

CHAPTER 7 - SYNTHESIS

The different topics of this thesis, giving it the title “Cenozoic tectonic evolution of the Iberian Peninsula: causes and effects of changing stress fields” have been addressed in the preceding chapters. This short final chapter integrates these aspects and arrives at some general conclusions.

Cenozoic tectonic evolution of the Iberian Peninsula

A detailed but fairly large-scale compilation of published data and new data on the geological evolution of the Iberian Peninsula reveals several stages of increased tectonic activity interrupted by periods of tectonic quiescence. In general, the Iberian Peninsula is going through a tectonically very active period since opening of the Mid-Atlantic Ocean started. After a long period in the Mesozoic, during which extension was the dominant mode of deformation, the Early Tertiary and the neotectonic M. Miocene to present-day were periods of major compression in the peninsula. During these periods, the main topographic features of the Peninsula (Pyrenees, Betics, Cantabrian Cordillera and Central System) developed. Active extension occurred contemporaneously along the eastern margin of the peninsula.

The large-scale tectonic evolution of Iberia was known (e.g. Ziegler, [1988]), in a limited number of time slices. This thesis adds previously unavailable temporal and spatial detail to this evolution by a combination of active tectonic structures, stress fields, sedimentary environments and reconstruction of blocks as presented in Chapter 4 and 5. Based on this combination of different types of geological data, more coherent reconstructions in narrower time slices were obtained. The changes between the different episodes, the correlation between the mentioned large-scale tectonic processes and local intraplate deformation and their extent in time and space can be readily observed in the presented detailed reconstructions.

Changing stress fields

Reconstructed tectonic activity, such as (a) horizontal and vertical motions, (b) polyphase (re)activation of faults, (c) (de)formation of sedimentary basins and (d) erosion and sedimentation, is indicative of the stress field and changes of this stress field in time. Based on both fault-slip data and above-mentioned general tectonic activity, it was possible to reconstruct the stress field in reasonable detail for several time spans. The new data on the evolution of the stress field presented in this thesis (Chapter 3) improved the data coverage in time and space. Analyzing the development of the stress field through time in certain regions (compare consecutive Figures in Chapters 4 or 5) reveals considerable spatial and temporal variation of the stress field in the Iberian Peninsula throughout the Tertiary.

In general, the Iberian portion of the Eurasian plate was dominated by compressive regimes with Sh_{max} orientations ranging between NW to NE. The dominant stress regimes range from uniaxial compression to compressive strike-slip; the latter occurred when compression was reduced or when sources of extension increased. Compression can be easily transmitted over thousands of kilometers: the collision between Iberia, Eurasia and Africa is observed throughout the entire microplate. On the other hand, tension is not affecting the entire peninsula. The extension in the Valencia Trough area only affects the eastern border of the peninsula, the effect is small in the internal parts of

the peninsula (change in stress field from nearly pure compression to strike-slip): normal faulting accommodates extension over short distances. This is due to the smaller strength of rock under tension as compared to compression [Ranalli & Murphy, 1987]. Although these general patterns can be distinguished, the local state of stress is strongly dependent on time and place and changes gradually from one stress regime to the other (see e.g. Simón-Gómez [1986]). A concept of different and subsequent 'tectonic phases' seems not to be valid over the entire plate; in the same time interval (e.g., the present), the stress regimes can change from pure extension (Central and northeastern Iberia) to pure compression (Betics) within a few hundreds of kilometers.

In order to analyze the evolution of the stress field in even more detail, additional information would be required on:

