

MESOZOIC BASIN DEVELOPMENT AND STRUCTURAL
INVERSION IN THE EXTERNAL FRENCH ALPS

By

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À Kate.



Frontispiece: The French Alps in winter.

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Abstract

This study considers the structure and stratigraphy of the Pelvoux and adjacent crystalline massifs, and their sedimentary cover in the Dauphinois zone of the external French Alps. On a larger scale it is concerned with the geodynamic evolution of the European-Tethyan margin, and aspects of the Western Alpine thrust structure in the Vercors and Devoluy Subalpine Chains.

During the Mesozoic, SE France evolved as a rifted continental margin to the Ligurian Tethys. Structural and stratigraphic studies disclose Triassic to Jurassic age normal faults affecting pre-rift basement in the study area. Two different ages are recognised: 1) late Triassic to early Liassic faults rotated to gentle SE-dips during extension, and 2) Upper Liassic to Mid-Jurassic steeply SE-dipping faults which locally breach structures of the earlier system. The steeper faults bound the larger half-graben basins. A subsidence analysis across the margin reveals a differential subsidence pattern with a pronounced negative subsidence anomaly, the Briançonnais uplift. A model of heterogeneous lithospheric stretching, consistent with normal fault geometries documented from the Pelvoux, is proposed to explain this. Thermal subsidence of the margin occurred with oceanic spreading in the Ligurian Tethys.

Contraction and inversion of the European-Tethyan margin commenced with Tertiary Alpine compression. Early normal faults strongly influenced the geometry and structural style of later Alpine thrusts and the study area displays a full range of inversion geometries which are described. Gently dipping faults were reactivated as thrusts and a decoupling of the syn- and post-rift sequence occurred. Buttressing of Alpine thrust displacements by the steeper normal faults resulted in tight upright folds and complex back-fold-thrust systems in the larger half-grabens. Overall, the geometries encountered do not comply with the thin-skinned thrust geometries documented from Rocky Mountain style thrust belts.

The timing of the Alpine deformation is complex and

both pre- and post-Nummulitique events are constrained by syn-tectonic unconformities in the southern and eastern Pelvoux. In the central Pelvoux, post-Nummulitique break-back thrusts slice through earlier fold and thrust structures of uncertain age.

Fold and thrust trends in the study area parallel pre-existing faults and a right-angle bend in the Western Alpine thrust belt may reflect the passive margin geometry. Pre-existing faults also influenced the kinematics of Alpine thrusting. In the Pelvoux tectonic transport directions vary, as they do around the Alpine arc, and post-Nummulitique structures show a progressive anti-clockwise rotation of transport direction from WNW to SW. Such variations are attributed to the interaction of passive margin fault geometry, indentation and rotation, reflecting Oligocene to Recent plate motions in the Mediterranean.

The ECORS-CROP deep seismic reflection profile and this study, show that the reactivation of pre-existing passive margin faults and the formation of new thrusts, both constitute important mechanisms of crustal shortening and thickening in the Western Alps.

Résumé

Cette étude porte sur les structures qui affectent, dans la zone dauphinoise des Alpes externes, le massif cristallin du Pelvoux et sa couverture sédimentaire. A plus large échelle, l'évolution géodynamique de la marge Européenne de la Téthys, et certains aspects des structures chevauchantes du Vercors et du Devoluy sont discutés.

Durant le Mésozoïque, l'évolution du SE de la France est celle d'une marge passive associée à la Téthys Ligure. Dans la région du Pelvoux, les études structurales et stratigraphiques mettent en évidence, du Trias au Jurassique, le jeu de failles normales qui affectent également le socle anté-triassique. Deux étapes sont reconnues par le présent travail: 1) durant l'extension se produisant du Trias au Lias inférieur, les accidents sont bascules supérieur au Jurassique moyen, se développent des accidents plus raides qui recoupent localement le système précédent. Dans la région étudiée, les failles les plus raides, sont de grand accidents pentés vers le SE, et qui limitent des demi-grabens.

L'analyse de la subsidence à travers la marge révèle un schéma de subsidence contrasté, avec une forte anomalie négative liée à la surrection du Briançonnais. Un modèle d'étirement hétérogène de la lithosphère, compatible avec la géométrie des failles normales observées dans le Pelvoux, pourrait expliquer cette anomalie. La subsidence thermique de la marge se produit lors de l'expansion océanique de la Téthys Ligure.

La contraction et l'inversion de la marge européenne de la Téthys commence avec la compression alpine tertiaire. Les failles normales pré-existantes influencent alors fortement la géométrie des structures chevauchantes alpines: dans la région étudiée, nous avons pu ainsi décrire une grande variété de géométries d'inversion. Les failles faiblement pentées sont réactivées en chevauchement et se poursuivent dans la couverture par un découplage entre séries syn-rift et séries post-rift. Les déplacements induits à l'alpin viennent buter contre les failles normales les plus raides et sont absorbés dans les grands demi-

graben par des pli serrés ou des structures rétrochevauchantes. Au total, les géométries reconstruites ici s'éloignent du style structural décrit dans les grands chevauchements superficiels des montagnes rocheuses.

La séquence de la déformation alpine est complexe, et les événements pré et post-nummulitiques sont contraints par des inconformités syntectoniques dans le sud et l'est du Pelvoux. Dans le centre du Pelvoux, des accidents post-nummulitiques recoupent des structures chevauchantes antérieures mais d'âge incertain.

L'orientation des axes de plis et des chevauchements est parallèle, dans la région étudiée, à celle des failles pré-existantes. Aussi à l'échelle des Alpes occidentales, l'angle droit limitant au SE le massif du Pelvoux pourrait correspondre à la géométrie de la marge téthysienne. La cinématique des chevauchements alpins est également influencée par les accidents pré-existants. Ainsi dans le Pelvoux les structures post-nummulitiques montrent une rotation anti-horaire de la direction de transport, qui passe de l'WNW au SW. En fait ces variations peuvent être attribuées à l'interaction entre la géométrie de la marge passive et le poinçonnement lié depuis l'Oligocène aux mouvements des plaques en Méditerranée.

Cette étude, apparemment confirmée par les résultats du profil sismique ECORS-CROP, montre que la reactivation de failles associées à l'ancienne marge passive et la formation de nouveaux chevauchements interviennent l'une et l'autre comme mécanisme de l'épaississement et du raccourcissement crustal dans les Alpes.

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Chapter 1: Introduction

1.1 Regional setting and location

The external crystalline basement massifs, the main subject of this investigation, inscribe a major arc around the external part of the Western Alpine thrust belt (Figures 1.1 & 1.2). They are exposed in late structural culminations developed during Tertiary Alpine compression. These culminations comprise pre-rift crystalline basement and Triassic sediments, and Jurassic syn-rift sediments. External of the basement massifs are the Subalpine Chains comprising thrustured and folded Mesozoic to Tertiary cover sediments mainly of the post-rift phase, and farther W, the remnants of a Tertiary Alpine foreland basin sequence superposed on Eocene to Oligocene sediments of the Rhone-Bas Dauphine extensional graben. The principal structural elements of the Western Alpine thrust belt and its adjacent foreland area of SE France are shown in Figure 1.2.

The precise area of study is located to the SE of Grenoble, France (Figure 1.3), and straddles the Isere (38) and Haute Alpes (05) departments. A considerable part of the field area lies within the confines of the "Parc National des Ecrins", and road access is limited (Figure 1.3). What is commonly referred to as the Pelvoux massif in the geological literature is in fact the "Massif des Ecrins", named after its highest peak the "Barre des Ecrins" (4102 m). The map geometry and general disposition of the different crystalline basement massifs (Taillefer-Belledonne, Rochail, Grandes Rousses, and Pelvoux) in the field area, is shown in Figure 1.4. Overall the terrain is of a rugged and mountainous nature, providing excellent 3-D exposure of the structures concerned.

1.2 Previous work

The Western Alps comprise one of the longest and best studied mountain chains, a consequence being the accompaniment of a voluminous "mountain" of literature. Any

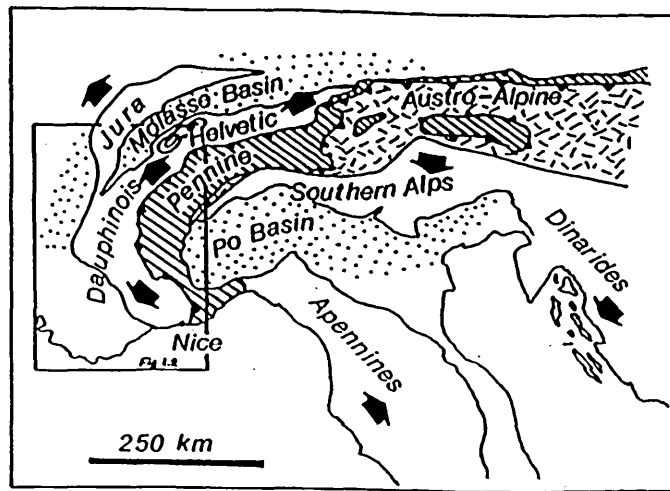


Figure 1.1: Principal tectonic and stratigraphic divisions of the Alps (modified after Trumphy, 1960; Debeltmas & Lemoine, 1970). The filled arrows indicate the direction of Alpine overthrusting, and the area indicated in Figure 1.2 is defined.

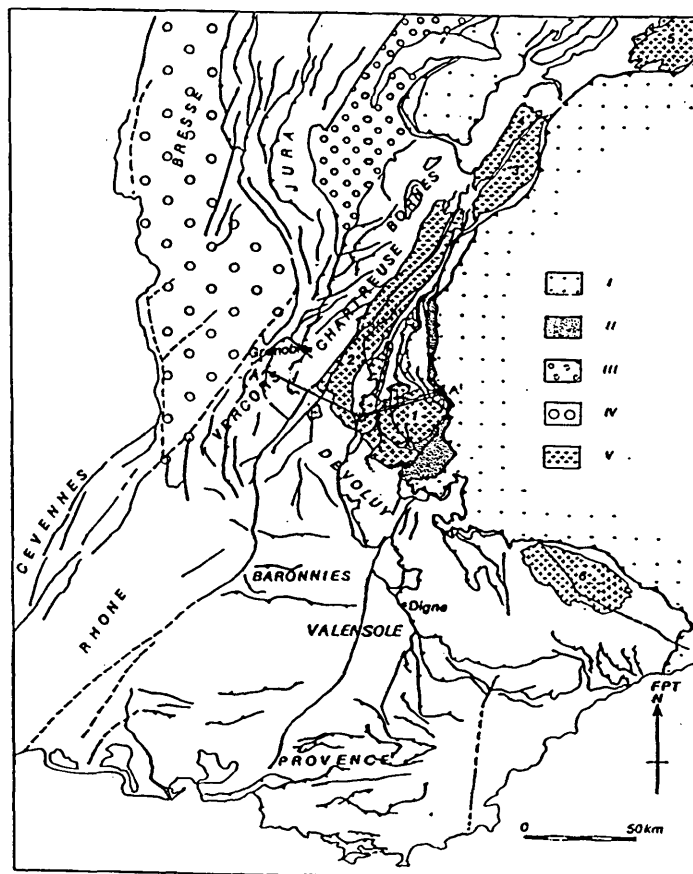


Figure 1.2: Simplified tectonic map of the Western Alps and SE France showing the main faults and the major tectonic units, I. Internal Penninic zones, above Frontal Pennine thrust (FPT), II. Eocene-Oligocene flysch rests unconformably on basement at the eastern and southern edge of the Pelvoux massif (1), III. external molasse basins, IV. Bresse and Rhone (Bas-Dauphine) graben fill, V. external basement massifs, 1. Pelvoux (this study), 2. Belledonne, 3. Mont Blanc, 4. Aiguilles Rouges, 5. Aar, 6. Argentera.

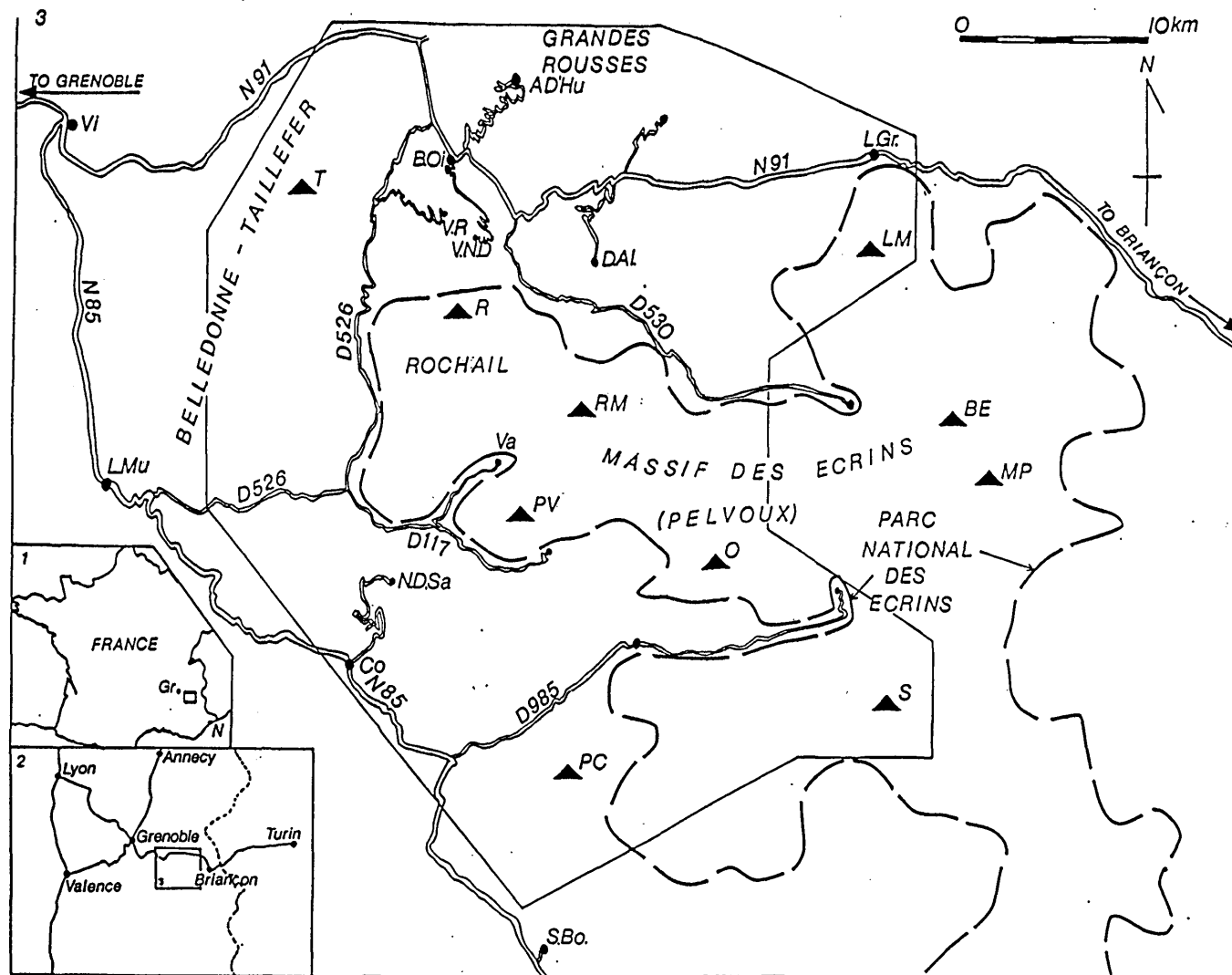


Figure 1.3: Location of the field area (3), to the SE of Grenoble (2), in SE France (1). On (3) the approximate limits of the mapped field area, the boundary of the Parc National des Ecrins, and the main roads are shown. The filled triangles represent the main mountains, T-Taillefer, R-Rochail, RM-Roche Muzelle, LM-La Meije, PV-Pic de Valsenestre, O-OLan, BE-Barre des Ecrins, MP-Mont Pelvoux, S-Sirac, PC-Petite Chaillol. Filled circles are villages and towns, Vi-Vizille, LMu-La Mure, Co-Corps, SBo-St Bonnet, NDSa-Notre Dame de la Salette, Va-Valsenestre, VR-Villard Reymond, VND-Villard Notre Dame, BOi-Bourg d'Oisans, DAL-Deux Alpes, A'DHu-Alpe d'Huez, LGr-La Grave.

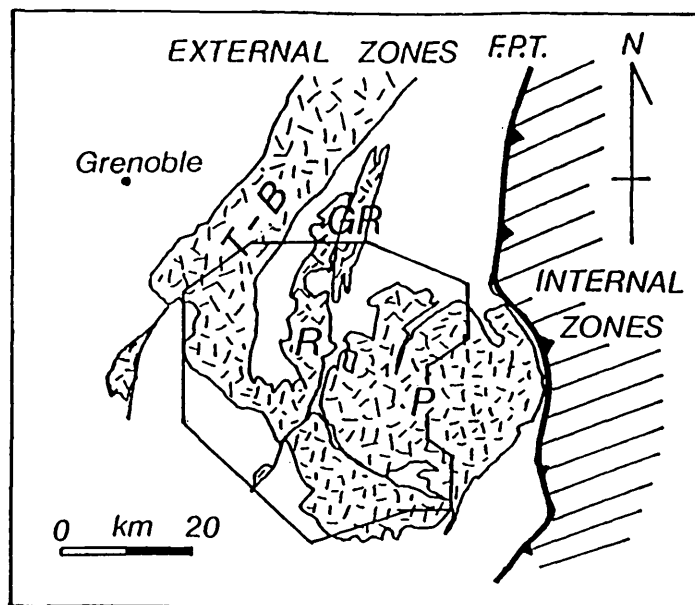


Figure 1.4: General location of the different basement massifs (speckled ornament) in the study area, TB-Taillefer-Belledonne, GR-Grandes Rousses, R-Rochail, P-Pelvoux.

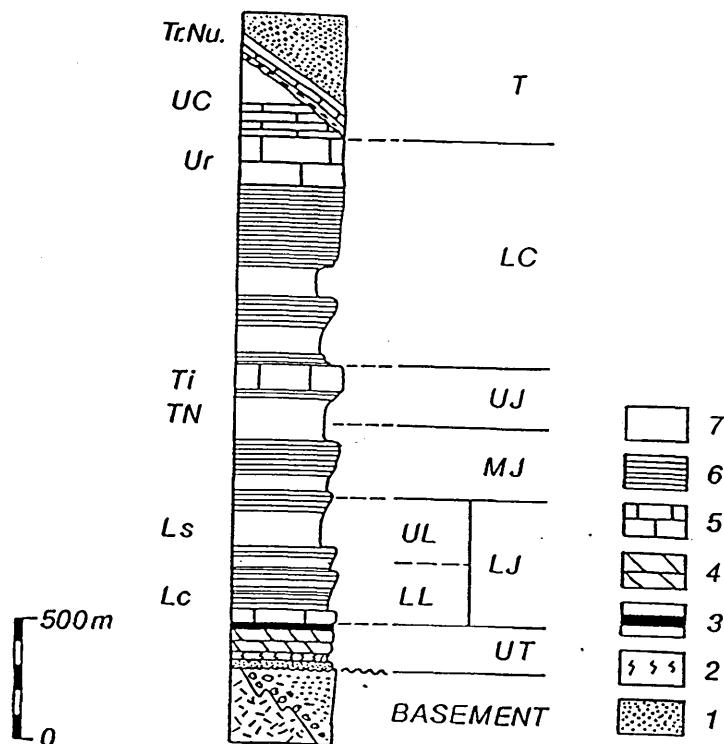


Figure 1.5: Generalized stratigraphy of the Dauphinois zone (external zones) (modified after Ramsay, 1963; Lemoine et al., 1986). Lithology types, 1. sandstones, 2. evaporites, 3. continental basalts, 4. platform dolomites, 5. limestones, 6. interbedded limestones-shales/marls, 7. shales/marls. Stratigraphy, UT-Upper Triassic, LJ-Lower Jurassic, LL-Lower Liassic, UL-Upper Liassic, MJ-Middle Jurassic, UJ-Upper Jurassic, LC-Lower Cretaceous, UC-Upper Cretaceous, T-Tertiary; local names, Lc-Lias calcaire, Ls-Lias schisteux, TN-Terres Noires, Ti-Tithonique, Ur-Urgonian, TrNu-Trilogie Nummulitique.

summary must, out of necessity, be selective.

The early Alpine works date back to the turn of the century and before, with the works of Termier (1903), Albert Heim (1921), Argand (1916), Collet (1927) and many others. It is a long established concept that the Alps formed by continental collision between Africa and Eurasia with the closure of an intervening oceanic area called Tethys (Argand, 1924). More recently Trumphy (1960) documented the necessary elements for continental collision: an overriding continent - the Austroalpine nappes, an intermediate ophiolitic melange located directly under the Austroalpine units, and the underthrust continent and its detached cover sequence in which are found some of the most complicated structures of the Western Alps (Hsu, in press).

More upto date, although now classical review works, are provided by Trumphy (1960; see also 1973; 1980), Goguel (1963), Ramsay (1963), Debelmas & Lemoine (1970), and Debelmas & Kerchove (1980). Trumphy (1960) and Debelmas & Lemoine (1970) proposed a basic tectono-stratigraphic division of the Western Alps into an external zone, separated from an internal "Penninic" zone, by the Frontal Pennine thrust (Figure 1.1). The external zone (where the external crystalline basement massifs are found) is subdivided into the Jura, Helvetic and Dauphinois domains which represent continental margin sediments deposited on both thinned and normal thickness European crust. The internal Penninic zone comprises the Valais domain of deeper water basinal sediments which are absent to the S in the Western Alps, the Sub-Brianconnais and Brianconnais domains characterized by a thick sequence of platform carbonates, and the Piemont domain with its now strongly metamorphosed "Schistes Lustrés" or "Bundnerschiefer", dated as Lower Cretaceous from microfossil discoveries (De Wever & Caby, 1981; Lemoine et al., 1984) and corresponding to deep water sediments including radiolarian cherts. These sediments are often intercalated with ophiolitic masses.

Numerous attempts have been made to relate the Alpine deformation to the relative displacement between the African and European plates (e.g. Smith, 1971; Dewey et al., 1973; Biju-Duval, 1977; Smith & Woodstock, 1982; Savostin et al., 1986) constrained by the history of N Atlantic spreading (i.e. see kinematic solution of Pitman & Talwani, 1972). No attempt will be made to discuss all of these models, however they can be grouped according to the number of independant plates involved and the vector motions of Apulia or Italy relative to those of Africa. Dewey et al. (1973) postulate the existence several independant micro-plates between the Eurasian and African continents. This is widely contested by many authors who invoke a simpler two-plate model in which the formerly independant micro-plate of Apulia (Italy) remains fixed to the African plate as a N African promontory (Dercourt, 1972; Channell et al., 1979; Bernoulli & Lemoine, 1981). Although there is general agreement that Tethyan extension from Triassic to Cretaceous reflects a sinistral displacement of Africa relative to Europe, while Cretaceous to Recent compression reflects a dextral displacement of Africa, Hsu (in press) points out several inconsistencies in the two-plate collisional models which assume the motions of Apulia to be the same as those of Africa. Seeing no need, or evidence for several independant micro-plates, Hsu (in press) proposes a three-plate refinement of the two-plate model in which Italy is the third and independant micro-plate. This third plate is referred to as the Mediterranean or M-plate, which is now segmented by the post-collision fragmentation of the Alpine orogen, such that the areas now underlain by the Western Mediterranean basins were formerly a part of the M-plate.

The timing of the Alpine deformation has long been a controversial subject matter, both on a plate scale (i.e. the commencement of continental collision) and the migration of deformation towards the more external parts of the belt (see Chapter 5).

Traditionally the Alpine deformation has been divided

into Palaeo or Eo - Austroalpine, Meso - Penninic, and Neo - Helvetic, events or episodic phases (Trumphy, 1960). Hsu (in press) contests this hypothesis of episodic orogeny proposed by Trumphy (1960) and argues for a continuous and gradual process dependant on rates of sea-floor spreading in the N Atlantic (i.e. the driving force). Hsu (in press) proposes the postulate of continuous orogeny since the early Cretaceous and uses a late Eocene collision event as a time marker; the timing of this event is itself a contentious point (see Tricart, 1984). Deformation occurring prior to this event is designated as Eo-Alpine and is related to the consumption of Tethyan oceanic crust. Deformation post-dating the late Eocene collision, normally referred to as Meso- and Neo-Alpine events (Trumphy, 1960), is referred to as only Neo-Alpine, there being no evidence for the two-fold division proposed by Trumphy (1960). This Neo-Alpine deformation is related to the underthrusting of the European continent, resulting in intra-crustal thrusting and a decoupling of the cover sequence from its basement, producing thin-skinned deformation out in the European foreland.

As mentioned in the foregoing, this work is mainly concerned with the more external parts of the orogen, specifically that of the Dauphinois domain with its external crystalline basement massifs. These massifs have in the past been considered as autochthonous or para-autochthonous Variscan basement (e.g. Ramsay, 1963; Gratier & Vialon, 1980). However, with the recent advent of thrust tectonics the external basement massifs were re-interpreted as far-travelled and re-orientated allochthonous thrust sheets of basement (Beach, 1981a; 1981b; 1981c; Boyer & Elliott, 1982; Butler, 1983; Butler et al., 1986). Recent French interpretations (e.g. Lemoine et al., 1981; Lemoine et al., 1986) have emphasised the importance of syn-sedimentary extensional tectonics and considered the external basement massifs as crustal scale tilted fault blocks of extensional origin which controlled Mesozoic sedimentation patterns throughout the evolution of the European Tethyan margin (Graciansky et al., 1979; Lemoine

et al., 1981; Tricart, 1984; Tricart & Lemoine, 1986). Comparisons have been made with the structures found at the Atlantic margin, such as in the Bay of Biscay (Montadert et al., 1979). The interpretation of geophysical data (Menard, 1979; 1980; Perrier & Vialon, 1980; Thouvenot & Perrier, 1980; Bayer et al., 1987) has emphasised the importance of detachment and thrusting in the European crust. Nevertheless, pre-existing structures may have influenced many aspects of the Alpine deformation, such that interpretations which consider the external basement massifs as preserved extensional fault blocks little modified by later Alpine compression, or those arrived at from the application of thin-skinned thrust principles only, are both probably not correct in their entirety.

The stratigraphy of the Western Alps has been considered in the context of a passive margin rift sequence (Graciansky et al., 1979; Lemoine et al., 1981; 1986; Lemoine & Trumpy, 1987) and is extremely variable across the former European margin. For details of these variations the reader is referred to Chapter 4. The generalized stratigraphy of the Dauphinois domain in external zones has been summarized by Ramsay (1963) and will be described briefly here. It is stressed that even on a local scale, the distribution of syn-rift sediments is controlled by extensional faulting which is manifest as variations in lithological type and thickness. For this reason only a generalized description of the stratigraphy is presented. Detailed accounts of the stratigraphy, including information on these fundamental variations, are presented in Chapters 2 and 4.

1. Basement: The pre-rift basement to the Mesozoic cover succession was consolidated and metamorphosed during the Variscan orogeny and comprises largely micaschists, gneisses, migmatites, and syn-orogenic late Variscan granites, otherwise known as crystalline basement. Late Variscan (Stephanian) and Permian sediments (conglomerates, sandstones and shales) make up the remainder of this basement suite and are commonly referred to as the

Carboniferous Houiller.

2. Triassic: Triassic sandstones and platform carbonates and dolomites with sporadic evaporitic horizons were deposited onto a subsiding erosional surface. Crustal extension of this age is manifest as basaltic lava flows which are often intercalated with the platform dolomites. Continental rift activity was initiated during the Triassic.

3. Liassic: This can be broadly divided into the Lower Liassic, informally referred to as the Lias calcaire, comprising limestones interbedded with calcareous shales and the Upper Liassic, informally called the Lias schistéux, comprising predominantly black calcareous and non-calcareous shales. Extreme thickness and facies variations are evident, reflecting a basin-and-shoal depositional environment.

4. Dogger and Malm (Middle and Upper Jurassic): A return to calcareous sedimentation occurred and the Dogger formations are characterized by limestones interbedded with marls. Continental rift activity ceased during the Callovian and the dominantly pelagic facies of the Callovian to Oxfordian black shales, informally called the Terres Noires, and a distinctive Tithonian pelagic limestone were deposited. The Malm sequence marks the post-rift phase of thermal subsidence and is therefore extensive across the European margin.

6. Cretaceous: The Dauphine domain is characterized by deep water pelagic shale and limestone deposition, relating to the general subsidence of the margin. However, a carbonate platform developed outside the margin (Jura-Paris Basin, Massif-Central, and Provence areas) (Lemoine et al., 1986). The platform-basin boundary prograded eastwards such that an Urgonian platform spread out across a large part of the formerly pelagic domain. A Cretaceous extensional event with fault controlled subsidence is also evident over much of SE France.

7. Tertiary: With continental collision and crustal thrusting, loading of the European crust induced a crustal flexure and foreland basin initiation. A marine flysch sequence was deposited into this deepening foreland flexure. This flysch sequence, locally known as the Trilogie Nummulitique, overlies a basal conglomerate and comprises 1) Nummulitic bearing limestones, 2) globigerina bearing marls, and 3) a turbidite flysch sequence infilling the foreland basin. This trilogy signifies a progressive deepening of the foreland basin and its infilling with the arrival of clastic flysch deposits. The Trilogie Nummulitique, in the region of the Pelvoux massif, was deposited on previously deformed and uplifted basement and Mesozoic cover sediments.

The stratigraphy of the external Dauphinois domain is summarized in Figure 1.5.

1.3 Aims of the research

The general aims of this research are as follows:

1. To examine the extensional development and subsidence-history across a transect of the European Tethyan margin at the latitude of Grenoble, from the onset of continental rifting to sea-floor spreading and passive margin formation. Emphasis is placed on understanding the dynamics and the geometry of normal faulting with detailed studies in the Pelvoux area.

2. To determine the importance and control of pre-existing faults or basement structure over the development of later Alpine thrust structures in the Pelvoux area.

3. When pre-existing extension faults reactivate as contraction faults, this phenomenon is commonly referred to as structural inversion. A substantial part of the thesis concerns itself with this aspect, defining the characteristic structural styles, geometries and strain

patterns, and the mechanical factors influencing the reactivation of pre-existing faults.

4. To determine the timing of the deformation and the sequence of collapse and thrusting of the more external parts of the European margin (the Pelvoux and adjacent Subalpine Chains) during Tertiary Alpine compression.

5. To define the geometry and direction of overthrusting in the Pelvoux region and on a larger scale, trace continuity of structures around this part of the Western Alpine arc and account for the origins of the arc.

6. To consider the geometry of crustal deformation by comparison of the surface structure of the Pelvoux and adjacent basement massifs, with the preliminary interpretations of the ECORS-CROP (Bayer et al., 1987) deep seismic reflection profile across the Western Alps.

Aside from the aims outlined above, a comprehensive understanding of the overall structure and history of geological development of the Pelvoux region and its adjacent Subalpine Chains, forms the primary objective of this investigation.

1.4 Methods of investigation

The basic data for this thesis derive mainly from geological observation by mapping the area outlined in Section 1.1. Mapping was undertaken at scales of 1:25 000 and 1:15 000, using maps published by the Institut Geographique National de France. In mapping, emphasis was placed on understanding the 3-D structural geometries, the continuity of structures and structural styles, and the overall regional structure.

The standard techniques of structural analysis in thrust belts (Dahlstrom, 1969; 1970; Hossack, 1979; Elliott & Johnson, 1980; Boyer & Elliott, 1982) were engaged, these include:

1. Construction of detailed structural maps and stratigraphic cut-off maps, although caution must be exercised in the interpretation of the latter to determine thrust transport directions, if pre-thrust structures are identified.

2. Construction of balanced sections where feasible. It is often necessary to modify such constructions so as to incorporate the effects of pre-existing extension.

3. Analysis of the deformation associated with the emplacement and displacement of rock masses. This included a systematic analysis of kinematic indicators and structural fabrics in the field.

In order to unravel the pre-existing basin configuration and the extensional history of the Tethyan rift system, a subsidence-history analysis of the syn-rift and post-rift sediments of the Tethyan European margin was combined with studies of fault geometry and stratigraphy. Certain deductions concerning pre-existing fault configurations are made on the basis of facies distributions and syn-sedimentary thickness variations, fault geometry and the mechanical behaviour of the faults during subsequent compression. Where earlier extension faults are incorporated into the later Alpine thrust system, a removal of the contractional displacements can sometimes be used to restore and constrain the pre-contractional template geometry and early fault history.

An integrated or interdisciplinary approach should provide the most efficient method to study such an area, taking into consideration all of the different data aspects; structural, sedimentological, stratigraphic and geophysical. Many of the models proposed cannot be constrained solely from the use of geometric analytical methods; the mechanical aspects of the deformation should also be contemplated together with textural and metamorphic studies, providing a further test on the viability of the

interpretations and models proposed. Where possible, such an integrated approach has been adopted in this thesis.

1.5 Organization of the thesis

This thesis is divided into two parts: Part I presents the basic data and observations with some local interpretation, while Part II is more theoretical and interpretative, pertaining to models of regional development supported by data presented in Part I. Because of the similarity in content, instead of being subdivided into several Chapters, Part I comprises solely of one Chapter (2). Part II comprises Chapters 3,4,5,6, and 7.

Chapter 2 is divided into several sections, each section covering a different part of the field study area. This division and the coverage of the different sections is summarized in Figures 1.6 and 1.7. Hopefully no confusion should arise as to the relative positions of, and area covered by, the different maps presented in Chapter 2. Chapter 3 is concerned with the phenomenon of structural inversion and the controls which govern the reactivation of pre-existing faults, with applications to the study area. A publication on this subject (Gillrist et al., 1987) is enclosed at the rear of the thesis. Chapter 4 is devoted to the development of the European Tethyan margin and includes a tectonic subsidence-history analysis of the syn- and post-rift Tethyan sediments and a study of the passive margin fault geometry in the Pelvoux area. A model of lithospheric stretching is proposed to explain patterns of differential subsidence and uplift across the European margin. Chapter 5 examines the timing of the Alpine deformation, and the kinematics of Alpine thrusting in the Pelvoux area and the effect of pre-existing basin configuration on the development of the Western Alpine arc. Chapter 6 considers the deep structure of the Western Alps and discusses the ECORS-CROP (Bayer et al., 1987) deep seismic reflection profile, and its implications for the crustal structure of the Western Alps.

