CHAPTER 4 - PALEOGENE GEOLOGICAL EVOLUTION OF THE IBERIAN PENINSULA AND WESTERN MEDITERRANEAN

Introduction

The European Alpine System is a classic zone of continent-continent convergence and collision. Numerous studies over the last decades addressed the tectonic and paleogeographical evolution of the closure of the Tethys Ocean and the evolution of the Alpine orogeny (amongst others [Savostin *et al.*, 1986]; [Biju-Dival *et al.*, 1977]; [Dewey *et al.*, 1989]; [Yilmaz *et al.*, 1996]; [Ziegler, 1988] and [Stampfli *et al.*, 2000]).
Southern and Central Europe is characterized by large-scale inversion structures, arcuate fold-and-thrust belts (Western Alps, Carpathians, Betic-Rif), rotating crustal blocks, and limited subduction and extension within an overall convergent setting. Many authors have studied in great detail the puzzling contemporaneous development of all these features related to the evolution of the Western and Central Mediterranean. The wealth of observations has provoked the proposition of a wide range of ideas, models, concepts and hypotheses. The general outline of plate tectonic evolution in the Mediterranean region has been the topic of several studies over the last decades (amongst others [Yilmaz *et al.* 1996]; [Ziegler, 1988]; [Dercourt *et al.*, 1986]; [Dewey *et al.*, 1989]). In spite of the large amount of publications on the development of the region through time, all of the publications turned out to not have the temporal or spatial accuracy required for my purpose. Yilmaz *et al.* [1996] does not include maps for the period between 33.5 to 10.5Ma, during which major plate reorganizations occurred in the region. Some are at a very large scale [Stampfli *et al.*, 2000] or concentrate on the central, oceanic part of the system [Dewey *et al.*, 1989] and do not provide information on the Iberian Peninsula. Others pay attention only to a part (post-25Ma) of its Tertiary evolution [Gueguen *et al.*, 1998], or on paleoenvironments [Dercourt *et al.*, 1993].

Therefore, a compilation, which is presented in this Chapter and Chapter 5, was made from data from literature and additional own unpublished data. The compilation will be presented in maps showing (a) the paleo-position of present-day coastlines, (b) active tectonic structures (faults/folds), (c) sedimentary facies provinces (eroded continent/continenal deposition/shallow marine/deep marine/eroded continent) (d) Paleostress results from kinematic indicator data and stress trajectories (direction of maximum horizontal compression) and for a few time spans (e) estimated paleo-topography and vertical motions. The many studies mentioned before have been very useful and convenient as starting points for this compilation. In contrast to the Iberian Peninsula, my personal knowledge of e.g. Northern Africa and Southern France is limited mainly to literature. Thanks to several excellent overviews of the tectonosedimentary evolution of regions outside the Iberian Peninsula (e.g. Sissingh [2001] for SE France or Wildi [1983] for Northern Africa), the study area could be extended. This incorporation of the areas surrounding the Iberian Peninsula is relevant to understand the large-scale tectonic processes that caused the intraplate deformation of the Iberian Peninsula and to put its evolution in a broader tectonic context.

The reconstructions presented should be regarded as a state of the art reference database, which will guide the interested reader to more detailed descriptions of regions and features rather than an authorative and final model for the evolution of the region.

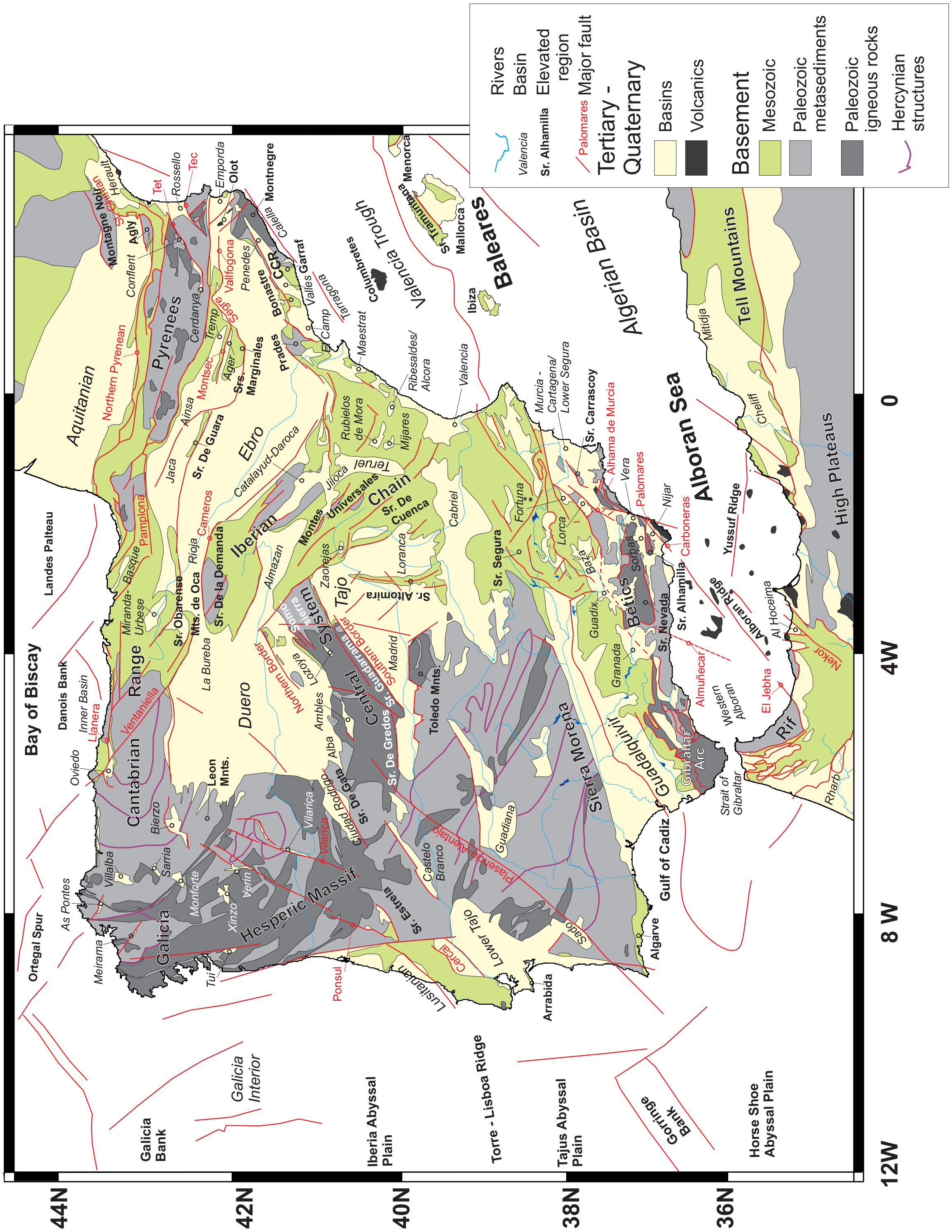


Figure 4.1
Overview of the Iberian Peninsula with the names of basins, major faults and regions that are used in the description of the geological evolution of the area.
(see www.geo.vu.nl/~andb/iberia for full color version)

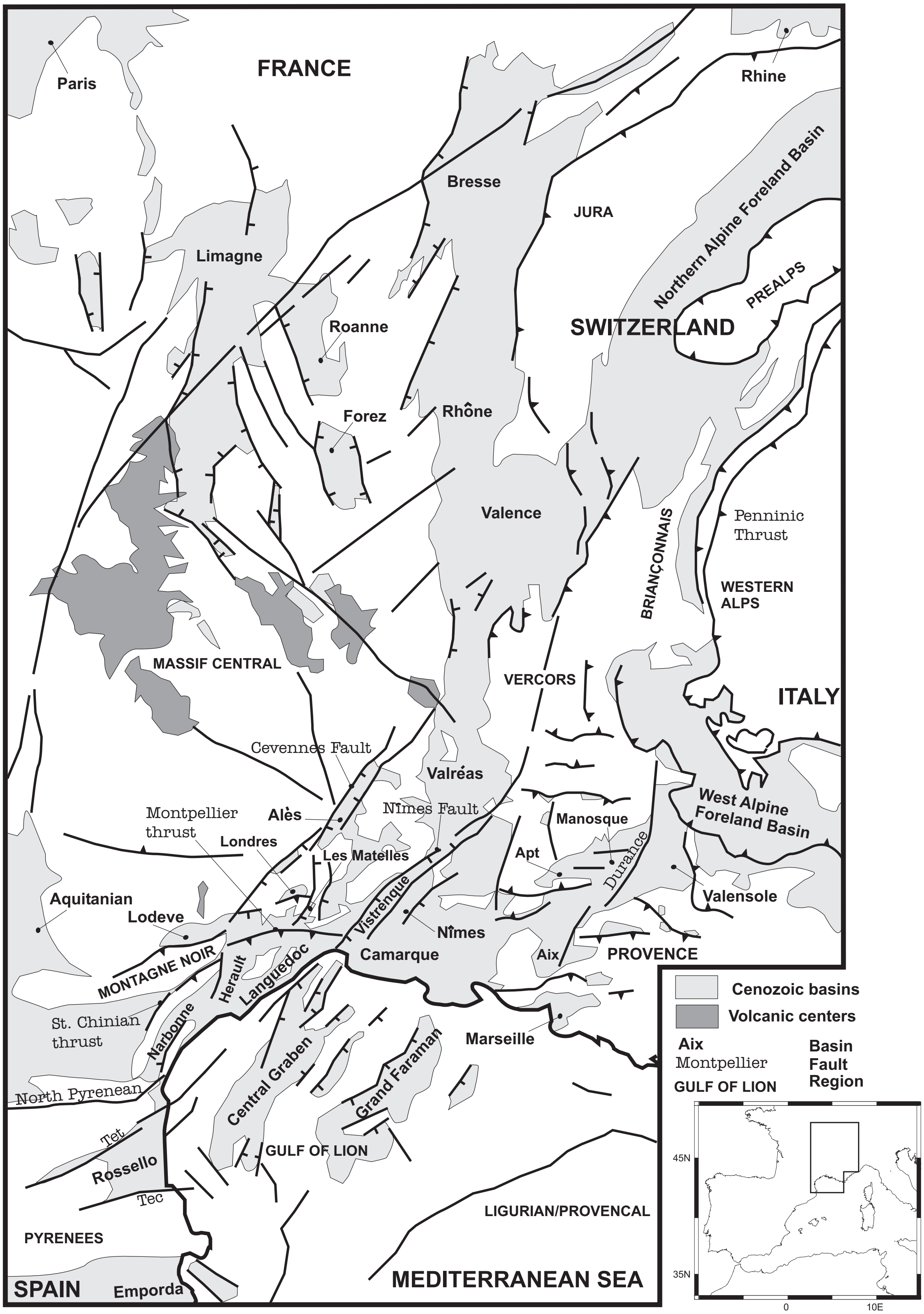


Figure 4.2
Overview of the Southern France with the names of basins, major faults and regions that are used in the description of the geological evolution of the area. Compiled from: Sissingh [2000], Sérrane [1995]; Vially and Trémolières [1996]; and Roure and Coletta [1996].

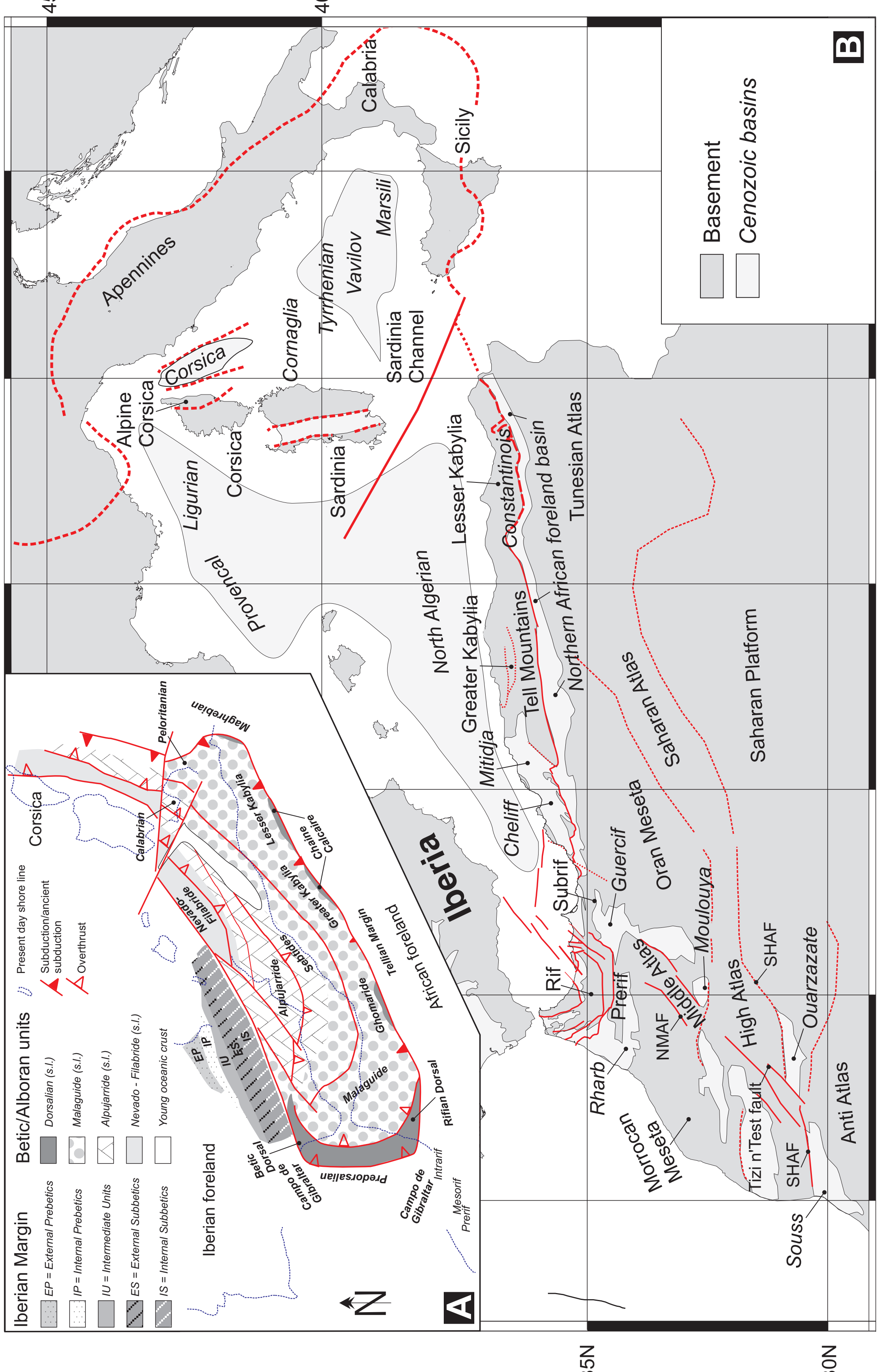


Figure 4.3
Panel A: ~Middle Miocene reconstruction of the Alboran/Betic block and surrounding margins. Division after Sanz de Galdeano [1990] and Fontboté and Vera [1983]. Panel B: Basins and structural domains in northern Africa and the central Mediterranean basin.

The general timescale used is the Cenozoic timescale by Berggren *et al.* [1995]. The paleo-position of the different crustal blocks/tectonic plates and their motion through time are based on the database of the Ocean Drilling Stratigraphic Network (ODSN), which can be found at http://www.odsn.de/odsn. The rotation poles on which these reconstructions are based are included in the ODSN website. Although one might argue about the correct position in this reconstruction of the many minor crustal blocks that have constituted the Mediterranean region through time, a reference database on the block movements has been chosen to enable understanding of direct implications of future improvements and new insights on the tectonic evolution of the area.

In this chapter results of the compilation are presented for the Paleogene, the time interval during which the Pyrenean collision was the major plate tectonic event in Iberia. In 7 compilation maps (for 65, 54, 42, 36, 30, 27 and 24 Ma), the major trends and significant changes with respect to previous time slices will be discussed in detail. The many references on which maps are based allow for more detailed regional information.

Names used in the text for regions, faults, basin and other features are depicted in a few regional maps: Figure 4.1 for the Iberian Peninsula, Figure 4.2 for southern France, Figure 4.3 for northern Africa, the western Mediterranean and the Betic/Alboran region. Figure 4.4 serves as legend for all the plates of the reconstruction.

65 Ma, KT-boundary, Figure 4.5

General

Along northern Iberia and southern France, inversion of former Mesozoic normal faults occurs and subduction or underthrusting is active. The Pyrenean subduction system was linked with the Alpine subduction system [Stampfli *et al.*, 2000]. Only the western part of the present Iberian Peninsula was emerged and had a planar and low relief, because the Mesozoic development of the western Iberian margin had not caused important rift shoulder uplift [Stapel, 1999]. Large parts of central Iberia were around sea level, so subtle eustatic sea level changes caused significant shifts in the position of the coastline.

Detail

Western margin

The Algarve is general emerging and a diapiritic phase is active [Terrinha *et al.*, 1990]. In the Lusitanian area, the northern continental platform is about 500-1000m deep from Paleocene-Lutetian, sea covered northern Lusitania [Azevêdo, 1991]. Further north, onshore Galicia, peneplane surfaces are developing [Pagés Valcarlos & Vidal Romaní, 1998].

Northern margin

General thermal subsidence affected the northern Iberian margin [Espina *et al.*, 1996b], while subduction of the newly formed oceanic crust of the Bay of Biscay was active since Campanian [Ziegler, 1988]. The margin had a N120 trending structure and major sedimentation flux [Boillot & Malod, 1988], with erosion along the present-day coast and S.Cantabria [Rincon *et al.*, 1983].

Central Iberia

Since the Mesozoic, marginal environments of the Iberian Basin extended into Central Iberia [De Bruijne, 2001], gentle sea-rise or fall inundated or emerged large areas. At the transition towards the Tertiary, the central part of the Iberian Chain emerges finally [Adrover *et al.*, 1983], shallow sea/lagoons and lowlands covered the region. The area was tectonically relatively quiet; a general (thermal?) subsidence was maintaining marine environments in parts of the region [Capote, 1983]. In the Almazan basin a stable shallow carbonate platform of L. Cret. Is topped by brackish to freshwater coastal swamp deposits of Danian age. A very similar situation was observed in the Sierra de la Demanda with the onset of non-marine Tertiary deposition [Bond, 1996]. Large parts of Central Spain (e.g. the Loranca Basin [Diaz Molina & Lopez Martinez, 1979]) were covered by lagoon-beach environments [ITGME, 1990b] or just above sea level as for example in the northern Central System resulting in erosion or non-deposition [ITGME, 1991d].

S. Pyrenees and Ebro

In the Ebro/Pyrenees area a large basin was present. The western basin was filled with marine (flysch) sediments, in the eastern part the 'Garumniense' (the continental equivalent) was deposited [Riba *et al.*, 1983].

N. Pyrenees and SW France

The southern border of Eurasia at this time, showed the first signs of the Pyrenean collision: the Aquitane basin slowly subsiding (onset of the flexural foreland basin) [Desegaulx & Brunet, 1990] while in the Provence compressive structures were general [Vially & Trémolières, 1996]. The same situation is valid with regard to the Alpine foreland due to the Briançonnais entering the subduction of the incipient Alps. The resistance to subduction of this continental block resulted in stress transmission into the Helvetic margin (northern part Valais Trough), where lithospheric deformation, uplift and erosion occurs [Ziegler *et al.*, 1998] and even more distal in the European foreland (North Sea/Channel) [Ziegler *et al.*, 1998]. The Briançonnais resisted underthrusting until M-L. Eocene [Stampfli *et al.*, 1998], on its northwestern part ongoing sedimentation has been documented for the Paleogene [Schmid *et al.*, 1996].

