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2 *Linking process, dimension, texture and geochemistry in*
3 *dolomite geobodies: a case study from Wadi Mistal*
4 *(northern Oman)*

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14 **ABSTRACT**

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16 Understanding the distribution and geometry of reservoir geobodies is crucial for net-
17 to-gross estimates and to model subsurface flow. This paper focuses on the process
18 of dolomitization and resulting geometry of diagenetic geobodies in an outcrop of
19 Jurassic host rocks from northern Oman. Field and petrographic data show that a
20 first phase of stratabound dolomite is crosscut by a second phase of fault-related
21 dolomite. The stratabound dolomite geobodies are laterally continuous for at least
22 several hundreds of meters (~1000 ft) and probably regionally and are half a meter
23 (1.6 ft) thick. Based on petrography and geochemistry, a process of seepage reflux of
24 mesosaline or hypersaline fluids during the early stages of burial diagenesis is
25 proposed for the formation of the stratabound dolomite. In contrast, the fault-related
26 dolomite geobodies are trending along a fault that can be followed for at least 100

meters (328 ft), and vary in width from a few tens of centimeters up to 10 meters (~1 to 33 ft). Petrography, geochemistry and high homogenization temperature of fluid inclusions all point to formation of the dolomite along a normal fault under deep burial conditions during mid to Late Cretaceous times. The high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in the dolomite and the high salinity measured in fluid inclusions indicate that the dolomitizing fluids are deep basinal brines that interacted with crystalline basement. The dolomitization styles have an impact on dimension, texture and geochemistry of the different dolomite geobodies, and a modified classification scheme (compared to the one from Jung and Aigner, 2012) is proposed to incorporate diagenetic geobodies in future reservoir modeling.

Running head:

Linking process, dimension, texture and geochemistry in dolomite geobodies

Keywords:

Carbonate, dolomitization, geobody, heterogeneity, fault, stratabound

INTRODUCTION

Heterogeneities in subsurface reservoirs are controlled by depositional facies and diagenesis and can be described as architectural elements of finite dimensions, or geobodies. Several studies such as Harris et al. (2011) focused on quantifying dimensions of carbonate sand bodies and their spatial patterns. Recent efforts led to a hierarchical classification scheme of carbonate geobodies and a set of rules to characterize their dimension, shape, orientation and spatial distribution as a basis for reservoir modeling (Jung and Aigner, 2012). Components of this classification scheme includes geological time, system, zone, shape, element and facies of deposition (Jung and Aigner, 2012). Aside from shape, all of the components in the Jung and Aigner (2012) classification are related to sedimentology and depositional

environment; diagenetic geobodies are not taken into consideration (Table 1A). Diagenesis is however especially important in carbonate reservoirs, since carbonate minerals are generally more reactive than silicates and that diagenetic processes change the rock fabric and pore network rapidly and dramatically. This paper focuses on two types of dolomite geobodies and stresses the importance of a process-based classification of diagenetic geobodies. Dolomitization is one of the main diagenetic processes that affect carbonate reservoirs, and it can strongly impact on the petrophysical properties of carbonate rocks. Mole-per-mole replacement of calcite by dolomite results in a porosity increase of about 13%, making dolomite a favorable reservoir rock (Weyl, 1960). However, overdolomitization (i.e. precipitation of a dolomitic cement) is common and decreases the calcite to dolomite replacement porosity gain. Moreover, the complex dolomitization process changes the fabric and rearranges the pore network influencing permeability characteristics of the rock. Dolomitization is thus likely to form geobodies with different geometries and flow-unit properties, adding to the heterogeneity of the reservoir (e.g. Lapponi *et al.*, 2011). Dolomite bodies that formed during or soon after deposition, such as in a sabkha environment or seepage reflux dolomitization, can be predicted to have geometries similar to the original depositional geobodies. However, dolomite bodies formed by other dolomitization mechanisms, such as structurally-controlled dolomite bodies (e.g. Davies and Smith, 2006; Lonnee and Machel, 2006; López-Horgue *et al.*, 2010; Shah *et al.*, 2010; Sharp *et al.*, 2010), are not expected to have a close tie with the depositional environment, especially when the host rock is significantly compacted at the time of dolomitization. As a consequence, their dimension and spatial distribution cannot be predicted in subsurface based on stratigraphic principles alone. Hence, a better understanding of the dolomitization process and how this impacts the spatial distribution of dolomite geobodies will bring us a step closer towards predictive modeling of heterogeneity in carbonate reservoirs.

One approach to characterize the 2D and (when possible) 3D shape and distribution of dolomite geobodies is to study outcrops that enable the assessment of lateral extent and distribution of dolomite, as well as its link to large-scale structures. Hence, this study focuses on outcrops of both stratabound and fault-related dolomite bodies in the Jebel Akhdar tectonic window in Oman. The aim of our study was to explore how the geobody geometries, texture and geochemical signature were linked to the dolomitization process involved, and to test how these different diagenetic geobodies could be classified using current schemes. To achieve this goal, we first reconstruct the paragenesis, timing, formation and fluid geochemistry of the stratabound and fault-related dolomitization phases. We then combine this dataset with a structural analysis, and compare the above dataset to the geobodies dimensions, shape and distribution in order to assess how processes impact on geobody dimensions and geometries. This leads us to propose a simple modification to the Jung and Aigner (2012) classification for geobodies for the purpose of reservoir modeling.

GEOLOGICAL SETTING

Oman is located at the southeastern extent of the Arabian Plate (Fig. 1). The study area is situated in the central part of the 700 km long Oman Mountains. The Permian-Triassic tectonic context of Oman involves the breakup of Pangea, which led to the opening of the Neotethys Ocean between Gondwana and the Cimmerian continental blocks (Stampfli and Borel, 2002). This breakup also resulted in the formation of a passive continental margin on the northern edge of the Indian and Arabian plates (Chauvet *et al.*, 2009). As a consequence of the Cretaceous to Neogene convergence of Laurasia and Gondwana, the Neotethys Ocean was closed and the Arabian plate margin underwent local uplift during the flexure of the lithosphere (Warburton *et al.*, 1990; Searle, 2007). Intra-oceanic subduction started in the

Cenomanian and continued into the Middle Turonian to Early Campanian with the southwest-directed obduction of the Semail ophiolite (Boudier *et al.*, 1985; Hacker, 1994; Breton *et al.*, 2004). This Late Cretaceous convergence also caused the thrusting of the Hawasina volcano-sedimentary nappe complex in Oman (Hanna, 1990). A multi-phase deformation episode early after the nappe emplacement and later Tertiary compression gave rise to the anticline of Jebel Akhdar (Searle, 1985; Michard *et al.*, 1994; Poupeau *et al.*, 1998).

Jurassic rocks (of the Sahtan Group) outcrop on the mountain flank along Wadi Mistal in Jebel Akhdar, a tectonic window where the Arabian Platform outcrops below the Semail ophiolite (Fig. 1). Here, Permian to Cretaceous platform sediments unconformably overlie Neo-Proterozoic to Cambrian basement (Rabu *et al.*, 1990).

