**Interplay between depositional facies, diagenesis and early fractures in the
Early Cretaceous Habshan Formation, Jebel Madar, Oman**

Veerle Vandeginste^{*1}, Cédric M. John¹, Christina Manning²

- 1) Imperial College London, Dept. of Earth Science and Engineering, Prince Consort
Road, London, SW7 2BP, U.K.
- 2) Department of Earth Science, Royal Holloway University of London, Egham Hill,
TW20 0EX, UK

*e-mail: v.vandeginste@imperial.ac.uk; phone: +44 20759 47332; fax: +44 20759 47444

ABSTRACT

Diagenesis and fracturing can significantly alter petrophysical properties of subsurface carbonate reservoirs, but the impacts of these processes at the inter-well scale are hard to predict. However, the initial distribution of sedimentary facies is easier to predict, and could template future diagenetic transformation and potentially influence fracturing as well. This paper provides a case study investigating the mutual influence of depositional facies on diagenesis and interpreted early fractures in the Berriasian-Valanginian Habshan Formation in Jebel Madar (Oman). The Habshan Formation mostly comprises ooidal limestones but dolomitized beds a few meters in thickness and laterally correlatable over at least 5 km are present. Discordant (vertical) dolomite streaks are present in the limestone but are rare. The various sedimentary and diagenetic processes were investigated using fieldwork, XRD analysis, petrography and geochemistry (major and trace elements, stable carbon and oxygen isotopes and strontium isotopes). The lithologies consist of mud-lean ooidal grainstones, mud-rich ooidal packstones and bioturbated limestones. The changes in the amount of fine-grained sediment suggest changing energy conditions at a backshoal position of a carbonate platform. The mud-lean versus mud-rich depositional textures impacted on early diagenesis by promoting abundant calcite cementation in the mud-lean ooidal grainstones and dolomitization in the mud-rich packstones. The discordant bodies, interpreted as sediment-filled early fractures, are dolomitized in the same way as the mud-rich ooidal packstones. Dolomitization occurred probably during the Valanginian-Hauterivian in a shallow burial setting, and was most likely caused by seepage reflux. Salt-doming post-dated dolomitization of the Habshan Formation. The results of this study document the inter-relationship between sedimentary facies, early diagenesis and early fractures.

Keywords:

depositional facies, diagenesis, dolomitization, fracturing, ooid, recrystallization, salt dome,
Middle East

1. Introduction

Imaging carbonate reservoirs by seismic reflection is one of the main techniques used in hydrocarbon prospection and production (e.g. Brown, 1991; Chilingarian et al., 1996). For sub-seismic scale interpretations, such as predicting facies away from a well, concepts of sequence stratigraphy are used to predict the lateral extent of potential reservoirs and seals (Mitchum et al., 1977). An aspect of carbonate reservoirs that receives comparatively less attention is predictive diagenesis (e.g. Jones and Xiao, 2006), even though diagenetic changes are frequent due to the unstable nature of carbonate minerals (e.g. Stehli and Hower, 1961) and can result in e.g. early cementation, secondary porosity development, and can influence fracture density.

Part of the issue with predicting diagenesis is that it is the cumulative product of many different processes, which can make it a challenging problem to tackle. A potential promising approach to overcome this challenge is to tie facies distribution in a sequence stratigraphic context with diagenetic trends (e.g. Taghavi et al., 2006) and early fractures (e.g. Zahm et al., 2010). Conceptually, one can hypothesize that the depositional facies of a rock will act as a template for early diagenesis, and that both depositional facies and early diagenesis will influence the degree to which a rock unit will fracture.

Outcrops provide natural laboratories to explore the potential links between facies, diagenesis and early fractures in carbonates. The current study focuses on dolomitization of the Habshan Formation that is composed mainly of ooidal packstones to grainstones (Droste and van Steenwinkel, 2004) deposited during the Berriasian to Valanginian (Early Cretaceous), and currently outcropping in Jebel Madar (Oman). The Habshan Formation in Jebel Madar is an ideal candidate for this study because the depositional textures of the carbonate rocks are

vertically repetitive and laterally continuous and seem to have impacted on the distribution of dolomite. Furthermore, dolomite is often of high importance for reservoir quality (e.g. Westphal et al., 2004), and a deeper understanding of its mode of formation and lateral distribution is desirable.

The goals of this paper are to 1) reconstruct the environment of deposition of the Habshan Formation, and the early diagenetic processes that took place, 2) understand specifically the process of dolomitization in the Habshan Formation, and 3) demonstrate the relationship between depositional facies, diagenesis, and interpreted early fractures in this formation.

The study integrates field observations, petrography, mineralogy and geochemistry to reconstruct the environment of deposition and the diagenetic history of the rock, and combines this with observation and measurements of what is interpreted as early fractures.

The regional example presented here is a case study of how complex processes in carbonate reservoirs result in different reservoir properties, in this case dolomitized units versus tight limestone units, and it can serve as potential analogue for subsurface reservoirs.

2. Geological setting

Jebel Madar is a folded and fractured carbonate carapace above a salt dome in the Adam Foothills south of the central Oman Mountains in northern Oman (Mount et al., 1998; Fig. 1). The Jebel has a dimension of 8 km by 5 km and rises about 500 m above the surrounding plains. The age of the diapirism of the Precambrian-Cambrian Ara salt is not well constrained. Farzadi (2006) suggests that the diapirism started shortly after the initiation of ophiolite emplacement in the Late Cretaceous, whereas Terken et al. (2001) links the Ara salt diapirism to tilting of the eastern flanks of the salt basins during the Miocene Alpine compressional phase. Ericsson et al. (1998) and Peters et al. (2003) suggest that minor salt movement initiated in the Late Jurassic and was followed by two periods of significant accelerated diapirism associated with the ophiolite obduction and the Alpine Orogeny. Also

Grélaud et al. (2006) proposes that diapirism was active during the Cenomanian, influencing the pattern of incisions in the Natih Formation.

Outcrops of Triassic to early Late Cretaceous age, belonging to the Hajar Supergroup, are visible along the main wadis in Jebel Madar (Fig. 1). The Hajar Supergroup is divided into four groups: the Akhdar, Sahtan, Kahmah and Wasia Groups (Fig. 2). Triassic dolostone from the Akhdar Group and Jurassic sandstone and limestone from the Sahtan Group crop out in the center of the Jebel, at the triple junction of the main faults. Unconformities occur at the Triassic-Jurassic and Jurassic-Cretaceous contacts. Emergence of the Arabian platform at the end of the Triassic relates to flexure of the Arabian platform upon opening of Neo-Tethys 2 as a result of the splitting of the Kavr-Lut microcontinent (Glennie, 2005). The Late Jurassic unconformity is caused by uplift of the eastern margin of the Arabian platform due to rifting preceding the opening of the proto Indian Ocean (Glennie, 2005).

Oman was located at an equatorial position during the Cretaceous (Farzadi, 2006), favouring extensive carbonate platform development on the passive margin of the Neo-Tethys. The Lower Cretaceous Kahmah Group comprises the Rayda, Salil, Habshan, Lekhwair, Kharaib and Shuaiba Formations (Fig. 2). The Rayda Formation, composed of thinly-interbedded mudstone and marl and thin beds of nodular chert, was the result of a rapid subsidence and subsequent flooding of the Arabian platform. The Salil, Habshan and Lekhwair Formations form a prograding succession of shallowing-upward facies. The distal slope mudstone of the Salil formation are characterised by shaly limestone and marls, mid-ramp shoal deposits of mainly ooidal and bioclastic grainstone and packstone characterise the Habshan Formation, and finally the Lekhwair formation is characterised by inner platform algal mudstone, wackestone, bioclastic-peloidal grainstone and rudist-bearing packstone and floatstone (Droste and van Steenwinkel, 2004). An emersion surface (sequence boundary) marks the top of the Lekhwair Formation, which is overlain by inner platform bioturbated, bioclastic

(rudist-bearing) and peloidal packstone of the Kharaib Formation and (on top) the Shuaiba Formation.

An unconformity marks the contact of the Shuaiba Formation and the overlying Wasia Group (Aptian-Turonian). The unconformity relates to the opening of the South Atlantic Ocean, which then led to the convergence in the Neo-Tethys, resulting eventually into intra-oceanic subduction and associated thrusting and obduction of the Semail ophiolite onto the Arabian Platform during the Campanian (Terken et al., 2001). Obduction of the ophiolite did not extend as far south as the Adam Foothills (van Buchem et al., 2002). The Wasia Group is composed of the Nahr Umr and Natih Formations. The Nahr Umr Formation is composed of calcareous shales and lime mudstones. The Natih Formation consists of interfingering carbonate platform and intrashelf basin deposits of thickly-bedded mudstone, bioturbated wackestone, packstone and grainstone with abundant rudists and *Thalassinoides* burrows (Droste and van Steenwinkel, 2004).

