Reconstructing the level of the central Red Sea evaporites at the end of the Miocene

Neil C. Mitchell1 | Wen Shi2 | A.Y. Izzeldin3 | Ian C. F. Stewart4

Abstract
Reconstructing the original depositional level of the Mesozoic and older c'salt giants' can reveal if their basins became filled to global sea level, but is complicated by dissolution, diapirism and because the time elapsed is so great. This is less of a problem in the Red Sea, a young rift basin that is transitioning to an ocean basin and where the evaporites away from coastal fringes are less affected by diapirism. In this study, we explore vertical movements of the evaporite surface of the central Red Sea imaged with deep seismic profiling, for the period of time after most evaporite deposition ended at 5.3 Ma (the Miocene-Pliocene boundary). This boundary is readily mapped across the basin as a prominent reflection in seismic data correlated with stratigraphy at three DSDP sites. We quantify changes in the average elevation of the evaporite surface due to (a) thermal lithospheric subsidence, (b) isostatic loading by Plio-Pleistocene sediments and water, (c) deflation needed to balance the volume of evaporites overflowing oceanic crust of 5.3 Ma age, (d) loss of halite by dissolution and (e) dynamic topography. Our best estimate of the evaporite level (−132 m air-loaded or −192 m water-loaded) lies below the range of estimated global sea level towards the end of the Miocene, suggesting that the basin remained under-filled. If geological interpretations of shallow water conditions existing at the end of the Miocene (Zeit Formation) are correct, this implies that the water level of the Red Sea declined and was unstable. These calculations illustrate how spreading of evaporites can enhance thermal subsidence to cause rapid development of accommodation space above major evaporite bodies, which in the Red Sea case has remained largely unfilled.

KEYWORDS
central Red Sea, Messinian, rift basins, Salt giant, sediment flux, tectonics and sedimentation

1 | INTRODUCTION

Evaporites (sedimentary rocks formed primarily of minerals precipitated from seawater and otherwise commonly referred to as ‘salt’ [Warren, 2010]) have been found with deep seismic profiling and drilling along many margins of the North and South Atlantic basins and Gulf of Mexico (Emery, 1977; Evans, 1978; Pautot et al., 1970; Rona, 1982). Such deposits can reach kilometres in thickness and arise from evaporation of seawater in highly restricted basins during their evaporite
phases (Warren, 2010). Working out whether the evaporites continued depositing until they completely filled their basins to global sea-level or the basins remained under-filled could be useful to help understand the environments of the basins at the ends of their evaporite deposition phases. An under-filled basin could potentially arise from gradual supply of seawater through a permeable barrier. In such a case, the water level within the basin would likely have been unstable, due to varying fluxes of seawater inflow, rainwater and evaporation. If a basin were identified as under-filled, stream profile analyses, which rely on assuming stable water levels and precipitation (Wilson et al., 2014), cannot be used to reconstruct solid-Earth uplift patterns. If evaporites rapidly fill basins to the existing global sea level, this may indicate a greater efficiency of seawater influx and high rates of evaporation relative to precipitation. The nature of the sediments immediately overlying the evaporites can potentially suggest whether the basin was filled or not. For example, shallow-marine sediments overlie the Jurassic evaporites in the Gulf of Mexico, suggesting that the evaporites filled that basin to global sea level (Hudec et al., 2013). However, there remains the possibility that the balance of inflow, evaporation and precipitation could potentially change to allow shallow marine carbonates and other sediments to deposit in a deep basin rather than evaporites, so this evidence may not be unequivocal.

An alternative to working out the degree of under-fill from the sediments overlying the evaporites is to reconstruct the original depositional level of the evaporites by correcting for effects of subsidence, flowage, isostasy, dynamic topography and dissolution, as far as they are known. For older deposits, this presents major challenges, as the history of rift development and transition to seafloor spreading is often not so well known. Subsidence can be locally rapid and otherwise complex during asymmetric continental rifting (e.g. Mohn et al., 2015). If evaporites were deposited on oceanic lithosphere, we can also expect rapid subsidence. Based on occurrences of dipping seismic reflection sequences under evaporites imaged off South America and West Africa, Jackson et al. (2000) suggested the evaporites were deposited onto the flanks of a subaerial mid-ocean ridge in the post-rifting phase of the basin. Dipping reflections are the result of large widespread lava flows like those found around subaerial spreading centres of the Afar and Iceland (Mutter et al., 1982). In contrast, Torsvik et al. (2009) interpreted results of plate reconstructions as showing that the South Atlantic evaporites were deposited onto extended continental crust during the Aptian rift phase and were subsequently split by seafloor spreading. In the Gulf of Mexico, Imbert (2005) interpreted seismic and magnetic data as showing the evaporites there had been deposited on oceanic crust, whereas Pindell and Kennan (2007) interpreted the presence of evaporites overlying oceanic crust in the Gulf of Mexico as caused by flowage, not deposition. The diversity of these opinions arises partly because the deposits lie very deep beneath the surface and are difficult to image seismically. Evaporite movements could in principle be dated from the ages of sediments deposited in front of and subsequently over-ridden by the evaporites, but in practice recovering samples would be prohibitively expensive. In addition, recognizing oceanic crust and the ocean-continent boundary is often not straightforward (Eagles et al., 2015), hence some Aptian evaporites previously thought to have lain on oceanic crust (Contrucci et al., 2004) have subsequently been suggested to have been deposited on hyper-extended continental crust (Cowie et al., 2017).

The Red Sea presents one of the best opportunities to reconstruct the original level of the surface of a major evaporite body or ‘salt giant’ as evaporite deposition in deep waters ended more recently, at the end of the Miocene at only ~5.3 Ma (Hughes & Beydoun, 1992). Because the Red Sea evaporites are much younger than the Mesozoic evaporites mentioned above, they are likely to have been less affected by dissolution, subsidence calculations should generally be more accurate and correcting the evaporite surface elevation for halokinetic movements is more straightforward. The underlying crust has been demonstrated to be largely oceanic (Shi et al., 2018), making subsidence calculations more straightforward than for continental rifts. The amount of evaporites of 5.3 Ma and older that have overflown crust of 5.3 Ma and younger age can be quantified from a series of deep-seismic reflection profiles (Izzeldin, 1982, 1987) (Figure 1). In particular, the top of the evaporites are well-imaged in seismic reflection data (‘S-reflection’ (Ross & Schlee, 1973)) allowing their present elevations to be mapped out easily across the basin. These data show that the S-reflection in deep water has less topography of diapirs than it does in shallow coastal waters (e.g. Savoyat et al., 1989), where the evaporites are loaded by denser terrigenous sediments and mobilised to form diapirs. The loading of the evaporites near the coasts presents a problem for this reconstruction because it may have forced the evaporites seawards by some poorly known amount, which we call ‘shunting’. However, as this shunting would be balanced by the evaporite body thickening away from the coast or further overfloving oceanic crust, it ensures the final estimate of the 5.3 Ma elevation of the evaporites is

---

**Highlights**

- Evaporite elevation adjusted for subsidence, flowage, dissolution and dynamic topography.
- End-Miocene evaporite basin was underfilled compared with global sea level.
- Unconformity formed during Messinian Salinity Crisis in Mediterranean and/or global lowstands.
an upper bound and hence reinforces the results here suggesting that the basin remained under-filled. As there are many dimensions and displacements involved in the calculations, we summarise the average values in Figure 2 to help the reader.

1.1 Datasets used

1.1.1 Seismic reflection data

The deep two-dimensional seismic reflection data were collected along the lines shown in Figure 1 and described and interpreted by Izzeldin (1982, 1987, 1989). One example is shown in Figure 3a. These data were collected with a 2.4 km streamer to 5–6 s seismic two-way time, imaging below the evaporites. Processing included move-out corrections but not migration. The reflections interpreted by Izzeldin (1982) included the seabed, the S-reflection at the top of the Miocene evaporites (Ross & Schlee, 1973), and the basement reflection. As shown in Figure 3a, the basement appears as a set of diffuse hyperbolae and occasionally more discrete reflection packages, whereas the S-reflection is prominent and continuous almost everywhere.

The basement reflections were converted to depth below the S-reflection by Shi et al. (2018) using 4.21 km/s P-wave seismic velocity (Vp) for the evaporites. This velocity was based on the results of refraction experiments in the central Red Sea (Davies & Tramontini, 1970; Drake & Girdler, 1964; Tramontini & Davies, 1969), which show well-defined phases (linear branches on seismic time-offset graphs) for P-waves travelling within the evaporites (Davies & Tramontini, 1970). This 4.21 km/s is slightly lower than 4.2–4.4 km/s in velocity models of Egloff et al. (1991) from deep seismic refraction experiments off the Sudanese coast, so it is potentially a lower bound on Vp (the effect of uncertainties is discussed later). Some reflectivity lies within the evaporite section in their upper parts (‘layered evaporites’ in
Figure 3a), likely due to layered sequences of shale, anhydrite and halite as recovered at the DSDP sites, hence our velocity may be too high locally because of 2–3 km/s velocities of the shale component (Whitmarsh et al., 1974). However, this error affects only our estimates of thickness of the evaporites that have overflown the 5.3 Ma boundary and hence their profile cross-sectional areas described later.

To calculate depths of the S-reflection beneath the seabed reflection, a 1.9 km/s velocity was used for the Plio-Pleistocene sediments overlying the evaporites, which was based on sample measurements (Whitmarsh et al., 1974) and seismic refraction data (Egloff et al., 1991; Gaulier et al., 1988). The depth-converted reflections are shown versus distance along the profiles in Figure 4, where distance is centred on the spreading axis interpreted from the axial magnetic anomalies in Figure 5a and Bouguer anomalies in Figure 5b (note this axial position is not everywhere centred on the valley shown in Figure 1 because evaporite flowage is asymmetric in places). The Plio-Pleistocene sediments are typically only 200–300 m thick away from the coasts (Mitchell et al., 2017, 2019; Ross & Schlee, 1973). They can be locally thicker nearer to the coasts in Figure 4. In those areas, our conversion from seismic two-way time to depth will be underestimated somewhat as a higher Vp would be expected for the compacted terrigenous sediments, but in practice these are small parts of the profiles and will have little effect on the average thickness used in the following calculations. A 1.538 km/s velocity for converting the seismic travel time to the seabed reflection was used as explained in Shi et al. (2018).

We have also updated a map of the depth to the S-reflection (Mitchell et al., 2017, 2019) with these deep seismic data. The result in Figure 6 was computed from the datasets shown in the inset map. To improve visibility, the S-reflection surface was interpolated onto a grid using a surface with tension procedure (Smith & Wessel, 1990) and grid values more than 0.15° of latitude or longitude away from survey lines were masked out.

1.1.2 | Basement age

We justify the assumption that oceanic crust underlies the central part of the central Red Sea later. In order to estimate
the position of 5.3 m.y.-old crust, we used the rotation pole of Chu and Gordon (1998) and assumed continuous seafloor spreading at the same rates predicted by that pole. This pole was derived from magnetic anomalies of Chron 2A (3.2 Ma) and agrees with modern rates derived from GPS measurements (Reilinger et al., 2015). Other data suggest the rates implied by this pole should be valid to 5.3 Ma. Arabia-Somalia opening rates in the Gulf of Aden derived by Fournier et al. (2010) for 0–2.5 Ma are no different from those of 2.5–5.0 Ma. DeMets and Merkouriev (2016) combined information from seafloor spreading anomalies in the Gulf of Aden with anomalies along the Southwest Indian Ridge to resolve the Nubia-Arabian plate motion since 20 Ma, also finding no resolvable change in motion since 5.3 Ma.

1.1.3 Bouguer gravity anomalies

The terrain-corrected Bouguer anomalies in Figure 5b were computed using the Sandwell et al. (2014) free air gravity anomaly grid (their version 24) updated with additional marine gravity measurements (Cochran et al., 1991; Girdler & Southren, 1987). The gravitational effect of the seabed was corrected using a seabed density of 2.2 g/cm³ and a bathymetry grid constructed from a global model (Smith & Sandwell, 1997, version 18) shown in Figure 1, where the bathymetry has been updated with multibeam data where available (Augustin et al., 2014, 2016, 2019; Ligi et al., 2012). The 2.2 g/cm³ density was chosen as a compromise between halite (2.16 g/cm³) and anhydrite (2.9 g/cm³) densities (Wheildon et al., 1974). This compromise assumes the section above the basement is mainly halite based on the transparency of seismic reflection images typically observed (Izzeldin, 1987), though presence of reflective seismic data suggesting also a layered evaporite sequence like that found at DSDP Sites 225, 227 and 228 (Whitmarsh et al., 1974). Their overlying Plio-Pleistocene sediments sampled at the DSDP sites have a similar density to halite (Mitchell et al., 2010) and are typically a uniform 200–300 m thickness away from the coasts (Mitchell et al., 2015, 2017; Ross & Schlee, 1973), so 2.2 g/cm³ is suitable for those sediments also.

