Controls on the formation of stratabound dolomite bodies, Hammam Faraun Fault block, Gulf of Suez

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Abstract

Dolomitization is commonly associated with crustal-scale faults, but tectonic rejuvenation and diagenetic overprinting often makes it difficult to unravel the timing and mechanisms of dolomitization. Furthermore, there is often a mass imbalance between dolostone volume, fluid and Mg concentration. This study considers differential dolomitization of the Eocene Thebes Fm. on the Hammam Faraun Fault (HFF) Block, Gulf of Suez using combined outcrop, petrographical and geochemical data. The Thebes Fm. has undergone a simple history of burial and exhumation as a result of rifting that was initiated in the Late Oligocene, making it an ideal case study in which to study fluid flow and dolomitization within a rift basin. Stratabound dolostone bodies occur selectively within remobilized sediments (debrites and turbidites) in the lower Thebes Formation and extend into the footwall of, and for up to 2 km away from, the HFF. They are offset by the N-S trending Gebel fault, which was active during the earliest phases of rifting, suggesting dolomitization occurred between rift initiation (26Ma) and rift climax (15Ma).

Isotope and trace element data suggest that dolomitization occurred from evaporated (~1.43 concentration) seawater at temperatures of ~65-80^{o}C. Thermal convection is interpreted to have occurred as seawater was drawn down surface-breaching faults into the Nubian sandstone aquifer, where they were convected and discharged into the lower Thebes Formation via the HFF. Assuming an approximate 10My window for dolomitization, a horizontal velocity of ~1 m/yr into the Thebes Fm is calculated, with fluid flux and reaction facilitated by fracturing. Although fluids were at least marginally hydrothermal, stratabound dolostone bodies do not contain saddle dolomite and there is no evidence of hydrobrecciation. This highlights how misleading dolostone textures can be as a proxy for the genesis, spatial distribution and length scales of such bodies in subsurface studies. Overall, this study provides an excellent example of how fluid flux may occur away during the earliest phases of rifting, and the importance of crustal-scale faults on fluid flow, even during the earliest part of their growth. Furthermore, since dolomitization occurred from seawater, we present a mechanism for fault-controlled dolomitization that has none of the inherent mass balance problems of classical, conceptual models of hydrothermal dolomitization.

Key words:

Rift initiation, stratabound dolostone, dolomitization, geothermal convection

Introduction

Stratabound dolostone bodies associated with faults are often interpreted to be genetically related to massive so-called hydrothermal dolomite (HTD) bodies (e.g. Southern Italy (Cervato, 1990; Boni *et al.,* 2000), Northern Spain (Gasparrini *et al.,* 2006; López-Horgue *et al.,* 2010; Dewit *et al.,* 2014), Lebanon and Iran ((Nader *et al.,* 2007; Sharp *et al.,* 2010) and Western Canada (Davies and Smith, 2006; Lonnee and Machel, 2006)). In these examples, the dolomitizing fluids utilise deep-seated structural lineaments, usually normal or strike-slip faults, as conduits for the upward flow of fluids derived from beneath the dolomitized succession. Pervasive dolomitization is interpreted to take place adjacent to the fault, forming massive, non-stratabound dolostone bodies, whilst stratabound bodies formed as fluids fluxed along beds away from the fault. This results in a 'Christmas tree' shaped geobody, often hundreds of metres or even kilometres in lateral extent, which have been targeted for hydrocarbon exploration.

There has been some success in the exploration and production of such bodies within the West Canada Sedimentary Basin (e.g. Davies and Smith, 2006) but few examples outside this region (e.g. Panuke Field, Canada; Wierzbicki *et al.,* 2006). Overall, the play has not proven globally successful because of a lack of either reservoir presence or volume. Nonetheless, HTD still has enormous economic potential as a host for low temperature Mississippi-Valley type (MVT) deposits within the USA, Canada and UK (e.g. Gregg, 2004). Furthermore, the conceptual model for the formation of HTD has come under scrutiny, since many studies fail to adequately explain the source and drive mechanism for dolomitizing fluids and source of magnesium.

On the Hammam Faraun Fault (HFF) Block, Gulf of Suez, non-stratabound dolomitization is observed in proximity to the HFF, with numerous, discontinuous stratabound dolostone bodies extending into the footwall for up to 2 km. Hollis *et al.* (2017) demonstrated that in this well-exposed example, the stratabound dolostone pre-dates, and is not genetically related to, an associated, cross-cutting non-stratabound dolostone. This is contrary to the interpretation of many dolostone bodies formed in proximity to faults. This paper provides detailed characterisation of these stratabound dolostone bodies through field observations and measurements, petrography and geochemistry. It discusses and tests the flow mechanisms responsible for their formation, calculates the mass balance and presents a conceptual model for dolomitization in the context of the structural evolution of the fault block. The results provide an important insight into fluid flux and diagenesis on carbonate platforms in the earliest stages of rift initiation.

Geological setting

The Gulf of Suez is a failed intracontinental rift that forms the northern continuation of the Red Sea rift, which originated during the early Palaeozoic as a narrow embayment of the Tethys Ocean. The rift was rejuvenated 23 Ma, during the separation of the African Plate from the Arabian Plate in the Late Oligocene (Jarrige *et al.,* 1990; Patton *et al.,* 1994; Sharp *et al.,* 2000; Younes and McClay, 2002; Bosworth *et al.,* 2005). From the Middle Miocene (14–12 Ma), stress in the southern Gulf of Suez transferred to the Aqaba–Levant transform boundary, formed by the collision of Arabia with Eurasia, and dramatically slowed down extension rates in the Gulf of Suez (Montenat *et al.,* 1988; Bosworth *et al.,* 2005).

The Suez rift is bounded by two major sets of marginal faults that strike parallel to the length of the Gulf (Alsharhan and Salah, 1995). The normal faults are linked by shorter, slightly oblique transfer faults, resulting in an *en echelon* fault pattern (Gawthorpe and Hurst, 1993; Patton *et al.,* 1994; Sharp *et al.,* 2000). In cross section, the rift is characterised by large, widely spaced half grabens with eroded crests and deep hanging-wall basins (Moustafa and Abdeen 1992; Knott *et al.,* 1995; Sharp *et al.,* 2000; Khalil and McClay, 2001). Each half graben consists of several fault blocks (Robson, 1971), with the study area located in the Hammam Faraun Fault (HFF) Block (Fig. 1). During the Oligo-Miocene, basaltic dykes and sills (dated as 22 and 26 Ma; Montenat *et al.,* 1986; Moustafa and Abdeen 1992; Patton *et al.,* 1994) were emplaced in the study area and provide the first stratigraphic evidence for rifting. Overall, volcanism in the Gulf of Suez is minor, but a large basaltic province has been described in northern Egypt, with a geochemical signature typical of primary magmas extruded over 1-2 Myr at around 23 Ma (Bosworth *et al.,* 2015).

The HFF is a major crustal scale fault, with greater than 4 km of displacement (Jackson *et al.,* 2005), controlling the position of the present day shoreline. The HFF Block is the footwall of the HFF and is exposed on the western side of the Sinai Peninsula, bounded to the east and south by the broadly NW-SE trending Thal Fault (Fig. 1). Sediments within the fault block (Fig. 1B) dip north-eastwards, controlled by the orientation of the HFF (Robson, 1971; Moustafa and Abdeen, 1992; Jackson *et al.,* 2005; Leppard and Gawthorpe, 2006). During the earliest stages of rifting, movement was focussed primarily along the Thal Fault and its isolated fault segments, including a N-S fault (referred to as the Gebel Fault; Fig. 2) which intersects the HFF and has a maximum displacement of 300 m. The presence of early syn-rift (Miocene) sediments that unconformably overly pre-rift carbonates within the hanging wall of this fault date movement to rift initiation (~26 Ma; Woodman, 2009). From 17-15 Ma, displacement focussed and localised onto the HFF, with no further movement along the Thal Fault and its associated segments (Gawthorpe *et al.,* 2003). Since the HFF is located immediately offshore of the study area, the hanging wall is not exposed and has not been imaged by seismic data. The damage zone of the HFF is exposed in the footwall and comprises fabric-destructive, non-stratabound dolostone that is pervasively dissected by non-stratabound calcite, dolomite, gypsum and iron oxide cemented fractures (Korneva *et al.,* 2017; Hirani *et al.,* in prep). Hydrobrecciation was not observed, except within a 30 m wide fracture corridor that bounds the massive dolostone body (Hirani *et al.,* in prep). Away from the damage zone, fracture spacing increases with bed thickness, with a higher fracture density (narrower spacing) in dolostone compared to limestone beds (Korneva *et al.,* 2017).

The pre-rift stratigraphy of the HFF Block comprises a 2000 m thick sedimentary section, which overlies Precambrian igneous and metamorphic basement rocks (Fig. 1) (Moustafa and Abdeen, 1992; Moustafa, 1996). The basement is overlain by a thick, undifferentiated package of quartz arentic sandstones, the Nubian Sandstone, which has very high average porosity and permeability (26%, >1D; Nabawy *et al.,* 2009) and is an important modern day aquifer. The overlying Mesozoic succession (Sudr, Matulla, Raha, Wata and Esna; Fig. 1B) comprises mixed carbonate and siliciclastic sediments with a moderate porosity but low permeability (7-12%, <0,01mD; Zalat *et al.,* 2010; Hassanain, 2012). The upper part of the pre-rift succession comprises of fine-grained basinal carbonates (Thebes, Darat and Tanka Formation), that are unconformably overlain by early syn-rift mixed carbonate and siliciclastic sediments of the Tayiba and Abu Zenima Formation (Jackson *et al.,* 2005).

The primary focus of this study, the early Eocence Thebes Formation, has been informally subdivided into the lower and upper Thebes, comprising eight lithofacies (Fig. 2; Table 1). The lower Thebes Fm. is composed of skeletal packstone (S1) in which four remobilised facies are embedded (Hirani, 2014):

- Lenticular-shaped matrix-supported and clast-supported conglomerates (R1) with convex upper contacts. These bodies form along depositional dip and are up to ~250 m long, < 65 m wide and up to ~20 m thick and are interpreted as debrites.
- Upward-fining formainiferal grainstones (R2) that are up to 100 m in length and < 50 m wide and < 6 m thick often overlying R1 facies, which are interpreted as turbidites.
- A single, thin (< 2 m thick) sheet-like bed of matrix-supported conglomerate (R3) with a scoured base which can be traced for up to 1.5 km along strike and 1 km down-dip towards the top of the Lower Thebes member. It is interpreted to have formed by shelf margin or upper slope collapse.
- The transition between the upper and lower Thebes Fm. is characterised by a slumped skeletal grainstone (R4) which is < 3 m thick and up to 5 m long and is interpreted to have formed by plastic flow.

The upper Thebes Fm. dominantly comprises skeletal wackestones (B1) with abundant planktonic foraminiferal wackestone facies that are characteristic of low energy, pelagic deposition in a deep marine environment. They are interbedded with massive skeletal grainstone beds (R5), sometimes with channelized bases (R6). Overall, the Thebes Fm. is interpreted as an upward-deepening succession that was deposited on an unstable slope and an outer ramp – basinal environment.

Methodology

Field, petrographical and petrophysical methods

The stratigraphic distribution of the dolostone bodies and their geometries were determined by field mapping and logging supported by Google Earth © images. Detailed stratigraphic logs were collected and comprehensively sampled within the dolostone bodies, host limestone and the contact zones at the margins of the dolostone bodies. Thin and polished sections were prepared from 219 samples, following impregnation with blue-dye resin and staining with alizarin Red S and potassium ferricyanide to constrain the mineralogy (Dickson, 1966). All thin sections were analysed using standard petrographic methods (transmitted light with cold cathode luminescence microscopy), with limestone textures described using the Dunham (1962) classification and dolostone textures described using the Sibley and Gregg (1987) classification. The burial history curve for the lower Thebes Fm. in the footwall of the Hammam Faraun Fault (HFF) was constructed using the present-day thickness of the sediments overlying the Thebes Fm., with tie points at various points throughout the syn-rift stratigraphy (Armstrong 1997; Gawthorpe *et al*., 2003; Woodman 2009).

Standard helium porosity measurements were performed on 44 plugged samples with 1 inch diameter core plugs of variable length using a digital helium porosimeter (DHP 100), Permeability was measured using N_2 gas under a series of pressure conditions using a digital gas permeameter (DHP 200) to determine the standard Klinkenberg permeability, reported as millidarcy (mD).

Geochemistry

All geochemical data is summarised in Table 2. Powdered samples were obtained using a dentist drill. Bulk mineralogy and stoichiometry were determined by XRD analysis, using the Lumsden (1979) equation to determine the measured d[104] spacing ($M = 333.3 \times \text{d-spacing} - 911.99$). Inductively-coupled plasma mass spectrometry (ICP-MS) and Inductively-coupled plasma atomic emission spectrometry (ICP-AES) were used to quantify the concentrations of minor and trace elements (Fe, Mn, Al, Sr, Ni, Ba) within 18 representative samples, reported in ppm. Rare Earth Element (REE) analysis was conducted on 16 representative samples at the French Research Institute for Exploitation of the Sea (Laboratoire Géochimie et Métallogénie), using the methodology outlined by Bayon *et al.,* (2009). The REE concentrations were normalized to the post-Archean Australian Shales (PAAS) for comparison with previous studies using REE seawater proxies (Nance and Taylor, 1976; Haley *et al.,* 2004; Nothdurft *et al.,* 2004).

C and O isotope ratios were collected at the SUERC research facility (East Kilbride, Scotland) using the carbonate CO_2 extraction line. The samples were analysed using the method outlined by (McCrea, 1950), where 7 mg of powdered sample was digested in 100% phosphoric acid in a 25°C water bath (calcite samples left to react for 3 hours and dolomite samples for 24 hours). The resulting CO_2 gas was analysed on a VG OPTIMA mass spectrometer. All stable isotope values are reported in per mil (‰) relative to the Vienna Pee Dee Belemnite (VPDB) standard. A marble standard was used to calibrate the results, as well as replicate analyses, which were reproducible to ±0.1‰ (2σ). The oxygen isotopic data was adjusted for isotopic fractionation associated with the phosphoric acid reaction using the Rosenbaum and Sheppard (1986) calibration.

Strontium isotope analyses were undertaken at the French Research Institute for Exploitation of the Sea (Laboratoire Géochimie et Métallogénie). The carbonate powders were weighed and dissolved in 3M HNO_3. After reacting, the liquid was heated on a hot plate to evaporate and eliminate excess ions and re-dissolved in 3M HNO_3. The sample was purified using Sr spec extraction chromatic resin and Sr column preparation followed the procedure of Horwitz and Bloomquist (1975). The Sr was collected and analysed using the Triton Thermal Ionization Mass Spectrometer, with precision of individual runs at 0.000005 (1σ). All results are reported to the NBS 987 standard.

Clumped isotope analysis was carried out on 3 dolomite samples at the Qatar Stable Isotope Laboratory at Imperial College London. Procedures followed Dale *et al.* (2011). Briefly, three replicate measurements were done per sample, with replicates measured after a month to correct for errors over time. For each replicate measurement, 5 mg of powdered sample was digested and reacted in 2 ml of 104% phosphoric acid at 90°C for 20 minutes. The resulting CO_2 gas was purified in a manual vacuum line through a poropak column and analysed on a MAT 253 mass spectrometer to simultaneously measure Δ_{47}, $\delta^{18}O$ and $\delta^{13}C$. Data correction were done using Easotope (John and Bowen, 2016): non-linearity corrections were applied using the heated gas method (Huntington *et al.*, 2009), and the acid fractionation factors for Δ_{47} were applied using Guo *et al.* (2009). All values are reported in the absolute reference scale of Dennis *et al.* (2011; Carbon Dioxide Equilibrated Scale, or 'CDES'), using the following standards for the conversion: heated gas (0.027‰ CDES, Wang *et al.*, 2004), ETH3 (0.700‰ CDES, Meckler *et al.*, 2014) and Carrara marble (ICM, 0.399‰ CDES, Dale *et al.*, 2014). The Δ_{47} values were converted into a temperature using the Kluge *et al.* (2015) calibration, and errors are reported as ±1S.E., corresponding to a confidence interval of 68%. To calculate the $\delta^{18}O$ of the dolomitizing water, we used the dolomite-water fractionation equation from Horita (2014).

Mass balance calculation

In order to calculate the number of moles of dolomite per m^3 of stratabound dolostone an average porosity of 19% was assumed based on an average present-day porosity of 13% plus an average 6% to account for porosity occluded by telogenetic cements. The input and output data for the mass balance calculation are shown in Table 3. For dolomite with a density of 2.84 g/cm^3 and a molar weight of 184.4 g, $1m^3$ of dolostone contains some 12500 moles of dolomite ((density /molecular weight) * (1- fractional porosity)). Mass balance calculations provide a minimum estimate of the total volume of fluid required to dolomitize the observed volume of stratabound dolostone based on this molar concentration of dolomite. The dolomitizing fluid is assumed to derive from Eocene-Oligocene seawater, the chemistry of which was calculated using PHREEQC (Parkhurst and Appelo 2013) based on compositional data in Brennan *et al.* (2013). This seawater has a salinity of 33.8$^{o}/_{oo}$, a density at

25°C of 1021 kg /m^3 and concentrations of calcium and magnesium of 16 and Mg 36ppm respectively. We use the Pitzer database (Plummer et al., 1988), which takes into account specific ion interactions that occur in high ionic strength solutions (Krumgalz, 2001). The dolomitization potential is the volume of fluid required to dolomitize 1m^3 of limestone, and was derived from the mass of dolomite estimated by PHREEQC to precipitate from the solution in the presence of excess calcite. From the minimum total volume of fluid required the minimum flux based on the estimated range of the age of dolomites from field relationships and strontium isotope data is estimated. Sensitivity to both temperature, using constraints from clumped isotope paleo-thermometry, and also the degree of evaporative concentration of the Eocene-Oligocene seawater, based on the $\delta^{18}O$ of the dolomitizing water were investigated.