3. Methods

We sampled the lower part of the Habshan Formation in three different outcrops, two of them at the base of a cliff face at the southern side of the Southwest Gully and one at the northern side of the East Gully (Fig. 1). We collected 40 samples in total. Rock slabs were cut and finely polished using silicon carbide grit 220 and subsequently grit 600. The polished surfaces of the hand specimens were etched with 1M HCl and stained with Alizarin Red S and potassium ferricyanide to distinguish calcite and dolomite and their ferroan equivalents following a procedure modified from Dickson (1966). In addition, thin section halves were stained using the same solution.

A thin section of each hand specimen was examined using a Zeiss Axioskop 40 microscope as well as a CITL Cathodoluminescence Mk5-2 stage (operating conditions of about 280 μ A and 13 kV) mounted on a Nikon Eclipse 50i microscope. For X-ray diffraction analysis, one

to two grams of bulk rock samples were prepared by crushing small pieces of host rock (cleaned with distilled water and dried overnight) using an agate mortar and pestle. The samples were spiked with halite to serve as an internal standard. The analyses were performed on a Philips PW 1830 diffractometer system using $\text{CuK}\alpha$ radiation at 45 kV and 40 mA. The powders were scanned over a sampling range of 2.5 to $70^\circ 2\theta$ with a sampling interval of $0.01^\circ 2\theta$ per step and a scan speed of 2 seconds per step. Artificial mixtures of calcite and dolomite with internal halite standard were run to calibrate the XRD system and quantify the abundance of calcite and dolomite present in the samples. The maximum error of the calculated percentages at the 95% confidence level is about 2% where there is up to 10% of a component present and about 7% for higher percentages of a component. The stoichiometry of the dolomite was determined using the equation of Lumsden (1979). Since the Fe content in the dolomite is < 5500 ppm, no correction is applied (as the shift of the d_{104} peak due to Fe would be < 0.00065 , equivalent to < 0.2 wt% CaCO_3). For elemental analysis, bulk sample powders were prepared from limestone host rock ($>95\%$ calcite based on XRD) and dolostone ($>50\%$ dolomite). In addition, dolomite was separated by dissolving calcite in bulk rock powder using a buffered solution composed of 5% acetic acid with ammonium acetate added to obtain a pH of 5 for one hour. During this process a small part of the dolomite can dissolve along with the calcite. However, this process does not affect the chemical signature of the remaining dolomite, as demonstrated by the results (elemental ICP-AES as well as stable carbon and oxygen isotopes) of both the bulk sample and the extracted dolomite sample of JMA16 (a 100% dolomite sample) and dolomite standard (GBW07114). An aliquot of 250 mg of each bulk carbonate and separated dolomite sample powder was dissolved in 50 mL of a 5% HNO_3 solution. In this process, 30 mL of the acid was added in four steps, two of 5 mL and two of 10 mL. The solutions were then heated to about 80°C for one hour and subsequently filtered and more acid was added to obtain a volume of 50 mL. Element concentrations (Ca, Mg, Mn, Fe, Sr, K, Al) were measured in the solution by ICP-

AES. Analytical precision at the 95% confidence level determined on replicate analyses is about 15% for Ca, Mg, Al and Sr, and 10% for Fe and Mn. A powdered limestone (CRM752) and dolomite standard (GBW07114) were used for internal calibration of the method, and prepared in the same way as the samples. A total of 12 samples was also studied by microprobe to cross-check that no leaching of iron sulphides or oxides or clays had a significant effect on the concentration of the elements measured by ICP-AES in the samples and to determine in addition the Na, S and Cl content. The geochemical composition was determined using a CAMECA SX-100 electron microprobe at the Université de Montpellier II (France). A beam size of 15 μm was used and detection limits are about 130 ppm for Na, 80 ppm for Mg, 50 ppm for S, 120 ppm for Ca, 150 ppm for Sr, 75 ppm for Fe, 90 ppm for Al, 50 ppm for Cl and 50 ppm for K. The standards were albite for Na, forsterite for Mg, cadmium sulfide for S, wollastonite for Ca, celestine for Sr, hematite for Fe, corundum for Al, apatite for Cl and orthoclase for K. Microprobe analysis correction is based on Merlet (1994). Stable carbon and oxygen isotope analyses were carried out on carbonate powders that were sampled with a dental drill, where pure limestone or dolomite samples could be drilled. In limestone samples, the ooids were targeted besides mud. Dolomite for stable isotope analysis was obtained both by drilling and by chemical separation using acetic acid as explained above. The samples (100 to 150 μg) were reacted with phosphoric acid at 70°C in a Thermo Scientific automated Kiel IV carbonate device and the resulting CO_2 gas was analyzed on a MAT253 mass spectrometer in the Qatar Stable Isotope Lab at Imperial College. Drifts in the values were corrected using NBS19 and an internal standard (ICCM, Imperial College Carara Marble). Reproducibility is better than 0.03‰ (1 σ) for $\delta^{13}\text{C}$ and 0.06‰ (1 σ) for $\delta^{18}\text{O}$. The oxygen isotopic composition of dolomite was corrected for acid fractionation using the fractionation factors given by Rosenbaum and Sheppard (1986).

A part of the samples that were analysed for stable carbon and oxygen isotopes, were also selected for Sr isotope analysis. The pure calcite samples were dissolved in ~1 mL dilute

HNO₃. In contrast, dolomite samples were pre-leached in sufficient dilute acetic acid to remove a third of the sample, then a further third dissolved in dilute acetic acid, with just the latter fraction analysed. The pre-leaching is designed to remove exchangeable Sr and Sr in acid-soluble contaminant phases before dissolution of the cleaned residue in the second acid-leach step (Bailey et al., 2000). Strontium was separated from the solutions using Eichrom Sr-spec resin, and analysed on a VG354 thermal ionization mass spectrometer at Royal Holloway University of London (UK). Samples were loaded on single Re filaments with a TaF emitter, and run using the multidynamic procedure of Thirlwall (1991). The SRM987 standard ran with these samples gave a mean value of 0.710256 ± 20 (2σ , $N=3$), within error of the longer term mean (0.710259 ± 17 ; 2σ , $N=62$).

4. Results

4.1. Field observations and mineralogy

The outcrops are characterized by alternating grey-coloured, more recessive limestone beds and reddish brown-coloured, less recessive (calcitic) dolostone beds (containing about 40 to 100% dolomite; Fig. 3). These stratabound dolostone beds are usually 0.5 to 2 m thick, but can be traced laterally over several kilometers in the outcrops of Jebel Madar (Fig. 4A). In addition, metre-scale discordant reddish-coloured calcitic dolostone bodies (calcitic dolostone cf. classification by Leighton and Pendexter, 1962) cross-cutting several beds occur in the two Southwest Gully outcrops Site 1 and 2 but not in the East Gully outcrop Site 3 (Fig. 1). The largest discordant dolostone body observed is 15 m long and up to 1.5 m wide (Fig. 4B). Other discordant dolostone bodies are 2 to 3 m long and 30 cm wide and some fractures of 1 to 2 m long have dolomitized halos of 10 to 20 cm wide (Fig. 4C). The discordant dolostone bodies cross-cut limestone beds and connect with the overlying and underlying stratabound dolostone beds. Components in limestone and dolostone beds are very similar, but the limestone beds are more cemented with calcite compared to the dolostone beds which have a

similar composition as the discordant dolostone bodies. The largest discordant dolostone body has a similar structural orientation (N135E) as small calcite veins in this outcrop, in contrast to the fracture with dolomitized halo (N10E) which has a different structural orientation than that of the small veins and the large veins (N45E; Fig. 5). The calcite veins crosscut the limestone, stratabound dolostone as well as the discordant dolostone.

Discordant calcitic dolostone is composed of about 75% dolomite at Site 1 and of about 50% dolomite at Site 2. Dolomite stoichiometry of the samples falls into two clusters, i.e. a cluster at 50 to 52 mole% CaCO_3 and a second cluster at 56-57 mole% CaCO_3 . Both clusters contain samples from the stratabound and discordant dolostone (Fig. 6). All samples from the East Gully site (Site 3 on Fig. 1), where only stratabound dolostone is present, contain stoichiometric dolomite.

