Structure and evolution of the northern Oman margin: gravity and seismic constraints over the Zagros–Makran–Oman collision zone

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Abstract

The obduction process in Oman during Late Cretaceous time, and continental-to-oceanic subduction along the Zagros–Makran region during the Tertiary are consequences of the Arabian–Eurasian collision, resulting in construction of complex structures composed of the Oman ophiolite belt, the Zagros continental mountain belt and the Makran subduction zone with its associated accretionary wedge. In this paper, we jointly interpret Bouguer anomaly and available petroleum seismic profiles in terms of crustal structures. We show that the gravity anomaly in northern Oman is characterized by a high-amplitude negative-positive couple. The negative anomaly is coincident with Late Cretaceous (Fiqa) and Tertiary (Pabdeh) foreland basins and with the Zagros–Oman mountain belts, whereas the positive anomaly is correlated to the ophiolite massifs. The Bouguer anomaly map indicates the presence of a post-Late Cretaceous sedimentary basin, the Sohar basin, centred north of the Batinah plain.

We interpret the negative/positive couple in terms of loading of the elastic Arabian lithosphere. We estimate the different Cretaceous-to-Recent loads, including topography, ophiolite nappes, sedimentary fill and the accretionary prism of the Makran trench. A new method, using Mindlin’s elastic plate theory, is proposed to model the 2D deflection of the heterogeneous elastic Arabian plate, taking into account boundary conditions at the ends of the subducted plate. We show that remnant ophiolites are isolated from Tethyan oceanic lithosphere in the Gulf of Oman by a continental basement ridge, a NW prolongation of the Saih-Hatat window. Loading the northward-limited ophiolite blocks explains the deflection of the Fiqa foredeep basin.

West of the Musandam Peninsula, the Tertiary Pabdeh foredeep is probably related to the emplacement of a 8-km-thick tectonic prism located on the Musandam Peninsula and in the Strait of Hormuz. Final 2D density models along profiles through the Oman mountain belt and the Gulf of Oman are discussed in the framework of Late Cretaceous obduction of the Tethys and synchronous subduction and exhumation of the Oman margin.

Keywords: flexure; elasticity; gravity; obduction; Oman; ophiolites
1. Introduction

The northern Oman mountain range (Fig. 1) is mainly characterized by the Cretaceous ophiolite belt of 600 km long, from the Strait of Hormuz to the Indian Ocean. High topography, up to 3000 m in the Jebel-Akhdar, composed of Palaeozoic to Tertiary rocks, is the other major feature of the northeastern Arabian mountain range (Fig. 1). If previous studies on ophiolites led to a better understanding of fast-spreading ridge mechanisms, they have been mainly focused on the obduction process as a stacking of oceanic material sheets (Glennie et al., 1973; Nicolas, 1988; Michard et al., 1994).

Obduction is accompanied by subduction and rapid exhumation of the Arabian continental margin and by formation of foredeep basins in external zones. The obduction event was followed by a Tertiary compressive regime in relation with Arabian–Eurasian convergence. This period is characterized by building of continental mountain belts in Zagros and Oman and creation of foredeep basins. The closing process is still active along the Zagros belt and the Makran subduction zone.

This tectonic evolution has built the actual northern Oman margin whose deep structures remain almost unknown. Are the ophiolites massifs northward-rooted? How did the convergence process model the passive northern Oman margin? And how can we explain the preserved foreland basins? Foreland basins are evidences of the large-scale deformations of the lithosphere due to loading during the convergence process. An elastic behaviour is classically adopted to estimate the deflection due to loading in mountain belt (Karner and Watts, 1983; McNutt et al., 1988). This approach is used in this paper, considering that the proposed models will give effective elastic properties for a more complex mechanical behaviour of the Arabian lithosphere (McNutt et al., 1988). Gravity and seismic data are merged over the Zagros–Makran–Oman to constrain the geometry of sediment deposits and ophiolite nappes. Then, 2D elastic models are developed to explain observed deflection of the Arabian lithosphere due to loading effects (ophiolites, topography and/or sedimentary loads). Finally, the consequences of our present-day structural model for the tectonic evolution of the region are discussed.

2. Geological setting

Northern Oman was a Tethyan passive continental margin initiated by pulsed rifting beginning in Permian time. The formation of the Hawasina and Hamrat-Duru basins and opening of the Tethys ocean followed in Triassic time (Béchennec et al., 1988; Pillevuit, 1993). The compressive regime, associated with Arabian–Eurasian convergence, started during the Late Cretaceous. It corresponds to a low-angle intra-oceanic subduction initiated at mid-oceanic ridge in the neo-Tethyan domain (Nicolas, 1988). Evidence of intra-oceanic thrusting includes the high-temperature (HT) metamorphism in the ophiolitic sole during the Cenomanian (95–90 Ma, Hacker, 1994). In the Campanian (80 Ma), southward migration of nappes (ophiolites, Hawasina and Hamrat-Duru sediments) reached the margin (Glennie et al., 1973; Lanphere, 1981; Boudier et al., 1985; Montigny et al., 1988). This stage was accompanied by downflexing and thrusting of the margin, with the deposition of the Muti formation in a first foredeep basin. Simultaneously, the Arabian continental margin was subducted northward, as it is shown by the high-pressure (HP) metamorphism in the Saih-Hatat window (20–23 kbar, 500°C) (Wendt et al., 1993; Searle et al., 1994) and was quickly exhumed as indicated by the unconformity of Maastrichtian oceanic transgressive sediments (70 Ma) onto the metamorphic unit. Uncertainties on the Early Cretaceous ages obtained by the 40Ar/39Ar method (Montigny et al., 1988; El Shazly and Lanphere, 1992) do not allow to consider an older high-pressure event in our geodynamic framework (Searle et al., 1994). Last motions of nappes associated with exhumation are recorded in the subsequent Fiqa foredeep basin, filled by nappé detritus.

