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Seismic Evidence of a Crustal Overthrust in the Western Alps[1])

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Abstract – Seismic data provided by the ALP75 experiment in the northwestern part of the Alpine Arc has been re-evaluated using ray-tracing techniques in a laterally inhomogeneous medium, thus improving a previous interpretation. The structure of the crust definitely appears to be different in the Northern Subalpine Chains where it is layered with an ESE mean dip of 30° and in the inner part of the Alpine Arc where no clear stratification can be derived from the available data.

In the Northern Subalpine Chains the boundary between the upper and lower crust is found to be extremely uneven whereas the Moho discontinuity underneath as well as a boundary in the lower crust shows no evidence of unevenness. This pattern can be ascribed to the rheological properties of the crust which has a brittle behaviour at shallow depths and a more plastic behaviour underneath. A velocity reversal is found under the inner part of the Alpine Arc at a depth of 11 to 23 km.

An attempt is made to synthesize deep seismic sounding data which have been accumulated since 1956 in southeastern France. The crust is stratified to the north of a line Cévennes–Aiguilles Rouges and amorphous to the south. This line, which is the former boundary between the European plate and the Provence sub-plate, is thought to have acted as a weak zone during the Alpine orogeny. A crustal overthrust could have occurred in this region which would have been furthered by the higher plasticity of the low-velocity zone. Finally we speculate on the physical significance of intracrustal velocity reversals and suggest they are associated with active tectonic areas where dislocation densities are high.

Résumé – Une nouvelle interprétation des données sismiques ALP75 acquises dans la partie nord-ouest de l'arc alpin est présentée sur la base de la méthode du tracé de rai en milieu bi-dimensionel. La structure de la croûte est totalement différente sous les chaînes subalpines septentrionales et sous la partie interne de l'arc alpin. Dans le premier cas, la croûte présente des réflecteurs intermédiaires qui plongent vers l'est-sud-est à 30° environ alors que dans le second cas, aucune stratification n'apparaît clairement.

Sous les chaînes subalpines septentrionales, la limite croûte inférieure et supérieure est très irrégulière tandis que le Moho et une discontinuité dans la croûte inférieure ne présentent pas ce même caractère. On peut mettre cette particularité en relation avec la rhéologie de la croûte qui a un comportement plus cassant en surface et plus ductile en profondeur. Une zone à moindre vitesse est mise en évidence sous la partie interne de l'arc alpin entre 11 et 23 km de profondeur.

On présente une tentative de synthèse des résultats obtenus par sondages sismiques profonds dans le sud-est de la France depuis 1956. La croûte est stratifiée au nord d'une ligne Cévennes-Aiguilles Rouges et amorphe au sud. Cette ligne, qui est l'ancienne limite entre la plaque européenne et la sous-plaque provençale, s'est peut-être comportée comme une zone de faiblesse lors de l'orogenèse alpine. Elle a pu favoriser le développement d'un chevauchement crustal au niveau de la zone à moindre vitesse, de comportement plus plastique. La signification physique des inversions de vitesse dans la croûte est finalement abordée: elles n'apparaissent peut -être que dans les régions tectoniques actives, là où les densités de dislocations sont élevées.

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Key words: Deep seismic soundings; Crustal structure; Western Alps; Low-velocity zone; Overthrust; Alpine orogenesis.

1. Introduction

This paper is the fifth in a series in which the results of the Lithospheric Seismic Alpine Longitudinal Profile, 1975 (ALP75) are presented. Extending from France to Hungary the profile runs across the following geological units: the Northern Calcareous Subalpine Chains (Bauges Massif), the Crystalline External Massifs (Aiguilles Rouges, Mont-Blanc, Aar and Gotthard Massifs) with a cut across the Penninic Domain (Bernhard and Silvretta Nappes), the Austro-Alpine Domain (Oetztal Nappes) through the Engadin and Tauern windows, and finally the Pannonian Depression.

The joint experiment which was carried out by nine European nations has been described exhaustively by ALPINE EXPLOSION SEISMOLOGY GROUP (1976) where the very first results concerning the bulk structure of the crust can also be found. The second contribution to this series (MILLER *et al.*, 1977) proposes an improved structural model for the Eastern Alps while the next paper (MILLER *et al.*, 1978), gives the broad outlines of the velocity distribution along ALP75. The fourth paper (ITALIAN EXPLOSION SEISMOLOGY GROUP, 1979) sets out a preliminary interpretation of a fan profile across the Dolomitic Alps which was carried out jointly with ALP75.

The aim of this paper is to give a detailed picture of the crust in the westernmost part of ALP75 (Fig. 1) between shotpoint A (Mont Revard, near Aix-les-Bains, France) and shotpoint B (Nufenen Pass, near Gletsch, Switzerland). Up to now only

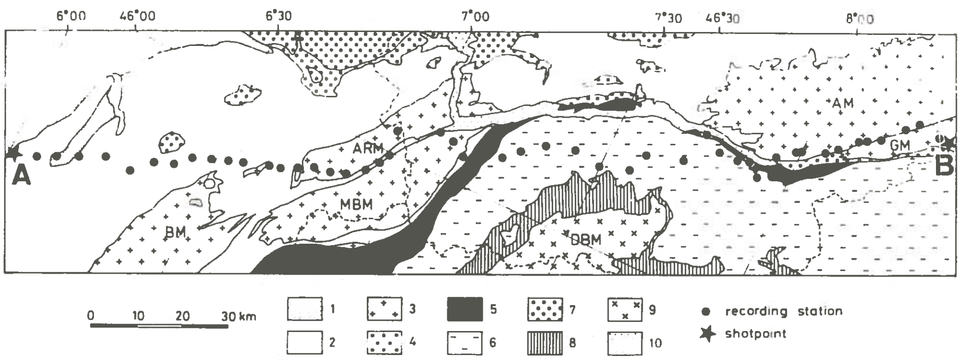

Figure 1
Location of the ALP75 recording stations along the segment AB on a background of geological units (geological map after COMMISSION GEOLOGIQUE SUISSE, 1972). (1) Extra-alpine sediments; (2) delphino-helvetic zone; (3) 'Massifs Externes'; (4) ultra-helvetic nappes; (5) Sion-Courmayeur zone; (6) Bernhard Nappe and Simplon–Tessin Nappes; (7) Klippen Nappes; (8) 'schistes lustrés'; (9) Dent Blanche Nappe; (10) Quaternary; BM, Belledone Massif; MBM, Mont Blanc Massif; ARM, Aiguilles Rouges Massif; GM, Gotthard Massif; DBM, Dent Blanche Massif.

