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N. Rollet, Jacques Déverchère, M.-O. Beslier, Pol Guennoc, J.-P. Rehault, et al.. Back-arc extension, tectonic inheritance and volcanism in the Ligurian Sea, Western Mediterranean. *Tectonics*, 2002, 21 (3), 1015 (23 p.). 10.1029/2001TC900027 . hal-00406718

HAL Id: hal-00406718

<https://hal.science/hal-00406718v1>

Submitted on 9 Feb 2021

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Back arc extension, tectonic inheritance, and volcanism in the Ligurian Sea, Western Mediterranean

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Received 17 April 2001; accepted 4 September 2001; published 15 June 2002.

[1] The Ligurian basin, western Mediterranean Sea, has opened from late Oligocene to early Miocene times, behind the Apulian subduction zone and partly within the western Alpine belt. We analyze the deep structures of the basin and its conjugate margins in order to describe the tectonic styles of opening and to investigate the possible contributions of forces responsible for the basin formation, especially the pulling force induced by the retreating subduction hinge and the gravitational body force from the Alpine wedge. To undertake this analysis, we combine new multichannel seismic reflection data (*Malis* cruise, 1995) with other geophysical data (previous multichannel and monochannel seismic sections, magnetic anomalies) and constrain them by geological sampling from two recent cruises (dredges from *Marco* cruise, 1995, and submersible dives from *Cylice* cruise, 1997). From an analysis of basement morphology and seismic facies, we refine the extent of the different domains in the Ligurian Sea: (1) the continental thinned margins, with strong changes in width and structure along strike and on both sides of the ocean; (2) the transitional domain to the basin; and (3) a narrow, atypical oceanic domain. Margin structures are characterized by few tilted blocks along the narrow margins, where inherited structures seem to control synrift sedimentation and margin segmentation. On the NW Corsican margin, extension is distributed over more than 120 km, including offshore Alpine Corsica, and several oceanward faults sole on a relatively flat reflector. We interpret them as previous Alpine thrusts reactivated during rifting as normal faults soling on a normal ductile shear zone. Using correlations between magnetic data, seismic facies, and sampling, we propose a new map of the

distribution of magmatism. The oceanic domain depicts narrow, isolated magnetic anomalies and is interpreted as tholeiitic volcanics settled within an unroofed upper mantle, whereas calcalkaline volcanism appears to be discontinuous but massive and has jumped in space and time, from the beginning of rifting on the Ligurian margin (~30 Ma), toward the Corsican margin at the end of the Corsica-Sardinia block rotation (~16 Ma). This space and time shift reveals the importance of the rollback of the Apulian slab and of the migration of the Alpine-Apennines belt front toward the E-SE for driving basin formation. We also state that initial rheological conditions and inherited crustal fabric induce important changes in the styles of deformation observed along margins and between conjugate margins. In the NE Ligurian basin the prerift Alpine crustal thickening together with slow rollback velocity likely contribute to distribute strain across the whole NW Corsican margin, whereas farther south the inherited Hercynian structural pattern combined with a faster rollback of the subducting plate tend to focus the extension at the foot of the margin, up to the Sardinian rift which ends within the SW Corsican margin. Therefore the mode of opening and the margin structures mainly depend on the balance between intrinsic, inherited crustal heterogeneity (fabric and rheological changes) and external conditions imposed by rollback of the subducting lithosphere. **INDEX TERMS:** 3040 Marine Geology and Geophysics: Plate tectonics (8150, 8155, 8157, 8158); 3025 Marine Geology and Geophysics: Marine seismics (0935); 8109 Tectonophysics: Continental tectonics—extensional (0905); 8159 Tectonophysics: Rheology—crust and lithosphere; **KEYWORDS:** passive margins, back arc extension, rollback, tectonic inheritance, subduction, Mediterranean Sea.

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1. Introduction

[2] The present-day western Mediterranean setting (Figure 1) results from an incomplete collision between the Africa and Eurasia plates that began during Cretaceous times. In this convergence context, back arc extension has widely developed behind the Apulian NW dipping arcuate subduction zone [e.g., *Cherchi and Montadert*, 1982; *Dewey et al.*, 1989; *Gueguen et al.*, 1997, 1998; *Jolivet et al.*, 1999, and references therein]. Extension has been

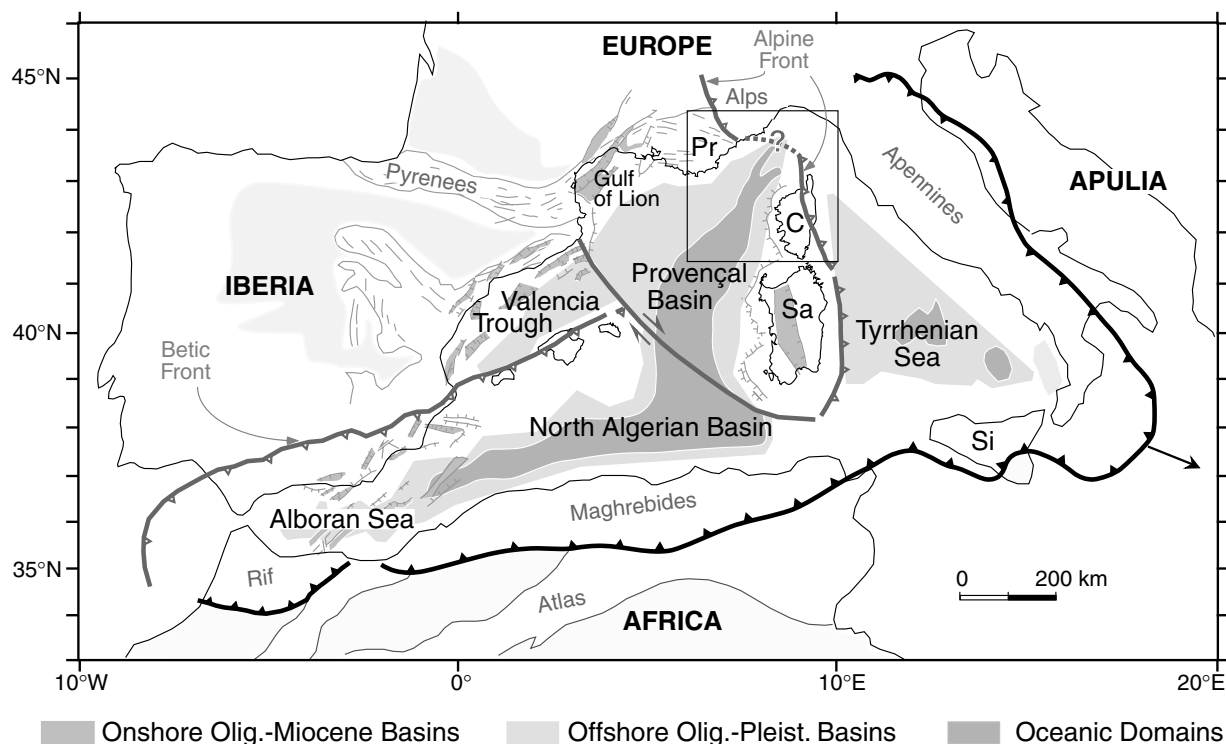


Figure 1. Simplified tectonic sketch of the western Mediterranean, modified after *Gueguen et al.* [1998]. Active extension has shifted eastward through time, following the rollback of the Ionian-Apulia subduction zone (thick solid line). The eastward migration of the Apennines trench and the back arc extension is symbolized by the dark arrow. Dashed line indicates the poorly known prolongation of the Alpine Front offshore. Inset depicts the Ligurian domain under study (location on Figure 2). Abbreviations are C, Corsica; Pr, Provence; Sa, Sardinia; Si, Sicily. (Reprinted with permission from Elsevier Science.)

triggered in the Ligurian domain, the Gulf of Lion, the Valencia trough, and the Alboran Sea, and has then propagated mostly from west to east, leading to the opening of the Algero-Provençal basin, and later, the Tyrrhenian Sea.

[3] This back arc extension has led to continental thinning, subsidence, and in the basins where continental breakup was achieved, to the formation of oceanic domains, the nature of which is reliably assessed only in the Tyrrhenian Sea [*Kastens and Mascle*, 1990]. In the case of the Ligurian Sea, the thinning is as large as ~ 18 km for the crust and ~ 50 km for the lithosphere relative to the “stable” European plate structure [*Kissling and Spakman*, 1996; *Waldhauser et al.*, 1998; *Chamot-Rooke et al.*, 1999], whereas a 30-km-thick continental crust and a 70-km-thick lithosphere are documented below Corsica [*Béthoux et al.*, 1999]. This makes this area the place where the highest topography and Moho gradients are preserved in the Mediterranean, with evidences for Moho steps as large as ~ 10 – 20 km between the European, Adriatic, and Ligurian Mohos [*Waldhauser et al.*, 1998; *Makris et al.*, 1999]. Another important feature of most western Mediterranean back arc basins (Alboran Sea, Valencia trough, Ligurian basin, Tyrrhenian Sea) is that they developed in sites partly thickened during previous orogenic phases (Betic, Pyrenean, Alpine, and Apennine orogenies, respectively). The reactivation of

mountain belts through back arc extension has resulted in widespread occurrence of exhumed and tectonically unroofed high-pressure metamorphic and plutonic rocks which have been especially well documented on Corsica and in the Tyrrhenian and Aegean seas [*Jolivet et al.*, 1990, 1999; *Fournier et al.*, 1991; *Gautier and Brun*, 1994; *Avigad et al.*, 1997; *Platt et al.*, 1998].

[4] The forces controlling the opening of several Mediterranean back arc basins have been intensively studied during the last decade. Three main types of processes have been proposed to explain the existence of contemporaneous and parallel extension regimes in the Tyrrhenian Sea and the Apennine orogen: (1) N-S shortening induced by the convergence of Africa and Europe [*Tapponnier*, 1977; *Boccalletti et al.*, 1982]; (2) passive retreating subduction of the Ionian-Adriatic domain [*Malinverno and Ryan*, 1986], later referred as “rollback” of the subducting slab; and (3) gravitational eastward collapse of the thickened Apenninic lithosphere [*Channell and Mareschal*, 1989]. A study using experimental modeling concludes that the three mechanisms must be combined in order to simulate the tectonics of the Tyrrhenian-Apennine system satisfactorily [*Faccenna et al.*, 1996]. However, several important questions are still a matter of debate in this respect: (1) the relative importance of tectonic forces acting during the evolution of Mediterra-

nean basins is often controversial, since they depend on complex, deep processes like delamination of lithospheric mantle, lateral slab detachment, variable buoyancy forces, or convective removal of thickened lithospheric roots, which are often poorly controlled [e.g., Carmignani et al., 1995; Lonergan and White, 1997; Carminati et al., 1998a, 1998b; Chamot-Rooke et al., 1999; Jolivet et al., 1999; Frizon de Lamotte et al., 2000]; (2) the role of tectonic inheritance within the heterogeneous upper plate undergoing extension behind the subduction zone has not yet been fully understood, in spite of recent modeling attempts [e.g., Faccenna et al., 1996; Becker et al., 1999]; and (3) the age, nature, and distribution of Neogene volcanism, which has accompanied and postdated back arc extension, are poorly constrained in many areas, and are not easily related to the various phases of back arc evolution [e.g., Mauffret et al., 1995; Lonergan and White, 1997; Jolivet et al., 1999; Frizon de Lamotte et al., 2000; Maury et al., 2000, and references therein]. The two latter points are particularly difficult to assess in the Ligurian Sea case, because little attention has been paid to this domain since the cruises performed in the 1980s [e.g., Réhault et al., 1984, 1985].

[5] This insufficient knowledge of the Ligurian domain evolution has therefore motivated the acquisition of new deep seismic reflection data (*Malis* cruise; lines located on Figure 2). This new data set is presented here, combined with previous data (seismic and rock sampling) and constrained by magnetic data and by rock samples newly dredged (*Marco* cruise) [Guenoc et al., 1995] or recovered during submersible dives (*Cylice* cruise) [Sossion and Guenoc, 1997]. They provide the unique opportunity to investigate in detail the superficial and depth structures of both margins and the basin and to document the extent of volcanism and its migration in space and time. The present paper analyzes the following: (1) the segmentation and the inheritance on the conjugate margins, in mapping the variations in structure and deformation of the thinned continental crust from SW to NE, i.e., along strike; (2) the tectonic style of the conjugate continental margins, and the extent and structure of the newly formed basin and of the transitional domain identified in between; and (3) the significance and origin of the volcanism and magmatism in the basin and on the margins. Hypotheses on the nature of the basement in the newly formed basin are proposed. These data also give some new insights into the relationships between extension and magmatism and the transition from continental rifting to seafloor spreading. From these observations we aim to better assess the relative contribution of the different mechanisms which led to the opening of the Ligurian basin.

2. Geodynamical and Geological Settings

2.1. Ligurian Basin Opening

[6] The opening of the first western Mediterranean basins was initiated ~30 Myr ago (see Jolivet and Faccenna [2000] for a review). They developed successively during late Oligocene–early Miocene in the Valencia, Ligurian, and Alboran basins [Maldonado et al., 1992; Roca and

Deségaulx, 1992; Fernandez et al., 1995], and during the middle-late Miocene in the eastern Balearic and Algerian basins [Roca and Deségaulx, 1992; Lonergan and White, 1997; Frizon de Lamotte et al., 2000], and until late Miocene and Plio-Pleistocene in the Tyrrhenian basin [Kastens et al., 1988; Gueguen et al., 1998]. All these basins were formed in a back arc position relative to the eastward or southward retreating Apulian (Apennines-Calabria-Maghrebides) subduction zone (Figure 1).

[7] Among these basins, it is worth noting that the oldest one, the Ligurian basin, has opened within the Alpine and northern Apennines collision belt in its northern part and is trending oblique to the main thrusts [Carmignani et al., 1995; Doglioni et al., 1997]. Clear compressive episodes have occurred successively from late Eocene to Pliocene in this part of the Alpine belt, comprising the southern sub-Alpine fold and thrust belt to the west and the internal Alps to the east. These phases are recorded on land, in particular at the contact of the foreland basin (the Provençal platform) with the Digne thrust sheet [Lickorish and Ford, 1998, and references therein] and as far as the Nice and Castellane arcs to the east, at the coastline [Ritz, 1991; Laurent et al., 2000] (Figure 3). Conversely, the southwestern Ligurian basin has developed within the eastern part of the Iberian plate [Carmignani et al., 1995], where no clear influence of the Pyrenean orogeny is documented. Whether the emplacement of the Oligocene mountain front near the present-day coastline of the north Ligurian basin implies significant crustal thickening close to it is questionable. The importance of prerift lithospheric thickening in the Liguro-Provençal zone is therefore highly variable in space and remains controversial [e.g., Mauffret et al., 1995; Chamot-Rooke et al., 1999]. However, note that calculated Oligocene uplift rates at the contact of the thrust sheets in the Digne arc are comparable to those recorded in active mountain belts [Artoni and Meckel, 1998]. Furthermore, the large part of the Ligurian domain located east of the main Frontal Penninic thrust (longitude 7.7°E, Figure 3) was necessarily located within the inner Alps at that time and therefore is likely to have undergone an important thickening.