Shi et al. (2018) have shown that Bouguer anomalies are strongly correlated with basement depths within 80 km of the axis. We have exploited this correlation to generate a grid
FIGURE 4 Profiles of seabed (green), S-reflection (top of the evaporites and other sediments of Miocene age, blue) and basement (red) interpreted from the seismic data. Plio-Pleistocene sediments lie between the seabed and S-reflection. Vertical bars mark 5.3 Ma basement estimated from distance to the spreading axis and the Nubia-Arabia rotation poles of Chu and Gordon (1998). Evaporites that have overflown this boundary are shaded grey. Evaporite top is shown in bold where used in the subsidence-corrected surfaces shown in Figure 11.
of basement depths shown in 3D in Figure 7 along with the bathymetry. This involved re-computing the correlation between the basement depths and the newly calculated Bouguer anomalies (0.0465 km mGal⁻¹) and applying it to the Bouguer anomaly grid within 80 km of the spreading axis. For regions further away from the axis, we used Delauney triangulation to interpolate using the Bouguer-derived values, basement depths from the any deep seismic data lying beyond 80 km and some points on land beyond the coastal sedimentary plain where basement appears exposed.

1.2 Background to the central Red Sea evaporites and basin structure

1.2.1 Tectonic framework

Extension of the African-Arabian shield started at ~27.5 to 23.8 Ma along the Eritrean margin, though more extensive normal fault activity was associated with a phase of volcanism at 24–23 Ma (Bosworth, 2015; Bosworth et al., 2005; Stockli & Bosworth, 2019). Plate-tectonic extension rates for the Nubia-Arabia boundary were derived by Chu and Gordon (1998) from identifications of magnetic anomaly Chron 2A. As the rotation pole lies in the Mediterranean, those rates increase southwards, from 11 to 15 mm/yr⁻¹ going from 23° to 18°N, and are broadly compatible with modern GPS measurements (Reilinger et al., 2015) suggesting that there has been little change in the Nubia-Arabia rotation rate since Chron 2A.

1.2.2 Evaporites

Most of our knowledge of the evaporites deposited within the basin comes from marginal commercial wells. According to Hughes and Beydoun (1992), the evaporites were deposited in the basin from about the start of

FIGURE 5 Potential field data for the central Red Sea. (a) Aeromagnetic anomalies originally shown in Izzeldin (1987) and updated by Zahran et al. (2003) reduced to the pole. Annotation as Figure 1. Asterisks mark an example pair of anomalies that are parallel and symmetrical about the axis. From the spreading rates of Chu and Gordon (1998), these are predicted to be Chron 5 or ~10 Ma. (b) Bouguer anomalies computed from the free-air gravity data of Sandwell et al. (2014) updated with marine gravity measurements carried out on RRS Shackleton (Girdler & Southren, 1987) and RV Conrad (Cochran et al., 1991). White cross symbols locate the interpreted ridge axis.
the Miocene, corresponding to the time of initial rifting (Bosworth, 2015; Bosworth et al., 2005), and comprised anhydrite interspersed with shales and carbonates through the Lower Miocene. The main body of halite-dominated evaporites was deposited in the Middle Miocene, followed by anhydrite-dominated evaporites in the Upper Miocene (known as the Zeit Formation by Egyptian geologists). Although the latter also contain significant clastic sediments, sediment cores recovered during Deep Sea Drilling Project Leg 23B in deep waters (Whitmarsh et al., 1974) from Sites 225 and 227 contained interbedded halite, anhydrite and shale, and those from Site 228 contained anhydrite and shale. The recovered lithologies are summarised in Figure 8a. The layered stratigraphy in the upper part of the evaporites found by drilling is compatible with the reflective segment of the seismic reflection data in Figure 3a and 3b immediately below the S-reflection, whereas a more transparent character in much of the deeper parts of the deep seismic data is compatible with the earlier largely halite-dominated evaporites (Izzeldin, 1987). Thus, layered evaporites containing mixed lithologies overlying more massive halite appear to extend across much of the central Red Sea away from the margins and the axial deeps. Biostratigraphy from commercial wells along the central Red Sea coasts (Bunter & Abdel Magid, 1989; Hughes & Beydoun, 1992) confirms a Pliocene age of the sediments immediately overlying the layered evaporites that were reported for the DSDP sites (Stoffers & Ross, 1974).

The origins of the solutes that formed the evaporites is important for our later discussion. They may have been supplied via Neotethys/Mediterranean (Bosworth et al., 2005; Coleman, 1974; Hughes et al., 1992; Orszag-Sperber et al., 1998). Evidence for this includes calcareous nanofossils identified by Boudreaux (1974) within the intercalated black shales of the evaporites drilled at Deep Sea Drilling Project (DSDP) Sites 225 and 227 (Figure 1), which have affinities with the Mediterranean (Orszag-Sperber et al., 1998), and Mediterranean fauna in some marginal commercial wells (Bunter & Abdel Magid, 1989; Hughes, 2014). Gargani et al. (2008) identified a structural high within the Gulf of Suez (Araba Wadi/ Zafarana Platform) as the likely sill between the Mediterranean and Red Sea domains. Their reconstructions of that area in the Messinian suggest that the sill lay 50 m below modern Mediterranean Sea level.

Alternatively, some or most solutes may have been supplied from the south via the Gulf of Aden. Hughes et al. (1991) interpreted the biostratigraphy of marginal wells as showing a progressive penetration of marine waters from the Gulf of Aden into the Red Sea from the late Oligocene onwards. Crossley et al. (1992) showed marine muds extending from the south as far as 19°–20°N at 25 Ma and along the whole Red Sea by 20 Ma (Lower Miocene). Orszag-Sperber et al. (1998) mentioned that the Upper Miocene units contain fauna with Indian Ocean affinities. As the southern area of the Red Sea around the present Bab el Mandab Strait (Figure 1) is complicated by active tectonics (Eagles et al., 2002) and volcanism (Gass et al., 1973), it is difficult to assess the morphological connection with the Gulf of Aden during the Miocene, so the possibility that the solutes were supplied from the Gulf of Aden cannot be ruled out from the paleogeography.

1.2.3  |  The present geometry of the evaporites and flowage

The S-reflection appears prominently in most seismic data collected in the Red Sea and was recognized in early reconnaissance surveys (Knott et al., 1966; Phillips & Ross, 1970; Ross & Schlee, 1973; Uchupi & Ross, 1986). Where it was sampled by drilling at DSDP Sites 225 and 228, it was found to correspond with the top of anhydrite beds marking the termination of the Miocene evaporites,
whereas at Site 227 it was suggested potentially to correspond with hard claystones lying up to 32 m above the evaporites, slightly within the Early Pliocene (Whitmarsh et al., 1974). The S-reflection therefore appears to represent an abrupt seismic impedance contrast at the Miocene-Pliocene boundary or very near to it.

In the map of the S-reflection surface topography (Figure 6), near to the coasts, where the evaporites are loaded by denser terrigenous sediment, the reflection is commonly strongly depressed. As a result of that loading, diapirs are common in seismic data near to many of the coasts of the Red Sea (Bosworth & Burke, 2005; Colombo et al., 2014; Davison et al., 1996; Gordon et al., 2010; Heaton et al., 1995; Ligi et al., 2019; Miller & Barakat, 1988; Mitchell et al., 2019; Mougenot & Al-Shakhis, 1999; Richter et al., 1991; Rowan, 2014; Savoyat et al., 1989). However, the Plio-Pleistocene sediments overlying the evaporites in deep water away from the coasts in the central Red Sea are typically hemipelagic. Measured densities of samples recovered at the DSDP sites are similar to or commonly less than that of halite (Figure 8b). Consequently, the surface of the evaporites shows more modest effects of diapirism away from the coastal areas compared with the coastal evaporites (Figure 3a).

Away from coasts and away from the axial deeps (closed-contour bathymetric depressions), the S-reflection surface undulates with a depth shallower than 1,000 m, though with broad depressions deeper than 1,000 m. Some of these depressions lie either side of the inter-trough zones (ITZs), which are shallower sills lying between the deeps. For example, depressions can be seen to either side of the Hatiba-Atlantis II ITZ, with its easterly depression extending nearly as far as the coastline. The Erba-Port Sudan ITZ is flanked by a depression immediately to its NE. These depressions have been proposed to be caused by evaporite flowage into the ITZs based on multibeam sonar data revealing giant evaporite flows with morphologies similar to ice glaciers (Augustin et al., 2014; Mitchell et al., 2017). Examples can be seen in Figure 9, e.g. at the Erba-Port Sudan ITZ flows occur from both sides and curved fabrics immediately north of the easterly flow where it has been obstructed by elevated basement. As ITZs are absent south of 20°N, the evaporites have not
flowed so extensively towards the axis and hence the evaporite surface remains at higher elevations to either side of the axis. Furthermore, loading by terrigenous sediments at the Tokar Delta has likely displaced or ‘shunted’ the evaporites, contributing to the higher elevation of the evaporites seawards of the delta.

Although diapirism is subdued, the seismic data in Figure 3a reveal some pillow-like features in reflections within the evaporites. The high-resolution image (Figure 3b) reveals an evaporite anticline flanked by growth stratigraphy, suggesting that vertical differential movements occurred while the layered evaporites were being deposited. Some growth stratigraphy and disrupted reflections also above the S-reflection suggest that vertical differential movements continued after the Miocene as well.

Also shown in Figure 3a,b, some reflections within the upper evaporites terminate at the S-reflection to form angular unconformities. Unconformities at the S-reflection can be observed in other datasets as well (Bonatti et al., 1984; Cochran et al., 1991; Ehrhardt & Hübscher, 2015; Feldens et al., 2016; Guennoc et al., 1988; Mitchell et al., 2017; Ross & Schlee, 1973), are reported in stratigraphic summaries of the region (Bunter & Abdel Magid, 1989; Hughes & Beydoun, 1992) and have been described in Saudi Aramco seismic reflection data (Colombo et al., 2014). They have been suggested to be associated with the reflooding of the Red Sea at the end of the Miocene (Izzeldin, 1987), wave erosion during declining or rising sea level (Mitchell et al., 2017) or regional exposure of the evaporites to rainwater (Colombo et al., 2014). Stoffers and Kühn (1974) interpreted some sections of anhydrite at DSDP Site 228 as indicating reworked particles. Abundant breccias and lack of halite at that site may also be due to dissolution of its halite components, such as caused by rainwater.

1.2.4 | Nature of crust

The central Red Sea axial deeps contain rocks of mid-ocean ridge basalt geochemistry (Altherr et al., 1988; Bonatti et al., 1984; Haase et al., 2000; Volker et al., 1993; van der Zwan et al., 2015), volcanic geomorphology in sonar data (Augustin et al., 2014, 2016; Ligi et al., 2012), identifiable seafloor-spreading magnetic anomalies (Chu & Gordon, 1988; Izzeldin, 1987; Ligi et al., 2012; Miller et al., 1985) and a magma chamber in seismic reflection data (Ligi et al., 2018), so there is little doubt that oceanic crust exists immediately under them. Away from the deeps in the inter-trough zones and areas towards the coasts, magnetic anomalies have low amplitudes and are difficult to associate with seafloor...
spreading chron. This has led to suggestions that these areas may contain stretched continental crust rather than oceanic crust (Bonatti, 1985; Cochran, 1983; Ligi et al., 2011, 2012).

