PERGAMON

Journal of Structural Geology 23 (2001) 521–530

JOURNAL OF STRUCTURAL GEOLOGY

www.elsevier.nl/locate/jstrugeo

Neotectonics at the Arabian plate margins

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Received 12 June 2000; accepted 4 July 2000

Abstract

Opening of the Red Sea is accompanied by convergence between the Arabian plate and Eurasia. Regional topography and structure favour gravity glide as the main driving force of plate translation. At the leading edge of the plate, the Zagros Mountains undergo coseismic serial folding which is equivalent to Holocene shortening by ~20 mm/year and which has led to major episodes of coastal uplift of which the last was ~1700 years BP. At the Jordan Rift transform, which bounds the Arabian plate on the west, a recurrence interval of ~1600 years is reported for events of $M_L \geq 5.5$. The palaeomagnetic record for the last 3.2 Ma indicates an average spreading rate for the Red Sea of ~20 mm/year; there is some evidence that hydrothermal activity in the Red Sea is pulsatory, with a period of ~2000 year, and that it reflects discontinuous spreading. The Holocene neotectonic records of the Zagros, the Jordan Rift and the Red Sea are the product of complex plate interactions and of the accumulation and release of strain in the crust along the plate margins. But they also reflect elastic strain energy storage and release within the Arabian plate, whence parallels in the period of major deformation episodes in the three deforming zones and the apparent discrepancy between the seismic moment predicted by plate kinematics and that recorded in the Zagros. Any associated intraplate deformation, if detected geodetically, would thus help the assessment of seismic hazard.

1. Introduction

It has long been observed (e.g. Falcon, 1974) that the extent of shortening in the Zagros Range, at the leading edge of the Arabian plate, is about the same as the separation between Arabia and Africa during development of the Red Sea axial trough (Fig. 1). Similarly, the amount of northeastward movement of the Arabian plate is usually linked to displacement along the Dead Sea transform (e.g. Searle, 1994). This paper explores the pattern of Holocene deformation at the three plate boundaries in an attempt to establish whether they are in any way mechanically coupled by the Arabian plate.

Even if allowance is made for a short and selective instrumental record, the distribution and level of present-day seismicity in Iran and Turkey (Fig. 2) can not be explained simply by crustal generation in the Red Sea (McKenzie et al., 1970). The kinematic picture is complicated by the multiplicity of plates that account for displacement at any plate boundary: the evolution of the Red Sea, for example, has been influenced by the motion of the Indian Ocean and of Africa west of the Rift (Somalian sub-plate) as well as Arabia and the Nubian sub-plate. In any case, seismicity at plate boundaries is strongly influenced by the character of the individual faults and may vary substantially from one fault segment to another.

Yet theoretical considerations favour transplate–margin interaction. Bott and Dean (1973) showed that, although instantaneous strain energy release at a compressive plate boundary is periodic, it is felt only close to the boundary and the plate interior acts purely as a reservoir of strain energy; but they also demonstrated that the sudden application or release of stress can diffuse across a representative plate 1000 km wide in 10^3–10^4 years.

The effect is likely to be primarily one of timing, possibly through triggering or the inhibition of imminent earthquake events. Deformation on the shores of the Persian Gulf and the Red Sea and along the Dead Sea rift is sufficiently active for a preliminary assessment of this proposal for the last 6000 years. The time period is long enough to combine the instrumental seismic record with field evidence for folding and faulting. This record lacks the resolution of short-term geodetic studies, although problems arise when data at conflicting timescales and of varying quality are compared. But the attempt seems justified by the need to counter hazardous seismicity as well as the opportunity it provides to test some of the premises of the plate model.

