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Normal faulting and tectonic inversion driven by gravity in a thrusting regime

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Abstract—Normal fault occurrences in high plateaux reflect an extensional regime which develops in mountain belts where convergence is still active. However, normal faults are also created in active belts, within the compressive edges which theoretically only undergo thrust faulting. The meaning of such normal faults is studied for a structure in a French external subalpine massif where the still active process of thickening mainly occurs by thrust faulting. The heterogeneous deformation which involves basement slices and sedimentary cover, results in progressive tilting of weak interfaces. In this structure gravity sliding processes induce normal faulting which, in this case, is compatible with a thrust faulting regime.

INTRODUCTION

IN MOUNTAIN ranges where convergence is still active, high plateaux often display normal faulting (see Molnar & Lyon-Caen 1988 for a review). However, normal faults are also known to exist in the compressive edges of mountain belts, for example in the southern French Alps.

Models for high plateaux normal faults have been proposed (Molnar & Lyon-Caen 1988), the main points of which are summarized below. The aim of this paper is to examine and understand normal faults which appear in a foreland thrust belt setting. The field example studied is located in the Vercors dauphinois subalpine massif. This massif is in the still thickening edge of the French External Alps. A build-up sequence inferred from a balanced cross-section through the Vercors massif is presented. It suggests that the inhomogeneous structure of the foreland thrust belt, where deformation induces progressive tilting of weak interfaces, may induce gravity sliding processes. These may account for the occurrence of normal faults in a thrusting regime. It is also suggested that thrust faults created by compression may be reactivated in an extensional sense, that gravitational sliding may amplify slip on thrust faults, and that some shallow thrusts may have no roots.

THE MEANING OF NORMAL FAULTS IN MOUNTAIN BELTS

The structure of collision belts generally reveals two types of normal faults: those which are related to an extension phase pre-dating collision, and those which form while the collision belt builds up. The former correspond to tilted blocks of passive margins and are often reworked as reverse faults during the collision which is the classical model of tectonic inversion (Gillchrist *et al.* 1987). The latter, which form within still active collision belts, have been explained in terms of the physical model of Molnar & Lyon-Caen (1988). These authors stated that "when the forces driving the plates together can no longer supply the potential energy needed to increase a belt in height, this is likely to grow laterally, generating high plateaux". The model referred to represents a simple mountain range isostatically compensated and in mechanical equilibrium (Fig. 1). This equilibrium implies that the mean horizontal force on a vertical cross-section normal to the strike is constant across the range (Dalmayrac & Molnar 1981), and therefore the mean horizontal stress (quotient of this constant force by the thickness of the crust) is greater under the lowlands than under the plateaux: $\sigma_{xx(\mathrm{L})} > \sigma_{xx(\mathrm{P})}$. Furthermore, the vertical stress σ_{zz}, equal to the lithostatic pressure, depends on the horizontal position, so that, at a given depth z, σ_{zz} is greater under the plateaux than under the lowlands: $\sigma_{zz(\mathrm{P})} > \sigma_{zz(\mathrm{L})}$. Since one cannot *a priori* know the ratio of vertical to horizontal stresses across a mountain range, three cases can be considered. In the first case (Fig. 1a), the vertical and horizontal stresses are identical under the plateaux. It follows that $\sigma_{xx(\mathrm{L})}$ is greater than $\sigma_{zz(\mathrm{L})}$ and that normal faulting is likely to develop in the wedge of the range. In the second case (Fig. 1b), the vertical and horizontal stresses are similar under the lowlands; thus $\sigma_{zz(\mathrm{P})}$ is greater than $\sigma_{xx(\mathrm{P})}$, and normal faulting has to be expected under high elevation areas. The third case (Fig. 1c) is an intermediate one where σ_{xx} is smaller than σ_{zz} under the plateaux but exceeds the vertical stress in the lowlands. This suggests that normal faulting in the plateaux can occur simultaneously with thrust faulting in the wedge of the range. These different cases show that in the accretionary wedge of a mountain range, between the plateaux and the lowlands, one must expect mainly reverse faults to occur.

However, in some mountain range wedges such as the southern French Alps, normal faults are clearly observed in the Embrunais–Ubaye nappes (Debelmas 1953, Kerckove 1969) and the Digne nappe (Labaume *et al.* 1989, Ritz 1991), while others in the Alpes Maritimes

