Submarine salt flows in the central Red Sea

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Abstract

The central Red Sea, an oceanic basin floored by Miocene evaporites reaching kilometres in thickness in places, is at an early stage of development, where seafloor spreading has geologically only recently replaced continental rifting. Surveying with a high-resolution multibeam echo-sounder around Thetis Deep, a new spreading centre, has revealed a remarkable series of structures resembling viscous gravity flows, which are

here interpreted as originating from flowage of the evaporites laterally unloaded by axial rifting and other processes developing the relief of the deep. The flow margins are marked by stream-wise lineaments and some apparently rotated markers. Their fronts in the floor of the deep are rounded in plan view and profile. Their surfaces contain small closely spaced features resembling extensional faults. In one area below declining gradients, the surface contains along-slope ridges and valleys typical of compression folds (ogives). Flow-parallel lineaments and extensional faults lie, respectively, parallel and orthogonal to the direction of maximum seabed gradient. Movement is apparently heterogeneous, at least in part by varied blocking by relief of underlying basement observed protruding between flows. Flowage is currently transporting materials into the floor of the deep where it has the potential to become incorporated into the young oceanic crust by repeated eruption of axial lavas over them. In the light of these new data, we re-examine the possibility and implications of flowage in the South Atlantic marginal evaporites, in particular whether flowage contaminated early oceanic crust in such areas.

Introduction

During the early stages of the Atlantic basin's development, widespread evaporites were deposited off parts of Angola, Congo, Gabon, Brazil, Morocco and North America at times when sea water communication with the global oceans was restricted and continental climates were dry (Emery, 1977; Evans, 1978; Pautot et al., 1966; Rona, 1982). In an early review of extant seismic data, Emery and Uchupi (1984) showed that many Atlantic marginal evaporites lie near the stratigraphic boundary between the faulted sediments contemporaneous with active continental rifting (the syn-rift) and the later-

deposited post-rift sediments that are commonly less deformed. In places, the evaporites extend onto oceanic or proto-oceanic crust (Contrucci et al., 2004; Emery and Uchupi, 1984; Meyers et al., 1996a). Although the common role of evaporites in tectonic deformation and diapirism following their loading by terrigenous sediments is well known (e.g., Gemmer et al., 2005), how they behaved in the earlier stages shortly after their precipitation from seawater is less well understood. Stratigraphic evidence for the early evaporite behaviour in the Atlantic is complicated, however, as seismic resolution is poor beneath kilometres of later terrigenous sediments and the loaded evaporites are commonly remobilised.

Since the 1970s, it has been known that the central Red Sea evaporites were deposited in a pre-existing continental rift valley (Coleman, 1974; Girdler and Whitmarsh, 1974). The topography of axial deeps there is thought to have enlarged with continued tectonic extension and subsidence, and dissolution of rift valley evaporites promoted by seafloor exposure and hydrothermal circulation associated with volcanism. This removed the lateral constraint on evaporites outside the deeps, allowing them to flow into the new depressions. Although the detailed geometric correspondence remains to be worked out, the Red Sea has been considered to be an analogue for the earlier evaporite-hosting margins of the Atlantic.

The salt flows described in the new sonar data presented here are a type of sedimentary flow in which viscous fluid-like behaviour of evaporites originates from the weak rheology of their halite (rock salt) component. They occur in subaerial environments with arid climates such as in Iran where they are called *namakiers* (Talbot and Jarvis, 1984). Otherwise described as salt glaciers, tongues or allochthonous sheets,

salt flows have been interpreted from seismic reflection data where rock salt has extruded onto the seabed (e.g., Amery, 1969; Demercian et al., 1993; Diegel et al., 1995; Fletcher et al., 1995; Hall, 2002; McBride et al., 1998; Mohr et al., 2007; Nelson, 1991; Talbot, 1993; Wu et al., 1990). In such environments, salt dissolution may be impeded by a surface layer of sediment and dissolution residue (Fletcher et al., 1995; Jackson et al., 1994; Wu et al., 1990).

After providing background information on the geology, we present the new data, which were collected on the Italian research vessel *Urania* in 2005. We describe the flow-like features and argue why their movements are likely to originate from the evaporites. We explore in the discussion issues complicating understanding of their dynamics, such as role of pore fluids and mylonitisation of halite. The possibility that some flowage occurred during the early stages of the Atlantic basin marginal evaporites is then re-examined in the light of these new results.

Regional structure

The Red Sea is a classical nascent oceanic basin, with mainly extended continental crust in the north (Cochran, 2005) giving way to continuous sea-floor spreading between Nubia and Arabia in the south (Bonatti, 1985; Cochran, 1983). Outside the deeps, basement seismic refraction velocities (Gaulier et al., 1988) are slower than those typical of oceanic crust but are in places somewhat fast for continental crust, which may be explained if basement of generally continental composition is variably intruded with dykes (Cochran and Martinez, 1988; Cochran, 1983). Accordingly, the Red Sea is an extended and subsided continental basin, with sea-floor spreading along

parts of its axis. With continued extension, axial deeps subside and their evaporites dissolve because of water penetrating along faults, tectonic exposure on the seabed, basaltic injections and hydrothermal circulation promoted by intrusions, creating deep depressions (Bonatti et al., 1984; Guennoc et al., 1988; Ross and Schlee, 1973). Losing their lateral constraint, the evaporites around the deeps then flow into them (Coleman, 1974; Girdler and Whitmarsh, 1974). Their subsequent flowage can potentially explain evaporites now overlying older magnetic anomalies (Girdler and Whitmarsh, 1974), halokinetic deformation away from the rift axis (Richter et al., 1991), and tectonic structures in Deep Sea Drilling Project (DSDP) cores (Whitmarsh et al., 1974).

