Constraints on the diagenesis, stratigraphy and internal dynamics of the surface-piercing salt domes in the Ghaba Salt Basin (Oman): A comparison to the Ara formation in the South Oman Salt Basin

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ABSTRACT

In the South Oman Salt Basin the Ara carbonates form an extensively cored, deeply buried ‘intra-salt’ hydrocarbon play. Six surface-piercing salt domes in the Ghaba Salt Basin (North Oman) provide the only outcrop equivalents for carbonates and evaporites of the infra-Cambrian Ara Group (uppermost Huqf Supergroup). Based on data from fieldwork, satellite images and isotope analysis it is concluded that most of the carbonate bodies in the Ghaba salt domes are time equivalent to the stratigraphically uppermost stringer intervals in the South Oman Salt Basin (A5-A6). Maturity analyses demonstrate that the carbonate “stringers” in the salt domes were transported with the rising Ara salt from a depth between ~ 6 to 10 km to the surface. Petrographic and stable isotope data show that their diagenetic evolution during shallow and deep burial was very similar to the Ara carbonate ‘stringer’ play in the SOSB. However, during the retrograde pathway of salt diapir evolution, the carbonate inclusions were exposed to strong deformation in the diaper stem and diagenetic alterations related to dedolomitisation. As the salt domes contain facies that are in all aspects identical to the deeply buried Ara play in the South Oman Salt Basin, this study provides substantial additional information for hydrocarbon
exploration in South Oman. In addition, our work has implications for the hydrocarbon prospectivity of the Ghaba Salt Basin and possibly of other ‘infra-Cambrian’ evaporite basins in the Middle East such as for the time-equivalent ‘Hormuz’ salt basins.

1 INTRODUCTION

The study of large rock inclusions (so-called ‘rafts’, ‘floaters’ or ‘stringers’) in salt diapirs is of broad economic interest as they constitute profitable hydrocarbon reservoirs but may also represent potential drilling risks due to their high fluid overpressures. The uplift mechanisms and associated controls on the internal deformation of a salt diapir has been subject of controversial debates in the past 20 years (Talbot and Jackson, 1987; Kupfer, 1989; Talbot and Jackson, 1989; Gansser, 1992; Talbot and Weinberg, 1992; Gansser, 1993; Weinberg, 1993; Koyi, 2001; Callot et al., 2006).

The most extensively studied stringers are the ones exposed in the Iranian salt domes. They are up to 6 km² in area and ~80 m thick, comprising volcanic and igneous rocks or sediments such as anhydrite and carbonates. These stringers are commonly interpreted as interlayered with the primary rock salt lifted from up to 10 km with the rising salt diapir to the surface (Kent, 1979; Gansser, 1992; Talbot and Jackson, 1987). Talbot and Weinberg (1992) argued that the concentric shapes of many salt domes observable in the Kavir area (central Iran) in air photographs (Jackson et al., 1990) are a result of circulation within the diapir. By this process, the initial internal stratigraphy develops into a mushroom-shape, resulting in gentle dips of the intra-salt inclusions for most portions of the upper level of a diapir.

Results from analogue modelling have shown that brittle layers passively follow the vertical ascent of ductile layers from the earliest stages until cessation of diapir evolution (Escher and Kuenen, 1929; Koyi, 2001; Callot et al., 2006). During this evolution, the embedded inclusions underwent stretching, leading to boudinage and rotation against the vertical boundaries of the diapir. As the diapir growth and salt supply stops, the inclusions start to descend within the diapir, which initiates an internal (secondary) flow within the salt diapirs. During this process, the inclusions undergo folding and create shear zones at the immediate contact with the salt. From these insights, one could expect that the structural configuration of inclusions at the surface is strongly influenced by the
internal kinematics (flow patterns) of a salt diapir. However, detailed mapping of mining
galleries in structural shallow levels of various salt diapirs revealed highly complex
arrangements of isoclinal and overturned folds with all geometrically possible shapes and
orientations (see Talbot and Jackson, 1987 and references therein). Besides the
complexity of the internal structural geology developing during the rise of (natural)
diapirs, strong dissolution near and at the surface leads to a structural reconfiguration of
the inclusions (Talbot and Jackson, 1987; Weinberg, 1993), which masks interpretations
on the style of salt tectonics.

Most of the insights on the structural evolution of salt-encased inclusions during diapir
rise emerged from the interplay of information from satellite images, air photographs and
fieldwork (primarily from the Iranian salt domes) by the workers cited above. These
observations give valuable insights into the processes occurring during final diapir
evolution. However, none of these works aimed to investigate the internal processes of
intra-salt inclusions during diapir rise.

Six surface-piercing salt domes in the Ghaba Salt Basin (GSB) of northern interior Oman
provide the unique opportunity to compare data on structural geology, diagenesis and
geochemistry of carbonate inclusions, which are the only lithological (outcrop)
equivalents to deeply buried intra-salt carbonates (so-called ‘stringers’) in the South
Oman Salt Basin (SOSB, Fig. 1). Several field surveys to the salt domes of the GSB were
made by petroleum geologists since the early 1950s with the aim to constrain the
stratigraphic position of the exposed exotic blocks. Currently, the only work published on
the surface-piercing salt domes clearly documented that the major constituent rocks are
carbonates and evaporites of the Late Precambrian to Early Cambrian Ara Group in the
SOSB, which constitutes one of the largest hydrocarbon plays in Oman (Fig. 1; Peters et
al., 2003). The Ara Group in the SOSB was deposited as a six-cyclic sequence of
carbonates and evaporites, which developed into large salt diapirs due to passive
downbuilding. The carbonates are buried to depth from 3 to 6 km and are fully enclosed
within the salt, forming a perfect seal for hydrocarbons in the carbonate ‘stringers’. The
understanding of the sedimentary facies and platform geometries of this unusual, deeply
buried play was considerably improved through the application of an outcrop analogue
from the terminal Proterozoic carbonate ramps of the Nama Group in southern Namibia
(Grotzinger, 2000, Grotzinger et al., 2000, Grotzinger and Amthor, 2002, Adams et al., 2004). However, some aspects of the Ara group reservoir architecture can not be sufficiently explained by the facies analogue from Namibia. Synsedimentary salt tectonics strongly influenced the sedimentary geometries in the Ara Group, causing e.g. the considerable thickness expansion of the basinal carbonate facies. Based on seismic character alone these thick basinal facies can easily be misinterpreted as reefal build-ups (Al-Siyabi, 2005). Moreover, the distribution of diagenetic phases, which govern the reservoir properties of the Ara carbonates, can not be studied at the outcrops in Namibia. Hence an analogue model based on the surface piercing salt domes of the GSB could potentially complete the existing reservoir model for the SOSB with respect to processes related to (synsedimentary) salt tectonic and diagenesis. Hence, the Ara carbonates from the surface piercing salt domes of the GSB could provide a useful completion to the subsurface facies and reservoir model of the SOSB.

The aim of this paper is therefore twofold: First, to better constrain the dynamics of salt tectonics in North Oman. Second, to compare and contrast data on lithology, organic matter maturity and diagenesis with existing data from the deeply buried Ara Group in the SOSB. This will contribute to a better understanding of reservoir quality evolution and diagenetic processes occurring during the uplift of salt diapirs.
Figure 1. Overview map of the Late Precambrian to Early Cambrian salt basins of interior Oman (shaded in grey) (modified after Peters et al., 2003 and Schröder et al., 2005) with the location of the six surface-piercing salt domes Qarn Sahmah (QS), Qarat Kibrit (QK), Qarat Al Milh (QM), Qarn Nihayda (QN), Qarn Alam (QA) and Jebel Majayiz (JM) in the Ghaba Salt Basin (GSB). Note that JM and QM are positioned at the Maradi fault zone and QK at the Burhaan fault zone.
2 GEOLOGICAL SETTING