METHODOLOGY

Dolomite geobodies were identified by the brown (stratabound) or red-brown (fault-related) weathering color at the outcrop, and measured using a surveyor's tape. The position of the fault was recorded, when visible, as well as the extent of the dolomite geobody into the stratigraphic layers. A fracture analysis was carried out on two beds chosen based on both accessibility and representativity of the beds compared to the entire outcrop. One of the two fracture analysis windows is a 22 m (72 ft) long transect in a gray limestone host rock bed (Fig. 2A). The second fracture analysis window is a 15 m (49 ft) long transect in a brown stratabound dolostone bed (Fig. 2A). The presence and position of all macroscopic fractures was recorded to assess changes in fracture density with distance to the main fault (see Appendix 1). The strike, dip, width, filling, and geometry of the fractures were also recorded. The average fracture width is 7 mm in both beds and the width ranges from submillimeter up to 10 cm (3.9 inch). The structural data were plotted onto stereographic

projections using OSXStereonet, a freeware by Nestor Cardozo. The structural data are presented in Appendices 1-3.

We investigated 136 hand specimens that were sampled in a zone extending about 10 m (33 ft) laterally and 100 m (328 ft) along the main fault. Host rock, stratabound dolomite and fault-related dolomite were sampled using a combination of hammer and chisel and a motor-driven rock corer. Rock slabs were cut and finely polished using silicon carbide grit 220 and subsequently grit 600. The polished surfaces 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 total of 139 thin sections were examined using a Zeiss Axioskop 40 polarization microscope (with a connected Zeiss AxioCam ICc1 digital camera for microphotographs) and a CITL Cathodoluminescence Mk5-2 stage mounted on a Nikon Eclipse 50i microscope (with an attached Nikon DS-Fi1c digital camera) for cathodoluminescence (CL) microscopy. Operating conditions for the CL stage were about 280 μ A and 13 kV. The CL color descriptions reported are based on unstained thin section halves.

Small rock pieces of carbonate phases were cut, cleaned with distilled water, dried overnight and then crushed to a fine powder using an agate mortar and pestle. One to two grams of this rock powder was spiked with halite to serve as an internal standard for X-ray diffraction (XRD) analysis. The analyses were carried out on a Philips PW 1830 diffractometer system (at the Natural History Museum, London) using CuK α radiation at 45 kV and 40 mA. The XRD system is fitted with a PW 1820 goniometer and a graphite monochromator. The powders were scanned over a sampling range of 2.5 to 70°2 θ (θ is the diffraction angle) with a step of 0.01°2 θ 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 calcite

and dolomite amounts present in the samples. The analytical precision is about 7% (2σ standard deviation). The stoichiometry of the dolomite was determined using the equation of Lumsden (1979), after correction for the shift in d-value associated with the iron concentration in the ferroan dolomite.

For elemental analysis, 250 mg of each carbonate sample powder of either dolomite or calcite (pure, not containing any iron oxides or hydroxides based on XRD) was dissolved in 50 ml of 5% HNO_3 . In detail, 30 ml of the acid is added in four steps, two of 5 ml and two of 10 ml. The solutions were heated to about 80°C and kept at this temperature for one hour and subsequently filtered (to determine the HNO_3 insoluble residue) and mixed with the remaining 20 ml of 5% HNO_3 to obtain a final volume of 50 ml. The HNO_3 insoluble residue varies from 2 to 29%. Elemental concentrations for Ca, Mg, Mn, Fe, Na, Sr, K, and Al were measured by ICP-AES at the Natural History Museum (London). Analytical precision determined on replicate analyses is about 15% for Ca, Mg, Al and Sr, and 10% for Fe and Mn (2σ standard deviation). Concentrations of Na appeared to be affected by the filtering process, and hence, the Na values are not used in this study. A powdered dolomite standard (GBW07114) was used for internal calibration of the method, and prepared in the same way as the samples. As an additional test, the geochemical composition was determined on several thin sections using a CAMECA SX-100 electron microprobe at the University de Montpellier II (France). Microprobe analysis correction is based on Merlet (1994). The microprobe analyses were consistent with the ICP-AES data for Ca, Mg, Mn, Fe, Sr and K, whereas the Al content determined by microprobe was slightly higher than that by ICP-AES.

Carbonate samples (100 to 150 μg) for stable carbon and oxygen isotope analyses were taken with a dental drill. This microsampling approach allowed us to sample specific diagenetic phases. The samples were subsequently reacted with phosphoric acid 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 London. All carbonate values in the text (both for oxygen and carbon isotope) are reported in per mil relative to the Vienna Pee Dee Belemnite (VPDB), and were corrected by running NBS19 ($\delta^{13}\text{C}$ value of +1.95‰ and a $\delta^{18}\text{O}$ value of -2.20‰) and an internal laboratory standard (Imperial College Carrara marble, ICCM, $\delta^{13}\text{C}$ value of +2.09‰ and a $\delta^{18}\text{O}$ value of -2.03‰). The oxygen isotopic composition of dolomite was corrected for acid fractionation using the fractionation factors given by Rosenbaum and Sheppard (1986) and Kim *et al.* (2007). External reproducibility was determined by replicate analysis of laboratory standards and NBS19 and was better than 0.06‰ for $\delta^{13}\text{C}$ and 0.12‰ for $\delta^{18}\text{O}$ (2σ standard deviation).

For Sr isotope analyses, a few mg of carbonate powder were dissolved in 500 μl 3M HNO_3 at 80°C for one hour and centrifuged. The solution was loaded onto a pre-cleaned and conditioned 250 μl Sr spec columns with Eichrom Sr spec resin. A total of 2 ml of 3M HNO_3 was used to elute the sample matrix. Strontium was subsequently collected in 5 ml 0.03M HNO_3 and evaporated at 70°C to dryness. All chemical preparation steps were carried out under a laminar flow hood. In preparation for analyses, samples were loaded onto single W filaments with 6M HCl and TaCl_5 activator. Analyses were performed on a Thermo Scientific TRITON in the MAGIC laboratories at Imperial College London using static collection mode and amplifier rotation after each of 10 blocks of 20 cycles. Mass fractionation was corrected using an exponential law and a $^{86}\text{Sr}/^{88}\text{Sr}$ ratio of 0.7219, and Rb interferences were corrected for using $^{87}\text{Rb}/^{85}\text{Rb} = 0.386$. A ^{88}Sr intensity of $3.5 \text{ V} \pm 10\%$ was maintained throughout the measurement. Repeated analyses of NIST standard SRM987 during the course of this study yielded a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.710255 ± 0.000014 (2σ standard deviation; $n = 18$).

Fluid inclusions were studied in doubly polished wafers on a Linkam THMSG600 heating-cooling stage. Calibration of the stage was performed by measuring phase

changes in synthetic fluid inclusions of known composition. Reproducibility of the final melting temperature of ice (T_m) is within 0.2°C and of the homogenization temperature (T_h) within 2°C . The fluid inclusion results presented in this paper are based on measurements from multiple fluid inclusion assemblages among 5 different dolomite samples and 1 calcite sample. Although T_h measurements from inclusions within single fluid inclusion assemblages vary generally $<5^{\circ}\text{C}$, the range is much larger among different assemblages and among samples and the range is considered as more representative for the studied dolomite or calcite phase (thus not limited to a micro-scale zonation of that phase).

