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Low-Pressure Regional Metamorphism in the Pyrenees and its Implications for the Thermal Evolution of Rifted Continental Crust [and Discussion]

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Low-pressure regional metamorphism in the Pyrenees and its implications for the thermal evolution of rifted continental crust

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During late Palaeozoic (Hercynian) low-pressure regional metamorphism in the Pyrenees, exceptionally high thermal gradients existed within the upper crust, and temperatures as high as 700 °C were attained at depths as shallow as 10 km, resulting in large-scale crustal anatexis. Stable isotope studies indicate that the crust was flushed by circulating ground waters to depths of 12 km, but the amount of fluid involved below 8 km was probably not much greater than 50% of the rock mass, and this fluid apparently did not penetrate the pre-Palaeozoic basement below 12 km. There is no evidence for continental collision in the region at that time, and these data, together with other geological and geophysical constraints, suggest that the most plausible tectonic setting for the metamorphism is a zone of continental rifting, possibly associated with strike-slip movement. Thermal modelling suggests that a transient, high-temperature heat source in the lower crust is required to account for the observed metamorphic *P–T* arrays. Among a range of possible solutions, a basaltic sill, 6–8 km thick and emplaced at 14 km could generate a maximum temperature array similar to those observed in the Pyrenees.

Introduction

High-temperature–low-pressure regional metamorphic terrains are relatively common in the geological record, and are particularly notable in the Hercynian of Western Europe (Bard *et al.* 1980; Zwart 1979; Guitard 1970; Wickham 1986*a*). The mineral assemblages of such terrains imply that over wide areas the crust became unusually hot at shallow depth, and that exceptionally high thermal gradients were developed in the upper crust. The Hercynian basement rocks exposed in the Pyrenees are a good example of this phenomenon; their mineral assemblages imply that temperatures of 700 °C were attained at depths of as little as 10 km and that, locally, thermal gradients exceeded 100 °C km^{-1} (Zwart 1979; Fonteilles 1981; Soula *et al.* 1986; Wickham 1986*a*).

In previous papers (Wickham & Oxburgh 1985, 1986, we have suggested that the tectonic setting best suited to explain the metamorphic phenomena in the Hercynian of the Pyrenees is that of rifted continental crust. Here we synthesize field and petrological data from the metamorphic rocks constraining the thermal structure of the Hercynian crust, with stable isotope data constraining the circulation of aqueous fluid within it, in an attempt to model quantitatively its thermal evolution. We conclude that an influx of mantle-derived mafic magma into the lower crust, at the time of rifting, provided the main heat source for the metamorphism.

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2. Hercynian metamorphism in the Pyrenees

Early Tertiary uplift in the Pyrenees has exposed an extensive Hercynian basement terrain (figure 1) composed of Palaeozoic metasediments, volcanics, high grade ortho- and paragneisses and a variety of granitoids. All these lithologies were metamorphosed or intruded during the late Carboniferous–early Permian Hercynian orogeny (Bard *et al.* 1980). Unlike the Alps, Tertiary deformation and metamorphism of the basement in the Pyrenees was mainly limited to the major fault zones, between which most of the rocks remain in relatively pristine condition (Zwart 1979). Exposure is excellent, and the differential elevation of the various basement blocks (e.g. the North Pyrenean Massifs in figure 1), allows access to a variety of structural levels, representing a plausible composite section of the Hercynian crust (see, for example, Vielzeuf 1984). This section ranges from granulite facies gneisses ('basal gneisses') to late Carboniferous and early Permian supracrustal sediments and volcanics. The uppermost parts of the Palaeozoic section were still being deposited at the time that the deeper units were undergoing regional metamorphism (Wickham & Oxburgh 1986). These rocks provide a comprehensive record of the response of the middle and upper crust to low-pressure regional metamorphism. Only a brief description of Hercynian geology in the Pyrenees is given here, and the reader is referred to Zwart (1979), Bard *et al.* (1980), Vielzeuf (1984), Wickham (1986*a*) and Wickham & Oxburgh (1986) for more detailed accounts.

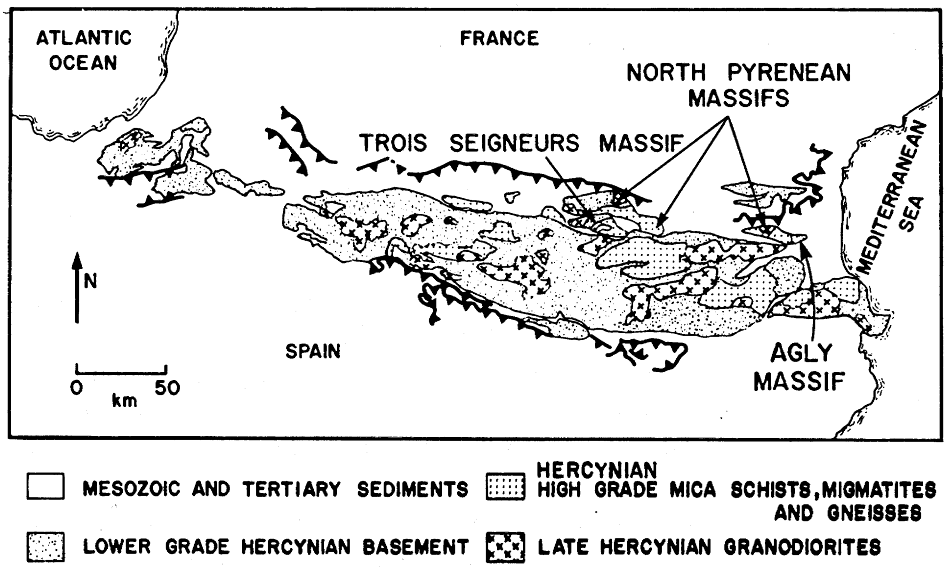

Figure 1. Hercynian basement outcrop in the Pyrenees. The location of the Agly, Trois Seigneurs and other North Pyrenean Massifs is indicated. The boundary between low- and high-grade Hercynian metamorphic rocks corresponds approximately to the appearance of biotite in pelites.

A typical Hercynian metamorphic sequence is exposed in the Trois Seigneurs Massif (figure 1), where a continuous section comprising mainly pelitic lithologies ranges from low-grade shales, through andalusite and sillimanite schists to migmatites and peraluminous granitoids (Wickham 1984, 1986*a*). A true-scale section through part of the metamorphic sequence is shown in figure 2 and indicates the proximity of the low- and high-grade rocks. Similar sequences are exposed in many other parts of the Pyrenees (see high-grade regions in

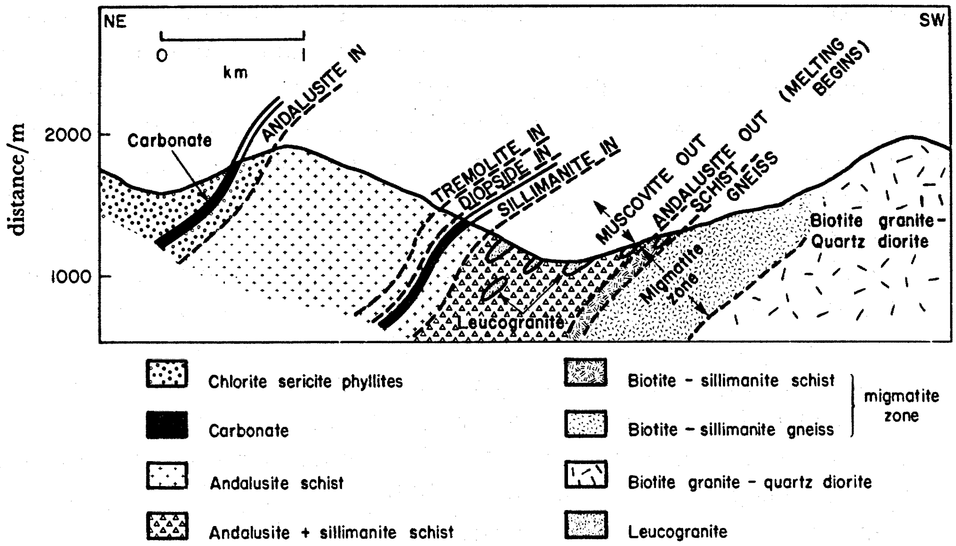

FIGURE 2. Schematic true-scale section through the Trois Seigneurs metamorphic sequence, indicating the progressive gradation from low-grade phyllites through high-grade mica schists and migmatites to granitoids derived mainly through anatexis of pelitic metasediment. Note the proximity of the low and high-grade rocks.

figure 1) and all show a remarkably rapid gradation from low to high grades over distances of only a few kilometres, implying the existence of very high temperature gradients during metamorphism. In other places, such as in the Agly and other North Pyrenean massifs (figure 1) granulite facies gneisses containing higher-pressure mineral assemblages occur, probably representing deeper structural levels in the Hercynian crust than those exposed at Trois Seigneurs (Vielzeuf 1984). Although structural and geochronological evidence is ambiguous, it has been suggested that these 'basal gneisses' represent the old metamorphic basement on which the Palaeozoic sediments were deposited, and which has been further metamorphosed during the Hercynian (Guitard 1970; Vitrac-Michard & Allegre 1975; Zwart 1979).

In the Trois Seigneurs Massif, petrological evidence indicates that the migmatite zone equilibrated at pressures of 3–4 kbar† and temperatures of 670–700 °C (Wickham 1984; 1986*a*). This pressure range is, in part, implied by the absence of kyanite and staurolite from the pelitic rocks and the presence of only very Mn-rich garnet. Other constraints include (1) the isograd sequence 'sillimanite in' – 'muscovite out' – 'melting begins' that is permitted over a relatively narrow range of pressures, approximately between 3 and 5 kbar for water activities greater than 0.5; (2) the stability of biotite during melting that implies pressures greater than 3 kbar; (3) biotite–cordierite geobarometry (Holdaway & Lee 1977), which indicates pressures of 3–4 kbar for Trois Seigneurs pelitic migmatites. Experimental data indicate that the onset of melting in pelitic rocks at these pressures and with high water activity would be between 670 and 700 °C (Hoffer 1978; Wickham 1984).

