Structural controls on non fabric-selective dolomitization within rift-related basin-bounding normal fault systems: Insights from the Hammam Faraun Fault, Gulf of Suez, Egypt

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Abstract
Fault-controlled dolostone bodies have been described as potential hydrocarbon-bearing reservoirs. Numerous case studies have described the shape and size of these often non fabric selective dolostone bodies within the vicinity of crustal-scale lineaments, usually from Palaeozoic or Mesozoic carbonate platforms, which have undergone one or more phases of burial and exhumation. There has been little attention paid, however, to fault-strike variability in dolostone distribution or the preferential localization of these bodies on particular faults. This study focuses on dolostone bodies adjacent to the Hammam Faraun Fault (HFF), Gulf of Suez. This crustal-scale normal fault was activated in the Late Oligocene, coincident with the onset of extension within the Suez Rift. Dolomitization in the pre rift Eocene Thebes Formation occurred in the immediate footwall of the HFF forming two massive, non facies selective dolostone bodies, ca. 500 m wide. Facies-controlled tongues of dolostone on the margins of the massive dolostone bodies extend for up to 100 m. The geochemical signature of the dolostone bodies is consistent with replacement by Miocene seawater, contemporaneous with the rift climax and localization of strain along the HFF. A conceptual model of dolomitization from seawater that circulated within the HFF during the rift climax is presented. Seawater was either directly drawn down the HFF or circulated from the hanging wall basin via a permeable aquifer towards the HFF. The lateral extent of the massive dolostone bodies was controlled by pre-existing HFF-parallel fracture corridors on the outer margins of the damage zone of the fault. The behaviour of these fracture corridors alternated between acting as barriers to fluid flow before rupture and acting as flow conduits during or after rupture. Multiple phases of dolomitization and recrystallization during the ca. 10 Ma period in which dolomitization occurred led to mottled petrographical textures and wide-ranging isotopic signatures. The localization of dolomitization on the HFF is interpreted to reflect its proximity to a rift accommodation zone which facilitated vertical fluid flow due to perturbed and enhanced stresses during fault interaction. It is possible that the presence of jogs along the strike of the fault further focused fluid flux. As such, it is suggested that the massive dolostones described in...
this study provide a window into the earliest stages of formation of fault-controlled hydrothermal dolostone bodies, which could have occurred in other areas and subsequently been overprinted by more complex diagenetic and structural fabrics.

1 | INTRODUCTION

The occurrence of non stratabound, non facies selective dolomitization within the vicinity of faults has been widely described in the literature (e.g. Braithwaite, Rizzi, & Darke, 2004; Davies & Smith, 2006; Dewit et al., 2012, 2014; Duggan, Mountjoy, & Stasiuk, 2001; Lonnee & Machel, 2006; López-Horgue et al., 2010; Martin-Martin et al., 2018; Sharp et al., 2010). Most of these studies emphasize the importance of faults as conduits for hot, magnesium-enriched brines responsible for dolomitization and describe an asymmetric “Christmas Tree” geometry attributed to fluid flow up and away from faults. Consequently, prediction of the distribution of fault-controlled dolostone bodies requires understanding of the source, composition, timing and mechanism of fluid flux within the context of the tectono-stratigraphic evolution of the basin. There is little consensus, however, as to what controls the preferential localization of dolostone along certain faults, or what controls along-fault-strike dolostone distribution. Since many case studies are of Mesozoic and Palaeozoic platforms that have undergone a complex history of structural rejuvenation and it is possible that the initial controls on the localization of dolomitization have been obscured.

The Eocene Thebes Formation occurs in the footwall of the Hammam Faraun Fault (HFF), a rift-related, basin-bounding normal fault with ca. 5 km of maximum displacement (Suez Rift, Egypt; Figure 1). Dolomitization occurs in the form of discrete, non fabric selective bodies that cross-cut stratigraphy and older, stratabound dolostone bodies (Hirani et al., 2018; Hollis et al., 2017). The outcrop provides an opportunity to understand the controls on the localization of dolomitization within a relatively young (Tertiary) extensional basin that has not undergone subsequent tectonic rejuvenation. It also provides the opportunity to determine the importance of structural complexity on the evolution of a rift system at the basin scale, in the form of transfer zones (sensu Morley, Nelson, Patton, & Munn, 1990), or accommodation zones (sensu Younes & McClay, 2002), and at the scale of individual fault arrays, since the HFF is composed of a series of segments linked by complex relay zones (e.g. Peacock, Nixon, Rotevatn, Sanderson, & Zuluaga, 2017).

Structurally complex settings related to growth and interlinkage of extensional faults are commonly associated with local perturbation of stresses (Crider & Pollard, 1998; Kattenhorn, Aydin, & Pollard, 2000), increased faulting and fracturing (e.g. Rotevatn & Bastesen, 2014; Morley et al., 1990) and highly complex damage zones (e.g. Bastesen & Rotevatn, 2012; Fossen, Johansen, Hesthammer, & Rotevatn, 2005). These structural complexities and heterogeneities result in a lasting susceptibility to dilatancy on faults and fractures under variable in situ stresses through time, during and after rifting (Ferrill & Morris, 2003; Gar treff, Zhang, Lisk, & Dewhurst, 2004), increasing the tendency to localize fluid flow and fluid–rock interaction (e.g. Fossen & Rotevatn, 2016; Johansen, Fossen, & Kluge, 2005). In this study we aim to understand how structures may control the localization and formation of non fabric selective dolostone bodies in rift basins via the following objectives:

(1) Describe and assess the extent of dolomitization in the footwall of the HFF.
(2) Constrain the structural style and deformation affecting the immediate footwall of the HFF.
(3) Determine the timing of dolomitization with respect to rift evolution and growth of the HFF based on field relationships, petrographical evidence and geochemical proxies.

The results have implications to understanding the fundamental structural controls on fluid flow and fluid–rock interaction in rifts, and have economic ramifications, given the role of hydrothermal dolomites as hydrocarbon reservoirs and mineral hosts.
FIGURE 1  (a) Map showing the major structural elements of the Gulf of Suez, and distribution of prerift and syn-rift sediments along the western Sinai coast (Khalil & McClay, 2001; adapted from Younes & McClay, 2002). The location of the Gharandal (GAZ) and Morgan (MAZ) accommodation zones are noted according to Moustafa (2003), with the location of other dolomitized successions along the Gulf highlighted. (b) Geological map of the wider study area within the Hammam Faraun Fault block (modified from Moustafa & Abdeen, 1992 and Moustafa, 2003). The Hammam Faraun Fault parallels the present-day coastline, with the Gebel Fault intersecting through the study area. Gebel HF denotes the locality at the top of the fault block. The stratigraphic column shows the pre- and syn-rift succession within the area. The location of the study area shown in Figure 2 is highlighted. (c) Cross-section through the wider study area illustrating the subsurface geology (modified from Moustafa, 2003). Colours correspond to the colours shown for the Formations listed in (b).

2 | GEOLOGICAL SETTING

The Gulf of Suez rift is the NW–SE-trending continuation of the Red Sea rift system, which formed during the Late Oligocene to Early Miocene (Jarrige et al., 1990; Patton, Moustafa, Nelson, & Abdine, 1994; Sharp, Gawthorpe, Underhill, & Gupta, 2000; Younes & McClay, 2002; Bosworth, Huchon, & McClay, 2005; Figure 1a). In the Middle Miocene (14–12 Ma), stress transferred to the NE–SW-trending Aqaba transform boundary, formed by the collision of Arabia with Eurasia, dramatically slowing extension rates in the Gulf of Suez (Bosworth et al., 2005; Montenat et al., 1988). The Suez rift is c. 300 km long, 30–80 km wide, and is bounded by large-scale normal fault zones (Figure 1a; Alsharhan & Salah, 1995). This geometry resulted in the formation of classic half-graben tilted fault blocks, with eroded crests and deep hanging wall basins (Khalil & McClay, 2001; Knott, Beach, Welbon, & Brockbank, 1995; Moustafa & Abdeen, 1992; Sharp et al., 2000). The dips of the half-grabens subdivide the rift into three asymmetric dip provinces, separated by two major accommodation zones that are oblique to the rift trend (Figure 1a; Moustafa, 1996; Khalil & McClay, 2001). These accommodation zones contain a complex array of normal faults linked by relay ramps and local cross-faults which enable strain to be transferred between the contrasting dip domains (Bosworth, Crevello, Winn JR, & Steinmetz, 1998; Moustafa & Abdeen, 1992; Younes & McClay, 2002). The onset of oblique-slip movement on the Aqaba and Levant transform in the late Miocene (Serravalian) led to a change from rift-normal faulting to oblique extension in the Suez Riff (Bosworth et al., 2005).