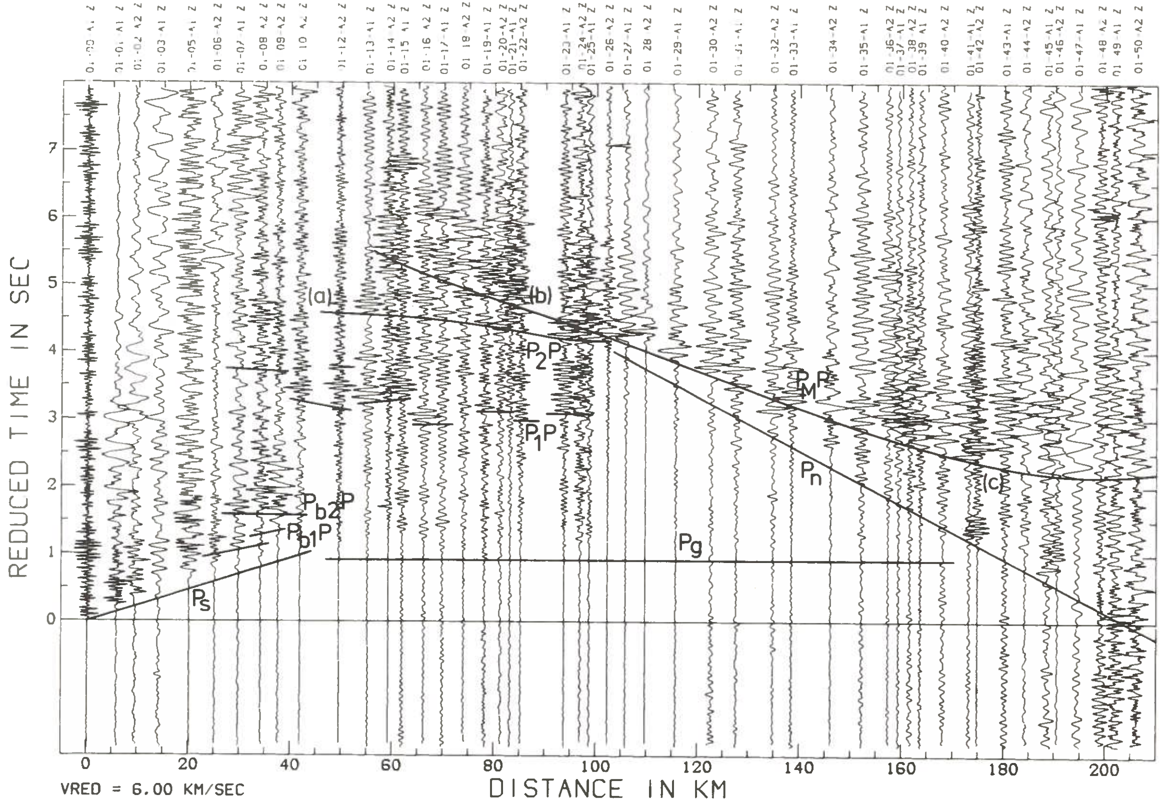

Figure 2

Record-section from shotpoint A (Mont Revard). Ps, sedimentary wave; Pg, wave propagating in the upper crust; PbP, waves reflected from the basement; P1P and P2P, intracrustal discontinuities; PMP (branches b and c), wave reflected from the Moho; Pn, wave propagating in the upper mantle.

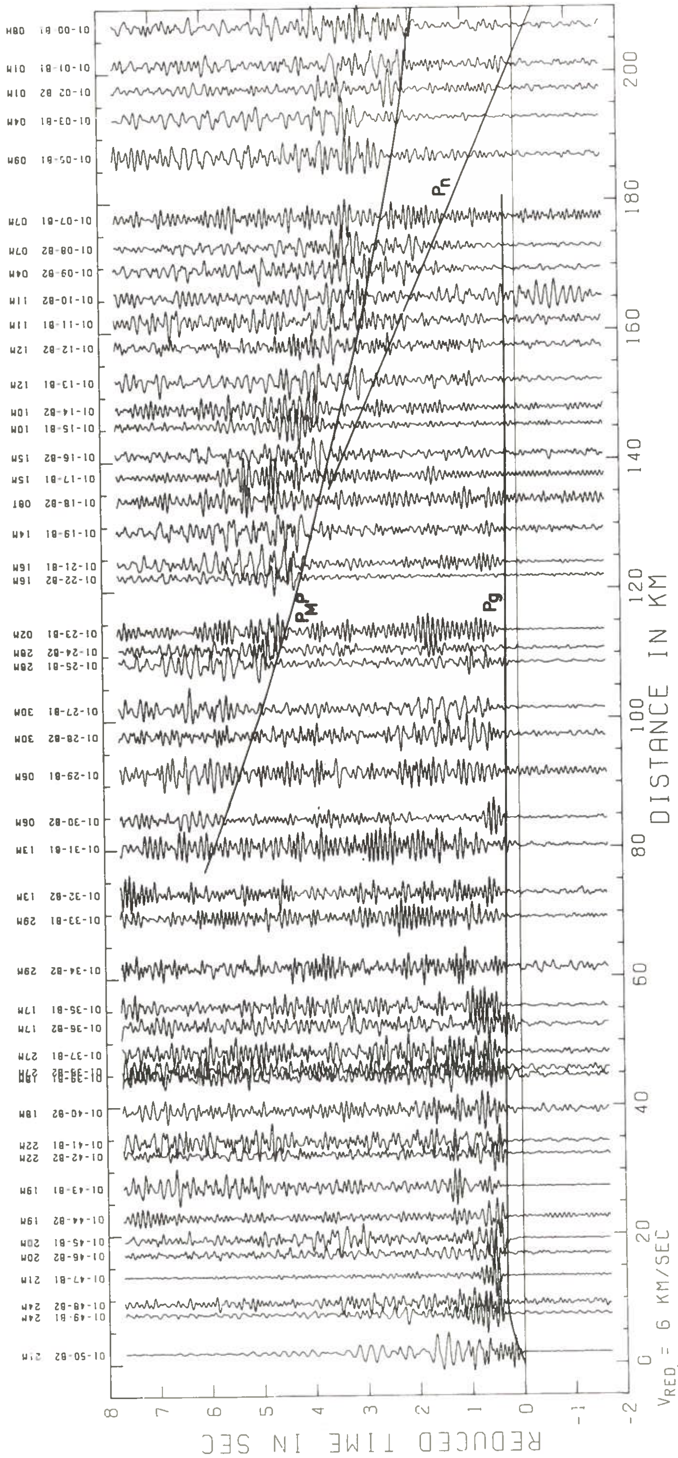

Figure 3
Record-section from shotpoint B (Nufenen Pass). Refer to Fig. 2 caption for abbreviations.

preliminary results have been obtained for this segment (THOUVENOT, 1976; OTTINGER, 1976) using flat-layered models with velocity gradients. Velocity-depth functions showed some discrepancies according to these authors but three points were agreed on: the presence of a velocity reversal in the upper crust; a mean crustal velocity of 6.2 km/sec; and a Moho depth of about 43 km in the northeastern part of the segment. The Moho discontinuity was indeed found to be shallower (37 km) towards the southwest, thus suggesting the presence of a rather steep dip between the two levels which could occur in the Penninic Domain. The original model (THOUVENOT, 1976) was checked through the computation of the Bouguer anomaly, with the assumption of two-dimensional structures. The deviation between the calculated anomaly and the observed data (15 mgal at most) was not found to be significant, given that the profile runs obliquely to the strike of the Alps and that the two-dimensional assumption is reaching its validity limit.

One only needs to have a look at Figs. 2 and 3, which display the record-sections along segment AB for shotpoint A and shotpoint B respectively, to realize that such an attempt to invert the data can lead to some misinterpretation. The flat-layered model which can be looked at as a valuable approximation in poorly tectonized regions is obviously insufficient here because of the fact that some layers appear to exist only locally. Furthermore, dipping interfaces cannot be excluded and should moreover be considered as a general rule. A third factor making the need for a two-dimensional seismic re-evaluation more necessary was the bent, or even broken, pattern of the observed travel-time curves which could not be explained at all by the flat-layer approximation.

Two approaches were used to model the crust: accepting the 1976 preliminary results, we proceeded to a quick 2-D adjustment on a desk-top calculator which afforded more pliability than a time-consuming computer program. In this first step a constant velocity was assumed in each layer and the height of the stations was not taken into account. We utilized then a slightly modified version of Will's PROMOS program which allows the introduction of velocity gradients in any direction and computes height corrections (WILL, 1975; WILL, 1976). Any crustal geometry can be attained by this ray-tracing program which proved very helpful in the present interpretation.