As documented elsewhere (Wickham 1986*a*), metamorphism and anatexis evidently occurred under water-rich conditions with a_{H_2O} buffered near unity. This is suggested by the rapid, gradational transition from 0% to 50% partial melting of the pelitic rocks over only 200–300 m of the metamorphic section. High a_{H_2O} is also implied by mineral assemblages in the metacarbonates (Wickham 1984).

† 1 bar = 10^5 Pa.

The narrow separation of the isograd surfaces shown in figure 2 suggests the existence of high thermal gradients during metamorphism, of about 80–100 K km^{-1} (Wickham 1986*a*). If we assume that the migmatite zone equilibrated at 700 °C and at 12 km depth (Wickham 1986*a*) then, by using this *P–T* fix as a datum, a unique depth can be assigned to each of the other isograds in the metamorphic sequence. The temperature represented by each isograd can then be estimated from the relevant phase equilibria (Wickham 1986*a*). The array of *P–T* points thus generated suggests that temperatures ranged from about 700 °C at 12 km to 450 °C at 9.5 km. However, it must be emphasized that there is considerable uncertainty in the absolute depth values, and in the gradient of the array because of possible post-metamorphic deformational effects, although these are not believed to have been large (Wickham 1984; 1986*a*). Metamorphism and anatexis were accompanied by deformation, generating mesoscopic tight, flat-lying isoclinal folds in the high-grade rocks, with axial surfaces parallel to the isograds (Zwart 1979; Soula 1982; Verhoef *et al.* 1984; Wickham 1984, 1986*a*, *b*; Soula *et al.* 1986). At Trois Seigneurs, boudinage of leucogranite pods within the main schistosity (which parallels the isograds) suggests that attenuation of the sequence may locally have been as high as 50%. It appears, however, that many of the leucogranites were intruded, and deformed, *before* the thermal maximum at high structural levels (see Wickham 1986*b*). Other evidence suggesting little post-metamorphic deformation includes the petrographic observation that some andalusite prophyroblasts overgrow the main schistosity, implying that prograde heating continued after this schistosity had formed. The lack of deformation of coarse, high-temperature retrogressive muscovite, which is a common phase within the high-grade pelites and granites, also indicates limited late attenuation (Wickham 1984, 1986*a*).

On balance, post-metamorphic shortening of the sequence appears to have been substantially less than 50%, averaged throughout the section. However, in the thermal models presented in this paper, we have attempted to satisfy both the *P–T* array observed in the area today (700 °C at 12 km to 450 °C at 9.5 km) and also a hypothetical, grossly expanded *P–T* array (700 °C at 18 km to 450 °C at 12 km) to allow for the most extreme attenuation of the sequence that is conceivable.

Fault-bounded, granulite facies gneisses in the Agly Massif and other North Pyrenean massifs have mineral assemblages implying equilibration at pressures of 4–7 kbar based on garnet–plagioclase–sillimanite (Newton & Hasleton 1981) and garnet–plagioclase–orthopyroxene (Newton & Perkins 1982; Bohlen *et al.* 1983) geobarometry, and temperatures in the range 750–850 °C (see, for example, Vielzeuf 1984). The 'basal gneisses' from the Agly Massif (see below) probably equilibrated at 5 kbar, and these and the other granulites have experienced higher pressures than any other metamorphic rocks in the Hercynian of the Pyrenees. They may therefore be representative of that part of the crust immediately below the Trois Seigneurs sequence (see, for example, Zwart 1979). This is also supported by geochemical and litho-stratigraphic evidence that the basal gneisses are the pre-Palaeozoic basement on which the Palaeozoic sediments were deposited (Zwart 1979; Guitard 1970; Vitrac-Michard & Allègre 1975). However, further geochronology is required to provide a clear correlation between the low-pressure granulite facies metamorphism in the 'basal gneisses' and that observed in the Palaeozoic metasediments.

The scale of melting seen in the low-pressure metamorphic–anatectic sequences was much less than that which occurred deeper in the crust. In addition to the peraluminous granitoids that were generated mainly through anatexis of pelitic material at 10–12 km depth (such as are exposed at Trois Seigneurs (figure 2) and elsewhere in the Pyrenees), by far the most significant phase of Hercynian magmatism is represented by the widespread, mainly granodioritic bodies (e.g. Maladeta, Mont Louis), which have invaded virtually unmetamorphosed Upper Palaeozoic sediments throughout the Pyrenees (Zwart 1979; Autran *et al.* 1980). At Trois Seigneurs, contact relations suggest that granodiorite intrusion postdated the regional metamorphic peak (Wickham 1984, 1986*a*). This, and the relatively high levels of emplacement, make it unlikely that the granodiorites contributed to the thermal effects within the metamorphic sequences. Isotopic data (Vitrac-Michard *et al.* 1980; Ben Othman *et al.* 1984; Wickham & Taylor 1985) imply that these plutons are crustal melts, but their bulk chemistry is inconsistent with their being derived from either the Palaeozoic metasediments or the 'basal gneisses'. Their existence implies very extensive melting of the Hercynian lower crust, occurring on a much larger scale than the effects observed at higher structural levels (e.g. the Trois Seigneurs migmatite zone). This has important implications for Hercynian thermal structure and will be discussed later.

At higher structural levels, the strong, shallowly dipping fabrics typically seen in the mica schists and migmatites pass into weaker, upright structures in the lower-grade rocks. Structural interpretation of these relations has been offered by Zwart (1979), Soula (1982), Soula *et al.* (1986) and Verhoef *et al.* (1984). Analysis of structures present in the metamorphic sequence developed in the Aston Massif led Verhoef *et al.* (1984) to recognize that the 'main' Hercynian folding episode was in fact a complex superposition of several distinct phases of deformation, which overprint successively at higher metamorphic grades. It is difficult to assign a specific tectonic setting to the Hercynian events based only on the deformational style, and more detailed work is required in a number of regions, coupled with accurate geochronology, before this can be done with any certainty. However, any tectonic model must be capable of explaining the characteristic, widely observed change from low-grade upright structures to high-grade shallow structures, and the general increase in the intensity of deformation with structural depth (Zwart 1979). The upright structures have been related to compressional tectonics predating the Hercynian metamorphic peak (Verhoef *et al.* 1984). The roughly synmetamorphic, flat-lying structures appear to have been superimposed on the steep structures only within the high-grade areas (Soula 1982; Wickham 1984; Verhoef *et al.* 1984). This deformation sequence may reflect extension (and metamorphism) in localized regions following an episode of more widespread compression and crustal thickening, although other interpretations relating to diapirism of granitoids have also been suggested (Soula 1982; Soula *et al.* 1986; see also discussion following this paper).

The precise age of the Hercynian metamorphism in the Pyrenees is not well known and may, to some extent, be regionally diachronous. Available geochronological data imply that the age lies somewhere between 335 and 280 Ma before present (Vitrac-Michard & Allègre 1975; Vitrac-Michard *et al.* 1980; Ben Othman *et al.* 1984; Bickle *et al.* 1985; Bickle & Wickham, unpublished results). Some of these data are summarized in Wickham & Oxburgh 1986, figure 2. If metamorphism occurred within this time-span, the Upper Carboniferous–Permian sedimentary record in the Pyrenees must provide important information about surface processes *during* metamorphism. It also underscores an important difference between the Hercynian metamorphism in the Pyrenees and that observed in tectonically buried regions associated with continental collision, where the termination of sedimentation typically precedes the metamorphic maximum by some tens of millions of years (see, for example, England &

Thompson 1984). In a rift zone, sedimentation might be expected to continue throughout the duration of any thermal perturbation associated with rifting.

3. Late Palaeozoic regional setting

The Palaeozoic metasediments that record the Hercynian metamorphism in the Pyrenees comprise a thick sedimentary sequence ranging from the Cambrian to the Upper Carboniferous. During this time deposition was more or less continuous (Zwart 1979). The Lower Palaeozoic is dominated by shales, while in the Devonian and Lower Carboniferous carbonates are more common. Condensed, andalusite–sillimanite-type metamorphic sequences are typically developed within Cambro-Ordovician metapelitic rocks. The Upper Carboniferous metamorphism therefore occurred at the end of a prolonged period of basin subsidence, during which time deposition had continued over several hundred million years.

In the Pyrenees, the stratigraphic progression from Carboniferous to Permian sees a change from marine sedimentation to continental sedimentation interrupted by angular unconformities, but there is no major gap in stratigraphic ages (Zwart 1979). Andesitic and rhyolitic volcanic horizons are common, but are probably related to late granodiorite magmatism, rather than representing extrusive equivalents of melts generated during metamorphism of the Palaeozoic sequence. In general, folded, cleaved Upper Carboniferous turbiditic flysch is unconformably overlain by Stephanian ignimbrites and andesitic volcanics, which are in turn overlain by Lower Permian continental deposits (red beds, calcretes and pisolitic limestones).

Some of the occurrences of Stephano-Permian volcanics and sediments in the Pyrenees appear to have been deposited in narrow, fault-bounded basins or staggered, en-echelon half grabens (Bixel 1985), suggesting a strike-slip setting with deposition in local pull-apart basins. The same tectonic setting was suggested by Reading (1975) for Westphalian to Permian sediments in the Cantabrian Mountains (the westerly extension of the Pyrenees). Here there are thick sedimentary deposits formed in small, deep basins, showing rapid facies changes and common angular unconformities. Although it remains to be conclusively demonstrated that these rocks were deposited at the same time as the metamorphism in the Pyrenees, they imply that extensional regions existed within a wide belt of strike-slip movement, operating throughout the Pyrenean–Cantabrian region.