2.1 | Structure of the Hammam Faraun Fault block

This study focuses on outcrops in the northern part of the Hammam Faraun Fault (HFF) block, a NW-trending, NE-titled crustal-scale fault block that is ca. 20 km wide and 40 km long (Moustafa & Abdeen, 1992; Patton et al.,
1994; Alsharhan & Salah, 1995; Figure 1b). It is located on the eastern rift flank in the central dip province, south of the Gharandal accommodation zone (GAZ on Figure 1a), which marks the transition between the northern and central dip provinces (Moustafa, 2003; Moustafa & Abdeen, 1992; Robson, 1971); towards the southeast, the fault zone loses displacement and links with the Baba-Markha Fault Zone (e.g., Jackson, Gawthorpe, Leppard, & Sharp, 2006; Jackson, Gawthorpe, & Sharp, 2006). The HFF block is bounded by basement-involved SW-facing extensional master faults of the Hammam Faraun Fault (HFF) to the SW, and the Thal Fault to the NE (Moustafa & Abdeen, 1992), and is broken up by multiple subsidiary NW-trending extensional intra-block faults. During the early stages of rifting, extensional strain was accommodated on the Thal Fault and distributed intra-block faults; as rifting progressed, most of the distributed early-rift faults were abandoned, and strain localized onto the HFF (15–17 Ma; Gawthorpe et al., 2003). The study area is located in the north-western corner of the HFF block (Figure 1b), in the footwall of the HFF, where there is exceptional pseudo-3D exposure of prerift Eocene carbonates. The HFF has ca. 5 km of displacement at the prerift/syn-rift contact (Gawthorpe et al., 2003). The subsurface geology of the Hammam Faraun Fault Block has been extensively studied by Moustafa (2003), with the cross-section in Figure 1c illustrating how the faults present in the study area dissect the stratigraphy.

The HFF is located offshore of the study area (Figure 1b) and controls the position of the present-day shoreline and a modern hot spring system (Sturchio et al., 1996). Overall it is northwest trending in the study area, although on a more local scale, it may comprise subordinate north–south-, northnortheast–southsouthwest- and east–west-trending segments (Gawthorpe et al., 2003). The HFF intersects with the north–south-oriented Gebel Fault, which is ca. 7 km long with a maximum throw of ca. 300 m (Figures 1b and 2). Early syn-rift deposits in the hanging wall of the Gebel Fault are unconformable on prerift carbonates of the Tanka Formation, indicating that the Gebel Fault nucleated and grew from rift initiation (Woodman, 2009), but was deactivated at some stage before the onset of the main rift climax (Gawthorpe et al., 2003). The present-day footwall elevation at Gebel HF (Figure 1b) is at 500 m above sea level.

2.2 | Sedimentology and stratigraphy

The prerift stratigraphy of the HFF block comprises a 2,000 m thick sedimentary succession, which overlies Precambrian igneous and metamorphic basement rocks (Figure 1b–c). The Palaeozoic to Lower Cretaceous stack of sedimentary rocks lie unconformably on the Precambrian basement (Alsharhan, 2003; Moustafa, 1996). The Cretaceous succession comprises shaley sandstones and limestones, principally deposited in a marine environment, overlain by Palaeocene to Lower Eocene Esna Formation shale that is itself unconformably overlain by the Thebes Formation (Moustafa & Abdeen, 1992). The partially dolomitized, prerift Early Eocene Thebes Formation is the

![FIGURE 2](image-url) Distribution and extent of massive dolostone bodies in the study area. The north massive dolostone body (NMD), south massive dolostone body (SMD) and their associated dolostone tongues (DT) are illustrated relative to the position of the Hammam Faraun Fault. Also highlighted are an earlier phase of stratabound dolostone bodies as described in Hollis et al. (2017) with evidence of cross-cutting relationships between the earlier stratabound dolostone bodies and the massive dolostone bodies described in this contribution. Illustrations are superimposed onto a Google Earth image, with the wider special context of this figure highlighted in Figure 1b.
primary focus of this study and is informally subdivided into the lower and upper Thebes. The lower Thebes Formation comprises remobilized slope deposits of matrix-supported (R1) and clast-supported conglomeratic debrites (R3) and turbiditic (R2) and slumped grainstones (R4) embedded within in situ slope packstone deposits (S1). The upper Thebes Formation consists of grainstone beds (R5), sometimes with channelized zones (R6), embedded in skeletal basinal wackestones (B1; Table 1). Most limestone facies have been recrystallized and have a microcrystalline calcite texture, although skeletal (basinal) wackestones (B1) appear unaltered (Hirani, 2014). Basaltic dykes and sills provide the first stratigraphic evidence for rifting and have been dated as Oligo-Miocene (24–22 Ma; Bosworth & Stockli, 2016).

3 | METHODOLOGY

The stratigraphic distribution and geometries of the dolostone bodies were determined by field mapping, logging and measuring (using a Jacobs Staff and a laser rangefinder) and using Google Earth© satellite imagery. Detailed stratigraphic logs (Figure 2) were collected and samples selected throughout logged sections within the dolostone bodies, host limestone and the contact zones at the margin of the dolostone bodies. About 261 thin sections were prepared with blue-dye resin impregnation to identify porosity, and stained with alizarin Red S and potassium ferricyanide (Dickson, 1966). All thin sections were studied under plane polarized light (PPL) and cross-polarized light (XPL) using a Nikon Optiphot 2 microscope. Limestone textures are described using the Dunham (1962) classification and dolomite textures are described using the Sibley and Gregg (1987) classification scheme. Polished sections were studied under a Cambridge CITL cold cathodoluminescence microscope at pressures of 350–500 μA.

bulk mineralogy and stoichiometry were determined by XRD analysis using a Bruker D8 Advance diffractometer. The diffractometer operated at 40 kV and 30 mA, and scanned the samples in the 20 range from 5.0 to 70° with increments of 0.02° [20]/s. The dolomite stoichiometry mol% CaCO3 was determined by applying Lumsden (1979) to the measured d[104] spacing (M = 333.3 × d-spacing – 911.99).

The Agilent 7500cx inductively coupled plasma mass spectrometer (ICP-MS) and Perkin-Elmer Optima 5300 dual view inductively coupled plasma atomic emission spectrometer (ICP-AES) were used to quantify the concentrations of trace elements (Fe, Mn, Al, Sr, Ni, Ba) within 56 samples, reported in ppm. Rare earth element (REE) analysis was conducted on 24 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 (Haley, Klinkhammer, & McManus, 2004; Nance & Taylor, 1976; Nothdurft, Webb, & Kamber, 2004).

Stable isotope analysis (δ¹⁸OVPDB and δ¹³CVPDB) was conducted at the SUERC research facility (East Kilbride, Scotland). Ten milligram of powdered sample was digested in phosphoric acid (McCrea, 1950), and the resulting CO2 gas analysed on a VG OPTIMA mass spectrometer. All stable isotope values are reported in per mil (%o) 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).

87Sr/86Sr ratios were determined on Sr isotope extracted from a nitric acid solution and purified onto chromatographic resin using standard cation exchange procedures (Horwitz & Bloomquist, 1975) at the French Research Institute for Exploitation of the Sea (Laboratoire Géochimie et Métallogénie) using a Triton Thermal Ionization Mass Spectrometer. All results were reported to the NBS987 standard, and the precision of individual runs was better than 0.00005.