[8] The timing of the so-called “synorogenic” to “post-orogenic” extension (i.e., the extension occurring after the main Alpine collision in the hinterland) is only approximate [see, e.g., Jolivet and Faccenna, 2000, and references therein]. Nevertheless, from synrift sediments sampled or observed in the Gulf of Lion and the Ligurian basin [Réhault et al., 1984, 1985; Gorini et al., 1993; Séranne, 1999] and from timing of uplift of rift shoulders [Jakni et al., 2000], it appears that rifting occurred from ~30 Ma to ~21 Ma and is followed by a ~30° counterclockwise drift phase of the Corsica-Sardinia block around a rotation pole located in the Gulf of Genova, from ~21 Ma to ~16 Ma [Vigliotti and Langenheim, 1995]. Recent paleomagnetic results coupled with $^{40}\text{Ar}/^{39}\text{Ar}$ dating in Corsica and Sardinia evidence a rotation of the Corsica-Sardinia block of 45°–55° during the drift phase [Gattacceca, 2001], thus confirming the prerift position from Gueguen [1995]. Although approximate, this timing shows that the Ligurian basin opened after the Eocene phase but almost contemporane-

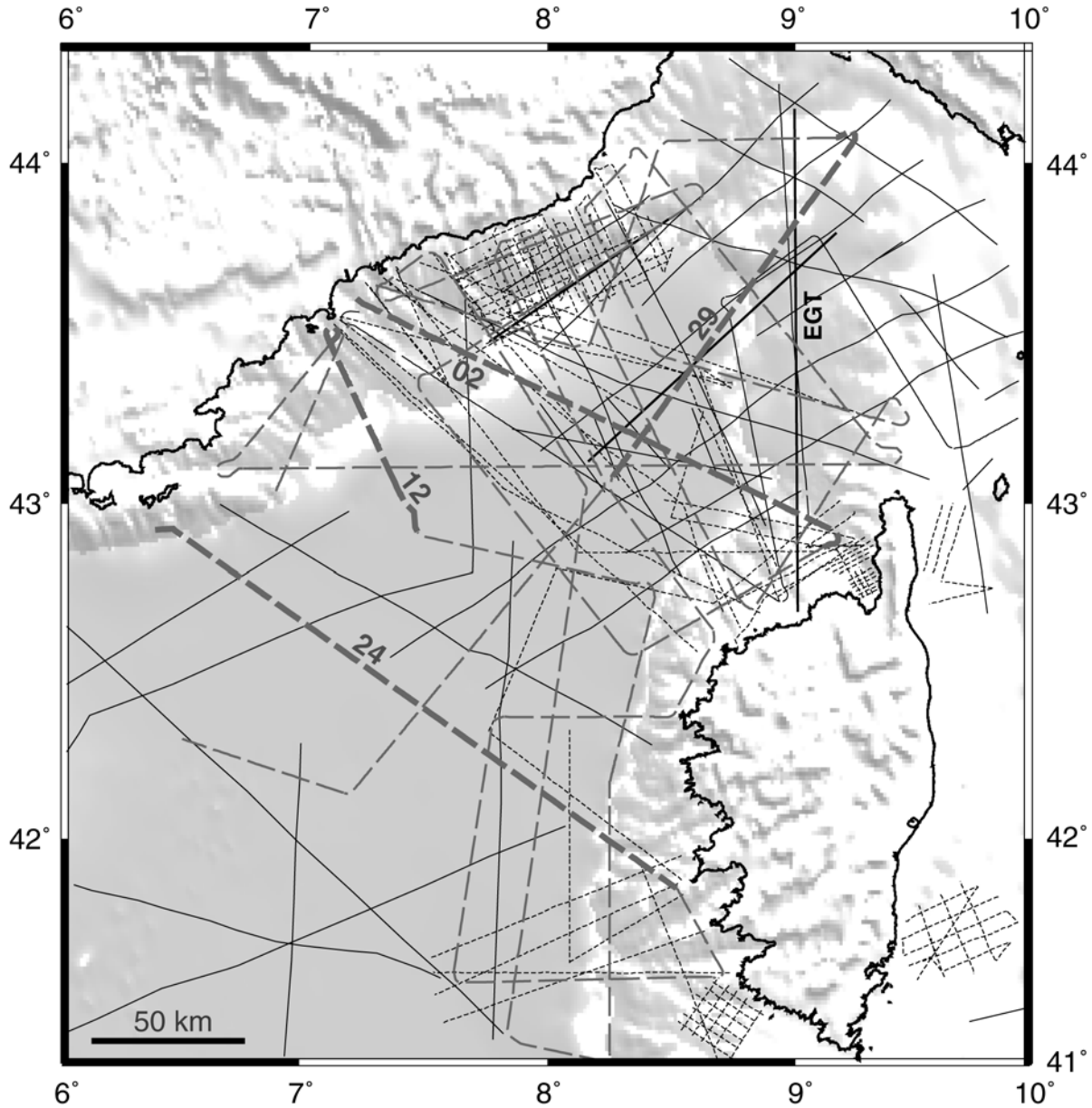


Figure 2. Location of multichannel seismic reflection profiles (long-dashed lines) acquired during *Malis* cruise [1995], together with previous wide-angle data (solid dark lines, including European GeoTraverse (EGT) [see *Egger et al.*, 1988]), other multichannel seismic reflection profiles (thinner dark lines) [see *Réhault et al.*, 1984], and monochannel seismic profiles (short-dashed lines). Numbered long-dashed lines are seismic reflection profiles shown in this paper. The background is shaded topography and bathymetry with coastline shown.

ously to the nappe emplacement phase of the Alpine orogeny which took place from late Eocene to early Miocene times [*Fry*, 1989].

2.2. Volcanic Activity and Back Arc Extension

[9] Rift-related volcanism is widespread in the western Mediterranean basins (Figure 4). The oldest samples recognized (Oligocene in age) are located in the Valencia trough, the Sardinian rift, and along the Liguro-Provençal margins

[e.g., *Coulon*, 1977; *Bellon*, 1981]. According to kinematics reconstructions [e.g., *Réhault*, 1981; *Burrus*, 1984; *Le Cann*, 1987; *Gueguen*, 1995; *Gueguen et al.*, 1998], this volcanism roughly trended NE-SW, i.e., parallel to the paleo-subduction zone before the rotation of the Corsica-Sardinia block. Therefore this volcanic activity seems at least partly linked to the subduction process. It is particularly well expressed on the western margins of the lithospheric continental blocks (Corsica-Sardinia-Mallorca), behind the Alpine-betic belt and the Apennines-Maghre-

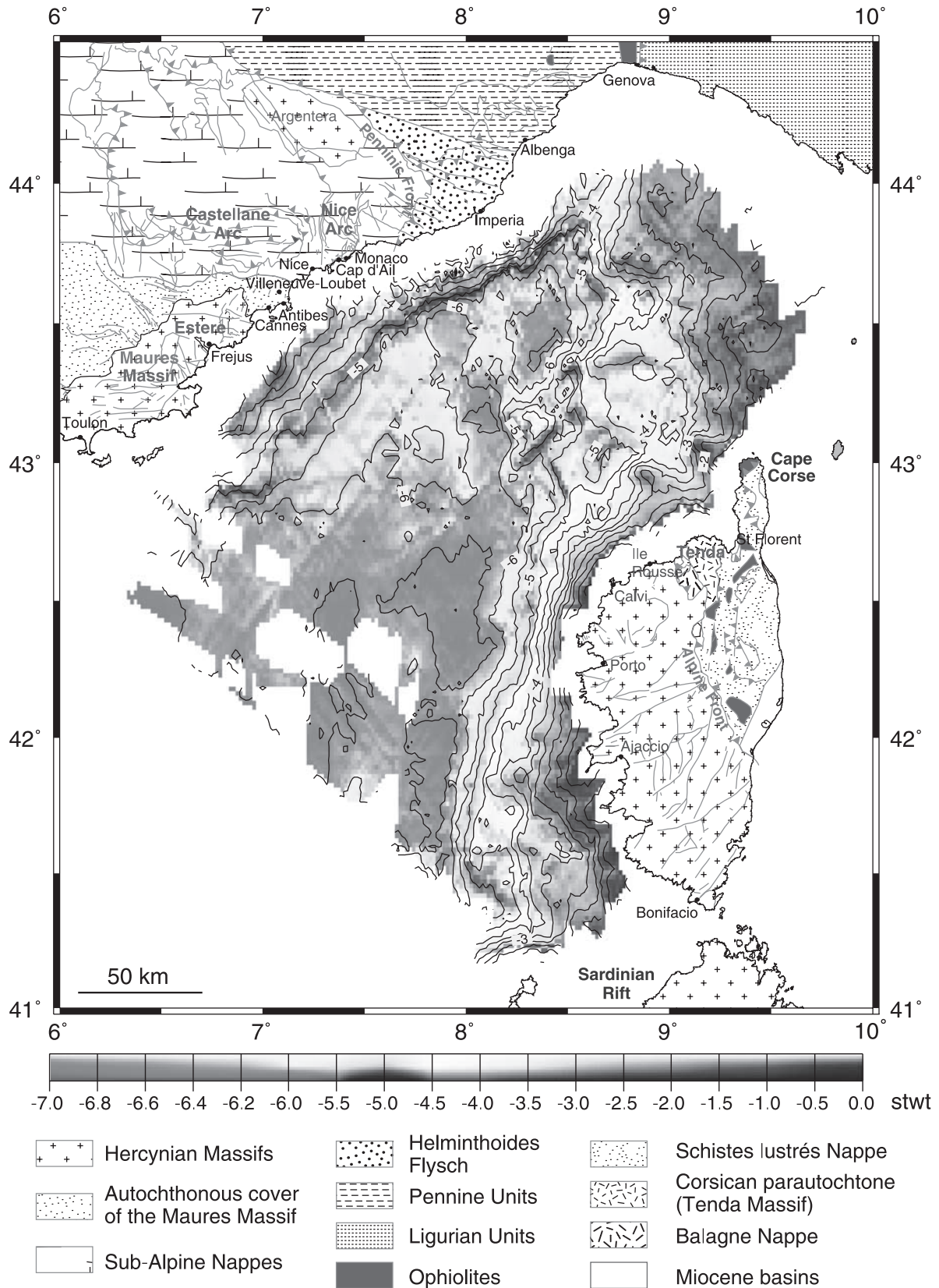
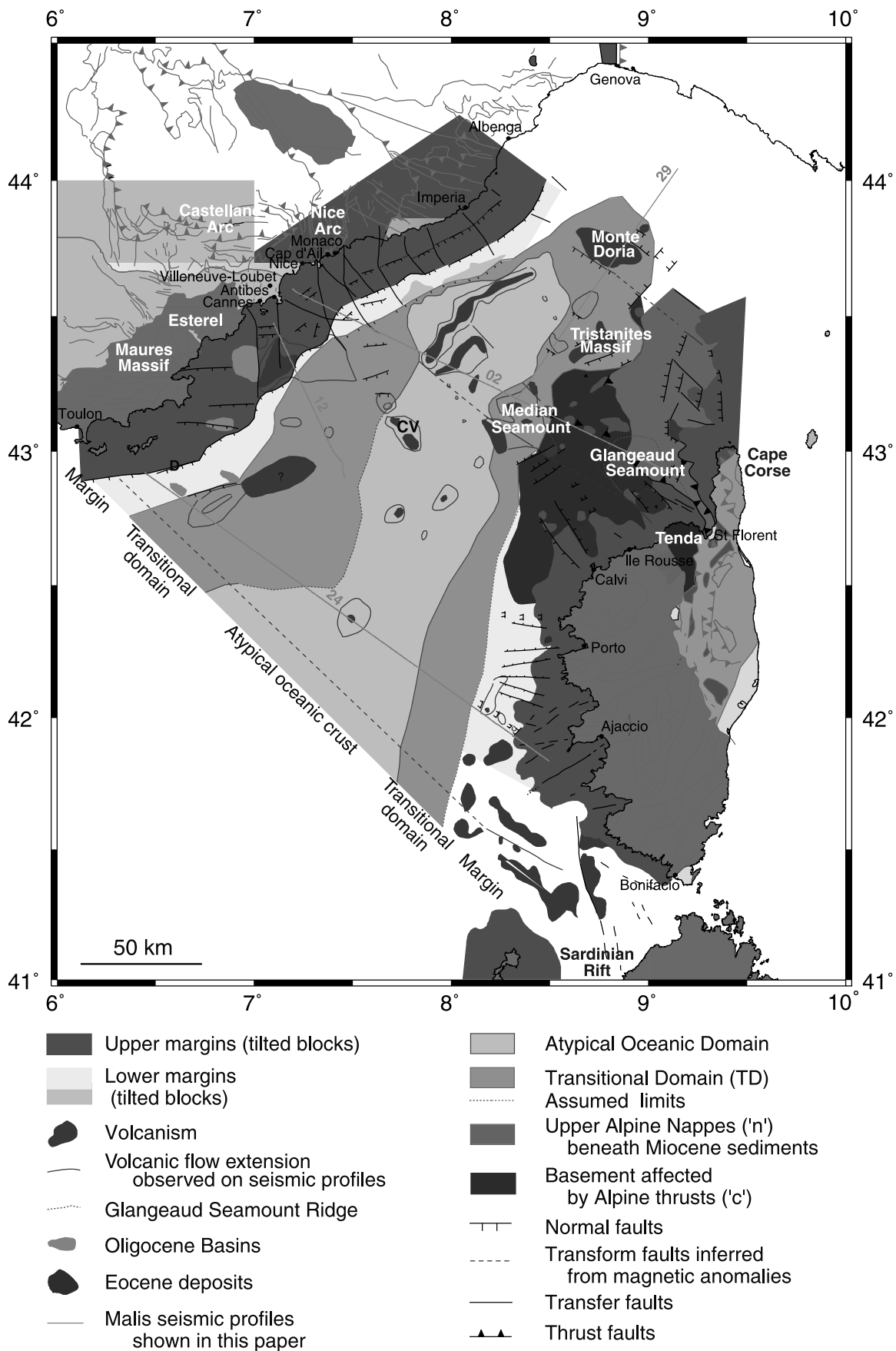


Figure 3. Isochrons of the top of the prerift basement (in seconds two-way time (s TWT)) and distribution of the geological domains observed onshore (see text for details). See color version of this figure at back of this issue.



bides subduction zone [Mauffret et al., 1995; Gueguen et al., 1998]. Nevertheless, the extent of this volcanic activity onshore and offshore, and its relationship with the alkaline magmas recognized within some of the basins like the Tyrrhenian [Kastens et al., 1988; Serri et al., 1993], are poorly known, especially within the first basins formed during Oligocene-Miocene times, since they have undergone a strong subsidence.

[10] Carminati et al. [1998a, 1998b] propose that the opening of the western Mediterranean in the Oligo-Miocene is driven by the propagation of slab detachments as a consequence of continental collision in the Alps and in the Betic-cordillera, and that the Alpine collision locked the Corsica-Sardinia drift to the east. They consider a slab break-off to have occurred along the northern African margin [Maury et al., 2000], which led to the end of the subduction zone retreat on the plate and the “jump” of extension toward the Tyrrhenian area. Jolivet and Faccenna [2000] prefer to explain the synchronism of extension in the Mediterranean region 30–25 Myr ago by the sharp decrease in the absolute velocity of Africa. Determining the origin of slab retreat is beyond the scope of this study. We will merely consider here the sudden rollback of the slab at ~30 Ma as one of the main processes on which to build our tectonic regional interpretation. We note that the volcanic rock ages (31–24 Ma, 24–17 Ma, 17–11 Ma) recorded during the depocenter migration in the foreland/foredeep domain onshore Italy [e.g., Patacca et al., 1990; Van der Meulen et al., 1998], possibly because of the excess of detachment-related slab pull, correspond to ages recorded on volcanic rocks around the Ligurian domain and in Sardinia (31–30 Ma for Estérel Massif and SW Sardinia; 26–24 Ma in Nice area and central Sardinian rift; 20–16 Ma in SW Corsica and northern Sardinian rift; 12–11 Ma in the Gulf of Genova and NW Corsica [Coulon, 1977; Coulon et al., 1978; Bellon, 1981; Montigny et al., 1981; Rossi et al., 1998]). A detailed analysis of the volcanism in the Ligurian basin and its relationship with deep structures is therefore an important target for a better understanding of the early stages of opening in the western Mediterranean Sea.