Such movement is compatible with widespread Stephanian–Permian wrench tectonics occurring throughout Western Europe (Ziegler, 1982; Arthaud & Matte 1977). Movements to accommodate shortening at this time in the Urals and the Appalachians were linked by a right-lateral transform fault system causing the development of a complex pattern of conjugate faults and related pull-apart structures that cross the Hercynian belt in Europe (figure 3). Rifting in the Pyrenean region may be a local expression of a very widespread phenomenon in Europe at that time, including early rifting in the North Atlantic, and the formation of narrow but deep wrench basins in various parts of the Bohemian Massif, the Massif Central and in the Saar–Nahe Trough, where over 3500 m of late Carboniferous sediments and volcanics were deposited (Ziegler 1982).

In the Pyrenees, oxygen and hydrogen isotopic data imply that the mica-schists and migmatites (zone 2 of Wickham & Taylor, 1986) equilibrated with surface-derived seawater at temperatures in excess of 500 °C during prograde Hercynian metamorphism (Wickham & Taylor 1985, 1986). This suggests that the ground surface was below sea level at the time. Taken

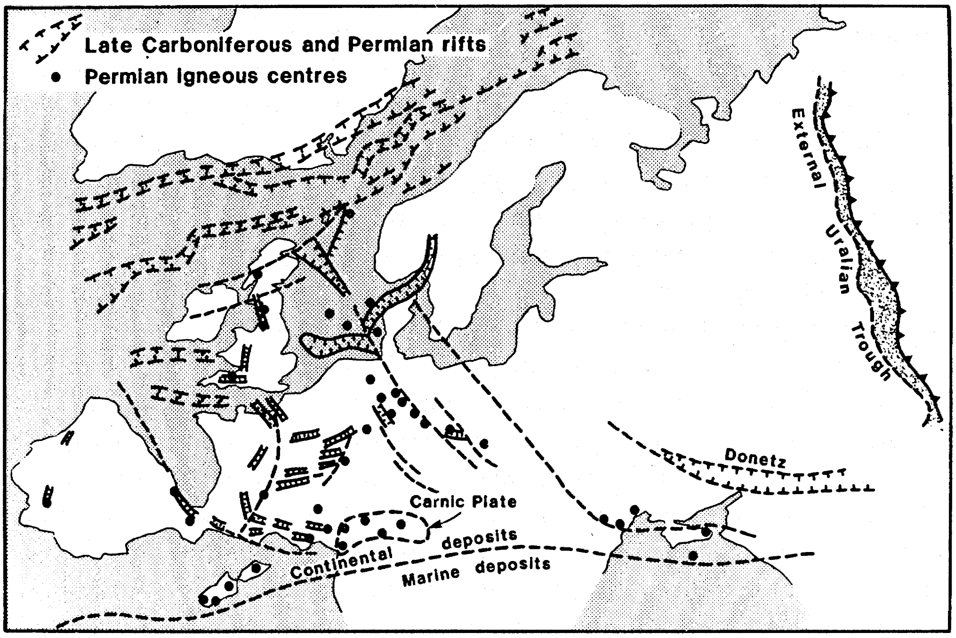

FIGURE 3. Late Carboniferous and Permian rifting in Western European after Ziegler (1982); Arthaud & Matte (1977).

with the striking contrast with collisional settings (no high-pressure metamorphic assemblage, no ophiolitic or oceanic lithology preserved, no large-scale overthrusting, and apparently continuous sedimentation during metamorphism) the combined geological evidence suggests that the metamorphism and its associated anomalously high thermal gradients may have been generated by a rifting event, associated with intrusion of mafic material (Wickham & Oxburgh 1985, 1986). The similarity between the inferred temperature structure of the Hercynian crust and that of modern continental rift zones is striking (see, for example, Lachenbruch *et al.* 1985; Baldridge *et al.* 1984), although in the Pyrenees the exact chronology of deformation, metamorphism, melting and fluid infiltration at depth, and volcanism and sedimentation at the surface remains to be established.

4. Isotopic constraints on fluid flow

(*a*) *Constraints on fluid volume*

Any attempt to understand the thermal evolution of metamorphic belts must take into account the influence of fluid flow on temperature distribution (see, for example, Oxburgh & England 1980; England & Thompson 1984). This is because even slow movement of fluid within the crust may involve heat transfer and the consequent modification of geotherms.

In the Hercynian of the Pyrenees, consideration of fluid movement is especially important because it has been shown that in the Trois Seigneurs Massif and elsewhere, the metamorphic rocks have been flushed by surface-derived aqueous fluid while they were at high temperatures (Wickham & Taylor 1985, 1986 and unpublished results). This conclusion was based on a study of the oxygen and hydrogen isotopic compositions of the metamorphic and igneous rocks and minerals. However, petrological evidence (see above and Wickham 1986*a*) also indicates that conditions were water-rich during metamorphism, and that a net influx of water into the anatectic region was necessary to promote the generation of the observed quantity of melt from the pelitic rocks, and account for the mineral assemblages observed in the metacarbonate rocks.

The oxygen isotope study of Wickham & Taylor (1985) enabled quantitative estimation of the amount of water involved in infiltration. Using a simple, closed-system mass-balance model (see Wickham & Taylor 1985 and Taylor 1977 for details), the variation in whole rock $\delta^{18}O$ of the metasediments with increasing water:rock ratio was calculated, and this is plotted in figure 4 for two likely extreme limiting values of -1 and $+8$ for $\delta^{i}_{H_2O}$, the initial $\delta^{18}O$ of the infiltrating water. This diagram can be used to constrain the minimum quantity of water involved in the homogenization process. The fields shown in figure 4 are the ranges of values shown by the mica schists at Trois Seigneurs, and indicate that a water:rock ratio of at least 0.2 is required to shift them by the observed amount ($+14.5$ to $+12.5$ in the metasediments), even in the unlikely event of the infiltrating water being pristine seawater with $\delta^{18}O$ of -1. In the more likely case of the infiltrating fluid's having $\delta^{18}O$ of $+4$ to $+6$, the water:rock ratio would be 0.3 to 0.5. The homogeneous $\delta^{18}O$ of $+11$ to $+12$ shown by most of the metasediments, and virtual absence of rocks with values lower than this, reflects this relatively well-defined upper limit to the water:rock ratio. It is clear that a large but finite volume of water passed through the rocks during metamorphism, and that this water, probably had a mass less than 50% of the total rock mass affected by homogenization (figure 4).

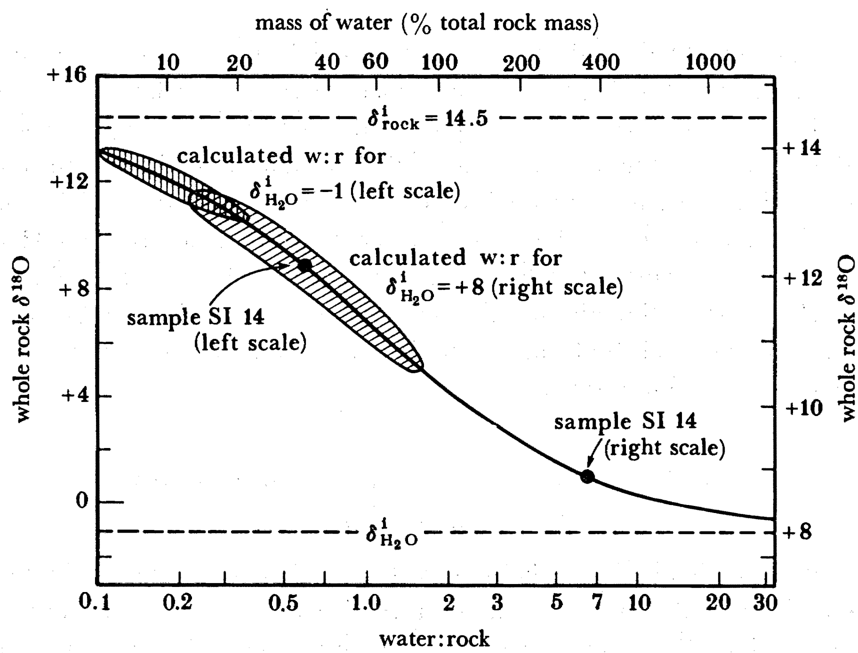

FIGURE 4. Whole rock $\delta^{18}O$ against calculated water:rock (w:r) ratio (closed system) for two different assumed initial $\delta^{18}O$ values (-1 on the left-hand scale and $+8$ on the right-hand scale) after Wickham & Taylor (1985). The same curve applies to both sets of calculations, and the two fields indicate the general range of $\delta^{18}O$ values shown by the mica schists. For an intermediate-composition water ($\delta^{i}_{H_2O} = +4$ to $+6$), water:rock ratios would be in the range 0.2 to 0.7. Sample SI 14 is an anomalously low-^{18}O metasediment, and if we assume that it also started with $\delta^{18}O = +14.5$, it must have experienced a higher water:rock ratio than any other sample.

(*b*) *Constraints on temperature gradient during metamorphism*

The homogeneity of whole-rock $\delta^{18}O$ values in a variety of lithologies from the Trois Seigneurs metamorphic sequence suggests that the rocks were pervasively flushed with aqueous fluid at some stage during metamorphism. This is shown by considering the data for quartz from pelites, granites and carbonates in the Trois Seigneurs sequence, as shown in figure 5. In this diagram, four new data points have been added to the data set of Wickham & Taylor (1985, figure 8). Complete oxygen isotopic equilibration at constant temperature throughout the metamorphic pile would result in quartz having the same isotopic composition everywhere, regardless of its lithological environment. However, if the values reflect equilibration with a large volume of water at the peak metamorphic temperature experienced by each point in the section, then the quartz $\delta^{18}O$ values should vary systematically across the area. This is because the close spacing of the isograds implies that a steep thermal gradient existed across the metamorphic sequence. Since the equilibrium $\Delta^{18}O$ quartz–water fractionation varies with temperature, $\delta^{18}O$ quartz should decrease by between 2.0 and 3.6‰ (Friedman & O'Neil 1977; Matthews *et al.*, 1983) over the inferred 250 K temperature increase between the 'andalusite in' isograd (450 °C) and the biotite granite (700 °C).