From the Maastrichtian to Early Miocene, subsidence and transgression occurred in northern Oman, with filling of the Sohar basin in the Oman embayment (Fig. 1). During this epoch, compressive tectonic events were recorded in the Musandam area with the formation of a third foredeep basin (Pabdeh) in the Sarjah area (Brown, 1971; Ricateau and Riché, 1980; Searle, 1985; Searle, 1988). During the Early Miocene–Pliocene, thrusts were reactivated in the whole range with large foldings and uplifts (Carbon et al., 1996). These structures are still preserved in
Fig. 1. Geodynamic framework (a) and simplified geological map (b) of the northern Oman area (modified from Coleman, 1981, and D. Carbon, pers. commun., 1996). Ophiolite massifs: 1 = Ibra; 2 = Maqsad; 3 = Muscat; 4 = Rustaq; 5 = Haylayn; 6 = Bahla; 7 = Wuqbah; 8 = Salahi; 9 = Fizh; 10 = Aswad.
Fig. 2. (a) Density of stations on the studied area. Origin: Petroleum Development Oman (onshore Oman), Bureau Gravimétrique International (Gulf of Oman: ships and derived satellite altimetry data, Sandwell and Smith, 1992), onshore Iran (Dehghani and Makris, 1984), Open University (northern Oman mountains, Shelton, 1984), Laboratoire de Géophysique et Tectonique–ISTEEM (northern Oman mountains, Ravaut et al., 1993, and this paper). (b) Bouguer anomaly map of the Zagros–Makran–Oman area. Gravity data at each station have been projected on a regular 2.5 × 2.5 km size mesh using a kriging interpolation method. Bouguer anomaly was calculated in the IGSN71 reference system using the GR67 ellipsoid formula; topographic corrections have been applied in a radius of 167 km using a 2600 kg/m$^3$ density reduction for relief. For recent stations, IGSN71 reference gravity base is Seeb Airport, $g = 978923.59±0.05$ mGal (Shelton, 1984). Contour interval: 10 mGal. Projection system: Universal Transverse Mercator (UTM); meridian origin: 57°E.
Fig. 2 (continued).
the present-day Oman mountains (3000 m in Jebel-Akhdar).

Pliocene deformation is characterized by continental collision in the Zagros mountains (Bird, 1978). In Oman, compressive and extensive structures coexist (Carbon et al., 1996). N–S continental subduction beneath the Zagros belt and oceanic subduction beneath Makran control the tectonic and sedimentary evolution of the area: the deflection of the Arabian platform allows the creation of a NW–SE foreland basin in the Arabian Gulf, dipping to the northeast. In the Strait of Hormuz, west of the Zendan fault, this deflection becomes E–W and Recent sediments are tilted to the east. East of this fault, subduction of the Tethys is associated with deflection of oceanic basement, active sedimentation, and formation of the Makran accretionary prism (White and Ross, 1979).

3. Bouguer anomaly map

For several decades, gravity surveys have been carried out in Oman for petroleum exploration. Petroleum Development Oman (PDO) collected data at about 45,000 stations from surveys realized between 1955 and 1969. These data were gathered south of the northern Oman mountains and on the Batinah coastal plain. They are supplemented by around 900 measurements covering the coastal plain and the ophiolite massifs (Manghnani and Coleman, 1981; Shelton, 1984). In 1992, we mapped the gravity field in the Jebel-Akhdar area (Ravaut et al., 1993) and this survey was extended eastward and westward in 1993, resulting in 1012 new stations with a mean spacing of 4 km. All the observed gravity values have been tied to the International Gravity System Network 1971 (IGSN71) from Seeb Airport base (Shelton, 1984). For these data, the topographic corrections were computed using Hammer’s table and 1:100,000 topographic maps up to 10 km. The amplitude of this correction reaches a maximum of 20 mGal in the Jebel-Akhdar area with a standard deviation of 1 mGal for the mountain area (Rey, 1989).

In order to map the gravity anomaly over the Zagros–Makran–Oman zone, gravity profiles from cruises (White and Ross, 1979) and Iranian surveys (Dehghani and Makris, 1984) were extracted from the Bureau Gravimétrique International (BGI) data bank. The marine data were processed with SEAVALID software (Adjaout and Sarrailh, 1992) by adjusting the cross-over errors observed at intersecting ship tracks.

In all, the surveys represent 52,077 stations (Fig. 2a). We computed free-air anomaly using the Geodetic Reference System 1967 ellipsoid formula. In the Gulf of Oman, 30 km offshore, the data were densified by adding the gridded (5 min×5 min) free-air data calculated by Sandwell and Smith (1992) from sea altimetric measurements. Bouguer anomalies (Fig. 2b) were mapped using gridded (2.5 km ×2.5 km) values after applying topographic corrections up to 167 km using the ETOPO5 digital elevation model from the Defense Mapping Agency (USA), with 2600 kg/m³ for relief density. On land, the Bouguer anomaly standard deviations are mainly related to station elevation uncertainties and topographic corrections, and are estimated to be 2.5 mGal.