However, the low magnetic anomaly amplitudes likely arise because of the ultra-slow spreading rates and because the basement is deep (Dyment et al., 2013), having been depressed by isostatic loading by the thick evaporites. The aeromagnetic anomalies in Figure 5a reveal a number of anomalies that are parallel to the axis and symmetrical about it (Izzeldin, 1987; Rasul et al., 2015) (such as those marked with “*” symbols). Seismic refraction experiments using ocean-bottom seismometer arrays off the central Red Sea African coast (Egloff et al., 1991) have revealed crustal velocity structures and thicknesses typical of oceanic crust within 70 km of the axis. Earlier seismic refraction experiments using sonobuoys and explosives were more
shallowly penetrating but also revealed high velocities typical of gabbro (Davies & Tramontini, 1970; Drake & Girdler, 1964; Tramontini & Davies, 1969). More recently, Shi et al. (2018) derived the basement structure under the evaporites from the deep seismic data of Izzeldin (1982) presented here, showing that, within ~80 km of the axis, it is typical ridge of a slow-spreading oceanic ridge near to a hotspot and very unlike continental rifts. Its surface has a rugged morphology typical of an ultra-slow ridge (Shi, 2019). The free air gravity field derived from satellite altimetry reveals a series of lineaments crossing the ridge parallel to the Nubia-Arabia plate motion, which are typical of fracture zones or other discontinuities of slow-spreading ridges (Mitchell, 2015; Mitchell & Park, 2014). Exhumed mantle as found at the Iberia margin (Whitmarsh et al., 2001) has been ruled out as basement velocities are not as varied as would be expected from varied hydration of mantle rocks. The weight of evidence therefore now strongly supports the existence of oceanic crust underlying the central Red Sea within 80 km of its axis.

Although the crust is oceanic under the axial ridge, other evidence suggests that thinned continental crust occurs towards the coasts. In the deep-seismic refraction experiment of Egloff et al. (1991) off the Sudanese coast, one line near to the coast revealed a crust 12–15 km thick with continental-like velocities. Another line perpendicular to the ridge revealed continental crust after 70 km from the spreading axis and a transition zone 10–20 km wide. Shi et al. (2018) found that Bougner gravity anomalies are strongly correlated with basement depths from deep-seismic data, as expected for oceanic crust, but that correlation breaks down towards the coasts. While the basement rises towards the coasts beyond ~80 km from the axis (Figure 4), that rise is not accompanied by much change in the anomalies (Figure 5b), suggesting the basement comprises lower density material or thickens landward. The deepest basement along the profiles in Figure 4 essentially represents the final stage of continental rifting and start of seafloor spreading. It coincides with the symmetrical anomalies marked with asterisks (‘*’ symbols) in Figure 5a. We assign this to Chron 5 based on its distance from the axis and the spreading rates of Chu and Gordon (1998). Although axis-parallel magnetic anomalies exist around the coast on the Arabian side (Hall, 1979), they are interpreted as being caused by dyke swarms within continental crust (Stockli & Bosworth, 2019), because seismological estimates of Moho depth along the Arabian coast of the central Red Sea are >20 km (Al-Damegh et al., 2005; Hansen et al., 2007; Sandvol et al., 1998) and mid-crustal industry seismic reflection data have been interpreted as showing continental crust (Colombo et al., 2014). The central Red Sea margins are essentially regarded as arising from amagmatic rifting (Bosworth & Stockli, 2016). Our interpretation of continental crust near the coasts is somewhat different from the original idea that oceanic crust floored the whole Red Sea based on the strong similarity of opposing coastlines (McKenzie et al., 1970), though is similar to interpretations of Izzeldin (1987) and Rihm and Henke (1998). The volume of this extended continental crust must be very limited, however, as structures in the shield rocks on either side of the Red Sea can be matched by closing the two coasts with only limited intervening stretched continental crust (Kozdroj et al., 2012; Sultan et al., 1993).

Searle and Ross (1975) suggested that, if a mid-ocean ridge underlies the central Red Sea, the elevated topography of the ITZs could arise from flowage of evaporites into fracture zone valleys, which are typically found at slow-spreading mid-ocean ridges. This indeed appears to be the case, as basement valleys crossing the axial high have been found to underlie the ITZs (Izzeldin, 1989; Mitchell et al., 2017). The 3D view of the basement under

---

**FIGURE 10** Schematic representation of the inferred central Red Sea structure at the end of the Miocene. Before later subsidence, flowage and dissolution, the evaporites were deposited to a high elevation relative to the spreading ridge, so the spreading centres were likely more greatly overlain by evaporites. Extension at those spreading centres (double-headed arrow (1)) and rotation caused by differential subsidence (curved arrow (2)) tilted the evaporites seawards and caused flowage seawards.
FIGURE 11   Top of the evaporites (S-reflection) of Figure 4 shown enlarged (green lines) along with that surface corrected for thermal subsidence (blue) and further corrected for isostatic loading by the Plio-Pleistocene sediments and water (black). Horizontal axis shows present-day distance from the spreading axis. Graph to right shows cross-sectional areas of evaporites that have overflown 5.3 Ma seafloor (x-symbols) and areas above the subsidence-corrected evaporites (circles). Red symbols represent data from west of the axis, green from east of the axis.
the central Red Sea in Figure 7 clearly shows this. The geomorphological structure of the central Red Sea axis, comprising deeps of >2,000 m depth separated by shallower inter-trough zones, is therefore the inverse of the basement structure, whereby basement highs underlie the deeps and basement lows underlie the ITZs.

Our working conceptual model for the central Red Sea at the Miocene-Pliocene boundary is as shown in Figure 10, with oceanic crust underlying its central ridge and thinned continental crust towards its coasts. Because the rate of thermal diffusion through oceanic lithosphere is greatest where the lithosphere is thinnest, its subsidence rate is greatest near the axis and decreases away from the axis towards the coasts. This causes a continual tilting towards the axis of any units emplaced above it, represented by the rotation (2) in Figure 10. That tilting and extension at the axis (1) has caused the flowage of evaporites continually towards the deeps, though generally at rates that are slow and comparable with plate-tectonic rates, as flowage has been unable to flood the floors of the deeps except at the ITZs. Slow rates may ultimately be due to the strengthening effects on the evaporite body of the laminated upper evaporites, which contain lithologies that are more rigid than pure halite.

### 1.2.5 Lithosphere dynamic topography

Dynamic topography of Earth's lithosphere arises from stresses originating in Earth's mantle. The high topography of the Africa-Arabia region has been suggested to have originated from such deep influences (Burke, 1996). Extensive volcanism in the western half of Arabia (Bosworth, 2015) occurs over a mantle regime with anomalously slow seismic velocities (Ritsma & van Heijst, 2000; Park et al., 2007; 2008; Priestly et al., 2008; Sicilia et al., 2008). For example, Chang et al. (2011) show shear waves up to 300 m/s slower than average mantle at 150 km depth. Polarization directions from shear waves suggest that olivine crystals in the upper mantle are oriented north-south beneath Arabia (Wolfe et al., 1999; Hansen et al., 2006), compatible with northward mantle flow away from the Afar hotspot. In some seismological studies extending to the central Red Sea, the effect of the hot mantle appears more subdued than under Arabia. For example, shear wave velocity perturbations reach a more modest 100 m/s (Chang et al., 2011) and shear-wave splitting falls below 1s below the central Red Sea, compared with >1.5 s beneath Arabia (Qaysi et al., 2018).

Whether dynamic topography is important for our reconstruction of the evaporite surface level in the central Red Sea depends on whether dynamic topography has changed over the past 5.3 m.y. Relevant to this, Wilson et al. (2014) evaluated the uplift history of Arabia from stream longitudinal profiles. Those profiles are convex-upwards, as expected from geologically more recent uplift (Whipple, 2001). Their analysis suggests that uplift rates have generally declined from 0.05 to 0.01 mm y⁻¹ over the past 10 m.y., implying ~160 m of uplift since 5.3 Ma.

The results of Wilson et al. (2014) require verification as their method relies on precipitation having remained constant and because the possibility that the Red Sea level drew down during the Miocene cannot be excluded. Many uplifted marine terraces have been found around the Red Sea coasts and could be useful for this. We excluded the youngest Holocene terraces because their interpretation is complicated by hydro-isostatic adjustments (Lambeck et al., 2011). Interpreting terraces associated with Marine Isotope Stage (MIS) 5e is also not so straightforward as their tops typically lie at only 3–6 m above sea level (asl) on both sides of the Red Sea. Uranium-series dating of this feature along the coasts of both Arabia (Bantan et al., 2015; Dawood et al., 2013; Dullo, 1990) and Africa (Arvidson et al., 1994; Choukri et al., 2007; El-Asmar, 1997; Gvirtzman & Friedman, 1977; Hoang et al., 1996; Hoang & Taviani, 1991; El Moursi et al., 1994; Plaziat et al., 2008) confirms that its upper surface was likely eroded during MIS 5e. The 3–6 m altitude of this terrace is similar to the altitude of sea level during MIS 5e estimated from the Huon Peninsula terraces (Lambeck & Chappell, 2001). This similarity has suggested to some researchers that the tectonic uplift of the coasts has recently slowed (Arvidson et al., 1994). Alternatively, uplift has continued and this similarity of elevation could partly reflect uncertainty in the MIS 5e eustatic level.

The older terraces are potentially more useful for resolving dynamic uplift but have not been so extensively dated. Taviani (1998) outlined problems of dating Pliocene reefs and suggested that many that have been attributed to the Pliocene may actually be early Pleistocene. The following refers to the numbered sites marked with red stars in Figure 1 and elevations are relative to modern sea level. We avoid the most tectonically active coasts, such as Gulf of Aqaba, Sinai and Gulf of Suez, which have terrace heights that differ from the rest of the Red Sea (Plaziat et al., 1998). Coral surfaces dated by uranium-series to MIS 7 have been reported at sites 1 (4 m (Dawood et al., 2013)), 2 (8.5 m (Hoang et al., 1996)), 3 (17 m (Hoang & Taviani, 1991)) and 4 (8 m (Arvidson et al., 1994)). A reef surface at 10 m at site 5 was dated using 18O measurements by Hamed et al. (2015) as MIS 9 and they have suggested by analogy that site 2 may also be MIS 9, as also suggested by Plaziat et al. (2008). If all the sites are assigned to MIS 7 and we use the median estimate of eustatic sea level (~15 m) of Rohling et al. (2014) and another estimate (~5.8 m) based on deep-water ocean 18O values (Miller et al., 2011), we can predict uplift rates and use them to estimate elevation...
change since 5.3 Ma assuming uplift rates have been constant. The results suggest the coasts have uplifted by 244–663 m for sites 1, 2, and 5, and up to 848 m for site 3 (this site lies on Zabargadh Island, which may still be tectonically active). If instead we assign all these terraces to MIS 9 (with highstand estimates of +5 m and +2.67 m), we estimate more modest uplift since the Miocene of −16 to +117 m (+193 m for Zabargad Island).

The maximum elevations of coral terraces that are not radiometrically dated but known to be Pleistocene provide further constraints on vertical motions. Based on extensive field observations particularly on the Egyptian Red Sea coast, Pleistocene coral reefs extend to 50 m elevation (Mansour & Madkour, 2015; Plaziat et al., 1998, 2008). The literature reveals elevations for the following sites also located in Figure 1:6, 7 and 8 (30, 16 and m (Dullo, 1990)), 9 (70–130 m (Dadet et al., 1970)), 10 (50 m (Jado & Zötl, 1984)), 11, 12 and 13 (all 32 m (El Moursi et al., 1994)), 14 and 15 (9 and 7 m (Roobol & Kadi, 2008)). Platform carbonates of Miocene age were mapped along the coast of Sudan north of Port Sudan by Schroeder et al. (1998). Although they didn’t give their elevations above sea level, their outcrops clearly exceed 100 m in elevation. Writing about similar reefs along the Egyptian coast, Perrin et al. (1998) suggested they were likely emplaced in the Middle Miocene before the hypersalinity associated with the main evaporite phase.

Some of the reason for such a large diversity of reef elevations for any given MIS could be due to continued movement on older synrift faults (Roobol & Kadi, 2008) or movements associated with weak evaporites underlying some of the coastal plains (Purkis et al., 2012; Purser & Hötzl, 1988). Given that the older uranium-series dates are potentially strongly affected by diagenesis by younger seawater (Plaziat et al., 2008), the claim of terraces having been mis-assigned to MIS 7 (Hamed et al., 2015) seems plausible. Putting more weight on the data with longer timescales, even where poorly dated, suggests to us that the amount of uplift attributable to broadly distributed dynamic topography since 5.3 Ma has been of order 100 m and no smaller than 50 m.