2. The Arabian Plate

The Arabian plate extends from the Red Sea to the Zagros

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PII: S0191-8141(00)00117-6

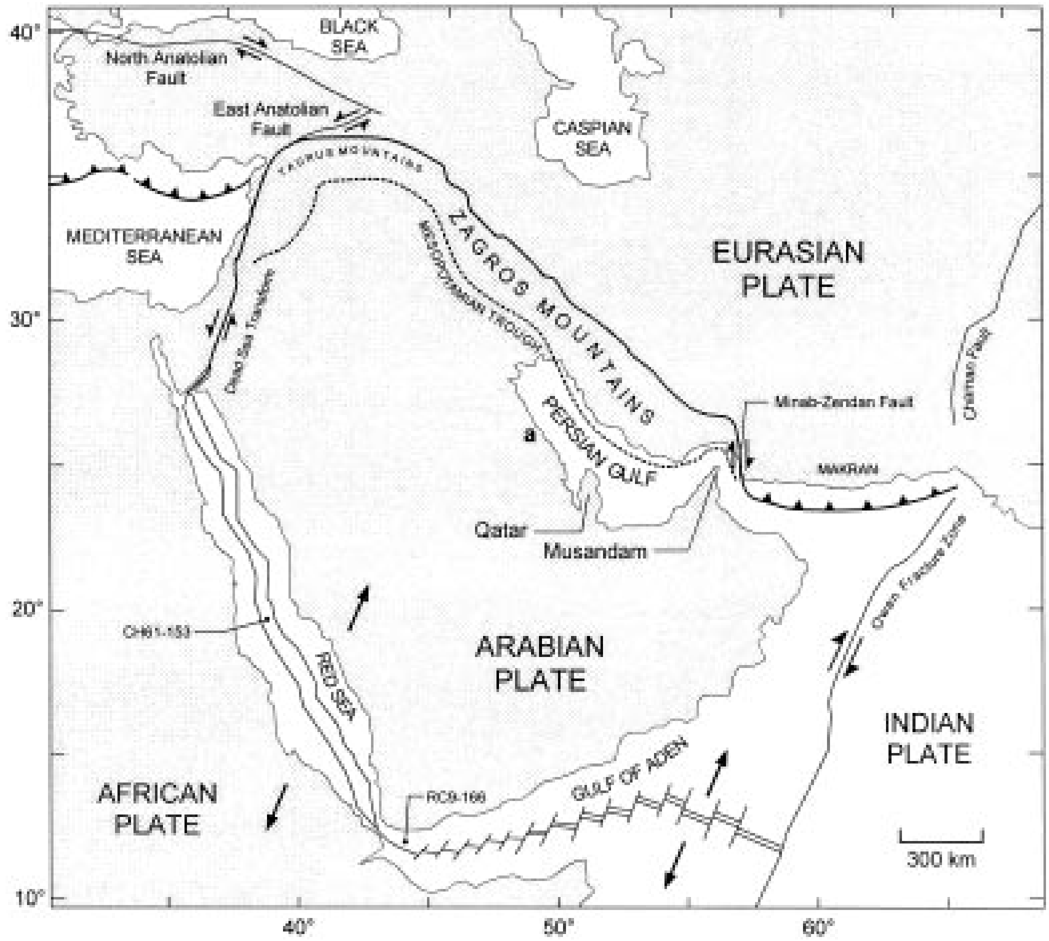

Fig. 1. Location map showing tectonic setting of the Arabian plate. **a** indicates part of coastal Saudi Arabia discussed in the text.

and from the Gulf of Aden to the Taurus Mountains. It consists of Precambrian basement overlain by Tertiary and later volcanics in the west and marine and continental sediments in the east (Reches and Schubert, 1987). Its interior lacks any trace of intense late Tertiary deformation (Al Kadhi and Hancock, 1980). The Mesozoic–Cainozoic sediments in the eastern part of the plate are horizontal or slightly tilted, and they display subdued Late Cretaceous–Eocene structures (Hancock et al., 1984). The seismic record for the interior (Fig. 2) shows no large earthquakes (Reches and Schubert, 1987) although medium magnitude earthquakes have occurred fairly continuously during the last 1200 years (Ambraseys and Melville, 1983; 1994).

Reches and Schubert (1987) modelled the Arabian plate as lithosphere 120 km thick. In order to obtain the extensional and compressive forces required to displace it they had to supplement lateral density variations in the lithosphere with a basal shear stress towards the Zagros. An alternative which does away with hypothetical basal shear is some version of gravity spreading from the Red Sea Rift, which has been shown to be plausible by Bird (1978). Indeed, his finite element modelling suggests that the continental crust may be mechanically decoupled from the stronger mantle lithosphere. Gravity glide (Price et al., 1988) can provide well in excess of the force required to sustain the Zagros folds (Mann and Vita-Finzi, 1988).

There is prima facie evidence for the requisite gravitational gradient in the seismic refraction profile of Mooney et al. (1985), which shows a northeastward deepening of the base of the crust (as postulated by Bird in 1978) abruptly from 18 km beneath the coastal plain to 38 km at the Red Sea escarpment and then gradually down to 45 ± 5 km under the Zagros. Using the same data, Prodehl (1985) shows the Moho descending more gradually to a depth of 45 km, which is, however, attained a mere 300 km east of the Red Sea axis. The depths for the Moho proposed by Snyder and Barazangi (1986) are 40 km beneath Mesopotamia and the Persian Gulf and 65 km beneath the Main Zagros Thrust at the rear of the range.

What happens to the leading edge of the plate is uncertain. Any assumption that convergence implied subduction was challenged by the absence of earthquakes deeper than 20 km in the expected area (Niazi et al., 1978). Two decades of seismic observation as well as finite element modelling (Bird, 1978) have extended this verdict to the Zagros Mountains as a whole. Falcon (1969) inferred