Structural characterization of the footwall of the HFF included qualitative/descriptive fracture characterization as well as spatial scanline-based characterization of the damage zone of the HFF. Spatial and dimensional attributes of fractures were collected within limestones and dolostones by means of scanlines acquired in the vicinity and 1.5–2 km away from the HFF (Wadi Wasit and Y-shaped Wadi areas; Figure 1b). Data recorded included fracture type, orientation, intensity (number of fractures per metre), fracture termination (stratabound or non stratabound fractures) and cement fill (Eker, 2013).

4 | CHARACTERIZATION OF THE MASSIVE DOLOSTONE BODIES

Dolomitization within the study area comprises non stratabound (so-called “massive”) dolostone bodies within the lower Thebes Formation, in the footwall of, and adjacent to, the HFF and the present-day hot spring system (Figures 2 and 3a). Stratabound dolostone bodies that are not associated with the massive bodies occur for up to 2 km away from the HFF. These dolostone bodies vary in length from 5 m up to 300 m along depositional dip, and range in thickness from 25 to 5 m (Figure 2; Hollis et al., 2017). Field relationships indicate that they predate the formation.
| 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*, *Dysoxocyclina*, *Alveolinids*, *Operculina*, planktonic foraminifera, echinoid fragments, bryozoan fragments | Microporosity in matrix, intragranular | Commonly recrystallized skeletal material and mud matrix |
| Remobilized 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*, bryozoan 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 dolomitized |
| Remobilized 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 |
| Remobilized 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 |
| Remobilized 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 |
| Remobilized 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 |
| Remobilized deposit | Channel grainstones in basinal deposits | R6 | Scoured | Sharp | Grainstone | Dasyclad algae, *Nummulites*, *Miliolids*, *Alveolinids*, *Soritids*, red algae, bryozoan 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 |
of the massive dolostone bodies. Specifically, they are offset by the Gebel Fault, which was active during the earliest phases of rifting but died prior to the rift climax (Gawthorpe et al., 2003). Petrographical and geochemical data indicate that the stratabound dolostone bodies formed from slightly evaporated seawater at temperatures of ca. 60–80°C during the late Oligocene–early Miocene (Hirani et al., 2018; Hollis et al., 2017).

The massive dolostone bodies have a distinctive reddish-brown weathering colour, relative to the cream colouration of the host limestone and can be divided into:

1. North massive dolostone body (NMD)—500 m wide and 75 to 80 m thick
2. South massive dolostone body (SMD)—280 to 400 m wide and 75 to 80 m thick
3. Dolostone tongues (DT) which extend away from the edge of the massive bodies for up to 100 m, with thickness ranging between 50 cm and 2 m.

X-ray diffraction analysis (Table 2) confirms that dolomite is the dominant mineral. The NMD and SMD are not apparently connected in outcrop (Figure 2), although it is possible that they may be in the subsurface. The northern boundary of the NMD body is diffuse with the host limestone, whereas the upper contact is bed-bound and the basal contact is not exposed. The southern boundary of the NMD is structurally bound by a NNE–SSW-trending fracture corridor that dips 70–80° in an eastward direction.

The eastern extent of the SMD is diffuse into the host limestone and both the upper and basal contacts are bed-bound, with the base of the dolostone body terminating against the mud-rich Esna Formation. The SMD is also bounded to the north by a WNW–ESE-trending fracture corridor that dips 70° to the west (Figures 2 and 3).

In outcrop the NMD and SMD principally comprise red-orange, crystalline, fractured, weakly fabric-preserving and fabric-destructive dolostone. Precursor slope packstone facies (S1; Table 1) can be distinguished by more bedded

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**FIGURE 3** (a) Massive dolostone bodies adjacent to the coastline and a present-day hot spring system. The south massive dolostone body (SMD) is bound to the west by a NW–SE-trending fracture corridor. (b) Dolostone tongues (DT) extending away from the north massive dolostone body (NMD), with sharp upper and basal contacts, and abrupt lateral terminations within the host limestone. For spatial context of these figures, refer to Figure 2.
and less vuggy textures (Figure 4b) compared to the remobilized matrix-supported conglomerates (R1; Table 1). The latter contain ghosts of, or mouldic and vuggy pores after clasts (Figure 4c and d). In thin section, the NMD and SMD bodies have fabric-destructive non-planar textures, with crystal sizes ranging between 10 and 30 μm (Figure 5). Under CL, the matrix-replacive dolomite exhibits a mottled red and orange luminescence (Figure 5b, d and f), with little evidence of zonation within individual crystals.

The majority of mouldic (after Nummulities and Alveolinidae foraminifera and echinoderm fragments) and vuggy pores are partially occluded by coarsely crystalline calcite, dolomite and gypsum cements, which also occur within fractures (Figure 5e–h). Dolomite cements have a bright red luminescent core proceeded by a thick non-luminescent zone, a thin bright red zone, a thick dull green luminescent zone and an outer thin bright red luminescent zone (Figure 5f). The composite paragenetic sequence

### Table 2

Summary of geochemical data collected from unaltered limestones, northern massive dolostone body (NMD), southern massive dolostone body (SMD) and dolostone tongues (DT). 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 CO₂ extraction line, and the resulting gas is analysed by mass spectrometry. Strontium isotopic ratios are measured with a VG 354 TIMS.

| Phase                        | Stoichiometry (mole%) | Mg (ppm) | Ca (ppm) | Fe (ppm) | Mn (ppm) | Al (ppm) | Sr (ppm) | Ni (ppm) | Ba (ppm) | d¹³C (‰ VPDB) | d¹⁸O (‰ VPDB) | d¹⁸O (‰ VSMOW) | ⁸⁷Sr/⁸⁶Sr |
|------------------------------|-----------------------|----------|----------|----------|----------|----------|----------|----------|----------|----------------|----------------|----------------|----------|
| Unaltered limestones         |                       | 3        | 3        | 3        | 3        | 3        | 3        | 3        | 3        |                |                |                |          |
| Min                          | –                     | 2,127    | 304,691  | 1,753    | 14.5     | 1,928    | 9,33     | 29.5     | 8.0      | –1.00         | –3.50          | 27.30          | –         |
| Mean                         | –                     | 6,327    | 313,390  | 2,182    | 16.9     | 2,150    | 1,240    | 40.3     | 20.7     | –0.17         | –3.35          | 27.45          | 0.707814  |
| Max                          | –                     | 12,362   | 320,615  | 2,648    | 21.5     | 2,305    | 1,775    | 48.8     | 32.5     | 0.70           | –3.10          | 27.70          | –         |
| SD                           | –                     | 5,359    | 8,064    | 449      | 4.0      | 197      | 465      | 9.8      | 12.3     | 0.85           | 0.22           | 0.22           | –         |
| Neomorphosed limestone       |                       | 3        | 3        | 3        | 3        | 3        | 3        | 3        | 16       | 16             | 16             | 16             | 4         |
| Min                          | –                     | 2,523    | 349,901  | 97       | 42.9     | 134.0    | 415      | 2.6      | 4.6      | –0.69          | –7.80          | 22.87          | 0.707773  |
| Mean                         | –                     | 5,108    | 358,407  | 145      | 138.1    | 219.5    | 576      | 6.6      | 11.7     | 0.35           | –6.37          | 24.32          | 0.70849   |
| Max                          | –                     | 6,334    | 367,722  | 217      | 272.7    | 372.1    | 664      | 12.9     | 17.8     | 1.04           | –5.12          | 25.65          | 0.708111  |
| SD                           | –                     | 2,243    | 8,938    | 64       | 119.8    | 132.5    | 140      | 5.5      | 6.6      | –0.59          | –0.71          | –0.77          | –0.00109  |
| North massive dolostone body (NMD) |               | 27       | 28       | 26       | 28       | 28       | 28       | 28       | 23       | 23             | 23             | 23             | 5         |
| Min                          | 53.07                 | 70,117   | 185,580  | 132      | 246      | 87.0     | 1.6      | 6.7      | –2.8     | –12.0          | 18.5           | 0.708133       |
| Mean                         | 54.72                 | 839,88   | 205,291  | 1,300    | 619      | 422      | 175.1    | 7.0      | 99.2     | –0.5           | –7.7           | 23.0           | 0.708361  |
| Max                          | 55.52                 | 101,556  | 260,396  | 4,277    | 1,204    | 1,087    | 303.7    | 21.0     | 513.9    | 0.9            | –4.9           | 25.8           | 0.708539  |
| SD                           | 0.61                  | 6,089    | 17,964   | 957      | 219      | 263      | 56.8     | 4.8      | 153.9    | 1.1            | 2.0            | 0.91           | 0.000191  |
| South massive dolostone body (SMD) |               | 10       | 10       | 7        | 10       | 10       | 10       | 10       | 7        | 7              | 7              | 7              | 6         |
| Min                          | 53.07                 | 56,251   | 189,652  | 849      | 356      | 247      | 91.8     | 3.1      | 24.7     | –2.33          | –11.30         | 19.30          | 0.708125  |
| Mean                         | 54.35                 | 81,279   | 197,897  | 2,233    | 834      | 551      | 201.7    | 8.6      | 205.5    | –0.12          | –6.32          | 24.59          | 0.708316  |
| Max                          | 54.90                 | 88,416   | 215,724  | 4,234    | 1,697    | 1,223    | 334.7    | 14.6     | 459.7    | 1.20           | –4.40          | 26.40          | 0.708576  |
| SD                           | 0.61                  | 9,303    | 8,720    | 1,351    | 505      | 305      | 72.8     | 8.6      | 205.5    | 1.26           | 2.36           | 2.42           | 0.00016   |
| Dolostone tongues (DT)       |                       | 16       | 12       | 9        | 12       | 12       | 12       | 12       | 14       | 14             | 14             | 14             | 6         |
| Min                          | 54.90                 | 78,898   | 194,812  | 432      | 192      | 227      | 145      | 2.0      | 7.1      | –1.90          | –10.00         | 20.60          | 0.708063  |
| Mean                         | 55.06                 | 84,105   | 207,916  | 782      | 379      | 597      | 201      | 5.5      | 85.3     | 0.10           | –7.47          | 23.20          | 0.708193  |
| Max                          | 55.52                 | 88,827   | 241,538  | 1,540    | 770      | 1,296    | 292      | 10.4     | 296.9    | 1.30           | –5.00          | 25.70          | 0.708419  |
| SD                           | 0.28                  | 3,338    | 18,056   | 333      | 186      | 275      | 49       | 2.3      | 107.3    | 0.91           | 1.82           | 1.87           | 0.00014   |
determined from thin section can briefly be described as dolomitization, dolomite cements, non luminescent calcite cement, luminescent calcite cement, quartz cement/silicifcation and haematite–goethite–sulphates (Figures 5 and 6).