RESULTS

Host rock limestone and stratabound dolostone texture

The Jurassic host rock in the studied outcrop consists of alternating gray limestone and brown stratabound dolostone beds of 20 to 50 cm (0.7 to 1.6 ft) thick that we traced for about 300 meters (984 ft) across the outcrop (Fig. 2A). One of the stratabound dolomite beds underlies a sandstone bed. The limestone texture is difficult to recognize due to microsparitization, i.e. micrite transformed to coarser crystals, and varies from mudstone to wackestone and packstone. Relicts of bioclasts are marked by cleavage twinned calcite zones commonly without clear original fossil structure (Fig. 3A). Wackestones and packstones contain shell fragments, crinoids, gastropods, benthic foraminifera or sponge fragments. Some packstones consist mainly of well-rounded peloids (Fig. 3B). There are cm-scale nodules (Fig. 4A), filled with coarse, cleavage-twinned, calcite (Fig. 3C) in some limestone beds. In a few cases, dolomite or quartz is present among the calcite in these nodules and small dolomite rhombs and stylolitic seams occur at the nodule rims (Fig. 3C).

The limestone beds can contain scattered dolomite rhombs, i.e. planar-p dolomite (Fig. 3D) sensu the classification of Sibley and Gregg (1987) used throughout this paper. Small dolomite rhombs (30 to 200 μm in size) can occur aligned along bedding. Rhombs are cut by bedding parallel stylolites (Fig. 3D). In some samples, small quartz crystals are present. The limestone contains stylolitic seams and stylolites (Fig. 3D) and is crosscut by thin veins (mostly $< 1\text{ cm}$) filled with red dolomite or white calcite and by thicker veins ($> 1\text{ cm}$) filled with both minerals.

The brown stratabound dolostone beds have a planar-s to planar-e texture with crystals generally 50 to 160 μm in diameter that are referred to as 'D1' (Fig. 3E). These planar dolostones are marked by a mosaic extinction pattern under crossed polarized light. The CL characteristics of these dolostones are a dull red luminescence for the main (cloudy) part of the dolomite rhombs, while some inclusion-poor rims and thin veins are much darker red luminescent (Fig. 3F).

The stratabound dolostones contain bedding parallel stylolites and are crosscut by red coarse (partially calcitized) dolomite veins. In contrast to the limestone beds, some brown dolostone beds contain quartz-calcite veins with calcite preferentially filling the center part of the vein (Fig. 4B) or sigmoidal or nodular quartz cement patches (Fig. 4C). In general, the quartz-calcite veins (up to a few centimeters thick) are thicker than the calcite veins ($< 1\text{ cm}$) and most dolomite veins. The stratabound dolostone beds commonly overlie (laminated) mudstone and underlie limestone containing calcite nodules or, occasionally, sandstone.

Structures and fault-related dolomite texture

Several discordant reddish brown dolomite geobodies crosscut the limestone and stratabound dolostone host rock. These red dolomite geobodies are a few meters thick, showing a "Christmas tree" pattern (Fig. 5), and follow a large, WNW trending normal fault which dips 50S (or subvertical WNW fault after bedding tilt correction)

over a hundred or more meters. The majority of the fractures in the two fracture analysis windows have a similar strike and dip as the main fault (Appendices 2-3). Some of the red dolomite and limestone host rock layers have been displaced by reverse movement on the large fault (Fig. 2B). Two types of breccia occur within the fault zone, i.e. red dolomite breccia with brown dolomite clasts (Fig. 4D) and a calcite cemented breccia with red and brown dolomite and gray limestone clasts (Fig. 4E). The contacts between red dolomite and limestone host rock are abrupt (Fig. 4F). The extent of the fault-related (red) dolomite away from the fault is often larger in limestone than in brown stratabound dolostone host rock (Fig. 4F). Instead, the brown dolostone beds are cross-cut by small red dolomite veins (Fig. 4F). The red fault-related dolomite contains zebra dolomite textures and is altered by dedolomitization (Vandeginste and John, 2012), causing the reddish colour, whereas the brown stratabound dolostone has a fine crystalline texture and is crosscut by red dolomite veins.

The reddish brown fault-related dolomite geobodies generally have a coarser fabric than the stratabound dolostones and are ferroan (based on the blue color upon staining). The original texture of the fault-related dolomite is commonly obliterated because of surface weathering (Vandeginste and John, 2012). The fault-related dolomite is referred to as 'D2'. Less weathered samples commonly display an alternation of gray finer and white coarser dolomite bands, typical of zebra dolomites (e.g. Emmons *et al.*, 1927). The finer gray dolomite consists of subhedral crystals of about 150 to 400 μm diameter, whereas the coarser white dolomite is characterized by both subhedral and anhedral crystals of 500 to 2000 μm in diameter that generally display a sweeping extinction under crossed polarized light (Fig. 6 C). Some crystals of the gray dolomite within zebra textures have a dark core with dull red luminescence in contrast to the non-luminescent crystal rims, and other crystals that are completely non-luminescent (Fig. 6 D). In the center of white zebra dolomite bands, there is evidence of former vugs, now filled with coarse calcite. The white

dolomite crystals lining these calcite-cemented former pore spaces are commonly non-luminescent saddle dolomites with alternating inclusion-rich and inclusion-poor zones (Fig. 6 E, F). There are bedding parallel stylolites that crosscut the finer gray as well as the coarser white dolomite.

Carbonate Geochemistry

The main carbonate phases used for geochemical analysis include limestone, brown stratabound dolomite D1, fault-related dolomite D2 and calcitized D2 (D2 that transformed from dolomite into calcite by the weathering process of dedolomitization, Vandeginste and John, 2012). Major and minor elements were determined on bulk samples of these phases, whereas stable isotopes were measured on microdrilled samples, i.e. on more detailed phases of D2, i.e. gray and white zones of D2 dolomite, red altered D2 (dolomite affected by weathering but not transformed into calcite, see Vandeginste and John, 2012) and calcite cement, in addition to limestone, D1 and calcitized D2.

Dolomite crystal stoichiometry

The stratabound dolomite has a stoichiometric composition (50.4 to 50.6 mole% CaCO_3). The fault-related dolomite is a ferroan dolomite (ankerite) and a stoichiometric composition is calculated for D2 (48.5 to 51.1 mole% CaCO_3).

Major and Minor Elements

The limestone has a much lower Mg (0.1-2.3%) and Fe content (36-1829 ppm) than the dolomite (7.9-11.7% Mg; 0.4-9.4% Fe). The decreasing Mg trend fits the higher Fe values (Fig. 7A). Dolomite D1 lies at the high Mg and low Fe end (11.7-11.8% Mg; 0.29-0.56% Fe) of this trend, whereas D2 has higher Fe (3.2-9.4%) and lower Mg values (7.9-10.7%). The Fe and Mn (0.28-0.82%) values co-vary in D2. Limestone

samples show higher Sr concentrations (130-268 ppm) than dolomites D1 and D2 (42-129 ppm Sr; Fig. 7B). In detail, the average Sr concentration is higher in D1 (129 ppm) than in D2 (69 ppm). The Al concentrations roughly co-vary with K concentrations whereby K (60-881 ppm) is about 2.5 times higher than Al (14-419 ppm).

