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POST-COLLISIONAL FORMATION OF THE ALPINE FORELAND RIFTS

E. Craig Jowett

Department of Earth Sciences, University of Waterloo, Waterloo, Ontario Canada N2L 3G1

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Abstract: A series of Cenozoic rift zones with bimodal volcanic rocks form a discontinuous arc parallel to the Alpine mountain chain in the foreland region of Europe from France to Czechoslovakia. The characteristics of these continental rifts include: crustal thinning to 70-90% of the regional thickness, in cases with corresponding lithospheric thinning; alkali basalt or bimodal igneous suites; normal block faulting; high heat flow and hydrothermal activity; regional uplift; and immature continental to marine sedimentary rocks in hydrologically closed basins. Preceding the rifting was the complex Alpine continental collision orogeny which is characterized by: crustal shortening; thrusting and folding; limited calc-alkaline igneous activity; high pressure metamorphism; and marine flysch and continental molasse deposits in the foreland region. Evidence for the direction of subduction in the central area is inconclusive, although northerly subduction likely occurred in the eastern and western Tethys.

The rift events distinctly post-date the thrusthing and shortening periods of the orogeny, making "impactogen" models of formation untenable. However, the succession of tectonic and igneous events, the geophysical characteristics, and the timing and location of these rifts are very similar to those of the Late Cenozoic Basin and Range province in the western USA and the Early Permian Rotliegendes troughs in Central Europe. This analogy suggests a similar origin by tensile stresses caused by viscous drag of mantle material which is mechanically and thermally coupled to the subducting plate. This model is broadly compatible with published estimates of lithospheric yield strengths and of tensile shear stresses produced by descending slabs, especially if the foreland was fractured by prior collision. It is hypothesized that compression at the trench before and during the orogeny counteracts this tensile shear stress until subduction ceases, whereupon the descending plate breaks off and continues to sink, causing extension in the overlying foreland crust and lithosphere and creating the foreland continental rift zones.

Key words: Alpine foreland, rifts, subduction.

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INTRODUCTION

The Cenozoic volcanic and rift system in the foreland of the Alpine orogeny forms an arc trend parallel to, and 100-400 km from, the Alpine thrust front (Fig. 1). The arc is formed by the Rhodanian, Limagne, Bresse, Rhine, Ruhr, Leine, Eger, and Labe rifts in France, Germany, and Czechoslovakia. This system or parts of it, has been variously considered to be related to (1)

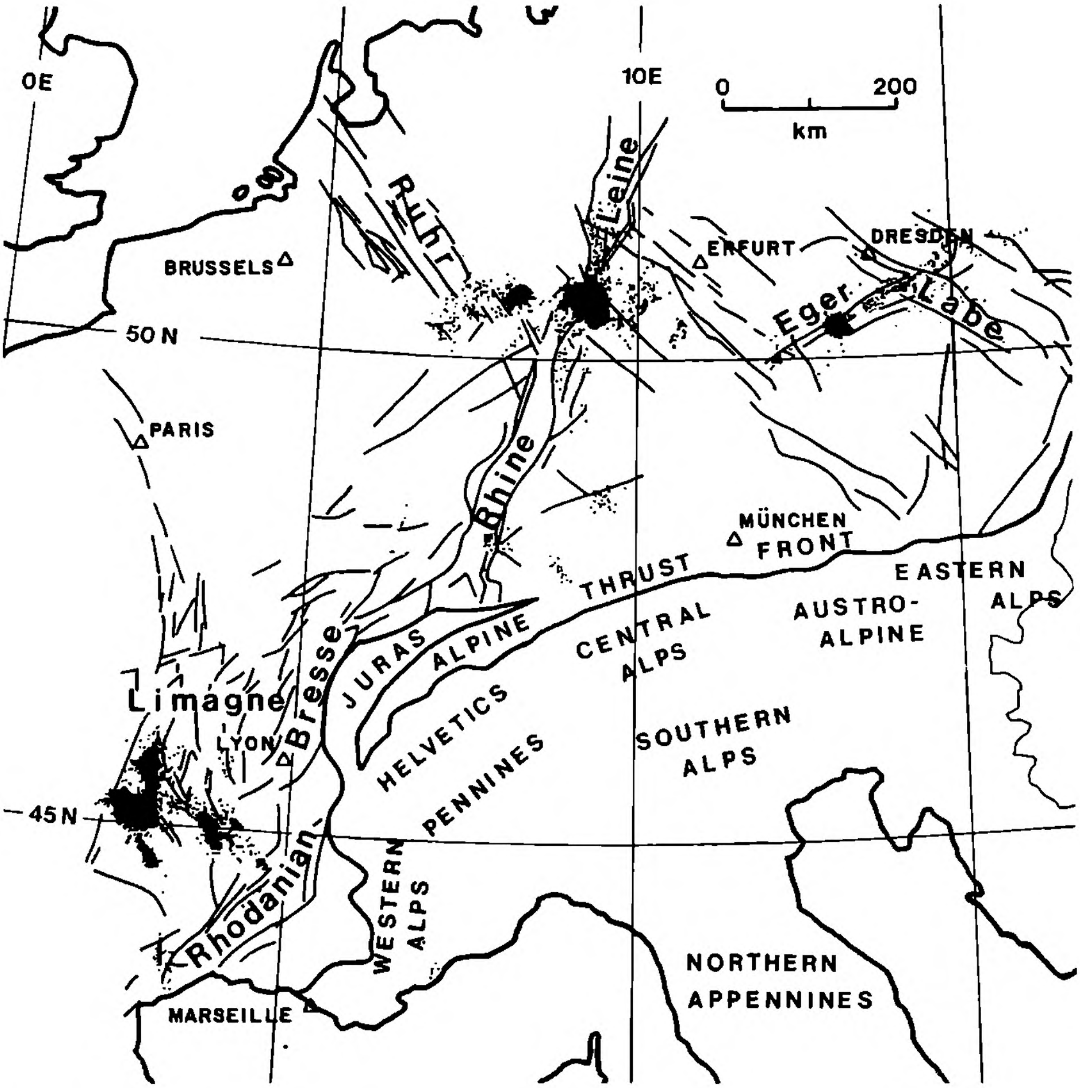

Fig. 1 Location map of the Alpine rift system and Alpine-related tectonic elements discussed in the text. The post-collisional Alpine rifts are located 100-400 km from the Alpine thrust front and outlined by major faults and Neogene bimodal igneous rocks (stipple) (after Illies, 1974; Maurey & Varet, 1980; Ziegler, 1982)

drift of Europe over a mantle plume (Duncan *et al.*, 1972; Burke *et al.*, 1973), (2) upwelling mantle and asthenosphere plumes (Brousse & Bellon, 1983), (3) Greenland-Eurasia spreading and North Sea rifting (Voight, 1974), (4) a larger rift system extending from the North Sea to East Africa (Richter-Bernburg, 1974); and (5) lateral redistribution of crust during collision of a continental promontory (Sengör, 1976; Tapponnier, 1977; Sengör *et al.*, 1978; Gordon & Hempton, 1986). The last model has been the most popular because the timing and position of the rift arc do indicate a genetic link with the Alpine orogeny (Illies, 1974; Ziegler, 1982; 1987). This "impactogen" model has, however, been inadequately supported by geologic evidence that demonstrates synchronism of rifting an collisional orogeny. In addition, the model has been used primarily to explain the Rhine rift only; not the whole rift system.

The purpose of this paper is to document the chronological succession of sedimentary, igneous, and tectonic events in the Alpine foreland region in order to establish the timing and location of rifting within a plate-tectonic context (Fig. 2). These data provide the geologic constraints necessary to compare with other foreland rifts and to propose a genetic model for the Alpine rift system as a whole.

COMPRESSIONAL TECTONISM

Plate convergence in the eastern Tethys (Greece, Turkey) dates to Late Triassic or Early Jurassic time (Sengör & Yilmaz, 1981), while in the western Tethys, convergence and subduction likely began in Early Cretaceous (Trümpy, 1973; Ziegler, 1982). In the Penninic basement nappes of the western sector, high-pressure metamorphic rocks are dated radiometrically at 130 to 110 Ma, and initial folding related to compressional convergence is dated stratigraphically as Aptian and Albian (115 to 100 Ma). Ziegler (1982) contends that convergence during Late Cretaceous and Early Tertiary (~ 80 to 60 Ma) between the Italo-Dinarid promontory (Adria) and the Central Carpathian Massif to the south and the European craton to the north was essentially north-south. This direction is supported by the drift reconstructions of Pitman and Talwani (1972) and Dewey *et al.* (1973) and by the paleostress orientations of Letouzey and Trémolieres(1980).