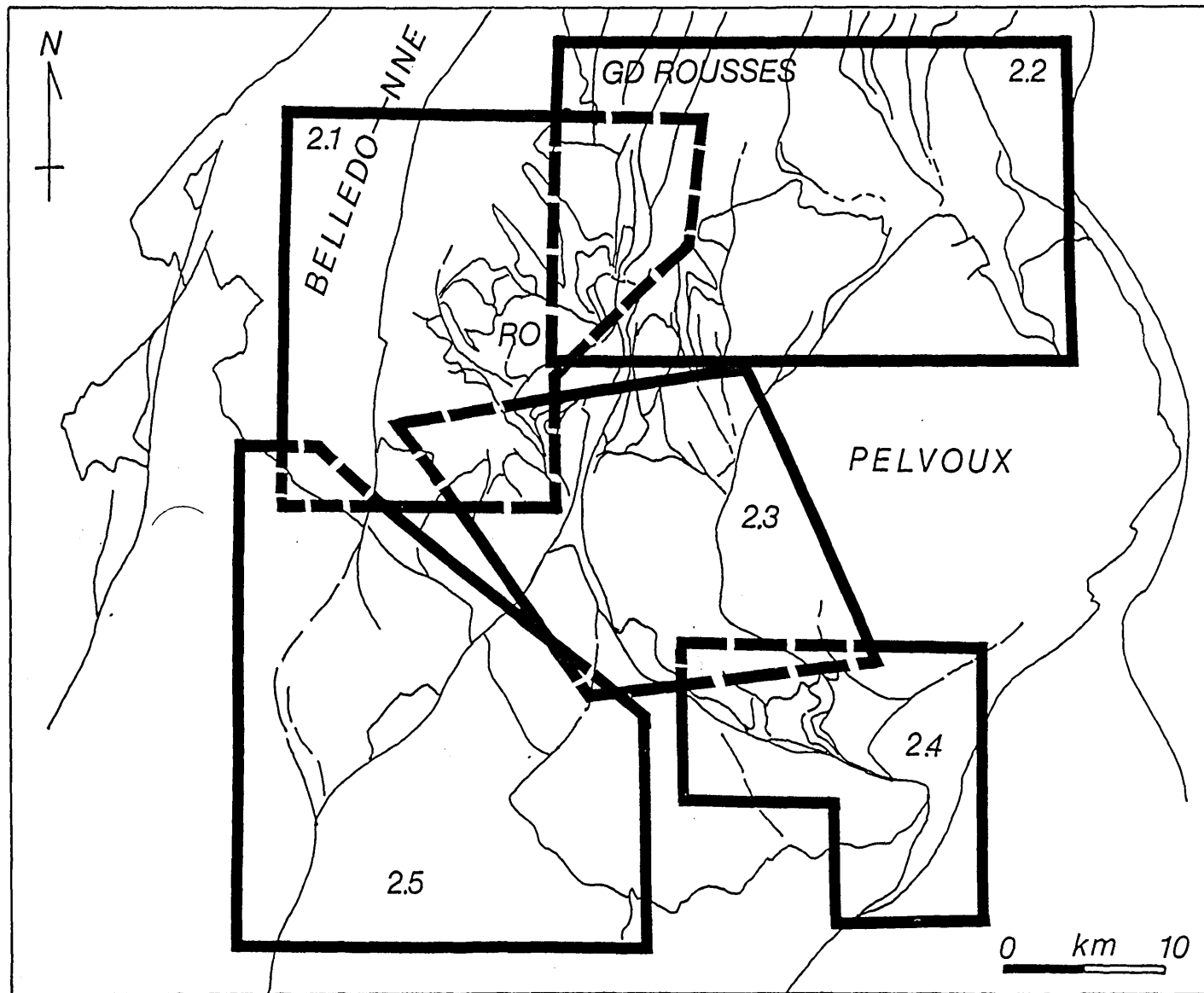


Figure 1.6: Areas comprising Chapter 2; 2.1 Bourg d'Oisans, 2.2 Northern Pelvoux, 2.3 Central Pelvoux, 2.4 Aiguille de Morges, 2.5 SW Pelvoux, 2.6 Vercors and Devoluy are off the map area. The outline of the basement massifs is indicated, Ro-Rochail massif.

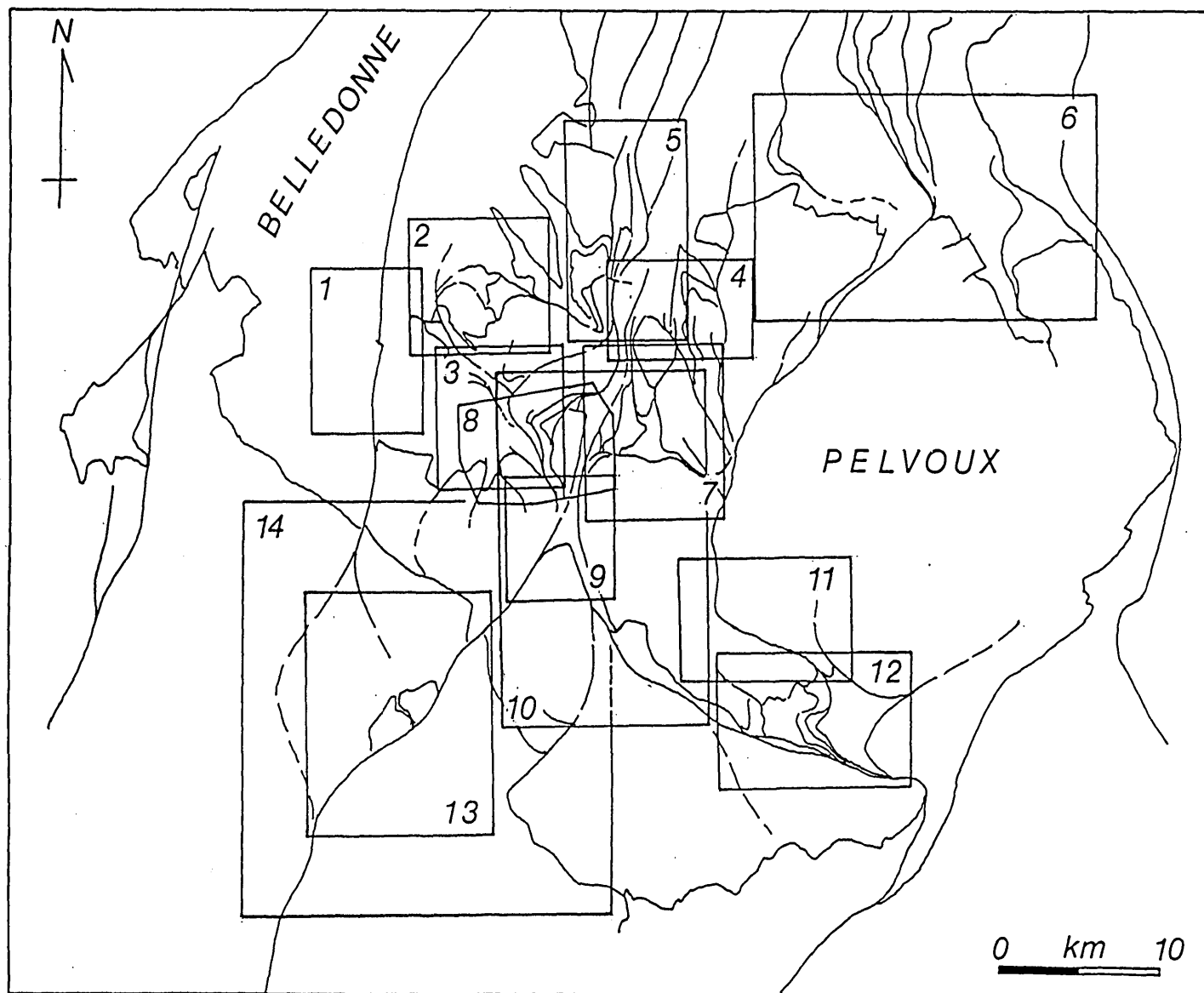


Figure 1.7: Location of maps presented in Chapter 2; 1-Fig. 2.1.5, 2-Fig. 2.1.42, 3-Fig. 2.1.46, 4-Fig. 2.2.2, 5-Fig. 2.2.5, 6-Fig. 2.2.11, 7-Fig. 2.3.1, 8-Fig. 2.3.5, 9-Fig. 2.3.8, 10-Fig. 2.3.9, 11-Fig. 2.3.15, 12-Fig. 2.4.1, 13-Fig. 2.5.8, 14-Fig. 2.5.1.

BASIC LEGEND

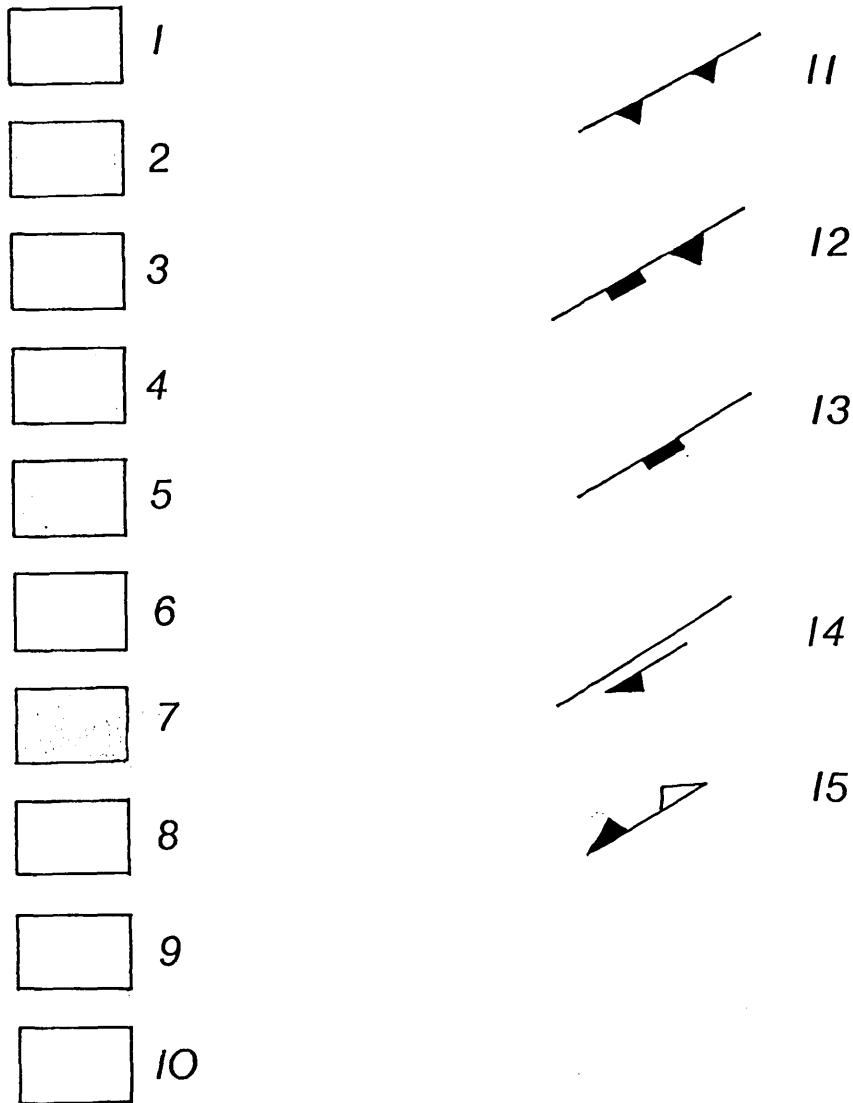


Figure 1.8: Basic legend for the maps and sections presented in this thesis, unless stated otherwise. 1. Crystalline basement, 2. Carboniferous Houiller, 3. Triassic platform dolomites, 4. Triassic continental basalts, 5. Lower Liassic, 6. Upper Liassic, 7. Middle Jurassic, 8. Terres Noires, 9. Tithonian, 10. Lower Cretaceous, 11. thrust faults, barbs indicate the direction of overthrusting, 12. early normal fault reactivated as a thrust fault, 13. normal fault, 14. strike-slip fault, arrows indicate sense, 15. reactivated fault symbol used on cross-sections, filled arrow denotes last sense of movement.

Concerning the nine-figure grid references used in Part I: these refer to the grid based on degrees from the international meridian (numbers shown in black on the exterior margins of the Institute Geographique National de France maps).

The majority of maps and sections presented in Part I are colour coded. The standard legend used for maps and sections throughout this thesis, unless stated otherwise elsewhere, is shown in Figure 1.8.

PART I:

Presentation and local interpretation of the field data

Chapter 2: Field data

2.1 Structure and stratigraphy of the Bourg d'Oisans area

2.1.1 Introduction

The location of the Bourg d'Oisans area is shown on Figure 2.1.1 (see also Figure 1.6). The characteristic feature is a N-S trending structural depression filled with Mesozoic sediments juxtaposed between two uplifted basement massifs; the Taillefer-Belledonne and the Rochail-Pelvoux massifs. This structural depression can be traced northwards along the eastern margin of the Belledonne massif, N of the present study area (see Figure 1.2), and in the literature is often referred to as the Bourg d'Oisans syncline (e.g. Barfety et al., 1979; Gidon & Barfety, 1983). The contact between Mesozoic sediments and the Taillefer basement in the W corresponds to a steep N-S trending normal fault, called the Ornon fault (Barfety et al., 1979). To the W, a similar NNE-SSW trending depression filled with Mesozoic sediments, plunges southwards off the uplifted Taillefer basement massif and is limited on its western margin by a NNE-SSW trending normal fault, the La Mure fault (Figure 2.1.41 for location), which juxtaposes Mesozoic sediments against the La Mure basement. Based on sedimentological arguments, these sediment filled depressions have been considered as Jurassic half-grabens which developed on the hanging-walls of crustal-scale tilted fault blocks, undergoing subsidence during extension (Barfety et al., 1979; Lemoine et al., 1981; 1986; Barfety & Gidon, 1983; Bas, 1985). The structure was also addressed by Ramsay (1963), who considered these structural depressions as narrow cover synclines pinched between broader basement anticlines, an interpretation later revived by Tricart & Lemoine (1986). Detailed meso- and micro-structural investigations of the deformed Bourg d'Oisans sedimentary cover have been carried out by Gratier et al. (1973), Gratier & Vialon (1975; 1980), and Vialon (1968; 1979).

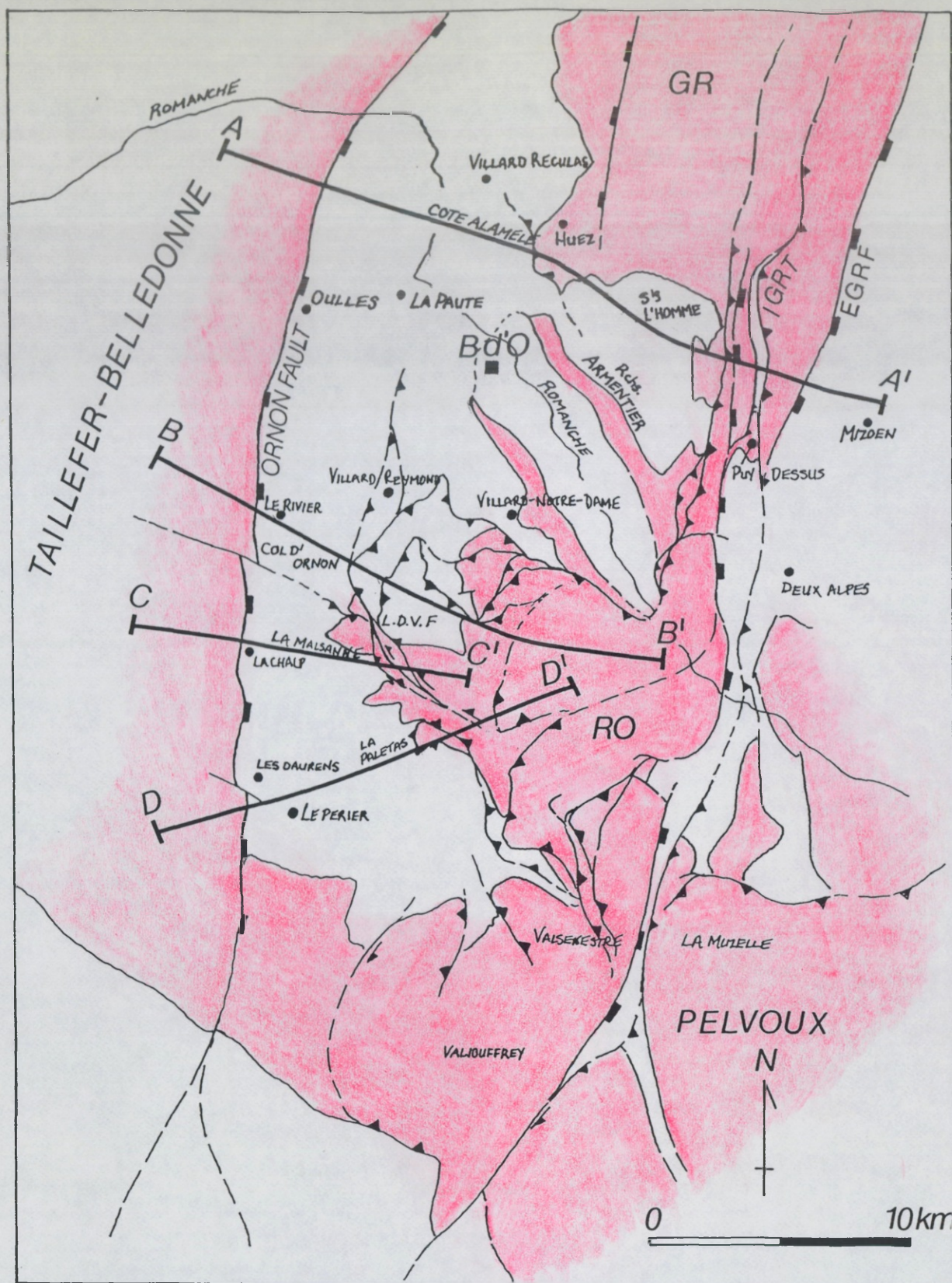


Figure 2.1.1: Simplified structural map of the Bourg d'Oisans area showing the principal faults and the locations of the section lines discussed in the text. GR-Grandes Rousses massif, RO-Rochail massif, EGRF-Eastern Grandes Rousses fault, IGRT-Internal Grandes Rousses thrust, LDV.F-Lac du Vallon faults, Bd'O-Bourg d'Oisans, AA'-Oulles - Mizoen cross-section, BB'-Grand Renaud cross-section, CC'-La Malsanne - Le Neyrarel cross-section, DD'-La Paletas - Le Perier cross-section, lines BB' and DD' also correspond to the Grand Renaud and Le Paletas - Le Perier stratigraphic sections respectively.

2.1.2 Mesozoic stratigraphy

The Mesozoic sediments have been studied in detail by French workers, notably Lemoine et al. (1981), Barfety & Gidon (1983) and Barfety (1985). These studies indicate that important vertical isostatic movements governed the pattern of Mesozoic sedimentation. The following descriptions of the Mesozoic stratigraphy are largely based on these works.

Two E-W stratigraphic sections through the N-S trending Bourg d'Oisans sedimentary cover will be presented.

2.1.2.1 Grand Renaud stratigraphic section:

The location of the Grand Renaud stratigraphic section, in the northern part of the Bourg d'Oisans area, coincides with the Grand Renaud structural cross-section (see Figure 2.1.1). The generalized stratigraphy and important stratigraphic features are shown in Figure 2.1.2. Liassic and Dogger syn-rift formations of significant thickness, progressively thin eastwards pinching out onto a depositional high over the Rochail basement, which was transgressed by the Malm post-rift sequence.

1. Malm post-rift sequence: The youngest sediments exposed on the Grand Renaud and the Pic du Col d'Ornon (G.R.7379,49869), are deep water pelagic shales of Berriasian (Neocomian) age (Rouboul, 1962). These form a part of the post-rift sequence which includes Tithonian pelagic limestones, and Kimmeridgian pelagic shales (locally referred to as the Sequanian and Argovian). In the eastern part of the section, towards the Rochail basement, the Sequanian and Argovian shales locally rest on shallow water bioclastic limestones, informally called the Calcaire du Petit Renaud (Barfety & Gidon, 1983). The shallow water limestones themselves overlies Lower Liassic limestones (Lias calcaire). The sedimentary contact between the Lias calcaire and the Calcaire du Petit Renaud is

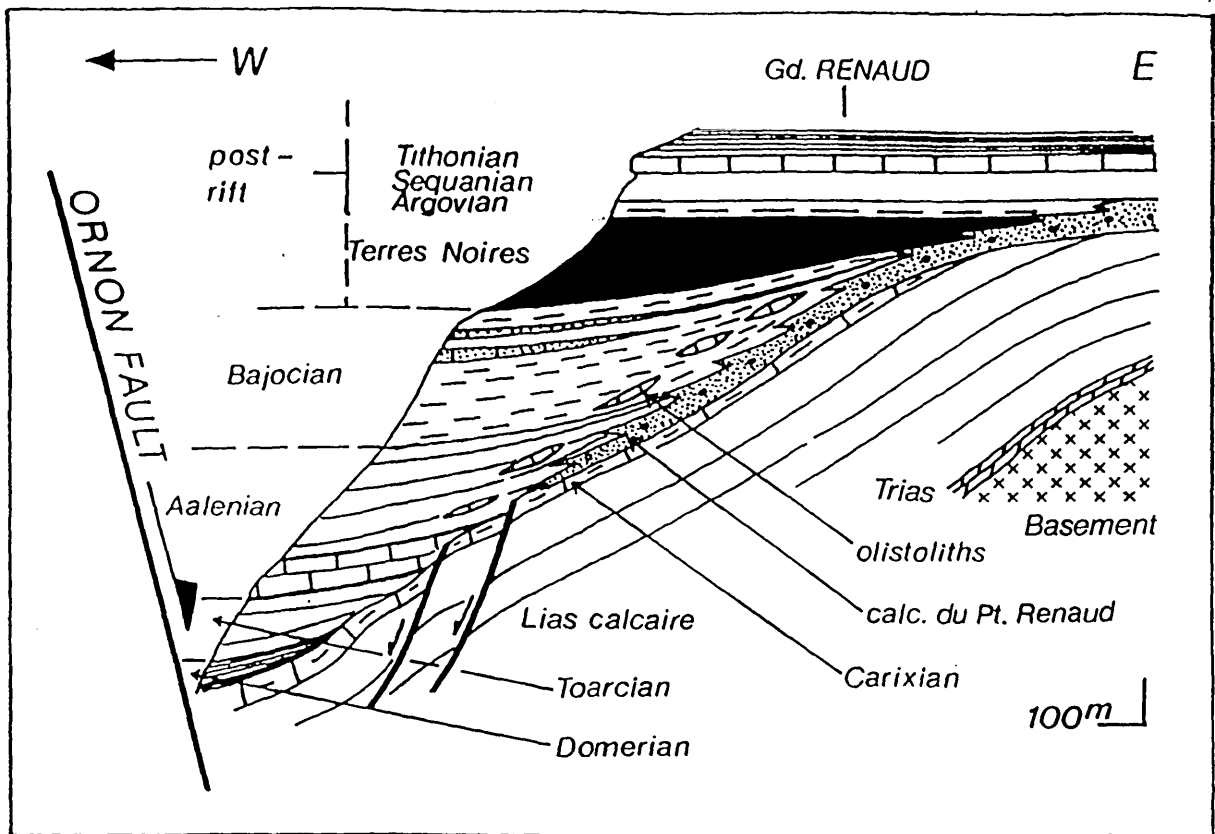


Figure 2.1.2: Schematic stratigraphic section through the northern Bourg d'Oisans area, the Grand Renaud section (see Fig. 2.1.1 for location), showing the effects of faulting, tilting, onlap and olistostrome development. Note the thickening of syn-rift sediments towards the Ornon fault (modified after Barfety & Gidon, 1983).

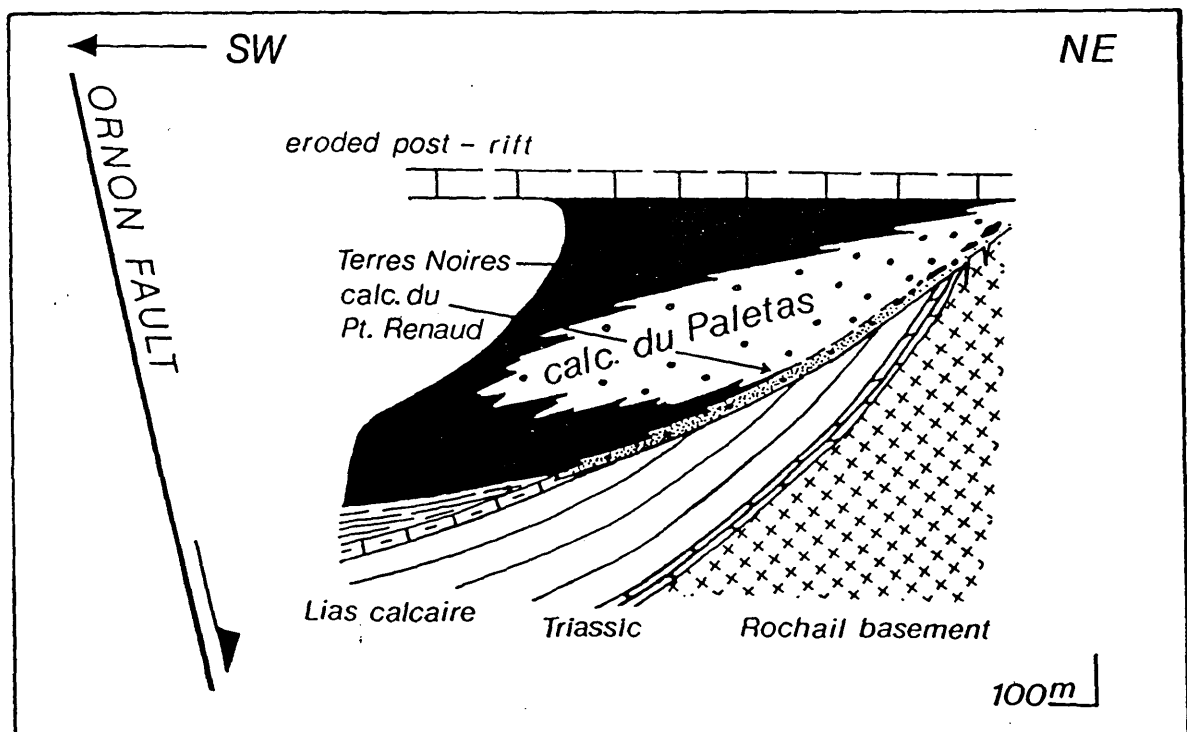


Figure 2.1.3: Schematic stratigraphic section through the southern Bourg d'Oisans area, the La Paletas-La Perier section (see Fig. 2.1.1 for location). Note the uplift and erosion of basement and earlier syn-rift sediments in the E (modified after Barfety & Gidon, 1983).

marked by an angular discordance at the Pic du Col d'Ornon (Barfety & Gidon, 1983) (see Figure 2.1.2). The Calcaire du Petit Renaud is not laterally extensive and is transitional with Oxfordian black shales (Terres Noires) and a Bajocian limestone-shale sequence to the W.

The Oxfordian black shales locally display substantial thickness variations ranging from 100m in the W to 0m the E, where they are transitional with the Calcaire du Petit Renaud. Minor syn-sedimentary growth faults which affect these black shales are sealed by overlying Argovian and Sequanian shales (Barfety & Gidon, 1983).

2. Liassic and Dogger syn-rift sequence: In the W, towards the Ornon fault, is a thick sequence of Toarcian and Aalenian deep water shales and Bajocian carbonates. To the E the Bajocian carbonates, and to a certain extent the Aalenian shales, pass laterally into the Calcaire du Petit Renaud which overlies the Lias calcaire. This E-W facies transition is marked by Carixian (Pliensbachian) age olistoliths lying within the Bajocian and Aalenian sediments. The presence of these Carixian age olistoliths lead Barfety & Gidon (1983) to infer the existence of a westward dipping Jurassic palaeoslope, reflecting a westward tilting of the Lias calcaire in the eastern part of this section.

From stratigraphic relationships and palaeontological age dating, the Calcaire du Petit Renaud can be interpreted as a condensed sequence ranging in age from possibly Early Bajocian to Callovian (Barfety & Gidon, 1983).

3. Triassic pre-rift sequence: Platform dolomites of uniform thickness, and continental alkaline basalts, directly overlie the Rochail crystalline basement in the eastern part of this section and constitute the pre-rift sediments of this area.

All of the relationships mentioned in the foregoing are shown in Figure 2.1.2.

2. La Paletas - Le Perier stratigraphic section:

The La Paletas stratigraphic section is located in the southern part of the Bourg d'Oisans area and coincides with the La Paletas - Le Perier structural section (Figure 2.1.1).

The post-rift sequence has been largely stripped-off by erosion, and it is mainly the Triassic pre-rift and Jurassic syn-rift sediments that are preserved. The generalized stratigraphy and important stratigraphic features are shown in Figure 2.1.3.

The Jurassic syn-rift stratigraphy closely resembles that described from the Grand Renaud section to the N. Towards the W, in the Ornon valley are exposed Domerian (Pliensbachian) and Toarcian age limestones and shales (Liassic syn-rift). At the eastern end of the section towards La Paletas (G.R.7380,49821), these Upper Liassic limestones and shales and most of the Dogger section are missing. An important eastward thinning of Oxfordian black shales (Terres Noires) is also associated with this eastward pinch out of the syn-rift sediments. To the E the Oxfordian black shales pass laterally into a newly differentiated formation, the Calcaire du Paletas (Barfety & Gidon, 1983), named after the peak in which it outcrops. The exact nature of this transition is obscured by later Alpine back-thrusting (Figure 2.1.37). Sedimentologically, this deposit closely resembles the Calcaire du Petit Renaud, corresponding to a shallow water bioclastic limestone comprising crinoidal debris swept into banks or shoals. The Paletas limestones are probably of basal Callovian or Oxfordian age, although they may include Bathonian and Upper Oxfordian age sediments (Barfety & Gidon, 1983).

The Calcaire du Petit Renaud occupies the same stratigraphic level as in the Grand Renaud section, separating the Terres Noires and Calcaire du Paletas from the underlying Lias calcaire. The Calcaire du Petit Renaud

is palaeontologically dated as being younger than Aalenian (Barfety & Gidon, 1983) and contains reworked clasts of Liassic limestones. The base of this transitional facies stratigraphically cuts down into older sediments to the E: firstly Lias calcaire, then Triassic spillites and dolomites, and eventually crystalline basement at the Rochers de Grande Eglise (G.R.7390,49814) and the Breche de Lauvitel (G.R.7400,49830) (see Figure 2.1.46). Where the Calcaire du Petit Renaud is in direct contact with crystalline basement, it is difficult to differentiate from the Calcaire du Paletas.

In the Bourg d'Oisans area, the general N-S alignment of deep and shallow water facies, uplift and erosion along a N-S axis in the E, and the presence of onlaps and olistoliths all suggest that sedimentation was tectonically controlled. The inferred deepening of the depositional environment towards the Ornon fault implies that Jurassic movements on this fault may be responsible. The erosion down to basement levels on the western margin of the Rochail massif, is interpreted to be the result of foot-wall uplift related to extension on the Ornon, and other Jurassic faults to the E, now obscured by Alpine thrusting.

Post-dating the active tectonics of the syn-rift phase, the European margin subsided and the earlier extensional fault blocks were subsequently "drowned". The post-rift sequence was deposited during this phase of thermal subsidence (Lemoine et al., 1986).

2.1.2.3 La Mure stratigraphy:

The La Mure stratigraphy is characterized by a westward thickening of syn-rift sediments towards the La Mure fault, and pronounced lateral facies variations; shallow water platform limestone facies on the uplifted foot-wall of the La Mure fault block pass basinward into deeper water pelagic shales. The transition is marked by calci-turbidites indicative of a steep depositional slope

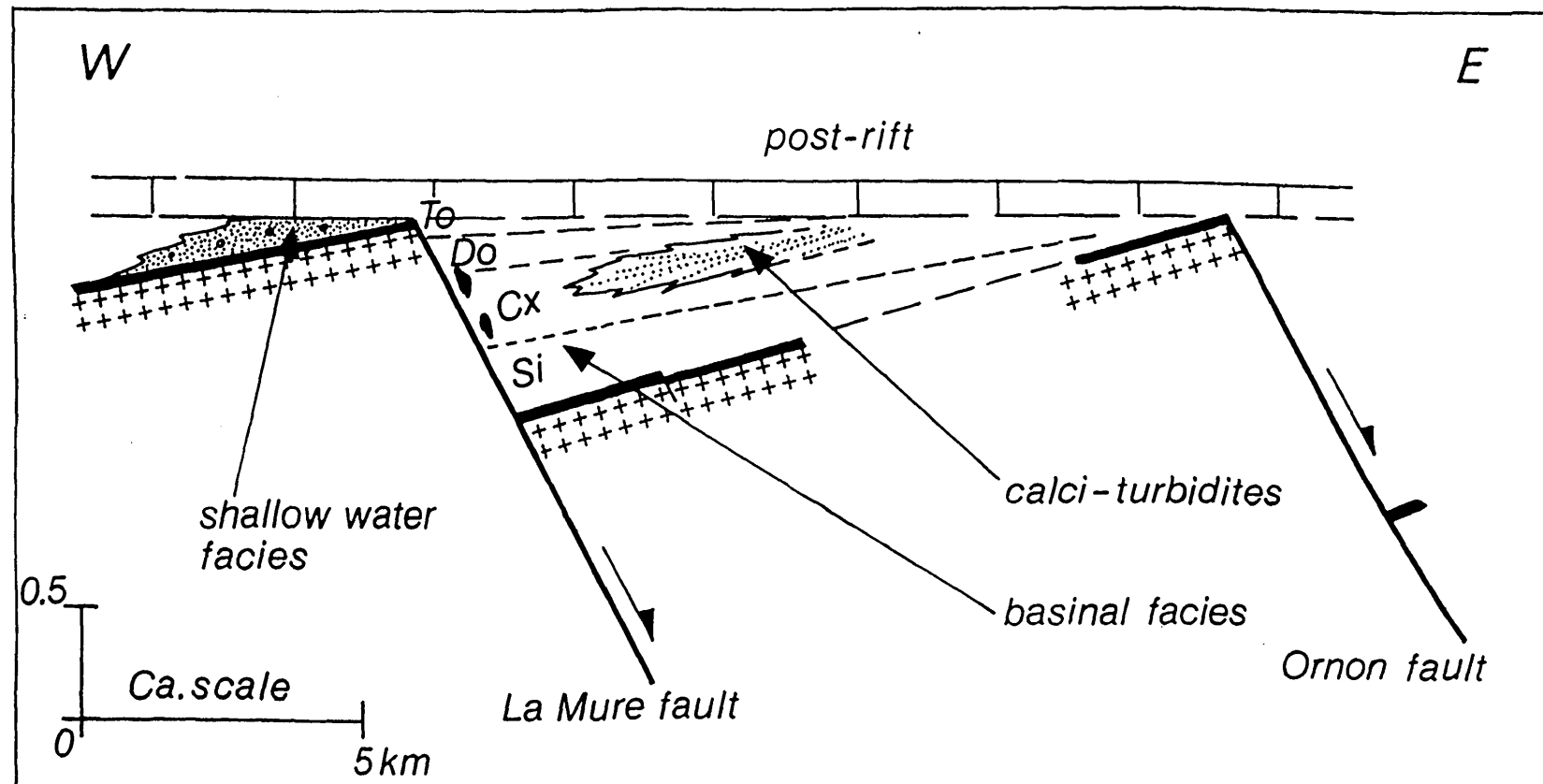


Figure 2.1.4: Schematic reconstruction of the stratigraphy across the La Mure half graben, To-Toarcian, Do-Domesian, Cx-Carixian, Si-Sinemurian & Hettangian, black- pre-rift Triassic, crosses- pre-rift basement (after Bas, 1985).

from the uplifted foot-wall to the deeper parts of the basin (Bas, 1985). The La Mure basin stratigraphy is summarized in Figure 2.1.4.

