Origin of the Permian–Triassic Iberian Basin, central-eastern Spain

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Abstract

The Iberian Basin was an intracratonic rift basin in central-eastern Spain developed since Early Permian times. The basin boundary faults were normal, listric faults controlling an asymmetric extension propagating northeast with time.

Hercynian or older lineaments controlled the orientation of the Iberian Basin and extension was accommodated basically in the hanging wall block by the formation of secondary grabens and a central high. The basin was related with the coeval Ebro, Catalan and Cuenca–Mancha Basins and their connections are discussed.

Subsidence curves show that the Early Permian–Early Jurassic period of extension can be subdivided into three rifting episodes and a flexural one. Extension factor increases from 1.17 in the northwest to 1.29 near the Mediterranean coast. The increasing extension rate was accommodated by transfer faults trending NNE–SSW, more important in the Levante area. The rift evolution is intermittent and seems to reflect distinct stress fields.

The collapse of the late Hercynian orogen and related increased heat flux, extension and rifting is the most probable origin of the Iberian Basin and related basins. The origin of the Catalan and the Valencia–Prebetic Basins is related to the southwards migration of the Hesse–Burgundy Rift along the eastern margin of the Iberian Microplate.

Keywords: Iberian basin; rift basin; asymmetric extension; subsidence curves

1. Introduction

The Iberian Ranges are a linear structure trending NW–SE in the northeast edge of the Iberian Microplate; it is an intracratonic, folded segment of the Alpine Chains bounded to the northeast by the ancestral Ebro Massif, now under Tertiary sediments, the Pyrenees and the Catalonian Ranges. It started to develop as a rift basin during the Early Permian and experienced several extensional periods during the Late Permian, the Mesozoic and the Cenozoic and at least two major compressive events that caused structural inversion, folding and thrusting, but no magmatism nor metamorphism (Fig. 1).

The Iberian Basin was related to other contemporaneous basins such as the Pyrenean, Ebro, Catalan and Cuenca–Mancha. This paper deals with the early phase of extension (Early Permian–Late Triassic) of the Iberian Basin, the nature of the sedimentary infilling, the geometry of the basin boundary faults and speculates on the causes of its origin and orientation. Detailed sedimentologic and stratigraphic descriptions of the Permian and Triassic sediments of the Iberian Ranges can be found in Virgili et al. (1983); Sopeña et al. (1983, 1988) and López-Gómez and Arche (1992a).

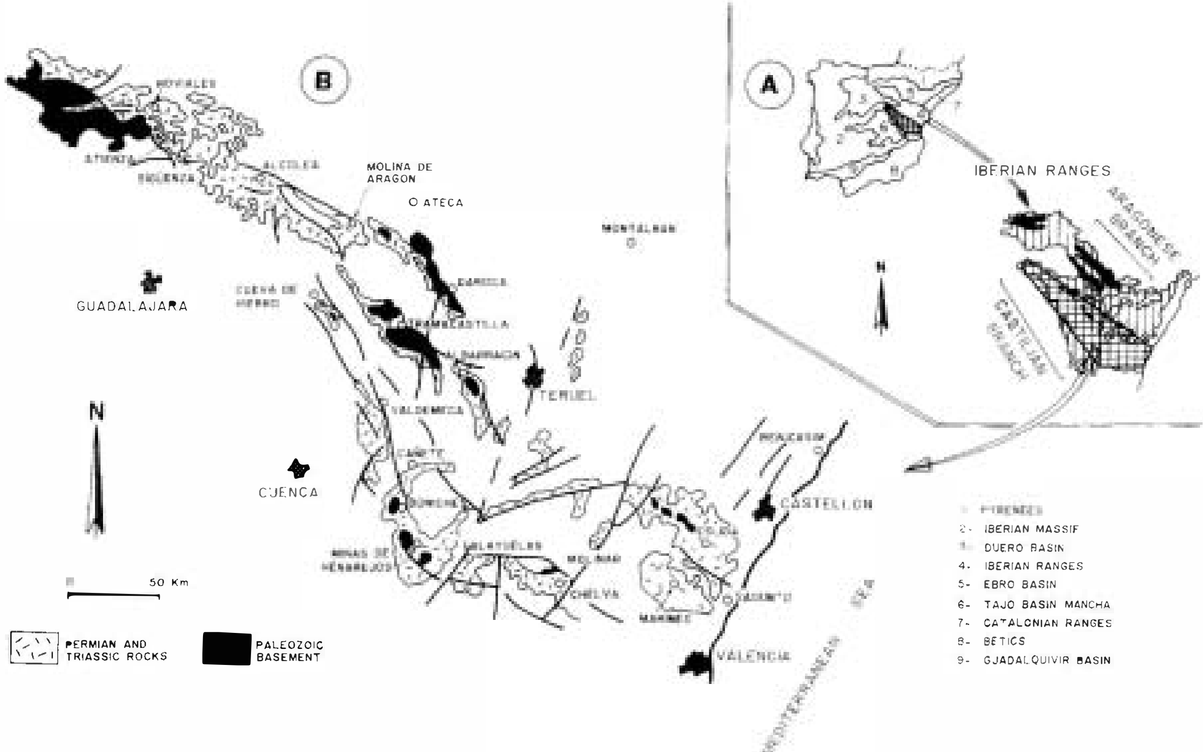


Fig. 1. The Iberian Range and its main Permian–Triassic outcrops.

2. The Hercynian basement

At the end of the Carboniferous, the Iberian Microplate was part of the Hercynian (or Variscan) Belt, a wide linear structure affected by intense deformation and magmatism during the Carboniferous, stretching from the Southern Appalachians to the Bohemian Massif. This megastructure was originated during a continent–continent collision between the North America (Laurentia) Plate, Fennoscandia Plate, several South Europe Microplates and Africa Plate. Deformation took place in several phases (Late Devonian–Late Carboniferous) (Arthaud and Matte, 1977; Dewey and Burke, 1973; Lorenz and Nicholls, 1974, 1984) and led to crustal shortening and thickening, regional metamorphism and widespread granitization (Matte, 1986).

The Iberian Basin developed on a Hercynian basement of Ordovician–Silurian slates deformed in kilometre-sized folds trending NW–SE to N–S and with eastwards vergence (Capote and Gonzalez-Lodeiro, 1983), affected by very low-grade metamorphism (chlorite–pyrophyllite zone). These materials are related petrologically and structurally with the so-called West Asturian–Leonese Zone of the Iberian Massif, the main part of the Paleozoic continental Iberian Microplate. Late Hercynian or older structures trending NW–SE (faults and folds) in the suture of the Iberian Massif and the Ebro Massif were inherited by the Iberian Basin master faults.

To the west of the Iberian Basin, the Hercynian basement is of a different nature: low P–high T metamorphism rocks of Precambrian–Ordovician age and extensive S-type granitic batholiths of the Central Iberian Zone. Ibarrola et al. (1987) differentiated two main types of Hercynian granitic rocks in this area, one, small in volume, of autochthonous peraluminous granites and another, the bulk of the intrusions, of allochthonous intrusive massifs, intermediate to acidic (mainly adamellites) and some posthumous postkynematic leucogranites. The age of the adamellites ranges from 344 ± 8 to 275 ± 11 m.y. and that of the leucogranites from 305 ± 6 to 286 ± 18 m.y. As will be described later, there are some calc-alkaline volcanic rocks (mainly andesites) in the base of the infilling of the Iberian Basin, dated by Hernando et al. (1980) as 282 ± 12 m.y.; there must be a genetic relationship between the andesites and the granites (Muñoz et al., 1985), but it is not definitely proven.

Dyke swarms of younger age have been found in the Central System, east of the Iberian Ranges (Galindo et al., 1994). The oldest family, trending E–W, have a tonalitic to leucogranitic composition and an age of 296 ± 3 m.y. (Late Carboniferous) (Galindo et al., 1994), contemporaneous with the latest granitic bodies composition and the younger dyke family, trending N–S, have a granitic composition and an age of 245 ± 7 m.y. to 220 ± 5 m.y. (Anisian–Karnian) (Galindo et al., 1994) and are related to an extensional episode coeval with the main Triassic rifting period in the Iberian Basin.

Crustal thickness is rather uniform in central Spain (Zeyen et al., 1985; Banda, 1987; Dañobeitia et al., 1992), about 32 km with three levels superimposed of 12, 11 and 9 km; a thickening of about 4 km (36 km) was detected in the Central System area and a thinning of 7 km (25 km) in the southeast Iberian Ranges, near the Mediterranean. There are no deep roots in the 1100-km-wide Hercynian Belt in Iberia and some of the main crustal reflectors, dipping about 30° near the surface, flatten out at about 11–12 km, as in other segments of the Hercynian Belt in Central Europe.

Two compressive periods (Early Cretaceous and Oligocene–Early Miocene) caused folding, fracturing and tectonic inversion in the Iberian Basin, along the NW–SE fault system; this is the origin of the present-day Iberian Ranges. These late Hercynian lineaments were important in the geologic history of the Ranges as was described earlier by Sopeña et al. (1988).

