it was dealt with alongside the Ghadir Manqil Formation as the ‘Abu Mahara Formation/Group’ (eg. Blendinger and Teyssen, 1989; Kapellos et al., 1992). Bell (1993a), however, revised the stratigraphic nomenclature and split the Abu Mahara Group into the Ghadir Manqil and Masirah Bay Formations (Fig. 6.8). He further subdivided the Masirah Bay Formation into
a Basal Carbonate Member (Hadash Formation herein), a Clastic Member and an upper Transition Member (Fig. 6.8). Bell (1993a) interpreted the Masirah Bay Formation as forming in a shallow marine or fluvial environment.

In this study, the base of the Nafun Group is taken to occur at the boundary between the Ghadir Manqil and Hadash Formations (Chapter 5). The Masirah Bay Formation is therefore also now placed within the Nafun rather than the Abu Mahara Group (Fig. 6.8).

**6.4. Subdivision of the Masirah Bay Formation**

In the Huqf area, only the upper part of the Masirah Bay Formation is well-exposed. This contains a record of two major progradational coarsening-up sequences. Both of these culminate in coarse-grained trough cross-bedded sandstone; the uppermost of which passes up into deeper water deposits that underlie the Khufai Formation. These two coarsening-up sequences allow the upper part of the Masirah Bay Formation to be subdivided into three members (Table 6.1). The lower member (Member 1) passes from the wave-rippled shoreface deposits of Facies Association A into the coarse-grained trough cross-beds that dominate Facies Association B. The middle member (Member 2) includes offshore storm-influenced deposits (Facies Association C) that pass up into another coarse-grained trough cross-bed dominated unit (Facies Association D). The upper member (Member 3) comprises the finer-grained deposits below the Khufai Formation that are included within Facies Association E.

In the Jebel Akhdar the Masirah Bay Formation is represented by deeper-water deposition than in the Huqf area. The deposits are dominated by argillaceous siltstones and sandstones and the three members and five facies associations proposed for the Huqf cannot be recognised. The Masirah Bay Formation in the Jebel Akhdar will therefore be dealt with separately to that in the Huqf area.
6.5. Facies analysis of the Masirah Bay Formation in the Huqf area

6.5.1. Facies Association A: Wave-dominated shoreface deposits

Facies Association A outcrops in the centre of the Khufai Dome forming the lower part of Member 1 and the lowest part of the Masirah Bay Formation that is well-exposed in the Huqf area. A further 250m of fine-grained siliciclastics was recorded from a corehole (CH-17) drilled directly below the lowest outcrops of Facies Association A (Fig. 6.2). However, these lower parts of the Masirah Bay Formation are only very patchily exposed more than 80km to the north of the rest of the Masirah Bay outcrops and have not been dealt with as part of this study.

Facies Association A is distinctive in the field due to its generally fine-grained and thickly-laminated to very thinly-bedded nature. It forms the lower 60m of the exposed Masirah Bay Formation in the Khufai Dome; the lowermost outcrops of Facies A1 occurring directly above Corehole-17. It is dominated by the wave-rippled sandstones of Facies A2, within which much more minor beds of massive sandstone only a few cm thick occur (Facies A3). Facies Association A is very homogenous for much of its 60m thickness, but shows a gradual coarsening-up before passing into Facies Association B. A summary of the facies occurring within Facies Association A is given in Table 6.2.

6.5.1.1. Facies A1: Wave-rippled sandstone and siltstone

Facies A1 forms the lowermost few metres of Facies Association A, and gradationally passes up into Facies A2. Facies A1 is composed of grey, fine-grained, moderately sorted quartzitic sandstones that are interbedded with fissile recessive siltstones on a few mm-scale (Plate 6.1). The sandstones show relatively planar to slightly undulatory structures. Upwards these become increasingly undulatory and rippled. These ripples are usually only partially developed, and show no internal structure. Undulatory bases and generally symmetrical profiles can be observed, however. Rare scour structures can also be seen. Sorting increases upwards, and the siltstones coarsen into fine-grained sandstones that become less recessive as Facies A1 passes up into Facies A2.
This facies represents deposition above storm wave base, probably in the offshore-transition or lower shoreface. The undulatory bases and symmetrical profiles to the ripples suggest that they were formed by the action of waves (De Raaf et al., 1977). Their increasing prevalence towards the top of the facies suggests increasing wave reworking and a move into the lower shoreface.

6.5.1.2. Facies A2: Wave-rippled sandstone

Facies A2 forms the majority of Facies Association A. It gradationally overlies Facies A1, and is in turn overlain by Facies B1 and B2. Facies A2 is composed of grey-to-yellow coloured, very well-sorted, fine-grained quartzitic sandstone. This sandstone forms few mm thick beds showing ripple lenses with undulatory bases and locally developed symmetrical profiles (Plate 6.2). These are usually a few mm in height with a wavelength of a few cm (ripple indices of ~10). Cross-lamination is rarely preserved. Where it is, no consistent palaeocurrent direction is shown. Between successive ripple lenses, slightly finer-grained recessive sandstone occurs. This is reduced upwards as the number of rippled lenses increases. The rippled sandstone often fills very low-angle scours up to a metre across that erode the tops of underlying ripples. The rippled sandstone of Facies A2 remains remarkably similar through its entire thickness (nearly 50m). Over this vertical distance the only change it shows is a slight coarsening-up to medium-grained sandstone, and a slight increase in the number and scale of the ripple lenses.

Facies A2 represents wave-dominated deposition on the shoreface. The ripples again show features indicating formation by waves (undulatory bases, symmetrical profiles, variable palaeocurrents, cf. De Raaf et al., 1977; Allen, 1981). The extensively rippled nature of the sandstone and its relative homogeneity over a thickness of ~50m suggests that it was affected by fairweather waves and was deposited on the shoreface in depths of <25m (Reading and Collinson, 1996). The well-sorted and texturally mature nature of the sandstones indicates probable shoreface reworking of sands derived from the high-energy beachface and the transport of finer particles offshore. The low-angle scours probably represent erosive storm events that were later filled or reworked by waves during fairweather conditions. The fact that a thickness of 50m of deposits formed in <25m water depth is preserved suggests that subsidence kept pace with sedimentation during the
deposition of Facies A2, allowing facies belts to be stacked subvertically. However, a shallowing towards the top of the facies indicated by the increased grain size suggests some progradation.

6.5.1.3. Facies A3: Thinly-bedded planar-laminated sandstone

Facies A3 is a relatively minor facies within the upper part of Facies Association A, occurring within stratigraphy dominated by Facies A2. Facies A3 is composed of sharp-based beds, usually less than 10cm thick, that are composed of well-sorted, medium-grained sandstone. A planar lamination is commonly developed in the lower few cm and this passes up into cross-lamination in the upper part of the beds. These beds can be traced laterally for at least a few metres.

Facies A3 probably represents storm deposits, when slightly coarser-grained sand that was worked in the beach-zone would have been driven offshore by storm-induced currents (eg. Johnson, 1977; Brenchley et al., 1993). The lower planar-laminated part represents upper stage plane beds formed during the main part of the storm, when combined oscillatory and unidirectional flows dominated (Cheel, 1991; Duke et al., 1991). The upper parts of the beds were probably reworked during the waning part of the storm as current and wave power waned and indicate a return to lower flow regime oscillatory currents (Brenchley, 1989). These are overlain by the wave-rippled sandstones of Facies A2, indicating return to fairweather conditions after storm deposition.

The storm events recorded within Facies Association A are mostly purely erosional (forming the low-angle scours filled by the rippled sandstone of Facies A2). Only occasionally is a record of the sand moved by these events left within the shoreface deposits of Facies Association A.

6.5.1.4. Facies Association A: Synthesis

The most likely setting for the sediments of Facies Association A is on a wave-dominated shoreface punctuated by infrequent storm events. Throughout the Facies Association, a shallowing-up is recorded. Deposits of the offshore-transition to lower shoreface (Facies
A1) pass up into the extensively rippled shoreface sandstones of Facies A2. The shallow scours and sandstone beds of Facies A3 indicate the influence of storms at this time. The coarsening-up through Facies A2, probably indicates continued shallowing. This is supported by the overlying deposits of Facies Association B, which indicate the influence of tidal currents and deposition in the upper shoreface to foreshore (Fig. 6.9A).

The progradational signature of Facies Association A suggests early relative sea-level highstand conditions (Figs. 6.4; 6.10). As Facies Association A is only exposed in the centre of the Khufai Dome, further analysis of lateral variations across the Huqf was not possible.

6.5.2. Facies Association B: High-energy current dominated deposits

Facies Association B gradationally overlies Facies Association A, forming the upper part of Member 1. It is most completely exposed in the centre of the Khufai Dome, where it is nearly 120m thick. The upper part of the facies association is also exposed in the Buah and Mukhaibah Domes and at Al Harf in the south of the Huqf area. Facies Association B is sharply overlain by Facies Association C, but this boundary is only seen at Al Harf.

Facies Association B is composed of medium- to coarse-grained sandstone that preserves a number of different types of cross-stratification. The lower part of this facies association is only exposed in the centre of the Khufai Dome. Here it comprises cycles of hummocky/swaley cross-stratified sandstone (Facies B1), passing through sigmoidal cross-stratification (Facies B2) into planar cross-stratified sandstone (Facies B3). These alternating facies are then overlain by the coarse-grained trough cross-stratified sandstone of Facies B4 that dominates for the next c. 50m. Granular lags (Facies B6) occur within this facies, both as bands a few grains thick in the lower parts of trough cross-sets and in beds more than a metre thick dominated by granular material. Facies B4 and B6 are seen at a number of localities within the Huqf area (Khufai, Buah, and Mukhaibah Domes). At Al Harf at the same level, large-scale planar cross-stratified sandstone dominates (Facies B5). A summary of the facies present within Facies Association B is given in Table 6.2.
6.5.2.1. Facies B1: Hummocky/swaley cross-stratified sandstone

Facies B1 forms a relatively minor facies, occurring in units 1 to 2m thick near the base of Facies Association B. It passes up into sigmoidally and then planar cross-stratified sandstones (Facies B2 and B3), which it is then interbedded with, forming B1-B2-B3 cycles. Facies B1 comprises moderately well-sorted sandstone which exhibits dominantly swaley (Plate 6.3) and occasionally hummocky cross-stratification. These bedforms are typically <20cm in height, with wavelengths >0.5m. The basal set boundary is erosional and is draped by the internal laminations, which often pinch and swell on a dm-scale. Rarely, these laminations pass laterally into accretionary hummocks. In the main, however, swale structures and internal erosion surfaces dominate. The top parts of some of the cross-stratified beds display symmetrical ripple forms.

The detailed conditions under which hummocky cross-stratification forms have sparked a large amount of debate over the years. Hummocky cross-stratification has been thought by many authors (eg. Walker et al., 1983; Duke, 1985, 1987) to form beneath flows overwhelmingly dominated by powerful wave-orbital motions. Other workers, however (eg. Allen, 1985; Nottvedt and Kreisa, 1987; Swift and Nummedal, 1987), have emphasized the involvement of a strong steady current superimposed on wave motions. Whatever the precise mechanisms of its formation, however, hummocky cross-stratification has long been considered to be associated with storm deposition (eg. Johnson and Baldwin, 1996), and is now generally accepted to form from combined oscillatory and unidirectional flow (eg. Brenchley, 1989; Duke et al., 1991; Cheel and Leckie, 1993; Midtgaard, 1996; Myrow and Southard, 1996) typified by storm conditions. Within Facies B1, it seems likely that the hummocky cross-stratification formed on a storm-dominated shoreface. This is indicated by a number of features. Firstly, the hummocky cross-stratification in Facies B1 occurs alongside the dominant swaley cross-stratification, which is often considered to be characteristic of storm-dominated shoreface deposits (Johnson and Baldwin, 1996). Secondly, Facies B1 overlies the shoreface sediments of Facies A2 and is in turn overlain by the tide-influenced deposits of Facies B2, further suggesting shoreface deposition by association. Initially, during storm conditions, sand would have been transported from the shoreline or beach-zone onto the shoreface. This would then have been reworked on the shoreface by powerful oscillatory storm waves and
unidirectional currents associated with the storm (eg. wind-driven, storm-surge ebb, rip, or geostrophic currents). This would have produced the hummocky and swaley cross-stratification. Finally, as the storm waned and wave power was reduced, the symmetrical wave ripples at the tops of some of the beds would have formed.

6.5.2.2. Facies B2: Sigmoidally cross-stratified sandstone

Facies B2, composed of dominantly sigmoidally cross-stratified sandstone, overlies the storm-dominated shoreface deposits of Facies B1. It is associated with the planar cross-stratified sandstones of Facies B3 and commonly passes up into this facies. The tops of sigmoidal co-sets are commonly rippled, and indeed, the sigmoidal cross-stratification is often interspersed with wave-rippled sandstone. The sigmoidal cross-stratification of Facies B3 is composed dominantly of well-sorted, medium-grained sandstone. This forms bedforms up to 50cm in height with consistently uni-directional palaeocurrent directions in which top-sets are preserved (Plate 6.4). Foresets are up to a few cm thick and are separated by few mm thick drapes of darker-coloured, finer (muddy), recessive material. These drapes reveal the sigmoidal shape of the cross-sets. A slight change in foreset angle above mud-drapes is common and indicates repeatedly interrupted progradation followed by reactivation (De Raaf and Boersma, 1971). The drapes thicken towards the toe of the cross-set, where wave-ripples are occasionally preserved. The alternating fine and coarser material in the bedform defines bundles (Visser, 1980). These bundles thicken and thin periodically on a dm scale along the bedform and where they are thinner, the finer drapes tend to be more prevalent.