The map exhibits negative anomalies on continental domains (Iran and Arabia) and positive anomalies on the oceanic Gulf of Oman. A spectacular minimum is observed in the Fiqa and Pabdeh foredeeps (−60 to −120 mGal). Toward the Strait of Hormuz, this gravity minimum assumes an arcuate shape and is connected northward to the negative anomaly coinciding with the Zagros belt (Snyder and Barazangi, 1986). From the Musandam promontory to Muscat, the negative trend is bordered by high anomalies (80 to 150 mGal) located on ophiolite outcrops, except between Muscat and Sohar where anomalies are shifted northward under the Batinah plain.

The Oman coast line is roughly the limit between a high ‘ophiolite anomaly’ and the negative anomaly (−50 mGal) located eastward on the Mastrichtian-to-Recent Sohar basin. The 90–120 mGal Bouguer values in the Gulf of Oman combine effects of the oceanic lithosphere and of the sediments recently deposed and piled into the accretionary prism of the Makran trench (White and Ross, 1979) as it is attested by the relative minimum along the Iranian Makran coast (Fig. 2b).

4. Elastic response of Arabian lithosphere to Cretaceous-to-Recent loading

In continental thrust belts, the Bouguer anomaly is classically used to analyze large deformations of
the lithosphere. It gives information about rheological behaviour of the continental lithosphere and of applied forces. Elastic plate models are favoured by many authors to explain the deformation of continental lithosphere (Karner and Watts, 1983; Royden, 1988; McNutt et al., 1988; Kruse and Royden, 1994; Lyon-Caen and Molnar, 1989). In our study, the asymmetric gravity low, extending from the Zagros to northeast Oman, may be interpreted as evidence of the elastic deflection of the Arabian lithosphere and its Moho. The actual deformation is a consequence of complex superposition of topographic loads (Zagros and Oman mountain belts), internal density heterogeneities, and associated subsurface loads (ophiolite nappes of Oman) and bending moments and/or vertical shear forces acting at the ends of the subducted Arabian plate beneath the Zagros range and Makran area.

We now estimate the active loads, inherited from previous tectonic events or related to present-day convergence, and foredeep geometry, inferred from gravity data and seismic exploration results. Syn-to-post obduction sedimentary basins (Fiqa, Pabdeh and Gulf of Oman) geometry may be defined by seismic data and well-log information. They are representative of the deflection of the Arabian lithosphere during compressive events. Correcting Bouguer anomaly for basins aids in modelling the ophiolite massifs and their associated loads.

4.1. The model (Fig. 3)

The classical equilibrium equation of a 2D thin elastic plate with variable elastic thickness, caused by vertical forces $F$ (in N/m$^2$) and overlying an inviscid fluid of density $\rho_m$ (kg/m$^3$) is given by:

$$\Delta [D \Delta (w - w_0)] + \rho_m g (w - w_0) = F$$  \hspace{1cm} (1)

where $D$ is the flexural rigidity (N m), $w$ (m) is the lateral displacement normal to the plate (deflection) and $\Delta$ is the Laplacian operator. In our case, $w$ is a deflection calculated with respect to an initial reference surface $w_0$. Eq. 1 is solved using Mindlin plate theory and a finite elements formulation (see Appendix A). The mechanical conditions at the ends of the subducted plate are simulated, at each node with a 10 km spacing, by vertical shear forces, $Q$ (N), and bending moments $M$ (N m). No earthquakes or large deformations are now observed in the Gulf of Oman, excepted along the Makran trench. Therefore, we bound the elastic plate 200 km north of the main Zagros thrust fault (MZT). In Makran, the ends correspond to the Iranian Lut block, north of the Makran prism (Fig. 1). The southern and western limits of our model are characterized by free boundary conditions and chosen far enough (3000 km) to avoid edge effects on the studied area. The eastern border located at the Owen fracture zone is a free boundary.

4.2. Geometrical constraints on sedimentary basins and ophiolite massifs

4.2.1. Syn-to-post obduction sedimentary structures:

(a) The Fiqa and Pabdeh basins. The numerous seismic profiles were merged during petroleum exploration of the Late Cretaceous Fiqa and Tertiary–Quaternary Pabdeh basins and several isopach maps have already been published (Searle, 1983; Patton and O’Connor, 1988; Boote et al., 1990; Warburton et al., 1990). The bottom of the Fiqa basin is marked by an onlap above the Muti formation whereas Maastrichtian–Paleocene overlying series form the top (Fig. 4a). The thickness of the southern Fiqa basin is detailed by adding results of ten unpublished seismic lines from Ministry of Petroleum and Minerals of Oman (MPMO).
3500 m thick Fiqa deposits are characterized by an asymmetric shape and are eastward-limited by the syn-obduction thrust nappes (Fig. 4b). The maximum depth is in front of the highest gravity anomalies created by ophiolite blocks. The location of the Lekhwair forebulge has been estimated at 100 km from the allochthonous wedge (Fig. 1) (Boote et al., 1990).