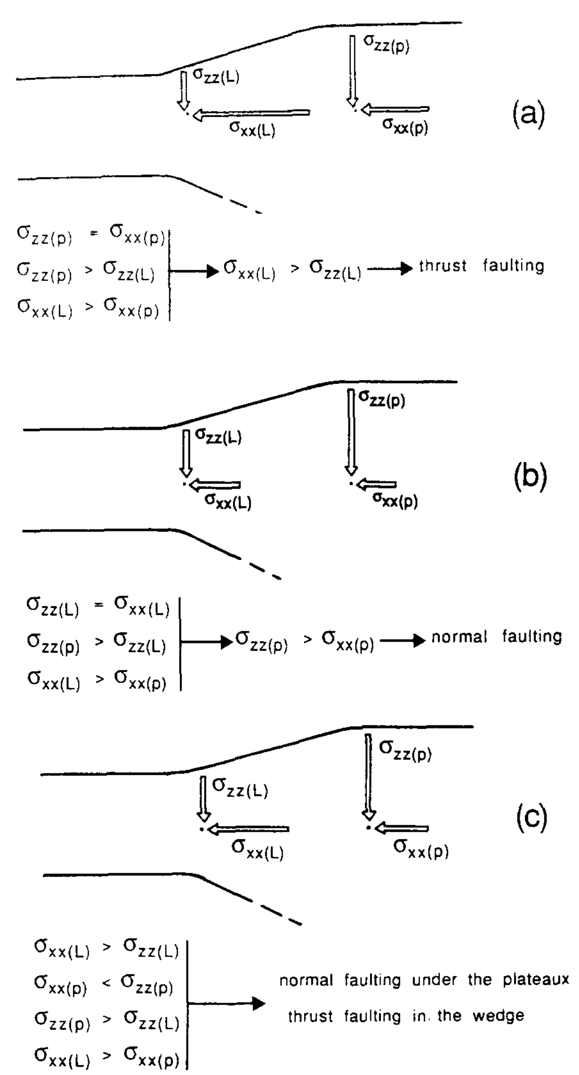

Fig. 1. Faulting regimes in the plateaux and the wedge of a mountain belt, after the model of Molnar & Lyon-Caen (1988). Whatever the type of stress distribution (a, b or c), normal faulting cannot, theoretically occur in the accretionary wedge of a mountain belt. $\sigma_{xx(L)}$, horizontal stress at depth z under the lowlands; $\sigma_{xx(P)}$, horizontal stress at depth z under the plateaux; $\sigma_{zz(L)}$, vertical stress at depth z under the lowlands; $\sigma_{zz(P)}$, vertical stress at depth z under the plateaux.

nappes (Graham 1981) may only be interpreted as such. The aim of this paper is to understand the meaning of such structures in areas undergoing shortening, from a field example in the French External Alps, on the eastern border of the Vercors subalpine massif.

THE VERCORS MASSIF AS A PART OF THE ACTIVE ALPINE WEDGE

The Vercors massif is part of the western-most French subalpine chain and its place in the French external Alps is shown in Fig. 2. This massif is the first elevated area above the low elevation unfolded foreland basin filled with molassic Miocene deposits. Its western limit also corresponds to the deformation front to the east of which the complete Mesozoic cover is involved in a compressive deformation through folds and thrusts. The edge of the Vercors massif increases in height from 200 to 1000 m over a few kilometres horizontal distance. The topographic envelope is slowly but regularly rising to the east with a 3° slope changing to 2°: plateau des Coulmes (1000 m), Cornafion (2000 m), Grand Armet (2900 m) and La Muzelle (3700 m). In this external zone of the belt comparisons of levelling survey data allow a calculation of the recent uplift rates. Thus, the uplift amount of the French External crystalline massifs between 1886 and 1969 ranges from 1 to 1.5 mm year^{-1} (Fourniguet 1977, Ménard 1988), as it is sketched by the arrow on Fig. 2 for the Belledonne massif, and is distributed over a 70 km bulge. This recent uplift is also observed at the range scale in the Swiss Alps (1.3 mm year^{-1} between 1870 and 1919, Schaer & Jeanrichard 1974, Rybach *et al.* 1977, Gubler *et al.* 1981) and Austrian Alps (1 mm year^{-1} between 1906 and 1971, in Levallois 1973).

A glacio-isostatic origin to these uplifts (after melting of the Alpine glaciers, 12,000 BP) is rejected since equilibrium has probably already been restored for 5000 years (Schaer & Jeanrichard 1974). A thermal origin is also rejected (Ménard 1988) since zones being uplifted