3. Sedimentology

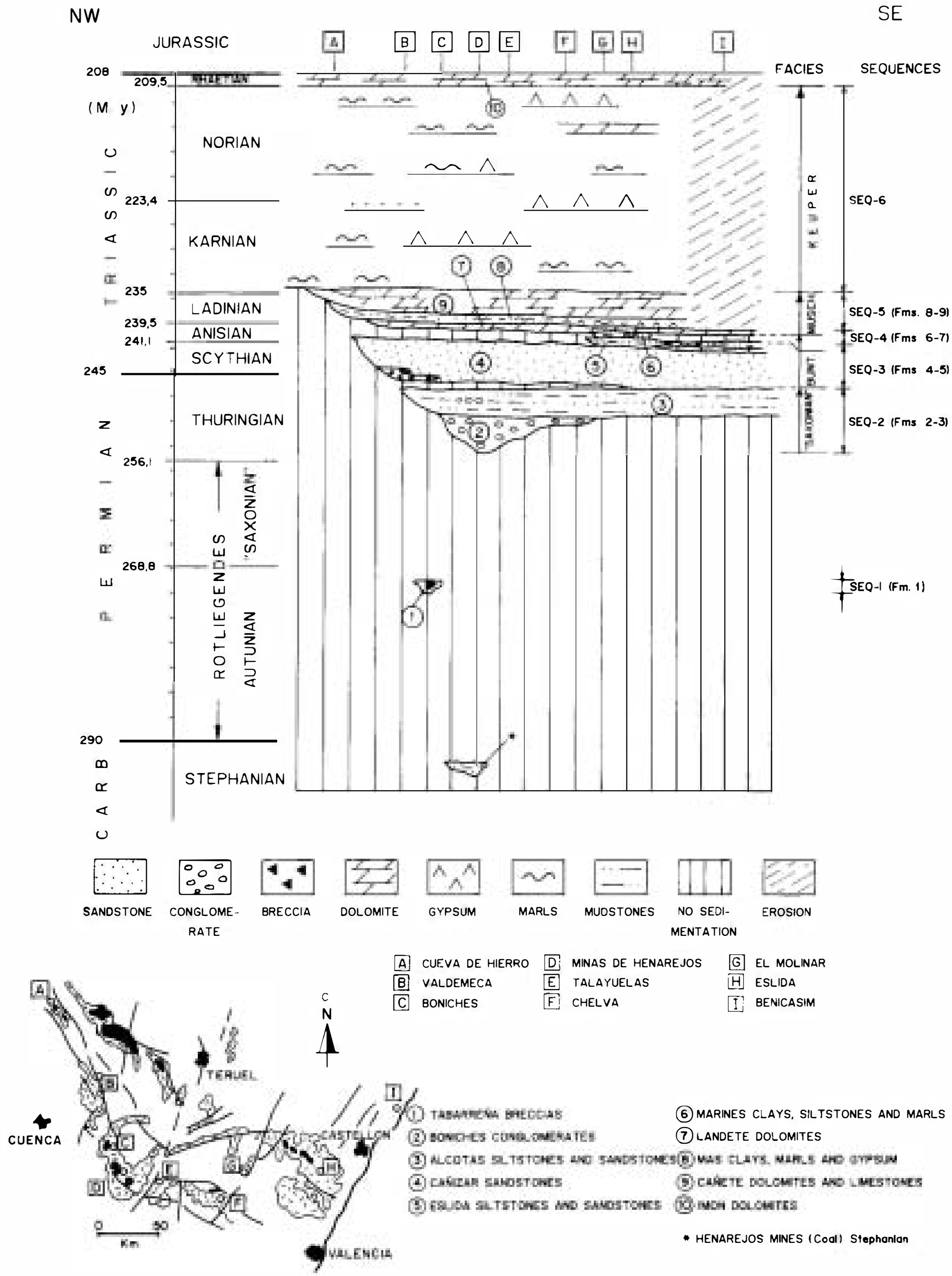
As we deal in this paper with the early stages of evolution of the Iberian Basin, only the sediments of Early Permian to Late Triassic age will be briefly described here. Detailed sedimentological and stratigraphic studies of these rocks can be found, among others, in: Ramos (1979); Sopeña (1979); Ramos et al. (1985); López (1985); Arche and López (1989) and López-Gómez and Arche (1992a,b, 1993).

The Permian–Triassic sedimentary infilling of the basin has been subdivided in this paper into six sequences bounded by unconformities or the equivalent paraconformities, so they are alloformations (Fig. 2). From base to top they are: Sequence 1 (Autunian, Fm. 1), Sequence 2 (Thüringian, Fms. 2–3), Sequence 3 (Thüringian–early Anisian, Fms. 4–5), Sequence 4 (early-middle Anisian, Fms. 6–7), Sequence 5 (late Anisian, Fms 8–9) and Sequence 6 (Karnian–Norian). The sequences show marked vertical and lateral changes of facies and thickness (Fig. 3) and all of them have been dated by means of pollen and spores, Foraminifera and Ammonites of biostratigraphic relevance (Boulouard and Viallard, 1971, 1982; Visscher et al., 1982; Doubinger et al., 1990; Márquez et al., 1994). The ages in Fig. 2 are those of Harland et al. (1990). The main characteristics of the sequences are as follows.

Sequence 1 (Fig. 3A). Found only in the northwest and central part of the Iberian Basin. Continental deposits in small, isolated semigrabens with internal drainage, but up to 2000 m of sedimentary and volcanic rocks accumulated in some of them. They consist of a lower interval with volcanic and volcanoclastic rocks and an upper interval of coarsening-upwards sequences of red siltstones, sandstones and conglomerates of lacustrine and alluvial fan origin. In the southeast, the volcanoclastic rocks are overlain by lacustrine siltstones and dolomites (Ramos, 1979; Pérez-Arlucea and Sopeña, 1985) and only scree deposits are present in the Boniches area (Tabarreña Fm.) (López-Gómez and Arche, 1994). The age of this sequence is Early Permian (Autunian?).

Sequence 2 (Fig. 3B, C). It consists of two formations, the Boniches Conglomerates Fm. and the Alcotas Siltstones and Sandstones Fm. They can be correlated with the 'Saxonian' Facies described by Ramos (1979) in the northwest domain of the Iberian Ranges. Its age is Thüringian (Late Permian) (López-Gómez et al., 1985) and was deposited in about 8–10 m.y.

The Boniches Fm. consists of well-rounded to subangular quartzite pebbles, up to 40 cm, organised in fining-upwards cycles and up to 85 m thick (Fig. 3) (Arche and López, 1989). Paleocurrent towards the northeast in the lower part, with a radial dispersion of



NW
SE
JURASSIC
A B C D E F G H I
FACIES
SEQUENCES
208
209,5
RHAETIAN
(My)
TRIASSIC
NORIAN
223,4
KARNIAN
235
LADINIAN
239,5
ANISIAN
241,1
SCYTHIAN
245
PERMIAN
THURINGIAN
256,1
ROTLIEGENDES
"SAXONIAN"
268,8
AUTUNIAN
290
CARB
STEPHANIAN
KEUPER
MUSCH.
BUNT
SEQ-6
SEQ-5 (Fms. 8-9)
SEQ-4 (Fms 6-7)
SEQ-3 (Fms 4-5)
SEQ-2 (Fms 2-3)
SEQ-1 (Fm. 1)
SANDSTONE
CONGLOMERATE
BRECCIA
DOLOMITE
GYPSUM
MARLS
MUDSTONES
NO SEDIMENTATION
EROSION
A CUEVA DE HIERRO
B VALDEMECA
C BONICHES
D MINAS DE HENAREJOS
E TALAYUELAS
F CHELVA
G EL MOLINAR
H ESLIDA
I BENICASIM
N
TERUEL
CUENCA
CASTELLON
VALENCIA
0 50
Km
1 TABARREÑA BRECCIAS
2 BONICHES CONGLOMERATES
3 ALCOTAS SILTSTONES AND SANDSTONES
4 CAÑIZAR SANDSTONES
5 ESLIDA SILTSTONES AND SANDSTONES
6 MARINES CLAYS, SILTSTONES AND MARLS
7 LANDETE DOLOMITES
8 MAS CLAYS, MARLS AND GYPSUM
9 CAÑETE DOLOMITES AND LIMESTONES
10 IMON DOLOMITES
* HENAREJOS MINES (Coal) Stephanian

40° and to the southeast in the upper part. They are interpreted as alluvial fan deposits coming from the Paleozoic highlands on the southwest margin of the Iberian Basin. Their marked cyclicity is rather simple (fining-upwards, thinning-upwards cycles) and could represent reactivations of the depositional systems by backfaulting in the basin margin and progradation and entrenching of the individual fan lobes. A climatic control would cause much more complex cycles and changes from water-laid sediments to mass-laid deposits not observed in this area.

The Alcotas Fm. consists of red siltstones and associated lenticular sandstone and conglomerate bodies, up to 170 m thick. Paleocurrents towards the south and southeast (Fig. 3). It is interpreted as floodplains of low sinuosity, ephemeral rivers with some temporary lakes in an interior drainage basin.

Sequence 3 (Fig. 3D, E). It consists of two formations, the Cañizar Sandstones Fm. and the Eslida Siltstones and Sandstones Fm. Its age is late Thüringian (Permian)–early (?) Anisian (Triassic) and the sequence was deposited in about 15–16 m.y.

The Cañizar Fm. is a characteristic pink to red sandstone interval found all over the Iberian Ranges. It consists of arkosic sandstones cemented by illite, K-feldspar and quartz up to 170 m thick with minor conglomerate and siltstone levels. It is interpreted as sandy braided river deposits, with paleocurrents to the southeast (López-Gómez and Arche, 1993).

The Eslida Fm. is found only along a narrow zone of the southeast part of the Iberian Ranges Basin (Fig. 3). It consists of red siltstones and intercalated, decametric, lenticular sandstone bodies of arkosic composition. It is up to 660 m thick, but decreases rapidly to the northwest and southwest. The sandstone bodies are channelised, of the simple and composite types of Friend (1983), and are interpreted as sandy braided river deposits of flashy discharge, alternating with floodplain deposits with ephemeral lakes and soil profiles (López-Gómez and Arche, 1993).

