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Uplift and subsidence history of the Alboran Basin and a profile of the Alboran Diapir (W-Mediterranean)

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Abstract

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The cooling and subsidence history of the Alboran Basin suggests that it is underlain by a diapiric mantle bulge. Numerical models of crustal doming by Cloetingh & Nieuwland (1984) are applied to investigate the deep lithospheric structure beneath the Alboran Basin. At least two 300 km long high angle faults of east-west strike would be expected directly north and south of the Alboran Sea if the Alboran Diapir had penetrated far into the lithosphere. As there is no record of such faults, it is probable that the Alboran Diapir is a shallow bulge in the lithosphere-asthenosphere interface, which deformed the overlying MOHO. The suggested profile for the deep lithosphere in the Alboran area allows discussion of the details of Weijermars' (1985) model for the tectonic evolution of the Alboran Basin and peripheral Betic-Rif orogens during the past 30-25 Ma.

Introduction

Crustal cooling is a mechanism that transforms either lithospheric domes or thinned lithosphere into depressions. A *lithospheric dome* is initiated by a diapiric swell, which replaces the cold lithosphere with hot asthenosphere (Neugebauer 1978, 1983; Turcotte & Emerman 1983). Cessation of the diapiric rise implies stagnation of the thermal perturbation, and subsequent cooling of the diapir may cause subsidence of the overlying rocks. By contrast, *lithospheric thinning* may be caused by several mechanisms, *e.g.*: 'subaerial erosion' (Sleep 1971), deep crustal metamorphism (Falvey 1974), or extensional tectonics (Artemjev & Artyuskov 1971; McKenzie 1978; Le Pichon et al. 1982) due to stretching by so-called plate reorganisation forces (Cloetingh & Nieuwland 1984). The stretching model has been convincingly applied to explain the North Sea Graben (Jarvis & McKenzie 1980), the break-up of the Farallon plate and the evolution of the Nazca plate (Wortel & Cloetingh 1981, 1983).

Lithospheric stresses induced by lithospheric doming have recently been quantified in finite element studies of Cloetingh & Nieuwland (1984). They verified the validity of their failure predictions by using the estimated geometry of the lithospheric bulges underneath the Baikal and Ethiopian Rifts, the Rhine Graben and the Cape Verde, Hoggar, Darfur and Hawaii Swells. Cloetingh & Nieuwland's (1984) theory will be applied here to the Alboran Basin at the western end of the Mediterranean Sea. This paper shows how a deep lithospheric profile can be inferred even without direct observations of such depths.

Alboran diapir: evidence

It has been suggested that the European and African plates impinge near Gibraltar (Dewey et al. 1973; Bourrouilh & Gorsline 1979) but the expected thickened suture zone is not recogniseable between Morocco and Spain (Bonini et al. 1973). Instead, the floor of the Alboran Sea consists of a continuous 13 km thick continental crust (Fig. 1; cf.: Banda et al. 1983) and gravity anomalies suggest the presence of a mantle diapir (Van Bemmelen 1973; Bonini et al. 1973; Loomis 1975). Crustal thicknesses to the north and south of the Alboran Sea increase rapidly to about 30 and 40 km under Morrocco and Spain, respectively (Banda et al. 1983).

Although lithospheric stretching preceeds basin subsidence by cooling in many areas (*e.g.*: North Sea, Jarvis & McKenzie 1980), the subsidence in the Alboran Region is more likely to be due to lithospheric doming. The observed bulge of the MOHO beneath the Alboran Basin (Fig. 1) is interpreted here to reflect lithospheric doming and not stretching or thinning for the following two reasons:

1) There is much evidence to suggest that the nappes of the Betic-Rif orocline (Fig.1) are due to nappe-shedding from a former topographic high at