4.2. Petrography

The components in the grey limestone beds consist mainly of ooids (200 to 1000 μm in diameter), some aggregate grains and a few echinoid fragments, intraclasts and peloids. Low amounts of iron sulphides or oxides are generally present in the ooids and peloids. The ooids are often extensively micritized, and some of the echinoid fragments have a micritized rim whereas other echinoid fragments are overgrown by syntaxial calcite cement (SC). Limestone beds are either massive, bioturbated or laminated. The bioturbations are often filled by poorly sorted and coarser allochems than the surrounding limestone, and the mud-supporting matrix in the bioturbation features is generally dolomitized. The limestone beds are composed of grainstones with ooids that are indented at grain-to-grain contacts resulting in generally concavo-convex contacts (Fig. 7A, B) or pressure solution contacts. The ooids are in some samples enveloped by a very thin (15-30 μm) rim of small crystals of bladed calcite cement (Fig. 7C, D), which is also generally present between indented ooids. This circumgranular cement (CGC) has a nonluminescent core and brown dull luminescent rim under CL. The rest

of the space between the particles is filled with coarser blocky calcite cement that has a brown dull luminescent first main zone (BC1) and a second orange dull luminescent main zone (BC2) under CL, both zones displaying thin stripes of varying intensity (Fig. 7A-D). The same blocky calcite cement is also present in intraparticle pore space. Several types of thin veins filled with calcite cement postdate the blocky calcite cement (Fig. 7C, D). In chronological order, there are 1) thin veins with orange dull luminescence under CL (BC2), 2) thin irregular dissolution zones filled with nonluminescent calcite cement (DVC, Fig. 7C, D), and 3) veins with large calcite crystals that have brown orange dull luminescence under CL (LC). All phases are presented in the paragenetic sequence (Fig. 8).

The stratabound calcitic dolostone beds correspond to originally ooidal packstones and grainstones. Commonly, the micrite between the allochems is preferentially dolomitized (with relatively dark coloured euhedral to subhedral crystals of 30 to 100 μm in diameter). Besides ooids, aggregate grains, intraclasts and peloids are also common, whereas bioclasts such as echinoid fragments are less frequent. A few allochems contain some intraparticle calcite cement. Centimeter-scale calcite nodules are present in some beds (Fig. 9). At Sites 1 and 2, the ooids vary in size from 300 to 700 μm , but at Site 3 (where no discordant dolostone was found), the ooids are between 500 and 1300 μm in diameter. The particles in the dolomitic beds seem to be less well-sorted than in the limestone beds. In rare cases, a part of the sample contains calcite cement (zoned brown and orange dull luminescent under CL) between the particles instead of the more common micrite; in these cases, clear euhedral dolomite crystals occur at the rim between allochems and calcite cement and the dolomite crystals are coarser (up to 350 μm in diameter) than the micrite-replacing dolomite crystals. These coarse dolomite rhombs generally have a dark (sometimes dedolomitized) zone at the core to rim boundary. Some fine dolomite zones are red dull luminescent with mottled or corroded pattern under CL and other dolomite zones have a CL pattern of nonluminescent crystal cores with red and dark red dull luminescent outer zonations (Fig. 10A, B). In Site 3,

the dolomite crystals seem to have corroded rims and predate an inclusion-free dolomite with zoned CL pattern (Fig. 10C, D) that fills dissolution vugs along with CL zoned calcite (ZC). Samples from both Sites 1 and 3 show evidence for dedolomitization where dolomite is partially replaced by nonluminescent calcite or by calcite with a CL pattern of alternating nonluminescent and yellow bright luminescent zonations. Nonluminescent calcite cement also occurs in veins that postdate brown dull luminescent calcite veins in the stratabound calcitic dolostones. There are bedding parallel stylolites in the stratabound dolostones. The meter-scale discordant calcitic dolostone bodies correspond to ooidal packstones to grainstones (Fig. 10E). Besides abundant ooids (400 to 700 μm in diameter), some aggregate grains, intraclasts, echinoid spines and rare mollusc and rudist fragments are present. The interparticle zones are generally dolomitized (Fig. 10E, F), but in a few samples some micrite can be recognized. The transition between dolomitized and undolomitized limestone is generally gradual, lacking a sharp contact, or locally the transition can coincide with a stylolite or vein contact. Dolomite in the discordant dolostone bodies has the same textural and CL characteristics as dolomite in the stratabound bodies. Dolomite texture seems linked to locality (site) rather than to the stratabound versus discordant nature of the body. Replacement of ooids by dolomite is almost absent at Site 2 (where the bulk rock contains about 50% dolomite), but quite extensive at Site 1 (where the bulk rock consists of about 75% dolomite). Dolomite crystals partially replacing ooids are commonly slightly coarser (up to 50 μm) and more transparent than the finer, darker dolomite that replaces the micrite matrix. Partially dolomitized ooids commonly have a dolomitized core and calcitic rim. In some cases, the ooid center may not be dolomitized or could consist of coarse, limpid calcite. Echinoid fragments, whether in the core of ooids or in the surrounding sediment, are generally not dolomitized. Bedding parallel stylolites occur in the discordant dolostone and seem to cut dolomite crystals (Fig. 11A-D).

4.3. Major and minor element geochemistry

The limestone samples have significantly lower Mg (0.3-0.9%), Fe, Al and Mn and higher K and Sr contents than the dolostone samples (Fig. 12). The element signature in discordant dolostone is very similar to that in stratabound dolostone. The calcite-vein samples generally show similar elemental concentrations as the limestone samples, with the exception of the Mg (<0.3%), Fe (<42 ppm) and K (<140 ppm) contents that are significantly lower.

4.4. Stable carbon and oxygen isotope geochemistry

The limestone samples in this study have $\delta^{13}\text{C}$ values ranging between +1.7 and +2.7‰ VPDB and $\delta^{18}\text{O}$ values from -5.6 to -2.7‰ VPDB (Fig. 13). Samples of micrite, ooids and blocky calcite cement between the ooids have a similar signature. These values are similar to those of Berriasian-Valanginian carbonate from a well about 30 km north of Jebel Madar (1 to +2.5‰ VPDB for $\delta^{13}\text{C}$ and -5.5 to -4.3‰ VPDB for $\delta^{18}\text{O}$, omitting the most depleted values that were interpreted as diagenetically altered by meteoric diagenesis related to the paleo exposure surface at the top of the Habshan Formation; Wagner, 1990). Dolomite in stratabound beds is characterized by values of +3.1 to +4.2‰ VPDB for $\delta^{13}\text{C}$ and -3.7 to -1.0‰ VPDB for $\delta^{18}\text{O}$. Dolomite in the discordant bodies has a similar signature with $\delta^{13}\text{C}$ between +3.1 and +3.7‰ VPDB and $\delta^{18}\text{O}$ of -3.0 to -1.2‰ VPDB. Calcite cements in thick veins (LC cement) were also drilled for stable isotope analysis; their composition is marked by $\delta^{13}\text{C}$ of +1.4 to +2.6‰ VPDB and $\delta^{18}\text{O}$ of -8.4 to -4.3‰ VPDB. Similarly, calcite nodules have a composition of +2.1‰ VPDB for $\delta^{13}\text{C}$ and -6.8 to -4.8‰ VPDB for $\delta^{18}\text{O}$.

4.5. Strontium isotope geochemistry

The Sr isotope ratios are enriched in ^{87}Sr in most of the samples. Only stratabound dolostone from Site 2 has a ratio comparable to that of Early Cretaceous marine water (Veizer et al.,

1999). The other two samples (limestone and discordant dolostone) in Site 2 as well as the limestone sample from Site 1 have a moderate ^{87}Sr enrichment, whereas all other samples from Site 1 and Site 3 have a high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (Fig. 14). There is a general correlation between higher $^{87}\text{Sr}/^{86}\text{Sr}$ values, lower $\delta^{18}\text{O}$, lower mole% CaCO_3 (thus stoichiometric versus non-stoichiometric dolomite), lower Fe, lower Na and slightly higher $\delta^{13}\text{C}$ in the dolomite samples.

