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1

Introduction

The location of the deep seismic profile and the geological and geophysical framework

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I.A. – WHY A GEOPHYSICAL TRAVERSE THROUGH THE WESTERN ALPS AND ITS BORDERLAND?

The major question posed by the existence of mountain belts is how shortening is obtained between colliding plates and how the thickening occurs in the belt itself. Is it by shortening (pure shear) at the scale of the two lithospheric plates, or by vertical disappearence of (lower) crust ("Verschlukung" of some Swiss authors), both mechanisms implying an overall symmetrical structure? Or is it by subduction of one lithosphere beneath the other, with or without asymmetrical shallow thrusts? Another question deals with the location of the main decoupling levels and movement zones; clearly, the base of the lithosphere must be a major decoupling surface. Surface geology and oil drilling suggest the existence of another level at shallow depth in the crust. Other levels could be located, for instance, at the base of the crust, as suggested by thermomechanical models. The lithosphere implied in the collision had a previous zone of weakness, the decoupling could occur at that level.

By analogy with sedimentary accretionary prisms (fig. I-3a) in front of subduction zones, it has been suggested [Mattauer, 1986] that a collision belt could create a continental accretionary prism (fig. I-3b). This raises the question of which part of the continental crust would be accreted (i.e. upper crust only, entire crust or, why not, flakes of the entire lithosphere thus involving mantle material (fig. I-3c).

Such vast problems cannot be handled by geologic studies alone. Because the whole lithosphere is implied, the insight at depth given by geophysical techniques becomes necessary. This is why geophysical traverses through mountain belts are so important. After a first ECORS deep seismic reflexion profile across the Variscan belt in northern France, the next profile in continental crust was undertaken in the Pyrenees [Choukroune *et al.*, 1989].

All the major problems mentioned above are encountered in the western Alps. For instance, the model of a vertical and symmetrical subduction ("Verschluckung") has been proposed in the central Alps, while an eastward or southward subduction of the European plate below the Adria plate preceding collision is generally accepted for the western Alps, except for Caby *et al.* [1978] who proposed the alternative of westward subduction. However, the amount of backthrusting is still largely discussed and seem to vary from the western to the central Alps, as can be seen by comparing the results of the ECORS-CROP profile with those of the NFP 20 profile [Pfiffner *et al.*, 1988; Bernoulli *et al.*, 1990; Schmid, 1993; Marchant *et al.*, 1993; Pfiffner, 1993]. Even within the ECORS-CROP profile, there are various possible interpretations which are presented in the conclusion to this volume.

Another question is whether only the European continental crust is stacked within the tectonic wedge [Mattauer, 1986] or there is also some involvement of the European mantle [Ménard and Thouvenot, 1984; Polino, 1986; Vialon, 1986]. Related to this is the amount of shortening between the colliding plates and how it is obtained. One extreme model implies underthrusting of the European plate far below the Po Basin [Butler, 1986; Roeder, 1989]. Other

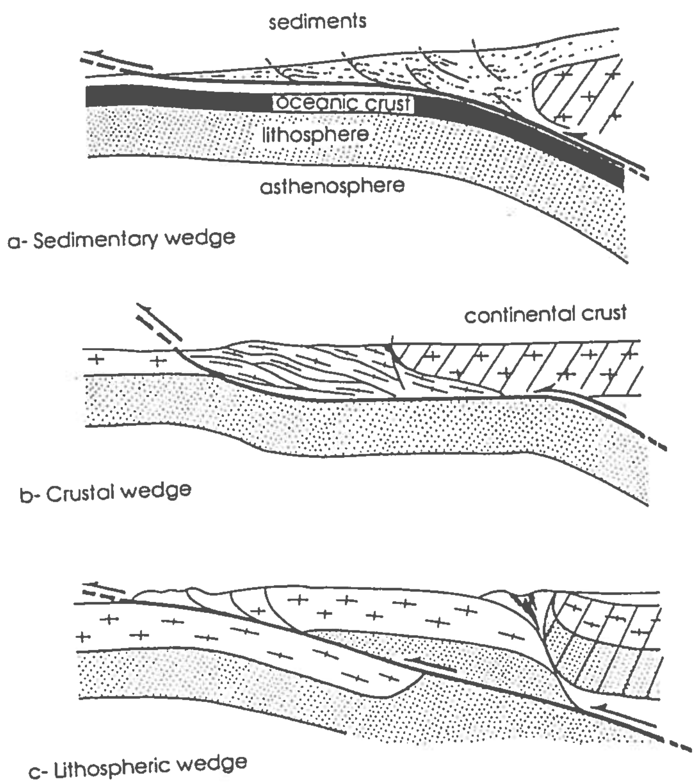


FIG. I-3. – Sketches of distinct accretionary wedging mechanisms during plate-convergence, showing how they depend on the location of the main decoupling horizon (bold line with arrows) [after Nicolas *et al.*, 1990]. The upper plate has been hatched in the three sketches. a : sedimentary accretion in an oceanic subduction zone. b : continental crust accretion in the Himalaya collision belt [inspired from Mattauer, 1986]. c : lithospheric accretion in a collision belt.

FIG. I-3. – *Schéma montrant l'importance des niveaux de décollement dans l'architecture des prismes d'accrétion liés à la convergence de plaques lithosphériques [d'après Nicolas* et al., *1990]. Dans les trois hypothèses, la plaque chevauchante est hachurée : a = accrétion de sédiments dans une zone de subduction ; b = accrétion crustale dans la chaîne de collision himalayenne [d'après Mattauer, 1986] ; c = accrétion lithosphérique dans une chaîne de collision.*

models restrict the eastward underthrusting to the Insubric line. This does not constrain adequately the amount of shortening for two reasons. First, the crust may have been partly transformed to eclogites, thus loosing its "crustal" geophysical properties and, being no longer imaged by geophysical techniques, it could extend eastward without a visible signature. Alternatively, the crust may have been stacked in the accretionary wedge, and then uplifted and eroded to an unknown extent [Platt, 1986; Polino *et al.*, 1990]. Finally, Laubscher [1971], Tapponnier [1978], Ricou [1984], Lacassin [1989] and Ménard [1988] have proposed that a large amount of shortening is accounted for by lateral extrusion of the plates.

In order to avoid the difficulties of 3-D reconstruction implied by the last point, we have decided to locate the ECORS-CROP profile in an area of the belt which is two-dimensional (cylindrical), in structure. This also helps us to project all data which have been obtained in a nearby area (see chapters III and IV) onto the plane of the profile.

Eventually, we need to solve palaeogeographic problems and reconstruct the history of the considered area in order to understand how the belt was formed. The Alpine belt and its continental sutures result from the closure of the Neo-Tethyan ocean which once separated the European and Adria plates. Alternative models stress the importance of other oceanic domains, such as the Valaisan trough [Lemoine and Trümpy, 1987], which could be associated with other weakness zones in the lithosphere. These domains may have been used later, during convergence, as movement zones separating lithospheric flakes.

Major listric faults may also have affected the European lithosphere during rifting, leading to the opening of the Piemontese Ocean. Such faults could have been reactivated during the collision, thus controlling the location of major thrusts. In this respect, it was important that the ECORS-CROP profile through the western Alps should be extended inside the European plate to the border of the Massif Central (Alpes II and Bresse profiles).

The problem of locating the main suture zone is important in view of the different palaeogeographic reconstitutions. Most authors consider that this suture coincides with the eastern limit of ophiolite outcrops in the Alps, and thus of oceanic rock-type derived from the Tethyan domain. This corresponds to the Viu-Locana zone. In such a case, the Sesia Lanzo unit, which is located east of this limit, belongs to the Adria plate. However, this unit has been affected by high pressure Alpine metamorphism and, consequently, by the subduction. This opens up the possibility that the suture is located eastward of the Sesia Lanzo unit, along the Insubric line. Sesia Lanzo kindred rocks may thus represent the margin of the Adria plate or the margin of the European plate, or some microplate in-between. The peridotite massif of Lanzo can similarly be considered as the mantle from below the Adria or European plates, or from beneath the Piemontese Ocean itself.

The location of the profile had to be chosen in order to give the best chance of solving these scientific problems, crossing also an area where the structures are as cylindrical as possible. But other constraints had to be considered: the profile should be as straight as possible, perpendicular to the main structural lines, while avoiding highly populated valleys. The chosen profile location satisfies most of these requirements.

I.B. – Geologic framework

The Alps are usually interpreted as the result of a continent-continent collision involving the European plate, the Adria microplate and the interposed Tethyan Ocean.

Alpine orogenesis was directly related to the relative motion of Europe to Adria, and started with the subduction of the intervening oceanic lithosphere. The oceanic subduction was active at least from 130 Ma, as indicated by radiometric ages on Alpine H P/L T mineral assemblages, and acted for at least 80 Ma, as suggested by the long lasting eclogite to blueschist facies metamorphic conditions, from the Upper Cretaceous (130 Ma) to the Eocene (60 Ma) [for a review, see Hunziker *et al.*, 1993].

Petrological estimates on subduction related Neo-Alpine mineral assemblages suggest that both basement and ophiolithic nappes reached high P/T conditions with depth ranging from 50 up to 100 km [Chopin, 1984]. The continent-continent collision occurred later (60-40 Ma). Continuing compression generated lithospheric wedging and crustal ruptures below the External Crystalline Massifs during the late Tertiary (40-25 Ma), the uplift of internal collided zones and, at more superficial levels, decollement nappes and detachment of the Jura cover.

From the foreland towards the hinterland, that is from west to east, distinct structural domains are thus defined according to their specific and independent histories.

According to surface criteria, we distinguish (plate I):

- the foreland *sensu stricto* corresponds to the stable fragment of the European plate, onto which detached and deformed tectonic units of the Alpine collisional belt were thrust. Along our transect, it corresponds essentially to the Massif Central and the Bresse Basin;
- the Jura Mountains constitute the frontal part of the Alpine belt, which overrides Mio-Pliocene (Pontian) sediments of the Bresse Basin;
- the Molasse Basin represents in part a foredeep basin developed above the bent European lithosphere (flexural basin created by loading of internal nappes) in Oligocene and early Miocene times, but this sedimentary basin as a whole was later translated above a basal detachment located in the Triassic and emerging at the Jura front;
- the Subalpine Massifs constitute a thrust and fold belt which affects the Mesozoic and Cenozoic sediments. Along the line of transect, it corresponds to the Borne Massif;
- the External Crystalline Massifs occur as basement windows beneath the detached Mesozoic cover of the Subalpine Massifs. They are usually regarded as late-stage ramp anticlines that outline the crustal shortening of the European plate.

All the above zones are grouped together into the foreland domain *sensu lato*, where Alpine compressional effects begin to be felt. Further east, the crust is thickened and the following zones can be recognized:

- the Axial metamorphic belt is bordered to the west by the Penninic thrust front (major crustal suture zone), and to the east by the Insubric line (Canavese fault). This metamorphic belt is made up of a pile of basement nappes, with additional sedimentary and ophiolitic units whose tectono-metamorphic structure is directly linked to early subduction and collisional stages;
- the South-Alpine domain represents the hinterland of the belt [Bernoulli *et al.*, 1990].
- the Po Plain represents the foreland for the Milano belt and southern Alps [Ménard, 1988], and also for the Neogene Apennines [Roure *et al.*, 1990].