The DT extend away from the massive bodies and are confined to specific strata. They exhibit sharp and planar upper and basal contacts with the adjacent limestone, with abrupt lateral terminations (Figure 3b). These bodies are facies selective, forming preferentially along beds of matrix-supported conglomerates (R1; Table 1) and grainstone turbidite facies (R2; Table 1) within the lower Thebes Formation (Figure 6a and b). This is evident in the field by the presence of mouldic pores after limestone clasts and centimetre-scale vugs. Petrographically, the DT comprise non planar dolomite, but with a wide range of crystal sizes (10 µm and 75 µm; Figure 6c–f). Under CL, matrix-replacive dolomite exhibits mottled dull and bright red luminescence. Mouldic pores after Nummulites and Alveolinidae foraminifera (Figure 6c and e), are partially cemented by dolomite and calcite (Figure 6e). Dolomite

**FIGURE 4** (a) Heavily fractured internal structure of the north massive dolostone body (refer to Figure 3a for spatial context). (b) Dolomitized slope packstone facies with consolidated beds intercalated with thin, less consolidated grainstone beds (arrows). (c) Mouldic porosity after clasts (arrows) within dolomitized conglomerate, with vuggy porosity also common. (d) Dolomitized matrix-supported conglomerate, evident by the presence of clasts (arrows), which are also dolomitized. (e) Mineralization within fractures composed of gypsum, anhydrite, barite, halite and haematite, with brecciated fragments of host dolostone also present within the mineralization. (f) Abrupt contact between host limestone and north massive dolostone body, with iron- and sulphate-rich mineralization (arrow) at the contact (refer to Figure 3b for spatial context). Hammer as scale = 30 cm length.
FIGURE 5  (a–d) Photomicrographs of the north massive dolostone body (NMD), in plane polarized light (PPL) and cathodoluminescence (CL). The paragenesis can be illustrated as dolomitization (d1), dolomite cements (d2), non luminescent calcite (c1), luminescent calcite (c2), quartz cement/silicification, haematite–goethite–sulphates (g). (a) Dolomite matrix (d1) with non planar texture. Vuggy porosity partially cemented by coarser cloudy core clear rim dolomite cements (d2), in PPL, (b) corresponding CL image of non planar dolomite matrix (d1) exhibiting mottled bright red and orange luminescence. The pore filling dolomite cements (d2) exhibit bright red luminescent cores (arrow) proceeded by a thick non luminescent zone, a thick bright red zone, a thick dull green luminescent zone and an outer thin bright red luminescent zone. (c) Sedimentary lamination of the precursor slope packstone facies preserved during dolomitization (d1), in PPL, (d) corresponding CL image of dolomitized slope packstone with a non planar texture and mottled bright red and orange luminescence. (e–h) Photomicrographs of the south massive dolostone body (SMD), in plane polarized light (PPL) and cathodoluminescence (CL). (e) Non planar dolomite matrix (d1) with mouldic porosity after Alveolinidae foraminifera partially occluded by coarse dolomite cements (d2) and gypsum laths (g), in PPL, (f) corresponding CL image showing mottled bright and dull red luminescent dolomite matrix (d1). Pore filling dolomite cements (d2) exhibit a bright red luminescent core, proceeded by a thick non luminescent zone and a thin bright red outer zone (arrow), whereas the gypsum lath (g) within the pore is non luminescent. (g) Vuggy porosity filled with dolomite cement (d2) proceeded by coarse blocky calcite cement (c2), and gypsum laths (g), in PPL, (h) corresponding CL image with bright luminescent cores of dolomite cement (d2), calcite cement (c2) with bright orange and subtle dull orange luminescent concentric zones and non luminescent gypsum laths (g)
Cements are zoned, with a mottled bright and dull red luminescent core and a thick dull red luminescent outer zone (Figure 6f). The calcite cements have concentric, non luminescent, bright and dull orange luminescent subzones (Figure 6f).

4.1 Geochemical characterization

Whole rock isotopic values for the unaltered limestones (B1) average $\delta^{18}O = -3.35\%_o$ VPDB and $\delta^{13}C = -0.17\%_o$ VPDB (Figure 7a; Table 2). In all other facies, limestone has been recrystallized by microcrystalline calcite (Hirani, 2014) prior to dolomitization; here carbon isotopic values are apparently unchanged ($\delta^{13}C = -0.69\%_o$ to $1.04\%_o$ VPDB) but the oxygen isotopes are depleted ($\delta^{18}O = -5.12\%_o$ to $-7.80\%_o$ VPDB; Table 2). The average isotopic composition of the NMD body is $\delta^{18}O = -7.7\%_o$ VPDB ($-4.9$ to $-12.0\%_o$ VPDB) and $\delta^{13}C = 0.5\%_o$ VPDB ($+0.9$ to $-2.8\%_o$ VPDB). The SMD body has an isotopic composition in a similar range, with an average $\delta^{18}O = -6.3\%_o$ VPDB and $\delta^{13}C = -0.1\%_o$ VPDB. The DT bodies have an average $\delta^{18}O = -7.5\%_o$ VPDB ($-10$ to $-5\%_o$ VPDB) and $\delta^{13}C = +0.1\%_o$ VPDB ($1.3\%_o$ to $-1.9\%_o$ VPDB) (Figure 7a). The stable isotopic composition of dolostone samples closest to the fracture corridors are more constrained than the range exhibited in the dolostones closer to the fault ($\delta^{18}O$ between $-4.4\%_o$ and $-6.6\%_o$ VPDB).
FIGURE 7 (a) Stable isotope (δ18O VPDB vs. δ13C VPDB) plot for unaltered and recrystallized limestones, north and south massive dolostone bodies (NMD and SMD, respectively) and dolostone tongues (DT). 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 dolostone samples exhibit a wide range of δ18O values between −4.4‰ and −12.0‰, and δ13C between 1.3‰ and −2.8‰. The isotopic values for samples within each dolostone body type do not cluster. (b) Temperature of dolomitizing fluids vs. δ18O dolomite. This plot is constructed using the Matthews and Katz (1977) fractionation factor to determine the minimum and maximum temperature of the dolomitizing fluid. Assuming a δ18O seawater of +1‰ to 0‰ (Veizer & Prokoph, 2015), fluid temperatures between 50 and 100°C would be expected.