2.1 The Ghaba Salt Basin

The six-surface piercing salt domes are located in the GSB, which is one of three evaporitic basins constituting the deep subsurface of interior Oman (Fig. 1). The Ara salt in Oman is thought to be time-equivalent to the Hormuz salt, which forms a number of basins from the Persian Gulf region to the Salt range of Pakistan (Gorin et al., 1982; Edgell, 1991). The salt basins of interior Oman follow a NE-SW alignment, which acted as left-lateral strike-slip faults during Late Precambrian to Early Cambrian times (Loosveld et al., 1996). In the SOSB and the GSB the basement is overlain by the Late Precambrian to early Cambrian Huqf Supergroup, which comprises the Abu Mahara Group, the Nafun Group and the Ara Group (Fig. 2). The carbonates and evaporites outcropping in the six surface-piercing salt domes of the GSB belong to the infra-Cambrian Ara Group, which also forms the subsurface hydrocarbon plays of the SOSB (Peters et al., 2003; Al-Balushi, 2005).
Figure 2. Chronostratigraphic summary of rock units in the subsurface of interior Oman (from Peters et al., 2003). The infra-Cambrian Ara Group is highlighted in the lithostratigraphic column and in cross section A-A’ (5 times vertical exaggeration); see Fig. 1 for location.
In the SOSB and partly in the GSB, the development of salt diapirs (Fig. 3) is a product of post-depositional salt movement triggered by differential loading of thick continental clastics onto the mobile substrate of the Ara salt (Loosveld et al., 1996). In the SOSB, the Ara salt forms the top, side and bottom seal for the Ara hydrocarbon carbonate ‘stringer’ play. While the Ara salt diapirs in the SOSB remained unchanged after the phase of passive downbuilding, the Ara salt diapirs in the GSB underwent reactivation followed by active piercing of the strata overlying the Ara salt and ultimately the surface (Loosveld et al., 1996; Peters et al., 2003; Filbrandt et al., 2006). Active piercing was initiated by the development of deep-rooted sinistral strike-slip faults, such as the far-reaching Maradi fault zone and the Burhaan fault (Figs. 1, 2 and 3a). The salt domes Jebel Majayiz and Qarat Al Milh are situated within the basement-involved Maradi fault zone (Fig. 1; Peters et al., 2003). Movements along the Maradi fault zone started during the Late Cretaceous, at the time of thrusting of the Oman ophiolites (Filbrandt et al., 2006), and continued until the Late Neogene as shown by folded Plio-Pleistocene wadi gravels along the fault (Hanna and Nolan, 1989). In addition, the diapirs are surrounded by a complex network of normal, strike-slip and reverse faults, pointing to a temporarily compressive stress field.

Figure 3. Geometry of the surface-piercing salt dome Qarat Kibrit from seismic data. a) Isopach map between Natih and top Gharif, showing the NNW striking Burhaan fault with location of salt dome Qarat Kibrit (QK) and Maradi fault zone. Section AB is shown in b) Seismic cross section through Qarat Kibrit diapir with
interpreted tops (see Fig. 2 for stratigraphy) and outline of diapir shape. Nimr = Early Cambrian, Gharif = Permian, Natih = Upper Early Cretaceous.

2.2 South Oman Salt Basin

The lithologies of the six surface-piercing salt domes are mainly composed by Ara Group evaporites and carbonates, which are clearly correlated to the Ara intra-salt hydrocarbon plays of the SOSB (Peters et al., 2003). From work on drill cores from the SOSB, it is known that the Ara Group spans the Precambrian-Cambrian boundary (Amthor et al., 2003) and consists of marine platform sediments, representing at least six third-order cycles of carbonate to evaporite sedimentation. Each cycle is characterized by the sedimentation of Ara salt at very shallow water depths, followed by the deposition of 20 to 220 m thick isolated carbonate platforms (so-called “stringers”) in deeper basins during transgressive periods (Mattes and Conway Morris, 1990). Bromine geochemistry of the Ara salt (Schröder et al., 2003) and marine fossils (Amthor et al., 2003) clearly indicate a seawater source for the Ara evaporites. In the SOSB, the Ara Group cycles are termed A0 to A6 from bottom to top (Al-Siyabi, 2005 and Fig. 2).

Cores from the Ara carbonate cycles A1C to A4C contain up to 5 sequences and display very repetitive facies patterns. The A1C shows one sequence of finely laminated dark carbonates interpreted as basinal facies to peritidal grainstones and thrombolites in the top section. The A2C is typical laterally very extensive and contains 4-5 sequences with a breccia surface recognizable in most A2C stringers, which represent the most prolific hydrocarbon reservoirs in the SOSB. The A3C interval contains 2-3 sequences and shows very similar facies to the A2C but is always thicker than 100 m (person. commun. J. Grotzinger). The main reservoir facies of the A4C in the Birba Area (see Fig. 2) comprises ‘crinkly’ laminites of the outer ramp and stromatolites and thrombolites of the inner ramp (Schröder et al., 2005). The A5C and A6C intervals are generally composed of similar facies as the A1C to A4C and partially show isolated ‘pinnacle’-like thrombolite reefs. Both intervals have been shown to be mostly non-producing reservoirs. Seismic sections show that the A5C consists of several structurally segmented slabs, which are in some parts of the basin overlain by the A6 clastics (Al-Siyabi, 2005, Fig. 2).
Generally, the shallow-platform carbonate facies of the Ara Group consists of grainstones and laminated stromatolites, while the platform barrier is formed by stromatolites and thrombolites (Fig. 4). The slope facies includes organic-rich laminated dolostones and the basinal facies is dominated by sapropelic laminites (Al-Siyabi, 2005). During seawater highstands, prolific oil source rocks have been formed in the deeper (some hundreds of metres), periodically anaerobic to dysaerobic parts of the basin (mainly slope and basinal mudstones in Fig. 4). Density stratification of seawater allowed preservation of a sufficient amount of organic material in the bottom layers as high productivity of algal material in the upper water layers was present (Mattes and Conway Morris, 1990).

Figure 4. Generalized facies distribution of an Ara carbonate platform with interpreted reservoir and source rock occurrences from the SOSB (Al-Siyabi, 2005).

The Ara evaporites include halite and anhydrite, which replaced primary gypsum (Mattes and Conway Morris, 1990; Schröder et al., 2003). Increasing seawater salinity due to basin regression led to the depositional succession carbonate-sulphate-halite, which in turn proceeded into the succession halite-sulphate-carbonate during the following sea level rise. Anhydrite overlying the carbonate is termed “roof anhydrite”, while anhydrite above the halite is referred to as the “floor anhydrite”. Both anhydrite horizons can be up to 20 metres thick. The thickness of the Ara Salt is 10 - 150 m in the A1 to A4 cycles and can exceed 1000 m in the A5 and A6 sequences (Schröder et al., 2003).
3 METHODOLOGY

3.1 Samples
In total, we investigated 107 carbonate samples from all of the six surface-piercing salt domes. Out of the 107 samples, 40 thin sections were prepared and stained with Alizarin Red S to differentiate dolomite from calcite. Bulk rock powders of 36 samples were analysed with X-ray diffraction (XRD) analysis using a D 5000 diffractometer. Total organic carbon was analysed for 60 samples, 43 carbonate samples for maturity assessment using solid bitumen reflectance (BR_r) and 101 carbonate samples for stable isotope analysis.

3.2 Stable Isotopes
Oxygen and carbon isotope analyses were performed at the laboratory of IFM-GEOMAR, Kiel (Germany). Carbonate powder was dissolved with 100% H₃PO₄ at 75 °C in an online, automated carbonate reaction device (Kiel Device) connected to a Finnigan Mat 252 mass spectrometer. Isotope ratios are calibrated to the Vienna Pee Dee Belemnite (V-PDB) standard using the NBS-19 carbonate standard. Average standard deviation based on analyses of a reference standard is < 0.07‰ for δ¹⁸O and < 0.03‰ for δ¹³C.

3.3 Solid bitumen reflectance (BR_r)
Maturity analysis requires a sufficient number (~ 50) of reflectance measurements on vitrinite or solid bitumen particles to provide reliable data. Reflected light microscopy under immersion oil revealed that all of the GSB samples contain solid bitumen particles, which are too fine-grained (~ 1 µm in diameter) to apply the conventional photometer method. Therefore, we measured the grey values (value of brightness) of the solid bitumen. For this method a Zeiss Axioplan microscope was interfaced with a Zeiss Axio digital camera and a desktop computer using a “Hilgers” instrument and the relevant software FOSSIL and DISKUS. The programs allow the direct conversion of grey values into mean random reflectance values (R_r in %). The calibration was applied using an Yttrium-Aluminium-Garnet (0.889% R_r) and a Gadolinium-Gallium-Garnet (1.714% R_r) at 10 V for a 40x/0.85 n.a. lens under immersion oil (n_e = 1.518).
3.4  Total organic carbon

The assessment of the carbon content was performed with a LECO multiphase C/H/H₂O analyser (RC-412). This instrument operates in a non-isothermal mode with continuous recording of CO₂ release during oxidation, thus permitting the determination of inorganic and organic carbon in a single analytical run. All analyses were performed in duplicate and the results were averaged.

4  FACIES EVOLUTION AND DEFORMATION

This chapter provides detailed observations on the sedimentological and structural architecture of constituent lithologies in the Qarn Nihayda and Jebel Majayiz salt domes from satellite images and fieldwork.