Field observations

The Thebes Formation is partially dolomitized in the footwall of the HFF (Fig. 3). The dolostone bodies are easily identified by their distinctive dark brown weathering colour, relative to the cream colour of the host limestone. Two main types of dolostone bodies have been observed in the field (Fig. 4):

a) Stratabound dolostone bodies, varying in length from 5 m up to 300 m along depositional dip, and ranging in thickness between 25 cm and 5 m. These dolostone bodies extend for up to 2 km away from the HFF, before disappearing beneath the sedimentary cover.

b) Massive dolostone bodies, up to 500 m wide (distance from HFF) and 60 to 80 m thick. Dolostone tongues associated with these massive bodies extend for up to 100 m away from the massive dolostone bodies, forming 1 to 5 m thick bands (Hirani, 2014; Hollis *et al.,* 2017).

This paper focuses on the stratabound dolostone bodies and the major controls associated with their formation. Stratabound dolostone only occurs in the Lower Thebes member (Fig. 2B), forming only within the debrite (R1) and turbidite (R2) facies. Debrite and turbidite facies may be completely dolomitized, or dolostone bodies may terminate within the debrites and turbidites (Fig. 4). There was no evidence in the field for a sedimentological control on the distribution of dolostone within facies R1 and R2, and no consistent observations to suggest a geological control on the termination of dolostone. The dolostone bodies are offset by the Gebel Fault, by approximately 300 m (Fig. 3). The dolomitized beds are often more prominent within the landscape than the limestone as they are more resistant to weathering. The upper and lower contacts of these bodies with the adjacent limestone are usually sharp and planar, with abrupt lateral terminations (Fig. 4).

Pre-dolomitization diagenesis and geochemistry

The texture and geochemical signature of the precursor limestone was described to determine the likely controls of rock reactivity and petrophysical properties on the distribution of dolostone. A large proportion of the slope and remobilised deposits have been recrystallised by microcrystalline calcite, but there is less evidence of recrystallization of basinal wackestone facies, except where the internal structure of the pelagic foraminifera is replaced by calcite (Fig. 5). From herein, the basinal wackestone facies are referred to as unaltered limestones, and the slope and remobilised deposits are referred to as recrystallised limestones.

The rare-earth element (REE) signatures of 3 unaltered limestone and 4 recrystallised limestone samples exhibit positive La and negative Ce anomalies (Table 2). They are also relatively enriched in the heavier rare earth elements (HREE), although the increase in the concentration of the HREE is suppressed in the recrystallized limestone (Fig. 6A). $\delta^{18}O$ and $\delta^{13}C$ ratios were measured on 3 unaltered limestone and 16 recrystallised limestone samples (Table 2). Whole rock isotopic values for the unaltered limestones average $\delta^{18}O$ = -3.35‰ $_{VPDB}$ and $\delta^{13}C$ = -0.17‰ $_{VPDB}$ whilst the recrystallised limestones exhibit lower $\delta^{18}O$ values (average = -6.4‰ $_{VPDB}$; -5.12‰ $_{VPDB}$ to -7.80‰ $_{VPDB}$) (Fig. 7A). The $\delta^{13}C$ values for the unaltered limestone (average = -0.2‰ $_{VPDB}$;-1.0‰ $_{VPDB}$ to +0.7‰) and the recrystallised limestone have a similar range (average = 0.4‰ $_{VPDB}$; -0.69‰ $_{VPDB}$ to +1.04‰ $_{VPDB}$). The $^{87}Sr/^{86}Sr$ ratio of 1 unaltered limestone sample is 0.707814 while the recrystallised sediments are more enriched (average = 0.707849; 0.707773 to 0.708011; Table 2).

Dolostone macroscale features

The stratabound dolostone bodies occur almost exclusively within the matrix-supported debrite facies (R1) and, to a lesser extent, the turbidites (R2) (Fig. 8). They typically exhibit fabric preserving textures, with a dark grey dolomitized matrix surrounding dolomitized clasts with a distinctive rusty brown colour (Fig. 8). Locally, dolomitized clasts within the debrites (R1) are wrapped by zebra-like dolomite fabrics (Fig. 8), which can also occur in random orientations within the dolostone strata, often beneath chert nodules. These visually striking fabrics are made up of alternating bands (2-4 mm thick) of dolostone, with elongated vuggy pores that are partially cemented by coarse calcite cements (Fig. 8) and, occasionally, gypsum. In addition, vugs and mouldic pores after clasts are often occluded by coarsely crystalline calcite cements with botryoidal, fasicular and dogtooth morphologies (Fig. 8). These cements have depleted $\delta^{13}C$ (-6.1‰ $_{VPDB}$, -3.6 to -10.1‰ $_{VPDB}$) and $\delta^{18}O$ (-11.1‰ $_{VPDB}$; -5.1 to -16.8‰ $_{VPDB}$). Iron and sulphate rich minerals, including haematite and gypsum, post-date the pore filling calcite cements.

Dolostone microscale textures

In thin section, the matrix of the stratabound dolostone bodies exhibit both planar-s and non-planar textures (Sibley and Gregg, 1987; Fig. 9). The planar-s textures have a cloudy core and some have clear rims (CCCR), ranging in size between 20 and 60 μm, with minor fabric preservation evident as ghosts of *Nummulites*. Under cathodoluminescence (CL), they exhibit zoned luminescence, with mottled bright red and orange luminescent cores, an intermediate thick dull green to yellow luminescent zone and a thin bright red luminescent outer zone (Fig. 9). The non-planar dolomite textures are composed of very cloudy crystals between 10 and 30 μm in size with no evidence of fabric preservation (Fig. 9). These dolomites are not zoned in CL, with mottled bright red and orange luminescent crystals.

Dolomitized clasts within the stratabound dolostone bodies comprise 75-150 μm CCCR dolomite crystals with a planar-s texture. Overall, the dolomitization is non-mimetic, but some precursor limestone texture is preserved as ghosts of benthic foraminifera. The contact between the clast and matrix is characterised by an irregular corrosion rim in which the dolomite crystal cores appear to be preserved relative to the crystal rims (Fig. 10). Under CL, clast-replacive dolomite has bright red and orange mottled luminescent cores with dull orange luminescent outer zones (Fig. 10). The matrix adjacent to the clasts has a cloudy, non-planar texture with rhombic crystals ranging in size between 50-80 μm that are predominantly non-luminescent with patches of bright red and orange luminescence.

Dolostone geochemistry

The geochemical composition of the stratabound dolostone bodies is summarised in Table 2. X-ray diffraction analysis confirms the mineralogy to be stoichiometric dolomite (mean $CaCO_3$ = 51.1%, 50.0-53.1%) with up to 5% calcite in some samples, and minor quantities of non-carbonate phases including quartz and gypsum. Rare Earth Element analysis of 9 samples reveals a positive La and negative Ce anomaly, with a decrease in the concentration of the HREE giving a flattened profile compared to unaltered limestone (Fig. 6A). All samples cluster tightly within the marine quadrant of a Pr/Pr* ($[Pr/(0.5Ce + 0.5Nd]_{SN}$) versus Ce/Ce* ($[Ce/(0.5La + 0.5Pr]_{SN}$) plot (Bau and Dulski, 1996; Webb and Kamber, 2000; Fig. 6B). $^{87}Sr/^{86}Sr$ ratios of are more radiogenic than that of the unaltered limestone and range between 0.707992 and 0.708241, coincident with Oligo-Miocene seawater (Table 2; Fig. 11). Concentrations of Fe (140 ppm, 37-411 ppm) and Sr (151 ppm, 58-232 ppm) are moderate, whilst Mn concentrations are enriched (469 ppm; 208-991 ppm) (Table 2).

The isotopic composition of the stratabound dolostone is constrained to a narrow range ($\delta^{18}O$ average = -5.5‰$_{VPDB}$; -4.0‰$_{VPDB}$ to -8.0‰$_{VPDB}$ and $\delta^{13}C$ average= -0.1‰$_{VPDB}$; -1.4‰$_{VPDB}$ to +1.2‰$_{VPDB}$; Fig. 7A; Table 2). The estimated temperature by clumped isotope paleo-thermometry (Table 4) overlaps at the 68% confidence level for the three samples in a range of 69-78°C (Fig. 7C). Using these temperatures, $\delta^{18}O_{dolomite}$ and the dolomite-water fractionation equation of Horita (2014), the fluid composition for the three samples at the 68% confidence level is +2.5±0.1‰ (Fig. 7C).

Petrophysical properties

Overall, porosity is higher in the unaltered limestones (13%; 1 – 39%;Table 1) than in the dolomitized facies (3%, 1-7%) whilst permeability is uniformly low (0.47mD, 0.01-3.5mD in limestone; 0.1mD average, all values <2mD in dolostone; Table 1). Within limestone facies, the best porosity is within basinal wackestone facies, (B1 average= 24%) and the channelized grainstone facies (R6 average = 18%). Pore types include interparticle and mouldic macropores and intraparticle microporosity. Dolomitized facies are dominated by intercrystalline macro- and micropores in thin section but also exhibit large vugs (10-30 cm diameter) and sub-spherical mouldic pores after clasts (5-20 cm diameter) which were too large to sample for core analysis.

Interpretation

Timing of limestone recrystallisation

To constrain the composition of the dolomitizing fluids, it is necessary to consider the composition of the precursor limestone, since the geochemical fingerprint of the stratabound dolostone could be inherited from the host rock (e.g. Banner *et al.,* 1988). The $^{87}Sr/^{86}Sr$ ratio of the unaltered and recrystallised limestones is less radiogenic than the stratabound dolostone, and coincides with late Eocene seawater (Koepnick *et al*., 1985; Fig. 11). The $^{87}Sr/^{86}Sr$ ratio for the unaltered limestone samples would be expected to match Early Eocene seawater, and the slightly younger age may either reflect minor recrystallization that is not visible petrographically or contamination by later-precipitated calcite cements. The unaltered limestones have a REE signature that is typical of seawater, with a positive La and negative Ce anomaly, as well as relative enrichment in the heavier rare earth elements (HREE) (Nothdurft *et al.,* 2004; Wyndham *et al.,* 2004). Whole rock isotopic values for the unaltered limestones lie within the range that is expected for deposition from Eocene seawater (Pearson *et al.,* 2001). Recrystallised limestones also display a positive La and negative Ce anomaly, but with slightly depleted REE concentrations (Fig. 6). Their $\delta^{13}C$ signature is similar to that of the unaltered limestones, however they have a significantly depleted $\delta^{18}O$ (mean = -6.4‰ $_{VPDB}$ compared to -3.6‰ $_{VPDB}$ for unaltered limestones). Assuming recrystallization from seawater, estimated as $\delta^{18}O_{SMOW}$ = -1 to -0.75‰ for early-mid Eocene seawater (Pearson et al., 2001) and using the fractionation factor of Friedman and O'Neil (1977), this would equate to temperatures of ~42°C. Based on a normal geothermal gradient of 25°C/km and a seawater temperature of 25°C, this temperature could have been reached at <750m burial. In this case, the depleted HREE signature could be a result of mineral stabilization under suboxic conditions (Haley *et al.,* 2000; Melim *et al.,* 2002).

Timing of dolomitization

Field relationships indicate that stratabound dolostone bodies are offset by the Gebel Fault, which became inactive prior to the rift climax, indicating that dolomitization must have occurred before the mid-Miocene (~15-17 Ma; Fig. 12; Gawthorpe *et al.,* 2003). Dolomitization is therefore constrained to post-deposition (35 Ma) and pre-15 Ma. It seems unlikely that dolomitization was pene-contemporaneous with sedimentation on the HFF Block because deposition during the Eocene took place in a deep-water basin (Moustafa and Abdeen, 1992), an environment that is not normally associated with dolomitization. Furthermore, the abundance of stratabound dolostone decreases with distance from the HFF, which was not initiated until 26 Ma (rift initiation), rather than along depositional dip; indeed no stratabound dolomite is seen down-dip of the study area, to the south (Hirani, 2014). It is unlikely that up-dip dolomitization occurred as only a very minor volume of dolostone is described from platform top and upper slope facies in age-equivalent, proximal sediments outcropping in the Northern Galala Plateau (Hontzsch *et al.,* 2011). Furthermore, no clasts of dolomitized limestone are found within otherwise undolomitized remobilized facies, ruling out the possibility that dolomitization in the field area is a down-dip extension, or remobilisation of, dolomitization on the platform top. Since deposition of deep water (outer neritic), slope-basinal limestone persisted into the late Eocene (Tanka Formation; Fig. 1; Jackson *et al.,* 2006), dolomitization at any point in the Eocene is considered highly unlikely. This constrains formation of the stratabound dolostone to the Oligo-Miocene (~26-15 Ma), a period of ~10 My that was coincident with rift initiation, but predated the rift climax.

Origin and temperature of dolomitizing fluids

Stratabound dolostone is stoichiometric and exhibits planar-s and non-planar textures. Although crystal texture is not an unequivocal indicator of the temperature of dolomitization, the presence of both fabrics would suggest that fluid temperatures ranged from just below to a little above the critical roughening temperature of 55ºC (Sibley and Gregg, 1987). The mottled orange and red luminescent cores observed in CL (Fig. 9) may represent remnants of the host limestone, perhaps due to rapid nucleation during the earliest stages of dolomitization. The clear rims that surround the core have a well-defined concentric zonation under CL, typical of a slower rate of growth as the crystal passively filled interparticle porosity. The ratio of Mn to Fe, based on mean values of Mn = 469 ppm and Fe = 140 ppm is high (3.35), suggesting luminescence is activated by Mn incorporated into the lattice under suboxic conditions from a fluid with low concentrations of Fe. Zonation would then most likely occur as a result of changes in the concentration of trace elements during crystal growth (Brand and Veizer, 1980; Machel and Burton, 1991). The thick, dull green to yellow luminescent zones within the planar-s dolomite could be activated by higher concentrations of REE (Machel and Burton, 1991) or Mn^{2+} substitution for Ca^{2+} within the crystal lattice (Gillhaus *et al.,* 2010).

The $^{87}Sr/^{86}Sr$ ratio of the stratabound dolostone coincides with late Oligocene to early Miocene seawater ((0.708062 to 0.70824; Koepnick *et al*., 1985, Jones *et al.,* 1994; Reilly *et al.,* 2002; Fig. 11). Average concentrations of Sr are 151 ppm, with the lowest concentrations (58 ppm) in the most stoichiometric dolomite and increasing Sr concentration with decreasing dolomite stoichiometry (Fig. 6C). Based on Tertiary dolostones formed from seawater, Vahrenkamp and Swart (1990) noted that

Sr concentrations increase as dolomite stoichiometry decreases, since Sr is preferentially hosted in the calcite lattice; our results are consistent with their observations. The relatively low concentrations of Fe (< 500 ppm; Table 2) within stratabound dolostone and average $\delta^{13}C_{dolomite}$= -0.1‰ $_{VPDB}$ are also consistent with formation from seawater ($\delta^{13}C$ ~ 0‰; Prokoph *et al.,* 2008) although $\delta^{13}C$ have a slightly greater range (-1.4 to 1.2‰ $_{VPDB}$) than unaltered and recrystallised limestone (Fig. 7A; Table 2). The REE profile of dolostone shows that the HREE are slightly suppressed compared to unaltered and altered limestone, although they cluster in the marine quadrant of the Pr/Pr* vs Ce/Ce* plot (Fig. 6B). Typically, $\delta^{13}C$ and REE profiles are inherited from the precursor limestone under all but the highest fluid-rock ratios (~1:10^{4}; Banner *et al.,* 1988). The subtle differences between the host limestone and stratabound dolostone in this dataset suggest that fluid-rock ratios could have been high enough for the compositional signature of the dolomitizing fluid to be retained.

The $\delta^{18}O_{dolomite}$ measured by conventional analysis have tightly clustered values (mean = -5.5‰ $_{VPDB}$; median = -5.6‰ $_{VPDB}$; mode = -6.5‰ $_{VPDB}$). From the three clumped isotope analyses, $\delta^{18}O_{water}$ ranges from +0.6 to +3.6‰. Applying these data to the conventional isotope data (Table 4b) using Matthews and Katz (1977) indicates that dolomitization from the lightest fluid, which approximates penecontemporaneous seawater (+0.6‰), would occur at 36 to 61°C, *i.e.*, lower temperatures than calculated by clumped isotope analysis. If dolomitization was from a heavier fluid (+3.6‰), fluid temperatures are calculated as 54 – 83°C; *i.e.* closer to the range of clumped isotope-determined temperature. In reverse, using the clumped isotope temperatures to calculate the oxygen isotopic composition of the dolomitizing fluid from the average $\delta^{18}O_{dolomite}$ of the conventional data ($\delta^{18}O_{dolomite}$ = -5.5‰ $_{VPDB}$; Table 4b) gives $\delta^{18}O_{water}$ = +4 to +5‰ $_{SMOW}$.