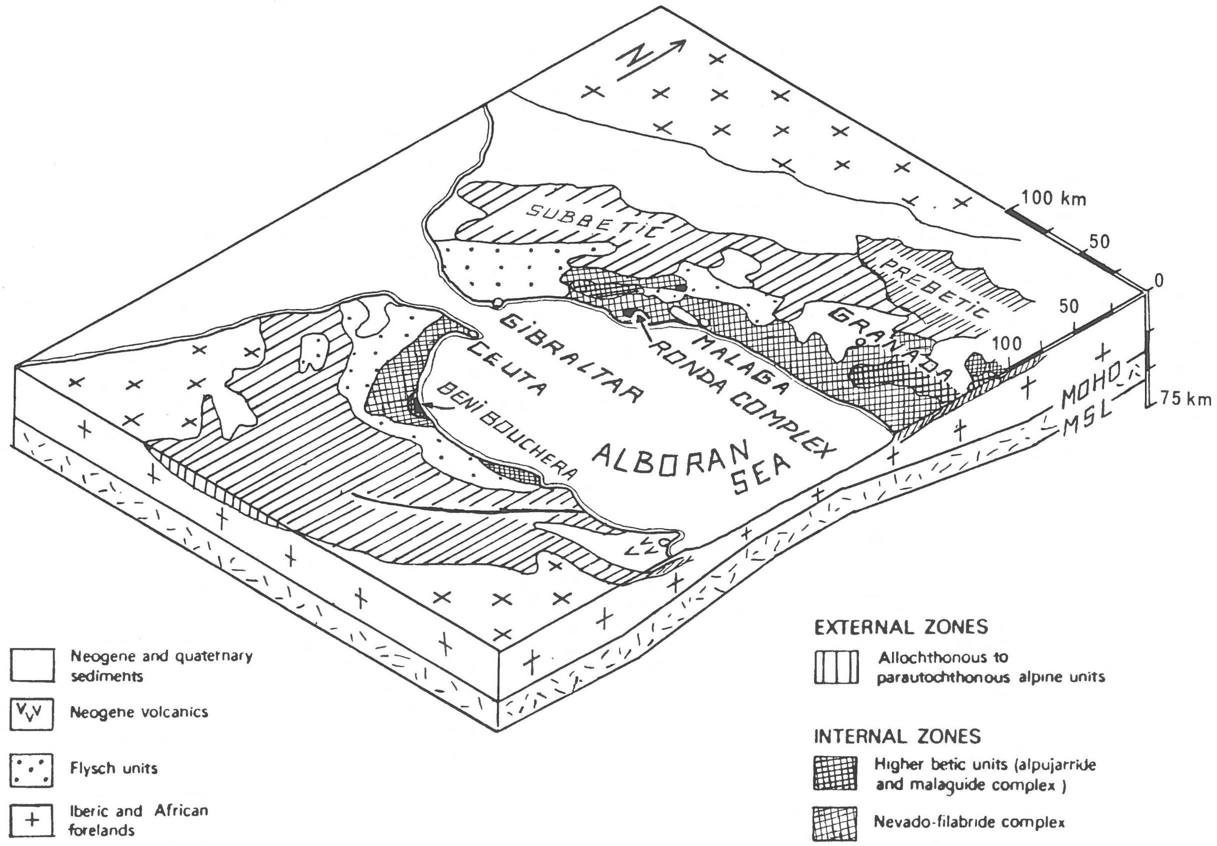

Fig. 1. Isometric block diagram of the mechanically strong upper part of the lithosphere (MSL) of the Alboran Basin and the peripheral Betic-Rif orocline, which comprises nappes translated from a former domal rise in the Alboran Sea. The legend refers to the surface geology. The basis of the MSL is drawn congruous with the MOHO. The bulge of 17 km amplitude and 300 km half wavelength in the normally 30 to 35 km deep MOHO is inferred from the seismic section of Banda et al. (1983). The surface geology is compiled after Vissers (1981, fig. 1).

