zones on the convex side of the ridge are larger than on the concave side, that is, the isochrons to both sides of the axis will not fit exactly when brought together. This situation probably did not exist in the early stage of opening of the Atlantic. Only later had the relief due to cooling developed sufficiently to produce a substantial ridge push. Continued field studies are needed to verify this and to give us the experimental data to refine the proposed model.

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# Teleseismic prospecting of lithospheric contrasts beneath the Pyrenees and Alps

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**The possibility of deriving the three-dimensional structure of the lithosphere by inversion of arrival times from distant earthquakes has previously been assessed using data from permanent seismic arrays[1]. Temporary emplacement of arrays in regions of key interest furnishes a practicable tool for investigating their structure[2,3], but the data needed for resolving complex three-dimensional structures are difficult to collect. We have therefore selected here regions where cylindrical symmetry reduces the problems to two dimensions and where the geological structure of interest is likely to produce large, easily detectable, relative arrival time anomalies. In this respect the depth variation of an interface with a large velocity contrast such as the Moho, as it is known to exist on average perpendicularly to the axis of alpine mountain ranges like the Alps and the Pyrenees, is of interest. In the case of the Pyrenees, explosion seismology profiling on parallel east–west lines in the North Pyrenean Zone (NPZ) and the Palaeozoic Axial Zone (PAZ)[4–7] provides evidence of an increase of over 15 km in Moho depth across the range. The variation in this depth beneath the North Pyrenean Fault (NPF)[8], a feature clearly expressed at the surface in the east as separating the NPZ and PAZ, can be shown to be very sharp, occurring over approximately 5 km. These results call into question the hitherto assumed smoothness of deep structure.**

The temporary array of seismic stations (Fig. 1) was maintained from the end of October 1982 to the beginning of January 1983. Ten earthquakes at distances of between 50° and 150° and with magnitudes between 5.1 and 6.3 were recorded by 8 to 18 stations. Arrival times undergo such sharp variations within the array that we cannot define a finite value of the wave slowness which could be used to infer smooth and deep structural variations[9]. Instead the data can be reduced to arrival time delays for each single station with respect to a standard propagation table of seismic waves[10]. Most stations were situated either on outcropping crystalline rocks or on basement with a thin cover of high-velocity pre-Cretaceous sediments. The northernmost stations were underlain by several kilometres of Cretaceous flysch and more recent sediments. Sedimentary delays can be accurately computed, reaching values as high as 1.0 s. For an event in Kazakstan (distance 53°, azimuth 53°) a sharp change in the relative delays of the order of 1 s occurs between the northern (early) and southern (late) stations over a distance of 15 km (station IZE to LAR). For a steeper incidence (Hebrides, distance 148°, azimuth 23°) a similar delay exists but this time IZE falls in the group of late stations. When projected back to the place where the rays cross the Moho (Fig. 2*a*) a cross-section is obtained which represents the Moho topography assuming

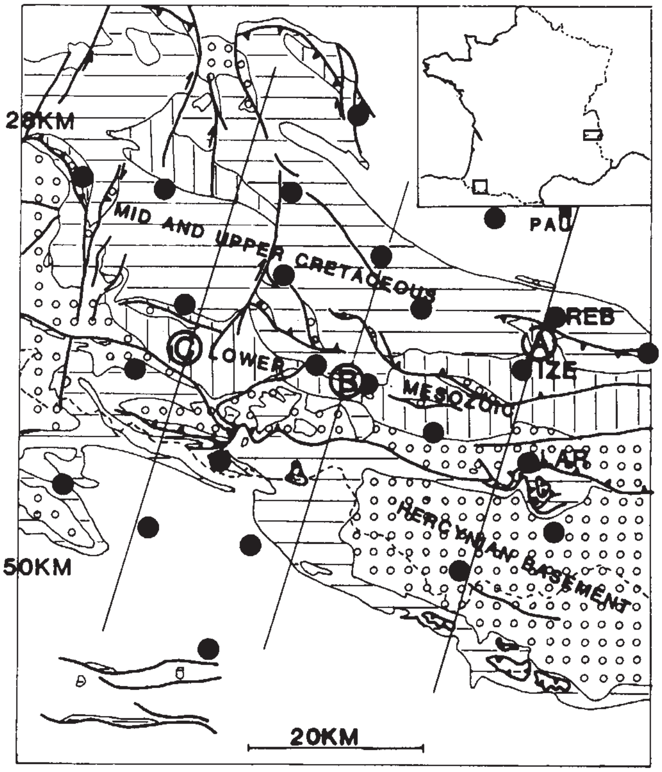


**Fig. 1** Temporary seismic stations in the Western Pyrenees. Depths to the crust–mantle interface, from east–west refraction profiles, are 28 km in the north, 50 km in the south. Epicentres of the three largest earthquakes in the region over the past 20 yr are indicated as A, B and C. Relative teleseismic residuals projected on lines through A, B and C are given in Fig. 2.