Consequently, although strontium isotope and trace element data are consistent with dolomitization from seawater, oxygen isotope data suggests dolomitization occurred from an oxygen-isotopically enriched fluid. This could occur if seawater was slightly evaporated. Using the technique of Swart (1989) and assuming a $\delta^{18}O_{seawater}$ = +0.6‰, $\delta^{18}O_{water\ vapour}$ = -2‰ (equivalent to modern day rainfall within the study area) and a humidity of 50% (equivalent to the modern day), evaporation of a 30% fraction of seawater would form a fluid with$\delta^{18}O_{water}$ = +5.5‰ and a salinity of 48 ppt. With an increasing fraction (40 to 50%) of evaporation to a fluid of 56 to 68‰ salinity, the $\delta^{18}O_{water}$ decreases to +3.0 and +2.0‰ respectively (Table 5). A mixed carbonate-clastic Oligocene succession on the HFF Block records deposition within a shallow water coastal environment (Jackson *et al.,* 2006). Evaporites have not been described, but given a palaeo-latitude similar to today, and therefore an arid to semi-arid climate, it is possible that seawater was mesohaline, and therefore a 30% evaporated fraction of seawater seems reasonable. The stratabound dolostone bodies are consequently interpreted to have formed from slightly evaporated (mesohaline) seawater, at temperatures up to ~80°C, during the Oligocene to early Miocene, prior to movement on the Gebel Fault and onset of the rift climax. This interpretation is evaluated further in the discussion.

Mass Balance

In order to calculate the volume of seawater required for formation of the stratabound dolostones, we consider a 1 m wide slice of the 200m thick lower Thebes member within which the stratabound dolomite bodies extend for a distance of 2000 m perpendicular from the HFF (Table 3). The distribution of dolostones could only be viewed in the field in pseudo-3D, but based on field logs, photos and mapping and modelling they were estimated to comprise 19% of the total rock volume. This gives an estimated volume of 400 000 m^3 carbonate per 1 m wide section with approximately 76,000 m^3 dolostone. An equal volume can be assumed in the hanging wall, even though this cannot be observed as the hanging wall is offshore and displaced vertically from the footwall by several kilometres.

The calculated temperature of dolomitization based on average $\delta^{18}O_{dolomite}$ and assuming a fluid of +3.6‰ SMOW is 64°C. Assuming dolomitization by Oligo-Miocene seawater at 65°C, up to 1.30 x 10^{-2} moles of dolomite can form from each Kg of seawater. Thus 77 Kg of seawater (ie. 1/1.30 x 10^{-2} moles/kg), or 0.077 m^3 is required to precipitate 1 mole of dolomite. At an estimated porosity of 19% the stratabound dolostone bodies are comprised of 12475 mol dolomite / m^{-3}, so multiplying by 0.076 m^3/mol gives a minimum of 957 m^3 of Oligo-Miocene seawater to precipitate 1m^3 of dolomite at 65°C. Multiplying by the total volume of dolostone gives 73 x 10^7 m^3 fluid necessary to dolomitize each 1m slice of the footwall. Assuming a maximum dolomitization period of 10 My (ie. from rift initiation to cessation of movement on the Gebel Fault), then a minimum flux of 7.3 m^3 per year would be required into the Lower Thebes in footwall for each 1 m wide section of the fault, with at least an equivalent flux into the hanging wall. Assuming that porosity was 19% and entirely effective, this gives a velocity of at least 77 m/yr within the fault, and a lateral flux of at least 1.0 m/year.

Considering mesohaline brines formed from Oligo-Miocene seawater at concentration factors of between 0.3 and 0.5 increases the dolomitisation potential at 65°C to 1.85 x 10^{-2} and 2.62 x 10^{-2} moles of dolomite respectively. This reduces the volumetric requirement, suggesting minimum fluid fluxes within the fault of 53 and 37 m/yr, equivalent to lateral fluxes of only 0.70 and 0.49 m/yr. At 80°C, although kinetics may favour more rapid dolomitisation, the dolomitisation potential is actually lower, with a reduction in the moles of dolomite formed per Kg by a factor of 0.76 compared to that at 65°C, independent of the degree of evaporative concentration (Table 3). The estimated fluid flow rates both within the fault and within the lower Thebes Formation (~1m / yr) are rather low compared with those in modern active hydrothermal systems (1000 m/yr) and modern confined aquifers and (1 to 30 m/yr) (Giles, 1987), but are consistent with rates calculated by Corbella et al (2014) for a similar system. Also, rather than a steady flow system operating over 10 My, it is likely that significantly higher flow occurred over numerous shorter periods associated with transient increases in fault zone permeability.

Post-dolomitization diagenesis

The stratabound dolostone bodies contain localised vugs and zebra-dolomite like features which are occasionally occluded by botryoidal, fasciular and dogtooth calcite (Fig. 8). The zebra-like fabrics are highly localised, often beneath chert clasts and are strikingly different from 'typical' zebra dolomite fabrics (e.g. Davies and Smith, 2006) because they lack saddle dolomite cements. There are many theories to explain the formation of zebra dolomite, but the absence of a relationship to facies and the absence of pressure solution features suggest that they most likely formed under elevated pressure and perhaps trans-tension (e.g. Lopez-Horgue *et al.,* 2009; Juerges *et al.,* 2016). It is very tentatively suggested that such conditions occurred during the late syn-rift when movement on the HFF waned in response to movement along the Aqaba- Levant transform (Montenat *et al.,* 1988; Bosworth *et al.,* 2005). At this point, dolomitization along the HFF is interpreted to have stopped (Hollis *et al.,* 2017) and therefore perhaps a limited flux of dolomitizing fluids inhibited cementation of the zebra-like fabrics by dolomite.

Continued uplift and rotation on the HFF from the Miocene to Recent has brought the HFF Block to surface (Fig. 12). During this time, the Thebes Formation re-entered the marine realm before it became emergent. Calcite cementation could have occurred within vugs at this time. Some textures (e.g. botroyoidal calcite) are suggestive of marine cementation, but the highly depleted carbon and oxygen isotopic values measured in these cements are suggestive of cementation by, or recrystallization from, meteoric fluids. Locally, this has also resulted in patchy dedolomitization of stratabound dolostone (Hirani, 2014).

Discussion

Dolomitization mechanism

The decrease in the volume of stratabound dolostone away from the HFF suggests that this major, crustal-scale fault played an important role in controlling the flow of dolomitizing fluids into the Thebes Fm. Dolomitization is interpreted to have occurred from seawater or slightly (<2x) evaporated seawater prior to the rift climax, when deformation became focused on the HFF, and before termination of movement on the Gebel Fault (i.e. before 15 Ma). At this time, the base of the Thebes Formation was at ~600 m (early Oligocene) to ~850m, when movement on the Gebel Fault was initiated (Fig. 12). Clumped isotope data indicates that dolomitization occurred at temperatures of 67-78^{o}C. Consequently, if dolomitization was from seawater, then it must have been heated and circulated towards and along the HFF before it had breached the surface.

From the early Oligocene, a series of incipient intrablock faults across the study area controlled the distribution of shallow marine sediments of the Tayiba Fm. (Early to Middle Oligocene), indicating that the proto-HFF footwall was flooded (Refaat and Imam, 1999; Jackson *et al.,* 2006). Where these intraplatform faults breached the surface, they could have acted as conduits for the drawdown and circulation of seawater. The low permeability of the underlying Mesozoic and Tertiary succession could have maintained seawater circulation along the fault plane to depth. On reaching the highly permeable Nubian Sandstone Formation (Fig. 13), however, seawater could have been entrained within free convection cells. Down depositional dip, fluids would have encountered the proto-HFF, where they could have discharged. Elevated heat flux associated with rift initiation would have both heated fluids and provided a conduit for egress of buoyant fluids through the proto-HFF (Fig. 13). Jackson *et al.* (2005) indicate that the tip of the HFF and steeply dipping normal fault splays developed in the Thebes Formation during rift initiation, and this would have allowed fluids to migrate into both the proto- hanging wall and footwall within the Thebes Formation. This would have been facilitated by an order of magnitude permeability contrast between the Thebes Fm. and the bounding formations. Flux may have been further enhanced by development of a narrow damage zone in proximity to the evolving HFF.

Such a conceptual model is not without precedent. Pearson and Garven (1992) modelled fluid flow within continental rift basins, and noted draw-down of fluids at fault escarpments with discharge in the centre of the rift. Jones and Xiao (2013) also modelled downward fluid flux along faults as controlling carbonate dissolution and cementation within pre-rift lacustrine carbonates, with more extensive carbonate dissolution in the vicinity of faults, where fluid is drawn down, and cementation where hot fluids are vented to the surface. Much of this work is model-based, however, and has seen application in only a few field studies. For example, using field, petrographical and geochemical proxies, Rusticelli *et al.* (2016) and Wilson *et al.* (2007) proposed dolomitization from seawater drawn down surface-breaching faults, whilst Corbella *et al.* (2014) modeled a basin-scale convection of seawater down faults and via a fracture basement, constrained by field observations and geochemistry. It is likely, therefore, that the conceptual model we present for fault-controlled fluid flux is more common than has been recognized within the literature. The limited number of examples perhaps reflects the fact that many case studies of fault-controlled dolomitization are in older sediments that have been overprinted by later phases of dolomitization (Hollis et al., 2017). However, it could also demonstrate the persistence of a conceptual framework whereby fluids are only considered as moving up faults. Given that many rift basins evolve through the localization of deformation from numerous small faults to a single, deep-seated crustal lineament is highly likely that similar patterns of fluid flow occur on carbonate platforms in extensional basins.

Implications to the composition of dolomitizing fluids

Seawater is the most volumetrically abundant Mg-enriched fluid for dolomitization. On geochemical evidence, it has been argued that stratabound dolomitization on the HFF took place from slightly evaporated seawater. Given that we propose convection of seawater via a basal clastic aquifer, however, the potential for modification of seawater during fluid migration has to be considered. In particular, fluid-rock interaction within the Nubian Sandstone aquifer, where fluid residence times were high, could have modified $\delta^{18}O_{water}$, $^{87}Sr/^{86}Sr$, and trace element compositions. Both REE profiles and $\delta^{13}C$ imply high water-rock ratios, and so any modification of seawater composition must involve significant fluid-rock interaction. The Nubian Formation is dominated by quartz arenites (Pomeyrol, 1968; Weissbrod and Nachmias, 1986; Nabawy *et al.,* 2009), however, offering little opportunity for isotopic or trace element enrichment. Importantly, the concentration of Sr relates strongly to dolomite stoichiometry, suggesting that the any apparent enrichment in Sr is a result of a higher concentration of Ca (Fig. 6C). Indeed, if $^{87}Sr/^{86}Sr$ had been enriched, then the actual age of dolomitization must be older than the apparent (Oligo-Miocene) age since the $^{87}Sr/^{86}Sr$ of post-Miocene seawater is more enriched than the values measured for the stratabound dolostones (Fig. 12). Given the narrow constraints on the timing of dolomitization (26-15 Ma) from field relationships, and an absence of evidence for pre-Oligocene dolomitization, it seems unlikely that this is the case.

Mid Ocean Ridge basalts have differentially been reported as both a source and a sink for Mg in seawater (Ligi *et al.*, 2013). Hence, it is possible that Mg-enrichment of seawater occurred during rift initiation, when basaltic dykes and sills were emplaced in proximity to the HFF, and in the Northern Gulf Basaltic Province (22-26 Ma; Montenat *et al.,* 1986; Moustafa and Abdeen 1992; Patton *et al.,* 1994; Bosworth, 2015). A linear mixing of > 50% seawater ($\delta^{18}O_{water}$ =0‰ and $\delta^{13}C_{water}$ =0‰) with a magmatic fluid of $\delta^{18}O_{water}$ = +7‰ and $\delta^{13}C_{water}$ = -7‰ would result in a fluid with $\delta^{18}O_{water}$ = 0 to +3.5‰ and $\delta^{13}C_{water}$ = -3.5 to 0‰ (Zheng and Hoefs, 1993), close to values determined for the stratabound dolostone. Normally, magmatic fluids are depleted in $^{87}Sr/^{86}Sr$, but Bosworth (2015) measured enrichment in radiogenic strontium, uranium and lead as well as low concentrations of MgO (5.6-6.3wt%) in the basalts from the Northern Gulf Basaltic Province. Consequently, enrichment in $^{87}Sr/^{86}Sr$ might have occurred as a result of seawater mixing with magmatic fluids, but this would mean that the actual age of dolomitization is older than interpreted, which is inconsistent with field observations. Furthermore, the volume of basalts within the study area is relatively low (Bosworth *et al.,* 2015) meaning it is unlikely that they were able to significantly modify fluid composition. The low concentrations of iron within the dolostone supports this notion, since any contribution by magmatic fluids to dolomitizing fluids would most likely increase concentrations of iron in the resultant dolomite.

Controls on dolomitization

Preferential dolomitization of the debrite and turbidite facies suggests that there was a facies control on the occurence of dolomitization, either because a) debrites and turbidites had higher permeabilities, and so focused fluid flux or b) there were differences in the mineralogy and texture of facies that controlled their dolomitization potential. The formation of discrete stratabound dolostone bodies has been recognised in many studies of fault-controlled dolomitization, and is often interpreted to reflect preferential dolomitization of highly permeable layers (Davies and Smith, 2006; Sharp *et al.,* 2010; Gomez-Rivas *et al.,* 2014; Corbella *et al.,* 2014). In this study, petrophysical measurements of remobilised limestone facies indicate that their present day matrix permeability is only moderately higher than the undolomitized, interbedded, slope packstones and less than the basinal wackestone facies of the upper Thebes Formation. Since dolomitization took place within sediments that had reached maximum burial, the measured porosity and permeability of the limestones is considered to be largely representative of their matrix properties during dolomitization, although there was some subsequent porosity, and therefore permeability, loss as a result of telogenetic cementation.

Korneva *et al.,* (2017) describe the distribution of fractures within limestone and stratabound dolostone within the damage zone of, and at 2 km from, the HFF. During formation of the stratabound dolostone, fractures and faults would have been present at the tip of the HFF (Jackson *et al.,* 2005), but offset on the HFF was minimal and therefore the damage zone would have been narrow. Therefore present day fracture density at 2km distance from the HFF might give some indication of likely fracture density during dolomitization. The highest fracture densities (narrowest spacing) is in matrix-supported conglomerates (R1), followed by slope packstones (S1) with the lowest density in turbidite grainstones (R2). It is possible, therefore, that the fracture network played an important role in the fluid flux away from the proto-HFF during dolomitization, perhaps facilitating dolomitization in the R1 facies.

Nevertheless, sediment texture and composition may also have played a role. Since the debrites and grainstone turbidites (R1 and R2) are coarser grained than the slope packstones (S1) it seems unlikely that reactive surface area influenced the loci of dolomitization, although micritization of grains could have increased the reactive surface area of some grains. It is possible that the diverse range of shallow water grains within R1 and R2 facies, compared to S1 facies, facilitated dolomitization preferentially in remobilized beds. Since mineralogical stabilization would have occurred prior to dolomitization it is unclear how this control might have operated, but one possibility is that remnant microdolomite inclusions within former high magnesium calcite grains acted as seeds for dolomitization (e.g. Juerges *et al.,* 2016).

Evidence for hydrothermal dolomitization

Conceptual models of fault-controlled dolomitization typically interpret dolomitization to have occurred from hydrothermal, evolved, crustal fluids that are expelled at high pressure along faults, by seismic pumping, into the host rock (Davies and Smith, 2006). They form a number of characteristic features, including non-planar replacive dolomite, zebra dolomite and hydraulic breccias cemented by saddle dolomite, interpreted to reflect high pressure fluid expulsion and fluid effervescence (Davies and Smith, 2006; Gasparrini *et al.,* 2006; Nader *et al.,* 2007; López-Horgue *et al.,* 2010; Sharp *et al.,* 2010; Dewit *et al.,* 2014). Machel and Lonnee (2002) criticised the misrepresentation of the term ‘hydrothermal’ in many of these conceptual models and defined its use as strictly referring to dolostone bodies where dolomitizing fluids were > 5°C warmer than the host rock. On the HFF block, we interpret the stratabound dolostone bodies to be fault-controlled in the sense that dolomitizing fluids entered the Thebes Formation from the proto-HFF. Assuming that the base of the Thebes Fm. was at ~600 – 850 m during dolomitization, an elevated geothermal gradient of 48°C km^{-1} and an ambient seawater temperature of 30°C, the Thebes Formation would have been < ~59 - 71°C. Since the temperature of dolomitizing fluids has been determined from clumped isotopes as 67-76°C, and from conventional isotopes as up to 83 °C, then we can consider the stratabound dolostone bodies to marginally to fully hydrothermal.