Deposits containing reactivation surfaces, mud drapes, and bundles that periodically thicken and thin laterally have been ascribed to tidal environments by many previous workers (eg. De Raaf and Boersma, 1971; Visser, 1980; Terwindt, 1981; Allen and Homwood, 1984; Yang and Nio, 1985; Reading and Collinson, 1996). During a tidal cycle, two opposing current directions (the ebb and flood) affect the sea-bed, separated by periods of slack water (high- and low-tide). One of the two opposing current directions (either the ebb or the flood) is commonly dominant (eg. Johnson, 1977; Levell, 1980a; Davis and Flemming, 1995), and indeed the separation of ebb and flood currents has been observed in many places (Visser, 1980). This would explain the dominantly uni-
directional palaeocurrents observed in this facies. The migration of the sigmoidal bedforms present in Facies B2 would have occurred during the dominant, probably flood current stage. During the subsequent slack water stage (either low- or high-tide) the few mm thick mud drape would have been deposited. The subsequent subordinate current would usually have been too weak to transport sand, but may have eroded the mud layer at the top of the lee side of the bedform formed during the dominant current stage. It is possible that the relatively rare ripples observed in some of the toe-sets could have been formed by agitation during the subordinate current stage. A period of slack-water would follow this, depositing more mud before the cycle was repeated, and migration of the bedform was reactivated during the dominant current stage. Such a cycle would be expected to produce mud-layer couplets reflecting the occurrence of two slack-water stages (Visser, 1980). Single mud-drapes (as in Facies B2) are also common in tidal deposits (eg. Allen and Homewood, 1984). The periodic thickening and thinning of the bundles produced by these tidal cycles probably reflects the effect of spring and neap tides. Tidal currents would be expected to intensify during springs, producing the thicker, sand-dominated bundles. During neaps, the amount of sand transported would be reduced, and an increasing amount of mud would be deposited. This would produce the thinner bundles in which the mud-drapes are more prevalent.

All these features taken together provide good evidence of a tidally-influenced depositional environment for Facies B2. Whether this was a subtidal or intertidal environment is less easy to say. The lack of evidence for emergence in Facies B2 suggests that subtidal deposition on the upper shoreface is possibly more likely. The close association of the sigmoidal cross-stratification with wave-rippled deposits suggests that wave activity was locally dominant over tidal currents during deposition of Facies B2.

6.5.2.3. Facies B3: Planar cross-stratified sandstone

Facies B3 is associated with Facies B1 and B2 below the trough cross-stratified sandstone of Facies B4. Cycles passing from Facies B1 to B3 occur, before the low-angle cross-stratified sandstone of Facies B3 dominates for ~10m below Facies B4. Facies B3 is composed of well-sorted, medium-grained sandstones. This has mainly been worked into low- to moderate angle (10-25°), planar to very slightly concave-up, cross-sets <20cm high
with erosional tops (Plate 6.5). The foreset surfaces are regularly spaced and very consistently show palaeocurrent directions towards the north-east. Individual cross-sets can be traced for 10m or more in some cases. Planar-laminated sandstone and sandstone containing amalgamated ripple-lenses (similar to Facies A2) are interbedded with this planar cross-stratified sandstone, becoming increasingly prevalent towards the top of Facies B3 below the trough cross-stratified sandstone of Facies B4. Throughout Facies B3, there is very little to no mud or finer-grained material present.

The well-sorted nature of Facies B3, the lack of any significant mud or silt, and the wave-rippled sandstones indicate continuing shoreface or shallower deposition from Facies B1 and B2. It is possible that Facies B3 represents shallower deposition than Facies B1 and B2 in the nearshore or top of shoreface. This is suggested by the low- to moderate-angle tabular cross-stratification interbedded with parallel-laminated and rippled sandstones. Similar structures have been reported from the nearshore and top of shoreface zones of modern wave-dominated shorelines (eg. Davidson-Arnott and Greenwood, 1974; Howard and Reineck, 1981), and have been interpreted from ancient deposits as forming within the shoreline zone (eg. Johnson and Baldwin, 1996). The planar cross-stratification may thus have formed through the migration of a top of shoreface bar (Fig. 6.9A). The associated planar laminations that become increasingly prevalent towards the top of Facies B3 represent deposition in an upper flow regime, possibly in the shallow flows that would be expected to wash across the foreshore as tides ebbed and flowed. If these interpretations are correct, Facies B3 represents shallower-water deposition than Facies B2. The B1 to B3 cycles are therefore shallowing-up. The prevalence of Facies B3 at the top of the cycles therefore indicates a continuation of the shallowing-up sequence observed through Facies Association A. However, the B1 to B3 cycles suggest that this was not simply a continuous shallowing and was at times punctuated by minor deepening events.

6.5.2.4. Facies B4: Trough cross-stratified sandstone

Facies B4 is composed of coarse-grained, red to light-grey coloured, trough cross-stratified sandstone. This dominates the stratigraphy for approximately 50m in the Khufai Dome (76-128m log KD2; Plate 6.6), overlying Facies B3 and passing up into the shale of Facies C1 (although this contact is not exposed). Facies B4 is also exposed in ridges in the
Mukhaibah and Buah Domes where its base and top are not exposed. At Al Harf in the south of the Huqf area, coarse-grained sandstone occurs at the same level, but forms the high-angle large-scale planar cross-beds of Facies B5. It is therefore apparent that this coarse-grained sandstone body (comprising Facies B4 and B5) has a lateral extent of at least 80km in a roughly north-south direction.

In the Khufai Dome, the trough cross-beds initially occur interbedded with planar cross-stratification, similar to that described from Facies B3. However, these quickly pass up into the stacked purely trough cross-stratified sandstone that dominates Facies B4. The coarse-grained sandstone is composed almost entirely (95%) of sub-rounded to rounded quartz grains with a locally developed calcite cement. Other grains include rare felsic volcanics and shales, as well as well-rounded, often chloritised, mica fragments. The sandstone is only moderately to poorly sorted, with grains varying from 0.1 to 3mm in size; most falling in the range 0.4-0.6mm. Granular material (Facies B6) also occurs associated with the sandstone. This is commonly concentrated into the lower parts of cross-sets, or occurs in bands up to a few metres thick within the trough cross-stratified sandstone.

As mentioned above, the cross-stratification present in Facies B4 is dominated by three-dimensional trough-shaped bedforms. In the lower 15m of Facies B4, these cross-sets are up to 50cm in height with erosive bases that can locally be seen to cut down 30cm (Plate 6.7). These pass up into smaller bedforms in which the trough cross-sets are reduced to <25cm height, before larger-scale bedforms are again established in the uppermost 20m of Facies B4. Despite this change in scale the overall shape of the cross-stratification remains similar throughout Facies B4. Similarly, fining-up within individual cross-sets (on less than a few cm scale) can also be seen throughout the trough cross-stratified facies (Plate 6.8). Palaeocurrent directions within Facies B4 were hard to establish because of the three-dimensional nature of the structures. However, general trends of trough axes and cross-stratification were orientated towards the west.

Trough cross-stratification is produced by the migration of three-dimensional bedforms (eg. Allen, 1968). This can occur in a number of different environments, and indeed, the trough cross-stratified sandstones of the Masirah Bay Formation have been previously
ascribed to fluvial deposition by some authors (Kassler, 1965; Gorin et al., 1982), and to shallow marine by others (eg. Pilcher and Buckley, 1995). It has previously been noted (eg. Levell, 1980; Collinson, 1996) that in Precambrian sandstone successions, where fossils are lacking and soils appear much less commonly, it is often difficult to distinguish fluvial, shallow marine, and even aeolian elements. However, the lack of bioturbation leads to the excellent preservation of primary sedimentary structures (as described above). Together with lateral facies variations, these allow a depositional setting to be postulated.

The coarse grain-size, poor sorting, and occurrence at the top of a shallowing-up sequence above the top of shoreface/shoreline deposits of Facies B3 initially suggests that a fluvial setting would be most likely for the trough cross-stratified sandstones of Facies B4. Fluvial and braided river systems could easily produce deposits as laterally widespread as Facies B4 (eg. Campbell, 1976; Ramos et al., 1986; Singh and Bhardwaj, 1991), and in the macroplant-free world of the Neoproterozoic, braided fluvial processes in particular, would have been much more common than today (Dott et al., 1986). Although relatively constant discharge in a braided river could explain the homogeneity of the sandstone body (Bristow, 1993), a number of features suggest that a shallow marine origin may be more likely: 1) There is no evidence for subaerial emergence; 2) no channels were identifiable in the field, although exposure (especially in the centre of the Khufai Dome) is good; 3) there is a lack of any fine (mud and silt grade) material; and 4) although sorting is commonly poor, the grains are rounded, and are mineralogically mature (95% quartz). This suggests significant reworking of the grains, or possibly derivation from an aeolian or beach source, before incorporation into Facies B4.

Such a laterally extensive, dominantly trough cross-stratified sandstone is probably more likely to have formed in an estuarine setting through the migration of subtidal sand shoals/bars (eg. Harris, 1988; Dalrymple et al., 1990; 1992; Reading and Collinson, 1996). In a tide-dominated estuary (where tidal-current energy exceeds wave energy at the estuary mouth) elongate sand bars are typically developed seaward of the tidal-energy maximum, where tidal energy is funnelled (Dalrymple et al., 1992). Sand bars are typically composed of medium to coarse-grained cross-stratified sandstone. They have been reported from the marine-dominated portion of a number of estuaries including Moreton Bay, Australia; Bristol Channel, Wales; Thames Estuary, England (Harris, 1988); and Cobequid Bay —
Salmon River Estuary, Bay of Fundy (Dalrymple et al., 1990). The trough cross-stratified sandstone of Facies B4 in the Huqf area may represent the deposits of more than one estuary. Alternatively, as estuaries with mouth widths of >100km have been reported (e.g. Gulf of Cambay, India (Off, 1963)), Facies B4, which stretches from the Buah Dome to Al Harf (a distance of more than 80km), could also have been deposited in one large estuary.

Deposition in a marine-dominated estuarine environment would allow for reworking of the grains in a fluvial or shallow marine environment prior to deposition, helping to explain the rounded and mineralogically mature nature of the sandstone. An estuarine interpretation would suggest a continuation of the shallowing-up trend observed below the base of Facies B4, probably indicating the development of late highstand conditions.

The dominantly uni-directional cross-stratification preserved probably reflects the dominance of one tidal direction over another (e.g. Levell, 1980a). Uni-directional cross-stratification could also be the product of preferential bedform migration during relatively short-lived, high-energy conditions (e.g. Johnson, 1977), possibly when the dominant tidal current was enhanced by storms (Amos et al., 1995). The lack of evidence for bi-directional palaeocurrents is therefore not a hindrance to adopting an estuarine tidal sand bar interpretation for Facies B4.

In the face of the arguments above, an estuarine tidal sand bar interpretation seems more likely than a fluvial one for the trough cross-stratified sandstone of Facies B4. This would indicate a continuation of the progradation observed below Facies B4 from Facies A1 to B3.

**6.5.2.5. Facies B5: Large-scale planar cross-stratified sandstone**

Large-scale planar cross-stratified sandstones of Facies B5 are only exposed at Al Harf in the south of the Huqf area. The base of Facies B5 is not exposed, but its top surface is sharply overlain by the shale of Facies C1, suggesting it may be a lateral equivalent of Facies B4. Facies B5 is at least 15m thick. It is composed of coarse-grained to granular sandstone, which forms planar cross-sets up to 2m in height (Plate 6.9). Grains are subrounded and are dominantly composed of quartz (70%), with minor feldspar, shale, and
chlorite. The sandstone is moderately sorted, with grains (mostly <2mm) occurring up to 5mm in size. Cross-set height varies from 0.5m to 2m and is commonly of moderate to high-angle (~25°). The cross-stratification fills erosive channels ~10m wide. Palaeocurrents recorded from Facies B5 consistently trend towards the east/north-east.

The planar cross-stratified sandstone of Facies B5 possibly represents the migration of tidal sand bars in a marine-dominated estuarine environment, in a similar manner to that described for Facies B4. If this is the case, the migrating bedforms were two-dimensional rather than three, and possibly record the dominance of a different tidal current to Facies B4.

The other possibility is that these represent fluvial sandstones. This is suggested by the coarse-grained, poorly-sorted nature of the sandstones, the higher mineralogical immaturity compared to Facies B4, and also by the presence of channels. However, the facies lying above Facies B5 suggest that Al Harf was in a more distal position than the other Huqf localities during Masirah Bay time. This makes it seem unlikely that Al Harf would experience fluvial conditions when the Khufai and Mukhaibah Domes were in a marine-dominated estuarine setting.

The exposure at Al Harf is limited, and a more detailed interpretation has not been attempted here.