Catalan-Sardinian margin

Stacking of ophiolitic thrust nappes 'westward' onto the continental margin of Corsica and Sardinia [Carmignani *et al.*, 1995]. A subduction system dipping under the Iberian plate was activated along its eastern and southeastern margin during the latest Cretaceous and Paleocene [Ziegler *et al.*, 2001]

S. Iberia

The southeastern margin of Iberia was under compression [Vera, 2001], resulting in erosion/non-deposition in the southern Prebetics [Kenter *et al.*, 1990]. At the end of the Mesozoic, the Mulhacen complex (Internal Betics) reaches a peak in metamorphism (eclogite facies), increasing from east to west [Nieto Liñan, 1996].

N. Africa

Carbonate platforms in a deeper marine basin covered large parts of northern Algeria. The stress field was N-S compression, resulting in E-W folding, NE-SW trending sinistral and NW-SE trending dextral strike-slip faults [Aris *et al.*, 1998]. In the S.C. High Atlas more or less continuous subsidence and sedimentation occurred, but more to the west and south the first signs of uplift are documented [Görler *et al.*, 1988]. Between the Middle and High Atlas, the Moulouya region is being uplifted [Morel, 1989]. The Oran Meseta constitutes an uplifted platform between the Atlas Troughs from Triassic and Miocene [Giese & Jacobshagen, 1992] and in the Tellian-Riffian margin erosion of the present-day foreland of Morocco and Tunisia [Wildi, 1983].

54 Ma, L. Paleocene - E. Eocene (Ypresian), Figure 4.6

General

A change along the active plate boundary at chron 18 [Roest & Srivastava, 1991] is leading to large-scale deformation along the northern plate boundary. Clockwise rotation of Iberia continues in conjunction with Pyrenean-Cantabrian subduction, rapid convergence between Iberia/Africa with respect to Eurasia [De Jong, 1990] and the onset of sea-floor spreading between Greenland and Norway [Ziegler, 1988]. Collision occurs in all of the Pyrenean, Sardinia/Corsican, and Betic domains, and coincides with the full development of northwestward subduction of Tethyan oceanic crust under the Betic/Alboran and Corsica/Sardinia domains.

Detail

Western margin

In the northern Lusitanian margin a 500 –1000m deep marine basin persisted until Lutetian [Azevêdo, 1991]. Minor inversion off shore in the Lusitanian Basin can be correlated to onshore erosion of ~300m of U. Cretaceous sediments [Rasmussen *et al.*, 1998]. Tectonic activity is observed in the Algarve/Lusitanian margin (NS directed compressional strike-slip [Lepvrier & Mougenot, 1984]) and southern Portugal as well. In the latter where alluvial fans are interfingering with lacustrine carbonates, first Tertiary movement of the Plasencia fault is inferred: the southeastern block upthrown, the Sado Basin down [Pimentel & Brum da Silveira, 1991]. Emplacement and uplift of the Gorringe Bank can be related to the change in plate boundary setting [Le Gall *et al.*, 1997].

Northern margin

The Asturian-Cantabrian margin is the locus of active southward underthrusting of the Bay of Biscay margin under Iberia, creating a fold-and-thrust belt in the Danois Bank area. Deep-sea sediments and basement are deformed and uplifted to even above sea level [Boillot *et al.*, 1979]. Within the Cantabrian Range the first sedimentation occurs in the Oviedo Basin. Conglomerate is supplied from the north and onlapping to the south. The basin is endorheic at least for the Paleogene [Alonso *et al.*, 1996]. Along the present-day Cantabrian coast (Santander) Alveolina limestone (shallow marine) is being deposited [Riba *et al.*, 1983].

Central Iberia

During the Paleocene, the Hesperic Massif (as the basement in western and Central Iberia is called) in western Iberia is being uplifted due to ENE-WSW extension with perpendicular compression, as inferred from the Duero Basin [Santisteban *et al.*, 1996b]. Alluvial fans in the western part of the basin grade into littoral sedimentation in the eastern part [Corrochano & Armenteros, 1989], indicating that the eastern extreme of this basin was the coastline. The Sierras Obarense (S. Cantabrian) was an environment of marine-transitional-continental [Pol & Carbeillera, 1983], the Rioja area, a little more to the south, was at around sea level. Although the latter was still connected to the north with marine deposits, from the L Mesozoic onwards marine sedimentation never occurred! Paleocurrents of detritus in the Rioja area come from the south [Jurado & Riba, 1996], where gentle deformation of the Cameros Basin is testified by unconformities in a Paleocene regressive sequence in combination with open folding [Platt, 1990] and erosion of the Cameros massif [Muñoz Jiménez & Casas Sainz, 1997]. Increasing deformation of the Cameros Basin/de La Demanda can be observed towards the Almazan Basin (southwest of the Cameros) as well: 4 episodes of fluviatile sedimentation, coming from the NE (Massif De la Demanda) are evolving to more proximal [Pol & Carbeillera, 1983]. In the north side of the Almazan Basin, fluviatile sedimentation occurs in the synclinal Arganza Basin during the Ypresian [Floquet *et al.*,]. The rest of the Almazan was still under non-deposition or erosion [Bond, 1996].

The paleogeography of the Central System was completely different from its present-day configuration. Two separated basins existed: one in the Zamora/Salamanca/Avila/Toledo/NW Madrid and another E/SE of Madrid. The SCS did not exist as such, but another minor relief existed in between both basins [Portero Garcia & Olivé, 1983]. The continental basin in the western sector was wide and slowly subsiding distal flood plain with (continental) salt lakes [Portero Garcia & Olivé, 1983] in an open landscape without major relief nearby [Martín Serrano *et al.*, 1996]. Alluvial systems put sediments into this basin from several sides. In the Penaranda-Alba area proximal alluvial systems have a southern provenance [Pol & Carbeillera, 1983], coming from the Guadarrama region where erosion or small scale alluvial fans, extending over the present-day mountains is documented for this period [ITGME, 1991d]. Related to the uplift of the Hesperic Massif, conglomerate & sandstone of an alluvial fan system entered the western side of the basin (Zamora/Salamanca) from the northwest, and in the southwestern region (Ciudad Rodrigo Basin) erosion or non-deposition occurred [Jiménez Fuentes, 1983]. The second basin in the Madrid area, in the present-day eastern Madrid and Loranca Basin extending east/southeast ward, was the locus of marine and coastal sediments during the Late Mesozoic -Early Tertiary. This implies that the eastern areas have experienced more pronounced uplift during the Alpine orogeny [Santisteban *et al.*, 1996b]. Towards the southeast (Loranca Basin), unstable basins of transitional and/or continental deposition [de Torres Perezhidalgo *et al.*, 1983] with coastal environments in the Loranca basin [Muñoz Martín, 1993], connection to shallow marine is not well known.

In the Iberian Chain, detritic sedimentation occurs, but is not widespread [Adrover *et al.*, 1983]. This can be related to active transpressional deformation [Ziegler, 1988]. The Cabriel basin infill changes from detritic to marls and gypsum [Adrover *et al.*, 1983].

SE Iberia

Generally, the southern an eastern margin of Iberia was emerged or became emergent during the L. Paleocene. Erosion is documented in the southwestern Valencia Trough with an estimated total of 5km erosion from 60-40Ma [Fernàndez *et al.*, 1995], for the External Prebetics (in relation to deposition of red conglomerate) [Fontboté & Vera, 1983] and the Internal Prebetics, which was emerged during all of the Paleogene [Fontboté & Vera, 1983]. New topography was being formed to the south of the margin, related to extension? [Kenter *et al.*, 1990]. The coastline was located near the Subbetics testified by a (few dispersed outcrops!) pink micritic limestone, called ‘Capas Rojas’ [Fontboté & Vera, 1983]. Towards the Eocene, the coastline [HNPC, 1992] shifted northward [Fontboté & Vera, 1983] leading to deposition of shallow marine limestone and sandstone in the Internal Prebetics, further south deepening to a ~1000m deep basin [De Ruig, 1991]. Water depth increased to 1000-2000m in the Sierra Alamedilla [Lu *et al.*, 1998].

Betic realm

In the Internal Betics, the Malaguide was being emplaced upon the Alpujarride leading to uplift and crustal thickening [Balanyá *et al.*, 1997], related to the Balearic orogeny. To arrive at this important crustal thickening (up to 35 km [De Jong, 1990]) in the upper plate of the Corsican/Sardinian subduction system, a southeast ward shift of the subduction front might have occurred. Erosion of the Malaguide is evident from calcareous turbidites that enter the Predorsalian/Mauritanicas, which contain Malaguide fragments [Durand Delga & Olivier, 1988] such as limestone with Microcodium [Fontboté & Vera, 1983]. The Malaguide (including the Tellian-Rif arc) remains being eroded or an area of non-deposition until transgressive E. Eocene biogenetic sandy limestone is being deposited [Fontboté & Vera, 1983]; [De Jong, 1990], which can be correlated to the start of an extensional phase [Balanyá *et al.*, 1997]. SE of the eastern Malaguide a carbonate continent with low relief was present. On this continent red continental deposits fill in erosional troughs [Martín Martín *et al.*, 1998]. A distensive period is inferred from PTt data for the Mulhacen complex [Nieto Liñan, 1996] and the Ghomaride, which is a distensive margin until the L. Eocene [Maate, 1996].

S.Pyrenean and Ebro

In the Pyrenean region, southward submarine emplacement of the Upper Thrust Sheets occurs [Muñoz *et al.*, 1983] under high rates of south-directed shortening and widespread marine foreland deposition [Vergés *et al.*, 1995]. In the S. Pyrenees/ Cantabrian Cordillera contraction deformation is active [Muñoz Jiménez & Casas Sainz, 1997], due to rapid convergence [De Jong, 1990]. In the western Basque area a deep marine Eocene siliciclastic flysch trough develops [Riba *et al.*, 1983] in a WNW-ESE trending direction extending as far east as the Pamplona fault. This may be the eastward progression of downwarping of the N. Iberian Margin [Ratt, 1988] or the complex zone where the flip in vergence occurs (northward 'subduction' of Iberia in Pyrenees versus southward in Cantabrian). The Boixol and upper Pedraforca thrusts are fossilized prior to 55Ma, but the lower Pedraforca thrust starts its motion [Verges & Burbank, 1996] and in front of these thrust sheets a foreland basin starts to develop, most of which is presently overthrusted by the Pyrenees. Sedimentation patterns help to reconstruct the development of the foreland basin. In the S. Axial Zone of the Pyrenees Paleocene platform limestone are topped with Ypresian turbidites [Teixell, 1996] and in the present Eastern Ebro [Villena *et al.*, 1996]. In the Eastern Pyrenees/Sierras Marginales L. Paleocene Garumniense (continental) deposition is followed by an E. Eocene transgression depositing Alveolinas limestone [Muñoz *et al.*, 1983]. Both areas indicate progressive deepening or southward migration of the foreland basin, which can be related to rapid rising of the Pyrenees in the E. Eocene [Ziegler, 1988]. During the middle Lutetian the eastern foreland shifted from restricted marine sedimentation in the Ripoll basin to a broader and shallower basin [Verges & Burbank, 1996]. The present-day central Ebro basin (area of Zaragoza/Lerida) forms the bulge related to this foreland basin: Paleocene or Eocene sediments are hardly being deposited here [Riba *et al.*, 1983]. Platform carbonates in the southern Ebro basin indicate that this area was the southern margin of the foreland basin. Finally, the southwestern border of the proto-Ebro basin (between the CCR and the Baleares) is active as a left lateral fault system [Ramos-Guerrero *et al.*, 1989], leading to a sudden input of conglomerates into the southeastern Ebro basin [Verges & Burbank, 1996] and to collision between the Alboran/Betic and Corsica/Sardinia.

N. Pyrenees and SW France

The northern Pyrenean foreland is still relatively quiet with deposition of shelf platform carbonates [Vergés *et al.*, 1995], although subsidence in the Aquitanian Basin, after slow subsidence during the Paleocene, is now increasing (water depth 0-500m) with a major phase in the eastern basin [Desegaulx & Brunet, 1990]. 'Pyrenean' N020° compression that might be related to this increased subsidence is inferred from variations in sediment thickness in syndepositional folding [Rocher *et al.*, 2000]. Just north of the developing Pyrenees, a syntectonic breccia is being deposited in front of the Montpellier thrust [Sérrane *et al.*, 1995], related to a maximum of compressional deformation in the Gulf of Lions [Vially & Trémolières, 1996]. More external, the Marseille Basin/ Provence is an area of erosion or no deposition, sediments are absent until ~late Lutetian [Stampfli *et al.*, 1998], for the Limagne/ Massif Central/Bresse emergence is inferred from erosion of the Mesozoic cover [Bois, 1993]. Massif Central: Alkaline volcanism [Bois, 1993]. In front of the Alps, the Briançonnais has subducted entirely and Valais oceanic crust enters the subduction zone around E-M. Eocene [Schmid *et al.*, 1996]. In the present-day western Alps area, pre-Eocene west verging folding and thrusting occurs south of Pelvoux under NE directed horizontal extension and strike-slip against the edge of the Valais Zone [Coward & Dietrich, 1989]. The present-day Bresse Rift is uplifted and eroded in relation to this early Alpine compression [Bois, 1993], that is due to the incorporation of the Briançonnais in the accretionary wedge of the Alps [Ziegler *et al.*, 1998]. In the Alpine foreland, flysch deposition starts [Coward & Dietrich, 1989].

Catalan-Sardinian margin

Just as the southwestern border of the Ebro Basin, the CCR area is under NW-SE compression, but the Iberian Chain still has not been formed [Guimerà, 1984]. Deposition of Garumniense and conglomerates (red beds with high carbonate content, paleosols) point at erosion [Capdevilla *et al.*, 1996]. The entire eastern margin is uplifted and eroded, [Fernàndez *et al.*, 1995] inferred up to 5km of erosion during the Paleocene to M. Eocene in the southwestern Valencia Trough. Paleocene uplift and gentle deformation of Mallorca (no sedimentation) [Ramos-Guerrero *et al.*, 1989], the Baleares in general (erosion [Fontboté *et al.*, 1983]) or the Valencia Trough region is inferred from the fact that Eocene sub-crop map shows different ages [Ramos-Guerrero *et al.*, 1989].

The inferred start of the northwestward subduction of Tethys ocean floor under 'Iberia' (~55Ma [Zeck, 1996]) is contemporaneous with collision of the Calabrian/Corsica-Sardinia blocks [Ziegler, 1988], leading to stacking of ophiolitic nappes onto the continental margin. Ypresian sediments seal this tectonic activity [Carmignani *et al.*, 1995]. Most probably related to this, southern Sardinia was emerged while the rest was in shallow marine (~200m?) environments, forming the southern extreme of the southern Pyrenean foreland basin [Carmignani *et al.*, 1989].

Alpine Corsica: conglomerate deposition [Egal, 1992]. The south-dipping Alpine subduction system is linked with the north to northwest dipping Balearic-Corsican subduction zone via a left-lateral transform fault separating the Adriatic block from the western ensemble of blocks.

N. Africa:

The northern African Margin is in a relatively quiet setting; limited dextral motion of Iberia is documented by Brede *et al.* [1992]. An epicontinental sea covers large parts of the margin, in the Tellian part reef-limestone is being deposited [Aris *et al.*, 1998] and in its Moroccan part shallow water carbonate deposition occurs [Uchupi, 1988]. The hinterland of this epicontinental sea, in the present-day foreland of Morocco and Tunisia, such as the Oran Meseta and a paleo relief north of M. Atlas [Herbig, 1988] and two separated blocks in N. Algeria [Wildi & Huggenberger, 1993], form emerged non-depositional or eroded areas. The development of this couple of epicontinental sea and emerging hinterland might be related to the onset of northwestward subduction of the oceanic crust of Africa under Iberia [Zeck, 1996]. More distal in the foreland minor activity of N040^{o} thrusts (Middle Atlas) and N070^{o} dextral strike-slip (High Atlas) is documented [Brede *et al.*, 1992]. In the M. Atlas, the southern block of the North Middle Atlas Fault (NMAF) is continuously upthrown from ~Cenomanian to at least the Upper Eocene. This is maintaining minor relief that is being eroded [Morel *et al.*, 1993], while on the northwestern, down thrown block siliciclastic sediments are being trapped [Herbig, 1988]. Even further south, in the central High Atlas, shallow marine to littoral/lagoon environments prevail on the southern block during the Paleocene-Eocene, indicating both low relief and low erosion rates on the bordering Saharan Platform [Görler *et al.*, 1988].

42 Ma, M. Eocene (L. Lutetian -E. Bartonian), Figure 4.7

General

The Azores-Gibraltar fracture zone becomes active around chron 18 (42Ma), but relative motion between Africa and Iberia is limited until 36Ma [Roest & Srivastava, 1991]. Until the amalgamation of Iberia to Eurasia along the Pyrenean suture, Iberia moves as an independent plate from 42-24Ma [Roest & Srivastava, 1991]. The Kings Trough – Pyrenees boundary is a compressional active plate boundary from 44Ma until 25Ma [Srivastava *et al.*, 1990]. In addition to the formation of the Pyrenees and its conjugate foreland basins, the Balearic orogen developed between Iberia and Adria/Magrheb [Butterlin *et al.*, 1986] along the SE margin of the Iberian plate.

Detail

Western margin

The northern part of the western margin is affected by activity along the Bay of Biscay-Pyrenean subduction. Both offshore and onshore, uplift is documented. In the Galicia Interior Basin an important, sometimes erosional unconformity between M. and L. Eocene sediments is related to uplift of the Galicia Bank (estimated to be of the order of several kms) [Murillas *et al.*, 1990]. Deformation decreases southward, moving away from the active boundary. Onshore, in NW Galicia, uplift is detected by the relative descent of the base levels of rivers [Pagés Valcarlos & Vidal Romaní, 1998].

Northern margin

Along the northern margin, oceanic crust of the Bay of Biscay is subducting southward under Iberia, creating an accretionary wedge. Subduction is continuing into the Early Miocene; the M. Eocene is the principal episode of this process [Murillas *et al.*, 1990]. This leads to NNW-SSE compression [Lepvrier & Martínez-García, 1990] in the Iberian northern margin, documented by a couple of NE verging folds and thrusts and inverse faults and thrusts verging SW developing in a shallow marine environment along the northern border of the Ebro Basin [Muñoz *et al.*, 1983]. The Amorican margin experienced minor compression in relation to the Eocene activation of the subduction [Ziegler *et al.*, 1995, 1998].

Central Iberia

The stresses related to the ongoing subduction are not restricted to the northern margin, but are transmitted to the Iberian mainland as well. In the Duero basin NNE-SSW compression with a large perpendicular extension component favored fault-related lowering towards the northeast [Santisteban *et al.*, 1996b]. Alluvial fans from nearly all edges enter the basin, the border facies pass into lacustrine environments in the central part of the basin [Jiménez Fuentes *et al.*, 1983]. Alluvial fans enter the northern Duero from the Cantabrian Range [Alonso *et al.*, 1996], in the northwestern Duero (Zamora) distal sediments of 2 alluvial systems with NW and SW provenance are deposited and in the western Duero (Salamanca) paleocurrents indicate southwestern and southern provenance [Jiménez Fuentes *et al.*, 1983]. In the southwestern Ciudad Rodrigo Basin conglomerates and alluvial sandstones [Jiménez Fuentes, 1983] are deposited in fans with paleocurrents towards E/NE [Santisteban *et al.*, 1996a]. Along the southern border non-deposition, alluvial facies (Penaranda-Alba) [Corrochano & Carballeira, 1983b] or a series of conglomerate with no relationship to the present-day Guadarrama [ITGME, 1991b] is observed. The high activity of the alluvial fans indicates uplift of many of the borders of the Duero Basin. Further foreland deformation related to the activity along the northern plate boundary of Iberia is found along the Sierra de La Demanda thrust front that becomes active, overthrusting the Rioja trough northwards. In the front of this thrusting, at the connection between the Duero and Ebro basins, the Montes de Oca develop as a 'high' [Pol & Carbeillera, 1983], separating the two basins. The Rioja area remains at around sea level and connected with marine deposits in north [Muñoz Jiménez & Casas Sainz, 1997], paleocurrents show a southern (Cameros) provenance of the clastic input [Jurado & Riba, 1996]. To the south of the active chain, in the Almazan Basin, the start of clastic sedimentation is the response to NW-SE compression and uplift of Iberian Range [Bond, 1996]. The amount of uplift is still limited as shown by the bordering eastern Duero Basin where lacustrine environments prevail and only limited clastic input from the Iberian Range and Sierra de La Demanda is observed [Pol & Carbeillera, 1983].