Nevertheless, many of the 'classic' features of fault-controlled hydrothermal dolomite are not observed. Zebra dolomite textures (Fig. 8) are rare and mostly occur beneath chert nodules. It has been suggested that these features formed during the onset of transtension along the Aqaba-Levant transform, forming perhaps by pressure build-up beneath low permeability lenses. They do not seem to be related to the main phase of stratabound dolomitization. Minor hydrobrecciation is seen within the damage zone of the HFF (Hirani *et al.,* in prep) but not in association with the stratabound dolostone. Saddle dolomite is conspicuously absent, despite fluid temperatures of >60°C, suggesting that crystal growth rates were too slow and/or fluid saturations too low for rapid precipitation of dolomite with a saddle morphology.

Consequently, the results of this study indicate that fault-controlled, hydrothermal dolomitization cannot be interpreted on the basis of what are commonly perceived to be characteristic textural features. Indeed, in a subsurface environment, the textures encountered within the stratabound dolostone might result in a misinterpretation of their origin, and hence spatial distribution and length scales. Our conceptual model, based on the integration of field mapping, structural relationships, petrographical and geochemical data instead provides evidence for fault-controlled convection of seawater at elevated temperatures but without the high pressure expulsion, rupture and effervescence commonly associated with fault controlled dolomitization (*sensu* Davies and Smith, 2006). Instead, the results suggest that geothermal convection, which is known to be an important control on the syn-depositional diagenetic modification of platform margins (e.g. Whitaker and Xiao, 2010; Jones and Xiao, 2006, 2013), can control diagenesis after burial in tectonically-active regions. Furthermore, it is possible for seawater to move up evolving, crustal-scale faults that have not undergone significant offset, or breached the surface, during the earliest phases of rifting. A significant advantage of this model is that it has fewer constraints with respect to fluid and Mg mass balance and so is able to more adequately explain the volume of in place dolostone than more commonly invoked models of fault-controlled dolomitization.

Overall, the relatively simple burial and uplift history of the Thebes Fm. provides an insight into a dolomitization process that might commonly develop within syn-rift basins. It might be obscured by subsequent diagenetic overprinting in basins that have a more complex, and long-lived, burial history. Alternatively, it might not be recognized as in some cases, such as this case study, the process will operate in pre-rift sediments during the upward-propagation of crustal-scale faults.

Conclusions

The results of this study provide an insight into fluid flux and diagenetic processes during rift initiation.

- Stratabound dolostone bodies on the HFF Block formed almost exclusively within debrites and grainstone turbidite facies of the lower Thebes Fm. from heated, slightly evaporated Oligo-Miocene seawater that circulated along the plane of the HFF during the onset of rifting.
- Faults appear to have provided permeable pathways that connected seawater with the deep, Nubian aquifer, enabling development of convective circulation driven by geothermal heating. These fluids ascended the proto-HFF, facilitated by heat generated during initial rifting. The resulting buoyant fluids utilised the fault as a conduit, escaping upwards and then flowing laterally into the lower Thebes Fm. Evidence for this is only exposed in the footwall of the HFF, but conceptual and flow models would support a similar distribution of stratabound dolostone in the hanging wall.
- The main control on the formation of stratabound dolostone could be the depth of the tip of the HFF, and the higher permeability of the Thebes Fm. compared to the bounding formations. Within the Thebes Fm., dolomitization appears to have been influenced by fracture permeability and potentially the chemical reactivity of the precursor dolomitized facies.
- Dolomitization has been constrained by field relationships and the structural evolution of the platform to a ~10 Ma period during the Oligo-Miocene, after rift initiation and prior to the rift climax.
- The conceptual model of dolomitization from seawater presented in this paper might be more common than recognized by the current literature. It has been observed in a few case studies from other basins, and it may well be a common process during the earliest stages of rifting, leading to selective dolomitization of pre-rift and possibly syn-rift carbonate platforms.

Acknowledgements

Stable isotope analysis was conducted under NERC award IP-1357-1112 at the NERC Isotope Geosciences Facility in East Kilbride. Strontium isotope analysis was conducted at University of Brest and Dr Stefan LaLonde is thanked for his help. This project was funded via ITF project 3310PSD by BG-Group, Saudi Aramco, Statoil and Total. Falcon of the Desert, Anja Eker, Michael Laukemann, Thomas Seers, Richard Newport and David Hodgetts are gratefully acknowledged for field support. Cedric John acknowledges the support of the Qatar Carbonate and Carbon Capture Research Center (QCCSRC) for instrumentation, sponsored by Qatar Petroleum, Shell, and the Qatar Science and Technology Park.

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Figure Captions

Figure 1 (A) Map showing the major structural elements of the Gulf of Suez and distribution of pre-rift and syn-rift sediments along the western Sinai coast (adapted from Younes and McClay, 2002 and Khalil and McClay, 2002). (B) Composite stratigraphic section of the Hammam Faraun Fault block (adapted from Sharp et al., 2000 and Jackson et al., 2006). This study focuses on the pre-rift Thebes Formation

Figure 2 (A) Geological map of the study area within the Hammam Faraun Fault block (modified from Moustafa and Abdeen, 1992 and Moustafa, 2003). The Hammam Faraun Fault parallels the present day coastline, with the Gebel Fault intersecting through the study area. (B) Stratigraphic column of the Thebes Fm. in the study area. The stratabound dolostone bodies are stratigraphically controlled, occurring primarily within the lower Thebes Fm.

Figure 3 Oblique view of study area (direction of view annotated in Fig. 2a) showing the distribution and extent of stratabound dolostone bodies relative to the Hammam Faraun Fault and the Gebel Fault, superimposed onto a Google Earth image. The origin of the massive dolostone bodies in the study area are further discussed in Hollis et al. (2017).

Figure 4. (A) and (B) Sharp, planar contacts (arrows) between dark brown dolostone bodies and adjacent cream coloured host limestone beds

Figure 5 Photomicrographs showing the petrographic features of the precursor limestones in plane polarised light. **(A)** Recrystallised grainflow grainstone, with both matrix and allochems replaced by micro spar calcite. **(B)** Unaltered basinal wackestones, with microporous mud matrix and minor recrystallisation of the internal structure of the pelagic foraminifera by mosaic calcite

Figure 6. (A) PAAS normalised REE concentrations comparing the REE patterns for averaged unaltered and recrystallised limestones to the stratabound dolostones. Positive La and negative Ce anomalies are present in each case, however a flattening in the HREE's is observed within the dolomite samples, as seen within the recrystallised limestone samples. A notable decrease in the total REE concentration of the stratabound dolomites is observed relative to the host limestones. (B) PAAS normalised (Pr/Pr*)SN versus (Ce/Ce*)SN cerium anomaly plot (Webb and Kamber, 2000) showing that all samples cluster tightly in the field of negative Ce and positive La anomalies. (C) Concentration of Sr compared to dolomite stoichiometry, showing an increase in Sr concentration with increasing CaCO3, as per Vahrenkamp and Swart (1990), shown by dotted line.

Fig. 7. (A) Stable isotope ($\delta^{18}O$ vs. $\delta^{13}C$) plot for unaltered limestones, recrystallised limestones, and stratabound dolostone samples. The whole rock isotopic composition for the unaltered limestones lie within the range that is expected for deposition from Eocene seawater (Pearson et al., 2001). In

comparison, the recrystallised limestone and dolostone samples exhibit lower δ18O values ranging between -3.62‰VPDB and -8.04‰VPDB. (B) Temperature of dolomitising fluids versus δ^{18}Oseawater plot for the stratabound dolostone bodies. This plot is constructed using the Matthews and Katz (1977) fractionation factor and can be used to determine the minimum and maximum temperature of dolomitising fluids using the heaviest and lightest δ18Odolomite values. Assuming a δ^{18}Oseawater of +1‰ to 0‰ (Veizer and Prokoph, 2015), a minimum fluid temperature of 40°C and a maximum fluid temperature of 78°C would be expected. (C) Clumped isotope temperatures (y-axis, ˚C) versus calculated δ^{18}Owater (x-axis, ‰ VSMOW) for three dolomite samples in this study. The shaded area represents the 68% confidence interval for each sample. Dashed lines represent the temperature range at which all three samples overlap at the 68% confidence interval (69-78˚C); this corresponds to a calculated fluid composition of approximately 2.5‰ VSMOW.

Figure 8 Macroscopic features of the stratabound dolostone bodies in the field. (A) Dark brown clast and grey matrix fabric within the stratabound bodies, both composed of dolomite. (B) Randomly orientated zebra dolomite-like fabrics, with elongated vuggy porosity. (C) Dogtooth calcite cement partially occluding vuggy porosity within zebra fabric. (D) Large botryoidal calcite cement partially cementing vuggy porosity. (E) Mouldic pores after clasts (arrow) of variable size. (F) Stratabound dolostone body lacking the presence of clast, clast pores or large vuggy pores, suggesting dolomitisation of grainstone turbidite (R2) facies.

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Figure 13. Conceptual fluid flow model which interprets that during the late Oligocene, seawater was drawn down incipient faults within the proto-footwall of the Hamman Faraun Fault and circulated by geothermal convection along the Nubian aquifer until they reached the proto-HFF. Fluids moved up the fault and were expelled laterally into the basal Thebes Formation where they migrated laterally.

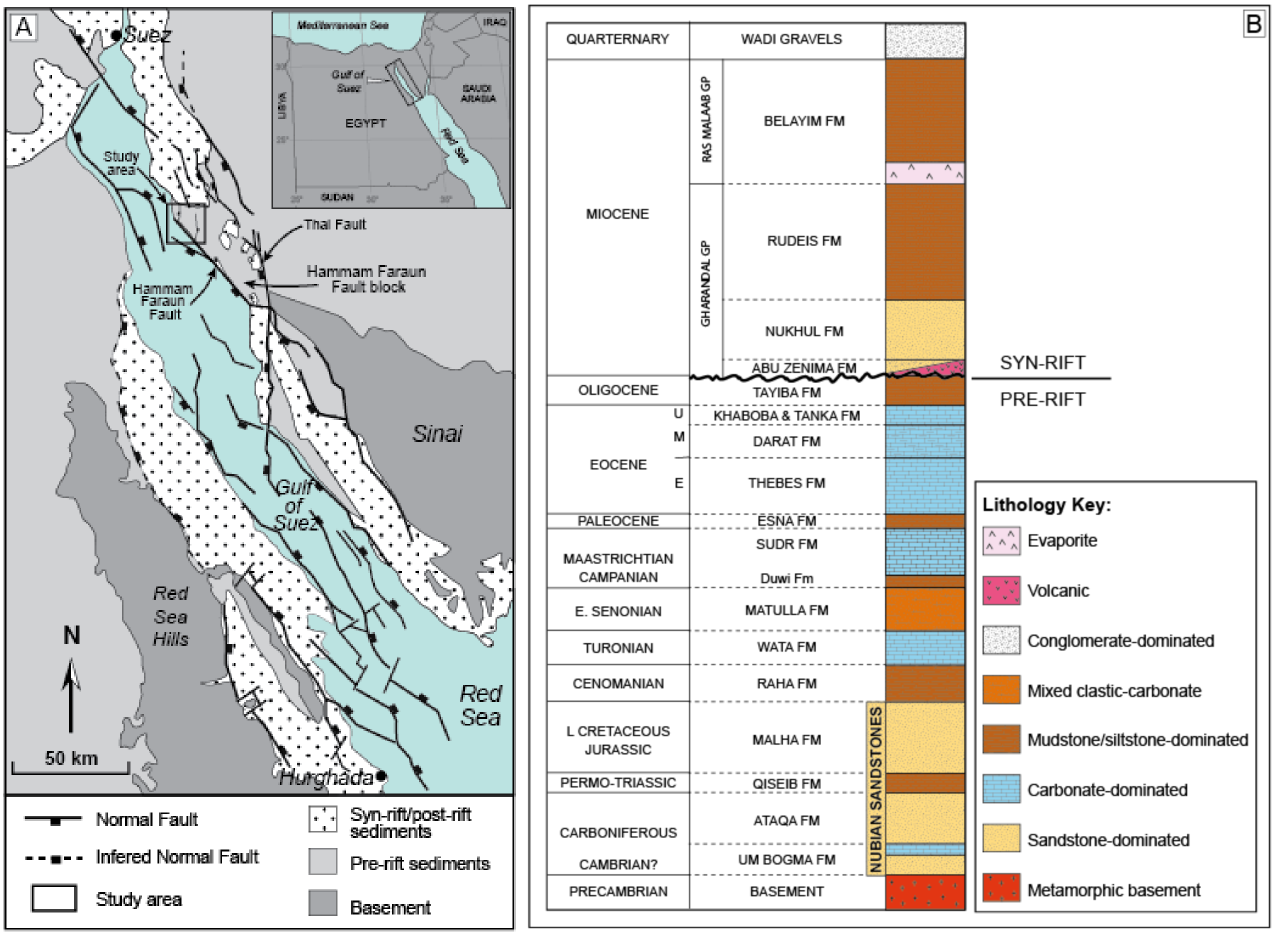
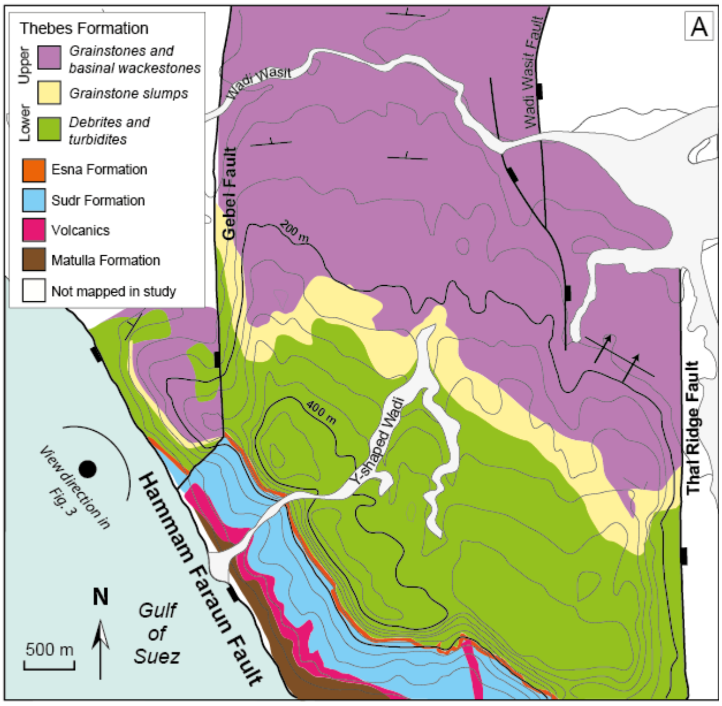


Fig. 1. (A) Map showing the major structural elements of the Gulf of Suez and distribution of pre-rift and syn-rift sediments along the western Sinai coast (adapted from Younes and McClay, 2002 and Khalil and McClay, 2002). **(B)** Composite stratigraphic section of the Hammam Faraun Fault block (adapted from Sharp *et al*., 2000 and Jackson *et al*., 2006). This study focuses on the pre-rift Thebes Formation.



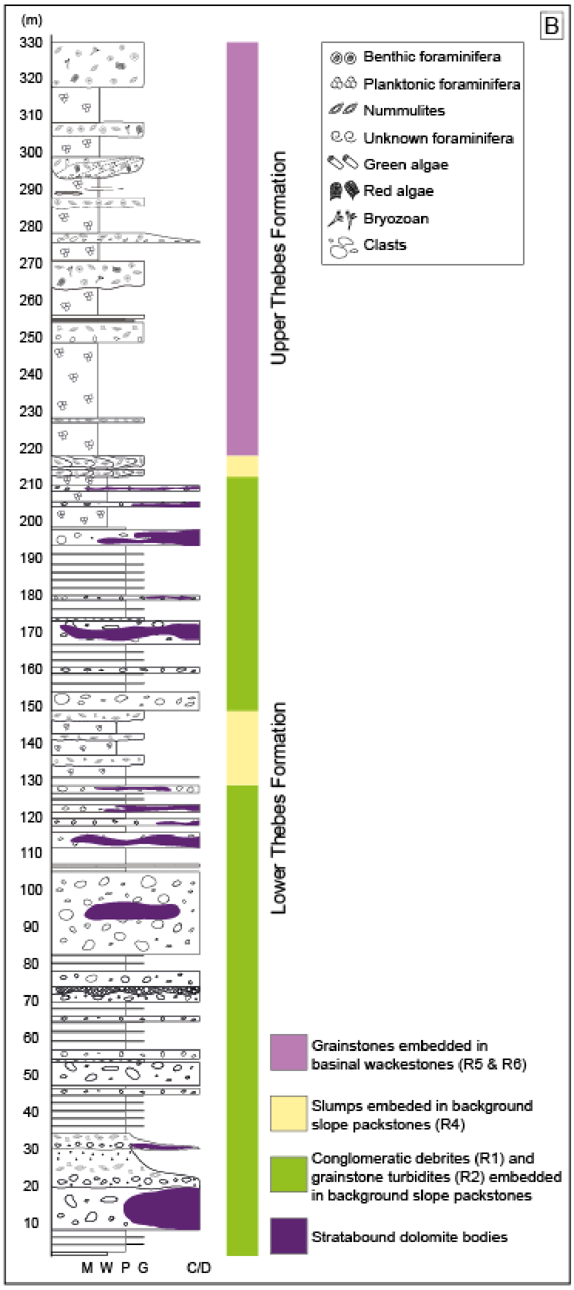


Fig. 2. (A) Geological map of the study area within the Hammam Faraun Fault block (modified from Moustafa and Abdeen, 1992 and Moustafa, 2003). The Hammam Faraun Fault parallels the present day coastline, with the Gebel Fault intersecting through the study area. (**B**) Stratigraphic column of the Thebes Fm. in the study area. The stratabound dolostone bodies are stratigraphically controlled, occurring primarily within the lower Thebes Fm.