1) The Alpine foreland

The foreland to the Alpine belt corresponds to a zone of the European platform where tectonic displacements and deformation effects related to collision are first observable. Under the weight of Alpine thrust sheets from more internal zones, the platform was subjected to a flexure on the scale

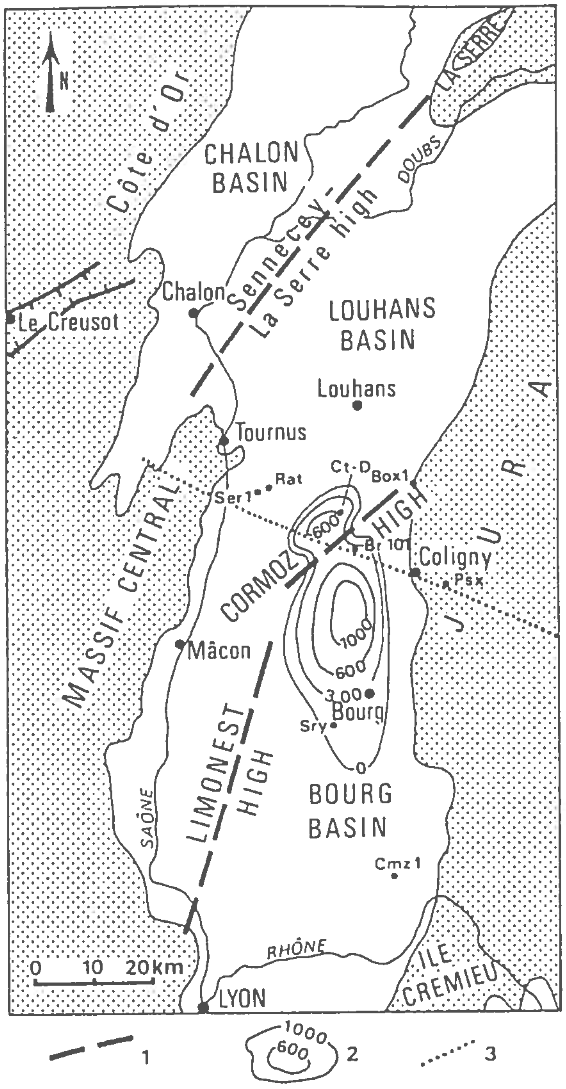


FIG. I-4. – Main structural features of the Bresse graben [after Bergerat *et al.*, 1990]. 1 : basement high ; 2 : Oligocene salt isopachs [after Curial, 1986 a, b] ; 3 : location of ECORS traverse. Box 1 : Bois-des-Chaux 1 borehole ; Br 101 : Bresse 101 borehole ; Cmz 1 : Cormoz 1 borehole ; Ct-D : Curciat-Dongalon borehole ; Psx : Poisoux 1 borehole ; Sry : St Remy borehole ; Ser 1 : Sermoyer borehole ; Rat : Ratenelle borehole.

FIG. I-4. – *Principales structures du fossé bressan [d'après Bergerat* et al., 1990]. 1 : *hauts de socle ;* 2 : *isopaques du sel oligocène [d'après Curial, 1986a, b] ;* 3 : *localisation du profil ECORS.* Box 1 : *forage Bois-des-Chaux 1 ;* Br 101 : *forage Bresse 101 ;* Cmz 1 : *forage Cormoz 1 ;* Ct-D : *forage Curciat-Dongalon ;* Psx : *forage Poisoux 1 ;* Sry : *forage de St Rémy ;* Ser 1 : *forage de Sermoyer ;* Rat : *forage de Ratenelle.*

of the lithospheric plate [Karner and Watts, 1983]. Detrital deposits and tectonic imbrications of the cover contributed to the infilling of this subsident zone ; in this way, a prism was formed which thickens toward the External Crystalline Massifs [Homewood and Lateltin, 1988]. The style and distribution of Alpine deformation in this zone is inherited from previous geodynamic regimes, and is related to lithological variations in the sedimentary cover as well as the specific tectonic evolution of the European platform [Chauve *et al.*, 1980].

a) The Bresse Basin

The Massif Central and the Bresse region correspond to parts of the foreland where the crust has not been thickened by the Alpine collision. However, the Bresse Basin also belongs to the West European Tertiary rift system which extends over more than 1,000 km from the North Sea to the Mediterranean (plate I). The overall N-S elongation of the Bresse Basin, in fact, results from a juxtaposition of blocks limited by N020° – and N050° – trending faults. These blocks correspond to a succession of highs and subsident zones [Rat, 1974 and 1984] which are, from north to south (fig. I-4) : the Chalon Basin, the Sennecey-La Serre High, the Louhans Basin, the Limonest High – extending northwards to the Cormoz High – and the Bourg-en-Bresse Basin. Finally, towards the south, the Vienne-Chamagnieu High separates the Bresse Basin from the Bas Dauphiné Basin.

Several of these highs (e.g. Sennecey-La Serre, Cormoz) have trends which are similar to those of Variscan structures in the nearby Massif Central (e.g. Stephanian basin of Saint Etienne, Permian basin of Le Creusot) and may result from basement reactivation. Even before the acquisition of the ECORS profile, the Bresse Basin was considered to be an asymmetric structure [Lefavrais-Raymond, 1962 ; Sittler, 1965] with a system of major normal faults (termed the Great Border Fault) in the east and the pre-Tertiary basement shallowing progressively to the west towards the Massif Central.

Two main periods of subsidence/sedimentation affected the Bresse Basin during the Tertiary ; the first episode is Eocene-Oligocene in age and the second is late Miocene to early Pleistocene [Rat, 1984 ; Bonvalot *et al.*, 1984].

The Cenozoic basin-fill may exceed thicknesses of 1,600 m. Locally, in the Bourg-en-Bresse Basin, there are important salt deposits [Curial, 1986 a, b ; Moretto, 1985]. This sedimentary fill is now well characterized due to the drilling of numerous boreholes in the Bresse Basin (plate I). However, the geometry of the fault system was still poorly known. The location and plunges of faults at depth – including faults making up the Great Border Fault – have been generally extrapolated (fig. I-5) from surface structures or from drillhole reconnaissance surveys. Broadly speaking, drillholes in the Bresse region go no deeper than the cover and thus, do not provide any information on the deep structure of the basin.

b) The Jura

The first units of the external Jura (fig. I-6), lie tectonically on top of material filling the Bresse trough [Michel *et al.*, 1951]. Classically, this superposition is interpreted as the thrusted frontal part of the Jura cover sequence which has been detached from its basement. The decollement is thought to take place over Triassic evaporites, pushing the Jura units forward onto the Bresse Basin [Buxdorf, 1916 ; Glangeaud, 1949]. Another interpretation considers the frontal Jura units as tilted blocks. In this context, the abnormal overlap of terrains does not imply a generalized detachment of the Mesozoic cover from its basement beneath the Jura [Mugnier and Vialon, 1986]. The Jura plateaux, making up the adjoining domain, are formed of weakly deformed rocks which may have been only slightly displaced with respect to their basement. The narrow belt of deformed rock bordering the plateaux are linked to a reactivation of Oligocene normal faults as wrenches [Caire, 1963 ; Chauve *et al.*, 1980] which are associated with movements in the West European rift system.

In the more internal zones of the Jura, contraction structures become more pronounced in the direction of the Molasse Plateau and are correlated with the sliding of the cover towards the NW. The detachment of the cover is well documented [Muller and Briegell, 1980].

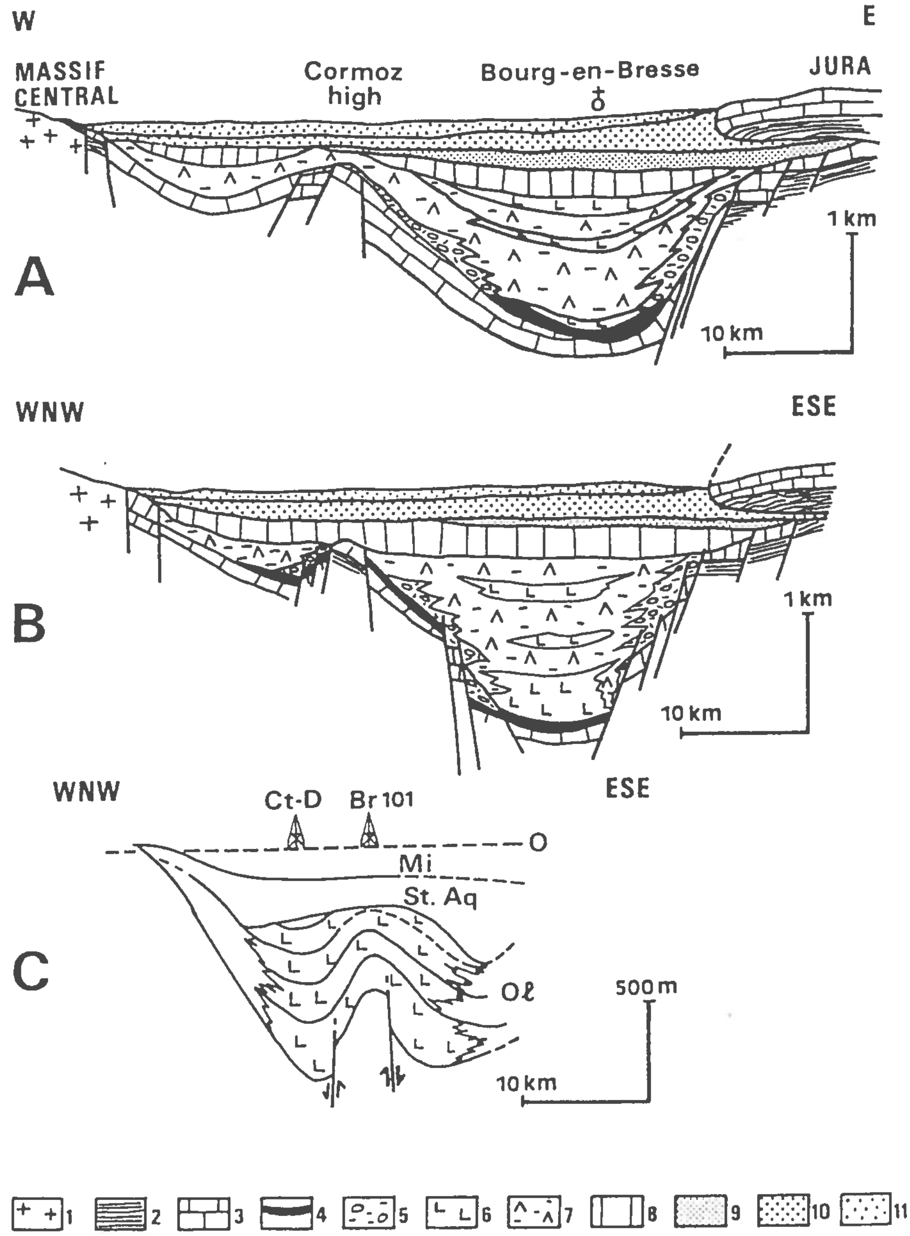


FIG. I-5. – Schematic geological cross sections in the Bresse graben before the ECORS profile [after Lefavrais-Raymond, 1962 (A); Rat, 1974 and Sittler, 1965 (B); Curial, 1986 (C)]. 1 : basement, 2 : Triassic-Liassic, 3 : Jurassic-Cretaceous, 4 : Eocene, 5, 6, 7 : Oligocene with detrital facies (5), salt (6), marls, clays and gypsum (7), 8 : lacustrine Aquitanian, 9 : marine Miocene, 10 : lacustrine Miocene (Pontian), 11 : Pliocene.

FIG. I-5. – *Coupes géologiques schématiques à travers la Bresse avant le profil ECORS [d'après Lefavrais-Raymond, 1962* (A)*; Rat, 1974 et Sittler, 1965* (B) *; Curial, 1986* (C)*]. Substratum prétertiaire :* 1 : *socle*, 2 : *Trias-Lias*, 3 : *Jurassique-Crétacé. Remplissage tertiaire :* 4 : *Eocène*, 5, 6, 7 : *Oligocène avec faciès détritique* (5), *sel* (6), *marnes, argiles et gypse* (7), 8 : *Aquitanien lacustre*, 9 : *Miocène marin* ; 10 : *Miocène lacustre (Pontien)*, 11 : *Pliocène.*

Drilling [Aubert, 1971] in this area has revealed important thrusts within the cover [Winnock, 1961]; their geometry has been partly elucidated by seismic surveys performed by oil companies. However, the recording time used in these different surveys does not exceed 5 s twt, so the seismic profiles obtained hardly penetrate beneath the Triassic succession. Despite all these difficulties in exploration, it is possible to define a tectonic style which brings together several factors which are fairly constant throughout the Jura foldbelt, i.e. :

- Mesozoic rocks composed of interbedded marl and limestone;
- sedimentary thicknesses increasing from the NW towards the SE (1,000-2,500 m);
- important erosion of the uplifted external margin, that has been emergent from the end of the Cretaceous;
- detachment of the Mesozoic cover above the Triassic salt (fig. I-7), proven at several points (e.g. the Mouthiers tunnel; drillholes situated near Lons and the Haute Chaîne),

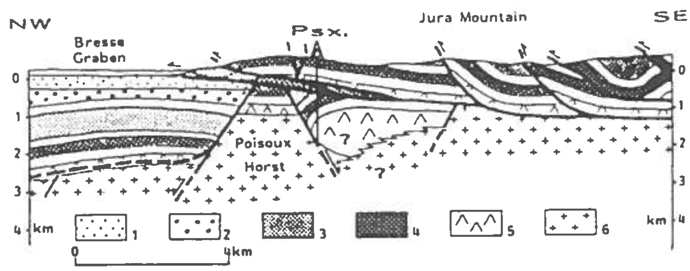


FIG. I-6. – Detailed interpreted section along the ECORS profile of the Jura Mountain front. Psx : Poisoux borehole ; 1 : Miocene ; 2 : Oligocene ; 3 : Upper Jurassic limestone ; 4 : Middle Jurassic limestone ; 5 : Upper Triassic ; 6 : basement and part of the cover beneath the décollement [after Guellec et al., 1990].