2 Observations and adjustments to the Miocene evaporite surface elevation

The seismic and multibeam data reveal that the evaporites have overflown the 5.3 Ma crustal boundary by varied amounts, tapering out towards the spreading axis (Figure 4). The top of the overflown evaporites is typically convex-upwards, whereas their base is more irregular, due to the relief of basement. The cross-sectional areas of the overflown evaporites vary from 19 km² for west line 17 to 65 km² for east line 7 (Figure 11). Although these areas could be affected by misplacement of the 5.3 Ma crustal boundary, the sum of east and west areas also varies about their average value somewhat less, by 43%. The average overflown area is 39.9 km². Varied overflow distances probably reflect varied influence of the more rigid anhydrite-dominated layered part of the evaporites, underlying basement morphology (Mitchell et al., 2010) and movements in 3D (Augustin et al., 2014, 2016; Mitchell & Augustin, 2017).

The top of the Miocene evaporites (S-reflection) tends to deepen away from the coast and towards the axes of the spreading centres (towards 0 km in Figure 4), although individual seismic lines (e.g. 11) can reveal a shoreward-deepening near the coasts where the evaporites are locally loaded by Plio-Pleistocene terrigenous sediment denser than hemipelagic sediment (also see Figure 6). The average depth of this surface on the longer lines (excluding line 27 and east line 29) was 908 m. We have applied the following series of corrections to this 908-m depth surface in order to reconstruct the level of the top of the Miocene evaporites at the Miocene-Pliocene boundary when most evaporite deposition ended.

2.1 Isostatic corrections

As two thirds of the profiles landward of the 5.3 Ma crustal boundary cross oceanic crust as explained earlier, we have used a thermal subsidence correction appropriate for oceanic lithosphere. This employed the half-space cooling model in which oceanic lithosphere subsides linearly with the square root of age. We have used a constant of proportionality (subsidence rate) of 342 m m.y.−0.5. This rate corresponds with the 370 m m.y.−0.5 expected from Crosby and McKenzie (2009) for a ridge axial depth of 2,200 m (the mean axial depth derived from the seismic data here), but includes an adjustment to allow for thermal blanketing by the overlying sediments. In the half space model, the contraction rate is proportional to the temperature difference across the lithosphere (Turcotte & Schubert, 1982). As the thermal modelling of Martinez and Cochran (1989) suggested the temperature of the upper surface of the crust in the northern Red Sea is ~100°C and somewhat similar temperatures were predicted by modelling a line of thermal conductivity measurements in the central Red Sea by Makris et al. (1991), the subsidence rate was multiplied by (1,300–100)/1,300, i.e., assuming a 1,300°C asthenosphere and 100°C upper boundary.

Thermal subsidence was calculated assuming all lithosphere is oceanic with ages derived using across-axis distances and rates from the Chu and Gordon (1998) pole and corrected the evaporite surface by 356 m to −552 m (Figure 2). As oceanic crust of Chron 5 and younger is present towards the axis, the rift margins of the central Red Sea are now in a post-rift thermal subsidence stage, so subsidence irregularities associated with syn-rift phases (e.g. Karner & Gamboa, 2007) do not apply here. A further indirect verification is provided by the coral terraces described in the background, which show that the central Red Sea coasts have been uplifting slowly over the
late Pleistocene as expected from changing broad-based dynamic topography, in contrast with the still active Gulf of Suez (Bosworth & Taviani, 1996). As stretched continental crust has finite lithospheric thickness at the onset of thermal subsidence (McKenzie, 1978), the thermal subsidence correction will likely be too large towards the coast and hence the 356-m average correction (Figure 2) is a maximum. For comparison, backstriping of data from a commercial well on the Egyptian Red Sea coast by Richardson and Arthur (1988) found only ~100 m tectonic subsidence there since 15.5 Ma. If there has been ~100 m broad-based dynamic uplift since the Miocene as explained earlier, this implies ~130 m of thermal subsidence over the Plio-Pleistocene.

Further isostatic corrections were made for loading by the Plio-Pleistocene sediments and water using the seismic reflections after the thermal subsidence corrections. These used densities of 2.0 g/cm$^3$ for the sediments (Whitmarsh et al., 1974) and 1.0 g/cm$^3$ for seawater, while 3.2 g/cm$^3$ was used for the mantle density, i.e., somewhat warm mantle beneath the axis (Makris et al., 1991). The results of both thermal and loading correction are shown in Figure 11. After carrying out these corrections, the corrected Miocene surface dips upwards towards the axis. As no corrections have been made for evaporite lateral movements and the original evaporite surface is more likely to have been flat (Gargani et al., 2008), this reveals that there is an excess mass of evaporite at ~40 to 60 km from the axis compared with areas further from the axis. That mass cannot have come from seaward of the 5.3 Ma crustal boundary (as that crust did not exist before 5.3 Ma, the maximum age of the evaporites), so the evaporites must have moved towards the younger oceanic lithosphere caused by greater subsidence rates there (Figure 10). The average elevation of this corrected surface for the longer lines (7, 9, 11, 15, 17, 19, 21, 25 and west line 29) is ~295 m (Figure 2).

The above isostatic calculations were only one-dimensional; no allowance has been made for flexural effects arising from lithospheric rigidity. This is because young, slow-spread oceanic lithosphere has an effective elastic thickness ($T_e$) of <13 km and typically $T_e < 5$ km (Cochran, 1979), whereas active continental rifts typically also have low $T_e$ of 5–15 km (Watts & Burov, 2003). Furthermore, we are ultimately seeking to reconstruct the spatially averaged elevations across the flanks, so any flexural effects have diminished impact on our final mean values.

### 2.2 Deflation of the evaporite surface caused by lateral movements

We mapped the boundary between evaporites and volcanic geomorphology in the multibeam data of Augustin et al. (2014), Augustin et al. (2016) and Augustin et al. (2019) (Figure 9). Using also the basement reflection and the reflection from the top of the evaporites (i.e. the S-reflection), we resolved the parts of the evaporites that had overflown the 5.3 Ma crustal boundary. Those regions are shown in grey in Figure 4. We calculated their cross-sectional areas (in the planes of these sections) and show their values by cross symbols in the right-hand graphs in Figure 11. For comparison, the cross-sectional areas above the subsidence and loading-corrected evaporite surfaces (black lines) to modern sea level in the left graphs of Figure 11 are also shown on the right-hand graph in Figure 11 (circles).

We then explored how the flowage of the evaporites beyond the 5.3 Ma crustal boundary has deflated the average level of the evaporites landward of the 5.3 Ma crustal boundary. The overflown evaporite cross-sectional area was divided by the 68.2 km average profile length of evaporite surface landward of the 5.3 Ma crustal boundary (using the longer lines 7, 9, 11, 15, 17, 19, 21, 25 and west line 29) to obtain the equivalent thickness $h$. As our earlier calculations already removed the isostatic effect of water loading, the isostatic movement $I$ due to the additional thickness $h$ (effectively restoring that mass landward of the 5.3 Ma boundary) was:

$$I = h(r_e/r_m)$$  \hspace{1cm} (1)

where $\rho_e$ and $\rho_m$ are the evaporite and mantle densities. The elevation of the evaporites with this additional mass was then:

$$Z_e = Z_0 + (h - I) = Z_0 + h(1 - r_e/r_m).$$  \hspace{1cm} (2)

Solving Equation (2) with $Z_0 = -295$ m and $h = 585$ m ($=39.9$ km$^2$ overflown area/68.2 km profile length) led to a corrected $Z_e = -75.6$ m (Figure 2).

### 2.3 Losses of evaporites by dissolution and diffusion

Diffusive loss of halite via the pore waters in the Plio-Pleistocene sediments was considered by Ranganathan (1991). Pore-water NaCl concentrations increase nearly linearly with depth for the 100 m above the evaporites at DSDP Sites 225 and 227 (Manheim et al., 1974), as would be expected for steady-state diffusion within constant permeability sediments. However, concentrations vary less with depth nearer to the seabed, which Ranganathan (1991) interpreted as a sign of compaction affecting porosity and hence permeability. In particular, a large pore water advection velocity is ruled out as the pore water salinity profiles are concave with depth, not convex with depth as would be expected with upwards advection. Furthermore, seismic reflection data show coherent folded structures within the upper evaporites (Mitchell et al., 2017, 2019) (Figure 3b), not breccias or other
loss from them will have been 0.3 km² (i.e. we assume that this loss on average over the Plio-Pleistocene, so the diffusive use half of the average 24.0 km length that has undergone the evaporites that have overflown the 5.3 Ma boundary, this implies a total loss of 1.7 km². For the evaporites that have overflown the 5.3 Ma boundary, we use half of the average 24.0 km length that has undergone this loss on average over the Plio-Pleistocene, so the diffusive loss from them will have been 0.3 km² (i.e. we assume that the evaporites have overflown at constant speed so half the length would represent the time-averaged evaporite extent). The combined area loss was therefore estimated to have been 2.0 km².

Sodium and chloride ion concentrations in the brine lakes of the Red Sea deeps are strongly correlated and follow a mixing line with Red Sea Deep Water as one end member, which suggests that most of the brine originates from dissolution of the evaporites (Schmidt et al., 2015). The precise manner in which the halite is dissolved is uncertain, although the Plio-Pleistocene sediments overlying the evaporites thin towards the floors of the deeps (Mitchell et al., 2010, 2017) and multibeam sonar data (Augustin et al., 2014, 2016, 2019) suggest large strains during evaporite flowage into the deeps so exposure through tectonics or slumping is likely. The brines vary in temperature and depths of their surfaces, so they are not strictly in a steady state, but the rate of loss of brine by diffusion into the overlying water might be useful as a rough estimate of this flux. Anschutz et al. (1999) estimated the loss of NaCl by diffusion for Suakin, Port Sudan, Nereus and Chain B deeps. Assuming the latter is representative of the Red Sea deeps are strongly correlated and follow a mixing line with Red Sea Deep Water as one end member, which suggests that most of the brine originates from dissolution of the evaporites (Schmidt et al., 2015). The precise manner in which the halite is dissolved is uncertain, although the Plio-Pleistocene sediments overlying the evaporites thin towards the floors of the deeps (Mitchell et al., 2010, 2017) and multibeam sonar data (Augustin et al., 2014, 2016, 2019) suggest large strains during evaporite flowage into the deeps so exposure through tectonics or slumping is likely. The brines vary in temperature and depths of their surfaces, so they are not strictly in a steady state, but the rate of loss of brine by diffusion into the overlying water might be useful as a rough estimate of this flux. Anschutz et al. (1999) estimated the loss of NaCl by diffusion for Suakin, Port Sudan, Nereus and Chain B deeps. Assuming the latter is representative of the Plio-Pleistocene surface represented by the S-reflection lay at Z_e=−132 m at 5.3 Ma. As we had corrected for water loading, this represents the level if the sea had completely drawn down (air-loaded). If the basin were water-filled to modern sea level, the Miocene-Plioene surface would have instead lain at Z_e=−192 m.

### 2.4 Reversing the change in lithospheric dynamic topography

We outlined earlier the evidence for the amount of broad-based lithospheric uplift originating from buoyant mantle. Subtracting the best estimate of dynamic topographic change (100 m) from the result of the earlier calculations (Figure 2) suggests that the Miocene-Pliocene surface represented by the S-reflection lay at Z_e=−132 m at 5.3 Ma. As we had corrected for water loading, this represents the level if the sea had completely drawn down (air-loaded). If the basin were water-filled to modern sea level, the Miocene-Pliocene surface would have instead lain at Z_e=−192 m.

### 3 Uncertainty estimates

Quantifying the uncertainty in Z_e is difficult given the nature of the data and reconstructions used. We have estimated the uncertainties in individual components of the reconstruction and then combined them allowing for whether they are systematic or random (Taylor, 1982). We follow the sequence of corrections outlined in our methods section. Random errors in digitising the S-reflection are ignored due to the large number of points digitized over the 11 lines ~3,280 points. Although we do not know the random errors of individual picks of the S-reflection, the standard error of their mean elevation will equal the standard deviation of those random errors divided by a factor of 57 (=√3280) and hence is likely to be small. However, aspects of the seismic data processing could have affected the wavelet shape and delay, and are unknown, hence introducing non-random uncertainties. There could also be biases in our horizon picks. To assess this, we sampled the bathymetry in Figure 1 along the seismic lines to test the seabed reflection, which has a similar character to the S-reflection. We found the seismic-derived depths are deeper by only 41 m on average. This is an imperfect assessment as some of the bathymetry in Figure 1 is interpolated, so our bias could be under-estimated. We therefore allow for a slightly larger ~50 m potential bias in the S-reflection depth (uncertainty 1).