Deposit type	**Facies type**	**Facies Code**	**Basal contact**	**Upper contact**	**Depositional texture**	**Allochems**	**Porosity**	**Notes**
In-situ deposit	Basinal wackestones	B1	Regular	Irregular or scoured	Wackestone	planktonic foraminifera, small *Nummulites,* uni- and bi-serial foraminifera, echinoid fragments	Microporosity in matrix	Sedimentary laminations may be visible with layered foraminifera
In-situ deposit	Slope packstones	S1	Regular	Irregular or scoured	Packstone, intercalated with thin grainstone beds	*Nummulites*, *Dyscocyclina, Alveolinids, Operculina*, planktonic foraminifera, echinoid fragments, bryozoan fragments	Microporosity in matrix, intragranular	Commonly recrystallised skeletal material and mud matrix.
Remobilised deposit	Matrix-supported conglomerate in slope deposits	R1	Irregular	Rounded	Matrix - Foraminifera wackestone Clast - Skeletal pack-grainstone	Matrix - mixed *Nummulites, Alveolinids*, planktonic foraminifera, echinoid fragments. Some clasts Clasts - *Nummulites, Operculina*, *Alveolinids*, bryozoa fragments, echinoid fragments	Matrix - microporosity Clasts - low to none	Subangular to subrounded clasts. Some clasts are composed of neomorphic calcite cement, fabric destructive and/or retentive. Some clasts are dolomitised.
Remobilised deposit	Foraminifera grainstone turbidite in slope deposits	R2	Irregular	sharp	foraminifera grainstone	*Nummulites, Alveolinids, Miliolids, Operculina*, serpulid worm tubes, echinoid fragments, red algal fragments, bryozoan fragments	intergranular	Little to no matrix. Intergranular porosity is commonly plugged by calcite cement. Evidence of fining upward cycles
Remobilised deposit	Clast-supported debris sheet flow in slope deposits	R3	scoured	sharp	Matrix - wackestone Clasts - Algal foraminifera grainstone	Matrix – small Nummulites Clasts - *Miliolids, Soritids, Nummulites, Alveolinids,* dasyclad algae, red algae, algal filaments, bryozoan fragments .	Matrix - high microporosity Clasts - Minor mouldic and intercrystalline	rounded to subrounded clasts. possibly debris sheet flow
Remobilised deposit	Slump grainstones in slope deposits	R4	scoured	regular	pack-grainstone	dasyclads, *Nummulites, Operculina, miliolids, echinoid fragments, green algae*	little to none	convolute bedding. Porosity often cemented by calcite. Contains little to no mud matrix

Deposit type	Facies type	Facies Code	Basal contact	Upper contact	Depositional texture	Allochems	Porosity	Notes
Remobilised deposit	grainflow grainstones in basinal deposits	R5	Irregular	sharp	grainstone	*Nummulites*, *Alveolinids, Miliolids, Operculina*, echinoid fragments, red algal fragments, bryozoan fragments	mouldic and intergranular	Little to no matrix. Geobodies often mimetically silicified
Remobilised deposit	Channel grainstones in basinal deposits	R6	scoured	sharp	grainstone	dasyclad algae, *Nummulites, Miliolids, Alveolinids, Soritids,* red algae, bryozoans fragments, serpulid worm tubes, echinoid fragments, mollusc fragments	minor intercrystalline	Lateral accretionary surfaces visible in cross-section. Channels bypass grainflow deposits, no evidence of cross-cutting relationships

Table 1. Facies descriptions for the Thebes Formation in the study area (after Corlett *et al*., in prep). The Lower Thebes Formation is composed primarily of facies S1, R1, R2, R3 and R4, and the Upper Thebes Formation is made up of facies B1, R5 and R6.

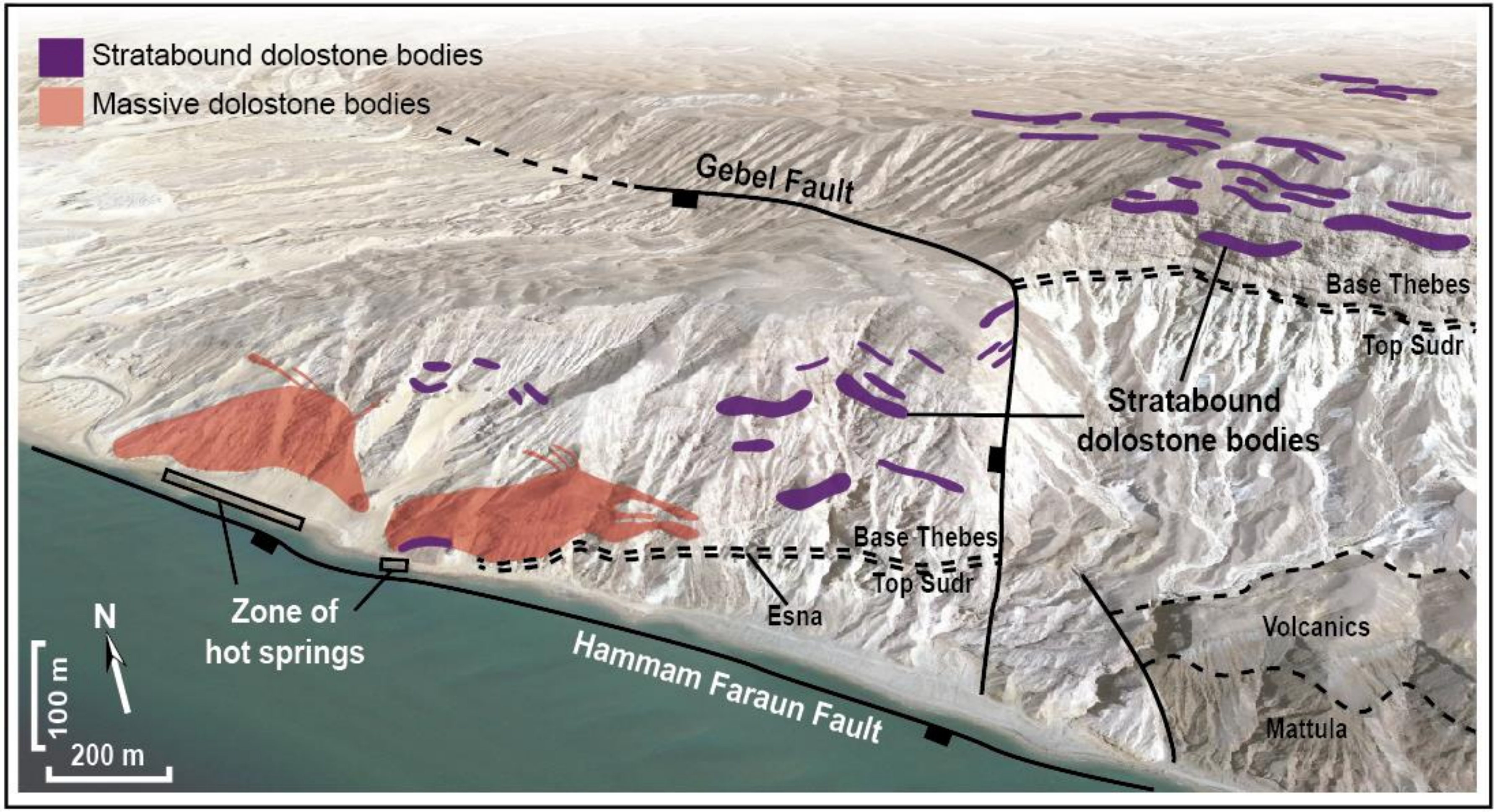


Fig. 3. Oblique view of study area (direction of view annotated in Fig. 2a) showing the distribution and extent of stratabound dolostone bodies relative to the Hammam Faraun Fault and the Gebel Fault, superimposed onto a Google Earth image. The origin of the massive dolostone bodies in the study area are further discussed in Hollis *et al.* (2017).

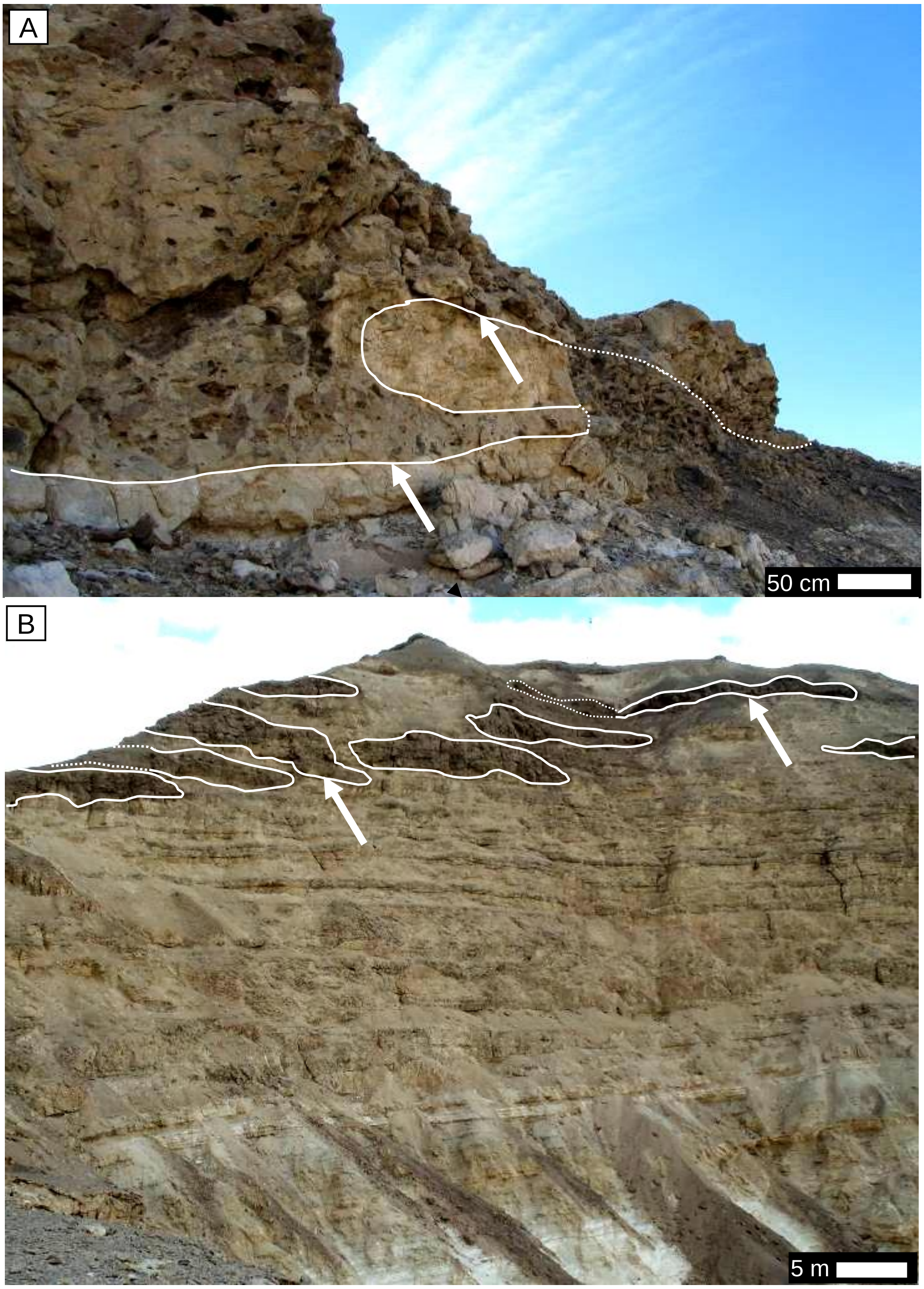


Fig. 4. (A) and **(B)** Sharp, planar contacts (arrows) between dark brown dolostone bodies and adjacent cream coloured host limestone beds.

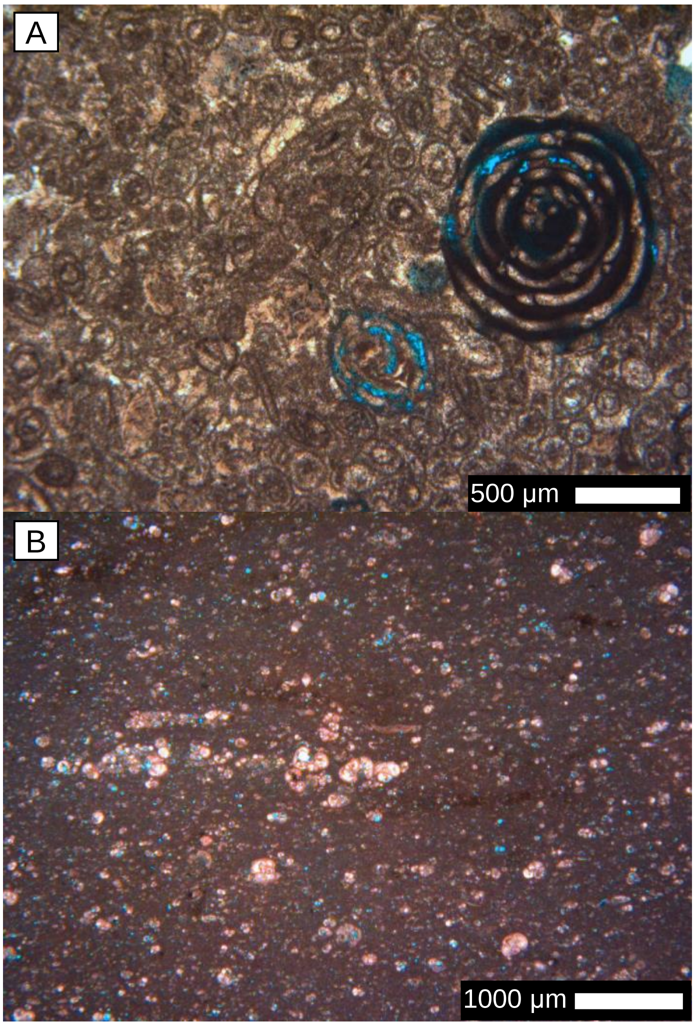


Fig. 5. Photomicrographs showing the petrographic features of the precursor limestones in plane polarised light. **(A)** Recrystallised grainflow grainstone, with both matrix and allochems replaced by micro spar calcite. **(B)** Unaltered basinal wackestones, with microporous mud matrix and minor recrystallisation of the internal structure of the pelagic foraminifera by mosaic calcite.

Phase	Stoichiometry (mole% $CaCO_3$)	Mg (ppm)	Ca (ppm)	Fe (ppm)	Mn (ppm)	Al (ppm)	Sr (ppm)	Ni (ppm)	Ba (ppm)	$d^{13}C$ ‰ VPDB	$d^{18}O$ ‰ VPDB	$d^{18}O$ ‰ VSMOW	87Sr/86Sr
Unaltered limestones													
n	-	3	3	3	3	3	3	3	3	3	3	3	1
min	-	2127	304691	1753	14.5	1928	933	29.5	8.0	-1.00	-3.50	27.30	-
mean	-	6327	313390	2182	16.9	2150	1240	40.3	20.7	-0.17	-3.35	27.45	0.707814
max	-	12362	320615	2648	21.5	2305	1775	48.8	32.5	0.70	-3.10	27.70	-
st dev	-	5359	8064	449	4.0	197	465	9.8	12.3	0.85	0.22	0.22	-
Recrystallised limestones													
n	-	3	3	3	3	3	3	3	3	16	16	16	4
min	-	2523	349901	97	43	134	415	2.6	4.6	-0.7	-7.8	22.9	0.707773
mean	-	5108	358407	145	138	220	576	6.6	11.7	0.4	-6.4	24.3	0.707849
max	-	6534	367722	217	273	372	664	12.9	17.8	1.0	-5.1	25.7	0.708011
st dev	-	2243	8938	64	120	132	140	5.5	6.6	-0.6	-0.7	-0.8	-0.000109
Stratabound dolostones													
n	14	18	10	18	18	18	18	18	18	25	25	25	5
min	50.03	32287	191817	37	208	23	75	0.6	8	-1.67	-8.04	22.62	0.707992
mean	51.13	77819	223841	140	469	107	151	5.2	137	-0.22	-5.61	25.12	0.708123
max	53.07	92590	313765	411	991	191	232	25.0	590	1.09	-3.62	27.17	0.708241
st dev	1.10	16149	35897	112	232	60	46	6.1	163	0.82	0.94	0.97	9.79E-05

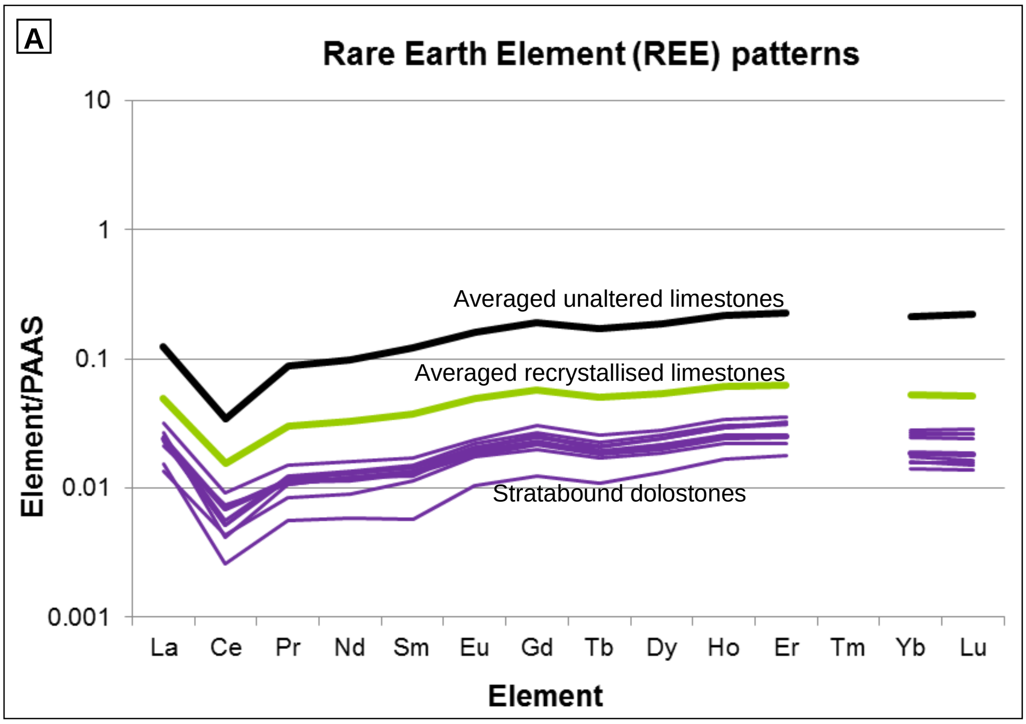
Table 2. Summary of geochemical data collected from unaltered limestones, recrystallised limestones and stratabound dolostones. Dolomite stoichiometry is determined using Lumsden's equation (1979). Trace element concentrations are measured using ICP-MS and ICP-AES. Stable isotopic concentrations are determined using a carbonate CO_2 extraction line, and the resulting gas is analysed by mass spectrometry. Strontium isotopic ratios are measured with a Triton Thermal Ionization Mass Spectrometer.