The so-called ‘Pabdeh basin’, including all the post-obduction sediments in our terminology, is not superimposed on the Fiqa foredeep. Maximum thickness (Fig. 4c), up to 4000 m, is shifted northward at west of the Musandam Peninsula. The lower part of this section is mainly related to intense folding and thrusting during the Early Tertiary (Ricateau and Riché, 1980; Searle, 1983; Dunne et al., 1990). The upper part corresponds to gently northward-dipping sediments in agreement with the actual subduction geometry. The Pabdeh basin is bordered by the 2000 m Musandam–Jebel-Akhidar mountains but its variation in depth is not correlated with the mean altitude of the neighbouring relief.

The N–S extension of both foredeeps is quite surprising in the frame of a N–S convergence context between Arabia and Iran since the Late Cretaceous.

(b) The Sohar basin and the oceanic basin. The Sohar basin was identified from previous academic and petroleum marine seismic surveys in the Gulf of Oman (MPMO, unpublished data). It is associated with a negative gravity anomaly (Fig. 2b). A 5000 m thick sequence of Maastrichtian-to-Recent sediments is locally observed in available wells. The basin geometry is constrained by sixteen unpublished petroleum seismic lines integrated with well log data and seismic cross-sections from White and Ross (1979) (Fig. 5a).

We focus on the NW–SE line 15 carried out during the 1975 French CEPM (Consortium d’Etude Petrolière des Marges) project which has been reprocessed on a length of 180 km (Fig. 5b). The Sohar basin exhibits a thick sedimentary sequence whose bottom coincides with an energetic reflector at 4 to 5 s (two way traveltime, TWT) depth. Two stages of deposit may be identified by important reflectors: a 2.5 s TWT thick carbonate unit from Maastrichtian to Late Miocene and a syn-tectonic compressive epoch from Late Miocene to Recent. Eastward, the observed basement rise may correspond to extension of the Oman ophiolite nappes or of the metamorphic Permo–Triassic continental margin. In the deepest part of the Oman Gulf, gently dipping layers are associated with the downflexed oceanic plate along the Makran trench and are folded at the front of the accretionary wedge.

The two main reflectors described on the profile are identified on all seismic lines including profiles published by White and Ross (1979) and were digitized. Well-log velocity records (Fig. 6a) were used to convert the TWT seismic reflectors to depth. We have extrapolated northward the sediments thickness behind the Makran front, in order to avoid border effects during sedimentary stripping of the Bouguer anomaly. The basement reflector is prolonged in the subduction direction (N–006°) with a 5° slope deduced from earthquake distributions (Byrne and Sykes, 1992) and from OBS seismic profiles (Niazi, 1980). The sedimentary cover is limited southward by ophiolite outcrops of Oman. The resulting isopach map (Fig. 5c) exhibits at Muscat a N–W–elongated basement ridge separating the Sohar basin from the oceanic domain of the Gulf of Oman. Along the Makran coast line, the sedimentary column of the prism reaches a maximum thickness of 13 km.

4.2.2. The ophiolite massifs

The wavelength and amplitude of anomaly created by the sedimentary cover partly hides the gravity effect of the ophiolite nappes. Stripping these effects requires knowledge of the density distribution in sediments. In the Oman foreland, bulk rock measurements realized during previous work (Manghani and Coleman, 1981; Shelton, 1984) indicate a 2400 kg/m³ density for Fiqa sediments and a 2550 kg/m³ density for Tertiary units. In the Gulf of Oman, well-log density records (Fig. 6b) are used to define a more refined density–depth relationship by adjusting data by the following exponential law:

\[ \rho = \rho_b - \phi_0 (\rho_b - \rho_l) \exp\left(-z/\lambda\right) \]  

where \( \rho_b \) is the matrix density (\( \rho_b = 2670 \) kg/m³), \( \rho_l \) is the fluid density (\( \rho_l = 1000 \) kg/m³), \( \phi_0 \) is the near surface porosity (\( \phi_0 = 0.5 \)) and \( \lambda \) is the characteristic depth (\( \lambda = 6000 \) m). A stripped map representing the effect of deep structures and of the ophiolites (Fig. 7a) is obtained by removing the sediment contribution to the Bouguer anomaly map. Ophiolite gravity highs are enhanced to a maximum...
Fig. 4. (a) An example of seismic profile crossing the foreland domain in Oman, from Boote et al. (1990). Seismic data from the Ministry of Petroleum and Minerals of Oman (MPMO) are integrated into previous re.

(b) A map showing the Tertiary deposits over the foreland of the northern Oman mountains, from published documents (Patton and O'Connor, 1988; Boote et al., 1990). Unpublished seismic data from the Ministry of Petroleum and Minerals of Oman (MPMO) are integrated into the map.
See Fig. 5a for location. (b) Isopach map of Fiqa deposits over et al., 1990; Warburton et al., 1990). Southward, unpublished results. Contour interval is 500 m. (c) Isobath map of Tertiary 1988; Boote et al., 1990; Warburton et al., 1990). Southward, previous results. Contour interval is 500 m.
of 155 mGal and are seaward shifted. These anomalies are globally still separated from the oceanic positive domain (220 mGal) by a relative minimum. In consequence, we assume that the preserved ophiolites along the Oman coast are not connected to Tethyan oceanic lithosphere of the Gulf of Oman, except for the Muscat ophiolites. Anomalies created by the ophiolites were analytically extracted from the stripped map (Fig. 7a). A low-order polynomial regional anomaly was computed and removed to obtain the ophiolite contribution to the regional field (Fig. 7b). 2D density models along three profiles crossing the northern Oman margin (Fig. 7c) confirm the rootless character of the ophiolite blocks as it was already proposed by Manghnani and Coleman (1981) and Shelton (1984). West of Sumail Gap, nappes are northward-dipping beneath the Batinah coast. A maximum depth of 10 km is observed on the Rustaq–Haylayn block.