Both AFT data (1700 m uplift between 45-30Ma, cooling 2.5-5 degrees/Ma) [Sell *et al.*, 1995] and sedimentary/structural data (mass flow conglomerates deposited in N. Madrid Basin [De Bruijne *et al.*, 2001]) suggest general but limited uplift of at least the NE-SCS. In the Central System erosion or non-deposition supports this uplift. The trend of the SCS is not at all identical to the present-day geomorphology: alluvial fans and localized small basins in the Sierra de Guadarrama (Turegano/Segovia) cross over the present-day mountains [ITGME, 1991d]. The Loranca Basin shows a period of quietness just before activation SE part of the basin [Muñoz Martín, 1997] during which lagoon environments are deposited or non-sedimentation/erosion creates an unconformity between Lower Tertiary and Upper Eocene sediments [de Torres Perezhidalgo *et al.*, 1983].

S.Pyrenean and Ebro

A period of major convergence between Eurasia and Iberia, with rates up to 6mm/year [Vergés *et al.*, 1995]. As a consequence, the Pyrenean belt is thrusting southward onto the Iberian foreland, the first important relief of the Pyrenees is being formed [Teixell, 1996] as documented by AFT (start of exhumation at ~50Ma) [Fitzgerald *et al.*, 1999], an increasing thrust rate [Puigdefàbregas *et al.*, 1991] and breakback thrusting, deforming the Pedraforca thrust sheet [Verges & Burbank, 1996]. The stacking pile of the Sierras Obarense (S Cantabria) is about to break through water surface [Muñoz Jiménez & Casas Sainz, 1997]. Around the Pamplona the south dpping Bay of Biscay subduction system stops and is replaced by the north dipping Pyrenean subdcution system [Engeser and Schwentke, 1986]. The foreland basin is marine, turbeditic and widening under the advancing load, motion along the Vallfagona thrust was initiated at around 43.5Ma [Verges & Burbank, 1996]. Together with the migration of the thrust sheets and foreland basin axis in southern direction, the related bulge invokes retraction of carbonate platform along the distal southern margin that is located in the central Ebro basin as shown by absence of deposition in this region [Villena *et al.*, 1996]. In the western, marine Jaca basin only small alluvial fans enter from the south [Vincent & Elliott, 1996]. The sediments derived from the advancing imbricated western Pyrenean thrust belt [Millan Garrido, 1995] are shelf and slope marl and sandstone in the western External Sierras [Teixell, 1996] and major fans [Muñoz Jiménez & Casas Sainz, 1997] in front of the arising belt.

Catalan-Sardinian margin

NW-SE compression [Guimerà, 1984] in northeastern Iberia is thrusting the CCR over the southwestern border of the Ebro Basin, which evokes increased subsidence in the SE part of the basin [Vergés *et al.*, 1998]. Under the same compression, the CCR hinterland, the later Valencia Trough, experiences NW-verging thrusting that is creating local relief and deposition of red molasse type sediments [Martínez del Olmo, 1996]. Erosion of this uplifted hinterland creates huge conglomeratic wedges, e.g. the St Llorenç del Munt, active from ~50Ma and Montserrate, active from ~46Ma [HNPC, 1992]. The conglomerates enter a marine Pyrenean foreland basin and therefore are being deposited in a deltaic environment. Marine sandstone and conglomerate dominate the deltaic fronts but sometimes include fringing reefs and prodeltaic platforms [Capdevilla *et al.*, 1996]. The elevation of the catchment area for the St Llorenç del Munt is estimated at around 700-1250m, *Marzo, Oliana guide_7,* which means that the highest summits in the area might have been well over 1500m. Mallorca forms the southwestern most part of this elevated region; a transgression towards the northwest is leading to the first Tertiary sedimentation in large parts of the Balearic domain. While northern Mallorca is still under continental conditions indicated by some fluvial intercalations, its southern domain is dominated by near shore platform sediments containing Nummulites [Ramos-Guerrero *et al.*, 1989]. Sardinia is still connected to Iberia. The marine conglomerate of the southwestern Sardinian Cixerri formation contains clasts from Mesozoic levels that are restricted to the Provence and Iberia [Cherchi, 1979]. Moreover, northwestern Sardinia forms part of the southern (in this region continental) foreland of the Pyrenees [Stampfli *et al.*, 1998]. Southeast of Sardinia and Corsica the Penninic accretionary prism develops, due to the northwestward subduction [Stampfli *et al.*, 1998]. The thrust belt that develops in the Sardinia Channel area (until latest Oligocene) is just the prolongation of this system. Marine flysches are being shed from these accretionary zones towards the east/south (Alpine Corsica) [Egal, 1992].

N. Pyrenees and SW France

The Pyrenean collision has its effect on the northern (European) foreland as well, the increased loading leads to major subsidence and deposition in the Aquitanian Basin, especially in central and western parts, water depth remains at ~0-200m [Desegaulx & Brunet, 1990]. Eastward, the Pyrenean belt was continuing, which can be observed in the Gulf of Lions [Vially & Trémolières, 1996], attested by overthickened and elevated crust [Sérrane *et al.*, 1995] and in the Camarque Basin, where high topography (1000-1500m) [Sérrane *et al.*, 1995] is estimated based on erosion of the Mesozoic and Paleogene succession. In the Gulf of Lions rocks similar to the Pyrenean axial & northern crystalline zones thrust over Permian and Mesozoic [Bois, 1993] and [Mauffret *et al.*, 1995]. The northern foreland shows important compressional deformation as well, such as E-W trending folds/thrusts and broad synclines in the Camarque and Provence areas [Mauffret & Gorini, 1996], syntectonic continental sedimentation in the Alès Basin (in front of Montpellier thrust?) [Sérrane *et al.*, 1995] and the Northern Provencal Cover Block where limited detritic continental deposits are documented [Villeger & Andrieux, 1987]. The compression is further recognized in the Languedoc [Ziegler, 1988], Nimes Basin (NS), [Villeger & Andrieux, 1987] and Ardeche area (~NE-SW) [Bonijoly *et al.*, 1996]. Deposition of middle Eocene limestone in the Languedoc [Sérrane *et al.*, 1995] and Marseille basin area is considered to be part of the northern foreland sequence [Stampfli *et al.*, 1998]. In general, southeastern France and the complete southwestern European margin are under N-S compression [Roure & Coletta, 1996], although the first signs of rifting in southern France are observed [Sissingh, 2001].

S. Alps

Pre- L. Eocene deformation (folding and thrusting) is predating the intrusion of Adamello. Local submarine fans deposit from Maastrichtian until L. Eocene marls in the Lombardian foreland [Bernoulli *et al.*, 1989].

SE Iberia and Betic realm

The effects of the Pyrenean collision and possibly activity of the Betic-Balearic orgeny can be observed in the southern part of Iberia as well in the Valencia-Alicante region. A Late Paleogene tectonic phase that can be related to early Pyrenean collision is inferred from uplift and erosion of the region creating an unconformity between Maastrichtian and Eocene sediments [De Ruig, 1991b]. The southern Iberian margin is deepening southward and probably thrust loaded by Internal Betic units [Vera, 2001]. The External Prebetics are being eroded, the Internal Prebetics are shallowing [Kenter *et al.*, 1990] and the isolated outcrops in the Subbetics show marly facies with frequent turbidites, locally olistostromes [Fontboté & Vera, 1983]. Coastline SE Iberia: [HNPC, 1992]

The Alboran units do not seem to show any evidence of compressional deformation related to the Pyrenean collision, which shows that for this epoch, the Alboran was not attached to/had not yet collided with mainland Iberia. The end of the emplacement of the Malaguide over the Alpujarride is followed by subsidence under vertical shortening of the Malaguide/Alpujarride pile [Balanyá *et al.*, 1997]. This subsidence is documented by a general transgression depositing biogenetic sandy and marly limestone on the Malaguide [Fontboté & Vera, 1983]. The transgression cannot have affected all of domain, because the sediments contain clasts of the Mesozoic cover of the Malaguide, so parts must have been emerged and eroded [Fontboté & Vera, 1983]. Crustal thickening still occurs in the Maghrebide (Greater Kabylia) in relation to oblique collision [Saadallah & Caby, 1996]. In the Dorsalian and nearby domains tilting of blocks created considerable relief [Durand Delga & Olivier, 1988] and combined with the aforementioned transgression over the Internal Dorsalian and Ghomarides (Malaguide) lead to laterally strong varying depositional sequences. From shallow water limestone deposition [Durand Delga & Olivier, 1988] to polygene conglomerates in the Predorsalian [Fontboté & Vera, 1983]. In the External Dorsalian marl with pelagic foraminifers, turbiditic biocalcarenites and conglomerates of local or more internal origin were deposited [Durand Delga & Olivier, 1988] and in the Mauritanicas conglomerates, conglomeratic limestone and marls dominated [Fontboté & Vera, 1983]. To the south of these units, south of Lesser Kabylia, subduction-related LT-HP metamorphosis occurs [Fontboté & Vera, 1983].

N. Africa

The northern part of the margin shows evidence for limited compressional deformation: folding of the Algerian [Aris *et al.*, 1998] and Moroccan Prerif with related conglomerates deposited on the Riffian and

Tellian margin [Wildi & Huggenberger, 1993]. Moreover, in the external western Tell a compressive phase is documented for the L. Lutetian [Wildi & Huggenberger, 1993]. Further south, in the High Atlas a last subsidence stage [Görler *et al.*, 1988] occurs and the western and central parts are flooded by sea [Giese & Jacobshagen, 1992]. Northern Africa is being deformed and uplifted, this is the main period of inversion of the Atlas Trough related to the start of collision with the Alboran units [Ziegler, 1988], compression reached into the Shara platform: a main phase of inversion of the Saoura-Ougarta chain is observed [Ziegler, 1995], contemporaneous with the inversion of the Triassic-Jurassic Atlas troughs and emplacement of the internal Kabylian and ultra-Tellian nappes [Vially *et al.*, 1994].

36 Ma, L. Eocene (Priabonian), Figure 4.8

General

The Pyrenean collision reaches its peak and the Pyrenees are very rapidly uplifted. Deformation related to this collision starts to migrate westward and into the interior of Iberia and results in uplift of the northern border and isolation from the sea of the Ebro basin. Reactivation of basement faults occurs in the Iberian Range, Catalan Coastal Range and the Central System. The Rhine Graben starts to form and links up with the graben in SE France. A fully developed NNW-ward subduction system is active south of Corsica/Sardinia and the Betic/Alboran. Major inversion in the Atlas ranges and even the Saharan Platform and subsidence along the northern African margin are related to the onset of (or accelaration of) subduction along the Corsican/Sardinian/Balearic margin [Frizon de Lamotte *et al.*, 2000].

Detail

Western margin

In the Sado Basin area, NNW-SSE compression inverts NW-SE trending faults, creating 3 alimentation areas. Marine environments prevail, but the basin is close to sea level [Pimentel & Azevêdo, 1994]. Northern Lusitania (north of Nazare) undeepens the marine basin to very reduced depth (<200m) due to a regression [Azevêdo, 1991].

Northern margin

Active subduction in the Bay of Biscay is continuing, compressional deformation continues until L. Eocene-Oligocene [Grimaud *et al.*, 1982] or the Middle Miocene. The crust is significantly thickened under the NE part of the Cantabrian Range [Fernandez Viejo, 1997]. Deformation is accommodated by strike-slip reactivation of NW-SE trending Hercynian faults that bisect the accretionary prism and fault propagation folding [Espina *et al.*, 1996b]. The Oviedo Basin is now fully developed as the Llanera thrust (EW) is thrusting southward and erosional deposits are southward onlapping a karstified paleorelief [Teixell, 1996]. Along the Asturian Coast where NNW-SSE compression is likely [Lepvrier & Martínez-García, 1990], deep marine facies prevail during the Eocene-Oligocene [Muñoz *et al.*, 1983]. In the Basque Basin, open marine environment dominate in the external basins [Ratt, 1988], but closer to the Pyrenees (Navarese Basin) marine influences become less important. The Pamplona fault is active, throwing the eastern block down and the western relatively up [Ratt, 1988].

Central Iberia

To the north of Rioja, in the still marine Miranda-Urbasa syncline, synsedimentary deformation and the first erosion of the Sierras Obarenses occur [Muñoz Jiménez & Casas Sainz, 1997]. The uplift and active deformation in this linking zone between the Cantabrian and Pyrenean ranges separates the Rioja area and the southern Pyrenean foreland from marine environments. The N-S compression responsible for this deformation is transmitted towards the interior of Iberia, inverting the Cameros Basin [Bond, 1996], which is the onset of Paleogene sedimentation in the Castellian Branch of the Iberian Range [ITGME, 1991c]. The eastern part of the Duero (next to the inverted Cameros Basin) witnesses a tectonic phase by an Eocene-Oligocene karst surface covered by conglomerates [Corrochano & Armenteros, 1989]. The same stress field imposes ENE-WSW extension with NNE-SSW compression within the Duero Basin, leading to the generation of NE-SW trending horst and graben structures associated with major uplift of the basin borders and source areas [Santisteban *et al.*, 1996b]. This uplift leads to renewed input of erosional materials by alluvial fans. Along the northern edge of the basin these enter from the Cantabrian Range, in the northwestern (Zamora) from both the NW and SW, in the western Duero (Salamanca) paleocurrents show SW to S provenance [Jiménez Fuentes *et al.*, 1983] and finally, in the Ciudad Rodrigo Basin paleocurrents towards the NE/E [Santisteban *et al.*, 1996b].

Conglomerates in the SW Duero containing Mesozoic clasts, deposited in a fining up alluvial fan sequence grading into distal alluvial facies proceeding from S & SW [Corrochano & Carballeira, 1983b] seem not yet related to the present-day Guadarrama [ITGME, 1991c], but document uplift and erosion of the Mesozoic cover. It is clear that part of the SCS is being uplifted; several facts even document the onset of individualisation of the Central System. The start of the configuration of the SCS is dated at the Eocene-Oligocene boundary by [Portero Garcia & Olivé, 1983], a period for which AFT data show an important period of uplift [De Bruijne *et al.*, 2001]. Furthermore the start of deposition of lacustrine and sebkha deposits in the Madrid Basin is related to embryonic movements of the Guadarrama thrust [Ziegler, 1988]. Prograding alluvial fans with a SCS provenance in the northwestern Madrid Basin (Beleña/Pinilla) [Portero Garcia & Olivé, 1983], depositing thick piles of detritic sediments in the Loranca and Madrid Basin [IGME, 1975], evidence that at least the northeastern SCS is being constructed. Sedimentation in the Iberian Range is indicating uplift as well: change from marl towards more detritic [Adrover *et al.*, 1983].

S. Pyrenees and Ebro

Folding of the still marine Miranda-Urbasa syncline separates the Pyrenean southern foreland basin from marine waters. The Priabonian Cardona marine salt marks the end of a major cycle in the basin evolution [Puigdefàbregas *et al.*, 1991]. The Eastern Ebro Basin is closed from world oceans as well [Verges & Burbank, 1996], indicating that the eastern connection was closed already at an earlier stage. The southern foreland basin of the Pyrenees becomes lacustrine in its center with alluvial fans building in from the margins [Puigdefàbregas *et al.*, 1991]. Break-back thrusting and increased thrust rate [Puigdefàbregas *et al.*, 1991] cause rapid building of antiformal stack (within ~1Ma), increasing the relief significantly, which leads to coarse alluvium entering the foreland basin from the north. In the western Pyrenees the Axial Zone (Gavarnie Thrust nappe) is emplaced and important folding and thrusting of the Jaca Basin occurs [Teixell, 1996]. The frontal thrusts of the Sierra Guara are thrusting southward contemporaneously, making the Jaca basin a piggyback basin [Teixell Cacharo, 1992]. In this basin terrestrial sandstone and shale dominate the southern half of the basin [Teixell, 1996], being deposited by a sandy fluvial system draining W-WNW [Vincent & Elliott, 1996]. The present-day southern border of the Ebro Basin is emerged [Riba *et al.*, 1983], and the first northward overthrusting of the Iberian Range over the edge of the basin takes place under N010^{o} directed compression [Guimerà, 1984]. The same compression maintains an elevated region located to the SE of the Ebro Basin. Elevation estimated for the CCR is on average 500-1000m above the level of the basin, in the north slightly higher (> 700-1250m) than in the central (500-800m) and southern (500-700m) part [*Roca pers.*]. The area od the future Valencia Trough is folded and emerged as well, 200-500m elevated above the L.Eocene sea level, which equals around 350-650 above present-day sea level [Morgan & Fernandez, 1992]. Sediments are being transported north and northwestward from this elevated region.

SE Iberia and Betic realm

The southern margin of Iberia is stable and deepening southward. The External Prebetics are emerged and eroded just as the Internal Prebetics are largely emerged until the early Oligocene [Fontboté & Vera, 1983]. The coastline of SE Iberia [HNPC, 1992] is located over the south of the Internal Prebetics: in the Intermediate Units and the Subbetics marl and marly limestone dominate with turbiditic intervals in the oriental sector [Fontboté & Vera, 1983]. In the Gulf of Cadiz, the first compressional features related to African-Iberian collision develop [Maldonado *et al.*, 1999]. Alpujarride/Malaguide: emplaced as a complex over the Nevado-Filabride [Balanyá *et al.*, 1997] at least before 22Ma: undeformed granites of that age and Oligocene-Miocene sediments unconformable over the Malaguides and Alpujarrides contact [Martínez-Martínez & Azañón, 1997]. These sediments are being deposited in a transgressive basin in the Internal Betics, to its eastern part deepening until hemipelagic environments[Fontboté & Vera, 1983]. Stacking of the Internal Betics crustal segments to climax from L. Eocene-M. Oligocene [Monié *et al.*, 1994], an intermediate PT metamorphic peak is documented for the Mulhacen complex in the Eocene-Oligocene [Nieto Liñan, 1996]. This over-thickening resulted in extensional deformation of the Internal Betics [Durand Delga & Olivier, 1988]. Until the L. Eocene, the Ghomaride was a distensive, passive margin. From now until L. Burdigalian, compression dominates, culminating in the Aquitanian [Maate, 1996]. In the zones fringing the Alboran/Betics/Kabylia Block, sedimentation suggests activity of the block as well. In the Mauritanicas this consists of marine conglomerates, limestone and conglomeratic marls, in the western part of the Campo de Gibraltar fine detritic limestone, marls and limestone and in the Predorsalian red marl with sandstone beds are being deposited [Fontboté & Vera, 1983]. Block tilting in the Dorsalian considerable relief [Durand Delga & Olivier, 1988] and both the Internal Kabylian and the Ultra-Tellian nappes show deformation [Ziegler, 1988].

Catalan-Sardinian margin

The St Llorenç del Munt conglomerate fan is basinward passing into more fully marine environments [Capdevilla *et al.*, 1996]. On Mallorca smaller conglomeratic wedges prograde to the SE, showing tectonic activity (ext??). An ongoing transgression to the N-NW leads to passing of the conglomerate wedges into coastal clay-rich marls and sandy limestone [Fontboté *et al.*, 1983]. The clastic input is associated with renewed relief in N. Mallorca [Ramos-Guerrero *et al.*, 1989]. This northern highland was connected with the CCR and the European mainland, as derived from mammal occurrence [Pomar Goma, 1983]. On Corsica/Sardinia a possible start of transpressive tectonics that is active until ~ Aquitanian [Carmignani *et al.*, 1995], related to the start of continental collision associated with thrusting of part of Corsican continental margin with obducted oceanic crust onto Eocene/Oligocene syntectonic deposits [Carmignani *et al.*, 1995]. In Alpine Corsica active northwestward thrusting is documented with a northwards increasing displacement [Egal, 1992].