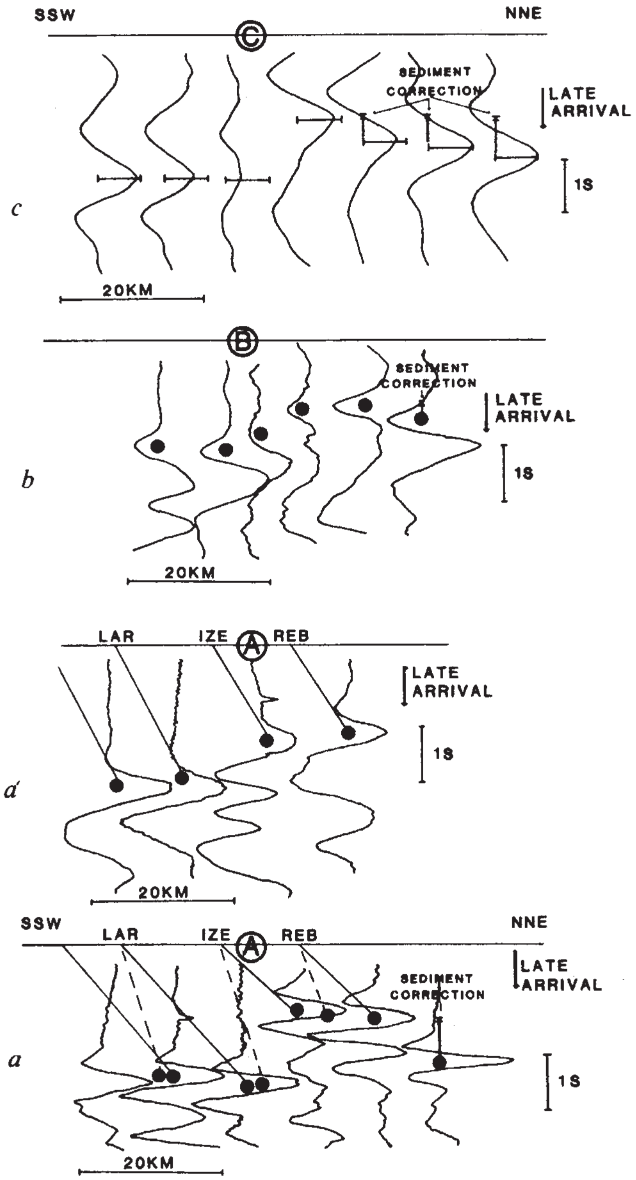


**Fig. 2** Relative teleseismic residuals across the Pyrenees. One second of relative residual corresponds to a change of ~20 km in Moho depth (vertical dimensions are thus compressed by 2.5 with respect to horizontal). *a*, Section through point A (see Fig. 1): P waves from Kazakstan event projected back to where they emerge from the mantle. Dots corresponding to New Hebrides earthquake with steeper incidence are added to show the sharpness of spatial variation in structure as data for IZE belong to early arrivals for the first and late arrivals for the second event. Correction for known sediments beneath the northernmost station account for 0.8 s later arrival. Note the change in delay beneath point A, indicating more than 20 km of variation in Moho depth. *a′*, Section through point A: seismograms of Salomon Islands earthquake intermediate in azimuth and incidence between the previous ones, with complex recording at IZE indicative of multipathing. *b*, Section through point B: seismograms are from Ecuador earthquake. The step to early arrivals (thinner crust) for northern stations is less sharp as here waves approach from the WSW where the crust is thicker. *c*, Section through point C: cross-correlograms of emergent waves from Kamchatka earthquake with a reference station. Note the change in delay beneath point C indicating also more than 20 km change in Moho depth. Taking into account known sediments beneath the three northern stations would bring arrivals as early as for the fourth one, indicating an almost flat shallow Moho.

the relative delays are wholly caused by its variation in depth. The 1-s difference between the southern and northern stations is consistent with the 20–25 km difference in Moho depth derived from explosion seismology[4–8], if the slightly larger