4.1  Facies

4.1.1  Jebel Majayiz
Figure 5. High resolution Quickbird image of Jebel Majayiz showing the interpreted strike of the stringers (red lines). Although no overall configuration is recognizable, there are several clusters with consistent strike. Similarly to Qarn Nihayda, the dome flanks are defined by outwards dipping stringer ridges, while the dome interior shows a much higher density and a chaotic juxtaposition of stringers. Circles denote locations of samples mentioned in the text.

The Jebel Majayiz salt dome is the second largest in size and is characterized by a high density of carbonate stringers showing an overall chaotic juxtaposition (Fig. 5). Generally, the lithofacies is dominated by stromatolites and thrombolites. The stringers in
the southern most third and the central eastern part of the dome are almost completely composed of various types of crinkly and pustular laminites, cherty stromatolites and thrombolites. This lithofacies association is exemplarily illustrated in two stratigraphic sections (Fig. 6; see Fig. 5 for location). Both sections show various transitions between crinkly laminites, pustular laminites and massive thrombolites, which can gradually proceed into stromatolite facies. The mesoscopic appearance of pustular laminites can be very similar to thrombolites, but with an overall laminar structure. In contrast to other parts of the dome, most stringers in the central northern parts (around sample JM42 in satellite image) are mainly composed of finely laminated carbonates, which can gradually pass into crinkly laminites and stromatolitic layers.

Stromatolites and pustular laminites are usually interpreted as peritidal facies of the inner ramp. The depositional environment of thrombolites ranges from shallow-subtidal to the platform margin and slope (Schröder, 2000). Finely and crinkly laminated carbonates are attributed to slope and outer ramp settings (Schröder et al., 2003). Siliciclastic sandstones in the GSB are often associated with stromatolites and therefore are assumed to be intertidal. Alternations between pustular laminites, thrombolites and crinkly laminites (Fig. 6) were also observed in the Birba area of the SOSB (Schröder, 2000). However, the facies in Jebel Majayiz differs from the other salt domes and the SOSB in the high proportion of siliciclastic material, chert and stromatoliths. In the SOSB siliciclastic material is most commonly found in the A6 interval.
Figure 6. Detailed stratigraphy of two sections, showing exemplarily the lithological composition of most stringers in the southern, central eastern and northeastern part. Various types of sandy and cherty stromatolites generally dominate the carbonate facies in Jebel Majayiz.
4.1.2 Qarn Nihayda

Figure 7. Satellite image of salt dome Qarn Nihayda (source: Google Earth, accessed in August, 2007). The outer rim of the dome is defined by tilted whitish Tertiary carbonates showing periclinal strike. The axis of the salt dome is oriented NW-SE,
which corresponds to the overall strike of several stringer ridges. The field photos highlight the complex and chaotic architecture of the stringers in the dome interior (black lines denote direction of dip). Note that the southern part is relatively smooth and essentially formed by anhydrite cap rock. Sampling (for solid bitumen reflectance and stable isotope analysis) was performed along the profiles A-A’ and B-B’, which are shown as simplified cross sections in Fig. 10. The green circle represents the location of the stratigraphic profile in Fig. 8, the red circle marks the location of sample QN 14 showing the potential negative $\delta^{13}$C excursion representative for the A4C interval of the SOSB.

In Qarn Nihayda, a number of profiles record gradual transitions from regressive conditions with elevated salinity to transgressive open-marine conditions (Fig. 7 and 8). At the base of the succession shown in Figure 6, crinkly laminites are overlain by pustular laminites, indicating a shallowing-upwards trend. Displacive anhydrite nodules in the pustular laminites probably indicate a supratidal environment. An overlying residual breccia consisting of crinkly-laminite clasts in an anhydrite-bearing, laminated, micritic matrix, is followed by a residual anhydrite layer. These layers represent restricted marine conditions that possibly led up to rock salt formation. A subsequent flooding event likely lead to the dissolution of evaporites, formation of the residual layers and the subsequent deposition of a ~ 15 m thick interval of light-grey, finely-laminated carbonates. Towards the top, these carbonates show an increasing amount of anhydrite rosettes and nodules, which partially display ‘chicken wire’ structures. These textures most likely indicate a gradual relative sea-level fall that culminated in supratidal conditions. Upsection, a change to more open marine conditions is indicated by a decreasing content of lath anhydrites in finely-laminated to massive carbonates. A further deepening is indicated towards the top of the succession by alternations of finely-laminated, dark-grey carbonates with massive, light-grey mudstones. The massive beds, which are interpreted as turbidites (Peters et al., 2003), show slump folds, dewatering channels and load casts; the latter two structures clearly indicate that this stringer is structurally inverted.
Figure 8. Detailed stratigraphy of a 30 m thick stringer interval in the western part of Qarn Nihayda. Sedimentary structures indicate that this stringer is inverted. See Fig. 7 for location.

4.2 Deformation-related structures

4.2.1 Jebel Majayiz

The structural configuration of stringers, as observed from the satellite image, does not indicate an overall trend (Fig. 5). However, the outer stringers tend to strike more or less parallel to the dome axis (N-S). Within the dome, clusters of stringers show consistent strike. One cluster, which covers an area from the centre towards the northeast, is
characterized by dominantly NE-SW striking stringers. Another cluster in the central-western part contains stringer with N-S orientation. This may suggest that some clusters of rock salt and stringers behaved coherently during rise and rotated against other adjacent clusters. The interior of the dome is characterized by a very high density of stringers, which are chaotically juxtaposed and superimposed against each other.

Deformation-related structures are very abundant in Jebel Majayiz. Brecciation is common in all lithofacies (Fig. 9a). Cataclasites form along normal faults (up to 2m wide), that dissect the stringers perpendicular to bedding. The association between the cataclasites and breccias is difficult to determine. In some cases, the cataclasites are veined and in other cases, the breccias constitute the cataclasites. Large-scale open folding of the stringers is very common. Isoclinal folding accompanied by strong brecciation and thrusting is also present (Fig. 9b-d; for location see JM2 in Fig. 5). Deformation, i.e. folding and thrusting, was most likely accompanied by hydrofracturing as e.g. indicated by irregular-shaped fractures containing highly fragmented material from adjacent layers (Fig. 9d).

The occurrence of anhydrite caprock in Jebel Majayiz is limited to the northern and southern parts and forms an outward dipping periclinal rim around the salt dome (Figs. 8e). Rarely, anhydrite caprock occurs in-between stringers in the dome interior, where it shows a pronounced foliation with a flow-like texture around embedded carbonate clasts (Fig. 9f).
Figure 9. Deformation-related structures observed in the salt dome Jebel Majayiz. 
a) Breccias are associated with veins oriented parallel and vertical to lamination. b) 
Thin and less competent (organic-rich) limestone layers show isoclinal folding. The 
lower bound of the uppermost thick layer (above pen) probably represents a thrust 
plane highlighted in c). d) Irregular-shaped fracture (stippled outline), containing 
highly fragmented material from adjacent layer; the structure is interpreted as a 
hydrofracture (due to high fluid pressures), propagating through the less competent 
thin layers and branching below the competent thick layer above. e) Whitish 
outward dipping anhydrite caprock at the southern margin of Jebel Majayiz. South
is to the right. This layer of anhydrite caprock corresponds to the white periclinal margin observable in Figure 7. f) Internal fabric of anhydrite caprock shows a strong foliation, ‘flowing’ around a carbonate clast (20 cm in diameter).

**4.2.2 Qarn Nihayda**

The satellite image of Qarn Nihayda shows that the dome axis trends NW-SE. Detailed mapping revealed that the strike of most stringers is oriented parallel to the dome axis. This is most apparent along laterally continuous stringer ridges along the western and eastern flank, which are subdivided into several individual blocks (Fig. 7). In most parts of the dome centre this general trend is still discernible but in the northern part of the dome centre the stringers show a more chaotic juxtaposition. The dip of most stringers ranges from 45° to 90° with the steepest dipping stringers at the dome margins and flat dipping stringers in the centre (cross section A-A’, Fig. 10). The simplified cross section B-B’ (Fig. 10) shows open folded stringers in the dome centre with the fold axis oriented parallel to the dome axis. In this cross section, stringers in the eastern part dip to the east, while stringers in the western part dip to the west. Despite of this overall configuration (common strike), the individual stringers in the western part can neither be correlated by the use of sedimentological nor structural features with the stringers in the eastern part.