N. Pyrenees, SW France and N. Alps

Generally, the French Alpine foreland is under near N-S compression [Bergerat, 1987]. In the Aquitanian Basin a major phase of subsidence is observed especially in the central and western parts [Desegaulx & Brunet, 1990], while synsedimentary folding occurs in the southern basin [Rocher *et al.*, 2000]. This is related to the rapid build up of the Pyrenean chain [Vergés *et al.*, 1995]. In the western part of the basin open marine environments prevail, towards the east an evaporitic event occurs, which is related to the Cardona salt in the Ebro foreland [Puigdefàbregas & Souquet, 1986]. The development of the foreland basin in the east has reached its maximum width: in front of the Northern Pyrenean Front the foreland sequence stops onlapping the Montagne Noir [Roure & Coletta, 1996]. In the northern foreland near the Pyrenean collision zone compression still dominates, as for example (NW-SE directed) in the Ardeche area [Bonijoly *et al.*, 1996] and in the Languedoc [Ziegler, 1988]. Moving away from active belt the deformation style changes. The Durance and Cevennes faults were reactivated in a transpressional way during the Eocene [Roure & Coletta, 1996] but in the Les Matelles Basin left lateral pre-Oligocene strike-slip causes down throw of SE block which points at extensional strike-slip [Sérrane *et al.*, 1995]. The Bresse Rift starts opening (Bergerat90), in the Alès Basin extension starts (earlier than in the rest of Languedoc) [Sérrane *et al.*, 1995], as well as in both the Apt basin (105 directed) and the Manosque Basin (120 directed) [Hippolyte *et al.*, 1993]. These are the preludes of the extensional regime that is invading southern France from the Rhine graben area, where E-W extension is active [Bois, 1993]. In the present-day Jura, extensional faulting is inferred [Guellec *et al.*, 1990], no sediments are being deposited [Bois, 1993]. A continental marine incursion enters the Bresse Graben and even the Rhine Graben through a marine connection along the southern Jura [Sissingh, 2001]. In the western Alps, the foreland basin sedimentation shows a rapid deepening to flysches. The Helvetic margin is being incorporated in the deformation as shown by Helvetic erosion products in these flysches, dated at 37-34Ma [Stampfli *et al.*, 1998]. Now the Valais oceanic crust is subducting to the south, at 35-33Ma slab detachment of the Alpine root is inferred. Closure was most likely not complete, since flexure in the Provence is not important, not forming a large foreland basin [Stampfli *et al.*, 1998].

S. Alps - Adriatic domain

To the south of the Alps a second foreland basin, the Piemonte Basin witnesses its very start of continental sedimentation [Bersezio *et al.*, 1993]. Uplift of the south Alpine margin is inferred from the erosion of upper Eocene limestone [Bernoulli *et al.*, 1989].

N. Africa

The last marine sediments are being deposited in the Middle Atlas (shallow marine limestone) [Herbig, 1988]. In the Central Constantinois, NE Algeria folding of the Tellian domain occurs under E-W compression (090-120) [Aris *et al.*, 1998]. Uplift of Central Morocco is evidenced by the fact that the eustatic E. Oligocene transgression did not inundate the region [Brede *et al.*, 1992]. A suite of alkaline volcanism develops in the junction area of the Middle and High Atlas. Local conglomerate fans with High Atlas clasts [Brede *et al.*, 1992] demonstrate the first uplift related to overthrusting of the High Atlas over its southern margin.

Late Eocene compressional structures are unknown in the Tell-Rif external zones [Frizon de Lamotte *et al.*, 2000]

30 Ma, M. Oligocene (Rupelian - Chattian), Figure 4.9

General

Major E-W trending directed extension invades southeastern France. This process is breaking up the former easternmost Pyrenees and starts limited rotation of the Corsica/Sardinian block. Compressional deformation is still occurring in the Pyrenean range and its southern foreland basin towards the west. Crustal thickening and related metamorphism in the Betic/Alboran units reaches a maximum at around 25Ma. The inversion of the former Atlas Trough is culminating.

Detail

Western margin

The present-day physiography of Kings Trough is formed around 32 Ma from intraplate volcanism [Srivastava *et al.*, 1990]. N040^o directed strike-slip compression is documented [Lepvrier & Mougenot, 1984] in the Portuguese margin from the Lisboan area to the Algarve. Offshore S Portugal lows and highs start to develop and are interpreted as the first compressional features in the eastern Atlantic related to the African-Iberian collision [Torelli *et al.*, 1997]. Under the same stress field the Lower Tajus Basin develops [Ribeiro *et al.*, 1990] and is correlated to a post-Eocene but pre-Aquitanian unconformity on- and offshore [Lepvrier & Mougenot, 1984]. The large extensional component in the strike-slip zone offshore causes pure extension: the offshore part of central Portugal subsides by normal faulting in the outer shelf/slope area [Rasmussen *et al.*, 1998]. The Sado Basin is emerged during large parts of the Oligocene but close to sea level. An erosional surface develops on its Eocene marine deposits [Pimentel & Azevêdo, 1994]. In northwestern Portugal a similar Paleogene erosional surface is creating planar topography [Cabral, 1989]. Tectonic activity in Galicia is demonstrated by normal faulting until M. Chattian in the As Pontes Basin [Huerta *et al.*, 1996].

Northern margin

Along the Asturian coast N-S compression is documented. A reorientation of the stress field from NW in Eocene to NE in the Oligocene is inferred [Lepvrier & Martínez-García, 1990]. Under this compression, ENE-WSW trending Hercynian thrusts are being reactivated in and around the Oviedo Basin and thrusting southward [Teixell, 1996] and conglomerate is being deposited in the Oviedo Basin [IGME, 1973]. Tectonic activity along the southern border of the Cantabrian Cordillera –Ubierna Fold Belt- is inferred from growth synclines that developed before the late Oligocene [Espina *et al.*, 1996a]. Parts of the Cantabrian coast are under marine conditions [Lepvrier & Martínez-García, 1990].

Central Iberia

Compression related to the Pyrenean collision is transmitted into the central part of Iberia. In the Duero Basin NNE-SSW compression and perpendicular extension generates a NE-SW trending horst and graben structure with major uplift of the borders and source areas. Oligocene arkoses are restricted to the S and E of the line Zamora/Salamanca, which implies uplift of the hills on the northwestern side of this line [Santisteban *et al.*, 1996a]. Further evidence for reactivation of the surrounding relief from the western and southwestern Duero is the influx of coarser detritus (conglomerate and sandstone) and retraction of the basin edge. The source material of these deposits is Paleocene deposit of earlier stages [Jiménez Fuentes *et al.*, 1983]. In the SSW-Duero (Penaranda-Alba) no sediments are known of Oligocene age, which suggests reactivated borders as well. In the southern Duero Basin tectonic activity of the Guadarrama Mountains can be deduced from a fining up sequence of breccia and conglomerate that is deposited discordantly over Paleogene sediments [ITGME, 1991c]. At least partially, the SCS was formed progressively [Portero Garcia & Olivé, 1983], the northeastern SCS N-S strike-slip compression [De Bruijne *et al.*, 2001] is documented.

The former Cameros Basin is thrusting northwards onto the Rioja Basin and southwestwards over the Almazan [Platt, 1990]. For the latter, this rejuvenation of deformation of the northeastern basin margin is related to N-S to NNE-SSW compression. Uplift of the border causes an increase in erosion rates, which results in building out of alluvial fans and fluvial sedimentation with a northern provenance, parallel to the axis of the basin [Bond, 1996]. To the eastern side of the Madrid Basin inversion of the Iberian Basin starts [Muñoz Martín, 1997], although very limited and not leading to deposition. Main activity of the border of the Iberian Range and the Loranca Basin is dated as middle to lower Oligocene. This resulted in an important angular unconformity between uppermost Eocene-lower Oligocene and upper Oligocene sediments Oligocene [de Torres Perezhidalgo *et al.*, 1983]. The lower sequence of detritic sediments shows paleocurrents without any relation to the present-day bordering chains and remainders of Mesozoic relief influence sedimentation patterns [Diaz Molina & Lopez Martinez, 1979].

SE Iberia and Betic realm

Stresses related to the Pyrenean collision and Tethys subduction benath the Balearic system cause deformation as far in the foreland as the Valencia area, where folding and SW thrusting under ~050 compression is observed [De Ruig, 1991]. In the External Prebetics tilting and folding is inferred. No sedimentation occurs until the E. Burdigalian and these sediments onlap different Eocene units [Kenter *et al.*, 1990]. Erosion of the Prebetics is sourcing the marly and turbiditic facies of the Subbetics [Fontboté & Vera, 1983] that demonstrate a northeastern provenance [Geel, 2000]. The northern margin of the eastern Internal Betics starts to rotate while foreland basin sediments are being deposited [Allerton *et al.*, 1993]. In the eastern Malaguide an E. Oligocene transgression deposits conglomerate, during the M. Oligocene fan deltas develop. Both show proximity of Sardinia [Geel, 1996]. The transgression is related to the onset of SE-NW extensional deformation and heating [De Jong, 1990]. Crustal thinning starts to invade the region south of the Balearic domain, entering the eastern Internal Betics. To the west, compression prevails and nappes are being emplaced, especially in the Ghomaride/Malaguide domain. Erosion of this internal basement follows upon its gradual uplift [Durand Delga & Olivier, 1988]. The maximum stacking of the Malaguide/Alpujarride complex onto the Nevado-Filabride domain [Balanyá *et al.*, 1997] resulted in a peak in metamorphism [Monié *et al.*, 1994]. The emergent landmasses establish a land connection between Africa and Iberia [Pomar Goma, 1983]. Detritus from essentially the emergent Ghomaride/Malaguide zone is deposited in the Dorsalian and Predorsalian [Durand Delga & Olivier, 1988], as up to 1200m Oligocene flysches in the Mauritanicas [Fontboté & Vera, 1983] and as deep water Numidian clastics in narrow basins parallel to the incipient Rif and Kabylian fold belts [Ziegler, 1988]. To the west in Campo de Gibraltar red clay with some limestone levels is sedimented [Fontboté & Vera, 1983].

S. Pyrenees and Ebro

The linking zone between the Pyrenean and Cantabrian ranges is actively being deformed and uplifted. In the Basque Basin molasse sediments become more and more conglomeratic [Ratt, 1988]. The first occurrence of continental deposition in the synsedimentary folded Miranda- Urbasa syncline is related to major uplift of its southern border from 35-32 Ma [Muñoz Jiménez & Casas Sainz, 1997]. However, the southern Sierras de Obarense front has not yet reached its present-day position, terrigenous sediments in the Rioja still show provenance from the south (Cameros) only [Jurado & Riba, 1996]. The development of the External Sierras in the Western Pyrenees during the Rupelian, results in an influx of terrestrial sandstone and shale into the Ebro Basin [Teixell, 1996]. Rapid exhumation of the axial zone (Maladeta) occurs from 35-30Ma, as inferred by AFT-data [Fitzgerald *et al.*, 1999] and maximum denudation rates of 240mm/kyr occur during the same period [Morris *et al.*, 1998]. Under N-S compression, the Iberian Range is thrusting oblique onto the southern Ebro basin [Guimerà, 1984]. At the same time the CCR is thrusted onto the SE margin of the Ebro Basin.

Catalan-Sardinian margin

The Valencia Trough area forms a peneplane with very little relief at the time of an E.Miocene transgression [Martínez del Olmo, 1996], therefore erosion is likely during the Oligocene. Fracturing, subsidence and rift shoulder uplift indicate the very onset of rifting [Fontboté *et al.*, 1983], developing a NE-SW trough bounded to the NE by relieves at the present-day CCR, Garraf and Malgrat highs. This continental basin forms the northwestern margin of an Oligocene marine basin [Roca & Deselgaulx, 1992]. The basin forms within a still compressive setting for the CCR [Roca & Deselgaulx, 1992], which is obliquely thrusting onto Ebro foreland, under NS-compression [Guimerà, 1984]. This suggests a piggyback type of basin between Baleares and CCR, verging to NNW [Roca & Deselgaulx, 1992].

Southeastward prograding conglomeratic wedges are building out along the Mallorca margin forming a regressive cycle. In the southern part of Mallorca, marine conditions are prevailing [Ramos-Guerrero *et al.*, 1989], while in the north lacustrine sediments with important detritic influx are being deposited, which evidences unroofing at large scale of nearby elevated regions (active tectonics) [Fontboté *et al.*, 1983]. Rift shoulder uplift could be the process causing the Mid-late Oligocene erosive denudation of a massif situated NW of Mallorca [Pomar Goma, 1983] and renewing of relief in the NE [Ramos-Guerrero *et al.*, 1989]. The Baleares are in communication with Africa during mid.-upper Oligocene [Pomar Goma, 1983]. Ongoing active subduction of the African Plate under Sardinia is demonstrated by 29Ma calcalkaline volcanism in Sardinia [Hippolyte *et al.*, 1993].

Alpine Corsica is folded under continuing shortening, producing N-S and NE-SW trending folds [Egal, 1992].

Pyrenees and SW France

In the northern Pyrenean foreland, the last deformation features related to the Pyrenean compression are formed [Viallard, 1985]. In the Aquitanian Basin subsidence of the ~200m deep basin continues, but is of minor importance [Desegaulx & Brunet, 1990]. The loading of the European foreland by the Pyrenean belt has ended and a limited part of the foreland is incorporated in the deformation. In the Londres and Lodeve syntectonic basins, northward thrusting (E of Cevennes fault) or backthrusting of Montagne Noir (west of Cevennes fault) starts erosion of the flexural sequence in the northern Pyrenean foreland [Roure & Coletta, 1996]. South of the Marseille Basin basement erosion documents a paleohigh [Hippolyte *et al.*, 1993]. An estimate for the elevation of this paleohigh comes from the Provence, where the pre-rift compressional relief is estimated to have been of the order of 1000-2000m, at least south of the Camarque basin [Sérrane *et al.*, 1995]. The dominant deformation style in the region becomes extension, which is invading the area from the north, where in the Limagne and Bresse Rift shale with sand layers document active rifting [Bois, 1993]. With an overall extension direction in SE France of N110 [Roure & Coletta, 1996], the extension is migrating towards the Valencia Trough area. In the Gulf of Lions alluvial/lacustrine basins evolve under NW-SE extension [Bois, 1993], where the first marine sediments occur in the southeastern sector, pointing at a marine connection to Tethys [Roca *et al.*, 1999]. Rifting begins in Ligurian Provencal region [Hippolyte *et al.*, 1993] resulting in the first, very limited, formation of oceanic crust in the Provencal Basin [Lonergan & White, 1997], although full development will not start before the E. Miocene [Roca, 2001].

(CONTINUED ON PAGE 114)

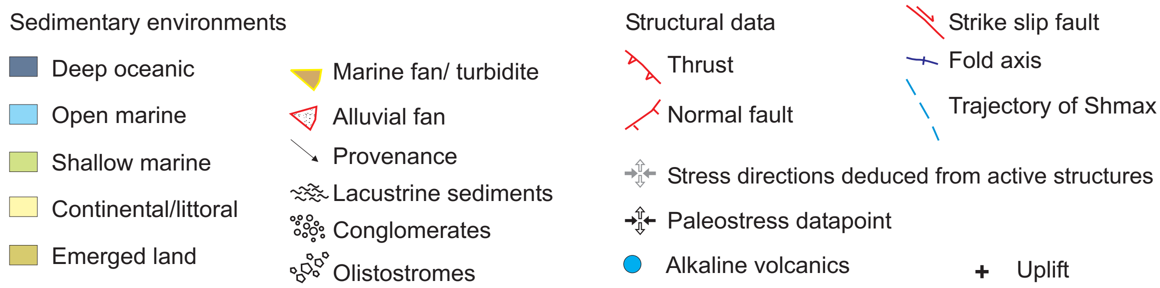


Figure 4.4
Explanation of colors, symbols and annotations used in Figures 4.5 – 5.8.

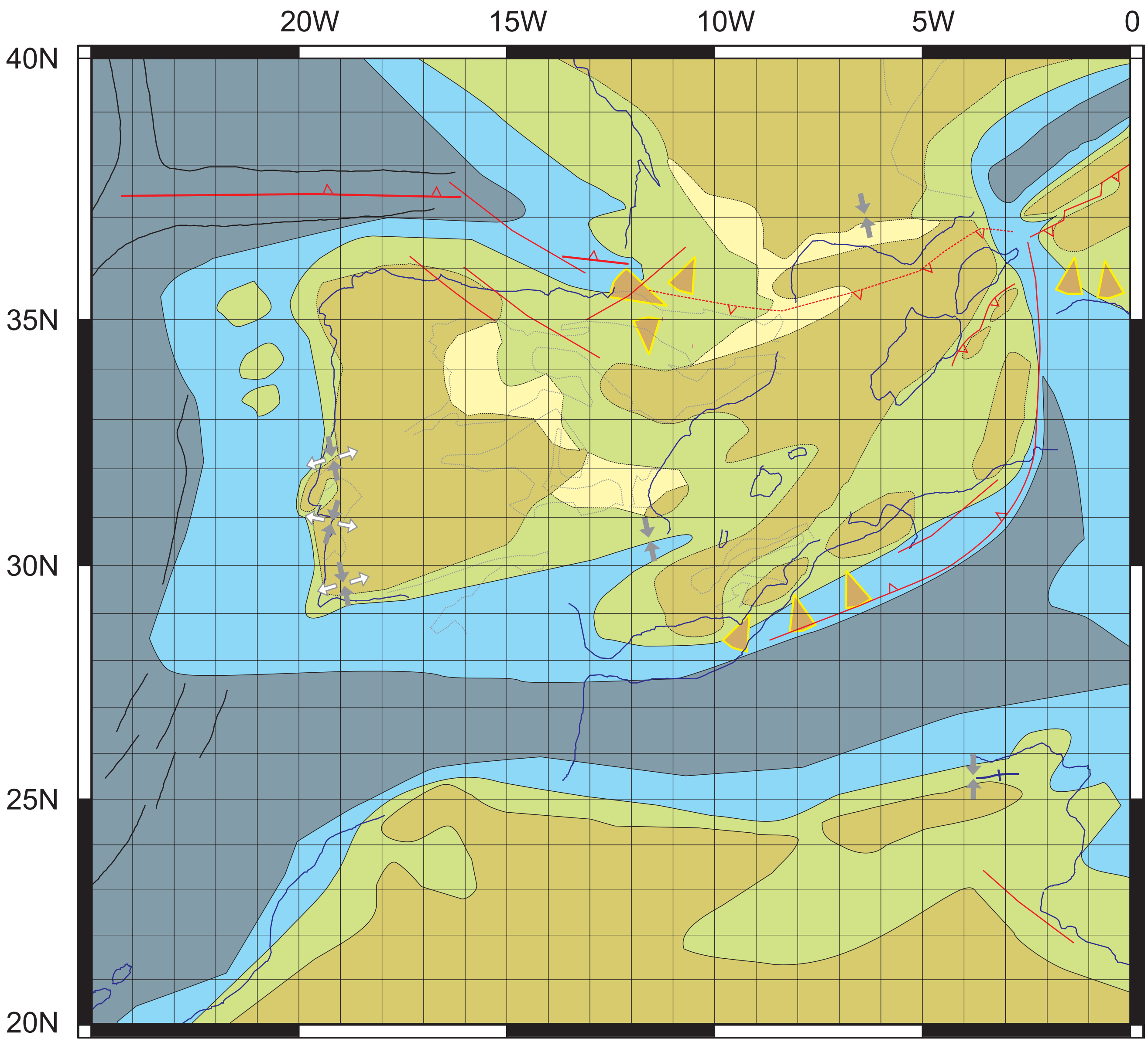


Figure 4.5 (65Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the Cretaceous-Tertiary boundary (65 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description. Based on reconstruction by the Ocean Stratigraphic Drilling Network (www.osdn.de).