Water column and sediment velocities are known well or their uncertainties have little impact. As oceanic subsidence...
rates are predictable once axial depth is accounted for (Crosby & McKenzie, 2009), we ascribe the main cause of subsidence uncertainty to our use of oceanic rates nearer to coasts where the crust is probably stretched continental crust (Egloff et al., 1991; Shi et al., 2019). The transition to stretched continental crust lies landward of an inflection in the basement profile and changed correlation between basement depths and Bouguer gravity anomalies, which was identified as the likely ocean-continent boundary (Shi et al., 2019). If subsidence had been only half the oceanic amount since the Miocene for the landward one third of those lines, $Z_e$ has been over-estimated by 50 m (uncertainty 2). We conservatively assume half the profile could be stretched continental crust, rather than the one-third mentioned earlier, to acknowledge potential uncertainty in the continent-ocean boundary.

As with the $S$-reflection digitising, random uncertainties have little effect on overflown evaporite areas due to the large amounts of data used (~1,000 points along each of the $S$-reflection and basement interfaces). However, evaporite velocity used to convert two-way times of basement reflections to depth could vary by up to 0.2 m/s comparing the different refraction results (Davies & Tramontini, 1970; Egloff et al., 1991; Tramontini & Davies, 1969), implying an uncertainty in overflown evaporite area of 1.9 km$^2$. This implies a 10-m uncertainty in $Z_e$ (uncertainty 3). A bias could arise from picking basement towards the upper envelope of diffraction hyperbolae (Mitchell, 1995). We estimate this bias could be 100 m, hence affecting average overflown area by 2.4 km$^2$, which implies a 13-m underestimation in $Z_e$ (uncertainty 4). Some parts of the basement are only interpolated, for which we assign a 1 km$^2$ overflown evaporite area uncertainty or 5 m uncertainty in $Z_e$ (uncertainty 5). The age of basement affects our estimated overflown area. From the diversity of model Nubia-Arabia rates in DeMets and Merkouriev (2016), we suggest a 5% error in the average cross-axis distance to the 5.3 Ma crustal boundary is possible, i.e., $\pm$2 km, implying $\pm$8 km$^2$ uncertainty in overflown area by 44 m uncertainty in $Z_e$ (uncertainty 6). For uncertainty in the loss of halite by pore water, we allow for the 10–40 m range of loss estimated by Ranganathan (1991), i.e., an uncertainty of 15 m. Allow for isostasy, this implies an uncertainty in $Z_e$ of 6.5 m (uncertainty 7). The uncertainty in dynamic topography is more difficult to estimate, though we suggest a range of 50–150 m about our 100 m best estimate is reasonable (50 m for uncertainty 8).

Deposition of terrigenous sediments along the coasts that are denser than halite was mentioned earlier as responsible for the depressions of the $S$-reflections in Figure 6 and abundant diapirism near the coasts. Such deposits create high lithostatic stresses that displace the evaporites seawards, a process we have called ‘shunting’. Shunting of the evaporites away from the coasts would have increased their volume in areas further towards the axis and hence makes our estimates of the end-Miocene evaporite level upper estimates. As deep seismic data are not freely available under the coastal sediments, we cannot easily quantify the shunting effect for all our lines, but the Tokar Delta is a particularly large terrigenous deposit formed by the only permanent river in the Sudanese coast, so we use it to estimate this extreme example. Bunter and Abdel Magid (1989) showed the Plio-Pleistocene reaching 1,800, 1,330 and $\geq$740 m in wells Durwara-2, Bashayer-1A and S. Suakin-1 located in Figure 1, but only 500 and 230 m thick at Abu-Shagara-1 and Maghersum-1 further north. Miocene evaporites lie beneath the Plio-Pleistocene sediments in all these wells. Assuming these sediments taper out linearly from an average 1,600 m at the Tokar Delta wells to the more normal $\sim$200 m thickness where shown by the reflections in Figure 4 for lines 25 and 29, the tapering distance is roughly 30 km. Hence, the along-profile cross-sectional area of Plio-Pleistocene sediments displacing evaporites and hence shunting is $\sim$30 km times an average 0.8 km thickness, or $\sim$24 km$^2$. Such an area distributed over the 68.2 km average length of our lines implies 352 m of elevation change. This is a large value, but as mentioned this is an extreme case, so the effect is likely much smaller than 352 m generally.

How poorly known this uncertainty is, we have not applied it below but nevertheless we suggest that it makes our final estimates elevation maxima.

Six of the uncertainties mentioned above are effectively random (uncertainties 1, 3, 5, 6, 7 and 8), whereas two are potentially systematic (uncertainties 2 and 4). We combine the random uncertainties first, assuming that they are uncorrelated, yielding a net uncertainty of 84 m. To estimate conservatively the upper bound on $Z_e$, we add this 84-m uncertainty to the 13-m uncertainty arising from diffraction hyperbolae effect and to the $\sim$132 m best estimate of $Z_e$. This yields an upper bound of $\sim$48 m. Similarly, for the lower bound, we add the 84-m random uncertainty to the 50-m uncertainty arising from potential wavelet offsets, and subtract that sum from $\sim$132 m. That yields a lower bound of $\sim$266 m. As before, $\sim$48 and $\sim$266 m represent water-free conditions. With water loading, these would instead be $\sim$70 and $\sim$387 m. These uncertainties are represented by the vertical bars in Figure 2.

4 | DISCUSSION
4.1 | Environment of deposition

Hughes and Beydoun (1992) interpreted Mid-Miocene faunas in central Red Sea commercial wells as ‘deep marine’. Our interpretation of Chron 5 ($\sim$10 Ma) in the magnetic anomalies suggests that oceanic crust extends to the Mid-Miocene and that the centre of the basin would have been deep at that time. It would have been closer to the $\sim$2,000 m depths of the present volcanic deeps than to the 192–132 m depths below modern sea level reconstructed for the end of
the Miocene (Figure 2). Hence, the emplacement of mainly massive halite in the Mid-Miocene occurred when the central Red Sea was deep, in keeping with the Hughes and Beydoun (1992) interpretation.

Towards the end of the Miocene, the basin gradually filled towards global sea level but not fully (Figure 2). Hughes and Beydoun (1992) reported barren to locally poor microfauna on both margins of the late Miocene Red Sea in commercial wells. They inferred a shallow marine environment or even sabkha conditions towards the end of the Miocene. Stoffers and Kühn (1974) interpreted petrographic and geochemical data of the evaporites recovered at the DSDP sites as indicating a shallow water environment and occasional subaerial exposure. Orszag-Sperber et al. (1998) summarising the literature available concluded that a shallow water depositional environment could explain the layered evaporites of the Upper Miocene Zeit Formation. A. Caruso (Università degli studi di Palermo, Italy, pers. comm. 2020) has also suggested that gypsum indicated in the DSDP cores by polymorphs (Stoffers & Kühn, 1974) was likely deposited in a very shallow sea.

Warren (2010) has suggested that the globally occurring ‘salt giants’ were most likely supplied by sea water by porous flow through a permeable barrier rather than a continuously open channel. Deposition of halite-rich evaporites can occur rapidly, e.g., over 400–600 ky in the South Atlantic (Tedesci et al., 2017) and 300 ky in the Mediterranean (Krijgsman et al., 1999), which provokes the question of why they stop accumulating. Although we cannot know much about the nature of the barriers modulating the flow of external seawater into the central Red Sea basin, our reconstructed evaporite elevations suggest that the hydraulic gradient along the basin from its barriers would have been small at the end of the Miocene. For example, if evaporites in the central Red Sea lay at 100–300 m depth relative to global sea level (values encompassing the 132 and 192 m depth, but allowing another 100-m correction for shunting) and the barrier to external inflow lay at around the either modern Bab Al Mandeb or Suez (~1,000 km distant, Figure 1), gradients were ~0.0001 to 0.0003 m/m. If the basin was water-filled in the late Miocene, it may have been stratified by water densified by evaporation. However, it may also have been susceptible to over-turning by winds; wind-driven gyres in the modern Red Sea reach ~200 m depth (Sofianos & Johns, 2007).

Perhaps the end of the evaporite phase was caused by a change in climate? Based on high deposition rates of sediments in the Gulf of Suez and Red Sea during the Messinian, Griffin (1999) suggested a wetter climate prevailed at that time. Pollen data from a well in the submarine Nile Delta show no change towards the latest Miocene (Fauquette et al., 2006) and Poud et al. (2012) interpreted paleobotanical data of the Messinian from two sites in northern Egypt as indicating ‘tropical xerophytic shrubland’ and ‘tropical grassland’ environments.

We suggest that the shallow gradients of the basin towards the end of the Miocene led to the replenishment by global seawater becoming more easily reduced (e.g. by volcanic or tectonic changes to barriers or channels) and may have helped to prevent the basin completely filling. Anhydrite forms roughly one third of the sections sampled at the DSDP sites (Figure 8a), whereas complete evaporation of seawater should lead to the deposition of 90% halite (Warren, 2010). Low halite proportions could arise if non-halite solutes were supplied by runoff from the adjacent continents rather than from global seawater. Water supplied to balance evaporation may therefore have been dominated by rain runoff from land rather than Mediterranean or Indian Ocean water towards the end of the Miocene. Stoffers and Kühn (1974) noted that the layered evaporites cored at DSDP Sites 225, 227 and 228 do not follow a classical lithological sequence expected from progressive evaporation of seawater, but instead appeared random. Such an outcome is explained by varied rainwater input (Gargani et al., 2008) along with marine water input likely fluctuating because of the ~30 to 40 m eustatic sea-level fluctuations (Figure 12) varying flow through a permeable barrier. If drawdown occurred at any point during the Mid-Miocene, the deep basin would have had >10°C higher temperatures because of adiabatic effects exacerbating evaporation and dryness (Hudec et al., 2013), and isostatic uplift could have uplifted the barrier further isolating the basin (Govers, 2009). However, such effects would have been much less dramatic if drawdown occurred towards the end of the Miocene. At that time, the ~60 m isostatic effect of desiccation (the difference between (i) and (ii) in Figure 2) would have had a modest effect of further isolating the basin. The combination of isolation and rainwater input may explain why the evaporites did not fill the basin entirely.

4.2 Origin of the unconformity at the S-reflection

Given that the reconstructed evaporite levels lie below the lowstands of eustatic sea-level in the Late Miocene (Figure 12), the unconformity was not created simply by a lowering in a Red Sea connected directly to the global oceans. However, two indirect causes are possible. First, if evaporation in the basin was balanced by inflow via the Mediterranean, the drawdown of the Mediterranean during the Messinian Salinity Crisis at 5.5–5.6 Ma (Krijgsman et al., 1999) would have shut off that supply and allowed the Red Sea level to draw down by evaporation, leading to wave erosion and rainwater corrosion creating the unconformity (Afifi et al., 2014). Alternatively, if the supply of seawater was via the Bab El Mandeb area through a permeable barrier, a lowstand of global sea-level may have restricted flow into the basin, also leading to drawdown of the
Red Sea, desiccation and erosion. Given the different dates in Figure 12 of the low-stands over the Messinian Salinity Crisis period of Miller et al. (2020) and the peak of that crisis in the Mediterranean at 5.4–5.5 Ma (Krijgsman et al., 1999), it might become possible in the future to discriminate these possibilities with accurate dating of sediments deposited immediately above the unconformity.

The eustatic sea-level curve of Miller et al. (2020) contains somewhat shallower low-stands in the Miocene prior to the Messinian Salinity Crisis (Figure 12). The high-resolution seismic reflection records in Figure 3b and others we have viewed show no prominent angular erosional unconformities within the uppermost evaporites, apart from the S-reflection. A similar impression is provided by other seismic reflection datasets from the central and northern Red Sea (Bonatti et al., 1984; Cochran et al., 1991; Cochran, 2005; Ehrhardt & Hübscher, 2015; Feldens et al., 2016; Izzeldin, 1987; Mitchell et al., 2017, electronic supplement) and interpretations of earlier paper records (Phillips & Ross, 1970; Ross & Schlee, 1973; Uchupi & Ross, 1986). This would seem to suggest the evaporites remained largely below wave base until the final erosional event that created the S-reflection unconformity.