Sequence 4 (Fig. 3F, G). It marks the beginning of the marine sedimentation in the Iberian Basin, and it is the equivalent to the Röt Facies and part of the Muschelkalk Facies of the German Basin.

It consists of two formations, the Marines Clays, Siltstones and Marls Fm. and the Landete Dolomites Fm.; its age is early-middle Anisian (Triassic) and the sequence was deposited in about 1 m.y. (Figs. 2 and 3).

The Marines Fm. consists of red siltstones and green and yellow marls, up to 100 m thick. The presence of linsen bedding, oscillation ripples and halite pseudomorphs and the upwards transition to marine carbonates could indicate an estuarine or transitional environment of sedimentation (López-Gómez and Arche, 1992b).

The Landete Fm. consists of laminated dolomites and some marls and evaporites up to 145 m thick. Stromatolites, bioclastic bars and tepee structures are present and Bivalves, Foraminifera, Echinoids and Conodonts have been found. It is interpreted as the evolution of a shallow marine carbonatic ramp.

Sequence 5 (Fig. 3H). It represents the period of drowning of the Paleozoic basin margins and the interconnections of the Ebro, Iberian and Cuenca–Mancha Basins. It consists of two formations, the Mas Clays, Marls and Gypsum Fm. and the Cañete Dolomites Fm. Its age is late Anisian–Ladinian, and the sequence was deposited in about 5.5–7 m.y. (Figs. 2 and 3)

The Mas Fm. presents evaporitic and variegated marls facies, very similar to the classical Keuper Facies, but of older age. Observed thickness is up to 80 m, but substantial dissolution of evaporites has taken place on outcrops and a much larger original thickness is probable. It is interpreted as a lateral evolution from estuarine facies in the northwest to coastal saline mud-flats and shallow carbonatic intertidal complexes to the southeast.

The Cañete Fm. is very similar to the Landete Fm.

Fig. 2. Permian and Triassic sediments of the southeast Iberian Ranges. Numbers on the right indicate major sedimentary sequences: *1* = Autunian (?) Sequence (Tabarreña Breccias); *2* = Thüringian Sequence (Boniches Conglomerates and Alcotas Siltstones and Sandstones Fms.); *3* = Early Triassic Sequence (Cañizar Sandstones and Eslida Siltstones and Sandstones Fms.) (Buntsandstein Facies); *4* = Middle Triassic Sequence (Marines Clays, Siltstones and Marls and Landete Dolomites Fms.); *5* = Late Triassic Sequence (Mas Clays, Marls and Gypsum and Cañete Dolomites and Limestones Fms.) (Muschelkalk Facies); *6* = Latest Triassic–Early Jurassic Sequence (Keuper Facies and Imón Dolomites Fm.). Sequences are bounded by unconformities and/or hiatuses and names correspond to formations.

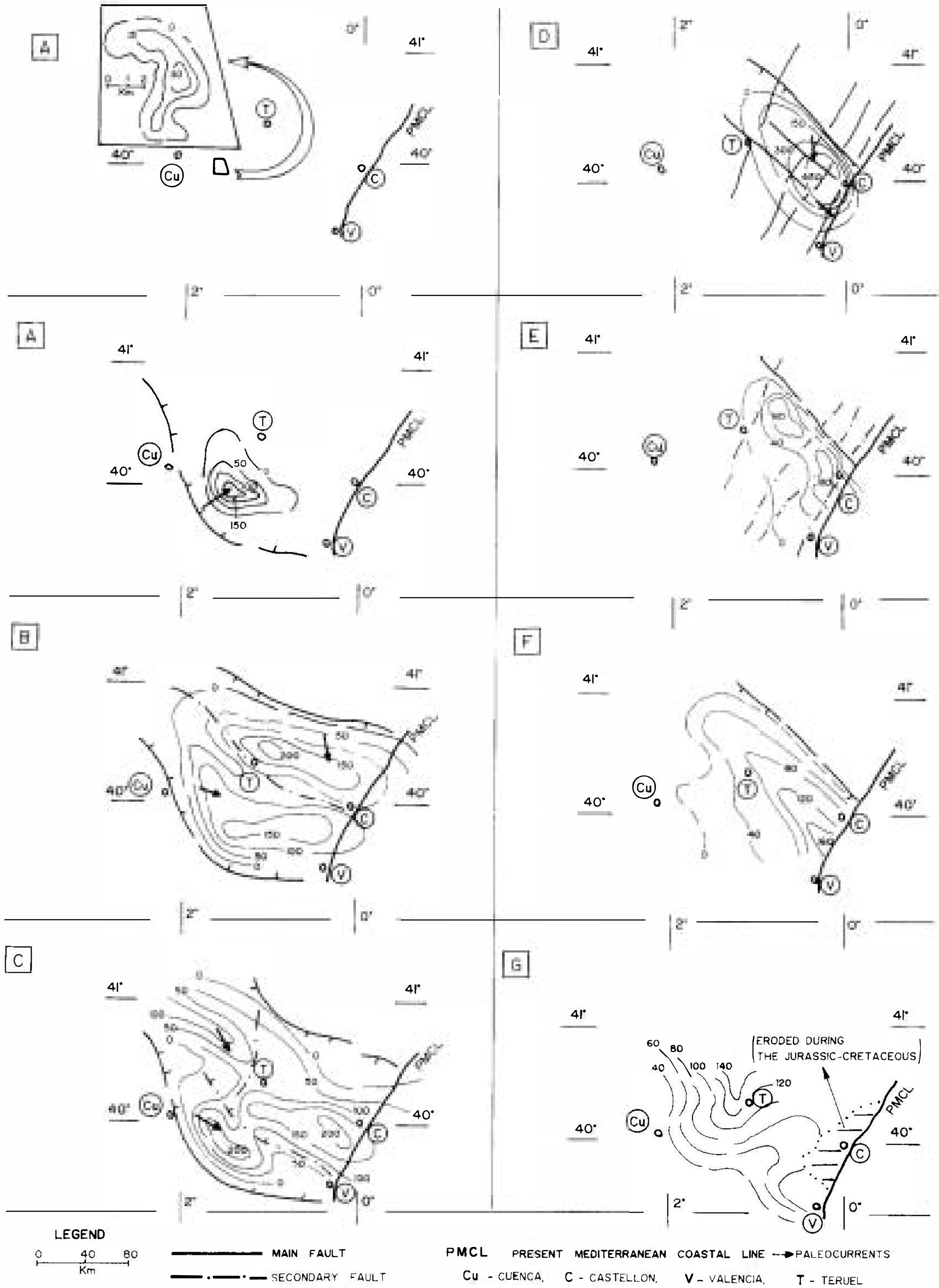


Fig. 3. Isopach maps of selected Permian and Triassic formations in the southeast Iberian Ranges. Arrows indicate mean palaeocurrent directions in the siliciclastic sediments. See Fig. 5 for the names of the main faults.

as well as its facies and environmental interpretation. It is up to 95 m thick and contains rich marine faunas (Bivalves, Ammonites and Conodonts).

Sequence 6 (Fig. 2). Not studied in detail in this paper. It consists of the Keuper Facies and the Imón Dolomites Fm. Its age is Karnian–Norian, and was deposited in 20–21 m.y. (Fig. 2).

The Keuper Facies is a complex interval composed of five formations in the southeast Iberian Ranges (Ortí, 1974), representing two transgressive–regressive cycles, with clays, marls and evaporites. They are interpreted as coastal sebkhas and shallow marine hypersaline inner shelf facies. Their original thickness is very difficult to estimate due to substantial dissolution of evaporitic intervals but several hundred metres seem plausible.

The Imón Fm. consists of laminated and cross-bedded dolomites and represents an open inner shelf. It is up to 20 m thick and of a Norian (Late Triassic) age.

4. Basin boundary faults

The main alpine structures of the Iberian Ranges trend NW–SE and NNE–SSW (Fig. 4) and transversal sections show a fan-like divergent structure (Guimerà and Alvaro, 1990), the result of a thin-skinned compressive deformation processes during the Oligocene and the Early Miocene. The alpine structures are parallel to the grain of the Paleozoic basement and some of these lineaments of Hercynian or older age played a crucial role in the development of the Iberian Basin and the subsequent inversion tectonics during the Cenozoic compressive periods. This is a fundamental fact well illustrated, for example, in the East African Rift System (Dunbar and Sawyer, 1988; Versfeld and Rosendahl, 1989; Morley et al., 1992).

Some of these fault systems have been identified as the Iberian Basin boundary faults since the early tectono-sedimentary synthesis and interpretation of Alvaro et al. (1979), like the North-Iberian and Hesperian Faults of these authors; they are considered as rectilinear, almost vertical, deep-seated faults more than 300 km long (Fig. 5).

Field mapping reveals a more complex picture for these fault systems. If the tectonic inversion that took place along them is balanced, there are three main normal fault systems that exert a control on the shape and location of the Permian–Triassic infilling of the Iberian Basin (Fig. 5): the Serrania de Cuenca Fault, the Molina–Teruel–Espadán Fault and the Ateca–Montalban–Maestrazgo Fault, all trending NW–SE, and the subordinate Teruel and Requena–Castellón Faults, trending NNE–SSW.

The NE–SW fault systems consists of laterally linked, arcuate sequences, 50–70 km long each, offset by transversal faults trending at high angles. If isopach and paleocurrent data are superimposed (Figs. 3 and 5), it is evident that they had a close control on the geometry of the basin, and the position of depocentres; in other words, from the Early Permian to the Early Triassic they were the active basin boundary faults. Similar tectonic control on the geometry and infilling of contemporaneous basins such as the Catalan and Pyrenean ones has been demonstrated by Marzo and Calvet (1985) and Gisbert et al. (1985). The combination of palinspastic reconstructions, isopach maps, paleocurrent maps and proximal–distal facies relationships demonstrate that the Iberian Basin was formed during the Early Permian as a series of isolated half-grabens of asymmetric infilling, evolving during the Late Permian into wider half-grabens and into a more or less symmetric graben in the Early Triassic; the Basin is coeval with a series of contemporaneous basins separated by Paleozoic highs. A marine transgression onlaps on these basins, overstepping every previous formation and drowning most of the highs during the Late Triassic; the Iberian Basin showed a well established 'steer's head' transversal profile at the end of the Triassic (Fig. 6).