FIG. I-6. – *Coupe interprétative de détail du profil ECORS au front du Jura.* Psx : *forage de Poisoux ;* 1 : *Miocène ;* 2 : *Oligocène ;* 3 : *calcaires du Jurassique supérieur ;* 4 : *calcaires du Jurassique moyen ;* 5 : *Trias supérieur ;* 6 : *socle et couverture tégumentaire sous décollement [d'après Guellec* et al., *1990].*

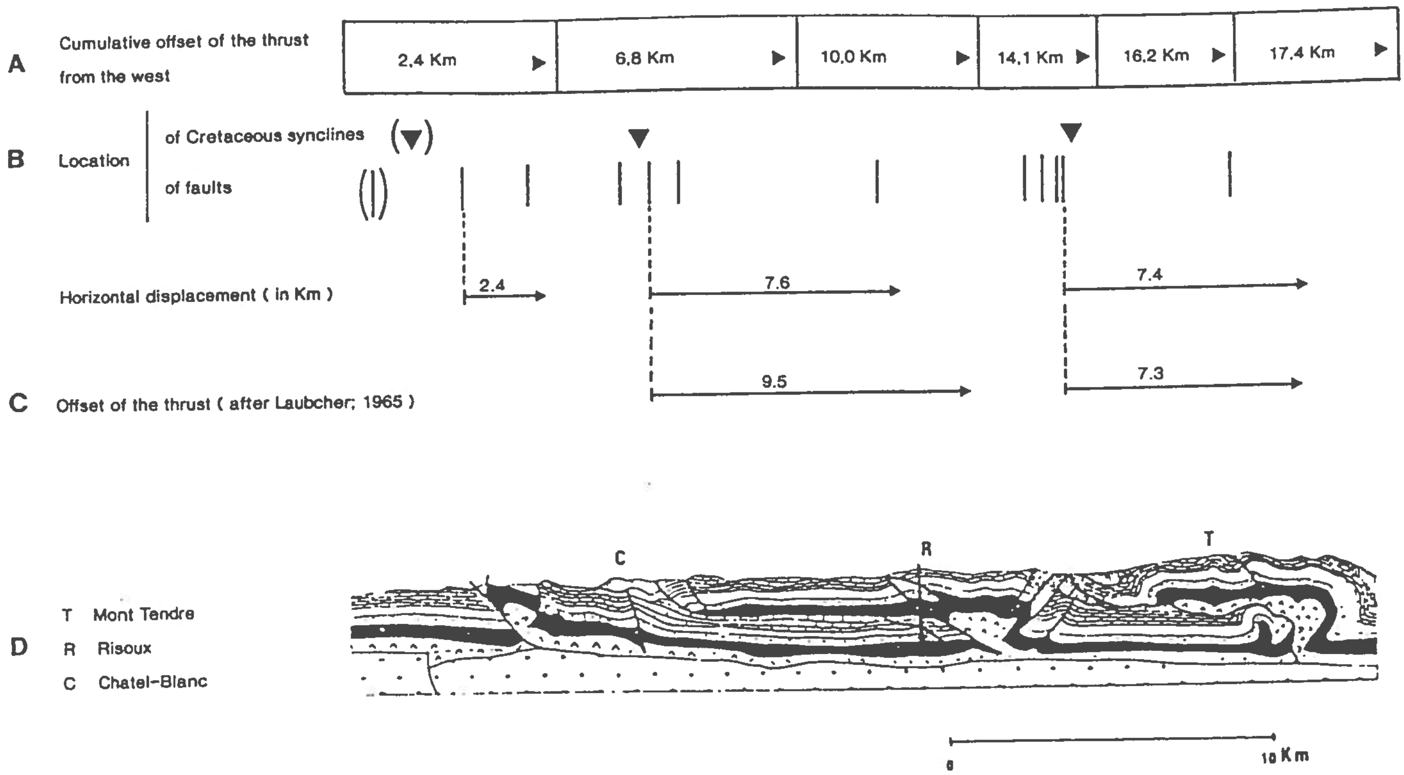


FIG. I-7. – Geologic cross-section in the inner Jura Mountains [after Laubscher, 1965].

FIG. I-7. – *Coupe géologique dans le Jura interne [d'après Laubscher, 1965].*

and presumably occurring beneath the entire Jura belt [Laubscher, 1965].

The Jura fold belt displays a prominent set of structures which merge to the south into the Subalpine ranges near the Chartreuse Massif with a large bend [Debelmas and Kerckhove, 1980; Debelmas *et al.*, 1980].

c) The Molasse Basin

The Molasse Basin, which extends in front of the Alps from Austria to France, separates the Jura Mountains from the Subalpine ranges. It consists of a segment of the European platform where flexuring is at a maximum.

The internal, Subalpine Molasse contains Oligocene continental facies sliced up at the Alpine thrust front. In the west, the external part is less deformed and is mainly composed of Miocene marine sediments [Homewood *et al.*, 1985].

The thickness and generally prism-like geometry of the detrital deposits lead to a rigid behaviour [Laubscher, 1961] for the cover sequence in this zone, which is made up of molasse and underlying Mesozoic rocks. As the rigid cover unit sled over the pre-Triassic basement, it indented against the internal units of the Jura belt [Vialon *et al.*, 1984; Philippe, 1994], which were more readily deformed because of stratigraphic thinning and the absence of molasse type deposition in the west [Mugnier and Vialon, 1986]. The occurrence of molasse between the Jura and the Alpine front fades away progressively to the south of Geneva, thus emphasizing the difference in structural organization which exists between the northern branch of the western Alps and the Alps of Dauphiné and Provence.

d) The Subalpine zones

The northern Subalpine zone exhibits regular SW-NE trending folds (fig. I-8) that are slightly oblique to the Belledonne-Mont Blanc-Aar crystalline axis. These folds are recumbent and overthrust toward the NW. The intensity of deformation increases transversally towards the basement massifs, and also along the axial traces of structures in a northeasterly direction. Formations are generally more marly near the basement. These rocks show schistose metamorphic textures which are well developed in equivalent units occurring in the upper valley of the Rhône in Switzerland [Gratier and Vialon, 1980]. Locally these microstructures evidence a top to the south east-verging backthrust.

The upper parts of the Subalpine units are capped by *klippen*. These erosional remnants of the highest tectonic units, become more and more important in a northeasterly direction from about the longitude of Annecy towards Sulens, Annes and the Pre-Alps. Beyond the Pre-Alps, along the zone of frontal molasse, there is a continuous superposition of Subalpine and internal thrust sheets.

To the south of Grenoble, the southern Subalpine zone widens out and is no longer directly related to the outcrop pattern of basement massifs (fig. I-8).

In a transition zone, folds are at first broadly N-S trending, but less regular and tight than folds observed in the northern Subalpine zone. Farther south, folds become affected by major bends which are superimposed on early E-W folds occurring beneath the late Cretaceous unconformity. Other features are added to this already complex tectonic pattern by the Mio-Pliocene detachment of the Digne nappe and the activity of its limiting tear-faults.

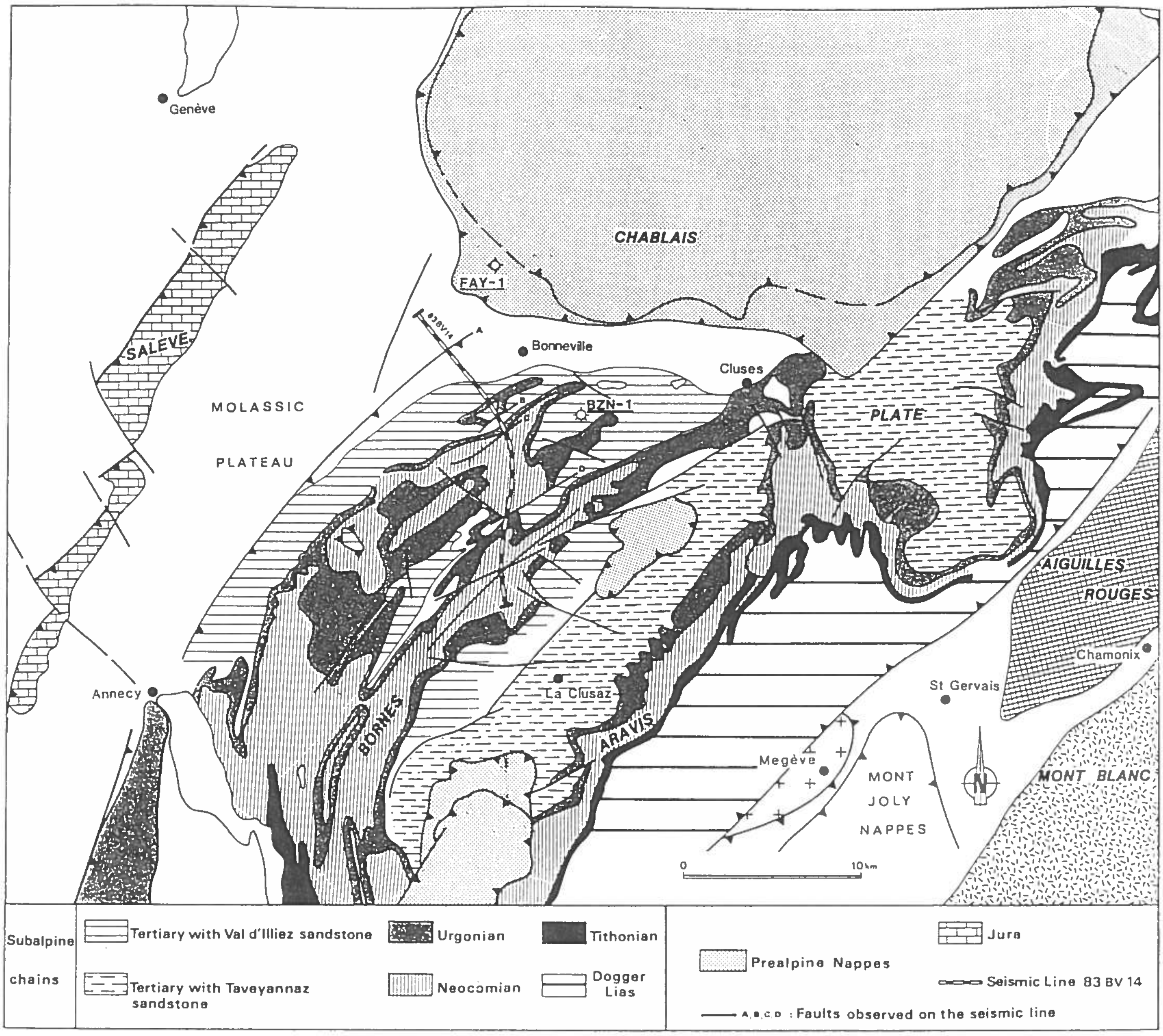


FIG. I-8. – Structural map of the Bornes Massif [after Charollais and Jamet, 1990]. Bzn 1 : Brizon 1 borehole, Fay 1 : Faucigny 1 borehole.

FIG. I-8. – *Carte structurale du massif des Bornes [d'après Charollais et Jamet, 1990].* Bzn 1 : *forage de Brizon 1 ;* Fay 1 : *forage de Faucigny 1.*

These displacements towards the west or south are partly due to the outward movements of flysch nappes which involved slices of the Briançonnais zone in the SW of the area between Pelvoux and Argentera. There, the Penninic thrust emplacement favoured dextral shear faults which variably affected those structures parallel to the trend of the External Crystalline Massifs [Ricou, 1984].

A further complexity derives from the fact that the foreland in the vicinity of Die (Drôme) lacks any post-late Jurassic (including Urgonian) limestone. The passage from the southern Subalpine zone to the Provencal domain occurs in a context where E-W trending structures are pointed out by major limestone units which, once again, make up a predominant part of the succession.