6.5.2.6. Facies B6: Granular lag deposits

The deposits of Facies B6 are coarse-grained sandstones rich in granular material that occur closely associated with the light-grey coloured sandstones of Facies B4 and B5. The granules that occur in Facies B6 are usually well-rounded and are dominated by milky white quartz grains. Granular material occurs at the base of some of the trough cross-sets in Facies B4, fining-up into coarse-grained sandstone on a cm to few cm scale. This is developed locally throughout Facies B4, but may become more prevalent for a few metres at certain levels (eg. 98-101m log KD2). Granular material of Facies B6 forms a bed two metres thick towards the top of Facies B4 in the Khufai Dome (Plate 6.10). In this bed, almost all the material is granular, and no fining-up into sandstone occurs. Trough cross-
stratification is developed locally within this bed on a similar scale to that described from Facies B4. Similar beds dominated by granular or very coarse sandstone material are observed towards the top of the exposed outcrop of the trough cross-stratified Facies B4 in the Mukhaibah and Buah Domes, and at the top of the planar cross-stratified Facies B5 at Al Harf.

The close association of Facies B6 with Facies B4 suggests that it was also deposited in a marine-dominated estuarine setting. The granular material concentrated at the base of the trough cross-sets of Facies B4 was also transported during dune migration. It may have undergone size-sorting in the scour pits of the migrating megaripples prior to its deposition (Levell, 1980b). It is also possible however, that the granular material was concentrated by winnowing currents (Channon and Hamilton, 1976) that removed the finer material prior to transportation in the megaripples of the tidal sand bar. If this is the case, the winnowing flows must have been less powerful than the depositional currents. Facies B4 only preserves the predominance of a single palaeocurrent direction. This may be the product of preferential bedform migration during relatively short-lived, high-energy conditions, possibly when tidal currents were enhanced by storms (eg. Johnson, 1977; Amos et al., 1995). The winnowing may therefore have been achieved during subordinate tidal currents or lower-energy conditions that were unable to cause bedform migration.

An alternative explanation for many of the granule-rich levels is that they formed through winnowing associated with transgression. This is the most likely explanation for the 2m thick granular bed of Facies B6 that occurs less than two metres below the top of the trough cross-stratified sandstone of Facies B4. The fact that this granular bed occurs so close to the contact with the overlying deeper-water shales of Facies C1, and the fact that it can be found throughout the Huqf, suggests that it is linked to a significant flooding event. Winnowing associated with this major transgression could have removed the finer sandstone and left the concentrated granular material of Facies B6. This 2m thick granular bed therefore probably represents a marine transgression ravinement.
6.5.2.7. Facies Association B: Synthesis

The lowermost deposits of Facies Association B (Facies B1, B2, B3) are only exposed in the centre of the Khufai Dome and represent dominantly shoreface deposition (Fig. 6.9A). The repeated B1 to B3 (storm-dominated to top of shoreface) minor shallowing-up cycles that occur in the lower part of Facies Association B indicate fluctuations in relative sea-level in an overall continuation of the shallowing-up observed through Facies Association A. Above these B1 to B3 cycles, deposition in an estuarine environment is recorded by Facies B4 (Fig. 6.9). The trough cross-stratification present in this laterally extensive facies was probably formed through the migration of three-dimensional bedforms in tidal sand bars in the marine-dominated outer portion of the estuary. Facies B4 therefore represents a continuation of the progradation recorded through the lower portion of Facies Association B, and reflects the development of estuaries and late highstand conditions across the Huqf area (Figs. 6.4; 6.10). The planar cross-stratified sandstone of Facies B5 that occurs at Al Harf at the same level as Facies B4 was probably deposited in a similar environment to Facies B4, but formed through the migration of two-dimensional rather than three-dimensional bedforms. Above Facies B4 and B5, a flooding event is recorded before the deeper-water deposition of Facies Association C is established. It seems likely that winnowing associated with this transgression was responsible for producing the granular bed of Facies B6 that occurs at the top of Facies Association B.

Facies Association B records dominantly shoreface to estuarine deposition and an overall shallowing-up. Two minor progradational sequences are recorded by the two B1 to B3 cycles, and this progradation is continued in Facies B4, when late highstand conditions were established (Fig. 6.10). As in Facies Association A, a significant thickness (~70m) of shallow water sediments are preserved and subsidence must therefore have been keeping reasonable pace with deposition. The flooding at the top of Facies Association B is recorded across the Huqf area. It must have been a significant event and is taken here to mark the top of Member 1.
6.5.3. Facies Association C: Offshore storm-influenced deposits

Facies Association C sharply overlies Facies Association B at the one locality where this boundary is exposed (Al Harf), forming the lower part of Member 2. There is a marked change in grain-size across this boundary from coarse-to-granular sandstone into shale, reflecting significantly deeper-water deposition compared to Facies Association B. Facies Association C is exposed in the Khufai Dome (where it is over 50m thick) and at Al Harf (where it is 32m in thickness). In both of these localities it gradually becomes increasingly sand-dominated upwards and passes gradationally into the wave-rippled or cross-stratified sandstones of Facies Association D.

The base of Facies Association C is only exposed at Al Harf, where the dark grey to red shales of Facies C1 can be seen. Passing up through this facies, the shales are punctuated occasionally by the thinly-bedded to lenticular sandstones of Facies C2. The sandstones of Facies C2 are more dominant in the Khufai Dome, where the interbedded shale quickly coarsens up into siltstone, before the wave-rippled sandstones at the base of Facies Association D are established. This suggests that the Khufai Dome was in a more proximal position compared to Al Harf.

6.5.3.1. Facies C1: Dark grey to red shales

The shales of Facies C1 are exposed at Al Harf, where they directly overlie the granular sandstone of Facies Association B. Here, Facies C1 dominates the stratigraphy for ~30m, before the thin sandstone beds of Facies C2 become more prevalent. Facies C1 is composed of fissile, mm-scale planar laminated shale (Plate 6.11). This has a red colouration for a few metres above the contact with Facies Association B (possibly caused by oxidising fluids moving through the permeable Facies B6). For the most part Facies C1 is grey to dark grey in colour.

The homogeneity and lack of any coarse-grained material within Facies C1 suggest that it was deposited in a quiet, probably deep (offshore) setting. The planar lamination suggests that it formed through the settling of very fine-grained particles from suspension. The
fissile and easily eroded nature of Facies C1 perhaps explains why it is not exposed elsewhere within the Huqf area.

6.5.3.2. Facies C2: Interbedded shales/siltstones and sandstones

Facies C2 is locally developed within the shales of Facies C1 at Al Harf (28m log AH1), as well as gradationally overlying them. It is also exposed in low-lying gullies in the Khufai Dome where the base of Facies Association C is not exposed. At both Al Harf and in the Khufai Dome, Facies C2 becomes increasingly sand-rich upwards before passing into the sandstones of Facies Association D. Where Facies C2 occurs within the shales at Al Harf it is composed of several laterally continuous (for at least 10m), planar, fine-grained sandstone beds less than a few cm thick, each separated by a few cm of shale. In the Khufai Dome, and at the top of Facies C1 at Al Harf, Facies C2 is composed of more laterally variable/lenticular fine to medium-grained sandstone beds interbedded with grey shales (Plate 6.12). The sandstones are <few cm thick, have slightly undulatory bases and pinch and swell laterally on a dm-scale (partially due to layer parallel extension). The sandstones are commonly red in colour, locally show fining-up, and contain mainly planar laminations, which in places can be seen to pass up into ripple cross-lamination towards the top of the bed. The shales interbedded with these sandstones commonly contain thinner (few mm thick) sandstone streaks. Moving upwards through Facies C2, the sandstone beds gradually become thicker and coarser-grained, and the thin interbedded shales coarsen-up into siltstone. Ripple cross-lamination also becomes more common within the sandstone beds. Eventually, the sandstones become amalgamated at the expense of the siltstone, and wave-ripples and planar cross-stratification become widely developed. This change marks the boundary between Facies Association C and Facies Association D.

The association of Facies C2 with Facies C1 and the continuing presence of shale and siltstone suggests that Facies C2 was also deposited in an offshore environment. The presence of the interbedded sandstones, however, indicates that deposition was not purely from suspension settling. The laterally extensive sandstones that occur at Al Harf (28m log AH1) were probably deposited by relatively low-energy turbidity currents transporting sand offshore and interrupting the background sedimentation. However, the interbedded sandstones and shales/siltstones which occur in the Khufai Dome and which overlie Facies

159
C1 at Al Harf, were probably formed by other processes. The lenticularity of these sandstones, and the planar-laminated lower parts of the beds that locally pass up into ripple cross-lamination, suggest that these sandstones instead represent the deposits of storms. Similarly to the sandstones of Facies A3, the sand may originally have driven offshore by storm-induced currents (eg. Johnson, 1977; Brenchley et al., 1993). Alternatively, plumes associated with high river discharges could have delivered the sand into the offshore environment (eg. Stow et al., 1996). The sand would then have been worked in an upper flow regime during the main part of the storm, when combined oscillatory and unidirectional flows dominated (Cheel, 1991; Duke et al., 1991). This would have produced the planar-laminated lower parts of the beds. As the storm waned, the current ripple cross-lamination at the top of some beds would have formed. These indicate a return to lower flow regime conditions (Brenchley, 1989). The interbedded shales/siltstones represent fairweather deposition dominated by suspension fall-out, with the sandstone streaks possibly representing more minor events. The lack of rippling within the siltstone indicates deposition below fairweather wave-base.

Moving up through Facies C2, the coarsening and increasing dominance of the storm-deposited sandstones indicates shallowing as the finer-grained fairweather deposits were probably more frequently eroded by storms and transported further offshore. Eventually the amalgamation of the sandstone beds and increasing prevalence of wave-ripples and cross-stratification suggests a move from a storm-dominated offshore shelf environment onto a storm-dominated site on the lower shoreface (Brenchley, 1989; Johnson and Baldwin, 1996).

6.5.3.3. Facies Association C: Synthesis

The shales of Facies C1 at the base of Facies Association C represent relatively quiet, deep-water offshore deposition. Their presence indicates that a significant flooding event occurred at the top of Facies Association B between Members 1 and 2 (Fig. 6.10). The shales contain rare distal turbidite beds (perhaps triggered by storms), but are otherwise dominated by deposition from suspension settling. The interbedded sandstones and shales/siltstones of Facies C2 that overlie Facies C1 indicate an increasing influence from storms, and a move into higher-energy and possibly shallower-water conditions moving up
through Facies Association C. The increase in storm sandstone beds and associated reduction in the amount of shale and siltstone present towards the top of Facies C suggests that overall there is a shallowing-up signature through Facies Association C. Indeed, by the time the boundary between Facies Association C and Facies Association D is reached, deposition has moved from a quiet offshore environment to a storm-dominated lower shoreface.

The move from offshore to shoreface deposition recorded by Facies Association C indicates progradation becoming re-established above the significant flooding event that occurs at the boundary between Facies Associations B and C. The flooding is accomplished almost entirely at this boundary, and there is no transgressive systems tract developed (Fig. 6.10).

6.5.4. Facies Association D: High-energy current-dominated deposits 2

Facies Association D gradationally overlies Facies Association C, forming the upper part of Member 2. Facies Association D is most completely exposed in the Khufai Dome where it is 40m thick. The facies association is also well-exposed in the Mukhaibah Dome, and at Al Harf in the south of the Huqf area. It is overlain by the thinly-bedded, microbially-wrinkled sandstones that occur at the base of Facies Association E.

Facies Association D is composed of dominantly medium- to coarse-grained sandstone that preserves a number of different structures. The sequence these structures are developed in, and the relative dominance of each one varies between different localities. In the Khufai Dome, wave-rippled and planar cross-stratified sandstones of Facies D1 and D2 pass up through a number of granular-rich beds (Facies D3) into trough cross-stratified sandstone (Facies D4). The trough cross-stratified sandstone is interbedded with planar-bedded and wave-rippled sandstones (Facies D6 and D1 respectively), before the tidally-bundled sandstones of Facies D7 occur at the top of the Facies Association. In the Mukhaibah Dome, the wave-rippled and planar cross-stratified sandstones of Facies D1 and D2 pass up into trough cross-stratification (Facies D4) interbedded with bi-directional cross-bedding (Facies D5). Finally, at Al Harf, only the wave-rippled and planar cross-stratified
sandstones of Facies D1 and D2 are developed. A summary of the facies present within Facies Association D is given in Table 6.2.

6.5.4.1. Facies D1: Thinly-bedded undulatory to wave-rippled sandstone

Thinly-bedded sandstone of Facies D1 is present in all of the logged sections. It dominates the base of Facies Association D in the Khufai and Mukhaibah Domes, but also occurs higher up in the section interbedded with Facies D4, D5 and D6. At Al Harf, thinly-bedded undulatory to wave-rippled sandstone dominates for the entirety of Facies Association D.