A single unaltered limestone sample of facies B1 has a strontium isotopic ratio of 0.707814. The dolostone bodies have a wider ranging radiogenic 87Sr/86Sr values (NMD body = 0.708113 to 0.708539, and SMD body = 0.708125 to 0.708576). The DT has 87Sr/86Sr ratios in a similar range to the massive dolostone bodies, averaging 0.708193 (Table 2).

Comparison of the Fe and Mn concentrations within the dolostones relative to the host limestone indicates that the concentration of Mn is greater in the dolostone samples and highest in the massive dolostone bodies, compared to the dolostone tongues (Table 2; Figure 8a). In contrast, the Fe content is highest within the SMD (average = 2.232 ppm) with lower concentrations in the unaltered limestone (facies B1; average 2182), NMD (average = 1300 ppm) and DT (average = 782 ppm; Table 2). Some variability in the Sr concentration within the NMD (average = 175 ppm), SMD (average = 202 ppm) and DT (average = 201 ppm) is noted; however, it is significantly lower than that measured in the unaltered limestone (average = 1,240 ppm).

The REE within the unaltered limestones (facies B1) have a positive La and negative Ce anomaly, and enrichment in the heavier rare earth elements (HREE) (Nothdurft et al., 2004; Wyndham, McCulloch, Fallon, & Alibert, 2004; Figure 8b). Recrystallized limestone and dolostone samples from all bodies exhibit lower total REE concentrations than the unaltered limestone, a positive La and a negative Ce anomaly with a flattened MREE and HREE profile (Figure 8b). Sample clustering within the marine quadrant of a Pr/Pr* ([Pr/(0.5Ce + 0.5Nd)SN] vs. Ce/Ce* ([Ce/(0.5La + 0.5Pr)SN]) plot confirms that these anomalies are true (Bau & Dulski, 1996; Webb & Kamber, 2000; Figure 8c).

5 | STRUCTURAL CHARACTERIZATION OF THE HAMMAM FARAUN FAULT

The studied outcrops are located in the immediate footwall of the HFF. The HFF itself is not seen in the study area as it is located offshore and therefore discrete fault strands, or elements of a fault “core” (sensu Caine, Evans, & Forster, 1996), are not exposed. Across the study area, approximately N–S-trending fracture corridors have been mapped (Figure 9a). In addition, two types of predominantly non-stratabound fractures dominate: non-mineralized, open-mode joints, described in detail by Korneva et al. (2017) and mineralized, open-mode syntactical veins with elongate-block and blocky cements and inclusion bands (sensu Bons, Elburg, & Gomez-Rivas, 2012) that are cemented by calcite, dolomite, gypsum, and iron oxides (Figures 4e and 9d–e). No saddle dolomite was observed, nor was there any evidence for chaotic, saddle dolomite cemented hydrobrecciation typical of many fault-controlled dolostone bodies (e.g. Davies & Smith, 2006). The orientation of fractures with limestone and dolostone bodies are shown in Figure 9a. Fractures within the NMD body and its associated DT strike north–south to northnorthwest–southsouthwest, with a secondary fracture trend striking northwest–southeast. Fractures within the SMD body and its associated dolostone tongues strike predominantly northwest–southeast, with some north–south- and east–west-trending fractures. Fractures measured in the undolomitized host limestone in Wadi Wasit strike northwest–southeast towards the eastern side of the wadi, whereas fractures towards the western side of the wadi strike north–south.

A damage zone (sensu Childs et al., 2009; Caine et al., 1996; Kristensen et al., 2016) associated with the HFF
extends ca. 450 m into the footwall (Figure 9a–b) interpreted on the basis of enhanced frequency of joints and veins within the host rock (Figure 10 and Korneva et al., 2017). It comprises two main components i) an outer damage zone of elevated fracture intensities (5–20 fractures per metre) that extends from ca. 150 to 450 m from the shoreline and ii) an inner damage zone of intense fracturing and brecciation in the ca. 150 m zone nearest to the shoreline (fracturing 20–40 fractures per metre, locally in excess of 50–100 fractures per metre), with near-complete obliteration of bedding. The highest fracture densities within the inner damage could only be estimated in the field because of their close spacing, short length, highly variable orientations and local cover of the outcrop by sand (Figure 9a–b). The massive dolostone bodies are principally located within this inner damage zone and are dissected by joints and veins that have apertures ranging from a few millimetres up to 30 cm, many of which are cemented (Figure 9c–e). Minor faults, mineralized by haematite and/or gypsum and with throws of <50 cm, are present within the inner damage zone, but are not common.

The transition between the outer and inner damage zone of the HFF is dominated by joints, with calcite-cemented veins occurring locally in larger fracture corridors (see below) close to the contact of massive dolostone and limestone. Further into the footwall, away from the HFF, the fractures are dominantly bed-confined opening mode fractures (joints and calcite filled veins). Nevertheless, it is possible to differentiate a dolomite crystal size/facies control on fracture density within the massive dolostone. An increase in fracture density corresponds to a decrease in crystal size (Facies R1 and R2; Table 1) whereas a decrease in fracture density relates to an increase in crystal size (Facies S1; Table 1) (Korneva et al., 2017).

5.1 Structural contact relationship of the massive dolostones and limestones

Throughgoing, steeply dipping (near-vertical) fracture swarms, or fracture corridors (sensu Gabrielsen & Braathen, 2014; Questiaux, Couples, & Ruby, 2010) may be up to 30 m wide and occur on the eastern contacts between the massive dolostone bodies and the limestone (Figures 9a and 10). They dissect the massive dolostone bodies and all beds within them and exhibit no evident shear displacement. The
fracture corridors are dominated by long joints (10–15 m vertical direction) and veins (Figure 9c–d), which are significantly wider (up to 20 cm) than elsewhere in the field area. Chaotic, highly dilatant breccias (sensu Woodcock, Dickson, & Tarasewicz, 2007) with highly angular limestone clasts (0.5–3 cm) in a matrix of finer grained carbonate are observed, as are a range of fracture cements that includes haematite, barite, gypsum and anhydrite cements (10%–30% of the brecciated volume; Figures 4e and 10b) with zonation and abundant syntaxial textures.

Partial dolomitization occurs for 2–4 m eastwards of the fracture corridors, away from the HFF fault (Figure 10d). This partial dolomitization occurs preferentially within the mudstone matrix and comprises euhedral to subhedral crystals, 50 μm to 200 μm in size, with cloudy cores and clear rims (Figure 10c). The cloudy cores are often dissolved to produce intracrystalline porosity (Figure 10d).