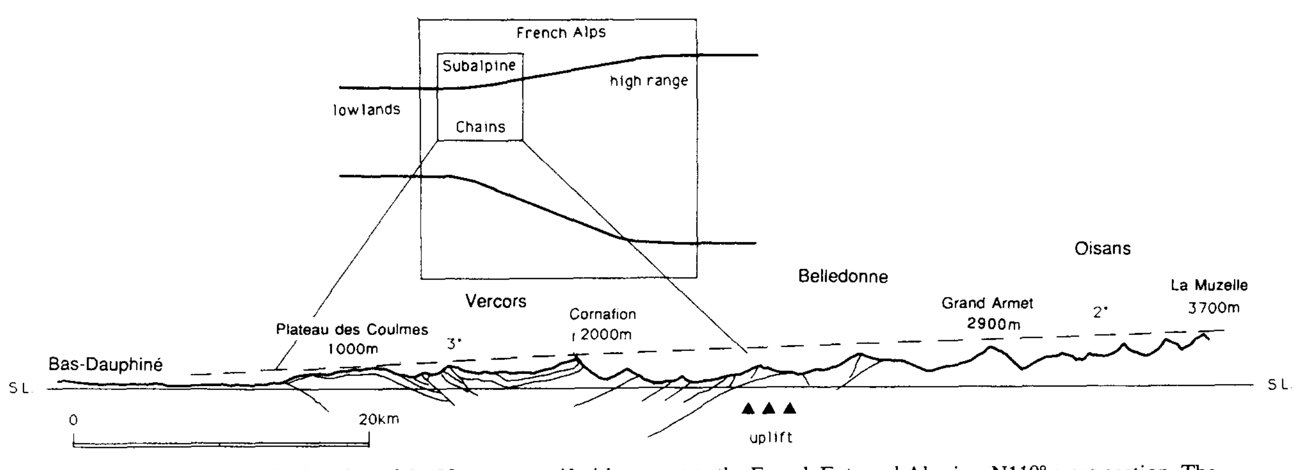

Fig. 2. Geodynamic situation of the Vercors massif with respect to the French External Alps in a N110° cross-section. The regularly W-dipping topographic envelope and the still active uplifting of the crystalline massifs (Belledonne) shows that the French External Alps belong to the accretionary wedge of the belt. As a part of this wedge, the Vercors massif is likely to undergo mainly thrust faulting.

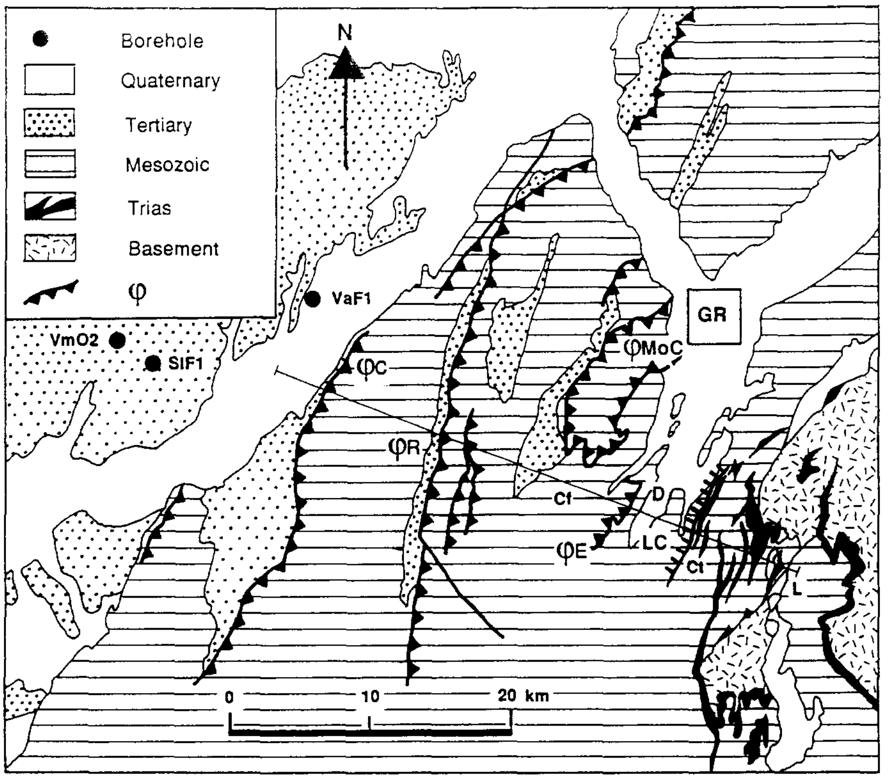

Fig. 3. Simplified geological map of the northern Vercors, showing the main structural features as well as the location of the cross-section of Fig. 5. L, Laffrey lakes; Ct, Conest mountain; LC, Les Combes normal fault; D, Drac river; ϕE, Eperimont thrust; Cf, Cornafion mountain; ϕMoC, Moucherotte–Cornafion thrust; ϕR, Rencurel thrust; ϕC, Cherennes thrust.

such as the French External crystalline massifs and Simplon area (Steck 1992), are narrow in comparison to the scale where the thermal processes are active in the range.

There is also evidence of past uplift in several areas of the Alps, according to the cooling ages of some minerals (muscovite, biotite, apatite) which date the time they go through an isothermal level. The data compiled by Ménard (1988) show the uplifts to occur as gently curved wide bulges which suggest continuous movements over large spans of time. Among the French External crystalline massifs, those of Mont Blanc and Belledonne have experienced a mean uplift rate of 0.7 mm year^{-1} since -14 Ma (Gasquet 1979, Demeulemester 1982, Mugnier & Ménard 1986). Taking into account a 33° km^{-1} normal geothermal gradient, the Belledonne massif was probably buried 9 km deep. The order of magnitude of the mean past uplift rate is similar to that of its recent uplift. This similarity suggests that there is a common tectonic origin to these uplifts (Schaer & Jeanrichard 1974, Ménard 1988) and, as a result, the hypothesis of thermal origin can be rejected.

The Vercors massif, at the western front of the Alpine area compressional deformation, has a W-dipping topographic envelope. This massif is the sedimentary cover lying on a crystalline massif which has been uplifted and continues to uplift at present. These features indicate that this massif is part of an external wedge as in the mountain range model of Molnar & Lyon-Caen (1988). As such, the only syntectonic deformation structures in this massif should be folds and reverse faults.