The acquisition of an ECORS profile in the foreland has enabled us to address the following problems :

• the geometry of the Tertiary subsident segment in the Bresse area ;

• the search for Stephanian basins associated with major Variscan crustal shears ;

• the amount of detachment involved in the Mesozoic cover during Alpine compression.

FIG. I-10. – Structural and palinspastic cross-sections in the Sub-Alpine Massifs [after Gratier *et al.*, 1989].

FIG. I-10. – *Coupes structurales et palinspastiques à travers les massifs sub-alpins [d'après Gratier* et al., *1989].*

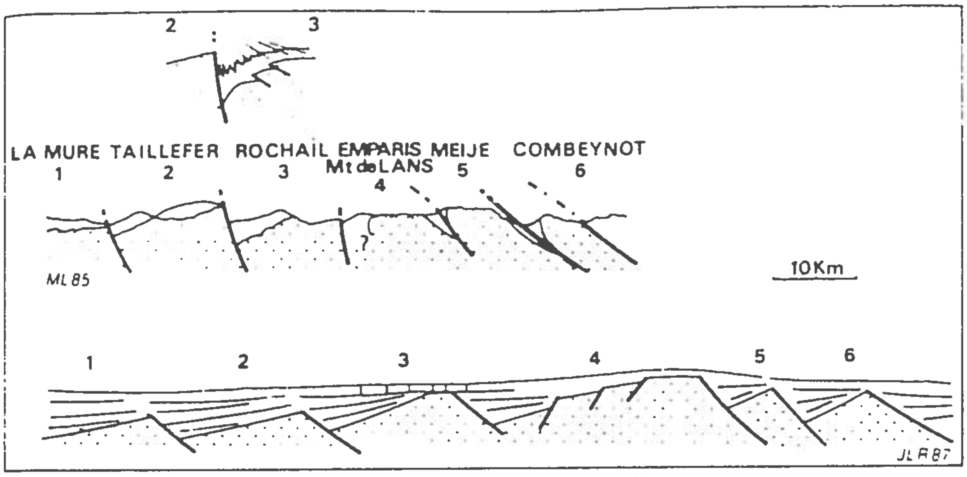
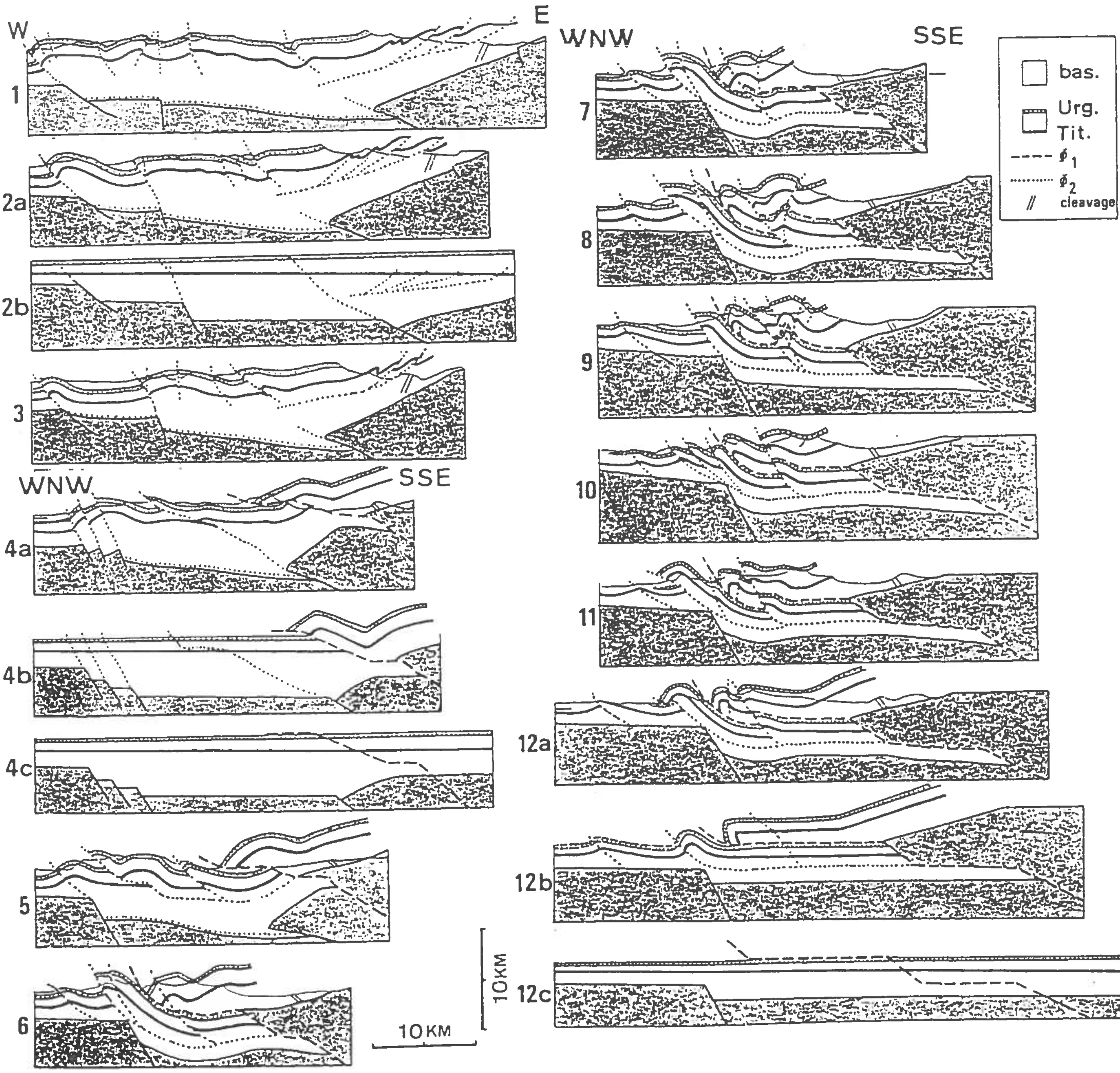
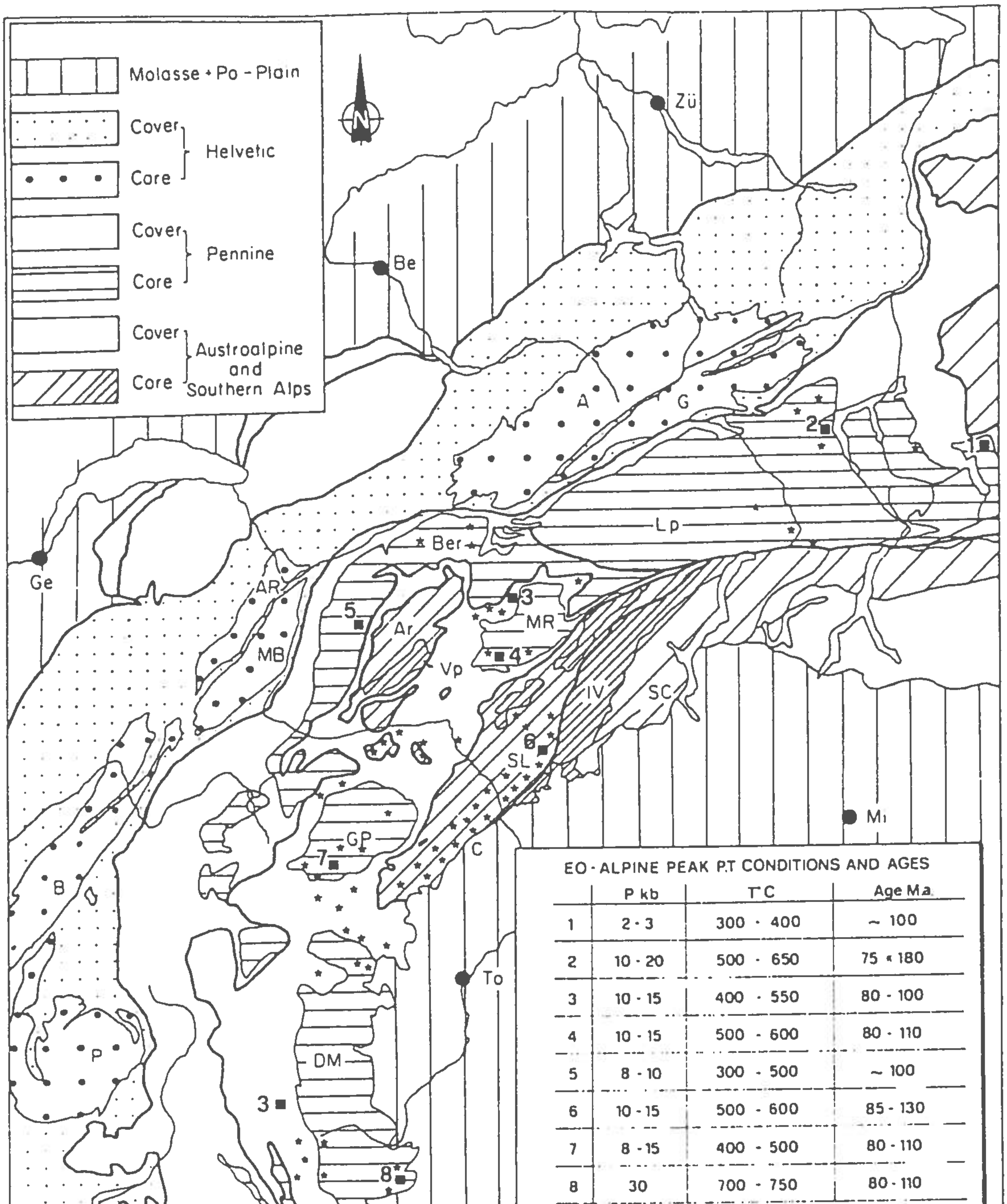


FIG. I-9. – Palinspastic section in the western Alps [after Rudkiewicz, 1988]. Note the present geometry of tectonically inverted Jurassic tilted blocks.

FIG. I-9. – *Coupe palinspastique des Alpes occidentales [d'après Rudkiewicz, 1988]. Noter la géométrie actuellement inversée des blocs basculés jurassiques.*





EO-ALPINE PEAK P.T CONDITIONS AND AGES

	P kb	T°C	Age Ma.
1	2 - 3	300 - 400	~ 100
2	10 - 20	500 - 650	75 « 180
3	10 - 15	400 - 550	80 - 100
4	10 - 15	500 - 600	80 - 110
5	8 - 10	300 - 500	~ 100
6	10 - 15	500 - 600	85 - 130
7	8 - 15	400 - 500	80 - 110
8	30	700 - 750	80 - 110

FIG. I-11. – Tectonic sketch map of the central and western Alps. The numbers refer to the locations of rock units in high-pressure facies radiometrically dated as Eoalpine [after Hunziker *et al.*, 1989]. Peak P-T conditions and ages are shown in the inset table. Stars indicate localities of eclogites. Abbreviations : A = Aar Massif; Ar = Arolla series; AR = Aiguilles Rouges Massif; B = Belledonne Massif; Be = Berne; Ber = northern Bernhard; C = Canavese; DM = Dora-Maira Massif; G = Gotthard Massif; Ge = Geneva; GP = Gran Paradiso Massif; IV = Ivrea-Verbano zone; Lp = Lepontine area; MB = Mont Blanc Massif; Mi = Milan; MR = Monte Rosa Massif; P = Pelvoux Massif; SC = Strona-Ceneri zone; SL = Sesia-Lanzo zone; To = Turino; Vp = Valpelline nappe; Zü = Zürich.

FIG. I-11. – Carte tectonique des Alpes occidentales et centrales. Les chiffres font référence aux localisations des unités de faciès haute-pression datées radiométriquement de l'Eoalpin [d'après Hunziker et al., *1989]. Les conditions P-T extrêmes et les âges correspondants sont indiqués dans le tableau. Les étoiles indiquent la localisation des éclogites. A = massif de l'Aar; Ar = séries de l'Arolla; AR = massif des Aiguilles Rouges; B = massif de Belledonne; Be = Berne; Ber = Gd St Bernard; C = Canavese; DM = massif de la Dora Maira; G = massif du St Gotthard; Ge = Genève; GP = massif du Grand Paradis; IV = zone d'Ivrée-Verbano; Lp = zone lépontine; MB = massif du Mt Blanc; Mi = Milan; MR = massif du mt Rose; P = massif du Pelvoux; SC = zone de Strona-Ceneri; SL = zone de Sesia-Lanzo; To = Turin; Vp = nappe valpelline; Zü = Zürich.*

2) The zone of crustal thickening

The zone of crustal thickening is characterized at the surface by elevated relief. On the basis of palaeogeographic and metamorphic criteria, this zone is classically subdivided into two parts.

a) The External Crystalline Massifs

The elevated relief of the External Crystalline Massifs follows a regular alignment corresponding to the Belledonne-Mont Blanc-Aar axis in the north and a less well defined trend associated with the Pelvoux and Argentera massifs in the south. These massifs are composed of Variscan and pre-Variscan metamorphic and granitic rocks which sometimes enclose Permo-Carboniferous sediments bounded by major Variscan faults. These rocks, similar to formations in the Massif Central, represent the continental basement which underlies sediments on the European margin of the Alpine Ocean.