![Figure 10. Simplified cross sections through Qarn Nihayda. Section A-A’ shows that most stringers dip to the west, while section B-B’ displays an overall symmetric stringer configuration with strike parallel to the dome axis. Note horizontal alignment of some stringers on steeply dipping NW striking stringers. Location of cross sections is indicated in Fig. 7.](image-url)
Two fold generations can be observed in Qarn Nihayda. For example, the most northwestern stringer dips with 70° to the east and is characterized by gentle to tight asymmetric folds with east and west dipping limbs and a N-S trending fold axis (Fig. 11a). The second fold generation represents the gentle bending of the whole stringer with a steeply dipping and E-W trending fold axis, which is common in all of the six salt domes. This fold style is difficult to observe in the field, since interlimb angles are in many cases up to 170°. It becomes more apparent in the satellite image by the slightly curved shape of some stringers (Fig. 7).

The deformation-related structures are in large parts very similar to Jebel Majayiz. The most obvious deformation-related structures in the field are breccias associated with numerous whitish, mostly calcite cemented veins (Fig. 11b-c). These breccias are very common and not tied to a specific lithology. The associated veins are often oriented parallel and perpendicular to bedding, displaying ‘dike-and-sill’ structures (Fig. 11c), which generally form by hydrofracturing due to high fluid overpressures (Mandl, 2005). In some cases, the breccias define up to 2 m wide damage zones, which are oriented perpendicular to the strike of the stringers. These fault zones likely promote the tectonic dissection of stringers in separated blocks. As the overall fold style of the stringers is open, it is difficult to observe crosscutting relationship between the tectonic breccias and folding. However, in some cases it appears that the breccias are clearly incorporated into folding. A few stringers show bedding-parallel and oblique (< 15°) thrust faults, which crosscut tectonic breccias (Fig. 11d). Because marker horizons are lacking, the displacements are not assessable. In addition, a number of faults oriented perpendicular to the bedding are defined by up to 7 cm thick cataclasites (Fig. 11e and f), showing slickensides on their fault planes (Fig. 11f). A slaty and fine-grained fabric typically characterizes these cataclasites.

The occurrence of massive blocks of anhydrite caprock is common in Qarn Nihayda and covers about 90% of the northern and the southern part (see white areas in Fig. 7). As observed in the Qarn Sahmah salt dome, whitish to yellowish anhydrite often forms in-between single stringer ridges in the centre of the dome. The mesoscopic fabric is characterized by thin laminae of pink to whitish anhydrite alternating with dark ochre-
coloured calcite laminae (Fig. 11g). This fabric is often highly fractured and cemented by gypsum or anhydrite (Fig. 11h).
Figure 11. Overview of deformation-related structures in Qarn Nihayda. Pen is 15 cm in length. a) Open fold indicated by stippled line with a N-S trending fold axis. b) Highly brecciated laminite showing angular fragments of various size cemented by white calcite. c) Photograph shows breccias with veins oriented perpendicular and parallel to (crinkly) lamination. d) A few stringers show thrust faults (stippled lines), indicating that the original thickness of a stringer can be extremely modified. e) typical orientation of brittle faults, defined by cataclasites, perpendicular to bedding and strike of a stringer. f) Cataclasite composed of a sandy stromatolite showing cm-sized slickensides. g) Anhydrite caprock showing typical stratification of pure calcite layers (dark layer at bottom) and calcite-rich anhydrite. Long side of image is 1.5 m. h) Anhydrite caprock with dense sets of gypsum- or anhydrite-cemented fractures.

4.3 Summary: Facies and Deformation

In both salt domes, vertical facies changes in relation to relative sea-level change can be observed (Fig. 6 and 8). On the other hand, the relatively chaotic arrangement and strong tectonic modification of stringers in the dome centre of Jebel Majayiz makes it difficult to trace lateral facies variations over more than ~200 m (Fig. 5). This is especially true for the northern part of the dome, which is affected by locally strong brecciation. The stringers with the largest lateral continuity can be found at the dome flanks. In the interior of the Qarn Nihayda dome, both in the central and in the northern part, is also more intensively deformed than the dome flanks, which is suggested by a higher density of tectonic breccias and veins. The relatively high continuity of the stringers at the dome flanks allows a mapping of lateral facies change over several hundred meters. In contrast to Jebel Majayiz, the Qarn Nihayda salt dome comprises carbonate facies that are very similar to the Ara carbonate facies described from the SOSB (e.g. Mattes and Conway Morris, 1990; Schröder et al., 2005). Hence, a detailed facies map of stringers along the flanks of the Qarn Nihayda salt dome will likely provide the best analogue for the subsurface reservoirs in the SOSB.
5  DIAGENESIS

5.1 Paragenetic Sequence

Based on 40 thin-sections, a common paragenetic sequence was developed for the six surface-piercing salt domes (Fig. 12). Adjacent stringers within the individual salt domes were exposed to different maximum burial temperatures, as indicated by bitumen reflectance (see below), and hence might have experienced slightly different diagenetic histories. On the other hand, the paragenetic sequence of all salt domes encompasses the same succession of diagenetic processes from shallow to deep burial and subsequent uplift. The term *shallow burial* refers to diagenetic processes that occur until the carbonate stringers were completely sealed by the Ara salt, which can be assumed to occur at a burial depth of around 30 m for rock salt (Casas & Lowenstein, 1989). Consequently, the term *deep burial* denotes the broad field of diagenetic alterations after the carbonate stringers were fully encased by the Ara salt. Uplift related processes become active when the diagenetic system becomes at least partially open to external fluids. The general paragenetic sequence developed for the salt domes can be directly compared to the general paragenetic sequence from the SOSB (Schönherr et al., 2008), which is also based on a basin wide analysis including the whole range of burial depth. The paragenetic sequence is presented in Figure 12 and supported with photographs in Figure 13 and 14.
Figure 12. Paragenetic sequence deduced from microstructural observations of samples from various Ara carbonate stringers of the six surface-piercing salt domes in North Oman.
Figure 13: Thin-section photographs A) drape of laminae around calcite cemented (stained red) mouldic pore, which is interpreted as early diagenetic anhydrite nodule; Qarat Al Milh, ppl = plane polarized light, Alizarin Red S.

B) Thrombolite displaying a clotted-grumelous microstructure. Dolomite is only partially replaced by calcite (stained red). Highly birefringent anhydrite cement fills the growth framework porosity between the microclots. Anhydrite was partially converted to gypsum; Qarn Sahmah, xpl = cross polarized light, Alizarin Red S.

C) Crinkly laminite with largely preserved dark, organic-rich laminae. Light laminae are dissolved forming mouldic to vuggy porosity, partially filled by solid bitumen (SB). The reflectivity of the solid bitumen indicates a burial temperature of close to 200°C.; Qarat Kibrit, ppl, unstained.

D) Cap rock of anhydrite, quartz and calcite (stained red). Solid bitumen (SB), indicating burial temperatures of close to 110°C, is present in pores. Presence of solid bitumen indicates that oil was present during cap rock formation. P = open porosity, ppl

E) Crinkly laminites characterized by mm-scale alternation between dark organic-rich and light organic-poor laminae. The fine crystalline dolomite in the organic-rich laminae is preserved, whereas the dolomite of the organic-poor laminae was replaced by sparry calcite (stained red). Replacive calcite (Cal-R) shows abundant small dolomite inclusions. A calcite-cemented fracture (Cal-F) running parallel to the laminae is devoid of dolomite inclusions; Qarat Kibrit, ppl, Alizarin Red S.

F) Breccia with floating clasts of dedolomitised laminitic clasts. The clasts are cemented by blocky calcite crystals that show no indication of a dolomitic precursor phase; Qarn Nihayda, xpl, Alizarin Red S.

G) The thrombolite facies shows fabric selective replacement of dolomite by calcite (stained red). Former microclot (Cl) now shows a geopetal infill indicating that microcrystalline dolomite was preferentially dissolved and the void subsequently cemented by sparry calcite (stained red). In contrast, coarse to medium crystalline, subhedral dolomite surrounding the former microclot is well preserved. Qarn Sahmah, ppl, Alizarin Red S.
H) In stromatolites, dolomite (Dol) is preserved where it is cemented by microcrystalline quartz (Qtz). Without cementation dolomite is replaced by blocky calcite (stained red). Inclusions of calcite (Cal) in euhedral quartz crystals indicate that quartz growth continued during dedolomitisation; Jebel Majayiz, ppl, Alizarin Red S.