Stable Carbon and Oxygen Isotopes

Dolomite D1 in stratabound dolostones shows a less depleted $\delta^{18}\text{O}$ composition (-4.8 to -2.9‰) than the gray and white dolomite D2 in the fault-related dolomite bodies (-10.4 to -4.6‰ and -10.3 to -6.3‰, respectively). Both dolomites, however, share a similar $\delta^{13}\text{C}$ composition (+0.1 to +3.2‰ for D1; +1.0 to +2.7‰ for gray D2, and +0.6 to +2.2‰ for white D2; Fig. 8A). The stable isotopic composition of the limestones, in contrast, is less well defined and ranges from -10.4 to -3.9‰ in $\delta^{18}\text{O}$ and from -2.2 to +3.3‰ in $\delta^{13}\text{C}$ (Fig. 8B). Most calcite cements in the vugs and in veins have a similar stable isotope composition to that found for the white D2 dolomite. However, a few of these samples reveal strongly depleted $\delta^{18}\text{O}$ values (-17.7 to -16.0‰), coupled with relatively depleted $\delta^{13}\text{C}$ signatures (-6.5 to -1.5‰); these cements were sampled from pore space between fault-related clasts of dolomite in the breccia along the fault zone. Finally, the stable isotopic signature of the weathering products, i.e. red altered D2 and calcitized D2 are similar to other altered and calcitized D2 samples presented in Vandeginste and John (2012) with $\delta^{18}\text{O}$ values of -7.2 to -3.2‰ for altered D2 and -8.2 to -5.2‰ for calcitized D2, and $\delta^{13}\text{C}$ value of +0.8 to +2.7‰ for altered D2 and -6.1 to -4.1‰ for calcitized D2 (Fig. 8B).

Strontium Isotopes

The lowest $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios were measured in the limestones and stratabound dolostone (D1) samples. Limestone $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary from 0.7075 to 0.7083 (one

outlier of 0.7105), while the stratabound dolostones show a wider range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from 0.7074 to 0.7098 (one outlier of 0.7137). In the fault-related dolostone, D2, the gray dolomite (zebra bands or more massive zones) yielded $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that fall mainly between 0.7142 and 0.7166, with the exception of two samples (out of 8 total) that yielded lower ratios of 0.7125 and 0.7127. The white dolomite (zebra bands or veins), on the other hand, define the radiogenic end of the $^{87}\text{Sr}/^{86}\text{Sr}$ range, with ratios between 0.7146 and 0.7195. Two samples of weathered reddish dolomite are characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7103 and 0.7107.

Combining the Sr isotope data with the $\delta^{18}\text{O}$ data (Fig. 9) depicts the different sample groups very well (Fig. 9). Stratabound dolomites are characterized by relatively enriched (high) oxygen isotopic compositions and relatively unradiogenic (low) strontium isotopic compositions. A broad negative correlation can be observed when combining the stratabound dolomite results with the results from fault-related dolomites. In detail, the weathered fault-related dolomites ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7103$ to 0.7107; $\delta^{18}\text{O} = -3.7$ to -3.2‰) reveal isotopic characteristics similar to the stratabound dolomite signature ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7074$ to 0.7098; $\delta^{18}\text{O} = -3.4$ to -1.3‰). The gray and white fault-related dolomites are characterized by higher, more radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and lower, more depleted $\delta^{18}\text{O}$ values ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7142$ to 0.7195; $\delta^{18}\text{O} = -10.3$ to -5.4‰). The limestone samples are not consistent with the negative correlation described by the other diagenetic phases. They reveal relatively uniform $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7075 to 0.7083) combined with a $\sim 3\text{‰}$ spread in $\delta^{18}\text{O}$ (-8.0 to -4.7‰).

Fluid Inclusion Characteristics

The coarse, fault-related, dolomite (D2) contains mainly two-phase fluid inclusions, whereas the pore-filling calcite in the dolostone contains quite abundant single-phase

fluid inclusions. No clear growth zones of fluid inclusions could be observed, and hence the main selection criteria was to avoid fluid inclusions in secondary trails. Fluid inclusions in the dolomite samples are typically about 2 x 1 µm in diameter, while they are commonly 3 x 2 µm and up to 6 x 10 µm in diameter in the calcite samples. First melting is observed at about -52°C, indicating an aqueous CaCl₂-NaCl system. Final melting of ice (T_m) occurs at around -24°C in both mineral phases (Fig. 10A). While T_m values for most inclusions show a normal distribution clustering around -24°C (Fig. 10A), ~9% of the inclusions show significantly less negative values (-8 to 0°C; not included on Fig. 13). These inclusions with T_m approaching 0°C fall at the rim of the calcite crystal and this rim is marked by a CL pattern of alternating non-luminescent and bright yellowish luminescent zones. The distribution of homogenization temperature (T_h) is more variable than the melting temperature. Overall, pore-filling calcite samples yield temperatures from 110 to 210°C, while dolomites are characterized by higher values of 150 to 240°C for dolomite (Fig. 10B). Besides the abundant type of fluid inclusions described above, three inclusions were found to contain CO₂ in addition to CaCl₂-NaCl-H₂O and they have a clathrate dissociation temperature of 6.1 to 11.3°C.

DISCUSSION

Nature and timing of fracturing

Based on field observations, several macro-scale diagenetic phases can be reconstructed and placed in a paragenetic sequence (Fig. 11). Crosscutting relationships indicate that stratabound dolomite predates fault-related dolomite, in turn predating calcite cement present in the pore space on top of coarse fault-related dolomite crystals. The occurrence of two different breccia types testify to two brecciation events, related to the activity of the main fault. No crosscutting

relationship was observed between these breccias, but their relative chronology is derived from the composition of the breccia clasts and of the cementing matrix. The first brecciation event, which resulted in the red dolomite breccia with brown dolomite clasts, is linked to the main fault activity and associated fault-related dolomitization. Based on its high dip angle relative to bedding, the fault probably originated as a normal fault. The first brecciation event occurred after the formation of the brown stratabound dolomite. The second brecciation event, which caused the calcite-cemented breccia with red and brown dolomite and gray limestone clasts, is thought to be related to the reactivation of the main fault as a reverse fault (see deflection of the bedding in Fig. 2). Several phases of fracturing and cementation took place. A series of quartz-calcite veins that are only present in the stratabound dolostone are bed-confined and most likely predate the main, fault-related, dolomitization event. Based on their similar texture and structural orientation, several red dolomite veins present in the limestone are thought to be closely related to the main fault-related dolomitization event. White calcite cement filled the fractures later than the red dolomite, but the time gap between them is not clear.

Timing of Stratabound and Fault-related Dolomitization Processes

Since D1 is cut by bedding parallel stylolites, D1 is interpreted to have developed before stylolitization (Fig. 11). The stratabound dolomite bodies are also crosscut by fault-related dolomite bodies, and thus D1 predates D2. In addition, a low temperature origin can be inferred for D1 based on the relatively high $\delta^{18}\text{O}$ (see also discussion below). All these arguments suggest D1 has an early diagenetic origin, i.e. near surface or shallow burial.