5. Discussion

5.1. Depositional facies

A dominance of ooids in the Habshan Formation and the presence of intraclasts in the ooidal packstones suggest a moderate-to-high energy, shallow-water environment, most likely the tidal environment of a shoal (e.g. Chow and James, 1987). The fact that well-sorted ooidal grainstones lacking micrite alternate with ooidal packstones and grainstones containing micrite suggests slight changes of the energetic level of the depositional environment. The grainstones lacking micrite were deposited above fair-weather wave-base in constant agitation. For the micrite-bearing packstones and grainstones, the depositional environment must have been slightly less energetic (e.g. a protected section at the back of a sand bar, possibly between fair-weather and storm-weather wave base, e.g. Brigaud et al., 2009) as to not sweep the lime mud away, but trap it between the particles. This is consistent with the fact that well-sorted ooidal grainstones are concentrated in the shallowest parts of the sand shoals and pass laterally into less well-sorted peloid-skeletal-ooid packstones in the slightly deeper water parts where tidal flow velocities are lower (Harris, 2010). In the Exumas of the Great Bahama Bank, aggregate grains are found platform-ward of the tidal delta lobes and locally in the tidal channels (Harris, 2010).

As documented in previous work (e.g. Droste and van Steenwinkel, 2004), the Habshan Formation represents ooidal shoal deposits at the rim of a prograding carbonate platform,

1 while slope and deeper water carbonate of the Salil Formation were deposited towards
2 northeastern Oman and the inner lagoon carbonate of the Lekhwair Formation were deposited
3 behind the shoal to the southwest. The Berriasian-Valanginian sea level fall (Haq et al. 1987)
4 corresponds to an erosion surface with karstic features at the top of the Habshan Formation
5
6
7
8
9 (Razin et al., 2007).

10 11 12 13 14 *5.2. Origin of the discordant bodies*

15
16 The discordant bodies represent subvertical small-scale heterogeneities of mud-supported
17
18 packstones that cross-cut both mud-rich and mud-lean ooidal grainstones. Based on their
19
20 geometry at the outcrop (Fig. 4), the discordant bodies could represent syndimentary or
21
22 early diagenetic fractures filled with sediment similar to the host rock. Fracturing in ooidal
23
24 grainstones can start as early as at the surface leading to a polygonal pattern of cracks caused
25
26 by dessication (Glumac et al., 2011). Larger scale fractures caused by brittle failure can also
27
28 develop at an early stage in the diagenetic history due to early lithification and cementation of
29
30 carbonates, such as in steep-sided platforms (e.g. Stanton and Pray, 2004) or in intertidal
31
32 ooidal grainstone deposits (Friedman, 1998). If these early fractures were opened to the
33
34 seafloor, unlithified younger marine sediments could fill them; such sediment-filled fractures
35
36 are known as neptunian dikes (Smart et al., 1988). The fracture infills would be characterized
37
38 by abundant fine-grained sediment mixed with coarser sediments resulting in mud-rich ooidal
39
40 packstone texture. An alternative hypothesis to explain the discordant vertical bodies would
41
42 be to interpret them as karstic dissolution features subsequently infilled with marine sediment.
43
44
45
46
47
48
49
50 Karstic features are possible in view of the paleo-exposure at the top of the Habshan
51
52 Formation (Wagner, 1990). However, no evidence was found to support this hypothesis, such
53
54 as vadose silt infill, alteration patina, or dissolution of grains at the contact of the discordant
55
56 body with the host rock. The orientation of the largest (i.e. meter scale) discordant dolostone
57
58 body (N135E) is parallel to the platform margin (Ziegler, 2001), similar to early fractures in
59
60
61
62
63
64
65

1 the Capitan reef formation where flexing and fracturing was triggered by the compaction of
2 unconsolidated slope and basinal sediment (Stanton and Pray, 2004; Kosa and Hunt, 2005).
3
4 However, our case is different because of the ramp setting (in contrast to the steep rim
5
6 platform of the Capitan reef) and the origin of the neptunian dikes is unknown, but could also
7
8 be related to earthquakes or salt movement.
9

10 11 12 13 14 *5.3. Paragenesis* 15

16 The alternation between micrite-rich and mud-lean depositional textures documented above
17
18 resulted in an alternating pattern of limestone and dolostone beds in the Habshan Formation.
19
20 Dolomitization of the Habshan Formation is most likely a diagenetic overprinting since the
21
22 ooidal packstone/grainstone texture precludes primary dolomite formation, for instance in a
23
24 sabkha environment. The earliest diagenetic process, observed in all beds, is micritization
25
26 (Fig. 15) partially masking the original ooid texture of concentric bands of radial mineralized
27
28 zones. The rims of some bioclasts are also micritized probably due to boring endolithic
29
30 organisms. Micritization of the ooids might suggest early diagenetic stabilization of
31
32 originally metastable carbonate, such as aragonite or high-Mg calcite (Cantrell and Hagerty,
33
34 1999). However, an aragonite precursor is not supported by the Sr concentrations (less than
35
36 350 ppm) or evidence of significant ooid dissolution (e.g. Tucker, 1985). There are only few
37
38 ooids that show dissolution of the nucleus. Near-surface early diagenesis in the mud-lean
39
40 ooidal limestones is characterized by the formation of a thin circumgranular cement rim
41
42 (CGC) enveloping ooids and of syntaxial cement (SC) surrounding echinoid fragments.
43
44 These two earliest cement phases (which are too small to be sampled for stable isotopes) are
45
46 probably linked to meteoric phreatic diagenesis based on the dog tooth crystal shape, the non-
47
48 luminescent zone of the calcite crystals under CL and the fact that these cement types are
49
50 common in such environment (Flügel, 2004; Taghavi et al., 2006; Walkden and Berry, 1984).
51
52 The limestones show evidence of mechanical compaction, such as concavo-convex ooid
53
54
55
56
57
58
59
60
61
62
63
64
65

grain contacts as well as chemical compaction, such as microstylolites. A third cement phase of calcite (blocky calcite, BC) fills the remaining pore-space in between the compacted carbonate grains. Since compaction is expected to be less prominent if the ooidal grainstones are completely cemented by a framework stabilizing interparticle calcite cement, we infer that the BC cement formed only after the onset of chemical compaction. Given favourable fluid circulation, the onset of formation of microstylolites can be as shallow as 30 m deep, but is usually a few hundreds of meters deep (Railsback, 1993). The BC cement probably predates dolomitization, but the petrographic textures are inconclusive. The BC cement displays the characteristics of a (shallow) burial cement with thick CL zones indicating some variation in the geochemistry of the cementing fluids or subtle changes in redox conditions. The fact that the micrite, ooids and BC cement have a similar stable isotope signature suggests that blocky calcite could be formed by marine water at a temperature within a range of 5 to 10°C of the limestone formation temperature. Some microfractures are also filled with shallow burial calcite cement (BC2) and predate some thin irregular dissolution vugs that are filled with nonluminescent calcite (DVC).

A dissolution phase creating vuggy porosity affected the mud-bearing ooidal carbonate before or contemporaneous with the onset of dolomitization. The dolomitization occurred mainly by fabric-preserving replacement. Minor dolomite cement formed in the vuggy porosity and predates a blocky CL zoned calcite cement (ZC). Dolomitization occurred before deepest burial since dolomite crystals are cut by bedding parallel stylolites. Dolomitization predates also the deformation associated with halokinesis at Jebel Madar, since the dolomite is crosscut by thick calcite veins (LC) that seem to be linked to the large-scale fractures and faults related to salt doming (John et al., 2012). Also the limestones contain bedding parallel stylolites and halokinesis-related fractures. Locally, the dolostone is affected by dedolomitization which is interpreted as a late-stage near-surface process with involvement of meteoric fluids (this interpretation is supported by the transparent,

undeformed nature of the crystals and the typical nonluminescent and yellow bright luminescent zonations observed under CL).

5.4. *Impact of facies and structures on diagenesis*

The original depositional texture, i.e. mud-rich ooidal packstones/grainstones versus mud-lean ooidal grainstones, significantly impacts on diagenesis by favouring calcite cementation in the high interparticle porosity beds and dolomitization in the mud-rich beds. First, the distribution of both CGC and SC early calcite cements are controlled by the presence of pore space between the components in the depositional host rock. Later, calcite-cemented, well-sorted ooidal grainstones are not affected by dolomitization, mud-rich ooidal packstones (both in stratabound and discordant bodies) are highly dolomitized, and bioturbated beds show dolomitization mainly restricted to the interparticle mud in between the coarser sediment in burrows. Dolomitization in the bioturbated beds is interpreted to relate to fluid focusing in the unlithified (and more permeable) burrows in contrast to the cemented/lithified nature of the surrounding host rock at the time of dolomitization (Kendall, 1977).