6 | DISCUSSION

6.1 | Fluid composition and source

Unaltered limestone samples have a REE composition typical of unmodified seawater (Figure 8b and c) (Koepnick

FIGURE 9  (a) Geological map of the study area, showing fracture orientations collected within the massive dolostone bodies and the adjacent host limestones. Fractures associated with the NMD body strike N–S to NNE–SSW direction, whereas fractures associated with the SMD body strike NW–SE, with secondary N–S and E–W trends. Fracture trends within the massive dolostone bodies appear to be controlled by the strike of the fracture corridors that bound these massive bodies. (b) Fracture frequency vs. distance from HFF for the dolostone bodies and host limestone. The inner damage zone is typically characterized by fracture densities of >50 per metre, whereas the outer zone is characterized by 5–20 fractures per metre. (c) Field photo of the inner damage zone within the north massive dolostone (NMD) body, highlighting intensely fractured dolomite, with inset showing calcite (c) cemented fractures. (d) PPL image of calcite (c2) cement fracture fill within dolomite (d1), (e) cathodoluminescence (CL) image of bright orange luminescent calcite (c2) cement fill.
et al., 1985; Reilly, Miller, & Feigenson, 2002). Recrystallized limestones and dolostones have lower total REE concentrations and a flattened HREE profile than unaltered limestones, potentially as a result of precipitation under suboxic conditions (Haley et al., 2004). All dolostone bodies have Fe and Mn concentrations >100 ppm (Figure 8c), also consistent with dolomitization under suboxic conditions (Brand & Veizer, 1980; Land, 1980). The similarity

**FIGURE 10** Characterization of fracture corridors. (a) Overview of fracture corridor adjacent to the north massive dolostone body (NMD). (b) Throughgoing nature and dense network of fractures within the corridor, (c) with the inset highlighting the calcite cement fill within some of these fractures. (d) Fracture corridor associated with the south massive dolostone body (SMD), with the photomicrographs showing the degree of dolomitization relative to the fracture corridor (see Figure 9a for special reference). (i) Plane polarized light (PPL) image of host limestone displaying minimal dolomitization of the mudstone matrix. (ii) Plane polarized light (PPL) image showing degree of dolomitization by euhedral to subhedral crystals increases of both the matrix and allochems within the fracture corridor. The dolomite crystal cored are often dissolved to create intracrystalline porosity. (iii) Plane polarized light (PPL) image showing complete dolomitization by euhedral to subhedral dolomite crystals with cloudy cores and clear rims.
in the REE profile of the recrystallized limestone and dolostones could indicate rock-buffering during dolomitization (Banner, Hanson, & Meyers, 1988). However, $\delta^{13}C_{\text{dolomite}}$ have a wider range (from $+1.3^{\circ}$ to $-2.8^{\circ}$ VPDB) than $\delta^{13}C_{\text{limestone}}$ ($+1.0^{\circ}$ to $-1.0^{\circ}$ VPDB) and Miocene seawater ($\delta^{13}C > 0^{\circ}$ VPDB; Prokopf, Shields, & Veizer, 2008) suggesting that fluid–rock ratios were high (>ca. $10^4$; Banner et al., 1988). The source of the lightest carbon isotopic values is unclear; they could reflect an input of organic carbon or mantle-derived CO$_2$, although volcanism within the study area was minor and dated to have occurred for only a short period, at rift initiation (Bosworth, 2015).

Given this, the REE composition of the dolostone is interpreted to be indicative of dolomitization from seawater. This seems reasonable since (i) seawater is the most potent near-surface fluid for dolomitization as it contains high concentrations of magnesium and is volumetrically abundant and (ii) the $^{87}Sr/^{86}Sr$ ratios measured in the NMD and SMD coincide with Oligo-Miocene seawater $^{87}Sr/^{86}Sr$ ratios. There is no evidence of data clustering for individual dolostone body types

![Figure 11](image-url)

**Figure 11** Strontium isotopic ratios for the host limestone and dolostone bodies, relative to the seawater strontium curve (Koepnick et al., 1985) and tectonic events of the Hammam Faraun Fault block. The unaltered and recrystallized limestones correspond to Late Eocene to Early Oligocene seawater. The $^{87}Sr/^{86}Sr$ ratios for all dolostone bodies ranges between 0.708063 and 0.708576, which coincides with late Oligocene to middle Miocene seawater $^{87}Sr/^{86}Sr$ ratios. There is no evidence of data clustering for individual dolostone body types.
The only other potential source of fluid for dolomitization would be compactionally expelled brines from the syn-rift, hanging wall succession of siliciclastic sediments (Nukhul and Rudeis Formation). This seems unlikely since such a mechanism usually involves a low-velocity, single-pass expulsion of fluid which rarely has sufficient volume or Mg-enrichment for massive dolomitization (see discussions in Machel & Lonnee, 2002; Frazer, Whitaker, & Hollis, 2014; Gomez-Rivas et al., 2014).

In thin section, replacive dolomite in the NMD and SMD has a mottled luminescence (Figure 5b, d and f). Mottling is also sometimes seen in the DT (Figure 6d) but here usually crystal fabric is usually better preserved (e.g. Figure 6f). This implies recrystallization (i.e. more than one phase of dolomitization) within the NMD and SMD, decreasing within the DT. The oxygen isotopic composition of all dolostone samples is wide-ranging, suggesting either a range of fluid compositions and/or fluid temperatures,

**Figure 12** (a) Conceptual model of fluid flux on the HFF Block during formation of the massive dolostone bodies. Seawater was drawn down surface-breaching faults within the footwall and migrated up-dip towards the HFF by geothermal convection along the Nubian sandstone aquifer. Once the HFF breached the surface, it is also possible that seawater was drawn in, and convected along, the HFF. The hypothetical function of the fracture corridors is illustrated with them either ponding fluids or transporting fluids (after Hollis et al., 2017). (b) A schematic conceptual diagram showing a segmented normal fault system with soft-linked (i) and breached (iii) relays as well as simpler single-segment portions of the fault system (ii). The figure illustrates in map-view (top) and 3D (bottom) the presence of damage zone anomalies at the relays, where the damage zone exhibits greater width, higher fracture intensities, and the fractures show a greater range of orientations. Note from the 3D block diagrams how the relays and their anomalously wide and complex damage zones are expressed in three dimensions. Where such damage zones form in areas of fault interaction they are commonly good conduits for fluid flow and tend to preferentially localize fluid flow (e.g. Dimmen et al., 2017; Gartrell et al., 2004). In the HFF area, we suggest that massive dolomitization is preferentially localized where these damage zones occur at relays or fault jogs like this.
such as would occur during numerous passes of fluid. It is proposed, therefore, that multiple phases of dolomitization occurred from numerous passes of seawater under suboxic conditions. Fluid temperatures could have varied as a result of footwall uplift, whereas fluid composition might have evolved through time by fluid–rock interaction or by changes in seawater chemistry. The latter is considered particularly important, as increasing isolation of the hanging wall basin from the open ocean resulted in desiccation and evaporite precipitation by the middle Miocene (ca. 16 Ma; Langhian Kareem Formation; Gawthorpe et al., 2003). This could have introduced denser, oxygen-isotopically enriched seawater to the HFF during the latest stages of dolomitization.

6.2 Timing and temperature of dolomitization

The close proximity of the massive dolostone bodies to the HFF, within its damage zone and bounded by HFF-parallel fracture corridors, suggests that this crustal lineament strongly controlled fluid flux. Although movement along the HFF began along isolated fault segments at rift initiation (ca. 26 Ma), other faults remained active until around 17 Ma when deformation became localized on the HFF (Gawthorpe et al., 2003). At this point, the linkage of fault segments along the HFF localized displacement along its length as it became a major, surface-breaching crustal-scale fault with km-scale offset, whereas other faults (e.g. the Thal Fault) died (Gawthorpe et al., 2003; Jackson, Gawthorpe, Leppard, et al., 2006; Jackson, Gawthorpe, & Sharp, 2006). Since the massive dolostone bodies are confined to the damage zone of the HFF, this would suggest that they started to form after rift initiation and that dolomitization continued during the rift climax, once the damage zone of the HFF had formed.