Thetis Deep (Figure 1) lies in the central Red Sea, where the axial deeps are nuclei of initial oceanic crust emplacement, with sea-floor spreading anomalies developed only for the last 2-3 m.y. at most (Bonatti, 1985; Chu and Gordon, 1998; Cochran and Martinez, 1988; Searle and Ross, 1975) and where the spreading rate is a slow 12 mm/yr (Chu and Gordon, 1998). Magnetic field measurements over Thetis Deep show prominent axial dipole anomalies and lesser secondary anomalies (Chu and Gordon, 1998; Izzeldin, 1987; Ligi et al., 2008), suggesting that basement magnetisation was acquired since only one to three major reversals of Earth's main field. Sea-floor spreading *sensu stricto* began not much earlier than the Brunhes-Matuyama boundary at 0.78 Ma. The central Red Sea is at an intermediate development stage, with an axial trough that is discontinuous along-strike. Deepes are separated by shallower "inter-trough" areas lacking axial magnetic anomalies and with a continuous but tectonically disturbed sedimentary cover in seismic reflection data (Bonatti et al., 1984). Other studies have reported the possibility of evaporite flowage in the central Red Sea, but lack

details of surface structures. For example, Bicknell et al. (1986) mentioned sidescan sonar images of Nereus Deep showing evaporite wedges overflowing faults. Flow-like lobate features occur in coarsely contoured bathymetry around the base of slopes within Vema, Nereus and Atlantis II Deep (Bicknell et al., 1986; Pautot, 1983).

Stratigraphy

Figure 2 summarises the stratigraphy at DSDP Sites 225 and 227 (Whitmarsh et al., 1974; Stoffers and Ross, 1974), which were drilled on the flanks of Atlantis II Deep (Figure 1). Although Thetis Deep has not been sampled by scientific drilling, continuous seismic reflection profiling (Bonatti, 1985; Bonatti et al., 1984; Phillips and Ross, 1970; Ross and Schlee, 1973) suggests a laterally consistent stratigraphy along the centre of the Red Sea. In particular, a regionally traceable reflector ('S'), described by earlier workers and confirmed in our seismic reflection data introduced below, suggests that the shallow stratigraphy around Thetis Deep is likely similar to that of Atlantis II Deep. At each of the DSDP sites, three units overlying the evaporites were identified. From the sea floor downwards (Figure 2, left), these are: (1) detrital silty clay, nanno ooze and chalk, (2) calcareous silty clay and clayey silt and siltstone and (3) dolomitic siltstones and shales. Features interpreted to have been caused by salt tectonics include deformed intervals (Girdler and Whitmarsh, 1974) (vertical bars to the right of each column in Figure 2) and steep 10°-60° dips of anhydrite within the evaporites (Whitmarsh et al., 1974). Measured sample densities (Figure 3) are similar to or less than halite densities, so we do not expect density inversion and diapirism of the evaporites.

The drilled evaporites are of Late Miocene age (Whitmarsh et al., 1974; Stoffers and Kühn, 1974). Their maximum thickness sampled, 133 m at Site 227, strongly under-represents their potential full stratigraphic thickness, which reaches 2-3 km around the farthest seawards commercial wells (Miller and Barakat, 1988; Ross and Schlee, 1973) and more than 4 km inferred from seismic refraction data (Avedik et al., 1988; Egloff et al., 1991; Gaulier et al., 1988; Rihm et al., 1991). In the DSDP cores, the evaporites include mainly halite and anhydrite, intercalated on various scales from < 1 cm to 10s of m, with intervening minor shale. Halite is medium to coarse grained and commonly cloudy from brine inclusions, with pore spaces filled with polyhalite needles. Based on various evidence, Stoffers and Kühn (1974) argued that the anhydrite was deposited in shallow marine and, in places, supratidal environments.

Seismic reflector S correlates to anhydrite beds at the top of the evaporites or to an immediately overlying claystone (Whitmarsh et al., 1974). Away from coasts, S is folded but otherwise conformable with the seabed, except around the margins of deeps where the sedimentary carapace typically thins into them (Bonatti, 1985; Whitmarsh et al., 1974). Whereas in the north S is deformed but can be traced continuously across deeps (Cochran et al., 1986; Ehrhardt et al., 2005; Guennoc et al., 1988; Martinez and Cochran, 1988; Phillips and Ross, 1970), in the south and centre of the Red Sea, evaporites apparently outcrop in seismic reflection data along the walls of deeps (Bonatti, 1985; Bonatti et al., 1984; Egloff et al., 1991; Guennoc et al., 1988; Phillips and Ross, 1970; Ross and Schlee, 1973; Searle and Ross, 1975), although in practice they are probably covered by seismically unresolved sediment that prevents dissolution (Eissen et al., 1989; Ross and Schlee, 1973).

Seismic refraction experiments

Refraction experiments using explosives (Tramontini and Davies, 1969; Davies and Tramontini, 1970) were carried out along the white lines in Figure 1, with solid circles marking sonobuoy locations. Despite the limits to interpretation of these data due to the celestial navigation of the experiments, they broadly show the evaporite geometry around Thetis Deep. Away from the axial trough, the travel time data reveal a refractor corresponding to the interface between two layers of strongly differing velocity. Its depth is represented by solid squares in Figure 4, plotted at the sonobuoy locations projected onto a line oriented perpendicular to the rift shown in Figure 1. The refractor's upper and lower layers have velocities of 3.99-4.55 and 5.72-8.64 km/s, respectively. Scatter of arrival times about trends expected of plane layers suggest that the interface has significant relief.