In the Khufai Dome and at Al Harf, Facies D1 gradationally overlies the lower-shoreface storm deposits of Facies C2. Both the planar cross-stratified sandstone of Facies D2, and the granular beds of Facies D3 form interbeds within sections of the stratigraphy dominated by Facies D1. Facies D1 is overlain by Facies Association E at Al Harf, and by the trough cross-stratified sandstone of Facies D4 in the Khufai and Mukhaibah Domes (within which it forms minor interbeds). Facies D1 is composed of moderately well-sorted, fine- to medium-grained sandstone, with rounded grains that are mainly composed of quartz (70-80%), with more minor weathered feldspars and rare micas. The sandstone is thinly-bedded and although some beds are planar, undulatory and ripple structures are widely developed. At Al Harf, cross-stratification is also locally developed (Plate 6.13). The ripples that occur tend to have undulatory bases and internal lamination is only locally developed. Where it is, variable palaeocurrent directions are indicated. Few mm-thick mudstone and siltstone layers occur between some of the sandstone beds. These are more prominent at some localities (KD1, KD3) than at others (KD2). Where the mud/siltstone is developed in the Khufai Dome, bedding tends to be more planar and spindle-shaped structures are developed on the bedding surfaces of the sandstones (Plates 6.14, 6.15). These are composed of sandstone and occur as isolated structures that only rarely intersect. They are few cm in length, <1cm wide, and have a relief of <0.5cm that can be both positive and negative on the bedding surface. In cross-section they may be V- or U-shaped. Rarer, longer examples up to 15cm in length are also locally developed. Although an alignment of these spindle structures can occasionally be seen, they are most commonly randomly orientated.
An accurate interpretation of Facies D1 in the Khufai Dome has to take into account the sediments lying above and below, and also explain the presence of the spindle-shaped structures described above. Its position between storm-dominated lower shoreface deposits (Facies C2) and tidally-influenced sandstones (Facies D4 and D5 – see below) suggests deposition on the shoreface or shallower. The presence of ripples with undulatory bases and variable palaeocurrents suggests wave action, and further indicates deposition above fairweather wave-base (eg. De Raaf et al., 1977; Allen, 1981). The spindle-shaped structures that occur on some bedding surfaces in the Khufai Dome bear a strong resemblance to subaqueous shrinkage cracks. These have been shown to form in response to changes in salinity (eg. White, 1961; Burst, 1965; Donovan and Foster, 1972; Plummer and Gostin, 1981). Such structures are distinguished from polygonal desiccation cracks on the basis of their discontinuous, spindle or sinuous-shaped nature. A subaqueous origin for similar features has, however, been the focus of some debate, with the assertion that some examples could be subaerial (Astin and Rogers, 1991; 1992; 1993) being refuted by other authors (Trewin, 1992; Barclay et al., 1993). If the spindle-shaped structures of Facies D1 are subaqueous, the salinity changes driving the shrinkage could have been caused by intermittent freshwater input. Taken together with the small-scale undulatory and wave ripple cross-lamination, this suggests either shoreface deposition close to the freshwater discharge of a river, or wave-influenced pro-delta slope/delta front deposition.

At Al Harf, however, where no shrinkage cracks occur and the overlying tidal deposits are not present, the interpretation of Facies D1 is slightly different. The wave-ripples indicate deposition above fairweather wave-base, but there is no evidence for a tidal influence in the overlying deposits as there is further to the north in the Khufai and Mukhaibah Domes. It therefore seems likely that Al Harf experienced lower-energy, deeper-water conditions compared to the Khufai and Mukhaibah Domes at this time and represents more distal deposition. The lack of shrinkage cracks is consistent with this interpretation, suggesting deposition further from freshwater river discharges.
6.5.4.2. Facies D2: Planar cross-stratified sandstone

Facies D2 occurs interbedded with the sandstones of Facies D1 at Al Harf and in the Khufai Dome, as well as occurring interbedded with the trough and bi-directionally cross-stratified sandstones of Facies D4 and D5. Where Facies D2 is interbedded with the undulatory and wave-rippled sandstones of Facies D1 the planar cross-stratified sandstones are medium-grained and moderately sorted, with dominantly sub-rounded grains composed of quartz (60-70%) and corroded feldspars. Cross-sets are typically less than 10cm in height and commonly lens out laterally over a few metres. Limited exposure restricts palaeocurrent measurement, but overall trends suggest variable directions. Where Facies D2 occurs interbedded with Facies D4 and D5, it tends to be of similar grain-size and composition to these sandstones (ie. medium to coarse-grained, 70-80% quartz with sub-rounded grains – see below) and forms more laterally continuous cross-sets up to 30cm in height.

Where the planar cross-stratified sandstones of Facies D2 occur as smaller, laterally discontinuous structures within Facies D1, they probably formed in a similar shoreface or wave-influenced delta front environment. The fact that these cross-stratified sandstones only occur as relatively minor interbeds suggests that the currents forming them were only intermittently developed. They may have formed through the transportation of sand offshore by currents linked to storm conditions or increased fluvial discharge. Alternatively, the cross-stratified sandstones may represent sand transported by tidal currents, possibly during spring cycles or storms, when the tidal currents would have been enhanced.

Where Facies D2 occurs interbedded with Facies D4 and D5, it probably formed in a similar marine-dominated estuarine environment through the migration of tidal sand bars. If this is the case, the planar cross-stratification simply represents the migration of two-dimensional bedforms under tidal flows rather than the more common three-dimensional forms that would have been responsible for forming the dominant trough cross-stratification of Facies D4.
6.5.4.3. Facies D3: Granular deposits

Granular deposits of Facies D3 occur towards the top of Facies D1 in the Khufai Dome below the trough cross-stratified sandstones of Facies D4 (e.g. between 192 and 200m log KD2). They are not seen at this level in Al Harf, nor in the Mukhaibah Dome (although this is probably due to a lack of exposure). Granular beds also occur within Facies D4, commonly occurring at the top of trough cross-stratified ridges, below finer-grained more recessive beds. Towards the top of Facies D1, the granular beds are sharp-based, up to 20cm thick and are laterally continuous for tens of metres (the extent of the outcrop). They are poorly sorted with sub-angular to rounded grains up to 1cm in size (most commonly <0.5cm). Grains are composed dominantly of quartz, with feldspars and lithic clasts including dark shales also occurring. Low-angle cross-stratification with few cm high cross-sets is locally developed. Where the granular beds occur within Facies D4, they are up to 10cm thick and commonly occur at the top of trough cross-stratified ridges. These ridges show coarsening-up within them and are overlain by slightly finer-grained, recessive beds that often contain smaller-scale sedimentary structures compared to the ridges. Granular material is less commonly also present, concentrated at the base of trough cross-sets within Facies D4. The grains that occur are sub-rounded to rounded, up to 0.5cm in size and have a comparable mineralogical composition to the trough cross-stratified sandstones (80% quartz, weathered feldspars and microcrystalline lithics).

The granular beds of Facies D3 that occur associated with Facies D1 possibly represent storm beds, when currents were strong enough to move coarser-grained material offshore. This is suggested by the immature nature of the deposits and the sharp lower bedding surfaces. The granular beds that occur within Facies D4, however, were probably formed by winnowing processes associated with transgression. This is indicated by their common occurrence at the top of minor coarsening-up sequences and by the rounded nature of their grains, suggesting significant amounts of reworking prior to deposition. If this is the case, it suggests a number of relatively minor flooding events during the deposition of Facies D4. It is also possible that such winnowing associated with transgression was also responsible for the granular beds associated with Facies D1. The occurrence within these granular beds of more angular grains, however, suggests this may be less likely.
The granular beds that occur at the base of individual trough cross-sets probably simply represent the incorporation of granular material as the coarse-grained bases of channels or scours.

6.5.4.4. Facies D4: Trough cross-stratified sandstone

Trough cross-stratified sandstones of Facies D4 dominate the stratigraphy for ~30m throughout the Khufai Dome (eg. 2-31m log KD1; Plate 6.16), as well as occurring associated with the bi-directional and planar cross-stratified sandstones of Facies D2 and D5 in the Mukhaibah Dome. Facies D4 overlies the top of Facies D1 in the Khufai Dome, and is in turn overlain by the finer-grained tidally-bundled deposits of Facies D6. Facies D6 is absent in the Mukhaibah Dome, however, and the trough cross-stratified sandstone passes up into the base of Facies Association E. Although there are similarities between Facies D4 and the trough cross-stratified sandstone of Facies B4, there are also some significant differences. In the west of the Khufai Dome (log KD1), Facies D4 comprises a series of metre to few metre-scale coarsening-up cycles; the top of each commonly being overlain by granular beds of Facies D3. Medium-grained sandstone occurs at the base of each cycle (Facies D4a). This is recessive and contains trough cross-sets <30cm high associated with the undulatory and wave-rippled sandstone of Facies D1 (Plate 6.17). Facies D4a passes up into medium to coarse-grained trough cross-stratified sandstone (Facies D4b), which often becomes increasingly granule-rich upwards. This is locally associated with the planar cross-stratified sandstone of Facies D2. These coarser-grained tops to the sequences contain erosively-based trough cross-sets up to 60cm in height and 2m or more in width (Plate 6.18). Grading can be observed in some individual trough cross-sets. Palaeocurrent directions taken from both the trough cross-beds and the planar cross-stratification at this level consistently trend towards the west (between north-west and south-west). In the west of the Khufai Dome at KD1, at least eleven coarsening-up cycles can be seen. These cycles initially thicken upwards (reaching a maximum thickness of just over 4m), and then thin again towards the top of Facies D4 (Plate 6.16). Moving further to the north-east within the Khufai Dome, the coarser-grained upper parts containing the larger-scale cross-stratification (Facies D4b) become increasingly dominant, and well-developed cycles passing from Facies D4a into D4b are rarer. At KD3, for example, a ridge almost 10m high dominated entirely by the trough cross-stratified
sandstone of Facies D4b occurs and only six or seven coarsening-up cycles can be seen. In the Mukhaibah Dome, further to the south, only the smaller bedforms of Facies D4a are developed, and the sandstone body at this level is dominated by bi-directional and planar cross-stratification (Facies D5 and D2 respectively), and by wave-rippling (Facies D1). Trough cross-bedding at this level is not developed at all at Al Harf even further to the south.

The sandstone that Facies D4 is composed of is commonly yellow in colour, containing moderately sorted, dominantly sub-rounded grains. 70-85% of the grains are quartz, with up to 20% composed of feldspars that are commonly weathered. Minor fine-grained (shale) lithics occur and glauconite is also locally developed. The grains are mainly 0.2 to 0.3mm in size, with rarer clasts up to 2mm and even larger. A calcite cement is locally developed.

As was discussed for Facies B4, trough cross-stratification can be formed in a number of different environments. In Precambrian successions lacking fossils, distinguishing between these can often be difficult (eg. Levell, 1980a; Collinson, 1996). However, a number of features present in Facies D4 allow a reasonable interpretation to be made. Its occurrence between shoreface/delta front (Facies D1) and intertidal deposits (Facies D6) immediately suggests shallow water deposition. This is supported by its coarse-grained nature and by its lateral association with the bi-directional cross-bedding of Facies D5. The moderate sorting, mineralogical maturity, lack of evidence for subaerial emergence, and most importantly (Deer et al., 1992), the presence of glauconite indicate that this is a marine rather than a fluvial deposit.

Similarly to Facies B4, such a dominantly trough cross-stratified sheet sandstone could have formed in the marine dominated part of an estuary through the migration of subtidal sand bars in a high-energy environment (eg. Harris, 1988; Dalrymple et al., 1990; 1992; Reading and Collinson, 1996). This suggests a continued shallowing from the underlying, probable shoreface deposits of Facies D1 in the Khufai Dome. The fact that trough cross-stratification is much less dominant at this level in the Mukhaibah Dome can perhaps be explained by the presence there of the bi-directional cross-stratification of Facies D5, possibly indicating the presence of a tidal channel (see below). In the Khufai Dome, uni-
directional palaeocurrents again dominate, suggesting one part of the tidal cycle was dominant, or that bedform migration could only occur when the dominant tidal current in the estuary mouth was enhanced, possibly during storm conditions (Amos et al., 1995). The fact that at the same level further to the south-west at Al Harf, more distal, lower shoreface deposits occur (Facies D1), suggests that the south-westerly palaeocurrents represent a dominant ebb tide.

The coarsening-up sequences that can be seen were not present in the lower trough cross-stratified sandstone of Facies B4. The probable explanation for their occurrence in Facies D4 is that they represent progradational cycles or 'genetic units' caused by high-frequency changes in relative sea-level (Fig. 6.11). In this case, Facies D4a would represent lower-energy wave-dominated deposits with a minor, relatively weak tidal influence. As progradation progressed, the influence of stronger tidal currents would become more prominent and the trough cross-stratification of Facies D4b representing shallower, higher-energy deposition would form. The dominance of Facies D4b in the north-east of the Khufai Dome suggests that deposition there was more proximal, and that shallower, high-energy conditions were more prevalent. The granular lags at the top of each shallowing-up cycle could be explained by winnowing associated with each transgression and the subsequent move back to slightly deeper deposition (see Section 6.5.4.3). The initial thickening of the genetic units, followed by subsequent thinning possibly represents the initial availability of accommodation space that gradually became filled as overall highstand conditions continued. Alternatively, the thinning of the genetic units towards the top of Member 2 could reflect a reduction in the supply of sediment.

6.5.4.5. Facies D5: Bi-directionally cross-stratified sandstone

Bi-directional cross-stratification (Facies D5) occurs in the Mukhaibah Dome at the same level as the dominantly trough cross-stratified sandstones of Facies D4 in the Khufai Dome. In the Mukhaibah Dome, Facies D5 dominates for up to 6m and is associated with trough cross-stratification as well as wave-rippling (Facies D1) and planar cross-stratification (Facies D2). In both of the sections logged in the Mukhaibah Dome, bi-directional cross-stratification of Facies D5 overlies sections dominated by Facies D1 in which wave-ripples, undulations and accretionary micro-hummocks occur. Facies D5 is
then in turn overlain by a metre or two of trough cross-bedding before Facies Association E is established. The bi-directional cross-stratification occurs as 'herring-bone' structures in which mainly planar cross-sets trending in opposite directions lie directly on top of each other (Plate 6.19). The planar cross-sets are up to 12cm in height and have erosive tops and bases. More sigmoidally-shaped cross-sets in which top-sets are preserved also occur. These are up to 20cm in height, commonly erode into underlying rippled beds and have wave-rippled toe-sets and tops. Reactivation surfaces can often be seen within cross-sets.