The strontium isotopic composition of the NMD, SMD and DT shows a wide range of values (0.708063–0.708576; Table 2), although the NMD and SMD mostly coincide with early to mid-Miocene seawater (Figure 11). This would suggest either (i) dolomitization from contemporaneous seawater over a period of ca. 8–10 Ma (prior to the mid-Miocene) or (ii) dolomitization from penecontemporaneous seawater that had undergone compositional modification by fluid–rock interaction. Although fluid–rock interaction cannot be ruled out, the coincidence of the strontium isotope data to the estimated timing of dolomitization established from field relationships, along with good petrographical evidence for multiple phases of dolomitization, suggests that the massive dolostone bodies formed by multiple fluxes of relatively unmodified seawater during the late Oligocene–mid-Miocene. This is supported by the interpretation that massive dolostone bodies post-date the stratabound dolostone bodies, which are dated from field relationships and strontium isotope data as rift initiation (28–24 Ma; Hollis et al., 2017).

The massive dolostone bodies exhibit non planar textures which infer dolomitization at temperatures of >50–60°C (Sibley & Gregg, 1987). Assuming that dolomitization occurred from seawater, fluid temperatures of ca. 50–100°C are calculated using δ18O$_{\text{dolomite}}$, late Oligocene to middle Miocene seawater and the fractionation factor of Matthews and Katz (1977) (δ18O$_{\text{SMOW}}$ of +1‰ to 0‰ from Veizer & Prokoph, 2015; Figure 7b). If the dolomitizing fluid was more evolved (i.e. heavier δ18O$_{\text{water}}$) then fluid temperatures would have been higher; i.e. we consider our calculated temperatures to be a minimum estimate. The present-day geothermal gradient is 47°C/km (Boulos, 1990), and hence a temperature of 100°C would require a burial depth of about 2 km. The Thebes Formation was buried to a maximum depth of ca. 900 m in the early Miocene (Hirani, 2014), but fluids of 45°C are brought to the surface at the hot springs today along the HFF (Figures 1 and 2; Sturchio et al., 1996). It is likely, therefore, that fluid temperatures were >5°C hotter than the host rock, i.e. hydrothermal, as per the definition of Machel and Lonnee (2002).

6.3 Dolomitization mechanism

The position of the massive dolostone bodies within the damage zone of the HFF is indicative of a genetic relationship between the HFF, fluid flux and dolomitization. Geochemical fingerprinting suggests dolomitization from seawater, which could have entered the Thebes Formation in two ways. Firstly, cool seawater could have been drawn-down faults within the hanging wall basin and circulated via a permeable aquifer towards the HFF (Hollis et al., 2017). The most likely aquifer is the Palaeozoic Nubian aquifer, which is dominated by quartz arenite and is therefore both geochemically inert and highly permeable (Hirani et al., 2018; Hollis et al., 2017; Nabawy, Geraud, Rochette, & Bur, 2009). Given that water depths up to 1,000 m persisted within the hanging wall basin throughout the Miocene (Youssef, 2011), and the mapped presence of the permeable Nubian Sandstone, fluid flux by this mechanism could have persisted throughout the syn-rift period, driven by high heat flux along the HFF. Once the HFF had tipped out at the sea floor (by the mid-Miocene—Gawthorpe et al., 2003; Jackson, Gawthorpe, Leppard, et al., 2006), a second mechanism could have been established, whereby seawater was drawn down and circulated along the plane of the HFF by geothermal convection (Figure 12). A combination of both mechanisms, not necessarily occurring simultaneously, could explain the wide range of fluid compositions and temperatures implied by isotopic signatures.
The absence of saddle dolomite, zebra dolomite fabrics and hydrobrecciation, within the core of the massive dolostone bodies is notable. The genesis of zebra dolomite textures has been ascribed to numerous mechanisms but is most commonly interpreted to occur by the expulsion of overpressured fluids (e.g. Davies & Smith, 2006; Dewit et al., 2014; Swennen, Vandeginste, & Ellam, 2003), for example along zones of prior structural weakness (e.g. Gaspirrini, Bechstadt, & Boni, 2006). In many cases, it occurs in basins that have undergone fault reactivation and basin inversion along crustal-scale, strike-slip lineaments. It is possible that pockets of pressurized fluids occurred within the fault segments, and the margins of the damage zone, as discussed below. Overall, however, the absence of zebra dolomite and hydrobrecciation means that there is little evidence within the massive dolostones for pervasive pore pressure build-up, rupture and rapid fluid expulsion (e.g. by seismic valving), consistent with convection of seawater along the HFF.

**6.4 Role of fracture corridors in the formation of the massive dolostone bodies**

The massive dolostone bodies occur within the damage zone of the HFF and are structurally bounded by fracture corridors (Figures 9a and 10) that display evidence of dilatant brecciation (Figure 10b). Their relationship and timing is therefore critical to understanding the process of dolomitization. Five hypotheses are proposed to explain their coincidence:

1. The fracture corridors post-date the massive dolostones; their localization at the limestone–dolostone boundary reflects a mechanical contrast at the interface.
2. The fracture corridors, or precursor structural heterogeneities, separate the massive dolostone bodies from pristine limestones because they were effective seals. This may have been because the fracture corridors were cemented, which would have hindered the transport of dolomitizing fluids across them.
3. The fracture corridors were very effective conduits. Dolomiting fluids entering from the HFF accessed the damage zone, leading to dolomitization. However, upon entering the bounding fracture corridor, the fluids were evacuated vertically rather than laterally into the footwall and, hence, did not travel across the fracture corridors to dolomitize the adjacent limestones.
4. The fracture corridors alternated between being transiently open and sealed through time.
5. A combination of one or more of the above, where later reactivation of a syn-dolomitization corridor dominates its present-day expression.

Overall, the first hypothesis is inconsistent with field observations that the NMD and SMD, but not the DT, terminate abruptly against the fracture corridor. If the fracture corridor had localized on the dolostone–limestone interface after dolomitization, we would expect the interface to be more irregular, mimicking the termination of the DT, whereas the contact is planar-tabular. Furthermore, there is no shear displacement associated with the fracture corridor; hence, the juxtaposition of limestone and massive dolostones across the fracture corridor is imposed by diagene-
sis, not by shear offset.

The second hypothesis, that the fracture corridors acted as effective seals during dolomitization, is contradicted by the observation of extensive multi-phased calcite cementation in the fracture zones (Figure 4e). This suggests that the fracture corridors at least periodically formed open fluid conduits, probably with repeated crack-seal events. Although the calcite cements post-date dolomitization, their presence indicates that the fractures within the fracture corridors could have been transiently opened and then sealed (Figure 9c–e), rather than acting as fully effective conduits (hypothesis 3). This supports the fourth hypothesis that the fracture corridors underwent phases of being an active fluid conduit and then sealing. The presence of dilatant breccias also suggests episodic and localized high pressure events, perhaps triggered by seismic movements along the HFF (sensu Woodcock et al., 2007).

It has been noted that the DT show preferential dolomitization of the remobilized matrix-supported conglomerates and grainstone turbidites (R1 and R2), indicating that there was a facies control on their formation, compared to the massive dolostone bodies, which are non facies selective. Furthermore, the DT comprise more planar, and larger, dolomite crystals (up to 75 μm) compared to the NMD and SMD (10 and 30 μm) and are more depleted in Fe and Mn (Figure 8a). The field relationship between the DT and the fracture corridors is ambiguous, and so their relative timing is equivocal. On the one hand, strontium isotope data imply that the DT are the oldest dolostone bodies in proximity to the HFF (mid-late Oligocene; Figure 11), suggesting that they may predate the formation of the NMD and SMD. In this case, the absence of recrystallization within the DT could reflect either a retreat of the dolomitization front towards the HFF or subsequent ponding and evacuation of dolomitizing fluids by the fracture corridors during the rift climax. Alternatively, the DT represent small volumes of fluid that migrated laterally across the fracture corridors within specific beds, whereby the efficiency of dolomitization decreases.

Overall, the fracture corridors are interpreted to have prevented large volumes of dolomitizing fluids from accessing limestones further into the footwall by alternate
(i) fracture corridor sealing and (ii) vertical fluid evacuation from the exposed structural level in question. Therefore, we conclude that the fracture corridors were important boundaries during dolomitization, with hypothesis 4 being the preferred scenario (Figure 12). The structural trend defined by this corridor, enhanced, by the mechanical contrast at the interface between the massive dolostones and the limestones, was then likely to have been reactivated during later fault slip. The abundance of uncemented joints in the fracture corridor means that at least part of the expression of the fracture corridor is a result of later (post-dolomitization and calcite cementation) reactivation of the fracture corridor, as suggested by hypothesis 5. In summary, the fracture corridors are interpreted to have acted as seals and prevented the flux of dolomitizing fluids further into the footwall, but with transient vertical fluid flux (hypothesis 4) with overprinting during later movement along the HFF (hypothesis 5).