5. The origin and tectonic evolution of the Iberian Basin — possible mechanisms

The Iberian Microplate was bounded since early times by two major dextral strike-slip fault systems: the Chedabucto–Gibraltar System to the south and the Bay of Biscay–Pyrenees system to the north, and an incipient rift along the eastern margin, trending roughly NE–SW, related to the Cevennes Fault System and the southward-propagating North Atlantic and Rhine Rift Systems.

The Permian–Triassic basins of Iberia trend E–W (Pyrenean–Asturian) or NE–SW (Catalonian,

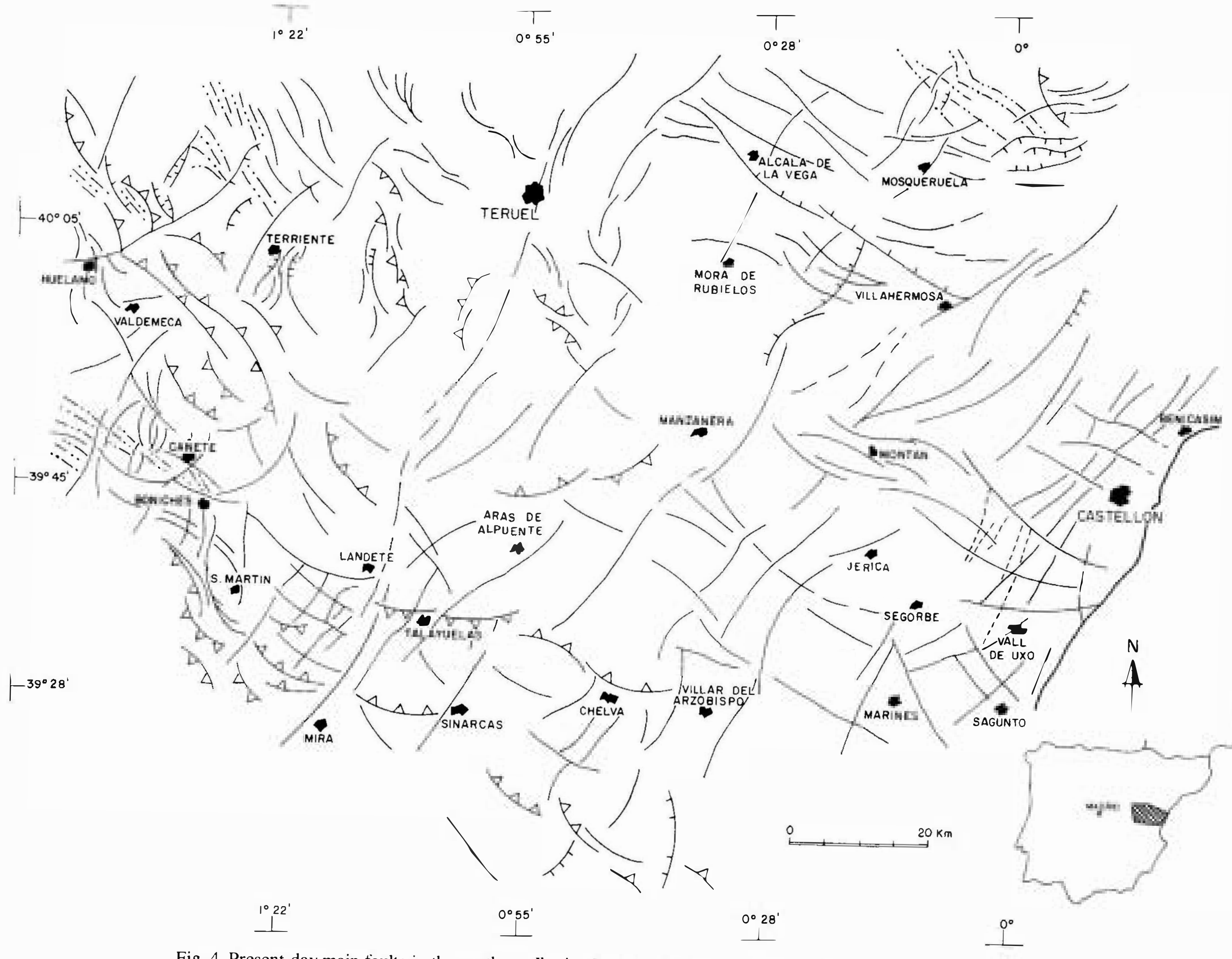


Fig. 4. Present-day main faults in the southeast Iberian Ranges with two main dominant directions: NW–SE and NNE–SSW.

Fig. 5. The three main NW–SE thrust systems and their reconstruction as the Iberian Basin's main basin boundary faults. (A) Present-day configuration. (B) Reconstruction of the basin boundary faults.

Valencia–Prebetic), but the Iberian Basin in unique in its NW–SE orientation, parallel to the fundamental Hercynian structures of the northeast Iberian Microplate.

Several hypotheses have been proposed to explain the origin and evolution of the Iberian Basin (Alvaro et al., 1979; Sopeña et al., 1988). All of them suppose that the Iberian Basin was the failed arm of a triple junction around a hot spot situated in the Valencia area, propagating from southeast to northwest; the other two arms were the Catalan and Valencia–Prebetic Basins. A McKenzie type of pure-shear graben, limited by linear, vertical, deep seated faults resulted from these hypotheses. However, there are three main problems not solved by them.

(1) Why are there no Early Permian volcanic rocks in the alleged hot spot area? They are found only 200 km to the northwest in the Iberian Basin and are unknown in the Catalan and Valencia–Prebetic Basins.

(2) Why are the basin boundary faults arcuate segments, not linear structures, and why were the syntectonic Early Permian–Early Triassic sediments deposited in asymmetric half-grabens whose depocentres migrated to the northeast in time?

(3) Why was extension taking place simultaneously in basins at right angles (NW–SE and NE–SW) (Iberian, Catalan, Valencia–Prebetic) and E–W (Pyrenees) during the Late Permian and Triassic? The large area affected by extensional tectonics during the timespan considered (more than 900 km) imply a crustal-scale-driving mechanism for its origin. Which is it?

We will try to propose a suitable alternative hypothesis to overcome these difficulties.

6. The geometry of the Iberian Basin boundary faults

Extensional models based on pure shear deformation, as first proposed by McKenzie (1978) predict vertical, deep-seated basin boundary faults, symmetric development of the basin, an extension factor quite considerable (β = 1.6 to 2.4) and voluminous early volcanism because the faults reach the asthenosphere.

In contrast, extensional models based on simple shear deformation (Wernicke, 1981; Wernicke and Burchfield, 1982; Lister et al., 1986) predict a wide variation of configurations with listric faults as basin boundary faults, sequential development of antithetic and synthetic faults in the hanging wall migrating away from the original boundary fault, highly asymmetric half-grabens, moderate extension factor (β up to 1.5) and very small amounts of volcanics because the faults very seldom reach the asthenosphere.

Field evidence and paleogeographic reconstructions favour an extensional process based on simple shear, brittle deformation of a basement under tensile stress. However, simple shear may be distributed over the lithosphere in such a way that the overall picture is one of pure shear (Reston, 1990), and surface data are not totally conclusive about the type of shear stress. High subcrustal geothermal gradients prevailed during the early phases of extension, requiring a multilayered stretching model (Royden and Keen, 1980) instead of a McKenzie model. The trace of the faults is arcuate (Fig. 5), so they must be listric, becoming horizontal in depth. As Gibbs (1984) has demonstrated, listric faulting creates highly asymmetric half-grabens because folding (roll-over) on the hanging wall is a consequence of this type of faults. The need to thin the roll-over

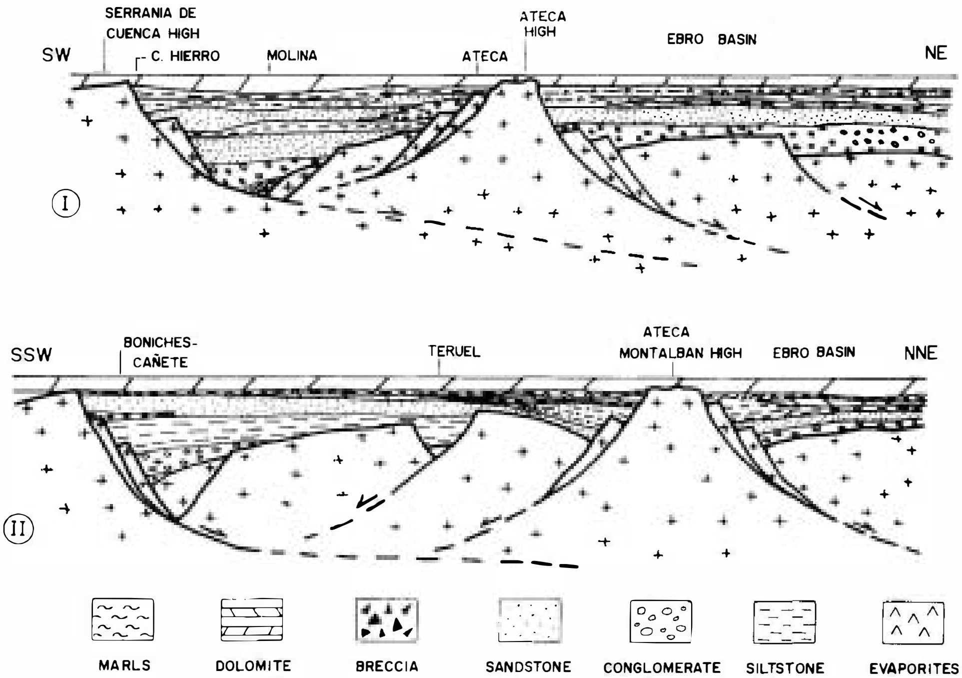


Fig. 6. Hypothetical reconstruction of the Iberian Basin and the southwest Ebro Basin after the sedimentation of the Cañete Dolomites and Limestones Fm. (Middle Triassic, Ladinian). The Central High (Ateca–Montalbán High) and the Serrania de Cuenca High are the basin boundaries. Note the asymmetrical infilling of the Basin. See Fig. 5 for locations.

leads to the formation of antithetic or counter faults in a fan migrating towards the hanging wall and, with further extension, to a crustal collapse leading to the formation of secondary grabens and central highs bounded by antithetic and synthetic faults.

