CHAPTER 1 - INTRODUCTION TO THE TECTONIC SETTING OF THE IBERIAN PENINSULA

Differential motions between tectonic plates create zones of intense deformation along their boundaries. In most cases large parts of the deformation is localized along these borders but forces related to the plate interactions also transmit stresses to the interior of the plates [Cloetingh, 1992]. If these intraplate stresses become sufficiently high to overcome the strength of the material, of which the plate is composed, then intraplate deformation related directly to plate interaction will occur (e.g. Ziegler *et al.* [1995]). Thus observations of intraplate deformation provide indirect information about the magnitude and type of activity along the plate boundaries. For example, interaction between the African/Arabian and Eurasian plates has generated a broad collision zone, the Himalayan-Alpine chain, running from SW Europe to SE Asia. For the Iberian Peninsula, located at the western end of this zone of convergence, the gradual opening of the North Atlantic is the most important factor in the complex pattern of differential motion between Eurasia, Africa, and Iberia over the past 120Ma. A condensed outline of the geological evolution of the Peninsula is provided in this chapter. The Cenozoic geological evolution will be elaborated in great detail in Chapter 4 and 5. The reader interested in more details on the pre-Cenozoic evolution is referred to references cited.

General geological evolution

The time frame this thesis on focuses is the Cenozoic (65Ma ago to present-day). Sedimentary basins that developed during this period (Figure 1.1) cover large parts of the Iberian Peninsula. The basement underlying these basins in the western part of Iberia consists mainly of Hercynian metasediments and igneous rocks; whereas in the eastern part the basement is composed of Mesozoic series. Pre-Tertiary structures (faults, suture zones) played a major role in the Cenozoic deformation of the Peninsula.

Pre-Mesozoic

Iberia was still attached to Armorica (N. France) prior to the Late Mesozoic opening of the Bay of Biscay [Garcia Mondejar, 1988]. At the time of the Hercynian orogeny (E. Carboniferous – L. Permian) Iberia was part of the Variscan arc running through Belgium, Northern France, Southern England [Ziegler, 1989]. The large-scale structural domains such as suture zones, that formed during this orogeny, can still be traced in the basement of western Iberia and play an important role in later deformation phases [Stapel, 1999]. The origin of the arc-shape of the Hercynian structures is still a matter of debate; see Dias & Ribeiro [1995] for a discussion of several models. Doblas *et al.* [1994] present a detailed description of the tectonic evolution from the Variscan to Alpine tectonic stages.

Mesozoic

Progressive opening of the Atlantic Ocean between the Americas and at first Africa, later Iberia and finally Europe caused large differential motions between these continents (see Figures 1.2a and b). Active extension resulted in a stage of major rifting during the Mesozoic as documented by extension on all of the margins of Iberia at one time or another:

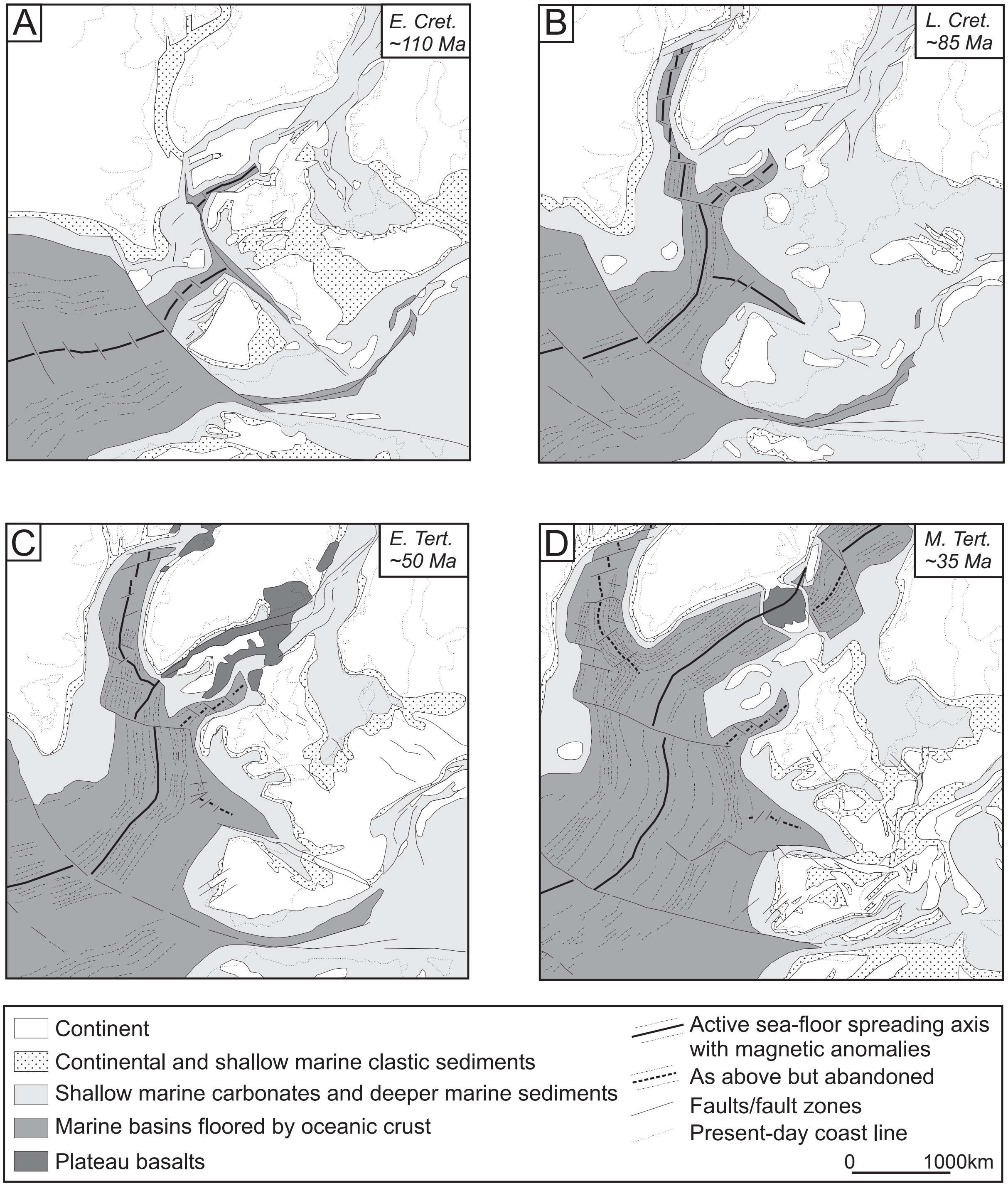

Figure 1.2
Progressive opening of the North Atlantic caused several periods of differential motion between the Eurasian, African, and Iberian plates. A = Early Cretaceous (~110 Ma), B = Late Cretaceous (~85 Ma), C = Early Tertiary (~50 Ma) and D = Middle Tertiary (~35 Ma). See text for explanation, after Ziegler (1988).

(a) The western Iberian margin. Subsidence history prior to and after break up of the western Iberian margin have been described by amongst others Stapel *et al.* [1996a] and Rasmussen *et al.* [1998]. Two major stages of rifting have been recognized: the first during Late Triassic and the second in Early Jurassic. This last stage of rifting is followed by an important regional hiatus, probably related to opening of the Central Atlantic Ocean coinciding with the beginning of oceanic spreading at the Iberian Abyssal Plain, dated at 126Ma [Whitmarsh & Miles, 1995]. A third important phase of extension started in the Late Jurassic and a final phase of extension, poorly dated in the shallow parts of the margin, occured during the latest Late Jurassic to earliest Early Cretaceous. The final separation of Galica Bank and Flamish Cap has been dated at around 118 Ma. Subsequently, the area experienced regional (thermal) subsidence [Stapel, 1999].
(b) The eastern Iberian margin. The Alpine Tethys opened during the Mesozoic and, therefore, the eastern part of Iberia was locus of important rifting as well: the Iberian Basin was formed [Salas & Casas, 1993]. The westernmost traces of Mesozoic sediments related to this rift can be found as far in the continent as the Central System, where patches are preserved in 'pop-downs' or along major structural contacts [De Vicente *et al.*, 1996c]. This area constituted the western extreme of the basin: thickness of the Mesozoic series increases rapidly eastward from the eastern Central System towards the Iberian Chain/Sierra de la Demanda. Subsidence analysis of the basin [Van Wees *et al.*, 1998] as well as thermal modelling [Fernàndez *et al.*, 1995] shows several rifting stages in eastern Iberia, the major ones being Permian-Early Triassic and Late Jurassic-Early Cretaceous of age.
(c) The southern Iberian margin. From Triassic until early Liassic (E. Jurassic) the southern Iberian margin was the locus of a large shallow-water carbonate and clastic shelf development. Rifting along this margin started approximately in the Toarcian (E. Jurassic, 190Ma), resulting in a break-up of the carbonate shelf and a deepening of a part of the basin. Depressions bounded by listric faults were filled with synrift sediments. Active rifting changed to post-rift during the Early Malm (L. Jurassic, 160Ma) and related thermal subsidence lasted during the rest of the Malm and Cretaceous [Vera, 1988]. This change in rifting activity can be correlated to the development of sinistral transtensional motion along the transform zone between Africa and Iberia [Bakker *et al.*, 1989; Biermann, 1995], along which local oceanic pull-apart basins were formed [Vera 2001].
(d) The northern Iberian margin. Here extension resulted in opening of the Bay of Biscay towards the end of the Early Cretaceous, which continued until 85 Ma [Ziegler, 1988].