The medium-grained sandstone these structures occur within is yellow-to-orange in colour. It is composed of 60-70% quartz grains, with up to 30% weathered feldspar, and rare shale clasts. The grains are dominantly sub-rounded and sorting is moderate to good. Rare quartz granules up to 0.5cm in size also occur.

Bi-directional or 'herring-bone' cross-stratification is diagnostic of tidal deposits (eg. Conybeare and Crook, 1968; De Raaf and Boersma, 1971). It records reversals of currents reflecting bedload transport and deposition during both the ebb and flood tides. Although no tidal bundling was observed, the reactivation surfaces within the sigmoidal cross-sets also indicate tidal deposition; probably when one phase of the tidal cycle was dominant (Section 6.5.2.2). This facies differs from most of the other tidal facies described from the Masirah Bay Formation in that both the ebb and flood tides caused the migration of the small dunes. It seems likely that the bi-directional cross-beds of Facies D5 represent deposition in a tidal channel that would have been swept daily by both the ebb and flood tides. The sigmoidal cross-beds and association with facies displaying uni-directional palaeocurrent directions may indicate times when either the ebb or the flood was dominant, but mostly, both tides were capable of sediment transport. Such a tidal channel probably formed in a similar tidally-influenced estuarine environment to Facies D4, representing one of many such channels that would have occurred between the tidal sand bars preserved in Facies D4.

6.5.4.6. Facies D6: Tidally-bundled sandstone

Facies D6 is developed within the Khufai Dome at the top of Facies Association D. Here it is a few metres thick (eg. 31-35m log KD1, 242-245m log KD2; Plate 6.20). It overlies
the last coarse-grained to granular bed at the top of the trough cross-stratified sandstone of Facies D4, and is overlain by the lower shoreface to offshore deposits at the base of Facies Association E. Facies D6 is composed of reasonably well-sorted fine- to medium-grained sandstone. Grains are subrounded to rounded and are dominantly quartz (~80%). This sandstone forms planar laminated beds that commonly pinch out laterally on a metre-scale. More undulatory to rippled beds occur interbedded with these planar laminated sandstones. A parting lineation on these bed surfaces trending between 160/340° and 180/360° is developed. Planar cross-stratification up to 15cm high can also be seen. The separation of the foresets in this cross-bedding varies periodically from closely-spaced (<1cm) to more widely-spaced (few cm) and back again (Plate 6.20). The tops of the cross-sets are commonly rippled.

The ‘bundled’ nature of the cross-sets of Facies D6 indicate a tidal influence; the periodicity reflecting neap-spring cycles (De Raaf and Boersma, 1971; Visser, 1980; Allen and Homewood, 1984; Reading and Collinson, 1996) (see Section 6.5.2.2 for more detailed discussion). More closely spaced foresets indicate neap tides when tidal currents were weaker, and more widely spaced ones indicate transport and deposition under the stronger currents associated with spring tides. The dominantly planar to slightly undulatory beds with parting lineations that dominate Facies D6 represent deposition in an upper flow regime (ie. by high velocity/shallow flows). These could have formed in the shallow flows that would be expected to wash across the tops of intertidal shoals/bars as tides ebbed and flowed. This would then suggest that Facies D6 represents deposition in an intertidal environment. Therefore, although there is a reduction in grain-size (and associated reduction in energy) compared to the underlying trough cross-stratified sandstones of Facies D4, this does not indicate deeper-water deposition. The flooding event indicated by the granular lag at the top of Facies D4 in the Khufai Dome is therefore unlikely to represent a major transgressive surface. Instead, it probably reflects a minor change in relative sea-level, and tidally-influenced deposition continues until the top of Facies D6. The major flooding that occurs before the onset of Facies Association E must therefore occur above this level.
6.5.4.7. Facies Association D: Synthesis

In the north of the Huqf area (Khufai and Mukhaibah Domes), the lowermost deposits of Facies Association D (dominated by Facies D1 and D2) indicate a continued shallowing from the underlying lower shoreface deposits of Facies C2 (Figs. 6.9B; 6.10). In the Khufai Dome, the subaqueous shrinkage cracks that occur within Facies D1, taken together with the small-scale undulatory and wave ripple cross-lamination, suggest either shoreface deposition close to the freshwater discharge of a river, or wave-influenced pro-delta slope/delta front deposition. The series of granular beds (Facies D3) that occur towards the top of Facies D1 possibly represent storm beds, or alternatively may reflect winnowing associated with a series of minor flooding events before the onset of the deposition of Facies D4 in a broad tide-dominated estuary mouth. Facies D4 indicates an overall continuation of the shallowing trend through the lower part of Facies Association D and probably indicates late highstand conditions. This trough cross-stratified facies dominates Facies Association D in the Khufai Dome. The coarsening-up cycles within Facies D4 represent progradational cycles caused by high-frequency changes in relative sea-level (Fig. 6.11). In this case, Facies D4a would represent lower-energy wave-dominated deposits with a minor, relatively weak tidal influence. As progradation progressed, the influence of stronger tidal currents would become more prominent and the trough cross-stratification of Facies D4b representing shallower, higher-energy deposition would form. The dominance of Facies D4b in the north-east of the Khufai Dome suggests that deposition there was more proximal, and that shallower, high-energy conditions were more prevalent. The granular beds at the top of each coarsening-up cycle probably represent winnowing of the finer material during the minor transgressive events between each cycle. The initial thickening and then subsequent thinning of the cycles possibly reflects highstand conditions in which initially available accommodation gradually became filled with continuing deposition. Above the trough cross-stratified tidal sand bars in the Khufai Dome the intertidal deposition of Facies D6 indicates continued shallow water deposition.

While deposition in Facies Association D was dominated by tidal sand bars in the Khufai Dome, deposition in a similar estuarine mouth environment, but in tidal channels swept by both the ebb and flood tides was prevalent further to the south in the Mukhaibah Dome (Facies D5) (Fig. 6.9B). At Al Harf, some 40km to the south-west, the wave-rippled
sandstone of Facies D1 dominates Facies Association D and the tidal influences that affected deposition further to the north-east are not apparent. This suggests that Al Harf was in a more distal position than the Khufai and Mukhaibah Domes during this time. This is further supported by the proximal to distal relationship that is seen moving from the north-east to the south-west in the Khufai Dome. The shallowest deposits at Al Harf were probably formed on a wave-dominated middle to lower shoreface.

The top of Facies Association D is overlain by a flooding event that can be seen across the entire Huqf area. This flooding marks the boundary between Facies Association D and Facies Association E, and between Members 2 and 3.

A continuation of the progradation recorded by Facies Association C is indicated at the base of Facies Association D by the continued coarsening- and shallowing-up (Fig. 6.10). In the Khufai Dome, this shallowing is continued further by Facies D4 which progrades over the underlying sediments and indicates the onset of late highstand conditions. Within Facies D4, a series of minor progradational cycles are recorded. The fact that intertidal sediments (Facies D6) are preserved overlying Facies D4 indicates the continuation of these cycles to the top of Facies Association D.

The more distal, wave-dominated sediments that are preserved at Al Harf are similar to the sediments at the base of Facies Association D further to the north-east. In the Khufai Dome these are covered by the progradation of Facies D4. At Al Harf a progradational shallowing-up is recorded through the sediments, but the fact that the overlying trough cross-stratified sandstones of Facies D4 are absent indicates that the estuarine tidal sand bars did not prograde this far to the south-west before the flooding at the base of Facies Association E occurred (Fig. 6.10).

6.5.5. Facies Association E: Lower shoreface to offshore deposits

Facies Association E forms Member 3 and is exposed overlying Facies Association D in the Khufai and Mukhaibah Domes and at Al Harf. It occurs at the top of the Masirah Bay Formation in the Huqf area, and is conformably overlain by the Khufai Formation. Although the outcrop is commonly partially covered in rubble, Facies Association E can be
seen to vary in thickness from 32 to 45m and to contain a remarkably similar series of facies throughout the Huqf area.

At the base of Facies Association E very thinly-bedded sandstones with *Arumberia*-like structures dominate (Facies E1). These pass up into the planar-bedded fine-grained carbonate and siltstone of Facies E2 that occurs below the Khufai Formation. Some of these beds have been remobilised and show slump structures. This sequence of facies in Facies Association E is recorded right across the Huqf area. A summary of the facies present within Facies Association E is given in Table 6.2.

6.5.5.1. **Facies E1: Very thinly-bedded sandstones with microbial 'wrinkle' structures**

Facies E1 occurs at the base of Facies Association E, directly overlying the top of Facies Association D. It can be seen in the Khufai and Mukhaibah Domes, and is probably only absent at Al Harf because of a lack of exposure at the appropriate level. It is overlain by the quieter and deeper water deposits of Facies E2 at all the localities at which it is exposed. Facies E1 varies in thickness from 12m in the Mukhaibah Dome to 18m in the Khufai Dome. Facies E1 is composed of red-coloured, very thinly-bedded, fine- to medium-grained sandstone. Beds are typically few mm to cm in thickness and are intercalated with finer recessive material (Plate 6.21). Sorting is moderate with sub-angular to sub-rounded grains dominantly composed of quartz. Glauconite is locally developed. In the lowermost few metres Facies E1 displays small (<2cm height) asymmetrical ripple structures, low-angle scours (Plate 6.22) and locally developed slump folding. The ripples mainly suggest palaeocurrents moving approximately east to west. Erosively-based lenses of coarse-grained sandstone up to 4cm thick also occur. Moving up through Facies E1, the sandstones become increasingly planar with only locally developed ripples and scours. On many bedding surfaces a series of sub-parallel grooves and ridges occur. These are only a mm or so in height and width, but are persistent lengthwise (with some branching and joining) on at least a metre-scale (Plates 6.23, 6.24). These structures can be observed in both the Khufai and Mukhaibah Domes at the same level and often completely cover bedding surfaces. At both localities the subparallel grooves and ridges consistently trend between 250° and 300°. The undersides of many of the beds the grooves
and ridges occur on preserve rare mm scale skip and prod marks. These indicate palaeocurrents acting at right-angles to the ridge and groove structures.

The finer-grain size of Facies E1 immediately suggests that these sandstones were formed in lower-energy conditions and probably deeper-water than the underlying tidally-influenced deposits at the top of Facies Association D. The ripples at the base of Facies E1 are asymmetrical and dominantly uni-directional suggesting that they were formed by the action of currents rather than waves. The lack of wave-rippling possibly indicates deposition below fairweather wave-base in the offshore environment. The scoured surfaces and coarser-grained beds that occur are probably the result of storms when more powerful currents could erode the existing sediments and transport coarser material offshore. The fact that these become increasingly rare up-section suggests that the site of deposition became progressively removed from the influence of storms. The very thinly-bedded sandstones that dominate the upper part of Facies E1 therefore probably represent deposition by currents (indicated by the skip marks on the bedding surfaces) in an offshore setting during fairweather conditions. The slumped beds that occur in this facies may reflect sedimentation on unstable slopes or dewatering after deposition.

The subparallel grooves and ridges bear some resemblance to *Arumberia banksi* that was first described from the Arumbera Sandstone, Northern Territory, Australia (Glaessner and Walter, 1975). Similar structures have since been reported from the latest Proterozoic elsewhere in the world (Bland, 1984; McIlroy and Walter, 1997). *Arumberia* was initially interpreted as a biogenic structure, firstly as an upright cup-like creature (Glaessner and Walter, 1975), and then as a colony of flexible thin-walled tubular elements firmly adpressed or possibly fused into a dense sheaf or bundle where it was in contact with the sediment surface (Bland, 1984). *Arumberia*-like structures are often associated with, or even superimposed upon, flute marks. This led to a subsequent re-interpretation of *Arumberia* as a non-biogenic structure, forming through the action of currents on cohesive muddy substrates which may have been microbiially bound (McIlroy and Walter, 1997). Its global occurrence at a similar stratigraphic level could then be explained by the subsequent rise of bioturbating metazoa, which may have destroyed such firm muddy substrates in the post-Cambrian by bioturbation and microbial cropping (Walter and Heys, 1985). Flute marks were not observed in Facies E1, however, and the skip and prod marks
that do occur indicate currents perpendicular to the ridges and grooves, rather than parallel as would be expected from the model of McIlroy and Walter (1997) for the formation of Arumberia. Facies E1 may well, however, owe its origin to a microbially bound sediment surface. Although modern microbial mat communities are typically only found in non-marine or stressed environments, in the Proterozoic, before metazoans became well established, they are commonly found in subtidal deposits (Hagadorn and Bottjer, 1997). A possible explanation for the structures in Facies E1, therefore, is that they formed on or just below the lower shoreface in cohesive, microbially bound sediments. The overlying shear stress of offshore currents moving over these sediments would have caused the sediment surface to become wrinkled into compressional ridges orientated perpendicularly to the current. This explanation fits with the depositional environment suggested by the sediments, and also explains the skip and prod marks orientated perpendicularly to the ridge and groove structures. The fact that these structures are only observed within Facies E1 in the Masirah Bay Formation is therefore probably a reflection of the occurrence of unique conditions when microbial mats were not only established, but were also wrinkled by currents moving across them.