Alternatively, dolomitization could be enhanced by the presence of organic matter in the bioturbations (Horbury and Qing, 2004; Mason, 1980), but this is not supported by the $\delta^{13}\text{C}$ signature of the dolomite in this study. We propose two hypotheses to explain dolomitization in the mud-rich ooidal limestone and not in the mud-lean ooidal limestone. First, dolomitization could have been favoured in the mud-rich beds if blocky calcite cement predates dolomitization and had completely cemented the mud-lean beds, drastically reducing their permeability. In this case, the mud-rich layers would be the most permeable layers, at time of dolomitization. Second, dolomitization could have been favoured by the higher nucleation potential in the mud-rich beds compared to the mud-lean beds even if the latter were the most permeable in case blocky calcite postdates dolomitization. In that case, the

circumgranular and syntaxial cements and the high fluid flow in the mud-lean beds could have inhibited nucleation of dolomite.

5.5. Possible mechanisms for the dolomitization of the Habshan Formation

Dolomitization occurred in the shallow burial environment. Geochemical data adds further evidence, but this data needs to be interpreted with caution since cross-correlation between several proxies measured in dolomite indicates that the original geochemical signature is altered by recrystallization (Table 1). This recrystallization is confirmed by petrography: the Ca-rich dolomite crystals have a zoned, dark CL pattern or nonluminescent core, whereas the stoichiometric dolomite has a mottled (or corroded), red luminescent pattern under CL. The most pristine end of the dolomite trend (Fig. 16) is represented by the lowest $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7076) and $\delta^{13}\text{C}$ (+3.1‰ VPDB) values, but highest mole% CaCO_3 (56.6 mole%; non-stoichiometric dolomite), $\delta^{18}\text{O}$ (-1.0‰ VPDB), Fe (5478 ppm) and Na (382 ppm) values. The Sr isotope signature of the most pristine dolomite is consistent with a Valanginian-Hauterivian marine signature (Veizer et al., 1999). If the Sr content in the dolomitizing fluids was very low, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio could have been buffered by the host rock. However, the elevated Fe and Na content in the pristine dolomite do not suggest host rock buffering. The characteristics of the dolomite at the base of the Habshan formation are mainly thin (1-2 meters) stratabound beds of dolostone with a few discordant vertical streaks of dolomite. Two main conceptual models are considered for dolomitization of the Habshan Formation. The first model would be to tie dolomitization with halokinesis, and thus assume that the dolomitizing fluid interacted with the salt and was expelled during salt tectonics. Although the relatively high Na content in the pristine dolomite could suggest involvement of fluids that dissolved the Ara salt, we argue that this hypothesis is not likely because a) we would expect more abundant fracture-related dolostone bodies linked to salt tectonic, whereas the dolomitization is focused on stratabound dolostone geobodies, b) the Ara salt is composed of

halite and anhydrite (Mattes and Conway Morris, 1990) and thus not an abundant source of the Mg needed for dolomitization (however, we cannot completely rule out that formations present on the path of the diagenetic fluid between the Ara salt and the Habshan could be a source of Mg), and c) dolomite crystals are cross-cut by bedding parallel stylolites and thus dolomitization predates maximum burial and halokinesis. Nevertheless, fluids linked with salt diapirism could have played a role in the resetting of the geochemical signature (for example for Sr isotope ratios) in both limestone and dolostone (the recrystallized, stoichiometric dolomite, as explained above), an interpretation mainly based on the ^{87}Sr enriched signature of the recrystallized dolomite (and ~ 0.7092 in Ara salt; Burns et al., 1994). Circulation of fluids related to salt tectonics is likely because the sites are located along two of the three major faults linked to salt diapirism at Jebel Madar (Fig. 1).

The second mechanism for dolomitization of the Habshan Formation in Jebel Madar is seepage reflux (Fig. 17). The relatively high Na content is consistent with an origin of evaporated seawater involved in the dolomitization process (highest Na values of >350 ppm, Sass and Bein, 1988). Moreover, a relatively low temperature, shallow burial origin for the dolomite is also supported by a) the $\delta^{18}\text{O}$ value (high temperature would cause a low $\delta^{18}\text{O}$ value, Matthews and Katz, 1977), b) the planar dolomite texture (indicating formation at $<50^\circ\text{C}$, Gregg and Sibley, 1984) and c) the non-stoichiometric nature of the dolomite (Warren, 2000). The Fe enrichment (and also slight Mn enrichment) in the dolomite versus limestone can relate to reducing conditions during shallow burial when dolomite was formed (Warren, 2000) or to the presence of more iron-rich formations at the source or on the pathway of the dolomitizing fluids. A higher iron content in certain intervals could relate to a higher clay content (Van Der Kooij et al., 2007) or to trapping of iron in microbial mats (Dahanayake and Krumbein, 1986). In the seepage reflux dolomitization model (Fig. 17), evaporated seawater would be formed probably behind the ooidal shoals, in the inner lagoon where isolation of the basin favoured high evaporation in the arid conditions prevailing at the

equatorial position of northern Oman at that time (Farzadi, 2006). The heavy brine would sink through the lagoonal mud, flow basinward exploiting the most permeable layers and not penetrate deeper than the contact with the marls of the Salil Formation (Fig. 17). Although no evaporite deposition has been reported in Valanginian-Hauterivian layers in the study area, some calcite nodules observed in this study could indicate anhydrite nodule formation based on their cauliflower outer shape (Maliva, 1987). Valanginian-Hauterivian dolomite linked with evaporite deposits is documented in central-eastern U.A.E. with a similar paleofacies setting of prograding carbonate platform as in Jebel Madar (Ziegler, 2001). Anhydritic nodular limestone and peloidal and ooidal grainstone compose the Habshan Formation in central and eastern Abu Dhabi, whereas the same Habshan Formation contains mainly mudstone, peloidal limestone, dolostone and carbonate containing anhydrite or chert nodules in western Abu Dhabi (Alsharhan and Kendall, 1991). It is thus not unlikely that anhydrite would be present in the subsurface in the nearby region around Jebel Madar. Even if evidence for evaporite would be completely lacking in the subsurface around Jebel Madar, it is still possible that mesosaline (instead of hypersaline) brines drove seepage reflux dolomitization. For instance, fabric-preserving dolomite in the Capitan Formation forereef facies was explained by seepage reflux of mesosaline fluids (Melim and Scholle, 2002). Seepage reflux of meso- or hypersaline fluids is thus the most likely model to explain dolomitization in the Habshan Formation. As mentioned above, dolomite in the stratabound and discordant bodies has the same origin.

A geochemical proxy that remains intriguing in this study is $\delta^{13}\text{C}$. Commonly, the carbon isotope value of dolomite is buffered by the host rock signature, which does not seem to be the case in this study. Because the shift between pristine dolomite (+3.1‰ VPDB for the most pristine end member dolomite measured) and limestone (+2.4‰ VPDB $\pm$ 0.2 ‰) is small, an origin from peculiar environments, such as perennial saline lakes (Cerling and Hay, 1986) or by CO_2 degassing (Frisia et al., 2011) is unlikely. Also if methanogenesis was

involved, a documented diagenetic process in Middle Eastern Cretaceous rocks (Raven and Dickson, 2007), a larger discrepancy between $\delta^{13}\text{C}$ in limestone and that in pristine dolomite would be expected. A small increase in $\delta^{13}\text{C}$ can however be expected during evaporation of seawater, and thus the $\delta^{13}\text{C}$ in the dolomite could reflect the evaporative origin of the dolomitizing fluid. Still, what remains unresolved is the fact that the least pristine dolomite (stoichiometric dolomite with enriched $^{87}\text{Sr}/^{86}\text{Sr}$ and depleted $\delta^{18}\text{O}$, cf. alteration trend of dolomite as interpreted and discussed above) has a slightly enriched $\delta^{13}\text{C}$ value and that, in contrast, the altered limestone (with enriched $^{87}\text{Sr}/^{86}\text{Sr}$ and depleted $\delta^{18}\text{O}$) has more depleted $\delta^{13}\text{C}$ values. The alteration or recrystallization signal causing either slightly enriched $\delta^{13}\text{C}$ in recrystallized dolomite or depleted $\delta^{13}\text{C}$ in recrystallized limestone could mean that dolomite recrystallization was caused by different fluids than the limestone recrystallization, but that both fluids must have had high $^{87}\text{Sr}/^{86}\text{Sr}$ and either elevated temperatures or a low $\delta^{18}\text{O}$.