Apart from widespread rifting along all of the margins of Iberia, another effect of the onset of active seafloor spreading in the Azores part of the North Atlantic (~126 Ma) and the Bay of Biscay (~115 Ma) (see Figure 1.2a) was an anti-clockwise rotation of Iberia with respect to Eurasia [Savostin *et al.*, 1986]. This induced left lateral motion between Iberia and Eurasia coincided with collision and Late Cretaceous subduction of the Ligurian Basin onto the eastern side of Iberia [De Jong, 1990], developing the stack of the Betic nappe units [Biermann, 1995]. Towards the end of the Mesozoic at about 85 Ma (see Figure 1.2b), the opening of the Atlantic propagates between Greenland and Ireland (first along the subsequently abandoned Rockall Trough), leaving the Bay of Biscay as a failed rift [Ziegler, 1988; Srivastava *et al.*, 1990]. The new dynamic setting led to a clockwise rotation of Eurasia with respect to Iberia causing approximately north-south convergence. This resulted in inversion of the northern margin of Iberia, even developing into northward subduction/underthrusting of Iberia (starting in the Campanian [Puigdefàbregas & Souquet, 1986]), creating the Pyrenees.

Cenozoic

In contrast to the Mesozoic, the Tertiary and Quaternary in the Iberian Peninsula are periods dominated by compressional deformation. Deformation related to the closure of the Bay of Biscay-Pyrenean zone progressed to the west through time causing inversion of Mesozoic extensional basins [Garcia Mondejar, 1988]. Development of the Cantabrian Cordillera was related to a short-lived southward subduction of the previously formed oceanic crust in the Biscay region during latest Cretaceous to Early Eocene [Boillot & Malod, 1988] and ongoing convergence. The termination of this subduction coincided with the separation of the rotation poles of Africa and Iberia (with respect to Eurasia) as proposed by Savostin *et al.* [1986] at around 54 Ma (see Figure 1.2c). The start of limited differential motion between Africa and Iberia can be related to the first occurrence of large basalt flows on the western side of Greenland (pointing at the onset of seafloor spreading in this area) and is reflected in the pattern of ocean floor ages. Stresses related to the collision along the northern margin of Iberia and Eurasia were transmitted to the interior of the Iberian plate and resulted in major inversion of the Iberian Basin (see Figure 1.2d), forming the Iberian Chain [Alvaro *et al.*, 1979], Sierra de la Demanda and Sierra de Gredos (western Spanish Central System).

Final amalgamation of Iberia to Eurasia at around 30 Ma coincided with a major change in active plate boundary: the left lateral Azores-Gibraltar zone south of Iberia [Srivastava *et al.*, 1990] is activated. Africa continued moving eastward with respect to Eurasia (including Iberia), causing an active left lateral motion plate boundary to the south of Iberia and contributing to the opening of the Valencia Trough and Balearic Basin to the east [Sabat *et al.*, 1995]. Extension in this region started as early as Oligocene on shore in southern France (related to the Rhine-Bresse Graben system) and shifted progressively southwestward, beginning by E. Miocene (23-20 Ma) in the Alboran domain [Sanz de Galdeano, 1996]. The driving process behind the development of this extensional basin is still a matter of debate, see Ziegler *et al.* [2001]. Opening of the Provencal and Valencia basins has been related to the subduction of African plate beneath the Iberian-European plate [Roca, 2001]. However, the system is located at the prolongation of the Cenozoic rift system of western and central Europe, that cannot be related to back-arc extension [Ziegler, 1994; Ziegler *et al*. 2001].

A change in direction of convergence from NNW to NW between Africa and Eurasia in Tortonian [Mazzoli & Helman, 1994] leads to the major development of the Betics. Inversion tectonics is observed in the interior of the Iberian Plate in the Spanish Central System [De Vicente *et al.*, 1996c] and in L. Tortonian-Messinian in the Alboran Basin [Lonergan & White, 1997]. Seismic activity in Pliocene and Pleistocene in central Iberia has been considerable [Giner Robles *et al.*, 1996; Rodríguez Pascua, 1997]. A high level of internal deformation is demonstrated by crustal scale folding [Cloetingh *et al.*, 2001], large-scale Pliocene uplift of several hundreds of meters in coastal areas [Janssen *et al.*, 1993], present-day seismicity [Buforn *et al.*, 1988] and the development of new crustal shear zones in the Alboran basin [Andeweg & Cloetingh, 2001]. All of these effects accompany the ongoing convergence between Africa and Iberia [Argus *et al.*, 1989].

A very detailed description of the Cenozoic tectonic evolution of the Iberian Peninsula and the western Mediterranean will be presented in Chapter 4 and 5.