Basement outcrops of the Subalpine zone occurs in faulted blocks. Numerous basement thrust units pinch out against their cover, causing its displacement towards the NW. This basement sometimes displays a tectonic pattern inherited from the period of Tethyan rifting, i.e. half-grabens separated by horsts where the cover remains more or less attached (fig. I-9). In most cases, the basement is upthrusted and shows Alpine metamorphic effects, that increases towards the northeast. On the eastern flanks of these crystalline massifs, thrust-slices are frequently steeply dipping and even overturned to the east.

Thrusting phenomena persist in cover rocks occurring beyond the internal flanks of the crystalline massifs and are comparable to effects seen in the more external parts of the Subalpine cover. This thrusted complex represents one of the main sutures of the Alpine belt and forms the boundary between the external and internal zones (fig. I-10).

b) Axial metamorphic belt (AMB)

The axial part of the Alpine chain is bordered by two major tectonic structures, the Penninic thrust front to the west and the Canavese line to the east. This part of the belt was affected by the well known HP/LT (high pressure/low temperature) mineral assemblages developed during the Cretaceous-Palaeogene metamorphic phases and ductile deformations of the Alpine orogeny.

The first evidence of European lithosphere flexing or deformation is recorded by the middle to late Cretaceous or Palaeocene turbiditic sequences, which lie either on oceanic crustal units [Lemoine, 1984; Lagabrielle, 1987] or on the Apulian (Adria) margin [Castellarin, 1977; Doglioni and Bosellini, 1987; Caron *et al.*, 1989]. Moreover, since at least Turonian time, fragments of Alpine HP/LT metamorphic rocks (blueschist) and Mesozoic sediments, as well as debris of south Alpine basement sourced from the hinterland, are reworked in the terrigenous flysch sandstones which lie along or directly upon the northern border of the Apulian foreland [Winkler and Bernoulli, 1986; Bischel and Haring, 1981].

The AMB is made up of an ophiolitic and basement nappe pile, largely thrust toward the north and west onto the European foreland. Metamorphic facies range from high-P/low-T eclogite or blueschist assemblages to H-T greenschist or amphibolitic assemblages. Along the western Alpine ECORS-CROP traverse, only the upper part of the nappe pile is exposed. Rocks, that are similar to those outcropping in the central Alps, are supposed to exist at deeper levels if we assume a lateral prolongation of the major structural trends (fig. I-11). Classically, each geometrically consistent segment (e.g. Valais, Briançonnais, Piemontais, southern Alps, etc.) is related to its own palaeogeographic domain in models using simple unfolding or "card-deck" deimbrication of the nappe pile to achieve kinematic reconstructions [Trumpy, 1980; Lemoine, 1984; de Graciansky, 1993].

The AMB has been interpreted in terms of present-day geodynamic environments. It is now thought to result from the formation of an orogenic wedge in which coupling of basement and ophiolitic nappes was ensured at different crustal levels by the operation of a unique long-lasting subduction zone at the Adria plate margin.

In this interpretation, the axial high-P/low-T metamorphic belt is completely detached from the other parts of the orogen and has been considered as the deep remnant of a Cretaceous orogenic crustal wedge, implying that the major lithospheric suture emerges as the Penninic thrust front [Polino *et al.*, 1990].

3) The Po Plain

The Po Plain is composed of a 500 km long E-W trending Neogene basin, bordered to the north by the south-directed nappes of the southern Alps, and to the south by north-directed structures of the Apennines (fig. I-12). Due to the thick Neogene infill of the Apennines foredeep, the geometry and the extent of compressive structures beneath the Po Plain can be assessed only by subsurface methods, including exploration wells and conventional seismic surveys [Pieri and Groppi, 1981; Pieri, 1983; Pieri and Mattavelli, 1986; Castellarin *et al.*, 1986].

The ECORS-CROP profile crosses the western part of the Po Plain from Canavese to Monferrato (plate I), but the Apenninic thrust front is not actually intersected. This segment has been partially characterized by oil exploration reconnaissance (Sali Vercellese and San Benigno Canavese wells); [AGIP, 1977] (fig. I-13) and belongs, in fact, to part of the flexural basin of the Apennines. The Plio-Quaternary deposits of this zone are well known from the work of Pieri and Groppi [1981], Ménard [1988], Ricci Lucchi [1985 and 1986], Rizzini and Dondi [1978] and Cassano *et al.* [1986], but older parts of the succession are masked.

a) Geophysical background

Oil exploration has been very active in the Po Plain. Hundreds of kilometres of conventional seismic surveys, gravimetric and magnetic surveys, as well as modelling correlated with deep wells – have been published [AGIP, 1977; Errico *et al.*, 1980; Pieri and Groppi, 1981; Pieri, 1983; Cassano *et al.*, 1986]. Tectonic and structural reconstructions [Castellarin, 1981; Castellarin *et al.*, 1988], as well as neotectonic studies, seismotectonic models and focal-mechanism determinations have given some constraints on the deep crust of the south Alpine belt [Zanferrari *et al.*, 1982; Bartolini *et al.*, 1988; Slejko *et al.*, 1987]. Deep refraction measurements have added consistent informations about the Moho geometry beneath and around the Po Basin [Gebrande *et al.*, 1978; Giese *et al.*, 1982; Müller *et al.*, 1982; Deichman *et al.*, 1986], especially those performed during the European Geotraverse Project [Thouvenot *et al.*, 1985; Giese, 1985; Nadir, 1988; Blundell *et al.*, 1992]. These data have confirmed the presence of a very complex Moho topography beneath the Po Plain, with the presence of several reflecting interfaces and a distinct deepening of the Moho towards the internal Apenninic and Ligurian domains. Finally, the deep

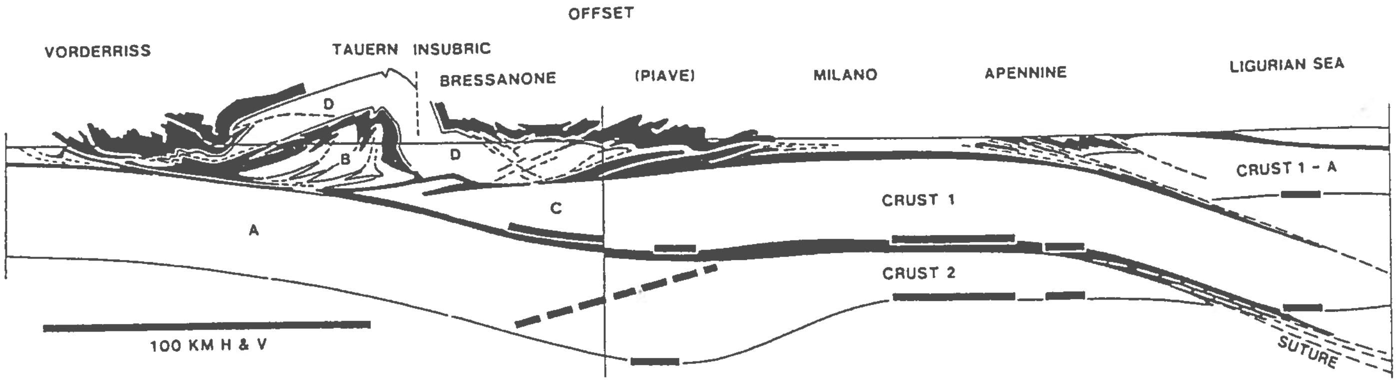


FIG. I-12. – Crustal cross-section of eastern Alps and Po Plain [after Roeder, 1989]. Vertical line is an E-W offset of 250 km. Shaded : crust; black : Mesozoic sediments. Present topography is not shown. M : Moho.
FIG. I-12. – *Coupe crustale à travers les Alpes orientales et la Plaine du Pô [d'après Roeder, 1989]. La ligne verticale marque un déport latéral de 250 km. En grisé : croûte; en noir : sédiments mésozoïques. La topographie actuelle n'est pas indiquée.* M = *Moho.*

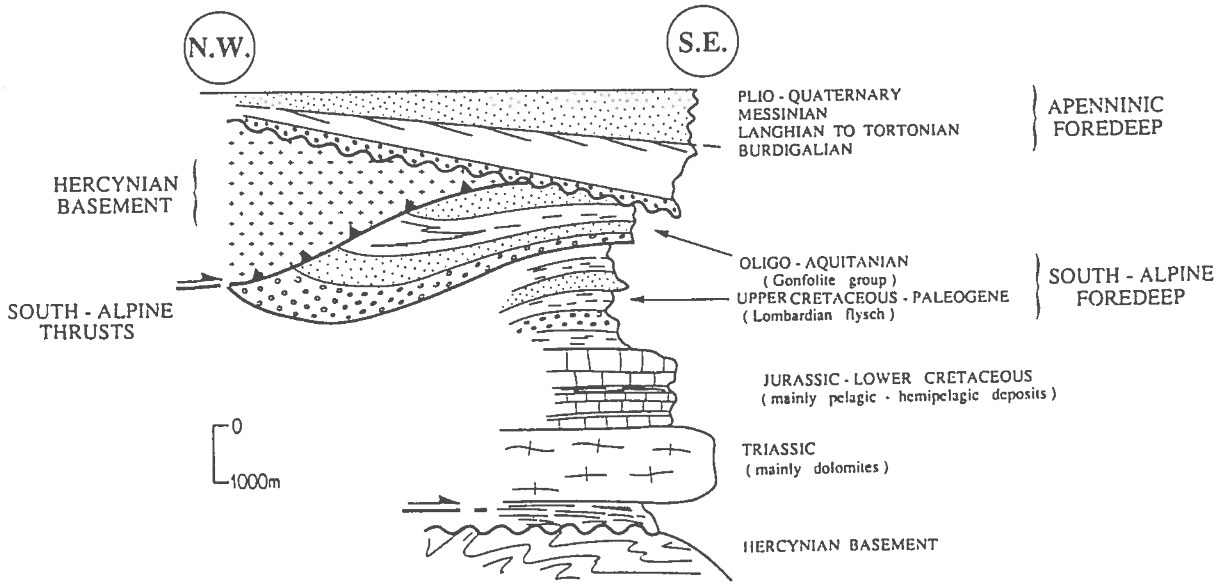


FIG. I-13. – Lithologic log of the Apulian/Insubric sedimentary sequence in the southwestern Alps [after Roure *et al.*, 1990].
FIG. I-13. – *Colonne litho-stratigraphique des séries sédimentaires apuliennes/insubriennes des Alpes sud-occidentales [d'après Roure* et al., *1990].*

seismic reflection results of the ECORS-CROP and NFP-20 profiles across the western and central Alps have given the most outstanding information about the crustal structure of the area [Bayer and ECORS Team, 1987; ECORS-CROP DSS Group, 1989; Rey *et al.*, 1990]. Results of the ECORS-CROP profile beneath the Po Plain are presented and discussed in the following chapters.

b) Stratigraphic evolution of the Po Plain region

The evolution of the Po Plain is characterized by two chronologically distinct steps, directly related to the evolution of the bordering Alpine and Apennine chains. The first episode occurred from Upper Cretaceous to Oligocene and relates to the south-vergent Alpine backthrusts, whereas the second episode relates to the Apennine fronts and thus, is post-Burdigalian in age.