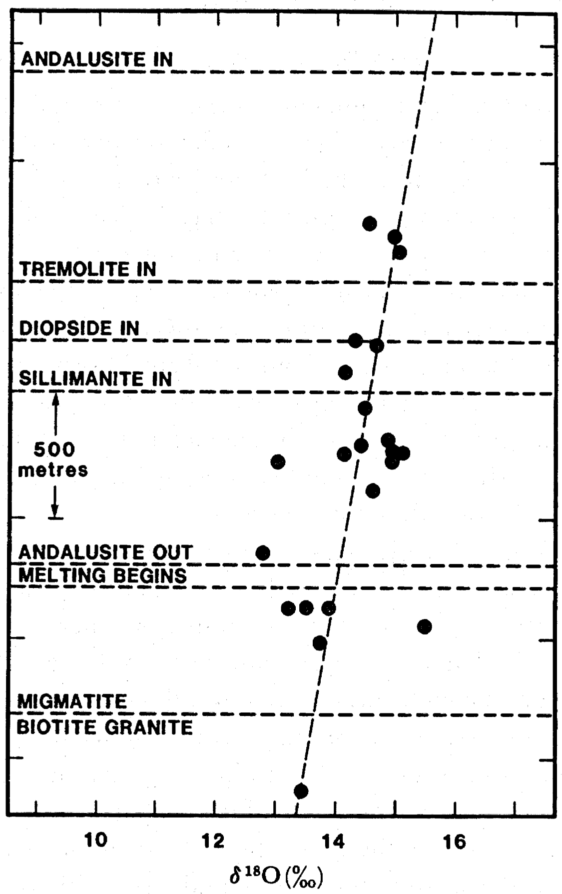

FIGURE 5. $\delta^{18}O$ of all quartz samples analysed throughout the higher-grade part of the metamorphic sequence. Note the relative homogeneity of quartz isotopic compositions in a variety of lithologies. A least-squares regression through the data points shows a change in $\delta^{18}O$ of about 2‰ between the 'andalusite in' isograd and the 'biotite granite'. However, the correlation coefficient is only 0.49. ●, Quartz from all lithologies.

A least-squares regression fitted through the data of figure 5 shows that there is some tendency for the quartz $\delta^{18}O$ values to become progressively lighter with increasing metamorphic grade by about 2.0‰ over this part of the sequence, although the data show considerable scatter and the correlation coefficient is only 0.49. The variation is rather less than that predicted by experimental measurements, if the isotopic values were equilibrated at peak metamorphic temperatures. Taken with other considerations, this caused Wickham & Taylor (1985) to propose that most of the circulation of fluid and $^{18}O:^{16}O$ homogenization occurred just before the metamorphic peak, when the thermal gradient was lower than that recorded by the isograds. However, the approximately 2‰ variation in $\delta^{18}O$ quartz shown in figure 4 could be compatible with a temperature difference of 150–250 K between the 'andalusite in' isograd and the deep biotite granite, within the general temperature range 450–750 °C, based on the experimental quartz–water calibrations of Clayton *et al.* (1972), Bottinga & Javoy (1973) and Matthews *et al.* (1983). The quartz isotopic variation is thus consistent with a pressure–temperature array of at least 60 °C km^{-1} through the metamorphic sequence at Trois Seigneurs. This value is not substantially different from the array of 80–100 °C km^{-1} implied by the separation of the metamorphic isograds (Wickham 1984, 1986*a*), although on balance it is likely that most fluid circulation had ceased before peak metamorphic temperatures were reached (see, for example, Wickham & Taylor 1985).

(*c*) *Constraints on the lower limit of fluid penetration*

The groundwater circulation documented at Trois Seigneurs and elsewhere in the Pyrenees must die out at some depth in the crust where the effective permeability becomes too low for fluid flow. High-grade paragneisses from the La Pège region in the Trois Seigneurs Massif (see, for example, Wickham & Taylor 1985, figure 2) may indicate this, because they lie at the deepest exposed structural levels in this region and appear to have been less thoroughly isotopically homogenized than the other metamorphic rocks in the area (Wickham & Taylor 1985, 1986). More convincing data indicating the lower limit to groundwater penetration come from granulite facies 'basal gneisses' from the Agly Massif (see figure 1). As discussed above, these rocks probably represent a deeper structural level of the Hercynian crust than the section exposed at Trois Seigneurs. They are lithologically heterogeneous, mostly comprising granitic and pelitic gneisses with thin carbonate bands, and are therefore good monitors of the degree of isotopic homogenization that has occurred.

Oxygen isotope data from the Agly Massif indicate that the 'basal gneisses' are far less homogenized than the Trois Seigneurs metasediments. This is illustrated in figures 6 and 7, where isotopic profiles across thin metacarbonate units from each area are shown. The initial sedimentary $\delta^{18}O$ value of the Palaeozoic carbonate rocks in the Pyrenees was probably about +25, but at Trois Seigneurs the metacarbonates now all have whole rock $\delta^{18}O$ values of +11 to +13, almost completely overlapping the range of values shown by the pelitic rocks (Wickham & Taylor 1985). The uniformity of values within an individual carbonate unit from the sillimanite zone is shown in figure 6. The $\delta^{18}O$ value of calcite shows little variation across the entire 15 m thick band, regardless of the percentage of calcite in each sample, and is close to the isotopic composition of the surrounding pelitic and psammitic rocks.

The data of figure 6 contrast with those of figure 7, where a similar isotopic profile through metacarbonate units within a sequence of granulite facies pelitic gneisses in the Agly Massif is shown. Although these units are substantially thinner than the one shown in figure 6, and

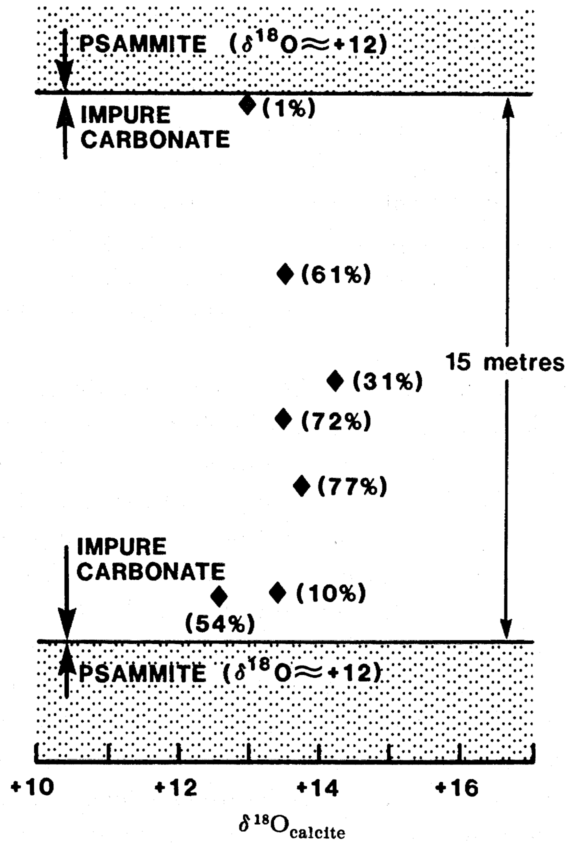

FIGURE 6. Oxygen isotope profile through a metacarbonate band SI 13 from the Trois Seigneurs Massif. The lithologies to either side are sillimanite bearing mica schists and psammites with whole-rock $\delta^{18}O$ of about +12. The isotopic composition of calcite within the entire 15 m unit is fairly constant at about +13.5, indicating pervasive isotopic equilibration with the surrounding rocks. The figures in parentheses denote the percentage calcite in each sample.

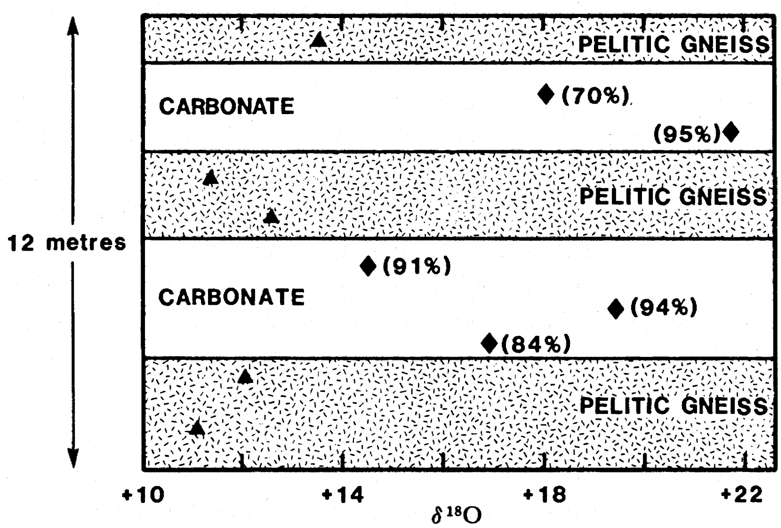

FIGURE 7. Isotopic profile through two thin metacarbonate bands within granulite facies pelitic gneisses in the Agly Massif. The pelitic gneisses have $\delta^{18}O$ values in the range +11 to +14, while calcite from the carbonate units has retained much heavier values which are sometimes as much as 10‰ higher. Steep isotopic gradients are preserved at the margins of the units (especially in the case of the upper one) and isotopic homogenization has been very limited. The percentage calcite in each sample is shown in parentheses and does not appear to correlate with $\delta^{18}O$ calcite. The data contrast strongly with those from Trois Seigneurs and imply that pervasive fluid circulation did not occur within the Agly gneisses. ◆, Calcite; ▲, whole rock.

despite the fact that the Agly rocks were metamorphosed at higher temperatures (Fonteilles 1981; Vielzeuf 1984), there has been little homogenization of $\delta^{18}O$ values. The isotopic composition of calcite from the metacarbonates is much richer in ^{18}O than the surrounding pelitic gneisses, and steep isotopic gradients are preserved at the margins of each unit. This implies that there has been little aqueous fluid transfer between the carbonates and adjacent pelitic gneisses during metamorphism, because such oxygen-bearing fluids are very effective at promoting isotopic exchange, especially at high temperatures. Clearly, the Agly 'basal gneisses' were metamorphosed under very different conditions from the Trois Seigneurs metasediments, and did not experience pervasive flushing with aqueous fluid (Wickham & Taylor 1986).