6. Conclusions

This study provides evidence of the inter-relationship between depositional facies, diagenesis and fracturing. In the case of the Habshan Formation, the initial control on permeability was the relative proportion of ooidal grains versus micrite that resulted in dramatically different diagenetic histories: the initially permeable ooidal grainstones were affected by calcite cementation, whereas the mud-rich lithologies were dolomitized. Discordant bodies interpreted as sediment-filled early fractures are dolomitized in the same way as the stratabound beds based on their similar mud-rich nature. Dolomitization happened by seepage reflux of evaporated seawater during the Valanginian-Hauterivian. Diagenesis of carbonate rocks is a complex problem involving multiple factors, but this study highlights that some of the late processes are pre-determined by early processes. In the Habshan Formation, sedimentological textures, the history of fluid flow and the sediment

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

infill of early fractures all play an important role in controlling the final distribution of dolostone bodies and tight cemented limestone. This implies that further work on statistical distribution of early fractures and the link with facies may provide geological rules to help predict the distribution and relative proportion of reservoir to non-reservoir units.

Acknowledgements

This research project is part of the Qatar Carbonates and Carbon Storage Research Center (QCCSRC), funded by Qatar Petroleum, Shell and the Qatar Science and Technology Park. Conxita Taberner is thanked for fruitful discussions and scientific advice. We would like to thank Martin Gill for running the X-ray diffraction analyses. We thank Assylzhan Dauletov for sample preparation and data treatment for XRD and stable isotopes.

References

- Allen, J.R.L., 1987. Desiccation of mud in the temperate intertidal zone: studies from the Severn Estuary and eastern England. *Phil. Trans. Roy. Soc. B* 315, 127-156.
- Alsharhan, A.S., Kendall, C.G.St.C., 1991. Cretaceous chronostratigraphy, unconformities and eustatic sealevel changes in the sediments of Abu Dhabi, United Arab Emirates. *Cretaceous Research* 12, 379-401.
- Bailey, T.R., McArthur, J.M., Prince, H., Thirlwall, M.F., 2000. Dissolution methods for strontium isotope stratigraphy: whole rock analysis. *Chemical Geology* 167, 313-319.
- Brown, A., 1991. Interpretation of three-dimensional seismic data. Third Edition. American Association of Petroleum Geologists Memoirs 42.
- Brigaud, B., Durlet, C., Deconinck, J.-F., Vincent, B., Pucéat, E., Thierry, J., Trouiller, A., 2009. Facies and climate/environmental changes recorded on a carbonate ramp: A sedimentological and geochemical approach on Middle Jurassic carbonates (Paris Basin, France). *Sedimentary Geology* 222, 181-206.

- Burns, S.J., Haudenschild, U., Matter, A., 1994. The strontium isotopic composition of carbonates from the late Precambrian (~560-540 Ma) Huqf Group of Oman. *Chemical Geology* 111, 269-282.
- Cantrell, D.L., Hagerty, R.M., 1999. Microporosity in Arab Formation carbonates, Saudi Arabia. *GeoArabia* 4, 129-154.
- Cerling, T.E., Hay, R.L., 1986. An isotopic study of paleosol carbonates from Olduvai Gorge. *Quaternary Research* 25, 63-78.
- Chilingarian, G.V., Mazzullo, S.J., Rieke, H.H., 1996. Carbonate reservoir characterization: a geologic – engineering analysis, part II. Elsevier, *Developments in Petroleum Science* 44.
- Chow, N., James, N.P., 1987. Facies-specific, calcitic and bimineralic ooids from Middle and Upper Cambrian platform carbonates, Western Newfoundland, Canada. *Journal of Sedimentary Research* 57, 907-921.
- Dahanayake, K., Krumbein, W.E., 1986. Microbial structures in oolitic iron formations. *Mineralium Deposita* 21, 85-94.
- Dickson, J.A.D. 1966. Carbonate identification and genesis as revealed by staining. *Journal of Sedimentary Petrology* 36, 491-505.
- Droste, H. 2001. Oman subsurface stratigraphy. *Petroleum Development Oman*.
- Droste, H., van Steenwinkel, M., 2004. Stratal geometries and patterns of platform carbonates: The Cretaceous of Oman. In: Eberli, G.P., Masferro, J.L., Sarg, J.F.R. (Eds.) *Seismic imaging of carbonate reservoirs and systems*. American Association of Petroleum Geologists Memoir 81, 185-206.
- Ericsson, J.B., McKean, H.C., Hooper, R.J., 1998. Facies and curvature controlled 3D fracture models in a Cretaceous carbonate reservoir, Arabian Gulf. In: Jones, G., Fischer, Q.J., Knipe, R.J. (Eds) *Faulting, fault sealing and fluid flow in hydrocarbon reservoirs*. Geological Society of London, *Special Publications* 147, 299-312.

- Farzadi, P., 2006. The development of Middle Cretaceous carbonate platforms, Persian Gulf, Iran: constraints from seismic stratigraphy, well and biostratigraphy. *Petroleum Geosciences* 12, 59-68.
- Flügel, E., 2004. *Microfacies of Carbonate Rocks. Analysis, Interpretation and Application*. Berlin, Heidelberg, New York: Springer-Verlag.
- Friedman, G.M., 1998. Rapidity of marine carbonate cementation – implications for carbonate diagenesis and sequence stratigraphy: perspective. *Sedimentary Geology* 119, 1-4.
- Frisia, S., Fairchild, I.J., Fohlmeister, J., Miorandi, R., Spötl, C., Borsato, A., 2011. Carbon mass-balance modeling and carbon isotope exchange processes in dynamic caves. *Geochimica et Cosmochimica Acta* 75, 380-400.
- Glennie, K.W., 2005. *The geology of the Oman Mountains: an outline of their origin*. Great Britain, Cambrian Printers, Aberystwyth, 110 p.
- Glumac, B., Curran, H.A., Motti, S.A., Weigner, M.M., Pruss, S.B., 2011. Polygonal sandcracks: Unique sedimentary desiccation structures in Bahamian ooid grainstone. *Geology* 39, 615-618.
- Gregg, J.M., Sibley, D.F., 1984. Epigenetic dolomitization and the origin of xenotopic dolomite texture. *Journal of Sedimentary Petrology* 54, 908-931.
- Grélaud, C., Razin, P., Homewood, P.W., Schwab, A.M., 2006. Development of incisions on a periodically emergent carbonate platform (Natih Formation, Late Cretaceous, Oman). *Journal of Sedimentary Research* 76, 647-669.
- Haq, B., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea-levels since the Triassic (250 million years to present). *Science* 235, 1156-1167.
- Harris, P.M., 2010. Delineating and quantifying depositional facies patterns in carbonate reservoirs: Insights from modern analogs. *AAPG Bulletin* 94, 61-86.

- Horbury, A.D., Qing, H., 2004. 'Pseudobreccias' revealed as calcrete mottling and bioturbation in the Late Dinantian of the southern Lake District, UK. *Sedimentology* 51, 19-38.
- Immenhauser, A., Dublyansky, Y.V., Verwer, K., Fleitman, D., Pashenko, S.E., 2007. Textural, elemental, and isotopic characteristics of Pleistocene phreatic cave deposits (Jabal Madar, Oman). *Journal of Sedimentary Research* 77, 68-88.
- John, C.M., Jourdan, A.-L., Kluge, T., Dale, A., Vandeginste, V., 2012. Clumped isotopes applied to carbonate diagenesis and high temperature systems. *Proceedings of the AGU 2012 Fall Meeting, San Francisco (California), 3-7/12/2012.*
- Jones, G.D., Xiao, Y., 2006. Geothermal convection in the Tengiz carbonate platform, Kazakhstan: Reactive transport models of diagenesis and reservoir quality. *AAPG Bulletin* 90, 1251-1272.
- Kaufman, J., 1994. Numerical models of fluid flow in carbonate platforms: implications for dolomitization. *Journal of Sedimentary Research* 64, 128-139.
- Kendall, A.C., 1977. Origin of dolomite mottling in Ordovician limestones from Saskatchewan and Manitoba. *Bulletin of Canadian Petroleum Geology* 25, 480-504.
- Kim, S.-T., O'Neil, J.R., 1997. Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochimica et Cosmochimica Acta* 61, 3461-3475.
- Kosa, E., Hunt, D.W., 2005. Growth of syndepositional faults in carbonate strata: Upper Permian Capitan platform, New Mexico, USA. *Journal of Structural Geology* 27, 1069-1094.
- Laubach, S.E., Eichhubl, P., Hilgers, C., Lander, R.H., 2010. Structural diagenesis. *Journal of Structural Geology* 32, 1866-1872.
- Le Métour, J., Platel, J.P., Béchenec, F., Berthiaux, A., Chevrel, S., Dubreuilh, J., Roger, J., Wyns, R., 1993. Geological map of Oman, scale 1:1,000,000: Directorate General of Minerals, Oman Ministry of Petroleum and Minerals.