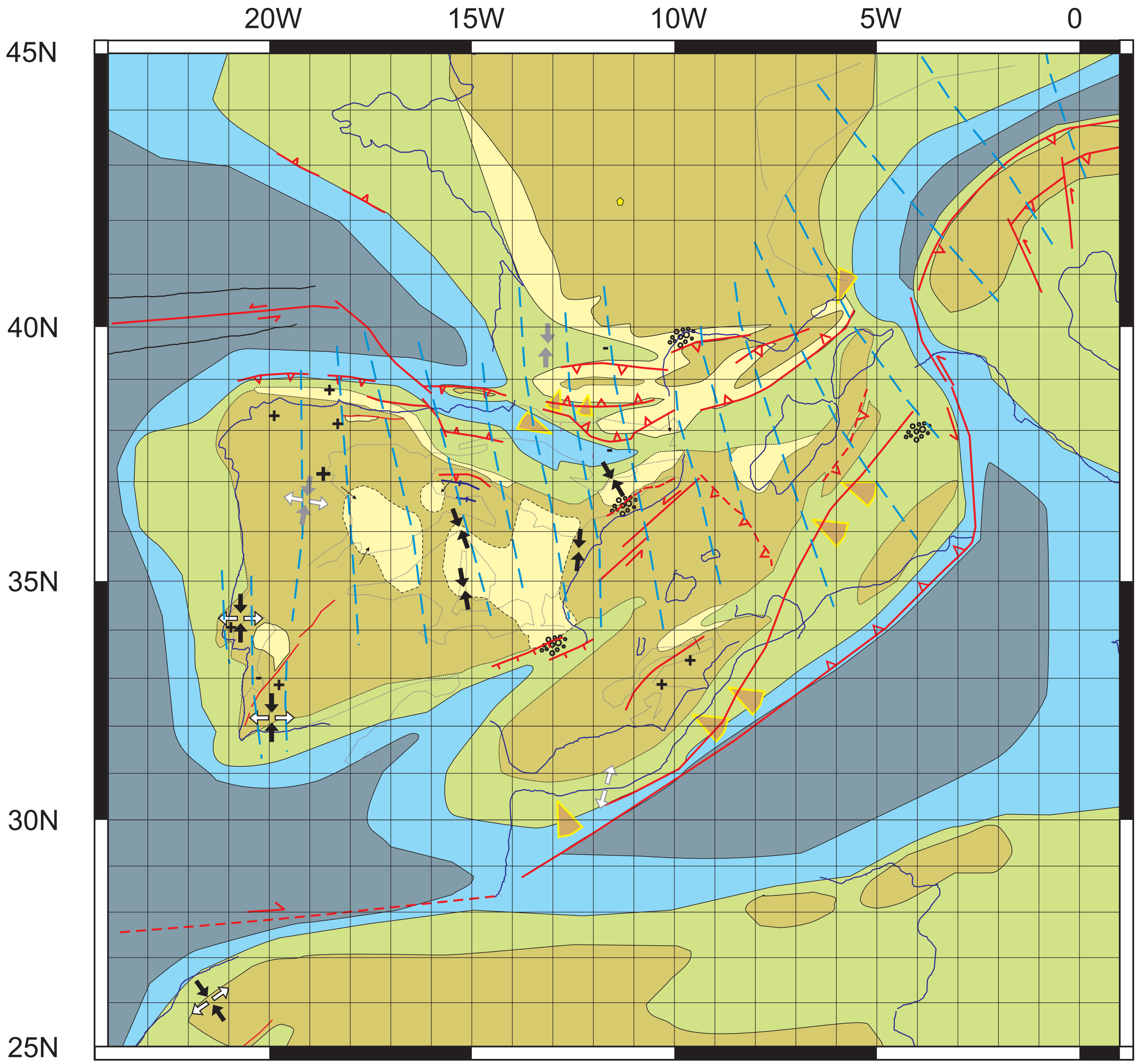


Figure 4.6 (54Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at L. Paleocene - E. Eocene (Ypresian, 54 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.

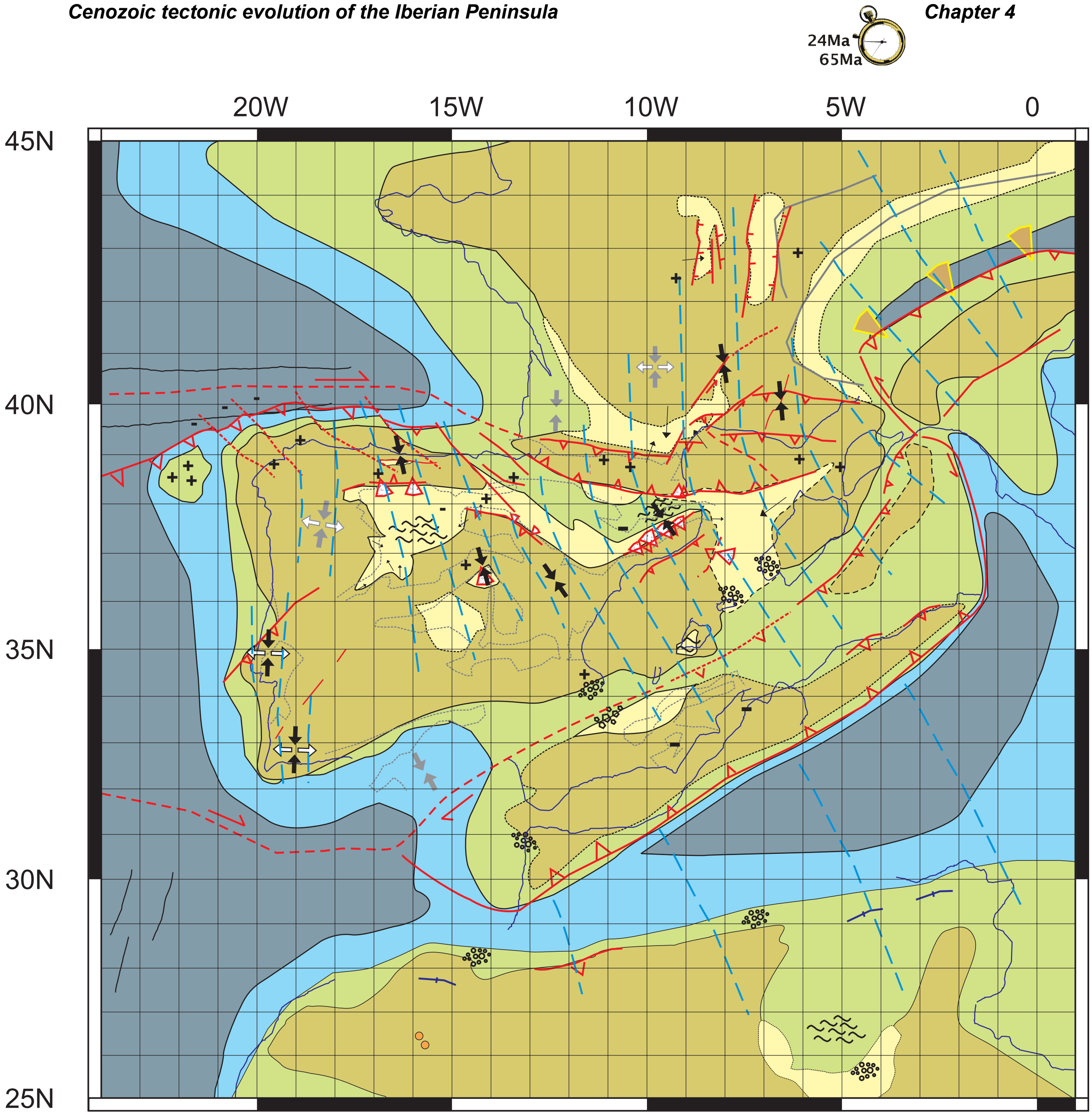


Figure 4.7 (42Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at M. Eocene (L. Lutetian -E. Bartonian, 42 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.

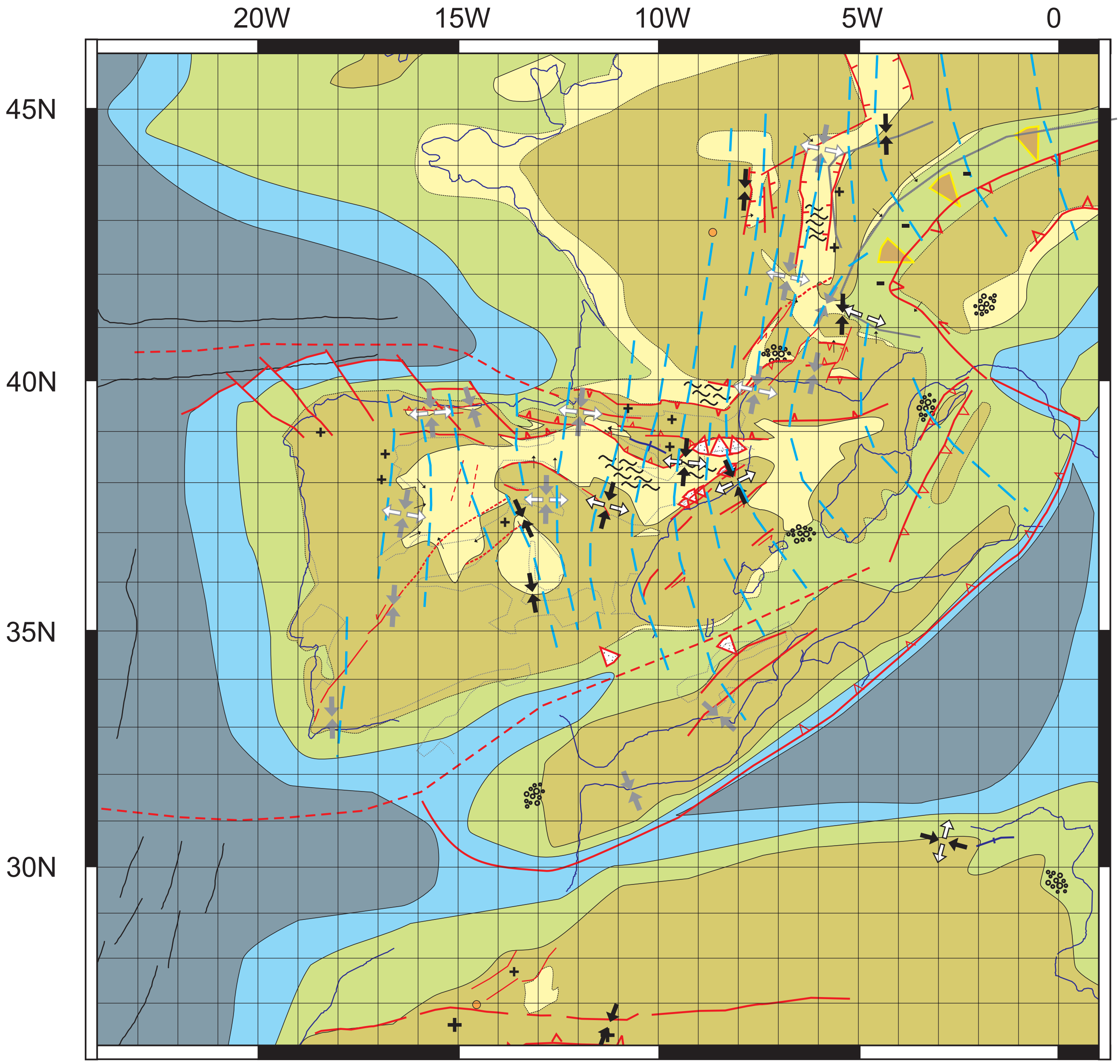


Figure 4.8 (36Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at L. Eocene (Priabonian, 36 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.

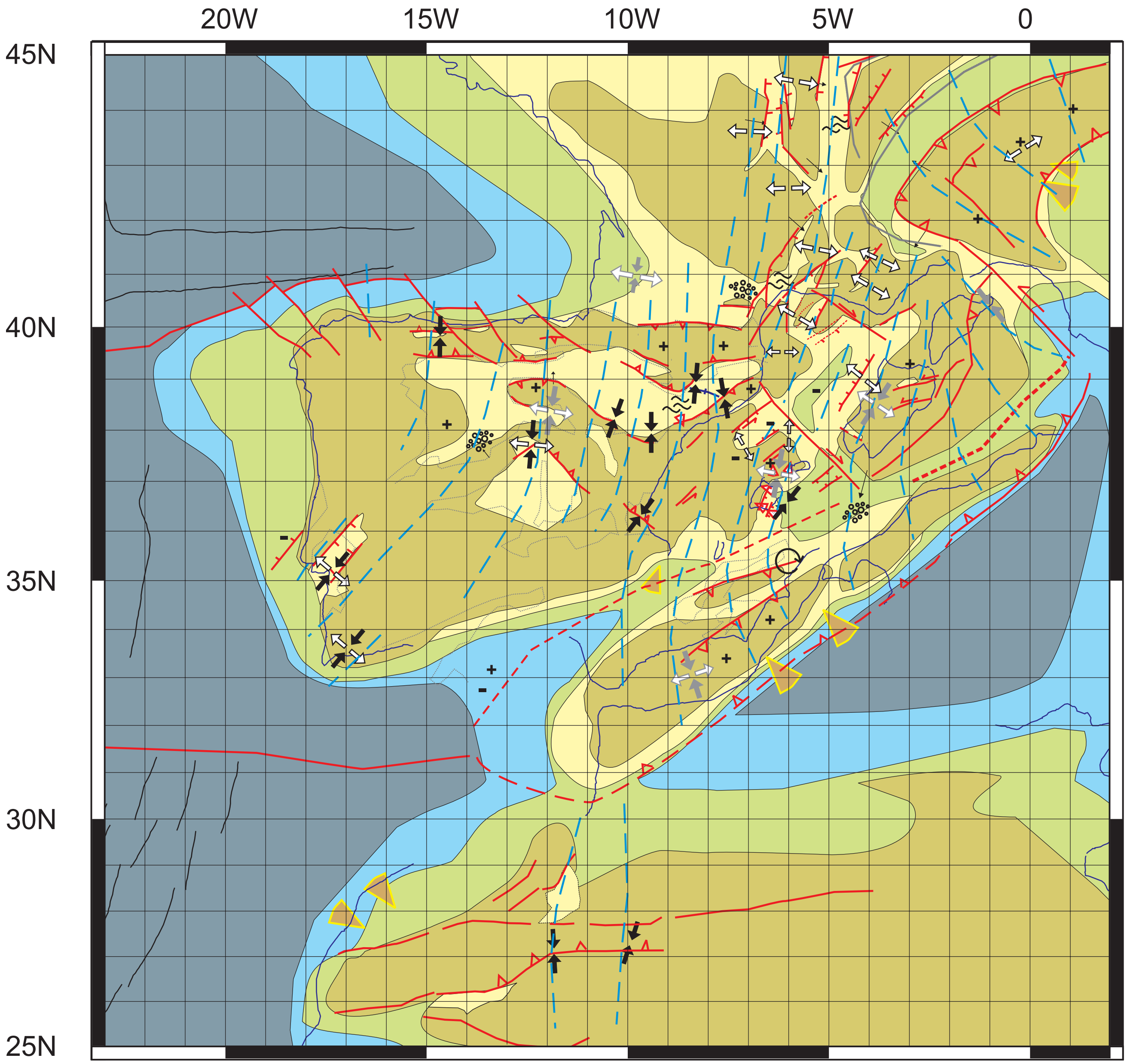
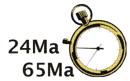


Figure 4.9 (30Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at M. Oligocene (Rupelian – Chattian, 30 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.



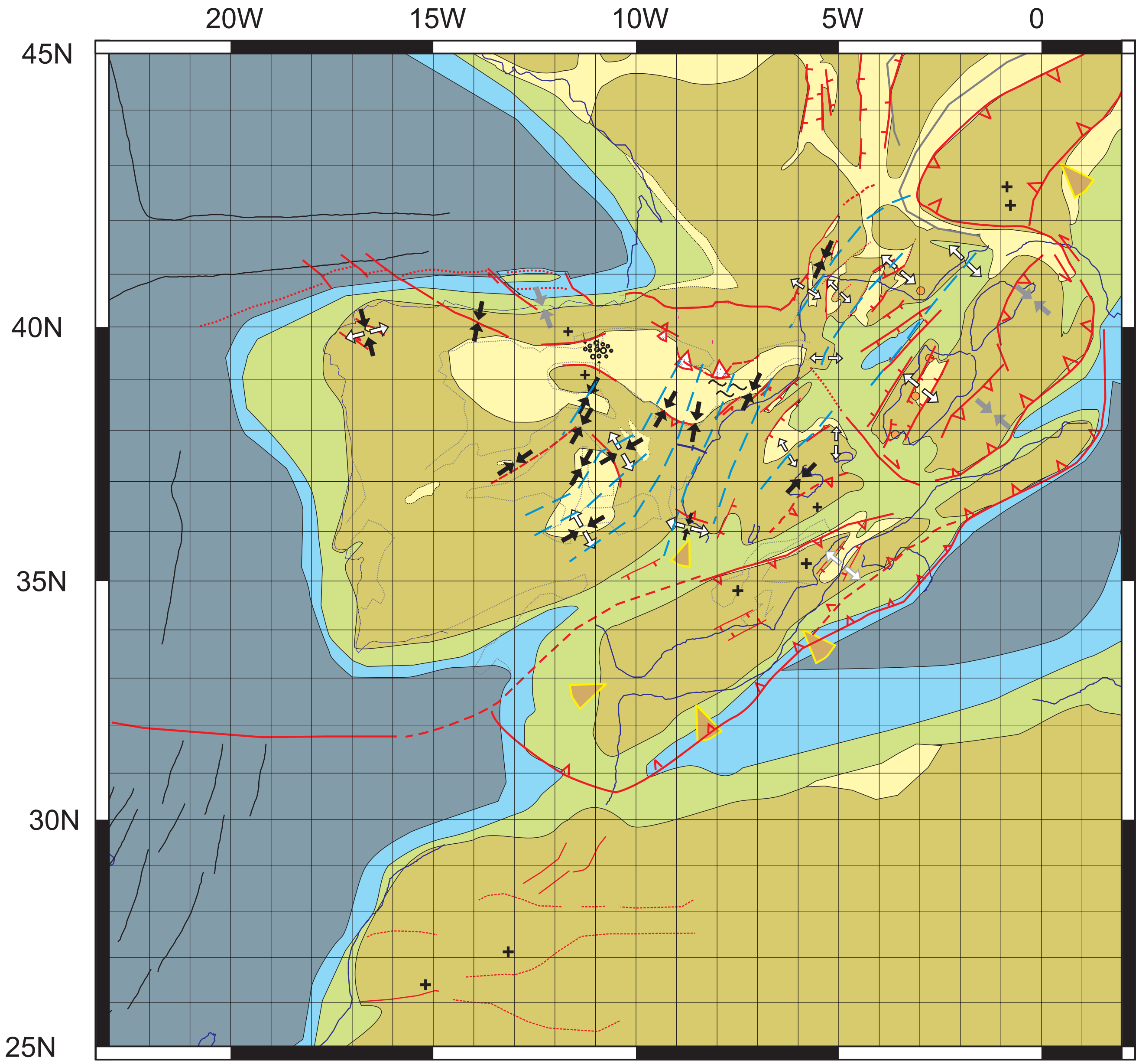
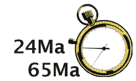


Figure 4.10 (27Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the L. Oligocene (Chattian, 27 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.