The upper layer (light grey bars in Figure 4) thickens with distance away from the deep. Outside the deep, it is likely sedimentary rather than volcanic in origin because of a lack of magnetic anomalies that would be expected to be associated with lavas. Its mean velocity of 4.26 km/s there is as expected for evaporites comprising a mixture of halite, anhydrite and shale (compared with 4.2 km/s of halite and 4.8 km/s of anhydrite samples (Wheildon et al., 1974) and less than 4 km/s of shale samples (Whitmarsh et al., 1974)). The exact composition of the evaporites cannot be easily estimated from these data but it is unlikely to be dominated by anhydrite, which would displace the velocity away from 4.2 km/s.

As 4.26 km/s overlaps with seismic velocities of volcanic extrusives (Hammer et al., 1994; Mitchell, 2001), the interface between the two layers potentially marks a velocity change within igneous basement rather than the basement-evaporite interface exactly. Based on the depth to comparable velocities in typical velocity-depth profiles of slow-spread oceanic crust (Minshull et al., 2006; Minshull et al., 1991), the bias in the off-axis evaporites is expected to be < 1 km, so they are likely > 1 km thick.

Refraction lines within the deep traverse volcanic morphology in our multibeam sonar data described later, so lavas and intrusive volcanics likely comprise a significant part of the top layer there. It probably also contains varying amounts of sediment, suggested by flat-lying deposits typical of turbidites and as the flowage described later will have carried sediments into the deep. Within the inter-trough zone, the upper layer (Figure 4, left graph) is thicker than elsewhere along axis and has a lower mean velocity of 4.13 km/s, consistent with a more intact evaporite body (Bonatti et al., 1984; Searle and Ross, 1975). A 30-40 mgal depression in the free air gravity anomaly there was interpreted (Tramontini and Davies, 1969) as caused by a >1 km depression of the pre-evaporite surface.

New acquisition of sonar and seismic data

Multichannel seismic refraction data (Figure 5) were acquired along three lines perpendicular to Thetis Deep and along two axis-parallel lines (located in Figures 6 and S1) using two synchronized 105 cubic inch GI airguns. They were collected with a 50 m shot spacing on a 600 m analogue streamer, digitized and recorded at a 1 msec sampling rate. Their processing included time-migration.

A hull-mounted 50 kHz Reson multibeam sonar was used to acquire bathymetry data with an acoustic spatial resolution of 25-50 m (varied by beam spreading with increasing water depth) along tracks run parallel to the rift. The sonar's high frequency limits seabed penetration (Mitchell, 1993). Data were collected with satellite-broadcasted Global Positioning System (GPS) differential corrections. For refraction corrections to the data, a depth-profile of acoustic velocity was acquired with a sound velocity probe and velocity changes monitored using occasional expendable bathythermographs. The sonar data were processed using software which allowed manual removal of obvious noise. Bathymetry grids were then produced using public (Wessel and Smith, 1991) and in-house software. The sonar data are shown as a shaded relief image and contour map in Figure 6 and as a 3D image in Figure 7. For later discussion on the origin of seabed flow features, Figure 8a shows a map of the direction and magnitude of seabed gradient. This was computed from the sonar data after filtering them over 3 km, a length-scale chosen as a compromise between preventing the gradient map appearing noisy from fine-scale features if the length-scale is too small and the need for sufficient data around the data margins to compute seabed curvatures shown in Figure 8b.

Some parallel lines of fine mottled texture running up the page in Figure 6a (marked "sonar artefacts") are caused by remaining noise in the sonar's outer beams so surface texture needs to be interpreted bearing these in mind, as well as some fine cross-track striping from the effect of rough weather on the system. Verifying the depths measured by multibeam systems is difficult so interpretation relies usually on observer experience and knowledge of known artefacts (de Moustier and Kleinrock, 1986; Hughes Clarke et al., 1996) but data collected without obvious blunders commonly resemble a

filtered version of the true seabed topography overlain with some noise (Goff and Kleinrock, 1991; Schmitt et al., 2008), which is here mostly attenuated by filtering during the data gridding. Correspondence of data in overlapping tracks and crossing seismic lines provide further confidence in these results. Interpretation is also simplified here as deep bottom currents are expected to be generally weak in the central Red Sea, so that abyssal current transport can be largely ignored. For example, deep thermohaline current speeds are only centimetres per second (Woelk and Quadfasel, 1996). A lack of strong currents is also implied by the common presence of stable brine layers in closed-contour deeps (e.g., Pautot et al., 1984).

Observations

Seismic reflection images

Within each image (Figure 5), a strong reflector occurs at similar two-way time to S identified previously (Phillips and Ross, 1970; Ross and Schlee, 1973) and at roughly similar depth to the evaporite surface at DSDP Sites 225 and 227, so we also interpret this as the S-reflector here. A line of data collected in 1979 crosses one of the flow-like features described later in the wall of Thetis Deep (Discovery 103). Although suffering from a strong source bubble oscillation, the record shows a prominent reflector at 0.17s two-way time or around 200 m sub-bottom depth, which is also identified as S.

On the right of line 21M, fold-like features occur. In places just lower-left of "NE" in Figure 5, fold amplitude is greater below S than at the seabed. On the right of 23M, similar features can be observed and that below "Syncline" has larger amplitude

below S than above it. On 25M, lower structures do not have much different fold amplitude to those above S.

Below 'f' in line 25M, a discontinuity between adjacent reflectors can be observed, with reflectors either side of a boundary differing in two-way time and the interface between them showing diffraction hyperbolae (not removed by migration). Somewhat similar features can be observed in the rift-parallel line 24M, which also includes a prominent fold-like feature on its right side. Crossing the inter-trough area, line 26M shows prominent shallow reflectors ('S') on the left and where it nears the floor of the deep on the right, but stratigraphy in the intervening region is confused, as observed in other inter-trough seismic lines (Bonatti et al., 1984).