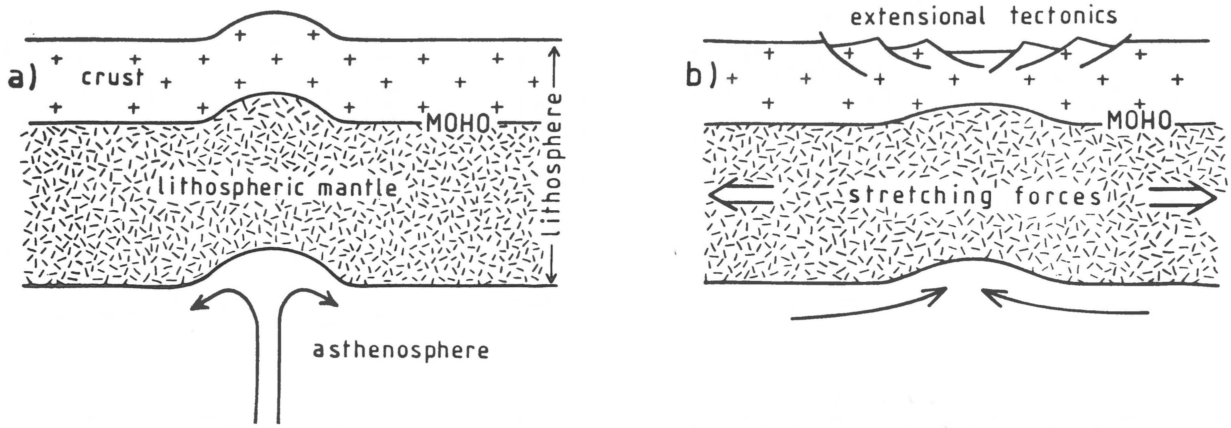

Fig. 2. Basin subsidence as a result of the cooling of deeply emplaced asthenospheric swells. The early swell may be caused by two different mechanisms: a) lithospheric doming above a primary diapiric swell; b) plate reorganisation forces may locally cause large tensile or stretching forces and crustal thinning by extensional tectonics which is initially compensated by a secondary swell of the asthenosphere.

the site of the present Alboran Sea (Torres-Roldan 1979; Platt et al. 1983). Such tectonic unroofing explains the rapid denudation and removes the need for extensive sedimentary erosion before subsidence which would be necessary otherwise. A former Alboran topographic high due to crustal thickening by plate collision suggested by Platt et al. (1983) is unlikely because it cannot explain the topographic reversal from high to depression.

2) Large ultramafic bodies, *i.e.* the Spanish Ronda and Moroccan Beni Bouchera ultramafic complexes, occur high in the crust of the Alboran Region (Fig. 1). Such ultramafic bodies might be expected in association with the forceful emplacement of a large diapiric body in the lithosphere beneath (Fig. 2a). The heavy ultramafics cannot be emplaced by a Rayleigh-Taylor overturn mechanism but must have been driven upwards by a fluid pressure from below (Loomis 1972). The ultramafic bodies would be unnecessary if the Alboran Basin formed due to lithospheric thinning by extension (Fig. 2b).

Numerical doming models

The numerical models for the doming of continental lithosphere of Cloetingh & Nieuwland (1984) are summarised in Figure 3. Important features are the assumptions about the structure and rheological properties of the mechanically strong upper part of the lithosphere (MSL) which extends to a depth corresponding to a lithospheric strength of 50 MPa (see Cloetingh 1982; Cloetingh & Wortel 1982; Cloetingh et al. 1984). The viscosities of the three different layers of the MSL (Fig. 3a) depend only on the temperature profile if a constant strain rate ($\dot{e} = 10^{-14}\,s^{-1}$) is adopted (*cf.* Pfiffner & Ramsay 1982). Consequently, the strength of the MSL is also a function of the temperature. For continental lithosphere it reaches a maximum of 700 MPa at the 35 km deep brittle-ductile transition. The strength decreases rapidly near the base of the MSL (Cloetingh & Nieuwland 1984).

Cloetingh & Nieuwland (1984) distinguished two doming models with different geometries. The temperature distribution near the base of the MSL does not change significantly in one model and the upwarping of the crust is purely mechanical (Fig. 3b). In the other model, the asthenospheric upwelling causes a significant thermal perturbation which changes the rheology of the overlying MSL resulting in thinning by inhomogeneous flow (Fig. 3c).

The minimum uplift required to cause brittle failure in the MSL-layer above either of the two types of lithospheric domes of Figures 3b and 3c was quantified by Cloetingh & Nieuwland (1984). The critical uplift for a particular half wavelength of the bulge in the base of the MSL is given in Figure 4.