Shallow burial: The earliest diagenetic phase is botryoidal cement that occludes some of the growth framework porosity in the thrombolite facies. Today these cements are either dolomitic or, if calcitic, contain small dolomite inclusions. This suggests multiple phases of replacement. However, their botryoidal form and blunt crystal terminations suggest an aragonitic mineralogy prior to dolomitisation. The drape of laminae around calcite cemented or open moulds is interpreted as displacive growth of anhydrite nodules in soft sediment before compaction and dolomitisation (Fig. 12, Phase 2; Fig. 13A). Dolomitisation seems to have occurred simultaneously with or slightly earlier than the first phase of carbonate leaching. The vuggy porosity created by this first phase of carbonate leaching is often filled by anhydrite cement (Fig. 12, Phase 5). Anhydrite cement also occluded most of the remaining growth framework porosity in the thrombolite facies (Fig. 13B). The timing of the development of clusters and single anhydrite lath replacing matrix dolomite, predominantly in the thrombolite and crinkly laminite facies, is equivocal. In many cases the anhydrite lath formed prior to significant stylolitisation. They are therefore attributed to the same phase of anhydrite growth as the shallow burial anhydrite cements (Fig. 12, Phase 5).

Deep burial: In its earliest form silica was only observed in the stromatolite facies of Jebel Majayiz, where it occurs as angular fine siliciclastic grains and as clear to light-brown silica cement (Fig. 12, Phase 6). In stromatolite, microcrystalline quartz-cement occludes the intercrystalline porosity of laminae consisting of fine crystalline sub- to euhedral dolomite (Fig. 13H). Silica cement is also present as equant micro- to megaquartz crystals and length-slow chalcedony cementing fenestral pores in stromatolites. The presence of length-slow chalcedony and the textural relationship with dolomite indicates that the bulk of this silica phase precipitated from solutions high in
Mg$^{2+}$ and SO$_4^{2-}$ (Folk and Pittmann, 1971) shortly after dolomitisation. A second carbonate leaching phase (Fig. 12, Phase 7), which is only of local importance, affected the coarser grained organic-poor parts of crinkly laminites (Fig. 13C). In contrast, the darker, organic-rich laminae are preferentially preserved (Fig. 13C). This leaching phase seems to be associated with hydrocarbon migration, since solid bitumen is relatively abundant in this porosity type. In some rare examples solid bitumen is enriched along stylolites but is absent from the matrix. This indicates that some stylolites have acted as pathways for hydrocarbon migration. The onset of stylolite formation hence predates oil migration. Where a textural relationship between stylolites and fractures can be observed, the fractures crosscut the stylolites. The main phase of fracture formation hence seems to postdate stylolite growth. This generation of fractures is normally cemented by dolomite or calcitic cement with small dolomite inclusions, indicating a former dolomitic mineralogy (Fig. 12, Phase 9). Microfracture-lining saddle dolomite, which is characterized by curved-shaped crystal facets and undulose extinction, occurs in few samples. Solid bitumen is relatively rare. For the bulk of the solid bitumen the bitumen reflectance data indicate a formation in the deep burial realm (Fig. 12, Phase 10).

Uplift related diagenesis: In the surface piercing salt domes of the GSB the carbonate stringers are intercalated within diapiric caprock, which forms through the dissolution of salt and the passive enrichment of insoluble impurities at the dissolution front (Posey and Kyle, 1988). In the Ghaba Salt Basin, caprock consists predominantly of anhydrite, quartz, feldspar and gypsum. As long as the carbonate stringers are sealed by salt, their diagenetic system is completely closed. When the salt around the carbonate stringers dissolves during caprock formation, the diagenetic system in turn becomes open to external fluids. In one sample, solid bitumen impregnated pores were observed in caprock (Fig. 13D), indicating that liquid hydrocarbons were present during or after caprock formation. The burial temperature recorded by the bitumen reflectance indicates that caprock formation started at temperatures close to or higher than 110°C, equivalent to a burial depth of more than 3,5 km (Fig. 12, Phase 12). Caprock formation can continue over long time periods whereby progressive dissolution along the salt cap-rock interface leads to underplating of the newly formed residue. Hence the cap-rock becomes younger
towards its base. This also implies that the change to an open diagenetic system did occur at the same time for all carbonate stringers.

The age relationship between diagenetic phases related to caprock formation and dedolomitisation is difficult to evaluate. Most of the uplift related processes seem to occur nearly simultaneously. One of the most important processes is dissolution of evaporites (Fig. 12, Phase 14). Halite cements were not observed petrographically, but halite was identified in trace amounts by XRD. Partially open growth-framework porosity in thrombolites often contains corroded remnants of former anhydrite cements. Moulds after anhydrite nodules and lath also point to extensive leaching of evaporites. Rehydration of anhydrite to gypsum (Fig. 13B) is seldom observed. Some former anhydrite nodules are now filled by calcite. In most cases the calcite is free of inclusions and is interpreted as cement filling the moulds of dissolved anhydrite nodules (Fig. 13A). In a few cases small anhydrite inclusions in the calcite point to a direct replacement of anhydrite by calcite (Fig. 12, Phase 16) without a void phase.

Calcitisation of dolomite is probably the most important uplift-related process (Fig. 12, Phase 17). Small dolomite inclusions in the coarse blocky calcite crystals indicate that calcitisation proceeded through a thin solution-film and not through a solution cavity-fill process (Fig. 13E). Independent of facies, calcitisation is most pronounced adjacent to strongly fractured and brecciated rocks. Outside such zones, calcitisation is facies and fabric selective. In mid ramp successions of interbedded laminated and massive dolostones, the laminated dolostones are much more pervasively calcitised than the massive ones. On the microscale, the organic-poor, light laminae of crinkly laminites are replaced by calcite, while the organic-rich, dark laminae are predominantly preserved as dolomite (Fig. 13E). The geopetal fill in former microclots of thrombolites (Fig. 13G) indicates that microcrystalline dolomite was preferentially dissolved (Fig. 12, Phase 18) and the void subsequently cemented by sparry calcite (Fig. 12, Phase 19). In contrast, coarse to medium crystalline, subhedral dolomite surrounding the former microclot is well preserved. This is one of the rare examples where calcitisation proceeded through a solution cavity-fill process instead of a replacement by thin-film solution without a void phase. Drusy and blocky calcite cement is abundant in fractures (Fig. 13E) and breccias (Fig. 13F). The absence of any dolomite inclusions indicates that these cements were
directly precipitated as calcite and did not form through a replacement of dolomite cement. The lack of any dolomite cement suggests that these fractures were formed during the uplift phase (Fig. 12, Phase 13; Fig. 14 e-f).

Authigenic euhedral quartz crystals of up to several mm lengths contain calcite and dolomite inclusions in the same crystal (Fig. 13H). Often the dolomite inclusions are concentrated in the core and the calcite inclusions in the rim of the quartz crystal. This pattern is interpreted as replacive growth during progressive dedolomitisation.

Dedolomitisation often is accompanied by calcite dissolution (Ayora et al., 1998). In contrast to many other studies on dedolomitisation, there is no evidence for a late uplift related phase of calcite dissolution, such as open rhombohedral pores (Evamy, 1967) or solution enlarged fractures. In general, the sedimentary fabric is well preserved (Fig. 14 a-d) and the different primary facies, such as laminites and thrombolites can be distinguished despite pervasive replacement during dolomitisation and dedolomitisation.
Figure 14. Carbonate samples stained with Alizarin Red S, red indicates calcite. a) hand-specimen of a clotted thrombolite with roundish mesoclots and a laminated part at the bottom. b) corresponding lamipeel, matrix and mesoclots are composed of calcite. c) Crinkly laminit showing alternations of dark (but organic poor) and beige (organic rich) laminae. D) corresponding lamipeel, organic poor laminae consist of calcite, while the organic rich laminae correspond to unstained dolomite
and in contrast to the hand specimen appear dark. e) + f) brecciated laminites with antitaxial vein cements composed of calcite.

5.2 Petrographic differences between the GSB and the SOSB

Shallow burial: Some of the earliest diagenetic phases observed in some samples from the SOSB, such as micritic and drusy dolomite cement, were not observed in the GSB samples. In contrast to the SOSB no indication of a very early, shallow burial phase of fracturing was observed in the GSB. However, a much more important difference between the two basins is the complete absence of halite cements and the reduced abundance of anhydrite cements in the surface outcrops of the GSB. Apart from these differences, the shallow burial diagenetic processes are very similar for both basins and some of the early diagenetic phases, like the botryoidal cement, are surprisingly well preserved.

Deep-burial: The low abundance of solid bitumen in the GSB is one of the most striking differences in comparison to the SOSB. Coke like solid bitumen, which indicates a phase of hydrothermal alteration in the SOSB (Schönherr et al., 2007), was not present in the GSB.

Uplift-related: All processes related to an open diagenetic system during uplift are absent from the SOSB. Fractures are common in the subsurface of the SOSB but are even more abundant in the surface outcrops of the GSB. Most fractures in the SOSB were formed in the deep-burial environment. In the GSB fracturing started to form in the deep-burial environment but the main phase of fracture development is related to uplift.