If the Agly and Trois Seigneurs rocks can be thought of as different elements of a composite crustal section, then the data of figures 6 and 7 imply that the lower limit to the penetration of aqueous fluid during the Hercynian metamorphism lay somewhere between the structural levels represented by the Trois Seigneurs metasediments and the Agly 'basal gneisses'. However, as discussed by Wickham & Taylor (1986), the assumption of a composite section remains uncertain because we cannot prove conclusively that the isolated occurrences of 'basal gneiss' in the Pyrenees are, in fact, representative of the 14–25 km depth range in the Hercynian crust. If the Agly rocks do indeed represent the basement on which the Palaeozoic sediments were deposited (Zwart 1979; Vitrac-Michard & Allègre 1975), then it is tempting to suggest that the lower limit of groundwater penetration was the basement–cover interface.

The oxygen isotope systematics of the Hercynian rocks from Agly and Trois Seigneurs suggest limits to the volume of fluid that passed through the metasediments, the depth of the fluid flow system, and the temperature distribution in the metamorphic rocks during fluid circulation and isotopic equilibration. In the next section we use these constraints to establish a thermal model for the thermal structure of the Hercynian crust.

5. Thermal aspects of Hercynian metamorphism in the Pyrenees

(a) *The thermal problem*

The problem posed by the metamorphic–anatectic sequence exposed in the Trois Seigneurs Massif, and in many other parts of the Hercynian basement of the Pyrenees, is that the mineral assemblages provide evidence of *both* abnormally high temperatures *and* abnormally high temperature gradients at depths of 9–12 km in the crust. Such extreme P–T régimes could not be accommodated by any plausible steady-state model of the crust in conductive thermal equilibrium, and also satisfy reasonable geological and geochemical constraints. For example, the depth of the onset of melting in Trois Seigneurs pelites at 700 °C and about 10–12 km implies a mean thermal gradient to the surface greater than 55 K km^{-1}. For the crust below that depth to conduct sufficient heat to sustain that gradient implies temperatures in the middle and lower crust that exceed any reasonable melting temperature. Whether or not the lower crust under the Hercynian Pyrenees was substantially molten, we know enough of the physical properties of continents to know that this is not a normal condition for the lower crust.

A substantial amount of heat must be supplied by some geologically acceptable means to the base of the 'high-level', 12 km deep melting zone. The fact that the maximum temperatures attained in the overlying section decrease rapidly upwards suggests either that the heating was a relatively rapid, transient event, or that the temperatures in the upper crust were depressed by convective circulation of surface fluids. We attempt to quantify these ideas below; but first we consider to what extent the fluid movements documented by the evidence of stable isotopes may have influenced crustal temperatures.

(*b*) *Convection*

Perhaps the most important geochemical feature of the Trois Seigneurs area is the demonstration that surface fluids have penetrated to depths of at least 12 km within the continental crust (Wickham & Taylor 1985). It is therefore important to consider whether they have exercised a major influence on the thermal history of the area.

We argued earlier that the deep circulation existed *before* the thermal maximum and that the total mass of seawater that flushed through the sedimentary pile was probably less than 50% of the rock mass (see figure 4). As emphasized above, this estimate is subject to some uncertainty because neither the efficiency of the isotopic exchange between rock and water nor the precise isotopic composition of the water is well known.

Before the main metamorphic event, and in the absence of circulation, the temperature at 12 km could have been in the range 300–400 °C, depending in part on how much crustal stretching had taken place (Lachenbruch *et al.* 1985). If surface waters were able to circulate to this depth, their overall effect would be to lower the mean crustal temperature by convecting heat to the surface. Whether this effect was significant at any particular depth would depend on the total mass of water circulating to that depth and on its velocity. At present the problem is too poorly constrained for quantitative analysis, but an upper limit to the cooling effect of the surface water is provided by assuming that water at surface temperature was instantaneously introduced into twice its mass of rock at some higher temperature and allowed to equilibrate. In this case the rock temperature might be lowered by about 35%.

Any of the circulating water that was present in the sedimentary pile at the onset of heating would presumably be expelled rapidly upwards as metamorphism progressed. The migrating water would advect heat and perturb the overlying thermal regime. Again, the question is whether this effect is likely to be significant. Using the same kind of argument as before, we can take 10% by mass as an unrealistically high upper limit to the amount of introduced water likely to be present at any time in the lower part of the sedimentary pile. Instantaneous expulsion of this amount of water into a similar mass of cool overlying rock would raise the temperature of the cool rock by a maximum $0.1\,\Delta T\,C_{pw}/C_{pr}$, where ΔT is the temperature difference between the cool rock and the hot water, and C_{pw} and C_{pr} are the heat capacities of water and rock, respectively. Even for this extreme amount of water the effect is not likely to be greater than a few tens of degrees.

The rather limited deep circulation for which the isotopic studies provide evidence in the higher grade and anatectic parts of the sedimentary pile, probably represents only the deepest and slowest part of a circulation system that affected the whole of the upper crust. Fluid movement at shallower levels could have been much more vigorous (see, for example, Norton & Taylor 1979). If this was so, the thermal régime of the upper crust would reflect the scale, geometry, and other features of the flow. In particular, temperatures would be depressed in regions of descending flow and elevated where the flow ascended.

It is clear, however, that fluid circulation did not long outlast the metamorphic peak in the upper crust. The discordant granodiorites that were emplaced shortly after the metamorphic maximum (see above) have oxygen isotope ratios that were not disturbed by circulating fluids.

Because of these uncertainties in the influence of fluid flow on thermal structure we must accept, for the present, that we do not know the temperature at 12 km at the onset of the main heating event. It could have been anywhere in the range 200–400 °C (corresponding to mean thermal gradients of 15–30 K km^{-1}) and for this reason, in the analysis given below, we take pre-intrusion gradient as a free parameter.

(*c*) *The observations*

We take as the observations that need to be satisfied the array of temperature–pressure points derived from the metamorphic isograds, ranging from about 450 °C at 9.5 km to about 700 °C at 12 km (see, for example, Wickham 1986*a*, figure 11). As discussed above, considerable uncertainty is associated with these values; in particular the slope of the array may have been steepened by post-metamorphic deformational effects, although these are not believed to have been large (see above and Wickham 1986*a*). To allow for this possibility, we also obtain solutions for a grossly expanded *P–T* array ranging from 700 °C at 18 km to 450 °C at 12 km.

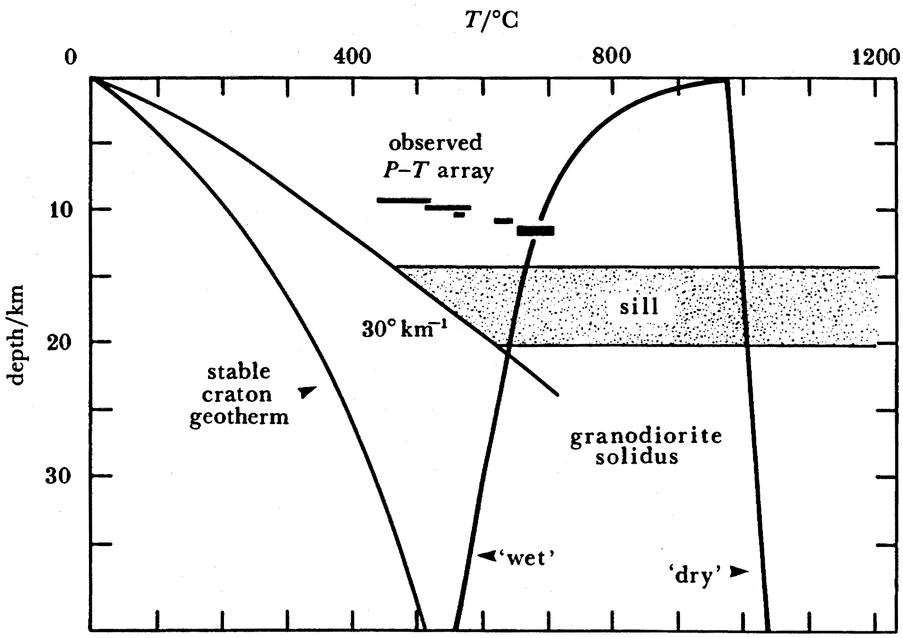

FIGURE 8. Various isograd constraints on the thermal structure of the Trois Seigneurs metamorphic sequence (after Wickham 1984, 1986*a*). The observed array of *P–T* points are shown as solid blocks, with the dimensions of the blocks denoting the approximate error in the location of each point (see Wickham 1986*a*, figure 11, for details). For reference, a 30 K km^{-1} geotherm, a stable carton geotherm and granodiorite solidi, wet and dry, are included. Also shown is the depth of a basaltic sill that could give the observed *P–T* array by its transient cooling.