- 1 Leighton, M.W., Pendexter, C., 1962. Carbonate rock types. In Classification of carbonate
2 rocks, a symposium. AAPG Memoir 1, 33-59.
- 3
- 4 Lumsden, D.N., 1979. Discrepancy between thin section and X-ray estimates of dolomite in
5
6 limestones. Journal of Sedimentary Petrology 49, 429-436.
- 7
- 8
- 9 Machel, H.G., 2004. Concepts and models of dolomitization: a critical reappraisal. In:
10
11 Braithwaite, C.J.R., Rizzi, G., Darke, G. (Eds.) The Geometry and Petrogenesis of
12
13 Dolomite Hydrocarbon Reservoirs. Geological Society of London, Special Publications
14
15 235, 7-63.
- 16
- 17
- 18 Machel, H.G., Anderson, J.H., 1989. Pervasive subsurface dolomitization of the Nisku
19
20 Formation in central Alberta. Journal of Sedimentary Petrology 59, 891-911.
- 21
- 22
- 23 Maliva, R.G., 1987. Quartz geodes: early diagenetic silicified anhydrite nodules related to
24
25 dolomitization. Journal of Sedimentary Petrology 57, 1054-1059.
- 26
- 27
- 28 Mason, T.R., 1980. Formation and diagenesis of Visean trace fossils from Ireland. Lethaia 13,
29
30 229-237.
- 31
- 32
- 33 Mattes, B.W., Conway Morris, S., 1990. Carbonate/evaporite deposition in the Late
34
35 Precambrian-Early Cambrian Ara formation of Southern Oman. In: Robertson, A.H.F.,
36
37 Searle, M.P., Ries, A.C. (Eds) The geology and tectonics of the Oman region. Geological
38
39 Society of London Special Publication 49, 617-636.
- 40
- 41
- 42
- 43 Matthews, A., Katz, A., 1977. Oxygen isotope fractionation during the dolomitization of
44
45 calcium carbonate. Geochimica et Cosmochimica Acta 41, 1431-1438.
- 46
- 47
- 48 Melim, L.A., Scholle, P.A., 2002. Dolomitization of the Capitan Formation forereef facies
49
50 (Permian, west Texas and New Mexico): seepage reflux revisited. Sedimentology 49,
51
52 1207-1227.
- 53
- 54
- 55 Merlet, C., 1994. An accurate computer correction program for quantitative electron probe
56
57 microanalysis. Mikrochimica Acta 114/115, 363-376.
- 58
- 59
- 60
- 61
- 62
- 63
- 64
- 65

- Mitchum, R.M.Jr., Vail, P.R., Thompson III, S., 1977. Seismic stratigraphy and global changes of sea level, part II: The depositional sequence as a basic unit for stratigraphic analysis. In: Payton, Ch. (Ed.) Seismic Stratigraphy: Applications to Hydrocarbon Exploration, American Association of Petroleum Geologists Memoirs 26, 53-62.
- Mount, V.S., Crawford, R.I.S., Bergman, S.C., 1998. Regional structural style of the central and southern Oman Mountains: Jebel Akhdar, Saih Hatat, and the Northern Ghaba Basin. *GeoArabia* 3, 475-490.
- Peters, J.M., Filbrandt, J., Grotzinger, J., Newall, M., Shuster, M., Al-Siyabi, H., 2003. Surface-piercing salt domes in the interior North Oman, and their significance for the Ara carbonate 'stringer' hydrocarbon play. *GeoArabia* 8, 1-40.
- Railsback, L.B., 1993. Intergranular pressure dissolution and compaction in a Plio-Pleistocene grainstone buried no more than 30 meters: Shoofly oolite, southwestern Idaho. *Carbonates and Evaporites* 8, 163-169.
- Raven, M.J., Dickson, T., 2007. Methanogenesis during Shu'aiba diagenesis: examples from Al Shaheen Field, Block 5, offshore Qatar. *GeoArabia* 12, 37-58.
- Razin, P., Grelaud, C., Mattner, J., 2007. The Cretaceous carbonate platform in Oman. Norsk Hydro field trip, v.1-4.
- Rosenbaum, J., Sheppard, S.M., 1986. An isotopic study of siderites, dolomites and ankerites at high temperatures. *Geochimica et Cosmochimica Acta* 54, 603-610.
- Sass, E., Bein, A., 1988. Dolomites and salinity: a comparative geochemical study. In: Shukla, V., Baker, P.A. (Eds) Sedimentology and geochemistry of dolostones, based on a symposium. SEPM Special Publications 43, 223-233.
- Smart, P.L., Palmer, R.J., Whitaker, F., Wright, P., 1988. Neptunian dikes and fissure fills: an overview and account of some modern examples. In: James, N.P., Choquette, P.W. (Eds) Paleokarst. Springer, Heidelberg, p. 149-163.

Stanton, R.J., Pray, L.C., 2004. Skeletal-carbonate neptunian dikes of the Capitan reef:

Permian, Guadalupe Mountains, Texas, U.S.A. *Journal of Sedimentary Research* 74, 805-816.

Stehli, F.G., Hower, J., 1961. Mineralogy and early diagenesis of carbonate sediments.

Journal of Sedimentary Petrology 31, 358-371.

Taghavi, A.A., Mork, A., Emadi, M.A., 2006. Sequence stratigraphically controlled diagenesis governs reservoir quality in the carbonate Dehluran Field, southwest Iran.

Petroleum Geoscience 12, 115-126.

Terken, J.M.J., Frewin, N.L., Indrelid, S.L., 2001. Petroleum systems of Oman: charge timing and risks. *American Association of Petroleum Geologists Bulletin* 85, 1817-1845.

Thirlwall, M.F., 1991. Long-term reproducibility of multicollector Sr and Nd isotope ratio analysis. *Chemical Geology* 94, 85-104.

Tucker, M.E., 1985. Calcitized aragonite ooids and cements from the Late Precambrian Biri Formation of southern Norway. *Sedimentary Geology* 43, 67-84.

van Buchem, F.S.P., Razin, P., Homewood, P.W., Oterdoom, W.H., Philip, J., 2002.

Stratigraphic organization of carbonate ramps and organic-rich intrashelf basins: Natih Formation (middle Cretaceous) of northern Oman. *American Association of Petroleum Geologists Bulletin* 86, 21-53.

Van Der Kooij, B., Immenhauser, A., Steuber, T., Hagmaier, M., Bahamonde, J.R.,

Samankassou, E., Tomé, O.M., 2007. Marine red staining of a Pennsylvanian carbonate slope: environmental and oceanographic significance. *Journal of Sedimentary Research* 77, 1026-1045.

Veizer, J., Ala, D., Azmy, K., Bruckschen, P., Buhl, D., Bruhn, F., Carden, G.A.F., Diener, A., Ebner, S., Godderis, Y., Jasper, T., Korte, C., Pawellek, F., Podlaha, O.G., Strauss, H., 1999. $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ evolution of Phanerozoic seawater. *Chemical Geology* 161, 59-88.

- Wagner, P.D., 1990. Geochemical stratigraphy and porosity controls in Cretaceous carbonates near the Oman Mountains. In: Robertson, A.H.F., Searle, M.P., Ries, A.C. (Eds.), *The Geology and Tectonics of the Oman Region*, Geological Society Special Publication 49, 127-137.
- Walkden, G.M., Berry, J.R., 1984. Syntaxial overgrowths in muddy crinoidal limestones: cathodoluminescence sheds new light on an old problem. *Sedimentology* 31, 251-267.
- Warren, J., 2000. Dolomite: occurrence, evolution and economically important associations. *Earth-Science Reviews* 52, 1-81.
- Westphal, H., Eberli, G.P., Smith, L.B., Grammer, G.M., Kislak, J., 2004. Reservoir characterization of the Mississippian Madison Formation, Wind River Basin, Wyoming. *AAPG Bulletin* 88, 405-432.
- Zahm, C.K., Zahm, L.C., Bellian, J.A., 2010. Integrated fracture prediction using sequence stratigraphy within a carbonate fault damage zone, Texas, USA. *Journal of Structural Geology* 32, 1363-1374.
- Ziegler, M.A., 2001. Late Permian to Holocene paleofacies evolution of the Arabian plate and its hydrocarbon occurrences. *GeoArabia* 6, 445-504.