In the SOSB the replacement of dolomite by calcite is of local importance, likely caused by thermochemical sulphate reduction in the deep-burial environment. Petrographically this dedolomite might be difficult to distinguish from dedolomite in the GSB. It therefore can not be completely ruled out that some of the dedolomite attributed to uplift related processes was already formed during deep-burial. However, the bulk of dedolomite in the GSB can clearly be attributed to uplift related processes.
**Porosity:** Growth-framework, fenestral and vuggy pores are more abundant in the GSB than in the SOSB. In total, this porosity increase outbalances the concomitant decrease in intercrystalline porosity in the GSB.

### 5.3 Stable isotope data

Stable oxygen and carbon isotopes of 101 carbonate stringer samples, mainly from Qarn Nihayda (n = 52) and Jebel Majayiz (n = 31), were analysed (Fig. 17a). The samples have $\delta^{18}O$ values that range between -10.3 and -0.5 ‰ Vienna Peedee belemnite (VPDB), and $\delta^{13}C$ values between -7.7 and 3.9 ‰ VPDB. The carbonate-matrix samples and primary dolomite cemented veins (Fig. 12, Phase 13) display mean $\delta^{18}O$ values of -4.0 ‰ VPDB ($s = 1.58$) and mean $\delta^{13}C$ values of 1.8 ‰ VPDB ($s = 1.74$). The lowest $\delta^{18}O$ values (-10.3 to -6.6 ‰ VPDB) derive from uplift related calcite vein cements (Fig. 12, Phase 13), which also show the most depleted $\delta^{13}C$ values in the range of -7.7 to -1.3 ‰ VPDB. The matrix of the same samples shows isotope values, which invariably plot in the average field of most samples.

The $\delta^{18}O$ values from the carbonate matrix samples of the surface-piercing carbonate stringers of the GSB are depleted by about 1.6 ‰ compared to the subsurface carbonate stringers of the SOSB (Fig. 15, Schoenherr et al., in press). In contrast, the mean values of $\delta^{13}C$, excluding the calcite vein cements, is nearly identical for both settings, although some GSB samples are depleted compared to the SOSB samples. With respect to $\delta^{13}C$ most of the matrix samples from the SOSB and many from the GSB plot close to the field of isotopic composition of the Neoproterozoic seawater (Derry et al., 1992; Jacobsen and Kaufman, 1999). Furthermore, two samples from the same stringer in Qarn Nihayda (QN14, Figs. 7 and 15) and two samples from two different stringers in Jebel Majayiz (JM2 and JM 16, Fig. 5 and 15) plot in the field of the short-lived negative $\delta^{13}C$ excursion, which marks the Precambrian-Cambrian boundary in the A4C interval of the SOSB (Amthor et al., 2003).
5.4 Interpretation and comparison to the SOSB

With respect to early and deep-burial diagenesis the development in the GSB seems to be nearly identical to the SOSB (Schönherr et al., in press). The lower abundance of solid bitumen would be explained most easily by a lower volume of hydrocarbons in the GSB compared to the SOSB. All other major differences are a consequence of diagenetic alterations in association with dedolomitisation in the GSB. Dedolomitisation, the
replacement of dolomite by calcite, is caused by Ca-rich solutions. Meteoric water, responsible for halite dissolution during caprock formation, likely is also undersaturated with respect to anhydrite (Posey and Kyle, 1988). When anhydrite is dissolved by meteoric waters, Ca\(^{2+}\) is expelled to the porewater and promotes calcite cementation and replacement of anhydrite by calcite. Calcite precipitation in turn decreases the pH and removes carbonate ions from the solution causing dissolution of dolomite (Back et al., 1983). The concurrent calcite precipitation and dolomite dissolution leads to dedolomitisation. This process could be enhanced by bacterial or thermochemical sulphate reduction, which removes sulphate from the porewater and increases alkalinity (Ben-Yaakov, 1973; Reuning et al., 2006), favouring anhydrite dissolution and calcite precipitation. The reduction of dissolved sulphate ions is accompanied by organic matter oxidation. The presence of pore filling solid bitumen in caprock demonstrates that liquid hydrocarbons were available for sulphate reduction during or after caprock formation. The bitumen reflectance of this solid bitumen indicates a burial temperature of about 110°C. At temperatures above 80°C thermochemical sulphate reduction is more likely to occur than bacterial sulphate reduction. Subsequently, bacterial sulphate reduction could have occurred during later uplift. The presence of saddle dolomite and solid bitumen, which are typical by-products of sulphate reduction (Machel et al., 2001), might indicate that this process was active in the GSB.

The influence of meteoric waters in the GSB is also supported by stable isotope values. The oxygen isotopic composition of carbonates is directly controlled by temperature and the isotopic composition of the precipitating fluid. The greater burial depth in the GSB, deduced from bitumen reflectance data, could have contributed to the shift towards lighter \(\delta^{18}O\) values compared to the SOSB. However, the fact that the most negative \(\delta^{18}O\) values occur in uplift-related, cemented fractures that formed late in the diagenetic history and the positive correlation to \(\delta^{13}C\) indicates that temperature is perhaps not the main controlling factor. A similar positive correlation between \(\delta^{13}C\) and \(\delta^{18}O\) was observed in Pleistocene phreatic cave deposits from the diapiric Jabal Madar dome in northern Oman (Fig. 15), which is underlain by Ara formation evaporites (Immenhauser et al., 2007). The authors attributed the trend towards lighter oxygen isotopes to the progressive mixing of saline, deeply circulating meteoric fluids that rose along the diapir.
stem with descending $^{18}$O depleted meteoric freshwaters. The accompanying shift towards lighter $\delta^{13}$C values was interpreted as the incorporation of variable amounts of $^{12}$C derived from the oxidation of soil-zone organic carbon derived from land plants (Meyers, 1997; Reuning et al., 2005). Descending meteoric freshwaters would be characterized by lighter $\delta^{13}$C values. The ascending basinal fluids in comparison would have obtained a less depleted $\delta^{13}$C signature through dissolution of marine carbonates during their circulation in the basin. Extensive fluid convection cells, where basinal waters are channelled upward along escape structures bounding the diaper stems, are known from the Gulf of Mexico (Posey and Kyle, 1988). A similar mechanism as proposed by Immenhauser et al. (2007) hence could explain the observed isotope trend in the surface piercing salt domes of the GSB. However, an additional source for depleted $\delta^{13}$C values in the surface piercing salt domes might be the oxidized organic matter from liquid hydrocarbons (with $\delta^{13}$C commonly between -25 and -30 permil) consumed during sulphate reduction (Machel, 2001, Reuning et al., 2002). The phreatic calcites in Jabal Madar (Immenhauser et al., 2007) have more depleted $\delta^{18}$O values than the carbonate stringers of the surface piercing salt domes. Likely the influence of deeply circulating basinal fluids, with higher $\delta^{18}$O values, was more important at the surface piercing salt domes than in the cave deposits of Jabal Madar. Alternatively, the meteoric waters which caused the dedolomitisation of the carbonate stringers at the surface piercing salt domes might have had a less depleted $\delta^{18}$O signature. Carbonate cements of quaternary conglomerates from south of the Oman Mountains vary between -8 and -1 permil $\delta^{18}$O PDB, indicating a less depleted isotopic composition of shallow groundwater (Burns and Matter, 1995).

The fact that vein filling calcite cements (Fig. 12, Phase 13 + 19) show the most negative oxygen and carbon isotope values suggests that faults acted as fluid pathways during dedolomitisation. In contrast to the calcite cemented veins, the bulk of samples from the carbonate matrix and primary dolomite cemented fractures (Fig. 12, Phase 9), show a large overlap with the $\delta^{13}$C values of the subsurface samples from the SOSB. This likely is due to a lower fluid to rock ratio outside of the uplift related fractures, which buffers the carbon isotope values towards the positive marine $\delta^{13}$C values of the host carbonate.
Four samples from the GSB plot close to the A4C samples from the SOSB that record the negative isotope excursion of the Cambrian/Precambrian boundary (Amthor et al., 2003). Since these samples from three different stringers in Jebel M and Qarn Nihayda are only weakly dolomitised, they could have recorded an unmodified secular isotope signal and hence would be time equivalent to A4C stingers in the SOSB. However, this hypothesis would need to be confirmed by additional carbon isotope analysis of these stringers.

5.5 Implication for reservoir properties

Dedolomites are often described as highly porous with a good hydrocarbon reservoir potential. Dedolomitisation can proceed through a direct replacement, where dissolution of dolomite and precipitation of calcite occur simultaneously (Evamy, 1967; Ayora et al., 1998). Alternatively, the dissolving dolomite leaves a void, which is subsequently filled by calcite cement either during the same overall process (GarciaGarmilla and Elorza, 1996), or from a different solution at a different time (Jones et al., 1989; James et al., 1993) Dedolomitisation has been described as i) porosity increasing due to the predominance of dolomite dissolution over calcite precipitation (Canaveras et al., 1996), ii) as porosity reducing because of cementation of open porosity by calcite (Munn and Jackson, 1980); or as iii) porosity preserving, where dolomite dissolution and calcite precipitation takes place pseudomorphically (Evamy, 1967).