In so far as a series of points of this kind reflects maximum temperature mineral assemblages, it is most unlikely that the *P–T* array defines a thermal gradient that ever existed in the crust. Each point simply records the maximum temperature attained at each depth and contains no information as to whether they were attained simultaneously, as would have to be the case if the array defined a geotherm (England & Richardson 1977; Oxburgh & England 1980).

Assuming no post-metamorphic deformation, the array of points shown in figure 8 is therefore a gradient of maximum temperature (∇T_{max}) and must represent the enveloping tangent curve to a family of cooling curves such as shown in figure 9. Note that the array of points in figure 8 is concave towards the temperature axis when extrapolated to surface temperatures and has a similar form to the enveloping tangent curve of figure 9.

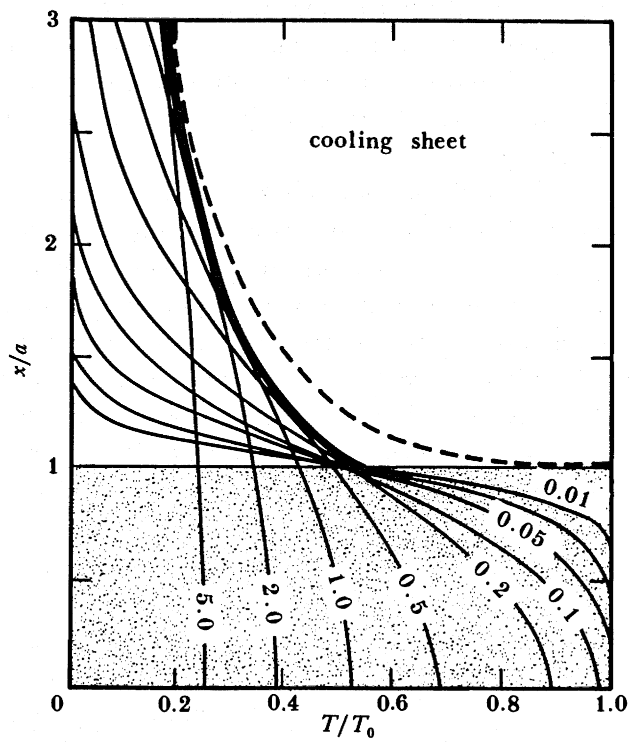

FIGURE 9. The relation between isotherms and the T_{max} array around a cooling sheet. The horizontal axis (T/T_0) shows temperature, T, as a fraction of the intrusion temperature, T_0. The vertical axis shows distance away from the centre of the sheet (thickness $2a$). Each of the lighter curves is for a different value of $\tau = \kappa t/a^2$ (i.e. time after intrusion; see text for discussion). The heavy enveloping curve is the locus of T_{max}, the maximum temperature experienced at a particular distance away from the sheet. The broken curve shows the T_{max} curve for a sill which was convecting during the earlier part of its cooling history (from $T/T_0 = 1$ to $T/T_0 = 0.8$).

(*d*) *The model*

In attempting to interpret the thermal evolution of the Trois Seigneurs Massif and other Hercynian terrains in the Pyrenees, the most important constraint is the very high interpreted value of ∇T_{max} in the approximate depth range 9–12 km. Even so, there can be no unique inversion of these observations and we therefore present families of solutions; choices between them must be based on their relative geological plausibility.

We consider, as the simplest case, the temperature field above an instantaneously emplaced hot sheet (i.e. a sill) that is assumed to be of infinite lateral extent. This is a problem that has been examined by Jaeger (1964, 1968), whose approach we follow here. The sill is of thickness $2a$. The country rock is taken to be at a uniform initial temperature, T_c, lower than that of the magma, T_m. All temperatures (T) are expressed in terms of $T_0 = T_m - T_c$. Distances (x) are measured vertically upwards from the centre plane of the sill. Then

$$\frac{T}{T_0} = \frac{1}{2}\left(\operatorname{erf}\frac{\xi+1}{2\tau^{\frac{1}{2}}} - \operatorname{erf}\frac{\xi-1}{2\tau^{\frac{1}{2}}}\right) \tag{1}$$

where $\xi = x/a$ and $\tau = \kappa t/a^2$. Thus, ξ is dimensionless distance and τ is dimensionless time (t).

The maximum temperature at any value of ξ occurs when $\tau = \tau_m = \xi/\ln[(\xi+1)/(\xi-1)]$.

This value of τ may then be substituted into (1) yielding T_{max}/T_0 as a function of ξ. In this simple model, a plot of $T/T_0 f(\xi)$ corresponds to the observed ∇T_{max} (i.e. the tangent curve of figure 9).

As this analysis applies only to an intrusion emplaced in country rock of uniform temperature, allowance must be made for the effects of the thermal gradient, β. The effect on temperatures would be additive (Jaeger 1984) and we therefore subtract a quantity, βd (where d is depth measured from the surface) from each observed temperature when comparing the observed ∇T_{Max} with the theoretical curve.

T_0 is the initial temperature of the sheet; for basaltic magma a temperature of about 1200 °C is appropriate. The theory predicts that at the instant of intrusion the temperature at the contact is $T_0/2$, and T_{max} decreases in the country rock further away from the contact in the manner shown by the heavy tangent curve in figure 9. In practice this estimate is too low, and leads to low estimates of temperature outside the sill because no account is taken either of latent heat of crystallization or of possible internal convection within the sill. Jaeger (1964) discusses various approximations that may be used to meet these problems; the effect of latent heat may be taken into account, by taking a higher value for T_0 and designating it T_0^*. For internal convection he presents a numerical result for a sill that remains well stirred until crystallization is complete. In figure 9 we also plot a curve approximated from Jaeger's numerical results (broken line). Subject to the numerous simplifications that have been made, selection of the 'correct' sets of parameters (i.e. those matching the natural situation) should cause the observed ∇T_{max} to plot on the tangent curve or between it and the convective case curve in figure 9.

In so far as we have no independent constraint on either sill thickness ($2a$), the depth to the top of the sill from the surface ($x-a$), or effective intrusion temperature (T_0^*), we present in figure 10 various families of curves for different sill depths (figure 10 *a–d*). For each sill depth we show, for different sill thicknesses, the combinations of T_0^* and β that will give ∇T_{max} curves close to that observed in figure 8.

In constructing figure 10 it was necessary to decide which combinations of parameters provided acceptable matches to ∇T_{max}. It was required that the upper temperature, 450 °C at 9.5 km, fell within 5 °C of the conductive cooling curve, and that the lower temperature, 700 °C at 12 km, fell either on the same curve or between it and the curve for a cooling sill with internal convection (see below and figure 9). The heavy line in figure 10 *a–d* is the locus of points that lie on the 'convective cooling' curve. Points to the left of that curve are not acceptable and, for any particular sill thickness, points further to the right are closer to the case of pure conductive cooling of a sill (without internal convection).

The approximation of the convective cooling curve from the numerical results given in Jaeger (1964, table 2) is crude, and amounts to constructing a curve for T/T_0 that exceeds the conductive values by about 20% close to the igneous contact ($\xi = 1.2$), falling to 3% at $\xi = 2.8$.

In the interests of brevity we shall not continue to repeat cautions about the crudity of this analysis. At best it can give only a semi-quantitative indication of the main physical constraints on the processes involved. These approximations are least reliable close to the igneous contact, at low ξ values (Jaeger 1964).

Note the negative slopes of the lines for each sill thickness in figure 10. For a particular intrusion thickness, the lower the pre-intrusion thermal gradient, the higher must be the intrusion temperature. For any particular intrusion depth, similar thermal effects are produced by thick and thin sills if the latter are either hotter, introduced under a higher thermal gradient,

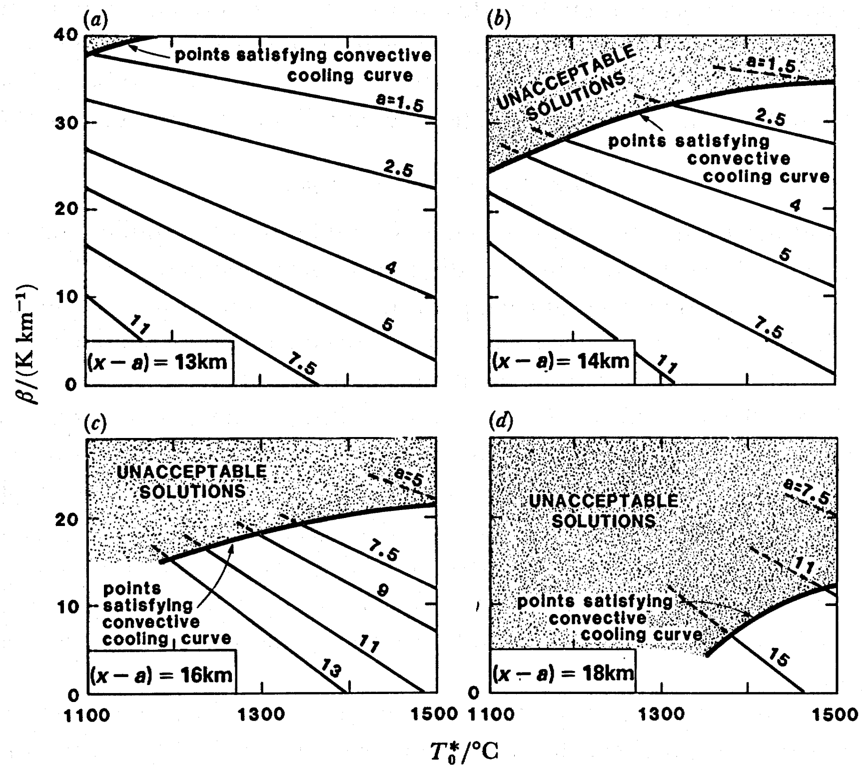

FIGURE 10. Families of curves showing 'acceptable' solutions (see text for definition) capable of generating the observed P–T array in the Trois Seigneurs Massif for different depths of sill emplacement, at (*a*) 13 km, (*b*) 14 km, (*c*) 16 km and (*d*) 18 km. For each depth of emplacement, the diagrams show that many solutions are possible for different combinations of pre-intrusion thermal gradient, β, effective temperature of intrusion, T_0^*, and sill half-thickness, a (see text for discussion). The heavy line in each diagram is the locus of points that lie on the convective cooling curve in figure 9, and any points to the left of this line (stippled field) are not reasonable solutions.

or both. For similar intrusion temperatures, and pre-intrusion thermal gradients, thin shallow sills may have the same thermal effect as deeper thicker bodies.