The isopach maps (Fig. 3) show clearly the asymmetric geometry of the Iberian Basin during the Permian and the Early Triassic and the close relationship between fault traces and geometry of the sedimentary infilling. The Serrania de Cuenca Fault System is believed to be the initial basin boundary fault, dipping to the northeast, as the Thüringian sediments of the second macrosequence come from a Paleozoic high to the southwest (the relief of the footwall rift flank). The sediments onlap largely the Paleozoic basement to the northeast, decreasing in thickness and antithetic faults in the hanging wall can be mapped extending as an antithetic counter fan (the Molina–Teruel Fault System), extending towards the northeast with time, away from the initial master fault.

The Serrania de Cuenca Fault developed with time a listric fan of faults cutting back towards the uplifted footwall as extension proceeded, as demonstrated by the fining-upwards nature of the correlative alluvial fan sequences and the more distal character of each successive sequence (Fig. 7), i.e., backfaulting and retrogradation of the apex of each fan system.

During the Anisian, further extension led to the formation of a 'Central High' (sensu Gibbs, 1984), the Ateca–Montalban High, limited by synthetic and antithetic faults (the Ateca–Montalban–Maestrazgo System). The Serrania de Cuenca Fault System became inactive and sedimentation shifted to the northeast margin of the Iberian Basin and the newly created Ebro Basin (Fig. 6) to the northeast during the Anisian. The synthetic fault system on the southwest margin of the Ebro Basin (the Northern Ateca–Montalban Fault) was linked with the sole fault and was energetically favoured, leading to active subsidence and the first marine transgression in this area.

These arguments favour an interpretation of the early stages of rifting in the Iberian Basin as a shallow structure developed probably in the brittle

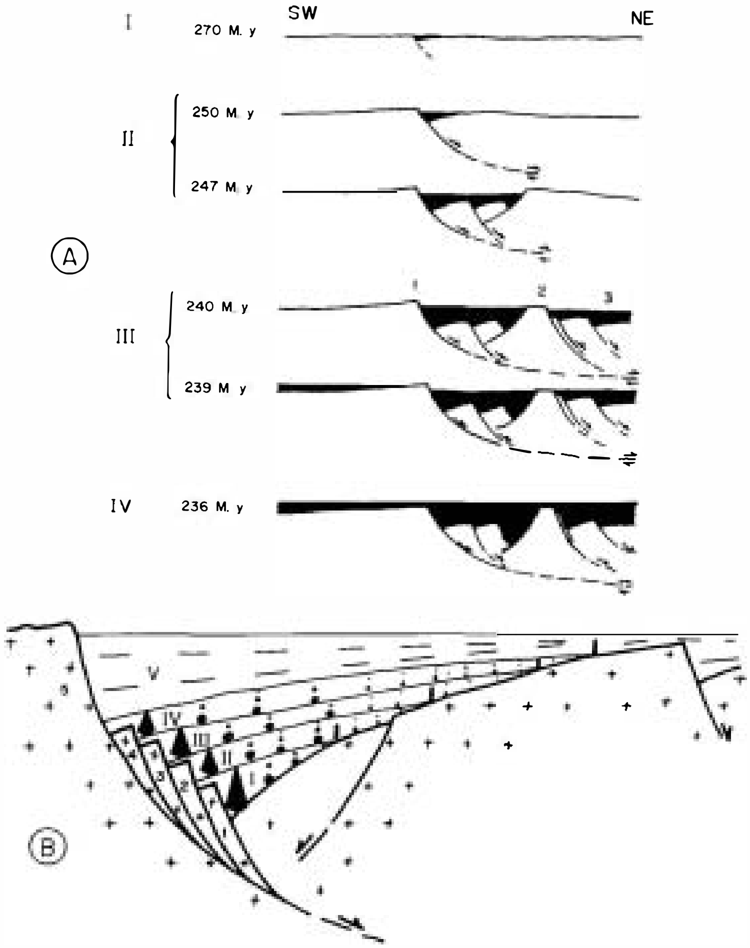


Fig. 7. (A) Schematic cross-sections of the Iberian, Ebro and Cuenca–Mancha Basins from the Early Permian to the Late Triassic. *I* = Autunian: very small basins along the initial break-up lineaments. *II* = Thuringian: development of the Iberian Basin in the hanging wall block of a listric fault as a supradetachment basin. *III* = Early Triassic: development of the Iberian Basin and creation of the Ebro Basin after further extension as an intra hanging wall basin; *IV* = Late Triassic: drowning of the footwall shoulder and creation of the Cuenca–Mancha Basin as a footwall subsiding basin; *1* = Serranía de Cuenca High; *2* = Ateca High; *3* = Ebro Basin. (B) Hypothetical evolution of the alluvial fan systems of the Boniches Fm. in stage II of Fig. 7A and Alcotas Fm. *I–V* = Depositional sequences. *1–5* = Stages of formation of the listric fan fault system. Notice the relationship between fault blocks and depositional sequences and the onlap disposition of the fan complexes against the footwall block and the reverse to normal grain size distribution in each fan sequence. Scale with vertical exaggeration.

part of the crust by tensile stress resulting in listric faults dipping to the northeast (and antithetic, subsidiary faults dipping to the southwest) and strain being accommodated by roll-over and extension of the hanging wall. This leads to a migration of the strain away from the initial basin boundary fault that becomes inactive with time (probably during the Early Triassic).

Fig. 6 shows two reconstructions of the Basin at the end of the Ladinian. Profile I, from the Cueva de Hierro High to the Ateca High and the Ebro Basin shows a fairly simple asymmetric graben with the main source of sediments in the southwest during the Late Permian (Ramos, 1979) and a shift to the northeast during the Early Triassic (García-Royo and Arche, 1987); the Ateca High was a more or less symmetric central high bounded by antithetic fault fans cutting down to the sole fault. The possibility of ramps and flats in this structure is high, but only seismic profiling could confirm or deny it.

Profile II, from Minas de Henarejos to the Ebro Basin, shows a more complex structure, with a secondary graben in Teruel due to crustal collapse of the roll-over and illustrates clearly the abandonment of the original basin boundary fault during the Anisian and the shifting of subsidence and sedimentation to the northeast margin of the basin as strain progressed in this direction. These differences are due to different extension rates in the northwest Iberian Basin and the southeast Iberian Basin. Alvaro (1987) obtained values of $\beta = 1.12$ for the first sector and $\beta = 1.26$ for the second one, and our data will be discussed later on.

7. The NNE–SSW fault system

The main basin boundary faults were offset by a series of linear, not arcuated faults. These were contemporaneous and synsedimentary as a along-strike reconstruction of the basin (Fig. 8) proves.

They are interpreted as transfer faults (Gibbs, 1984, 1990) linking the segments of the main NW–SE faults and accommodating different rates of extension in each compartment. They have a strike-slip movement and trend at about 60° of the main fault system; this implies that extension was oblique, not perpendicular to the NW–SE faults. Compressive (transpressional) structures may develop as oblique, secondary highs in the basin; the Cueva de Hierro–Tramacastilla High (see Fig. 8 for changes of thickness in sections G and H) could be one of these; extensional (transtensive) structures along transfer faults could explain the sudden increase in thickness of the sediments in the Teruel and Castellón areas (see Fig. 8). The main transfer faults were the Teruel, Espadán and Castellón Faults (see Figs. 4 and 5).

8. Subsidence history of the Iberian Basin: some constraints derived of the backstripping method

The subsidence history of the Iberian Basin can be studied in more detail and in some way quantified using the backstripping method. It allows an evaluation of the total subsidence of the basin through time and separates its tectonic and sediment load component.