- Exact dating of Tertiary sediments especially in the smaller basins, would improve the data set and enable lateral extension of the areas for which trajectories of $S_{h_{max}}$ can be reconstructed. Better age-constraints for the sediments and thus for the stress fields observed in these sedimentary sequences could be provided by application of magnetostratigraphic studies to these basins, as done for the Teruel Basins [Krijgsman *et al.*, 1996], or by biostratigraphic studies using mammal teeth [Daams *et al.*, 1996].
- Additional paleostress analyses are required, particularly for periods and areas that have been studied less extensively. The first is limited by the occurrence of sediments of the period under consideration (Cenozoic sediments in this case). However, in areas for which the recent deformation is well documented, stress regimes responsible for older tectonic events could be established. Due to overprinting by younger deformation that obscures earlier tectonics, it might be difficult to do so based on structural data only. This requires integration of additional data from other fields of geology. An example is the Spanish Central System, for which the Middle Miocene to present-day stress field was well studied [De Vicente *et al.*, 1996b]. By integration of different aspects (structural geology, sedimentary geology and fission track analysis) the Paleogene deformation has been reconstructed (section 3.2). Moreover, this integrated study provided new details about the most important mechanisms and deformation patterns in the late stage tectonic evolution of the area. Especially in Portugal data coverage could be improved by analyzing the stress field in northeastern (Vilariça and similar basins), eastern Portugal (Castelo Branco Basin), south and east of Coimbra, southcentral Portugal (Sado Basin and eastern Lower Tâmega Basin). In Spain data from the central west would be of great value.

A difference between local observations of stress orientation in the field and general patterns of stress at large scale will always exist, due to scale differences. However, features that have an important effect on local stress state (large faults, crustal inhomogeneities) appear to have minor influence on larger scale. Therefore, if abundant data on local stress observations is available, more reliable regional patterns may be constructed.

Causes of changing stress fields

The peculiar situation of Iberia, being located at the corner of active plate boundaries throughout the Cenozoic, provides an unique setting in which the relation between the changing plate boundary conditions and the stress field in the peninsula is relatively direct and straightforward. It offers possibilities to create numerical models to test the effect on the stress field by changing the boundary conditions. The main stages of intraplate deformation mentioned can be related to plate tectonic activity/plate interactions, as shown in this study. Conversely, changes in the stress field can yield

additional constraints on the same topics. For example, previously undocumented plate boundary activity along the southern boundary of the Iberian plate in the Early Tertiary is inferred in this thesis, based on the modelling results. In this way, local observations of the stress field can offer indications on the activity of plate boundary processes.

The Iberian plate got into the gradually closing Europe-Africa couple with the late Cretaceous onset of counter-clockwise rotational convergence of Africa-Arabia and Eurasia. The young, and thus weak, Biscay-Pyrenean rift wrench zone was compressionally reactivated during the Campanian. Increasing space constraints led to the end Cretaceous activation of the Balearic-Corsica-Sardinian subduction zone, along which older oceanic (Tethys) crust was consumed. The early Tertiary deformation resulted from collision between Eurasia and the Iberian microplate. The origin of the opening of the Mediterranean is matter of debate, but is related to subduction of Tethyan oceanic crust. During the last 12 Ma, collision between Africa and Eurasia is the most important source of stress. Internal deformation by folding [Cloetingh *et al.*, 2001] and faulting [De Vicente *et al.*, 1996b] seems to be the primary mode accommodating the progressive convergence between Africa and Eurasia. Africa hereby seems to be passing Iberia along the right, peeling off Morocco and pushing Iberia westward [Andeweg & Cloetingh, 2001]. The results of the models for the western margin of Iberia (section 2.4), suggest that high levels of stress in this region will increase even further. With the continuing opening of the Atlantic, this area might be prone to the start of subduction in future.

First order stress sources (plate boundary related) can create large-scale topographic features, which are effects of the stress field. In this way, the latter become 2nd order stress sources, which cause local modification of the regional stress field.

Although the models can predict several of the observed stress fields based on rather straightforward principles, additional and detailed knowledge about (a) the crustal composition and configuration (density and independent Moho depth observations), (b) paleo-topographic indications, (c) motion, interaction and timing of the several tectonic blocks in the Mediterranean area and (d) the geometries of these blocks, could further improve the models and their results.

In general I conclude that the new concepts in numerical modelling of stress fields presented in this thesis can help to gain a better understanding of the relationship between plate tectonic processes and the local state of stress. This can lead to information about regions that experience high levels of stress and are prone to seismic activity. The scale of the models presented in this thesis, however, is too large to predict local seismic activity. Nevertheless, the results of the large-scale models could be used as a first order approximation of the stresses to be applied as boundary conditions to smaller scale experiments where better constrain on the 3D-geometry is available.