Due to the loading of early emplaced south-directed nappes in the south Alpine realm, a flexural foredeep basin progressively developed along the southern margin of the collisional belt on the Apulian plate. Its sedimentary infill can still be observed on northern side of the Po Plain and, evidently, in the wells that have been drilled in the Po Plain itself [Sali Vercellese, AGIP, 1977; Pieri and Groppi, 1981] (plate II). There is some uncertainty as to whether the Cretaceous Lombardian flysch was derived from the erosion of early south Alpine structures. However, younger syntectonic sediments are clearly linked to south-directed thrusts of late Palaeogene-early Neogene age loading the Apulian foreland (Oligo-Aquitanian Gonfolite facies; [Forcella and Rossi; 1980; Andreoni *et al.*, 1981; Gunzenhauser, 1985; Gelati *et al.*, 1988; Bernoulli *et al.*, 1989]). They consist of deep-water turbidites and other gravity-flow deposits. Isopach maps of these terrigenous beds [Ménard, 1988; Bersezio and Fornaciari, 1988, 1989] outline the geometry of the late Palaeogene south Alpine foredeep confined to the northwestern border of the Po Plain (fig. I-14). In the early Miocene (Burdigalian to Langhian), the south Alpine foredeep migrated southeastward and was centered near Milano (fig. I-15) [Ménard, 1988]. Farther east, north of

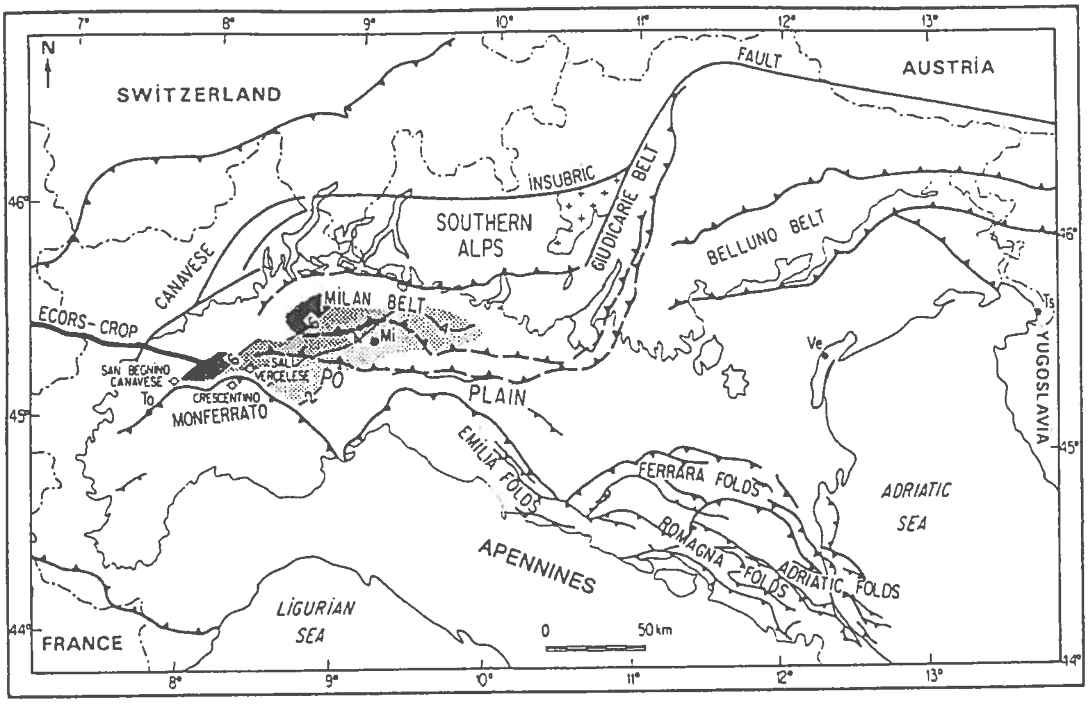


Fig. I-14. – Isopach map of the late Palaeogene south Alpine foredeep (Gonfolite facies) [after Ménard, 1988]. This map is not corrected palinspastically for later thrusts.

Fig. I-14. – *Carte isopaque du Paléogène terminal de l'avant-fosse sud-alpine (faciès gonfolitique) [d'après Ménard, 1988]. Cette carte ne prend pas en compte les rejeux ultérieurs de chevauchements.*

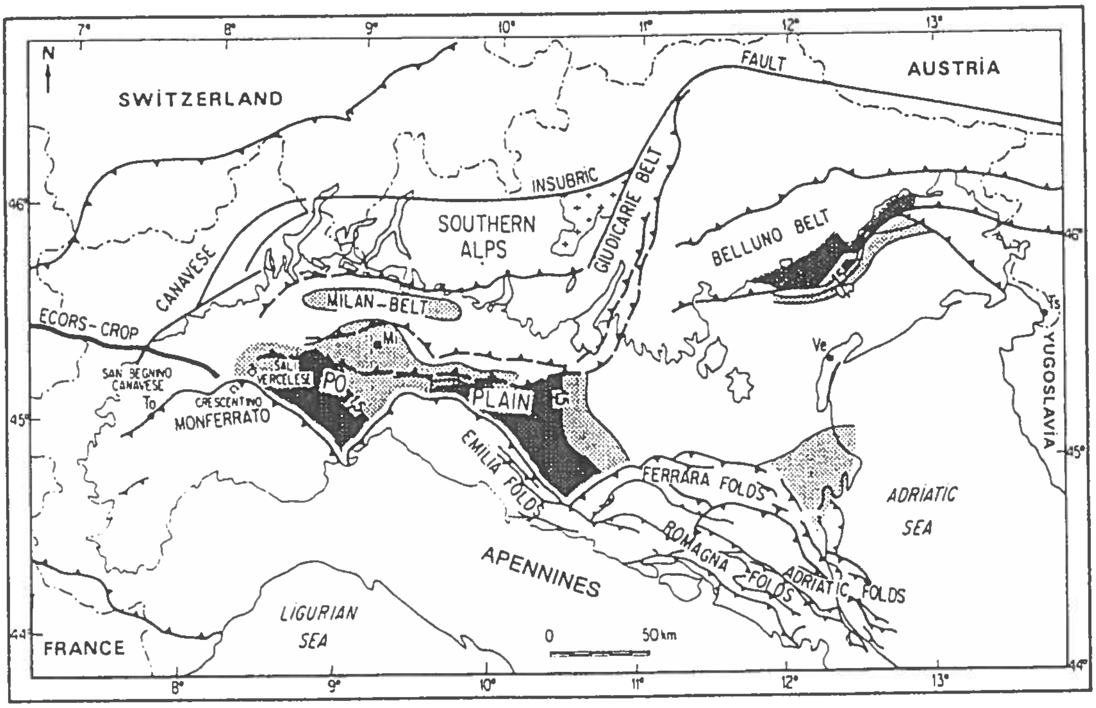


Fig. I-15. – Isopach map of the Middle Miocene south Alpine foredeep [after Ménard, 1988]. This map is not corrected palinspastically for later thrusts.

Fig. I-15. – *Carte isopaque du Miocène moyen de l'avant-fosse sud-alpine [d'après Ménard, 1988]. Cette carte ne prend pas en compte les chevauchements ultérieurs.*

Venice, late Miocene to Pliocene terrigenous sequences were still trapped in a remnant of the south Alpine foredeep (fig. I-16).

The present-day Po Plain reflects the extension of the Messinian to Quaternary Apenninic foredeep, in which the infill thickens southward [Pieri and Groppi, 1981]. Beneath

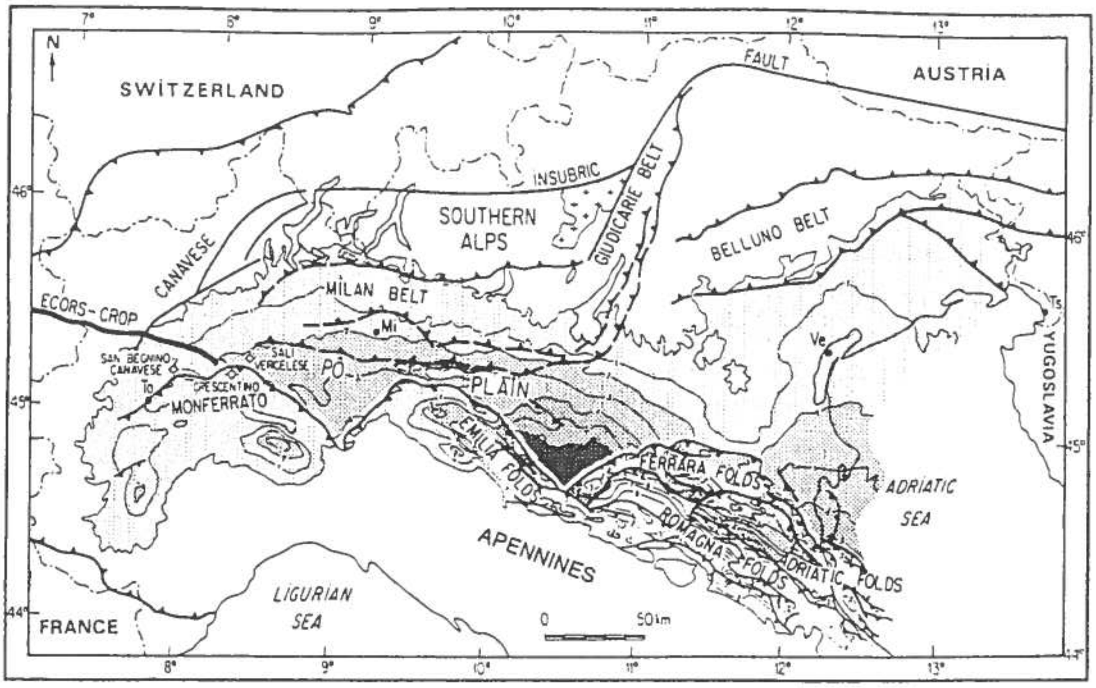


FIG. I-16. – Isopach map of the Messinian to Quaternary Apenninic foredeep [after Pieri and Groppi, 1981].
FIG. I-16. – *Carte isopaque des séries messiniennes à quaternaires de l'avant-fosse apenninique [d'après Pieri et Groppi, 1981].*

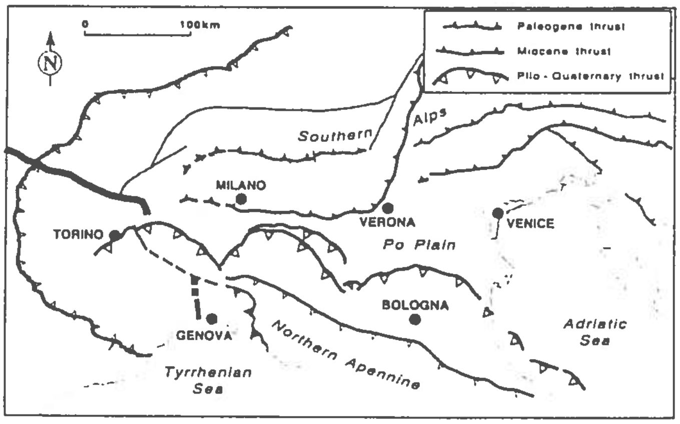


FIG. I-17. – Structural sketch of the Po plain.
FIG. I-17. – *Schéma structural de la plaine du Pô.*

a general Messinian erosional surface [Rizzini and Dondi, 1978], a locally more restricted Miocene (mainly Langhian to Tortonian) sedimentary prism can be mapped in the western Po Plain [Ménard, 1988; Ricci-Lucchi, 1985]. This location may also be related to early Apenninic deformation.

c) Tectonic framework of the Po Plain

Despite the fact that the south Alpine thrust front usually remains buried beneath the Plio-Quaternary sediment fill of the Po Plain, the timing of south-verging thrusting is well documented by surface and subsurface data. In the Brescian Alps, pre-late Eocene thrusts have been identified [Brack, 1981; Castellarin, 1992 and ref. therein] (fig. I-17). It is possible to correlate these early deformations with the development of a first south Alpine foredeep in the Eocene (Ternate Formation) [Bernoulli *et al.*, 1988] and Oligocene [Gonfolite trough]. Farther south in the Milan belt, the deformation affects beds as young as the Middle Miocene (Langhian to Serravallian) in the Malossa oil field [Pieri and Groppi, 1981; Errico *et al.*, 1980; Chierici *et al.*, 1979] and possibly also in the Como area [Bernoulli *et al.*, 1989]. North of Venice, in the Belluno belt and in Friuli, the Miocene Alpine deformation induced the development of a late Neogene foredeep, and shortening was still active there in Plio-Quaternary times [Zanferrari *et al.*, 1982; Slejko *et al.*, 1987].