Fault-related dolomitization forming D2 is linked to normal fault activity as evidenced by the specific occurrence of D2 along the fault and by the presence of host rock clasts in a breccia cemented by D2. Based on the presence of saddle

dolomite which indicates a formation temperature above 80°C (Machel, 2004), the chemical signature (ferroan), the high fluid inclusion homogenization temperature (of about 200°C), and its relative timing with respect to D1 in the limestone and stratabound dolostone, D2 must have a deep burial origin. High temperatures are reached in the Jurassic rocks in the study area only during the Upper Cretaceous tectonic events (Fig. 12) involving intra-oceanic subduction from the Cenomanian to early Campanian with southwest-directed obduction of the Semail ophiolite (Breton *et al.*, 2004) and possibly during subsequent onset of uplift. Most likely, the dolomitized normal fault developed as a result of extension in the subducting slab during the Upper Cretaceous. This is consistent with the WNW-ESE strike of the fault. Normal fault systems develop commonly in outer trench slopes of collisional orogens as a result of flexural bending, and thus, faulting in the upper part of the lithosphere (Chapple and Forsyth, 1979). Normal faulting in similar environment could also be caused by whole-lithosphere extension due to slab pull, though (Schoonmaker *et al.*, 2005). Another hypothesis links the normal faulting to the Maastrichtian post-obduction ENE-WSW extension or Oligocene NNE and NW extension (Fournier *et al.*, 2006). Although NNE extension could have caused the normal fault studied (see also discussion below), an Oligocene age is not consistent with the observed geochemical (mainly Sr isotopes) and fluid inclusion data discussed below showing that dolomitization occurred before deepest burial. The observed inversion of the normal fault is related either to the Late Cretaceous or to Eocene to Miocene episodes of shortening, after which the overall plate convergence was transferred to the Makran subduction system, leaving the Oman Range and its foreland tectonically quiet (Chemenda *et al.*, 1996; Tarapoanca *et al.*, 2010).

Diagenetic Fluid Evolution

Stratabound dolomite

The fine-crystalline D1 is petrographically and geochemically different to the fault-related D2. Although different geochemical signatures in the D1 and D2 could be linked to different degrees of fluid-rock interaction rather than to different geochemical signatures of the D1 and D2 dolomitizing fluids, several arguments allow us to discard this possibility. First, D2 crosscuts D1, suggesting two different dolomitization events. Second, if both D1 and D2 formed by the same geochemical fluid but by different fluid/rock ratios, then the fluid/rock weight ratio for D1 should be lower than about 10 based on the $^{87}\text{Sr}/^{86}\text{Sr}$ signature and lower than 0.1 based on the $\delta^{18}\text{O}$ signature (using equations from Banner and Hanson, 1990). However, to dolomitize limestone (mole-per-mole replacement), a fluid/rock ratio of 42 is needed for (halite-saturated) evaporated seawater (based on Carpenter, 1978). Moreover, most basinal brines have an Mg deficit (Lowenstein and Timofeeff, 2008), suggesting even higher fluid/rock ratios are needed. These arguments demonstrate that the geochemical differences observed in D1 and D2 must relate to a different geochemical signature of the D1 and D2 dolomitizing fluids.

The relatively low Fe and Mn contents in D1 indicate dolomitizing fluids that were depleted in Fe and Mn compared to those that formed D2 (see below). The slight ^{87}Sr enrichment in D1 compared to Jurassic seawater (0.7068-0.7077; Veizer *et al.*, 1999; Fig. 9) could be linked to the occurrence of quartz nodular patches and quartz-calcite veins in stratabound dolostone. The bed-confined nature of these quartz-bearing veins could suggest cementation of a fracture network confined to the early phase of dolostone (D1) and that predates fault-related dolomitization, based on the fact that dolostone is more prone to fracturing than limestone (Sinclair, 1980). Alternatively, a local quartz source may have triggered quartz cementation in fractures within stratabound dolostone beds, supported the occurrence of a sandstone layer overlying one of the main stratabound dolostone beds. In addition, quartz might have been present as nodules in the stratabound dolostone beds (possibly by filling of a few bioturbations, based on the bioturbated nature recognized in the underlying

limestone bed) before the formation of the quartz-bearing veins. Despite the uncertainty on the formation of the quartz-bearing patches and veins, the slight ^{87}Sr enrichment in the stratabound dolostone indicates interaction of the dolomitizing (or possibly dolomite recrystallizing) fluids with siliciclastics.

Using the oxygen isotopic composition of D1, and assuming a formation temperature between 30 and 50°C (based on near-surface or shallow burial origin in an arid climate), the derived dolomitizing fluid $\delta^{18}\text{O}$ is about -2‰ (based on the fractionation equation of Land, 1985; Fig. 13). This signature indicates seawater or evaporated seawater, assuming a Jurassic age (based on Veizer *et al.*, 1999).

Fault-related dolomite

High Fe and Mn concentrations in D2 indicate dolomitization by deep burial fluids that are highly enriched in Fe and Mn, since these elements are favoured by reducing conditions in burial fluids (Warren, 2000). In addition, the dolomitizing fluids must have obtained Fe and Mn from a non-carbonate source, i.e. siliciclastics or crystalline basement. Given the fault-related nature of D2 dolomite, the origin of the normal faulting, and the relatively radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (up to 0.720; Fig. 9), it is likely that the dolomitizing fluids interacted with Precambrian sandstone or granodioritic basement rocks (Allen and Leather, 2006). These fluids were subsequently expelled upwards along the normal fault during fault activity. The strongly enriched $^{87}\text{Sr}/^{86}\text{Sr}$ signature is inconsistent with models proposing that the dolomitizing fluids originated from interaction with the Semail ophiolite (Coy, 1997), which has much lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (i.e. ~0.705; Kawahata *et al.*, 2001). The inverse correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$ in D2 and the fact that this trend points towards the signature of marine limestone at the lower end (Fig. 9) indicate a partial relict signature of the original limestone, or an inhomogeneous interaction or partial buffering of the dolomitizing fluid with the host rock. In addition, based on the recrystallized limestone

trend showing depleted $\delta^{18}\text{O}$ values and the fact that the D2 trend correlates with the original Jurassic limestone signature and not with the recrystallized limestone signature, we suggest that D2 dolomitization took place before deep burial when high temperature recrystallization of the limestone took place. This is consistent with the timing proposed above, i.e. Late Cretaceous.

Considering the homogenization temperature of fluid inclusions as a minimum estimate of the formation temperature of D2, and combining this with the measured $\delta^{18}\text{O}$ of the dolomite, the $\delta^{18}\text{O}$ of the dolomitizing fluid is estimated at about +7‰ VSMOW (based on the equation by Land, 1985; Fig. 13). Taking into account the high salinity of the D2 dolomitizing fluids (about 24 wt% NaCl eq. based on fluid inclusion data), its strongly positive $\delta^{18}\text{O}$ is probably indicative of a highly evaporated brine source. Thus, a deep basinal brine was probably responsible for the formation of D2.

Alternatively, high $\delta^{18}\text{O}$ values were found in saddle dolomite which were influenced by fault-focused circulation of magmatic fluids in the Devonian West Point Formation in Gaspé Peninsula in Quebec (Lavoie *et al.*, 2010). Magmatic fluids are typically characterized by a $\delta^{18}\text{O}$ of +6 to +8‰ VSMOW (Hoefs, 2009). There is obducted ophiolite in the study area (Nicolas *et al.*, 2008), but no magmatic intrusions that could explain the stable oxygen isotopic signature of the dolomite studied. As mentioned above, the high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in D2 dolomites are inconsistent with an ophiolite-derived fluid source (Bosch *et al.*, 2004). Furthermore, magmatic fluids are often accompanied by significant degassing, producing CO_2 with depleted $\delta^{13}\text{C}$ values ($\sim -6\text{‰}$). D2 dolomites however show no evidence of such carbon isotopic signatures ($\delta^{13}\text{C}$: mostly +1 to +2.5‰), and only a few fluid inclusions contained minor amounts of CO_2 . Because of the rare occurrence of this type of fluid inclusions in the samples, these particular fluid inclusions are interpreted to be either secondary

or to have lost their primary fluid upon opening and subsequent refilling with a later diagenetic fluid, which is suggested to relate to the deep burial metamorphic process.