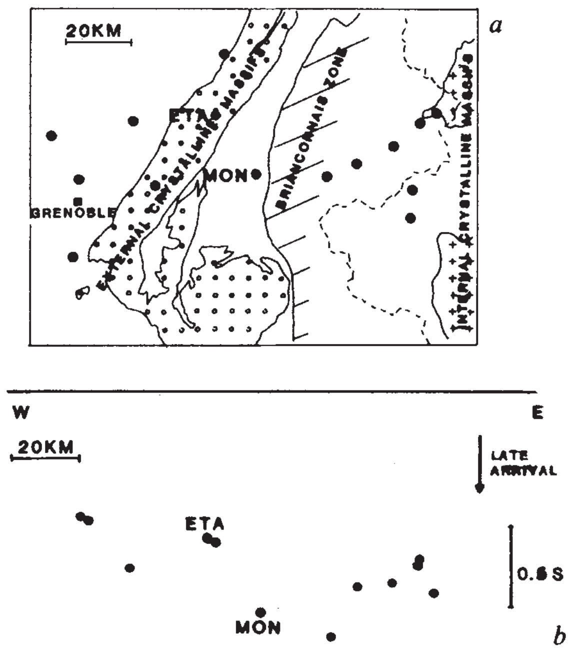


**Fig. 3** *a*, Temporary seismic array in the Western Alps. *b*, Relative residuals of waves arriving due north from Aleutian earthquakes on a west–east section of the array. Note the relative delay from ETA to MON.

average crustal velocity in the north is taken into account. Our data also indicate that this variation occurs very sharply on a north–south distance of the order of 10 km, near point A. This is additionally documented by a Salomon Islands earthquake, intermediate in azimuth and incidence between the previous two which at station IZE gives a complex seismogram indicative of multipathing (Fig. 2*a′*). This is evidence for a situation similar to that in the eastern part of the range—juxtaposition of two crustal segments of different thickness—and it illustrates the westward extension of the deep expression of the NPF. In Fig. 2*b* seismograms from an earthquake in Ecuador recorded by central stations in the array are displayed on a line through point B. Waves arrive from the WSW, almost opposite to the direction of approach from Kazakstan. The step to earlier arrivals for the northern stations is also documented here although it appears less clearly because the waves are approaching through thicker crust. Figure 2*c* presents a picture similar to Fig. 2*a* for the western group of stations in the array from an earthquake in Kamchatka (distance 82°, azimuth 12°); as the signals have progressive onsets and a longer period, the cross-correlograms have been plotted, projected back to where the rays cross the Moho. Beneath point C and within 10 km horizontally, a-1 s step is again documented, enabling extension of the deep expression of the NPF even further west. Tracing the NPF as a deep fracture further than seen at the surface provides additional support for evolutionary models emphasizing its fundamental regional role[11,12]. Extension of the NPF to the western half of the range is a subject of debate among geologists as surface evidence is not clear here although seismic activity is well documented[13].

As to the present state of the NPF, points A and B and C where we document the Moho disruptions were the exact epicentres of the three largest earthquakes which occurred in the western half of the Pyrenees over the past two decades—those at Arudy, 29 February 1980, Mb ~ 5.1 (ref. 14); at Arette, 13 August 1967, Mb ~ 5.5; and at Arbailles, 6 January 1982, Mb ~ 4.8. We therefore propose that the location of the main seismic activity in the present-day stress-field is controlled by preexisting heterogeneity in the inherited lithospheric structure.

Reconnaissance explosion seismology in the western Alps in the 'fifties and early 'sixties led to contour maps of Moho depth[15,16] which in this region showed an important but smooth dip of the Moho of 20% from west to east; however, the data constraining these contours are generally loose. All models of the deep structure of the western Alps have this smooth Moho with variable upper crustal structures thought to correspond to complicated overthrusts, with westward flaking of crustal segments, matching the surface decollements and nappes. Such models are used not only for the shallow mantle zone of Ivrea, but also for even further west than the external crystalline massifs[17]. We maintained our array for 6 weeks as an essentially west–east line from the foreland through the external crystalline massifs and the Briançonnais unit (Fig. 3*a*).