A simple model consisting of an equivalent horizontal stratum is chosen to calculate the mass by surface unit using the following relationship:

$$\sigma = \int_0^h \Delta \rho(z) dz \approx \Delta g / 2\pi G$$

(3)

where $\sigma$ is the surface density contrast, $h$ is the thickness of the ophiolite blocks, $\Delta \rho(z)$ is the density contrast of ophiolites with Arabian crust, $\Delta g$ is the ophiolites anomaly and $G$ the gravitational constant (see Table 1 for numerical values).

4.3. Applications

4.3.1. The obduction flexure

The loads associated with the ophiolite anomaly (Fig. 7b) is defined in Eq. 1 by $F_0 = \sigma g$, where $g$ is the gravitational acceleration. The resulting deflection is accompanied in the foreland domain by
Fig. 5 (continued). (b) Vertical seismic reflection cross-section, line-drawing and interpretation in two-way-traveltime (TWT) of the western part of line 15 from CEPM survey (bold part of the line 15 in (a)). Seismic line reprocessed in the frame of this work (geometry setup, muting, spectral analysis, source deconvolution (VAPORCHOC), deconvolution, CDP stack, velocity analysis (normal move out), final stack on Promax software). Sohar basin, continent–ocean transition zone, abyssal plain and Makran prism front are indicated on the section. 1 = Late Miocene to Recent; 2 = Paleocene to Early Miocene; 3 = Late Cretaceous.
infilling materials acting through a vertical force 
\[ F_i = g\rho_i(w - w_0), \]
where \( \rho_i \) is the density of the infilling material. Table 1 shows numerical values of the parameters used in Eq. 1. In the foreland area, the pre-obduction elevation of the plate is supposed to be null, i.e. \( w_0 = 0 \). As a first step, the rigidity is assumed constant over the plate. The theoretical deflection is calculated for various \( D \)-values ranging from \( 10^{18} \) N m to \( 10^{25} \) N m and results are illustrated on an E–W profile (Fig. 8a and b). An homogeneous elastic plate assumption is not realistic to explain the Fiqa foredeep basin. Our preferred values are \( 5 \times 10^{23} \) N m for the external Arabian platform and \( 10^{22} \) N m over the northern Oman belt and the coastal plain (Fig. 8a and b). The effective elastic rigidity under the Oman platform is close to Snyder and Barazangi’s estimation (Snyder and Barazangi, 1986) given for the Arabian foreland basin \( (10^{23} \) N m). The calculated deflection is in agreement with the observed Fiqa sediment distribution (Fig. 4b). Moreover, the preserved Lekhwair bulge is roughly coincident with our computed bulge at 100 km from the frontal folded–faulted zone. High-to-low effective rigidity variation, from external zones toward the northern Oman Mountains, is explained by a relative gently dipping basement (and Moho) that steepens abruptly.

4.3.2. Flexure related to Tertiary-to-Recent compressive events

The Late Tertiary structural evolution of the Oman and Zagros thrust belts has resulted in building of important relief (up to 3000 m in Jebel Akhdar and 2000 m in Musandam and Zagros). The corresponding load in the right hand side of Eq. 1 is represented by 
\[ F_r = (h_r + (w - w_0))\rho_c g \]
where \( h_r \) is altitude of the relief and \( \rho_c \) the density of the infilling material and topography. The initial deflec-
Fig. 6. (a) Velocity/depth well-log data of three boreholes in the Gulf of Oman (see location in Fig. 5a). (b) Well-log density data from three boreholes (see location in Fig. 5a).
Fig. 7. (a) 'Regional' gravity anomaly map from Bouguer anomaly map corrected for shallower effects. Gravity effect of sedimentary basins is computed using constant densities in foreland area ($\rho_1 = 2400$ kg/m$^3$ for Fiqa and $\rho_2 = 2550$ kg/m$^3$ for Tertiary-to-Recent deposits) and an exponential density/depth law in gulf of Oman (Sohar basin, Makran accretionary prism) (Fig. 6b). The blank area corresponds to an absence of data (see Fig. 2a). (b) 'Residual' gravity anomaly map. The long-wavelength anomalies are analytically removed (low-order polynomial approximation) from the map of (a). The residual map represents the ophiolites effect.
Fig. 7 (continued). (c) 2D ophiolites models along three profiles through the Oman mountain range. Observed anomalies are drawn from the residual map in (b). Density of ophiolites is 3070 kg/m³ (Manghnani and Coleman, 1981; Shelton, 1984).