2. *Field of investigation*

In the previous explosion seismology experiments which were carried out in the Western Alps (CLOSS and LABROUSTE, 1963; CHOUDHURY *et al.*, 1971) shotpoints were always situated either in the crystalline basement of the so-called 'Massifs Externes' (Roselend and Lac Nègre) or in the Penninic Domain (Mont Cenis and Lac Rond des Rochilles). Except for the quarry blasts in Roselend and Mont Cenis, this was mainly due to logistic reasons: the bore-hole firing techniques for a large amount of explosives were inefficient at that time and cheap support was available through armed forces. Two criteria were therefore instrumental in deciding the shotpoints: the vicinity of military shotfields and adequate highland lakes which allowed easy underwater firings.

Although these goals were achieved and the experiments could display a satisfactory picture of the Moho discontinuity, two disadvantages rested in the geographic setting of the shotpoint in the case of quarry blasts or in the low-frequency content of the source signal in the case of underwater firings. Moreover, the recording system used prior to 1969, which had low dynamics because of the use of photographic recording, consisted of one-sec short-period pendulums connected to galvanometers with a natural frequency of 2.2 Hz. The resulting bandwidth was consequently narrow with a sharp maximum between 1 Hz and 2 Hz which prevented the observation of high-frequency intracrustal reflections.

It should be stressed therefore that shotpoint A in Mont Revard was the first to be fired in the Dauphinois Zone, and therefore the first in the Northern Subalpine Chains, in association with wide band-pass recording systems such as the ones used since 1970 in conventional DSS surveys (PERRIER and RUEGG, 1973).

From the morphological point of view, Mont Revard is one of these fortress-like plateaux exposing an Urgonian cliff of massive reef-rock limestones which are the characteristic features of the landscape in this area. The stratigraphy of the Subalpine Chains is dominated by the intercalation of marly and limestone series, the most prevalent of the latter being, besides the Urgonian (Schrattenkalk of the Helvetian facies), the Tithonique (Hochgebirgskalk) represented here by fresh-water limestones. The most important marl horizons are those of the Oxfordian and the Valanginian, the latter forming a slope between the two calcareous steps.

The stratigraphy becomes more complex further below. The overthrust of these Mesozoic series over the Lac du Bourget Miocene resulting in a recumbent folding which would increase the thickness of the sediments to the east has been well stated (DEBELMAS, 1974). According to RUTTEN (1969), there is a general consensus that 'décollement' occurring in the evaporate horizons of the Triassic has never been very significant in the Subalpine Chains, but other 'décollement' horizons are likely to be found in the 'Lias schisteux' or the 'terres noires' series (GOGUEL, 1952). Therefore, as Rutten points out, 'a drilling program in the Subalpine Chains might well come up with some unexpected facts'. Drilling results in the Dauphiné-Savoie area (Fig. 4) could be of some interest but one should bear in mind that the drilling sites are all situated in the peri-Alpine Molasse Basin, in other words, they do not take account of the frontal overthrusting.

Accepting these results we can estimate that the depths of the Permian series as indicated in Fig. 4 are likely to be only minimum values further east. An expected depth of the ante-Triassic basement under Mont Revard would be therefore at least 3000 m[3]).

[3]) All depths given in this study refer to sea-level, except where otherwise stated.

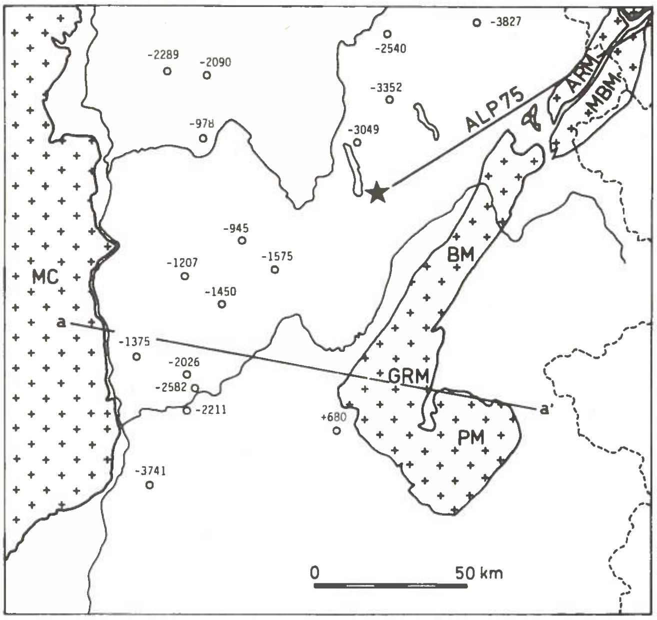

Figure 4
Position of the Palaeozoic basement (referred to sea-level) between the French Massif Central and the Alpine Arc on the basis of deep boreholes results (after MENARD, 1979). aa', Position of the cross-section Fig. 5; MC, Massif Central; BM, Belledonne Massif; GRM, Grandes Rousses Massif; PM, Pelvoux Massif; ARM, Aiguilles Rouges Massif; MBM, Mont Blanc Massif.

Nevertheless the position of the basement up to the crystalline outcrop of the Aiguilles Rouges Massif (Fig. 1) is far from being clear. In a recent work MENARD (1979) suggested that the basement is quite deep (about 8000 m) under the Subalpine Chain of Vercors and under the Alpine Groove near Grenoble, and that the Belledonne Massif is possibly overthrusting the Subalpine basement (Fig. 5). His approach is supported by a re-evaluation of the profile Lac des Rochilles-Rhône, 1956 in which Pg waves are delayed by about 2 sec. One could assume that the same feature occurs

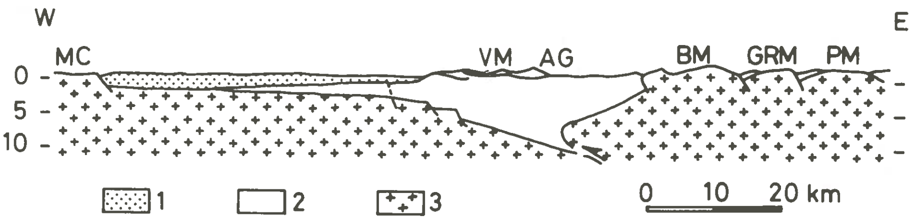

Figure 5
Deep cross-section of the central part of the French Alps, showing the structure of the main zones and the postulated high dip of the Belledonne Massif under the Northern Subalpine Chains (after MENARD, 1980). (1) Cainozoic; (2) Mesozoic; (3) Palaeozoic basement; VM, Vercors Massif; AG, Alpine Groove ('Sillon Alpin'). Refer to Fig. 4 caption for other abbreviations.

further north with the Aiguilles Rouges Massif and we will take account of this possibility later.

The tectonics of the frontal zone of the Penninic Domain (Bernhard Nappe) will not be discussed in detail here since, as will soon become evident, the distance between the shotpoints was too long to provide usable data on the sub-surface features of the Wallis area. We will rather concentrate the last part of this section on the next geological units encountered by ALP75: the Aar and Gotthard Massifs.