On some lines approaching the floor of the deep, S shallows towards the seafloor, so the overlying hemipelagic sediments taper out. This occurs, for example, on the left of 23M (where S is marked) and below the 10 km scale bar of 26M. Beneath flat-lying depressions within the floor of the deep (21M, 23M and 26M), some parallel sub-bottom reflectors can be observed but are of low amplitude.

Multibeam sonar

Within the floor of Thetis Deep, which is roughly encompassed by the 1500 m contour in Figure 6b, the topography is rugged. Many small elevated circular features there, some with central depressions (e.g., "cone" in Figure 6a), are typical of volcanic cones at mid-ocean ridges (Smith and Cann, 1992). One feature marked "ridge" in Figure 6a has a knobbly morphology typical of axial volcanic ridges (Parson et al., 1993). Abundant small-relief sharp-crested escarpments observed in the image are also typical of

normal faults found in mid-ocean ridge valley floors (Ballard and van Andel, 1977).

Some areas of Thetis Deep's floor are flat, such as where marked "basin" in Figure 6a, possibly due to either sedimentary deposits such as turbidites or ponded widespread lava flows.

Outside the floor, the topography is smoother and bathymetry contours are less convoluted (Figures 6 and 7). To the left of 'a' and 'b' in Figure 6a, features can be observed with fronts on the floor of the deep that are rounded in plan view and profile (Figure 9). Some long ridges and troughs on their surfaces are oriented perpendicular to bathymetry contours (Figure 6b) and parallel to the plan-view direction of maximum seabed gradient (Figure 8a). Surface textures of features 'a' and 'b' include shallow, closely-spaced ridges and troughs oriented roughly along-slope, with typical spacings of 200-500 m and angular relief <20 m, illustrated by the easterly 3 km of profile y-y' shown enlarged and vertically exaggerated at the top of Figure 9. The grey shading in Figure 6a picks out broader scale undulations, i.e., the steps in elevation also shown in Figures 7 and 9 (numbered in the latter). These steps correlate along-slope with more sharply defined escarpments in the adjacent more rugged terrain (Figure 7). Step two away from the floor of the deep within 'a' (marked '2' on Figure 6) contains a small area of rugged morphology within an embayment. From the contours in Figure 6b, these flow-wise features have 200-500 m of vertical relief above the terrain lying adjacent along-rift. Furthermore, the shading in Figure 6a and contours in 6b highlight a straight but smooth relief of a 100-m-high escarpment running parallel with the rift within 10 km of the top-right edge of the map (marked "Fault" along the top of Figure 6a), which suggests a buried normal fault escarpment associated with the axial rift.

In the right side of Figure 7, two areas of rugged morphology with similar roughness to that on the floor of the deep separate areas of smooth morphology 'a' and 'b' and a further smooth region to the north. At the southern edges of areas 'a' and 'b' in Figure 6a marked 'cf', two sets of convex-northeast curved fabric can be observed. Along the south side of feature 'a' immediately above profile x-x', features oriented rift-parallel are crossed by perpendicular normal faults. Though less well developed, flow-like features east of 'c' and 'd' in Figure 6a also have rounded fronts, some down-slope oriented ridges and troughs, and small surface lineaments oriented along-slope. Elsewhere, the contact between the floor of the deep and the smooth terrain commonly has a rounded, lobate morphology. In the top of Figure 7 and to the left of 'h' in Figure 6a, a 10-km wide region of small surface ridges and troughs gives way down-slope to a region of ridges and troughs with larger spacings and heights.

In the inter-trough area (northwest of 22°48'N, 37°33'E), some curvilinear ridges and troughs occur oriented parallel to the deep. At 'e', in contrast, small ridges and troughs form a fanning pattern. At 'f', larger-wavelength ridges and troughs, and some elsewhere to the upper-left of the floor of the deep, are oriented at high angles to the deep. This can be observed between 'g' and near the boundary with the volcanic morphology to the southeast.

Interpretation

A volcanic origin for the flow-like features of Figure 6 is ruled out by a lack of any volcanic edifice relief in regional bathymetry or magnetized source in the magnetic anomaly data (Izzeldin, 1987). In addition, the folding of layered reflectivity observed in

the seismic reflection images outside the deep is characteristic of sedimentary sequences. Although the alternative possibility of slippage within the superficial hemipelagics cannot be ruled out entirely, we observed no shallow-gradient reflectors of displacement surfaces in our seismic images, nor are there any embayments typical of landslide headwalls around the upper walls of the deep in Figure 6. Deformation structures in the Site 225 and 227 hemipelagics were described as small also with no evidence of large displacements (Girdler and Whitmarsh, 1974). Furthermore, the 200-500 m relief of the flow-like features is mostly greater than the hemipelagic thickness. Nevertheless, some slope failure of the hemipelagic beds may occur, such as around the steep margins of the flows. The flow fronts marked in Figure 9 are also difficult to interpret from surface morphology alone as the southern lobe of flow 'a', for example, could represent a shallow volcanic laccolith or cone. In agreement with earlier predictions (Coleman, 1974; Girdler and Whitmarsh, 1974; Ross and Schlee, 1973), however, we interpret the main flow-like features of Figure 6 (a, b, c, d, etc) as having been caused by evaporite flowage, which passively carried the overlying hemipelagic section.

Widely distributed extension within the thin hemipelagic layer by normal faulting or boudinage is interpreted here to have created the fine-scale along-slope ridges and troughs on the surface of the flows. The down-slope ridges and troughs could either originate from deformation of the flows as they move over underlying surface topography, much like how some streamwise marks form in ice streams (Gudmundsson et al., 1998; Stokes et al., 2007) and lava flows (Keszthelyi et al., 2004), or they arise from strike-slip movements of parts of the evaporites. The latter seems likely for at least some of these features based on the strike-slip-like features in the seismic lines (Figure 5)

and some differential movements could potentially explain the separate flow lobes in plan-view within 'a' and 'b' (Figure 6a).