4.3 The nature of the spreading centres at the Miocene-Pliocene boundary

The inter-trough zones in the present-day Red Sea are floored with evaporites (Bonatti et al., 1984), which have flowed in from the flanks of the ridge at points where the axial basement is presently deepest (Augustin et al., 2014, 2016; Mitchell et al., 2010, 2017; Searle & Ross, 1975) (Figure 7). From the basement relief in Figure 7 and the S-reflection data in Figure 6, we can evaluate the elevation difference between the evaporites away from the axis and the axial basement beneath the ITZs (effectively a measure of potential energy driving flowage). Within the three ITZs marked in Figure 7 from north to south, this relief is 3,000, 2,200 and 1,900 m. Hence, ~2,000 m or more elevation difference has been sufficient to cause the evaporites to flood the axis in these ITZs.

At the end of the Miocene, the parts of the spreading axis away from the fracture zones or other non-transform discontinuities were likely to have lain near to 2,000 m depth, as they do today within the deeps, while the basin was ~70 to 80 km narrower. Hence, if the evaporites lay near to modern sea-level, they would have had high elevations relative to the oceanic crust of the spreading axis. They are likely to have covered much more of the axis making the deeps more limited in size, if they had existed. If the spreading centres were buried under evaporites, this in turn implies that much of the axial volcanic activity would have occurred under sediments, as at other sedimented spreading ridges, potentially leading to emplacement of dykes and sills, rather than a layer of extrusives that more usually causes sea-floor spreading magnetic anomalies. This could further explain why the magnetic anomalies older than Chrons 1–3 have poorly formed oceanic characteristics (Dyment et al., 2013).

4.4 Application to the Mesozoic Atlantic and Gulf of Mexico margins?

The reconstruction of the evaporite elevation at the Miocene-Pliocene boundary in the Red Sea may be useful for...
considering the corrections needed to improve the accuracy of evaporite body reconstructions in other, older basins elsewhere (Garcia et al., 2012; Strozyk et al., 2017). Sample textures, mineralogy or other data may also suggest that those basins transitioned to shallower water conditions or subaerial exposure (Rodriguez et al., 2018), but without knowing the vertical position of the evaporites relative to global sea level we cannot know if those conditions arose because the basin filled completely or if instead barriers to inflow arose or became impermeable. The reconstruction corrections for subsidence and evaporite movements would be more uncertain due to poorer knowledge of the basement and long timescale involved. Whereas we have been able to use a model for pore-water diffusion loss of halite (Ranganathan, 1991), induration of more deeply buried and older strata overlying the Mesozoic marginal evaporites would render them more susceptible to fracturing, hence making long-term changes in permeability less easily predicted. Most attempts at evaporite reconstruction have ignored halite loss by dissolution, so this is a significant source of uncertainty (Rowan & Ratliff, 2012).

The Red Sea Miocene evaporite surface declined rapidly from 192–132 m to 908 m depth in only 5.3 m.y. by thermal subsidence (rapid due to the underlying mostly oceanic lithosphere), isostatic loading by seawater and later sediments, dissolution and spreading (flowage). This relatively quickly created a large water-filled depression (‘accommodation space’). As the Red Sea environment has remained dry and much of the drainage from the surrounding continents, particularly the Nile River, does not enter the Red Sea (Frostick & Reid, 1989), this depression has not been filled with sediments. Compared with other young basins, the subsidence here is enhanced by the evaporite spreading onto the new oceanic crust. This combination of effects may also have operated at some of the earlier evaporite-filled basins, leading to rapid development of accommodation space.

5 | CONCLUSIONS

After correcting the mean elevation of the evaporite surface as recorded by the seismic S-reflection for the effects of lithospheric thermal cooling, spreading onto crust younger than 5.3 Ma, dissolution and broad-based dynamic topographic change, we estimate that it lay at −132 m (air-loaded) or −192 m (fully water-loaded) at the Miocene-Pliocene boundary when most evaporite deposition stopped. We cannot easily also correct for shunting away from the coasts by tenuous sediment loading, though that may have elevated the evaporite surface by 100 m or more. Hence, −132 and −192 m are upper bounds on the elevation, so the basin ended the evaporite phase under-filled.

Several researchers have suggested that the uppermost evaporites at the DSDP sites and marginal commercial wells in the central Red Sea were deposited in shallow marine environments. A prominent angular unconformity is commonly found at the S-reflection, whereas high-resolution seismic reflection data typically show layered upper evaporites with no evidence of unconformities within them, suggesting that they were deposited below wave base. The shallow-marine conditions and unconformity suggest that the sea drew down fully towards the end of the Miocene. As the sea was shallow, any isostatic response to that drawdown would have been small (~60 m). The transition from massive halite deposition during the Mid-Miocene to layered evaporites in the Upper Miocene may have been a response to the gradual reduction in the hydraulic gradient driving seawater inflow associated with the shallowing and perhaps a change in climate.

ACKNOWLEDGEMENTS

Many of the figures shown in this article were created with the GMT free software system (Wessel & Smith, 1991). The Bouguer anomalies in Figure 5b were created from the free-air anomalies calculated by David Sandwell and co-workers and generously provided online. The authors thank Nico Augustin, Froukje van der Zwan, Martin Schade and others for their work in collecting the multibeam sonar data shown here. Author NCM thanks the Royal Society for funding his participation in that cruise (IES\R3\170081). He also acknowledges the NERC for earlier support in this research (NE/I52880X/1). The idea of a spreading ridge causing differential subsidence was originally suggested by Christian Hübscher. We thank Or Bialik for some very useful suggestions concerning the regional climate. For helpful comments that significantly improved this article, we also thank reviewers Bill Bosworth, Abdulkader Afifi and Marina Rabineau, and editor Craig Magee. The authors declare that they have no conflicts of interests in this work.

DATA AVAILABILITY STATEMENT

This study is based on the seismic data of Izzeldin (1982), which are not publicly available. However, the authors can respond to reasonable requests for a copy of the digitized horizons. The data in Figure 5a are not publicly available. Those in Figures 1, 5b, 7 and 9 were computed from publicly available data (www.ngdc.noaa.gov/mgg/, topex.ucsd.edu/index.html, doi.pangaea.de/10.1594/PANGAEA.912178, deepseadrilling.org/).

ORCID

Neil C. Mitchell https://orcid.org/0000-0002-6483-2450

REFERENCES

Afifi, A. M., Tapponnier, P., & Raterman, N. S. (2014). The Messinian Unconformity in the Red Sea: Evidence for Widespread Dessication? AAPG Search and Discovery Article #90188, 11th Middle East Geosciences Conference and Exhibition, 10-12 March 2014, Manama, Bahrain.
Al-Damegh, K., Sandvol, E., & Barazangi, M. (2005). Crustal structure of the Arabian plate: New constraints from the analysis of teleseismic receiver functions. *Earth and Planetary Science Letters*, 231, 177–196.

Altherr, R., Henjes-Kunst, F., Puchelt, H., & Baumann, A. (1988). Volcanic activity in the Red Sea Axial Trough - Evidence for a large mantle diapir? *Tectonophysics*, 150, 121–133.

Anschütz, P., Blanc, G., Chatin, F., Geiller, M., & Perrett, M.-C. (1999). Hydrographic changes during 20 years in the brine-filled basins of the Red Sea. *Deep-Sea Research Part I: Oceanographic Research Papers*, 46, 1779–1792.

Arvidson, R., Becker, R., Shanabrook, A., Luo, W., Sturchio, N., Sutlair, M., Lotfy, Z., Mahmood, A. M., & el Alfy, Z. (1994). Climatic, eustatic, and tectonic controls on quaternary deposits and landforms, Red Sea coast, Egypt. *Journal of Geophysical Research*, 99, 12175–12190.

Augustin, N., Devey, C. W., van der Zwan, F. M., Feldens, P., Tominaga, M., Bantan, R., & Kwasnitschka, T. (2014). The transition from rifting to spreading in the Red Sea. *Earth and Planetary Science Letters*, 395, 217–230.

Augustin, N., Mitchell, N. C., van der Zwan, F. M.; SHIPBOARD SCIENTIFIC PARTY. (2019). RV Pelagia Fahrbericht/Cruise Report 64pe-445: SALTAX: Geomorphology and Geophysics of Submarine Salt Flows in the Red Sea Rift, Limassol (Cyprus) – Safaga (Egypt). 27.08. – 21.09.2018. GEOMAR Report, N. Ser. 050, GEOMAR Helmholtz-Zentrum für Oceanforschung, Kiel, Germany, 46.

Augustin, N., van der Zwan, F. M., Devey, C. W., Ligi, M., Kwasnitschka, T., Feldens, P., Bantan, R., & Basaham, A. S. (2016). Geomorphology of the central Red Sea Rift: Determining spreading processes. *Geomorphology*, 274, 162–179. https://doi.org/10.1016/j.geomorph.2016.08.028

Bantan, R. A., Abu-Zied, R. H., & Haredy, R. A. (2015). Lithology, fauna and environmental conditions of the Late Pleistocene raised reefal limestone of the Jeddah coastal plain, Saudi Arabia. *Arabian Journal of Geosciences*, 8, 9887–9904.

Bonatti, E. (1985). Punctiform initiation of seafloor spreading in the Red Sea during transition from a continental to an oceanic rift. *Nature*, 316, 33–37. https://doi.org/10.1038/316033a0

Bonatti, E., Colantoni, P., & Delta Vedova, B. & Taviani, M. (1984). Geology of the red sea transitional zone (22°N-25°N). *Oceanologica Acta*, 7, 385–398.

Bosworth, W. (2015). Geological evolution of the Red Sea: Historical background, review and synthesis. In N. M. A. Rasul, & I. C. F. Stewart (Eds.), *The Red Sea: The formation, morphology, oceanography and environment of a young ocean basin*. Springer Earth System Sciences.

Bosworth, W., & Burke, K. (2005). Evolution of the Red Sea-Gulf of Aden rift system. In P. J. Post, N. C. Rosen, D. L. Olson, S. L. Palms, K. T. Lyons, & G. B. Newton (Eds.), *Petroleum systems of divergent continental margin basins*. 2005 Gulf Coast Section SEPM Foundation 25th Bob F. Perkins Annual Research Conference, Houston, 4–7 Dec, 2005 (pp. 342–372). Houston, Texas, USA.

Bosworth, W., Huchon, P., & McClay, K. (2005). The red sea and gulf of aden basins. *Journal of African Earth Sciences*, 43, 334–378.

Bosworth, W., & Stockli, D. F. (2016). Early magmatism in the greater red sea rift: Timing and significance. *Canadian Journal of Earth Sciences*, 53, 1158–1176.

Bosworth, W., & Taviani, M. (1996). Late Quaternary reorientation of stress field and extension direction in the southern Gulf of Suez, Egypt: Evidence from uplifted coral terraces, mesosopic fault arrays, and borehole breakouts. *Tectonics*, 15, 791–802. https://doi.org/10.1029/95TC03851

Boudreaux, J. E. (1974). Calcareous nannoplankton ranges, deep sea drilling project leg 23. In R. B. Whitmarsh, D. E. Wesser, D. A. Ross, et al. (Eds.), *Initial reports of the deep sea drilling project* (Vol. 23, pp. 1073–1090). US Govt. Printing Office.

Bunter, M. A. G., & Abdel Magid, A. E. M. (1989). The Sudanese Red Sea: 1. New developments in stratigraphy and petroleum-geological evolution. *Journal of Petroleum Geology*, 12, 145–166.

Burke, K. (1996). The African Plate. *South African Journal of Geology*, 56, 339–410.

Chang, S. J., Merino, M., van der Lee, S., Stein, S., & Stein, C. A. (2011). Mantle flow beneath Arabia offset from the opening Red Sea. *Geophysical Research Letters*, 38, article L04301, https://doi.org/10.1029/2010GL045852

Choukri, A., Hakam, O. K., Reyss, J. L., & Plazzi, J. C. (2007). Radiochemical data obtained by spectrometry on unrecrystallized fossil coral samples from the Egyptian shoreline of the north-western Red Sea. *Radiation Measurements*, 42, 271–280. https://doi.org/10.1016/j.radmeas.2006.12.005

Chu, D., & Gordon, R. G. (1998). Current plate motions across the Red Sea. *Geophysical Journal International*, 135, 313–328.