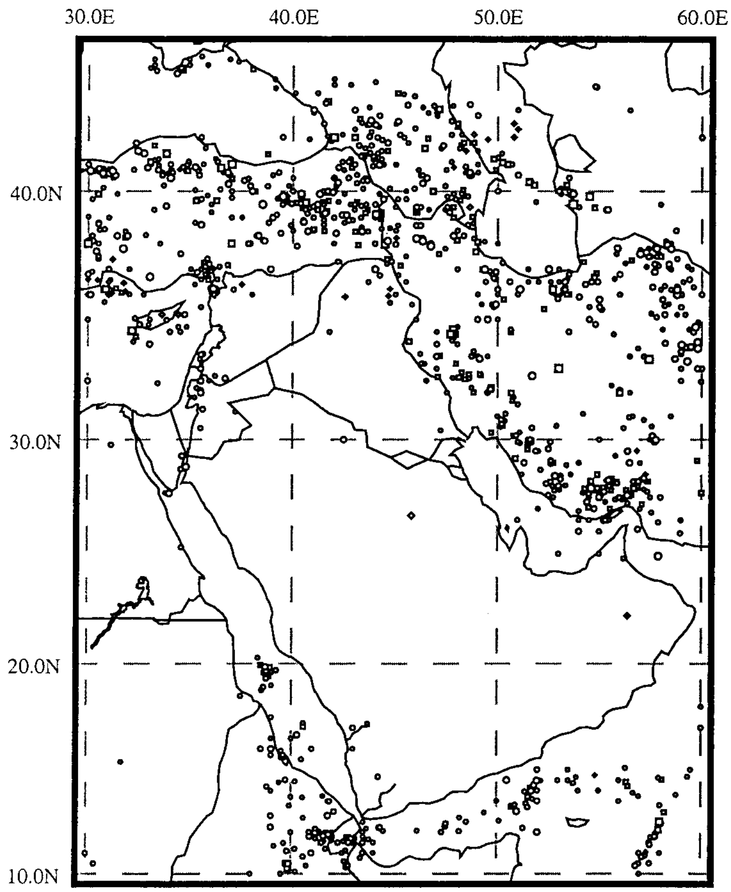

Fig. 2. Seismicity of the region ($M \geqslant 5$) during AD 10 to 1999 (courtesy of NDGC/NOAA).

from the predominance of shallow earthquakes between 53° and 50°E that the crust of the Arabian shield is moving northeast "over the mantle and under-riding central Iran … intermittently since Triassic times". The present consensus is that the basement is taking up convergence mainly by thickening associated with the reactivation of normal faults as reverse faults (Falcon, 1969; Jackson, 1980; Ni and Barazangi, 1986; Berberian, 1995); detachment between the sedimentary cover and the faulted basement is promoted by salt. Deformation of the Palmyride fold belt in Syria, at the rear of the Bitlis suture, is closely analogous (Searle, 1994; Alsdorf et al., 1995).

3. The northeastern margin

The Mesopotamian depression and its south-eastern prolongation, the Persian Gulf, are widely considered to occupy a downwarp in the Arabian plate which resulted from its tectonic loading by the Zagros Mountains (Uchupi et al., 1999). An earlier attempt to evaluate deformation during the Holocene (Vita-Finzi, 1982) embodied eustatic data derived from other parts of the world in order to highlight local tectonic effects. Using glaciohydroisostatic criteria, Lambeck (1996) modelled the position of successive shorelines in the Gulf since the last glacial maximum (~18 000 BP). In his view, the sea in the eastern part of the Gulf had risen to its present level by about 6000 years BP and in the western Gulf had exceeded it by 2–3 m before falling to its current position.

Shoreline chronologies for parts of coastal Oman, Qatar and eastern Saudi Arabia point to submergence of the northeast coast of Arabia 6000–3500 years ago by ~1.5 m (Vita-Finzi, 1982; McClure and Vita-Finzi, 1991) and thus, to judge from Lambeck's model, indicate little tectonic disturbance during the Holocene. In contrast, the Musandam Peninsula betrays clear subsidence whose extent can be estimated from the depth offshore of a series of cemented valley fills thought to date from ~35 000 to 10 000 years BP (Vita-Finzi 1982). The resulting amount is a minimum,

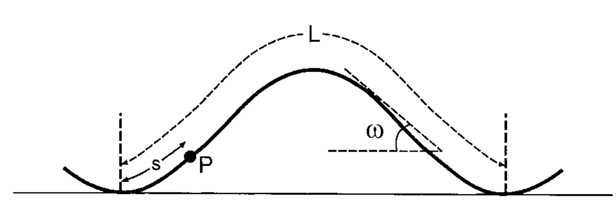

Fig. 3. Terms used to calculate fold shortening ΔX from uplift measurement Δy.