The compression resulting from the Sub-Hercynian and Laramide collisions (~ 85 to 65 Ma and 60 to 55 Ma) affected the foreland up to 1000 km north of the present thrust front by inverting sedimentary basins (Ziegler, 1982). For instance, the Polish Trough, a major NW-trending rift zone active in the Early Permian and again in the Triassic, contained 10 km of Late Paleozoic and Mesozoic sedimentary rocks (Pożaryski and Brochwicz-Lewiński, 1978) and must have once had a thinned crust to accommodate this subsidence (Ziegler, 1982). However, inversion of this basin by Alpine tectonism produced a thickened crust (Znosko, 1981) by folding of the sediment fill, reversal of movement along previous tensional faults, and deep wrench faulting. Variscan massifs (Harz, Fore-Sudetic Block, *etc.*) were tilted up along steep reverse and transcurrent faults (Ziegler, 1982), resulting in crustal shortening similar to that produced in the North American Cordilleran foreland by uplifted basement blocks (Laramide) (Hamilton, 1981). This fracturing during compression may have substantially weakened the crust and affected the location of subsequent foreland rifts.

The Alpine continental collision event is complex. Collision recurred repeatedly in several main events and at slightly different times along the Alpine chain. Further complications arise from the different names for the tectonic events along the chain. In the Central and Eastern Alps (Austroalpine), the Paleo-Alpine orogeny caused thrusting and folding from ~ 100 to 85 Ma (also called Pre-Gosauian stage) and from 75 to 68 Ma Intra-Gosauian) (Trümpy, 1973, 1980; Fauple *et al.*, 1980; Janoschek & Matura, 1980). Crustal shorte-

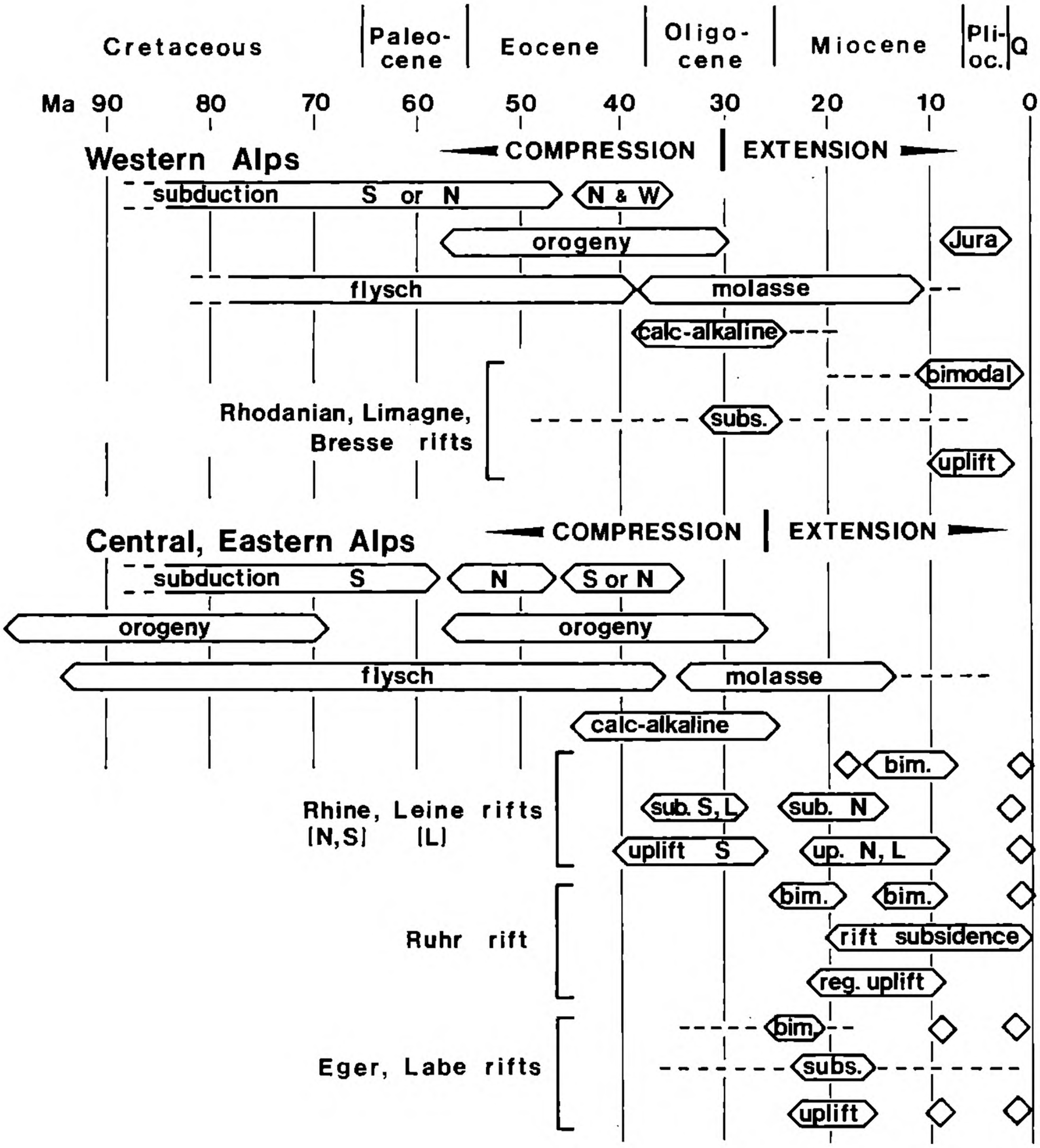

Fig. 2 Succession and timing of tectonic, igneous, and sedimentation events in the Western, Central, and eastern Alpine regions prior to and during formation of the Alpine foreland rift basins. Bimodal igneous activity, rift subsidence, and regional uplift mark the extensional stress regime of ~ 25-0 Ma (~ 30-0 Ma in the W. Alps), distinctly post-dating the main crustal-shortening collision event of ~ 40-32 Ma. Any model of genesis must explain the association with the orogeny, and the change from compression in the foreland to extension immediately following the orogeny. (Compiled from: Knetsch, 1963; Hsü & Schlanger, 1971; Trümpy, 1973, 1980; Illies, 1974, 1978; Lippolt *et al.*, 1974; Teichmüller, 1974; Coisy & Nicolas, 1978; Illies & Greiner, 1978; Channell *et al.*, 1979; Kopecky, 1979; d'Argenio *et al.*, 1980; Carte Géologique de la France, 1980; Cortesogno *et al.*, 1980; Debelmas *et al.*, 1980; Fauple *et al.*, 1980; Janoschek & Matura, 1980; Kornprobst, 1980; Letouzey & Trémolières, 1980; Lüttig & Walter, 1980; Malkovsky, 1980; Maury & Varet, 1980; Radulescu & Sandulescu, 1980; Ziegler, 1980, 1982, 1987; Brousse & Bellon, 1983; Homewood *et al.*, 1986; Pfiffner, 1986; Bachmann *et al.*, 1987)

ning likely ensued from continental collision between the Italo-Dinarid promontory (Adria) of Africa and the European craton during the Sub-Hercynian and later Laramide orogenies (Ziegler, 1982) (Fig. 2).

The main Meso-Alpine orogeny occurred between 40 and 32 Ma in the Western and Central Alps (Trümpy, 1973, 1980) and between ~ 48 and 30 Ma in the E. Alps (termed the Post-Gosauian stage) (Fauple *et al.*, 1980). This main stage coincided with the end of flysch sedimentation and resulted in much of the Alpine metamorphism, the Periadriatic calc-alkaline granitoids, and possibly 300 km of crustal shortening (a closing rate of 4 cm/a) (Trümpy, 1973, 1980; see Fig. 2). Thrusting cotinued in the Central and E. Alps until the Oligocene-Miocene boundary (~ 24 Ma).