Figure Captions

Figure 1. Geological map of northern Oman (simplified after Le Métour et al., 1993). Inset figure shows part of the Middle East, locating Oman (indicated in orange), with the green box delineating the northern Oman area presented on the geological. The bottom figure shows a Google Earth image of Jebel Madar with geological formations indicated (modified from Immenhauser et al., 2007). Location of the study sites are indicated.

Figure 2. Overview of stratigraphic units outcropping in Jebel Madar (based on Droste, 2001).

Figure 3. Lithologic log from Site 1. The base of this log corresponds to the base of the Habshan Formation, and thus the contact with the underlying Salil Formation. Dolomite content is based on XRD analyses on hand samples, collected as indicated.

Figure 4. Outcrop photographs. A) Laterally continuous stratabound dolostone (SBD). Tree as scale. B) Stratabound dolostone (SBD) and discordant dolostone (DD) in Site 1. Person (1.7 m tall) as scale. C) Dolomitized halo along fracture at Site 1. Lens cap (diameter of 5.8 cm) as scale.

Figure 5. Rose diagram showing structural orientation of calcite veins and dolomite bodies (total of 8 measurements).

Figure 6. Dolomite stoichiometry on the x-axis (mole% CaCO_3 calculated from XRD data) versus ordering on the y-axis (width of the dolomite peak d_{104} at mid-height, expressed in 2θ). The plot indicates no significant difference between the discordant and stratabound dolostone. FWHM: Full Width at Half Maximum

Figure 7. Microphotographs of ooidal limestone. Scale bar is 500 μm . (A) Concavo-convex contacts (cc) of ooid grains, blocky calcite cement (BC) filling interparticle pore space and calcite vein in grainstone sample from Site 1. (B) CL picture of same view as panel A. CL zoned cement (BC1 and BC2) in interparticle zones and orange luminescent calcite vein (BC2). (C) Circumgranular cement (CGC) enveloping ooid grains, blocky calcite cement (BC) filling interparticle as well as intraparticle (ip) pore space, thin irregular dissolution vugs filled with cement (DVC) in cemented grainstone sample from Site 2. (D) CL picture of same view as panel C. CL zoned cement (BC1 and BC2) filling interparticle and intraparticle

pore space, nonluminescent cement in dissolution vugs crosscutting ooids and cement in interparticle zones. The bluish luminescent dots are an artefact of thin section preparation.

Figure 8. Compiled paragenetic sequence derived from thin sections analysis of limestone and dolostone samples in the three sites studied. Colour indications refer to CL colour, DL = dull luminescent.

Figure 9. Photograph of calcite nodule with cauliflower outer shape. Lens cap (diameter of 5.8 cm) as scale.

Figure 10. Microphotographs of dolostones. Scale bar is 250 μm . (A) Stratabound dolostone from Site 2 with relict undolomitized micrite and dolomite rhombs being crosscut by calcite veins. (B) CL picture of same view as panel A. Dolomite rhombs have a CL zoned pattern and display an orange luminescence at the rims of calcite veins. (C) Stratabound dolostone from Site 3 with cloudy dolomite rhombs (cd), inclusion-free dolomite (d) and calcite cement (ZC) in dissolution vugs. (D) CL picture of same view as panel C. Corroded cloudy dolomite rhombs with orange-red luminescent rims and non luminescent dedolomitized zones. Inclusion-free dolomite displays a reddish CL zoned pattern and the vug-filling calcite has a brownish CL zonation. (E) Discordant dolostone from Site 1. Fine crystalline dolomite between ooid grains and some coarser dolomite within the grains. (F) CL picture of same view as panel E. Reddish luminescent dolomite that is affected by dedolomitization (light brownish luminescent calcite) to a minor extent.

Figure 11. Photographs of cross-cutting relation between discordant dolostone and bedding parallel stylolites. The stylolite is indicated by arrows. (A) Macrophotograph documenting bedding parallel stylolite present in both limestone and discordant dolostone. Lens cap

(diameter of 5.8 cm) as scale. (B) Microphotograph showing bedding parallel stylolite in discordant dolostone. (C) Microphotograph of bedding parallel stylolite cross-cutting the dolomite in partially dolomitized discordant body. (D) CL view of panel C. Note how the dolomite crystals are cut by the bedding parallel stylolite.

Figure 12. Box plots of minor element concentrations from ICP-AES measurements, except for the Na values which were measured by microprobe. Note that about half of the limestone samples had a Na concentration below the detection limit and that the box plot is only based on the values above the detection limit, thus representing an overestimate for Na concentration in limestone. L: limestone, DD: discordant dolomite and SBD: stratabound dolomite

Figure 13. Cross-plot of stable isotopes of carbon and oxygen. The light grey box indicates the signature of Berriasian-Valanginian carbonate in “Well E”, about 30 km north of Jebel Madar in Oman (Wagner, 1990). The dark grey box represents the range for Berriasian-Valanginian marine carbonate based on Veizer et al. (1999). The limestone values measured at Jebel Madar are similar to the subsurface values of Oman.

Figure 14. Cross-plot of stable isotopes of strontium and oxygen. The grey box indicates the signature of Berriasian-Valanginian marine carbonate based on Veizer et al. (1999). The label (number) next to each data symbol indicates the sampling site.

Figure 15. Schematic representation of the diagenetic changes occurring in the different lithologies, A) mud-bearing ooidal packstone/grainstone, B) mud-lean ooidal grainstones and C) bioturbated beds. The discordant dolomite bodies are not represented in these diagrams. See text for details. CGC: circumgranular cement, SC: syntaxial cement, BC: blocky cement,

DVC: nonluminescent calcite cement in dissolution vugs, ZC: CL zoned cement, LC: coarse calcite cement

Figure 16. Correlation of dolomite stoichiometry (in mole% CaCO_3 , calculated from XRD data) with several proxies, i.e. stable carbon and oxygen isotope and strontium isotope values and Fe, Mn and Sr content (ICP-AES) and Na content (microprobe) in dolomite. The arrow indicates the alteration or recrystallization trend.

Figure 17. Seepage reflux model explaining dolomitization at the base of the prograding Habshan formation. Evaporative conditions in the back shoal lead to the formation and sinking of a meso- or hypersaline brine.

Table 1. Isotopic data for the samples selected for strontium isotope analysis. Stable oxygen and carbon isotope values are presented in per mil VPDB. Also dolomite stoichiometry (expressed in mole% CaCO_3) is given for dolomite-bearing samples.

Figure 1

[Click here to download high resolution image](#)

Figure 2
[Click here to download high resolution image](#)

Figure 3
[Click here to download high resolution image](#)

Figure 4
[Click here to download high resolution image](#)

Figure 5
[Click here to download high resolution image](#)

Figure 6

[Click here to download high resolution image](#)

Figure 7

[Click here to download high resolution image](#)

Figure 8
[Click here to download high resolution image](#)

Figure 9
[Click here to download high resolution image](#)

Figure 10
[Click here to download high resolution image](#)

Figure 11
[Click here to download high resolution image](#)

Figure 12
[Click here to download high resolution image](#)

Figure 13

[Click here to download high resolution image](#)

Figure 14

[Click here to download high resolution image](#)

Figure 15
[Click here to download high resolution image](#)

Figure 16
[Click here to download high resolution image](#)

Figure 17

[Click here to download high resolution image](#)

Table 1

Table 1. Isotopic data for the samples selected for strontium isotope analysis. Stable oxygen and carbon isotope values are presented in per mil VPDB. Also dolomite stoichiometry (expressed in mole% CaCO₃) is given for dolomite-bearing samples.

Sample	Type	Site	⁸⁷ Sr/ ⁸⁶ Sr	2SE	d ¹⁸ O	d ¹³ C	Mole% CaCO ₃
JMA3	Discordant dolostone	1	0.709653	0.000009	-3.0	3.7	51.7
JMA7	Limestone	1	0.708134	0.000014	-5.2	2.3	
JMA16	Stratabound dolostone	1	0.709307	0.000016	-2.5	3.7	51.7
JMA26	Limestone	2	0.708259	0.000010	-5.3	2.3	
JMA27	Stratabound dolostone	2	0.707597	0.000010	-1.2	3.4	56.2
JMA28	Discordant dolostone	2	0.708250	0.000022	-1.3	3.2	55.7
JMC16	Stratabound dolostone	3	0.709426	0.000016	-2.0	4.1	50.5
JMC19	Calcitized dolostone	3	0.709596	0.000012	-3.8	1.7	