A NORMAL FAULT WITHIN THE EXTERNAL WEDGE IN VERCORS

The normal fault chosen as an example is located between the Vercors massif and the Laffrey lakes (labelled L on Fig. 3), to the east of Drac river, near the hamlet of Les Combes. The fault trace on the 1/50,000 geological map of Vif (BRGM 1967) has a N015° trend, parallel to the main Alpine structures in this area. It dips 75° W and puts Toarcian limestones in the hanging wall into contact with Triassic terrains, including evaporites, in the footwall. This fault constitutes the western boundary of a series of W-dipping slices of lower Liassic calcareous layers outcropping west of the Conest mountain and underlain by the Triassic evaporites cited above (Fig. 3). These evaporites thin to the east and are even completely lacking in the Laffrey area.

The aim of this work is to understand the reasons why such a normal fault formed in this accretionary environment. Thus, the role played by this fault in the progressive structural development of this region must be determined. The stages of the shortening process have to be restored by means of a well constrained balanced cross-section, taking into account the geological and geophysical data, and integrating the kinematics now widely accepted for this region.

Geological and geophysical data

The geological data cover the thickness variations in the sedimentary pile, the deep structures of the Jurassic rifting (Lemoine *et al.* 1981), the pre-Alpine structures (Arnaud 1973, Vialon *et al.* 1989), the basement depth (Orgeval & Rumeau 1955, Ménard 1980), the stratigraphic continuity between cover and basement in the eastern part of the cross-section (Barfety *et al.* 1970, Lory 1948) and the formations which are likely to behave as décollement surfaces (Gidon 1981, Ménard & Thouvenot 1984, 1987, Butler 1988).

Crustal-scale deformation framework

The history of the passive margin in the French Alps has been summarized in many published articles (Butler, 1987, 1989, Ménard & Thouvenot 1987, Ménard 1988, Coward & Dietrich 1989, de Graciansky *et al.* 1989, Guellec *et al.* 1990, Mugnier *et al.* 1990, Nicolas *et al.* 1990, Vialon 1990) which refer to such concepts as tilted blocks, inversion tectonics and thrust tectonics. The widely accepted framework is as follows. At the beginning of the Mesozoic, the pattern of the European passive margin was controlled by normal faults inherited from late Hercynian (or even older) structures. These faults formed the boundaries of various sized tilted blocks in the upper 5–10 km of the crust. When the Alpine deformation reached the external domain (during Oligo-Miocene times), the crustal shortening occurred by westward in-sequence basement involved thrusting, with the thrust geometries influenced by pre-existing normal faults.

In many published crustal balanced cross-sections, the Belledonne External crystalline massif is interpreted as the western-most basement thrust sheet. The westward movement of this massif is then considered to induce, in turn, the shortening of the cover by means of detachments along weak layers and the propagation of ramps. However, the way in which the basement thrust is drawn differs slightly from one author to the other. Although in previous works Butler (1987, fig. 19, 1989, figs. 12 and 15) drew the Belledonne basement massif as lying on a gently dipping footwall ramp merging at depth in the Moho detachment zone, in another paper (Butler 1988) he proposed that Alpine thrusts have a number of different detachment levels within cover, basement and lithosphere, and stated that the Alpine sole thrust beneath the Belledonne massif must have a staircase form with a mid-crustal detachment. A low velocity zone reported by Ménard & Thouvenot (1984, 1987) is fairly consistent with this detachment which may correspond to the lower boundary of basement titled blocks, and strongly supports the staircase geometry illustrated in the balanced crustal cross-section of Fig. 4.

Using recent data from geophysical profiles, the work of Arpin *et al.* (1988) brings new constraints with respect to the basement depth under the Vercors massif and enables the authors to balance several cross-sections in the subalpine chains. However, their northern Vercors cross-section does not take into account the Triassic and lower Liassic W-dipping slices of Conest mountain. Moreover, the uppermost Moucherotte thrusted unit (ϕMoC on Fig. 3) is assumed to root onto the back of the Belledonne basement block; however, until now, no trace of this thrust fault could be found in the field, in the area between Moucherotte and Conest mountain. Furthermore, the cross-section does not include the unusual normal fault which has to be explained in this shortening regime. In this region, however, a normal fault is also referred to by Butler to explain the westward-thickening of the Liassic succession from 200 m in Laffrey to 1500–2000 m in the Drac valley. In Butler's interpretation (1989, fig. 12) this fault is thought to be a pre-existing down-to-the-west fault bounding the external Belledonne basement block which was acting as a high while the Vercors region was an eastward-deepening low. However, in such a fault environment, particular deposition facies including slope breccias, mixed sedimentary and crystalline debris, olistolites (Barfety & Gidon 1984), should be found. But there is no field evidence pointing to the presence of such sediments in the Triassic and lower Liassic slices of Conest mountain. Furthermore, Butler's normal fault (1989, fig. 11) does not intersect the present topographic surface, contrary to what the geological map of Vif (BRGM 1967) states in Les Combes hamlet, on the western edge of Conest mountain. Therefore, following Arpin *et al.* (1988), it seems reasonable to suggest that the thickness variations in the Liassic sediments may be linked to an eastward-directed synsedimentary extensional fault. In this interpretation, the Belledonne block is a downthrown-to-the-east tilted block. The pelagic Liassic shales deposited in the deeper part of the half-graben steadily decrease in thickness towards the east and turn into the Laffrey condensed limestone sequence, on top of the tilted block (Fig. 5a, from D to L).