The recording stations were actually sited in the Aar Massif (Fig. 1). The Gotthard Massif, where shotpoint B was fired in Nufenen Pass, occurs only in the very last kilometres of the segment. The two massifs are independent since, after the formation of the Helvetic Nappes, the Gotthard Massif is believed to have approached the Aar Massif over a distance of 'at least 20 km' (TRUMPY, 1963), thus compressing and vertically tectonizing the Tavetscher Zwischenmassiv which can be looked at as the root zone of the Helvetic Nappes. Whether or not seismic rays from Mont Revard did travel much in the southwestwards continuation of the Tavetscher Zwischenmassiv is an open question, since the trend of the three massifs is the same as that of the profile and side effects of a few kilometres cannot therefore be excluded. It is obvious, however, that seismic rays leaving Nufenen Pass with a high take-off angle have to go through the remnants of this tectonized zone. Travel-time anomalies should therefore be expected for the Pg wave.

3. Sub-surface features in the vicinity of the shotpoints: Position of the Palaeozoic basement under the Northern Subalpine Chains and velocity-depth function in the Aar-Gotthard Complex

Shotpoint A provided the only data which will be used in this study to estimate the position of the ante-Triassic[4]) horizon under the Northern Subalpine Chains. As will be seen later on, the Pg wave from shotpoint B becomes too weak in this part of the profile to be relied on.

The most obvious feature of the record-section originating from shotpoint A in the first two seconds of reduced travel-time (Fig. 6) is the rather low-velocity wave observed over the first 40 km. This wave, denoted Ps, is superseded further on by the Pg wave with an intercept time of about 1 sec. The Ps wave, interpreted as a sedimentary wave, can be correlated in a straight line yielding a 5.4 km/sec velocity, although the rather high value could lead to some scepticism.

The doubt disappeared after we checked the velocity through further field measurements with the seismic hammer method in an Urgonian outcrop near Grenoble: two 20 m-long profiles yielded a 5.7 km/sec velocity, thus proving that

[4]) One should bear in mind that Carboniferous and Permian series are likely to be included in the sedimentary cover from the seismic point of view.

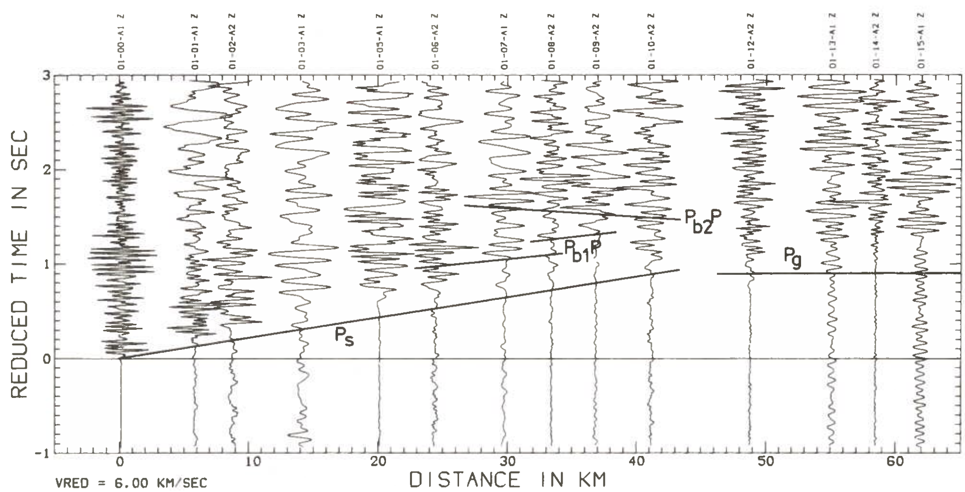

Figure 6
Main wave groups used to estimate the position of the basement under shotpoint A. Ps, Sedimentary wave; Pb1P, wave reflected from the dipping part of the basement (as seen from the shotpoint); Pb2P, wave reflected from the rising part of the basement.

highly compacted limestones might not be seismically distinguishable from weathered crystalline rocks[5]). This state of fact is also reported by MILLER *et al.* (1977) or WILL *et al.* (1978) in the Northern Limestone Alps, where similar and even higher P-wave velocities can be found in the sediments. As the tectonics of the Mesozoic cover are rather complex in the Subalpine Chains, the Ps wave is believed to have integrated the various layers, the expected velocity increase with depth being tempered by low-velocity series such as the Oxfordian marls, the 'Lias schisteux' or the Triassic evaporites. It is nevertheless rather puzzling to note the straightness of the Ps travel-time curve although the waves propagate in such an inhomogeneous medium as a jumble of extremely folded Mesozoic sediments.

The Ps wave could also have been interpreted as a head wave refracted in the roof of a dipping basement. But the minute intercept-time would imply that the basement is very shallow under Mont Revard – at least as high as sea-level – which is inconsistent both with geological considerations (Fig. 5) and with drilling results (Fig. 4). Furthermore such a dipping model would make rather critical the interpretation of strong late-arrival waves (P-basement-P or PbP) which are the prevailing phases in the 30–40 km range of the profile (Fig. 6).

These PbP waves can be split into two groups of branches, Pb1P and Pb2P. Because of their high-energy aspect both groups were interpreted as waves reflected from the ante-Triassic horizon, Pb1P being reflected from a dipping interface and Pb2P from a

[5]) Laboratory experiments on limestone samples from the same place yielded a 5.35 km/sec velocity (Thiercelin, private communication).