The Neo-Alpine event in the Helvetic Alps (30 to ? Ma) produced perhaps 60 km of shortening and indicates an external subduction zone (Trümpy, 1973). In Late Miocene and Pliocene time, the Jura thrust was formed in the W. Alps (Fig. 2).

FORELAND FLYSCH AND MOLASSE SEDIMENTARY ROCKS

As the Alpine nappes advanced northwards during the Late Cretaceous and Early Tertiary compression, narrow Alpine and Carpathian foredeep basins developed by loading and flexuring of the continental plate (Ziegler, 1982). The flysch may have been deposited in a trench or back-arc setting, or between two plates moving transversely to each other (Hsü, 1972). The associated ophiolites indicate that the depositional site was ocean rather than continental crust.

The whole foreland area emerged with the onset of major compression in the Late Paleocene (~ 56 Ma). By the Oligocene the Alpine foredeep shallowed to deposit shallow marine then continental foreland molasse sediments (Buchi & Schlanke, 1977; Homewood *et al.*, 1986; Pfiffner, 1986), although flysch sedimentation persisted in the Carpathian foredeep. Oligocene to Late Miocene molasse is also present in the W. Alps (Kerckhove, 1980; Homewood *et al.*, 1986).

The foreland flysch and molasse basins (>3 km thick) were partly overridden and covered by advancing thrust sheets which were active in the W. and E. Alps until earliest Miocene (~ 24 Ma) (Ziegler, 1982). Major thrusting occurred in the Late Oligocene (Fig. 2), as indicated by the presence of Oligocene sediments in the nappes and by the presence of Early Miocene sediments overlying thrust fronts.

CALC-ALKALINE IGNEOUS ACTIVITY

Compared to the E. Alps and Carpathians, andesitic or calc-alkaline igneous activity in the W. Alps was limited in volume, particularly for an orogeny of this magnitude and with the apparent subduction of significant oceanic plate (Dercourt, 1980; Maury & Varet, 1980). Volcanic and intrusive rocks, with their remobilized detritus, are aligned in a north-south trend parallel to the W. Alps (Fig. 3) and likely signify that an Oligocene volcanic arc existed between 35 and 30 Ma (Maury & Varet, 1980). A younger group of calc-alkaline breccias and conglomerates (34 to 20 Ma) are present near Nice on the Mediterranean coast.

Volcanic rocks occur in the W. Alpine foreland flysch sediments, mainly as detritus but also as pyroclastic layers (Kerckhove, 1980; Maury & Varet, 1980; Pfiffner, 1986). The Taveyannaz and Elm clastic formations of Late Eocene to Early Oligocene age (~ 43 to 33 Ma) consist of up to 60% andesite clasts. Similar lithic types occur in other sedimentary formations but volcanic flows are rarely in place (Kerckhove, 1980).) This calc-alkaline detritus infers the presence of an andesitic island arc which was subsequently covered by thrust sheets or was completely eroded.

In the Southern Alps, calc-alkaline rocks include the large Bregaglia granitoids (30 to 25 Ma; tonalite, granodiorite and granite) and the Adamello granitoids (45 to 28 Ma) (Trümpy, 1980). The Central Alps of Switzerland and Austria contain only the continuation of the Taveyannaz volcaniclastic sandstone (Debelmas *et al.*, 1980; Trümpy, 1980). No calc-alkaline rocks were associated with the Late Cretaceous (~ 100 Ma) Paleo-Alpine event (Trümpy,1980).

In the E. Alps, Late Eocene to Oligocene (40 to 30 Ma) magmatism produced tonalite and granodiorite bodies (the Periadriatic Intrusions), and Early – Middle Miocene (~ 18–14 Ma) andesitic and dacitic volcanism occurred in the Styrian basin in southeast Austria (Janoschek & Matura, 1980). Carpathian calc-alkaline igneous activity continued from Late Cretaceous (~ 85 Ma) to Pliocene (~ 4 Ma) time in continental magmatic arcs resulting from subduction of oceanic plate (Radulescu & Sandulescu, 1980). The relatively continuous line of calc-alkaline intrusive masses between the W. Alps and Carpathians (Debelmas *et al.*, 1980) indicates that oceanic-plate subduction occurred all along the Alpine chain (Fig. 3).

EXTENSIONAL TECTONISM

The N-trending compression between Europe and Africa during the major crustal shortening period of Eocene – Early Oligocene time changed to a NE trend between the Italo-Dinarid plate (Adria) and the western Mediterranean area in Late Oligocene – Early Miocene time (Letouzey & Trémolières, 1980) as Adria moved southwesterly relative to Europe. This timing also corresponds to the Apennine orogeny caused by the easterly rotation of Sardinia–Corsica into Italy or Adria (d'Argenio *et al.*, 1980). The ENE stress direction in North Africa during Late Oligocene and Early Miocene indicates sinistral strike-slip movement oriented E-W between the two continents and a possible decrease in N-S compression.

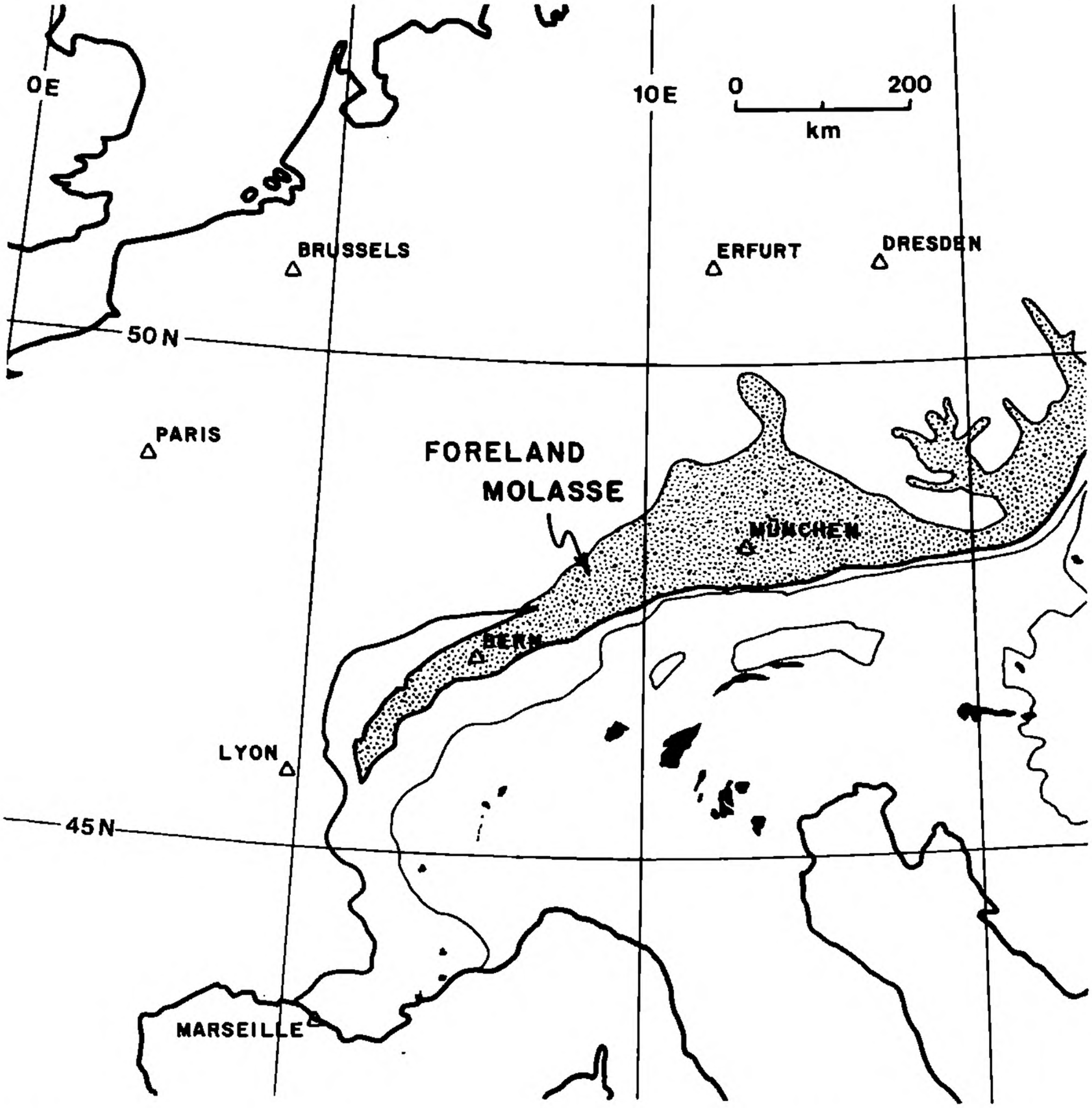

Fig. 3 Paleogene calc-alkaline igneous rocks and foreland molasse basin of the Alpine orogeny associated with collision (after Debelmas *et al.*, 1980; Ziegler, 1982)