2.3. Geological History Onshore the Ligurian and Corsican Margins

[11] We choose to refer to the Ligurian margin as the segment of the northern margin from Toulon to Genova, and the Corsican margin as the whole eastern margin from the Gulf of Genova to Bonifacio (Figure 3). Contrasting geological features are observed on the exposed parts of the Ligurian and Corsican margins. We briefly recall them here as they are important for an interpretation of the offshore structural units deduced from experimental seismology. Two main orogenic events have determined the structural pattern of both margins: the Hercynian and the Alpine orogenies [Réhault et al., 1984; Mauffret et al., 1995;

Gueguen et al., 1998]. The diversity of outcrops observed around the Ligurian basin emphasizes the important lateral variations in nature and fabric of the basement.

2.3.1. Ligurian Margin

[12] From west to east along the Ligurian coast, there are three major onland geological series.

1. The autochthonous metamorphic Hercynian basement of the Maures-Estérel Massif (Paleozoic) extends from west of Toulon up to Cannes. This basement is composed of Variscan metamorphic rocks intruded by Carboniferous granitoids. The Permian continental series surround the Maures Massif from Toulon up to Fréjus and unconformably overlie the Variscan basement. To the northeast of the Maures Massif a large Permian volcanic zone forms the present-day Estérel Massif [Rouire et al., 1980].

2. The autochthonous Meso-Cenozoic sedimentary cover is between Cannes and Nice.

3. The allochthonous outer alpine units are made of (1) the Castellane and Nice arcs, formed by a pile of Mesozoic and Tertiary nappes verging toward the north; these nappes are sliding on a major décollement above the crystalline basement [Fry, 1989; Laurent et al., 2000]; (2) the allochthonous Cenomanian-Maastrichtian Helminthoides Flysch, located between the outer and inner Alps; and (3) and the Schistes Lustrés nappe, including the ophiolites Massif of the Voltri region located near Genova.

2.3.2. Corsican Margin

[13] The onshore Corsican margin displays even more complex Cenozoic structures, with stacked nappes of oceanic material thrust upon the European margin and followed by an extension synchronous with rifting and spreading of the Ligurian basin [Jolivet et al., 1990; Fournier et al., 1991; Malavieille et al., 1998]. Three main units are observed [Durand Delga, 1978].

1. The allochthonous units of Alpine Corsica are mostly composed of two different stacks of ophiolitic units: the Schistes Lustrés nappe and the overlying Balagne-Nebbio nappe, which also thrusts upon the western autochthonous basement and Eocene sediments. The Balagne-Nebbio nappe is composed of Mesozoic unmetamorphosed rocks of “Ligurian” (Adriatic oceanic plate) affinity, whereas the Schistes Lustrés nappe is made of various Mesozoic metamorphic rocks including ophiolitic paleo-Ligurian remnants and eclogitic relics deformed under HP conditions;

2. The autochthonous Hercynian Corsica is mostly composed of an Hercynian basement (Carboniferous granites, Permian annular magmatic complexes) and pre-Hercynian rocks (Paleozoic series and metamorphic basement). This basement is affected by SW-NE late Hercynian faults, which later acted as sinistral strike-slip faults during the Alpine phase.

3. Allochthonous gneissic rocks represent two deformed parts of the European continental margin with contrasting styles of deformation: (1) a thrust unit recrystallized under medium-metamorphic conditions and slightly displaced

Figure 4. (opposite) Structural sketch map of the Ligurian domain after results from this study (see onshore geological domain captions on Figure 3). Continuous lines (numbered) refer to *Malis* cruise seismic profiles. Abbreviations are as follows: CV, central volcano; D, reflector “D” emergence (see text for details). See color version of this figure at back of this issue.

toward the west (the Tenda Massif), and (2) a supracrustal slice strongly deformed and metamorphosed under HP-LT conditions (the Sera di Pino and Farinole gneiss nappes) [see *Malavieille et al.*, 1998, and references therein].

[14] In Alpine Corsica, the exhumation (more than 50 km) of high-grade metamorphic rocks is believed to have occurred along a previously eastward dipping arc-continent subduction plane [*Malavieille et al.*, 1998], later followed by a cordillera-type metamorphic core complex denudation [*Jolivet et al.*, 1990, 1991; *Fournier et al.*, 1991; *Carmignani et al.*, 1995; *Daniel et al.*, 1996], in a single, westward dipping subduction context [*Brunet et al.*, 2000].

[15] According to $^{40}\text{Ar}/^{39}\text{Ar}$ dating, exhumation of eclogites, metamorphosed up to the blueschist facies conditions, has occurred between 45 and 38 Ma, whereas the late denudation phase took place between 30 and 12 Ma [*Brunet et al.*, 2000; *Jakni et al.*, 2000]. The progressive eastward migration of compression and extension evidenced in Alpine Corsica and north Tyrrhenian Sea has played a significant role in the thinning of the Corsican crust by reactivation of the thrust contacts as ductile normal faults [*Jolivet et al.*, 1990] between 33 and 22 Ma, and again eastward between 18 and 12 Ma, at the onset of the Tyrrhenian rifting [*Jolivet et al.*, 1990; *Van der Meulen et al.*, 1998; *Brunet et al.*, 2000, and references therein].

3. Data Acquisition and Processing

[16] This paper uses mainly multichannel (MCS) and single channel (SCS) seismic reflection profiles combined with a new set of MCS and wide-angle data recorded during the *Malis* cruise onboard the French R/V *Le Nadir* (IFREMER) in 1995. The seismic interpretations have been constrained by samples (*Marco*, 1995, and *Cylice*, 1997 cruises), reprocessed magnetic data (compilation of two aeromagnetic surveys and reduction to the pole) and altimetry data [*Sandwell et al.*, 1995]. The *Malis* cruise collected 37 MCS sections, totaling 3300 km in the Ligurian basin on both conjugate margins and in the deep basin (Figure 2). The acquisition system was composed of a 2.5-km-long streamer made of 96 channels, permitting a 24-fold coverage and a far offset of 3000 m, and by a source array of 10 air guns of GI-gun type (Generator-Injector), also called "pulse free" air guns [*Avedik et al.*, 1993, 1996]. The total array volume was 1935 cubic inches (31.7 L) and the signal-generating volume was 1140 cubic inches (18.7 L). Shot interval was set at 50 m, sample rate at 4 ms, and record length at 17 s. Seismic processing included spherical divergence, attenuation of multiple energy, normal move out (NMO) correction, stacking, FK-domain migration, depth-varying time domain filtering, and automatic gain control linked to water depth.

[17] When interpreting seismic reflection data we are facing two major problems. The first one is linked to the presence of a thick salt layer in the western Mediterranean deep basins, formed during the Messinian salinity crisis, at the time when the Mediterranean Sea was temporarily and repeatedly separated from the Atlantic Ocean [*Cita et al.*, 1978; *Ryan and Cita*, 1978; *Clauzon et al.*, 1996]. This salt layer creates an acoustic velocity inversion with depth which

does not favor the recording of acoustic waves from the layers just beneath it. Salt diapirs also generate large hyperbolas that strongly alter the sections even if time-migrated [e.g., *de Voogd et al.*, 1991]. The second problem is related to the time arrival of the first water column multiple from the seafloor, which comes close to the top of the acoustic basement reflector in the basin (at around 6.5–7 s two-way time (TWT)) and at shallower levels on the margins: therefore this additional signal interferes with the deepest reflections within the crust. Other encountered problems are (1) the very rough bathymetry which reflects the strong imprint of the Messinian desiccation [e.g., *Bellaiche*, 1993]; related deep canyons, which were avoided during MCS data recording, prevent us from an easy correlation between parallel lines transverse to the margins; and (2) the narrow and steep upper slope of the Corsican margins which has favored strong erosional processes and then contributed to the absence and/or scarcity of prerift, synrift, and even postrift sediments on large parts of the margin.

[18] In order to minimize problems arising from the salt layer and long-period multiple energy, we have taken particular care during the acquisition and processing phases. First, we have used a particular source combination made of a tuned air gun array with different volumes and depths in order to increase the acoustic energy and frequency content. All air guns were synchronized by using the "single bubble" pulse generating method, in order to exploit the strong energy at low frequency contained in the first bubble pulse [*Avedik et al.*, 1993, 1996]. This method has allowed to reach a good seismic penetration within deep layers, down to the top of the basement of the basin and the margins, and sometimes down to the Moho (around 8–10 s TWT). Second, a particular prestack processing method was applied on the data which attenuates the seabed multiple and some multiples of the uppermost sediment layers. This method, based on the TPMUL method of the Geovecteur SoftwareTM (*Compagnie Générale de Géophysique*, France), is designed for computing a model of multiple events and to subtract it from the input gathers. The program automatically picks up the traces for which the residual curve after normal move out (NMO) is above an upper bound imposed by the user, and it deletes them. At a given depth, the curves are characterized by a time difference referring to the maximum offset. This method was selected among several others, as it was particularly effective for eliminating the seafloor multiple energy in the deep basin and allows us to better identify the top of the basement in the whole basin and also locally until 10 s TWT under the margins.

4. Identification of the Various Domains in the Ligurian Basin

[19] The *Malis* MCS reflection profiles have been cross-correlated with other older MCS profiles, i.e., ~3500 km of seismic sections called MS (OGS Trieste), Augusta, Odysee, J (Institut Français du Pétrole (IFP)) [see *Réhault*, 1981; *Réhault et al.*, 1984; *Gennesseaux et al.*, 1989], and SCS profiles acquired on the upper margins [e.g., *Réhault*, 1981; *Bellaiche*, 1993; *Gaullier*, 1993;

Chaumillon et al., 1994]. Using all seismic data together, we attempt to draw the structure of the basement of the Ligurian basin and the spatial distribution of the main acoustic units below the sedimentary pile with as many details as possible.

[20] We first propose a new and updated map of the top of the prerift acoustic basement by interpolating the isochrons from one profile to another, using all the MCS reflection profiles available in the region (Figures 2 and 3). The top of the prerift sediments on the margins and the top of the acoustic basement in the deep basin have been systematically digitized. We have also tried to recognize the offshore continuation of the major geological domains mapped onshore and around the basin. This new map improves the accuracy on the morphology and nature of the acoustic basement compared to previous maps [*Réhault*, 1981; *Burrus*, 1984]. Moreover, on the basis of the interpretation of the whole data set available in the basin, including magnetic and gravity anomalies, we propose new boundaries for the three domains recognized in the acoustic basement in the Ligurian basin, namely, (1) the continental thinned margins, (2) the transitional domain, and (3) the central domain (Figure 4). We rely mostly on acoustic facies changes evidenced from our interpretation of the homogenous and deeply penetrative (at least down to the acoustic basement surface) *Malis* MCS data set.

4.1. Thinned Continental Margins

[21] Following previous authors [*Pascal et al.*, 1993; *Mauffret et al.*, 1995], the continental thinned conjugate margins are limited oceanward by the slope break of the basement acoustic facies. This domain is structured in grabens and a few half grabens.

[22] We identify offshore two major domains which are also recognized onshore (Figure 3): (1) the Hercynian basement and its sedimentary cover located in the Maures-Estérel Massif on the northern margin and in the western part of the Corsican margin, and (2) the Alpine nappes (Mesozoic cover, Eocene Flysch, ophiolites and Schistes Lustrés) located east of Antibes on the northeastern part of the Ligurian margin, and east of Ile Rousse on the eastern part of the Corsican margin. Changes of synrift structural trends in the prerift acoustic basement correlate quite well with changes in the geology of the outcrops observed along the coast (Figure 4): This suggests that the preexisting structures have influenced the mode of deformation of the acoustic basement during rifting. The morphology of the top of the acoustic basement and of the deeper reflectors also provides some indications on the width of the margins after rifting and on their detailed structure, from 0 to 6 s TWT, if we neglect at first order possible postrift tectonic deformation. The morphology of the prerift acoustic basement on the Corsican margin is better preserved than on the conjugate margin, where the active Messinian erosion event has strongly leveled the prerift and synrift structures. Nevertheless, on both margins, it is quite clear that the width of the margins is asymmetrical. Indeed, the Ligurian margin is about 40–50 km wide from northeast to southwest, in contrast with the

Corsican margin which is generally wider (~70–80 km width on NW and SW Corsican margins).

4.2. Transitional Domain

[23] The limits and nature of the so-called “transitional domain” (TD) identified between the continental margins and the deep basin have been revised. The existence of a TD in the Ligurian basin was proposed by *Gueguen* [1995], who has mostly based its interpretation on the correlation between the change in magnetic patterns [*Bayer et al.*, 1973; *Mauffret et al.*, 1995] and the Messinian salt pinch at the foot of the margins. We revise the position of the oceanward limit of the TD by combining two criteria: (1) a change in magnetic field expression between the quiet magnetic zone observed around the deep basin, corresponding to negative magnetic anomalies (up to 50 nT), and the disturbed magnetic field in the deep basin; and (2) a change in the acoustic basement facies systematically observed on seismic reflection data (Figures 5a and 5b). We have preferentially used this latter criterion, except where seismic data are absent or too sparse (like on the SW Ligurian margin). On the *Malis* cruise data, the TD along the Ligurian margin is characterized by a group of strong reflectors at the top of the acoustic basement slightly deepening toward the continent (Figure 5a), which generally represents the deepest seismic reflectors. On the Corsican margin, and especially on its northwestern part, the TD depicts a seismic facies typical of volcanic flows associated with large, circular magnetic anomalies (see section 6.2). We assume that the seismic facies changes between the TD and the deep basin (Figures 5a and 5b) likely indicate the transition to a “true” ocean-like domain, since they can be attributed to different genetic and evolutionary processes of the crust. Compared to previous studies, our TD oceanward limit is rather similar to the one proposed by *Réhault* [1981] and *Le Douaran et al.* [1984] on the Ligurian margin, except in NW Corsica where this limit is shifted ~20 km oceanward from *Réhault*'s [1981] one. Compared to previous studies, we therefore propose a slightly narrower oceanic domain in the northeastern half of the Ligurian basin, and a wider NW Corsican margin.