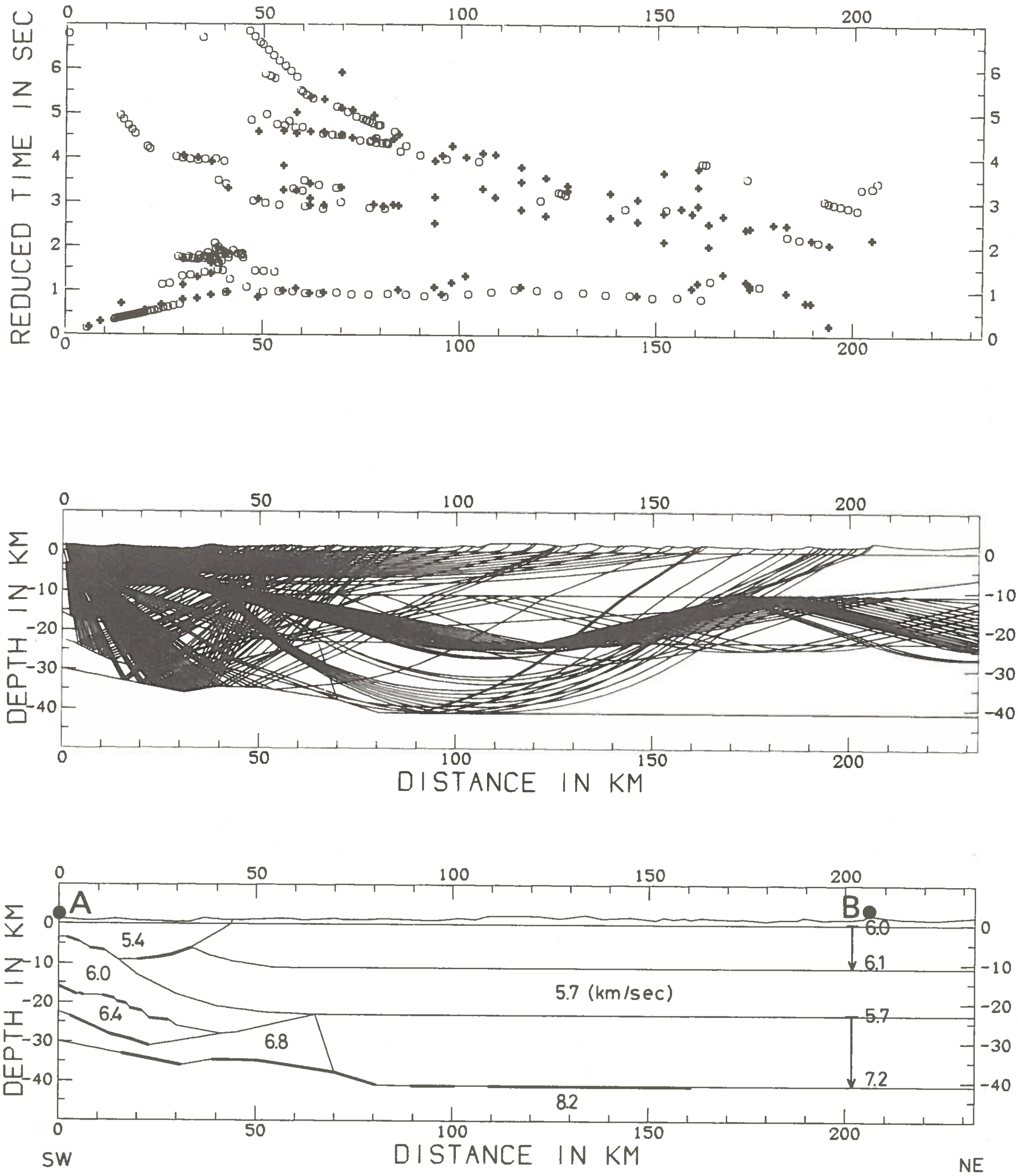

Figure 7
Results of model calculations for the segment AB (shotpoint A). Cross-section (lower part), ray-paths (middle part) and calculated travel-times (circles, upper part) with observed travel-times (crosses). In the cross-section, reflecting elements are shown in heavy lines. Discontinuities shown in thin lines are assumed.

rising interface (Fig. 7). Moreover the Pb2P signal looks rather like a single reflection at a distance of 30 km and becomes more complex and stretched in time further on. This feature can be explained by the lens effect of the concave rising part of the basement which focuses the reflected energy at distances of around 40 km.

To derive the shape and position of the Palaeozoic basement we assumed that the velocity found just under the surface (5.4 km/sec) was the mean velocity in the sedimentary cover. *The final model (Fig. 7) shows a 3500 m deep basement under the shotpoint, with a maximum depth of 9000 m being reached under the southern end of the Lac d'Annecy depression.*

This shape of the basement can be likened with the model postulated by MENARD (1980) for the boundary between the Vercors Subalpine Chain and the Belledonne Massif (Fig. 5) or with the trough-like feature described by RYBACH *et al.* (1978) to the north of the Aar Massif. But, although it is the first clue to a possible overthrusting of the Aiguilles Rouges Massif over the western basement, we would expect the rising part tö be convex rather than concave. There is probably not enough seismic data available to allow a thorough study of this point. Moreover a concave shape can also be explained by a palaeo-relief prior to the overthrust. In such a model (Fig. 7) the Pg waves which can be ovserved up to about 160 km cannot be looked at as true head waves, since Pg rays must travel in the upper part of the sedimentary cover before being refracted in the crystalline basement. This model is suitable because of the very poor quality of the Pg phase over long distances (Fig. 2).

On the other hand, the record-section originating from shotpoint B (Fig. 3) shows a Pg curve which is typical for a crystalline medium: over the first few kilometres in distances the apparent velocity is low and then increases up to a maximum of 6.0 km/sec. The corresponding velocity–depth function shows a strong velocity gradient (1 sec^{-1}) between the surface (+2000 m) and a depth of about 2 km (sea-level). The velocity found immediately under the surface (4.0 km/sec) seems to have quite a low value compared with the velocity–depth distributions which have already been arrived at for crystalline rocks in Western Europe.

For instance OTTINGER (1976) found a velocity–depth distribution conformable to the normal in the Simplon–Tessin area. The anomaly we are dealing with is ascribed to the fact that Nufenen Pass is in the Gotthard Massif whereas the stations are sited on the southern edge of the Aar Massif and that *over the first kilometres of the profile the Pg wave goes through the low-velocity Gotthard verrucano.*

4. The stratified western crust and the amorphous eastern crust

Another striking feature which can be observed on the record section A (Fig. 8) is the high-energy reflection at a reduced travel-time of about 3 sec. This reflection cannot be satisfied by a single 'hyperbola' but by several branches, with important offsets (up to 0.5 sec) occurring between each of them. These offsets could be due to

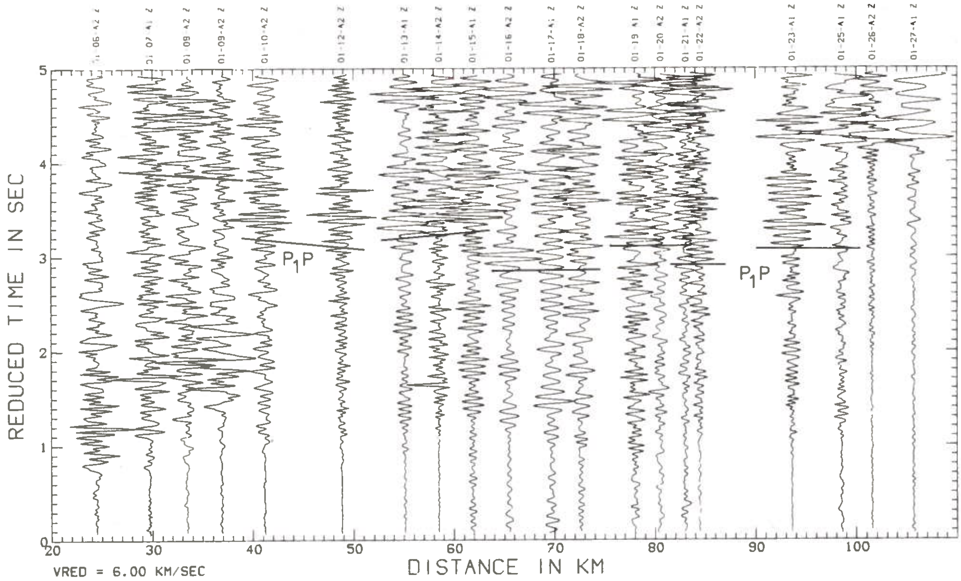

Figure 8
Broken pattern of the P1P reflection as observed on the record-section (shotpoint A) in the 30–90 km range. Travel-time jumps can reach 0.5 sec and can, with difficulty, be explained by crustal heterogeneities.

either lateral crustal heterogeneities or to local variations of the reflector. However, if heterogeneities do occur within the upper crust they could not be the sole reason for such travel-time jumps. In this interpretation, the split nature of the reflection (denoted P1P) has been entirely ascribed to various elements of a discontinuity between the upper and lower crust.

Moreover the almost horizontal trend of the branches, even over short distances, infers rather a strong dip of the reflector, and the 2 sec time-lag between the P1P asymptote and the Pg curve suggests the presence of a low-velocity zone (LVZ) in the upper crust, at least on a good deal of the ray-path[6]). The model shown in Fig. 7 was constructed assuming a constant velocity of 6.0 km/sec in the upper crust under the Subalpine Chains while an LVZ is introduced further east to fit the data.