APPENDIX A

In this Appendix is explained how ridge push force can be calculated. Ridge push forces have been included in two ways in the modelling presented in this thesis. In the first case, forces are applied as line forces perpendicular to boundaries of the models to represent the opening ocean. In the latter case, Ridge Push forces are applied in a physically more correct way as intraplate forces.

Ridge push as a line force

The Ridge push force (F_{rp}) is related to elevation -with respect to fully cooled sea floor- (e) and thickness (S) of the lithosphere. Both parameters are age-dependent and therefore ridge push forces are related to the age (t) of oceanic crust in a oceanic basin. The classical models applied to calculate age-depth relations for cooling oceanic crust are the Plate Model and the Boundary Layer model (see Parsons & Sclater [1977] for details). The Boundary Layer model yields better fits to observations for ocean floor younger than 70Ma, whereas the Plate Model is generally preferred for oceanic crust older than 70Ma. The lithospheric age dependency arises from the fact that oceanic crust is cooling from the moment it is being formed at a Mid Oceanic Ridge. With increasing age (moving away from the MOR) heat loss is gradually decaying, which results in a finite increase in density of oceanic crust and a maximum depth for fully cooled oceanic crust. Combining the age-depth relationship with this depth for fully cooled oceanic crust (6400m below sea level [Parsons & Sclater, 1977]) results in the next formulas to calculate elevation [$e(t)$, in meters] with respect to fully cooled sea floor:

Young ocean (<70Ma): $e(t) = 3900 - 350\sqrt{t}$

For older lithosphere (>70Ma): $e(t) = 3200^{-t/62.8}$

where t is age in Ma. For $t = 0$ Ma, the elevation with respect to fully cooled sea floor thus becomes 3900m (3900m above 6400m below sea level = 2500m below sea level). Once this theoretical depth of the top of the seabed is established, the thickness of the lithosphere can be determined using the standard Boundary Layer Model with a constant basal heat flux [Wortel, 1980]

$$S(t) = e(\rho_m - \rho_w) / (0.423\rho_m \alpha T_m)$$

where ρ_m and ρ_w represent density of mantle material and water respectively, T_m is the temperature at the lithosphere-asthenosphere boundary and α is the thermal expansion coefficient.

The age-dependency of ridge push is expressed by

$$F_{rp}(t) = ge \times (\rho_m - \rho_w) \times (S/3 + e/2)$$

where g is the gravity acceleration, ρ_m and ρ_w represent density of mantle material and water respectively.

From the formulas describing the age-depth relationship for cooling oceanic crust it is evident that for increasing age the elevation with respect to fully cooled seafloor