KINEMATICS PROPOSED FROM THE BALANCED CROSS-SECTION

Classical methods are used to balance this section (Fig. 5). The thickness of competent layers is assumed to be constant as well as the areas of the thrust units (plane

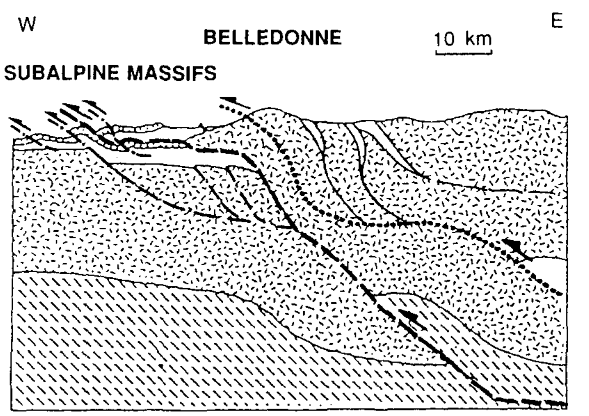

Fig. 4. Balanced cross-section from Ménard & Thouvenot (1987) showing the Miocene–Present Belledonne basement thrust (thick dashed line) as a result of tectonic inversion of tilted basement blocks. It is the westernmost crustal thrust fault involved in the process of thickening of the wedge in the external French Alps; it induces shortening in the subalpine massifs by thrust faulting.