4.3. Central Domain

[24] Although deep drilling data are lacking, the central Ligurian basin is considered as an oceanic-type domain by most authors [*Réhault*, 1981; *Burrus*, 1984; *Le Douaran et al.*, 1984; *Le Cann*, 1987; *Pasquale et al.*, 1994; *Gueguen*, 1995; *Mauffret et al.*, 1995; *Contrucci*, 1998; *Gueguen et al.*, 1998; *Contrucci et al.*, 2001], but with varying characteristics and shapes. These interpretations differ on the nature of the crust, which is poorly constrained and seems to be atypical according to geophysical data [e.g., *Mauffret et al.*, 1995; *Chamot-Rooke et al.*, 1999]. Indeed, the crust is characterized by much thinner than average thickness (<4 km) and ambiguous magnetic anomalies (too complex and discontinuous to correlate with isochrons) [see, e.g., *Bayer et al.*, 1973; *Burrus*, 1984]. We have based our interpretation of the atypical oceanic domain on the presence of a particular acoustic facies (Figure 5b) formed by basement undulations (~5 km length and more than 1 s TWT amplitude) often covered by diffractions which are similar to

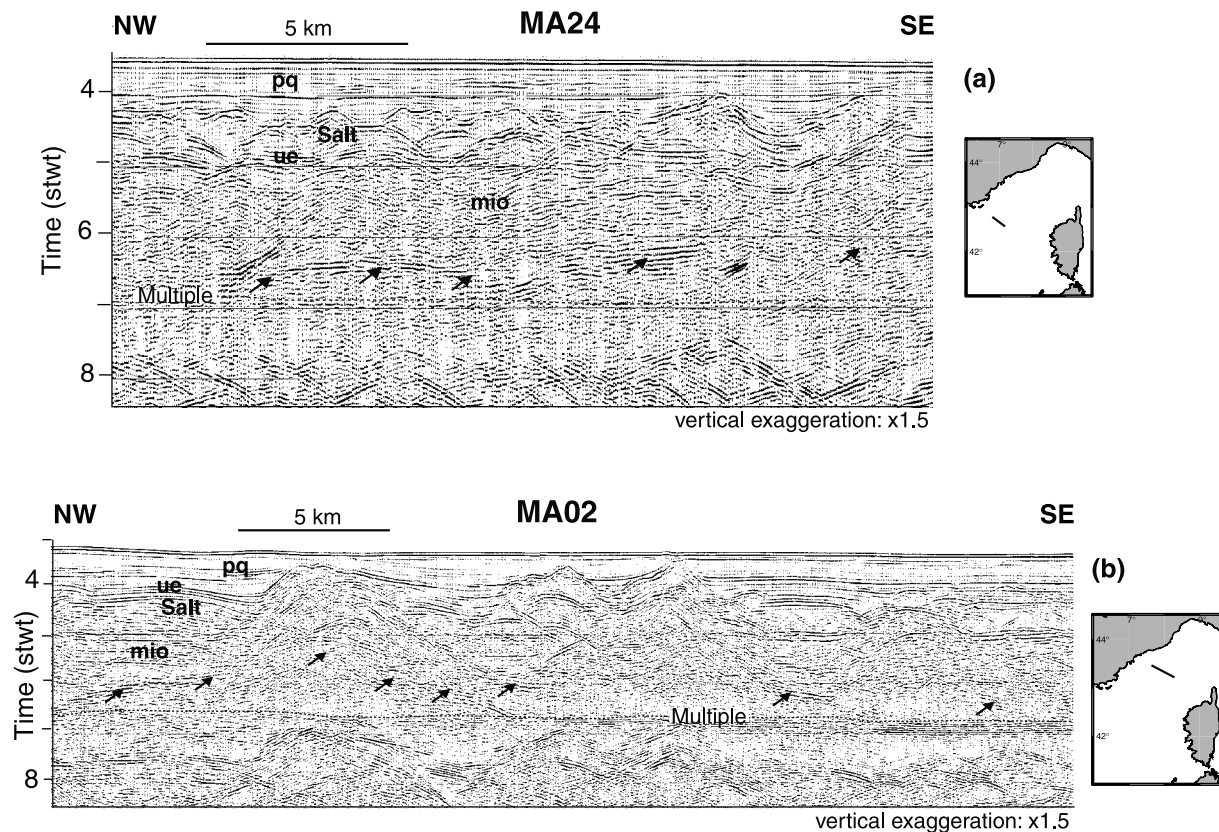


Figure 5. Typical facies of the acoustic basement: (a) in the transitional domain (TD), and (b) in the NW deep basin. In the TD we interpret the reflective top of the acoustic basement (arrows) as volcanic flows sourced from nearby lava centers. In the deep basin, we interpret undulations of the basement (arrows) as discontinuous volcanoes and dykes intruded in a disrupted crust (see text for details). Insets locate the *Malis* lines shown. Abbreviations are as follows: stwt, seconds two-way travel time; pq, Plio-Quaternary sediments; ue, Upper Evaporites (late Messinian at ~ 5.3 Ma); mio, Miocene sediments.

seismic facies identified in the Tyrrhenian Sea [Pepe *et al.*, 2000] and calibrated with Ocean Drilling Program (ODP) site 650 well data [ODP Leg 107 Shipboard Scientific Party, 1987]. This acoustic basement is overlain by sediments which form several small angular unconformities due to basement irregularities. This observation attests that these undulations are not the only consequence of pull-up effects due to thickness variations of the salt layer. Finally, we also note from the morphology of the deep basin that no clear axial ridge is present near the center of the basin. Instead, we observe only several small discontinuous ridges (up to 20 km across) trending N030–040 (Figure 3).

5. Segmentation and Style of Deformation on Conjugate Margins

5.1. Mapping the Margin Segmentation

[25] At large scale, prerift rheology and geological inheritance are known to strongly influence back arc extension, as

shown for instance in the central Mediterranean [e.g., Facenna *et al.*, 1997]. We attempt here to determine the role of inheritance and rheology in the case of the Ligurian basin. We call hereinafter “segment” a part of the margins characterized by the same trend and width on a significant distance.

[26] Despite the Messinian erosion event that leveled much of the prerift and synrift structures [e.g., Réhault *et al.*, 1984], we can recognize three major segments along the Ligurian margin (Figures 3 and 4): (1) one from Toulon to longitude 6.7°E , trending roughly E-W, poorly imaged because of the scarcity of seismic data, (2) one from longitude 6.7°E to Antibes, trending N30–40°, relatively wide and smooth, averaging 50 km width, and (3) one from Antibes to Albenga, trending 50°N , narrower and with an abrupt slope, averaging 40 km width. The two southern segments limit offshore the metamorphic Hercynian Maures-Estérel Massif, whereas the northern one is the offshore prolongation of the Alpine nappes. On the Corsican margin (Figures 3 and 4) [see Guennoc *et al.*, 1998], we also observe a clear segmentation from north to south into

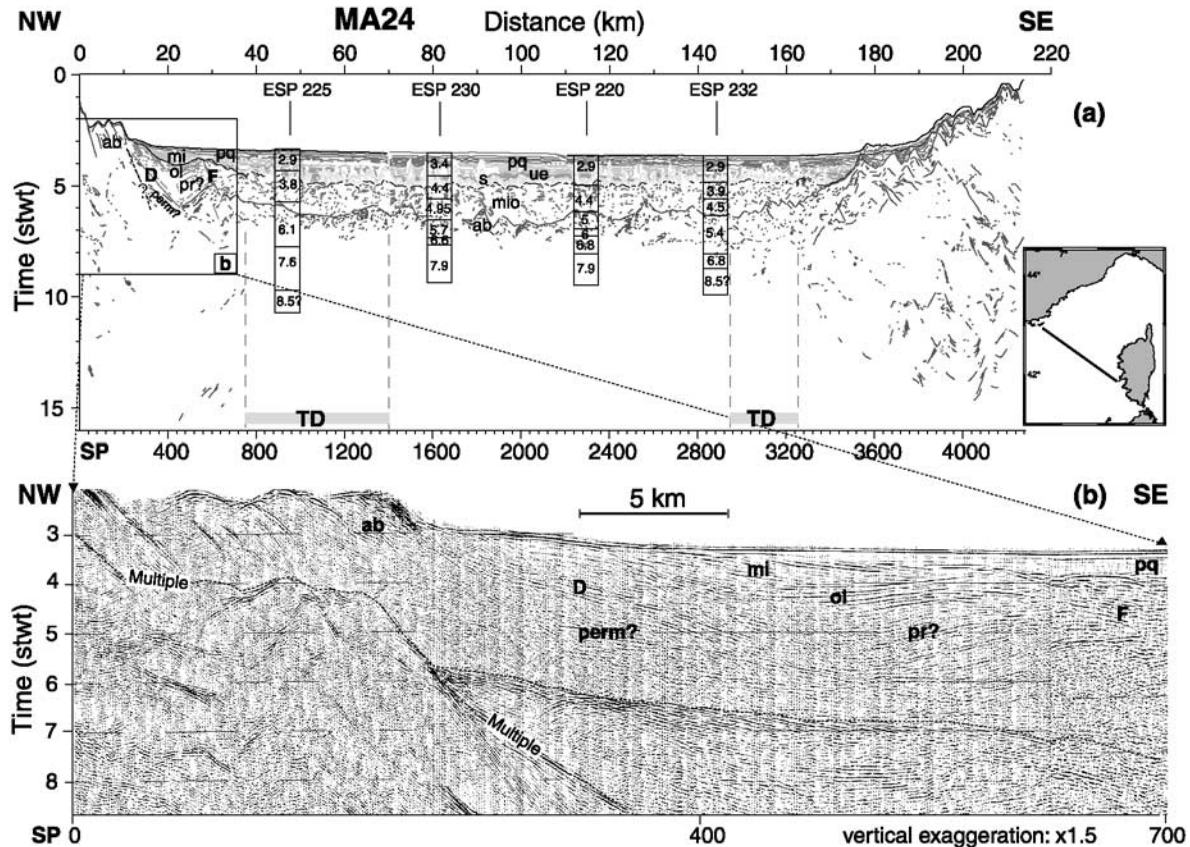


Figure 6. Multichannel seismic reflection line MA24, South Ligurian basin (location inset, right top). (a) Interpretation with ESP velocity logs from *Le Douaran et al.* [1984]; velocities in km/s. (b) Migrated seismic section of inset (inset “b” in Figure 6a). Abbreviations are as follows: D, strong reflector at the top of the acoustic basement dipping toward the basin offshore the Maures Massif; F, normal fault; ab, acoustic basement; perm, Permian deposits; pr, prerift sediments; ol, Oligocene sediments; mi, infrasalt Miocene or Messinian slope sediments; pq, Plio-Quaternary sediments; ue, Upper Evaporites (late Messinian at ~ 5.3 Ma); s, salt layer; mio, Miocene sediments; TD, transitional domain; SP, Shot Points numbers; and stwt, seconds two-way travel time. Dashed line and bold question mark on inset “b” in Figure 6a show the assumed crystalline basement top.

three main parts: (1) the first one, trending roughly N-S to N40°E in the Gulf of Genova up to Cape Corse-Calvi, is very wide (~ 80 km) and corresponds to the offshore continuation of Alpine nappes; (2) the second segment, on the western Corsican margin between Calvi and Ajaccio, is the narrowest (~ 40 km) and marks the prolongation of the Hercynian metamorphic basement observed onshore; (3) the third segment trends roughly N160° along the southwestern Corsican margin, and becomes wider (~ 80 km) from latitudes 42°00'N to 41°20'N. It may be related to the offshore prolongation of the Sardinian rift toward the north, which developed in the zone of unthickened crust reactivated at the end of the Ligurian opening phase and during a strong postrift magmatic event [*Coulon, 1977; Montigny et al., 1981; Gennesseaux et al., 1989; Carmignani et al., 1995; Rossi et al., 1998*]. Therefore this segmentation appears clearly correlated to onshore geological changes (see section 2.3) and thus suggests a clear control of inheritance and rheology on the morphological segmenta-

tion of the margins (Figure 3) when compared to other factors like differential erosion or tectonic flexure [*Chauvillan et al., 1994*].

5.2. Control of Multiphase Inheritance on Extensional Structures

[27] The crustal structure imaged on MCS lines, in good agreement with the morphological segmentation, confirms that the style of extensional deformation is variable along the margins and between the conjugate margins, and that it is directly dependent on the Hercynian or Alpine inheritance. We select one representative profile on each conjugate margin in order to depict the structures involved at depth in each case.

5.2.1. Ligurian Margin

[28] On the Ligurian margin, offshore the Maures Massif, the southward prolongation of the Hercynian basement outcrops on the NW part of line MA24 (Figure 6). The

top of the acoustic basement (“ab”) is underlined by a strong reflector (D) (between shot points 230 and 600, Figure 6) slightly deepening toward the basin ($\sim 30^\circ$ of apparent dip), down to 6.5 s TWT. Below reflector (D) we observe a series of continuous reflectors which deepens toward the basin ($\sim 16^\circ$ dip) and top laps below it. We interpret these reflectors as Permian and/or Mesozoic series equivalent to the Permian basins and their Mesozoic cover which surround the Maures Massif onshore. If this hypothesis is correct, this sedimentary cover must be separated from the diffracting crystalline basement by an unconformity or paleo-topography (dashed line on Figure 6a) which is not visible on profile MA24; note basement diffractions even on the migrated section in Figure 6b). On the basis of these observations and the correlation with other seismic lines, we suggest that the onshore Permian basin surrounds the whole Maures Massif and reaches its southern limit on the Ligurian margin. This hypothesis is consistent with the absence of Permian sediments in Corsica, on the conjugate margin. Above and along this reflector (D), a sedimentary cover (“ol/pr”), ~ 4000 m thick (assuming an acoustic velocity of 4 km/s), is tilted toward the continent. We suggest that this series is composed of slightly deformed prerift sedimentary cover of the Maures Massif and/or by thick synrift Oligocene sediments, which would be thicker in the central part of the rift than at its end, near Saint-Tropez. Indeed, upper sediments (“ol”) have comparable acoustic facies to those observed northward of the MA24 profile, in the Stoechades canyon, where Oligo-Aquitainian deposits (from 30–28 to 21 Ma), made of interlayered conglomerates and sandstones, have been identified by submersible with a thickness of at least 700 m [Bellaiche *et al.*, 1976; Angelier *et al.*, 1977]. On our section, the sedimentary pile (ol/pr) is bounded by two normal faults (D and F on Figure 6) dipping toward the basin (30° and 40° of apparent dip for D and F, respectively), and displays a rollover geometry above reflector D. We show that this reflector D, which is clearly imaged on several lines from Toulon to longitude 6.5°E , forms a continuous, nearly E-W trending fault along the margin (Figure 4). This trend is subparallel to the structural pattern observed onshore in the Maures Massif [Rouire *et al.*, 1980] (Figure 3), where faults affect the basement and have a multiphase history inherited from Stephanian time. Therefore we suggest that both onshore and offshore fault networks were inherited from the same multiphase tectonic history, and that offshore structures have been reactivated during the Oligocene opening of the Ligurian basin.

[29] This interpretation is supported by comparable structural features geometry observed offshore and onshore the Gulf of Lion [Gorini *et al.*, 1993; Guennoc *et al.*, 1994; Mauffret *et al.*, 1995; Benedicto *et al.*, 1996; Séranne, 1999]. Offshore, it is interpreted as a décollement associated with reactivated Pyrenean thrusts. Onshore, rollover geometries are observed in half grabens filled by 4000-m-thick Oligo-Aquitainian sediments, in and northward of the Vistrenque graben, along listric faults flattening at depth in southeastward dipping crustal ramps with low dip (25°). They are interpreted as synrift extensional features either

formed by inversion tectonics of inherited Pyrenean thrusts, or newly formed and controlled by crustal weakening associated with the previous Pyrenean thickening.

5.2.2. Corsican Margin

[30] On the Corsican margin, near the NW Cape Corse, the influence of Alpine tectonics in the structural pattern is clearly observed. Profile MA02 (Figure 7), trending perpendicular to the margin, depicts the acoustic basement (“ab”) slightly deepening toward the basin, from 4 to 6.5 s TWT, forming collapsed blocks of around 8 to 15 km width. The nature of this acoustic basement has been identified by sampling during submersible dives (*Cylice* cruise) [Sosson and Guennoc, 1997; Sosson *et al.*, 1998] on the outcropping summit of the Glangeaud Seamount located slightly southward of MA02 line (Figure 4). The pegmatitic granite sampled here indicates that the Glangeaud Seamount is the offshore prolongation of the Hercynian Corsican crystalline basement of the Tenda Massif. Wide-angle seismic data acquired in the region along the European GeoTraverse (EGT) profile [Egger *et al.*, 1988] (location on Figures 2 and 7, inset), provide some consistent basement velocities of 6.2–6.3 km/s.

[31] At the foot of the margin, the top of the acoustic basement is underlined by strong reflectors (c) gently dipping toward the continent with an apparent dip of 30° . These reflectors sole into a group of strong reflectors (d), subhorizontal and subparallel, between 6 and 7 s TWT, located between shot points 2900 and 3100 on Figure 7. They are imaged on seismic lines along the whole NW Corsican margin (from Porto to the northern Cape Corse, Figure 4), from the foot of the slope down to the transitional domain boundary. By analogy with onland features, we tentatively interpret reflectors (c) as a thrust at the base of units related to the Alpine orogeny and reflectors (d) as a décollement (or shear zone) within the Alpine nappes. We have no way to know whether reflectors (d) mark the limit between allochthonous and autochthonous rocks. As the basement is shifted and deepens toward the basin along these features, we propose that they were reactivated as normal faults during the Ligurian basin opening. Indeed, such reactivated features are described onland, on the eastern flank of the Tenda Massif, where normal ductile shearing reactivation occurred in the 32–25 Ma time interval along the east Tenda shear zone [Brunet *et al.*, 2000], which was a former Alpine thrust [Jolivet *et al.*, 1990, 1991; Daniel *et al.*, 1996]. An offshore analogue is described in the northern Tyrrhenian Sea, where comparable tectonic features are interpreted as related to the denudation of nappes associated with back arc exhumation [Jolivet *et al.*, 1998].