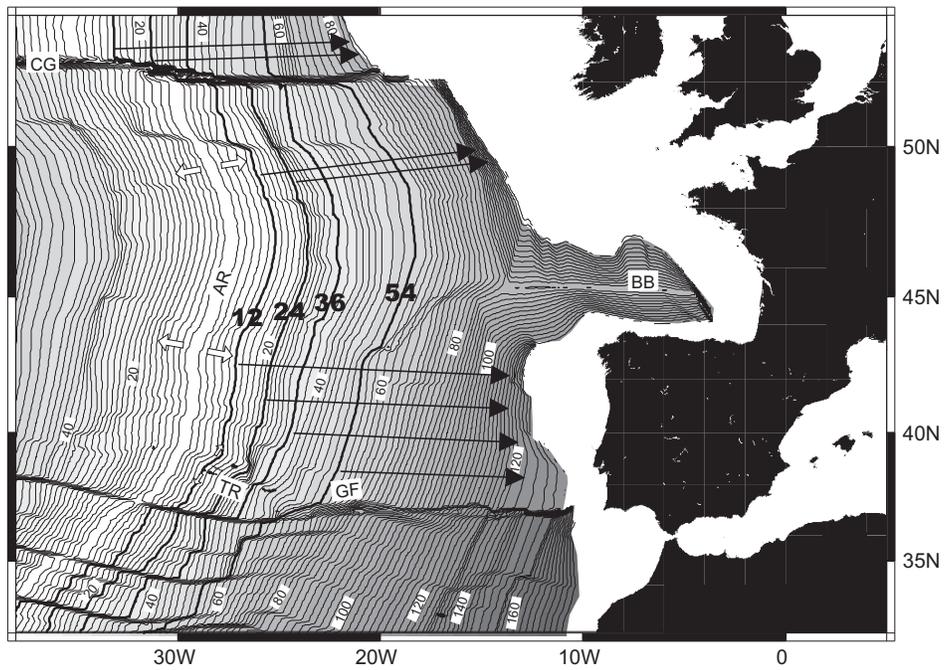
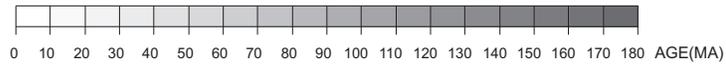
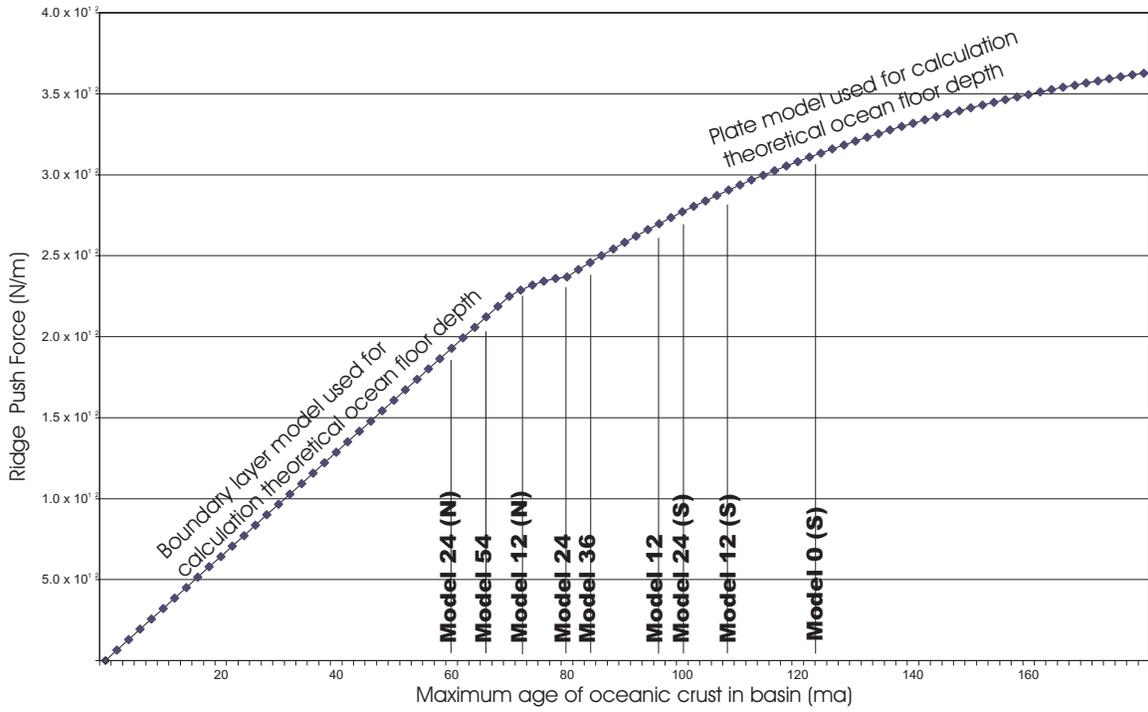


Figure A.1: Ridge push calculated for several segments in the North Atlantic for the different reconstruction times.

decreases. Therefore, the contribution to ridge push of oceanic crust older than ~120Ma can be neglected.

In the models presented in Chapter 5, F_{rp} has been incorporated in the first runs as a line force at the boundary of the model and have been calculated using the above-mentioned formulas and a modification of the method as applied by Janssen [1996]. Figure A1 shows that using these calculations, the ridge push exerted by the MOR at 54Ma was $\sim 1.0 \times 10^{12}$ N/m less than the value of $\sim 3.0 \times 10^{12}$ N/m for the same segment at present-day. Because the models in Chapter 6 represent the tectonic situation of Iberia since the Tertiary, it is important to include age-dependency of ridge push forces.

Ridge push as intraplate forces

In the model of Coblenz *et al.* [1994], which is described in section 2.3, ridge push force is calculated as the result of buoyancy forces related to the cooling lithosphere. This method is treating the ridge push physically more correct to be an intraplate force.