The Bourg d'Oisans and La Mure stratigraphic sections clearly record an early history of Liassic growth and basin formation, controlled by movements on the Ornon and La Mure faults. For future reference, they are referred to as the Bourg d'Oisans, and La Mure half-grabens respectively.

2.1.3 Structure

2.1.3.1 Ornon fault:

The Ornon fault has a present day normal offset, and trends NNE-SSW along the eastern margin of the Belledonne-Taillefer basement massif. The northern termination of this fault may correspond to where the Belledonne massif plunges northwards underneath its Mesozoic cover. To the S, the continuation of the Ornon fault in the Jurassic cover of the Beaumont mountains (see Figure 2.5.1), becomes increasingly difficult to follow. It may branch or link with the NE-SW trending Aspres-le-Corps fault to the SW of the Pelvoux massif (see Section 2.5).

In the Bourg d'Oisans area the Ornon fault changes strike (Figure 2.1.1). To the S of the village of Oulles, the Ornon fault trends N-S dipping 65-70 E. However, N of Oulles the trend of the fault changes to become NNE-SSW and its eastward dip steepens becoming marginally overturned at the village of Allemont (Grand, pers. comm.). This steepening and overturning of the Ornon fault may reflect movements on a deeper level thrust in the basement, the external Belledonne thrust (Menard, 1979).

In Ornon valley S of Bourg d'Oisans, the Ornon fault forms a present day topographic feature, and the steep-walled Ornon valley side trends parallel to the exhumed fault surface (Plate 2.1.1). The best exposures of the Ornon fault are at three consecutive localities along



Plate 2.1.1: Viewing SW towards the Ornon fault. The present valley side parallels the exhumed fault scarp and the La Chalp olistoliths adjacent to the fault can be discerned. The size of these olistoliths is evident from buildings in the adjacent village of La Chalp.



Plate 2.1.2: The Ornon fault plane at La Chalp with a cohesive ultra-cataclasite deformed by tensional veins. The camera lens cap is 5.5 cm in diameter. See text for locality and details.

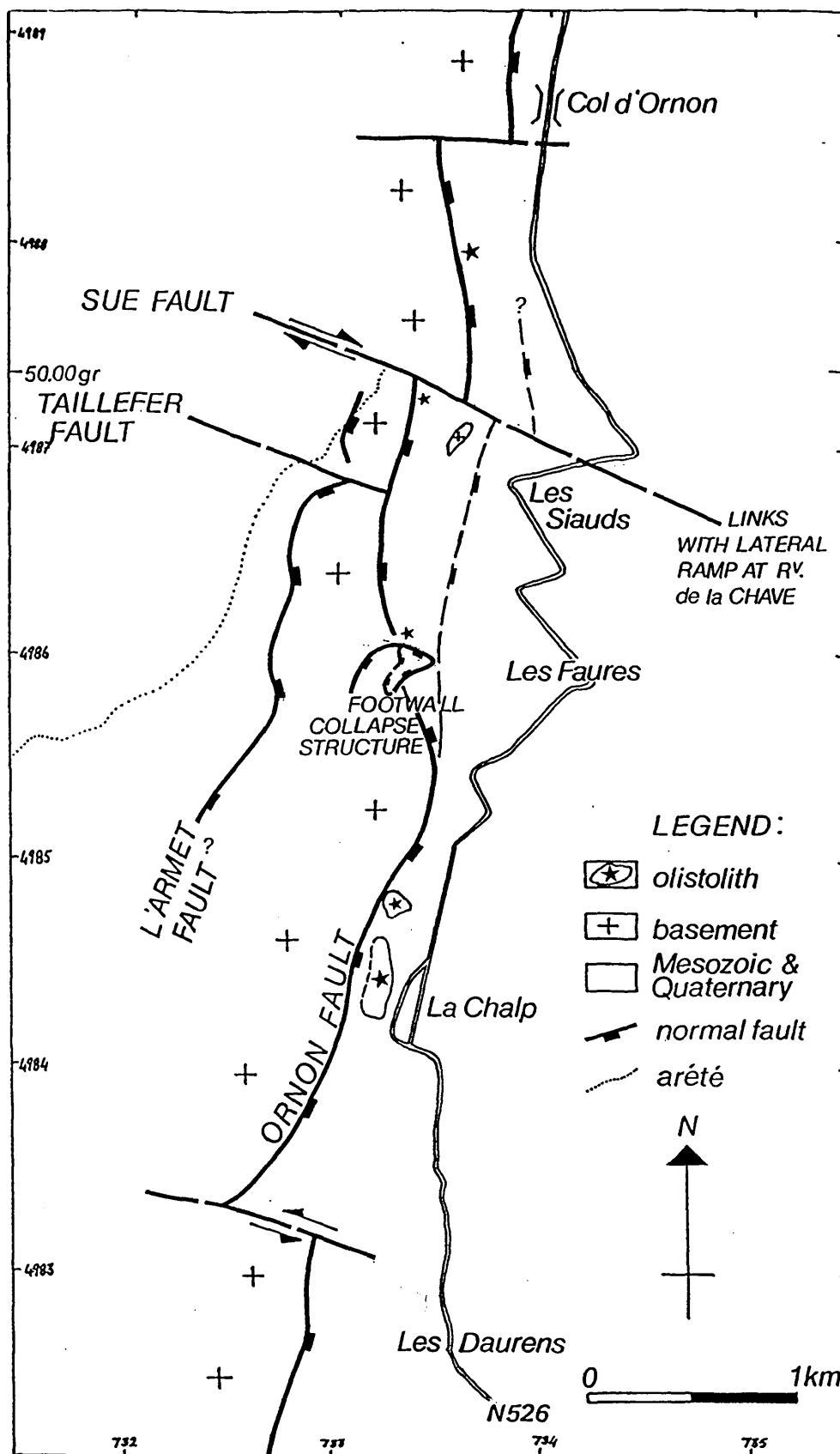


Figure 2.1.5: Simplified structural map of the Ornon fault along the Ornon valley (see Fig. 1.7 for regional location). Principal localities discussed in the text are indicated. Based on mapping by the author.

the western Ornon valley side (Figure 2.1.5). A description of each locality follows:-

1) La Chalp - Chantelouve:

In the Ornon valley, W of La Chalp, the Ornon fault and its contact relationship with the sedimentary cover crops out.

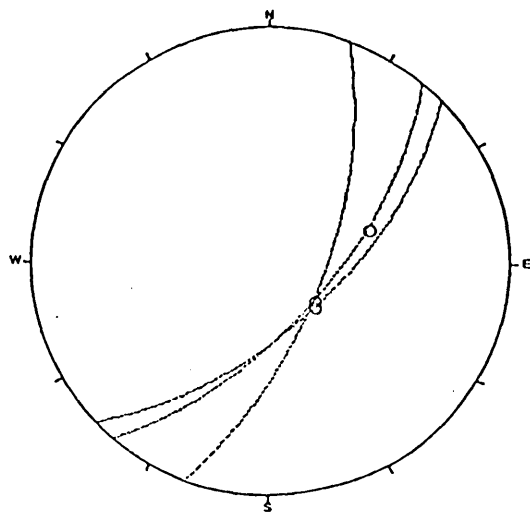
i) Fault plane description: Where exposed, the Ornon fault is typically a discrete, planar structure, dipping 60-70° E juxtaposing cover rocks against the Taillefer basement. At La Chalp (G.R.7331,49841), indeed for most its length, the trace of the fault plane is curvilinear (Figure 2.1.5).

The fault plane constitutes a zone of intense shear, marked by a thin veneer of fine grained fault rock resembling a cohesive ultra-cataclasite (Sibson, 1977). A primary banding or foliation is observed with green, presumably chlorite rich horizons, interleaved with darker fine grained fault rock. This fine grained fault rock is deformed by a number of brittle, tensile fractures containing quartz fibre growths (Plate 2.1.2). Perpendicular to these tensile fractures are fibrous slickensides resting on the surface of the ultra-cataclasite. These fibrous slickensides and tensile fractures are consistent with a normal dip-slip sense of movement.

In the immediate foot-wall of the fault - in the basement, the strain becomes distributed, with a progressive decrease in intensity away from the main fault. However, other faults are evident further into the foot-wall and the strain is again localized along discrete anastomosing shear zones. This suggests that although the exposed Ornon fault corresponds to a discrete break, the real width of the fault zone accommodating the extensional movement may be quite large.

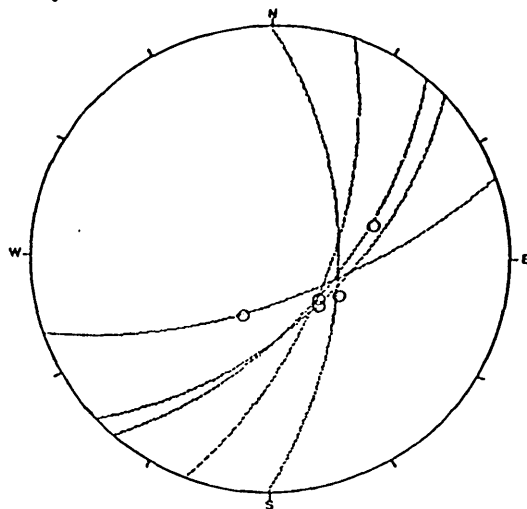
Where the cover sediments abutt against the fault

a. Ornon fault - La Chalp



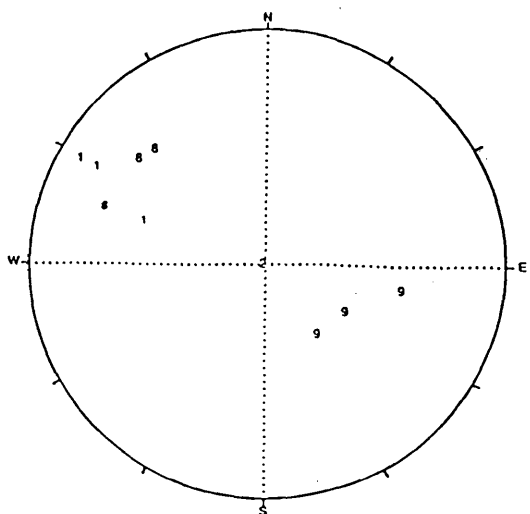
Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - FAULT PLANE	3	50.00
9 - SLICKENSIDES	3	50.00
Total number of data		6

b. Ornon fault



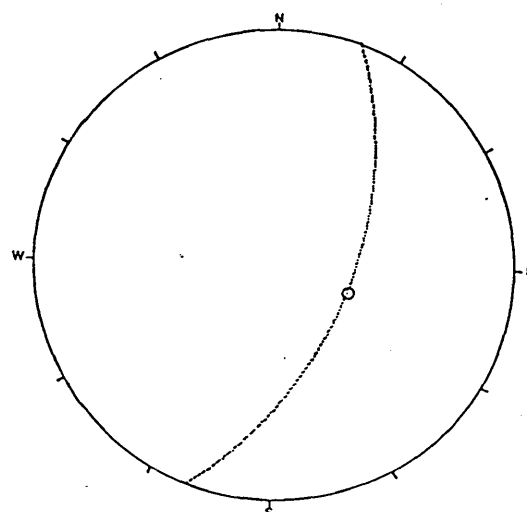
Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - FAULT PLANE	5	50.00
9 - SLICKENSIDES	5	50.00
Total number of data		10

c. Ornon fault - La Chalp (poles)



Codes	n°	%
1 - CLEAVAGE	5	38.46
2 - DILATIONAL FRAC	2	15.38
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - FAULT PLANE	3	23.08
9 - SLICKENSIDES	3	23.08
Total number of data		13

d. La Chalp olistolith



Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - FAULT PLANE	1	50.00
9 - SLICKENSIDES	1	50.00
Total number of data		2

Figure 2.1.6: Equal angle (Wulf net) plots of slip-lineations and fault plane orientations, a. for La Chalp locality, b. for all Ornon fault, d. for La Chalp olistolith, and c. equal area (Schmidt net) plot of linear and planar (poles) fabrics for the La Chalp locality.

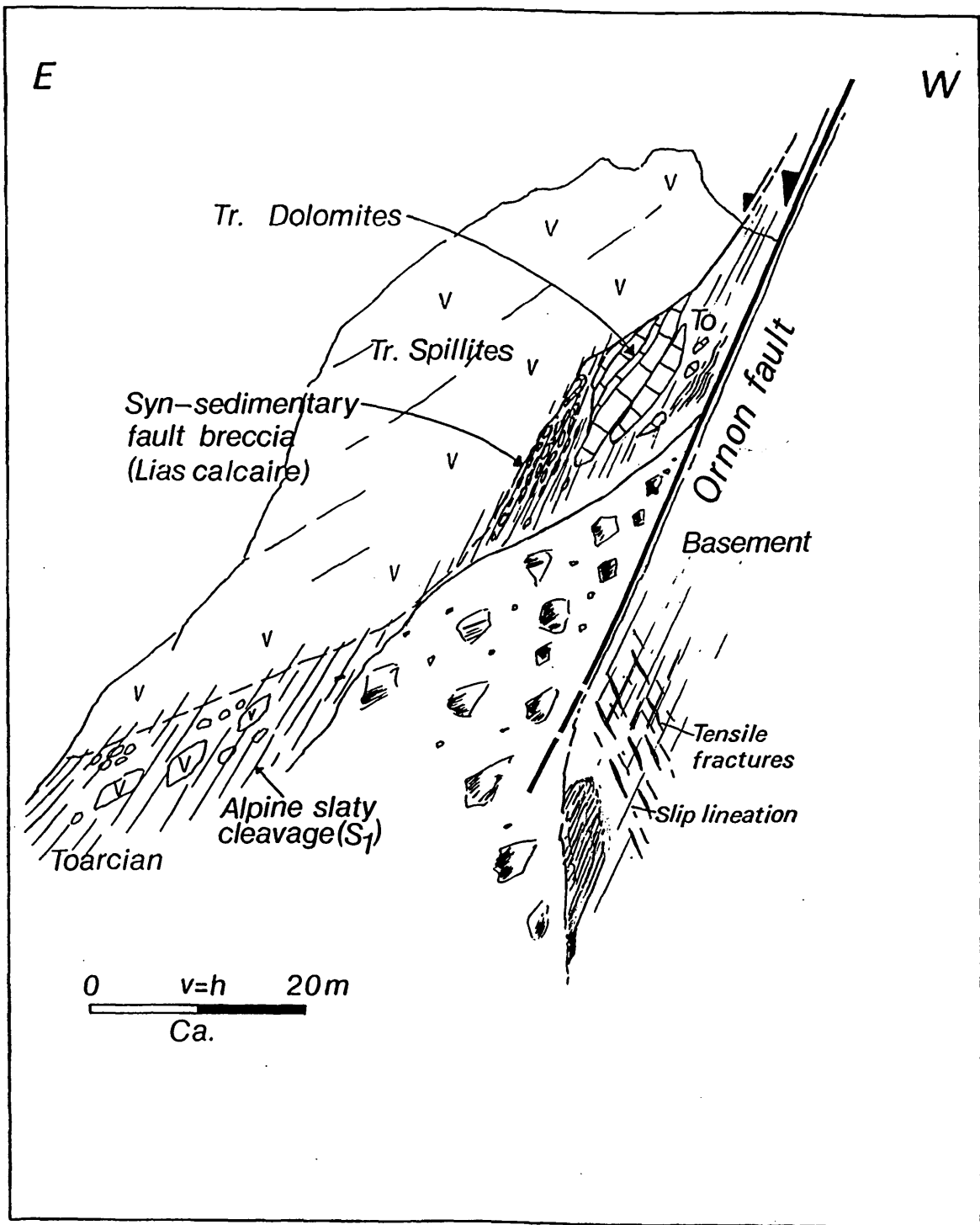


Figure 2.1.7: Field sketch of the Ornon fault and La Chalp olistolith showing principal sedimentological and structural fabrics. Tr-Triassic, V-spillites (basalts), To-Toarcian (see also Barfety & Gidon, 1984).

plane, a fine grained clay rich fault gouge has formed suggesting an overall marked intensification of slip develops preferentially along the interface between the gouge and the intact rock. It is difficult to obtain conclusive and consistent shear sense directions from this gouge zone. Shear bands which deform the steeply eastward dipping (020/62° E) Alpine cleavage suggest that both reverse and normal reactivations of this fault zone occurred, or alternatively, there was a dominantly pure shear deformation of the cover adjacent to the fault plane. A progressive decrease in dip of the main Alpine cover cleavage towards the E, suggests that the last phase of movement on this fault was extensional.

All of the structural orientation data recorded for this locality are shown stereographically in Figure 2.1.6.

ii) Jurassic age sedimentary olistoliths: To the NW of La Chalp are two large blocks of bedded Triassic sediments with length/width dimensions in the order of 80-100m x 20-50m. They are completely enveloped in younger Jurassic sediments, and have been previously documented by Barfety et al. (1979), and Barfety & Gidon (1983), who postulate a Jurassic age olistolith interpretation.

The most northerly of these olistoliths is shown in Figure 2.1.7. It is composed of predominantly Triassic spillite, with some Triassic dolomites at the olistolith base. The original bedding appears to be preserved in both units, and at the base of the spillite, on the SW facing side of the olistolith, a discrete detachment or shear plane is exposed (Figure 2.1.8). Slickenlines indicate that the spillite block slid down towards the E (Figure 2.1.6 d), presumably from the topographically higher foot-wall.

The olistoliths rest in Toarcian age sediments (Barfety et al., 1979). Smaller syn-sedimentary blocks of Triassic spillite and dolomite, and disrupted Lower Liassic limestone clasts also occur within the Toarcian matrix. Immediately adjacent to the Ornon fault, the bedding in

the Liassic limestones and shales is completely disrupted, giving the appearance of syn-sedimentary mass flow deposit.

2) Col d'Ornon:

Between La Chalpe and the Col d'Ornon, several exposures located along the foot-wall scarp of the Ornon fault are characterized by olistoliths and mega-breccias in Toarcian age sediments. The Ornon fault and its hanging-wall sediments are particularly well exposed S of the Col d'Ornon on the southern slopes of la Sue (G.R.7335,49872) (Figures 2.1.5 & 2.1.10).

i) Fault plane description: The exposed fault rock is a fine grained cohesive ultra-cataclasite which includes coarser brecciated fragments of wall rock - crystalline basement. Discrete brittle shears which intersect the principal foliation at low angles (15°) are interpreted as synthetic reidel shears (see Figure 2.1.9). These together with other shear sense indicators are kinematically consistent with a normal dip-slip sense of displacement.

The width of the shear zone affecting the foot-wall of basement is in the order of tens of metres, although restricted access does not allow for a more accurate thickness estimate. The fault zone is comprised predominantly of sheared basement, with occasional slivers of Triassic cover sediments caught up in the shearing. The orientation of the fault plane and principal foliation is $070/68^\circ$ E.

Adjacent to the fault plane, lenses of basement, limited by anastomosing shear zones, are introduced into the disturbed cover sequence. These are interpreted to be minor foot-wall collapse structures, plucked from the foot-wall fault scarp during or after its formation. To the E, these pass into irregularly shaped decametric blocks of Triassic dolomite and spillite. Further E, there is a progressive decrease in the size of the clasts making up this chaotic breccia, and the limestone clast content

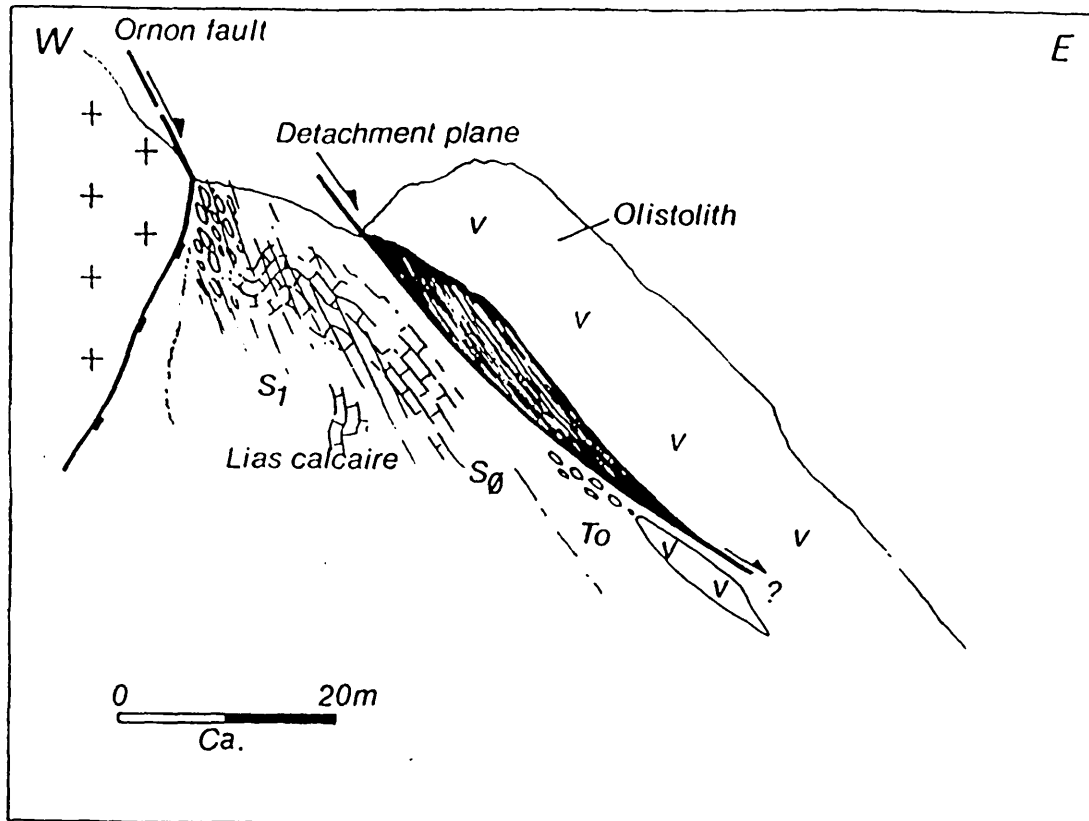


Figure 2.1.8: Field sketch of the detachment plane at the base of La Chalp olistolith, S0-bedding trace, S1-main Alpine cleavage, To-Toarcian, V-spillites.

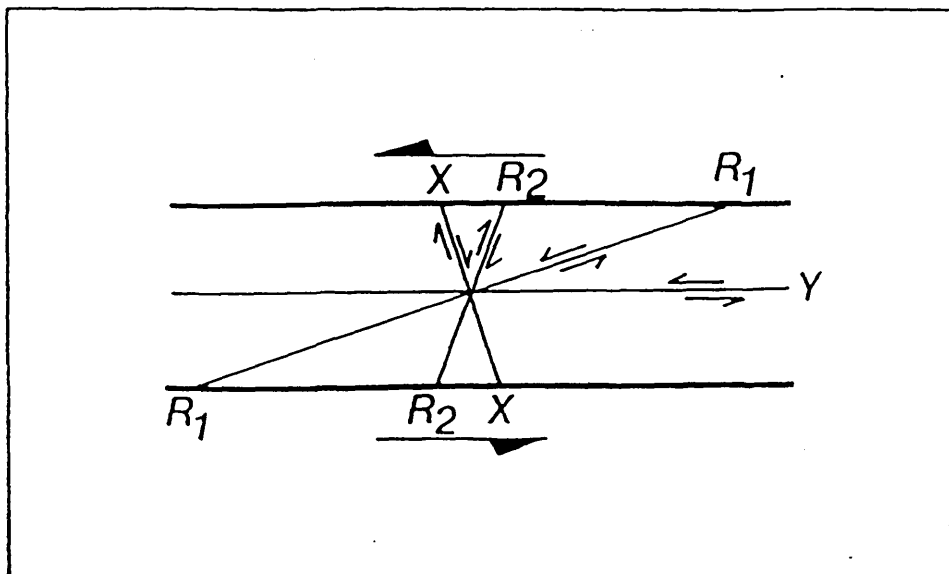


Figure 2.1.9: Micro-shear arrangements in gouge and cataclastic fault zones used to determine the sense of shear, the R1 extensional shear bands and Y-foliation are indicated. From Logan et al. (1979).

increases. The angularity of these breccia clasts, some of which show internal fracturing, suggests that sedimentologically they have not been transported very far and the presence of larger blocks (megabreccias) adjacent to the Ornon fault, suggests that they have been derived locally. These breccias, interpreted as debris flows by Barfety & Gidon (1984), progressively interfinger with a bedded Liassic shale sequence of Toarcian age (Figure 2.1.11). To the E this sedimentary sequence appears to rest conformably on poorly exposed basement (Figure 2.1.10).

At this locality the Ornon fault is offset by a WNW-ESE trending transfer fault called the La Sue fault (Figure 2.1.5). To the E this fault may link with, or be the same fault, as a compressional lateral/oblique ramp structure in the basement. Similar transfer faults offset the Ornon fault along its length. The exact age of these faults is difficult to determine; they may be Alpine, although a Jurassic age of initiation and Alpine age reactivation is preferred.

A series of major extensional faults which deform the foot-wall basement of the Ornon fault can be observed from the N562 road between the Col d'Ornon and La Chalp. The most impressive of these forms the E face of "L'Armet" mountain (Figure 2.1.12) and is interpreted as a splay fault from the main Ornon fault.

On a smaller scale, at the Cros de Qunchiol (G.R.7331,49860), a basement block is detached from the main massif along a gently dipping fault. In the foot-wall to this fault are both basement and cover sediments (Figure 2.1.13). The hanging-wall basement of this fault is deformed by a number of discrete extensional brittle shears with an R1 orientation which detach into the main extensional fault (Figure 2.1.13). The normal shear sense determined from fibrous slickensides along this basal fault is consistent with that determined from extensional shear bands (Figure 2.1.9). The cover sediments in the foot-wall to this fault form a chaotic breccia of Triassic dolomites,

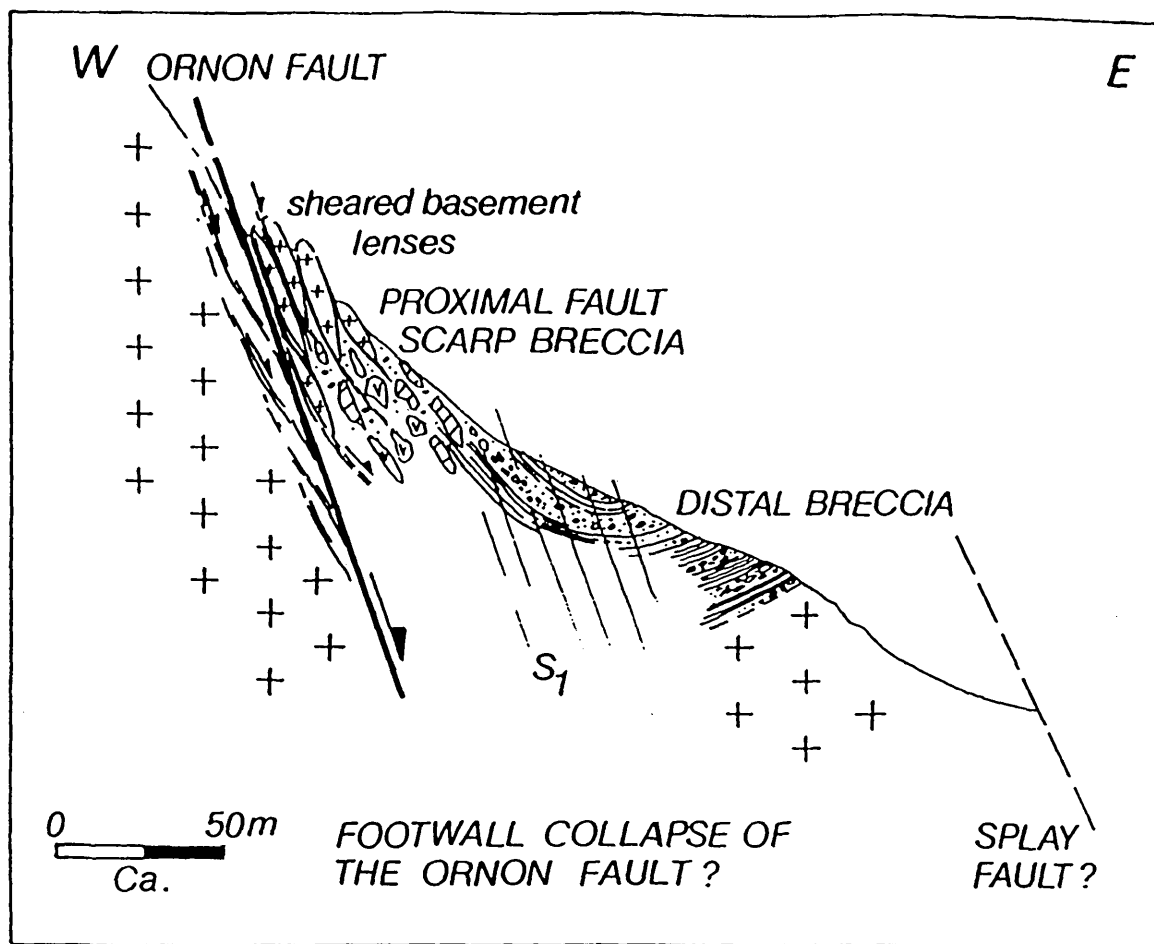


Figure 2.1.10: Field sketch of the Ornon fault at Col d'Ornon, crosses-basement, S₁-main Alpine cleavage (see also Barfety & Gidon, 1984).

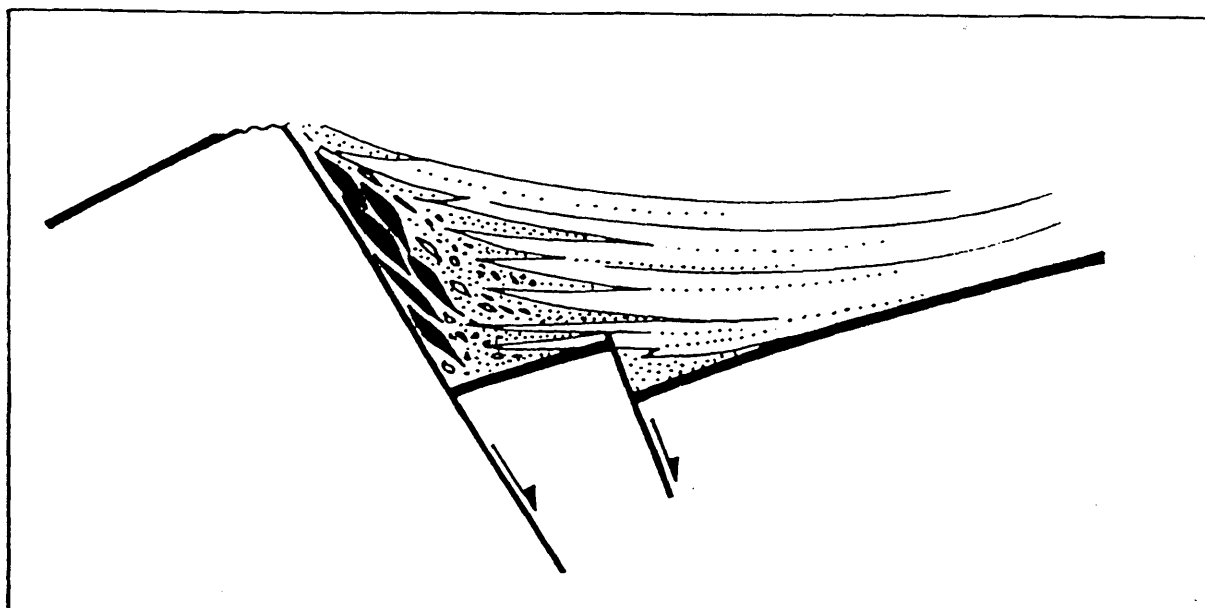


Figure 2.1.11: Suggested reconstruction of the Ornon fault and a hanging-wall splay fault at Col d'Ornon. Proximal fault scarp breccias grade E-wards into more distal breccias.

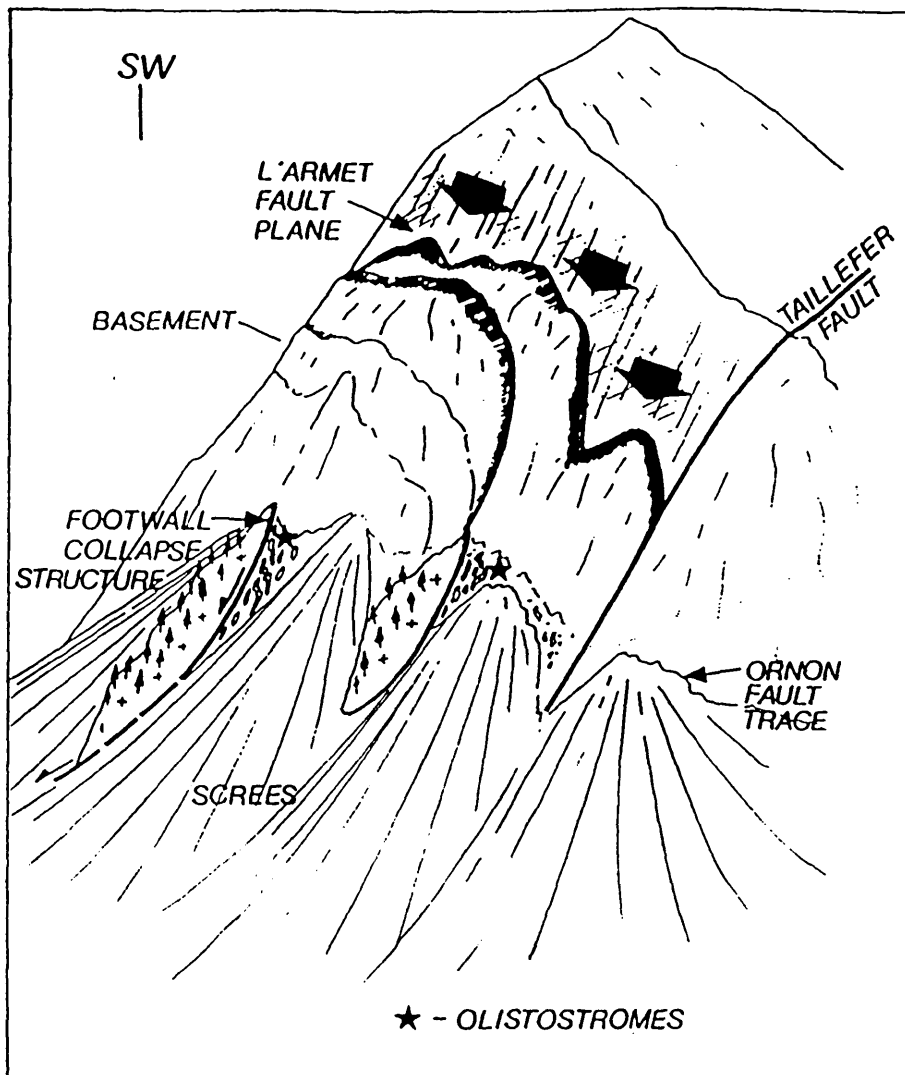


Figure 2.1.12: Field sketch of L'Armet mountain showing the Ornon and other extensional faults deforming the foot-wall basement. Note foot-wall collapse structures adjacent to the Ornon fault.