[32] In the upper margin, the acoustic facies (n), up to 1.5 s TWT thick, is formed by layered reflectors slightly dipping toward the continent. According to EGT data, velocities range in this series from 2.2 to 4.8 km/s [Egger *et al.*, 1988]. Both sediment series (n) and the underlying acoustic basement are affected by NE-SW trending normal faults dipping toward the basin with a 60° apparent dip (Figure 8). We interpret the series (n) as the offshore prolongation of the upper Alpine nappes which cover the acoustic basement onland. There, the Nebbio Nappe (upper unit like the

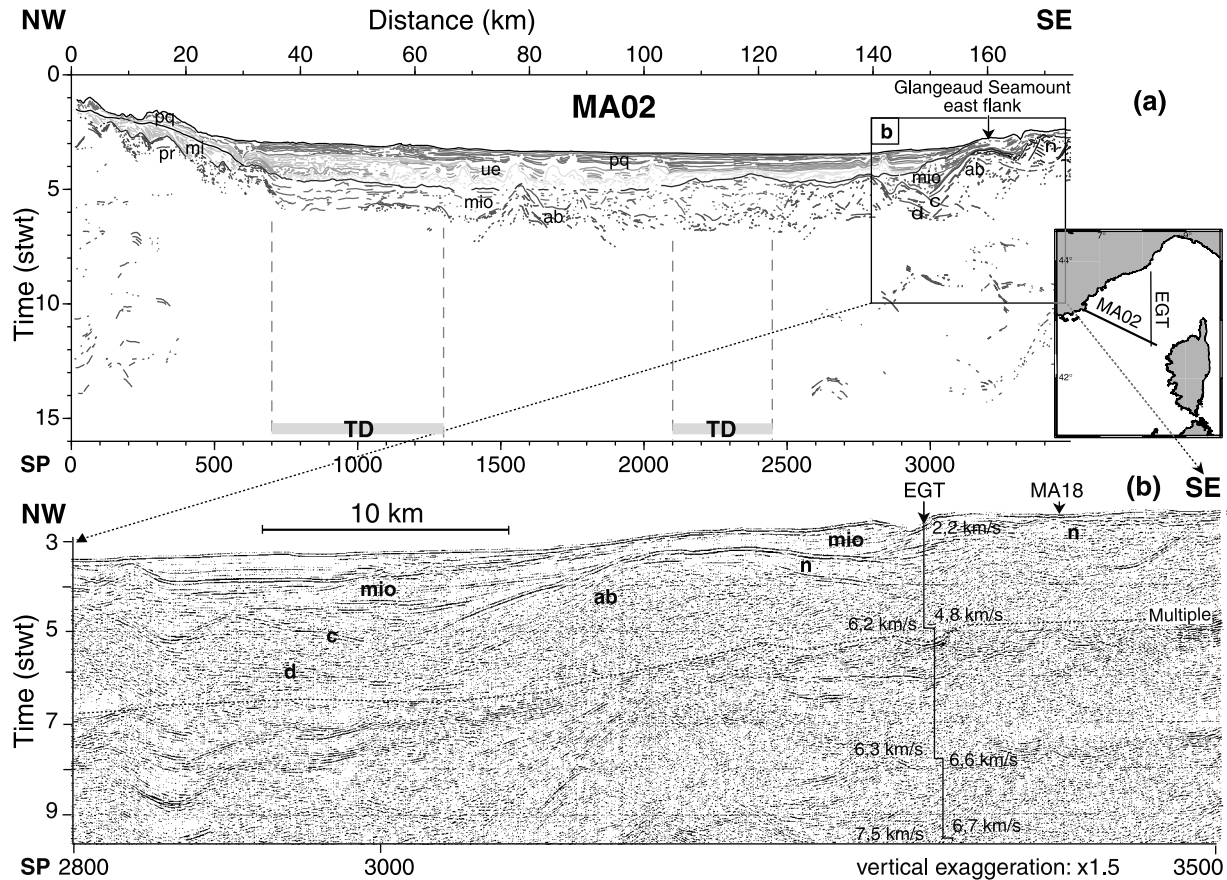


Figure 7. Multichannel seismic reflection line MA02, northern Ligurian basin (location on inset, top right). (a) Profile interpretation; (b) Migrated seismic section of inset “b” shown in Figure 7a. Abbreviations are as follows: ab, acoustic basement of the Glangeaud-Tenda unit; n, Alpine thrust nappes; c, reflectors “c”; d, reflectors “d” (see text). The European GeoTraverse (EGT) velocity determination from Egger *et al.* [1988] and the location of MCS line MA18 are shown. Other abbreviations are as in Figure 6.

Balagne nappe) overthrusts the Schistes Lustrés nappe (in a narrow sense), and these two Alpine units are unconformably overlain by the Miocene sediments (17–11 Ma) of the St. Florent basin [Ferrandini *et al.*, 1998]. This structural setting is similar to the one observed in the acoustic units in the upper margin. Our reflection seismic interpretation (Figure 8) suggests that normal faulting took place after the Alpine nappes development in the upper Eocene and before Miocene sedimentation. Therefore these faults are coeval with the Oligocene reactivation of former Alpine thrusts (reflectors (c) and east Tenda shear zone). Offshore central and southern Corsica, few steep fault scarps are shifting the top of the margin basement down the slope (Figures 4 and 6), and several transverse faults are in the prolongation of Hercynian structures onshore (Figure 4) which may have been reactivated as synrift transfer faults.

5.3. Asymmetrical Structure of Conjugate Margins

[33] The deepest structure (lower crust) of the conjugate margins is poorly documented by the available seismic

data, as wide-angle data are scarce, and MCS data provide only locally information down to the Moho. However, we have identified the main gross scale structural features of both margins down to midcrustal levels and have cross-checked them.

[34] The Ligurian margin is narrow (around 40 km wide), mainly formed by two major tilted blocks (15–20 km wide) with locally, offshore Imperia, a third smaller block (Figure 4). Because of Messinian erosion, the prerift geometry is poorly preserved on this margin (Figure 9). Indeed, the Messinian erosion surface (MS) leveled the upper margin and is clearly connected to the base of the salt (s) layer in the basin (Figure 9). The structural trend of the two main crustal blocks is subparallel to the main trend of the whole margin and varies from southwest to northeast (Figure 4), depending upon the tectonic inheritance: the Stephano-Permian tectonic phase to the southwest offshore the Maures-Estérel Massif, and the Alpine tectonic phase to the northeast. The conjugate west Corsican margin is wider at its northern and southern ends (70–80 km wide), where it is composed of numerous small

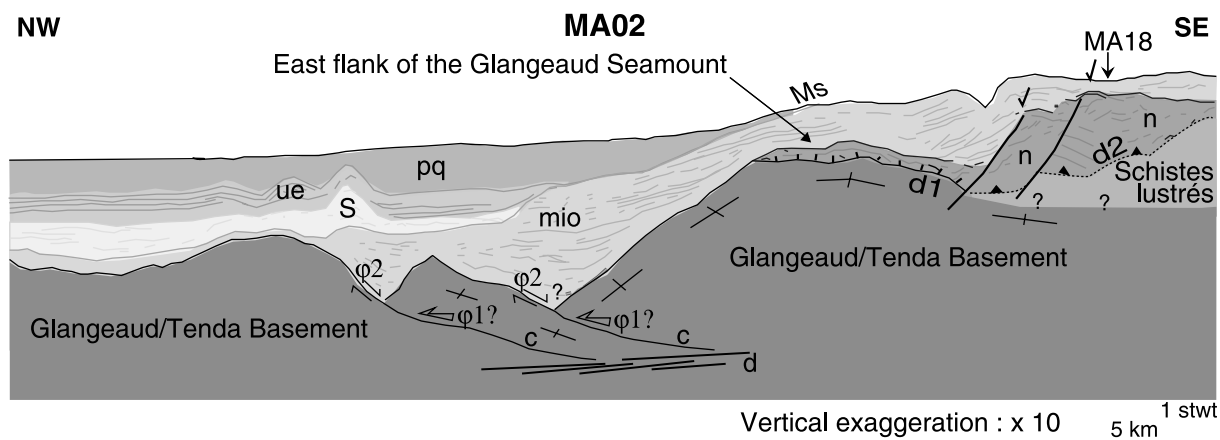


Figure 8. Interpretation of the northwestern Corsican margin on line MA02 (see Figure 7 for location). Solid arrows locate two synrift faults shifting the Nebbio nappes (n) in the upper margin. Abbreviations are the same as in Figure 6. Labels are as follows: $\varphi 1$, HP-LT Alpine compression; $\varphi 2$, later extension that started in Oligocene. The d1 and d2 reflectors are interpreted as Eocene thrust contacts at the bottom of the Upper Alpine nappes, overlying the granitic basement and the Schistes Lustrés nappe (in a narrow sense), respectively [Jolivet *et al.*, 1991; Daniel *et al.*, 1996].

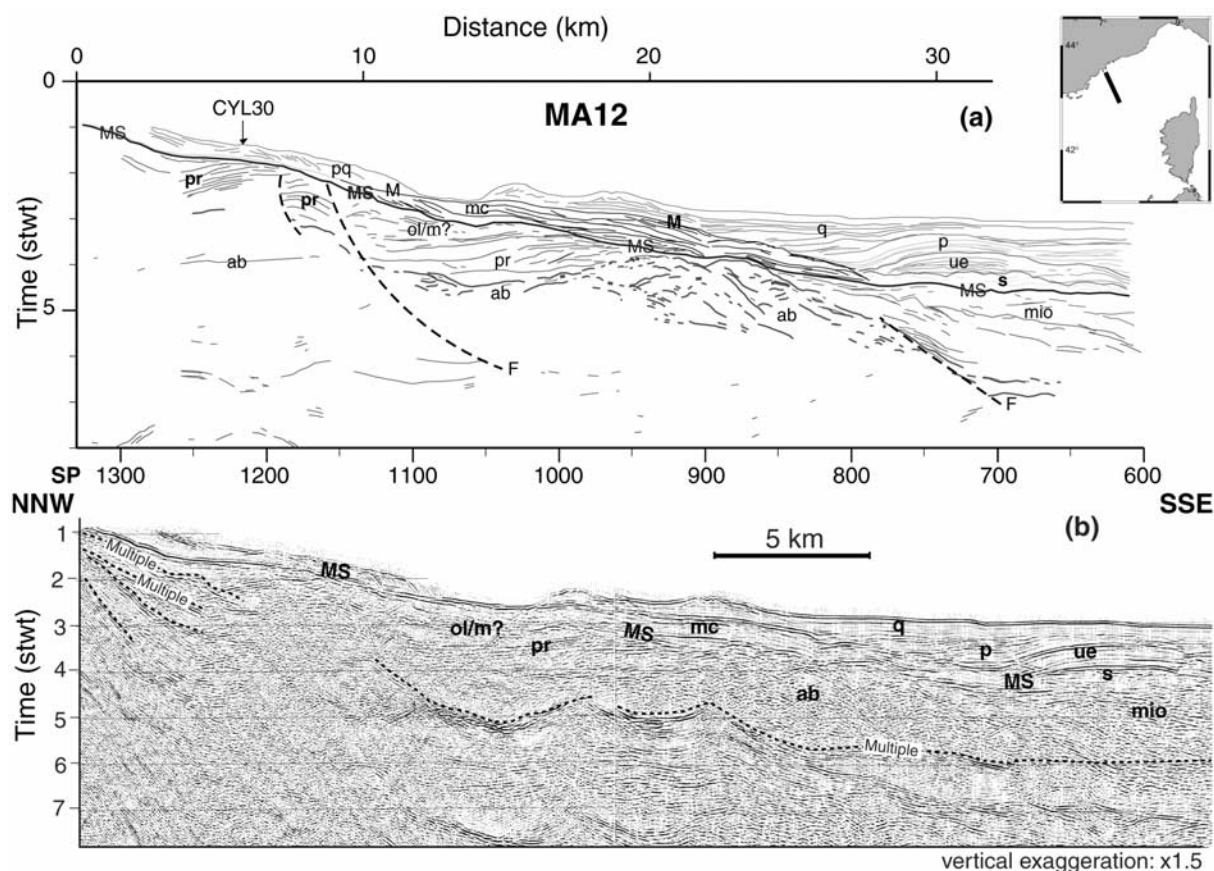


Figure 9. Multichannel seismic reflection line MA12 representative of the structure and geometry of the Ligurian margin. (a) Interpretation (MS, Messinian erosional surface; ol/m, Oligocene and/or Miocene sediments). “CYL 30” localizes *Cylice* cruise submersible dive number 30 [Sossou and Guennoc, 1997]. “F” dashed lines, assumed faults limiting the late Oligocene tilted blocks of the margin. See Figure 6 for other abbreviations. (b) Migrated *Malis* multichannel seismic reflection line corresponding to the line drawing in Figure 9a.

blocks (three to seven blocks between 10 and 17 km wide; see Figures 6 and 7). The structural style on this margin is influenced by two different tectonic patterns: At the NE end, the Alpine inherited structures extend across the entire margin, whereas at the SW end, the northward continuation of the N-S Sardinian rift localized the extension in a graben at the foot of the margin (Figure 4).

[35] In the sedimentary cover, the synrift sediment (ol/m) thickness decreases progressively toward the NE on the Ligurian margin. Synrift sediments are finally absent at the NE end of profile MA02 (Figure 7) and farther north. On the Corsican margin, no synrift deposits have been identified, except for some limited Oligocene sediments discovered onshore [Ferrandini *et al.*, 1999], but Miocene sediments are suboutcropping (Figures 7 and 8). This deposition asymmetry depicts a different subsidence evolution and/or a differential erosion history on both conjugate margins, and a more recent differential uplift since 10 Ma [Chaumillon *et al.*, 1994; Bigot-Cormier *et al.*, 2000; Jakni *et al.*, 2000]. From north to south, the top of the basement remains in a higher position further into the basin on the Corsican margin than on the Ligurian margin (Figure 6). Note that part of this difference in elevation could also be related to the recent differential uplift between the conjugate margins [Réhault, 1981; Chaumillon *et al.*, 1994; Bigot-Cormier *et al.*, 2000; Jakni *et al.*, 2000].

6. Nature and Evolution of Magmatic Events

[36] From our MCS data interpretation constrained by sampling, we have performed a systematic identification of volcanic or volcanoclastic bodies in the basin and the margins. Although more data are needed to accurately determine their extent at depth, we attempt here to gather information from MCS seismic sections, magnetic anomalies, and sampling on the seafloor, in order to roughly assess the geometry and extent of volcanic bodies and plutonic units in the Ligurian basin, which have been only scarcely explored previously.

6.1. Cross Correlation of Seismic Data

[37] Two seismic lines from *Malis* cruise offer the opportunity to cross-check our MCS interpretations with wide-angle data down to the Moho: profile MA24 in the SW part of the Ligurian basin (ESP data from *Le Douaran et al.* [1984]; Figure 6), and profile MA29 in the Gulf of Genova (wide-angle data from *Egger et al.* [1988] and *Makris et al.* [1999]; Figure 10). In both places, we note the presence of a high-velocity layer (HVL) at the base of the crust (7.5–7.7 km/s), located under peculiar seismic facies. In the Gulf of Genova, where the HVL is well constrained by three velocity logs, the HVL is located at depth under the Monte Doria volcano. The Monte Doria Seamount, on top of which basalts 11–12 Ma old were sampled by dredges and submersible dives (K/Ar method) [Sossion and Guennoc, 1997; Réhault *et al.*, 1998], depicts the representative acoustic facies of the volcanic and volcanoclastic units we have identified on the lines, which is made of short and wavy reflectors (Figure 10). This acoustic facies is charac-

terized by ovoid bodies, found in the TD, the central domain, and on the deep SW Corsican margin (Figure 4).