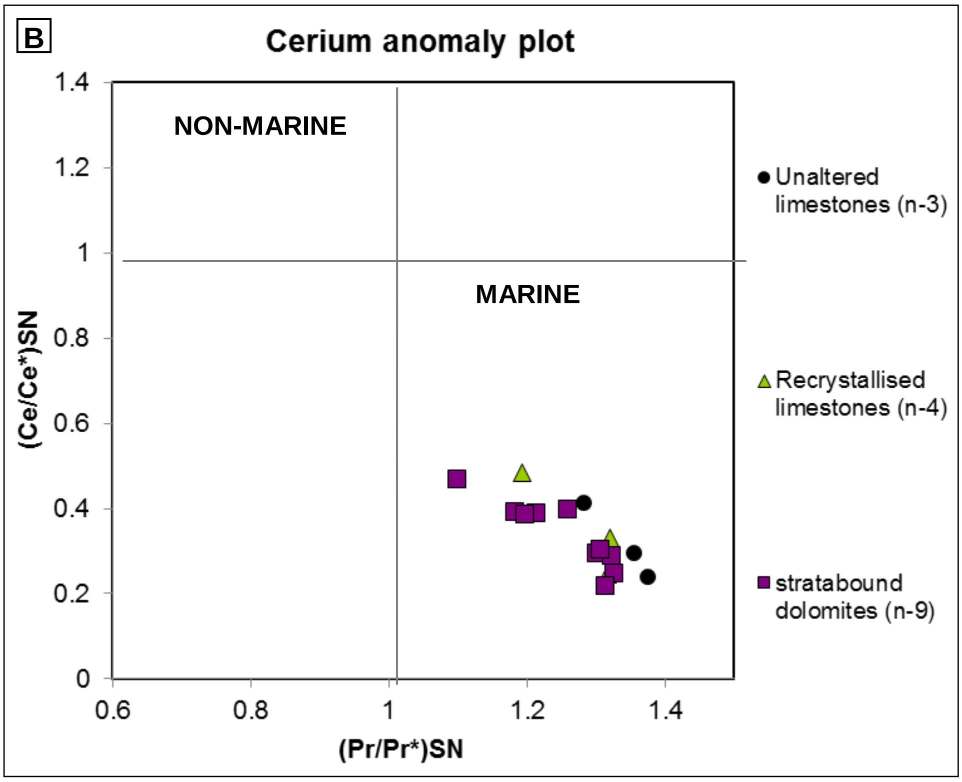
Sample	Stoichiometry (mole% $CaCO_3$)	$\delta^{18}O_{carbonate}$ (VPDB)[a]	$\delta^{13}C$ (VPDB)[a]	Δ_{47}[b]	T (°C)[b]	$\delta^{18}O_{water}$ SMOW
11B	50.6	- 4.59 (0.03)	- 0.72 (0.02)	0.561 ± 0.004	75 ± 2	4.17
34C	50.0	- 6.01 (0.36)	0.49 (0.02)	0.580 ± 0.008	65 ± 3	1.19
81A	53.1	- 4.86 (0.06)	1.06 (0.05)	0.575 ± 0.008	68 ± 4	2.85

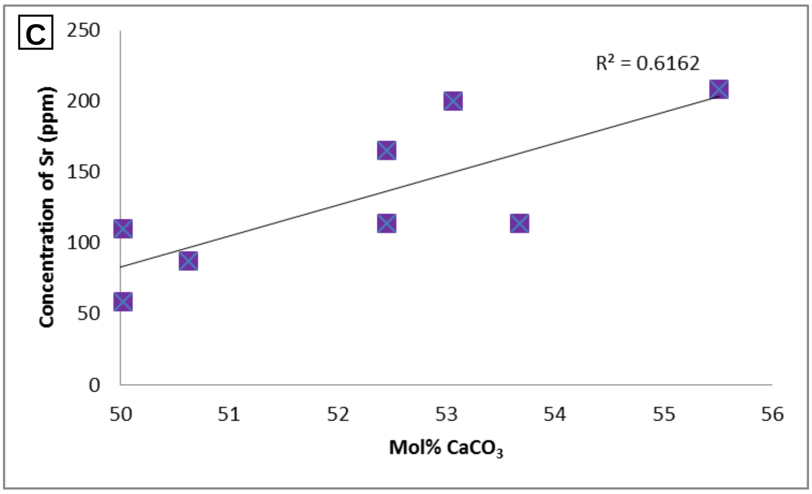
Table 3. Summary of isotopic composition, Δ_{47}, temperature and $\delta^{18}O_{fluid}$ for stratabound dolostones as determined from clumped isotope analysis. a= Error is 1σ for $\delta^{13}C$ and $\delta^{18}O$ carbonate, b= Error for Δ^{47} and temperature (T) is given as standard error, calculated by the replicate standard deviation divided by square root of n.

Formation	Age	Thickness (m)	Porosity (pu)	Kx (m^2)	Kx (mD)	Kz (m^2)	Kz (mD)	K
Nukhul	Miocene	200	0.20	1.00E-13	101	5.00E-14	50.7	5
Tanka	Upper Eocene	100	0.14	9.40E-16	0.952	9.80E-18	0.01	3
Darat	Middle Eocene	200	0.14	9.40E-16	0.952	9.80E-18	0.01	3
Thebes	Lower Eocene	400	0.20	2.20E-15	2.229	9.80E+18	0.01	3
Sudr	Upper Cretaceous	200	0.14	9.40E-16	0.952	9.80E-18	0.01	3
Matulla	Upper Cretaceous	100	0.33	3.80E-14	38.503	3.50E-16	0.35	3
Raha and Wata	Lower Cretaceous	300	0.18	7.90E-14	80	7.90E-14	80.00	3
Nubian	Carboniferous - Cambrian	600	0.32	3.90E-12	3952	2.00E-12	2026	4.5
Fractured basement	PreCambrian	1200	0.15	2.90E-13	294	2.90E-13	294	4
Basement	PreCambrian	1200	0.05	1.00E-19	0.000	1.00E-19	0.00	4

Table 4. Intrinsic rock properties used for input into fluid flow simulation. Data was derived from Hassanain *et al.* (2012) (Raha and Wata Formation), Mousa *et al.* (2011) (Matulla and Duwi Formation) and Nabawy *et al.* (2009) (Nubian Formation). Data for the Thebes Formation is from this study.

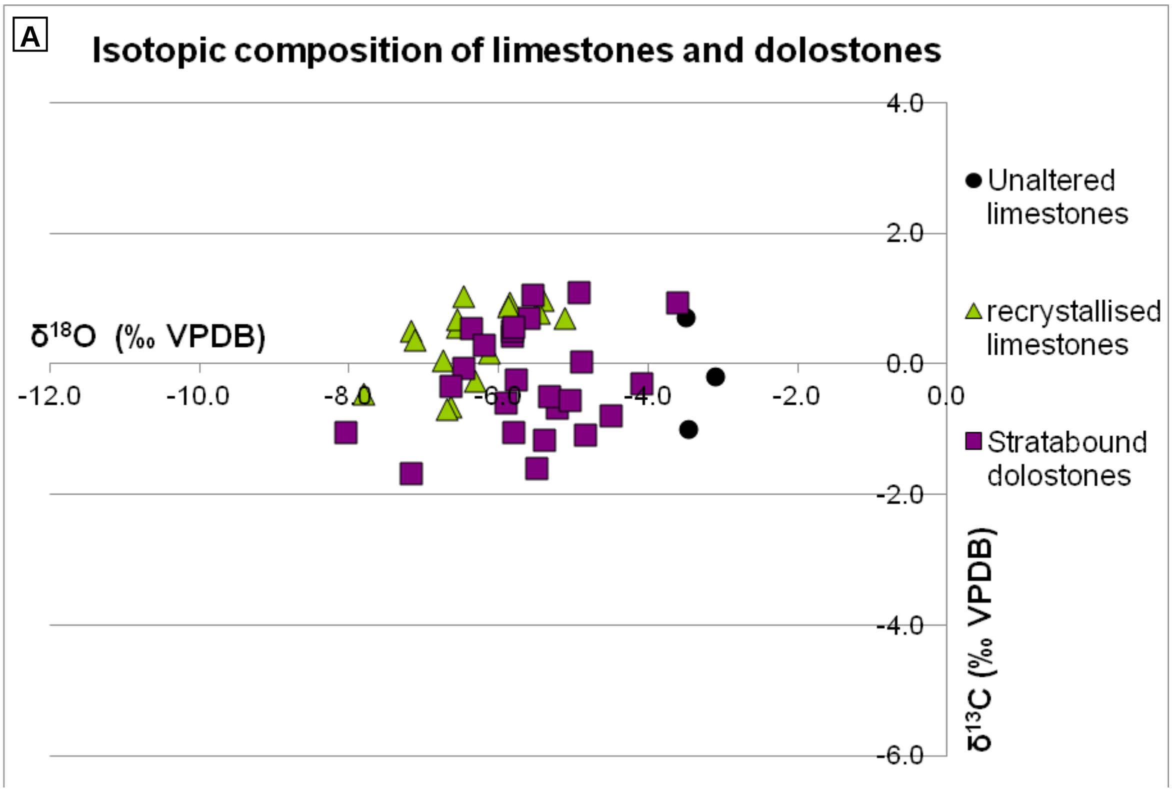

A
Rare Earth Element (REE) patterns
10
1
0.1
0.01
0.001
Element/PAAS
Averaged unaltered limestones
Averaged recrystallised limestones
Stratabound dolostones
La
Ce
Pr
Nd
Sm
Eu
Gd
Tb
Dy
Ho
Er
Tm
Yb
Lu
Element

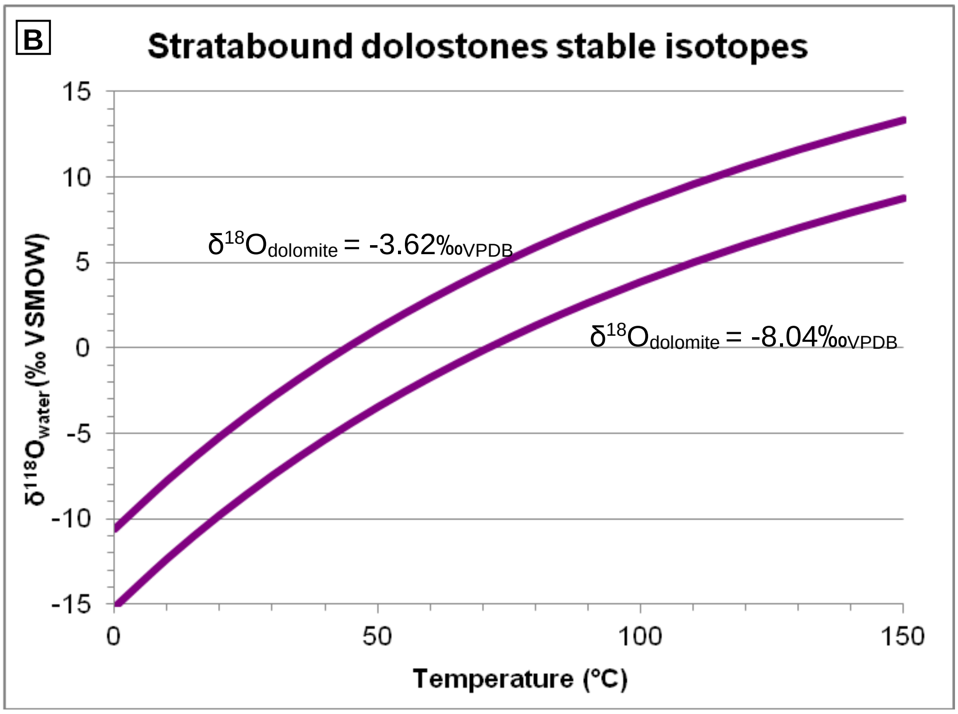

B
Cerium anomaly plot
1.4
1.2
1
0.8
0.6
0.4
0.2
0
(Ce/Ce*)SN
NON-MARINE
MARINE
0.6
0.8
1
1.2
1.4
(Pr/Pr*)SN
Unaltered limestones (n-3)
Recrystallised limestones (n-4)
stratabound dolomites (n-9)



GRAPH 6C – CHANGE TEXT FONT TO ARIAL AND COLOUR OF SYMBOLS TO PURPLE… BLUE CROSS WITHIN CAN BE SEEN THE SQUARE CAN BE SEEN.

Fig. 6. (A) PAAS normalised REE concentrations comparing the REE patterns for averaged unaltered and recrystallised limestones to the stratabound dolostones. Positive La and negative Ce anomalies are present in each case, however a flattening in the HREE's is observed within the dolomite samples, as seen within the recrystallised limestone samples. A notable decrease in the total REE concentration of the stratabound dolomites is observed relative to the host limestones. **(B)** PAAS normalised (Pr/Pr*)SN versus (Ce/Ce*)SN cerium anomaly plot (Webb and Kamber, 2000) showing that all samples cluster tightly in the field of negative Ce and positive La anomalies. **(C)** Concentration of Sr compared to dolomite stoichiometry, showing an increase in Sr concentration with increasing $CaCO_3$, as per Vahrenkamp and Swart (1990), shown by dotted line.