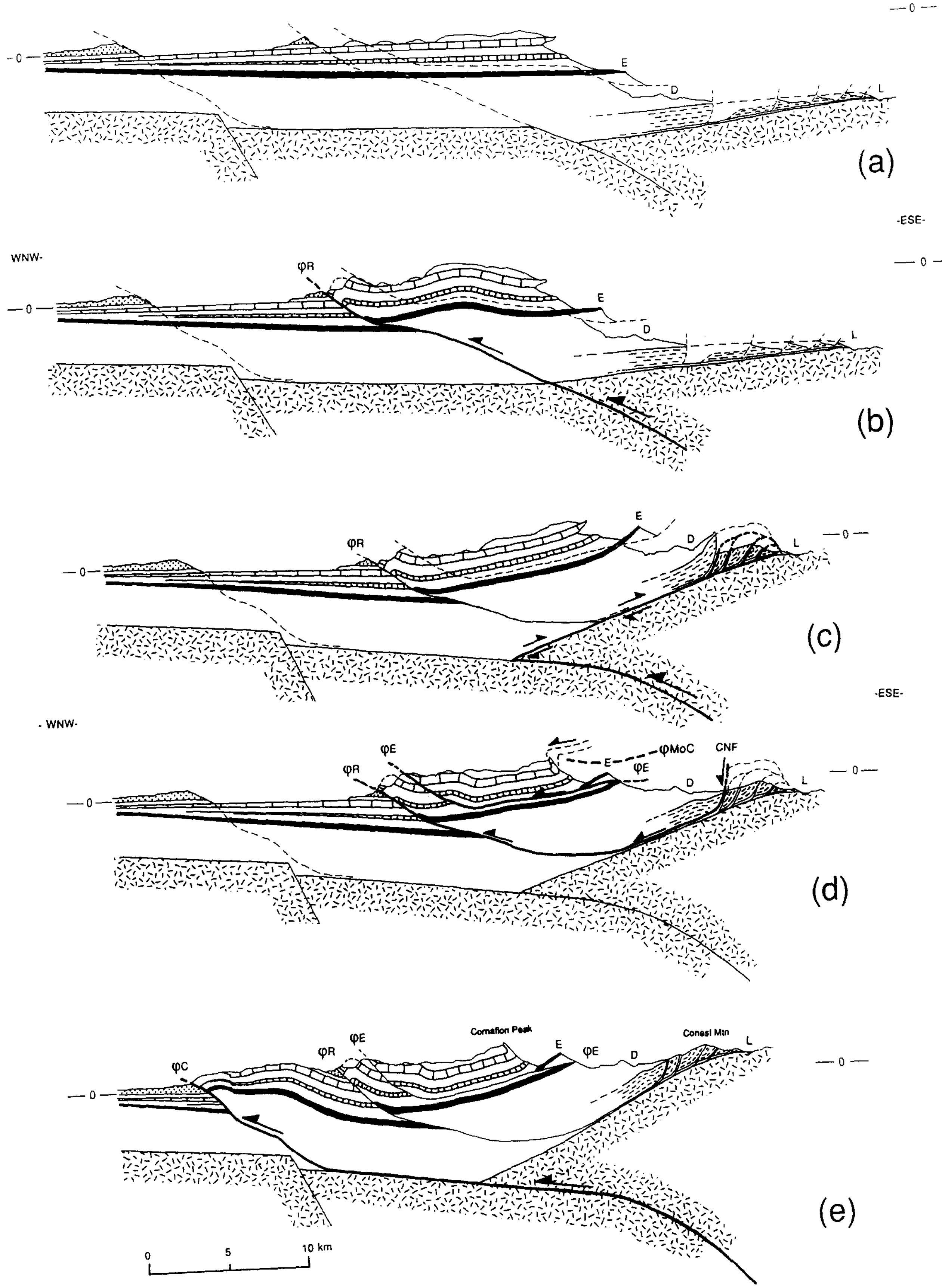

Fig. 5. Speculative restoration and balancing of a N110° cross-section through northern Vercors. Stippled, Miocene molasse; large bricks, mid-Cretaceous Urgonian limestones; small bricks, lower Cretaceous Valanginian limestones; black, upper Jurassic Tithonian limestones; dashed, lower and upper Lias; randomly dashed, crystalline basement. (a) Initial state. (b) Beginning of shortening; the ϕR propagates. (c) Underthrusting of the basement and appearance of the Conest Mountain backthrusts. (d) Gravitational gliding stage; the westernmost Conest backthrust is inverted as a normal fault and propagated upward as LC normal fault; ϕR is reactivated; creation of ϕE and ϕMoC. (e) Final stage; rear push of the basement creates ϕC.

strain). The competent Urgonian (mid-Cretaceous) and Tithonian (upper Jurassic) limestone formations are selected as marker layers for shortening measurements. However, in the eastern part of the section, where these formations have been removed by erosion, the upper limit of lower Lias ('calcareous Lias') is the only reliable marker.

In this westernmost part of the Alpine domain, the timing of deformation is not very well constrained. It is generally agreed to start about −14 Ma (middle Miocene) when the Belledonne massif starts uplifting, and to be continuing at present. As the maps show (BRGM 1967, 1975, 1978) for the Moucherotte, Rencurel and Cherennes thrusts, the Mesozoic sediments are thrusted onto Miocene molasse, thus roughly dating the thrusting events.

In the kinematic model proposed hereafter, the aim is not to state the absolute age of any event but give the relative chronology of the thrusts determined from a strict balancing of the cross-section. This includes the geological and geophysical data discussed earlier. Although it is slightly different from the sequential models described in the previous works by Gidon (1981), Arpin *et al.* (1988) and Butler (1988), it does not contradict these models and, in fact, emphasizes the significance of the backthrust structures between the Drac and Laffrey.

In Miocene times in the French external Alpine domain, when shortening began to lift up the Belledonne crystalline block to the west (Fig. 5a), the sedimentary rocks of the above half-graben were probably left relatively undeformed. Thus, it may be reasonable to assume that before the upper surface of the Belledonne block reached the level of the first basement step to the west, the overlying cover remained fixed on the basement and was likely to be passively uplifted. Consequently, the displacement relative to the sedimentary pile of the neighbouring western domain may then have been accommodated by the appearance of the Rencurel thrust fault (ϕR), with a 1 km offset (Fig. 5b). At this stage, the upper surface of the Belledonne block was probably still slightly dipping to the west.

As soon as the toe of the Belledonne block passed over the edge of the neighbouring basement step, it was thrust under the sedimentary pile (Fig. 5c) which, at this stage, was not displaced with respect to the foreland. This underthrusting was probably quite easy if the Triassic layers in this area included the same type of evaporites as those found in the Conest mountain, east of the section. However, to accommodate the shortening of the basement, the cover above ϕR must also have shortened. But, as mentioned earlier, in the Laffrey region (labelled L, east of the section) the cover has remained fixed on the crystalline rocks, due to the lack of evaporites. Therefore, the deformation had to occur elsewhere and it was thus achieved through a series of backthrusts (Conest mountain), involving 5–7 km of shortening as suggested by the speculative restoration of the Conest mountain Liassic sliced structure (Fig. 6). At this stage, since the wedge-shaped toe of the Belledonne sheet had just passed over the footwall ramp, the dip of its upper surface increased to 25–30° to the west.

As a result, this deformation sequence created a continuous fault surface, from the westernmost backthrust of the Conest mountain to the leading line of ϕR (Fig. 5c). This curved surface looks like a 'slip circle' at the bottom of a landslide. The weak mechanical properties of this pre-cut surface with respect to the unstrained rocks, coupled with its shape and the strong dip of the basement–cover interface, allowed the huge overriding mass to slide under the force of gravity. This slip caused the lower part of the westernmost Conest backthrust to be reworked as a listric normal fault and to propagate upward, thus creating the steeply dipping Combes nor-