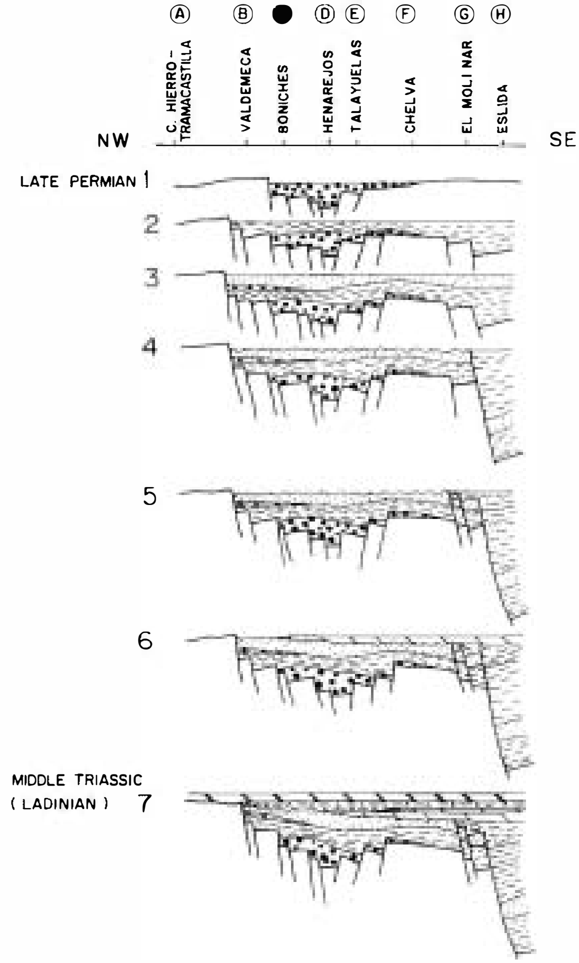


Fig. 8. Longitudinal reconstructions of the southeast Iberian Basin from the Late Permian (Thuringian) (*1*) to the Middle Triassic (Ladinian) (*7*). Synsedimentary faults belong to the NNE–SSW transfer fault system, and they ceased to be active during the Early Triassic (Anisian), except in the southeast extremity of the Basin. See *A–H* locations in Fig. 2.

Only a few attempts have been made to apply the backstripping method in the Iberian Ranges (Alvaro, 1987; Sánchez-Moya et al., 1992; Salas and Casas, 1993; Roca et al., 1994; van Wees, 1994) and in related areas (Ebro Basin, Brunet, 1984; Zoetemeijer et al., 1990; and Aquitaine BasinDesegaulx and Moretti, 1988).

The backstripping method (Watts and Ryan, 1976; Stekler and Watts, 1978) assumes local isostatic compensation (Airy isostasy) and the tectonic subsidence is considered the part of the total subsidence caused by the fundamental stretching lithospheric processes leading to the origin and evolution of the sedimentary basin. The method also assumes that during compaction only the porosity of the sediment diminishes (Perrier and Quibilier, 1974), that is, the volume of solid grains remains constant during the process. Loss of porosity is irreversible.

Compaction curves have been compiled for different types of sediments since the pioneer work of Athy (1930) and Hedberg (1936). In this paper, a combination of the curves proposed by Shinn and Robbin (1983), Bond and Kominz (1984) and Baldwin and Butler (1985) are used with the help of a modified 'Subside' program (Hsui, 1989).

In spite of the ample use of the method, some caution should be applied to the interpretation of the curves, as demonstrated by Gallagher (1989), concerning deviations of ± 100 m in the correct values due to error margins in stratigraphic ages and uncertainties in paleobathymetries or sea-level changes and Audet and McConnell (1994), due to the fact that porosity reduction depends not only on burial depth but also on time. The possible influence of intraplate stress is not considered in this paper (Cloetingh et al., 1985; Kooi and Cloetingh, 1989; Cloetingh and Kooi, 1992).

Six subsidence curves are presented here: four based on original geological profiles of the Iberian Ranges and two on commercial oil wells in the Ebro Basin (Mirambell-1) and the Cuenca–Mancha Basin (Belmontejo). Only the Permian–Triassic–Early Jurassic interval of the curves will be analysed in this paper. The curves have comparable shapes and show an initial period of rapid subsidence (270–237 m.y.) and a subsequent period of much slower subsidence (237–208 m.y.), interpreted as the tectonic or rifting subsidence period and the flexural or thermal subsidence period, respectively (Fig. 9).

A more detailed examination of the curves shows that the tectonic subsidence period can be subdivided into four episodes of rifting followed by the beginning of a thermal subsidence period:

(1) *Early Permian* (270–260 m.y.), represented only in the Cañete profile. Red beds deposited in a small half-graben. The episode ended abruptly and was succeeded by uplift and erosion of the basin.

(2) *Late Permian* (256–247 m.y.): sedimentation of the 'Saxonian' Facies in a broad rift basin trending NW–SE, subdivided into several sub-basins or depocentres. This episode also ended in uplift and partial erosion and has been observed in the Cañete, Teruel, Chovar–Eslida and Villafamés sections.

(3) *Late Permian–Middle Triassic* (245–237 m.y.): sedimentation of the continental Buntsandstein Facies and the shallow marine Muschelkalk Facies in rift basins. Rapid transition to the following period.

(4) *Middle Triassic–Early Jurassic* (237–208 m.y.): sedimentation of the shallow marine upper Muschelkalk Facies, Keuper Facies and Imón Dolomite Fm. It represents the thermal subsidence period of this tectonic cycle. The younger cycles (Middle Jurassic to Quaternary) are not analysed in this paper, but are obvious in the subsidence curves presented in Fig. 9.

Tectonic subsidence rate for the period considered increases from the northwest (Cañete) to the southeast (Chovar–Eslida) and can be measured by means of the β factor (Fig. 9) 1.17 for Cañete, 1.22 for Teruel and 1.29 for Chovar–Eslida. The Villafamés profile is incomplete due to partial erosion of the Triassic–Early Jurassic sediments prior to the Oxfordian and β factor cannot be calculated in this case. A possible explanation of this increase will be presented in the next chapter. An instantaneous McKenzie stretching model was developed for the Late Permian–Triassic extension period to calculate the β factor, and a crustal thickness of 32 km was assumed. We are aware that this is only a first-order approximation due to our limitations in calculating the β factor because the method used can not take into account a multiphase rifting model.

The subsidence rates compare well with data published in other domains of the Iberian Ranges (Sánchez-Moya et al., 1992; Salas and Casas, 1993), although our values are slightly bigger and the polyphasic type of the rifting stage is obvious and older than previously suspected.

From the two subsidence curves calculated for commercial oil wells previously cited (Mirambell and Belmontejo), the first one shows a well developed tectonic subsidence episode comparable to the youngest one observed in the Iberian Basin (Late Permian–Early Triassic); older sediments were not deposited in this area close to the northeast margin of the Iberian Basin. The flexural episode is also

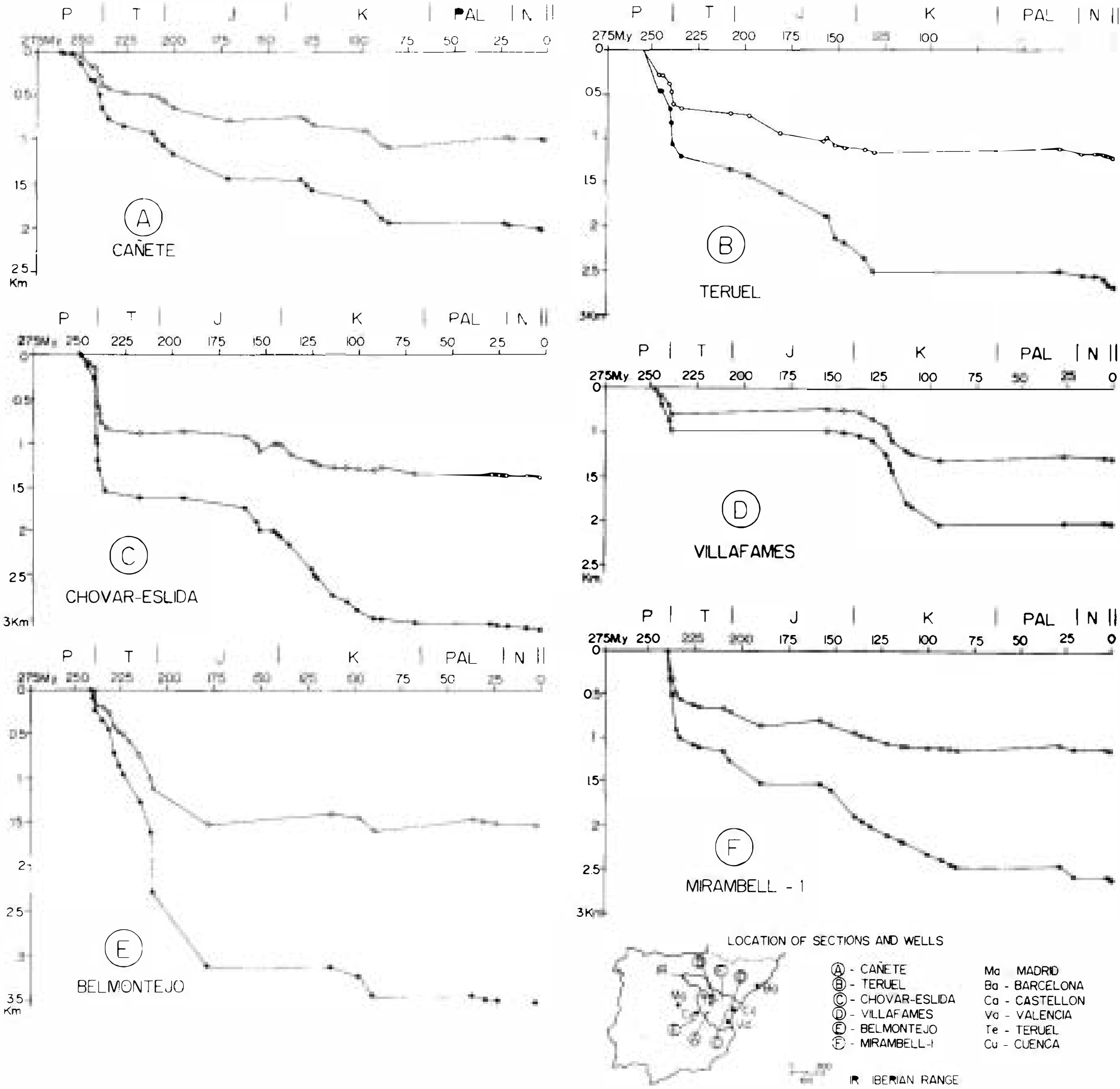


Fig. 9. Subsidence curves for six sections and wells in the Iberian Ranges, Ebro and Cuenca–Mancha Basins. (A) Cañete, (B) Teruel, (C) Chovar–Eslida, (D) Villafamés, (E) Belmontejo, and (F) Mirambell-1. Circles = tectonic subsidence. Dots = total subsidence.

comparable and a β factor of 1.22 for the Triassic–Early Jurassic subsidence period is obtained.