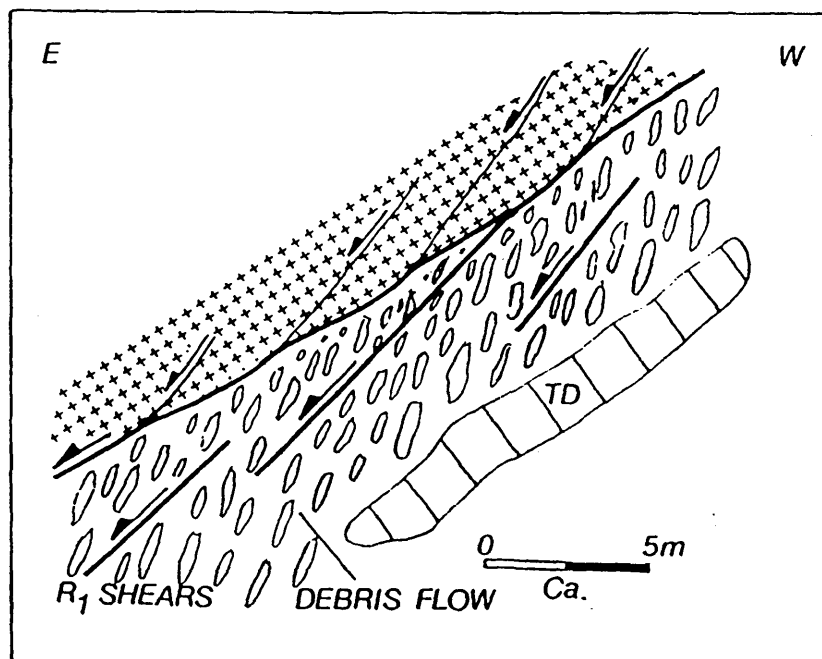


Figure 2.1.13: Field sketch of the foot-wall collapse detachment plane. Extensional R₁ shears deform the disturbed cover sequence in its foot-wall. Crosses-basement, TD-Triassic dolomites.

spillites and Liassic limestones, resembling a mass debris flow deposit. A crude shape fabric can be discerned and the larger dolomitic blocks have their long axis oriented parallel to this shape fabric (Figure 2.1.13). Pervasive extensional shears with an R1 orientation deform these cover sediments.

These basement slices are interpreted as slumped blocks detached from the foot-wall of the main Ornon fault scarp.

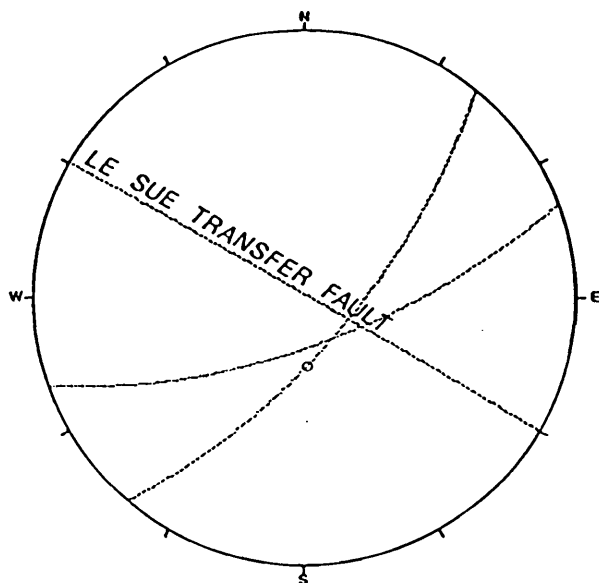
3) Les Daurens:

The Ornon fault plane is exposed to the W of Les Daurens (G.R.7329,49825) in the southern part of the Bourg d'Oisans half-graben (Figure 2.1.5). No sedimentary olistoliths of the type and dimensions previously described occur at this locality and the Upper Liassic shales are in direct contact with the Taillefer basement.

i) Fault plane description: Similar deformational fabrics as those described from the previous localities are observed. Pervasive brittle shear zones cut back into the foot-wall basement from the main fault. The main fault corresponds to a discrete break along which formed a cohesive ultra-cataclasite. This is deformed by tensional veins, some of which are fibre filled, and fibrous slickensides which are perpendicular to the tensional veins. These together with R1 shear bands in the adjacent sedimentary cover are consistent with a normal dip-slip sense of shear. Small lenses of Triassic dolomites are also sheared out parallel to the main fault plane foliation.

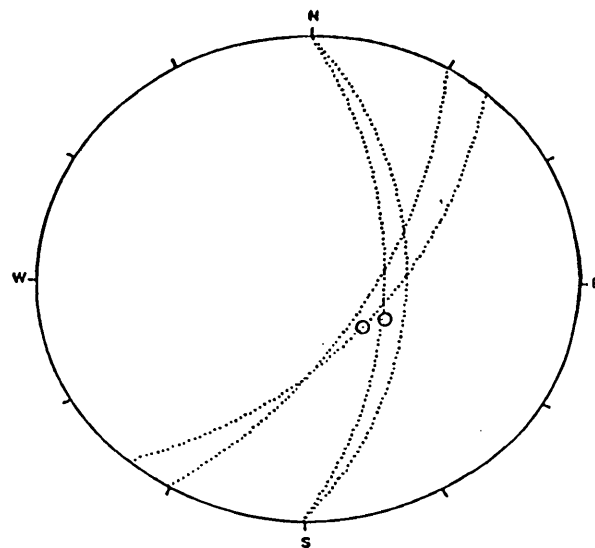
The Liassic cover sediments adjacent to the Ornon fault are deformed by a penetrative, upright Alpine cleavage which dips 72° E and strikes 020°. A cleavage refraction defines the bedding which is oriented 040°/56° E, and is otherwise difficult to recognise. The extensional shear fabrics of the Ornon fault deform this Alpine cleavage, implying that some of the normal movement post-

a. Ornon fault - Le Sue



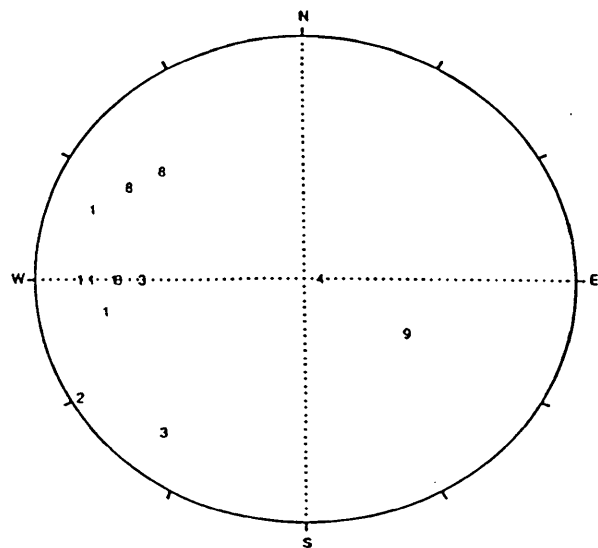
Codes	n°	%
1 .	0	0.00
2 .	0	0.00
3 .	0	0.00
4 .	0	0.00
5 .	0	0.00
6 .	0	0.00
7 .	0	0.00
8 - FAULT PLANE	3	75.00
9 - SLICKENSIDES	1	25.00
Total number of data		4

b. Ornon fault - Les Daurens



Codes	n°	%
1 .	0	0.00
2 .	0	0.00
3 .	0	0.00
4 .	0	0.00
5 .	0	0.00
6 .	0	0.00
7 .	0	0.00
8 - FAULT PLANE	4	66.67
9 - SLICKENSIDES	2	33.33
Total number of data		6

c. Ornon fault - Les Daurens (pole)



Codes	n°	%
1 - CLEAVAGE	5	38.46
2 - R1 SHEAR	1	7.69
3 - Y FOLIATION	2	15.38
4 - X - LINEATION	1	7.69
5 .	0	0.00
6 .	0	0.00
7 .	0	0.00
8 - FAULT PLANE	3	23.08
9 - SLICKENSIDES	1	7.69

Figure 2.1.14: Equal angle (Wulf net) plots of slip-lineations and fault plane orientations (a & b), and equal area (Schmidt net) plots of the structural data from the Ornon fault and adjacent cover at Les Daurens (see Fig. 2.1.5 for localities).

dates the Alpine compression. A progressive decrease in the amount of eastward dip of this Alpine cleavage away from the fault, supports this late movement interpretation - a dip change of 12° was recorded. All of the orientation data for this locality are shown stereographically in Figure 2.1.14.

2.1.3.2 Movement history of the Ornon fault:

From a study of the structural and sedimentary fabrics exposed along the length of the Ornon fault, the following movement history is proposed.

i) Jurassic movements: The existence of Toarcian age olistoliths along the length of the Ornon fault is consistent with other sedimentological criteria, arguing for Jurassic movements on the Ornon fault.

ii) Post-Jurassic movements: It is more difficult to ascertain with any degree of certainty the relative age of the shear fabrics documented from the Ornon fault. However, some of the extensional fabrics appear to post-date the main Alpine cleavage. This late normal movement is interpreted to be a consequence of gravitational instabilities induced by the thrust related uplift of the Belledonne-Tailleur massif on the external Belledonne thrust (Menard, 1979). During this uplift, and the passage of the Belledonne basement thrust sheet over a foot-wall ramp in the subsurface, extensional strains in the hanging-wall may cause the extensional reactivation of any pre-existing kinematic discontinuity such as the Ornon fault. A passive back-rotation of the Ornon fault in the hanging-wall of the external Belledonne thrust may also explain the overturning of the Ornon fault to the N of Bourg d'Oisans.

Having indicated from mainly sedimentological and stratigraphic arguments that important Jurassic extensional movements occurred on the Ornon fault, it is impossible to determine conclusively which, if any, of the extensional

fault plane fabrics can be assigned to this phase.

2.1.3.3 Four structural cross-sections:

Four E-W structural cross-sections were constructed across the Bourg d'Oisans half-graben. For each section line, detailed descriptions of the important and relevant structural relationships (i.e. the data) are presented.

1. Section A, Oulles-Mizoen cross-section:

The section line is shown on Figure 2.1.1, a simplified structural map of the Bourg d'Oisans area. As argued previously, the Ornon fault is shown with a normal dip-slip displacement of approximately 3 km. The depth of the hanging-wall pre-rift Triassic sequence was determined by projecting the exposed surface structures to deeper levels in the section. The foot-wall level of the Triassic pre-rift sequence was predicted by projecting the surface dips in the frontal part of the Belledonne -Taillefer massif and from isolated Triassic outliers exposed on its summits. However, the exact thickness of syn-rift sediments adjacent to the Ornon fault is unknown, and a substantial error in predicting the pre-rift hanging-wall cut-off may result.

i) Oulles - Lignarre valley road section: The meso- and micro-structural aspects of the cover deformation along this road section are described in detail by Vialon (1979), and Gratier & Vialon (1975; 1980). The results presented here largely accord with their findings.

The sedimentary cover adjacent to the Ornon fault at Oulles (G.R.7350,49955) is characterized by intense ductile strains, and the shortening is largely accommodated by ductile folding and penetrative cleavage formation. This slaty cleavage generally dips steeply to the E, or is vertical (average dip of 75-84°E, N-S strike), and is axial planar to tight upright N-S trending folds. These folds have been classed as similar folds (class 2) by Gratier &

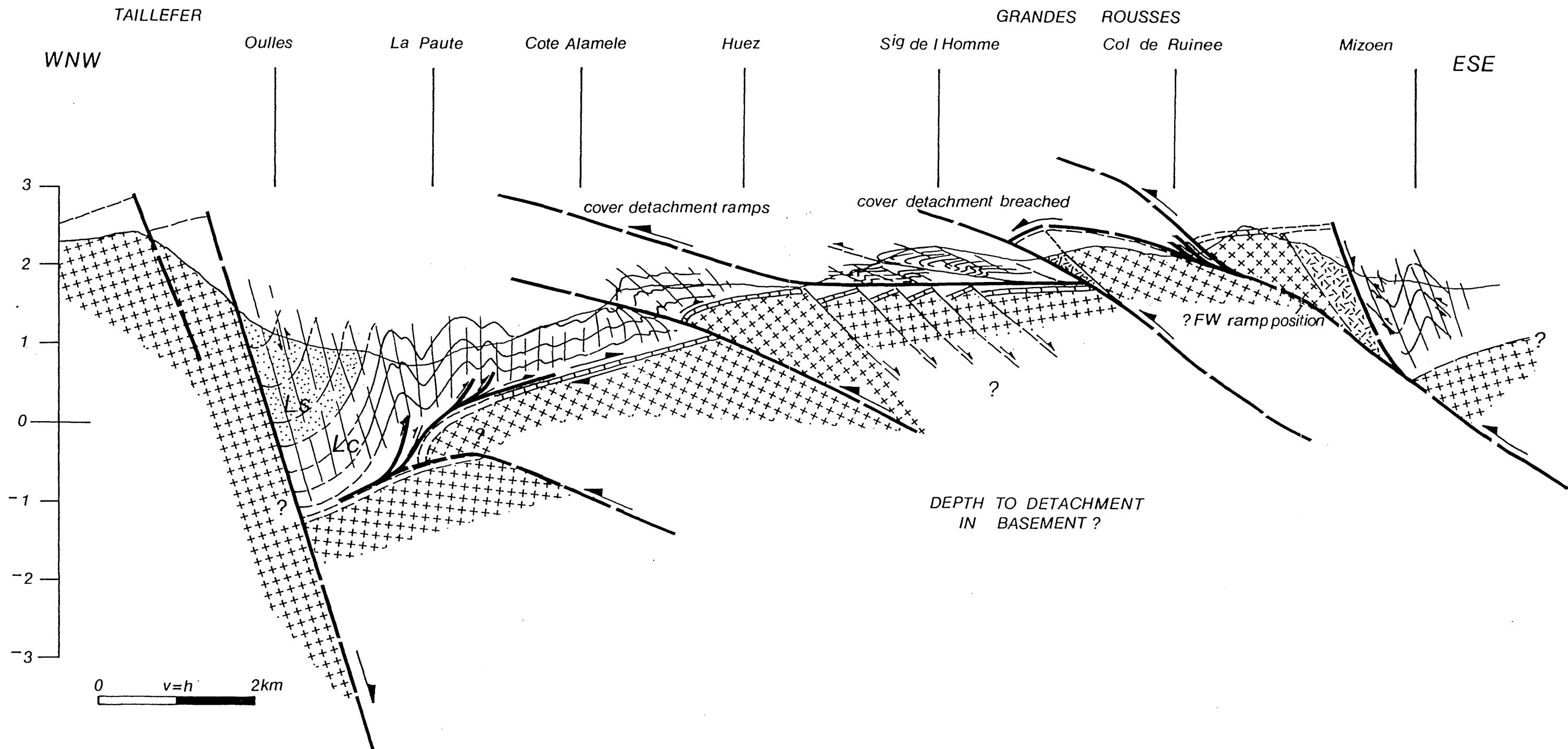


Figure 2.1.15: Oulles-Mizoen structural cross-section across the northern part of the Bourg d'Oisans half-graben (location shown as A-A' on Fig. 2.1.1), crosses-crystalline basement, speckles-Carboniferous Houiller, brickwork-Triassic dolomites and spillites, Lc-Lias calcaire, Ls-Lias schisteux.

Vialon (1980); they have sharp hinges and straight limbs, and the distance between beds measured parallel to the cleavage surface remains constant. The fold limbs show a strong vertical attenuation, best observed in sections perpendicular to the cleavage plane (XZ plane). A strong refraction of the cleavage occurs where it passes from competent limestone beds into less competent shales along the fold limbs. This arrangement of the cleavage, which subtends a low-angle with respect to the bedding, results in the development of micro-lithons. Continued shearing across these sub-vertical discontinuities appears to have accommodated continued horizontal contraction, and results in a strong flattening and modification of the fold shape in the XZ plane.

Two linear fabrics are observed on the cleavage surface. The first of these is a bedding-cleavage intersection lineation (S₀/S₁) which defines minor fold axes in the Oulles section and is often curved showing pronounced N-S plunge variations. The other lineation is a steeply plunging to vertical down-dip stretching lineation, manifest as a preferred mineral orientation and by the long axes of deformed and stretched objects such as belemnites and sponge spicules. In some cases stretched belemnites record a vertical extension exceeding 700 %. The X-direction of finite extension is also defined by calcite fibre growths in pressure shadows developed adjacent to framboidal or cuboidal pyrite bodies. These fibre growths record the incremental strain history (Durney & Ramsay, 1972) and their curved form suggests an anti-clockwise rotation occurred during the progressive deformation. Some pressure shadows show that extension occurred parallel to the Y direction of the XY deformation plane. The bedding and S₁ cleavage in the steeply dipping fold limbs are also transected by horizontal fracture planes infilled with quartz and calcite fibre growths, the long axis of which parallel the finite extension direction. These also indicate that an important vertical extension and horizontal contraction affected the cover sediments adjacent to the Ornon fault plane. All of the geometrical

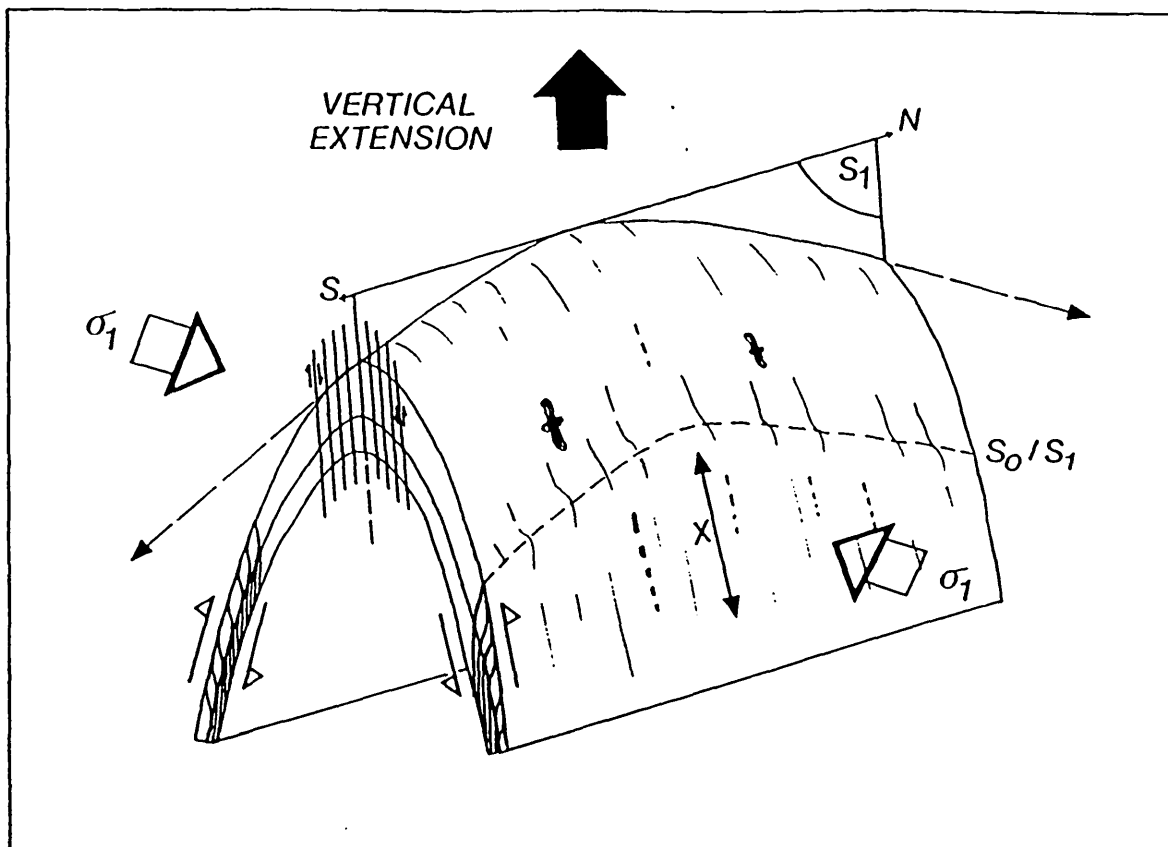


Figure 2.1.16: Simplified geometry of tightened folds adjacent to the Ornon fault at Oulles, showing the principal linear and planar structural fabrics, X-extension direction, S0/S1-cleavage/bedding intersection lineation, S1-main Alpine cleavage. Modified after Vialon (1979).

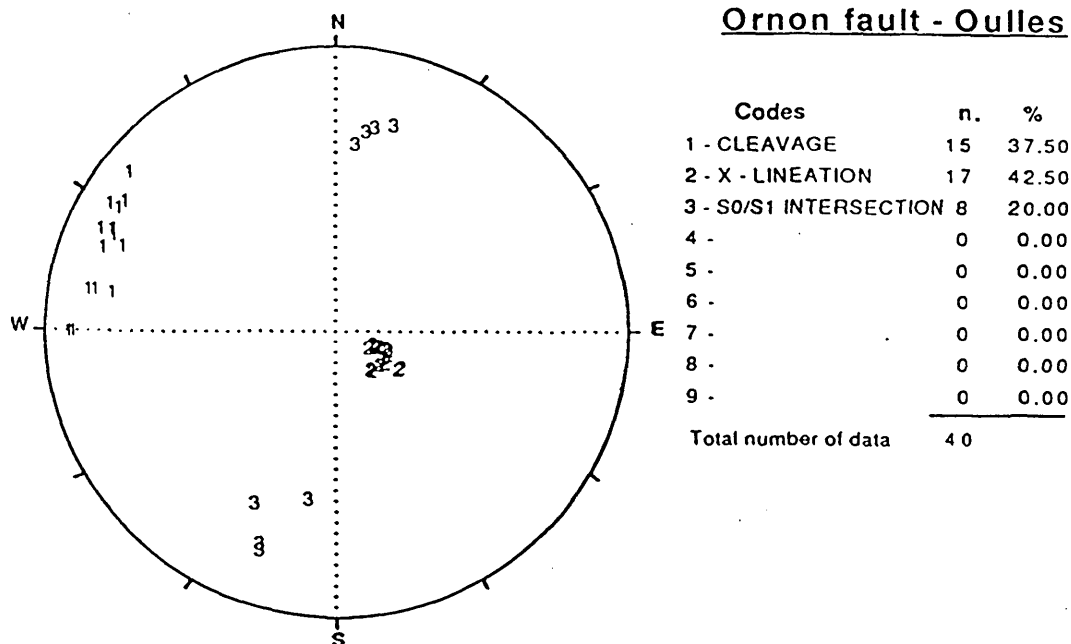


Figure 2.1.17: Equal area (Schmidt net) plot of the principal planar (poles) and linear fabrics in the sedimentary cover adjacent to the Ornon fault. Note plunge variations of fold axes defined by S0/S1 intersection lineation.

relations between the fold structure and the linear and planar fabrics are shown in Figures 2.1.16 and 2.1.17.

ii) La Paute - Cote Alamele: The tightly folded and highly strained zone of Oulles passes eastwards into a zone where the structural style is dominated by more open, longer wavelength and lower amplitude folds, exposed on both sides of the NW-SE trending Romanche valley.

At La Paute (G.R.7375,49955), the folds are tighter than those exposed on the Cote Alamele (G.R.7392,49962) to the NE (Figure 2.1.18). The La Paute folds show an eastward vergence, and the E-facing limbs are commonly deformed by minor eastward directed thrusts. The average axial trend is 020-200° and the cleavage vertical. The slaty cleavage is still penetrative, though not as penetrative as was observed at Oulles.

The structures used to continue this section line are exposed in the 700 m vertical cliff exposure of Cote Alamele (Figure 2.1.19). The wavelength of the folds increases eastwards and the cleavage becomes less penetrative. The folds are asymmetric towards the E and are commonly deformed by minor bedding-parallel thrust detachments, suggesting that they may have been initiated by a flexural-slip mechanism of folding. They are only marginally modified by the axial planar slaty cleavage which is vertical to steeply westward dipping. The fold axes still trend N-S.

Along the Oulles - La Paute - Cote Alamele traverse, a subtle fanning of the main upright slaty cleavage is observed. A steeply westward dipping cleavage which parallels the Ornon fault dip in the Oulles section, passes eastwards into a vertical to marginally westward dipping cleavage in the La Paute-Cote Alamele section.

At the eastern end of the Cote Alamele section, towards the D211 Alpe d'Huez road (G.R.7399,49950), the bedding progressively steepens assuming a westward dip.

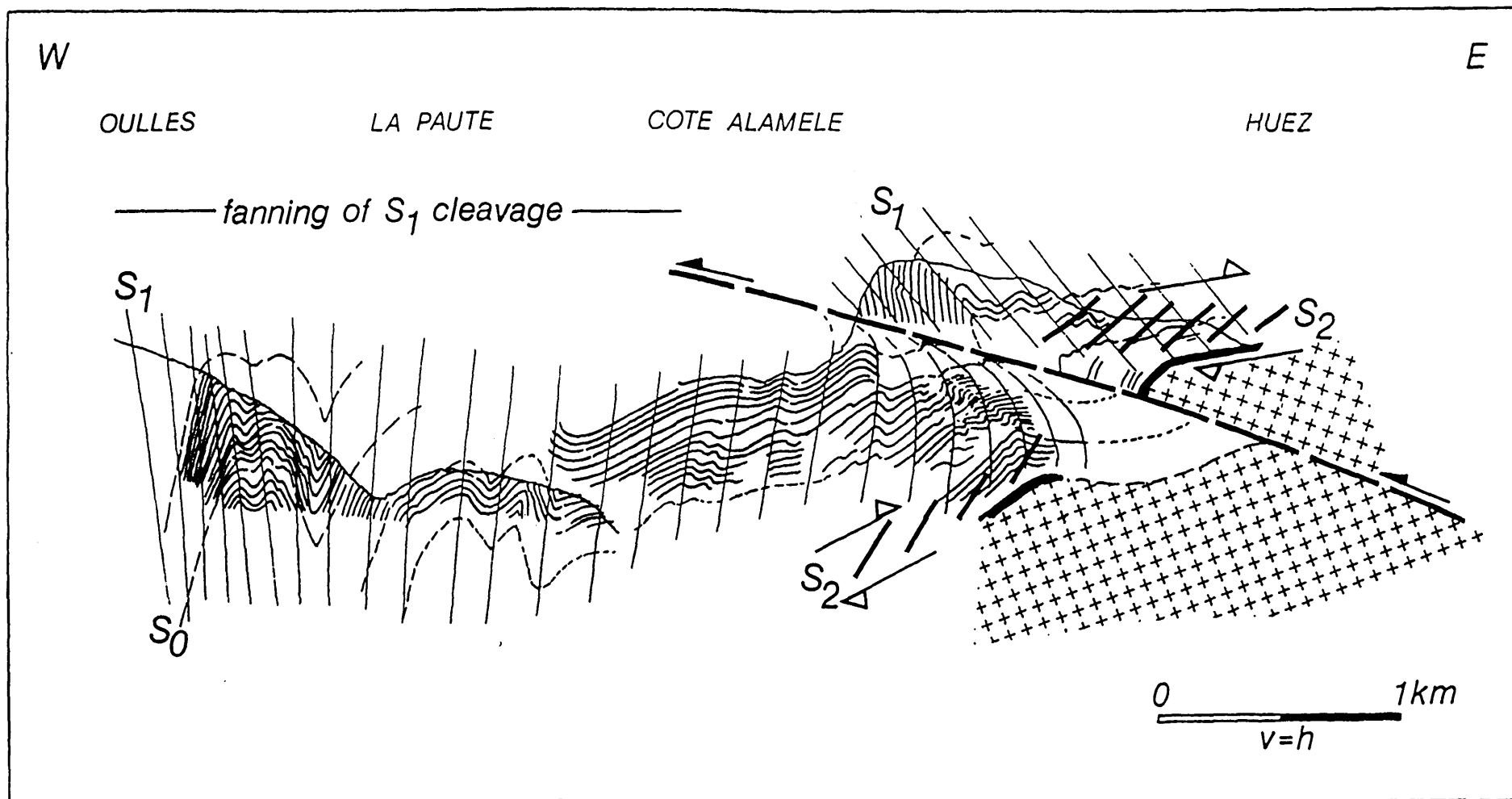


Figure 2.1.18: Structural cross-section showing the change in fold-style from Oulles to Cote Alamele, and the Huez basement-involving thrust which truncates and deforms earlier structures in its foot-wall. The main S_1 cleavage is locally deformed by a back-thrust cleavage (S_2) close to the basement-cover interface. For location see Fig. 2.1.15, crosses-basement, bedding in Liassic cover is indicated (modified after Gratier et al., 1973).

Associated with this is a progressive anti-clockwise rotation of the main upright cleavage so that it becomes eastward dipping, and the eastward verging back-folds previously described are refolded by asymmetric westward verging folds, generating a minor fold-facing confrontation or thrust triangle zone (Figure 2.1.19).

On the Alpe D'Huez road, where it crosses the basement-cover contact (G.R.7401,49950), a fairly low-angle of intersection between the main cleavage (S1) and bedding in the less competent shale units is observed. This may indicate a change in structural style along this section line, from the similar type folds formed adjacent to the Ornon fault, to concentric or flexural-slip folds formed farther to the E, the latter strongly suggesting that the half-graben sedimentary cover is decoupled from its underlying basement. At this locality, the S1 cleavage is also deformed and crenulated by a widely spaced fracture cleavage (020/70° NW). This second cleavage (S2) is most penetrative in the incompetent shale horizons, and implies late eastward directed bedding-parallel shear along the lower stratigraphic levels close to the basement-cover interface.

The fold-facing confrontation and westward shearing of earlier folds and existing upright cleavage, are interpreted to have formed in the foot-wall to a basement overthrust, all of the earlier structures progressively rotating into parallelism with the overthrust shear direction. This basement thrust is exposed in the Ravine de Sarenne (G.R.7400,49951), below the villages of Le Chatelard and Huez, and is called the "Huez" thrust (Gratier et al., 1973; Gratier & Vialon, 1980). At higher structural levels the thrust climbs upwards through the stratigraphic section slicing through the earlier eastward verging back-folds, and deformation in its foot-wall generates minor westward verging folds. In the upper levels of the Cote Alamele, where the thrust intersects the topography, a westward-verging hanging-wall fold is observed (Figure 2.1.18).

WNW

COTE ALAMELE

ESE

68

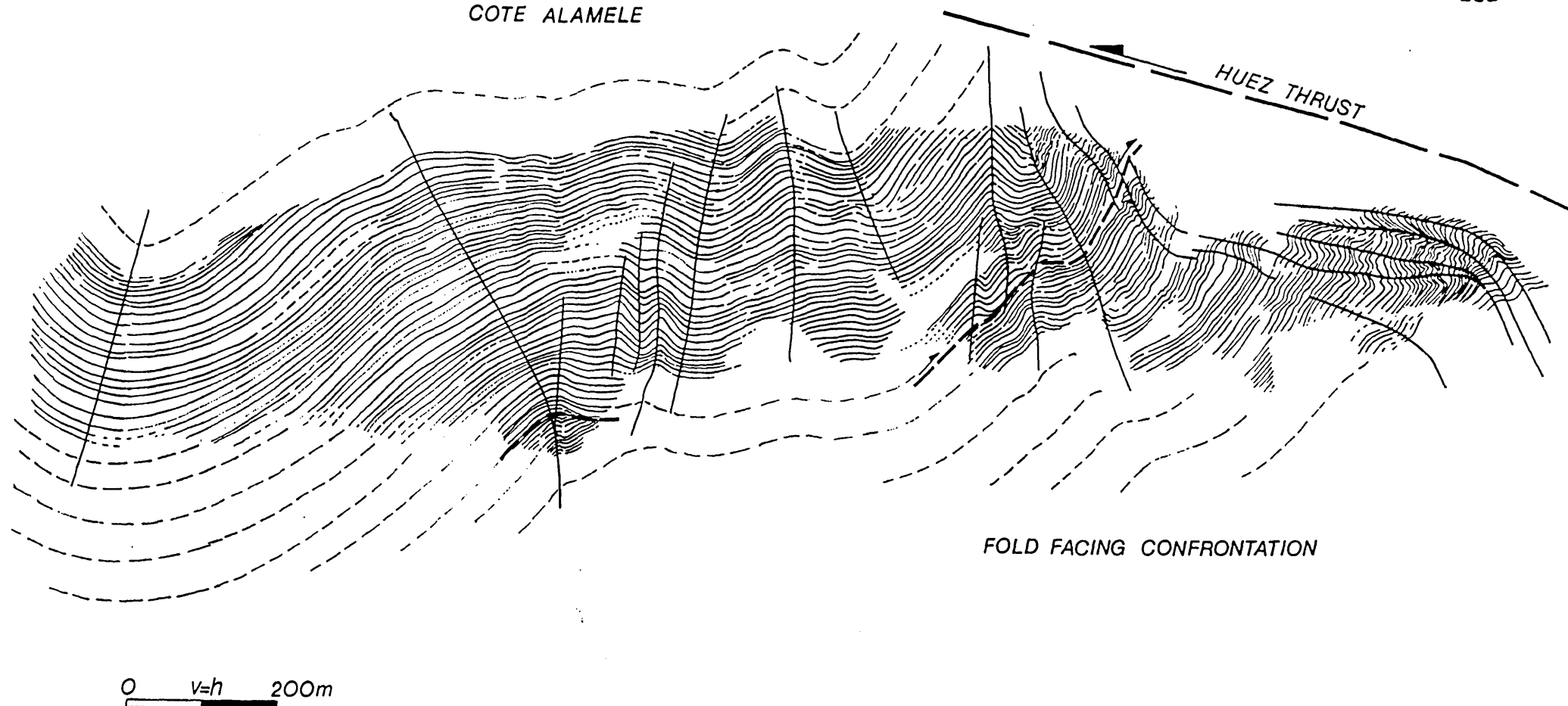


Figure 2.1.19: Field sketch across Cote Alamele showing incipient stages of back-fold-thrust system, and fold-facing confrontation (or thrust triangle zone) in the Lias calcaire, which results from the later emplacement of the Huez thrust. For location refer to main section line (Fig. 2.1.15).

The D211b road from Huez to Villard Reculas (G.R.7405,49963) transects the hanging-wall of the Huez thrust permitting a comparison of the deformation fabrics exposed in the hanging-wall and foot-wall of the thrust to be made. At a level close to the basement-cover contact, the S1 cleavage and bedding are deformed by a pervasive S2 fracture cleavage. This S2 fracture cleavage is localized around minor eastward-directed thrust detachments (back-thrusts with respect to the dominant westward tectonic transport direction). One such back-thrust, exposed on the Villard Reculas road section, is sketched in Figure 2.1.20.