The change to NE-oriented compression in Europe in Late Oligocene time coincides with the onset of major rifting, volcanism, and subsidence in the W. Alpine foreland (Fig. 1 and 2) followed by rifting in the central and E. Alps in the Miocene. In Late Miocene to Early Pliocene, a short reversal of direction moved the Italo–Dinarid plate to the WNW, forming the Jura thrust sheet in the W. Alps and inverting the Bresse and southern Rhine rifts, while the other rift basins continued to subside (Illies, 1978; Rat, 1978; Ziegler, 1982). This time period also coincides with major volcanic events in the Limagne area, the formation of the Rhine-Ruhr-Leine junction (Vogelsberg), and the second volcanic phase in the Eger area (Fig. 1 and 2). Present-day compressional stress directions are N-NW (Letouzey & Trémolieries, 1980; Philip, 1980; Illies *et al.*, 1981) allowing some subsidence in the NW-trending Ruhr rift. In the Jura mountain region, the direction is still W to NW, effectively keeping the Bresse rift inactive.

BIMODAL IGNEOUS ACTIVITY

Alkali basalt and bimodal (basalt-rhyolite) rocks erupted between ~ 24 Ma and the present in the French rifts, with most activity between 20 and 4 Ma (Maury & Varet, 1980; Brousse & Bellon, 1983) (Fig. 1 and 2). Earlier volcanism between 65 and 30 Ma was of minor importance (Brousse & Bellon, 1983). No volcanism is known from the Bresse rift.

Minor basalt was produced sporadically in the Rhine rift since the Late Cretaceous (Lippolt *et al.*, 1974; Mäussnest, 1974) and peaked in the Eocene (50 to 40 Ma). These earlier Rhine volcanics were not voluminous but typically formed dikes along NNE-trending fissures that formed as early as the Paleocene (60 Ma). Many of the fissures were not volcanic and likely represent rejuvenated Permian structures (Illies, 1974; Mäussnest, 1974). Ziegler (1980, 1982) regarded the Paleocene basalt dikes as products of wrench faulting during Alpine compression (which caused basin closure or inversion) rather than representating crustal extension. The main volcanism (tholeiitic basalt at the Rhine-Ruhr-Leine rift junction centred on the Vogelsberg pile) was Middle to Late Miocene (~ 12 to 8 Ma) in age, with a resurgence in the Quaternary (~ 1 Ma) (Teichmüller, 1974; Lüttig & Walter, 1980).

In the Ruhr (Lower Rhine) rift, alkali basalt and trachyte tuffs were extruded mainly between 25 and 18 Ma (Early Miocene) (Teichmüller, 1974) but continued to near-Recent time in the south near the Rhine rift (Eifel pile) (Lüttig & Walter, 1980). Rift volcanic rocks just east of the Rhine-Leine rifts (Rhön and Knülgebirge areas) are mainly 15 to 10 Ma in age (Knetsch, 1963).

In the Eger (Ohre) and Labe rifts, bimodal alkaline magmatic rocks formed in three distinct periods dated by K-Ar studies: 35 to 17 Ma (Oligocene to Early Miocene); 9 Ma (Late Miocene); and 1 Ma (Quaternary) (Kopecky, 1979). The first phase, which peaked in earliest Miocene (25 to 22 Ma), was the most extensive and voluminous (Malkovsky, 1980).

BLOCK FAULTING AND RIFT-BASIN FILL

Although some Late Eocene continental sediments were deposited in the Limagne region, the major period of tensional faulting and rift-basin fill began in mid-Oligocene time (~ 35 Ma) (Fig. 2). (Scarp-related fanglomerates, saline-lake deposits in hydrologically closed basins, and immature sediments in general signify fault-related deposits.) Differential subsidence was most important during the Oligocene but continued well into the Pliocene, depositing up to 2500 m of fluviatile and saline-lake deposits in a horst and graben environment (Carte Géologique de la France, 1980).

Likewise, the Rhodanian rift subsided rapidly from the end of the Eocene through the Oligocene. The fault-bounded basins accumulated up to 1-2 km of mainly saline-lake and restricted marine sedimentary rocks which pass laterally into the coeval W. Alpine foreland molasse (Carte Géologique de la France, 1980). In the Bresse (Rhone) rift, closed basins formed during the Early Eocene. In the Late Oligocene, intense tectonism and subsidence deposited thick alluvial-fan facies and marine evaporites followed by Miocene saline-lake deposits (Carte Géologique de la France, 1980). Formation of the Jura Mountains during the Late Miocene and Early Pliocene terminated subsidence (Illies, 1978; Rat, 1978).

In the Rhine rift, major normal faulting and differential subsidence ensued from Late Eocene to Middle Oligocene time in the south end of the rift (Illies, 1974), possibly with basinal uplift at ~ 10 Ma (Villemin *et al.*, 1986). The centre of subsidence moved northward to the central area in Middle to Late Oligocene and to the northern end in the Miocene. Rifting ceased between mid-Miocene and Early Pliocene time (Illies & Greiner, 1978) and Late Pliocene reactivation was concentrated in the northern area (Illies, 1974).

Subsidence in the Leine (Hesse) rift took place along Early Permian (late Variscan) structures during Late Eocene to Late Oligocene time (~ 38 to 28 Ma) (Illies, 1974; Schenk, 1974) and is presently being uplifted (Ziegler, 1982). Ruhr (Lower Rhine) rifting was active from Early Miocene (~ 20 Ma) to Recent time and formed perpendicular to Variscan trends. Eger rift sediments range in age from Eocene to Recent (Kopecky, 1979) but most subsidence occurred during Early to Middle Miocene time (~ 25 to 11 Ma) (Malkovsky, 1980).

The Vienna Basin rift zone began to subside in mid-Miocene, and by the Quaternary, nearly 2 km of marine then continental basin-fill had accumulated; no evidence for volcanism is present in the fill (Janoschek & Matura, 1980). It is not clear whether the Vienna and the Pannonian Basin to the east belong in the foreland-rift family and they will not be considered further.

REGIONAL UPLIFT IN THE RIFT ZONES

The areas of rifting were subjected to regional uplift during the Late Cenozoic (Ziegler, 1980, 1982) which, as shown below, was generally coincident with volcanism and rift subsidence. The Massif Central (Limagne rift) was uplifted in Late Miocene to Pliocene time (~ 8 to 3 Ma), coincident with volcanism but also with the Jura thrusting and wrench faulting in the Massif Armoricain to the northwest (Coisy & Nicolas, 1978; Ziegler, 1982).