The smooth surfaces of the flows suggests that halite is not generally exposed at the seabed, which otherwise variably dissolves to leave irregular topography (Pilcher and Blumstein, 2007), so the hemipelagics have hydraulically sealed the underlying evaporites to prevent their widespread dissolution. Small irregular areas, however, such as in the second step away from the floor of the deep along flow 'a', may have been caused by local breaching of the hemipelagic layer and dissolution.

The question of how basement affects the flow pattern needs further seismic data to address, but the peaks of rugged topography rising above the smooth flow surfaces in Figure 7 suggest that they are obstructing flowage in these areas. Those peaks lie 1 km above the floor of the deep, which is a typical axial valley relief for slow oceanic spreading centres (Small, 1994). The curved fabrics on the southeasterly side of flow 'b' may have originated in a more rift-parallel orientation, perhaps due to halokinetic deformation or from large displacements on underlying rift faults able to break through the evaporites. Although their origin is uncertain, their curvature suggests increasing displacement towards the centre of flow 'b' with their southerly margins pinned by a flow obstruction.

The configuration of small ridges immediately west of 'h' in Figure 6a, passing down-slope to larger wavelength ridges where seabed gradient shallows suggests that, whereas the former are caused by distributed extension of the flow surface on the fault escarpment slope much like in spreading failures (Micallef et al., 2007), the latter features are compression folds, much like ogives of glacier or volcanic flows slowing on

shallower gradients (de Silva et al., 1994). The larger ridges are less angular than the fine-scale features in profile, consistent with folding and a lack of faults offsetting the seabed. The disrupted stratigraphy in seismic line 26M (and in seismic profile 'g' of Bonatti et al. (1984) crossing this area perpendicular to 26M) favours a tectonic rather than sedimentary origin for these ridges. A sharp trough crossing these fold-like features immediately to the right of and sub-parallel to line 26M in Figure 6a offsets a lineament suggesting possible right-lateral strike-slip displacement. Along with further crossing features, the fanning of fabrics at 'e', curvature of their surrounding down-slope troughs towards the south and the rounded front of the inter-trough sedimentary terrain where it meets the floor of the deep, these observations suggest that the evaporites here are moving with a significant component along-axis. Along-axis movements are also suggested by ridges and troughs interpreted as folding perpendicular to the deep at 'f' in Figure 6a and by irregular orientations of normal faults east of them.

If the density of the evaporites were to vary greatly spatially, the flows might not necessarily be expected to move in a simple down-slope direction. (A fluid-like body of larger (or smaller) density than its fluid surroundings tends to spread (contract) laterally at the surface.) In Figure 8a, however, the flow structures mostly do lie parallel (strike-slip-like) or perpendicular (normal faults, folds and thrust fronts) to the plan-view direction of maximum seabed gradient. Flow therefore appears to be driven by potential energy gradients oriented with the same directions as the local maximum seabed gradients, so major lateral variations in density seem unlikely. Furthermore, if the underlying basement topography had only recently evolved while the overlying flowage structures preserved a different pre-existing basement configuration, discrepancies

between the gradient directions and flow structures would be expected. The good correspondence then implies that flowage is either more rapid than basement evolution or that basement relief has evolved in a geometrically similar fashion to any pre-existing structure.

Discussion

Style of flowage and the source of movement

Images of 3D seismic reflection data of Fletcher et al. (1995) showing allochthonous evaporites extruded onto the seabed suggest that the evaporites largely flowed *en masse*. Those images show evaporites forming well-rounded bodies as though created by a simple viscous fluid extruding and then spreading at the surface. In contrast, our data suggest a more heterogeneous flowage, with possible strike-slip movement between bodies of evaporite. The movement appears to be relatively shallow, as the vertical reliefs of the flows (200-500 m) are much shallower than the full >1 km vertical extent of evaporite.

Temperature should be expected to affect evaporite flowage. The evaporite surface at the Atlantis II DSDP Sites is estimated to be at 29-41°C (Erickson et al., 1975). If thermal profiles are steady state, the evaporites will be warmer by 10°-12.5°K at 500 m and 80°-100°K at 4000 m depth below their surface (using measured thermal gradients and evaporite conductivities (Girdler et al., 1974; Wheildon et al., 1974)). Strain rates of solution-transfer creep (Spiers et al., 1990) should consequently increase by a factor of four from 500 to 4000 m. Weakening of the evaporite base might be expected to promote

deep-seated movement, in contrast with the more superficial movement inferred from our data.

The origin of variable and shallow flowage appears to arise at large scale from the relief of basement in partially blocking the movement; evaporites are flowing as though through open sluices (Figure 7). The origins of finer-scale structures are more difficult to ascertain from our data. Whereas some stream-wise fabrics of ice and volcanic flows can be traced to underlying obstructions [Keszthelyi, 2004 #7417; Gudmundsson, 1998 #7539], we have not been able to trace the evaporite flow-wise fabrics to such originating sites with these new data. Alternatively, the original geometry of halite deposits affects the geometry of flowage (Schlöder and Urai, 2007). As even small amounts of water greatly weaken halite, by enabling solution transfer creep (Spiers et al., 1990; Urai et al., 1986) and enhancing grain boundary sliding (Watanabe and Peach, 2002), a lack of *en masse* deep seated movement is potentially also explainable if the buried halite has compacted and expelled much of its interstitial water. Varied shallow penetration of seawater along surface fractures could then lead to spatial heterogeneity of flowage.