This thrust fault (Figure 2.1.20) corresponds to a discrete brittle failure plane of listric shape which slices through earlier fabrics. Both the bedding and the earlier S1 fabric are refolded, forming an eastward-verging hanging-wall anticline and minor foot-wall folds. The S2 fracture cleavage is axial planar to these folds. The intensity of the back-thrust cleavage is variable, increasing towards the thrust fault. The smooth listric trajectory of the thrust fault, the ductile folding and fabric development which thickens the more incompetent beds in the immediate foot-wall and the intense hanging-wall cleavage, all suggest that ductile deformation in the form of folding and cleavage development preceeded the propagation of the thrust fault (Coward & Kim, 1981; Cooper & Trayner, 1986; Williams & Chapman, 1983).

At a higher structural level, towards Villard Reculas (G.R.7395,49972), a progressive diminishment in the intensity of the S2 fabric occurs. At Villard Reculas the average S1 cleavage dip is between 40-60° E, striking 170°, and minor folds display a westward vergence. The westward shearing of the S1 cleavage and the formation of the westward verging folds, may have been generated in the foot-wall to a higher level and partially out-of-sequence cover thrust ramp which slices through earlier folds in the sedimentary succession to the E of Alpe d'Huez (see Figure 2.1.15).

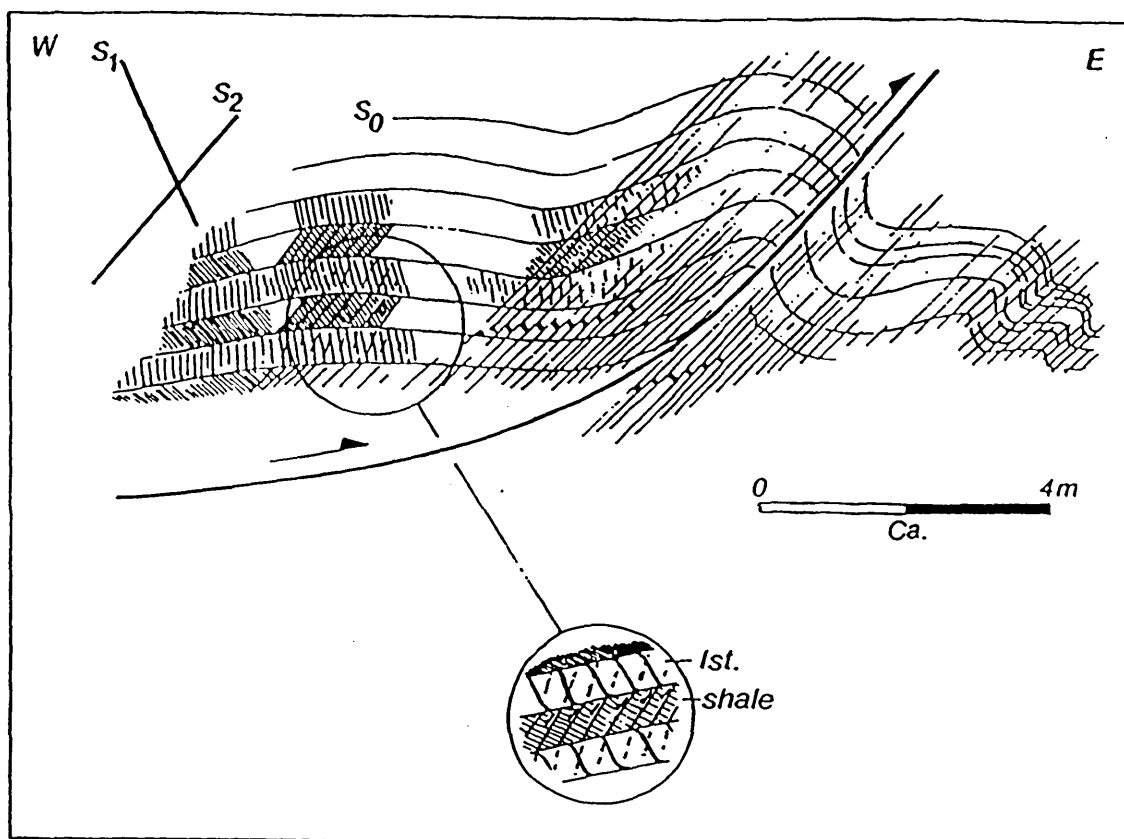


Figure 2.1.20: Field sketch of pervasive back-thrust cleavage (S2) in the Lias calcaire, from the Villard Reculas road section. See text for further details.

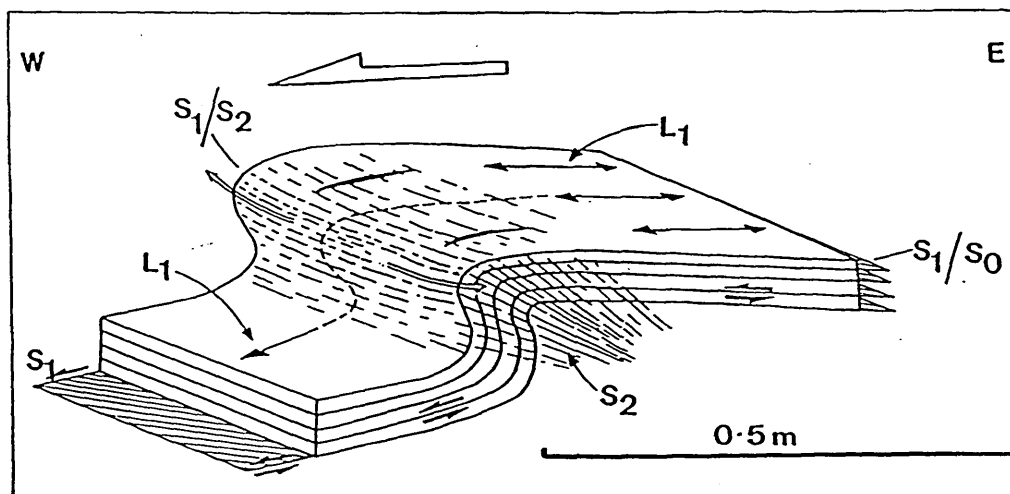


Figure 2.1.21: Geometry of secondary shear folds from the Liassic cover detachment zone exposed at Armentier. An early bedding-parallel fabric (S1) is refolded and crenulated. S0-bedding trace, S1-early fabric, S2-crenulation cleavage, L1-folded lineation. Arrow denotes sense of shear operative along detachment.

In terms of displacement, the back-thrust shear concentrated close to the base of the cover succession in the hanging-wall of the Huez thrust, is minor. From the disposition and cross-cutting relationships of the different structural fabrics, it is also clear that movements on the Huez thrust post-date an earlier folding of the cover sediments, and that other superimposed structural fabrics reflect movements on other structurally lower basement thrust sheets. In this case the Huez basement thrust breaches a higher level detachment on which the cover sediments are thought to have decoupled. The back-thrust sense of displacement is interpreted to result from the driving of a rigid basement wedge into the surrounding softer sedimentary cover (see Figure 2.1.6, Section 2.1.4.2). The estimated displacement on the Huez basement thrust is in the order of 600 m. Similarly the excess area encountered in the western part of the section (to the W of the Huez thrust) may be resolved at depth by a similar basement thrust sheet. Such a basement thrust wedge is also able to explain the back-thrust shear located at the base of the cover succession in the foot-wall to the Huez thrust (Figure 2.1.15).

iii) Montagne de l'Homme - Rochers d'Armentier: The section line is continued across the poor exposure of the Montagne de l'Homme (G.R. 7435,49952) (Figure 2.1.1). This absence of data is compensated by a projection of the structures exposed on the Rochers d'Armentiers and Cote du Seignet (G.R.7420,49940).

At the Rochers d'Armentier, there is a detachment of the Jurassic cover succession from its underlying basement. Unlike the Ultra-Dauphinois thrust system to the N, where the Mesozoic cover preferentially decouples along a Triassic gypsum easy slip horizon (Beach, 1981a), here the cover detaches along the first Lower Liassic shale horizons and the Triassic sediments remain fixed to their basement. The basal gypsum horizon observed to the N, with minor exceptions, is largely missing from the Pelvoux sequence. This absence and the local changeability of the Lower

Liassic detachment horizon inhibits the movement of the sedimentary cover across its basement, and a penetrative ductile shear strain imparted on the lower levels of the cover thrust sheets results (see Beach, 1981d).

This cover detachment is characterized by an intense ductile strain, with brittle faulting being subordinate. Early ductile deformation along the detachment resulted in the formation of a penetrative bedding-parallel cleavage with a pronounced mineral stretching lineation, the latter implying that an important component of bedding-parallel slip occurred. This stretching lineation trends ENE-WSW and is considered to be representative of cover thrust transport direction. Early bedding-parallel quartz and calcite veins are thinned, boudinaged and in places mylonitized implying a modification by bedding-parallel or cleavage-parallel extension and shear. These bedding-parallel veins are perhaps indicative of high pore fluid pressures during the incipient stages of detachment. Beach (1981d) and Beach & Jack (1982), document similar structures in the Ultra-Dauphinois thrust system to the E, and suggest a mode of initiation by hydraulic fracturing, where fluid pressures jacked up the thrust sheet overburden during the early thrust sheet movements (Price, 1977).

The early bedding-parallel fabrics and mineral stretching lineation are folded by minor secondary folds (Figure 2.1.21), producing a crenulation cleavage. The axial directions and asymmetry of these folds are consistent with a westward overthrust shear sense, although in the plane of the cleavage the hinge lines are often curvilinear. Folds similar to these are described by Gratier & Vialon (1980) from the Mont Blanc area, and the Mizoën area in the eastern part of this section line. Beach (1981d; 1982) also describes similar structures from the La Grave area along the northern margin of the Pelvoux massif. They are interpreted to have formed during a progressive non-coaxial deformation where the shear is concentrated parallel to the main planar anisotropy, here the stratification or cleavage.

At higher structural levels above the main cover detachment, the deformation becomes more heterogeneous and the Liassic sediments are deformed by asymmetric, and in places recumbent westward-verging folds and brittle thrusts. These folds have sub-horizontal axial planes and gently plunging hinge lines which are curved in map plan. The cutting up section of the thrusts towards the WNW, and the westward vergence of folds is consistent with a WNW thrust transport direction. The main cleavage is axial planar to the folds, which are interpreted to have formed during bedding-parallel simple shear, with buckling and eventual thrusting. The cover may have become detached in the foot-wall to a basement thrust farther to the E (Figure 2.1.15).

iv) Rochers d'Armentier half-grabens: N of Bourg d'Oisans on the Rochers d'Armentier (G.R.7415,49930), the basement-Triassic unconformity is offset by late Triassic to early Liassic faults, which downthrow the unconformity 20-30m to the E. In the hanging-walls of these faults the unconformity dips into the fault plane at an angle of 30° with little in the way of roll-over fold, suggesting that the faults were of the planar domino-type. In one of the half-grabens the hanging-wall dips are modified by minor folding which may be of Alpine age. The foliation of the Hercynian basement into which the faults pass downwards, is deformed by a number of minor extensional brittle shears (Figure 2.1.22).

The sedimentary infill is well exposed in only one of these half-grabens, the other infills being concealed by recent landslips and screes. In the well exposed example, the main basement fault displays an unusual shallowing upwards profile. Immediately adjacent to the basement fault scarp are sheared lenses of basement enveloped by the surrounding sedimentary cover. It is suggested that the shallowing upwards fault profile results from a foot-wall collapse of the basement fault scarp (Figure 2.1.23). A shear fabric was observed along the fault plane, and R1 shears which offset the main foliation (Plate 2.1.3)

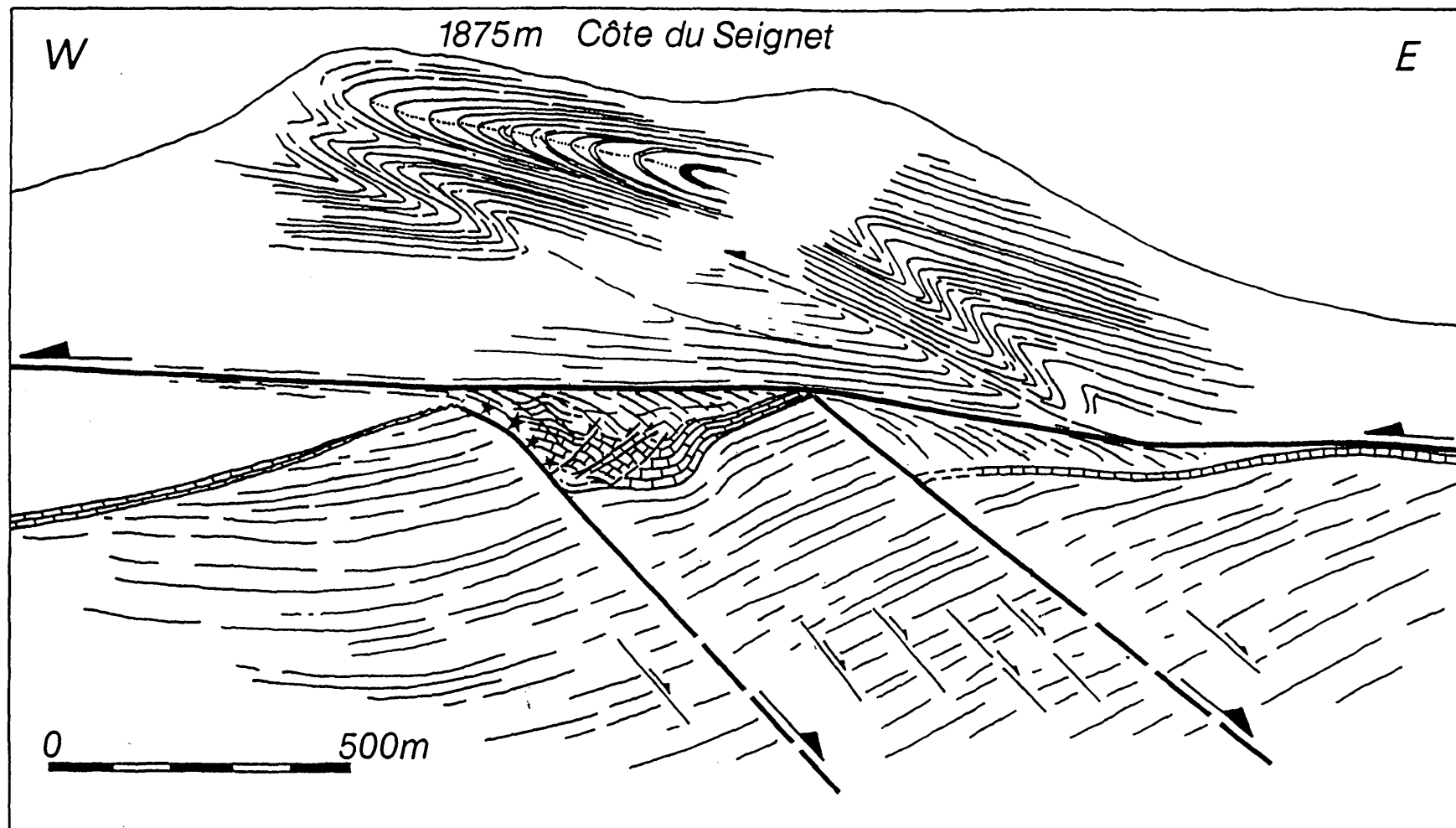
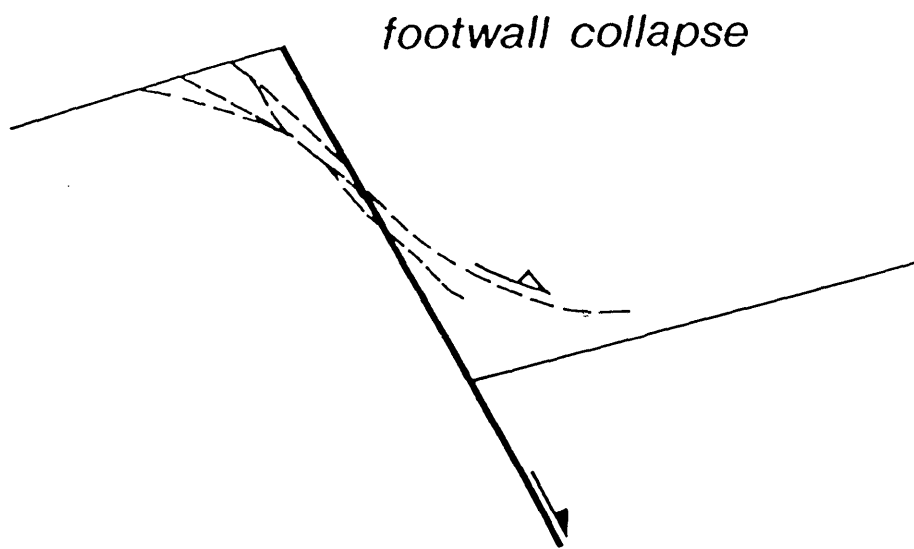


Figure 2.1.22: Exposed cross-section of Rochers d'Armentier half-grabens (locality shown on Figs. 2.1.1 & 2.1.15). Fine lines show bedding traces, heavy lines are faults. Brickwork indicates Upper Triassic rocks which thicken into the half-graben, but thin and are eroded on the foot-walls. The Liassic cover is isoclinally folded above the main thrust detachment. Syn-sedimentary breccias are indicated by stars, and the main Variscan basement foliation is shown.

A.



B.

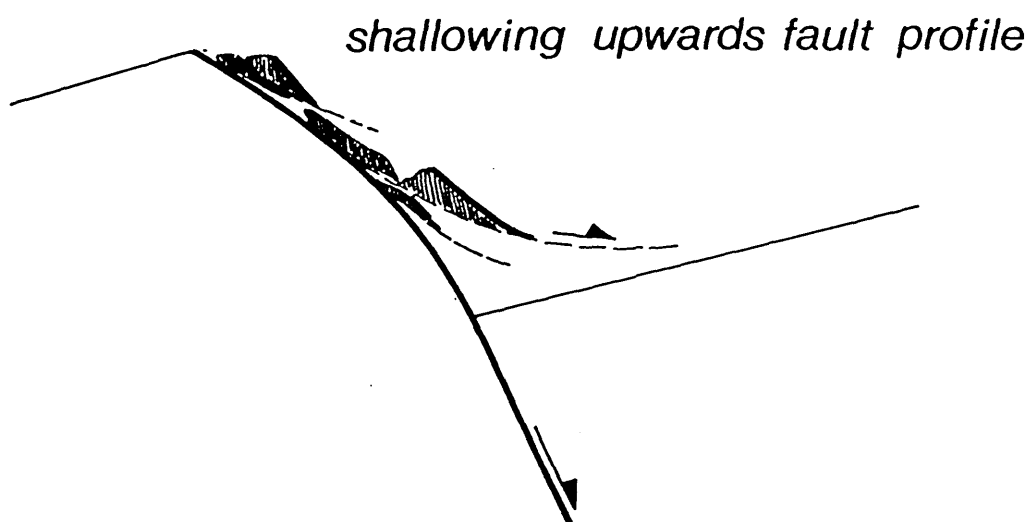


Figure 2.1.23: Schematic diagram to illustrate the formation of a shallowing upwards fault profile, by collapsing the foot-wall scarp (C.f. Rochers d'Armentier).



Plate 2.1.3: Extensional shear fabric from Liassic age normal fault at Rochers d'Armentier. The camera lens cap is 5.5 cm in diameter. The main fabric is crenulated by R1 shears (arrow).



Plate 2.1.4: Viewing S across the Romanche valley towards the Pregentil anticline with Bourg d'Oisans in the bottom left of picture. Note the "parapente" in the centre of picture for scale.

indicate a normal sense of movement. As this fabric does not affect the overlying Liassic cover sediments it is probably of Jurassic age, and not Alpine. In the immediate hanging-wall of the exposed normal fault is a disrupted sedimentary fabric, interpreted by Grand et al. (in press) to have been derived from the eroded foot-wall of the basement fault scarp. However, alternative tectonic explanations of this disrupted sedimentary fabric may also be forwarded.

This zone of disrupted sediments may correspond to a broad zone of ductile extensional strain, formed as the hanging-wall sediments are dragged against the resistant foot-wall of basement. The bedded units in the hanging-wall become smeared out parallel to the fault plane in the slip-direction. Evidence supporting this hypothesis is the shape fabric of the disrupted beds, where lenses of Triassic dolomites have their long axis parallel to the slip-direction on the eastward dipping fault plane. This suggests that during fault movements, individual beds suffered a bedding-parallel extension leading to the development of pinch-and-swell structures and where rupturing occurs, isolated tectonic lenses.

This interpretation provides an alternative explanation as to how these Jurassic fault scarp breccias are formed. Many probably result from the combined interaction of ductile extension adjacent to the fault scarp, and erosion of the adjacent foot-wall.

This zone of ductile extensional strain passes eastwards into a fault scarp breccia comprising mainly dolomitic clasts. A crude stratification, which progrades into bedded Triassic dolomites to the E, can be discerned.

The Triassic infill of the half-graben is deformed by a number of minor planar normal faults, antithetic to the master fault (Figure 2.1.22). The high bedding cut-off angles and low fault plane dips imply that the amount of extension is locally high, and is perhaps not compatible

with that on the main fault which passes downwards into the basement. The Triassic bedding is folded into an open E-W trending syncline, which as previously suggested is probably of Alpine age. The Lower Liassic sediments overlap, and are not deformed by the planar faults affecting the Triassic sediments. Higher up in the half-graben sequence, the Lower Liassic sediments progressively onlap the main fault. These basement faults do not affect the overlying Alpine cover thrust detachment, further substantiating a Mesozoic age interpretation.

The sedimentary infill previously described shows remarkably little in the way of internal deformation. In the upper part of the sedimentary infill, an eastward-dipping spaced cleavage devoid of any mineral stretching lineation, intensifies upwards and asymptotes into the overlying flat detachment of the sedimentary cover (Figure 2.1.22). This cleavage within the half-graben is axial planar to minor folds which buckle the sedimentary infill, and which verge toward the W, indicating that they were generated in the foot-wall to the overriding cover thrust sheet.

These documented relationships illustrate the strong lithostratigraphic control over the development of the cover thrust detachments, which exploit the uniform Liassic shale horizons unaffected by the late Triassic to early Liassic extensional faults.

A similar minor buckling of the cover sediments beneath the principal detachment is observed to the E of the Rochers d'Armentier half-grabens, where upright to westward-verging folds and their axial planar cleavage become sheared parallel to the main fabric of the detachment zone. Further E, this detachment steps down in the stratigraphic section to incorporate Triassic spillites and dolomites in the thrusting.

v) **Grandes Rousses basement (Herpie thrust):** Continuing the section line to the E (Figure 2.1.15), Carboniferous

Houiller sediments of the Grandes Rousses basement overthrust Liassic cover sediments of the Montagne de l'Homme on the Herpie thrust (G.R.7445,49935). A thrust sliver of Triassic spillites and dolomites lies in the immediate foot-wall to this thrust. The Herpie thrust can be traced northwards within the Grandes Rousses massif by a thin but laterally continuous band of Houiller sediments located on its hanging-wall (Figure 2.2.5). Mesozoic age extensional faulting documented in the Lac du Besson area (G.R.7430,50010), N of Alpe d'Huez (Grand, 1987), lies in the foot-wall of this thrust. To the S, this thin strip of Houiller sediments continues into the Deux Alpes area where it parallels the Pied Moutet fault, interpreted to be a Jurassic age fault (see Section 2.2). The Herpie thrust, which carries the Grandes Rousses basement in its hanging-wall, is interpreted to breach the basement-cover thrust detachment (Figure 2.1.15).

vi) Internal Grandes Rousses thrust: The displacements associated with the basement-cover thrust detachment which ramps in the Alpe d'Huez region (the partially out-of-sequence thrust ramp), are interpreted to have been partly generated in the foot-wall to the Internal Grandes Rousses basement thrust (Beach, 1981a), which is well exposed to the E of le Puy-Dessous (G.R.2730,49930). This thrust fault, like the Herpie thrust, trends N-S along the length of the Grandes Rousses massif (Figure 2.2.5). To the N, at the Col de Ruinee (G.R.2732,49952) and the Croix de Cassini (G.R.2740,49952) (Figure 2.2.5), imbricated Liassic and Triassic cover sediments lie in the immediate foot-wall of the basement thrust. Evidence for Jurassic age faulting has been documented by Rudkiewicz & Vibert (1982). To the N, at deeper structural levels, the Internal Grandes Rousses basement overthrusts a band of Carboniferous Houiller sediments of the external Grandes Rousses (BRMG La Grave 1:50 000 geology map, 1976). At le Puy-Dessous where the Internal Grandes Rousses thrust is exposed, the thrust relationships are more complex. A sketch section illustrates the contractional geometry exposed (Figure 2.1.24). In the W the Triassic-basement contact is

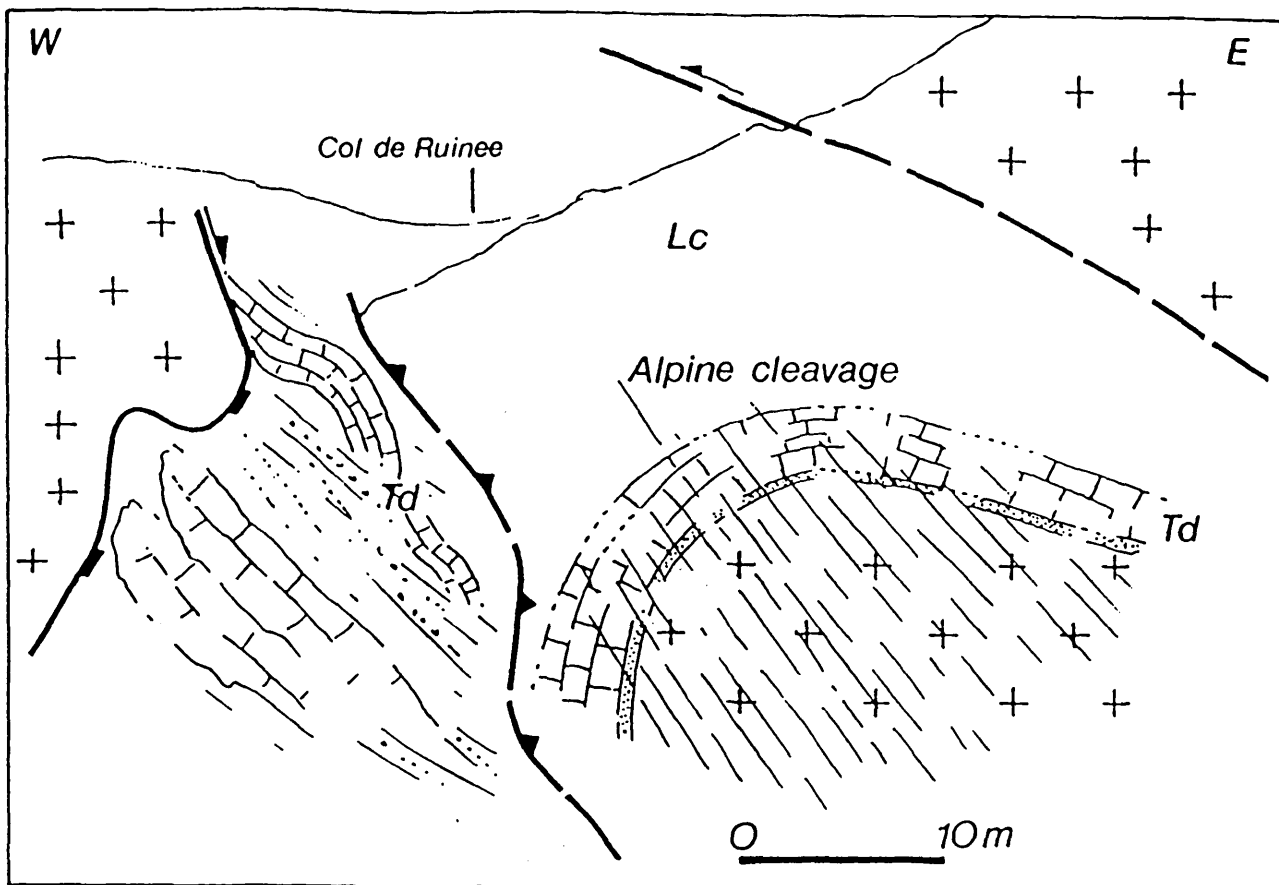


Figure 2.1.24: Field sketch of tight basement anticline formed in the foot-wall to the Internal Grandes Rousses thrust at Le Puy Dessous (see Fig. 2.1.1). This may have formed by the attempted reactivation of a Jurassic normal fault, one such fault being shown in left of diagram. Crosses-basement, Td-Triassic dolomites, Lc-Lias calcaire, main Alpine cleavage also indicated.

interpreted as a normal fault and an extensional shear fabric is developed in the basement along the contact. The fault plane dips at 68° E, striking N-S, and may or may not correspond to a Jurassic age normal fault. To the E, in the foot-wall to the Internal Grandes Rousses thrust, the basement and Triassic dolomites are folded into a tight westward-verging anticline, and the basement is deformed by a penetrative Alpine fabric which affects the overlying Triassic rocks. This basement fabric dips 38° E, striking 160° . The basement anticline trends N-S and plunges gently to the S. The Liassic cover sediments in the immediate foot-wall to the basement thrust contain a penetrative cleavage which dips 32° E, striking 140° . A mineral stretching lineation (plunging 26° towards 072°) gives the tectonic transport direction. The Triassic and Liassic sediments outcrop again to the E as a small outlier on the Croix de Cassini, and define the hanging-wall geometry of the Internal Grandes Rousses thrust sheet. The hanging-wall comprises a broad strip of Carboniferous Houiller sediments and basement gneisses (Figure 2.2.5). To the S the displacements associated with this thrust pass into the Deux Alpes area.

vii) Eastern Grandes Rousses - Mizoen: The eastern margin of the Grandes Rousses massif is limited by a steeply eastward dipping fault called the Eastern Grandes Rousses fault (Rudkiewicz & Vibert, 1982) (Figure 2.2.5). This fault has a present day normal offset and the presence of Triassic and basement olistoliths in its hanging-wall, lead Rudkiewicz & Vibert (1982) to propose a Jurassic age movement history.

The Jurassic cover sediments in the hanging-wall of this fault define a syncline, with Middle Jurassic sediments in its core. The Alpine deformation of the cover sediments adjacent to the Eastern Grandes Rousses fault is intense. N of the Lac du Chambon, between Mizoen and Singuineret (G.R.2750,49927), a penetrative S1 cleavage which intersects the bedding at a low-angle, is folded and crenulated by asymmetric westward verging shear folds, more

common in the less competent shale horizons. These fold early bedding-parallel calcite and quartz veins. The intensity of the deformation in this part of the section may result from a buttressing of the Alpine cover thrust displacements by the Eastern Grandes Rousses fault. The southward continuation of this structure into the Deux Alpes area is discussed in Section 2.2.

2) Section B, the Grand Renaud cross-section:

The location of the WNW-ESE section line, constructed parallel to the tectonic movement direction, is shown on Figure 2.1.1. The cross-section is shown in Figure 2.1.25.

i) Ornon fault: Where the section line intersects the Ornon fault S of the village, Le Rivier (G.R.7343,49905), the quality of exposure is poor and the structures described from Oulles are used to complete this part of the section.

Adjacent to the Ornon fault is a stratigraphically defined N-S trending syncline, with Upper Liassic (Toarcian and Domerian shales) exposed its core. This structure is continuous along the length of the Ornon fault.

ii) Fold styles in the Lias calcaire: To the immediate E of the Ornon fault, folds affecting the Lias calcaire are exceptionally well exposed on both sides of the Villard Raymond valley, along two natural WNW-ESE cliff-sections and along the NE-facing side of the Romanche valley, S of Bourg d'Oisans.

The most prominent structure is the Pregentil anticline (G.R.7370,49922) (Plate 2.1.4), which trends N-S and plunges gently to the N. This fold has a wavelength in the order of 1.5-2.0 km and its amplitude exceeds 1.5 km. The fold limbs are deformed by minor secondary folds. The cleavage is axial planar and vertical, striking 040°, and displays a weak vertically plunging mineral stretching lineation.

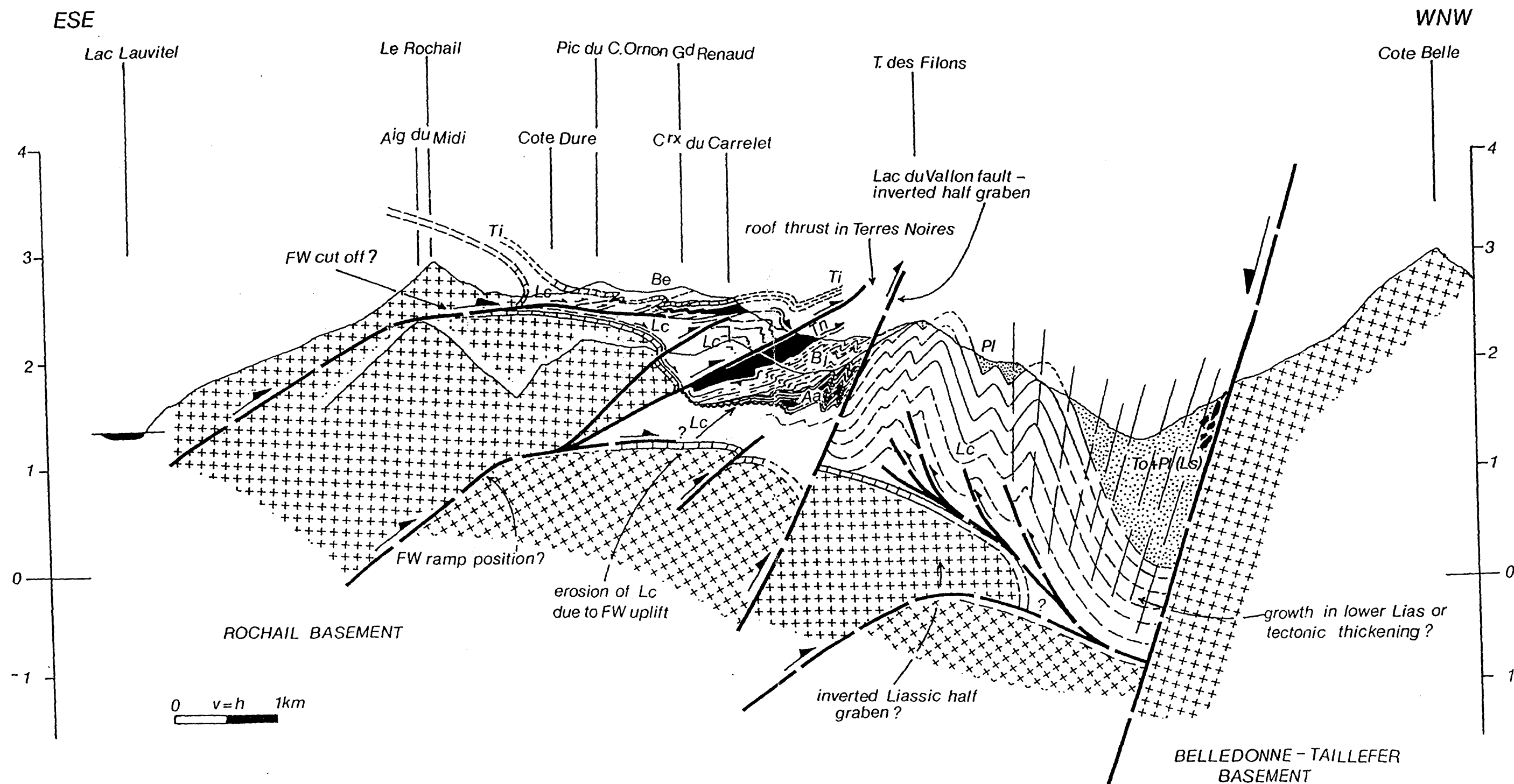


Figure 2.1.25: Grand Renaud structural cross-section across central part of Bourg d'Oisans half graben (location shown as B-B' on Fig. 2.1.1). Crosses-crystalline basement, brickwork-Triassic dolomites and spillites, Lc-Lias calcaire, Ls-Lias schisteux, To-Toarcian, Pl-Pliensbachian, Aa-Aalenian, Bj-Bajocian, Tn-Terres Noires, Ti-Tithonian, Be-Berraisian. Nb. Section sense reversed with respect to others (AA', CC', and DD').