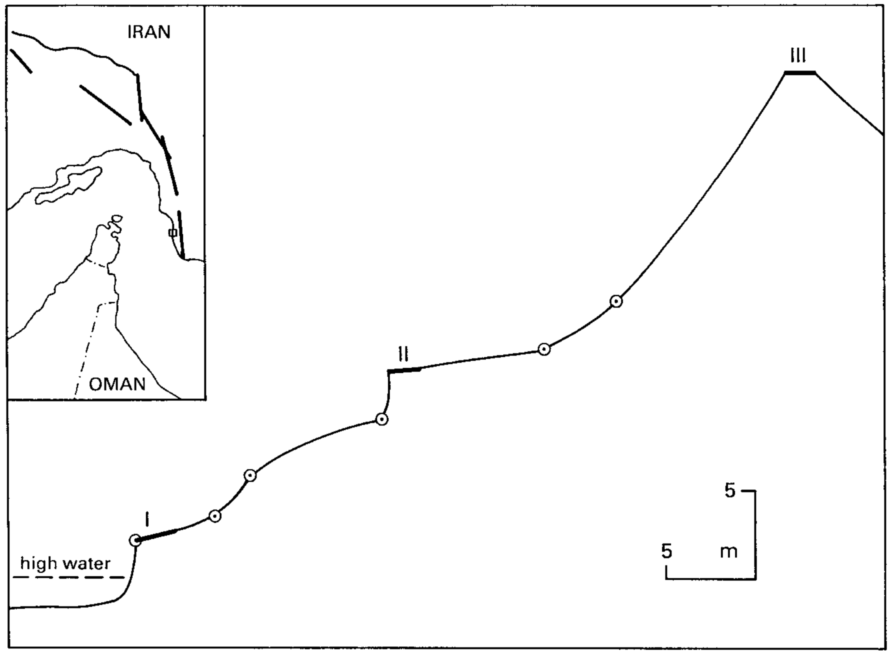

Fig. 4. Profile and location of Section I, SW Zagros.

as the fills may well extend beyond the depths recorded by the survey. The deepest example, off Ghubbat ash Shabus on the eastern coast of the peninsula, lie at −115 m. Allowing for the subsequent rise in sea level of 40m (Lambeck, 1996) the average minimum rate of subsidence is 7.5 mm/year rather than 8.5 mm/year as previously proposed (Vita-Finzi, 1982).

Some of the more sheltered inlets in the peninsula display horizontal intertidal undercuts up to 5 m deep, the combined effect of boring by marine organisms (especially the bivalve *Lithophaga cumingiana* Dunker and *L. obesa* Philippi) and wave action. A jetty built of the local limestone in 1943 has been undercut at a rate of 0.25 cm/year, which implies that the deepest undercuts have taken ~2000 years to form (Vita-Finzi and Cornelius, 1973) and that relative sea-level has changed little during this period, although a tidal range of as much as 2.8 m at the Straits (Defant, 1961) makes it difficult to decide between a pronounced reduction and a complete halt in the rate of subsidence.

There is an analogous hiatus in uplift across the Straits of Hormuz on the Zagros coast, where discontinuous uplift has long been inferred from the presence of raised marine terraces and river terraces (e.g. Falcon 1975). On the hypothesis that folding of the Zagros is serial and that the youngest and fastest growing fold is to be found on or just off the coast, an attempt was made (Vita-Finzi, 1979) to compute the shortening represented by Holocene uplift on coastal folds assuming decollement and sinusoidal, buckle fold geometry (Sattarzadeh et al., 2000). Following Mann and Vita-Finzi (1982),

$$\Delta y = \Delta X * 2\pi/\omega * (s/L)^2 \quad (1)$$

where Δy is the vertical displacement undergone by the dated palaeoshore at a point P on an anticline, ΔX the corresponding shortening, ω the maximum dip of the fold limb in radians, L the total fold length, and s the distance from P to the nearest trough (s/L being small) (Fig. 3). Data were obtained for two transects across a fold running roughly E–W (Fig. 4) and one for a transect across an anticline running roughly N–S in order to secure a resultant roughly parallel to the shortening azimuth ~NE. The calculation was based on uncalibrated ^{14}C ages and the global sea-level curves of Flint (1971). The recalculated ages (Table 1) were corrected for sea level after Lambeck (1996) and thus are similar to the values previously obtained using the hydroisostatic model of Clark et al. (1978).

Comparison of the results given in Table 1 with the convergence rate of 30 mm/year on an azimuth of 015° obtained by Jackson et al. (1995) suggests that, in accordance with the concept of serial folding (Price, 1975; Shearman, 1976), shortening is concentrated on the frontal fold of the sequence. This is not to ignore the evidence for

Table 1
Holocene deformation rates (mm/year) at Straits of Hormuz. Revised after Vita-Finzi (1982). Δy is the vertical displacement undergone by a dated shoreline, while Δx is the corresponding shortening. Age calibration after Stuiver and Reimer (1993), sea-level correction after Lambeck (1996)

	Least squares		Average	
Section	Δy	ΔX	Δy	ΔX
1 (W–E)	5	17.6	4.8	16.9
2 & 3 (S–N)	1	9.7	0.8	7.8
Resultant (azimuth)		20.1 (061°)		18.6 (065°)

Table 2
^{14}C terrace ages at Section 1, Straits of Hormuz. Revised after Vita-Finzi (1979)

Terrace	Age (cal. year BP)	Elev. (m)	Corr. elev. (m)
III	6715 ± 440	28.6	31.5
II	2410 ± 240	11.8	10.9
I	1675 ± 260	2	1.4

distributed deformation well inland indicated among other things by widespread seismicity (Berberian, 1995) but to emphasize that the locus of maximum folding will shift towards the deformation front owing to progressive rotation of the fold limbs (cf. Hardy and Poblet, 1994) and of associated reverse faults in the basement.