If the medium-coarse grain size of halite in the DSDP cores (Stoffers and Kühn, 1974) continues to depth, deformation should be dominated by grain boundary sliding rather than solution-transfer creep, which is effective with small grain sizes where diffusion path lengths are short (Urai et al., 1986). In contrast, Wenkert (1979) argued that solution-transfer creep may be needed to explain rapid movements of namakiers, which were difficult to explain in terms of other deformation mechanisms. According to Schlöder and Urai (2007), the fine grain sizes of mylonites formed within namakiers allow efficient solution transfer creep and, from theoretical rheological relationships

(Carter and Hansen, 1983; Urai et al., 1986), enable namakiers to deform 1-2 orders of magnitude faster than coarser domal halite. Although we lack *in situ* information, we suspect that the submarine flows probably also develop mylonites. The development of mylonites by nucleation of new grains and dissection by grain boundary migration (Schlder and Urai, 2007) or subgrain rotation and intergranular cracking (Talbot, 1981) may progressively weaken halite locally.

Compaction reduces porosity in halite pans to ~0% within only 45 m depth of burial (Casas and Lowenstein, 1989), but shale buried with the deposits may not so easily expel its pore water and compact if intercalated halite has low permeability. Based on laboratory experiments on halite with <0.5 wt% porosity (Peach, 1991), permeability at 800 m burial pressures should be only 10^{-21} m^2 . Those waters could then be released later as the shales deform, weakening the intercalated halite. Alternatively, if pore waters become over-pressured, slippage may occur within the shale itself.

The laboratory permeability data can be used to estimate the rate of shale compaction. Assuming for the sake of argument that the evaporites contain 10% shale (Figure 2) comprising 20% water after 200-300 m burial based on a Site 227 sample (Whitmarsh et al., 1974), 1000 m of evaporites would then originally have contained $20 \text{ m}^3/\text{m}^2$ of water. If low permeabilities caused pore pressures to adopt the vertical stress of solid over-burden (a conservative assumption as it leads to under-prediction of compaction time-scale), the pore pressure gradient ($\partial P/\partial z$) above hydrostatic available to drive vertical flow of pore fluids could reach 10^4 Pa/m . From Darcy's Law, the water volumetric flux (m^3/s) is

$$Q = -(\kappa/\mu)A\partial P/\partial z \quad (1)$$

where κ is permeability (m^2), μ is the interstitial fluid dynamic viscosity (Pa.s) and A is cross-sectional area of flow (m^2). From the values above, with $\mu=10^{-3}$ Pa.s, Q/A is calculated to be 10^{-14} m/s. Such a flux would drain 20 m of water over 60 m.y., an order of magnitude longer than the 5 Ma age of these deposits. The lower evaporites seem therefore unlikely to have fully dried, so there is a potential for pore pressures to affect deformation, either within the shales or by influencing deformation within adjacent halites. The exact origin of the deformation is thus uncertain.

Spatial pattern of flowage and seabed gradients

Whereas the flow morphology is largely as expected from the gross topography if gradient significantly affects flowage, extension of the carapace or dissolution might be expected where the seabed steepens, but steepening areas lack such evidence (e.g., on 'b' in Figure 8a). A simple flow model (Figure 8b, lower-left inset) can be employed to explore this further. In the model, a halite layer deforms viscously in a laminar fashion and without basal slippage, while carrying passively an overlying carapace of hemipelagic sediment. Although we strongly suspect the flows do not have this simple structure uniformly over the area, comparing the model predictions with the morphologic data is useful for highlighting how flowage is anomalous. This structure is similar to that of laminar flow of a viscous fluid lying on a slope, differing only by the presence of the passive hemipelagic layer. Analytical expressions for the velocity structure of such flows (e.g., Middleton and Wilcock, 1994) can be readily adapted to show that the down-slope specific transport flux of material is linearly proportional to gradient if the fluid rheology is linear.

If all properties of the mobile evaporite (density, thickness, viscosity...) except gradient were uniform over the whole map, the pattern of vectors in Figure 8a would then also represent the pattern of specific transport flux. From continuity, flowage converging where the terrain is concave upwards (gradient decreasing down-slope) and diverging where the terrain is convex upwards would lead to the evaporites accumulating and thinning, respectively (Mitchell and Huthnance, 2007). Figure 8b therefore shows $\nabla^2 H$ (second derivative of elevation H) computed from the multibeam data. In many areas where the flows are convex-upwards (white in Figure 8b), the surface is not strongly disrupted by normal faults or eroded (Figure 8a). This can be observed, for example, where the flows abruptly steepen crossing rift-parallel topography at 'b' in Figure 8a and other downward-curved areas around the deep's margins. The lack of extension would be even more surprising if the rheology were non-linear (Talbot and Jarvis, 1984), as the material flux would increase more rapidly with gradient, or if movements progressively increase mobility, such as reducing halite grain size (Schlöder and Urai, 2007) or progressive release of pore waters.

Run-out of the flows could potentially explain this discrepancy as thinning of the flows on slopes could compensate for increasing gradient, leading to more modest relative movement between steep and flat areas. Thinning of at least the hemipelagic layer is supported by the seismic data showing reflectors tapering out towards the seabed down-slope (Figure 5). If this explanation were not correct, we would need to appeal to mechanisms that instead stiffen the evaporites with progressive strain, which seem unlikely (Schlöder and Urai, 2007).