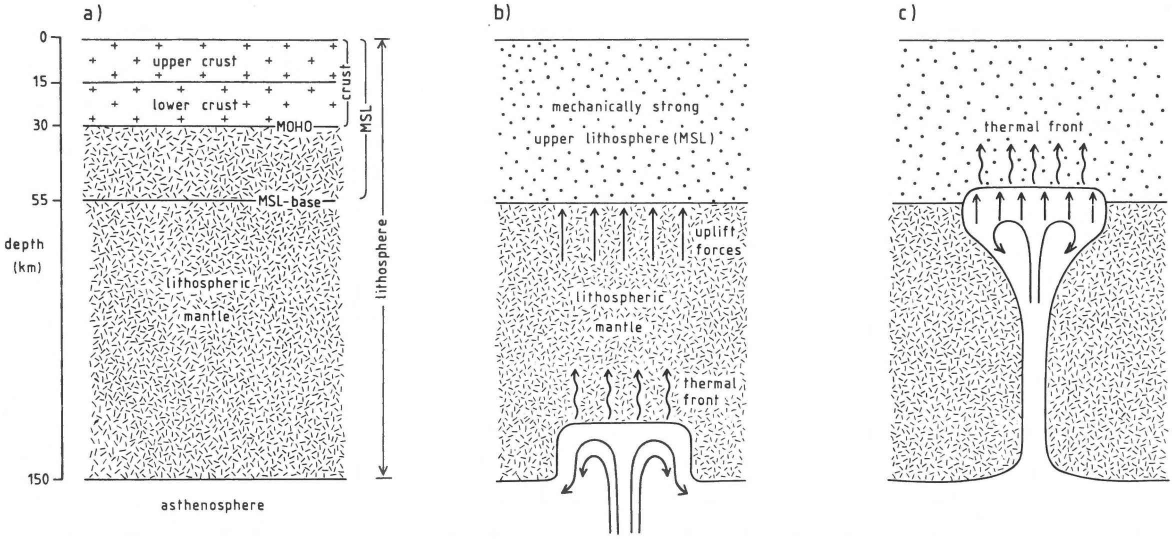

Fig. 3. The numerical models of Cloetingh & Nieuwland (1984) for doming of the continental lithosphere: a) The 150 km thick lithosphere comprises a 15 km thick upper crust, a 15 km thick lower crust, a 30 km deep MOHO, a 120 km thick lithospheric mantle and a 55 km thick MSL. The rheology of the lithosphere is incorporated by assuming an upper crust of quartzite, a lower crust of diabase and a lithospheric mantle of olivine (creep laws appllied: Shelton & Tullis, 1981, for quartzite and diabase; Goetze, 1978, for olivine). b) Immature diapiric bulge of the asthenosphere-lithosphere boundary cannot thermally affect the 55 km thick MSL and causes only mechanical uplift forces. c) Mature diapir penetrating the lithospheric mantle causes subaerial erosion of the MSL and the thermal front attenuates the rheology of the MSL above the area subject to the uplift forces. The continental lithosphere is thinned from 150 km to 50 km.

Application to the Alboran Basin

The thickness of the MSL beneath the Alboran Region before the doming episode is unknown.

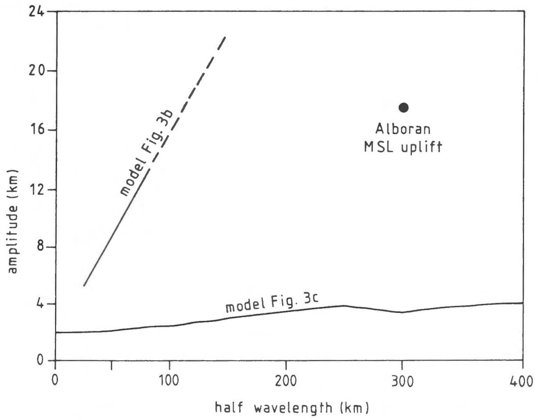

Fig. 4. The minimum uplift required to cause lithospheric failure at the points of maximum flexure in the MSL when domes develop according to the doming models of Figures 3a and 3b. The bulge in the lower boundary of the MSL underneath the Alboran Basin would have an amplitude of 17 km and a half wavelength of 300 km if it is geometrically congruous with the bulge of the MOHO. This plots as a point and indicates that faulting would have occurred if the model of Figure 3c is applied. Faults are absent according to the model of Figure 3b.

However, it is unlikely to have differed much from the general profile of Figure 3a, because the surrounding 'normal' Iberian crust is 30-35 km thick and can also be divided into lower and upper sections at a depth of about 15 km (Banda et al. 1983).

Seismic sounding experiments show a bulge in the MOHO of 17 km amplitude and 300 km half wavelength beneath the Alboran Sea (Fig. 1; Banda et al. 1983). However, the resolution of the profile is insufficient to establish whether or not the lower part of the MSL is thinned. Hence direct observations do not yet allow distinction between the two models for the lithospheric mantle beneath the Alboran Sea (*i.e.*: figure 3b or 3c), nor whether the MSL rheologically attenuated during the doming.