Pore-filling calcite

Fluid inclusion data collected from calcite cement filling pore space around D2 saddle dolomite crystals indicate high formation temperatures (~ 160°C) and high salinity (23 wt% NaCl eq.) of the $\text{CaCl}_2\text{-NaCl-H}_2\text{O}$ fluids. We hypothesize that the few inclusions that reveal low salinity data result from leakage and refill by fluids of later diagenetic events. Evidence demonstrates the presence of a CL-zoned calcite rim of pore-filling cement; this CL zonation is typical of precipitation from meteoric fluids and is consistent with the non-saline fluid inclusion measurements.

Taking the $\delta^{18}\text{O}$ of the pore-filling calcite (about -7‰ VPDB) and a formation temperature of about 160°C, the $\delta^{18}\text{O}$ of the original fluid is estimated at +12‰ VSMOW (based on the equation by Friedman and O'Neil, 1977). This value is higher than that the one calculated for the D2 forming fluids and is probably caused by either buffering of the $\delta^{18}\text{O}$ signature or an overestimate of the formation temperature. Buffering by the surrounding dolomite is the more likely explanation for the pore-filling calcite, since the main fracturing and thus vein-filling phase is related to the main fault activity and thus deep burial, high temperature Late Cretaceous tectonics. Based on the lower Th of the pore-filling calcite compared to that of D2, it is possible that both mineral phases are formed by the same fluid but at different temperatures. A temperature drop (e.g. from 200 to 180°C considering Ca/Mg activity ratio of 14) can result in a change from dolomite to calcite saturation in the fluid (Carpenter, 1980).

Breccia-cementing calcite

The most depleted stable isotope compositions ($\delta^{13}\text{C}$ of -6‰ and $\delta^{18}\text{O}$ of -16‰; Fig. 8) belong to calcite that cements the second breccia type. The depleted isotopic compositions for O and C suggests either involvement of meteoric water at about 60°C, or magmatic fluid at about 250°C (based on the depleted $\delta^{13}\text{C}$ signature, consistent with high fluid/rock ratio, the typical $\delta^{18}\text{O}$ signature of magmatic fluid or meteoric water in the region, Burns *et al.*, 1998, and the equation of Friedman and O'Neil, 1977). We favor the involvement of meteoric fluids, as no clear evidence is found for the presence of a magmatic source. In addition, the estimated temperature of 250°C would be untypically low for magmatic fluids.

Conceptual model for the genesis of stratabound dolomite geobodies at Wadi Mistal

Arguments presented above suggest a near-surface or shallow burial origin for the stratabound dolomite geobodies, with the involvement of seawater or evaporated seawater. Previous models for near-surface or shallow burial dolomitization by seawater were proposed for carbonate islands (e.g. Budd, 1997; Vahrenkamp and Swart, 1994), but cannot explain the stratabound dolomite in this study, i.e. on a shallow-water epi-continental ramp.

An evaporated seawater origin is more likely to explain the formation of D1. The stratabound dolostone beds commonly overlie (laminated) mudstone and underlie limestone containing calcite nodules or sandstone. The interpreted depositional facies is a restricted lagoonal environment, and the calcite nodules could represent replaced evaporite nodules. This interpretation is consistent with the interpreted depositional facies of similar Jurassic beds in the Musandam Peninsula (de Matos, 1997). The lack of conclusive evidence for the presence of evaporites in the study area may suggest that the dolomitizing fluids were mesosaline (Melim and Scholle, 2002). Most likely, Jurassic mesosaline (or hypersaline) fluids caused stratabound

dolomitization in a process of seepage reflux. Several Jurassic stratabound dolostone beds have also been reported in the Musandam Peninsula and in the subsurface of the U.A.E. and were interpreted as supratidal fine crystalline dolomite often associated with algal mats or as shallow burial dolomite (de Matos, 1997). In addition, Late Jurassic formations (with the Arab-D dolomitized strata) are important oil reservoirs in the Middle East; several episodes of dolomitization affected these rocks, including seawater, reflux and burial dolomitization (Swart *et al.*, 2005).

Conceptual model for the genesis of fault-related dolomite geobodies at Wadi Mistal

The fault-related dolomite in our study is most likely hydrothermal, i.e. at least 5 to 10°C warmer than the surrounding rocks (Machel and Lonnee, 2002), based on 1) the high trapping temperatures (average of 220°C +/- 20°C) derived from fluid inclusions (after applying a minimum pressure correction for 4 km depth, Zhang and Frantz, 1987) and 2) the fact that the rocks underwent deep high temperature diagenesis, also referred to as deep anchizone metamorphism with temperature not exceeding 200°C during deepest burial in Early Campanian (Breton *et al.*, 2004). Nevertheless, both temperature estimates bear some uncertainty because of a) potential stretching of the fluid inclusions in dolomite (however no pronounced skewness of the Th distribution suggesting stretching was observed) and b) the lack of an accurate temperature measurement of deepest burial. Moreover, rapid upward fluid flow from greater depths through high-permeability faults and fractures is a likely mechanism to supply hydrothermal fluids (Deming, 1992), and is consistent with the association of dolomite geobodies with the fault and the absence of underlying igneous intrusions. The fault-controlled fluid flow is potentially linked to basement-rooted fault structures under Jebel Akhdar that are interpreted as normal faults (Al-Lazki *et al.*, 2002) or as north-dipping, high-angle, blind reverse faults (Mount *et al.*,

1998). The reverse faults have been interpreted as basement-involved compressional structures active during the Oligocene compressional deformation event (Mount *et al.*, 1998).

As argued above, fault-related dolomitization probably occurred during Late Cretaceous deep burial related to normal faulting in the subducting slab during the regional compressive regime (Fig. 14) and the fault probably underwent reverse reactivation during the Oligocene. Brecciation associated with normal faulting most likely aided fluid flow along the fault. High Fe and Mn contents, high $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, all are consistent with fluids that interacted with the crystalline basement, and thus with channeling along basement-rooted faults. A deep-seated basement fault control with elevated heat flow along long-lived crustal scale faults is also an important mechanism for hydrothermal dolomitization in Iran (Sharp *et al.*, 2010).

Hydrothermal dolomite often occurs in association with thrust and normal faults (Sharp *et al.*, 2010) as well as basement-rooted wrench faults (Smith, 2006). Such dolomitization could be linked to increased heat flow associated with salt domes (Beavington-Penney *et al.*, 2008) or to convective flow caused by igneous activity (Lavoie and Chi, 2010) whether or not combined with tectonically induced flow related to fault reactivation (Wilson *et al.*, 2007). While faults and fractures are essential fluid pathways (e.g. Eichhubl and Boles, 2000; Davies and Smith, 2006), the magnitude of the shear offset determines the highest permeability zone (for faults in compressional and strike-slip settings), i.e. within the fault zone for small-offset faults (<3 cm) due to the many joints and high-porosity breccia, but adjacent to the fault zone for large-offset faults (>10 m) due to a wide zone of low-porosity breccia which can form permeability barriers (Antonellini and Mollema, 2000). Similarly, growth of cataclastic fault cores from fractured damage zones in extensional and strike-slip fault zones show an evolution from high to low permeability (Billi *et al.*, 2003; Micarelli *et al.*, 2006). It has also been reported that opening-mode fracture

porosity in dolostones is controlled by both the aperture size and post-kinematic cements (Gale *et al.*, 2004).