Time-scales of flowage

Iranian namakiers move at annually averaged rates of $\sim 1\text{--}8$ m/yr (Talbot and Jarvis, 1984; Talbot et al., 2000; Talbot and Rogers, 1980). As their movements occur primarily during two months of the rainy season (Talbot and Jarvis, 1984), continuous submergence of submarine flows of similar thickness, gradient and composition might be expected to lead to annualised velocities at least six times faster. On the other hand, buoyancy in water reduces the excess density (of evaporites and their carapace with ambient water) driving flow to only ~ 1 g/cm³, which is half that in air, and the namakiers are generally steeper, so these factors partly compensate. For example, the northerly namakier of Kuh E Namak, which has a ~ 1 m/yr annual velocity (Talbot and Rogers, 1980), occurs on a ~ 0.2 m/m gradient, which is double the typical submarine gradients in Figure 8a. Nevertheless, given their continual submergence and their greater vertical thicknesses than only 50–100 m of the Kuh E Namak (Talbot and Jarvis, 1984), these data imply that 1 m/yr could underestimate velocities of the submarine flows, in which case they would traverse ~ 10 km of the walls of Thetis Deep in less than only 10 ky.

A 10 ky time-scale is much briefer than either the Miocene age of the evaporites or the 0.7 Ma age of volcanism implied by the magnetic anomalies. The discrepancy might be explainable if the flows have been observed at a critical moment, for example if flowage requires a threshold shear stress or strain to initiate rapid movement (Schlder and Urai, 2007). Arguing against such a fortuitous association, however, is the evidence of flowage at other deeps such as Vema, Nereus and Atlantis II (Bicknell et al., 1986; Whitmarsh et al., 1974; Girdler and Whitmarsh, 1974; Pautot, 1983). If the Thetis Deep flows do travel at >1 m/yr, they may therefore be replenished by flow from a broader area

of the Red Sea and be destroyed repeatedly within the floor of the deep by dissolution and incorporation of insoluble components of evaporite and carapace into the oceanic crust by axial lavas repeatedly erupting over them. Alternatively, flowage may be less rapid, for example, if halite is a less abundant phase than suggested.

Early stages of opening of the Atlantic and implications for the ocean-continent boundary

Without high quality seismic data and stratigraphic information, the possibility of early flowage is difficult to evaluate, but some features of the Atlantic deposits are broadly comparable with those of the central Red Sea, which suggest that useful analogies may be drawn. For example, the South Atlantic Aptian evaporites have been suggested to have also originally precipitated out to >1 km stratigraphic thickness locally (Davison, 1999; Meyers et al., 1996a). Some Atlantic evaporites were precipitated during the syn-rift phase (Davison, 1999; Karner et al., 1997) and can be observed in seismic reflection images displaced by normal faults (Davison, 2007). Lateral unloading by exposure along fault escarpments may have allowed flowage to occur locally. However, the evaporite base is more commonly subdued (Cobbold and Szatmari, 1991; Contrucci et al., 2004; Davison, 1999; Demercian et al., 1993; Duval et al., 1992; Emery and Uchupi, 1984; Karner et al., 1997; Meyers et al., 1996b; Moulin et al., 2005) so we do not expect a fault exposure mechanism to have initiated flowage widely.

Some researchers have suggested that, in contrast to the central Red Sea, the South Atlantic Aptian evaporites formed when seafloor spreading had already developed and some precipitated directly over oceanic or proto-oceanic crust (Davison, 2007; Jackson et al., 2000). The oceanic or proto-oceanic basement in many areas near to

Aptian evaporites is subdued (Contrucci et al., 2004; Davison, 2007; Moulin et al., 2005; Wilson et al., 2003) consistent with either widespread lava flows of proto-oceanic crust or with low relief oceanic crust created at moderate-fast spreading rates (Malinverno, 1991; Small, 1994). Jackson et al. (2000) speculated that seaward-dipping reflectors underlie the South Atlantic evaporites but are unreported because they are poorly imaged. If correct, the lava flows generating the reflectors and their overlying surface would originally have dipped landward or horizontally (Mutter et al., 1982), suggesting that there would have been little or no potential energy gradient available to cause evaporites precipitated on them to flow seaward. On the other hand, ridges generating relatively smooth oceanic crust can later develop axial valleys (e.g., Klitgord and Mudie, 1974), so flowage may not be ruled out entirely for such areas, depending on how much thickness the evaporites accumulated compared with any subsequent axial valley relief.

Information from combined seismic refraction and reflection surveys can potentially help to address some of these issues. In particular, a dense array of ocean bottom instruments deployed along a line of 26 stations off Angola was used to collect wide-angle data, which allowed velocities of the oceanic crust and adjacent structures to be well-imaged (Contrucci et al., 2004, their Figure 6; Moulin et al., 2005). The resulting velocity model suggests an interesting configuration of evaporites lying over extended continental crust as far as a basement ridge at the ocean-continent transition (OCT). Beyond the OCT, oceanic crust is strongly heterogeneous, varying in seismic velocity by >1 km/s over 10-20 km. Although such variability might have arisen from normal processes associated with slow-spreading ridges (Contrucci et al., 2004), the new data collected at Thetis Deep suggest to us that flowage into the young spreading centre could

potentially have contributed heterogeneity if lavas erupted over the evaporites allowed their insoluble residues to have become variably incorporated into the oceanic crust. A sedimentary contamination occurs today in spreading centres close to continents, such as the Juan da Fuca Ridge by turbidites (Davis and Lister, 1977) originating from sediments supplied by northwest American rivers, but rivers are deflected away from the Red Sea by the rift topography (Steckler and Omar, 1994). Contamination can therefore potentially occur by evaporite flowage without large terrigenous input. Further geophysical datasets combined with drilling are needed to quantify the extent to which oceanic crust is modified immediately seaward of the OCT in such areas.