Ayora et al. (1998) argue based on reactive transport modelling and petrographic observation, that dedolomitisation can either take place as pseudomorphic replacement on a volume to volume basis or non-pseudomorphically on a mole to mole basis following the equation:

$$\text{CaMg(CO}_3\text{)}_2 + \text{Ca}^{2+} = 2\text{CaCO}_3 + \text{Mg}^{2+}.$$  

They state that pseudomorphic replacement will preserve the porosity, whereas non-pseudomorphic replacement will reduce the porosity since the molar volume of dolomite is less than two molar volumes of calcite.
In the case of the GSB the petrographic evidence clearly points to direct replacement as the dominant process. Evidence for solution-cavity fill, such as geopetal fill, has only been observed sporadically.

The preservation of many primary or early diagenetic structures, such as the botryoidal cements, points to pseudomorphic replacement as dominant process. Dedolomitisation itself hence would have been porosity preserving (Ayora et al., 1998). Direct calcite precipitation, independent of dolomite replacement, leads to the occlusion of fracture and intercrystalline porosity. A later phase of calcite dissolution, like it has been observed in many other dedolomites (Evamy, 1967; Ayora et al., 1998), is not present in the GSB. Hence, the overall increase in porosity does not seem to be due to the dedolomitisation itself but can be attributed to the dissolution of anhydrite and probably also halite mainly from growth-framework, fenestral and vuggy pores.

In summary, the study of the diagenesis of surface piercing Ara carbonates can in some respect be helpful for the interpretation of diagenetic relationships in the subsurface of the SOSB since many of the early to deep burial diagenetic phases were preserved by pseudomorphic replacement. A direct comparison of petrophysical (porosity and permeability) and geochemical (e.g. $\delta^{18}O$) properties on the other hand is hampered by the strong uplift related diagenesis in the GSB. To obtain an unaltered carbon isotope signal, e.g. to identify the A4C interval, it is recommended to use facies which are less affected by dedolomitisation, such as the massive dolostones of the mid ramp facies.

6 SOLID BITUMEN REFLECTANCE AND PALAEO-TEMPERATURES

Samples for both, total organic carbon (TOC) and mean random solid bitumen reflectance ($BR_r$) were selected in order to compare data between several ‘stringers’ within one salt dome and between the salt domes.

6.1 TOC

The majority of the samples selected for TOC determination derive from dark and fetid finely laminated carbonates of the basinal facies ($n = 55$), which macroscopically appeared to represent potential source rocks. In addition, some samples originate from the
crinkly laminite and thrombolite facies. Only 3 out of 60 samples have TOC values higher than 0.5% and that about 90% of all samples have TOC values < 0.1%.

### 6.2 Solid bitumen reflectance

The majority of the samples from which the TOC was determined, and some additional crinkly laminites, thrombolites and one diagenetic cap rock sample were selected for BR<sub>r</sub> measurements (n = 43 samples). The very low TOC values correlate very well with a very low content of solid bitumen in all samples. Reflected light microscopy has shown that ~ 95% of the 43 samples contain solid bitumen, however, besides a few exceptions, most samples contain only trace amounts of very fine-grained solid bitumen, mostly accumulated along laminae. The applied method allows to measure BR<sub>r</sub> of solid bitumen particles with a rectangular area of at least 0.42 µm². As most of the solid bitumen particles in all samples studied are smaller than this area only 25 out of the 43 samples could be used to calculate palaeo-temperatures from the BR<sub>r</sub> data. In order to obtain palaeo-temperature data, the BR<sub>r</sub> data were converted to VR<sub>r</sub> data (Schoenherr et al., 2007), which were converted to palaeo-temperatures using the equation of Barker and Pawlewicz (1994). The comparability of maturity data of samples from different locations within one salt dome is additionally hampered by the presence of 2 and locally of 3 generations of solid bitumen.

Despite of these difficulties hampering the use of BR<sub>r</sub> for correlation of individual stringers within a salt dome, some insights on the maturity of the six surface-piercing salt domes can be drawn.

In *Qarat Kibrit*, the BR<sub>r</sub> data for 5 samples from a 20 m thick stringer described in Peters et al. (2003) consistently show two generations, the first with a BR<sub>r</sub> of ~ 2% and the second with a BR<sub>r</sub> of ~ 1%, indicating burial temperatures of 200 °C and 150 °C, respectively. BR<sub>r</sub> data from a stringer ~ 200 m in the north of the first outcrop indicate a burial temperature of 170 °C.

In order to test, if the symmetric configuration of the stringer ridges observed in *Qarn Nihayda* matches with a possible symmetry trend of maturation isogrades, we measured the BR<sub>r</sub> of samples located along the WSW-ENE trending profiles (see Figs. 7 and 10). The dataset indicates highest burial temperatures of 200 °C to 250 °C of stringers located...
in the dome centre and of 280 - 300 °C of stringers outlining the eastern flank of the
dome (see Fig. 7). In one sample, solid bitumen in diagenetic caprock (Fig. 13D) records
a burial temperature around 110 °C. The solid bitumen indicates that liquid hydrocarbons
were present during or after caprock formation. However, it is not clear if the
hydrocarbons were sourced from within the carbonate stringers or migrated into the
caprock from an external source. A contribution from possible salt flank traps seems
possible, even so an exploration well just to the east of the diapir proved to be
unsuccessful.

In the *Qarn Alam* salt dome, two samples from outcrops about 300 m apart show near
identical BR_r values, which correspond to burial temperatures of 240 °C.

Palaeo-temperatures from 5 stringers of the *Qarn Sahmah* salt dome vary between 150 to
200 °C. Two thrombolite samples, which have been taken 2 m apart from a ~ 400 m long
and 1.5 m thick basalt dike indicate palaeo-temperatures of 330 - 370 °C. Interestingly,
microstructures of the basalt show cracks, which are cemented by calcite, quartz and
subordinate solid bitumen.

Palaeo-temperatures from 4 stringers of salt dome *Jebel Majayiz* are between 176 and
214 °C.

In summary, this dataset shows that the surface-piercing stringers in the GSB underwent
burial temperatures from 140 to 300 °C with the majority of burial temperatures at ~ 200
° C.

### 6.3 Interpretation and comparison to the SOSB

The extremely low TOC values (< 0.1%) of most samples from the basinal facies indicate
low organic matter productivity and hence poor source rock potential in the salt domes.
This is in contrast to the laminites of the A1C-A4C reservoir intervals of the SOSB,
which have TOC values in the range of 0.5 – 5.0 % (Schoenherr et al., 2007). The BR_r
values and the calculated burial temperatures can not be directly compared to the SOSB,
since the BR_r values in the SOSB are strongly overprinted by hydrothermal alteration
(Schoenherr et al., 2007). In the GSB hydrothermal alteration seems to be restricted to the
vicinity of the basalt intrusion in Qarn Sahmah. Other evidence for hydrothermal
alteration, such as coke like solid reservoir bitumen, was not observed in the GSB. Palaeo-burial temperatures between 140° and 300°C in the GSB indicate burial depths between 5 and 10 km, assuming a (randomly chosen) geothermal palaeo-gradient of 30°C/km. Hence most of the stringers were buried deeper than the deepest stringers in the SOSB (~ 6 km).

7 CONCEPTUAL MODEL FOR SALT DIAPIR EVOLUTION

This chapter discusses the depositional, diagenetic and structural evolution of the surface-piercing salt domes in 4 stages and integrates data from the subsurface of the SOSB to highlight differences occurring during the uplift of intra-salt carbonate stringers. The structural evolution is in large parts compiled from the work of Escher and Kuenen (1929), Loosveld et al. (1996), Callot et al. (2006) and Filbrandt et al. (2006).