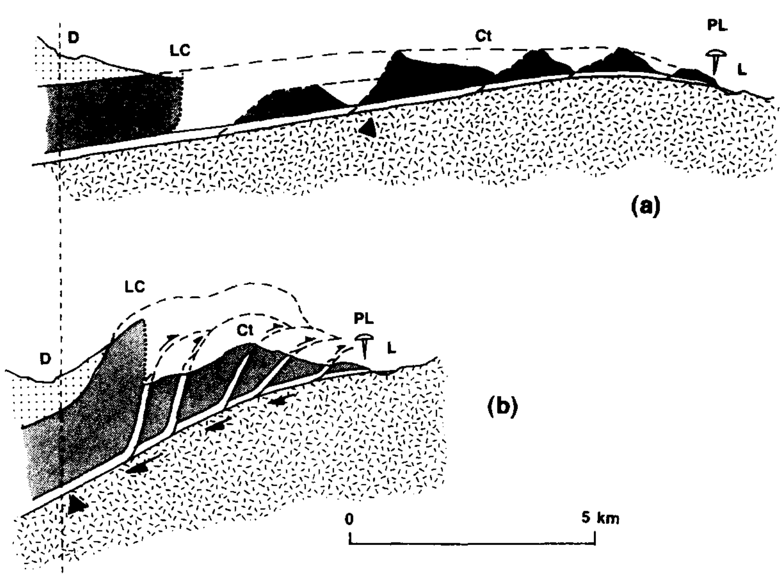

Fig. 6. Detailed speculative restoration and balancing of the eastern part of cross-section on Fig. 5. (a) Initial stage. (b) Underthrusting of the basement. Same symbol letters as on Fig. 5. Black triangle, marker point linked to the basement; randomly dashed, crystalline basement, white, Triassic rocks, shaded, Liassic limestones, stippled, Liassic shales; PL, pin line near Laffrey where the cover remains fixed to the basement.

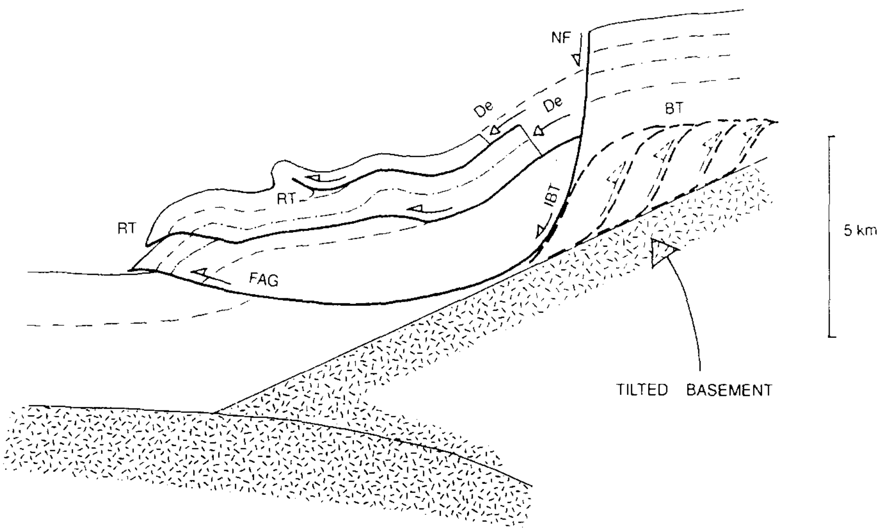

Fig. 7. Sketch of the gravitational readjustments induced in the cover by a basement thrust sheet tilting the structures which were created during the previous stages of shortening. BT, backthrust now inactive; IBT, gravity-inverted backthrust; NF, normal fault created by upward propagation of the inverted backthrust; FAG, forethrust whose slip is amplified by gravity gliding; DE, local denudation resulting in upper rootless thrusts (RT).

mal fault which has a 1 km vertical offset. This phenomenon constitutes a particular case of tectonic inversion.

On the Rencurel thrust, the 1 km increase in westward slip driven by gravity does not correspond, of course, to any further shortening of the basement.

During this deformation step, gravity may also have induced two other faults: the Epérimont thrust (ϕE) and the Moucherotte–Cornafion thrust (ϕMoC). In this interpretation, the latter appears on the section (Figs. 5d & e) as a fault propagation fold, into which the Moucherotte thrust (Fig. 3) dies out to the south (Donzeau *et al.* 1992). This explains why no field evidence of a plausible root structure has ever been found, either for the ϕMoC thrust, as mentioned before, or for the ϕE thrust. These faults are therefore assumed to have formed as décollement surfaces within the Neocomian (mid Cretaceous) argillaceous formations, thus providing local denudation processes as sketched in Fig. 7.

In the last stage (Fig. 5e), the Belledonne crustal thrusting was still active and moved westward until the basement slice reached its present position. In so doing, it created a new thrust in the cover, the Cherennes thrust (ϕC); the décollement propagated ahead at the bottom of the stratigraphic pile, before the frontal ramp was initiated on a basement normal fault scarp (the 'faille de l'Isère' fault in Arnaud 1973). The amount of slip on the ϕC is 2–3 km.

Discussion

One may wonder, however, why this particular sequence of events is the preferred solution for balancing this cross-section, and especially why it has been assumed that the Belledonne block underthrusts the cover (thus generating the Conest backthrusts), just after the ϕR formed and before the ϕC occurred.

If it had been assumed that the underthrusting of Belledonne basement and the formation of the Conest backthrusts had occurred before ϕR, ϕR would have to be explained only by gravitational gliding on a slip circle, quite similar to that defined earlier. But this would have implied a 2 km normal displacement on the Combes normal fault, and this is not consistent with the actual stratigraphic offset on this fault.

If the underthrusting of the Belledonne block, the backthrusts and the Combes normal fault had occurred after ϕR and ϕC, the final stage structure would not have been very different from stage (e). However, just before the underthrusting took place, the cover in the central part of the section would have shown a bulge with its upper surface gently dipping to the east. Since such a geometry is very unlikely for the surface of a tapering wedge, this sequence was not selected.