6.5.2. Facies E2: Interbedded carbonate and siltstone beds

Facies E2 occurs directly above the thinly-bedded sandstones of Facies E1 in both the Khufai and Mukhaibah Domes (the lower contact not being exposed at Al Harf) and forms the top of the Masirah Bay Formation in the Huqf area. It is conformably overlain by the Khufai Formation. Facies E2 has a similar nature right across the Huqf area, but varies in thickness from 18m in the Mukhaibah Dome to >25m in the Khufai Dome and at Al Harf. The base of the facies is composed of few cm thick beds of well-indurated red to grey dolomite, which on a few cm to dm scale are interbedded with slightly finer, more friable siltstone beds (Plate 6.25). The alternation of these beds gives the base of Facies E2 a distinctive ‘stepped’ appearance. Beds are planar and contain a lamination that initially also appears planar, but which is actually slightly ‘crinkly’ on a mm scale. This lamination is defined by finer-grained, <0.5mm thick, mud-rich layers, which are commonly rich in brown-coloured opaque minerals. Rare quartz grains up to 0.1mm in size also occur within the carbonate. These ‘stepped’ beds dominate for up to 11m before passing up into lighter-coloured <20cm thick planar beds of siltstone containing a laterally continuous
planar lamination (Plate 6.26). These beds become increasingly dolomitic close to the boundary with the Khufai Formation. They show the development of a few mm scale planar lamination defined by alternations of slightly coarser and finer grains. Rarely, the beds towards the top of Facies E2 contain rafted flakes, or may show tight folded structures within them. Less than 3m below the contact with the Khufai Formation, lenticular iron-rich nodules up to a few cm in size occur.

Six samples from this level of the Masirah Bay Formation were analysed for stable isotopes (Table 6.3). The results show positive $\delta^{13}$C values varying between 2.2‰ and 4.6‰ and slightly negative $\delta^{18}$O values varying between −0.5 and −2.7‰.

The fine-grained and dominantly planar-bedded nature of Facies E2 suggests it was deposited in a quiet offshore environment. The laterally continuous laminations in the siltstones suggest that they formed from fall-out from suspension. The ‘crinkly’ lamination developed in the carbonates possibly indicates the influence of microbial activity on the sediment surface (Tucker and Wright, 1990). However, their fine-grained and dominantly planar nature still suggests that fall-out from suspension was the main process in their formation. The ‘stepped’ nature of the carbonate beds at the base of Facies E2 is simply a product of varying degrees of cementation. The influx of pelagic carbonate in Facies E2 could be linked to the progradation of the platform carbonates of the overlying Khufai Formation (Pilcher and Buckley, 1995; McCarron, 2000). Some later remobilisation of the material settling from suspension is indicated by the rare rafted flakes and the locally developed tight folding within beds that probably formed from soft sediment deformation related to dewatering. The iron-rich nodules that occur a few metres below the Khufai Formation are possibly an early diagenetic feature (Pilcher and Buckley, 1995). They are interpreted as indicating a period of low sedimentation rates associated with a maximum flooding surface. Overall, Facies E2 represents a move to deeper and quieter water deposition dominated by suspension settling. The increasing prevalence of dolomite towards the top of Facies E2, and the outer-ramp setting for the base of the overlying Khufai Formation (McCarron, 2000), suggest a shallowing at the top of Facies Association E.
The stable isotope results from Facies E2 are comparable to those recorded at the base of the Khufai Formation (McCarron, 2000). They indicate a return to positive values after the negative excursion of the Hadash Formation. The δ13C results are also similar to those recorded from the top of the Masirah Bay Formation in the Jebel Akhdar (Section 6.7.1.2), further corroborating the correlation between these two areas. An in-depth study of the stable isotopes from the Masirah Bay Formation was not conducted, however, and further conclusions based on them have not been attempted.

6.5.5.3. Facies Association E: Synthesis

The deposits of Facies E1 at the base of Facies Association E indicate a deepening from the underlying tidally-influenced Facies Association D. Facies E1 was deposited in a storm-affected lower shoreface to offshore environment. The reduction in the number of scoured surfaces and storm beds moving up through Facies E1 indicates quieter conditions and a probable continued deepening. The ridge and groove structures on bedding surfaces indicate the microbial colonisation of submarine sediment surfaces at this time. The finer-grained and planar-bedded nature of the overlying Facies E2 suggests the dominance of quiet water settling from suspension and indicates a continuation of the deepening trend. The fact that Facies E2 is most thickly developed at Al Harf probably indicates an increase in tectonic subsidence and accommodation space available towards the west that was then at least partially filled by the sediment supply. The carbonate ramp interpretation for the Khufai Formation (McCarron, 2000), and the increase in carbonate material directly below it suggests a shallowing towards the top of Facies Association E. This may have not have occurred until after the iron-rich concretions a few metres below the top of Facies Association E, which possibly indicate a maximum flooding surface at this level.

The initial deepening at the base of Facies Association E was probably accompanied by the backstepping of the wave-rippled sandstones at the base of Facies E1 onto the top of Facies Association (Fig. 6.10). As flooding continued until a few metres below the Khufai Formation, retrogradational to aggradational, relatively deep-water deposition probably dominated much of Facies Association E in the uppermost part of the Masirah Bay Formation.
6.6 Synthesis of the Masirah Bay Formation in the Huqf Area

Only the upper part of the Masirah Bay Formation is preserved in outcrop in the Huqf area. It is most completely exposed in the centre of the Khufai Dome, and is dominated by shallow marine siliciclastic deposition. The exposures in the Khufai Dome record two major coarsening-up sequences separated by a major flooding event. The lowermost of these coarsening-up sequences comprises Member 1. Member 1 records dominantly progradational deposition as it passes from Facies Association A into Facies Association B (Figs. 6.9A; 6.10). Wave-rippled sandstones deposited on the shoreface (Facies Association A) pass up into shallower, tidally-influenced deposits exhibiting minor shallowing-up cycles (Facies B1-B3), before the overall shallowing is continued and the estuary mouth tidal sand bar deposits of Facies B4 and B5 are established (Fig. 6.9B). These sand bar deposits are extensively developed across the Huqf area. The top of Facies Association B (and top of Member 1) is marked by a flooding event (Fig. 6.10). Winnowing associated with this transgression accounts for the granular deposits (Facies B6) that occur at the top of Facies Association B. These mark the top of Member 1 and the lower coarsening-up sequence.

The second coarsening-up sequence incorporates Facies Associations C and D (Member 2). The offshore, quiet-water shale deposits at the base of Facies Association C contain the maximum flooding surface. They become increasingly storm-dominated upwards and shallow-up into the shoreface/delta front deposits that occur at the base of Facies Association D. This part of the stratigraphy records the progradation of Facies Association D over Facies Association C. Overall shallowing continues through Facies Association D, although this is possibly punctuated by minor flooding events. A coarse-grained, estuary mouth tidal sand bar deposit (Facies D4) progrades over the underlying Facies Association D deposits (Figs. 6.9B; 6.10). Unlike the lower tidal sand bar deposits of Facies B4, this sandbody contains coarsening-up cycles. These probably reflect high-frequency changes in relative sea-level (Fig. 6.11). The higher-energy coarser parts of the cycles are more prevalent to the north-east of the Khufai Dome, suggesting more proximal deposition there. The tidal sand bar deposits of Facies D4 are overlain by intertidal deposits of Facies D6 in the Khufai Dome. At the top of Facies Association D in the Mukhaibah Dome, the deposits of a tidal channel swept by both the ebb and flood tides are preserved. Further to
the south at Al Harf, no tidal influence is seen, and the coarsening-up recorded from Facies Association C into Facies Association D ends in wave-rippled shoreface deposits. This suggests deposition to the south was lower-energy and more distal, and follows the same proximal-distal direction observed within the Khufai Dome. It seems likely that the marine-dominated estuarine deposits that occur in the Khufai Dome did not prograde as far as Al Harf, which was in a more offshore position at the time (Fig. 6.10). The coarsening-up, and dominantly shallowing-up sequence, recorded by Facies Associations C and D is overlain by a significant flooding event. Deeper-water and lower-energy deposition then dominates in Facies Association E (Member 3). Deepening continues above the initial transgression as Facies E1 passes up into Facies E2. A maximum flooding is probably reached a few metres below the Khufai Formation (Fig. 6.10). At the top of the Masirah Bay Formation in the Huqf area an increased amount of carbonate material and a shallowing occurs prior to the establishment of the Khufai carbonate ramp. Facies Association E at the top of the Masirah Bay Formation is relatively uniform across the Huqf and represents retrogradational to aggradational, relatively deep-water deposition across the whole area (Fig. 6.10).

The coarse-grained to granular sandstones that occur at the top of the two major coarsening-up sequences are both good potential reservoirs. They are both >30m thick and have considerable lateral extent (at least 80km in outcrop, and more inferred from subsurface data). Better exposure of the upper sandstone at the top of Facies Association D allows a progradation and eventual dying-out of this body to the south-west to be inferred (Fig. 6.12). Although fewer data were available for the lower coarse-grained sandstone at the top of Facies Association B, subsurface data suggests that this is also limited in its extent to the south-west (Fig. 6.13) (Bell, 1993a). However, probable equivalent coarse-grained sandstones can be seen to the north in Shara South, and to the east in Masirah Bay 1, suggesting their extent in this direction may be considerable (Fig. 6.13). Porosity in the subsurface varies from 5-18%, but is commonly affected by compaction, carbonate cementation, quartz overgrowths, and leaching of unstable minerals (Borgomano, 1994). Although no potential sources were observed in the Huqf area, very organic rich black shales are seen in outcrop in the Masirah Bay Formation in the Jebel Akhdar (Section 6.7.1). The presence of good reservoirs and potential sources makes the Masirah Bay Formation a potentially interesting future prospect.
Overall, the dominantly shallow marine interpretation of the Masirah Bay Formation in the Huqf area adopted in this study, with only minor (if any) fluvial input, is in general accordance with the work of Pilcher and Buckley (1995). Little evidence was seen to support the fluvial interpretations adopted for much of the Masirah Bay Formation in the Huqf area by Gorin et al. (1982), Dubreuilh et al. (1992) and Platel et al. (1992b).

6.7 The Masirah Bay Formation in the Jebel Akhdar

The Masirah Bay Formation in the Jebel Akhdar is considerably different to that exposed in the Huqf area. In the Jebel Akhdar, the whole of the Masirah Bay Formation is exposed. Its base is defined as occurring directly above the last carbonate bed or ‘stringer’ of the Hadash Formation, and it is again overlain by the carbonate of the Khufai Formation. The most complete exposures occur in Wadi Sahtan, Wadi Bani Awf, and Wadi Hajir (Figs. 6.5; 6.7), with further outcrops occurring in Wadi Mistal (Plate 6.27), Wadi Bani Jabir and Wadi Mu’aydin. The Masirah Bay Formation in the Jebel Akhdar varies in thickness from 170m in Wadi Hajir to ~200m thick in Wadi Bani Awf and Wadi Sahtan. It is dominated by deep-water, fine-grained deposits. The five Facies Associations identified in the Huqf area cannot be recognised, and there is no evidence for the two major progradational sequences seen in the Khufai Dome. Three facies have been identified as part of this study:

i) Argillaceous siltstone to fine-grained sandstone
ii) Limestone
iii) Medium- to coarse-grained cross-stratified sandstone

6.7.1. Facies Analysis

6.7.1.1. Facies 1: Argillaceous siltstone

Facies 1 dominates the Masirah Bay Formation for almost all of its 150-200m thickness in the Jebel Akhdar. It is composed of argillaceous siltstone, which is laminated on a few mm scale (Plate 6.28). This lamination is dominantly planar, but is often obscured by the
well-developed cleavage. Although these siltstones are commonly white or very light-grey in colour with red, iron-rich weathering, this appears to be a product of later weathering and leaching, as in other localities they are dark grey in colour (Plate 6.29) and are very rich in organic material (up to 3% Total Organic Carbon) (McCarron, pers. comm.). This organic-rich siltstone is developed locally near Hadash in Wadi Mistal, but is more extensive (>100m thickness) in Wadi Sahtan and Wadi Hajir. In Wadi Mistal, and locally in Wadi Hajir, better indurated beds 5-10cm thick occur. These are typically darker in colour and are interbedded on a dm scale with more friable, lighter siltstones, giving the outcrop a distinctive ‘step-like’ appearance. X-ray analysis of the siltstone (Kapp and Llewellyn, 1965) showed it to be composed of quartz, mica, and some plagioclase with traces of alkali feldspar, hematite and chlorite.

The fine-grained, thick and homogeneous nature of this facies suggests that it was deposited in quiet, deep-water. This is further suggested by its position above the transgressive Hadash Formation. The planar lamination suggests that deposition was dominated by fall-out from suspension, possibly from turbiditic plumes. The organic-rich nature of this facies suggests that organic matter has been partially preserved, indicating the sediment accumulated in a sub-oxic depositional environment.

6.7.1.2. Facies 2: Limestone

Limestone forms a minor facies in the Masirah Bay Formation in the Jebel Akhdar. It occurs towards the top of the sections in Wadi Hajir, Wadi Bani Awf, and Wadi Sahtan, where it is interbedded with the argillaceous siltstones of Facies 1. In Wadi Bani Awf, beds of limestone up to 1.5m thick occur in the upper 50m of the Masirah Bay Formation. In Wadi Hajir, however, limestone beds are only present in the uppermost 10m and are at most 0.7m thick. The limestone tends to have sharp upper and lower boundaries with Facies 1. It is micritic, dark grey in colour, and is homogeneous throughout the bed. Minor opaques with rarer quartz and mica occur. It seems unlikely that the absence of limestone in the east of the Jebel Akhdar was due to deeper water conditions there as coarser deposits and palaeocurrent directions from Wadi Bani Jabir suggest the east was actually in a more proximal location than the west. The fact that limestone is not seen
further to the east in the Jebel Akhdar is therefore probably due to a lack of exposure at the appropriate level.