Perhaps more informative than an average rate, especially in highly seismic areas, is the tempo of deformation. The data for all the Arabian and Iranian sections discussed here are derived from palaeoshores, which indicate either a genuine pause in the relative movement of the sea against the land or one brought about by a coincidence in the displacement of both land and sea sufficiently prolonged for a recognisable shoreline to develop (Lajoie, 1986). The only sequence with sufficient resolution to make the analysis possible is Section 1 on the Zagros coast, but even here only three terraces have hitherto supplied ^{14}C ages. They are listed in Table 2.

The faunal evidence points to coseismic uplift, as the surface of all three terraces included a large range of intertidal and subtidal species in growth position (Fig. 5; Vita-Finzi, 1986). But it would be unreasonable to explain the sequence solely by three cataclysmic earthquakes resulting

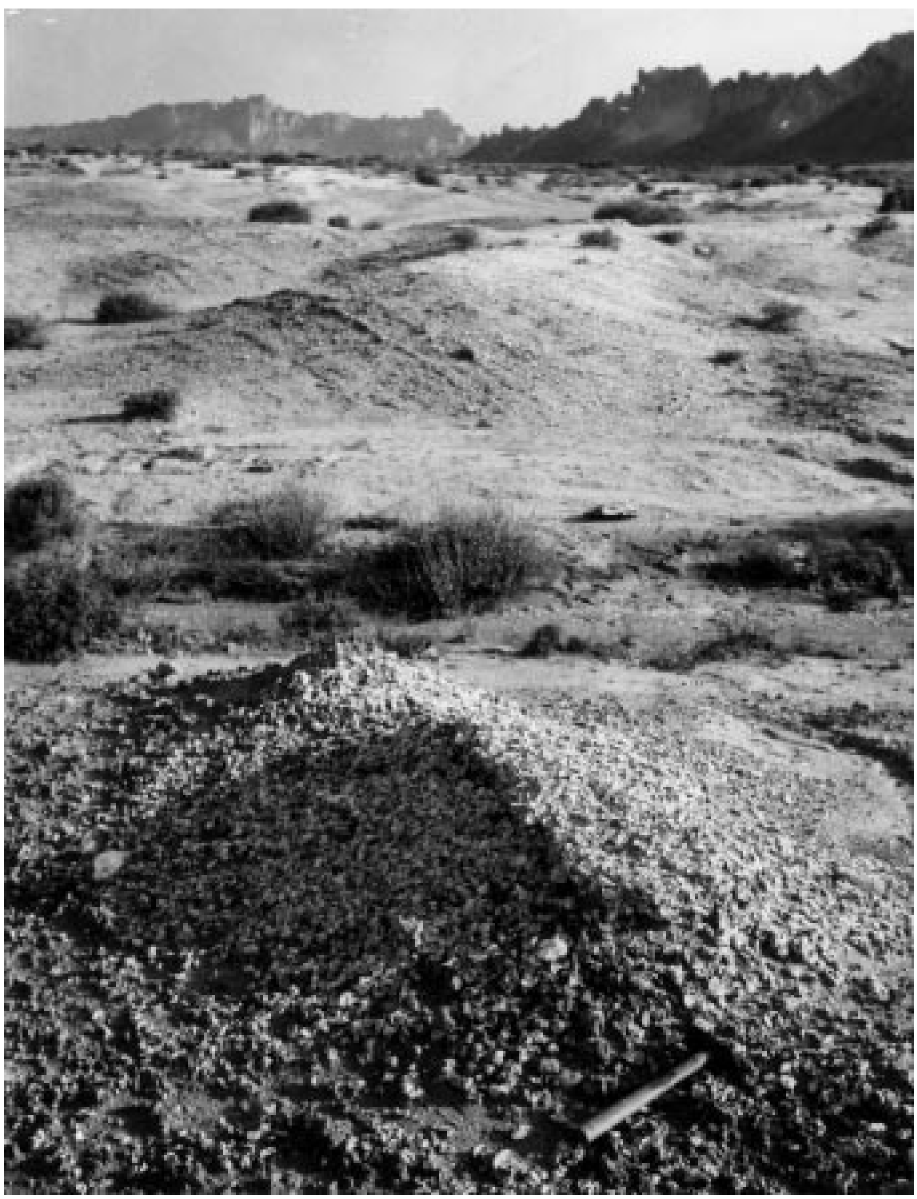

Fig. 5. Surface of Terrace I at Section 1 showing oyster banks indicative of coseismic emergence.