7.1 Ara Group deposition (Late Precambrian to Early Cambrian)

The fieldwork of Peters et al. (2003) clearly revealed that the carbonates and evaporites exposed in the six surface-piercing salt domes belong to the Late Precambrian to Early Cambrian Ara Group, well known from the SOSB. They describe an outcrop in Qarat Kibrit that displays the cyclic Ara Group succession of rock salt – anhydrite – carbonate – anhydrite – rock salt. Bromine contents of rock salt from this succession vary between 49 – 68 ppm (Schoenherr et al., submitted), indicating a normal marine feed for rock salt (Valyashko, 1956). This is consistent with bromine data in the range of 45 - 109 ppm for the Ara salt from the SOSB (Schröder et al., 2003). Although the affinity to the Ara Group is well established for the surface-piercing carbonate stringers, their exact correlation to the Ara cycles (A0C to A6C) as recorded in the SOSB is not straightforward. In the SOSB the stratigraphy is based on geochronology, bio- and chemostratigraphy. Dating of several ash beds suggest that the Ara Group in the SOSB was deposited between ca. 547 Ma and 540 Ma (Amthor et al., 2003; Bowring et al., 2007). Absolute dating of the stringers is not possible for the GSB, since no ash bed was identified in the surface piercing salt domes. In the SOSB the fossils Cloudina and Namacalatus are very abundant in the A1-A3 carbonates, especially in the thrombolite facies, but are absent from the A4-A5 level (Amthor et al., 2003). Despite the widespread
occurrence of thrombolites in the GSB, no Cloudina and Namacalatus fossils were found. Their absence suggests that the stringers in the surface piercing salt domes are younger than A3C.

The A4C stringer interval in the Birba area of the SOSB is characterized by a negative carbon isotope excursion (Amthor et al., 2003) coincident with a pronounced enrichment in redox sensitive trace elements (Schröder and Grotzinger, 2007) and a strong gamma ray signal (Mattes and Conway Morris, 1990). At least the negative carbon isotope excursion can be correlated world wide to the Precambrian-Cambrian boundary (Amthor et al., 2003). The presence of a gamma-ray response and trace element enrichment were not tested for the GSB, but two stringers in Jebel Majayiz and one stringer in Qarn Nihayda show the negative isotope signal typical for the A4C interval. Since these samples are only weakly dedolomitised, they could have recorded an unmodified secular isotope signal and hence would be time equivalent to the A4C stringers in the SOSB (see below). The fact that only three from ~100 isotope samples show a possible A4C signature leads to the conclusion that the bulk of the stringers in the GSB likely is time equivalent to the A5 and A6 interval in the SOSB. This interpretation is supported by the abundance of sandstone layers and siliciclastic rich carbonates in the Jebel Majayiz salt dome (Fig. 6), since clastic intervals occur between the A5C and the A6C in the SOSB (Fig. 2). The low hydrocarbon productivity of the A5C and A6C intervals in the SOSB would also fit to the low content of TOC and solid bitumen observed in the surface-piercing stringers.

### 7.2 Initial stage of passive salt diapirism (Middle Cambrian to Late Ordovician)

The differential loading of the thick Haima clastics onto the mobile substrate of the Ara salt caused passive diapirism (downbuilding) until the Haima grounded on the subsalt strata in the SOSB as well as in the GSB (Loosveld et al., 1996; Peters et al., 2003). Seismic sections presented by Peters et al. (2003) (their Fig. 22) indicate the possible presence of stringers in the subsurface of the Saih Nihayda field in the GSB in a structurally similar position as the intra-salt stringer reservoirs in the SOSB. So far, the intra-salt stringers in the GSB are not a target of hydrocarbon exploration, probably due
to difficulties in seismic detection and the deep burial of the stringers. The presence of solid bitumen generally indicates that oil was generated in the surface-piercing stringers, probably during the last periods of Haima deposition (as in the SOSB, Terken et al., 2001). The stringers exposed in the salt domes show very small amounts of solid bitumen and very low TOC values. On the other hand, the presence of buried hydrocarbon (gas?) bearing stringers, equivalent to the A1C to A3C in the SOSB, cannot be excluded. As indicated before, the presence of liquid hydrocarbons during or after caprock formation is indicated by solid bitumen (Fig. 13d). However, the hydrocarbons source is ambiguous. They could have been derived from within the carbonate stringers or migrated into the caprock from an external source.

7.3 Re-active diapir growth and surface piercement (Late Cretaceous – recent)

The Late Cretaceous activity along the Maradi fault zone and the Burhaan fault caused re-active diapir growth and finally the piercement of the surface by the salt domes Jebel Majayiz, Qarat Al Milh, Qarat Kibrit and Qarn Alam (Peters et al., 2003; Filbrandt et al., 2006). The association of deeply rooted faults with the position of the salt domes Qarn Nihayda and Qarn Sahmah is so far not confirmed by seismic interpretations. The very high amount of anhydrite caprock representing a dissolution residue from large quantities of rock salt and the presence of magmatic rocks and pre-Ara volcanics suggests a different evolution of Qarn Sahmah compared to the 5 more northern located salt domes (see Fig. 1).

The widespread occurrence of cataclasites and tectonic breccias in the surface-piercing stringers were not observed in the extensively cored Ara carbonate stringers of the SOSB. The most likely structural position of this deformation is the narrow diapir stem, where differential stress is highest (up to 5 MPa) throughout a salt diapir (Schoenherr et al., submitted). This uplift related structural deformation probably led to the second phase of calcite cemented fracture formation (Fig. 12, Phase 13 and 19; Fig. 14), and is coincident with the major phase in dedolomitisation.

The highly heterogeneous distribution of palaeo-temperatures within the salt domes shows that the stringers derive from different depths and thus different stratigraphic
intervals. It seems unlikely that all of the stringers were at the same structural level before piercement of the surface. The recent structural configuration of stringers and the widespread occurrence of anhydrite caprock in most salt domes (e.g. Fig. 9g-h) rather suggest strong dissolution of salt near and at the surface, which led to rotation and a chaotic juxtaposition of the stringers. These processes strongly masked the original structural configuration of stringers during diapir rise and thus insights into the internal kinematics of the north Oman salt diapirs. An exception represents the near symmetric and periclinal orientation of most stringers in Qarn Nihayda (Fig. 7), which points to the preservation of the original stratigraphy after piercement of the surface. However, the palaeo-temperatures from the Qarn Nihayda stringers do not show the symmetrical stringer configuration, i.e. increasing palaeo-temperatures towards the dome centre.

Insights from seismic sections and fieldwork suggest a relatively simple geometry of the six surface-piercing salt diapirs with far-reaching diapir stems and cylindrical terminations (Fig. 3, 7, 9). The lack of structures indicating lateral spreading (injection into adjacent strata) of the salt suggests constant confinement by country rock walls during emplacement. The steep dip of the salt (Qarat Kibrit) and of numerous carbonate stringers (e.g. Qarn Nihayda) contradicts a cut through the uppermost structural levels of a mushroom-shaped diapir as suggested by Talbot and Weinberg (1992) for some of the salt plugs in Iran. In contrast to many salt domes in Iran (Jackson et al., 1990) or in the subsurface of northern Germany (Mohr et al. 2007), there is no evidence for the presence of present or past salt glaciers.
Figure 16. Schematic sketch of the structural evolution of the surface-piercing salt domes in the Ghaba Salt Basin. a) Deposition of the Ara Group in infra-Cambrian times. b) Passive downbuilding due to differential loading of the Haima clastics. c) Late Cretaceous strike-slip faulting (e.g. Maradi Fault Zone) d) Re-active diapir growth along deep-rooted strike-slip faults. Intra-salt carbonate stringers underwent strong brecciation and folding. Salt dissolution led to the formation of an anhydrite caprock and allowed meteoric waters to enter the stringers, which is deduced from depleted $\delta^{18}$O values compared to the subsurface stringers of the
SOSB. e) Strong salt dissolution after periodic rainfalls caused removal of salt and rotation of stringers.

8 CONCLUSION

- The surface-piercing stringers generally comprise the same suite of facies as the deeply buried stringer play in the SOSB and thus represent an important outcrop analogue to study lateral facies development on a scale up to several hundred meters.

- Poor source rock development, the near absence of the A4C negative $\delta^{13}C$ isotope excursion and the lack of the Late Precambrian fossil Cloudina, as well as the abundance of clastics presumably point to the exposure of the stratigraphically uppermost stringer intervals (A5C and A6C compared to the SOSB) in the salt domes.

- Palaeo-temperatures point to very deep burial depths of up to 10 km due to passive downbuilding of the Ara Group.

- The stringers are internally much higher deformed than in the SOSB and depleted $\delta^{18}O$ isotopes of syn-tectonic veins and the carbonate matrix suggest deformation and dedolomitisation in a diagenetically open system due to the contact with meteoric waters within the diapir stem.

- Many of the early to deep burial diagenetic phases were preserved by pseudomorphic replacement. A direct comparison of petrophysical (porosity and permeability) and geochemical (e.g. $\delta^{18}O$) properties with the SOSB is hampered by the strong uplift related diagenesis in the GSB. Weakly dedolomitised carbonates, such as the massive dolostones of the mid ramp facies, should be used to identify the light carbon isotope shift associated with the A4C interval.

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