The most unexpected result of our teleseismic experiment for waves arriving from the north was a clear delay across the boundary between the crystalline massifs of the external zone and the Briançonnais unit of the Penninic zone to the east (Fig. 3*b*). The length of this delay (0.6 s) and the sharpness of the variation, which occurs between two stations 15 km apart, are similar to those across the NPF. Thus our data suggest that in the western Alps the deep structure may in places be similar to that of the Pyrenees, more closely resembling a juxtaposition of lithospheric segments than a smooth transition between them. Part of the anomaly might be explained by supposing that the crusts on either side differ in average velocity and composition, as has been proposed further to the west[17], or that mantle velocities are different. Nevertheless it seems difficult to account for the observations without supposing a step in an interface with a strong velocity contrast such as the Moho, which we consider better explains both the amount and sharpness of the anomaly.

Brief periods of recording a small number of teleseismic signals by a dense array in two geologically interesting regions of supposed lateral variation in crustal thickness have shown sharp and strong delay time anomalies. In the Pyrenees a problem is solved: the NPF remains a major fracture, with a step in the Moho in at least part of the western half of the range where its surface expression is not clear and where explosion seismology has not allowed precise definition. The coincidence of the surface projection of the Moho step, as detected from teleseismic anomalies, and the epicentres of the main earthquakes at upper and middle crustal depths imply that the lithospheric fracture extends almost vertically to the surface. In the case of the Alps, detection of similar variations in delay times brings into question the hitherto assumed smoothness of Moho depth variation and prompts the setting up of further specially designed explosion seismology experiments as used in the Pyrenees[4–7].

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# Time-dependent models of single- and double-layer mantle convection

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**One outstanding problem of great geophysical importance is the vertical extent of convection in the Earth's mantle[1]. Steady-state models of convection in the mantle have yet to produce a description of the mantle which is consistent with geochemical and geophysical observations. Here we present time-dependent numerical models of two-dimensional convection which simulate mantle convection. Starting from a fluid initially at rest with a purely conductive temperature profile, we now find that for a Rayleigh number (Ra) of $10^7$, the model experiences a transient period of double-layer convection, which lasts on the order of hundreds of millions of years when scaled to the Earth's mantle. It is suggested that transient periods of double-layer convection in the Earth's mantle may have existed in the past, whether or not such a period exists today.**

Models of mantle convection have been proposed with several different combinations of convecting layers, including convection confined to the upper mantle, convection in separate layers in the upper and lower mantle, and whole-mantle convection[2–10]. The models of double-layer convection have been motivated by geochemical analyses which suggest the presence of two (or more) sources of elementally and isotopically distinct rock, possibly the mantle above and below the 670-km seismic discontinuity[11–14]. This discontinuity may be caused by a phase transition or by a change in the chemical composition. Phase changes alone are unlikely to prevent convection across the 670-km discontinuity[15,16], but only a small increase in chemical density could prevent whole-mantle convection[16,17], though some mixing between the layers would still occur (P. Olson, personal communication). However, steady-state, two-layer convection implies a substantial temperature drop across the intervening conductive boundary layer, which requires a much lower viscosity in the lower mantle than in the upper mantle[9]. The latter is inconsistent with results which show the viscosity of the mantle to be nearly uniform[18].

Convection is in general a time-dependent phenomenon, rather than steady-state. Previous models of time-dependent convection have produced only a single layer of convecting cells[19–25]; often the time dependence consisted of an oscillation about a steady-state pattern. However, our models[26] show that as Ra is increased towards values appropriate for the present mantle, qualitatively different evolutions result, at least for certain initial conditions.

The models are based on the Boussinesq approximation at infinite Prandtl number, and treat convection in a two-dimensional, cartesian coordinate box (aspect ratio of 1.4), heated from below. The equations of motion are solved numerically by a computer code originally written by Lux[27]. The biharmonic equation is solved directly by the method of conjugate gradients, while the energy equation is solved by an alternating directions-implicit method with upwind differences for the advective terms. The viscosity is taken to be exponentially dependent on the temperature (proportional to $\exp[6(0.5-T)]$, in terms of the dimensionless temperature $T$)[28], so that a variation by a factor of 400 is possible between the hottest ($T=1$) and coldest ($T=0$) regions. This variation is intended to simulate at least partially the strong temperature and pressure dependence of the viscosity presumed to occur for mantle matter[1]. The Rayleigh number for these variable viscosity models is based on the viscosity for $T=0.5$, which occurs at the centre of the layer in steady state. The boundary conditions are free slip, with impenetrable, insulating side walls. The time step is at most a fraction of the Courant–Friedrichs–Lewy time step, ensuring temporal resolution of time-dependent flows. The spatial resolution of the models is sufficient to include at least three grid points in the