[38] Several authors have previously identified a HVL along the Ligurian margin from the Gulf of Lion to the Gulf of Genova [*Le Douaran et al.*, 1984; *Egger et al.*, 1988; *Pascal et al.*, 1993; *Contrucci*, 1998; *Chamot-Rooke et al.*, 1999; *Rollet*, 1999; *Contrucci et al.*, 2001]. Conversely, it has been found only locally on the Corsican margin [*Contrucci et al.*, 2001], but only few wide-angle data exist there. The HVL acoustic velocities range between 7.1 to 7.6 km/s. They correspond generally to 2 to 3 km thick lenticular layers located at the base of the crust under the foot of the margin and in the TD. They have been interpreted either as deep oceanic crust [*Le Douaran et al.*, 1984], serpentinized mantle [*Mauffret et al.*, 1995], or underplated material [*Makris et al.*, 1999]. The data set available in the Liguro-Provençal basin does not allow us yet to discriminate between these hypotheses.

6.2. Magnetic Anomalies Against Seismic Data

[39] The new map of magnetic anomalies reduced to the pole produced here (Figure 11) offers the opportunity to check our seismic interpretation and to increase the spatial coverage of magmatic fields. This map also allows for a completion toward the west of the map recently published by *Chiappini et al.* [2000] and for a better control on the overall stability of the magnetic anomaly pattern published more than 20 years ago [*Bayer et al.*, 1973; *Galdéano and Rossignol*, 1977]. Indeed, the pole reduction applied is likely to provide a more accurate location of magnetic anomalies. The superimposition of the volcanic acoustic facies determined (Figure 11, thick red lines) and the magnetic anomalies shows a fairly good fit. Volcanism identified on seismic profiles are generally above the largest positive magnetic anomalies, not only on the margins, but also (and mostly) in the transitional domain and in the deep basin (Figure 4). Only the Monte Doria Massif is slightly shifted northward compared to a moderate high-anomaly zone, whereas a large ovoid anomaly west of Cape Corse seems not related to any volcanism identified on seismic lines. We may distinguish two types of magnetic signatures: High positive magnetic anomalies are dominant on the margins and within the transitional domain, whereas low positive anomalies form two narrow, discontinuous lineaments subparallel to the basin axis in the deep basin. We propose that the first signature represents volcanic flows sourced from nearby lava centers located on the transitional domain or in the deep basin (section 4). Concerning the second magnetic signature, the nature of the assumed volcanism in the deep basin is unknown since it is hidden under 7–8 km of sediments and was never drilled. In this axial domain the relation between basement reliefs and magnetic anomalies is not simple and obvious. The sources of most magnetic anomalies are therefore to be searched into intrabasement magnetic bodies. Nevertheless, by analogy with the Tyrrhenian Sea we can hypothesize that the axial volcanism has a tholeiitic affinity and forms a thin basaltic crust overlying mantle rocks, since both types of rocks have been identified there in several boreholes (ODP Leg 107) [see, e.g., *Masclé*

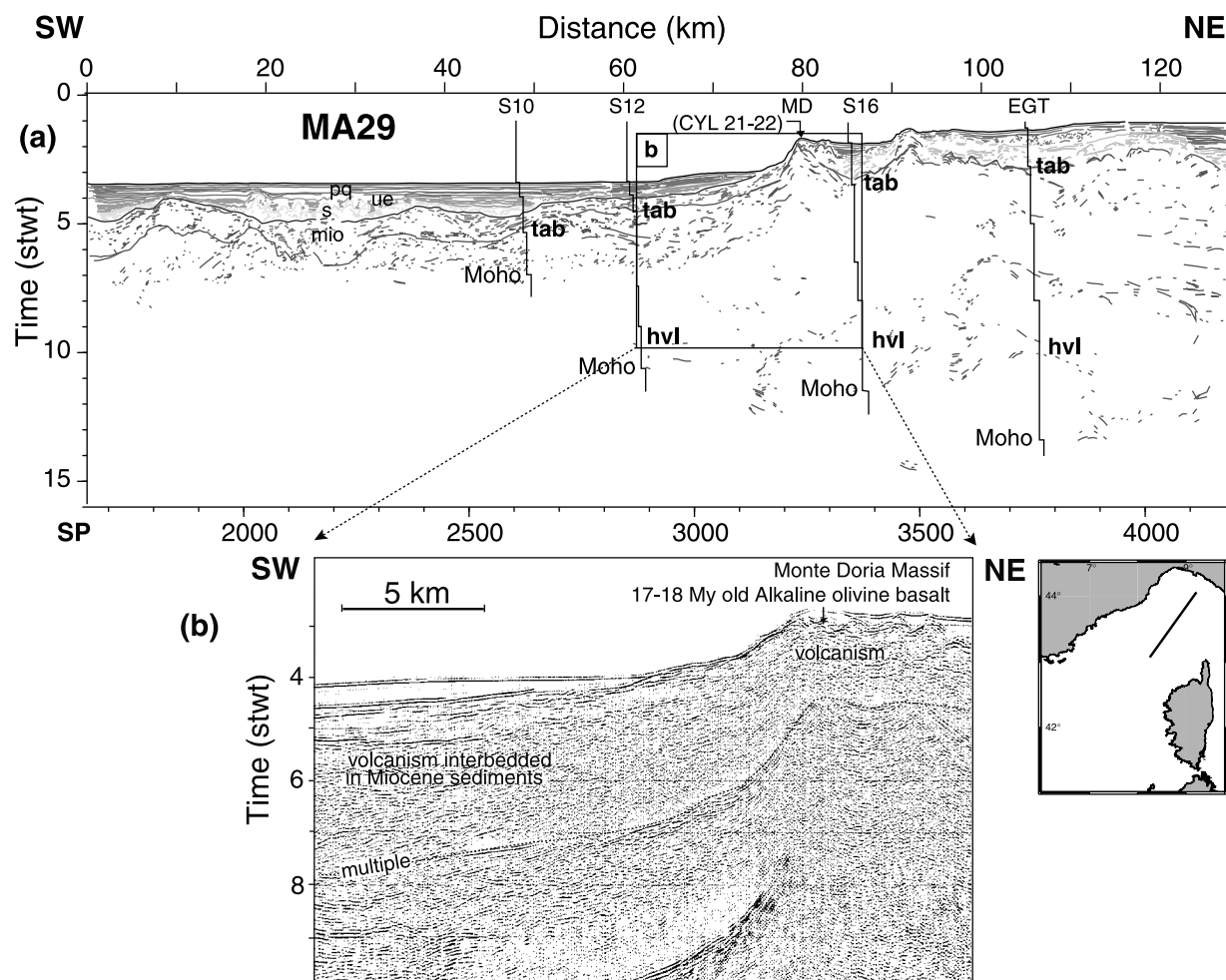


Figure 10. Multichannel seismic Line MA29 in the northern Ligurian basin. (a) Interpretation with three wide-angle velocity profiles (“S” profiles) [Makris *et al.*, 1999] and EGT profile [Egger *et al.*, 1988] converted in seconds two-way time. Monte Doria (MD) Seamount locates the position of *Cylice* cruise submersible dives number 21–22 [Sossion and Guennoc, 1997] and dredges from *Marco* cruise [Guennoc *et al.*, 1995, 1998]; (b) migrated seismic reflection line corresponding to the inset “b” in Figure 10a, illustrating the acoustic facies of the volcanism sampled at point MD (alkaline olivine basalt, 17–18 Myr old). Top of the acoustic basement is labeled “tab”; high-velocity level (7.5–7.7 km/s) in the lower crust is labeled “hvl.” See Figure 6 for abbreviations.

and Réhault, 1990; Bonatti *et al.*, 1990]. It was proposed that this oceanic-type seafloor could be the consequence of a synrift tectonic unroofing of the upper mantle in a continental rupture zone [Masclé and Réhault, 1990].

6.3. Space-Time Migration of Volcanism

[40] Three main magmatic fields are recognized on the margins and in the transitional domain (red bold curves on Figure 4): (1) offshore the Maures-Estérel Massif, (2) offshore the NW part of Corsica and in the Gulf of Genova, and (3) in the SW part of Corsica, the Bonifacio Strait [Guennoc *et al.*, 1995; Rossi *et al.*, 1998]. Furthermore, several spread volcanic fields are found in the deep basin (Figure 4).

[41] The large, ovoid magnetic anomalies observed at the foot of the Maures margin, and smaller anomalies

located farther north, form three discontinuous magmatic fields on the margin and in the transitional domain (Figures 4 and 11). They are found onshore as eruptive products (mainly, volcanic ash and andesitic breccia of Villeneuve-Loubet and Cap d’Ail) and intrusions (estérellites, sills at the eastern end of the Maures Massif) [Rouire *et al.*, 1980]. Although not sampled offshore, they are therefore likely to represent calcalkaline bodies, as in the Catalonia and Menorca margins [Mauffret *et al.*, 1995, and references therein], and to be related to the subduction setting. The two volcanic fields offshore Corsica have been widely sampled and provide clear indications on the age and the nature of the magmatism which is associated with the basin opening [Rossi *et al.*, 1998]. In the SW Corsican margin and the Sardinian rift prolongation onshore the calcalkaline lava flows have a strong

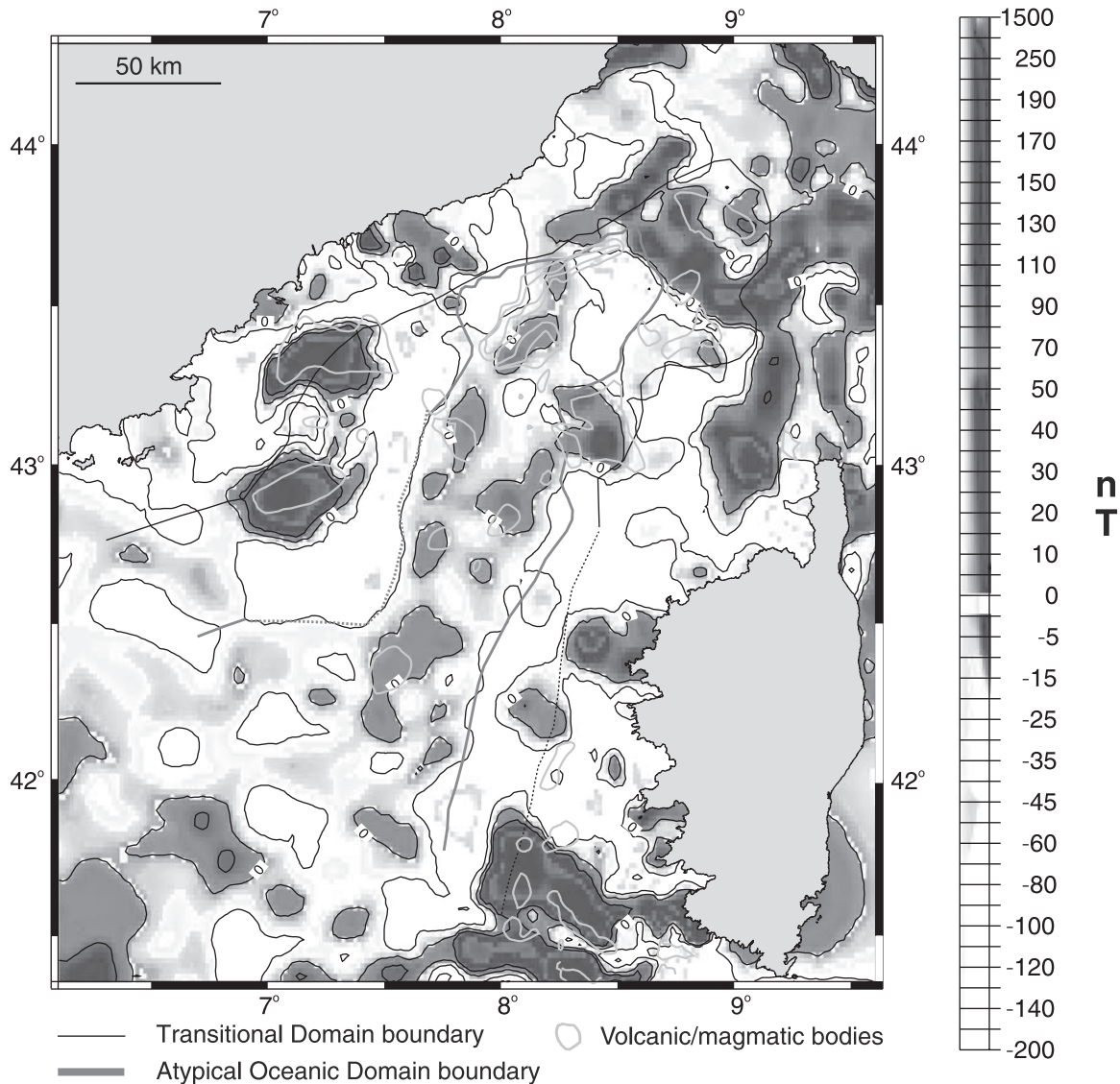


Figure 11. Superimposition of the volcanic and magmatic bodies determined on seismic profiles (red contours) above the magnetic anomalies newly determined by reduction to the pole. Coastlines (thin dark line) and boundaries of the transitional domain TD (thick dark line) and of the deep oceanic-type basin (gray line) proposed in the Ligurian basin are shown. Units are in nanoteslas (nT). See color version of this figure at back of this issue.

potassic nature (basalts, acid andesites and dacites). Dating suggests two distinct magmatic events: (1) the $^{40}\text{Ar}/^{39}\text{Ar}$ ages on amphiboles are 16 ± 0.4 Ma, i.e., close to the age obtained by fission tracks on apatite (17.2 ± 0.8 Ma), and K/Ar ages on total rocks with low loss-on-ignition (21–18 Ma and 16–15 Ma) [Coulon, 1977; G. Ruffet, personal communication] are corroborated by $^{40}\text{Ar}/^{39}\text{Ar}$ ages on separated biotites and amphibolites (20 Ma and 15.7 Ma; G. Ruffet, personal communication); (2) at Monte Doria (Gulf of Genova) and NW of Corsica, the lava flows are also mostly potassic and acid, but ages indicate a more recent event at 11–12 Ma (G. Ruffet, personal communication). All ages are thus systematically younger than the ones (Oligocene) recognized on the Ligurian coast, like in

Monaco, Cap d’Ail, and Villeneuve Loubet (Figure 3) [see Rouire *et al.*, 1980].

[42] Therefore the volcanism identified in the Ligurian domain has migrated from the Ligurian margin at the beginning of rifting, toward the Corsican margin until the end of the Corsica-Sardinia block rotation. From the compositions and ages found by sampling in the basin we may distinguish at least two distinct contributions: (1) calcalkaline and dominantly andesitic magmas, comparable to volcanics found in the Gulf of Valencia and in Sardinia, first emplaced in the western Ligurian basin, and later on the Corsican margin; and (2) intrusion of magmas derived from mixing of lithosphere involved in the root of the northern Apennines belt, as found in the Tyrrhenian Sea [Serri *et al.*,

1993]. Around 12–11 Ma, a third calcalkaline and alkaline volcanic phase is recorded only in the NE part of the basin, in the Gulf of Genova (Figure 4). This area has therefore undergone a long-lived volcanic activity (at 21–18 Ma, 16–15 Ma, and 12–11 Ma) which seems more or less steady in space and is close to the Toscane Province, where deformation is reinitiated at the onset of Tyrrhenian opening 12–10 Ma ago [Jolivet *et al.*, 1998].