6.5 | Implications

In summary, this study provides evidence for localization of dolomitization on the HFF. Dolomitization is predominantly fabric-destructive and fracture permeability highest (high density, length and connectivity) within the inner damage zone of the HFF, becoming more fabric-retentive and less densely fractured within the outer damage zone (Figures 9 and 10). This implies more frequent and pervasive dolomitization and fracturing close to the HFF. The preservation of stratabound, planar dolomite fabrics within the DT, on the margins of the massive dolomite bodies could support this notion of a retreating dolomitization front, towards the HFF, through time. It has been argued that fracture corridors played an important role in inhibiting the lateral flux of fluids away from the damage zone, but the absence of massive dolostone elsewhere on the HFF Block is also noteworthy and suggests a specific localization of fluid flux. Areas of fault intersection on normal faults, such as at fault jogs and relay ramps, commonly show an increase in fracture density and fracture connectivity that can facilitate vertical fluid flux (Dimmen, Rotevatn, Peacock, Nixon, & Naerland, 2017; Gartrell et al., 2004). The massive dolostone bodies described in this study occur at a jog on the HFF, immediately south of the Gharanal Accomodation Zone (GAZ; Figure 1a). It is suggested, therefore, that their location is related to a higher density of connected fractures at this position on the HFF, around the relay ramp on the fault jog, compared to other along-strike positions. This enhanced vertical permeability could have channelized fluids into the basal Thebes Formation where they ponded at the base of the Thebes Formation. This resulted in a localization of dolomitization at the base of the upper Themes Formation, above the mudrocks of the Esna Formation and beneath the fine-grained, low permeability upper Thebes Formation (Hirani, 2014).

This interpretation of localization of dolomitization at zones of structural complexity appears to be supported by observations elsewhere in the Suez Rift. Here, dolostones are found in the footwall of significant border faults in several places on both the western and eastern flanks of the rift such as the Abu Shaar Platform (Clegg, Harwood, & Kendall, 1998; Coniglio, James, & Aissaoui, 1988), Gebel Zeit (Winn, Crevello, & Bosworth, 2001) and also at El Tor, ca. 160 km south of the study area. In each case, the occurrence of dolostone coincides with jogs in the trace of major normal faults, or near the intersection of major normal faults and major, rift-wide accommodation zones (Moustafa, 2003; Figure 1a). The HFF lies in the transition zone between the northern and central dip domain at a regional cross-rift accommodation zone (termed the Gharandal transfer zone in Moustafa, 2003). The Abu Shaar and El Tor Platforms similarly occur on the margins of the cross-rift accommodation zones and have associated hot springs. The interaction of these accommodation zones with crustal-scale normal faults may have perturbed local stresses and enhanced fracture apertures, enhancing the vertical flow properties. This larger scale structural complexity is closely tied to complexity on the scale of fault-related damage (e.g. Rotevatn & Bastesen, 2014), itself related to perturbed and enhanced stresses during fault interaction (Crider & Pollard, 1998; Kattenhorn et al., 200; Maerten, Gillespie, & Pollard, 2002; Soliva, Benedicto, Schultz, Maerten, & Micarelli, 2008). Such damage may further ensure localized conduits for fluid flow by vertical permeability enhancement and, potentially, local dilation that promotes open fractures (e.g. Ferrill & Morris, 2003; Gartrell et al., 2004; Rotevatn & Bastesen, 2014).

Consequently, we propose that dolomitization occurred preferentially where heated fluids were vented within zones of structural complexity, such as relay ramps (e.g. Fossen & Rotevatn, 2016), fault intersections (e.g. Peacock et al., 2017) or rift-scale accommodation zones (e.g. Morley et al., 1990). There are differences in the size and shape of the dolostone bodies, however. On the HFF Block, massive dolostone bodies are restricted to the fault damage zone. On the Abu Shaar Platform, dolomitization is pervasive, ubiquitous and fabric-retentive (Clegg et al., 1998; Coniglio et al., 1988), implying that other controls are also important. The Thebes Formation is a deep-water succession of mudstones and wackestone with interbedded conglomerates and grains, and permeability prior to dolomitization appears to have been low (Korneva et al., 2017). In contrast, the Miocene carbonate platform on Abu Shaar dominantly comprised skeletal...
pack-grainstone, mostly composed of aragonitic and high magnesium calcite grains (Clegg et al., 1998; Coniglio et al., 1988). If the style of dolomitization was controlled by sediment composition and effective porosity, then pervasive, fabric-retentive dolomitization on the Abu Shaar Platform could reflect the higher permeability and greater reactivity of the sediments. Furthermore, the fault damage zone at Abu Shaar has not been characterized and it is possible that a different fracture distribution and/or density than that adjacent to the HFF offered a contrasting style of fluid flux.

In a global context, this implies that it might be possible to predict the likely occurrence of massive, non facies selective dolostone bodies in rift basins within zones of structural complexity. However, the petrophysical characteristics of those bodies may be variable and dependant on the precursor sedimentary texture and composition. In many case studies, fault-controlled hydrothermal dolostones exhibit more complex structural, petrographic and geochemical fabrics than observed in this study; this most likely reflects their history of multiple burial, exhumation and fault reactivation events (Hollis et al., 2017). The young age and burial history of this study area potentially provides a window into the earliest stages of massive, fault-controlled, hydrothermal dolomitization. This may provide a template that is subsequently overprinted by more texturally complex bodies formed during fault reactivation, overpressure release and rupture.

7 SUMMARY AND CONCLUSIONS

Integration of field-based data, petrography and geochemistry has been used to determine the controls on the formation of massive dolostone bodies adjacent to the HFF.

1 Massive, non facies selective dolostones and facies selective dolostone tongues within the Thebes Fm. on the HFF Block, Suez Rift. They formed contemporaneously by the convection of seawater within the damage zone of the HFF.

2 Although ghosts of some precursor facies can be seen in outcrop, dolomitization is mostly fabric-destructive. Within the inner damage zone, the dolostone is highly fractured and brecciated, whereas rock fabric is better preserved in the outer damage zone, particularly within the stratabound dolostone tongues. Consequently, the rock physical properties of the massive dolostone will vary significantly over the ca. 500 m width of the bodies.

3 Petrographical fabrics and wide-ranging geochemical signatures indicate multiple phases of dolomitization and recrystallization over a period of ca. 10 Myr, coincident with the rift climax. This demonstrates that even in a young dolostone body that has not undergone subsequent burial and exhumation, rock fabric and geochemical proxies can be highly heterogeneous.

4 The young age and simple burial history of the HFF means that the massive dolostone bodies observed within its damage zone in this study potentially provide a window into the earliest stages of fault-controlled hydrothermal dolomitization, which are often masked by more complex textures formed by subsequent periods of fault reactivation, seismic rupture and brecciation.

5 Fracture corridors played a fundamental role in controlling the localization of dolostone formation by alternately acting as barriers and conduits to fluid flux. In doing so, the massive dolostone bodies were confined to the damage zone of the HFF. Short dolostone tongues show that only minor volumes of fluids penetrated outwards, away from the fault damage zone, perhaps prior to the formation of the fracture corridors.

6 The position of dolostone bodies on the HFF may reflect localization of vertical fluid conduits and dilation due to perturbed and enhanced stresses on the HFF. This stress may form during fault interaction at zones of structural complexity associated with rift accommodation zones and jogs within the trace of the bounding crustal fault. This suggests that the occurrence of massive dolostone bodies should be predictable within rift basins, although the size and petrophysical characteristics of the dolostone body will vary in accordance with precursor limestone composition, facies architecture, fluid temperature and structural heterogeneity.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of the article.

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