Fig. 7 (continued). (c) 2D ophiolites models along three profiles through the Oman mountain range. Observed anomalies are drawn from the residual map in (b). Density of ophiolites is 3070 kg/m³ (Manghnani and Coleman, 1981; Shelton, 1984).
Table 1
Definition of symbols used in the text

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<th>Symbol</th>
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<td>w</td>
<td>depth after deflection</td>
<td>m</td>
<td>computed deflection is ( (w - w_0) )</td>
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<td>D</td>
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<td>length characteristic (density/depth law in Sohar basin)</td>
<td>m</td>
<td>6000</td>
</tr>
<tr>
<td>( \phi_0 )</td>
<td>surface porosity (density/depth law in Sohar basin)</td>
<td>-</td>
<td>0.5</td>
</tr>
<tr>
<td>( t_p )</td>
<td>thickness of Makran accretionary prism</td>
<td>m</td>
<td></td>
</tr>
<tr>
<td>( t_m )</td>
<td>thickness of Musandam tectonic prism</td>
<td>m</td>
<td></td>
</tr>
<tr>
<td>( Q_{2,1}, Q_{m} )</td>
<td>vertical shear force at ends of the subducted plate</td>
<td>N</td>
<td>Makran: ( Q_{2,1} = 1.5 \times 10^{17} ) Zagros: ( Q_{2,1} = 5 \times 10^{16} )</td>
</tr>
<tr>
<td>( M_{2,1}, M_{m} )</td>
<td>bending moment at ends of the subducted plate</td>
<td>N m</td>
<td>Makran: ( M_{2,1} = 5 \times 10^{20} ) Zagros: ( M_{2,1} = 5 \times 10^{20} )</td>
</tr>
</tbody>
</table>

5. Consequences on density models of the northern Oman margin

The deflection model does not simply explain the regional trend of the stripped anomaly (Fig. 7a). Density modelling is realized to clarify the deep structures of oceanic crust and the Arabian margin whose thinning is superimposed on the deflection. 2D models are proposed along three profiles crossing the northern Oman mountains and the Gulf of Oman (Fig. 11). The sedimentary structures are constrained by isopach maps (Fig. 4b and c) and ophiolite bodies were previously modeled (Fig. 6c). From seismic refraction results in Saudi Arabia (Mooney et al., 1985; Mokhtar and Al Saeed, 1994), a reference thickness of 38 km and a crust/upper-mantle density contrast of 400 kg/m\(^3\) were adopted for the Precambrian Arabian crust. Beneath the continental foreland domain and oceanic part of the Gulf of Oman, the Moho shape was approximated by calculated deflection summed for obduction and Tertiary loading (Fig. 8a, Fig. 9b).

Preliminary density models have shown the rootless character of the ophiolite massifs (Fig. 7c) whose loading is consistent with flexure of the Fiqa foreland basin. The ophiolite nappes are probably separated from the oceanic domain by a continental ridge previously identified on the seismic profiles. For the Gulf of Oman, a 6-km-thick oceanic crust is chosen from field work evidences on Oman ophiolites (Nicolas, 1988). Such a thickness is not sufficient to explain the low values, near 220 mGals, of the stripped anomaly for a typical oceanic lithosphere. The poor available geophysical constraints involve large undetermination for deep density solutions: an oceanic crustal thickening by a factor 2, two
Fig. 8. (a) Theoretical deflection caused by ophiolite loading. $D = 5 \cdot 10^{23} \text{ N m}$ for Arabian platform and $D = 10^{22} \text{ N m}$ for internal part of the northern Oman mountain range.

Oceanic lithospheres superimposed during the subduction process and/or northward extension of continental crust beneath the Makran accretionary wedge. Results of seismic experiments off the Makran coast (Niazi, 1980) have revealed a 6.7-km-thick 'volcanic basement' and allow us to reject the first assumption.

The regional increase of the stripped anomaly from the Oman mountains to the oceanic domain is interpreted in terms of thinning of the Arabian continental crust. For profiles A and B (Fig. 11), we propose to extend the thinned Arabian continental crust to 50 km off the Makran coastline. Northward, the proposed model, corresponding to the low-angle subduction hypothesis, is poorly constrained. The 2D models indicate a continuity of the Saih-Hatat window along the 'seismic basement ridge' separating the Sohar basin from the oceanic domain, and the external character of the ophiolite massifs (Ibra to Fizh, Fig. 1) with respect to the continental ridge. North of Muscat, extension of the continental margin...
along profile C is more reduced than on profiles A and B over the Batinah plain.

6. Discussion

The agreement between the predicted and the observed deflections for the obduction stage involves a flexural rigidity decreasing from the foreland area to the internal domain. The low rigidity in the internal zone may be the result of crustal basement thickening of the Arabian margin, consecutive to folding and thrusting during the obduction process. It may also reflect the initial elastic contrasts between a rigid Arabian shield and a weak Permo-Triassic margin.

The Tertiary-to-Recent deflection takes into account the major structural features:

(a) The 4-km-thick Tertiary-to-Recent N-S Pabdeh foredeep basin (Fig. 9c) is explained by: (1) a corner effect of the sediments piled in the Makran prism near the Strait of Hormuz; (2) subduction along Zagros-Makran; and (3) a 8-km-thick tectonic prism over the Musandam Peninsula and the Strait of Hormuz. In this area, compressive events have been described during the Tertiary and are characterized by intense folding and reverse faulting (Ricateau and Riché, 1980; Searle, 1988; Dunne et al., 1990). Such a tectonic prism is in agreement with the upthrust Mesozoic platform observed on seismic profiles (Ross et al., 1986; Michaelis and Pauken, 1990).