7. Discussion

7.1. Magmatism: Underplating or “Undercrusting”?

[43] High-velocity levels (HVL) are described on several Atlantic-type passive margins, which are interpreted either as underplating resulting from the partial melting of the upper mantle during rifting on volcanic margins [e.g., *White and McKenzie*, 1989; *Eldholm and Grue*, 1994; *Holbrook et al.*, 1994], or as partly serpentinized peridotites of the upper mantle exhumed in the continental breakup zone on nonvolcanic passive margins (West Iberia margin, “undercrusting” of *Boillot et al.* [1989], *Pinheiro et al.* [1992], *Whitmarsh et al.* [1993], *Brun and Beslier* [1996], *Pickup et al.* [1996]; Labrador Sea, *Chian et al.* [1995]; Bay of Biscay, *Thinon* [1999]). As underlined by *Mauffret et al.* [1995], the Liguro-Provençal margins do not display the main characteristics of volcanic margins, since neither seaward dipping reflectors nor thick HVL are found. Moreover, the Tyrrhenian Sea, which opened in a similar geodynamic context, depicts an atypical oceanic crust corresponding to a very thin basaltic crust overlying serpentinized mantle [*Masclé and Réhault*, 1990; *ODP Leg 107 Shipboard Scientific Party*, 1987; *Bonatti et al.*, 1990; *Kastens and Masclé*, 1990]. From our data set only, we cannot unambiguously discount previously proposed hypotheses. However, the above mentioned observations, together with the range of acoustic velocities and the geometry and the location of the thin HVL found almost systematically along the Ligurian margin, suggest that the HVL could represent partly serpentinized peridotites. Accordingly, volcanism would be limited to ovoid bodies in the TD of the Ligurian basin, which indeed is further documented by magnetic data (Figure 11). The synrift exhumation of the mantle may have occurred either by simple shear along reactivated inherited compressive structures [*Mauffret et al.*, 1995], or as a consequence of lithospheric boudinage [*Brun and Beslier*, 1996].

7.2. Contributing Forces in the Opening of the Ligurian Basin

[44] The formation of back arc basins has been highly debated in plate tectonics since a long time. Several mechanisms have been proposed to explain their evolution [e.g., *Martinez et al.*, 1995], many authors laying emphasis on the western Mediterranean case [*Dogliani*, 1995; *Dogliani et al.*, 1997; *Faccenna et al.*, 1997; *Gueguen et al.*, 1998; *Jolivet et al.*, 1999; *Jolivet and Faccenna*, 2000]. Our investigation of the Ligurian basin has focused on the structural pattern and the distribution of volcanism on the margins and in the

deep basin. From these results, we seek here to speculate about the relative contributions of forces to the formation of the Ligurian basin, and especially the pull force induced by the sinking Apulian oceanic slab and the gravitational collapse of the Alpine orogen at the time of rifting [e.g., *Dogliani et al.*, 1997; *Faccenna et al.*, 1997].

[45] It is widely accepted that the pull force induced by the subducting Adriatic (African) slab is the most important one for the Tyrrhenian back arc extension, which occurred in continuity with the Ligurian basin opening [e.g., *Malinverno and Ryan*, 1986; *Faccenna et al.*, 1996, 1997]. Considering the overall migration of volcanic activity, this force likely played an important role also in the dynamics of the SW Ligurian Sea opening, since a relatively long oceanic slab had already been subducted at late Oligocene time [*Jolivet et al.*, 1998; *Chamot-Rooke et al.*, 1999]. This could have allowed for relatively high extensional rates in the upper plate, as expected from experimental modeling [e.g., *Shemenda*, 1993; *Chemenda et al.*, 1996; *Faccenna et al.*, 1996]. The Hercynian tectonic inheritance together with a strong rollback effect from the hinge of the subducting plate could have therefore localized the extension at the foot of the margin, in the prolongation of the Sardinian rift toward the north (Figure 12b).

[46] Nevertheless, we believe that the Ligurian domain has undergone quite contrasting conditions from NE to SW (Figure 12). Indeed, the NE domain is located quite close to the rotation pole of the Corsica-Sardinia block [*Réhault et al.*, 1984], and the amount of subduction there is much less than in the south. Furthermore, the subducting lithosphere was continental since the Oligocene at this place, whereas it is oceanic until today south of the latitude of mid-Corsica, at $\sim 42^\circ\text{N}$ [*Malinverno and Ryan*, 1986; *Serri et al.*, 1993]. We therefore suggest that extension velocity (and hence, strain rate) and slab pull were strongly decreasing from SW to NE in the Ligurian domain at the time of opening. This has important consequences on the style of extension, since these parameters are known to play a key role in the cooling and strengthening of the rifted area and in the consequent shifting of the locus of extension [*Bassi*, 1995; *Faccenna et al.*, 1997]. Finally, the former Alpine-Apennines thickened crust, remnants of which have been found offshore NW Corsica and in the Gulf of Genova, could have been responsible for additional gravitational body forces (see section 2.1) and for a high Moho temperature ($650^\circ\text{--}800^\circ\text{C}$) in the NE Ligurian domain. Although the contribution of buoyancy forces versus the slab pull component cannot be quantified here, we suggest from our observations and from previous models of continental extension [e.g., *Buck*, 1991; *Bassi*, 1995] that this orogen has added a substantial contribution to produce distributed extension on the whole NW Corsican margin, and hence, even smaller strain rates. This interpretation is also evoked for the northern Tyrrhenian opening [*Faccenna et al.*, 1996, 1997], and more widely for the Mediterranean basins developed within the Alpine orogen [*Dogliani et al.*, 1997; *Jolivet et al.*, 1999].

[47] Whether a lateral slab detachment enhanced the migration of extension from the Ligurian basin toward the

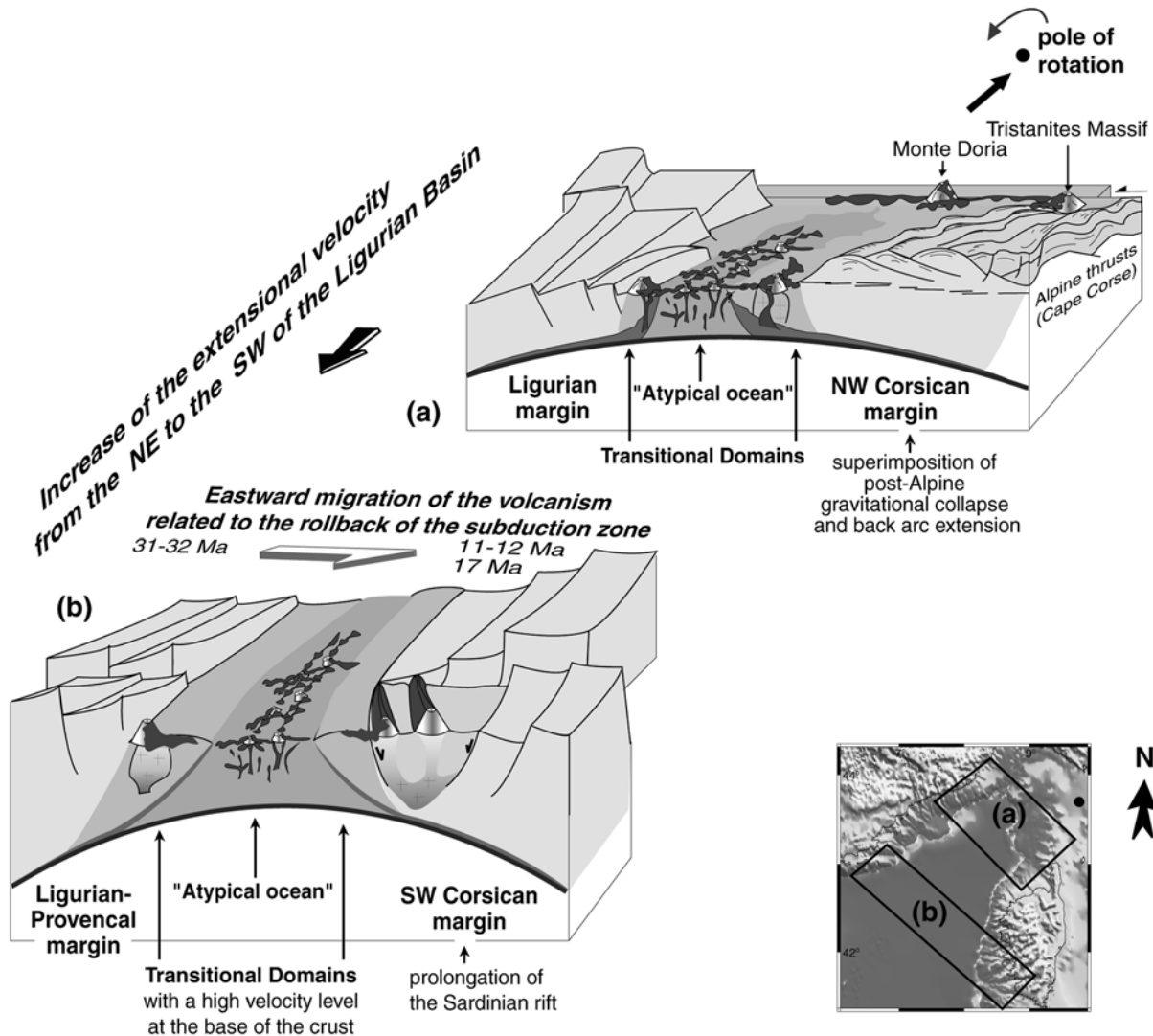


Figure 12. Synthetic block diagrams summarizing our interpretation of the present-day structure and evolution of the Ligurian basin. (a) View of the NE part where back arc extension and gravitational collapse of the Alpine orogenic belt have taken place, thus distributing the extensional deformation over the NW Corsican margin. (b) View of the SW part where margins are narrow, except at the foot of the SW Corsican volcanic margin, in the prolongation of the Sardinian rift.

Tyrrhenian Sea [Carminati *et al.*, 1998a, 1998b], or whether the increase in the retreat velocity of the African plate at $\sim 35\text{--}30$ Ma is only a consequence of the decrease in the absolute velocity of Africa [Jolivet and Faccenna, 2000] is beyond the scope of this study. Nevertheless, both processes are not excluding each other and may satisfactorily explain why back arc extension was triggered at that time.

7.3. Margins Asymmetry and Migration of Extension

[48] The rollback of the subducting zone hinge produces an extensional stress field in the upper plate which is believed to generate an asymmetrical mechanism of rifting, as long as the volcanic arc is steadily settled on the side of the subducting plate [Martinez *et al.*, 1995]. In this model,

back arc extension forms a narrow margin on the side of the volcanic arc, close to the subduction zone, whereas the conjugate margin is widening by tectonic extension and is accompanied by volcanism in the first stage of opening. This geometry is observed in the southern Tyrrhenian Sea where the Sardinian margin is wider than the Calabrian conjugate [Masce and Réhault, 1990; Wang *et al.*, 1989; Spadini *et al.*, 1995], and also in the Provençal basin where the Gulf of Lion margin is wider than the west Sardinian conjugate [Burrus, 1984; de Voogd *et al.*, 1991]. Conversely, we observe an opposite asymmetry of conjugate margins in the NE and SW Ligurian basin, with a Corsican margin located on the volcanic arc side wider than the Ligurian conjugate, except in the central part where both margins are narrow (Figures 3 and 4). This observation

suggests other factors to be accounted for and questioned the general validity of the *Martinez et al.* [1995] hypothesis.

[49] A first important parameter to examine is the strain rate. We have postulated in the previous section (section 7.2) that (1) prerift rheology is weak in the north Ligurian domain owing to geological inheritance; (2) the extensional velocity decreased from SW to NE during the Ligurian Sea evolution because of its scissors-shape opening. Duration of drifting in the Ligurian Sea, although poorly constrained, can be assumed to be ~ 5 Ma (see section 2.1). From the width of the “atypical” oceanic domains inferred in the northern and western basin (~ 50 and ~ 100 km, respectively; Figure 4), a crude estimate of the mean relative extension velocity is ~ 1 and ~ 2 cm yr $^{-1}$ in the north and south Ligurian basin, respectively. This latter value is close to the average extension velocity found in the north Tyrrhenian basin, estimated at ~ 1.5 – 2 cm yr $^{-1}$ [*Faccenna et al.*, 1997]. This is also close to the threshold value postulated by *Wright et al.* [1996] for a slow extensional back arc setting, whereby cross-arc seamounts and ridges are supposed to form and to be preserved during arc migration. Note that a similar, rough basement morphology is observed in the Ligurian and Tyrrhenian deep basins. Conversely, these extension rates are much smaller than rates of drifting speculated in the Algero-Provençal and the southern Tyrrhenian basins, which are ~ 4 – 5 and ~ 6 cm yr $^{-1}$, respectively (see a review by *Faccenna et al.* [1997] and *Chamot-Rooke et al.* [1999]).

[50] Two other key parameters determining the shaping of conjugate margins are the rheological properties of the lithosphere and the asymmetry of the boundary forces induced by the slab rollback. In the northern Tyrrhenian Sea, it is proposed that the magmatic arc focuses the extensional process in the weakened ductile crust and that the sense of shear is controlled by the lower crust partially melted under the active magmatic zone [*Jolivet et al.*, 1998]. Since the thermal anomaly moves eastward as a consequence of the migration of the slab, the crust goes back to a higher strength and extension jumps eastward, thus widening the eastern margin [*Jolivet et al.*, 1998; *Brunet et al.*, 2000]. This process is likely to occur for slow rollback, relatively weak lithospheric rheology, and low strain rates, since in these conditions, strengthening is large enough before yield stress is reached [*Bassi*, 1995]. This is indeed what is expected in the context of the north Tyrrhenian Sea, characterized by a continental-type subduction regime, a thick postorogenic Alpine crust, and a rotation pole as close as the Gulf of Genova [*Faccenna et al.*, 1997].

[51] From this example, it appears that other processes than the extension velocity obviously control the style of rifting in back arc settings, like for instance prerift crustal thickening (and consequently, rheological properties prior to rifting) and slab pull asymmetry. We have shown that in the northern Ligurian basin, ~ 30 – 20 Ma ago, (1) prerift thickening is likely; (2) extension velocity is low owing to the proximity of the pole; and (3) slab pull is weaker than in the south. Therefore a scenario similar to the northern Tyrrhenian one [*Jolivet et al.*, 1998] may have determined its evolution during rifting and drifting, resulting in wide-

spread extension on the Corsican margin, on the side of the migrating volcanic arc. Further south, we suggest that a moderate but nevertheless faster and more efficient slab rollback may have enhanced an overall strong strain localization during extension, thus giving birth to rather narrow Ligurian and Corsican (south and central) margins, since no significant crustal thickening is reported there before the Oligocene Ligurian rifting. In the Provençal basin, the previous outer Pyrenean thrusting may have modified the rheology of the overriding plate and favored the widening of the Gulf of Lion margin during rifting [*Burrus*, 1984; *Gorini et al.*, 1993; *Séranne*, 1999], whereas the western Sardinian margin remained relatively narrow [*Thomas and Gennesseaux*, 1986]. Finally, we have shown that a strong volcanic event at the end of the Ligurian basin opening (~ 17 Ma) is associated to the Sardinian rift extension up to the SW Corsican margin [*Thomas and Gennesseaux*, 1986; *Gennesseaux et al.*, 1989; *Guennoc et al.*, 1995, 1998]. We believe that this second event may explain why this SW Corsican margin segment became wide. Such an event is likely related to the slowing down of the subduction velocity which occurred at the end of the spreading episode in the Ligurian Sea (~ 15 Ma), soon relayed by a new back arc pulse in the Tyrrhenian Sea. This two-step opening is thought to result from the interaction between the slab and the 660 km transition zone in the mantle [*Faccenna et al.*, 2001].

8. Conclusion

[52] New multichannel seismic reflection data constrained by magnetic data and geological samples allow us to image the deep structure of the Ligurian basin and to describe contrasting tectonic styles of opening of the basin during its Oligocene-Miocene evolution. The Ligurian basin opening appears to result at first order from back arc extension linked to the rollback of the subducting Apulian slab and of the Alpine-Apennines belt front toward the E-SE in a way similar to what happened later in the Tyrrhenian Sea. Calcalkaline volcanism is recognized on both sides of the margins and in the transitional domains, whereas an atypical oceanic domain was formed during spreading in the deep basin, characterized by discontinuous magnetic anomalies.