Control on dolomite geobody dimensions, diagenetic geobody classification and impact on flow

The benefit of the process-based approach (investigating the diagenetic phases to gain a better understanding of the diagenetic processes) exemplified here at Wadi Mistal is that process (in this case dolomitization) impacts on distribution and shape of geobodies. The same approach can be applied in reservoir case studies. The stratabound geobodies are regularly shaped, less than 0.5 m (1.6 ft) thick but at least several hundreds of meters (~1000 ft) long and follow bedding laterally. In contrast, the fault-related dolomite geobodies are discordant and are at least a hundred meters (328 ft) long (along the fault plane) and up to a few meters wide. The thickness is more irregular (than the stratabound dolomite geobodies) since the extent of the fault-related dolomite body away from the fault varies along the fault plane. The link between the host rock texture and the extent of the fault-related dolomite away from the fault into the host rock is not distinct. The fault-related dolomite extent in a peloidal grainstone bed (0.7 m [2.3 ft] wide) is smaller than or similar to that in the bioclastic wackestones and packstones (0.5 to 3.5 m [1.6 to 11.5 ft] wide). This may be consistent with flow simulations showing that grain-dominated packstones are preferentially dolomitized over lower reactive surface area grainstones in interlayered grainstone-packstone units (Al-Helal *et al.*, 2012). However, the variations in dolomite extent are small and not consistent, and one of the long dolomite extensions crosscuts the bedding at a low angle. The latter might be the result of fractures, but also other lateral textural heterogeneities may have caused diverging fluid flow patterns. The lack of clear correlation between dolomite extent and host rock lithology is most likely caused by the strongly compacted nature

and diagenetic modifications that impacted the original host rock texture prior to dolomitization. Nevertheless, two features are significant. First, a laterally extent dolomite sliver tapers-up on an overlying mudstone containing siliciclastic material. In this case, the lower-permeability mudstone probably acted as a baffle to ascending dolomitizing fluid flow. Second, the fault-related dolomite extent is generally smaller in stratabound dolostone beds than in limestone beds.

The main control on the dimension of the stratabound and fault-related dolomite bodies is thus first and foremost the dolomitization process itself. However, the initial depositional texture also profoundly influences the stratabound dolostone geobodies; therefore, the depositional classification of geobodies of Jung and Aigner (2012) can be applied to this type of diagenetic geobodies (stratabound geobodies at Wadi Mistal would correspond in this classification to a Jurassic ramp system in the protected platform zone and with a “sheet” shape, see Table 1A).

By contrast, structural elements were the main control on the distribution of the fault-related dolomite bodies, which may be related to the late-diagenetic and deep-burial (> 4 km) timing, and thus strongly compacted, cemented, low-porosity nature of the host rock limestone. We propose that the Jung and Aigner (2012) classification can be modified to incorporate fracture-related geobodies by adding one new system to the three existing ones (“ramps”, “shelf” and “platform”): we call this system “structural” (Table 1B). Although it has been demonstrated that facies can control the density and distribution of fractures (Laubach et al., 2009), the resulting fractures tend to be interconnected and thus the “structural” system is not primarily constrained by facies or depositional environment, but overprints it. The orientation and extent of the fault needs to be constrained in the “structural” system, but all other elements of the classification can be maintained (for instance, timing would refer to the timing of dolomitization, and the zone would refer to the environment of deposition of the host rock). The extensive collection of “shape” terms from Jung and Aigner (2012) can to some extent apply to diagenetic geobodies if they are suitably generic (such as

“sheets”), but some additional terms are proposed (such as “oblongoids”, “lensoids” or “christmas tree”; Table 1B). The distribution of the dolomite geobodies along the fault can be classified as “aligned” in the Jung and Aigner’s (2012) terminology.

Ultimately, the benefit of understanding the distribution of depositional and diagenetic geobodies is to be able to predict the impact of these features on flow of reservoir fluids. The stratabound dolomite geobodies may have had a higher intercrystalline porosity than the replaced limestone, and thus could behave as a thin reservoir with important lateral extension, or a possible migration pathway if a structural dip exists. The fault-related dolomite bodies, on the other hand, have only little intercrystalline porosity (mainly in the center of the white zebra bands that was then cemented with calcite). It is thus likely that the fault-related dolomitization process resulted in a sealed fracture. Hence, in the Wadi Mistal case study dolomite geobodies distributed along the fault probably acted as a barrier to flow, resulting in possible compartmentalization of the reservoir. We expect that other dolomite bodies formed under similar conditions and by the same processes, i.e. deep-burial with normal fault related origin, could possibly have a similar effect on reservoir fluid flow in regional subsurface reservoirs. We stress, however, the importance of understanding individual processes for diagenetic geobodies, as it is equally likely that fracture-related dolomite geobodies formed under slightly different conditions would act as preferential pathways for fluids, or even as reservoirs (e.g. Smith, 2006).

SUMMARY AND CONCLUSIONS

The first part of our study, based on understanding the process of dolomite geobody formation at Wadi Mistal, has shown that:

744 - The main diagenetic phases in paragenetic order include: stratabound
745 dolomite, quartz-calcite veins, fault-related dolomite, pore-filling calcite, calcite
746 veins, breccia-cementing calcite, dedolomite.

747 - Stratabound dolomite is interpreted to have formed during early diagenesis by
748 mesosaline (or hypersaline) fluids in a process of seepage reflux based on
749 crosscutting relations with bedding parallel stylolites and fault-related
750 dolomite, the interpreted depositional facies and the geochemical
751 characteristics of D1.

752 - Fault-related dolomite is interpreted to have formed in deep burial and to be
753 controlled by normal fault activity along which fluids were focused based on
754 the structural context and the geochemical characteristics of D2. This
755 dolomitization probably happened in the mid Late Cretaceous when normal
756 faults developed due to extension in the subducting slab. The dolomitizing
757 fluid was most likely deep basinal brine that had interacted with the crystalline
758 basement.

759 Based on our new data we exclude the previously suggested model that the
760 dolomitizing fluids in northern Oman originated from interaction with the Semail
761 ophiolite. We interpret that dolomitizing fluids migrate along normal faults during
762 subduction in a convergent setting. This study hence proposes that there is focused
763 structurally-controlled ascending fluid flow in the subducting slab in addition to or as
764 part of tectonically-driven fluid flow towards the foreland and continental interior in
765 collision regimes. As such tectonic settings are also observed in other parts of the
766 world, structurally-controlled dolomite is expected to be found in many other regions.
767 Furthermore, we highlight that the control on dolomite body dimension is linked to the
768 dolomitization process itself, but depositional geobodies as well as structural
769 elements can act as templates for the distribution of diagenetic geobodies. This leads
770 us to propose a slightly modified classification of geobodies that accounts for both

depositional and diagenetic cases, and ultimately will help modeling of existing subsurface reservoirs.