Cochran, J. R. (1979). An analysis of isostasy in the world’s oceans 2. Mid-ocean ridge crests. *Journal of Geophysical Research: Solid Earth*, 84, 4713–4729.

Cochran, J. R. (1983). A model for the development of the Red Sea. *American Association of Petroleum Geologists Bulletin*, 67, 41–69.

Cochran, J. R. (2005). Northern Red Sea: Nucleation of an oceanic spreading center within a continental rift. *Geochemistry, Geophysics, Geosystems*, 6, art. Q03006, https://doi.org/10.1029/2004GC000826

Cochran, J. R., Gaulier, J. M., & le Pichon, X. (1991). Crustal structure and the mechanism of extension in the northern Red Sea: Constraints from gravity anomalies. *Tectonics*, 10, 1018–1037. https://doi.org/10.1029/91TC00926

Coleman, R. G. (1974). Geologic background of the Red Sea. In R. B. Whitmarsh, O. E. Wesser, D. A. Ross, et al. (Eds.), *Initial reports of the deep sea drilling project* (Vol. 23, pp. 813–819). U.S. Govt. Printing Office.

Colombo, D., McNeice, G., Raterman, N., Zinger, M., Rivetta, D., & Sandoval Curiel, E. (2014). Exploration beyond seismic: The role of electromagnetics and gravity gradiometry in deep water subsalt plays of the Red Sea. *Interpretation*, 2, SH33–SH53. https://doi.org/10.1190/INT-2013-0149.1

Contrucci, I., Matias, L., Moulin, M., GeLL, L., Klingelhofer, F., Nouzé, H., Aslanian, D., Olivet, J.-L., Réhault, J.-P., & Sibuet, J.-C. (2004). Deep structure of the West African continental margin (Congo, Zaire, Angola), between 5°S and 8°S, from reflection/refraction seismics and gravity data. *Geophysical Journal International*, 158, 529–553. https://doi.org/10.1111/j.1365-246X.2004.02303.x

Cowie, L., Angelo, R. M., Kuznin, N., Manatschal, G., & Horn, B. (2017). Structure of the ocean-continent transition, location of the continent-ocean boundary and magmatic type of the northern Angolan margin from integrated quantitative analysis of deep seismic reflection and gravity anomaly data. In *Petroleum Geoscience...*
of the West Africa Margin, Geol. Soc. Lond. Spec. Publ. 438 (Ed. by T. Sabato Ceraldi, R. A. Hodgkinson & G. Backe). Geological Society, London, Special Publications, 438(1), 159–176.

Crosby, A. G., & McKenzie, D. (2009). An analysis of young ocean depth, gravity and global residual topography. Geophysical Journal International, 178, 1198–1219.

Crossley, R., Watkins, C., Raven, M., Cripps, D., Carnell, A., & Williams, D. (1992). The sedimentary evolution of the Red Sea and Gulf of Aden. Journal of Petroleum Geology, 15, 157–172.

Dadet, P., Marchesseau, J., Millon, R., Motti, E., & Schürmann, H. M. E. (1970). Mineral occurrences related to stratigraphy and tectonics in tertiary sediments near Umm Lajj, Eastern Red Sea Area, Saudi Arabia. Philosophical Transactions of the Royal Society A, 267, 99–106.

Davies, D., & Tramontini, C. (1970). The deep structure of the Red Sea. Philosophical Transactions of the Royal Society A, 267, 181–189.

Davison, I., Bosence, D., Alsop, G. I., & Al-Awah, M. H. (1996). Deformation and sedimentation around active Miocene salt diapirs on the Tihama Plain, northwest Yemen. In: Salt Tectonics, Geol. Soc. Spec. Publ. 100 (Ed. by G. I. Alsop, D. J. Blundell & I. Davison). Geological Society, London, Special Publications, 100(1), 23–39.

Dawood, Y. H., Aref, M. A., Mandurah, M. H., Hakami, A., & Gameil, M. A. Rasul, & I. C. F. Stewart (Eds.), The Red Sea: The formation, morphology, oceanography and environment of a young ocean basin (pp. 99–121). Springer Earth System Sciences.

Demets, C., & Merkouriev, S. (2016). High-resolution estimates of Nubia-Somalia plate motion since 20 Ma from reconstructions of the Southwest Indian Ridge, Red Sea and Gulf of Aden. Geophysical Journal International, 207, 317–332. https://doi.org/10.1093/gji/ggw276

Drake, C. L., & Girdler, R. W. (1964). A geophysical study of the Red Sea. Geophysical Journal of the Royal Astronomical Society, 8, 473–495.

Dullo, W.-C. (1990). Facies, fossil record, and age of Pleistocene reefs from the Red Sea (Saudi Arabia). Facies, 22, 1–45. https://doi.org/10.1007/BF02536943

Dyment, J., Tapponnier, P., Afifi, A. M., Zinger, M. A., Franken, D., & Musaiyen, E. (2013). A new seafloor spreading model of the Red Sea: Magnetic anomalies and plate kinematics. American Geophysical Union 2013 Fall Meeting.

Eagles, G., Gloaeguen, R., & Ebinger, C. (2002). Kinematics of the Danakil microplate. Earth and Planetary Science Letters, 203, 607–620.

Eagles, G., Pérez-Díaz, L., & Scarselli, N. (2015). Getting over continental margins. American Association of Petroleum Geologists Bulletin, 62, 223–234.

Fauquette, S., Suc, J.-P., Bertini, A., Popescu, S.-M., Warny, S., Taoufiq, N. B., Villa, M.-J.-P., Chikhi, H., Feddi, N., Subally, D., Clauzon, G., & Ferrier, J. (2006). How much did climate force the Messinian salinity crisis? Quantified climatic conditions from pollen records in the Mediterranean region. Palaeogeography, Palaeoclimatology, Palaeoecology, 238, 281–301.

Feldens, P., Schmidt, M., Mücke, I., Augustin, N., Al-Farawati, R., Orif, M., & Faber, E. (2016). Expelled subsalt fluids form a pockmark field in the eastern Red Sea. Geo-Marine Letters, 36, 339–352.

Fournier, M., Chamot-Rooke, N., Petit, C., Huchon, P., Al-Kathiri, A., Audin, L., Beslier, M.-O., D’Acremont, E., Fabbri, O., Fleury, J. M., Khanbari, K., Lepvrier, C., Leroy, S., Maillot, B., & Merkouriev, S. (2010). Arabia-somalina plate kinematics, evolution of the Aden-Owen-Carlsberg triple junction, and opening of the Gulf of Aden. Journal of Geophysical Research, 116. https://doi.org/10.1029/2008JB006257

Frostick, L., & Reid, I. (1989). Is structure the main control of river drainage and sedimentation in rifts? Journal of African Earth Sciences, 8, 165–182.

Garcia, S. F. M., Letouzey, J., Rudkiewicz, J.-L., Filho, A. D., & Lamotte, D. F. (2012). Structural modeling based on sequential restoration of gravitational salt deformation in the Santos basin (Brazil). Marine and Petroleum Geology, 35, 337–353.

Gargani, J., Moretti, I., & Letouzey, J. (2008). Evaporite accumulation during the Messinian Salinity Crisis: The Suez Rift case. Geophysical Research Letters, 35, art. L02401, https://doi.org/10.1029/2007GL032494

Gass, I. G., Mallick, D. I. J., & Cox, K. G. (1973). Volcanic Islands of the Red Sea. Journal of the Geological Society of London, 129, 275–310. https://doi.org/10.1144/jgs1973.116

Gaulier, J. M., Lepichon, X., Lyberis, N., Avedik, F., Gely, L., Moretti, I., Deschamps, A., & Hafez, S. (1988). Seismic study of the crustal thickness, northern Red Sea and Gulf of Suez. Tectonophysics, 153, 55–88.

Girdler, R. W., & Southren, T. C. (1987). Structure and evolution of the northern Red Sea. Nature, 330, 716–721. https://doi.org/10.1038/330716a0

Girdler, R. W., & Whitmarsh, R. B. (1974). Miocene evaporites in Red Sea cores, their relevance to the problem of the width and age of oceanic crust beneath the Red Sea. In R. B. Whitmarsh, O. E. Weser, D. A. Ross, et al. (Eds.), Initial reports of the deep sea drilling project (Vol. 23, pp. 913–921). U.S. Govt. Printing Office.

Gordon, G., Hansen, B., Scott, J., Hirst, C., Graham, R., Grow, T., Spedding, A., Fairhead, S., Fullarton, L., & Griffin, D. (2010). The hydrocarbon prospectivity of the Egyptian North Red Sea basin. Geological Society, London, Petroleum Geology Conference Series, 7, 783–789. In: Petroleum Geology: From Mature Basins to New Frontiers - Proceedings of the 7th Petroleum Geology Conference (Ed. by B. A. Vining & S. C. Pickering). https://doi.org/10.1144/0070783
Govers, R. (2009). Choking the Mediterranean to dehydration: The Messinian salinity crisis. Geology, 37, 167–170. https://doi.org/10.1130/G25141A.1

Griffin, D. L. (1999). The late Miocene climate of northeast Africa: Unravelling the signals in the sedimentary succession. Journal of the Geological Society, 156, 817–826.

Guennoc, R., Pautot, G., & Coutelle, A. (1988). Surficial structures of the northern Red Sea axial valley from 23°N to 28°N: Time and space evolution of the neocanoeic structures. Tectonophysics, 153, 1–23.

Gvirtzman, G., & Friedman, G. M. (1977). Sequence of progressive diagenesis in coral reefs. In S. H. Frost, M. P. Weiss, & J. B. Saunders (Eds.), AAPG special volumes (Sg 4): Reefs and related carbonates-ecology and sedimentology (pp. 357–380). Tulsa, OK: American Association of Petroleum Geologists.

Haase, K. M., Mühe, R. & Stoffers, P. (2000). Magmatism during extension of the lithosphere: Geochemical constraints from lavas of the Shaban Deep, northern Red Sea. Chemical Geology, 166, 225–239. https://doi.org/10.1016/S0009-2541(99)00221-1

Hall, S. A. (1979). A Total Intensity Magnetic Anomaly Map of the Red Sea and Its Interpretation, U.S. Geological Survey Saudi Arabian Project Report 275, U.S. Geological Survey, Jeddah, Saudi Arabia, 260 p.

Hamed, B., Bussert, R., & Domink, W. (2015). Stratigraphy and evolution of emerged pleistocene reefs at the Red Sea coast of Sudan. Journal of African Earth Sciences, 114, 133–142.

Hansen, E., Rodgers, A. J., Schwartz, S. Y., & Al-Amri, A. M. S. (2007). Imaging ruptured lithosphere beneath the Red Sea and Arabian Peninsula. Earth and Planetary Science Letters, 259, 256–265.

Hansen, E., Schwartz, S. Y., Al-Amri, A. M. S., & Rodgers, A. J. (2006). Combined plate motion and density-driven flow in the asthenosphere beneath Saudi Arabia: Evidence from shear-wave splitting and seismic anisotropy. Geology, 34, 869–872. https://doi.org/10.1130/G22713.1

Heaton, R. C., Jackson, M. P. A., Bamahmoud, M., & Nani, A. S. O. (1995). Superimposed neogene extension, contraction, and salt canopy emplacement in the Yemeni Red Sea. In M. P. A. Jackson, D. G. Roberts, & S. Nelsen (Eds.), Salt tectonics: A global perspective. AAPG Memoir 65, 333–351. Am. Assoc. Petrol. Geol.

Hoang, C. T., Dalongeville, R., & Sanlaville, P. (1996). Stratigraphy, tectonics and palaeoclimate implications of uranium-series-dated coral reefs from uplifted Red Sea carbonates. Geology, 24, 47–51.

Hoang, C. T., & Taviani, M. (1991). Stratigraphic and tectonic implications of uranium-series-dated coral reefs from uplifted Red Sea Islands. Quaternary International, 35, 274–373.

Hudec, M. R., Norton, I. O., Jackson, M. P. A., & Peel, F. J. (2013). Jurassic evolution of the Gulf of Mexico salt basin. American Association of Petroleum Geologists Bulletin, 97, 1683–1710.

Hughes, G. W. (2014). Micropalaeontology and palaeoenvironments of the Miocene Wadi Waq carbonate of the northern Saudi Arabian Red Sea. GeoArabia, 19, 59–108.