Four samples from limestone beds in Wadi Bani Awf were run for preliminary stable isotope data. As in the Masirah Bay Formation in the Huqf area, the $\delta^{13}$C values are positive, varying from 1.4 to 5.2‰ (Table 6.3). Unlike the Masirah Bay Formation in the Huqf area, however, the $\delta^{18}$O values are strongly negative, varying from −5.6 to −7.3‰ (Table 6.3). Both the $\delta^{13}$C and $\delta^{18}$O results are remarkably consistent with those recorded from the base of the Khufai Formation in the Jebel Akhdar (McCarron, 2000).

These limestones are only developed at the top of the Masirah Bay Formation below the Khufai Formation. They share many similarities with the carbonates of the Khufai Formation and probably formed in a similar, quiet outer-ramp environment (McCarron, 2000). Their presence at the top of the Masirah Bay Formation indicates periodic carbonate deposition prior to the onset of Khufai proper sedimentation. Each carbonate possibly reflects a small-scale progradational cycle capped by a minor flooding event before the larger-scale progradation of the Khufai Formation over the top of the Masirah Bay Formation became well-established (McCarron, 2000).

The similarity between the $\delta^{13}$C results recorded here and those recorded from the top of the Masirah Bay Formation in the Huqf area suggests relatively open conditions in both areas at the time of deposition and supports the correlation between the two areas.

6.7.1.3. Facies 3: Medium- to coarse-grained cross-stratified sandstone

This facies only occurs in Wadi Bani Jabir in the far north-east of the Jebel Akhdar. Only the lower 20m or so of the Masirah Bay Formation is well-exposed here, but this is significantly different to that seen anywhere else in the Jebel Akhdar. Interbedded with the siltstones of Facies 1, a number of medium- to coarse-grained, and occasionally even granular sandstones occur. These are up to 30cm thick, sharply-based, and commonly show a fining-up. Many of the beds show no internal structures other than the grading, but in some a planar lamination is developed. A locally developed low-angled planar to swaley cross-stratification, commonly with an erosive base also occurs (Plates 6.30, 6.31).
This indicates palaeocurrents trending broadly towards the west/south-west. The sandstone is quartz-rich and grains are dominantly sub-rounded.

The sharp bases to beds, interbedding with siltstone, and locally developed swaley and low-angle cross-stratification probably indicate deposition by storms (Johnson and Baldwin, 1996) (see Section 6.5.2.1). Some of the graded beds may alternatively have been deposited by turbidity currents. Whatever the precise method of formation, the coarser-grained nature of this facies in Wadi Bani Jabir indicates a more proximal position to the palaeoshoreline compared to the rest of the Jebel Akhdar. The palaeocurrents preserved indicate flow towards the west/south-west and suggest that most of the coarse-grained material was deposited before the more distal areas in Wadi Mistal and further west were reached.

6.8. Synthesis of the Masirah Bay Formation in the Jebel Akhdar

The Masirah Bay Formation in the Jebel Akhdar is dominated by quiet, deep-water sedimentation. In the north-east, more proximal deposits are preserved. Palaeocurrent indicators preserved within these suggest that the fine-grained sediments further to the west were being derived from this direction. The thickening of the Masirah Bay Formation from Wadi Hajir to Wadi Sahtan perhaps indicates an increase in tectonic subsidence towards the west. Some of this accommodation would have been filled by the sediment supply, leading to the stratigraphic thickening observed in this direction.

Similarly to the Hadash Formation, there is no direct evidence for glaciation within the Masirah Bay Formation. This again indicates the major change in environment and depositional style that occurs at the top of the Ghadir Manqil Formation. Towards the top of the Masirah Bay Formation in the Jebel Akhdar limestone beds start to occur. These are more prominent in the more distal areas, and possibly indicate the punctuated progradation of the Khufai Formation from more proximal locations. If this is the case, the limestone beds at the top of the Masirah Bay Formation in the west could well be time equivalents to the base of the Khufai Formation in more proximal positions. In the more distal areas, carbonate would only be established intermittently and siliciclastic deposition would dominate for longer.
6.9. Summary of the Masirah Bay Formation

In this study, the term 'Masirah Bay Formation' has been used to include deposits in the Huqf and Jebel Akhdar regions lying between the Hadash and Khufai Formations. The Masirah Bay Formation is notably different in the two areas in which it is exposed. In the Huqf area, Masirah Bay deposition is dominated by shallow marine processes, and the sediments record two large-scale progradational sequences. In the Jebel Akhdar however, deep-water deposition dominates and no evidence for the sequences seen in the Huqf area is apparent. The Masirah Bay Formation is ~600m thick in the Huqf area (including data from Corehole-17), but is <200m in thickness in the Jebel Akhdar. This probably reflects differing rates of tectonic subsidence, and perhaps more importantly, differing rates of sediment supply between the two areas.

In both areas, more distal deposition occurs towards the west/south-west. This indicates that the Masirah Bay deposits in the Jebel Akhdar are not simply direct distal equivalents of those in the Huqf area lying much further to the south. The Masirah Bay sediments in the Jebel Akhdar may have been deposited in a separate basin to those in the Huqf area, or alternatively, may have been deposited in the same basin, but further to the north and further from the palaeocoastline (Fig. 6.14).
Chapter 7

Discussion and conclusions
CHAPTER 7. DISCUSSION AND CONCLUSIONS

7.1. Introduction

Each formation studied in this thesis has already been discussed to some extent within each relevant chapter. The aim of this chapter is therefore to summarise the main conclusions reached in the previous chapters, discussing each formation chronologically and analysing the depositional, tectonic, chronostratigraphic and global implications of this work. Finally, the burial/subsidence history of the Huqf Supergroup as a whole will be analysed in the light of these findings.

7.2. Discussion of the Huqf Supergroup

The nomenclature used for the lower Huqf Supergroup throughout the thesis has been slightly modified from previous workers. These changes are outlined in Figs. 3.7 and 6.8.

7.2.1. The Halfayn Formation

The Halfayn Formation comprises the volcanics and shallow marine/fluvial volcaniclastic sediments that overlie granodioritic basement near the village of Al Jobah in the Huqf area. New U-Pb zircon dates produced as part of this study suggest the basement at Al Jobah is c. 825 Ma in age, and that the Halfayn Formation is c. 800 Ma in age. The Halfayn Formation therefore pre-dates the volcanics of the Jebel Akhdar area, suggesting that the formation forms the oldest exposed deposits of the Abu Mahara Group (Fig. 7.1). The Halfayn Formation is unconformably overlain by a carbonate that occurs below the Masirah Bay Formation sediments of the area. This carbonate is probably an equivalent of the Hadash Formation in the Jebel Akhdar. This suggests the unconformity here represents a time-gap of ~200 Myrs, during which no deposition was recorded in the Huqf area.

Deposits from this time period are preserved in the Jebel Akhdar, however. The oldest of these comprise the Jabir/Ghubrah Formations.
7.2.2. The Jabir Formation

The Jabir Formation is a new formation, somewhat tentatively proposed in this thesis. It incorporates the immature sandstones and siltstones that unconformably underlie the Fiq Member of the Ghadir Manqil Formation in Wadi Bani Jabir in the north-east of the Jebel Akhdar. These are possibly Ghubrah Formation equivalents, but the lack of any evidence for glaciation within them can be used to suggest, albeit speculatively, that these sediments pre-date the Ghubrah Formation (Fig. 7.1). This suggestion is supported to some extent by sedimentary clasts within the Ghubrah diamictites that closely resemble the Jabir Formation deposits. If the Jabir Formation does represent the earliest phase of deposition exposed in the Jebel Akhdar, the lack of Ghubrah sediments could be explained if Wadi Bani Jabir was a high during Ghubrah time as it was for much of the Ghadir Manqil (indicated by the preservation of only 5m of Fiq deposits beneath the Hadash Formation). Alternatively, any Ghubrah deposits may have been eroded during Ghadir Manqil time, when Wadi Bani Jabir appears to have been situated on a rift shoulder. Further work looking to date horizons within the Jabir deposits, or possibly analysing heavy minerals across the Jebel Akhdar, would be needed before the Jabir Formation could be confidently placed within the Abu Mahara Group.

7.2.3. The Ghubrah Formation

The Ghubrah Formation comprises the oldest sediments exposed in Wadi Mistal, Wadi Mu’aydin, and Wadi Sahtan in the Jebel Akhdar. These sediments are dominated by ice-rafted diamictites that probably formed in a distal glaciomarine environment. The Ghubrah Formation can be correlated between Wadi Mistal and Wadi Mu’aydin by the presence of the overlying volcanics of the Saqlah Member. In Wadi Sahtan, where these are absent, the correlation is based on considerations of the overlying Fiq Member stratigraphy. An ash bed within the Ghubrah Formation that was previously dated through the U-Pb zircon method at 723 +16/-10Ma (Brasier et al., 2000; McCarron, 2000) was re-sampled as part of this study. The ash-bed yielded a new U-Pb zircon date of 711.8 ± 1.6Ma. This is comparable to the previously produced date and suggests it is highly unlikely that the zircons have been inherited. The new date therefore confirms that the Ghubrah Formation is of upper Sturtian age. This suggests it is an equivalent of the
Rapitan glaciation of North America (eg. Kaufman et al., 1997), the Chuos/Ghaub glaciations of the Kalahari and Congo cratons (eg. Kaufman et al., 1997; Hoffman et al., 1998a, b), and the lower ‘Varanger’ glaciation of Svalbard and Scandinavia (eg. Brasier and Shields, 2000). As this date actually comes from within a glacial diamictite, it provides one of the best constraints yet produced on the age of these Sturtian-equivalent glacials.

The homogenous nature of much of the Ghubrah Formation, coupled with the commonly pervasively cleaved and laterally discontinuous exposures makes correlations within it difficult to impossible. Lateral variability is also probably hidden by the incomplete exposure. The tectonic setting of the Ghubrah Formation is therefore uncertain.

7.2.4. The Ghadir Manqil Formation

The Ghadir Manqil Formation is a glacially-influenced formation. It is considered to be of Marinoan-age (c. 600-590 Ma) here through considerations of the overlying Nafun Group stratigraphy, and an age date of 544 ±3.3Ma from the Fara Formation. This would suggest it is an equivalent to the Blaubeker/Blaskrans of the Kalahari craton (eg. Saylor et al., 1998) and the Icebrook of North America (eg. Narbonne et al., 1994). Such a chronostratigraphic placement would make the Ghadir Manqil Formation considerably younger than the underlying Ghubrah Formation, and suggests that the boundary between the two formations represents a major unconformity encompassing almost 100 Myrs of time. Unfortunately this boundary is poorly exposed. Despite attempts to date horizons above this boundary (which only yielded inherited zircons), and attempts to differentiate between the Ghubrah and Ghadir Manqil diamictites based on clast type, the direct evidence for the suggested unconformity between the Ghubrah and Ghadir Manqil Formations remains scanty. However, using the new U-Pb zircon dates, at least 170 Myrs of time has to be incorporated into the Abu Mahara and Nafun Groups in the Jebel Akhdar. With current knowledge, it appears the most obvious way to account for a significant amount of this time is to place an unconformity between the Ghubrah and Ghadir Manqil Formations.
The Ghadir Manqil Formation is composed of the volcanic Saqlah Member and the glacially-influenced Fiq Member. The Saqlah Member occurs at the base of the Ghadir Manqil Formation, but is limited in its extent to the eastern part of the Jebel Akhdar (Fig. 7.1). The discontinuous lava flows it comprises vary in composition between alkali basalts and trachy-andesites. These probably belong to a differentiated alkali suite emplaced during extension in an intraplate continental setting. The volcanism of the Saqlah Member is probably related to the more extensive rift-related volcanics of the Hatat Formation in the Saih Hatat region. The presence of tholeiitic lavas in the Saih Hatat region indicates mantle melting and suggests a stretch factor of 1.5 with a raised mantle potential temperature, or a stretch factor of 3 with purely passive rifting (McKenzie and Bickle, 1988; Latin and White, 1990).

The Fiq Member of the Ghadir Manqil Formation overlies the Saqlah Member, or where this is absent, the Ghubrah Formation. The Fiq Member is well-exposed and preserves records of events that can be confidently correlated across the Jebel Akhdar. It comprises a sequence up to 1.5km thick that is dominated by sediment gravity flow deposits punctuated with at least four discrete glacial events. The variable thickness of the Fiq Member (1.5km in Wadi Sahtan compared to 5m in Wadi Bani Jabir ~60km further to the east), the large volume of commonly coarse-grained gravity flow deposits, and the volcanics of the Saqlah Member all provide good evidence that the Ghadir Manqil Formation was deposited in a rift basin (Fig. 7.1). The thin Fiq Member preserved in Wadi Bani Jabir cannot be due to post-Fiq erosion as it is overlain by the Hadash Formation. Wadi Bani Jabir was therefore probably situated on a rift shoulder during Fiq time. Palaeocurrent directions are dominantly towards the east in Wadi Sahtan, but are more commonly towards the west in Wadi Mistal. This suggests that material was being derived from both sides of a broadly north-south trending rifted basin approximately 60km across during Fiq time (Fig. 7.1). This would be comparable in size to rift basins imaged from the Abu Mahara Group from the subsurface of Oman (eg. Loosveld et al., 1996).