Regional uplift of the Vosges-Black Forest Massif around the Rhine rift began in the Eocene and continued to Late Pliocene-Early Pleistocene (~ 45 to 1 Ma), although the area may now be subsiding (Illies, 1978; Illies & Greiner, 1978; Villemin *et al.*, 1986). The Rhenish Massif (Rhine-Ruhr-Leine rift junction) uplift began in Early Miocene (~ 20 Ma) before the main volca-

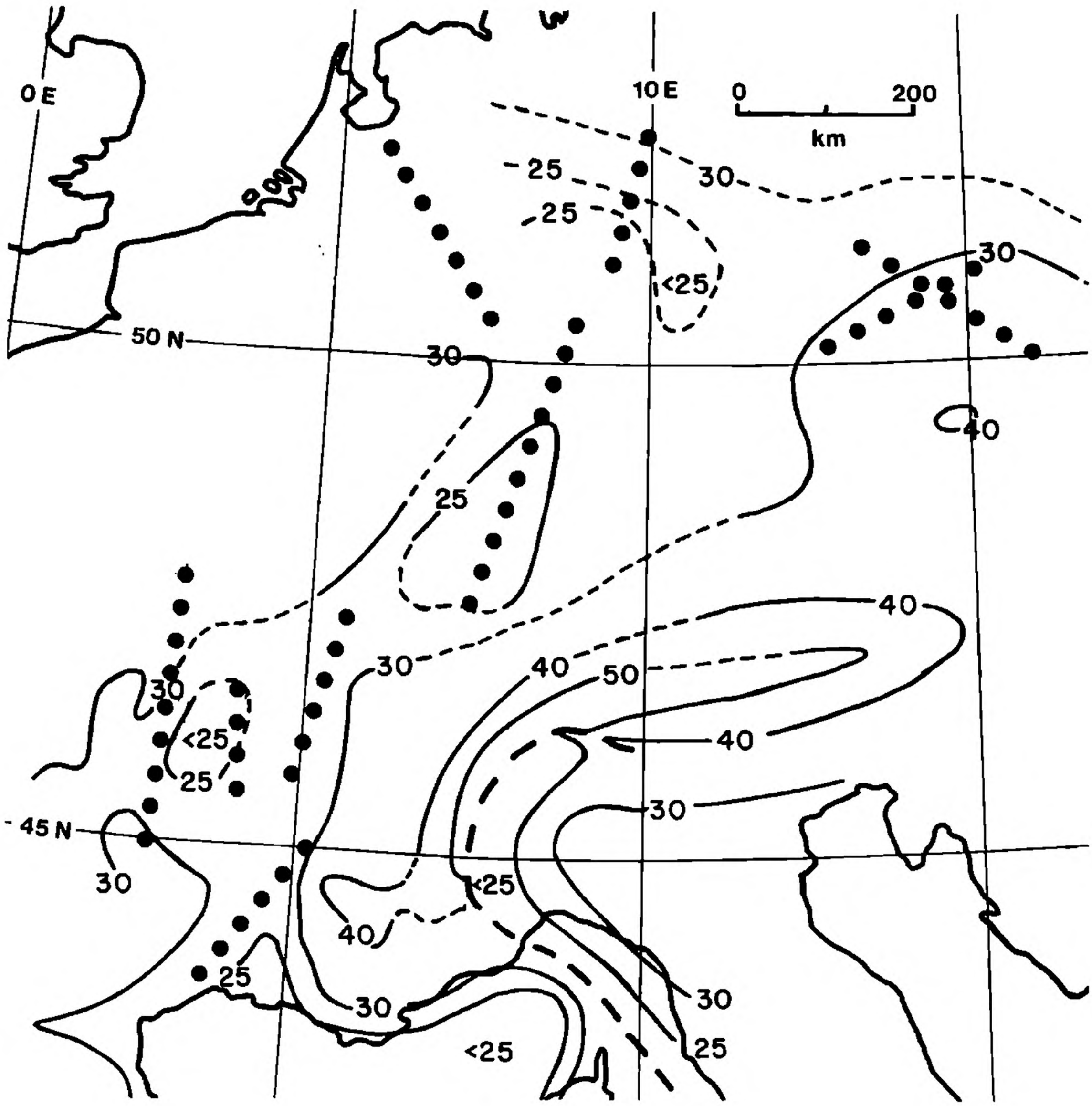

Fig. 4 Continental crust under the rifts is thinned relative to the regional 30-35 km and is thickened to 45-60 km under the Alpine orogen (after Giese, 1976; Zatopek, 1979; Hirn, 1980). Dots represent rifts shown in figure 1

nism later in the Miocene (~ 12 to 8 Ma). Uplift recurred in the Quaternary (1 Ma) accompanied by strong volcanism (Teichmüller, 1974).

The Bohemian Massif (Eger rift) was uplifted in the Early to Middle Miocene coeval with the first phase of volcanism (~ 24 Ma) and rapid subsidence of the rift valley (Malkovsky, 1980; Ziegler, 1982). Uplift and volcanism occurred again in the latest Miocene (~ 6 Ma) and in the Pleistocene (~ 1 Ma). Uplift in the Central Alps began in the Oligocene, accelerated in the Pliocene and is continuing at a measured rate of > 1 km/Ma in the middle of the chain (Illies & Greiner, 1978).

GEOPHYSICAL CHARACTERISTICS

CRUSTAL AND LITHOSPHERIC THICKNESSES

Seismic soundings show that the continental crust under the Limagne rift thinned to 20-24 km from a regional value of 30 km (Hirn and Penier, 1974; Hirn, 1980) (Fig. 4). The crust of the Bresse rift zone is less, but clearly, thinned (27-29 km) relative to the Limagne rift (Hirn, 1980), while the crust of the Rhodanian rift to the south has a thickness of 24 – 27 km (Fig. 4).

Similarly, the crust under the Rhine rift is thinned in a linear zone coincident with the rift to 22 km in the south and 26 km in the north from a regional 30 – 32 km thickness on the east and west flanks (Giese, 1976) (Fig. 4). The Eger rift has a 29 – 32 km thick crust (Beranek and Zounkova, 1979; Zatopek, 1979), although this may represent a thinning of only 2-3 km from the regional.

Assuming a pre-rift thickness of 30 – 35 km, the crust under these Cenozoic rifts has been thinned by a factor of 1.3 – 1.5 in France, 1.2 – 1.5 in the Rhine area, and ~ 1.1 – 1.2 in the Eger rift area. Under the Alpine orogeny, however, the crust is thickened to 45 – 60 km by thrusting and crustal shortening from a regional 35 km (Hirn, 1980; Trümpy, 1980).

The lithosphere appears to be thinner than normal in the Rhine and Ruhr areas but thicknesses calculated in the Limagne and other rift areas are contradictory (Souriau, 1981; Calcagnile & Scarpa, 1985).

HEAT FLOW

The Limagne, Rhodanian, Bresse, and Rhine rift zones are all characterized by anomalously high heat flow above 80-90 mW/m^2 (Fig. 5) (Gable, 1979). Similarly, the Ruhr, Leine, Eger, and Labe rifts are anomalous (Cermak, 1979; Hurtig and Oelsner, 1979). Because the trends of heat flow are the same as the thinned crust and in cases the lithosphere, the anomalies are likely caused by passive upwelling of asthenosphere during lithospheric thinning (Jarvis, 1984; Villemin et al., 1986). The background and presumably pre-rift heat flow is ~ 70 – 80 mW / m^2 (Fig. 5), a value high enough that significant extensional deformation is likely (Lynch & Morgan, 1987; Kusznir & Park, 1987).

EVIDENCE FOR SUBDUCTION DIRECTION

Oceanic plate subduction in the Alpine region can be inferred from evidence of closure of an ocean (Tethys) and the presence of at least a small calc-alkaline arc and obducted oceanic crust (ophiolites). Whether oceanic (Radulescu & Sandulescu, 1980) or continental (Debelmas *et al.*, 1980) plate was subducted, and especially the direction, is controversial. Continental crust