Hughes, G. W., Abidine, S., & Girgis, M. H. (1992). Miocene biofacies development and geological history of the Gulf of Suez, Egypt. Marine and Petroleum Geology, 9, 2–28.

Hughes, G. W., & Beydoun, Z. R. (1992). The Red Sea - Gulf of Aden: Biostратigraphy, lithostatigraphy and palaeoenvironments. Journal of Petroleum Geology, 15, 135–156.

Hughes, G. W., Varol, O., & Beydoun, Z. R. (1991). Evidence for Middle Oligocene rifting of the Gulf of Aden and for Late Oligocene rifting of the southern Red Sea. Marine and Petroleum Geology, 8, 354–358. https://doi.org/10.1016/0264-8172(91)90088-I

Imbert, P. (2005). The mesozoic opening of the Gulf of Mexico: Part 1, Evidence for oceanic accretion during and after salt deposition. In P. J. Post (Ed.), Transactions of the 25th Annual GCS-SEPM: Petroleum systems of divergent continental margins (pp. 119–150). SEPM, OK: Tulsa.

Izzeldin, A. Y. (1982). On the structure and evolution of the Red Sea. Strasbourg, France: PhD Thesis, Université Louis Pasteur, 163 p.

Izzeldin, A. Y. (1987). Seismic, gravity and magnetic surveys in the central part of the Red Sea: Their interpretation and implications for the structure and evolution of the Red Sea. Tectonophysics, 143, 269–306. https://doi.org/10.1016/0040-1951(87)90214-9

Izzeldin, A. Y. (1989). Transverse structures in the central part of the Red Sea and implications on early stages of oceanic accretion. Geophysical Journal, 96, 117–129.

Jackson, M. P. A., Cramez, C., & Fonck, J.-M. (2000). Role of subaerial volcanic rocks and mantle plumes in creation of South Atlantic margins: Implications for salt tectonics and source rocks. Marine and Petroleum Geology, 17, 477–498. https://doi.org/10.1016/S0264-8172(00)00006-4

Jado, A. R., & Zötl, J. (1984). Quaternary period in Saudi Arabia. Springer-Verlag.

Karner, G. D., & Gamboa, L. A. P. (2007). Timing and origin of the South Atlantic pre-salt sag basins and their capping evaporites. In: Evaporites through Space and Time (Ed. by B. C. Schreiber, S. Lugli & M. Babel). Geological Society, London, Special Publications, 285(1), 15–35.

Knott, S. T., Bunce, E. T., & Chase, R. L. (1966). Red Sea Seismic Reflection Studies. In: The World Rift System, Geol. Surv. Canada, Paper 66-14, pp. 78–97.

Kozdroj, W., Kattan, F. H., Kadi, K. A., Al Alfy, Z. S. A., Qweiss, K. A., & Mansour, M. M. (2012). SGS-EMRA Project for Trans-Red Sea Correlation between the Central Eastern Terrane (Egypt) and Midyan Terrane (Saudi Arabia): Saudi Geological Survey Technical Report SGS-TR-2011-5, 62 pages, 65 pls.

Krijgsman, W., Hilgen, F. J., Raffi, I., Sierro, F. J., & Wilson, D. S. (1999). Chronology, causes and progression of the Messinian salinity crisis. Nature, 400, 652–655. https://doi.org/10.1038/23231

Lambeck, K., & Chappell, J. (2001). Sea level change through the last glacial cycle. Science, 292, 679–686. https://doi.org/10.1126/science.1059549

Lambeck, K., Purcell, A., Fleming, N. C., Vita-Finzi, C., Alsharekh, A. M., & Bailey, G. N. (2011). Sea level and shoreline reconstructions for the Red Sea: Isostatic and tectonic considerations and implications for hominin migration out of Africa. Quaternary Science Reviews, 30, 3542–3574. https://doi.org/10.1016/j.quascirev.2011.08.008

Ligì, M., Bonatti, E., Bortoluzzi, G., Cipriani, A., Cocchi, L., Caratori Tontini, F., Carminati, E., Ottolini, L., & Schettino, A. (2012). Birth of an ocean in the Red Sea: Initial pangs. Geochemistry, Geophysics, Geosystems, 13, Paper Q08009. https://doi.org/10.1029/2012G004155

Ligì, M., Bonatti, E., Bosworth, W., Cai, Y., Cipriani, A., Palmiotto, C., Ronca, S., & Seyler, M. (2018). Birth of an ocean in the Red Sea: Oceanic-type basaltic melt intrusions precede continental rifting. Gondwana Research, 54, 150–160. https://doi.org/10.1016/j.gr.2017.11.002

Ligì, M., Bonatti, E., Tontini, F. C., Cipriani, A., Cocchi, L., Schettino, A., Bortoluzzi, G., Ferrante, V., Khalil, S., Mitchell, N. C., & Rasul,
Shi, W., Mitchell, N. C., Kalnins, L., & Izzeldin, A. Y. (2018). Oceanic-like axial crustal high in the central Red Sea. *Tectonophysics*, 747–748, 327–342. https://doi.org/10.1016/j.tecto.2018.10.011

Sicilia, D., Montagner, J.-P., Cara, M., Stutzmann, E., Debayle, E., Lépine, J.-C., Lévêque, J.-J., Beuclet, E., Sebai, A., Roul, G., Ayele, A., & Sholan, J. M. (2008). Upper mantlne structure of shear-waves velocities and stratification of anisotropy in the Afar Hotspot region. *Tectonophysics*, 462, 164–177.

Smith, W. H. F., & Sandwell, D. T. (1997). Global sea floor topography from satellite altimetry and ship soundings. *Science*, 277, 1956–1962.

Smith, W. H. F., & Wessel, P. (1990). Gridding with continuous curvature splines in tension. *Geophysics*, 55, 293–305. https://doi.org/10.1190/1.1442837

Sofianos, S. S., & Johns, E. W. (2007). Observations of the summer Red Sea circulation. *Journal of Geophysical Research*, 112, Paper C06025. https://doi.org/10.1029/2006JC003886

Stockli, D. F., & Bosworth, W. (2019). Timing of extensional faulting along the magma-poor central and northern Red Sea Rift margin - Transition from regional extension to necking across a hyperextended rifted margin. In N. M. A. Rasul, & I. C. F. Stewart (Eds.), *Geological setting, palaeoenvironment and archaeology of the Red Sea* (pp. 81–111). Springer Nature Switzerland.

Stoffers, P. & Kühn, R. (1974).Red Sea evaporites: A petrographic and geochemical study. In R. B. Whitmarsh, O. E. Weser, D. A. Ross, et al. (Eds.), *Initial reports of the deep sea drilling project* (Vol. 23, pp. 821–847). U.S. Govt. Printing Office.

Stoffers, P., Ross, D. A. et al (1974). Sedimentary history of the Red Sea. In R. B. Whitmarsh, O. E. Weser, & D. A. Ross (Eds.), *Initial reports of the deep sea drilling project* (Vol. 23, pp. 849–865). U.S. Govt. Printing Office.

Strozzy, F., Back, S., & Kukla, P. A. (2017). Comparison of the rift and post-rift architecture of conjugated salt and salt-free basins offshore Brazil and Angola/Namibia, South Atlantic. *Tectonophysics*, 716, 204–224. https://doi.org/10.1016/j.tecto.2016.12.012

Sultan, M., Becker, R., Arvidson, R. E., Shore, P., Stern, R. J., al Alfy, Z., & Guinness, E. A. (1993). New constraints on Red Sea rifting from correlation of Arabian and Nubian Neoproterozoic outcrops. *Tectonics*, 12, 1303–1319.

Taviani, M. (1998). Post-miocone reef faunas of the Red Sea: Glacioisostatic controls. In B. H. Purser, & D. W. J. Bosence (Eds.), *Sedimentation and tectonics of rift basins: Red Sea-Gulf of Aden* (pp. 574–582). Chapman & Hall.

Taylor, J. R. (1982). *An introduction to error analysis, the study of uncertainties in physical measurements*. Oxford University Press.

Tedesco, L. R., Jenkyns, H. C., Robinson, S. A., Sanjínés, A. E. S., Viviers, M. C., Quintaes, C. M. S. P., & Vazquez, J. C. (2017). New age constraints on apatite evaporationes and carbonates from the South Atlantic: Implications for oceanic anoxic event 1a. *Geology*, 45, 543–546. https://doi.org/10.1130/G38886.1

Torsvik, T. H., Rouse, S., Labails, C., & Smethurst, M. A. (2009). A new scheme for the opening of the South Atlantic ocean and the dissection of an Aptic Salt basin. *Geophysical Journal International*, 177, 1315–1333.

Tramontini, C., & Davies, D. (1969). A seismic refraction survey in the Red Sea. *Geophysical Journal of Royal Astronomical Society*, 17, 225–241.

Turcotte, D. L., & Schubert, G. (1982). *Geodynamics: Applications of continuum physics to geologocal problems*. John Wiley and Sons.

Uchupi, E., & Ross, D. A. (1986). The tectonic style of the northern Red Sea. *Geo-Marine Letters*, 5, 203–209. https://doi.org/10.1007/BF02233804

van der Zwan, F. M., Augustin, N., Devey, C. W., Bantan, R., & Kwasnitschka, T. (2015). New insights into volcanism and tectonics in the Red Sea rift. *Chemical Geology*, 405, 63–81.

Volker, F., McCulloch, M. T., & Altherr, R. (1993). Submarine basals from the Red Sea: New Pb, Sr, and Nd isotopic data. *Geophysical Research Letters*, 20, 927–930.

Warren, J. K. (2010). Evaporites through time: Tectonic, climatic and eustatic controls in marine and nonmarine deposits. *Earth Science Reviews*, 98, 217–268.

Watts, A. B., & Burov, E. B. (2003). Lithospheric strength and its relationship to the elastic and seismogenic layer thickness. *Earth and Planetary Science Letters*, 213, 113–131.

Wessel, P., & Smith, W. H. F. (1991). Free Software helps map and display data. *EOS, Transactions, American Geophysical Union*, 72, 441. https://doi.org/10.1029/90EO00319

Whelidon, J., Evans, T. R., & Girdler, R. W. (1974). Thermal conductivity, density, and sonic velocity measurements of samples of anhydrite and halite from sites 225 and 227. In R. B. Whitmarsh, O. E. Weser, D. A. Ross, et al. (Eds.), *Initial reports of the deep sea drilling project* (Vol. 23, pp. 909–911). U.S. Govt. Printing Office.

Whipple, K. X. (2001). Fluvial landscape response time: How plausibl is steady-state denudation? *American Journal of Science*, 301, 313–325. https://doi.org/10.2475/ajs.301.4.5.313

Whitmarsh, R. B., Mantatschal, G., & Minshull, T. A. (2001). Evolution of magma-poor continental margins from rifting to seafloor spreading. *Nature*, 413, 150–154. https://doi.org/10.1038/35093085

Whitmarsh, R. B., Weser, O. E., & Ross, D. A. (1974). *Initial reports of the deep sea drilling project*, 23b, Washington, D.C.: U. S. Government Printing Office.

Wilson, J. W. P., Roberts, G. G., Hoggard, M. J., & White, N. J. (2014). Cenozoic epeirogeny of the Arabian Peninsula from drainage modeling. *Geochemistry, Geophysics, Geosystems*, 15, 3723–3761. 3710.1002/2014GC005283

Wolfe, C. J., Vernon, F. L., & Al-Amri, A. (1999). Shear-wave splitting across western Saudi Arabia: The pattern of upper mantle anisotropy at a Proterozoic shield. *Geophysical Research Letters*, 26, 779–782. https://doi.org/10.1029/1999GL00056

Zahran, H. M., Stewart, I. C. F., Johnson, P. R., & Basahel, M. H. (2003). *Aeromagnetic- Anomaly Maps of Central and Western Saudi Arabia*. Saudi Geological Survey. Scale 1:2 Million. Saudi Geological Survey Open-File Report SGS-OF-2002-8, 6 p., 4 plates.

---

**How to cite this article:** Mitchell NC, Shi W, Izzeldin AY, Stewart ICF. Reconstructing the level of the central Red Sea evaporites at the end of the Miocene. *Basin Res* 2021;33:1266–1292. https://doi.org/10.1111/bre.12513