A
Isotopic composition of limestones and dolostones
δ18O (‰ VPDB)
δ13C (‰ VPDB)
4.0
2.0
0.0
-2.0
-4.0
-6.0
-12.0
-10.0
-8.0
-6.0
-4.0
-2.0
0.0
Unaltered limestones
recrystallised limestones
Stratabound dolostones


B
Stratabound dolostones stable isotopes
δ118Owater (‰ VSMOW)
15
10
5
0
-5
-10
-15
δ18Odolomite = -3.62‰VPDB
δ18Odolomite = -8.04‰VPDB
0
50
100
150
Temperature (°C)

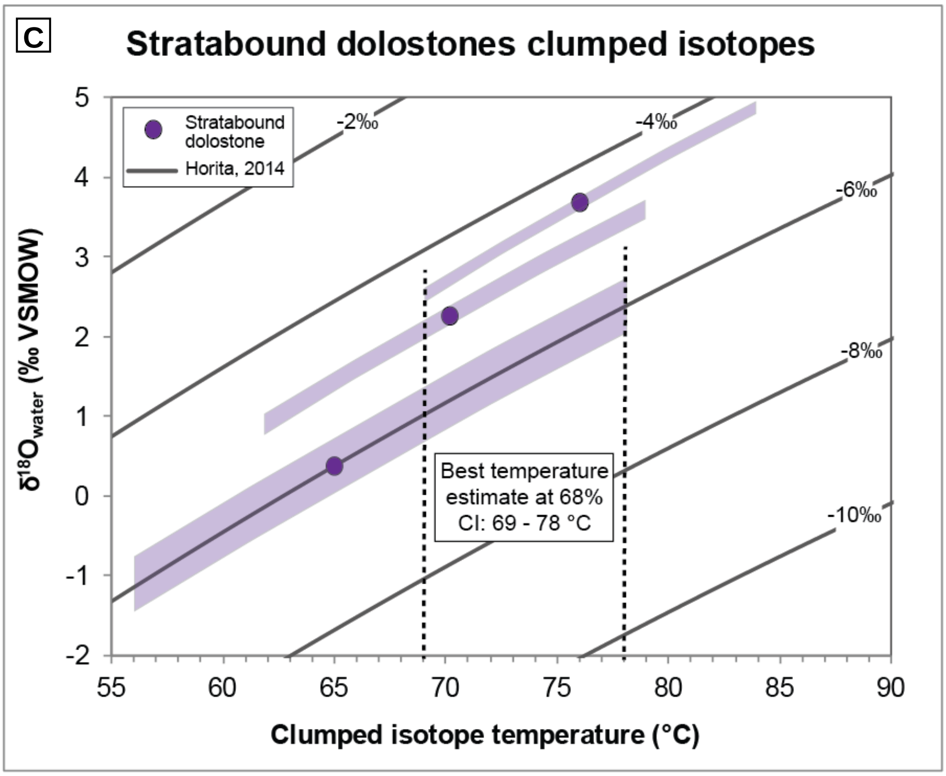
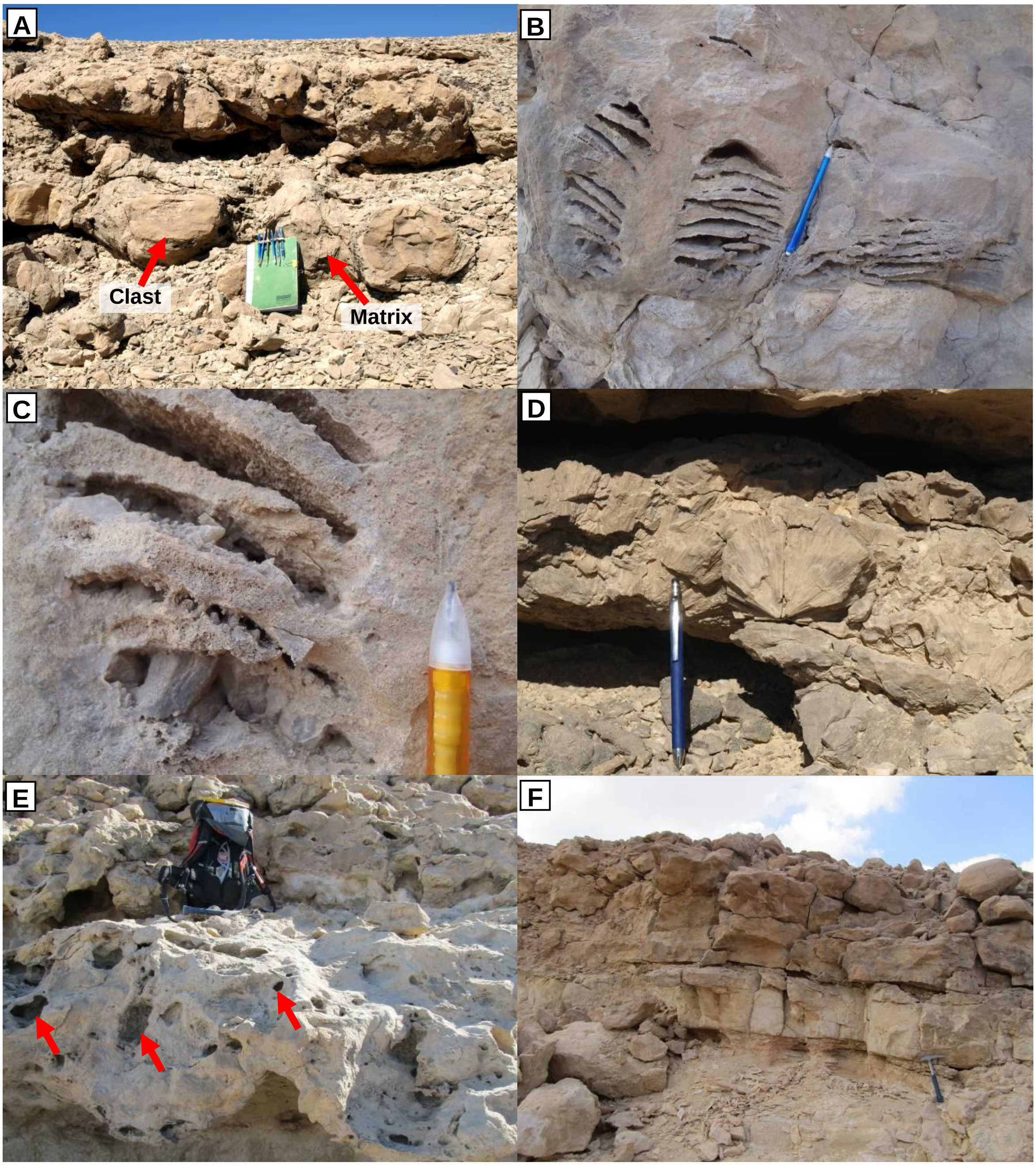


Fig. 7. (A) Stable isotope ($\delta^{18}O$ vs. $\delta^{13}C$) plot for unaltered limestones, recrystallised limestones, and stratabound dolostone samples. The whole rock isotopic composition for the unaltered limestones lie within the range that is expected for deposition from Eocene seawater (Pearson *et al.,* 2001). In comparison, the recrystallised limestone and dolostone samples exhibit lower $\delta^{18}O$ values ranging between -3.62‰$_{VPDB}$ and -8.04‰$_{VPDB}$. **(B)** Temperature of dolomitising fluids versus $\delta^{18}O_{seawater}$ plot for the stratabound dolostone bodies. This plot is constructed using the Matthews and Katz (1977) fractionation factor and can be used to determine the minimum and maximum temperature of dolomitising fluids using the heaviest and lightest $\delta^{18}O_{dolomite}$ values. Assuming a $\delta^{18}O_{seawater}$ of +1‰ to 0‰ (Veizer and Prokoph, 2015), a minimum fluid temperature of 40°C and a maximum fluid temperature of 78°C would be expected. **(C)** Clumped isotope temperatures (y-axis, ˚C) versus calculated $\delta^{18}O_{water}$ (x-axis, ‰ VSMOW) for three dolomite samples in this study. The shaded area represents the 68% confidence interval for each sample. Dashed lines represent the temperature range at which all three samples overlap at the 68% confidence interval (69-78˚C); this corresponds to a calculated fluid composition of approximately 2.5‰ VSMOW.



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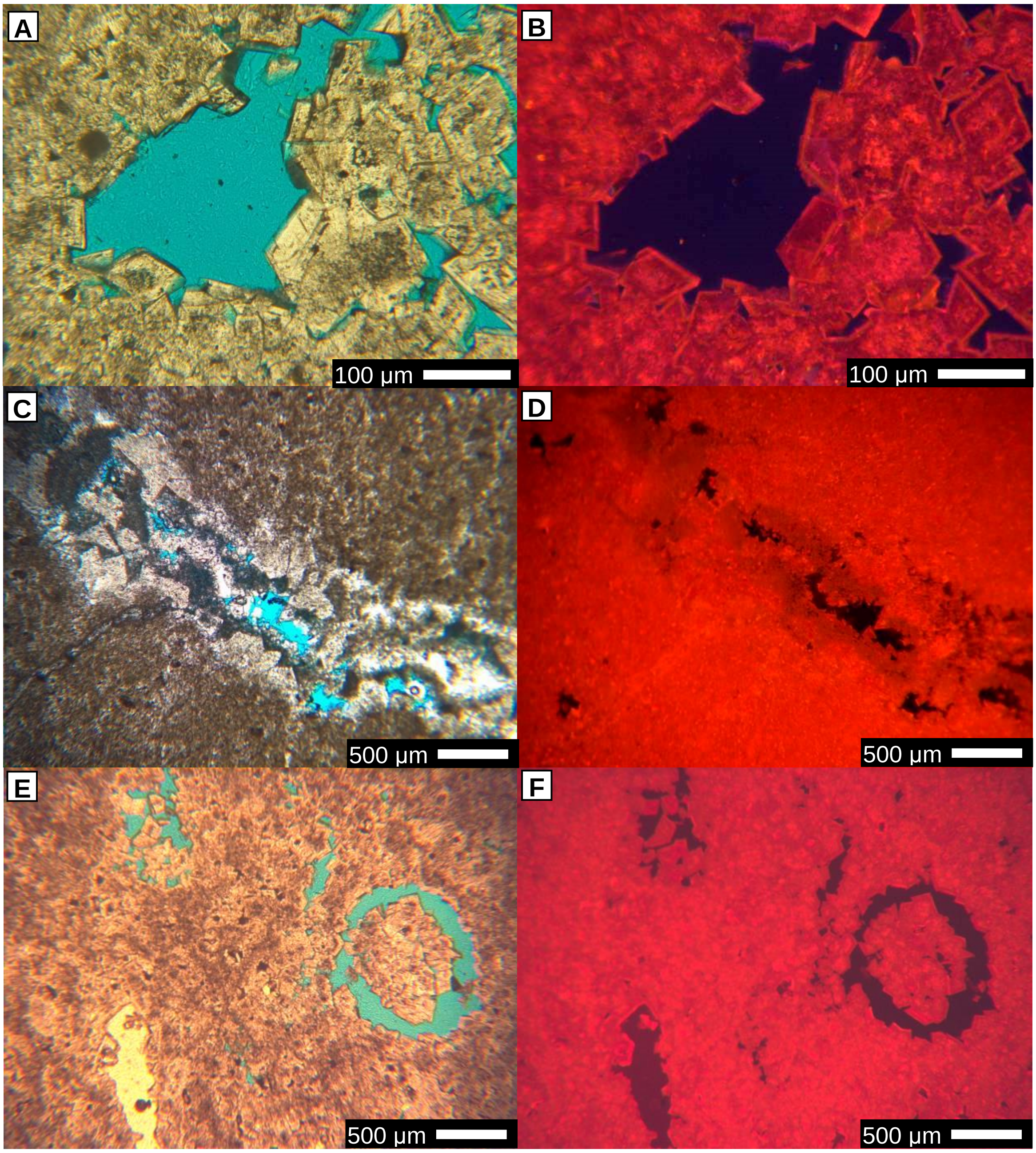

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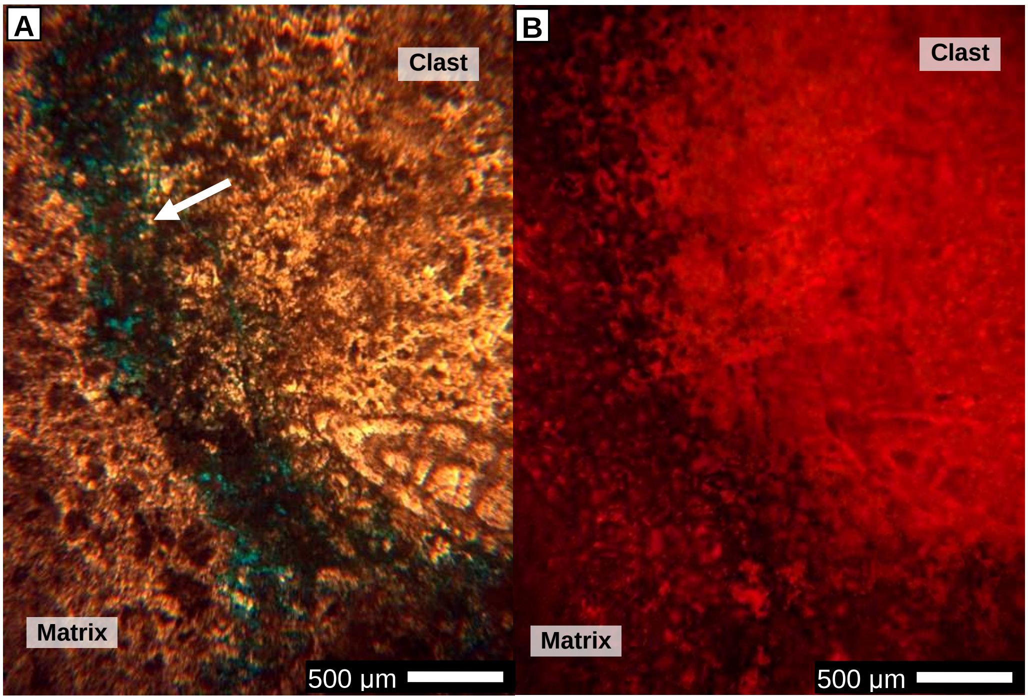

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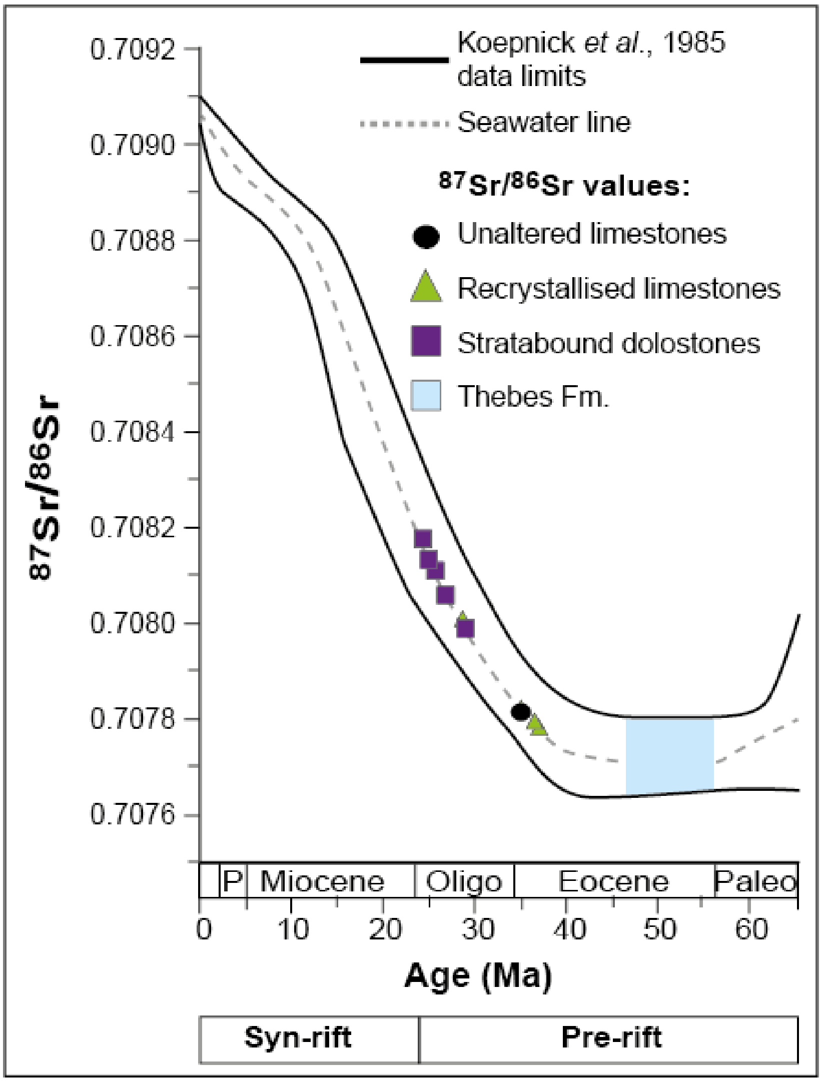

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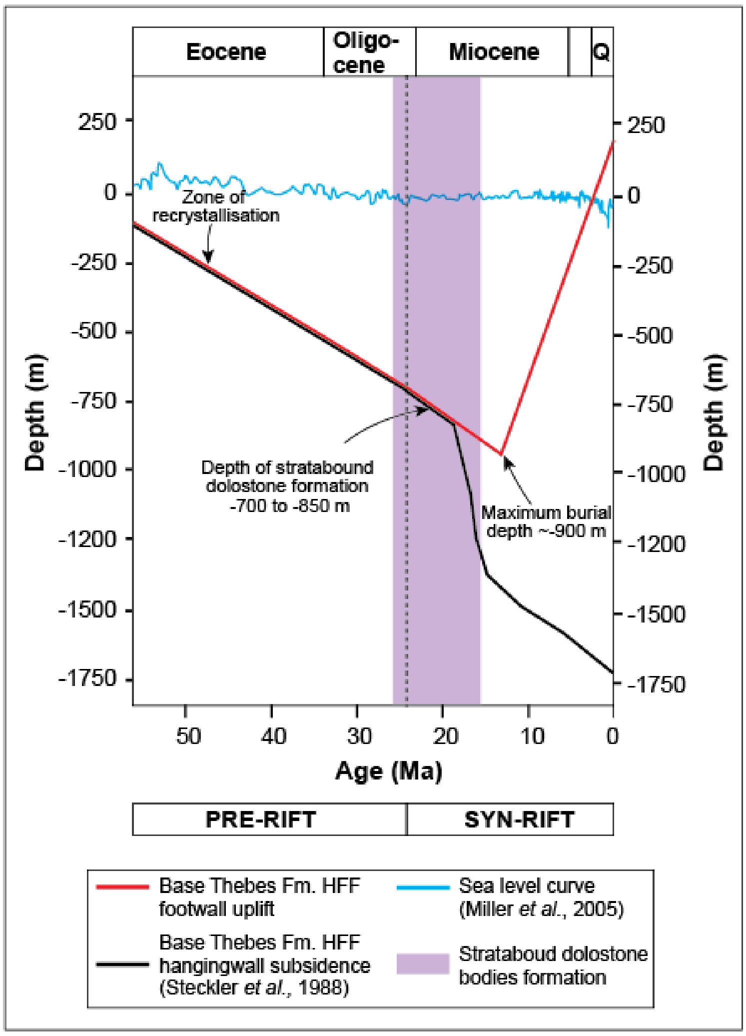