In the adjacent syncline to the E, and in minor synclines at higher structural levels on the Pregentil fold, Barfety & Gidon (1983) document the presence of Toarcian, Domerian and Aalenian age sediments. These overlie the Lias calcaire with a discordant contact. A thicker sequence of Domerian and Toarcian age sediments observed in the syncline to the W (the main Ornon syncline) progressively pinches out against the Lias calcaire on the E-facing limb of the Pregentil anticline. It is this relationship which lead Barfety & Gidon (1983) to postulate their Jurassic half-graben interpretation.

The southward continuation of the Pregentil anticline is exposed on the N-facing slopes of la Tete de Louis XVI (G.R.7367,49902) and the Tete des Filons (G.R.7365,49891) (see Figure 2.1.42). The fold axes trend N120 on the Tete de Louis XVI and N160 on the Tete des Filons. This change in fold trend was also documented by Barfety et al. (1970).

The folds deforming the Lias calcaire are concentric folds, the hinge zones of which are characterized by i) numerous bedding-parallel detachments exploiting the less competent shaly members and ii) a complementary thrusting of the more competent limestone members. The concentric nature of these folds is especially apparent along the cliff-section exposure on the S-side of the Romanche valley, SE of Bourg d'Oisans (Plate 2.1.5).

This concentric style of folding is interpreted to signify a decoupling of the sedimentary cover from its underlying basement. Such a decoupling may have been initiated in the foot-wall to basement overthrusts mapped to the E, or reflect a deeper level thrust shortening of the basement coupled with a buttressing effect of the Ornon fault. At least some basement thrusting is required to explain the present elevation of the folded Lias calcaire along this section line (Figure 2.1.25).

iii) Villard Raymond "syncline": To the E of the Pregentil anticline, several hundred metres of Aalenian, Bajocian,



Plate 2.1.5: Concentric folds in the Lias on the S side of the Romanche valley, see text for details. Looking S.



Plate 2.1.6: Folding and thrust imbrication (not shown) of the competent Tithonian limestones on the NE face of the Grande Renaud. See text for details. Looking SW.

and Oxfordian age sediments constitute the stratigraphically defined Villard Raymond syncline (Barfety et al., 1970; Barfety & Gidon, 1983). On the Vizille 1:50 000 geology map (1972) these sediments are portrayed as a N-S trending fault slice, bordered to the W and E by Lower Liassic sediments.

The abrupt westward termination of these sediments against the Lias calcaire of the Crete du Rochail (G.R.7367,49876), is interpreted by Barfety & Gidon (1983) to be a normal stratigraphic contact. However Barfety et al. (1970), for the same contact, proposed a sinistral strike-slip interpretation disagreeing with an earlier dextral strike-slip interpretation proposed by Vernet (1961). Barfety et al. (1970) also argued that the Lac du Vallon faults were active during the Jurassic, controlling the deposition of the mostly Dogger deep water shales that make up the Villard Raymond syncline. Such an early normal fault interpretation may explain the present high strain profiles encountered in the Villard Raymond syncline, and the juxtaposition of younger deep water sediments against shallow water Lias calcaire along a transpressional-reverse fault zone of Alpine age.

On the E-facing limb of the Pregentil anticline, along the contact between Liassic and Dogger sediments, Gratier et al. (1973) describe structures indicative of extension parallel to the Pregentil fold axis. Within the Dogger sediments, N of the Chalet Pastoral (G.R.7372,49894), two oblique intersecting fold directions deform a competent limestone marker horizon in less competent shales. The axes of earlier folds which trend N30 to N40 are locally refolded by a later set of folds trending N160. Gratier et al. (1973) suggested that two intersecting conjugate strike-slip zones trending N40 and N160, with the main movement corresponding to a dextral shearing along the N40 trend, might explain the above mentioned features. From the observations of the author, the intense strain and vertical fabrics marking the western contact between the Lias calcaire and Dogger sediments, may correspond to the

continuation of the Lac du Vallon faults in the sedimentary cover (see this section). The refolding of earlier structures may equally be explained by the superimposition of compressional fabrics relating more than one phase of movement on the Lac du Vallon faults.

The eastern border of the Villard Reymond syncline is thrust controlled. The Lias calcaire overthrusts Dogger and Upper Liassic shales along a thrust contact exposed mid-way on the slopes of Clot de la Selle (G.R.7378,48980) (see Figure 2.1.42), E of Villard Reymond. This thrust plane is a gently dipping feature which truncates earlier NW-facing folds. At the Croix du Carrelet (G.R.7388,49905) this thrust cuts back towards the the ESE to link up with the Villard-Notre-Dame basement thrust. This thrust which carries the Lias calcaire, the remainder of the reduced syn-rift sequence, and a part of the post-rift sequence in its hanging-wall, transfers its displacement along a thrust flat which exploits the easy slip horizon of the Oxfordian black shales (Terres Noires) (Figure 2.1.25).

iv) Pic du Col d'Ornon-Grand Renaud: This is mainly concerned with the deformation affecting the reduced syn-rift and the post rift sedimentary sequence carried on the hanging-wall of the Villard-Notre-Dame thrust.

In the hanging-wall of the Villard-Notre-Dame thrust (Figure 2.1.42), a second cleavage (S2) crenulates an earlier cleavage (S1). The intensity and form of the second cleavage (S2) is variable, depending on its vertical distance from the Villard-Notre-Dame thrust plane exposed on the W-face of Crete de Croix de la Garde (G.R.7380,49892). This relationship is depicted schematically in Figure 2.1.26.

Close to the Villard-Notre-Dame thrust plane, the early S1 cleavage is obliterated and the S2 cleavage is sub-parallel to the thrust surface, which dips gently to the E. With increasing altitude the S2 cleavage becomes less penetrative and assumes a steeper dip (50-60° E). At

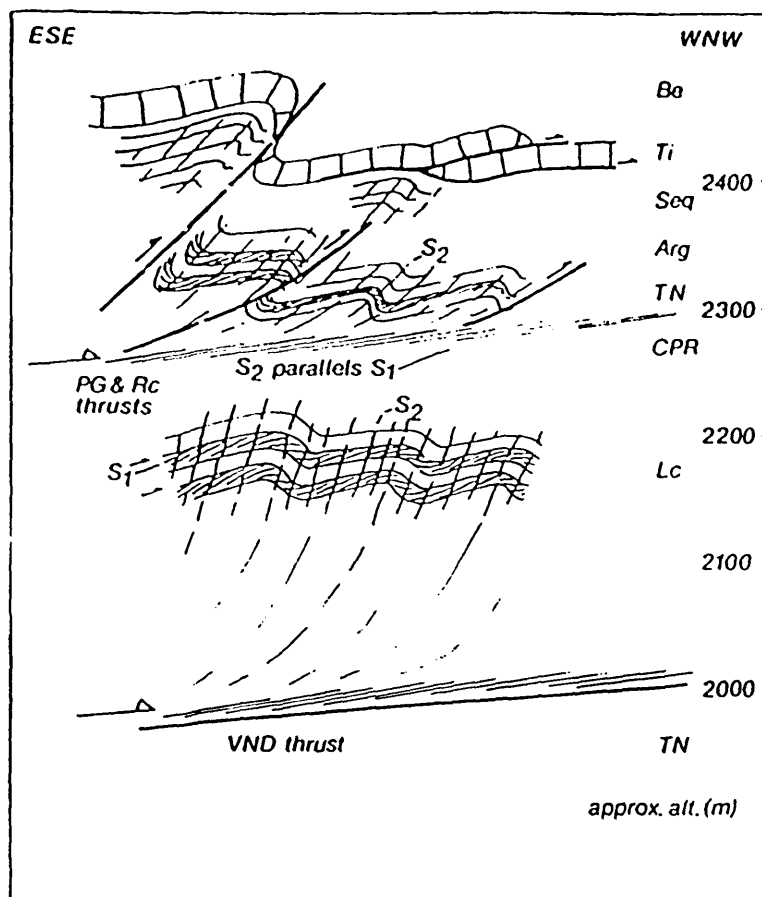


Figure 2.1.26: Structural fabrics in the hanging-wall to the VND thrust, taken from the W face of the Grand Renaud mountain. See text for explanation.

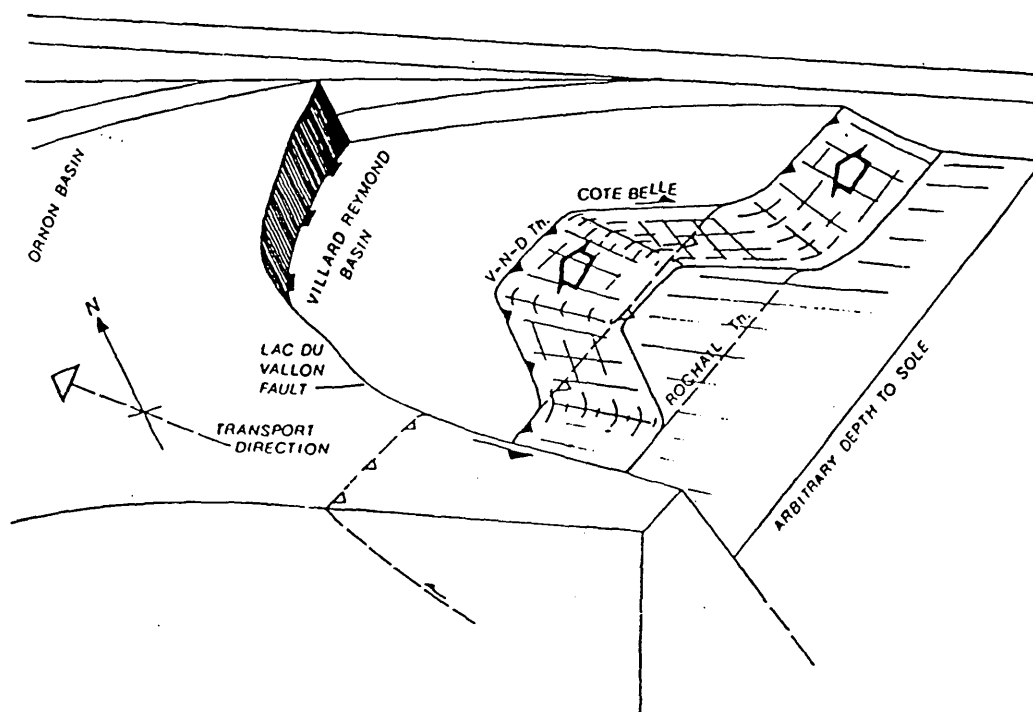


Figure 2.1.27: 3-D cartoon illustrating the foot-wall geometry of basement thrusts along the western Rochail, and the pre-existing Lac du Vallon faults which define the Villard Reymond basin. The Ornon fault is located to the W and is not shown.

higher altitudes the S2 cleavage is axial planar to F2 folds whose axes trend 050N, and are asymmetric towards the NW. These refold an earlier cleavage which sub-tended a low-angle to the bedding prior to the later folding, implying an early component of bedding-parallel shear (see also Gratier et al., 1973).

A change in deformation style occurs on passing from the Lias calcaire into the less competent Terres Noires, Sequanian and Argovian shales of the post-rift sequence. This reflects the contrasting mechanical properties of the syn-rift and post-rift sediments, and their control over the finite geometry of the deformation. In the Terres Noires, Sequanian and Argovian predominantly shale horizons, the S2 folds assume a much tighter geometry, and the S2 cleavage is less steeply inclined. The more competent limestone members within these formations display a tendency to buckle and imbricate and thrust detachment becomes increasingly more important along these horizons. The increasing shear strains and probable translation of the post-rift sequence by detachment along the Terres Noires, are interpreted to be the result of basement thrusts which ramp through the reduced syn-rift sedimentary sequence, and "flatten" along the Terres Noires and other post-rift shale horizons. However, in detail the geometry of the deformation is complex. This reflects the distributed and penetrative nature of the strain throughout the syn- and post-rift sequences.

v) **Thrust deformation of the Tithonian post-rift limestones:** A 3-D understanding of the thrust geometries affecting the Tithonian post-rift limestones can be obtained from the examination of exposures on three separate faces of the Grand Renaud mountain (Figure 2.1.42). At the head of the Pierre Grosse valley (G.R.7380,49885) a thrust imbrication of the Tithonian limestones is observed on the E-face of the Grand Renaud (Plate 2.1.6). The thrust planes dip between 20° and 25° S and strike approximately E-W. Asymmetric folds in the hanging-walls of these thrusts trend N120, and verge

towards the NNE. The geometries of these thrust imbrications show that displacements have ramped up from the lower more shaly formations, and are transferred along the base of the Tithonian so that in parts only the Tithonian limestone is involved in the thrust imbrication. A similar thrust imbrication of the Tithonian is encountered on the W-face of the Grand Renaud. These structures may be the same, or link with, the imbricates exposed on the E-face of the Grand Renaud (Figure 2.1.42).

The NE-facing direction of these folds and thrusts lead Gratier et al. (1973), and Barfety & Gidon (1983), to propose a northward direction of tectonic transport. However, these structures could equally be interpreted as lateral structures to a WNW direction of overthrusting, or oblique structures to a NW transport direction. Furthermore as these structures appear to be coeval with the F2 folds and thrust related fabrics described from the W-face of the Grand Renaud, to which can be ascribed a WNW to NW thrust transport direction, it is likely that they are oblique or lateral fold and ramp structures formed in the same WNW or NW transport system.

On the SW face of the Pic du Col d'Ornon, a more ductile folding and buckling of the Tithonian limestones is observed. The axes of these folds trend NE-SW and they are asymmetric towards the NW, further corroborating a general NW tectonic transport interpretation for the structures exposed on the E-face of the Grand Renaud (Figure 2.1.42).

There is evidence to suggest that the Tithonian limestones are deformed by late Jurassic to early Cretaceous age extensional faults. The Tithonian and Berriasian post-rift sediments are offset 40m, by an E-W trending normal fault which dips steeply to the S. The geometry described by the Tithonian limestones in the hanging-wall to this fault corresponds to an E-W trending roll-over anticline, suggesting that the fault was listric and presumably detached into a lower shale sequence. A late Jurassic to early Cretaceous age of faulting is proposed

because the hanging-wall Tithonian limestones are smeared out in a ductile fashion parallel to the fault plane. This ductile bedding-parallel extension presumably occurred during slip on the normal fault. Pinch-and-swell structures result, and isolated lenses of Tithonian limestones are formed parallel to the fault plane. The absence of brittle failure and the dominance of the ductile deformation in the hanging-wall sediments suggest that they were not fully lithified at the time of the extension.

It is impossible to determine whether or not this late Jurassic to early Cretaceous extensional faulting is part of a regional extension of this age, as documented in other areas of the Alpine foreland (Graciansky et al., 1987), or is of a gravitational origin related to compaction and subsidence dips produced by a thermal relaxation of the passive margin (Gibbs, 1984).

vi) **Basement-cover thrusting:** A large part of the deformation affecting the cover sediments of ^{the} ~~the~~ Bourg d'Oisans half-graben developed in response to basement-cover thrusting along its eastern margin.

These basement-cover thrusts are exposed to the SE of Villard-Notre-Dame (Figure 2.1.42), and on the western edge of the Rochail massif. They have been previously documented by a number of authors including Reboul (1962), Vernet (1965), Gratier et al. (1973), Barfety & Gidon (1983). The general S-dip orientation of some exposed thrust planes, in particular the Cote Belle thrust, led the above authors to postulate a northward direction of compression and thrusting. However, these basement-cover thrusts may equally be interpreted as forming part of a linked thrust system where the direction of tectonic transport was towards the WNW.

1. **Villard-Notre-Dame thrust (VND):** The frontal part of this thrust intersects the topographic surface on the W side of the Crete de Croix de la Garde (G.R.7380,49890), and emplaces a hanging-wall of Lias calcaire over a Mid-

Jurassic shale succession of Bajocian to Oxfordian age. To the E, the NNE-SSW trending Pierre Grosse valley erodes deeply into the hanging-wall of the VND thrust sheet. Steeply dipping (45° W) Lias calcaire and Triassic dolomites which overlies basement make up the W-facing valley side. The observed westward dip is clearly too steep to be compatible with a Jurassic age basinward dip generated in the hanging-wall of the Ornon fault. An increase in Triassic bedding dips along the river Pisse (G.R.7394,49887) confirms an Alpine basement-fold interpretation. This basement-fold of relatively large amplitude and long wavelength, trending N060, is interpreted to have formed over a basement foot-wall ramp in the subsurface. As the VND thrust developed, this hanging-wall fold was subsequently carried over a foot-wall of Lower Liassic cover sediments.

A cleavage which intersects the bedding with a low-angle is observed in the lower levels of the Lias calcaire (S1 ?) and is folded around the hanging-wall basement fold, suggesting some detachment of the cover occurred prior to basement thrusting and the formation of the Villard-Notre-Dame anticline. The surface trace of the VND thrust in the sedimentary cover cuts back laterally across the topography to link with a basement lateral ramp exposed on the Cote Belle (G.R.7402,49890).

The overall doming and elevation of the basement forming the basement culmination of the Rochail and Cote Dure, is therefore interpreted to have formed in response to both a passive folding of the basement over a foot-wall ramp in the subsurface (the VND ramp), and a thrust stacking of basement thrust sheets.

2. Pierre Grosse thrust (PG): The VND basement-fold is breached by a minor basement thrust - the Pierre Grosse thrust outcropping beneath the Pierre Gros pinnacle (G.R.7388,49882). This thrust has also been described by Barfety & Gidon (1983), who in terms of displacement, attach more importance to this thrust than to the

structurally lower Villard-Notre-Dame thrust. However the outcrop is such that both hanging-wall and foot-wall Triassic cut-offs can be mapped, and assuming a NW tectonic transport direction, a maximum displacement of 70-100m can be assigned to this thrust fault.

Although the displacement associated with this fault is minor, the indentation of the rigid basement into the overlying cover, previously folded in the hanging-wall of the VND thrust, generates a NW-facing syncline which trends N40 (further supporting a NW thrust transport direction).

3. Cote Belle thrusts: The Cote Belle thrust is interpreted to be a lateral ramp structure on to which branch the Villard-Notre-Dame, the Pierre Grosse, Rochail and other thrusts which imbricate the Rochail basement. The interpretation proposed here differs from the 1:50 000 Vizille geology sheet (1972), which shows two separate thrust faults on the Cote Belle. From detailed mapping, the fault exposed at le Peron (G.R.7396,49890) is the same as the fault exposed on the Cote Belle (G.R7406,49888) (Figure 2.1.42). The map relationships are complicated because subsequent erosion in the ravine du Vallon, has resulted in a deeper section through the thrust fault, exposing a basement-basement thrust ramp contact. Furthermore, the Cote Belle thrust has been domed upwards and back-steepened by a minor foot-wall imbrication of the basement and its Triassic cover (Figure 2.1.42).

At le Peron (G.R.7396,49890), where the Rochail basement overthrusts the Lias calcaire, it was hoped to date the relative age of the basement thrusting with respect to the folding and detachment of cover. However from field relationships the relative timing of these two events is not so clear. The Triassic dolomites and spillites clearly adhere to their basement, while folding in the lower levels of the Lias calcaire necessitates a decoupling of the overlying cover (Plate 2.1.7). The style of folding is disharmonic and the folds closest to the Triassic display no obvious asymmetry. However at higher



Plate 2.1.7: Fold and thrust deformation of the Lower Lias in the foot-wall to the Villard-Notre-Dame basement thrust. The horizontally bedded Triassic dolomites remain fixed to the basement. Looking WSW. See text for details.



Plate 2.1.8: Back-fold-thrust system of La Malsanne, looking S. See Fig. 2.1.29 for details.

levels, closer to the Cote Belle thrust, a distinct package of strongly asymmetric folds striking N060, are detached from the lower stratigraphic levels along a bedding-parallel thrust (Plate 2.1.8). The cleavage intensifies closer to the basement thrust contact. It is probable that these folds formed during the NW directed emplacement of the Cote Belle basement, however an earlier decoupling of the Liassic cover sequence in the foot-wall to basement thrusts farther to the E, cannot be totally discarded.

As mentioned in the foregoing, the interpretation of the Cote Belle thrust as a lateral ramp differs from the frontal ramp interpretations proposed by other workers (e.g. Gratier et al., 1973). The WNW movement along this fault is confirmed from the orientation of kinematic indicators exposed along the fault zone, and from the hanging-wall and foot-wall bedding cut-off orientations in the area of thrust imbrication exposed in the Ravine du Vallon (G.R.7398,49889).

The dip of the Cote Belle thrust steepens as it passes into the basement, and it can be traced to the SE by the occasional thrust slice of Triassic dolomites. The fault passes into the Veneon valley at le Sarret (G.R.7419,49881), close to a hydro-electric plant and its trace takes a sharp bend to the N becoming a frontal ramp on the lower slopes of Pied Moutet (G.R.7430,49885). To the N it eventually branches with thrusts in the Grandes Rousses massif.

The suggested foot-wall geometry of these basement thrust faults is shown schematically in Figure 2.1.27.

4. Rochail thrust: On the lower NE facing slopes of the Pic du Col d'Ornon, the Cote Dure basement and its covering of Triassic and Liassic sediments dip 45° to the SW. This steep dip reflects the lateral continuation of the Pierre Grosse hanging-wall fold and probably the deeper subsurface termination of the Villard-Notre-Dame basement thrust sheet against a lateral ramp. A cleavage in the first shale

horizons of the Lower Liassic intersects the bedding at a low-angle and is re-oriented around this larger lateral fold in the basement. The S-facing limb of the Cote Dure fold is parallel to the orientation of the inferred lateral ramp at depth. Two opposing lateral ramps, the Cote Belle and Cote Dure, are thus identified; the perfect situation from which to constrain the tectonic movement direction (Boyer & Elliott, 1982; Butler, 1982). On the Cote Dure SW-facing fold limb, at Jassirette (G.R.7380,49870), the Triassic and Lower Liassic sediments are offset by an E-W trending normal fault of approximately 100m displacement, downthrowing to the S. This fault does not affect sediments younger than Lower Liassic, neither does it post-date overlying Alpine deformation fabrics in the sedimentary cover. For these reasons it is suggested that the Jassirette fault may be of Liassic age.

At the Grand Clot (G.R.7382,49865), to the N of the Col du Rochail (G.R.7382,49857), is the Rochail basement thrust. The Rochail thrust plane and its hanging-wall are folded by the lower level PG and VND thrust sheets, such that ^{the} fold axis of the initially NE-SW trending, NW-facing Rochail hanging-wall fold with overturned lower limb, now plunges 40-45° to the SW giving the present map geometry (Figure 2.1.42).

The overturned lower limb of the Rochail anticline is deformed by a number of NE-SW trending vertical faults which displace the Triassic spillites and the lower most Liassic sediments by 2-4m. The high bedding cut-off angles are perhaps reminiscent of Liassic age extension faults, which were re-oriented during Alpine compression. Alternatively, these faults may accommodate an attenuation of the less competent cover sediments in the overturned limb of the Rochail anticline, and are of Alpine age. The Liassic cover sediments are also deformed by tight asymmetric, reclined chevron folds which trend NE-SW, and NW-verging thrusts. Their geometry and orientation suggests that they are contemporaneous with the overthrusting of the Rochail basement.

To the NE the Rochail thrust splays into a number of basement imbricate thrusts, which intersect the topography of the Crete du Rochail and the Rochers de Aiguilles de Midi, and then branch with the Cote Belle lateral ramp.

To the SW, the overturned limb of the Rochail anticline plunges beneath the Mesozoic cover and the upper limb dips gently toward the SW (30°). The Rochail thrust sheet and ramp anticline terminate against the NW-SE trending Lac du Vallon faults, interpreted as Jurassic age extensional faults which reactivate as oblique or lateral ramps during the Alpine compression. The Rochail thrust continues to ramp through the Liassic cover on the lower slopes of the Pic du Col d'Ornon, and eventually flattens along the Terres Noires. The distributed nature of the strain in the sedimentary cover makes the recognition of simple ramp-flat thrust geometries difficult. However, the localization of strain into zones of intense buckling and thrusting, as occurs on the SW face of the Pic du Col d'Ornon, and the correlation of such zones across the Ornon mountain predicts the zone of thrust ramping for the Rochail thrust (Figure 2.1.42).

3. Section C, La Malsanne - Le Neyrarel cross-section:

The section line is shown on Figure 2.1.1, the cross-section in Figure 2.1.28, and the structures are described generally from W to E.

i) Structural styles in the Lias calcaire of La Malsanne: To the immediate E of Les Faures (G.R.7342,49858), the Lias calcaire are spectacularly folded into tight chevron folds with sub-vertical axial planes. The overall geometry, and the meso- and micro-structures closely resemble that described from the Oulles section. These N-S trending chevron folds are deformed by a penetrative axial planar slaty cleavage with a pronounced down-dip mineral stretching lineation. To the E, these tight upright folds progressively become more open and assume an eastward vergence.

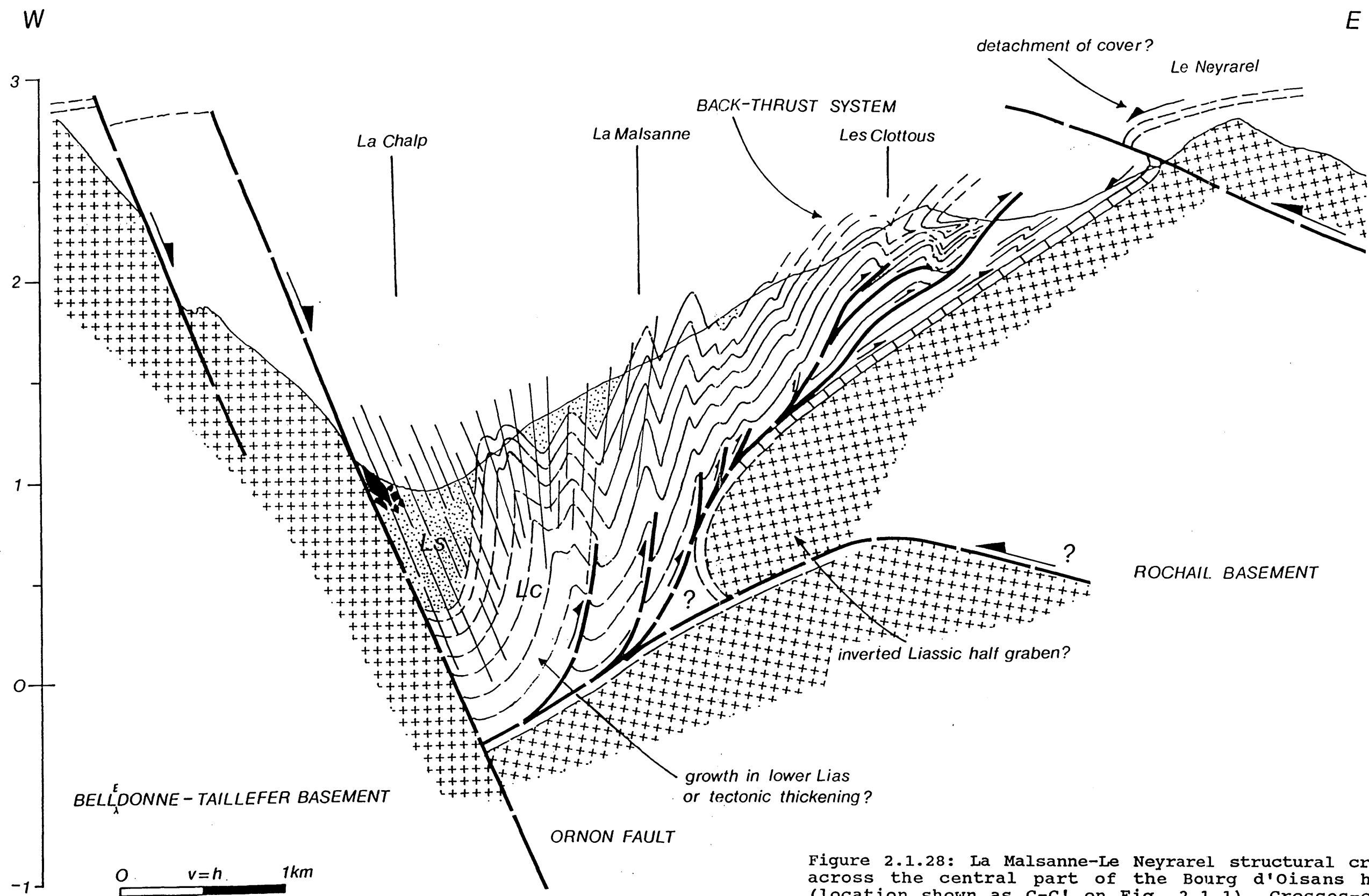


Figure 2.1.28: La Malsanne-Le Neyrarel structural cross-section across the central part of the Bourg d'Oisans half-graben (location shown as C-C' on Fig. 2.1.1). Crosses-crystalline basement, brickwork-Triassic dolomites and spillites, Lc-Lias calcaire, Ls-Lias schisteux, olistoliths adjacent to the Ornon fault also shown as is the main cleavage. The cut-off relationship of the pre-rift Triassic against the Ornon fault is poorly constrained, and the degree of W-dip may be substantially less than indicated.

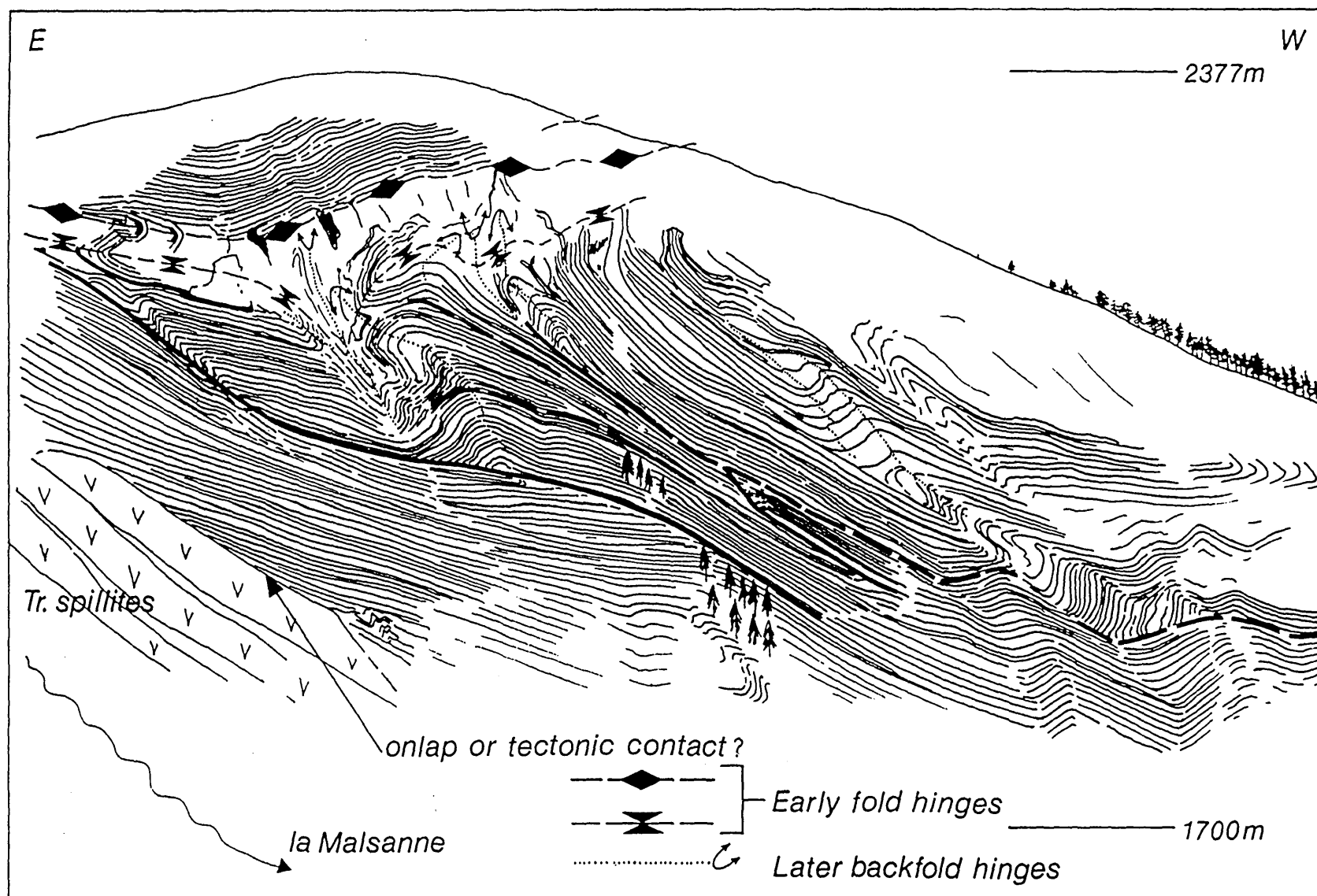


Figure 2.1.29: Panoramic sketch of La Malsanne back-fold-thrust system (see Plate 2.1.8). Complex fold-interference patterns where back-folds re-fold earlier NW-verging reclined chevron folds are shown (for location refer to Figs. 2.1.1 & 2.1.28). V = h.

Exposed on the E-W trending, N-facing valley side of La Malsanne (G.R.7360,49853) is a complex back-fold-thrust system where cover displacements are expressed as ductile eastward-verging folds and numerous bedding-parallel thrust detachments (Plate 2.1.8 & Figure 2.1.29). The importance of thrusting and detachment as opposed to folding, increases towards the eastern end of the exposed section.