The Belmontejo well shows a very rapid, polyphasic initial subsidence from the Early Triassic to the Late Triassic, specially during the sedimentation of the Keuper Facies. This fact cannot be explained solely by the completeness of the evaporite record in subsurface in comparison with the outcrops observed in the Iberian Ranges. The anomaly could be explained if the Anisian–Karnian thermal event found in the Central System by Galindo et al. (1994) took also place in the basement of the Cuenca Basin and fast thermal cooling took place at the end of the Triassic in this area, but this hypothesis remains to be proven by further geophysical studies of the basement of this area.

9. Discussion of the possible extensional processes during the Permian–Triassic period in the Iberian Basin

The first expression of the Iberian Basin was a series of isolated basins a few kilometres long of Autunian age in the area between Noviales and Boniches (Fig. 1). All of them are coeval with the latest granitic intrusions (Ramos, 1979; Sopeña, 1979; Hernando et al., 1980; Muñoz et al., 1985). No rocks of Autunian age have been recognized to the northwest or the southeast of this area.

These pockets of extension developed and linked in two larger continental basins during the late Permian (Thüringian), propagating towards the northwest and the southeast and finally, in the Early Triassic linked to form a laterally thoroughgoing continental rift system (Fig. 10). These phases of evolution and its timing cannot be explained by the hot spot-aulacogen hypothesis of Alvaro et al. (1979), but certainly extension was limited and repeated extension-tectonic inversion events took place in approximately the same area, a typical feature of the aulacogenic basins.

Bonnatti (1985) studied the onset of rifting and sea-floor spreading in the Red Sea and observed that rifting is preceded by the emplacement of regularly spaced hot mantle plumes, propagating to the northwest in time, along strike of the future rift system. A similar process was observed in a continental domain as the East African Rift System (Ebinger et al., 1984; Bosworth, 1985; Ebinger, 1989).

These rift segments or pockets of extension closely resemble the Autunian isolated half-grabens; they indicate that tensional strain varied along the rift system and started at the weakest points. The cause–effect relation between initial rifting and andesitic volcanism remains to be explored, but certainly existed. Did volcanic intrusion weakened certain spots initiating rifting or was the tensional stress concentrated in pockets and the ascent of magmas passively took place along tectonically weakened areas?

During the Late Permian strain propagated along strike both to the northwest and the southeast, creating two sub-basins separated by the Cueva de Hierro–Tramacastilla High (Figs. 8 and 10). This geometry reflects differential extension rates in the rift segments: high in the sub-basins and low in the high area.

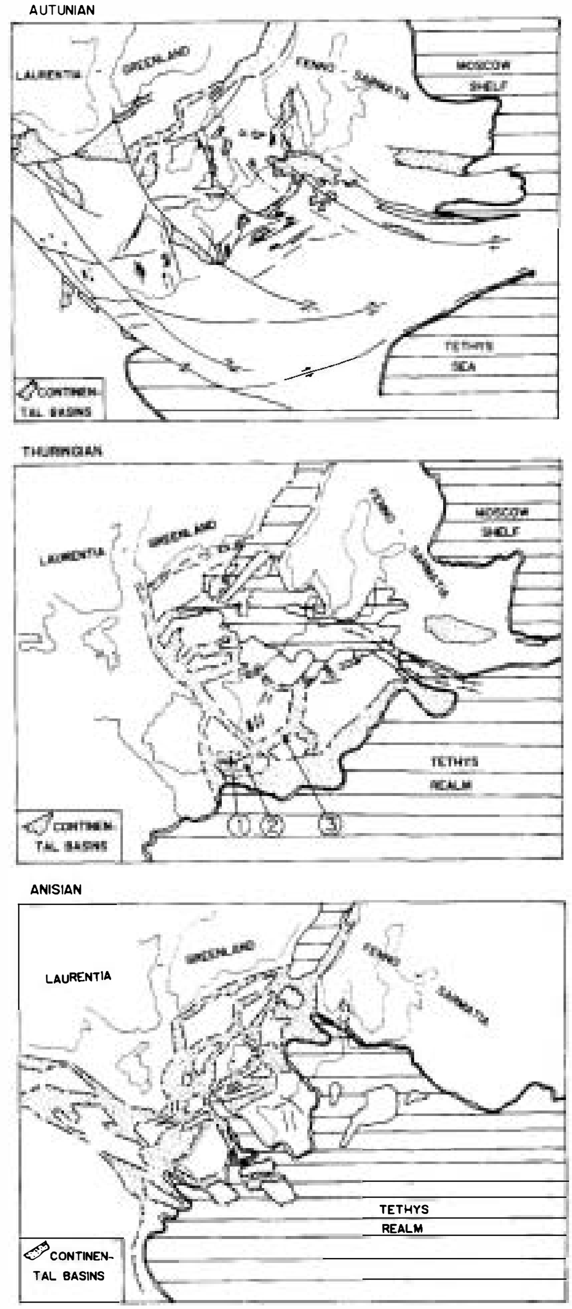


Fig. 10. Palaeogeographic reconstructions of western Europe for the Autunian, the Thuringian and the Anisian, based on Ziegler (1988) and modified by the authors. The Iberian Basin developed in response to tensional stress along the northern and southern margins of the Iberian Microplate, while the Catalan Basin and the Valencia–Prebetic Basins (*1*) bounded to the east by the Corsican–Sardinian–Balearic Block (2) are related to the southwards propagation of the Hesse–Burgundy Rift (*3*).

The rift segments were linked together during the Early Triassic (Fig. 10) to form a continuous structure, with transfer faults accommodating differences in extension rates.

The rates of rift propagation along strike are comparable to present-day examples in East Africa (Ebinger, 1989): the Kenya Rift propagated southwards 750 km in about 20 m.y., approximately the period between Early and Late Permian ($\approx$21–25 m.y.) and the resulting geometry was very similar: half-grabens infilled by lakes and alluvial fans separated by basement highs. The mature phase (Late Permian–Early Triassic) took about 20 m.y. to be completed and again time and geometry compare well with the present-day Zambezi Rift System, open to the Indian Ocean at the south end of the East African Rift System.

10. The orientation of the Iberian Basin and its relations with coeval basins

The NW–SE orientation of the Iberian Basin is an exception among coeval basins; some of them trend E–W (Asturian, Pyrenean) and others NE–SW (Catalan, Valencia–Prebetic). Some of them started to extend and subside during the Early Permian (Asturian, Pyrenees, Iberian and some small basins in the southern edge of the Iberian Massif), others during the Late Permian (Catalan) and the Early Triassic (Valencia–Prebetic). It is difficult to envisage a simple stress field causing coeval extension in at least three main orientations during the later period, but if we consider the fact that the eastern basins did not exist during the Early Permian, a two-stage development in response to different processes can explain some of these problems.

Since the works of Arthaud and Matte (1977), it is generally accepted that the Iberian Microplate was part of a dextral megashear zone between the Southern Appalachians and the Urals, just after the Hercynian Orogenesis (Stephanian–Autunian). Using a simple Riedel diagram (Fig. 11) of dextral strike-slip stress and strain, we observe that the Iberian Basin trends exactly as the expected tensional, normal faults, and the Asturian and Pyrenean Basins could be transtensional basins along the Bay of Biscay Fault. As the Iberian Microplate was moving westwards relative to Europe in this period and, possibly, rotating clockwise in response to the strike-slip stresses at its margins, extensions in the E–W and NW–SE directions were enhanced. If this hypothesis is correct, they can be considered a consequence or inheritance of the Hercynian Orogeny. It can also explain the orientation of the transfer fault system associated to the Basin Master Faults, because they were oriented parallel to the expected antithetic strike-slip faults in the Riedel diagram (Fig. 11).

According to this hypothesis, extension should be more important to the southeast; as this is the case (Fig. 9) illustrated by the subsidence curves, we accept this explanation even in the absence of detailed paleomagnetic studies demonstrating the independent evolution of the Iberian Microplate and the Ebro Block. However, this situation changed during the Tertiary, as demonstrated by De Ruig et al. (1991) and Geel et al. (1992).

The basins trending NE–SW along the eastern margin of the Iberian Massif are younger (no Autunian rocks), lack volcanics in their early stages, and the age of the first sediments lying on the Paleozoic rocks youngs to the southwest is Thüringian in the Catalan Basin, and Anisian–Ladinian in the Prebetic Basin (Pérez-López, 1991). Extension rates were moderate during the Mesozoic, not exceeding a β factor of 1.30 and no oceanic crust was created. In a general map of Western Europe for the Early Triassic (Fig. 10) it is obvious that they can be related to the southwards propagation of the Hessian–Burgundy Rift System along the eastern edge of the Massif Central, separating the Catalan Block from the Corsican–Sardinian–Balearic Block in Thüringian times and the southeast edge of the Iberian Massif from the Betic-Rift–Apulia Block in Anisian–Ladinian times. These basins were the consequence of an independent geologic phenomenon and related to the alpine basins to the northeast, not to the escape to the west of the Iberian Microplate (Fig. 10).