If rheological attenuation had taken place during the domal rise and if the bulge in the MOHO beneath the Alboran Sea were congruous with the 20 km deeper base of the MSL, then the bulge of the base of the MSL would also have an amplitude of 17 km and a half wavelength of 300 km (Fig. 1). Such an MSL-bulge would be represented by a point shown in Figure 4 and implies that extensional faults should transect the floor of the Alboran Basin.

Both the morphology of the subsiding Alboran Basin and its Bouguer gravity anomaly pattern (Bonini et al. 1973) suggest a 300 km long east-west ridge in the top of the mantle. The hypothetical faults therefore should be 300 km long and run east-west along the northern and southern borders of the Alboran Basin cutting through the points of maximum MSL flexure. Since such faults have not been reported, the model of a shallow asthenospheric uplift (Fig. 3b) appears to be most appropriate for the Alboran Basin.

Discussion: the late Alpine tectonic history of the Alboran Region

Research in the Betic-Rif orocline has until now been dominated by the dogma that Morocco and Spain are separated by the collision zone between the African and European plates. The collision between Africa and Europe would have occurred at the site of the modern Alboran Sea about 18 Ma ago in Burdigalian times (Dewey et al. 1973; Bourrouilh & Gorsline 1979). This plate collision has been previously considered as the cause of the formation of the former Alboran topographic high (Platt et al. 1983). However, this is unlikely since the subsequent subsidence of the Alboran Basin needs crustal thinning by stretching, sublithospheric erosion or doming with excessive surface denudation. It cannot be explained by the progressive crustal and lithospheric thickening that would occur between two colliding plates. I therefore introduced an alternative tectonic scenario (Weijermars 1985) of which the details are discussed below.

Between 20 and 30 Ma ago, the Alboran Region may have passed over the small-scale convective upwelling (Weijermars 1985) which occurs nowadays under the Atlas Mountains 450 km to the south-west (Liu 1983). The past position of this convective upwelling relative to the Alboran Region can be inferred from the absolute plate velocity of 1.5 to 2 cm a^{-1} to the NE attributed to both the European and African plates (Minster et al. 1974; Solomon et al. 1975). This convective upwelling could have caused a thermal bulge in the asthenosphere (Fig. 5a) if the movement of the overlying plates was sufficiently slow or if they stopped temporarily (Nakiboglu & Lambeck 1985).

Any rising thermal dome such as the Alboran

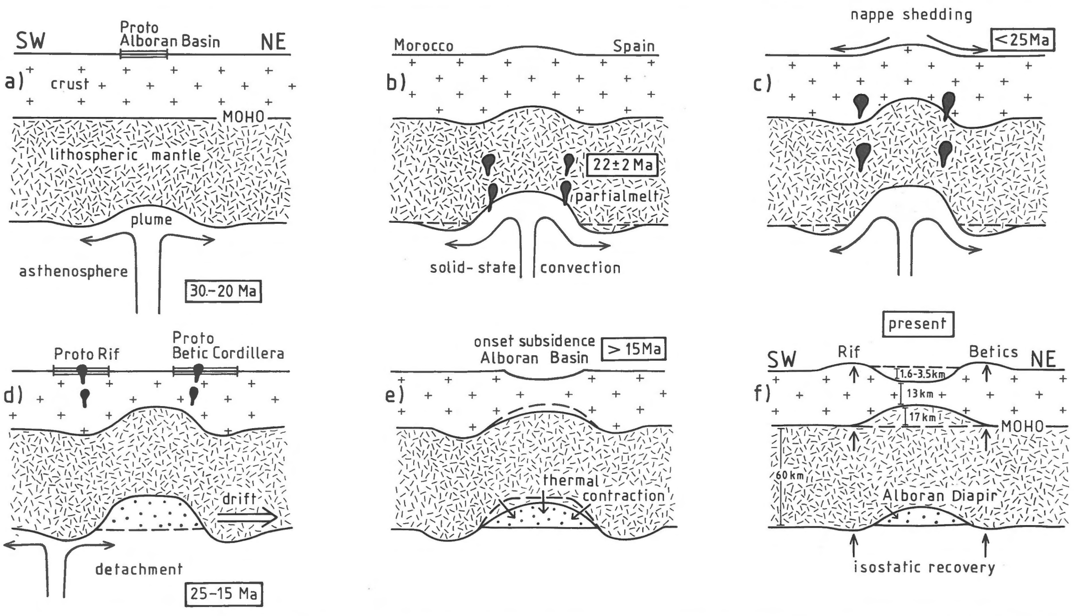

Fig. 5. a-f) Synopsis of the tectonic evolution of the Alboran region since it passed over an upwelling convection current which triggered the rise of a thermal instability at the asthenosphere-lithosphere boundary about 30 Ma ago. See discussion.