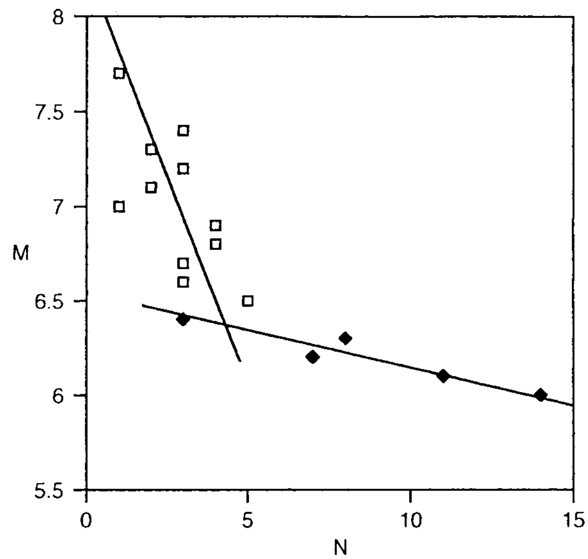

Fig. 6. Magnitude/ frequency plot of Zagros earthquakes 1909–1992 listed by Jackson et al. (1995). Symbols are used only to distinguish the two populations discussed in the text.

in net uplifts of as much as ~9 and ~20 m. The largest earthquake in the region which has resulted in a recorded episode of coastal uplift was on the Makran coast at Pasni (1945.12.27, $M_w = 8.1$), and it led to only ~2 m of emergence (Byrne et al., 1992). Although there are major differences in the tectonic setting of the Makran and of the Zagros, the Pasni evidence suggests that uplift of the lowest terrace at Site 1 represents the latest substantial earthquake to affect the structure.

The earthquake data presented by Jackson et al. (1995) for 1909–1992 have been plotted in Fig. 6 as a magnitude/ frequency graph. The break at $M \geq 6.5$ is doubtless partly an artifact of the record (cf. Yeats et al., 1997); but it is consistent with two preliminary conclusions from the fold data. First, the presence of only three major terraces shows that there have been few occasions when uplift was rapid and substantial enough to lift the intertidal zone clear of wave action, and that uplift had been preceded by the prolonged stillstand required for an extensive platform to develop. Second, granted that earthquakes recorded in the Zagros between 1909 and 1992 account for no more than 10–20% of the deformation predicted by plate tectonics (Jackson et al., 1995), not all the uplift resulting from folding is necessarily coseismic even if many of the Zagros folds are over blind reverse faults in the basement (Berberian, 1995; Sattarzadeh et al., 2000).

4. The Dead Sea transform

The Dead Sea transform exhibits ~107 km of left-lateral displacement since the middle Miocene (Quennell, 1958). Gravity data suggest that the transform can be divided into at least 15 segments measuring 25–55 km (ten Brink et al., 1999). The system is constantly evolving, and some of the segments are currently inactive. On the other hand, simultaneous rupture on a number of segments is required to account for the moment magnitude of the largest historical earthquakes, namely $M_w = 7.3$. ten Brink et al., 1999 argue that these adjustments were in response to small changes in the relative plate motion between Arabia and Africa. The locus of fault activity has likewise migrated in the transpressive segment of the transform in Lebanon (Butler et al., 1997).

Seismic moment calculations on the rift indicate between ~6mm/year and 10 mm/year of slip (Galli, 1999; Khair et al., 2000). The discrepancy with the ~10 mm/year predicted by plate-tectonic models (DeMets et al., 1990) has been ascribed, as elsewhere, to an inadequate instrumental record coupled with temporal clustering of substantial earthquakes and to the contribution of aseismic creep. The latter, according to Ben-Menahem (1981), increases in importance southwards, from zero near the collision front with Eurasia to a maximum of 65% at the Gulf of Eilat. Geodesy can not yet resolve the issue, doubtless in part because its record is too short and in part because movement is revealed as a complex pattern resulting from structural inhomogeneities, the interplay between the fault system and other major regional structures, and irregularities in the geometry of the opposed plate boundaries (Eyal, 1996).

The problem now arises of the relationship between slip, magnitude and frequency. Rotstein and Arieh (1986) report that there was no evidence of fault slip associated with an event of $M = 6.25$ with its epicentre 40 km north of the Dead Sea; at the Crusader castle of Vadum Jacob the AD 1202 earthquake, which had an estimated magnitude of 7.6, produced 1.6 m of offset (Ellenblum et al., 1998). Galli (1999) concludes that in the central part (Jordan Valley Fault) slip is mainly the product of events with $M \approx 6$–7 and in the southern rift (Wadi Araba) $M \approx 6.5$. The corresponding recurrence intervals are 800–1500 and 200–400 years.