Furthermore, the kinematics proposed from the balanced cross-section can be supported by some seismic data.

Seismic data

Only a few seismic events have been recorded in this area since 1959: the seismic activity is low and very few seismic stations existed in the field until recent years. Nevertheless, three events which occurred along the studied cross-section are worth considering.

In view of the difficulties in obtaining good data (inadequate distribution of stations, flat layer velocity model), the vertical location is given to within about 5 km. Despite this great inaccuracy, the nodal planes of the focal solutions may correspond to the major thrust faults in the above balanced cross-section (Fig. 8). The low dip reverse slip of the 25 April 1962 earthquake (**M** = 4.5) could be indicative of either a reworking of the ϕR under gravity, or an incremental westward slip on the sole thrust of the cover. Such a displacement at the basement cover interface can be linked to the 22 November 1979 earthquake (**M** = 4.2) whose reverse

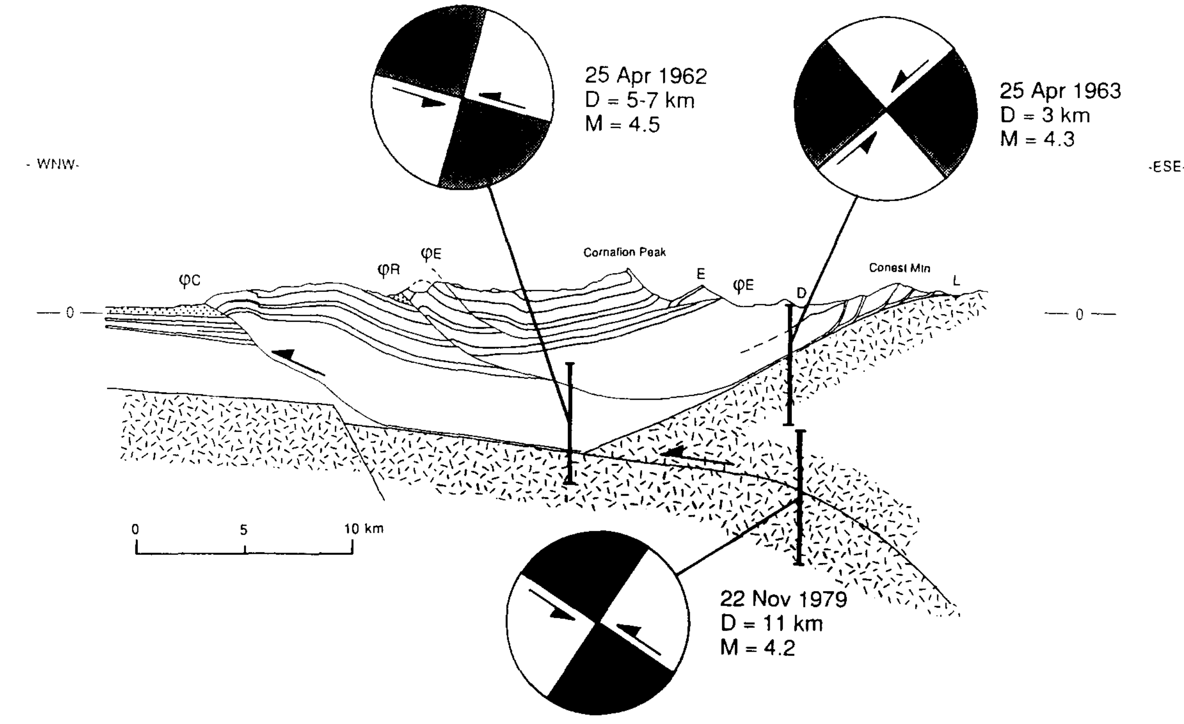

Fig. 8. The kinematics inferred from the balanced cross-section of Fig. 5 may be supported by the seismic data recorded in this area since 1962. Despite the great inaccuracy of the focal solutions, the nodal planes may correspond to some of the major faults described above; in particular, the 25 April 1963 earthquake could account for a gliding event on the W-dipping basement–cover interface.

slip may have occurred on the Belledonne crustal thrust fault. The 25 April 1963 event (**M** = 4.3) took place some 7 km above that of 1979. The normal fault slip inferred from its focal solution is *a priori* incompatible with the crustal shortening deduced from the 1979 event. However, slip on the W-dipping nodal plane may correspond to gravity gliding on the tilted basement–cover interface.

CONCLUSION

An analysis of the development of the Miocene to Present faults in a region of the French External Alps shows that, in the foreland thrust belt of a mountain range, where the crust is still thickening by active forethrust faulting, a gravitational instability may develop.

In this wedge, where the theoretical models predict only thrust faulting, gravitational readjustments may induce normal faults.

These readjustments are induced by the progressive increase of the slope of the wedge as well as that of the tilted weak layers (precut fore or backthrust surfaces generated by crustal shortening, weak sedimentary strata such as evaporites and marls). These processes may lead to a new type of inversion in which tilted backthrusts are reactivated as normal faults. They may also generate rootless forethrusts by denudation of the upper part of the tilted cover.

Such kinematic interpretations can help in studying collision tectonic areas where a part of the seismic activity involves normal focal solutions.

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