Conclusions

Flow-like features identified on the seafloor of Thetis Deep are interpreted as caused by flowage of evaporites based on their 200-500 m vertical reliefs (thicker than the superficial hemipelagic layer) and the presence of tectonically deformed evaporites in seismic reflection data from the flanks of the deep. From the seismic data, some flow-wise ridges and valleys can be attributed to strike-slip movements between bodies of evaporites that have moved independently, whereas others might have arisen from sub-flow topography. At large scale, the evaporites flow in separate corridors rather than *en masse* because of variably obstructing basement. Where the flows steepen over rift fault escarpments, flowage might be expected to speed up, causing extension, but the sonar data do not reveal evidence for any dissolution or extension in such areas. We speculate that the flows run-out on steep gradients so that thinning of the deforming layers compensates for the effect of gradient on flow speed.

Existing seismic reflection data from the South Atlantic suggest that flowage may have been initiated locally there by exposure along active normal faults but more commonly the evaporites lie at the boundary between the syn-rift and post-rift, and possibly on proto-oceanic crust (Davison, 2007; Jackson et al., 2000), where topographic gradients driving flow were probably limited. Where evaporites flow into the floors of volcanically active deeps such as Thetis, eruption of lavas over the evaporites can potentially incorporate their insoluble residues into oceanic crust. Data are too limited to address whether this has occurred more generally in evaporite-hosting oceanic basins, but one particularly dense array of ocean bottom stations deployed off Angola has revealed an interesting heterogeneity in oceanic crustal velocities of >1 km/s over 10-20 km (Contrucci et al., 2004) which we speculate may have partly arisen from flowage into the early spreading centre.

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Figures

Figure 1. Location of the Thetis Deep study area in the central Red Sea (rectangle locates the multibeam sonar data of Figure 6). Large black-outlined circles locate DSDP Sites 225 and 227 adjacent to Atlantis-II Deep and Site 228 farther south. Grey levels range below sea-level from 2774 m depth (black) to 0 m (white) and above sea-level from 0 m (black) to 1500 m altitude and above (white). White lines around Thetis Deep locate seismic refraction experiments (Tramontini and Davies, 1969) summarised in Figure 4, with solid white circles marking the sonobuoy locations. To create the cross-section in Figure 4, these sonobuoy locations were projected along-rift onto the line marked by the arrow marked "Fig. 4".

Figure 2. Stratigraphy at DSDP Sites 225 and 227 located in Figure 1 based on Figure 2 of Stoffers and Kühn (1974) and the site reports (Whitmarsh et al., 1974). Grey represents undifferentiated sediments and sedimentary rocks. White and black fill represents anhydrite and halite, respectively (intervening grey represents shale).

Numbers to left of each column are the units of the DSDP site reports. Vertical bars to the right mark intervals of deformation structures in cores attributed to halokinetics (Girdler and Whitmarsh, 1974). "S" marks the expected location of the S reflector in seismic data.

Figure 3. Laboratory measurements of bulk density (Manheim et al., 1974). Vertical grey bars show the range of 8 density measurements on halite samples recovered from Site 227 (Wheildon et al., 1974). Horizontal dotted lines mark the top of the evaporites, corresponding to the S reflector.

Figure 4. Results of seismic refraction experiments (Tramontini and Davies, 1969) along the white lines in Figure 1. The depths of an interpreted refractor are represented by solid squares, plotted at the sonobuoy locations (dashed lines connect alternate values derived with different but equally valid velocities). Sonobuoy locations have been projected N150°E onto the line oriented N060°E from 22°N, 38°E marked by the arrow in Figure 1 (i.e., "cross-axis distance" is parallel to this line), apart from the inter-trough values shown on the left. Seafloor depths at the buoys are from (black crosses) the multibeam bathymetry in Figure 2 or (grey crosses) bathymetry derived from satellite altimetry data and echo-soundings (Smith and Sandwell, 1997). Star symbols represent the mean bathymetry of each line derived from seabed reflections in the experiment.

Figure 5. Examples of seismic reflection lines located on Figure 6a. Top five panels are multichannel seismic data collected on *R/V Urania* in 2003 and the lower panel is single channel seismic data collected on *RRS Discovery* in 1979 (Searle, 1980). Multichannel

data are plotted at a common vertical scale. Seismic two-way times are given to the sides of panels. Line 24M runs parallel to the rift connecting lines 23M and 25M to the east of Figure 6, and is located in Figure S1. "S" is the interpreted S-reflector corresponding to the top of the evaporites.

Figure 6. (a) Shaded-relief image of the multibeam data (artificial sun from the upper right of map). Lines marked x-x' to z-z' locate the longitudinal sections in Figure 9.

White dashed lines marked 21M, 23M, 25M and 26M and white dotted line marked "Discovery 103" locate the seismic reflection profiles in Figure 5. Annotation '2' refers to the second step away from the deep referred to in the text. (b) Color-coded bathymetry contoured at 100 m intervals, with 500 m contours in bold.

Figure 7. Three-dimensional image of the multibeam bathymetry data from the southeast looking northwest over the east side of Thetis Deep. Annotation 'a' and 'b' corresponds to annotation in Figure 6a. Vertical exaggeration is 4:1.

Figure 8. (a) Shaded-relief image of the multibeam data overlain by vectors representing the direction and magnitude of seabed gradient calculated over a 3-km lengthscale. (b) Curvature of the bathymetry derived by calculating the Laplacian, $\nabla^2 H$ of elevation H over a 3-km lengthscale. Positive curvature (black) represents concave-up regions. Map coordinates are Universal Transverse Mercator distances in km (zone 37). Lower inset shows a reference model used in the text to discuss the pattern of surface deformation (plus symbols represent the model deforming evaporite layer).

Figure 9. Longitudinal sections down flows 'a' and 'b' ($x-x'$, $y-y'$ and $z-z'$) in Figure 6a.

Vertical exaggeration is 3.23:1. Annotation 1, 2 and 3 refer to the numbering of steps described in the text. Top profile shows surface texture in the last 3 km of profile $y-y'$ at 10:1 vertical exaggeration.