The upper boundary of the LVZ cannot in any account be shallower than 10 km because otherwise it would prevent the observation of the Pg wave from shotpoint B over large distances (Fig. 9). On the other hand, a maximum depth of the lower boundary can be estimated at about 23 km: if the LVZ were extending deeper into the lower crust it would be rather critical and brain-racking to make the model consistent with the deep reflections. *But it stands to reason that other ray-tracing models can be calculated for so many other parametrizations of the LVZ, each parameter – depth of the*

[6]) GIESE *et al.* (1970), in a first attempt to synthesize geological and geophysical results in the Western Alps, already suggested the presence of a velocity inversion in the upper crust under the Mont Blanc Massif.

upper boundary, depth of the lower boundary, mean velocity, shape of the frontal zone – being chosen within a domain of likelihood. Because these domains are rather tight we stress that the model proposed in Fig. 7 is stable though not unique.

The mean velocity in the inversion zone (5.7 km/sec) was chosen so that it is still possible to correlate this velocity inversion with the expected increase in temperature, as will be seen in the next section. Lower values (e.g. 5.5 km/sec) would make this correlation impossible while higher values are prohibited by the considerations above.

One should take care, when modelling the shape of the LVZ, that many of the rays can be trapped inside if the upper and lower boundaries are not carefully chosen. The shape of the LVZ shown in Fig. 7 was taken because it is in fact the simplest: the LVZ is present in the crust under the 'Massifs Externes' but not under the Subalpine Chains. The shape of the frontal zone was assumed having regard to surface tectonics as described for instance by MENARD (1979 and 1980). Because the upper and lower boundaries are parallel, the effect of the LVZ on the seismic rays is one of an optical dioptre. Such a shape allows no loss of energy and a lens effect in fact focuses the rays at around 60 km where the reflections are strongest.

Once the geometry of the LVZ has been chosen, ray-tracing techniques can be applied to derive the position of the P1P reflector, in order to explain the travel-time jumps observed on the record-section (Fig. 8). *We stress once more that, although the proposed model is not unique, the strong dip of the reflector as well as its unevenness should be considered as established results.*

The broken pattern of the P1P reflection 'hyperbola' contrasts strongly with the smooth reflection curves which can be observed from 50 km up to shotpoint B (Fig. 2). As was previously pointed out for the P1P reflection, the horizontal trend of branch (a) suggests again a dipping boundary. In an earlier evaluation (THOUVENOT, 1976), this boundary has been interpreted as the Moho discontinuity, but it seems more likely that branch (c) is linked to the less defined branch (b), with the two of them together being looked at as the actual PMP reflection. Branch (a) would then be only another intracrustal reflection (P2P).

Looking now at the model Fig. 7, one can note that all the dips of the crustal layers are consistent under the Subalpine Chains. The mean dip is 15°, but it is only an apparent dip, since the profile runs obliquely to the Alpine Arc in this area. Assuming that the dipping occurs straight towards the Arc – which follows from evident geological and tectonic considerations – and that the angle between the trend of the profile and the strike of the dipping crust is approximately 40°, *the corresponding true dip would be 30°. This considerable dipping of the crust at the western margin of the Alps is another important result.*

When observing the record-section originating from Nufenen Pass (Fig. 3) it is surprising to note the lack of any sharp reflection within the crust. If energy can be traced on a few isolated recordings around 3 sec in reduced travel-time, which suggests an intracrustal boundary in a 20–25 km depth range, it does not seem wise to try any

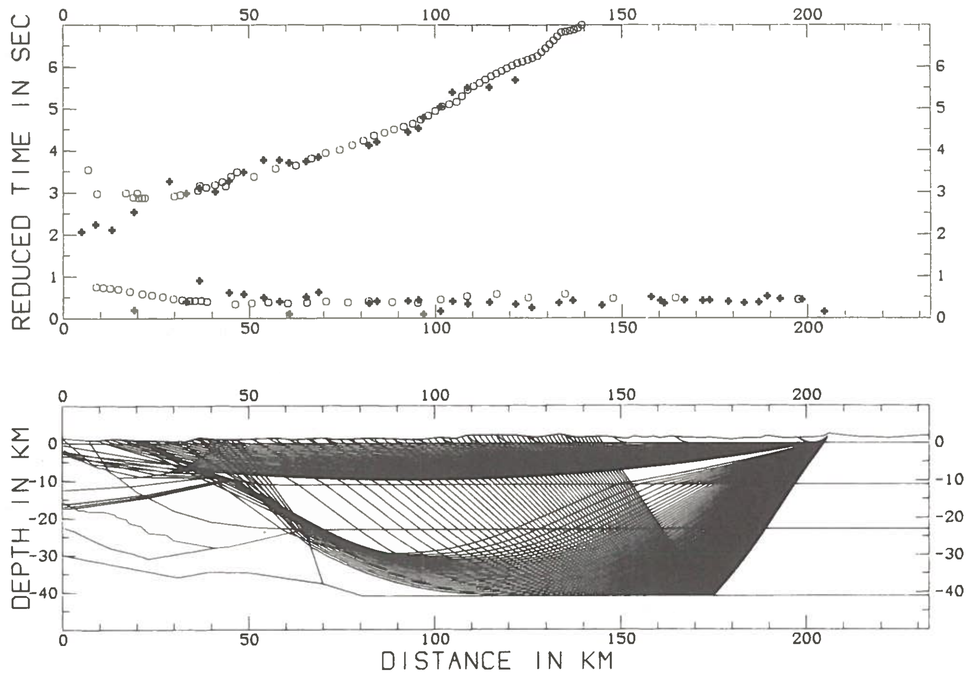

Figure 9
Results of model calculations for the segment AB (shotpoint B).

correlation. In our opinion the crust underneath the inner part of the Alpine Arc is rather 'amorphous'[7]) and the lower crust is probably a wide zone with a strong velocity gradient, although local boundaries allowing reflections of energy are likely to occur in the middle part of the crust. Such discontinuous reflectors have already been postulated by SMITHSON (1978) following COCORP surveys in the United States.

Contrasting with these evanescent reflections, the PMP reflection from the Moho discontinuity is very pronounced. In the ray-tracing model proposed for shotpoint B (Fig. 9), a layer with a velocity gradient of 0.75 sec^{-1} is introduced in the lower crust to fit the data. The Moho is approximately level at a depth of 41 km.

The crustal model shown in Fig. 7 (shotpoint A) or Fig. 9 (shotpoint B) is supported by the observation of the Pn waves refracted in the uppermost mantle where a mean velocity of 8.2 km/sec was calculated.

5. Geodynamic implications: Evidence of a crustal overthrust

The aim of this section is to draw speculative geodynamic conclusions from the very peculiar structural pattern arrived at in this study. The juxtaposition of a stratified

[7]) 'Unstratified' in this context.

crust in the southwest next to an amorphous crust in the northeast can be easily understood if we refer to VIALON (1974) or to MENARD (1979, 1980). VIALON (1974) suggested that an important strike–slip process took place during the late Cretaceous period along a belt running in a SW–NE direction from the Cevennes Mountains (southeastern part of the French Massif Central) towards the Northern Alps. There would be a direct connection between this SW–NE trend and that of the tear fault net which appeared in the Alpine basement as a result of the Hercynian orogenesis and which was re-activated during the Alpine orogenesis. The strike–slip was interpreted as a continental sinistral transform faulting process due to a general N–S compression. From the plate tectonics point of view this zone should be looked at as the boundary between the European plate and the Provence sub-plate (WESTPHAL *et al.*, 1978). It is generally agreed that the sinistral transform faulting process was initially activated by the rotations of the Iberic plate and of the Corsican and Sardinian sub-plates and that dextral processes possibly occurred afterwards following the differential motion of the