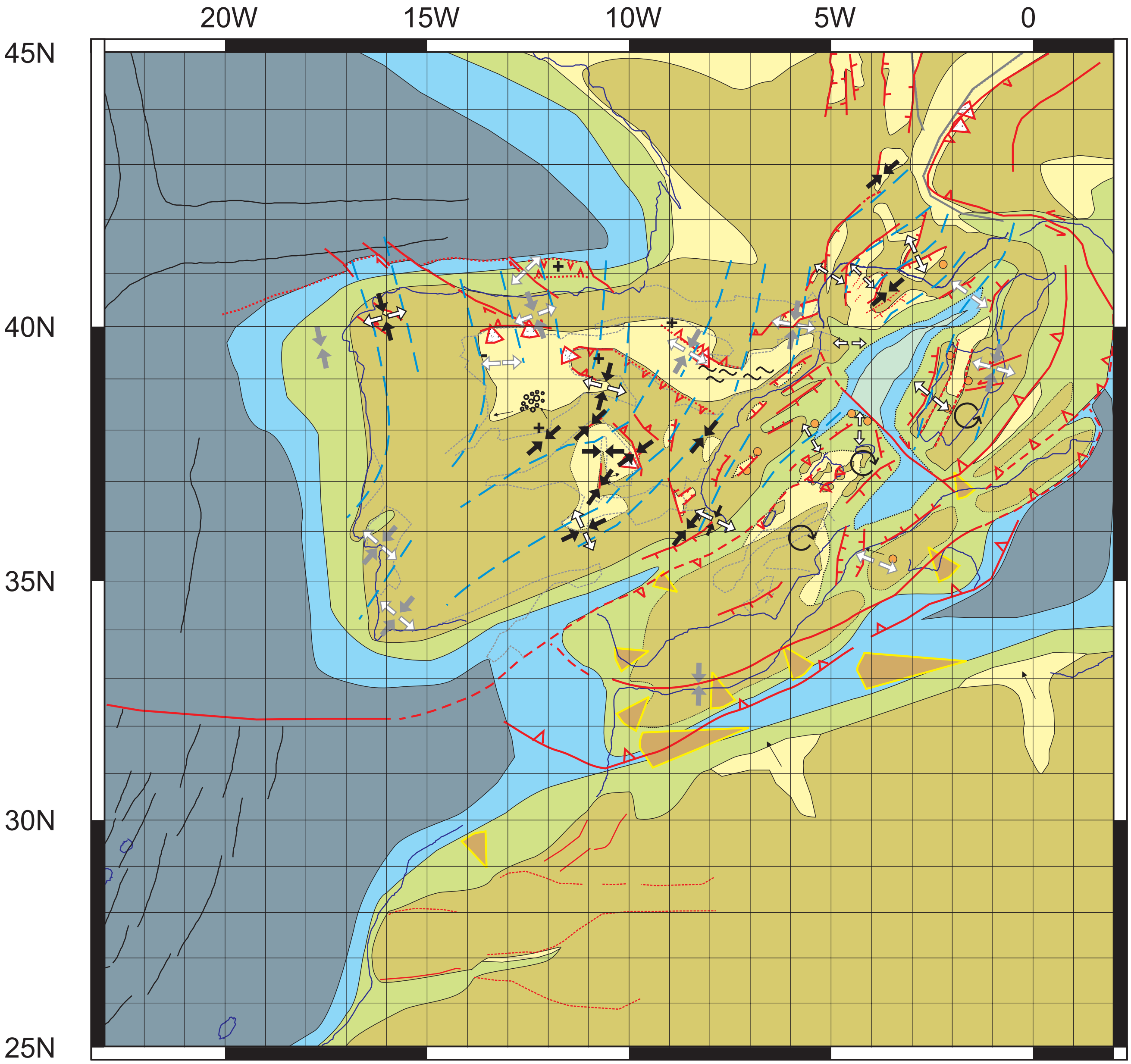


Figure 4.11 (24Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at L. Oligocene - E. Miocene (L. Chattian - E. Aquitanian, 24 Ma). See Figure 4.4 for explanation of symbols and colors and text for detailed description.

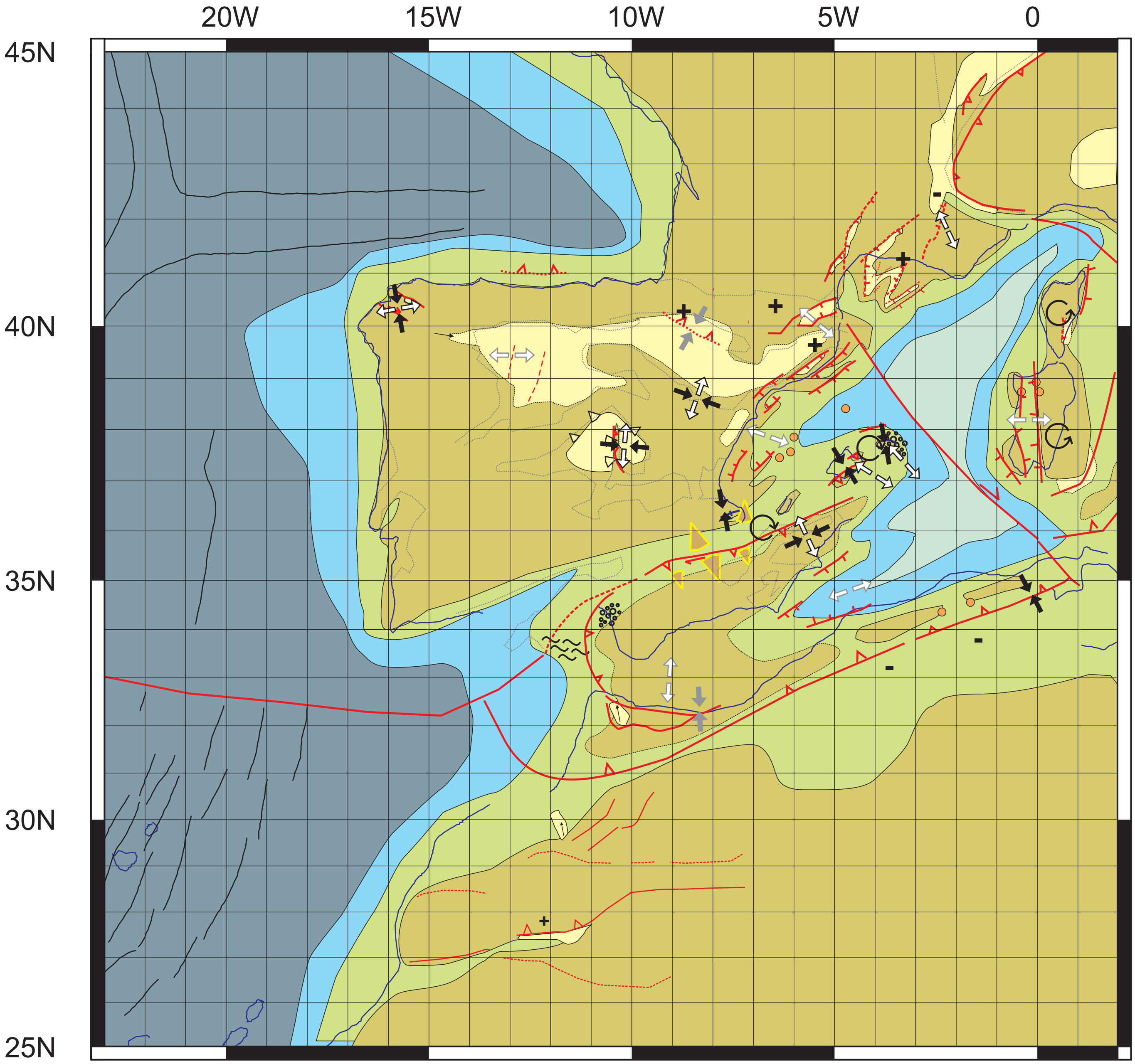
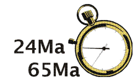


Figure 5.1 (21Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the E. Miocene (L. Aquitanian - E. Burdigalian, 21 Ma). See Figure 4.4 for explanation and text for detailed description.



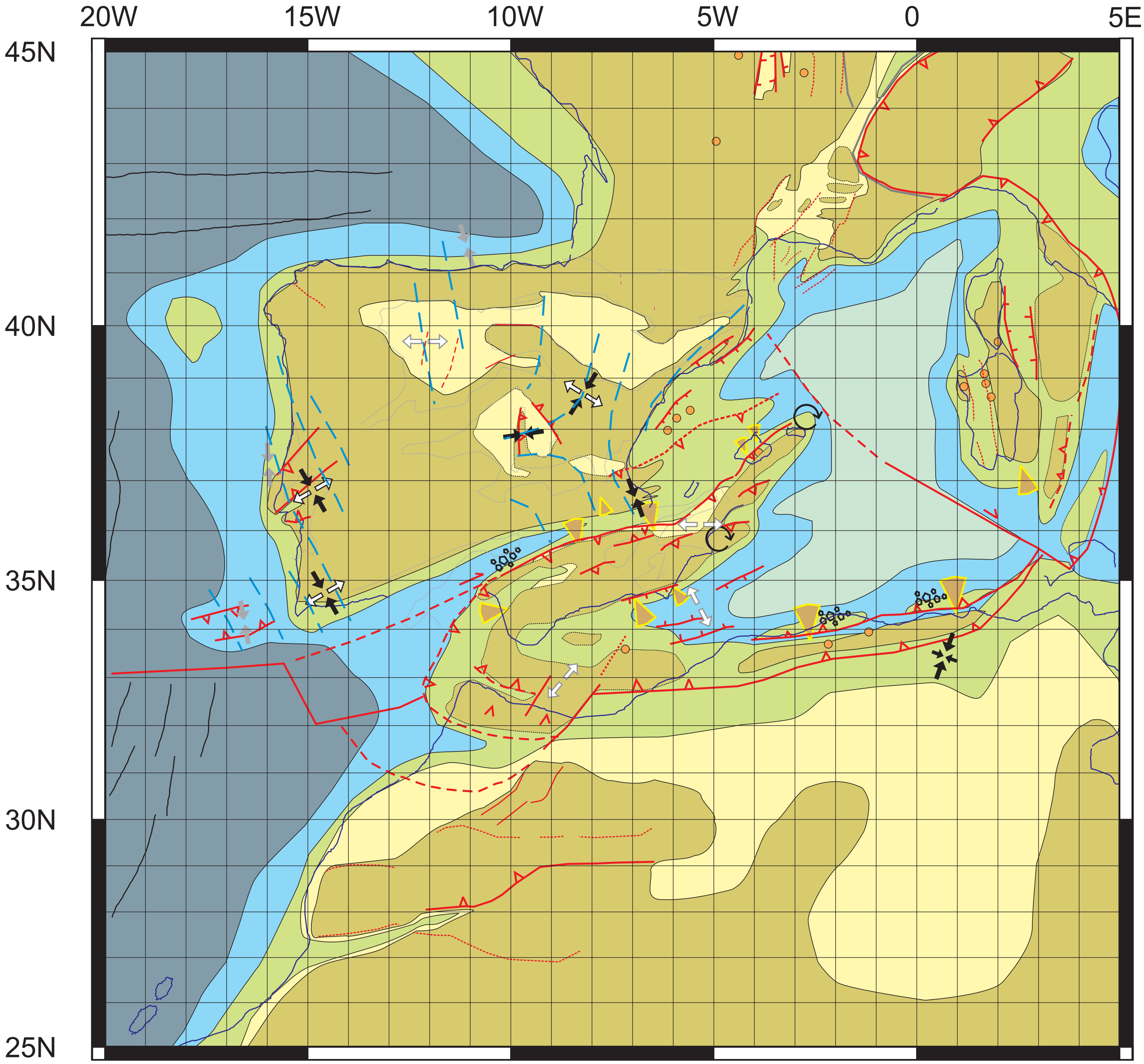
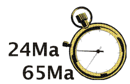


Figure 5.2 (18Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the E. Miocene (L. Burdigalian, 18 Ma). See Figure 4.4 for explanation and text for detailed description.



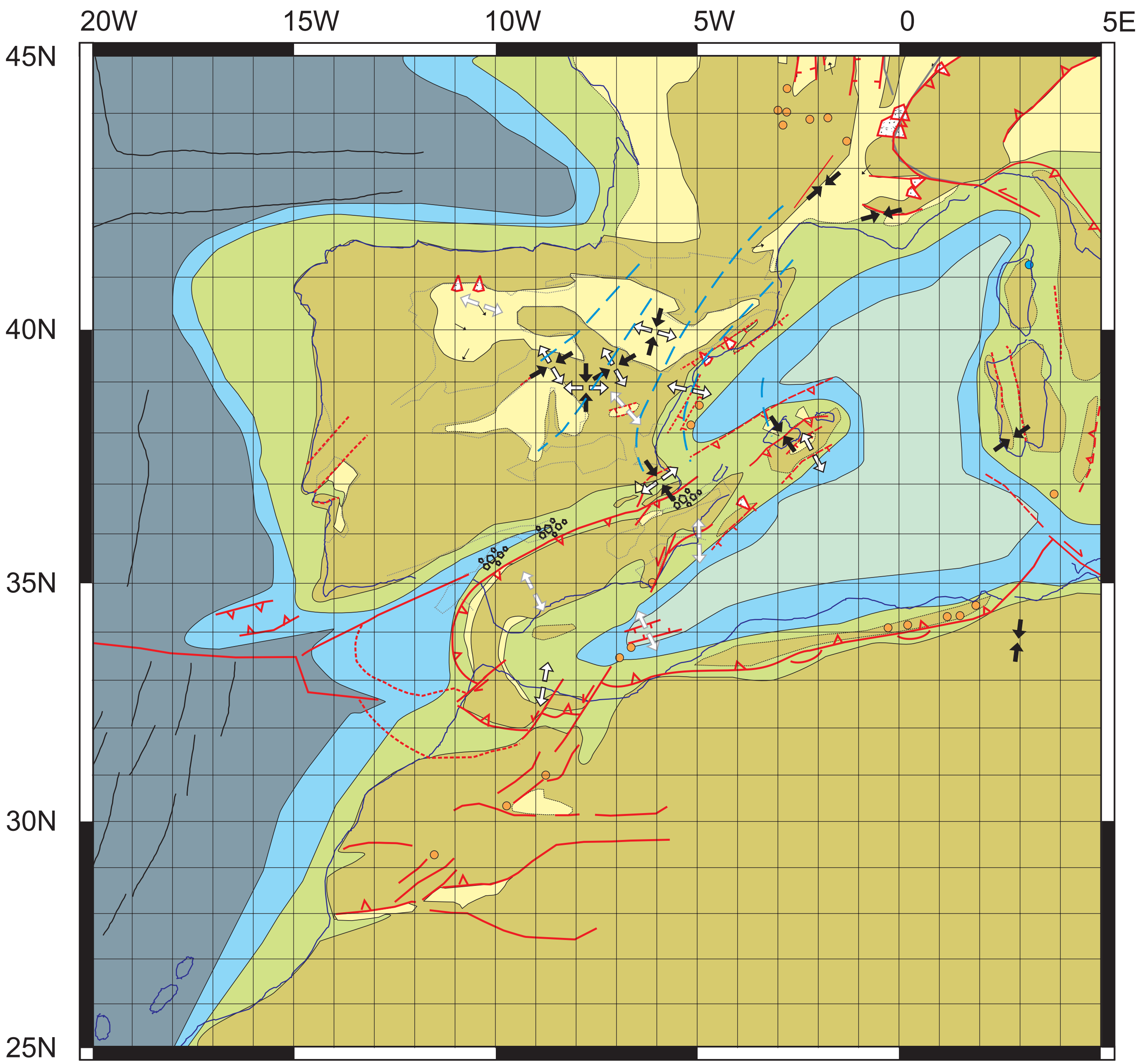


Figure 5.3 (15Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the M. Miocene (L. Langhian - E. Serravallian, 15 Ma). See Figure 4.4 for explanation and text for detailed description.

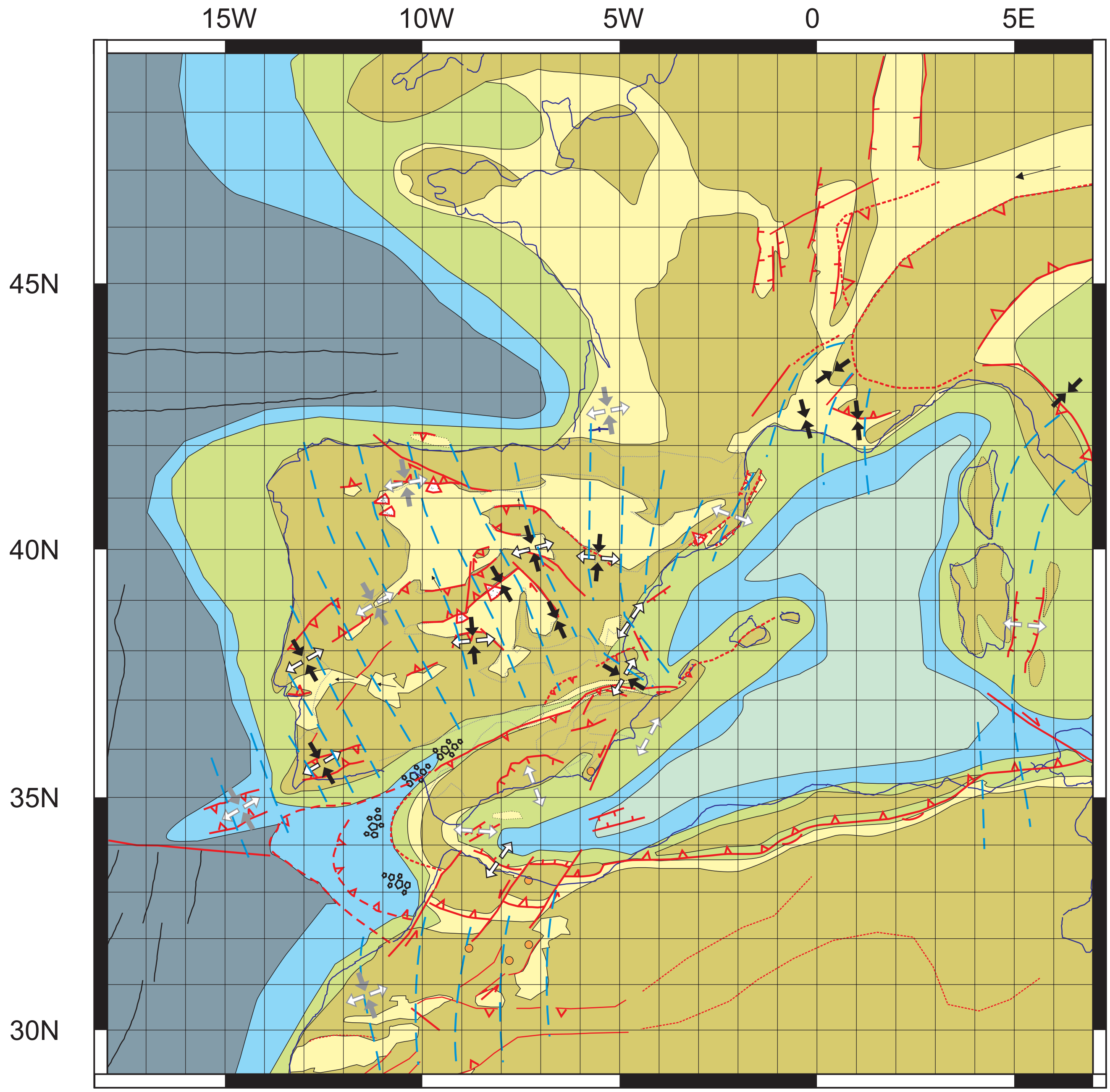


Figure 5.4 (12Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the M. Miocene (L. Serravallian, 12 Ma). See Figure 4.4 for explanation and text for detailed description.