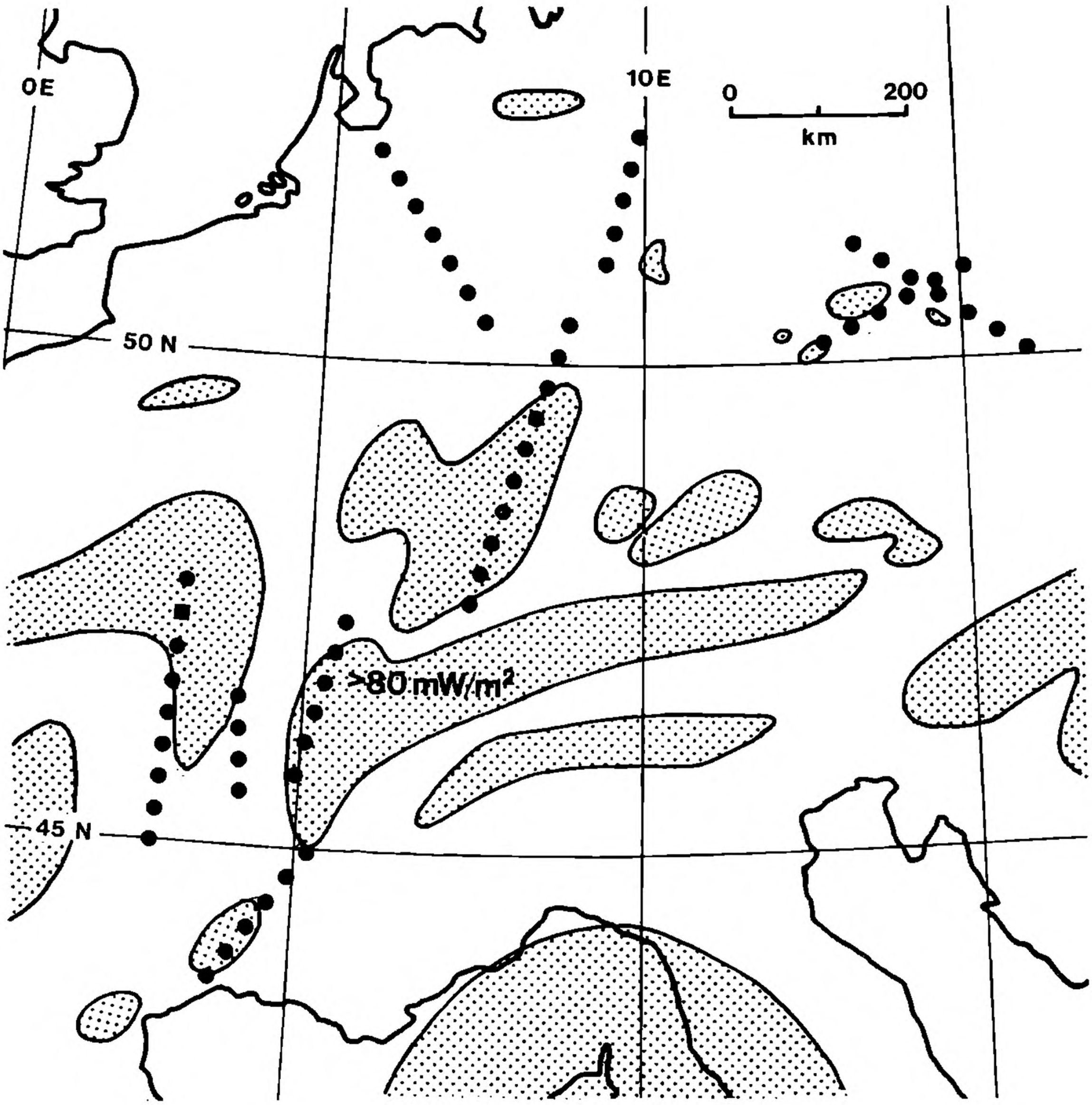

Fig. 5 Rifts generally coincide with elevated heat flow values of >80 mW/m^2 (stippled) compared to regional background of 70-80 mW/m^2 (from Cernak & Hurtig, 1979). Dots represent rifts shown in figure 1

subduction under continental crust (A-type) can also be interpreted as the effect of overthrusting and crustal shortening during collision. Subduction directions have generally been inferred to be the same as dip directions of thrust sheets (e.g., Janoschek & Matura (1980) for the Austroalpine belt). However, this criteria cannot be used implicitly because, for example, west-dipping thrust sheets in the N. American Cordillera were produced during eastward subduction (Monger *et al.*, 1972; Dickinson, 1976). Although many authors are uncommitted with regard to plate subduction, the following summarizes some of the arguments.

In the E. Alps, Dietrich (1976) and Frisch (1978) invoke southward subduction of oceanic plate under the North Calcareous Alps during the Late Cretaceous and Early Tertiary. Frisch (1978) considered the molasse foredeep as the zone of continental underthrusting. Dietrich (1976) cited the calc-alkaline magmatism to the south as an indication of southward subduction. Janoschek and Matura (1980) also favour subduction of the Pennine oceanic plate (located between the Bohemian Massif and the Austroalpine nappes) southward under the Austroalpine and South Alpine continental crust during Paleo-Alpine compression (100 to 70 Ma), apparently because of the south-dipping thrust sheets. After this event, the ophiolites were completely covered by northward advancing Austroalpine thrust sheets. They also advocate continuous southward subduction of the Pennine oceanic crust during the Early Cenozoic to account for the main phase of shortening and thrusting and the Periadriatic intrusions (40 to 32 Ma) along the southern margin of the Austroalpine segment. Channell *et al.* (1979) agree that the Pieniny Klippen flysch belt (the Carpathian equivalent of the Austroalpine Pennine belt) indicates southward subduction of the flysch trough in the Late Cretaceous and Early Tertiary. The ophiolite-flysch units of the Pennine-Pieniny klippen have been thrust up onto continental crust on either side of the suture in opposite directions, suggesting that the direction of obduction does not necessarily reflect the dip of the subduction zone. Coleman (1971) and Dewey (1976) both proposed models in which ophiolites may be obducted up onto the overriding continental plate (usually from a back-arc basin) to dip in the opposite direction of the subduction zone.

In the eastern Tethys, Tapponnier (1977), Channell *et al.* (1979) and Sengör and Yilmaz (1981) proposed, on the basis of ophiolite and calc-alkaline belts and geologic reconstructions, that the Tethys oceanic plate was subducted primarily northward under Europe from Early Cretaceous to the Neogene. Even though the plate tectonic history in this area is complex, Sengör and Yilmaz (1981) detailed adequate geologic evidence to determine the northerly direction with confidence. Similar substantive evidence is lacking in areas to the west.

Northward-dipping subduction from Late Paleocene to mid-Eocene was favoured for the Central Alps by Hsü and Schlanger (1971), who suggested that the flysch was situated behind (north of) an island arc. In their model, based on the occurrences of granitic magmas, the subduction direction flipped to a southerly one in mid-Eocene time. Channell and Horvath (1976), on the other hand, suggested that the direction may have flipped from south to north during the later stages of crustal shortening. Oxburgh (1972) advocated northward oceanic plate subduction and 'flaking off' of the top part of continental crust attached to the subducting oceanic plate. Thrusting of this flake onto the European plate during collision would give the false impression of southward-dipping subduction. Reutter *et al.* (1980) also proposed lithospheric splitting, based on a low-velocity zone at 30-60 km depth under Italy which suggests that continental crust was thrust under to the east, even though subduction was to the west. This flaking or splitting is also suggested by Trümpy (1980) for the Aar Massif in Switzerland, which he considers as upper crust sheared off at the first (granitic) zone of velocity inversion and then bent upward.

Aubouin (1980) concluded, on the basis of calc-alkaline rocks and attitudes of overthrust sheets, that southward subduction occurred from the Carpathians westward and northward subduction from the Balkans-Hellenides eastward. Tapponnier's (1977) reconstruction allows for little or no subduction north and west of the Adrian promontory.

In the western Mediterranean, Channell *et al.* (1979) illustrate a model of post-Eocene subduction of Tethyan oceanic plate northward under Calabria (S. Italy), Sardinia, and the European craton. They note that southerly subduction might have existed from the Late Cretaceous to the Eocene, as put forward by Boccaletti *et al.* (1976) and Reutter *et al.* (1978). However, Channell *et al.* (1979) suggest that the evidence for southward subduction in this area is unclear, and that the evolutionary picture of the western Mediterranean is better explained by continuous northward subduction between the Apennines (Italy) and Corsica (and Europe) from the Late Cretaceous onwards.

On the basis of pressure-temperature constraints of metamorphism, Kornprobst (1980) introduced two subduction-obduction models for the W. Alps in France. The model that better explained the metamorphism had eastward subduction of oceanic plate under the Austroalpine continent (equivalent to southward subduction in the E. Alps) until the Late Cretaceous whereupon the oceanic plate was obducted westward (or northward) onto the European plate during Paleo-Alpine compression. By the mid-Eocene (~ 45 Ma) subduction reversed to a westward (or northward) direction under the European craton until the orogeny was completed by the Early Oligocene.

In summary, the plate tectonic history of the Alpine orogeny is complex and controversial and there is rarely definitive evidence to favour one model over another. It is possible that subduction of the Tethys oceanic plate produced a calc-alkaline island arc which was infrequently preserved in place. Subduction was possibly directed southward in the E. Alps and Carpathians and northward (and westward) in the Central and W. Alps with the two sections separated by a north-trending transform fault. In the western Tethys, either there was a Tertiary reversal from southward to northward subduction, or the direction was always northward. It is also possible that in the E. Alps the direction switched from south to north, or was northerly all along: arguments used for exclusively southward subduction are just not conclusive. East of the Carpathians, northward subduction was the norm.