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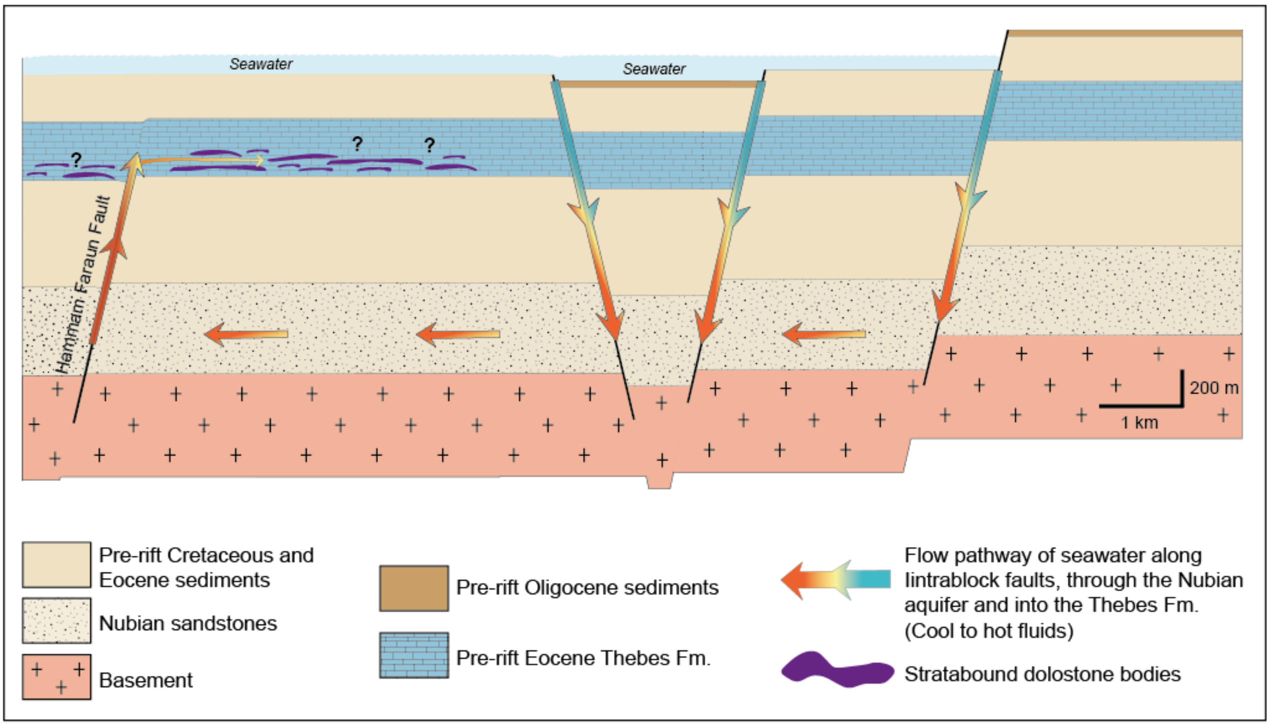

Fig. 13. Conceptual fluid flow model which interprets that during the late Oligocene, seawater was drawn down incipient faults within the proto-footwall of the Hamman Faraun Fault and circulated by geothermal convection along the Nubian aquifer until they reached the proto-HFF. Fluids moved up the fault and were expelled laterally into the basal Thebes Formation where they migrated laterally.

Deposit type	Facies type	Facies Code	Basal contact	Upper contact	Depositional texture	Allochems	Helium porosity (%)	Kh (mD)	Pore type	Notes
In-situ deposit	Basinal wackestones	B1	Regular	Irregular or scoured	Wackestone	planktonic foraminifera, small *Nummulites,* uni- and bi-serial foraminifera, echinoid fragments	24% (9-39%, n=7)	1.07 mD (0.04-3.5, n=7)	Microporosity in matrix	Sedimentary laminations may be visible with layered foraminifera
In-situ deposit	Slope packstones	S1	Regular	Irregular or scoured	Packstone, intercalated with thin grainstone beds	*Nummulites*, *Dyscocyclina, Alveolinids, Operculina*, planktonic foraminifera, echinoid fragments, bryozoan fragments	13% (9-35%, n=8)	0.30 mD (<0.01 - 1.7, n=8)	Microporosity in matrix, intragranular	Commonly recrystallised skeletal material and mud matrix.
Remobilised deposit	Matrix-supported conglomerate in slope deposits	R1	Irregular	Rounded	Matrix - foraminifera wackestone Clast - Skeletal pack-grainstone	Matrix - mixed *Nummulites, Alveolinids*, planktonic foraminifera, echinoid fragments. Some clasts Clasts - *Nummulites, Operculina*, *Alveolinids*, bryozoa fragments, echinoid fragments	9% (<1-17%, n=8) Dolomite = 4% (1-7%)	0.35 mD (<0.01 - 1.5, n=8) Dolomite = 0.12mD (<0.01 – 0.6 mD)	Matrix - microporosity Clasts - low to none	Subangular to subrounded clasts. Some clasts are composed of neomorphic calcite cement, fabric destructive and/or retentive. Some clasts are dolomitised.
Remobilised deposit	Foraminifera grainstone turbidite in slope deposits	R2	Irregular	sharp	foraminifera grainstone	*Nummulites, Alveolinids, Miliolids, Operculina*, serpulid worm tubes, echinoid	16.5% (16 -17%, n=2)	1.5mD (0.8 – 2.2, n=2)	intergranular	Little to no matrix. Intergranular porosity is commonly plugged by calcite cement. Evidence of fining upward cycles

						fragments, red algal fragments, bryozoan fragments	Dolomite = 2% (1-3%)	Dolomite = 0.2mD (<0.01 – 0.56mD)		
Remobilised deposit	Clast-supported debris sheet flow in slope deposits	R3	scoured	sharp	Matrix - wackestone Clasts - Algal foraminifera grainstone	Matrix – small Nummulites Clasts - *Miliolids, Soritids, Nummulites, Alveolinids,* dasyclad algae, red algae, algal filaments, bryozoan fragments .	Nd	nd	Matrix - high microporosity Clasts - Minor mouldic and intercrystalline	rounded to subrounded clasts. possibly debris sheet flow
Remobilised deposit	Slump grainstones in slope deposits	R4	scoured	regular	pack-grainstone	dasyclads, *Nummulites, Operculina, miliolids, echinoid fragments, green algae*	Nd	nd	little to none	convolute bedding. Porosity often cemented by calcite. Contains little to no mud matrix
Remobilised deposit	grainflow grainstones in basinal deposits	R5	Irregular	sharp	grainstone	*Nummulites*, *Alveolinids, Miliolids, Operculina*, echinoid fragments, red algal fragments, bryozoan fragments	10% (1 - 20%, n=16)	0.28mD (0.01 – 1.4, n=2)	mouldic and intergranular	Little to no matrix. Geobodies often mimetically silicified
Remobilised deposit	Channel grainstones in basinal deposits	R6	scoured	sharp	grainstone	dasyclad algae, *Nummulites, Miliolids, Alveolinids, Soritids,* red algae, bryozoans fragments, serpulid worm tubes, echinoid fragments, mollusc fragments	18% (3 - 10%, n=3)	0.3mD (0.01 – 0.8, n=3)	minor intercrystalline	Lateral accretionary surfaces visible in cross-section. Channels bypass grainflow deposits, no evidence of cross-cutting relationships

Table 1. Facies descriptions for the Thebes Formation in the study area (after Corlett *et al*., in prep). The Lower Thebes Formation is composed primarily of facies S1, R1, R2, R3 and R4, and the Upper Thebes Formation is made up of facies B1, R5 and R6.

Phase	Stoichiometry (mole% $CaCO_3$)	Mg (ppm)	Ca (ppm)	Fe (ppm)	Mn (ppm)	Al (ppm)	Sr (ppm)	Ni (ppm)	Ba (ppm)	d^{13}C ‰ VPDB	d^{18}O ‰ VPDB	d^{18}O ‰ VSMOW	87Sr/86Sr
Unaltered limestones													
n	-	3	3	3	3	3	3	3	3	3	3	3	1
min	-	2127	304691	1753	14.5	1928	933	29.5	8.0	-1.00	-3.50	27.30	-
mean	-	6327	313390	2182	16.9	2150	1240	40.3	20.7	-0.17	-3.35	27.45	0.707814
max	-	12362	320615	2648	21.5	2305	1775	48.8	32.5	0.70	-3.10	27.70	-
st dev	-	5359	8064	449	4.0	197	465	9.8	12.3	0.85	0.22	0.22	-
Recrystallised limestones													
n	-	3	3	3	3	3	3	3	3	16	16	16	4
min	-	2523	349901	97	43	134	415	2.6	4.6	-0.7	-7.8	22.9	0.707773
mean	-	5108	358407	145	138	220	576	6.6	11.7	0.4	-6.4	24.3	0.707849
max	-	6534	367722	217	273	372	664	12.9	17.8	1.0	-5.1	25.7	0.708011
st dev	-	2243	8938	64	120	132	140	5.5	6.6	-0.6	-0.7	-0.8	-0.000109
Stratabound dolostones													
n	14	18	10	18	18	18	18	18	18	25	25	25	5
min	50.03	32287	191817	37	208	23	**58**	0.6	8	-1.35	-8.04	22.62	0.707992
mean	51.13	77819	223841	140	469	107	151	5.2	137	-0.1	-5.5	24.90	0.708123
max	53.07	92590	313765	411	991	191	232	25.0	590	1.17	-4.05	26.30	0.708241
st dev	1.10	16149	35897	112	232	60	46	6.1	163	0.82	0.94	0.97	9.79E-05

Table 2. Summary of geochemical data collected from unaltered limestones, recrystallised limestones and stratabound dolostones. Dolomite stoichiometry is determined using Lumsden's equation (1979). Trace element concentrations are measured using ICP-MS and ICP-AES. Stable isotopic concentrations are determined using a carbonate CO_2 extraction line, and the resulting gas is analysed by mass spectrometry. Strontium isotopic ratios are measured with a Triton Thermal Ionization Mass Spectrometer.

Temperature (°C)	65	65	65	80	80	80
Concentration Factor	1	1.43	2	1	1.43	2
Density of dolomite	2840000	2840000	2840000	2840000	2840000	2840000
Moecular weight of dolomite (grams per mole)	184.4	184.4	184.4	184.4	184.4	184.4
mol dolomite	15401	15401	15401	15401	15401	15401
moles per m^3 at 19% porosity	12475	12475	12475	12475	12475	12475
moles precipitated per kg water	0.01	0.02	0.03	0.01	0.01	0.02
kg water to reciptate 1 mole dolomite	77.06	53.97	38.13	100.87	70.52	50.25
Density of fluid	1005	1015	1027	996	1006	1019
m^3 fluid to ppt 1 mole dolomite	0.077	0.053	0.037	0.101	0.070	0.049
Volume of fluid to precipitate $1m^3$ dolomite	957	664	463	1264	875	615
Distance from HFF	2,000	2,000	2,000	2,000	2,000	2,000
Thickness of Thebes	200	200	200	200	200	200
Width of block containing SBD	1	1	1	1	1	1
Total volume of 1m slide in which SBD is observed	400,000	400,000	400,000	400,000	400,000	400,000
Volume of SBD in 1 m slice (assume 19% dolomitized)	76,000	76,000	76,000	76,000	76,000	76,000
Total volume of fluid required to ppt dol over $1m^3$, assuming 363 m^3 seawater to precipitate 1 mol dolomite in footwall	72,725,808	50,432,610	35,182,467	96,030,421	66,477,165	46,770,682
Total volume of fluid required to ppt dol over $1m^3$, assuming 363 m^3 seawater to precipitate 1 mol dolomite in footwall and hanging wall	145,451,616	100,865,219	70,364,933	192,060,841	132,954,330	93,541,365
Time to dolomitise	10,000,000	10,000,000	10,000,000	10,000,000	10,000,000	10,000,000
Cubic metres fluid required per year into footwall	7.27	5.04	3.52	9.60	6.65	4.68

Velocity of flux into footwall (m/year) assuming 19% effective porosity	38.3	26.5	18.5		50.5	35.0	24.6
Velocity of flux into footwall and hanging wall (m/year) assuming 19% effective porosity	76.6	53.1	37.0		101.1	70.0	49.2
Horizontal velocity if 19% of vertical thickness is dolomite and 200 m vertical thickess m/yr	1.01	0.70	0.49		1.33	0.92	0.65

Table 3 Mass balance and velocity of dolomitizing fluids at 65^{o}C and 80^{o}C

Sample	Stoichiometry (mole% $CaCO_3$)	$\delta^{18}O_{carbonate}$ (VPDB)[a]	$\delta^{13}C$ (VPDB)[a]	Δ_{47}[b]	T (°C)[b]	$\delta^{18}O_{water}$ SMOW
11B	50.6	- 4.59 (0.03)	- 0.72 (0.02)	0.561 ± 0.004	76 ± 8	3.6±1.0‰
34C	50.0	- 6.01 (0.36)	0.49 (0.02)	0.580 ± 0.008	67± 11	0.6±1.0‰
81A	53.1	- 4.86 (0.06)	1.06 (0.05)	0.575 ± 0.008	70 ± 9	2.3±1.3‰

	Temperature at +0.6‰ SMOW	Temperature at +3.6‰ SMOW	Fluid at 67°C	Fluid at 76°C
Minimum $\delta^{18}O_{dol}$= -8.0‰ PDB	**61°C**	**83°C**	**+1.5‰**	**+2.7‰**
Mean $\delta^{18}O_{dol}$ = -5.5‰ PDB	**45°C**	**64°C**	**+4.1‰**	**+5.3‰**
Max $\delta^{18}O_{dol}$ l= -4.0‰ PDB	**36°C**	**54°C**	**+5.6‰**	**+6.9‰**

Table 4 Summary of isotopic composition, Δ_{47}, temperature and $\delta^{18}O_{fluid}$ for stratabound dolostones as determined from clumped isotope analysis. a= Error is 1σ for $\delta^{13}C$ and $\delta^{18}O$ carbonate, b= Error for Δ^{47} and temperature (T) is given as standard error, calculated by the replicate standard deviation divided by square root of n. Dolomite stoichiometry determined as per Lumsden (1979) **Table 3b**,Range of possible fluid temperatures and d18O water using clumped isotope constraints based on conventional isotope data, based on Matthews and Katz (1977)

Assuming 25^{o}C and 70% humidity

$\delta^{18}O_{water}$	1.03
T	25^{o}C
h	0.70
f	0.3
$\delta\alpha$	-2
δo	0.6
A	-1.83
B	-3.10

$\delta^{18}O_{water}$	1.19
T	25^{o}C
h	0.70
f	0.4
$\delta\alpha$	-2
δo	0.6
A	-1.67
B	-2.98

$\delta^{18}O_{water}$	0.98
T	25^{o}C
h	0.70
f	0.5
$\delta\alpha$	-2
δo	0.6
A	-1.59
B	-2.92

Assuming 30^{o}C and 50% humidity

$\delta^{18}O_{water}$	5.52
T	30^{o}C
h	0.5
f	0.3
$\delta\alpha$	-2
δo	0.6
A	-0.48
B	-2.19

$\delta^{18}O_{water}$	3.05
T	30^{o}C
h	0.5
f	0.4
$\delta\alpha$	-2
δo	0.6
A	-0.45
B	-2.16

$\delta^{18}O_{water}$	1.97
T	30^{o}C
h	0.5
f	0.5
$\delta\alpha$	-2
δo	0.6
A	-0.43
B	-2.15

$\delta^{18}O_{water}$	$\delta^{18}O$ of residual water body= $(\delta o\text{-}A/B)f^{B}+A/B)$
δo	$\delta^{18}O$ composition of original water
δa	$\delta^{18}O$ of atmospheric moisture
α	equilibrium fractionation between O and H_2O, based on Kakiuchi and Matsuo, 1979
αw	activity coefficient of water= -0.000543f-2 - 0.018521f-1 + 0.99931
h	humidity
τ	ratio of activity coefficients for $^{18}O/^{16}O$
	where (t^-1 - 1) x 10^3 = 0.47m$CaCl_2$ + 1.107m$MgCl_2$ - 0.16mKCl
$\Delta\varepsilon$	kinetic enrichment factor, $\Delta\varepsilon=k((1-h)/\alpha w)$
ε	α - 1
f	fraction of water evaporated
A =	$(((h/\alpha w)*\delta a + \Delta\varepsilon + (\alpha-t))/\alpha)/1-(h/\alpha w)+\Delta\varepsilon$
B =	$(h/\alpha w)- \Delta\varepsilon + (\alpha-t))/\alpha)/(1-h)/(\alpha w+\Delta\varepsilon)$

Calculation assumes molality $CaCl_2$= 0.4, $MgCl_2$= 1.48 and KCl= 0.28

Table 5 Calculation of the $\delta^{18}O_{water}$ of the dolomitizing fluid, based on Swart *et al.*, 1989