The higher of these ranges is favoured by other studies of coseismic deformation in both the southern and the central rift. Amit et al. (1999) report a recurrence interval of 2000–1200 years for $M > 6$ events during the Pleistocene–Holocene of the southern Araba. In the Dead Sea area the laminated deposits of Pleistocene Lake Lisan, which span 50 000 years and include seismites as well as faults, indicate an average recurrence period of 1600 ± 2800 years for earthquakes of $M_L \geq 5.5$, the magnitude required for the development of seismites by fluidization of the marls. The events cluster in periods of about 10 000 years separated by quieter periods of similar duration (Marco et al., 1996). A value of ~1500 years was proposed by Ben-Menahem (1991) for the 'maximum magnitude earthquake' (7.3) for the transform between 33.4° and 29.5°N. Events of $M_s > 7.2$ have been recorded in the Beqa′a and Karasu segments of the northern rift and are separated by quiescent intervals of 450–700 years (Khair et al., 2000), but the extent of associated deformation hereabouts is poorly documented and the tectonic picture is complicated by interaction with the East Anatolian Fault.

In short, the nature and rate of displacement on the Dead Sea transform varies significantly along strike, with creep decreasing in importance northwards in favour of large events. In the Araba and Jordan segments, however, there is some indication that coseismic deformation is associated mainly with events of $M \approx 6.5$ which have a periodicity in the region of 1400–1600 years.

5. The Red Sea

The early stages in the evolution of the Red Sea are commonly dated to the Oligocene (Girdler and Southren, 1987). Two episodes of sea-floor spreading have been recognized. The current phase is generally thought to have begun about 4.5 My ago (Girdler and Styles, 1974) and rifting is propagating northwards (Omar and Steckler, 1995), possibly in punctiform fashion (Bonatti, 1985). The magnetic sea-floor timescale yields a spreading half rate of ~10 mm/year (Makris and Rihm, 1991); the motion is equivalent to convergence at 28°E 46°N on an azimuth of 015° at 30 mm/year (Jackson et al., 1995). A recent recalculation of spreading rates since the middle of geomagnetic chron 2A (3.2 Ma) suggests that the half rate ranges from ~10 mm/year in the north to ~16 mm/year near latitude 18°N (Chu and Gordon, 1998). The Gulf of Aden is thought to be the product of a propagating rift which is spreading westwards at a little less than 3.0 mm/year (Courtillot, 1980).

The accepted spreading rates for the Red Sea as a whole can not be resolved better than the duration of the Brunhes palaeomagnetic interval, namely 690 000 years, but observations from the submersible Pisces XI in the axial zone near 18°N indicated a series of relatively short-lived changes: 20 mm/year for 1.71–0.9 Ma, 30 mm/year for 0.89–0.69 Ma, and 10.4 mm/year for 0.69 Ma to the present day (Juteau et al., 1983). The duration and timing of the episodes of magmatism and faulting on the ridge remain conjectural, and any geodetic measurements that are eventually made across the Red Sea will be of little value for assessing variations in Holocene extension rates.

An attempt was accordingly made to refine the extensional record by reference to the Late Quaternary deposits of the Red Sea floor. The circulation of sea water through hydrothermal systems at mid-ocean ridges influences the ^{18}O composition of the ocean by means of chemical reactions (Muehlenbachs and Clayton, 1976) and to a lesser extent through changes in the pattern of temperature and circulation. Discussion of the process generally focuses on the variations in the intensity of sea-floor spreading at the global scale (e.g. Holland, 1984) and therefore yields time constants measured in 10^6 years. At shorter time scales the ^{18}O signal is probably dominated by glacial–interglacial cyclicity. In a marginal sea with restricted access to the world ocean, such as the Red Sea, local climatic and hydrological conditions may obscure the global signal especially when water exchange is even more limited during periods of low sea level.

The fact remains that local changes in hydrothermal activity can leave an imprint in the deep-sea record; witness the bursts of hydrothermal activity that are thought to be responsible for the isotopic composition of carbonate cement recovered from Sites 716–719 on ODP Leg 116 (Boulègue and Mariotti, 1990). A similar effect has been postulated for Red Sea carbonates older than 10 000 years (Lawrence, 1972) and was suspected of distorting the isotopic composition of foraminiferal tests there (Deuser and Degens, 1969); but isotopic fluctuations in the Red Sea are generally ascribed to a combination of changes in sea level and in climate (Deuser et al., 1976; Almogi-Labin et al., 1991).

Most of the published palaeoclimatic data for the Red Sea lacks the resolution to reveal any such short-term pulses of activity. An exception is Core CH61-153, midway up the Red Sea (Fig. 1), which displays oscillations of ~1.5 per mil