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FIGURE AND TABLE CAPTIONS

Figure 1. Geological map of northern Oman (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 map. The bottom figure shows a detailed geological map of the Jebel Akhdar tectonic window (after Le Métour *et al.*, 1993). Location of the Wadi Mistal outcrop is indicated.

Figure 2. Field photographs. (A) The outcrop studied showing alternating gray limestone and brown stratabound dolostone beds and red discordant fault-related dolomite. Car and two persons as scale. The white lines labeled as FB1 and FB2 are the fracture analysis beds. (B) Detailed view of box indicated in panel A showing bending of limestone and fault-related dolomite beds by fault drag during reverse movement along reactivated fault.

Figure 3. Microphotographs of limestone and stratabound dolostone host rock samples. Top of the bedding is at the top of the microphotograph. Thin sections in B stained; the pink colour thus indicates a calcitic composition. (A) Recrystallized wackestone with fossil relicts characterized by calcitic cement zones. Note also the burial compaction layering and the irregular vein crosscutting the sample. (B) Peloidal packstone to grainstone with cleavage twinned calcite cement between particles. Note also the subvertical veins that crosscut the peloids and cleavage twinned calcite cement. (C) Recrystallized wackestone with large nodule filled with cleavage twinned coarse calcite. A subvertical vein (also with cleavage twinned calcite) stops on the stylolite at the rim of the nodule. Note the small dolomite rhombs at the rim of the nodule and along the stylolite. (D) Recrystallized wackestone with scattered small dolomite rhombs that are cut by bedding parallel stylolite. (E) Stratabound dolostone. (F) CL image of E.

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1099 **Figure 4.** Field photographs. (A) Calcite nodules in limestone bed. (B) Bedding-
1100 normal lensoid vein containing brownish altered quartz (mainly at vein rim) and white
1101 calcite (mainly in vein center) in stratabound dolostone bed. (C) Nodular patch of
1102 brownish altered quartz, crosscut by fractures, in stratabound dolostone bed. (D) Red
1103 dolomite breccia with brown dolomite clasts (Cl). (E) Calcite cemented breccia with
1104 red dolomite, brown dolomite and limestone clasts. (F) Brown stratabound dolostone
1105 (SBD) bed with red dolomite fractures; red dolomite (FRD) and limestone (L) above
1106 and below brown stratabound dolostone bed.

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1108 **Figure 5.** Dimension of studied fault-related dolomite body.

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1110 **Figure 6.** Microphotographs of fault-related dolomite samples. (A) Fault-related
1111 dolomite with relicts of D1 rhombs. (B) CL view of A. (C) Fault-related dolomite with
1112 saddle dolomite crystal and calcite filling pore space on top of coarse dolomite. (D)
1113 CL view of C.

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1115 **Figure 7.** Crossplots of (A) Mg versus Fe concentration and (B) Sr versus Mn
1116 concentration in limestone, dolomite D1, D2 and calcitized D2.

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1118 **Figure 8.** Stable carbon and oxygen crossplot. The gray box indicates the Jurassic
1119 marine seawater signature (Veizer *et al.*, 1999). (A) Dolomite D1 in stratabound
1120 dolostone beds and gray and white dolomite D2 in fault-related dolomite bodies. (B)
1121 Other phases sampled in the Jurassic samples, i.e. limestone, red altered dolomite
1122 D2, calcitized D2 and calcite cement in veins and vugs. The black box on this panel
1123 indicates the range of panel A.

1124

Figure 9. Crossplot presenting $^{87}\text{Sr}/^{86}\text{Sr}$ ratios versus $\delta^{18}\text{O}$ values for limestone, stratabound dolomite (D1), fault-related gray, white and reddish altered dolomite (D2). The gray box indicates the Jurassic marine seawater signature (Veizer *et al.*, 1999).

Figure 10. Histograms of (A) final melting temperature of ice and (B) homogenization temperature for two-phase aqueous fluid inclusions (excluding secondary low salinity inclusions) in dolomite and pore-filling calcite.

Figure 11. Sequence of paragenetic phases.

Figure 12. Reconstructed burial curve for the Jurassic outcrop in Wadi Mistal, based on Le Nindre *et al.* (2003) for Jurassic till earliest Late Cretaceous time; this part of the curve must be read on the depth scale. Uplift history is debated; the two main views are presented based on fission track data interpretations from Mount *et al.* (1998) and Saddiqi *et al.* (2006); these curves must be read on the temperature scale. The deepest burial during Late Cretaceous is estimated from Breton *et al.* (2004), but the general Late Cretaceous to early Cenozoic is uncertain and is indicated by a dashed line (two hypotheses related to the interpretations of Mount *et al.*, 1998 and Saddiqi *et al.*, 2006).

Figure 13. Graphical representation of the oxygen isotopic equilibrium between dolomite, $\delta^{18}\text{O}_{\text{water}}$ (on SMOW scale) and temperature (based on Land, 1985). The $\delta^{18}\text{O}_{\text{dolomite}}$ (on VPDB scale) is presented versus homogenization temperature (average $\pm 1\sigma$) for four D2 dolomite samples. For the presentation of D1, an estimated formation temperature is used.

Figure 14. Evolution sketch for Early Jurassic to Campanian time. The transect is parallel to the main compression direction (phase 3 and 4 modified from transects from Breton *et al.*, 2004). Note the different vertical scale for phase 1 and 2 versus phase 3 and 4. Stratabound and fault-related dolomitization is also illustrated at outcrop scale in the panel on the right.

Table 1. A) Existing classification of depositional geobodies of Jung and Aigner (2012). **B)** Proposed additions to the Jung and Aigner (2012) classification scheme to include diagenetic geobodies.

APPENDIX

Appendix 1. Histogram of fracture density (separated for fracture width ranges) in function of the distance from the main large fault. Histogram intervals are 1 m (3.3 ft). (A) The fracture analysis bed at the top of the outcrop in gray limestone host rock and some red dolomite. (B) The fracture analysis bed in the middle of the outcrop in the brown stratabound dolostone bed with some red dolomite close to the main fault.

Appendix 2. Lower-hemisphere equal-area projection of poles of uncorrected (left) and bedding tilt corrected (right) values of (A) 82 fractures measured in the fracture analysis bed at the top of the outcrop with red fault-related dolomite and gray limestone host rock and (B) 106 fractures measured in the fracture analysis bed in the middle of the outcrop with red fault-related dolomite and brown stratabound dolostone host rock.

Appendix 3. Lower-hemisphere equal-area projection of fractures (uncorrected for bedding tilt) represented as great circles. Different types of circles are used depending on the type of fracture filling. Data are presented over intervals of distance

1180 away from the main large fault, in the direction of the arrow; data were thus collected
1181 in the main fault zone and northward away from the main fault. (A) The fracture
1182 analysis bed at the top of the outcrop in red dolomite and gray limestone. Data are
1183 presented in intervals of 5 m (16.4 ft). (B) The fracture analysis bed at the middle of
1184 the outcrop in red dolomite and brown stratabound dolostone. Data are presented in
1185 intervals of 3 m (9.8 ft).
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Figure 1

Figure 2

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Figure 14

Key

—— fracture filled with brown dolomite or dedolomite
 ——— fracture filled with white calcite

— — - fracture with white calcite and brown dolomite or dedolomite
 - - - - - fracture without filling

— — fracture filled with calcite and quartz
 N total number of data per plot