[53] We argue that tectonic inheritance (crustal thickening and structural fabric) has a substantial influence on the different styles of deformation on the same margin and on the asymmetry between the conjugate margins. In the NE Ligurian basin the prerift Alpine thickening of the crust and a relatively slow rollback effect contribute to delocalize the extension on the NW Corsican margin, producing the gravitational collapse of this part of the Alpine orogenic belt through the reactivation of several oceanward faults soling on a relatively flat ductile shear zone. Farther south, the Hercynian structural inheritance combined with a more efficient slab retreat localizes the extension, first, during rifting in the central basin and second, at the end of drifting on the SW Corsican margin, in the prolongation of the Sardinian rift.

[54] Using the overall pattern of migration of volcanism toward the southeast, accurately located magnetic anomalies, and the structural pattern described, we propose that the evolution of the Ligurian basin took place in three main periods in an almost continuous process of extension:

1. In the late Oligocene–early Miocene, on the upper plate of the African–Apulian subduction, rifting initiates in the Ligurian basin area at relatively slow velocities (~ 1 and ~ 2 cm yr⁻¹ in NE and SW, respectively) and is associated with important calcalkaline magmatism on land and with large ovoid magnetic anomalies along the western Ligurian margin at sea. During this rifting period, a restricted number of tilted blocks (two or three blocks on the Ligurian margin) appeared along narrow margins, except in the previously thickened Corsican crust where extensional deformation is distributed over 120 km at least. This extensional deformation includes extensional features observed onshore in Corsica.

2. From 21 Ma (end of rifting generally inferred) until 16–15 Ma the anticlockwise rotation of the Corsica–Sardinia block toward the east allows for the beginning of oceanic spreading characterized by discontinuous tholeiitic volcanic edifices settled within unroofed mantle (as observed in the Tyrrhenian Sea). This explains the development of magnetic anomalies atypical of oceanic crust in the deep basin and underlined by two main discontinuous lineations sub-parallel to the basin axis.

3. Finally, the end of opening at 16–15 Ma is associated with a second important calcalkaline volcanic phase along the Corsican margin linked to the migration of the subducting lithosphere toward the E–SE and to the still ongoing gravitational collapse of the Alpine orogenic belt, with calcalkaline and dominantly andesitic magmas first emplaced along the west Corsican margin, and intrusion of magmas derived from mixing of lithosphere involved in the root of the northern Apennines belt. Around 12–11 Ma, a third calcalkaline and alkaline volcanic event is recorded in the Gulf of Genova, probably triggered by the gravitational collapse of the yet thick Alpine crust associated with the first stage of opening of the Tyrrhenian Sea.

[55] **Acknowledgments.** We are strongly indebted and deeply grateful to the crews of the R/V *Nadir* and *le Suroit* (GENAVIR-IFREMER) and to all other members of the scientific staffs during *Malis95*, *Marco95*, and *Cylice97* cruises. E. Gueguen, G. Ruffet, and N. Chamot-Rooke are thanked for their help in data interpretations in the course of this study, as well as N. Béthoux, I. Contrucci, J. Ferrandini, R. Gloaguen, D. Lahondère, P. Rossi, F. Sage, M. Schaming, and J. Virieux for help and fruitful discussions at different stages of the work, and W. P. Schellart and C. Faccenna for helpful comments on the manuscript. This research was funded by INSU (Institut National des Sciences de l'Univers, Centre National de la Recherche Scientifique, Marine Geosciences and PNRN Programmes), CNRS, and University Pierre et Marie Curie (Paris VI). Contribution 388 of UMR 6526-Géosciences Azur (CNRS/UPMC/UNSA/IRD).

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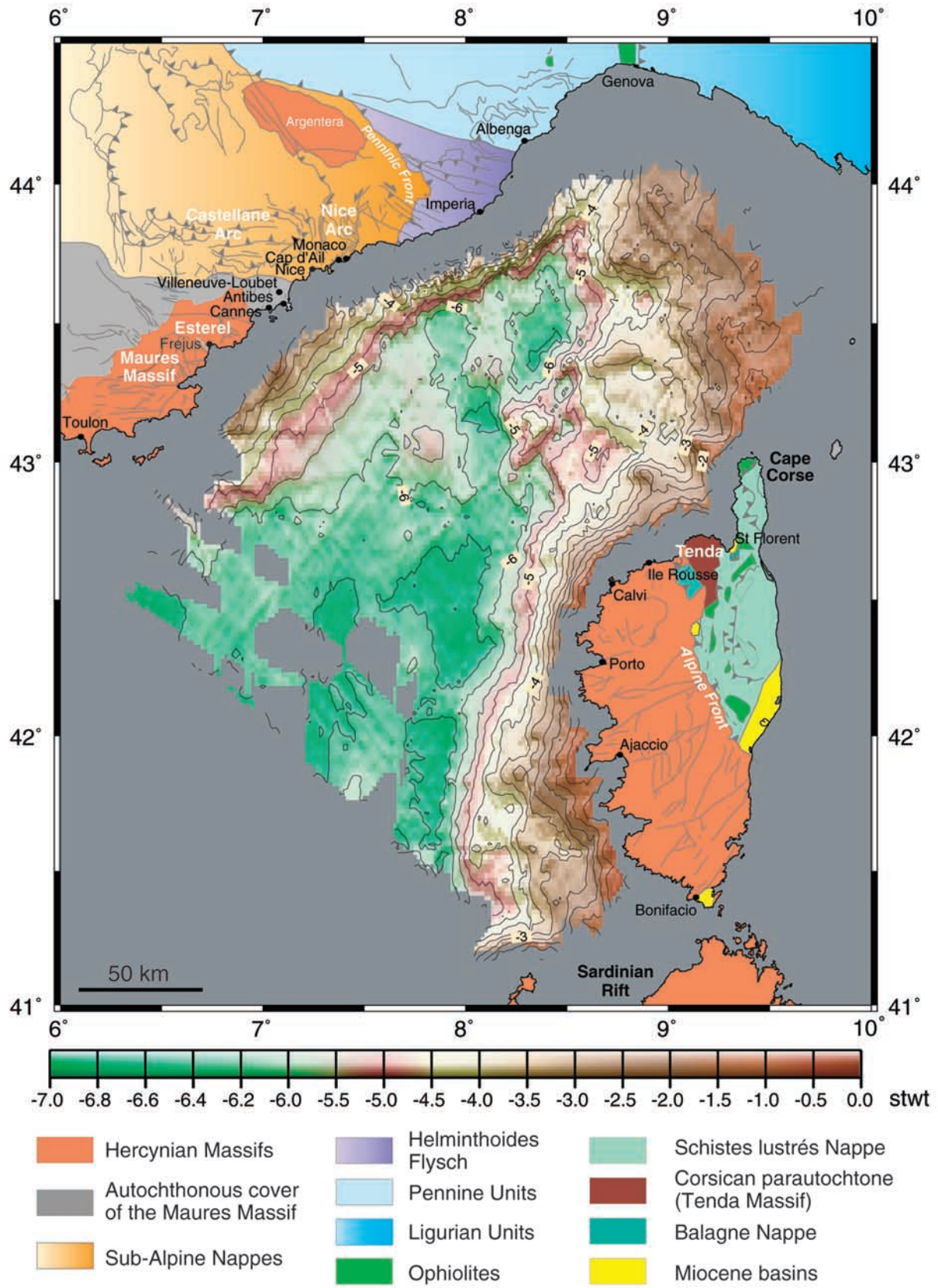
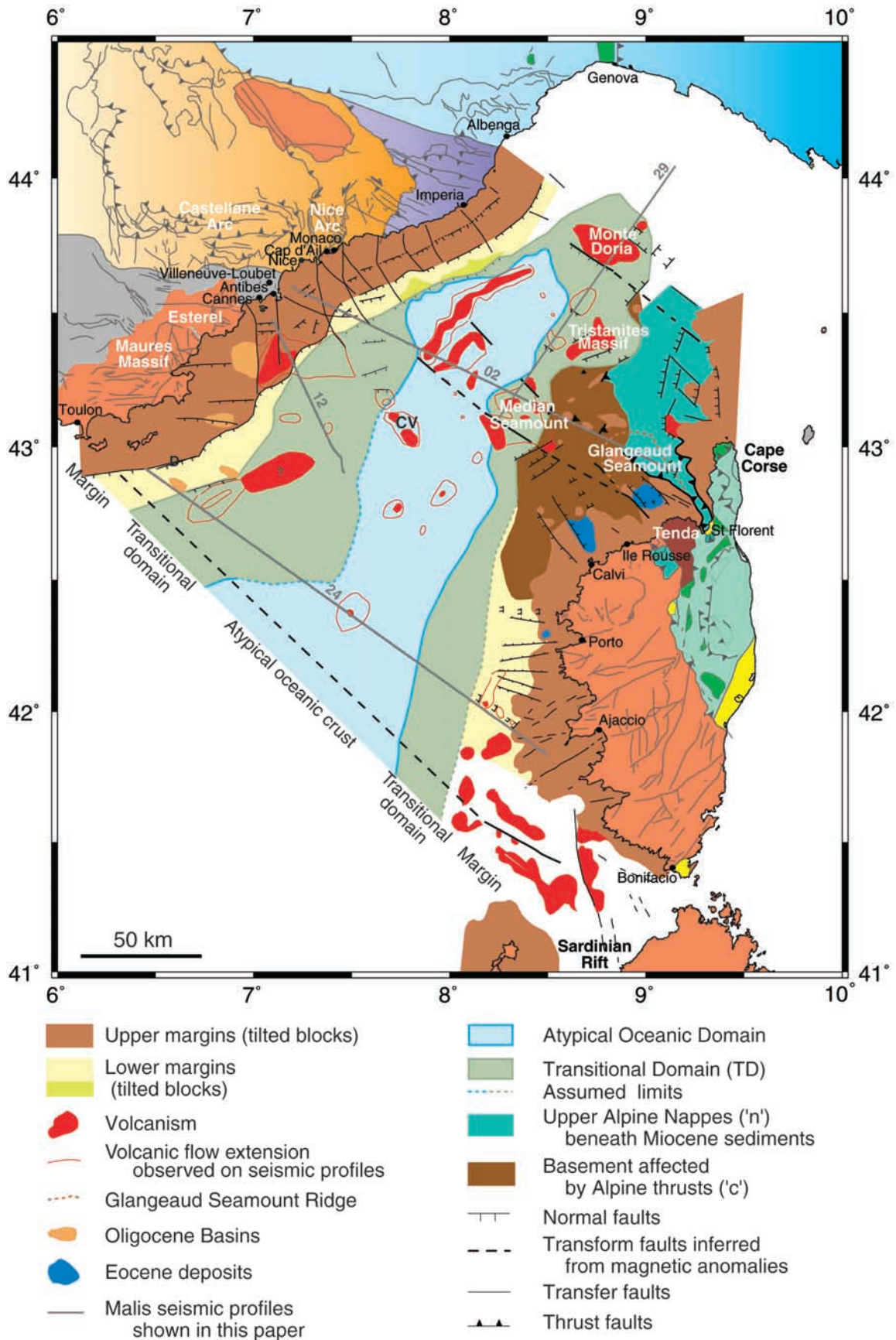


Figure 3. Isochrons of the top of the prerift basement (in seconds two-way time (s TWT)) and distribution of the geological domains observed onshore (see text for details).



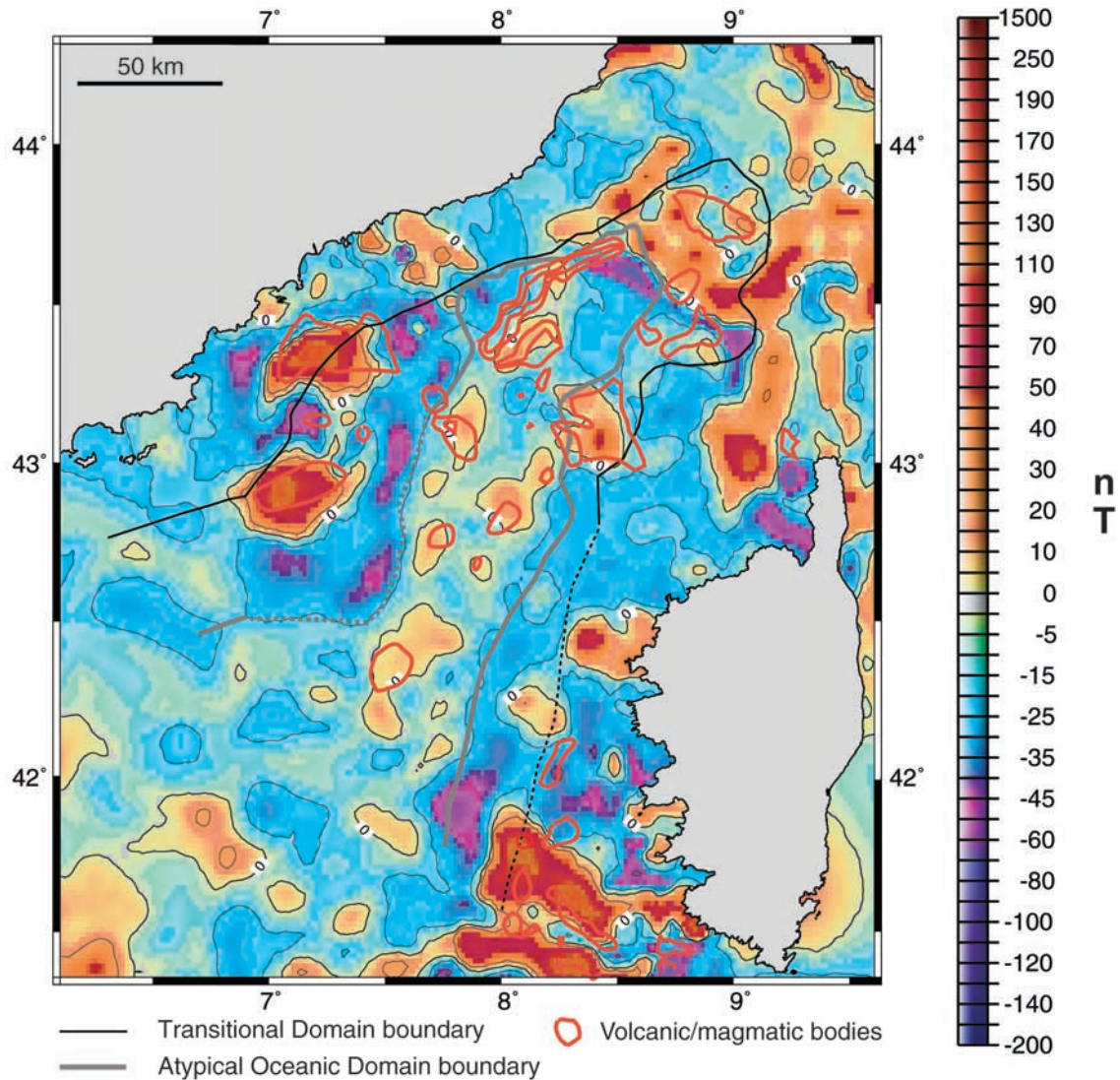


Figure 11. Superimposition of the volcanic and magmatic bodies determined on seismic profiles (red contours) above the magnetic anomalies newly determined by reduction to the pole. Coastlines (thin dark line) and boundaries of the transitional domain TD (thick dark line) and of the deep oceanic-type basin (gray line) proposed in the Ligurian basin are shown. Units are in nanoteslas (nT).

Figure 4. (opposite) Structural sketch map of the Ligurian domain after results from this study (see onshore geological domain captions on Figure 3). Continuous lines (numbered) refer to *Malis* cruise seismic profiles. Abbreviations are as follows: CV, central volcano; D, reflector “D” emergence (see text for details).