Deeply buried beneath late Neogene sequences, the north-directed Apenninic thrust front can be followed by subsurface data all along the southern border of the Po Plain, from the Monferrato belt to the Adriatic Sea (figs. I-18 and I-19) [Piana and Polino, 1995]. The emplacement of these Apenninic structures results mainly from post-Messinian shortening. It induced the development of the Alpine foothill monocline, a subsurface expression of the Apenninic flexure, outlined by the progressively southward deepening of the Messinian erosional or evaporitic surface [Pieri and Groppi, 1981; Rizzini and Dondi, 1978; Ricci-Lucchi, 1986]. In addition to this younger evolution of the Apennines, earlier deformations are documented by the occurrence of late Burdigalian or Langhian to Tortonian terrigenous sequences (Formazione Marnose Arenacea) which reworked pebbles derived longitudinally from either basement or Alpine basinal units, and may represent the distal sedimentary infill of an early foredeep [Ricci-Lucchi, 1986].

Due to the lack of subsurface data, the geometric relationships between the Apennine and the south Alpine structures beneath the western part of the Po Plain remained highly conjectural before ECORS-CROP. Nonetheless, the

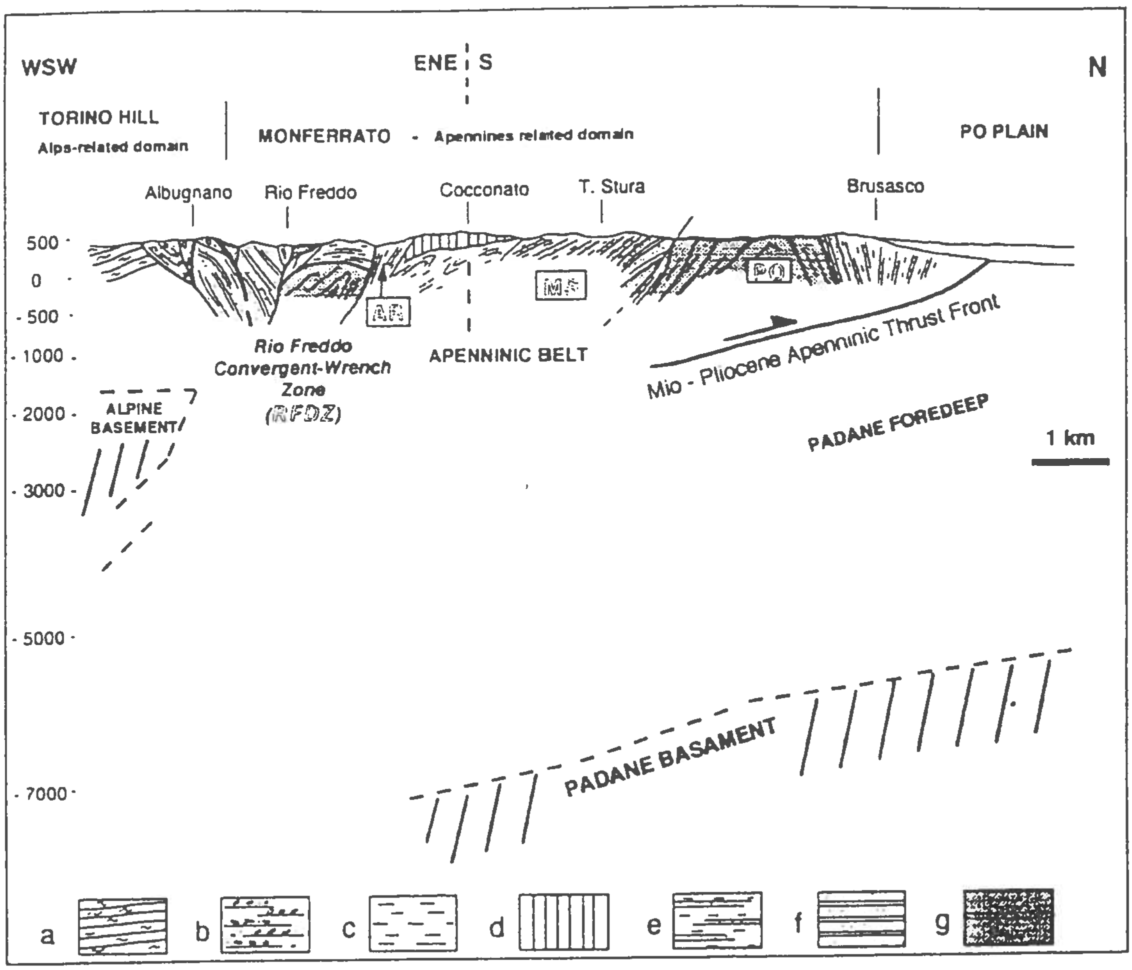


FIG. I-18. – Structural section across the Torino hills [after Piana and Polino, 1995].

FIG. I-18. – *Coupe transversale à travers les collines de Turin [d'après Piana and Polino, 1995].*

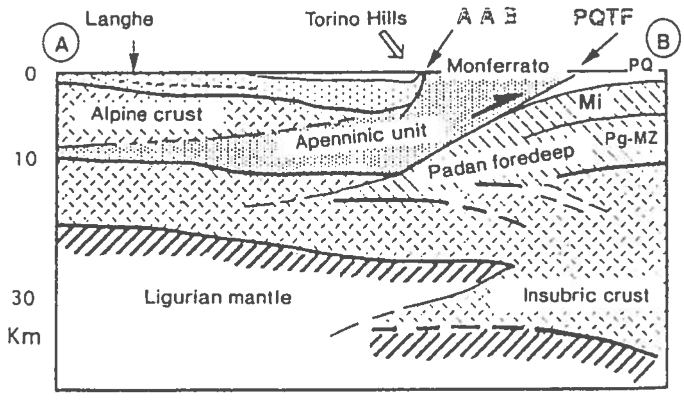


FIG. I-19. – Crustal scale cross section in the northern Apennines [after Polino *et al.*, 1995].

FIG. I-19. – *Coupe structurale aux confins de l'Apennin septentrional et de la plaine du Pô [d'après Polino* et al., *1995].*

occurrence of a major post-Aquitanian unconformity in the Monferrato area, separating deep-water Palaeogene flysch and shallow-water Miocene limestones [Gelati, 1968; Gelati and Gnaccolini, 1982; Lorenz, 1969; Haccard *et al.*, 1972], is an indication of early tectonic events in this domain, related either to Alpine or Apenninic evolution.

I.C. – THE DEEP STRUCTURE OF THE WESTERN ALPS: OUR STATE OF KNOWLEDGE PRIOR TO ECORS-CROP OPERATIONS

Studies of the deep structure of the Alpine arc and its bordering domains (figs. I-20 to 23) began at the end of the 1950's [Closs and Labrouste, 1963; Choudhury *et al.*, 1971; Perrier, 1973; Perrier and Ruegg, 1973; Giese and Prodhel, 1976]. As far as the Bresse, the Jura and the western part of the arc are concerned, investigations have been based on regional profiles which were parallel [Thouvenot, 1976; Michel, 1978; Ansorge *et al.*, 1979; Thouvenot and Perrier, 1981] or transverse [Thouvenot *et al.*, 1985] to the belt. The most important results that stand out from this mass of data are as follows.

1) Location of the pre-Triassic basement in the foreland and Po Plain areas

On the French side of the belt, basement is locally inferred from anomalies in the seismic wave velocities of the upper part of the crust [Ménard, 1979]. Late arrivals observed in the Subalpine domain have been interpreted as reflections from the base of the Mesozoic basin fill at 6-8 km depth. This peri-Alpine SW-NE-trending depression borders the northwestern External Crystalline Massifs (Belledonne and Aiguilles Rouges) and extends in front of the Aar Massif, where its existence is proved by drilling. The

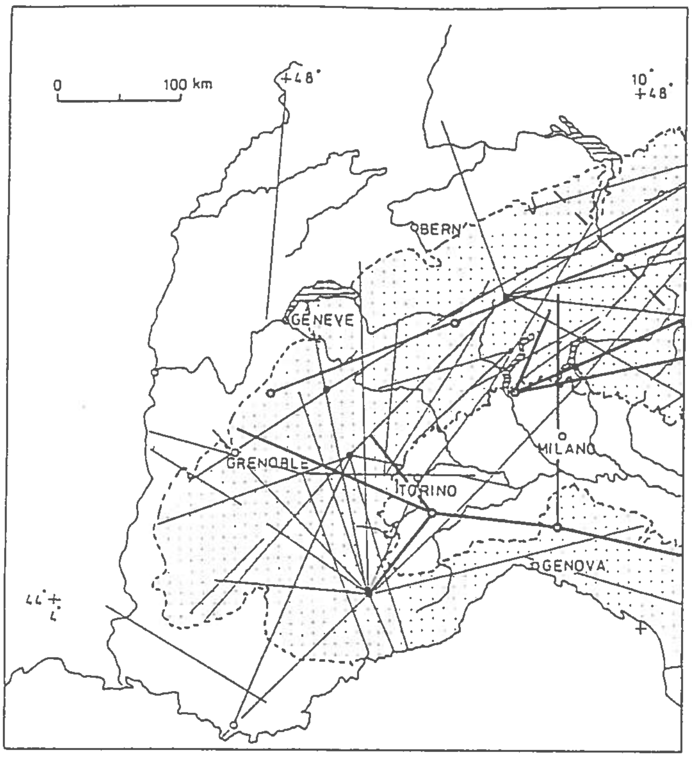


FIG. I-20. – Position of Deep Seismic Sounding profiles in the Alps : thin lines = before 1975 ; thick lines = after 1975 [after Giese and Prodehl 1976, updated].

FIG. I-20. – Position des grands profils alpins : traits maigres = antérieurs à 1975 ; traits gras = postérieurs à 1975 [d'après Giese et Prodhel 1976, complété].

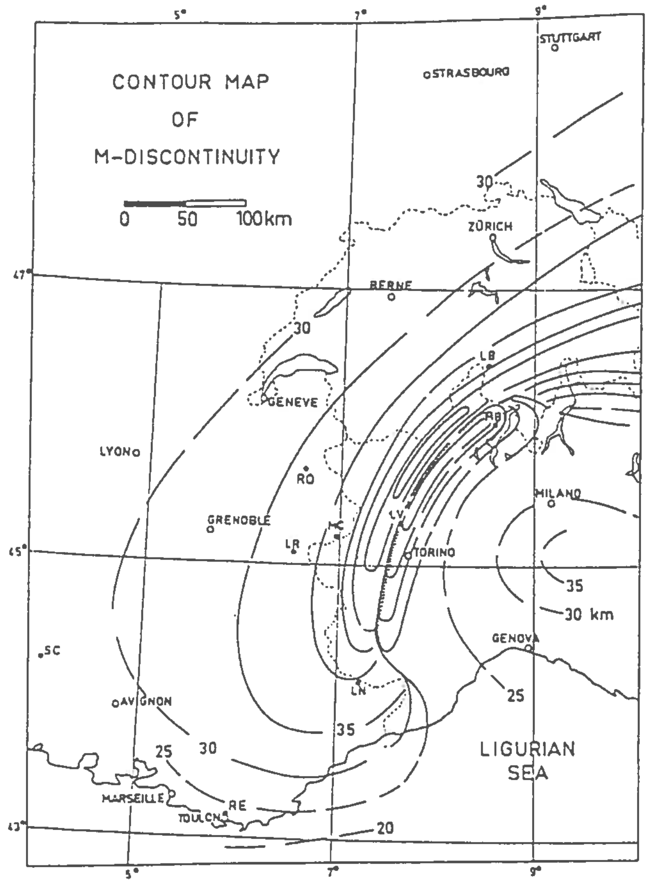


FIG. I-21. – Contour map of the Moho discontinuity in the western Alps [after Giese and Prodhel 1976].

FIG. I-21. – Schéma général de la profondeur du Moho sous les Alpes occidentales [d'après Giese et Prodehl 1976].

deepening of the basement in this zone has been interpreted as being related to frontal thrusting (towards the NW) of these external massifs, due to intracrustal decoupling (see section (5) below).

On the Italian side, the location of crystalline basement beneath the Po Plain has been derived mainly from magnetic anomalies, which suggest a depth of about 8-10 km. However, it should be noted that reflectors at this depth have never been detected during AGIP seismic reflection surveys. Approaching the belt, the basement rises rapidly and appears to be affected by marked compressive features ; in places, its structure is complicated by frequent igneous intrusions of Permian and Cenozoic age.