PROPOSED GENETIC MODEL OF THE ALPINE FORELAND RIFTS

In spite of the complexity of plate movements and the apparent insolubility in some areas of the direction of subduction, the definite pattern of type,

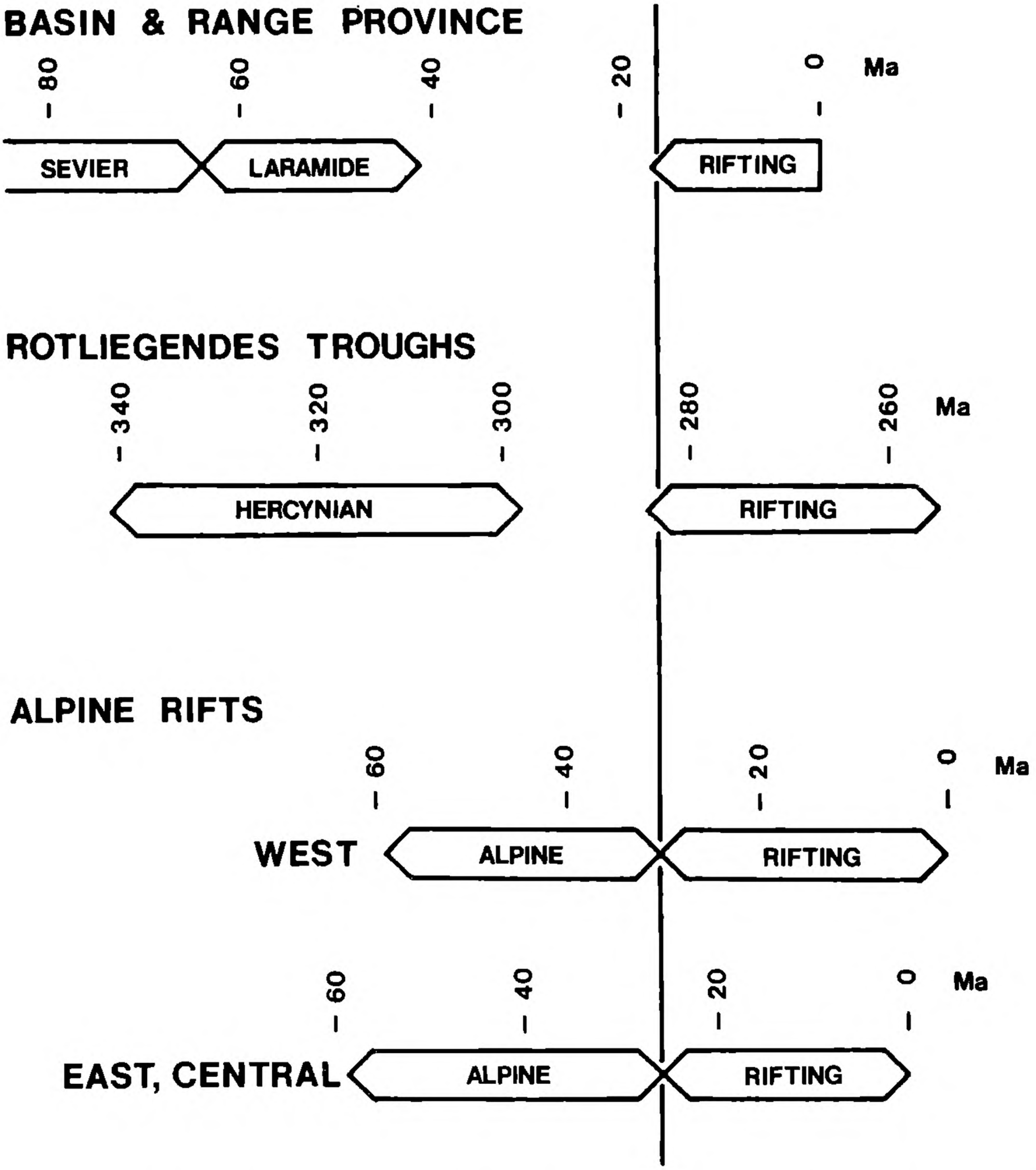

Fig. 6 Post-collisional extensional regimes of the Basin and Range and Rotliegendes rifts (Jowett & Jarvis, 1984) compared to the Alpine rifts. The striking similarities suggest a common origin for these foreland rifts

timing, and location of tectonic and igneous events shown in figure 2 sets constraints on any genetic model for the foreland rift basins.

The compressional stress regime from 100 to ~ 25 Ma (100 to ~ 30 Ma in W. Alps) is chronologically distinct from the extensional regime of ~ 25 to 0 Ma (~ 30 to 0 Ma in W. Alps). The succession of tectonic and igneous events is very similar to the succession of events in the Cenozoic Basin and Range province in western USA and Early Permian Rotliegendes rifts in Europe (Fig.

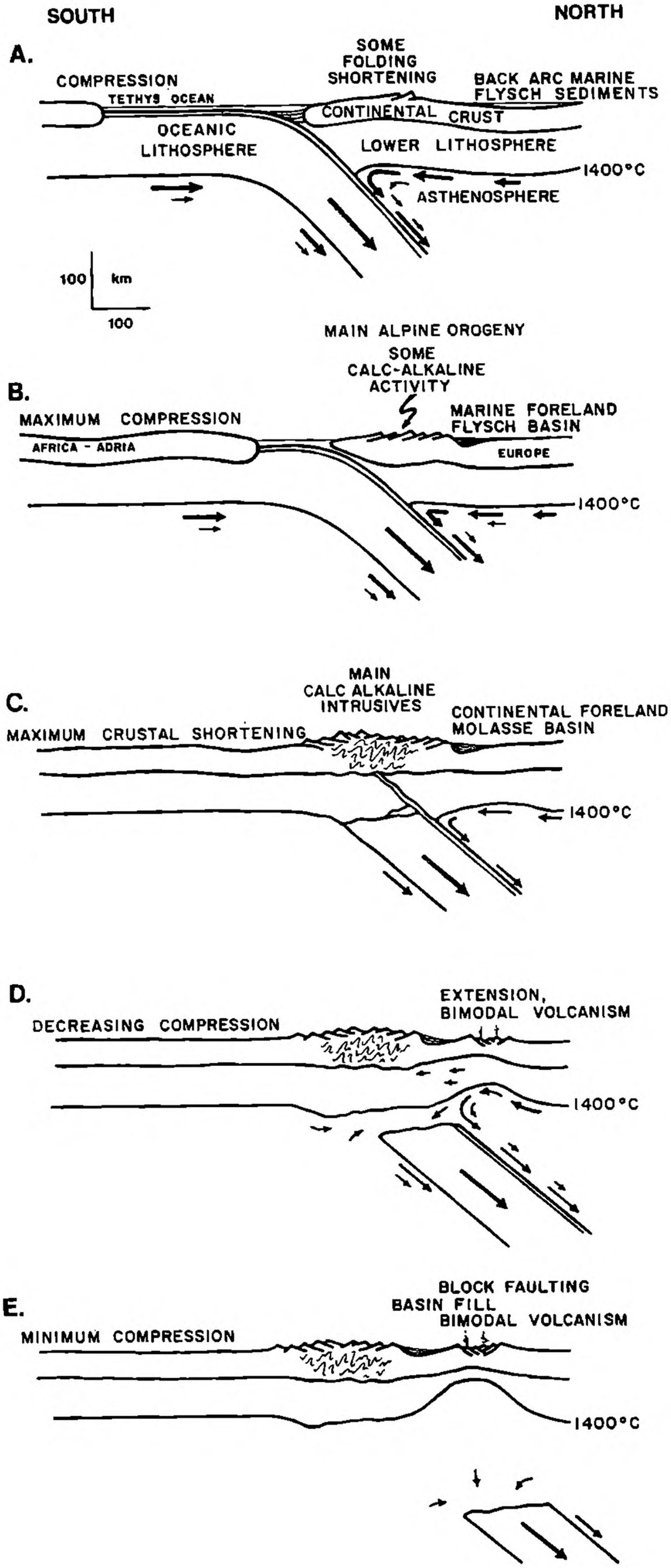

Fig. 7 A possible plate-tectonic model of formation of the Alpine foreland rifts.
A. Compressional oceanic plate subduction, possibly with calc-alkaline volcanic arc – foreland tensional stress of ~ 10 MPa due to secondary convention cell is counteracted by compression.
B. As Africa-Adria continent approaches, maximum lateral compression produces main Alpine thrusting.
C. Continental collision results in maximum crustal shortening, calc-alkaline intrusions, and molasse basin in foreland.
D. Change to bimodal igneous activity and extension as compression decreases.
E. Block faulting, basin fill, and further bimodal volcanism as sinking slab breaks off. Modified from Jowet and Jarvis (1984)