The Fiq Member of the Ghadir Manqil Formation contains a well-exposed and relatively complete record of a Neoproterozoic glaciation. These glaciations are recorded on every continent of the globe and were more widespread than any subsequent Phanerozoic glaciation. They have been the subject of much debate in recent years, as palaeomagnetic
work has revealed that sea-level glaciation reached into equatorial latitudes. To explain this, some workers (e.g. Hoffman et al., 1998a, b) favour a model involving global glaciation (snowball Earth model), whilst others (e.g. Williams et al., 1998) suggest that an increased obliquity of the Earth (high-obliquity model) would better fit the observed data. The Fiq Member preserves at least four discrete glaciations separated by non-glacial facies, some of which are only locally recorded within the basin. This suggests repeated glacial cycles and unfrozen seas that are not compatible with the snowball Earth hypothesis, which invokes long-lasting (~10 Myr) synchronous global glaciations during which time the hydrological cycle ceased. Although evidence for the extreme seasonality predicted by the high obliquity model was not observed in the Jebel Akhdar, the deposits here are marine. Equivalents of the Ghadir Manqil Formation in the south of Oman (Mirbat Sandstone) preserve shallower-water and continental glaciation, and contain ice wedges, suggesting that marked seasonal changes were occurring. The deposits of the Ghadir Manqil Formation therefore suggest that either a series of severe Phanerozoic-style glaciations, or an alternation of frozen and unfrozen conditions caused by high obliquity of the Earth are probably more characteristic of the Late Neoproterozoic glacial epochs than one long-lasting snowball Earth.

7.2.5. The Hadash Formation

The base of the Hadash Formation marks a major change in depositional environment. Its base is taken here to mark the boundary between the Abu Mahara and Nafun Groups (as was first suggested by McCarron, 2000). Above the top of the Ghadir Manqil Formation, no further direct evidence for glaciation is observed in Oman until the Phanerozoic. The Hadash Formation is a widespread, probably deep-water, transgressive deposit that occurs across the whole Jebel Akhdar, and which has been identified in the subsurface across the whole of Oman (the ‘basal carbonate member’ of Bell, 1993a). Similar thin carbonates capping Proterozoic glacial successions are well documented from around the world. They are commonly <15m thick, are laterally extensive on a basinal scale, and preserve a similar, highly negative $\delta^{13}C$ excursion. It has been suggested on the basis of these distinctive characteristics that such a ‘cap carbonate’ could be used to define a new ‘Terminal Proterozoic System’ (e.g. Christie-Blick et al., 1995). The Hadash Formation is therefore not only of importance within Oman, but is potentially also of global interest.
A detailed isotopic and geochemical analysis of the Hadash Formation was conducted as part of this study. Sampling for this study was conducted at a higher density than any previous analysis of Neoproterozoic cap carbonates. The results from this suggested that although the Hadash Formation does preserve a negative carbon isotopic signature similar to other Neoproterozoic cap carbonates, inflection points within the signature from this <15m thick carbonate unit that could be confidently correlated on a basinal, nevermind a global scale, could not be identified.

A number of models have been proposed to explain the occurrence of widespread and possibly even global cap carbonates and their associated negative isotopic excursions. Evidence from within the underlying Ghadir Manqil Formation suggests that the hydrological shut down during snowball conditions required by some models would be hard to envisage for the Omani section. Some evidence within the Hadash Formation for microbial activity (‘crinkly’ lamination, roll-up structures, and pyrite) could be used to support the model that cap carbonate deposition was linked to gas hydrate destabilisation (Kennedy et al., 2001). Whatever the exact mechanism for cap carbonate formation, however, the fact that a number of discrete glacial events occur in the underlying Ghadir Manqil Formation suggests that if cap carbonate formation is linked to deglaciation, cap carbonates do not have to be restricted to distinct levels in the stratigraphy. This, coupled with the difficulty found in trying to correlate $\delta^{13}$C curves at a <15m resolution, suggests attempting to define a new Terminal Proterozoic System based on the isotopic signatures of cap carbonate units may prove problematic.

The carbonate unit that occurs unconformably overlying the Halfayn Formation in the Huqf area has previously been correlated lithostratigraphically with the Hadash Formation. This Huqf carbonate preserves a significantly more positive $\delta^{13}$C signature than the Hadash Formation in the Jebel Akhdar, suggesting this correlation may not be valid. However, the fact that this carbonate marks a transgression onto the Huqf high that would correspond well with the Hadash transgression in the Jebel Akhdar prior to the onset of Masirah Bay deposition means the correlation remains attractive (Fig. 7.1). This leads to two possible scenarios: 1) the Huqf area was a high that was not flooded until a later stage in the transgression and deposition of the Nafun Group consequently began slightly later in the
Huqf area compared to the Jebel Akhdar; or 2) deposition was synchronous in the two areas, but the shallower water isotopic profile from the Huqf area does not preserve the basinal signature recorded in the Jebel Akhdar.

The fact that by the end of Hadash time broadly synchronous deposits are preserved in both the Huqf and Jebel Akhdar areas suggests basins were no longer partitioned to the same degree as they were during Abu Mahara times (Fig. 7.1). This probably indicates that a period of relative tectonic quiescence had become established.

7.2.6. The Masirah Bay Formation

The Masirah Bay Formation is well exposed in both the Huqf and Jebel Akhdar areas. These deposits can be correlated between the two areas based on lithological considerations as well as comparison of stable isotopes from the overlying Khufai, Shuram and Buah Formations (McCarron, 2000). In the Jebel Akhdar, the Masirah Bay Formation gradationally overlies the Hadash Formation. It is composed of <200m thickness of sediments dominated by fine-grained, deep-water siliciclastics. Coarser-grained deposits in Wadi Bani Jabir with palaeocurrent directions towards the west suggest that the material deposited in the Jebel Akhdar at this time was being derived from the east. In the Huqf area, only the upper part of the Masirah Bay Formation is well exposed. The sediments here preserve two major progradational sequences that each record a change from deposition in an offshore environment, through shoreface conditions, into marine-dominated estuarine deposition. These two progradational sequences and their associated changes in relative sea-level can not be identified in the deeper-water deposits of the Masirah Bay Formation in the Jebel Akhdar (Fig. 7.1). Palaeocurrent directions and facies in the Huqf area suggest that deposition became more distal towards the west/south-west. The upper parts of each progradational sequence in are composed of coarse-grained sandstone. Exposures in the Huqf area suggest that these sandstone bodies extend for at least 70km in a broadly north-south direction. From outcrop the sandstone bodies can be seen to thin in a south/south-westerly (increasingly distal) direction. A preliminary study of well data suggests that these sandstone bodies do not extend much further to the south or west, but that they may thicken further to the east. The top of the Masirah Bay
Formation in both the Jebel Akhdar and Huqf areas records the increasing prevalence of carbonate and a relatively gradual transition into the overlying Khufai Formation.

7.2.7. The Khufai, Shuram, and Buah Formations

The Khufai, Shuram and Buah Formations of the upper Nafun group are all well-exposed in both the Jebel Akhdar and Huqf areas. The carbonate Khufai Formation was probably deposited in a carbonate ramp setting, with more distal deposition in the north in the Jebel Akhdar compared to the Huqf area. The Shuram Formation is dominated by siliciclastics, and this has been suggested to indicate a change in climatic regime compared to the Khufai Formation (McCarron, 2000). The fact that carbonates within it preserve a strongly negative isotopic signature has been used to suggest it may represent a glacial period that is not recorded in the rock record of Oman. The Shuram Formation has therefore previously been correlated with deposits recording the younger Marinoan-age glaciation (Figs. 4.5; 4.6; Saylor et al., 1998; McCarron, 2000). The Buah Formation is a carbonate platform deposit. All three of the Khufai, Shuram, and Buah Formations maintain broadly similar lithologies from the Huqf area to the Jebel Akhdar suggesting the Nafun Group as a whole was deposited during a period of relative tectonic quiescence. Similarly to the Masirah Bay Formation, in all three, deeper water deposits are preserved in the Jebel Akhdar compared to the Huqf area (McCarron, 2000). At the top of the Buah Formation in the Jebel Akhdar, the ~600m thick Fara Formation composed of generally remobilised ash-rich deposits occurs. This indicates volcanism and suggests renewed tectonic activity at this time. Probable equivalents of the Fara Formation in the subsurface (the Ara Group) also contain volcanics and show large lateral variations in thickness. This indicates that towards the end of Huqf Supergroup time renewed tectonism had resulted in basin partitioning.

7.3. Subsidence history of the Huqf Supergroup

The new U-Pb zircon dates from the Huqf Supergroup and the time-frame suggested for the glacial of the Abu Mahara Group allow a subsidence/burial history to be constructed. The tectonic setting of the Ghubrah Formation is uncertain, but it is possible that the rifting in the Ghadir Manqil Formation represents reactivation of older (c. 720 Ma), Ghubrah
faults. If this is the case, the post-rift sag phase was possibly eroded prior to the onset of Ghadir Manqil deposition. Ghadir Manqil extension and rifting had ceased by the onset of the Hadash Formation at the base of the Nafun Group (c.590 Ma). The lateral continuity and lack of volcanism present in the formations of the Nafun Group suggest tectonic quiescence. They therefore probably record the post-rift sag phase of the Ghadir Manqil faulting. Another possibility is that the Nafun Group accommodation space was created by foreland flexure caused by a load situated to the west. This would tie in with the ‘Western Deformation Front’ and the suggestion that a major thrust belt arrived from the west in Ara time. The wide extent of the Nafun basin, however, suggests that it would have to be a rather rigid lithosphere if flexure related to a load situated some distance to the west was responsible for producing the basin. Above the top of the Nafun Group (c. 545 Ma), tectonic activity and basin compartmentalisation is renewed. This is possibly linked to the final stages of collision between East and West Gondwana that has been suggested to extend from Arabia to Antarctica at this time (eg. Dalziel, 1997; Fig. 1.7).

The volcanics of Saih Hatat and the Saqlah Member at the base of the Ghadir Manqil Formation were used to infer stretch factors greater than 1.5 for the extension during Ghadir Manqil times. The amount of subsidence resulting from a number of different stretch factors within the given time-frame can thus be compared to the actual stratigraphic thickness of sediments deposited in the Ghadir Manqil Formation and Nafun Group in the Jebel Akhdar (Fig. 7.2). Decompaction for the sediments has not been attempted, as the uncertainties on the age dates would still leave potentially large errors. A simple plot of stratigraphic thickness versus time for the Ghadir Manqil Formation and Nafun Group is shown in Fig. 7.2, alongside subsidence curves produced by varying stretch factors (calculated using equation 3.18, p58, Allen and Allen, 1990).

Calculated subsidence of 2340m for the Ghadir Manqil Formation and 900m for the post-rift Nafun group with a stretch factor of 1.5 (Fig. 7.2), compared to measured thicknesses of 1500m and 1100m respectively (this study and McCarron, 2000) suggests that the higher stretch factors producing greater amounts of subsidence are not applicable to the Jebel Akhdar. The volcanics in the Saih Hatat region that indicate some degree of mantle melting are possibly linked to greater stretching further to the east. Evidence for mantle melting was not apparent in the Jebel Akhdar, where stretch factors may have been lower.
The fact that the Nafun Group records a slightly greater thickness than would be predicted with a stretch factor of 1.5 may be indicative of a flexural influence. The good evidence for rifting within the Ghadir Manqil Formation does suggest, however, that at least some of the Nafun subsidence must have been linked to post-rift sag. Until time constraints within the Nafun Group are improved, and the shape of the subsidence curve can be inferred, the precise tectonic setting will remain uncertain.
References
REFERENCES


197


200


Dalrymple, R. W., Knight, R. J., Zaitlin, B. A. and Middleton, G. V. 1990. Dynamics and facies model of a macrotidal sand-bar complex, Cobequid Bay – Salmon River estuary (Bay of Fundy). Sedimentology, 37, p577-612


Deynoux, M. 1985. Terrestrial or waterlain glacial diamictites? Three case studies from the Late Precambrian and Late Ordovician glacial drifts in West Africa. *Palaeogeography, palaeoclimatology, palaeoecology*, **51** (1), p97-142


Glaessner, M. F. and Walter, M. R. 1975. New Precambrian fossils from the Arumbera Sandstone, Northern Territory, Australia. *Alcheringa, 1,* p59-69


Grey, K. and Corheron, M. 1998. Late Neoproterozoic stromatolites in glaciogenic successions of the Kimberly region, Western Australia: evidence for a younger Marinoan glaciation. *Precambrian research, 92*, p65-87


208


Kellerhals, P. and Matter, A. 2000. Facies analysis of a glaciomarine sequence, the Late Proterozoic Mirbat Sandstone Formation, Sultanate of Oman. Unpublished manuscript, Bern University, Switzerland


Levell, B. K. 1980b. Evidence for currents associated with waves in Late Precambrian shelf deposits from Finnmark, North Norway. *Sedimentology*, 27, p153-166


Morton, D. M. 1959. The geology of Oman. *5th World Petroleum Congress, 1959, section 1, paper 14*


Myrow, P. M. and Southard, J. B. 1996. Tempestite deposition. *Journal of sedimentary research, 66 (5)*, p875-887


Pearce, J. A. and Cann, J. R. 1973. Tectonic setting of basic volcanic rocks determined using trace element analyses. Earth and planetary science letters, 19, p290-300


219