We conclude that the orientation and coeval evolution during the Late Permian–Triassic of several rift basins in the Iberian Microplate were the consequence of two different stress fields, one related to the strike-slip faults related to the posthumous Hercynian movements (Autunian) and the other to essentially extensional stress propagating towards the southwest about 2300 km, from the Boreal Basin to the Prebetic Basin (Thüringian–Triassic).

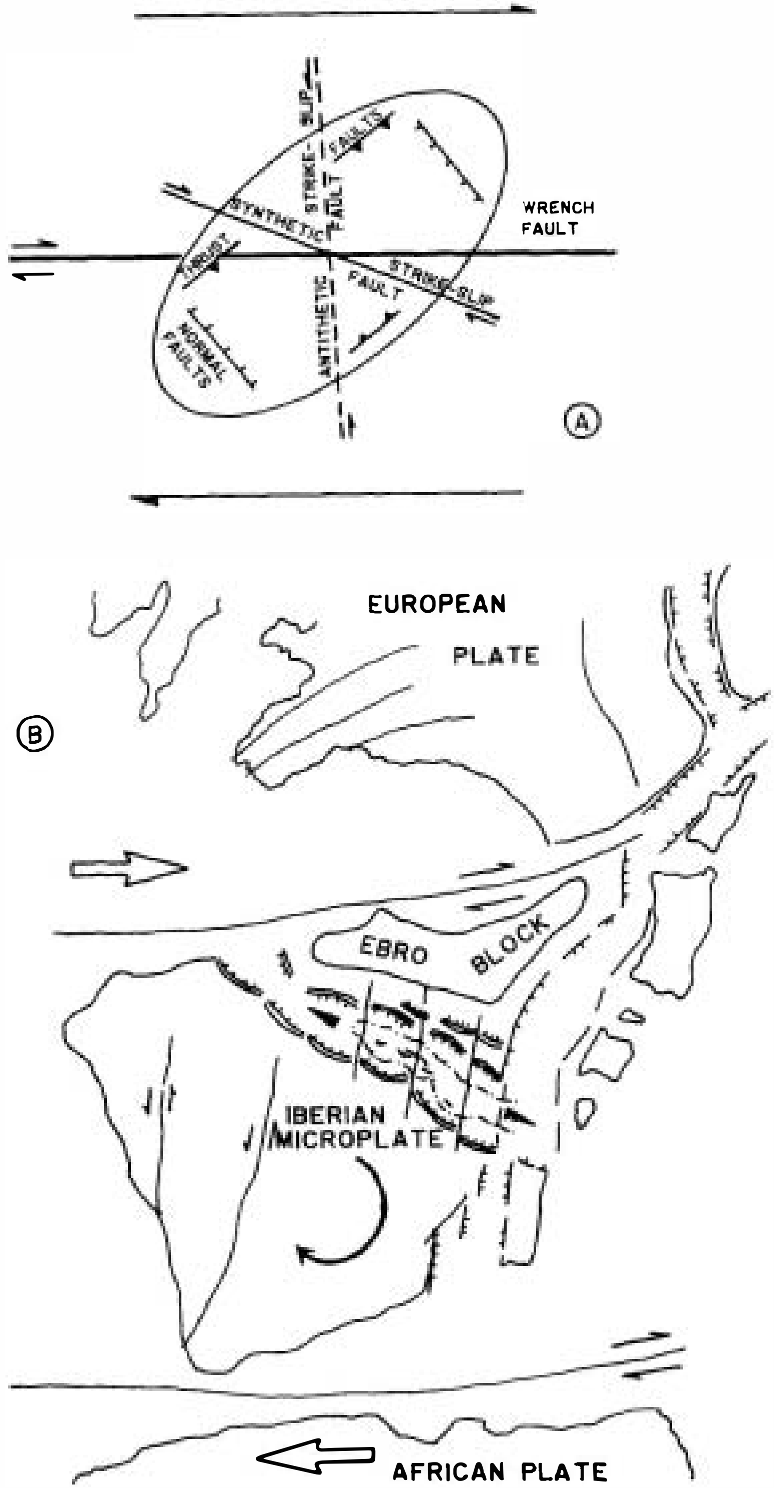


Fig. 11. (A) Riedel diagram summarizing the main deformations in a dextral wrench fault system. (B) Cartoon showing the development of the Iberian Basin from initial, isolated pockets of deformation (Autunian basins) prograding to the northwest and southeast in time. The Catalan and Valencia–Prebetic Basins trend almost perpendicular and cannot be the consequence of the same stress field. An arrow suggest the possible clockwise rotation of the Iberian Microplate during the Late Permian–Early Triassic.

11. The driving mechanism of the extensional processes

A crustal-scale mechanism should be invoked to explain simultaneous extension in a continental crust area more than 900 km wide (only in the Iberian Microplate).

The Iberian Microplate was a zone of continent–continent collision during the Hercynian orogeny resulting in crustal shortening and thickening, may be up to 60 km. S-type granites were intruded at the end of the process, between 330 and 275 m.y. Similar processes took place in other segments of the Hercynian Belt (Malavielle et al., 1990; Seranne, 1992). High relief and thick crust are stable while compression at the edges of the belt is maintained, but after its cessation, gravitational instability results in a short period of time (Bird, 1979; Houseman et al., 1981; Glazer and Bartley, 1985). Significant weakening occurs 15–20 m.y. after the main orogenic events, and the deep crustal root of the chain probably delaminates and collapses. This leads to rapid uplift and extension, increased heat flow and temperature gradient and, after 30–35 m.y. to thermal subsidence. If this process is completed, a 'normal' crust about 35 km thick will result and gravitational equilibrium will be restored.

This process has been used to explain the late orogenic extension of some orogenic belts (McClay et al., 1986; Menar and Molnar, 1986; Dunbar and Sawyer, 1988; Nelson, 1992) and could explain some features of the origin and evolution of the Iberian Basin and coeval basins and the lack of substantial crustal roots, a common feature in the northern European Hercynides and the Appalachians.

The concept of crustal collapse after continental collision followed by increased heat flux and extension was applied by Lorenz and Nicholls (1974, 1984) to the Late Paleozoic belt of Hercynian Europe and to the Iberian segment by Doblas (1991), Díez-Balda et al. (1992), Doblas et al. (1993), Escuder et al. (1994) and Díez-Balda et al. (1995), but the timing and tectonic regimes in the transition from late Hercynian events (Late Carboniferous) to the early alpine ones (early-Late Permian) are still open to discussion.

Present crustal thickness of the Iberian Massif is about 34 km, and, as the Alpine Orogeny did not affect greatly its central part, delamination and collapse of the roots should take place just after the main Hercynian movements. Intrusion of hot asthenosphere caused melting in the lower crust and emplacement of Stephanian–Autunian S-granites and andesitic intraplate volcanism. The sedimentary infilling of the Autunian basins, however, could represent the beginning of the alpine extension, after the volcanic intrusions, in relation with the punctiform type of extension observed in other basins, with a strong strike-slip component. Generalised extension and weakening of the upper crust took place during the Thüringian, 20 m.y. after the last granites were emplaced, as predicted by the delamination model.

Rifting and extension ceased in the Iberian Basin after another 20 m.y., during the late Anisian, and were re-emplaced by thermal sag and subsidence, again in good agreement with the model. Renewed extension took place during the Early Jurassic and the early-Middle Cretaceous.

We believe that, as was suggested by Menar and Molnar (1986) for the northern segment of the Hercynides, deep-rooted collapse led to the creation of the Iberian Basin and coeval basins and, in this sense, they are a Hercynian heritage and not an independent phenomenon, but tectonic regime was clearly different and the successive stages can be clearly differentiated.

12. Conclusions

The Iberian Basin was an intracontinental rift structure that appeared in the northeast margin of the Iberian Microplate during the Early Permian, and its first stages were contemporaneous with the emplacement of late Hercynic granitoids and andesitic continental volcanism.

The rifting or tectonic subsidence phase of the Iberian Basin stretched from the Early Permian to the Early Triassic (about 40 m.y.) and was succeeded by a thermal subsidence or flexural stage until the Early Jurassic.

Sediments in the rifting phase were of continental origin and accumulated in asymmetric half-grabens. The depocentres migrated steadily to the northeast with time. The sediments can be subdivided into six depositional sequences bounded by unconformities.

The basin boundary faults are interpreted as lystric faults trending NW–SE becoming horizontal perhaps at 12–15 km, with secondary transfer faults trending NNE–SSW offsetting the structure. Extension was accommodated in the hanging wall block by the secondary grabens and the formation of a central high.

The original orientation of the Iberian Basin is interpreted as a response to dextral strike-slip stress at the margins of the Iberian Microplate and deformation along old lineaments trending NW–SE.

The rifting started in the central part of the Basin as isolated pockets of extension during the Autunian that propagated in two directions with time: northwest and southeast. The hot spot hypothesis for the origin of the basin is rejected.

The Catalan and Valencia–Prebetic Basin were developed in a later stage and are related to the southwards propagation of the Hesse–Burgundy rift during the Late Permian.

The driving mechanism of the Early Permian–Early Triassic basins is thought to be a consequence of the crustal collapse of the deep roots of the overthickened Hercynian Belt and later extension, rifting and increased heath flow and thermal gradient.

The Iberian Basin, as well as the Asturian and Pyrenean Basins, can be considered a late Hercynian heritage, because they developed along Hercynian lineaments and are closely related to the late Hercynian events and their tectonic consequences. The Catalan and Valencia–Prebetic Basins were related to a different slightly younger tectonic process: the southwest migration of the Hesse–Burgundy Rift along the eastern margin of the Massif Central and the Cevennes Fault System.

Alpine compression and tectonic inversion in the Iberian Range took place along the same tectonic lineaments controlling the extensional events and this fact points out the role of the late Hercynian lineaments in the alpine history of the Range.

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