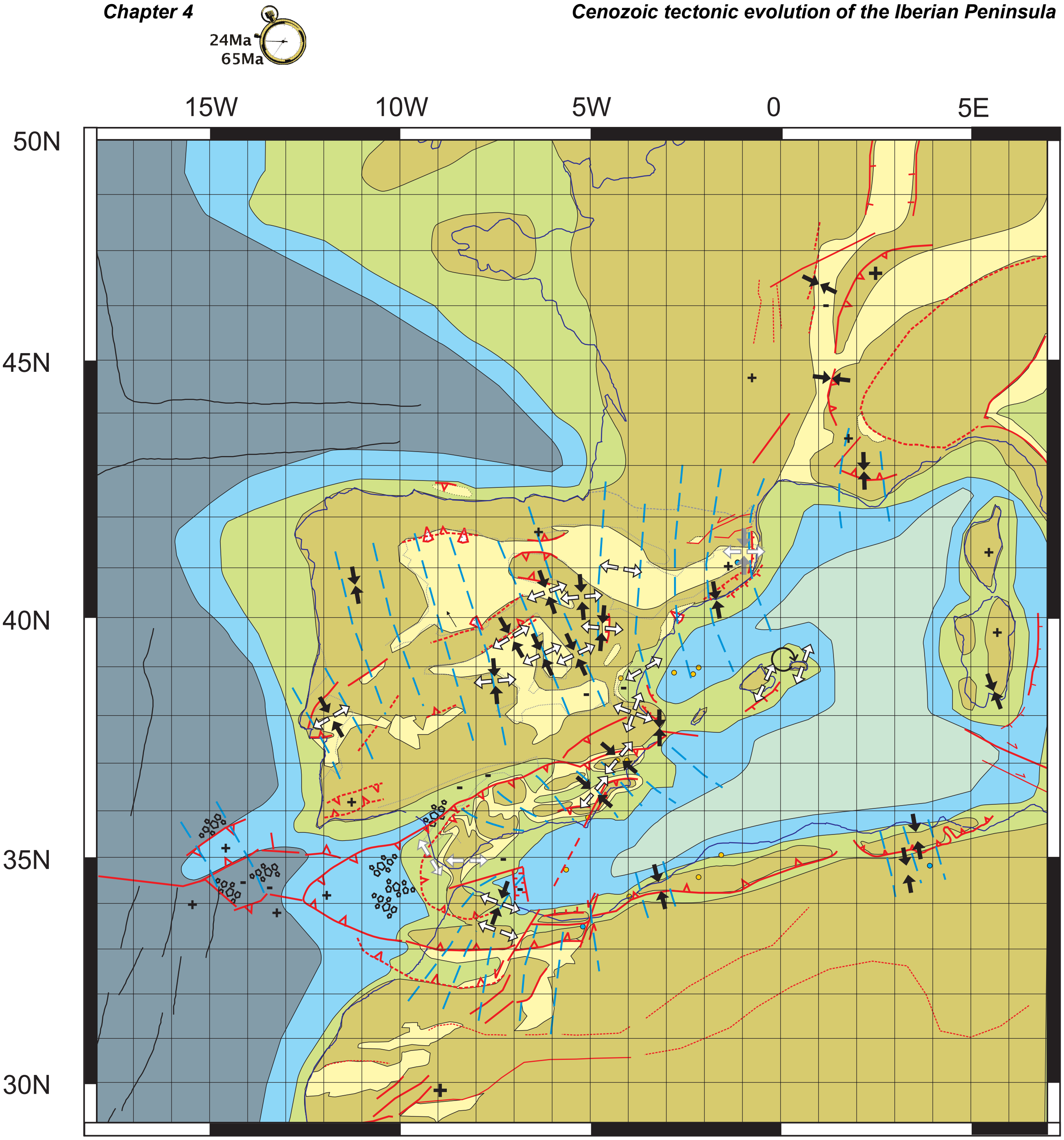


Figure 5.5 (9Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the L. Miocene (Tortonian, 9 Ma). See Figure 4.4 for explanation and text for detailed description.

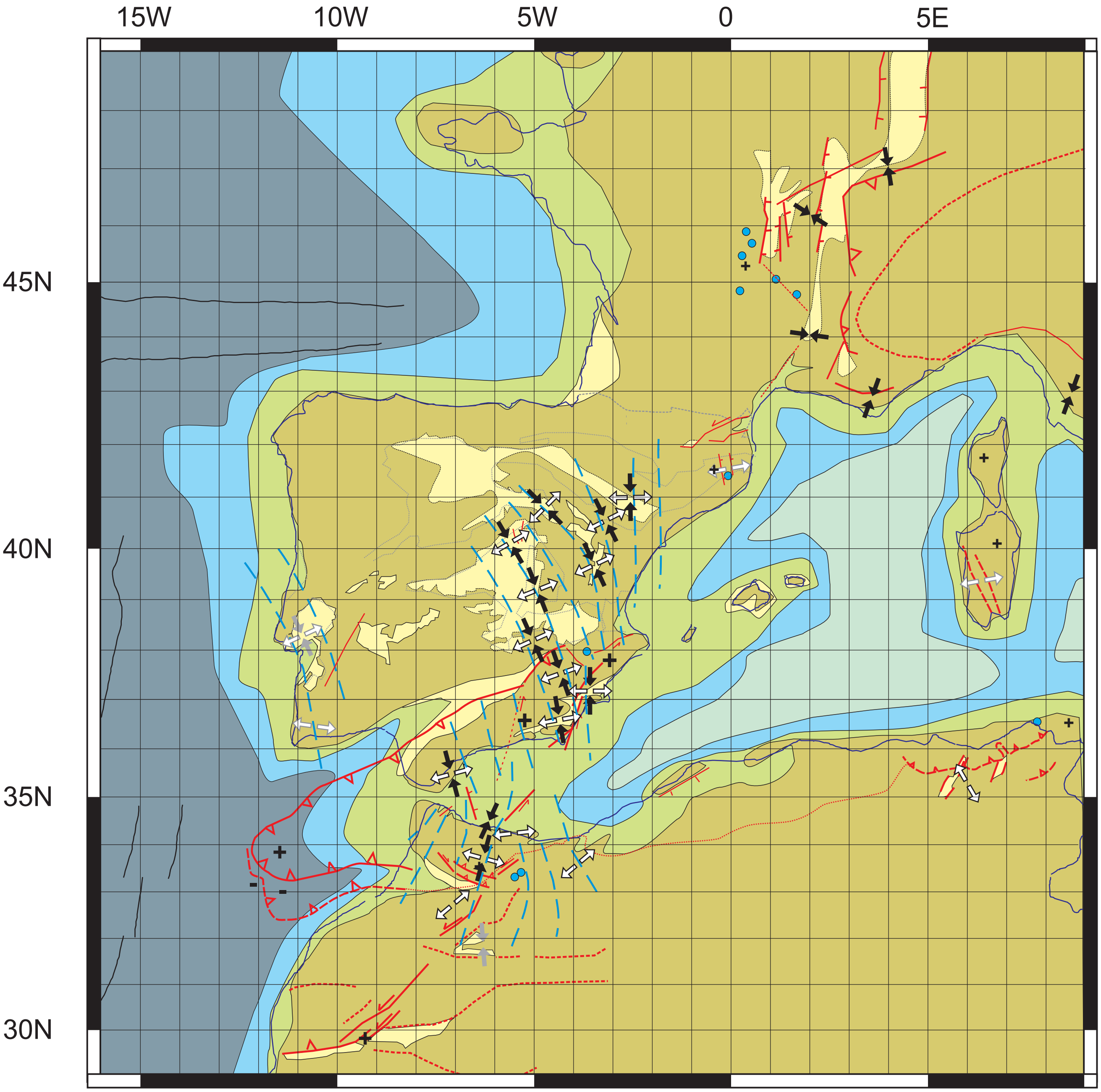


Figure 5.6 (6Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the L. Miocene - E. Pliocene (Messinian – Zanclean, 6 Ma). See Figure 4.4 for explanation and text for detailed description.

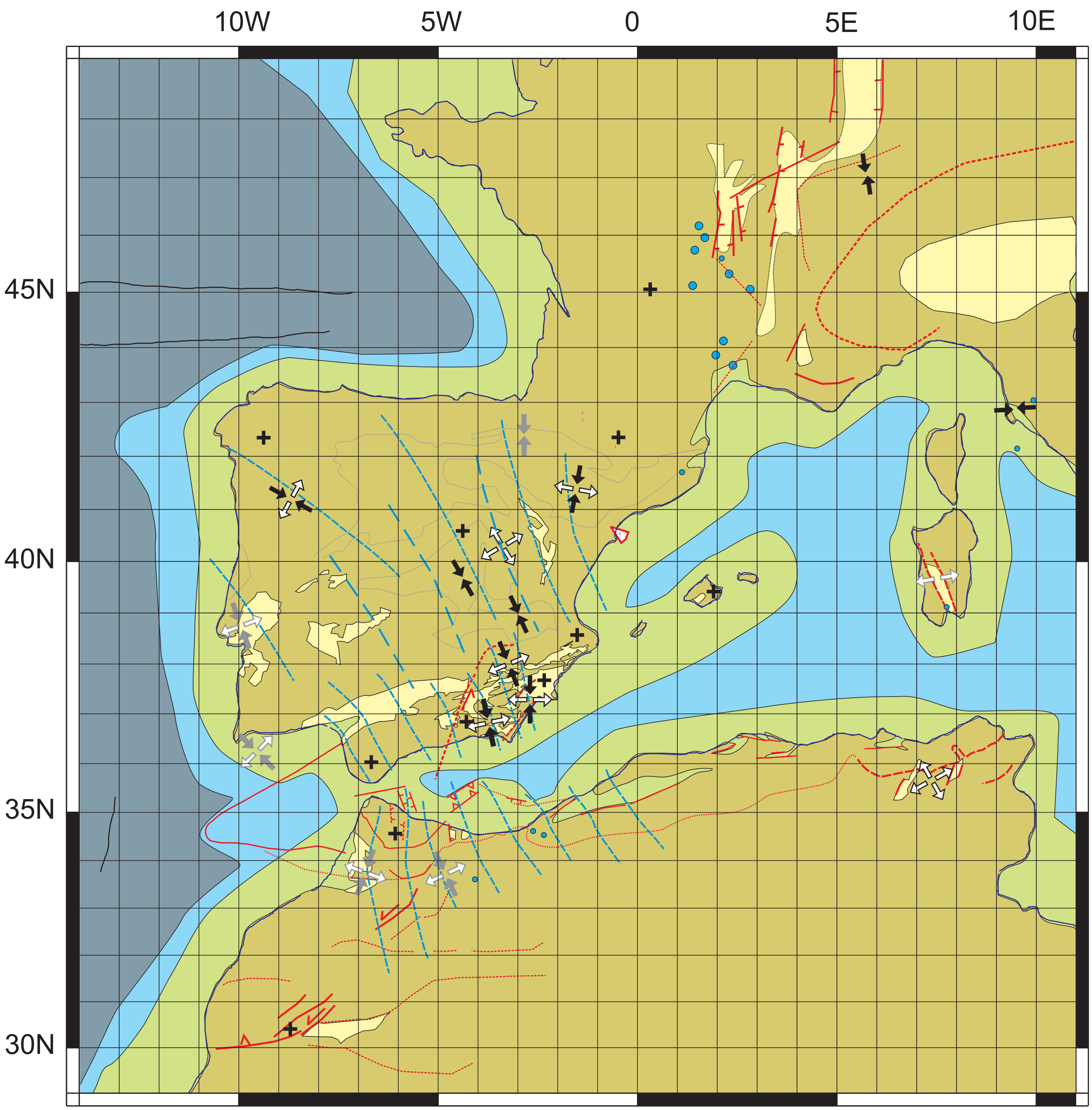


Figure 5.7 (3Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the L. Pliocene (3 Ma). See Figure 4.4 for explanation and text for detailed description.

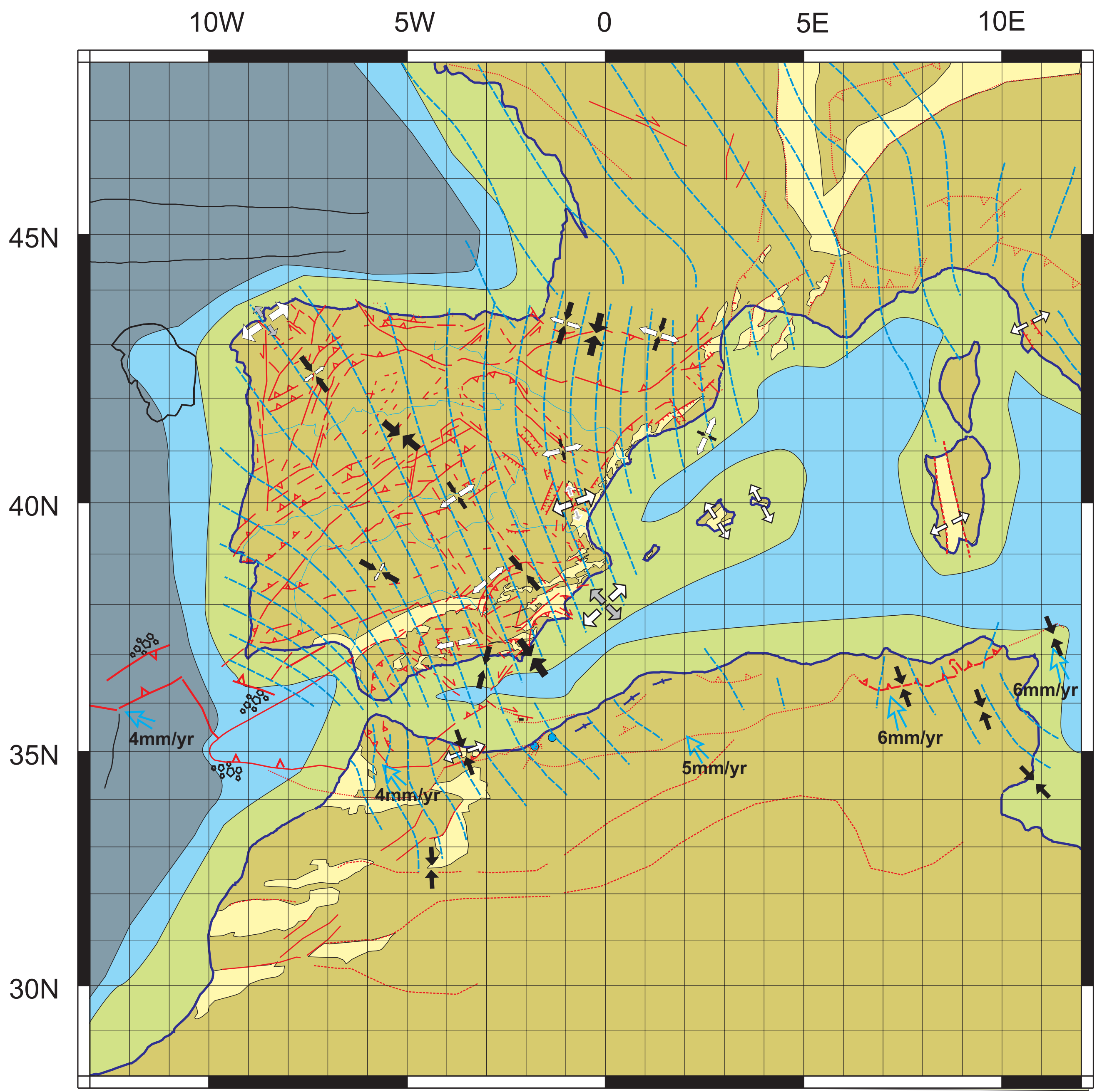


Figure 5.8 (0Ma)
Paleo-tectono-geological reconstruction for the Iberian Peninsula and the western Mediterranean at the Holocene (0 Ma). See Figure 4.4 for explanation and text for detailed description.

Several, presently onshore, extensional basins start their development with very limited sedimentation. In the Herault basin active rifting starts in small fault bounded basins, now buried under later synrift alluvial plain Aquitanian sediments [Sérrane *et al.*, 1995]. During all of the Oligocene, in the Marseille Basin shallow fresh water deposits or non-deposition occurs under N125-105 extension [Hippolyte *et al.*, 1993]. Similar extension direction (N120) is documented in the Nimes/Camarque Basin as well [Villeger & Andrieux, 1987]. Lacustrine sediments dominate in the E. Oligocene in the Manosque basin [Roure *et al.*, 1992].

Under the increasing Alpine crustal wedge, the northern Alpine foreland basin subsides rapidly and expands northwards [Ziegler *et al.*, 1996]. Large detritic fans start to enter the basin (Reuss and Sernf fans) [Sissingh, 1997]. Communication between the marine Alpine foreland basin and the Rhine and Bresse graben exists through two 'inlets', the Jura becomes an isolated island [Sissingh, 2001].

S. Alps - Adriatic domain

The Bergell intrudes in the Southern Alps under E-W orogen parallel extension [Schmid *et al.*, 1996], marking the end of the earliest stage of post collisional shortening related to slab detachment [Ziegler *et al.*, 1996]. To the south the Gonfolite basin develops [Stampfli *et al.*, 1998] and infill of the Piemonte Basin is turbiditic [Bersezio *et al.*, 1993].

N. Africa

In the Atlas ranges, first compressional deformation and related uplift is widespread. Relief is maintained in the Middle Atlas by continuous uplift of the southeastern block of the North Middle Atlas Fault from ~Cenomanian until ~Oligocene [Herbig, 1988]. For the South High Atlas first uplift [Herbig, 1988] related to N-S to N020^{o} compression [Fraissinet *et al.*, 1988] is assumed.

27 Ma, L. Oligocene (Chattian), Figure 4.10

General

The plate boundary running from Kings Trough to the Pyrenees remains active from 44Ma until 25Ma [Srivastava *et al.*, 1990], deformation along this plate boundary is waning. WNW-ESE extension is dominant in the Provencal-Ligurian basin and starts to propagate into the Valencia trough. The onset of drifting apart of the Baleares, Sardinia/Corsica and Iberia. Rotation of Corsica/Sardinia results in collision with 'Alpine Corsica'. Extension starts to invade the southern crustal segments of the Alboran block. Subduction of Tethys oceanic crust has ended in the west, after most of the oceanic crust has been subducted. The Alboran/Betics block is still not mechanically coupled to the southern margin of Iberia.

Detail

Western margin

The generation of a post-Eocene but pre-Aquitanian unconformity on- and offshore in the Arriba/Lisboa area is correlated to NW-SE directed compression [Lepvrier & Mougenot, 1984]. In northwestern Portugal the formation of a major Paleogene erosion surface results in deposition of clay-sandstone in planar topography [Cabral, 1989]. The Castelo Branco basin is the locus of deposition of conglomerate and sandstone [Dias & Cabral, 1989].

Northern margin

Activity along the northern Iberian margin is renewed [Boillot & Malod, 1988] resulting in the development of horst and graben parallel to the margin [Boillot *et al.*, 1979] under NE-SW compression, as documented in the Asturian Basin [Lepvrier & Martínez-García, 1990]. Along the Asturian coast marine environments are present in the San Vicente de la Barquera and Santander areas [Lepvrier & Martínez-García, 1990]. The As Pontes basin is being formed. Strike-slip activity along a curved fault plane leads to active thrusting and normal faulting from 28.7Ma [Huerta *et al.*, 1996].

Central Iberia

The first appearance of conglomerates with a northern (Pyrenean) source area in the Rioja [Jurado & Riba, 1996] show that the Cantabrian Cordillera is close to its present-day position. Activity and renewed uplift of both sides (Cameros and Sierras Obarense is dated at 27-26 Ma [Muñoz Jiménez & Casas Sainz, 1997]. For the rest, central Iberia is rather quiet. Deformation of the Iberian Range is inferred from the lateral restriction of the last phase of sedimentation (orange marls) and observed in the Maestrazgo/Serriana de Cuenca and Montes Universalès [Adrover *et al.*, 1983]. The monotone alternation of canalized sandstone and lutite developing into gypsum [Diaz Molina & Lopez Martinez, 1979] deposited in the neighboring Loranca Basin, interpreted as the distal front of a wet alluvial fan [de

Torres Perezhidalgo *et al.*, 1983], does restrict any deformation of the Iberian Range to very limited. For the northern and northeastern SCS deformation possibly related to activity of the Iberian Range [De Bruijne *et al.*, 2001] is inferred [Portero Garcia & Olivé, 1983]. N045°-055° directed compression results in sinistral reactivation of N060/090 trending and dextral reactivation of N140°/160° trending faults. These movements form two small basins (Corneja and Ambles) in the Sierra de Gredos [Babín Vich & Gómez Ortiz, 1997].