Figure 10 also suggests that, for a sill emplaced at 13 km, 1 km below the metamorphic sequence, there is a very wide range of acceptable combinations of sill thickness, initial temperature and thermal gradient. For sills emplaced deeper, the range becomes more restricted. At 16 km the possibilities are limited to thick, higher-temperature sills emplaced under medium to low thermal gradients, and for 18 km it is doubtful whether any of the limited possibilities are geologically reasonable. For depths greater than 20 km the initial temperatures required are very high and not geologically acceptable.

An emplacement depth of 13 km is regarded as unlikely on the grounds that more evidence of the magma should be seen (e.g. within the deeper parts of the Trois Seigneurs metamorphic sequence). However, the model suggests that a body emplaced in the 14–16 km depth range would satisfy the thermal constraints, and this would be more acceptable on geological grounds (see below). If it were basaltic in composition it could be relatively thin. If it were granodioritic, and therefore had a lower intrusion temperature, it would need to be thicker and would occupy much of the lower crustal thickness.

Figure 10 solutions and the preceding discussion relate to the *observed* P–T array exposed today in Trois Seigneurs Massif. Although we do not believe that post-metamorphic attenuation of the sequence has been large (see above), in the next section we discuss solutions obtained for an expanded P–T array, which assumes that the observed sequence has been shortened by more than 50% (probably a gross overestimate), giving a limiting minimum for ∇T_{max}.

(e) *Effects of simplifying assumptions*

The solution that has been used for the cooling of an infinite sheet applies only to the case where that sheet is emplaced in an infinite medium at uniform temperature. In fact we have a medium that is bounded by a constant-temperature surface, the surface of the Earth. This effect could readily be included (Jaeger 1964), but has been omitted for simplicity because the influence of the surface is unimportant at the depths of interest here.

We have made no allowance for the latent heat of melting in the anatectic zone. The effect of melting will be to buffer temperatures within the melting interval so that the observed values of T_{max} will be lower than expected in the absence of melting. In principle, this problem could be met by substituting, for the observed T_{max} values, T^*_{max} where $T^*_{max} = T_{max} + L/C_p$; a suitable value for L would be chosen on the basis of the degree of melting at each depth; T^*_{max} would thus decrease upwards and converge with T_{max} at the top of the melting zone. Similarly, in the metamorphic sequence, account should be taken of the heats of reaction of the new mineral parageneses.

It has been assumed that the thermal diffusivity of all parts of the system is the same; this is clearly not the case, but provided that our main concern is with ∇T_{max} and not precisely *when* T_{max} was attained at any particular point, diffusivity variation is not important. As pointed out above, the plot of T/T_0 against ξ does not depend on τ, the quantity that contains the thermal diffusivity.

Potentially the most serious source of error is the uncertainty associated with ∇T_{max}. If post-metamorphic shortening across the metamorphic sequence at Trois Seigneurs was significant, ∇T_{max} may have been rather lower than is observed today. To allow for this possibility we have made a similar analysis to that shown in figure 10 for an expanded P–T array ranging from 700 °C at 18 km depth to 450 °C at 12 km, corresponding to a much lower ∇T_{max}. A very wide range of parameter combinations can satisfy such an array, and these are illustrated in figure 11. The general features of the solutions are similar to those shown in figure 10. They involve sills of varying thickness and temperature, but emplaced at much deeper levels in the crust; some are equivalent to underplating the crust with a layer of mafic magma that is several kilometres thick. The thinner sills require very high pre-intrusion thermal gradients.

(f) *Discussion*

The hypothetical pressure–temperature–time (PTt) trajectories that would be followed by reference volumes of rock at different depths in the Trois Seigneurs sedimentary pile during Hercynian metamorphism are shown in figure 12. Much of the path is speculative because neither the rate of burial of the Palaeozoic units nor their rate of return to the surface, nor the precise time of intrusion of the mafic material relative to stretching, is known. There is nevertheless a marked contrast to metamorphic processes in collision zones, where a substantial thickness of crust may be heated and where metamorphism seems to be terminated through uplift and erosion. In the present situation metamorphism seems to occur over a limited time interval, particularly at the higher grades, and it seems likely to be terminated by conductive

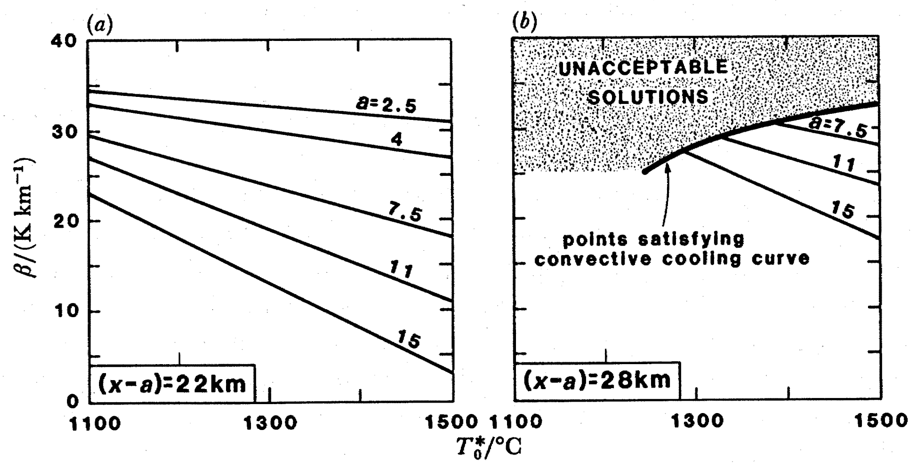

FIGURE 11. Two sets of curves showing 'acceptable' solutions that would generate an expanded ∇T_{max} (P–T array) from 700 °C at 18 km to 450 °C at 12 km (to allow for the possible effects of extreme post-metamorphic attenuation of the metamorphic sequence). Solutions are shown for sills emplaced at (*a*) 22 km and (*b*) 28 km and, as in figure 10, indicate the various combinations of β, T_0^* and a that are possible. The emplacement depths in this case are so deep that they could correspond to underplating the crust with several kilometres of basaltic magma, rather than intruding this material at higher structural levels within the crust.

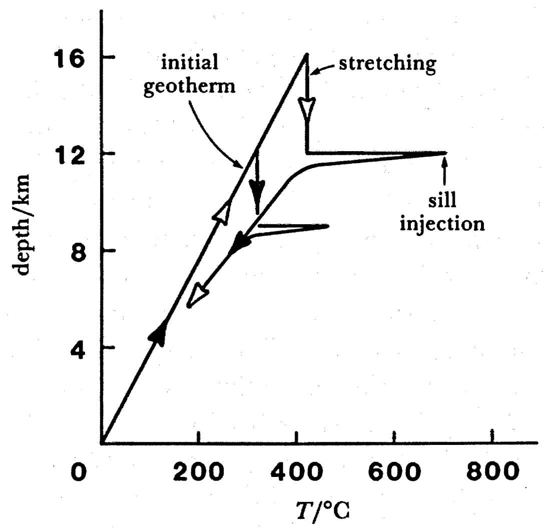

FIGURE 12. Qualitative pressure–temperature–time trajectories for points at the top (solid arrows) and bottom (open arrows) of the metamorphic interval analysed in the text, assuming that stretching entirely preceded intrusion. The amount of stretching is not constrained, nor are the times at which many parts of the paths are reached. See text for discussion.

cooling at constant pressure. This difference in cooling behaviour should be reflected in contrasts between the patterns of cooling ages of igneous and metamorphic rocks in collision zones, and those in low-pressure metamorphic terrains.

We have argued elsewhere that the most plausible setting for the Hercynian metamorphism in the Pyrenees is a rifting strike-slip zone (Wickham & Oxburgh 1985, 1986). We have extended the earlier observation that surface-derived seawater had circulated down to depths of at least 12 km in the crust, to show that this circulation seems not to have entered the old pre-Palaeozoic basement upon which the lower Palaeozoic sedimentation probably took place (Wickham & Taylor 1986).

The high temperatures and high gradients of maximum temperature reached in the 9.5–12.0 km depth interval provide a constraint on the range of thermal models that can be entertained. For example, the observed *P–T* array is satisfied by the emplacement of a basaltic sill, 6–8 km thick at a depth of about 14 km, although many other solutions are possible.

We have modelled the heat source as a sheet, and assumed that the present-day outcrop areas of the high-grade metamorphic rocks simply represent the elevated parts of a regionally developed zone of metamorphism and anatexis. If the various Hercynian metamorphic terrains exposed in the Pyrenees were in fact separate features each with its own heat source, our analysis would not be very much affected. A substantial lenticular body of magma would be required beneath each region, or alternatively a laterally continuous, thin, sheet-like body that thickened under each one.

Our inference about the regional tectonic setting leads us to suppose that the main intrusive thermal event was preceded by crustal stretching (Wickham & Oxburgh 1985, 1986). We have no independent evidence for this and are therefore unable to estimate the likely crustal temperature perturbation resulting from the stretching, but the pre-intrusion thermal gradient was probably not greater than 35 K km^{-1}.