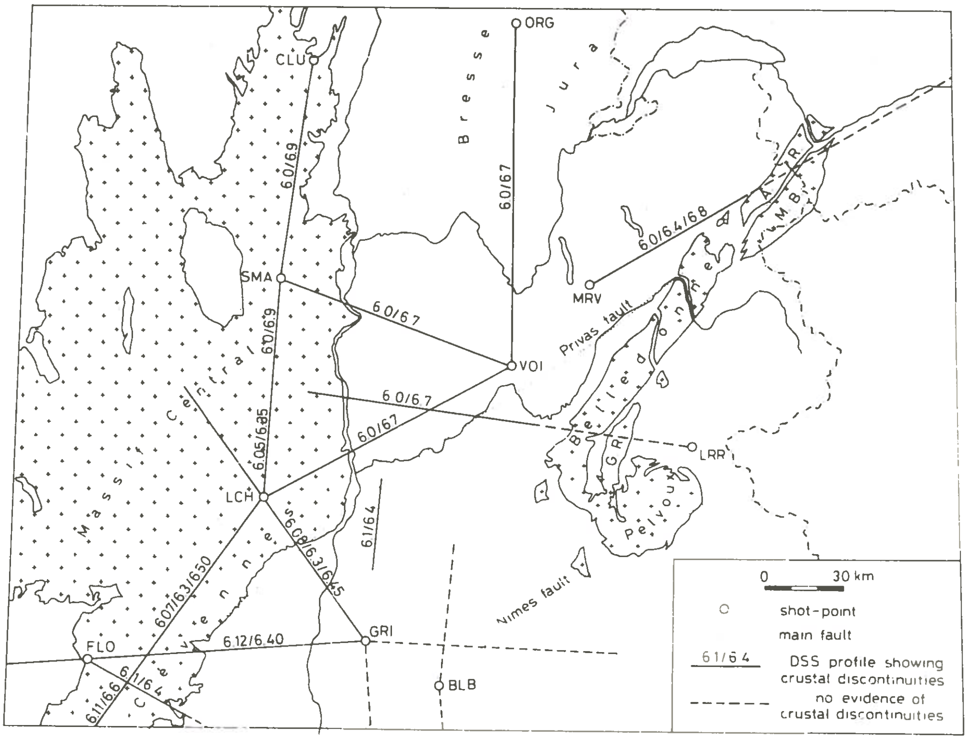

Figure 10

Synoptical map of the Rhône valley and the adjacent area showing the crustal structure as derived from deep seismic soundings. Compilation of results by TARDI (1957), PERRIER and RUEGG (1973), SAPIN and HIRN (1974), EGLOFF and ANSORGE (1976), MICHEL (1978) and THOUVENOT and PERRIER (this study). Main fault system after VIALON (1974). Shotpoints: BLB, Buis-les-Baronnies; CLU, Cluny; FLO, Florac; GRI, Grignan; LCH, Le Cheylard; LRR, Lac Rond des Rochilles; MRV, Mont Revard; ORG, Orgelet; SMA, Saint Martin; VOI, Voiron.

European plate and of the African plate. In the structural scheme proposed by Vialon, the northernmost fault of this strike-slip belt (Privas fault) has a N50 trend and extends up to the Chamonix synclinorium whereas the southernmost fault (Nîmes fault) runs through the Pelvoux Massif (Fig. 10).

MENARD (1979 and 1980) corroborated Vialon's assumptions and improved the previous model in the sense that his synthesis of available geophysical and geological data on the upper crust in southeastern France showed unequivocal differences in the seismic properties of the basement: if we put aside the L-shaped arc of the 'Massifs Externes' (Aiguilles Rouges, Mont Blanc, Belledonne, Grandes Rousses, Pelvoux, Argentera) which seems to have acted as a rigid wedge during the Alpine orogeny, the basement is shallow with high Pg-velocities (6.0 km/sec or more) to the northwest of the strike-slip belt, while the depth of the basement can reach drastic values to the southeast (10 km under the 'fosse vocontienne') together with low Pg-velocities (5.8 km/sec). This drop in velocity is believed to be the expression of a more plastic behaviour of the basement in this area.

Figure 10 is a synoptical map of southeastern France which shows the crustal structure derived from deep seismic soundings experiments from 1956 up to 1975: we differentiated profiles where no crustal stratification was found from profiles where such a stratification does occur. It is striking to note the consistency of the results with Vialon's structural scheme: *the crust definitely appears to be stratified to the northwest of the strike-slip belt and amorphous to the southeast.* This would only be a

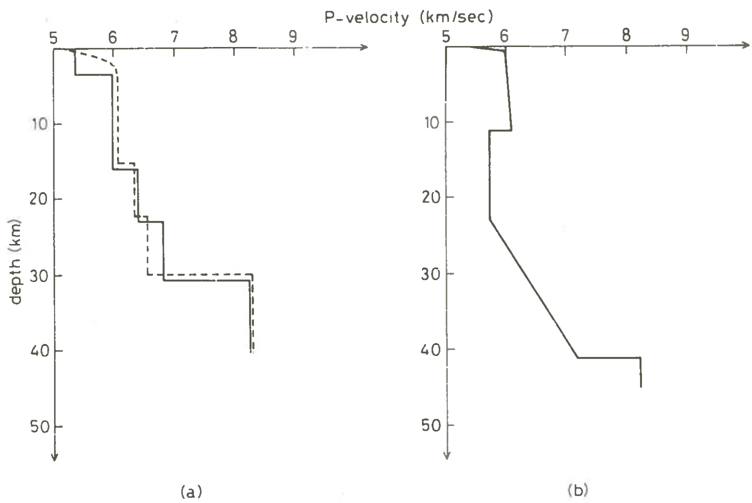

Figure 11
Evolution of the velocity-depth function along segment AB. (a) Comparison of V(z) under the Northern Subalpine Chains (full line) and under the Cevennes Mountains (south-eastern part of the French Massif Central, dashed line); (b) the velocity-depth function under the inner part of the Alpine Arc shows a low-velocity zone in the upper crust together with a wide-transition zone in the lower crust. There is no more evidence of intracrustal boundaries.

prolongation in the whole thickness of the crust of Ménard's differentiation of the basement.

A comparison between the crustal structure of the Northern Subalpine Chains and of the southeastern margin of the French Massif Central (PERRIER and RUEGG, 1973; SAPIN and HIRN, 1974) brings still more confidence in this structural scheme (Fig. 11). Both selected areas are part of the stratified crustal block and the corresponding velocity–depth functions show an undeniable similarity with regard to the velocity values as well as to the depth of the reflectors or to their number.

The second point which we would like to stress is the presence of a low-velocity zone under the inner part of the Alpine Arc (Fig. 11) and the important dip of the subalpine crust. We suggest that the differentiation found in the crustal structure north and south of the strike–slip belt does apply for the low-velocity zone: *there is no velocity inversion in the upper crust to the northwest of the belt while it possibly exists to the southeast.* The dip of the subalpine crust expresses the crustal overthrust which was furthered by the LVZ during the Alpine orogeny. Thus, *the strike-slip belt postulated by* VIALON (*1974*) *acted as a weak zone at that time and allowed a crustal overthrust furthered by the presence of the low-velocity zone.*

The physical and rheological meaning of this LVZ can give rise to many speculations. As for rheology, AHORNER *et al.* (1972) and FRECHET (1978) have already pointed out that most seismic events in the Western Alps are located in the first 10 kilometres of the crust. This is consistent with our crustal model if we assume that *the upper crust has a brittle behaviour whereas, at depths greater than 10 km, a more plastic behaviour occurs in connection with the low-velocity zone.* An evidence of such a rheological transition is found in the unevenness of the boundary between the upper and lower crust under the Northern Subalpine Chains, while the Moho discontinuity and the boundary in the lower crust are both very smooth (Fig. 7). An open question is whether this plastic behaviour should be looked at as an acting power in the crustal overthrust or as a consequence.