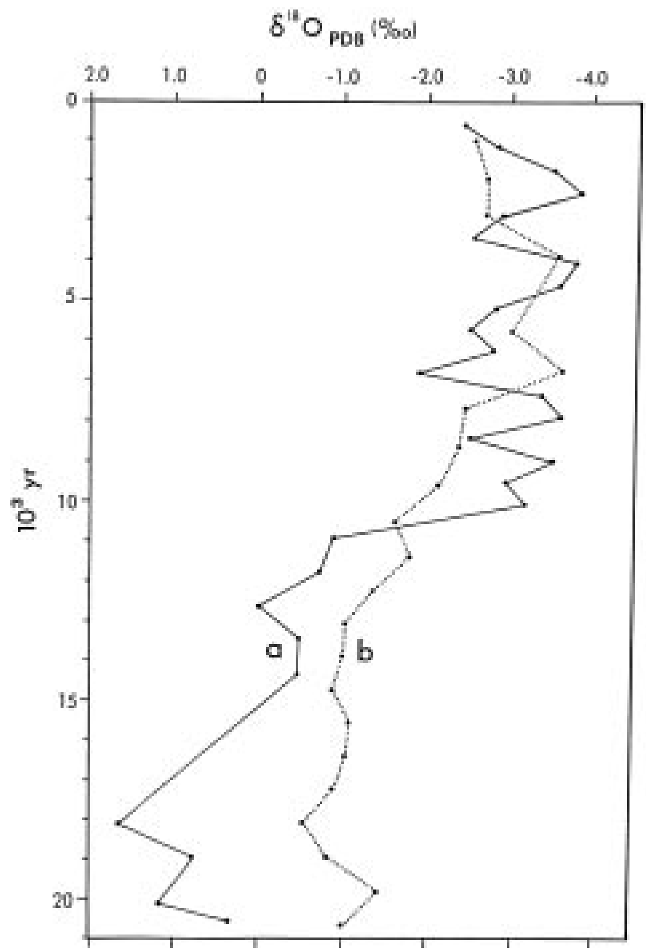

Fig. 7. $\delta^{18}O$ record of cores CH61 and RC9-166 (after Thunell et al. 1988). For location see Fig. 1.

in the ^{18}O composition of the planktonic foraminifer *Globigerinoides ruber* during the last 10 000 years with an average period of ~2000 years (Thunell et al., 1988). The periodicity is absent from the record for the Gulf of Aden during the corresponding period (Core RC9-166, Fig. 7). It has no obvious eustatic or climatic explanation and may reflect episodes of accelerated hydrothermal activity in the area represented by the core record.

6. Discussion

The Holocene deformation reported above at three margins of the Arabian plate is compatible with uniform plate translation: deformation at 10^2 years time scales can be representative of motions over 10^6 years (Lambeck, 1988). Any parallels in the periodicity of extension, strike-slip and uplift at the Red Sea, the Dead Sea rift and the Zagros respectively could be explained by the accumulation and release of elastic strain at faults within locations with broadly similar (especially elastic) crustal properties. And, as elastic strain is generally confined to a zone whose width is similar to the thickness of the elastic medium, events on, say, the Dead Sea transform will not be felt more than a few tens of kilometres away.

But there are two major reasons for not rejecting out of hand the possibility that these plate boundary interactions are at least partly interdependent (Hempton, 1987). The first derives from long-term chronologies. Red Sea extension has been episodic, with a prolonged hiatus ~15–5 Ma followed by faster extension, difficult to reconcile with imperturbable plate translation on its eastern margin. What is more, accelerated extension coincided with the onset of extrusion tectonics north of the Bitlis–Zagros suture (Hempton, 1987). A later, plate-scale rearrangement of the stress field left its traces in the Gulf of Suez rift as well as in the central Red Sea. It led during the late Pleistocene to a change in extension azimuth from 045–055° to its present orientation of 010–020° (Bosworth and Taviani, 1996).

Secondly, the gravity data suggest that "elastic flexure of a relatively stiff continental plate over an inviscid fluid layer may distribute vertical stress associated with the Zagros orogeny as far...as the Persian Gulf" (Snyder and Barazangi, 1986). The suggestion here is that, when stress is released in major earthquakes, the effects are transmitted from the Zagros as far as the Red Sea and vice versa. Such waves even if drastically attenuated during their traverse could trigger impending slip on the opposite margin by modifying the static stress field (cf. Anderson et al., 1994).

By the same token, the fact that the proposed extensional episodes at the Red Sea and the uplift events in the Persian Gulf bear on short tracts of plate boundary is not a serious obstacle to the notion of large-scale time correlation. Anderson (1975) has shown how trench earthquakes may trigger activity along adjacent parts of the arc because stresses are increased by accelerated plate motion and stress wave diffusion. Such an effect is to be expected along other kinds of plate boundary characterized by strain accumulation.

Recent developments in ^{14}C dating and satellite geodesy offer scope for testing these suggestions. The time lag between events at the Zagros and on the Dead Sea rift falls squarely within the range of palaeoseismology. The propagation of postseismic stress waves could already be detected by tiltmeters over two decades ago (Bott and Dean, 1973); it appears to lend itself to satellite altimetry and radar interferometry, and may provide a reliable precursor of destructive seismicity.

Acknowledgements

I thank J.R. Cann, Donald Mann, R. I. Walcott and R.S. White for discussions; Rob Butler, Mike Ellis, Stan Murrell, Iain Stewart and R.S. Yeats for their comments on a draft of this paper; and L. Whiteside and Simon Tapper for help with the figures.

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