SE Iberia and Betic realm

Erosion products of the External Prebetics arrive in the Internal Prebetics as detritic red continental series [Fontboté & Vera, 1983] and in the Subbetics where detritus has a NE provenance [Geel, 1996]. The coastline is situated over the Internal Prebetics [HNPC, 1992]; its eastern part is dominated by shallow marine limestone [Fontboté & Vera, 1983].

Within the Internal Zones of the Betics crustal thickening culminates resulting in a peak metamorphism before 25 Ma, coinciding with the onset of extensional thinning [Monié *et al.*, 1994]. The Greater Kabylian Naciria massif shows a low temperature event (Ar-Ar), which suggests an extensional phase [Monié *et al.*, 1988], before it starts overthrusting to the south at around 25 Ma. Extension of the Mulhacen complex from 28-23Ma is shown by PTt data [De Jong, 1991]. This extension, inferred from PTt modelling as well [Platt & Whitehouse, 1999], leads to breaking up of the Internal zones into several blocks, the western one of which becomes the Betic/Riffian or Alboran Block (bounded to the north by the North Betic fault, to the south by the Jebha fault) [Durand Delga & Olivier, 1988]. The extension within the Malaguide domain creates basin conditions that vary with respect to local tectonic activity. Series that are (still) submerged in parts of the domain are being eroded under continental environments in other parts [Fontboté & Vera, 1983]. This (L.Oligocene – E. Aquitanian) extensional rifting is only exposing the Malaguide realm because the developing grabens are filled with Malaguide detritus exclusively [Geel, 1996] (Alozaina formation). In the Eastern realm, similar sediments are deposited (Ciudad Granada). To the west of the Internal Zones, in the Internal Dorsal the transgressive series of "arenisca de Horca" (clay rich marl) indicates erosion of the Mesozoic cover and basement of the Malaguide and maybe even Alpujarride [Fontboté & Vera, 1983]. Towards the south and southwest, a deepening of the marine environment is suggested by the lateral change from Predorsalian microbreccia-limestone to up to 1200m Oligocene flysch in the Mauritanicas [Fontboté & Vera, 1983].

S. Pyrenees and Ebro

The southern front of the Pyrenees is active: the External Sierras develop, leading to erosion (end of sedimentation cycle) of the Jaca Basin [Teixell, 1996]. The exhumation of the Maladeta slowed down significantly [Fitzgerald *et al.*, 1999]. All along the border with the Ebro Basin series of progressive unconformities develop (last major tectonic?) [Muñoz *et al.*, 1983] and the southern foreland starts to be filled in with conglomerates [Fitzgerald *et al.*, 1999]. Towards the south in the Ebro Basin, these conglomeratic series pass into fluvial redbeds and lacustrine sediments [Teixell, 1996]. In the southern part of the Ebro Basin evidence for N020°-030° compression is widespread [Guimerà, 1984]. Although less pronounced, deformation related to this compression can be observed in the CCR as well, where fractures are being reactivated obliquely under sinistral shear of the CCR boundary faults [Alsaker *et al.*, 1996].

N. Pyrenees and SW France

Normal faulting in the Jura/Alpine foreland parallel to the Alpine belt and its foreland basin [Wildi & Huggenberger, 1993]. Infill of the Limagne and Bresse Rift with shale and sand layers [Bois, 1993] is related to active rifting. Active extension migrates southward through the Alpine foreland. Although Pyrenean compression is still observed in the Languedoc [Roure & Coletta, 1996] and folding in the Aquitanian basin [Viallard, 1985], contemporaneous ~N110 directed transtension is widespread in the same region [Roure & Coletta, 1996]. Closer to the Alpine front and in relation to this, more disperse directions of extension (155-015) occur in the Marseille Basin [Hippolyte *et al.*, 1993]. In the Narbonne basin, the St Chinian thrust is reactivated extensionally [Roure *et al.*, 1994] and active rifting in the Herault basin results in small fault bounded basins, now buried under later synrift alluvial plain Aquitanian sediments. Along strike to the north, the Alès Basin is still active as well [Sérrane *et al.*, 1995] and just as well, the Valensole/Manosque basin (around sea level) documents extension [Roure *et al.*, 1994].

S. Alps - Adriatic domain

The southern Alpine domain is tectonically very active, leading to steep gradients of the Southern Alpine wedge. The Bergell experiences rapid uplift and erosion due to backthrusting of the Central Alps over the Southern Alps [Schmid *et al.*, 1996]. Erosion of Bergell is witnessed by turbidites in the Gonfolite Basin with a clear Bergell Provenance [Bersezio *et al.*, 1993]. The entire Lombardian fore deep experiences deeper water environments [Ziegler *et al.*, 1996].

Catalan-Sardinian margin

Extension is entering the Valencia Trough area, but the Baleares are still linked with Eurasia (communication of mammals) [Pomar Goma, 1983]. The Sardinia Basin is being generated under N130-140 extension [Vially & Trémolières, 1996] and volcanic activity starts [Monaghan, 2001].

For northeastern Alpine Corsica the last compressive tectonics is being documented [Egal, 1992]. Alpine Corsica is now attached completely to the Corsica/Sardinia.

N. Africa

The south central High Atlas is uplifting [Görler *et al.*, 1988], but for the rest rather quiet [Frizon de Lamotte *et al.*, 2000].

24 Ma, L. Oligocene - E. Miocene (L. Chattian - E. Aquitanian), Figure 4.11

General

The active plate boundary is gradually relocated from the north (Kings Trough and the Pyrenees) to the south of Iberia (Azores-Gibraltar system) after 25Ma [Srivastava *et al.*, 1990]. A major sedimentary break in the Iberian Neogene basins is observed [Calvo *et al.*, 1993] and might well be related to the plate boundary reorganizations. Convergence rates between Eurasia (of which Iberia is a part now) and Africa change between 25-20 Ma from fast to slow [Lips, 1998]. Therefore, activity along the new plate boundary is limited and with a dextral wrench component [Ziegler *et al.*, 1996]. Extension starts to invade the internal zones of the Betics/Alboran, breaking the Kabylian and Calabrian/Peloritan blocks from the Balearic/Betics/Alboran.

Detail

Western margin

Along the Portuguese margin, the Sado Basin is emerged, stable and eroded [Pimentel & Azevêdo, 1994], while in the Lower Tajus basin a transgressive series is being deposited [Azevêdo, 1991]. Deformation off shore Galicia [Murillas *et al.*, 1990] and strike-slip activity onshore Galicia leading to the first sediments in the As Pontes Basin (L. Oligocene until ~E. Miocene) indicate N-S compression [Santanach Prat, 1994]. From geomorphological evidence, a new erosive period (due to compression?) [Pagés Valcarlos & Vidal Romaní, 1998] would provide theses sediments. Erosion of the western edge of the nearby Duero Basin supports this as well (see *Central Iberia*).

Northern margin

Last important compressional deformation of the Asturian margin took place until ~25Ma, resulting in an offshore horst-graben structure controlling the Cenozoic sedimentation [Boillot *et al.*, 1979] and the main inversion of the Penas Trough (between Le Danois Bank and Asturian Massif) [Ziegler, 1988]. The top of this Danois bank is very close to sea level during the E. Miocene [Boillot *et al.*, 1979]. The Asturian margin is the locus of transgressive sedimentation [Boillot *et al.*, 1979], and along the Asturian/Cantabrian coast compressional deformation Nummulitic Eocene limestone documents N-S compression, *own data.* The As Pontes basin shows strike-slip and thrust activity [Huerta *et al.*, 1996].

Central Iberia

The sedimentation pattern in the Duero Basin suggests a major tectonic phase. In the W/SW Duero E. Miocene sediments are absent, which suggests a further retraction of the basin edge [Portero Garcia *et al.*, 1983]. Along both the northern and eastern border conglomerate/coarse sandstone are being deposited discordantly over lacustrine carbonates [Mediavilla *et al.*, 1996], in the E. Duero alluvial fans are infilling from the Sr. De la Demanda that is located to the east of the basin. More evidence for tectonic activity comes from the nearby Almazan Basin, where N030° compression is demonstrated [Maestro *et al.*, 1997]. The entire Duero basin shows a gentle sinking towards N and W accommodated by small slip along normal faults under EW-extension [Santisteban *et al.*, 1996b]. This suggests limited foreland basin type development of the northern Duero Basin related to increased loading by southward thrusting of the Cantabrian Cordillera [Marín *et al.*, 1995]. The relative uplift of the southwestern edge was leading to incision of the Duero with paleocurrents towards the west for the first time in this area [Santisteban *et al.*, 1996a]. The present-day drainage patterns starts establishing [Santisteban Navarro, 1998], progressively changing a larger part of the Duero Basin into exoreic. Evidence for uplift of the southeastern border of the Duero Basin (the Spanish Central System) comes from 120m arcosic fans [Portero Garcia *et al.*, 1983] and the region of Penaranda-Alba where E. Miocene coarse conglomerate and sandstone proceed from the SE [Corrochano & Carballeira, 1983b]. At least the northeastern SCS is tectonically active due to N050°-090° directed compression. This results in strike-slip deformation along the southern border fault [De Bruijne, 2001] and limited thrusting and related strike-slip deformation of the Mesozoic/Paleogene cover [Sanchez Serrano, 1991]. In the central SCS (Sierra de Gredos) N050° directed compression is forming the Ambles and Corneja Basins [Babín Vich & Gómez Ortiz, 1997].

The Iberian Range shows major activity [Adrover *et al.*, 1983] and is thrusting over the southern edge of the Ebro Basin [Guimerà, 1984]. A widespread erosion surface that is covered by L. Oligocene deposits is being deformed thoroughly in this tectonic active period. After that no significant deformation, erosion or uplift occurs (another, Early Miocene erosional surface covered by Lower Miocene sediments is only gently deformed) [González *et al.*, 1998]. The Neogene basins within the IC are formed as small grabens with limited sedimentation. At the other side of the Iberian Range, progressive unconformities are formed at the border with the Loranca Basin [Muñoz Martín, 1993]. Large alluvial fans enter the Loranca Basin from the SSE (Tortola) and E. Partly related to this activity, the Sierra Altomira experiences its first uplift [Muñoz Martín & De Vicente, 1998]. Further south, the Guadiana, Jucar and southern Tajo basin are the locus of lacustrine limestone deposition [Adrover *et al.*, 1983].

S. Pyrenees and Ebro

In the western Pyrenees the External Sierras stop being developed [Teixell, 1996] just as in the southeastern Pyrenees thrusting terminates at around ~25Ma [Morris *et al.*, 1998], followed by a southeast shift of major denudation. A relatively quiet period occurs: a Partial Annealing Zone is developed in the Maladeta [Fitzgerald *et al.*, 1999]. Along the entire northern border of the Ebro Basin the last stage of the progressive unconformities takes place [Muñoz *et al.*, 1983]. In the Jaca/Gauss basins the Huesca and Tremp/Gauss fans start entering the basin [Vincent & Elliott, 1996].

Catalan-Sardinian margin

The northern Valencia Trough is starting to rift actively [Roca & Deselgaulx, 1992]. Calc-alkaline volcanism related to the rifting and the northwestward subduction is active in the area until Late Burdigalian [Bois, 1993]. A marine transgression is entering the Valencia realm [Roca & Deselgaulx, 1992]. In the Barcelona/Sant Feliu half grabens synrift sediments are deposited, restricted to deeper parts of the block faulted troughs [Roca *et al.*, 1999]. Throughout the entire Valencia Trough a major unconformity separating 2 main tectono-sedimentary stages is observed [Martínez del Olmo, 1996]. In the Catalan Coastal Ranges extension starts [Gueguen *et al.*, 1998], leading to a considerable decrease in relief. Estimates of elevation are of the order of ~300m, *Roca pers.*

The eastern margin of Menorca is rifting into a NW-SE trending basin [Rehault *et al.*, 1984] and leading to breaking of Kabylia from Menorca at ~23Ma. Syntectonic conglomerates are being deposited in this region [Pomar Goma, 1983]. Early volcanism occurs along the NW-SE oriented North Balearic Fracture Zone [Mauffret *et al.*, 1995]. Related to this breaking apart, Mallorca/Menorca show a 20-degree clockwise rotation since timing of magnetization (prior to Oligo-Miocene) but before upper Miocene. Related to active thrusting on Mallorca [Freeman *et al.*, 1989] unstable platform sediments, irregular topography, a non -linear coastline and mixed shelf sedimentation characterize Mallorca [Ramos-Guerrero *et al.*, 1989].

Between northern Sardinia and the Alps, a complex pattern of deformation is observed: sinistral strike-slip between Corsica and N. Apennines [Dewey *et al.*, 1989] along the major transfer zone linking the Alpine and Tethyan subduction fronts. Sinistral strike-slip along near E-W trending faults in eastern Sardinia/Corsica is related to transpressive tectonics [Carmignani *et al.*, 1995]. The rifted basin that developed in central and south Sardinia under N130-140 directed extension [Vially & Trémolières, 1996], is now the locus of important post-rift sedimentation [Monaghan, 2001], continental environments prevail [Carmignani *et al.*, 1989]. In the meanwhile, to the south of Sardinia, a fold-and-thrust belt is active in the Sardinia Channel because the Sardinian margin collides with the Maghrebian-Sicilian block [Catalano *et al.*, 1995]. SE-verging imbricated thrust units shed arcosic turbidites to the SE [Catalano *et al.*, 1989]. Most likely this belt is related to an accretion process in the ongoing NNW ward subduction. Related calc-alkine volcanics are observed in a volcanic arc running through Corsica/Sardinia [Bois, 1993], the Nice area and the Western Alps.

SE Iberia

The major rifting phase in the Valencia area [Roca & Deselgaulx, 1992] results with perpendicular ~170 extension in the development of NNW-SSE and ENE-WSW fault systems in the Prebetics [De Ruig, 1991]. Detachment between Iberia and the Internal Betics (Alboran) is still not active, as inferred from the fact that no compressional deformation that would be the effect of collision, is observed yet! Moreover detritus is arriving in the Subbetics with NE-provenance only [Geel, 1996]. Block faulting in the area leads to related complex sedimentation patterns. In the External Prebetics both continental redbed, conglomerate and sandstone deposition [De Ruig, 1991] and discordant shallow marine deposits [Fontboté & Vera, 1983]. In the eastern Internal Prebetics Aquitanian shallow marine shelf deposits onlap transgressive over red detritic continental Oligocene deposits, while in the western part, shelf – and slope deposition continued without a major sedimentary break [Geel *et al.*, 1992].

Betic realm

Extensional systems are thinning the Malaguide/Alpujarride stack [Lonergan & White, 1997], breaking the Internal Betics up into several blocks [Durand Delga & Olivier, 1988]. Synchronous with the opening of the W. Mediterranean, WNW-directed extensional shear is thinning Greater Kabylia at ~25Ma [Saadallah & Caby, 1996]. The Malaguide starts a strong rotation (up to 200 degrees to L. Miocene) [Allerton *et al.*, 1993] while the Alpujarride starts to be exhumed rapidly from below the Malaguide complex (until 19Ma) [Johnson *et al.*, 1997]. In the meantime, a peak in compressional deformation is inferred for the Ghomaride [Maate, 1996]. Separation between Kabylia and Alboran (Rif/Betics) [Wildi, 1983] cannot yet have advanced far: erosion products of Kabylian basement are observed in mass flows in the Velez Rubio corridor [Geel & Roep, 1998]. Erosion of Kabylia can have been produced by emersion related to shoulder uplift in the extensional setting in the Algerian Basin [Wildi, 1983]. Evidence for the end of subduction under the Alboran comes from PTt data [De Jong, 1991] and coincides with the estimate for slab detachment under the Alboran shortly before 22Ma [Zeck, 1996]. A decrease in the convergence rate between Eurasia and Africa occurs [Lips, 1998]. This suggests that ‘subductable’ oceanic crust has been consumed and the Alboran continental block is about to start overthrusting the northwestern margin of Africa. To the south and southwest, flysch depocenters are filled in rapidly. In the Campo de Gibraltar a flysch trough develops that is deepening to the west and in the Predorsalian a sequence of marl and clay with Numidian sandstone (flysch-type) is deposited [Fontboté & Vera, 1983]. In the Dorsalian the same type of sedimentation dominates, forming the clay-rich marl of the 'arenisca de Horca' in the internal part and sandstone and clay in the external part [Fontboté & Vera, 1983]

N. Pyrenees and SW France

Even though a new start of contraction is documented [Rocher *et al.*, 2000], the North Pyrenean Fault becomes inactive [Roure *et al.*, 1989] and even close to the Pyrenees deformation is transtensional or pure extensive in a general ~110 direction, rotating more NW-SE towards the east. In southwestern France, compression directions rotate clockwise to approximately N060^{o}, perpendicular to the western Alpine front [Bergerat, 1987]. These facts show that the remnant compression in southwestern France is related to the Alpine collision and waning of the Pyrenean collision results in general extension in north eastern Iberia/Gulf of Lions. The graben of Rosselo develops along the Tet and Tec faults with a continental infill interrelated to small marine fluxes [HNPC, 1992], and in the Narbonne basin extensional reactivation of St Chinian former frontal thrust occurs [Gorini *et al.*, 1991]. On the basinward part of the margin of the Gulf of Lions, synrift Oligocene-Aquitanian sediments are deposited in continental environments [Sérrane *et al.*, 1995]. A thick sequence of synrift alluvial plain Aquitanian sediments is accumulating in the Herault basin [Sérrane *et al.*, 1995]. More to the east, the former Pyrenean relief has been reduced; generally low relief is observed with sedimentary basins close to or at sea level. Infill of the Valréas basin even becomes marine due to thermal subsidence of the Oligocene rift basin [Roure & Coletta, 1996], the Camarque Basin is filled with synrift sediments close to sea level [Sérrane *et al.*, 1995]. Marine environments (evaporite) are dominant in the Aix area during the latest Oligocene but in the Marseille Basin shallow fresh water deposits are accumulating during major subsidence of 1.4m/1000 year. This basin is a large graben perpendicular to the 155 extension in this region [Hippolyte *et al.*, 1993]. And last, the Valensole/Manosque basins witnessed not much topography either, where around sea level and related to extension along Durance fault [Roure & Coletta, 1996], in the Manosque basin evaporites dominate [Roure *et al.*, 1992]. The shale with sand layers deposited in the Limagne and Bresse Rift form the last sequence for the Limagne, the Bresse rift will experience late subsidence in Miocene-Pliocene [Bois, 1993]. The northern Alpine Foreland Basin is deepening eastward. Non-deposition in the Jura [Wildi & Huggenberger, 1993] changes to freshwater molasse in the western parts and even marine environments towards the eastern extreme [Andeweg & Cloetingh, 1998]. Major alluvial fans enter the basin from the rising Alps in the south [Sissingh, 1997].

S. Alps - Adriatic domain

In the Bergell area (Western Alps) an end of the rapid uplift is observed [Schmid *et al.*, 1996], which led to a regional break in sedimentation in the Gonfolite Basin [Bersezio *et al.*, 1993]. A general deep marine unconformity is being formed, with the largest amount of erosion in the west [Bernoulli *et al.*, 1989].

N. Africa

In the North African realm, gentle compressional deformation is observed. The North African Flysch Trough shows south-directed contraction [Martínez-Martínez & Azañón, 1997]. The northern Moroccan Meseta is emerged and Numidian flysch from Algeria are arriving from the S [Wildi, 1983]. In the Prerif-Mesorif a syntectonic turbiditic sandstone (Zoumi) shows S-SE provenance [Morley, 1987], indicating uplift of the South Central Atlas. This uplift is related to subsidence in the southern foreland basin of the High Atlas, the Ouarzazate basin that is still in an initial marine stage [Görler *et al.*, 1988].