2) Alpine foreland crust (Massif Central, Bresse and Jura)

The Alpine foreland appears to have well-differentiated upper and lower crust with very stable average velocities (6.0 and 6.7-6.9 kms.s^{-1}, respectively). Their thicknesses vary from place to place and are of the order of 20 and 7-10 km, respectively.

3) Crustal thickening beneath the axial zone

Crustal thickening occurs vertically beneath the regional gravimetric minimum, i.e. offset by a few tens of kilometres from the outcrop of the External Crystalline Massifs which constitute the highest relief. The maximum thickness (50-60 km) of this root zone is poorly known because of the limited amount of data and the uncertainty on the average crustal velocity. The upper mantle has a "normal" velocity of 8.1-8.2 km.s^{-1}. More towards the convex side of the arc, beneath the External Crystalline Massifs, the Moho is rather shallower (35-40 km). However, it is still unclear how these variations in crustal thickness have come about (cf. « smooth » or « brittle » models for the Moho, see section 7).

4) The Ivrea body

The positive gravimetric anomaly which follows the border of the Po Plain from Locarno to Cuneo has long been interpreted as an uprised body of basic and ultrabasic rocks (N.B. the Ivrea body is **not** the equivalent of Ivrea zone rocks cropping out at the surface).

This anomalous structure (velocities of 7.4 km.s^{-1} at depths of 5-10 km) was first revealed by early seismic refraction studies in the Alps. Very late reflected wave arrivals were also observed on these profiles. In the most widely accepted interpretation, these late arrivals correspond to reflections from an extremely deep Moho (at least 50 km) combined with a velocity inversion beneath the Ivrea body. This inversion (down to velocities of 5 km.s^{-1}) is poorly constrained because it is difficult to establish the extension at depth of the Ivrea body. During recent surveys, these two seismic events – caused by refraction on the Ivrea body and reflection from the Moho – have both been clearly identified.

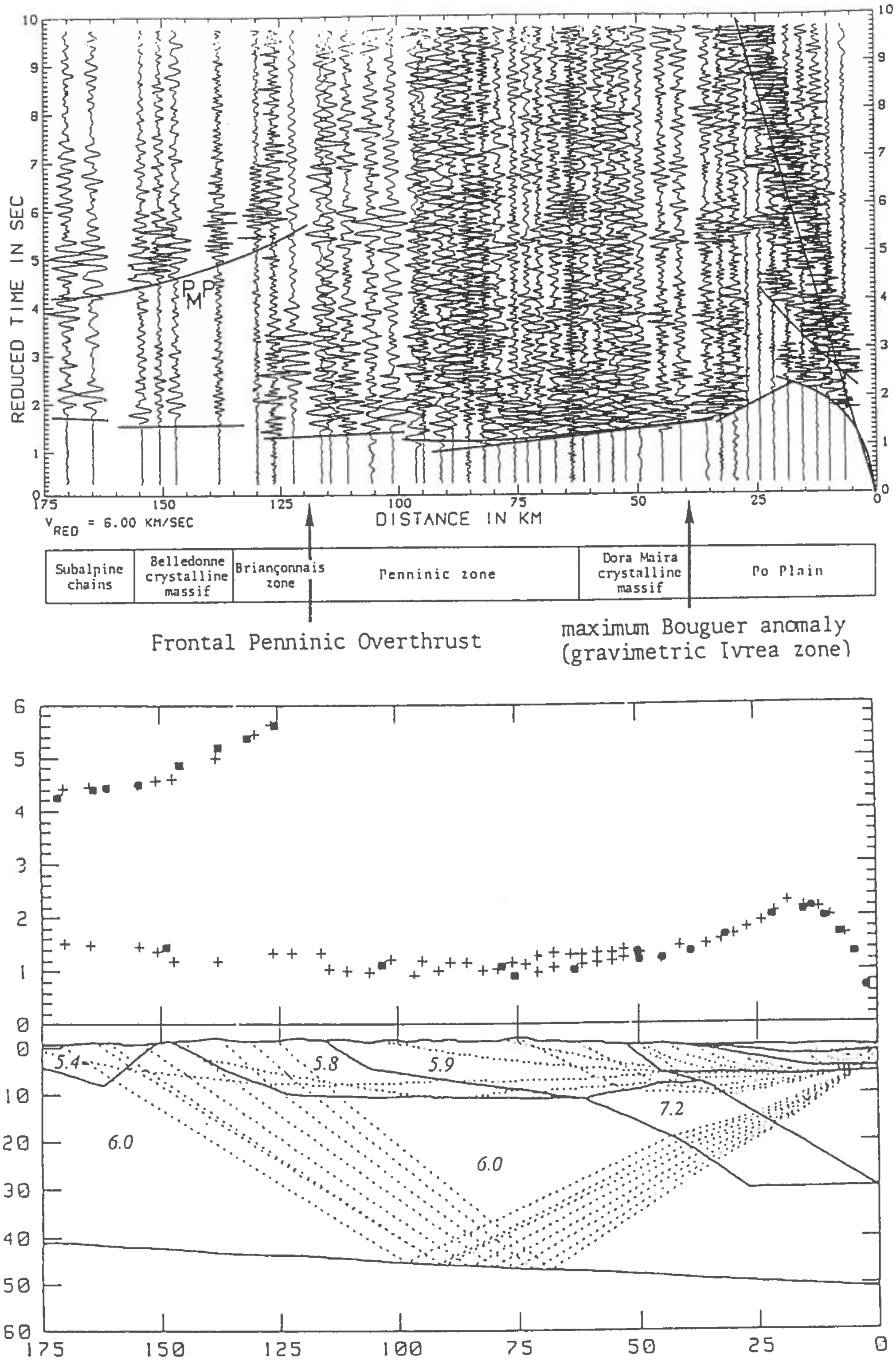


FIG. I-22. – Along a transverse profile from Grenoble to Torino, modelling of the Ivrea body and of a deep reflector [after Thouvenot *et al.*, 1985].
FIG. I-22. – *Sur un profil transversal Grenoble-Turin, modélisation du corps d'Ivrée et d'un réflecteur profond [d'après Thouvenot* et al., *1985].*

The origin of the Ivrea body remains a matter of debate : is it a "floating" unit composed of lower crust and oceanic mantle of south Alpine affinity enclosed in Alpine continental crust, or is it an obducted slice of south Alpine lithosphere ? In the latter case, this would imply a continuous connection to the east with the Moho under the Po Plain as well as a connection towards the west with the deep autochthonous Moho. In fact, we can speculate that this structure reflects a far more widespread upthrusting of the entire Alpine lithosphere.

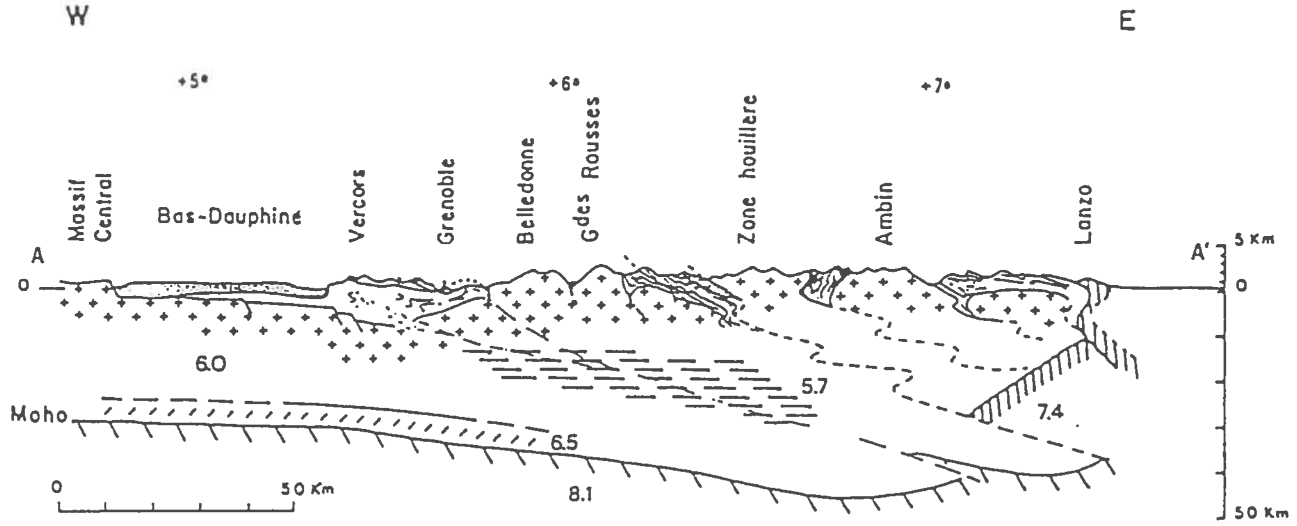


FIG. I-23. – Comprehensive E-W section of the Alpine domain at latitude of Grenoble from Pennine zones to the Massif Central [from Perrier and Vialon, 1980].

FIG. I-23. – *Coupe synthétique E-W du domaine alpin à la latitude de Grenoble, des zones penniques au Massif Central [d'après Perrier et Vialon, 1980].*

5) Lateral variations in velocity between the Alpine and peri-Alpine crustal segments

Leaving the Alpine arc in a northwesterly direction, the crust thins to a "normal" thickness of about 30 km beneath the Jura belt. This thinning is only one aspect of the transition which occurs across a line joining the Cévennes with the Aiguilles Rouges Massif. To the northwest of this line – in the peri-Alpine domain – all the profiles show a well-marked lower crust (cf. section (2) above), whereas in the Alpine domain to the south-east there are no discernable continuous markers present. Furthermore, a low velocity channel occurs within the crust at depths between 15 and 25 km according to the interpretation of certain profiles. Some authors go so far as to assert that a low velocity zone exists throughout the Alpine domain. We suggest here that its presence is very likely, at least at the local level. Such a zone would particularly favour the decoupling of the NW External Crystalline Massifs from the rest of the Alpine pile, associated with folding and detachment in the Subalpine ranges.

6) Moho(s) under the Po Plain

Under the western part of the Po Plain, the depth to Moho is of the order of 30-35 km. EGT experiments in 1983 have enabled an estimate of velocities beneath this interface (7.5 km.s^{-1}). However, these same data [Giese, 1985] reveal the existence of a second reflector at even greater depth (50 km, with velocities of 8.2 km.s^{-1} beneath the reflector). The geodynamic significance of these two horizons is still poorly understood. According to Giese [1985], Giese *et al.* [1991], Buness and Giese [1991] and Blundell *et al.* [1992], there has been a superposition of two crust – Adriatic and European – with complete continuity of the 50 km-reflector with the autochthonous Alpine Moho. In this hypothesis, the crustal root – which up to now has been relatively well localized beneath the Axial zone of the Alps – could extend for a great distance under the Po Plain.

7) Smooth or brittle Moho?

Due to the absence of clear observational evidence, possible steps in the Moho geometry have long been ignored in models for the deep structure of the Alps. Nevertheless, teleseismic records obtained along a transect of the Alpine arc have shown the existence of abrupt shifts in arrival time [Hirn *et al.*, 1984]. Such shifts are comparable with similar effects observed on either sides of the North Pyrenean Fault. This anomaly may be partly attributed to abrupt changes in the mean velocity of the crust or upper mantle. However, this phenomenon – which takes place in less than 20 km along the profile – is far more easily explained in terms of a stepped Moho. Even if such steps in the Moho remain speculative, our image of a smooth crust/mantle transition beneath the Alps may require some modification.

8) The Alpine lithosphere beneath the Moho

Our knowledge of the sub-Moho lithosphere is based essentially on the analysis of surface waves [Knopoff *et al.*, 1966; Panza and Muller, 1978]. The first mentioned authors introduce a very low velocity channel (4.1 km.s^{-1} instead of 4.5-4.6 km.s^{-1}) – with a roof at about 70 km – into the S-wave velocity model for mantle beneath the western Alps. Panza and Muller [1978] consider that "Verschluckung"-type phenomena have influenced the S-wave variations observed in the mantle beneath the entire Alpine belt. Otherwise, Poupinet [1976] has detected the trace of a very low velocity channel which fades out to the east of the Ivrea zone. This was achieved through the analysis of teleseismic residuals. Since a low velocity channel is anomalous in a continental collision zone setting, Poupinet [1976] has interpreted this structure as a remnant of the ancient mid-ocean ridge separating the European and Adria plates. Blockage on the ridge is thought to have led to an inversion in the direction of subduction.