An intrusive body of the composition and size suggested would also have an important effect on temperatures in the underlying crust. There would be a more prolonged elevation of temperature that could lead to very large-scale partial melting of the lower crust (depending on water content and composition). This may be the origin of the very large, 'late' granodiorite plutons, described above, and thought on the basis of geochemical data to be derived from the lower crust (see also Wickham & Oxburgh 1986; Wickham 1986*a*).

An objection to our proposed model must be the relative rarity of mafic rocks in the metamorphic–anatectic zone. It is necessary to postulate that the mafic magma was largely contained within the pre-Hercynian basement. There is, however, limited evidence of mafic intrusion in the deepest part of the anatectic zone, in the form of mafic xenoliths in the structurally deepest part of the biotite granite at Trois Seigneurs (Wickham 1984, 1986*a*, *b*) and mafic pods and sills, which occur commonly within the 'basal gneisses' at Agly, Castillon and St Barthelemy (see, for example, Wickham & Taylor 1986). A kilometre-sized mafic and ultramafic complex is associated with the high-grade rocks exposed in the Gavarnie region (Soula *et al.* 1986) and layered mafic rocks have also been described at Castillon (Roux 1977) and at Saleix (Vielzeuf 1984). Although these rocks have not been dated, they may represent part of the mafic magma responsible for the metamorphic effects.

It is interesting to compare our thermal model with others for similar areas. Based on studies of the Canigou and Agly Massifs, and other parts of the Eastern Pyrenees, Guitard (1970) proposed that the Hercynian high thermal gradient metamorphism was the result of the 'effet du socle'; essentially that local elevation of the basement ('basal gneisses') advected the isotherms closer to the surface (implicitly by erosion of the overburden), thereby generating transient, high and localized, near-surface thermal gradients. This model now appears to be quantitatively inadequate to explain the more tightly constrained *P–T* relations in the metamorphic sequences.

We have previously suggested (Wickham & Oxburgh 1985, 1986) that analogies may exist between the tectonic setting of the Hercynian Pyrenees and the present-day Gulf of California–

Salton Sea–Imperial Valley system. A detailed analysis of the thermal and mechanical aspects of this system has been presented by Lachenbruch *et al.* (1985). It is clear that the situation they describe is related to that which we deduce for the Pyrenees, but that there are important differences both in the nature of the evidence available to constrain the processes in the two areas, and in the detailed character of the processes themselves. In the Pyrenees, the maximum temperatures reached in the upper 12 km of the crust are well constrained, but the contemporary crustal structure and kinematics are not. In the Salton Sea–Imperial Valley (SSIV) there is good control of the present-day kinematics, crustal structure, and near-surface heat flow. In the Pyrenees there may have been a much smaller amount of crustal extension, and it is very unlikely that the crust reached any kind of steady thermal state such as seems reasonable in the SSIV, where continuous extension has gone on for at least 5 Ma. The areas have in common that their thermal characteristics require the addition of substantial volumes of magma to the lower crust, or to its base. Some of the thermal profiles proposed for SSIV are similar to those we deduce for the Pyrenees, but in the former case they depend on the contemporary deposition of cold sediment relatively close to the zone of magmatic heating, in a way that cannot have occurred in the Pyrenees, where the zones of deposition and heating were separated by at least 14 km of crust. The two areas provide complementary views of different modes of crustal extension.

During the final revision of this paper, Dr Scott Baldridge of the Los Alamos National Laboratory kindly brought to our attention a paper on the Rio Grande Rift (Baldridge *et al.* 1984). The model proposed in that paper for the Rio Grande Rift on the basis of geophysical and geological observations is strikingly similar to that which we have proposed here.

6. Conclusion

The Hercynian metamorphic terrains exposed in the Pyrenees record a thermal event in which both high temperatures and very high thermal gradients were developed in the upper crust. While it is likely that the metamorphism occurred within thinned continental crust, the Hercynian high thermal gradients are unlikely to represent a steady thermal state, and were probably generated by massive advective heat transfer from the mantle into the crust, in the form of mafic intrusions. A mafic sill several kilometres thick, emplaced into the middle or lower rust, could produce a P–T array similar to that observed in the Trois Seigneurs Massif by purely conductive cooling.

Available geochronological and field data suggest that there may be a short delay between the metamorphic peak in the 'high level' anatectic zone and the intrusion of the lower crust-derived 'late' granodiorite plutons (see above; Wickham & Oxburgh 1986 and references therein; Zwart 1979), although some workers have suggested that the two were roughly contemporaneous (Soula 1982; Soula *et al.* 1986). The thermal model that we prefer offers a possible explanation for this apparent time-lag. A mafic intrusion emplaced into the middle crust would generate a rapidly moving, short-lived thermal pulse in the region above it, while the region below would warm more slowly, but ultimately reach higher temperatures, and for a longer time. In this way, large volumes of magma could be generated by anatexis of the lower crust (even if this were relatively dry) but rather later than the thermal maximum experienced in the upper crust above the mafic intrusion. Thus the highly simplified model presented here appears capable of explaining quantitatively the thermal structure of the 'high level' metamorphic sequences, and qualitatively the chronology of emplacement of the various crustally derived granitoid magmas generated during Hercynian metamorphism in the Pyrenees.

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Discussion

H. G. Reading (*Department of Earth Sciences, Parks Road, Oxford, U.K.*). To those of us working on sedimentation and tectonics in ancient terrains, and on the Hercynian, it is critical to distinguish between a strike-slip tectonic pattern and an extensional one, or, in a transtensile régime, the degree to which the strike-slip or extension predominated. The essence of a strike-slip régime is that the rifted pull-apart basins are relatively small, sink with great rapidity to depths of up to 10 km or more, and are never far away from compressional, uplifted blocks that were being deformed simultaneously with basin sinking. Crustal levels may be moved vertically up to as much as 15 km over short distances (10–20 km) during periods of perhaps 10 Ma or less (i.e. at rates of 100 cm per 1000 years or more). Because of the proximity of localized extensional and compressional regimes the basins may be uplifted quite soon after their formation, as exemplified by the large number of exhumed ophiolites in strike-slip zones.

In contrast, true extensional rift basins (e.g. East African rift, Rheingraben, Central Graben, North Sea) are generally larger features, subside more slowly (about 10 cm per 1000 years), do not have associated compressional uplifted blocks, and the vertical contrast of crustal levels is less. They are likely to be followed by a long period of slow thermal subsidence (1–4 cm per 1000 years). Inversion, if it takes place, does so some time after basin formation.

Can Dr Wickham and Professor Oxburgh say how far their evidence, and perhaps structural evidence elsewhere in the Pyrenees, is more compatible with a regional strike-slip and hence pull-apart basin model, or with a regional rift basin model? Can they also expand on how they feel the metamorphic styles of pull-apart basins may differ from those of extensional rift basins?

S. M. Wickham. Dr Reading's question raises a number of very interesting and important topics. They are, however, ones on which, at this stage, we are able to do little more than speculate. So far, no attempt has been made to interpret the Upper Palaeozoic sedimentary record of the Pyrenees in the light of the tectonic setting that we believe is required by the Hercynian metamorphism. It will be important to do so.

In principle it should be possible to distinguish the regional metamorphism that develops in a purely extensional tectonic regime from that of a strike-slip–extensional setting; in the latter case, it is possible that the highest grades of metamorphism would be reached only in the pull-apart basins, giving a regional pattern of metamorphic 'hot-spots'. It would be tempting to interpret the distribution of high-grade Pyrenean massifs in this way, but at present this cannot be done with any confidence.

R. L. M. Vissers (*Institute of Earth Sciences, Utrecht, The Netherlands*). An extensional setting for the Variscan orogeny in the Pyrenees, as proposed here by Dr Wickham and a recently advocated by Wickham & Oxburgh (1985) to explain the high geothermal gradients involved in Pyrenean type high-temperature–low-pressure regional metamorphism, seems inconsistent with the structures observed in the axial zone of the Pyrenees. Structural mapping by the Dutch group of Zwart and co-workers during the 1960s has demonstrated that in the low-grade rocks in particular, tight folds with steep axial-plane slaty cleavages characterize large parts of the axial zone (Zwart 1979). These steep structures affect the entire Palaeozoic sequence including the Lower and Middle Carboniferous. It is only in relatively small areas of higher metamorphic grade that flat-lying structures are observed with a subhorizontal component of extension. Recent structural analysis by Verhoef *et al.* (1984) in the western part of the Aston massif has shown that these flat-lying structures are broadly synmetamorphic and that they are superimposed on the steep structures observed in low-grade areas.

The question arises how the steep tight folds, which demonstrably predate the peak of the Variscan metamorphism and consistently indicate something of the order of 50% or more horizontal shortening in sections as large as half the width of the axial zone, can be understood in terms of overall crustal extension.

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S. M. Wickham. We would make two points in reply to Dr Vissers. First, we are unable to comment on the relative importance of extensional and strike-slip movements in the evolution of the Hercynian Pyrenees (see our reply to Dr Reading). Whereas it is never a straightforward matter to attempt to relate the geometry of structures developed on one scale to a much larger-scale kinematic régime, it is possible that the steep axial-surface features to which he refers developed in response to strike-slip movements. It is also possible that they formed during an early phase of regional compressive deformation on to which rifting was later, locally superimposed in certain areas, generating flat-lying structures associated with the high-grade metamorphic rocks (see §2 of this paper).

Secondly, we would draw Dr Vissers's attention to an observation made by Professor Fyfe a number of years ago, when he pointed out that in the energy budget of an orogenic belt as a whole, the amount of energy required for the process of regional metamorphism was at least two orders of magnitude greater than that required to achieve even the most intense deformation. We do not wish to labour the point, but the first and most important feature to explain in the Hercynian Pyrenees is their metamorphism, and that we believe we have done.