(b) On the Zagros belt, bending moment and shear force allow to tilt the Arabian Moho from about 1° near the Zagros folded belt to 5° at the MZT, and to predict a 15 km maximum crustal thickening.

(c) Over the Makran, the chosen boundary conditions and the prism loading lead to a deflection in agreement with the geometry of the upper Campanian reflector (Fig. 9c). Over the oceanic domain, the predicted deflection is consistent with the depth of the upper Campanian basement, considering an initial depth \( w_0 = 2 \) km for this marker.

The 2–4 km amplitude of the deflection calculated over the Batinah plain and the northern Oman mountains is mainly a result of bending of the subducted lithosphere and accretionary wedge of the Makran. However, such a subsidence is not observed in Oman since Campanian time. It indicates that the elastic assumption is too simple to fully explain the deformation of the northern Oman region. Decreasing amplitude of the deflection could be obtained for instance by decoupling the continental part from the oceanic part of the Arabian lithosphere. Moreover, compressive events in northern Oman since the Late Tertiary have resulted in reactivation of a low-angle thrust ramp, crustal shortening and building of the northern Oman mountains (Carbon et al., 1996).

The density models have revealed a lateral variation of the continental margin structures from the Muscat area to Musandam Peninsula. This difference may be partly inherited from variations in extensional process during the Permo-Triassic Tethyan opening. Lateral variations of the models may also reflect the different conditions of the subduction and exhumation of the margin in the Late Cretaceous, as it is attested by restricted outcropping of the HP rocks within the Saih-Hatat window.

The rootless character of the ophiolite nappes and the large seaward extension of the Arabian margin may be understood in the frame of the tectonic evolution of the northern Oman region during the Late Cretaceous:

(a) At the Coniacian–Santonian stage (90–80 Ma) intra-oceanic obduction occurred, with thrusting and
Fig. 9. (a) Theoretical Tertiary deflection from topography and Makran accretionary prism loading. The low rigidity domain of the Oman mountain belt ($D = 10^{22}$ N m) issued of obduction modelling is extended northward beneath the Zagros belt. Rigidity of the plate over the Gulf of Oman is $5 \times 10^{23}$ N m. (b) Total deflection model for the Tertiary to Recent, including boundary conditions for the Zagros-Makran subduction (Zagros: $M_s = 5 \times 10^{20}$ N m; $Q_s = 5 \times 10^{16}$ N; Makran: $M_m = 5 \times 10^{20}$ N m and $Q_m = 1.5 \times 10^{17}$ N) and a thick tectonic prism over the Musandam Peninsula and the Strait of Hormuz.
Fig. 9 (continued). (c) The theoretical Tertiary deflection \((w - w_0)\) is compared to observed thickness of the post-Late Cretaceous sediments of the Pabdeh foredeep basin (profile II-III') and of the Makran accretionary prism (profile III1-II1'). We assume that \(w_0 = 0\) for profile II-III'. Knowledge of depth after deflection \(w\) depends on assumption about initial depth before deflection, i.e. bathymetry of the margin and Oman abyssal plain. Discrepancy between theoretical and observed curves from a distance of 100 km on profile III1-II1' may be explained by an initial depth increase of the bathymetry of 2 km in the continental–oceanic transition zone.

7. Conclusions

In this paper, we have shown that the Bouguer anomaly map on the northern Oman mountains is characterized by a negative–positive couple as in many collision belts. The large-scale gravity low over the Zagros–Oman mountains and foreland has been interpreted in terms of elastic flexure of the Arabian lithosphere as attested by the Fiq' and Pab-deh foredeep basins. The Late Cretaceous deflection in the Fiq' foredeep has been explained by the loading effect, during obduction, of the rootless ophiolite nappes. The Tertiary-to-Recent compressive events explain the important deflection observed in the Pab-deh basin and Makran trench. We have demonstrated that deflection in the Pabdeh foredeep is related to subsurface sedimentary loading, probably in the Early Tertiary, over the Musandam Peninsula and the Strait of Hormuz. Detailed studies on this area based on an accurate foredeep stratigraphy will allow to restore the compressive history of the Musandam–Hormuz area.

Gravity and seismic studies have revealed the existence of the Sohar basin and the complexity of the northern Oman margin geometry. The presence of a continental basement ridge prolonging the Saih-Hatat window must be confirmed by complementary seismic studies.

We have proposed density cross-sections of the northern Oman margin which are characterized by a large northward extension of the continental crust.
The low value of the Bouguer anomaly in the oceanic domain and the aerial character of a large part of the Makran prism indicate that a subducted Arabian plate over Makran must include a doubled oceanic lithosphere or a more extended continental crust. Here also, deep seismic soundings are necessary to solve this problem.
Fig. 11. Density models along three profiles crossing the northern Oman region and the Gulf of Oman (see location of profiles in Figs. 2 and 7a). The observed anomaly has been corrected for sediment effect. Geometry of ophiolites massifs is taken from models of Fig. 7c. Density contrast at continental and oceanic Moho discontinuity is chosen as 400 kg/m$^3$. 
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Appendix A. Finite elements formulation

Instead of a fully 3D theory, where (and without special procedure) the stiffness matrix can be ill-conditioned for small thickness, we use the Mindlin (or Reissner) plate theory which is closer to the 3D case (and so more accurate) than the classical Kirchhoff thin plate, since transverse shear strains are taken into account.