The geometry of this back-fold-thrust system is further complicated by the presence of pre-existing structures. In the upper part of the section, complex interference patterns occur where the back-folds refold earlier folds. The earlier folds are NW verging reclined chevron folds whose axes trend NE-SW (N040), while the later back-folds trend NNW-SSE (N160) and are ENE-verging. This near orthogonal arrangement fold axial trends facilitates a re-folding and buckling of the earlier fold hinges, which were oriented parallel to the later compression direction. The oblique axial trends of the two fold sets, and the resulting interference patterns are shown in Plates 2.1.8 & 2.1.9, and Figure 2.1.29. A schematic diagram (Figure 2.1.30) illustrates the 3-D geometry.

Gratier et al. (1973) suggested that these obliquely trending earlier folds were of pre-Senonian age, correlating them with pre-Senonian structures found in the Devoluy (see Section 2.6). Although this may be the case, these NE-SW trending folds exposed in the Malsanne cliff-section could equally have formed during the main NW directed Alpine thrusting event, and are later refolded by back-folds and thrusts belonging to the same fold-thrust system.

ii) **Superficial Liassic extension:** The eastward directed onlap of Lower Liassic limestones and shales onto Triassic spillites exposed along the Malsanne cliff-section (Figure 2.1.29) may be interpreted as evidence of tilting during the deposition of the Lower Liassic. The Triassic-Liassic

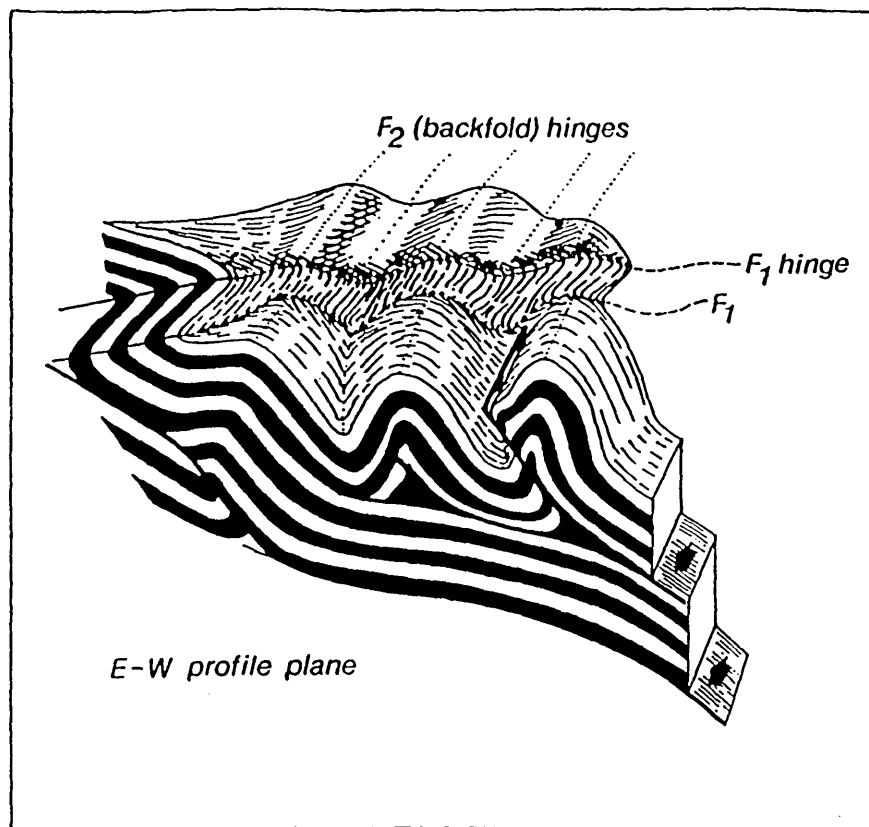


Figure 2.1.30: 3-D diagram illustrating the two-phase fold geometry along the Malsanne cliff section. E-verging back-folds (F₂) refold earlier reclined chevron folds (F₁).

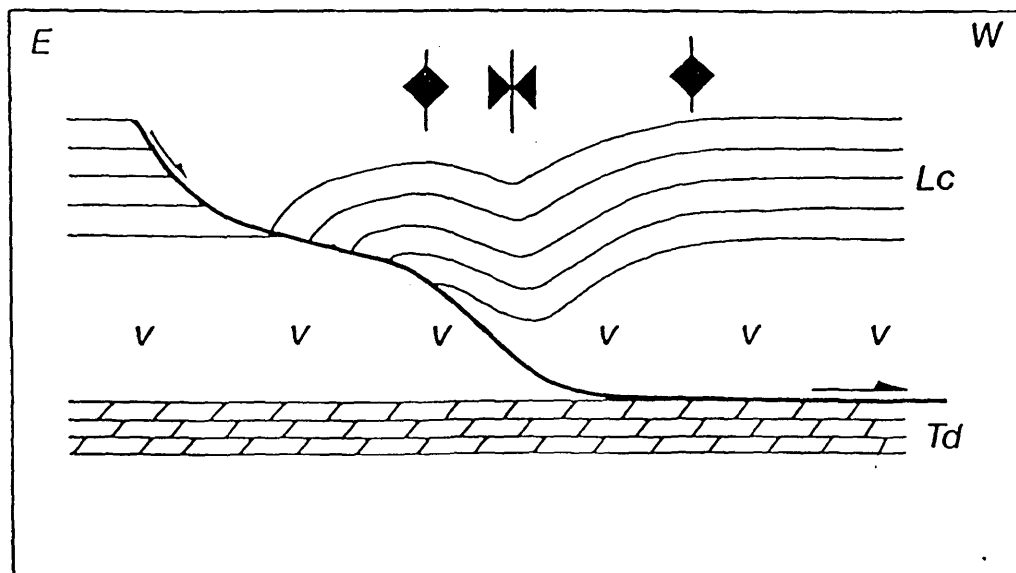


Figure 2.1.31: Schematic diagram illustrating the geometry produced as the hanging-wall moves over a ramp-flat detachment fault, Td-Triassic dolomites, V-Triassic spillites (basalts), Lc-Lias calaire.

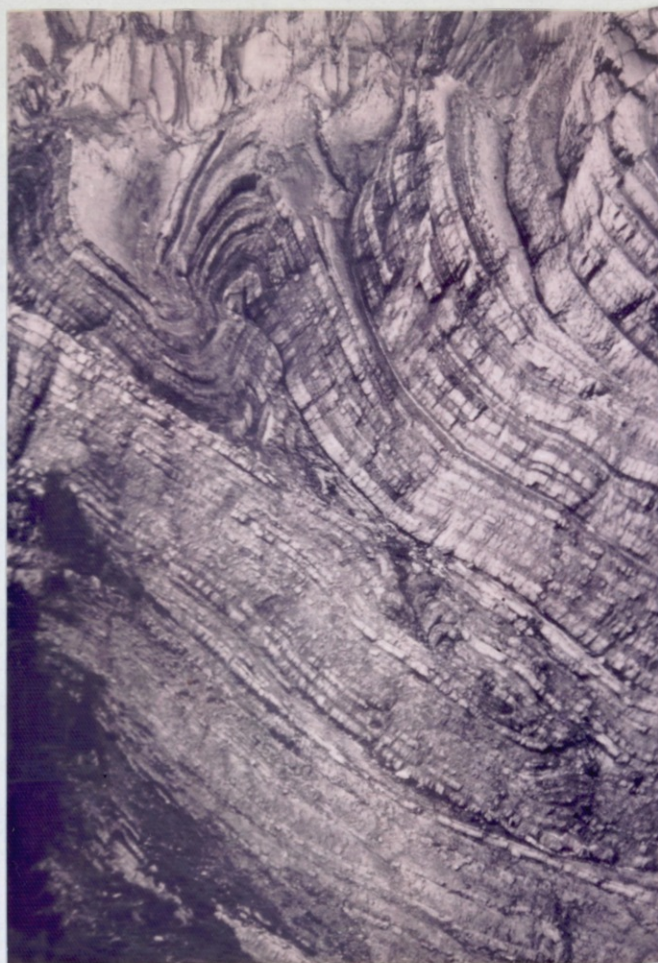


Plate 2.1.9: Close up of the La Malsanne back-folds showing folding of an earlier fold hinge. See text for details. Looking S.



Plate 2.1.10: Hanging-wall basement fold and inverted half-graben of La Paletas, see Fig. 2.1.40 for details. Looking NNW.

boundary is also offset by a low-angle extensional fault, antithetic to the main Ornon fault. The hanging-wall syncline and adjacent roll-over anticline geometry of this extensional fault suggests that the extension was largely thin-skinned, reflecting the movement of the hanging-wall over a ramp-flat staircase trajectory. This is depicted schematically in Figure 2.1.31. These hanging-wall folds trend N-S, suggesting an E-W direction of extension. Such thin-skinned, superficial structures in the hanging-wall of the Ornon fault may be gravitationally driven, reflecting steep hanging-wall dips generated during extension on the Ornon fault with a rotation of the hanging-wall (see Section 2.1.4.2).

iii) Ravine de la Chave thrust: The westward dip of the Triassic pre-rift sediments exposed in the Malsanne valley is too steep to be the original hanging-wall dips generated during Jurassic age extension on the Ornon fault. The explanation for these steep Triassic dips is evident in the exposure of the Ravine de la Chave (G.R.7366,49861), to the immediate N of the Malsanne section line. Here, the basement and its cover are folded into a tight hanging-wall ramp anticline. This anticline plunges 20° towards 120°, and is cut out against a basement thrust plane dipping 60° to the SSW, which emplaces the hanging-wall over a foot-wall of Lower Liassic sediments. Fibre-slickensides and hanging-wall cut-off trends measured in the field give a thrust transport direction towards 300° (see Figures 2.1.43 & 2.1.44), suggesting an oblique thrust ramp structure. The steep Triassic dips recorded in the Malsanne valley reflect the hanging-wall dips of the unexposed frontal part of the Ravine de la Chave thrust sheet. It has been suggested that the WNW-ESE trending part of this compressional fault is a reactivated Jurassic fault (Lemoine, pers. comm.), which links with, or is the same as the la Sue fault, interpreted as a Jurassic age transfer fault to the Ornon fault. The steep NW directed dips of the Triassic sediments exposed at la Combe (G.R.7370,49866), which are in the foot-wall of the Ravine de la Chave thrust, suggest another deeper basement sheet may have back-steepened the Ravine de la

Chave thrust.

iv) Lac du Vallon faults: At la Combe (G.R.7370,49866) are two sub-parallel faults, described by Barfety & Gidon (1983) and called the Lac du Vallon faults. Barfety & Gidon (1983) interpret both of these faults to have a Jurassic movement history which governed the distribution of the deeper water Dogger shales that make up the Villard Reymond syncline to the N. After the effects of the Alpine compression both faults have remained in net extension.

The western fault is exposed on the S-facing valley side of la Combe (G.R.7369,49866), and is sketched in Figure 2.1.32. In the hanging-wall, a large block of Triassic spillite parallels the fault plane, and to the E is partially enveloped by the Liassic cover sediments. Barfety & Gidon (1983) interpret these spillites to have been derived from the foot-wall during Jurassic extension. In the foot-wall the bedding in the Triassic dolomites is disrupted and folded. The random orientation of these folds suggests that they may be slumps related to gravitational instabilities, triggered by active syn-sedimentary extensional fault movements.

An examination of the deformation fabrics affecting the Triassic spillites adjacent to the fault plane reveals an Alpine movement history. A vertical foliation parallel to the fault plane is deformed by steeply plunging asymmetric folds, vertical (R1) shear bands, and striations and slickensides which are all kinematically consistent with an oblique-reverse to sinistral strike-slip reactivation of this fault during Alpine compression. The orientation of these structural fabrics is shown stereographically (Figure 2.1.33).

The Jurassic age cover sediments in the hanging-wall of this fault are intensely deformed by tightly asymmetric, and in places isoclinal folds, which trend NE-SW and verge towards the NW. To the immediate W of this fault the Liassic sediments have suffered only minor shortening and

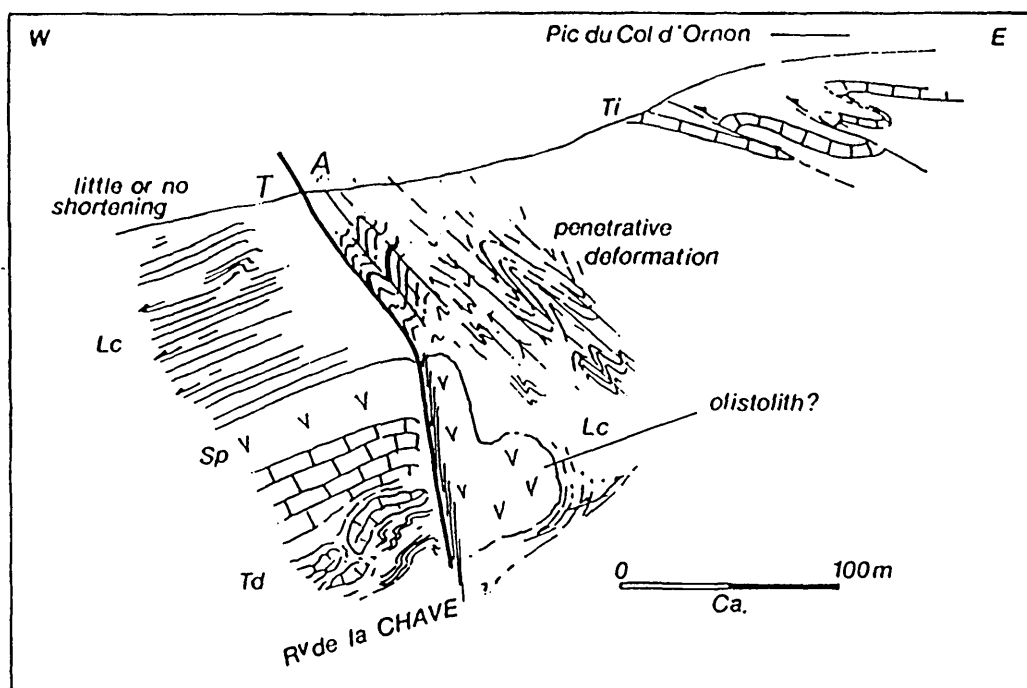
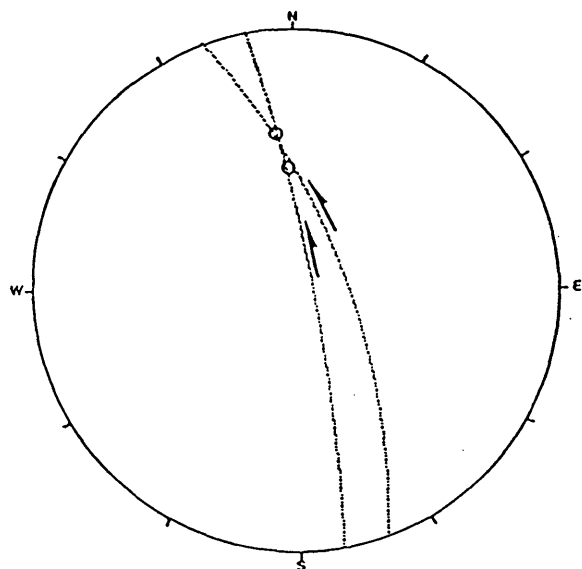


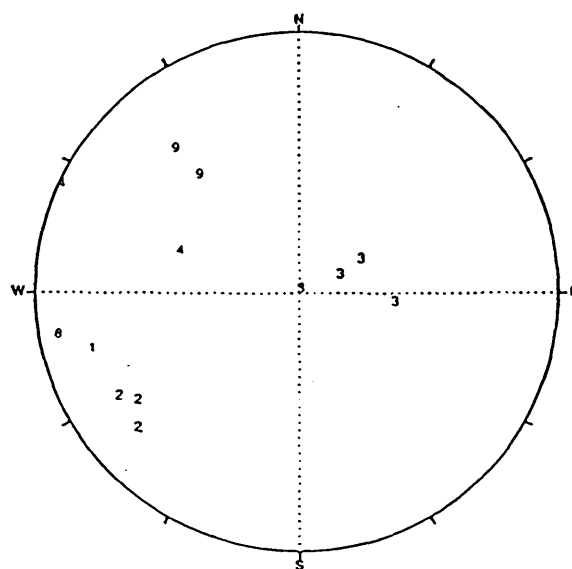
Figure 2.1.32: Field sketch of the western Lac du Vallon fault. Td-Triassic dolomites, Sp-spillites, Lc-Lias calcaire, Ti-Tithonian. T-towards viewer, A-away from viewer. For further explanation see text.

a. Lac du Vallon faults (sl)



Codes	n'	%
1	0	0.00
2	0	0.00
3	0	0.00
4	0	0.00
5	0	0.00
6	0	0.00
7	0	0.00
8 - FAULT PLANE	2	50.00
9 - SLICKENSIDES	2	50.00
Total number of data	4	

b. Lac du Vallon faults (poles)



Codes	n'	%
1 - Y FOLIATION	1	7.69
2 - R1 SHEARS	3	23.08
3 - SHEAR FOLD HINGES	4	30.77
4 - AXIAL PLANES	2	15.38
5	0	0.00
6	0	0.00
7	0	0.00
8 - FAULT PLANE	1	7.69
9 - SLICKENSIDES	2	15.38
Total number of data	13	

Figure 2.1.33: a. Equal angle (Wulf net) plot of Lac du Vallon fault planes and their slip-lineations. b. Equal area (Schmidt net) plot showing orientation of structures in the Lac du Vallon fault zone, all planes plotted as poles.

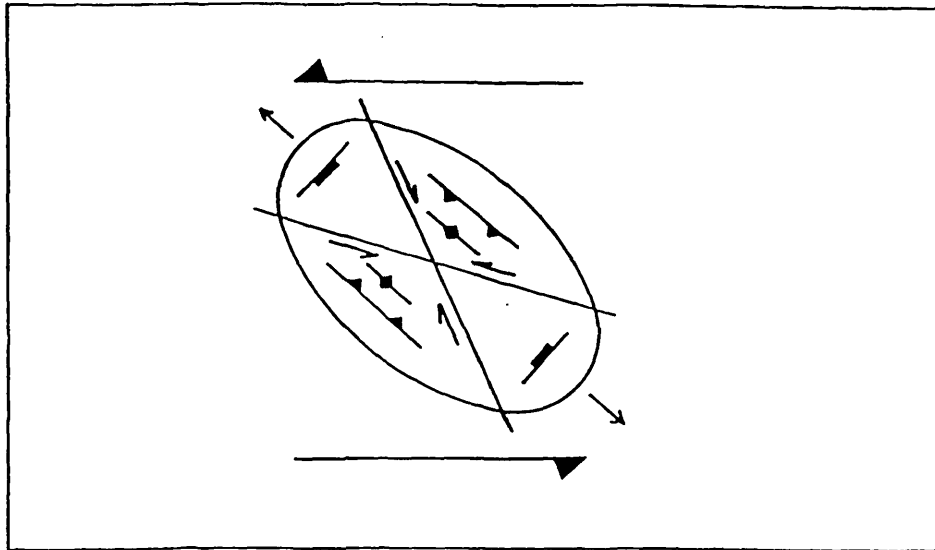


Figure 2.1.34: Arrangement of compressive and extensional elements within a strike-slip zone, after Harding (1974).

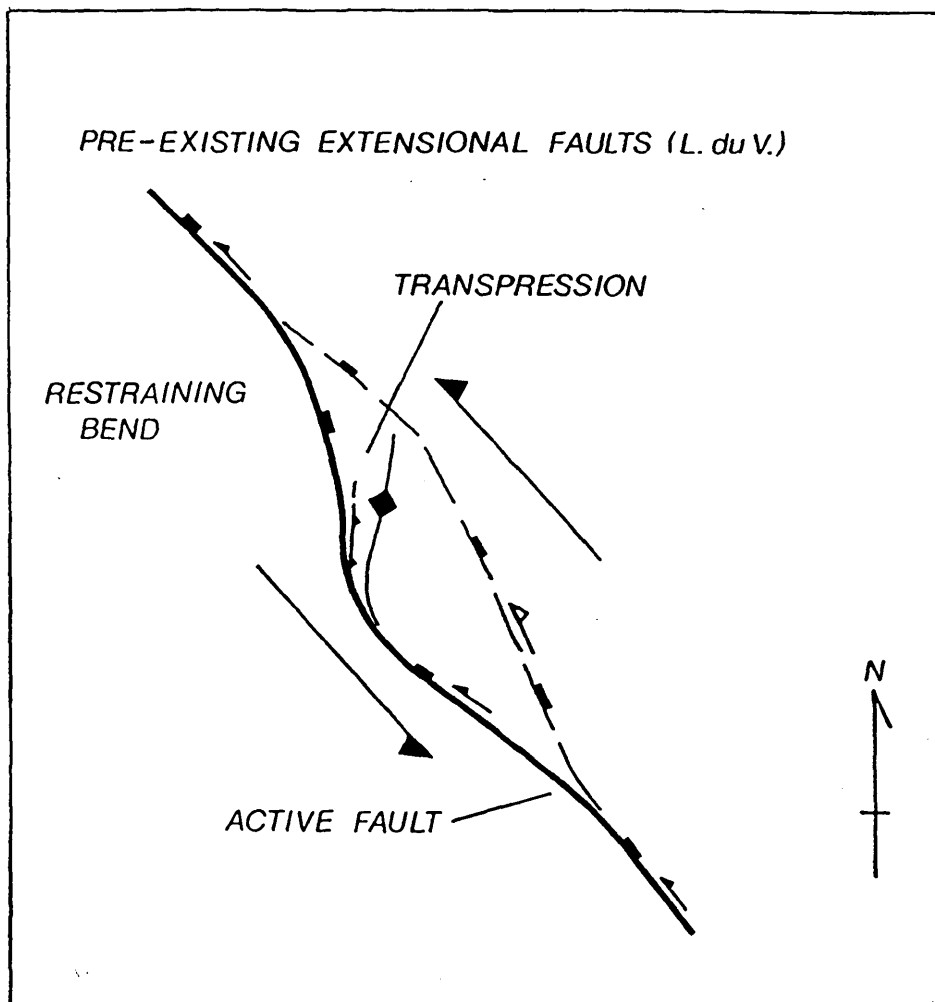


Figure 2.1.35: Restraining fault bend in the pre-existing western Lac du Vallon fault (heavy line), causing buckling and thrusting of the hanging-wall (for location see Fig. 2.1.42).

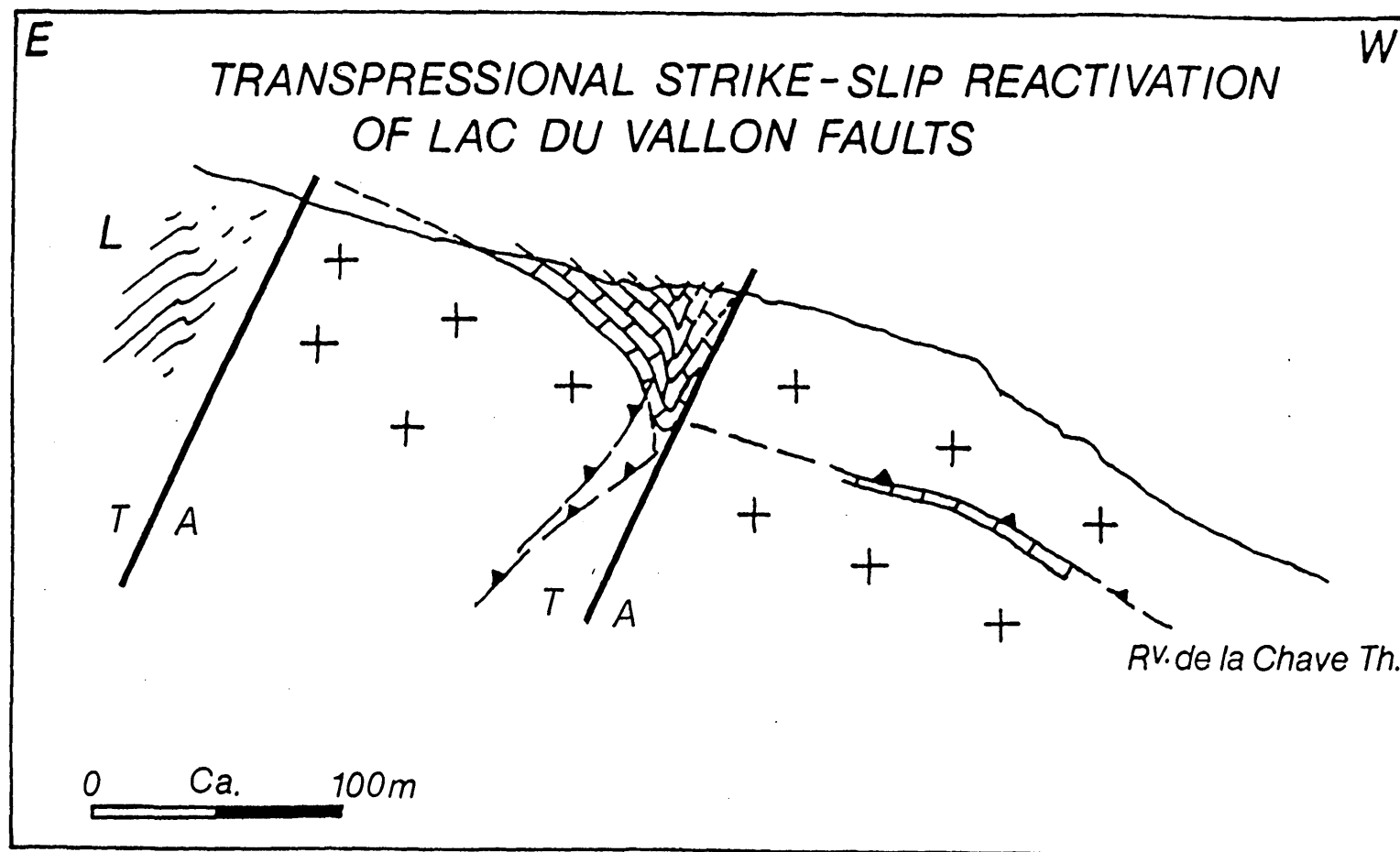


Figure 2.1.36: Field sketch of the Lac du Vallon faults. Triassic dolomites (brickwork) are folded into a tight syncline. Crosses-basement, L-Liassic cover, T-towards viewer, A-away from viewer.

deformation. This increase in intensity of the deformation affecting the downthrown eastern compartment and the absence of deformation to the W, may be explained by a buttressing effect of the fault on Alpine cover thrust displacements, forming a local zone of transpression adjacent to the pre-existing Lac du Vallon faults.

Traversing southwards across la Combe, the trace of the western Lac du Vallon fault takes a left-hand bend and then resumes its former N160 trend. This acted as a "restraining bend" during the Alpine sinistral strike-slip reactivation generating a zone of local transpression (Figures 2.1.34 & 2.1.35) and causing buckling and thrusting of the rigid basement and its cover (Figure 2.1.36). During this strike-slip reactivation, the restraining bend was "by-passed" by activating the more easterly of the Lac du Vallon faults (Figure 2.1.35). The Lac du Vallon faults can be traced southwards into the Rochail basement where they become increasingly difficult to follow (see Figure 2.1.46).

v) Neyrarel thrust: In the eastern part of this cross-section (Figure 2.1.28) Liassic and Triassic sediments are overthrust by the Rochail basement on the Neyrarel thrust (G.R.7380,49842). This thrust plane is exposed and dips 60 E, striking N-S. The orientation of striae and slickensides suggest a thrust transport direction towards the SW, contradicting to the WNW-NW transport direction recorded for the structures to the N. The relationship of this fault to the Lac du Vallon faults, is difficult to determine. The anomalous overthrust direction may reflect problems encountered during the Alpine reactivation of the Lac du Vallon faults. To the S the Neyrarel thrust cuts back into the Rochail basement where it is difficult to follow.

4) Section D, the Le Perier - La Paletas cross-section:

The syn-rift facies changes in the southern sector of the Bourg d'Oisans half-graben are described in Section 2.1.2, and the intense coaxial strains affecting Upper

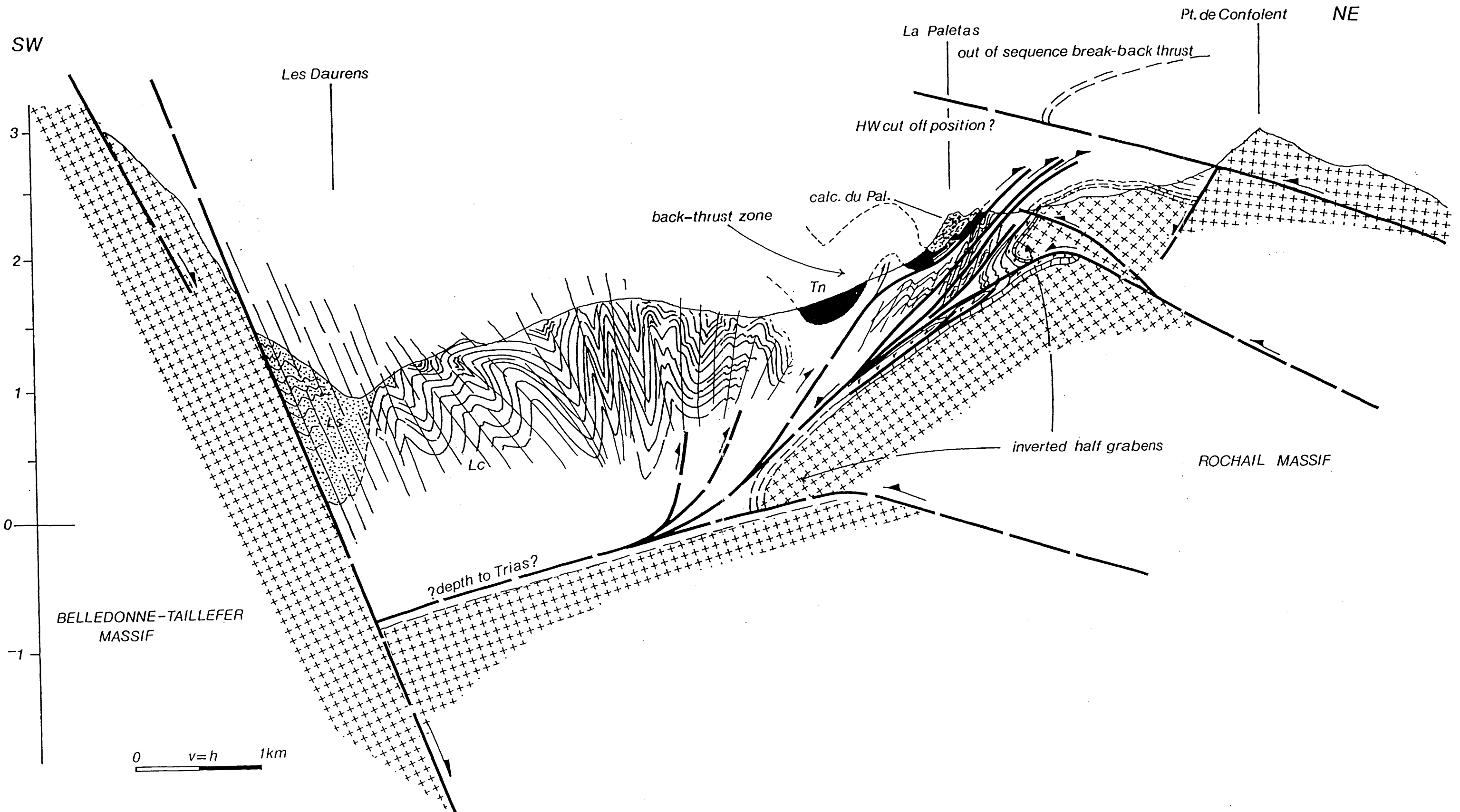


Figure 2.1.37: La Paletas-Le Perier structural cross-section across the southern Bourg d'Oisans half-graben (location shown as D-D' on Fig. 2.1.1). Crosses-crystalline basement, brickwork-Triassic dolomites, V-spillites (basalts), Lc-Lias calcaire, Ls-Lias schisteux, TN-Terres Noires, and shallow water Calcaire du Paletas limestones indicated. The subsurface geometry adjacent to the Ornon fault is poorly constrained as is the deep basement thrust geometry.

Liassic shales adjacent to the Ornon fault at les Daurens are discussed in Section 2.1.3.1. The line of the cross-section is shown in Figure 2.1.1, and the structural cross-section in Figure 2.1.37.

i) La Paletas back-thrust zone: To the E, away from the Ornon fault, the style of folding in the Lower Liassic sediments becomes more open, and the strain less penetrative. At Le Perier (G.R.7351,49810) the folds are upright to marginally westward-verging with an axial planar slaty cleavage and N-S trend. A weak vertical stretching lineation is manifest by calcite fibre growths around pyrites. Farther E beyond Confolens (G.R.7368,4981), these folds become eastward-verging forming a back-fold-thrust system. This back-fold-thrust system is correlated with the structures documented from the Malsanne section and another intermediate E-W section exposed on the S-face of the Sommet des Clottous (G.R.7370,49840) (see also Figure 2.1.46). At lower structural levels, associated with these back-folds is a W-dipping cleavage that is most pronounced in the less competent shale horizons and which intersects the bedding at a low-angle (cleavage, 65W/160; bedding, 44W/130). This back-thrust cleavage, together with fibre-slickensides exposed on the bedding surfaces, suggests a strong component bedding-parallel slip prior to, or during, flexural-slip folding.

Farther E on the southern slopes of La Paletas (G.R.7382,49820), the intensity of this back-thrust deformation increases and the Jurassic cover sediments are deformed by a number of superimposed westward-dipping fold and thrust fabrics. Back-thrusting events of different ages are identified.

All the structures exposed on the southern slopes of la Paletas have been tilted westwards by basement thrusting deeper in the section. Triassic spillites and dolomites dip 35-40°W, too steep to reconcile with Jurassic age hanging-wall dips to the Ornon fault. These steep dips are interpreted to reflect the forelimb dips of a deeper

basement thrust sheet in the subsurface.

Early back-folds are correlated with the back-fold-thrust structures described from the Malsanne section. However in the E these earlier back-folds assume a much tighter asymmetric form, and develop a penetrative westward dipping axial planar cleavage. This tightening may be in response to a later back-thrust shear, superimposed on these structures. These tight asymmetric eastward-verging back-folds are sliced through by later discrete back-thrust faults (Figure 2.1.38). The discrete fault surfaces are characterized by fibre growths diagnostic of a pressure solution-slip mechanism of deformation (Elliott, 1976a). Different generations of fibre-slickensides are identified, with the earlier sets becoming sheared as a result of renewed slip along the fault plane. Fibre-filled vein networks in the hanging-wall of the thrust sheet suggest that high pore fluid pressures probably facilitated movements along these later discrete back-thrust surfaces. The minimum displacement on one such back-thrust (Figure 2.1.38) is in the order of 100m, estimated by the measured offset of a zone of intense folding observed in both the hanging-wall and foot-wall of the fault. The whole of this section is penetratively deformed by such back-thrusts.

ii) Gravity structures: On the opposite, S-side of this E-W trending Confolens valley, the structural style exposed on the slopes of Les Terraces (G.R.7387,49812), is different from that of the zone of back-thrusting described above suggesting that either the back-thrusting dies out at a lateral tip, or the sole fault to this back-thrust system has climbed laterally to a higher stratigraphic level towards the S.