Diapir is likely to have been emplaced mainly in the solid state. The peridotitic intrusions exposed in the crust around the Alboran Sea are visualised as having risen as magma globules due to partial melting of the Alboran Diapir (Fig. 5b). Pyroxenite of the Beni Bouchera ultramafic complex is derived from a cotectic melt at 1873 K and a pressure of 2850 MPa (Kornprobst 1970) which implies melting at a depth of about 100 km. Emplacement of the Ronda Complex into the crust is dated at 22 ± 4 Ma by the Rb-Sr age of associated anatectic granites (Priem et al. 1979). The 22 ± 2 Ma Sm-Nd age of a garnet clinopyroxenite in the Ronda Complex probably also indicates a cooling age (Zindler et al. 1983). However, the possibility remains that this age represents the initial melting deep in the lithosphere (Zindler et al. 1983). The evolution of opinions about the age of Beni Bouchera and Ronda ultramafics has been reviewed by Loomis (1975) and Westerhof (1977).

The lithospheric dome above the Alboran Diapir shed gravity-spreading nappes to the north and south as it rose (Fig. 5c; *e.g.* Weijermars 1985; Platt et al. 1983). The Alpine Betic-Rif orocline is interpreted to consist of nappe sheets in which the youngest rocks are 25 Ma old Oligocene sediments (*i.e.* , in the Malaguide Complex of the Betic Zone, see Egeler & Simon 1969 a,b). The Alboran surface bulge must obviously be younger than the youngest rocks translated by the nappe shedding it caused. It therefore cannot be older than 25 Ma.

The Alboran Diapir appears to have bulged the overlying MSL but not to the extent of penetrating the lithospheric mantle to thermally affect the rheology of the MSL. Otherwise 300 km long east-west faults would have marked the northern and southern boundaries of the Alboran Basin and these are absent. The Alboran Diapir is therefore interpreted to have ceased rising at an immature stage.

The tectonic unroofing stopped when the Alboran Region ceased to rise. This presumably happened when the Alboran Diapir was detached from the convective upwelling as the Alboran Region was carried to the NE (Fig. 5d) at a velocity of 1.5 to 2 cm a^{-1} (*cf.* Minster et al. 1974). This mechanism has been termed lithospheric shifting in a geodynamic scenario for the evolution of the Alps (Vlaar & Cloetingh 1984).

The *hot* asthenosphere of the Alboran Diapir had temporaily replaced *cold* lithosphere and therefore started to cool when detached from its heat source; cooling caused basin subsidence. DSDP cores and seismic reflection profiles of the Alboran Basin suggest the occurrence of 15 Ma old Serravalian sediments (Mulder & Parry 1977). This implies that the Alboran Diapir was already cooling as early as 15 Ma ago (Fig. 5e).

Rise of the Alboran Diapir produced a complementary downwarping of the asthenosphere-lithosphere boundary at its margins (Figs. 5b and 5c). These downwarps of the asthenosphere-lithosphere interface peripheral to the Alboran diapir, which itself has a positive gravity anomaly of 50 mgals, are still visible as negative gravity anomalies (of -100 mgals) over the Betic-Rif orocline (Bonini et al. 1973). The rise of the Betic-Rif orocline could be due to (still active?) isostatic uplift of the downwarped lithosphere (Fig. 5f) which would have begun when the Alboran Diapir ceased to grow.

Alpine nappes are exposed in the Betic mountains in east-west striking ranges which define a system of basins comprising Neogene sediments (Kampschuur & Rondeel 1975). Such a relationship suggests that the Betic nappes were refolded and faulted during a Neogene tectonic event which may be associated with isostatic recovery (Fig. 5f). Large angular uncorformities locally observed between Tortonian and Messinian sediments in the Internal Zone (Th. B. Roep, pers. communication 18-01-85) and the thrusting of the Subbetic over the Prebetic (Fig. 1) in Tortonian times (Hoedemaeker 1973; Jerez-Mir 1973; Azema 1977) suggest that the tectonism peripheral to the Alboran Diapir was most intensive towards the end of Tortonian times, about 7 Ma ago.

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was inspired by the Betic Cordilleras after four months field mapping in 1978, when I started as a graduate student at the University of Amsterdam.

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