We briefly recall the expression of the virtual work principle which is a variational form of the local equilibrium equations more suitable for finite elements implementation. For more details, see for example Owen and Hinton (1980) (the slight difference in our work is the presence of an additional term in the stiffness matrix for taking into account the hydrostatic pressure acting on the bottom of the plate).

A.1. Mindlin plate equilibrium equations

Let A be the domain of interest:

\[ A = \{(x, y, z) \in I R^3 | (x, y) \in \Omega \subset I R^2, \ z \in [-h/2, h/2]\} \]

where \( h = h(x, y) \) is the plate thickness and let \( u \in I R^3 \) be the generalized displacements vector \( u = (w, \phi_x, \phi_y) \)

where \( w \) is the lateral displacement normal to the \( x^2 \)-plane and \( \phi_x, \phi_y \) the normal rotations in the \( xz \) and \( yz \)-planes. The body forces are designed by \( b \) and here, the only non-zero component of \( b \) is the lateral one:

\[ b = F e - \Delta \rho g w e, \quad e = (1, 0, 0)^T \]

where \( F \) is the lateral loading per unit area and \( -\Delta \rho g w \) is the restoring pressure associated with a depression \( w \) of the plate in the underlying fluid.

For any virtual generalized displacement \( \hat{u} = (\hat{w}, \hat{\phi}) \), the principle of virtual work can be expressed by:

\[ \int_{\Omega} \hat{\varepsilon}^T \sigma d\Omega + \int_{\Omega} \Delta \rho g w \hat{\varepsilon} u d\Omega = \int_{\Omega} F \hat{u}^T e d\Omega \]

where \( \hat{\varepsilon} = (\hat{w}, \hat{\phi})^T \in I R^5 \) stands for associated virtual curvatures and virtual shear strains:

\[ \hat{\varepsilon} = \left( \begin{array}{c} \hat{\phi}_x \\ \hat{\phi}_y \\ \hat{\phi}_z \end{array} \right) = -\left( \begin{array}{ccc} \partial^2 w/\partial x^2 & \partial^2 w/\partial x \partial y & \partial^2 w/\partial x \partial z \\ \partial^2 w/\partial x \partial y & \partial^2 w/\partial y^2 & \partial^2 w/\partial y \partial z \\ \partial^2 w/\partial x \partial z & \partial^2 w/\partial y \partial z & \partial^2 w/\partial z^2 \end{array} \right) \]

\[ \hat{\phi} = (\phi_x, \phi_y)^T = (\partial w/\partial x, \partial w/\partial y)^T \]

\( \sigma \in I R^3 \) is the vector of bending moments and shear forces:

\[ \sigma = [M, Q]^T = ([M_x, M_y, M_z], [Q_x, Q_y])^T \]

and is related to \( \varepsilon \) through the constitutive relationships \( \sigma = D \varepsilon \), where for an isotropic elastic material:

\[ D = \frac{E h}{12(1-\nu^2)} \begin{bmatrix} 1 & \nu & 0 \\ \nu & 1 & 0 \\ 0 & 0 & 1-\nu \end{bmatrix} \]

\[ D_i = \frac{G h}{\alpha} \begin{bmatrix} 1 & 0 \\ 0 & 1 \end{bmatrix} \]

\( \nu \) is the Poisson ratio, \( E \) the young modulus, \( G \) the shear modulus and \( \alpha \) a shear correction term (\( \alpha = 1.2 \)).

A.2. Finite elements implementation

In a finite elements representation, the displacements and strains may be expressed by:

\[ u(x, y) = \sum_i N_i(x, y) u_i \]

\[ \varepsilon(x, y) = \sum_i \theta_i(x, y) \]

where \( u_i \) is the vector of nodal displacements, \( N_i \) the shape functions and \( B_i \) the strain-displacement matrix (containing shape functions and their derivatives).

Writing in a similar way the virtual displacements and strains, and noting that the virtual work is true for any virtual displacement \( \hat{u}_i \), we have for each node \( i \):

\[ \int_{\Omega} B_i^T \sigma d\Omega + \int_{\Omega} \Delta \rho g w N_i u d\Omega = \int_{\Omega} F N_i e d\Omega \]

Introducing the constitutive law we can write for each element \( \Omega \), of the mesh:

\[ \sum_{j=1, nd} K_{ij}^\Omega u_j = F_{ri} \quad i = 1, \ldots, nd \]

where \( nd \) is the number of nodes in the element \( \Omega \), and \( K_{ij}^\Omega \) are given by:

\[ K_{ij}^\Omega = \int_{\Omega_j} (B_i^T \gamma B_j^T + \Delta \rho g N_j e \otimes e) d\Omega \]

and

\[ F^\Omega = \int_{\Omega} F N_i e d\Omega \]

The \( K_{ij}^\Omega \)-values form the local stiffness matrix \( K^\Omega \) whose dimensions are \( 3nd \times 3nd \). The global stiffness matrix and the
global loading vector are then obtained by assembling, respectively, all submatrix $K^-$ and all vectors $F$.

Elements used must be at least quadratic to obtain accurate solutions. Good results are performed with 6-node triangular elements or 9-node quadrangular elements. In addition, for better resolution we use a pre-processor associated with a mesh generator which performs a local mesh refinement such that element sizes are adjusted taking into account the loading distribution and/or its gradient.

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Résumé