Referring to RUTTER (1976) and to NICOLAS and POIRIER (1976), one can try to investigate the mechanisms which are believed to be potentially able to produce pseudo steady-state flow in natural rock deformation processes and therefore to determine which mechanism is prevailing in the LVZ. Besides cataclasis, dislocation mechanisms and solid state diffusion processes, Rutter points out the importance of the pressure solution process, where diffusive mass transport is assisted by the presence of water at grain boundaries. Effective diffusivities in the aqueous phase may, at temperatures of about 200–400°C, be of the order of solid state diffusivities at very much higher temperatures.

To go further into the rheology of the LVZ, one needs therefore to know the magnitudes of the local state and microstructural variables. Most of them are not known uniquely, and at best they can only be assigned upper and lower limits. A reasonable temperature distribution can be derived from HURTIG and OELSNER (1977) whose computations result from heat flow analysis in Europe: one can expect a temperature of 300°C at a depth of 10 km under the Swiss Alps while a temperature of 500°C can be found 10 km deeper. Their results are consistent with those by POTY *et al.* (1974) which are derived from temperatures and pressures of formation of Alpine fissures and which yield an average temperature of 400°C at a depth of 10 km *when the 'Massifs Externes' rose up.* This drop of 100°C in temperature seems to be quite reasonable given that the surrection took place some 15 My ago.

As for the likely differential stress in the crust, an upper bound is provided by faulting mechanisms of earthquakes and by theoretical dynamic model calculations for earthquake sources. Even the introduction of cracks-with-barriers models does not yield stress drops higher than 500 bars (BOUCHON, 1978), although HANKS (1974) computed a drastic value of 1500 bars for the San Fernando Earthquake. Accordingly a reasonable upper bound for likely stress drops in the crust can be estimated at 1000 bars while the lower bound is arbitrarily assigned the value of 1 bar, which seems to be typical of low-stressed areas.

Adopting now a likely quartz grain size of 0.1 mm and assuming that the rheology of the upper crust is governed by the steady-state flow in quartz, bounds for the temperature and stress can be used to plot a box onto the appropriate deformation mechanism maps for quartz as given by RUTTER (1976) and consequently one can investigate the processes which develop in the low-velocity zone. Without pressure solution the predominant mechanism is dislocation creep, although solid state diffusion processes – Coble creep and Nabarro–Herring creep – occur at low stress values but with very low strain rates. By the addition of a pressure solution, the dislocation creep field recedes and the prevailing mechanism is pressure solution diffusion. It is worth noting that, in this case, even at a low stress value of 10 bars, the strain rate is about four orders of magnitude higher than without pressure solution. This is why RUTTER and MAINPRICE (1977) speculated that pressure solution might be a significant process helping to determine the rate at which aseismic slip occurs in fault zones at shallow to moderate crustal depths. Using a grain size of 0.1 mm and sensible diffusivity parameters, they estimated that a fault zone 1 m to 10 m wide would slip at a rate of the order of 1 cm/yr under a shear stress level of 100 bars and at a temperature of 300°C. This proves that such a mechanism could be held responsible for the surrection of the 'Massifs Externes' and for the push forwards of the Subalpine Chains, as a result of the crustal 'décollement' occurring in the more plastic low-velocity zone.

From the physical point of view, the meaning of the LVZ is not well stated and it can give rise to much more speculation that its rheological significance. The main point is whether or not this zone can be likened to a layer with a strong attenuation. This question will remain open since we are not in a position to investigate the anelasticity of the LVZ – precisely because this is a low-velocity layer and no ray turning-point occurs in the relevant depth range. We cannot therefore use direct waves to estimate the quality factor in the LVZ and the sole way to achieve it would be eventually the use of under-critically reflected waves – e.g. from the Moho discontinuity. We have not gone more deeply into this problem in spite of its importance.

If there is no higher attenuation in the LVZ than anywhere else within the crust, a good deal of the velocity drop can be explained by a pure temperature-effect. Assume the upper crust to be granitic and suppose the temperature to be 400°C at a depth of 15 km (HURTIG and OELSNER, 1977). Then, according to FIELITZ (1976), the relative velocity-drop would be 2%. Since the study of the Pg wave from Nufenen Pass yielded a 6.1 km/sec at a depth of 10 km, it is likely that the temperature effect cannot lower the velocity to less than 6.0 km/sec in the middle of the LVZ. This is in agreement with compressional wave velocity measurements by CHRISTENSEN (1979) who stresses that, if crustal velocity inversions produced by high temperature are likely to be common within the crust, the velocity decreases usually reported for crustal LVZ in explosion seismology studies are much greater than the decrease which can be accounted for by temperature alone. As Fielitz's laboratory experiments refer only to dry rocks, the remaining 0.3 km/sec velocity-drop should be ascribed to the presence of fluid components in the crust. In an extensive survey of physics of high pressures and temperatures, DMOWSKA and HANYGA (1974), among others, have emphasized the strong influence on the strength of silicates which is exerted by the weakening behaviour of pore water. According to HOBBS *et al.* (1972) and to GRIGGS (1974) the hydrolysis of bonds Si—O—Si can occur even in the presence of a small amount of water. The result is a higher dislocation mobility in quartz which furthers an eventual plastic deformation, alters the attenuation and the elastic moduli, and therefore modifies the velocity (Gueguen, private communication).

If a high attenuation were to be discerned in the LVZ, another possible mechanism could be a relaxation process in dislocations such as the one analysed in olivine by GUEGUEN (1974) in connection with the upper mantle LVZ. This process yields a high attenuation and drastic drops in elastic moduli at low frequencies and high temperatures. Unfortunately, and although deformed quartz crystals obviously display high dislocation densities, no experimental data on dislocation relaxation in quartz is available at the moment. Explaining crustal low-velocity zones by such a mechanism could be nevertheless fascinating because *we feel that crustal low-velocity zones are associated with active tectonic areas where dislocation densities are high.*

6. *Conclusion*

A two-dimensional evaluation of seismic data provided by the ALP75 experiment in the northwestern part of the Alpine Arc shows that the crustal structure under the Northern Subalpine Chains is completely different from that under the inner part of the Arc. Intracrustal reflectors with an ESE mean dip of 30° are found in the first case, whereas no clear stratification is displayed in the second case. Moreover the boundary between the upper and lower crust is very uneven under the Subalpine Chains while the Moho discontinuity underneath shows no evidence of unevenness. In contrast with these stratification features, a low-velocity zone is found under the inner part of the

Alpine Arc at a depth of 11 to 23 km, with a wide transition zone between the upper and lower crust.

The structural cross-section arrived at shows a speculative overthrust of the intra-Alpine upper crust over the pre-Alpine upper crust. The high plasticity of the low-velocity zone could be the result of the temperature effect, hydrolytic weakening in quartz and high dislocation densities together, therefore furthering a 'décollement' in the basement around 15 km in depth. The front of this overthrust would be sited on a line with a SW–NE trend running along the western boundary of the 'Massifs Cristallins Externes'. This interpretation is consistent with former speculations by VIALON (1974) according to which the great suture lines of the Alpine basement, which were present at the end of the Hercynian orogenesis and which are believed to be the prolongation in the Alpine domain of the N50 Cévennes faults of the eastern margin of the French Massif Central, would have played a very prominent part during the Alpine orogenesis.

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