6). In addition, the characteristics of crustal extension and thinning, normal block faulting, hydrologically closed basins, immature continental to marine basin-fill, and bimodal volcanism are similar in the three areas. In both of the still-active Alpine and Basin and Range areas, regional uplift, high heat flow, and hydrothermal activity are characteristic. Noteworthy differences between the Alpine and the other two regions are: (1) bimodal volcanism generally followed basin subsidence in the W. Alpine system rather than preceding it as in the East African rifts (Baker, 1986) and the Basin and Range and Rotliegendes rifts; (2) the short elapsed time between compressional and extensional regimes in the Alpine rifts (in contrast with the ~ 20 Ma lag in the other regions). The similarities are striking enough, however, to suggest a general model of formation for all three rift zones.

Models of promontory impact (with sideways movement of crust) and "impactogen" models to form foreland rifts (e.g., Sengör *et al.*, 1978) are defensible when the collisional orogen and rifting are coeval. However, when applied to the Alpine rifts or parts of them (Sengör, 1976; Tapponnier, 1977; Sengör *et al.*, 1978; Gordon & Hempton, 1986; Hancock & Bevan, 1987), they are inconsistent with the distinct time separation of compressional and extensional regimes (Fig. 2). (The Leine subsidence and the early strike-slip movement and minor volcanism in the S. Rhine, however, do coincide with the main orogenic event (Fig. 2) and can be explained by impact compression.) Even more incompatible with the geologic evidence is the impact model suggested for the Rotliegendes rifts (Gordon & Hempton, 1986) where there is a 15 Ma time lag between orogeny and rifting (Fig. 6).

In contrast, the genetic model proposed in this paper for the Alpine foreland rifts (Fig. 7) is compatible with the geologic evidence, especially the timing and location of the rifts, and can be generally applied to these other areas. It requires oceanic plate subduction under the rifted continent (i.e., northward in the case of the Alpine) but not particular features such as promontories. The model assumes a tensile stress under the continental foreland resulting from mantle material being drawn back towards the subducting plate by viscous drag (see Jowett & Jarvis, 1984). Strong secondary convection cells do develop under overriding plates after ~ 5 Ma of subduction (Hsui & Toksöz, 1979) and couple the slabs thermally and mechanically to the overlying lithosphere (Hager & O'Connell, 1981; Rabinowicz *et al.*, 1983; Mitrovica & Jarvis, 1985, 1987; Garfunkel *et al.*, 1986). The cells impart vertical deflections and horizontal tensile shear stresses on the base of the overlying lithosphere (Hager & O'Connell, 1981; Mitrovica & Jarvis, 1987) in the general area where foreland rifts are located. The ~ 10 MPa lateral tensile shear stress on the base of the lithosphere calculated by Mitrovica and Jarvis (1987) does not decrease with time if a slab is substituted for the isolated cold plume used in their modeling. When lateral migration of the slab away from the trench (Chapple & Tullis, 1977) is considered, the convective velocity below the overriding plate increases (as well as the distance along which the shear stress acts) (Garfunkel *et al.*, 1986) and the tensile shear stress likewise increases (Hager & O'Connell, 1981). Therefore a shear stress of 10 MPa or more is a reasonable value.

With heat flow values of 70-80 mW/m^2, the critical tensile force necessary for extension as defined by Kusznir and Park (1987) is 1.5-2.5 x 10^{12} N/m applied laterally through the thickness of the lithosphere. Using the over-simplified assumption of a homogeneous and entirely brittle crust, this force is crudely equivalent to a tensile shear stress of 43-83 MPa applied to the base of a 30-35 km crust. Although detailed modeling with a heterogeneous lithosphere is necessary, the ~ 10 MPa tensile shear stress calculated by Mitrovica and Jarvis (1987) is within an order of magnitude of the stress necessary for extension and may be adequate in itself if the lithosphere was previously weakened by igneous activity or fracturing.

To explain why rifting does not occur earlier, the model assumes that the tensile stress in the foreland region is counteracted by the lateral compression imparted at the trench as the Africa-Adria continent approaches (Fig. 7a). Collision with Europe (Fig. 7b) forms thrust sheets and initiates foreland molasse deposition. Lateral crustal shortening follows (Fig. 7c) and the subducted plate breaks off, continuing to sink because of negative buoyancy. When lateral compression is decreased, either by cessation of subduction or by a change in plate motion (e.g., strike-slip between Africa and Europe), the tensile shear stress, maintained by the sinking slab for millions of years (Mitrovica & Jarvis, 1987) but now with no counteracting compressional stress, stretches and thins the foreland area (Fig. 7d). Extensional tectonics, high heat flow, and regional uplift result until the sinking slab is decoupled from the overlying continent (Fig. 7e). The decrease in overburden pressure and increase in heat flow induce partial melting in the lower lithosphere to produce basaltic volcanism.

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Streszczenie

POKOLIZYJNA GENEZA RYFTÓW PRZEDPOLA ALP

E. Craig Jowett

Seria kenozoicznych stref ryftowych z bimodalnym (bazalt – riolit) wulkanizmem tworzy na przedpolu Alp od Francji po Czecho-Słowację nieciągły łuk równoległy do łańcucha alpejskiego. Charakterystyczne cechy tych ryftów kontynentalnych obejmują: ścienienie skorupy do 70-90% jej regionalnej miąższości – niekiedy związane ze ścienieniem całej litosfery, wapienno-alkaliczne lub bimodalne asocjacje magmowe, obecność bloków ograniczonych uskokami normalnymi, wysokie wartości strumienia cieplnego i przejawy działalności hydrotermalnej, regionalne wydźwignięcie, skały osadowe od niedojrzałych kontynentalnych po morskie w odciętych hydrologicznie basenach. Powstanie ryftów było poprzedzone złożoną orogenezą alpejską, związaną z kolizją kontynentalną. Orogeneza ta odznaczała się: skróceniem skorupy, nasunięciami i fałdowaniem, ograniczonym wulkanizmem wapienno-alkalicznym, metamorfizmem wysokociśnieniowym, osadami morskiego fliszu i lądowej molasy na przedpolu. W części centralnej nie ma rozstrzygających dowodów na kierunek subdukcji, a we wschodniej i zachodniej Tetydzie mogła ona być skierowana ku północy.

Wydarzenia ryftowe są wyraźnie późniejsze od etapu orogenezy związanego z nasuwaniem i skróceniem, w związku z czym nie dają się utrzymać "zderzeniowe" modele ich powstania. Jednakowoż, następstwo zdarzeń tektonicznych i magmowych, cechy geofizyczne oraz czas powstania i rozmieszczenie tych ryftów są bardzo podobne jak w późnokenozoicznej prowincji Basin and Range na zachodzie USA i jak w późnopermskich rowach czerwonego spągowca w środkowej Europie. Ta analogia sugeruje podobną genezę – w rezultacie naprężeń rozciągających wywołanych lepkim ciągnięciem materiału płaszcza, który mechanicznie i termicznie łączy się z płytą pogrążaną. Model ten jest w ogólnych zarysach zgodny z publikowanymi ocenami wytrzymałości litosfery i naprężeń rozciągających wytwarzanych przez opadające płaty, zwłaszcza, gdy przedpole jest spękane w wyniku wcześniejszej kolizji. Można przypuszczać, że kompresja w rowie przed orogenezą i w trakcie niej przeciwdziała temu naprężeniu rozciągającemu tak długo aż ustanie subdukcja, a opadający płat odłamie się i dalej pogrąża, powodując rozciąganie w nadległej skorupie i litosferze przedpola oraz powstanie stref ryftowych przedpola.