The structures exposed on Les Terraces may reflect early gravity tectonics in the Bourg d'Oisans half-graben. The deformation is characterized by numerous zones of bedding-parallel slip into which small-scale extensional faults detach. The frontal parts of these are compressional, and characterized by zones of thrust

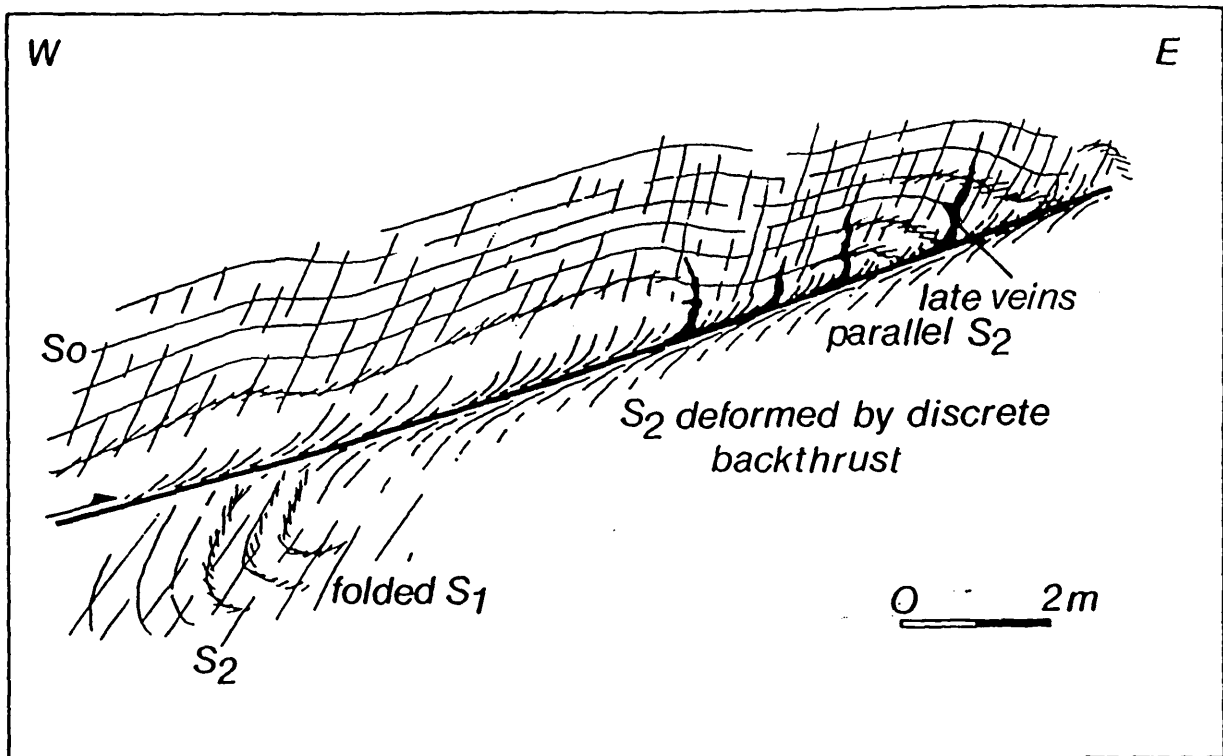


Figure 2.1.38: Late discrete back-thrust slicing through earlier back-folds and back-thrust cleavages in the Lias calcaire. S0-bedding trace, S1 and S2- earlier back-thrust cleavages. Location is beneath La Paletas peak, see Fig. 2.1.37.

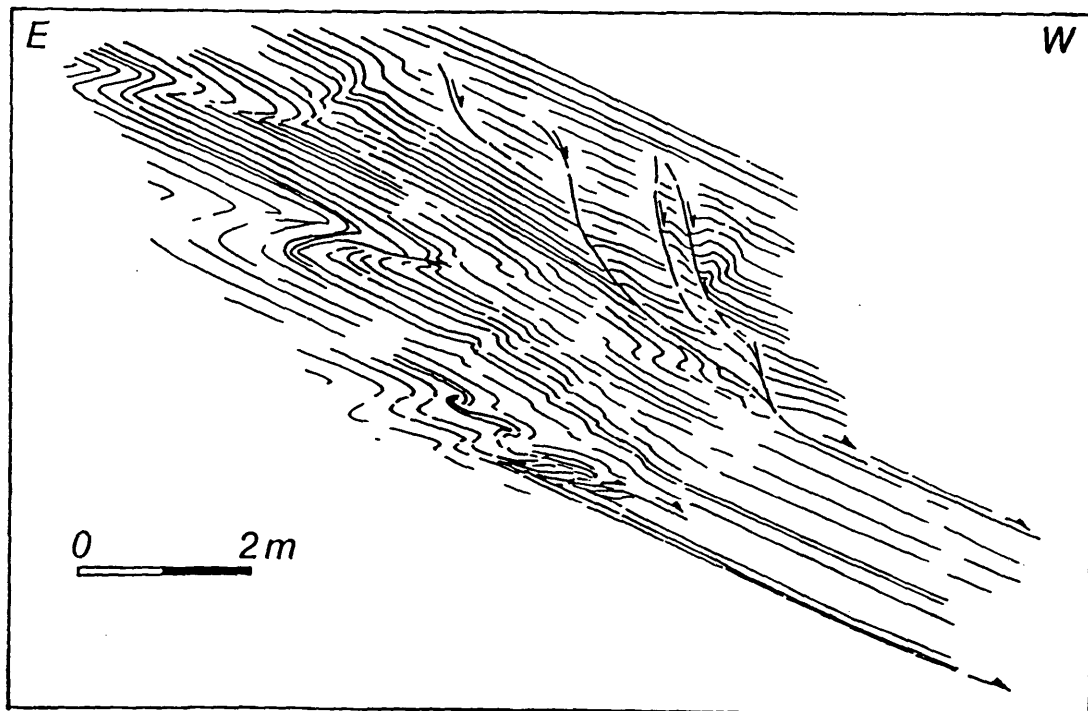


Figure 2.1.39: Field sketch showing early gravity structures in the Lias calcaire of Les Terraces. See text for further explanation.

imbrication and ductile asymmetric folds with rolling hinges (Figure 2.1.39). The compressional folds and thrusts trend N-S and are westward-verging, implying detachment occurred down the hanging-wall of the half-graben slope. We can expect to find similar structures in the zone of back-thrusting previously described, however the intensity of this deformation makes the recognition of such early structures extremely difficult.

iii) La Paletas half-graben: To the E of La Paletas (G.R.7382,49825), the cover sediments are elevated by a basement thrust which repeats basement and Triassic rocks. Both foot-wall and hanging-wall cut-offs of the basement-Triassic unconformity are exposed, and the thrust displacements are transferred along a flat detachment fault which exploits the pre-rift/syn-rift stratigraphic contact. The hanging-wall of this basement thrust sheet is folded into a tight SW-verging anticline with an overturned forelimb (Plate 2.1.10 & Figure 2.1.40). It is proposed that this basement thrust is a reactivated Liassic extensional fault. The following lines of evidence support this interpretation:

1. The anomalous bedding cut-off relationship of the Lower Liassic beds in the hanging-wall of the thrust against the thrust plane, and the stratigraphic pinch out of these beds against the hanging-wall basement anticline (Plate 2.1.10 & Figure 2.1.40). This geometry corresponds to the original prism-shape of the Liassic half-graben infill.

- 2) The nature of the deformation affecting this prism-shaped body of Lower Liassic sediments. Beds in the upper part of the prism have suffered more shortening by folding than beds in the lower part of the prism. This is interpreted to reflect the extra bed-lengths present in the upper parts of a half-graben growth sequence. During reactivation and inversion, these must shorten more than their lower level counterparts (see Chapter 3).

- 3) The exaggerated hanging-wall basement-fold. This

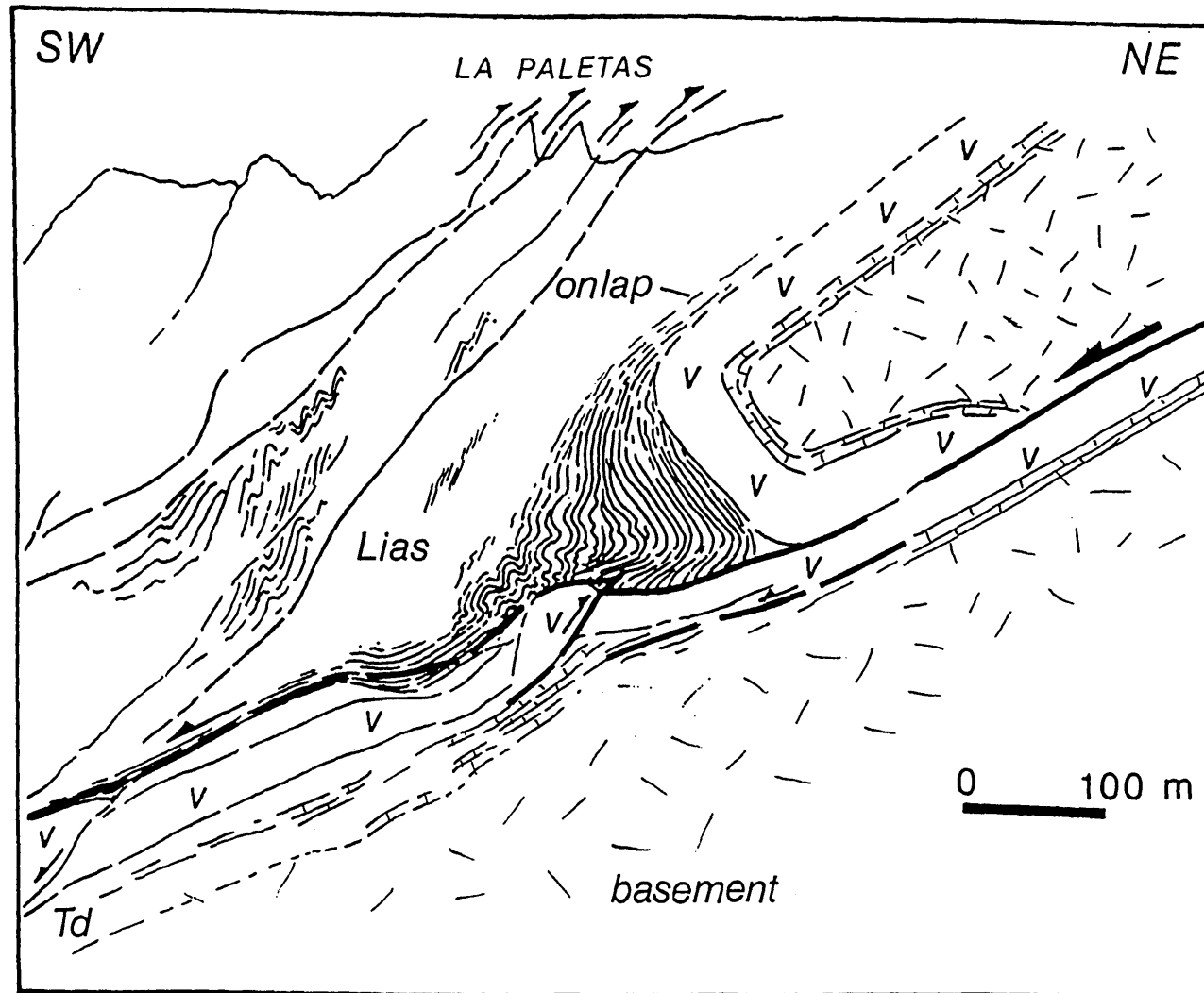


Figure 2.1.40: Inverted Liassic half-graben of La Paletas (see Fig. 2.1.37 for location). This line drawing shows the prismatic shaped infill of the Liassic half-graben thrust completely over its original foot-wall (see also Plate 2.1.10). Speckles-basement, Td-Triassic dolomites, V-spillites, L-Liassic cover.

reflects the mechanical problems encountered during the attempted reactivation of a non-planar normal fault profile (see Chapter 3).

The final contractional geometry could easily be mistaken for a hanging-wall ramp/foot-wall flat thrust relationship in a simple thin-skinned thrust system, deforming a non-tectonised layer-cake shelf sequence. However, detailed aspects of its geometry, and the style of the deformation implicates a more involved evolution with the reactivation of a pre-existing Liassic age extensional fault.

Striations recorded from the basement thrust plane suggest that the tectonic movement direction was towards the SW. This is different from the thrust transport directions recorded for the northern area, implying that we are not dealing with a simple uni-directional thrust system.

iv) Basement-cover thrusts of the Signal du Lauvitel - Breche de Perier: In the E, the sedimentary cover is overthrust by the Rochail basement (Figure 1.2.46). On the lower slopes of the Signal de Lauvitel (G.R.7400,49815) the La Paletas basement-fold is sliced through by the Signal de Lauvitel basement-involving thrust (Plate 2.1.11). Both structures are continuous into the Vallon de Valsenestre area to the SE (Section 2.3), where the Bourg d'Oisans half-graben presently terminates. To the NE by the Lac de Plan Vianney (G.R.7405,49835), the fault plane is eroded back and can be traced across to the Lac Lauvitel (G.R.7420,49850) where it localizes along a band of Carboniferous Houiller sediments in the basement. Along the Crete de la Montagne, beneath the Confolant peak (G.R.7395,49840), a gently dipping basement fault overlies Triassic and Liassic sediments. This is not the same thrust as the Signal de Lauvitel thrust and cuts back across the topography via a lateral structure into the Lauvitel basement. Kinematic indicators from the basement thrust fault suggest a SW direction of thrust movement. This



Plate 2.1.11: Gently NE-dipping Signal de Lauvitel basement thrust (arrow) slices through the La Paletas anticline. Folded Triassic dolomites occur in the foot-wall to the thrust to the left of the neve patch. See text for details. Looking SE.



Plate 2.2.1: Syn-sedimentary fault breccia of Triassic and Liassic clasts enveloped in Toarcian age sediments. From the Grande Cote. See text for details.

thrust also slices through what is interpreted to be a pre-existing Liassic extensional fault, which contains Triassic olistoliths along its length. This Liassic fault trends WNW-ESE, and its southern compartment is downthrown towards the SW (Figure 2.1.46).

2.1.3.4 Southern termination of the Bourg d'Oisans-half graben:

The present southward continuation of the Bourg d'Oisans half-graben Jurassic sediments is limited by a number of basement thrusts which climb laterally from the basement into the Jurassic cover. Details of these thrust geometries, and the southward termination of the Bourg d'Oisans half-graben in the Vallon du Valsenestre area, is addressed in Section 2.3.

2.1.3.4 Thrust transport directions:

Studies of the regional fault geometry were combined with studies of the kinematic indicator trends to determine the thrust transport direction.

1) Thrust transport direction of the sedimentary cover:

Evidence is presented for a decoupling of the syn- and post-rift sequence from the underlying pre-rift sequence and its basement. The cover sequence is detached in the foot-wall to basement thrusts on the western edge of the Rochail. However as this cover detachment surface is joined, folded or breached by lower level thrusts involving basement, it is difficult to determine in the foot-wall to which basement thrust the Bourg d'Oisans cover originally detached. These later basement thrusts, some of which are reactivated early extensional faults, refold or deform earlier fabrics in the sedimentary cover.

Moreover, in the western part of the area, fold orientations and strain patterns in the sedimentary cover may not be reliable indicators of the thrust transport

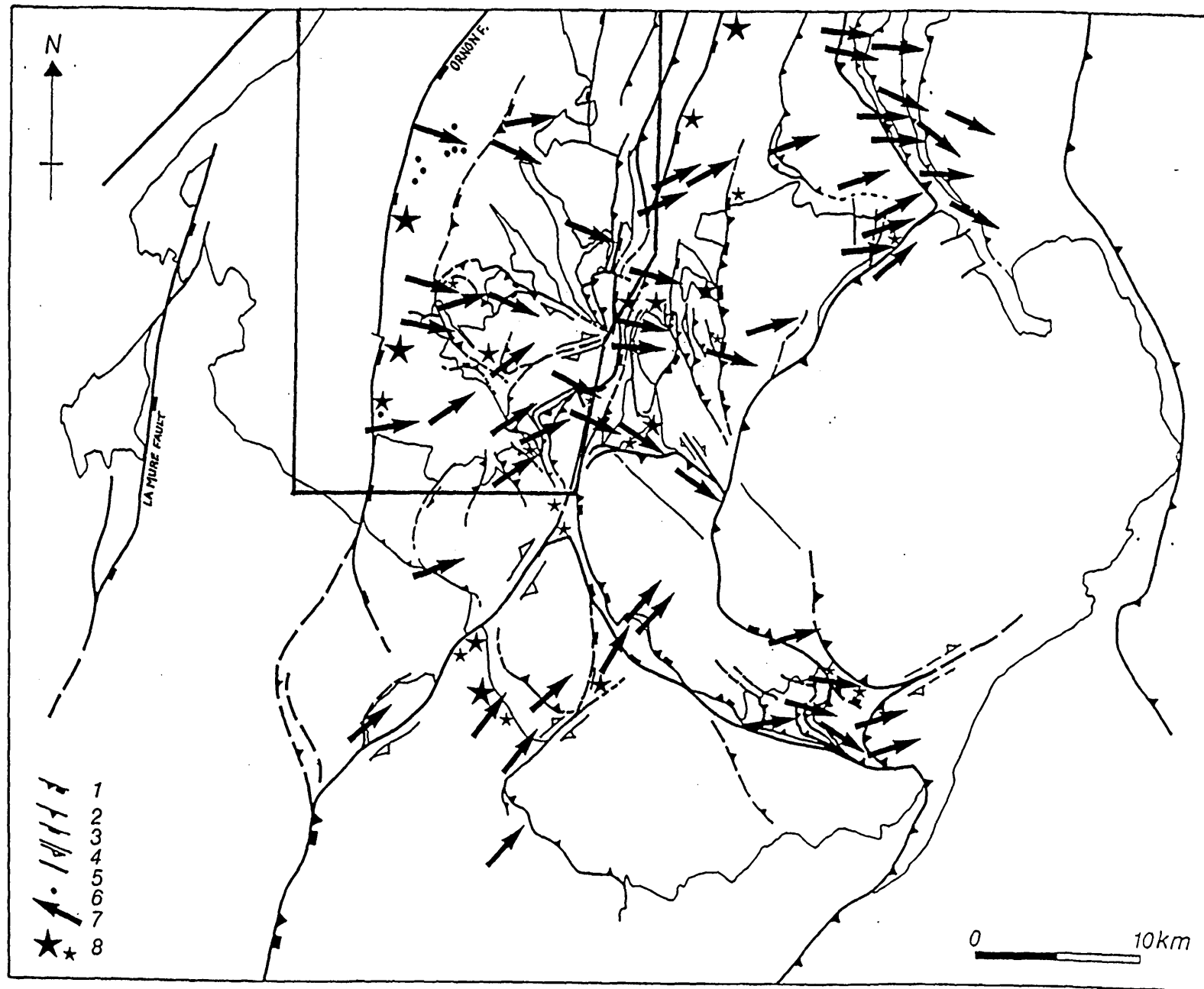


Figure 2.1.41: Map shows stretching lineation trends for all the Pelvoux area. Here we are concerned with the Bourg d'Oisans area (outlined), specifically the vertically plunging lineations (circles) adjacent to the Ornon fault. Based on mapping by the author.

direction because they strongly reflect the influence of pre-existing extensional fault scarps i.e. the Ornon fault. A map of the finite extension directions (Figure 2.1.41), normally a reliable indicator of thrust transport direction, shows that along a N-S trending zone parallel to the Ornon fault, the extensional axis of the strain ellipsoid is vertical, and from a semi-quantitative strain analysis (i.e. carried out in the field), the magnitude of strain increases dramatically approaching the Ornon fault.

In the E, fold orientations and kinematic indicator trends in the sedimentary cover are consistent with a WNW thrust transport direction. This differs only marginally from the 310° direction obtained from the analysis of basement-cover thrusts in the eastern part of the area (see below). Some of the fold trends, notably those recorded in the foot-wall to the Villard-Notre-Dame/Cote Belle thrusts are highly oblique to the regional N-S trend. These folds developed in the foot-wall to a basement lateral ramp structure, and may have formed with such an orientation, or were rotated into parallelism with the direction of thrusting.

Because of later intense back-thrusting, it is difficult to determine the initial thrust transport direction for the sedimentary cover in the central and the southern parts of the Bourg d'Oisans half-graben.

2) Thrust transport directions in the basement:

i) Northern Bourg d'Oisans area: The recognition of opposing lateral ramps is the ideal situation from which to constrain the thrust transport direction (Boyer & Elliott, 1982; Butler, 1982). The Cote Belle lateral ramp, and the hanging-wall expression of Cote Dure lateral ramp (Figure 2.1.42), represents such an opposing lateral ramp situation. This together with the NW-SE trending Lac du Vallon faults (see also Figure 2.1.42), constrains the direction of basement thrusting towards 310° (NW).

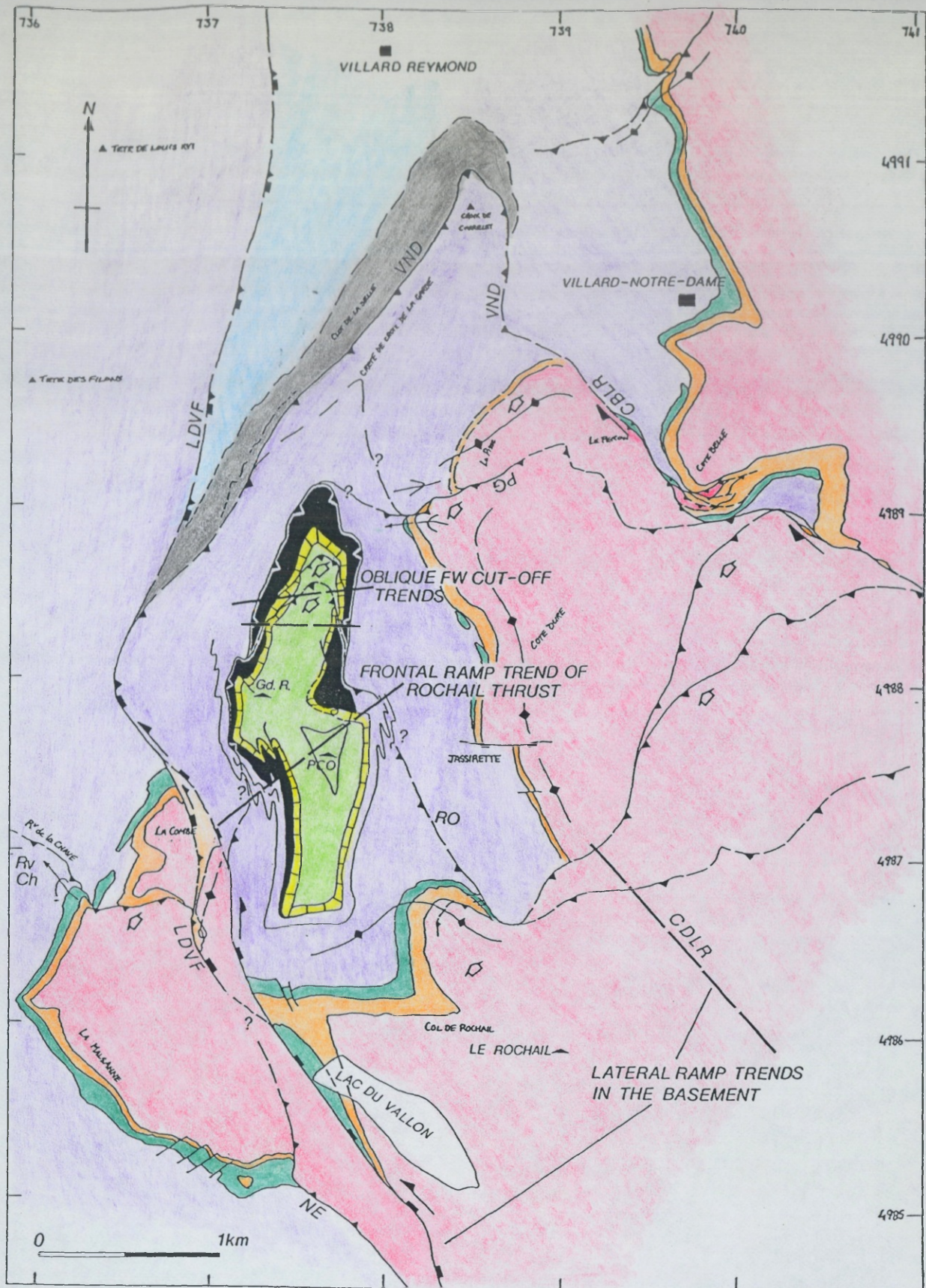


Figure 2.1.42: Simplified structural map of the northern Bourg d'Oisans area showing main faults, cut-off trends and lateral ramp trends in the basement. Location and legend shown in Figs. 1.7 & 1.8. VND-Villard-Notre-Dame thrust, PG-Pierre Grosse thrust, CBLR-Cote Belle lateral ramp, CDLR-Cote Dure lateral ramp, RO-Rohail thrust, LDVF-Lac du Vallon faults, RvCh-Ravine de la Chave thrust, NE-Neyrarel thrust. Arrows indicate direction of basement overthrusting, black-Terres Noires. Based on mapping by the author.

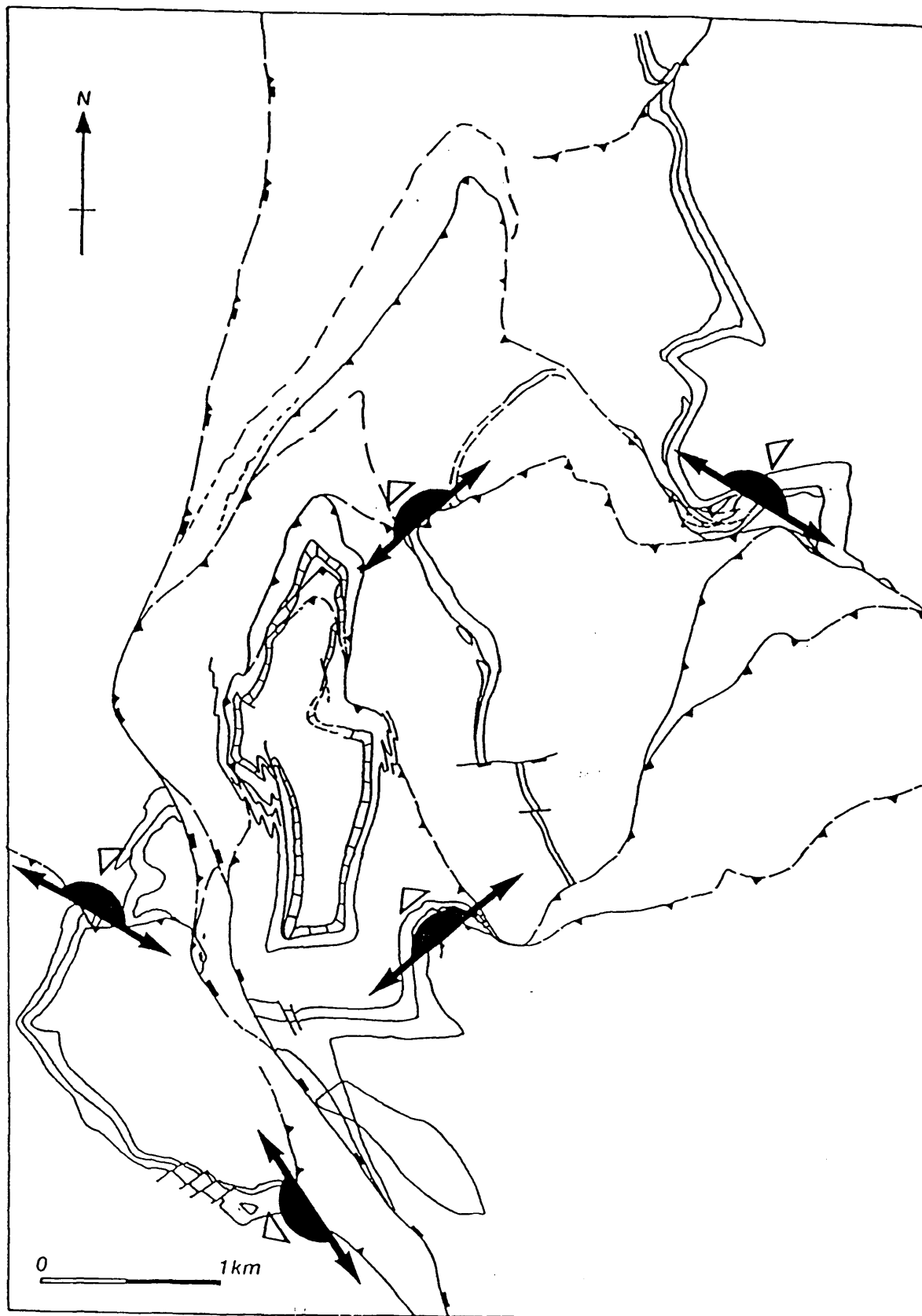


Figure 2.1.43: Basement-Trias unconformity cut-off trend map for the northern Bourg d'Oisans area. Apex of triangles indicates direction of thrust transport (i.e. frontal or lateral cut-off trends). Localities and legend as for Fig. 2.1.42.

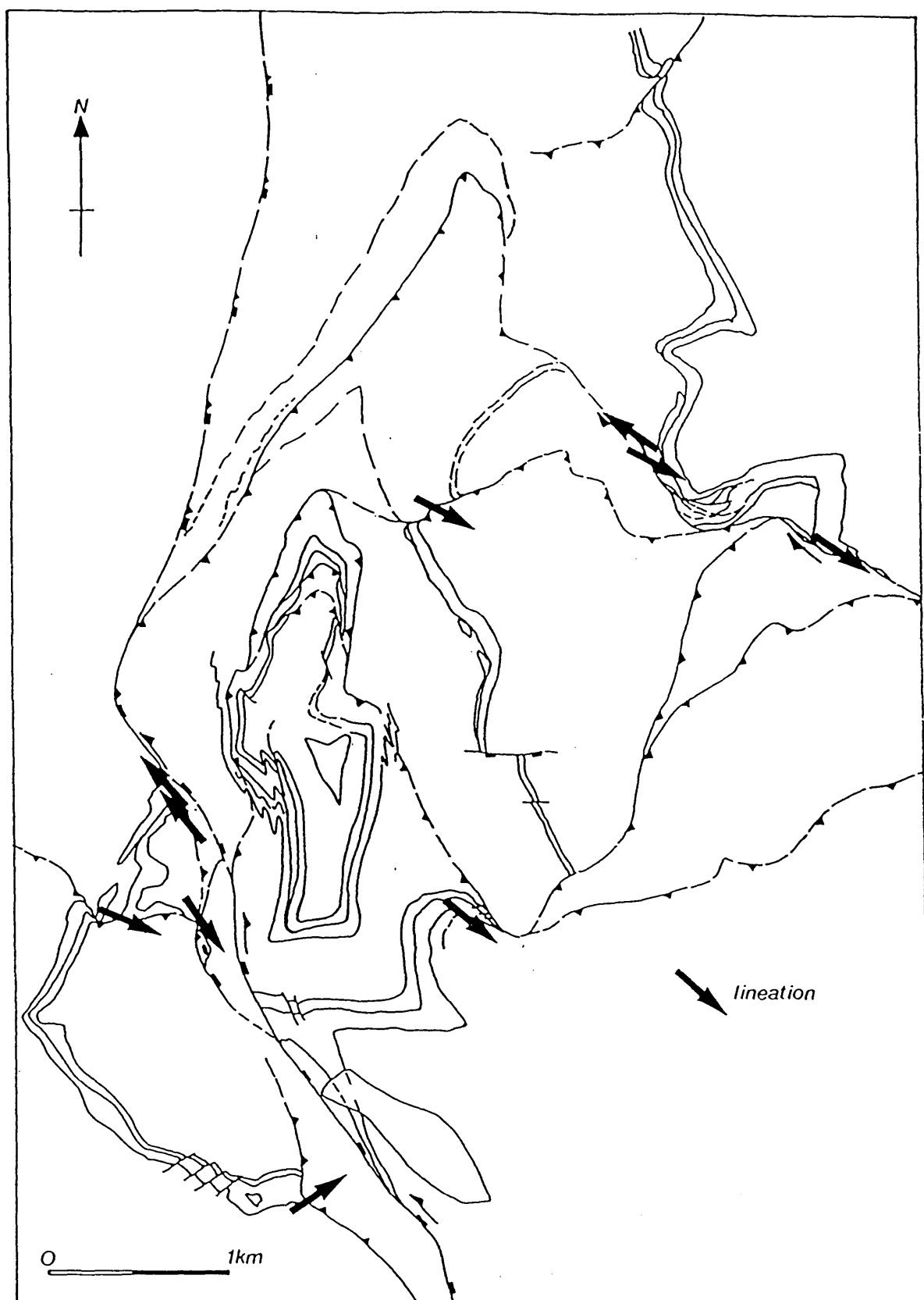


Figure 2.1.44: Slip-lineation data from basement-cover thrust faults in the northern Bourg d'Oisans area. A NW-SE trend is evident. Localities and legend as for Figure 2.1.42.

The thrust transport direction determined from fault cut-off trends of the Triassic pre-rift sequence (Figure 2.1.43), is consistent with the NW transport direction proposed from the lateral ramp orientations.

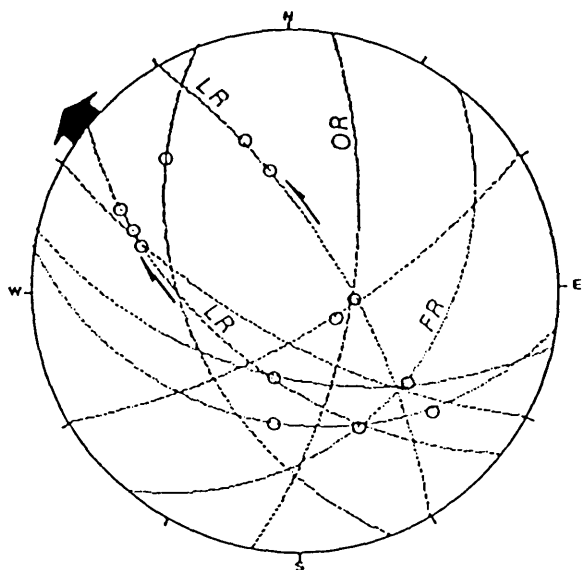
Lineation trends (fibre-slickensides, striations, and the preferred alignment of minerals - mineral stretching direction) are presented in Figure 2.1.44. The slip-lineation data and fault plane orientations are also plotted stereographically (Figure 2.1.45 a & b). A NW-SE lineation trend is evident, and is consistent with a thrust transport direction towards 310° . The motion along the NW-SE trending fault planes is dominantly strike-slip, this orientation of plane corresponding to the thrust lateral ramps. The thrust motion on the NE-SW trending, SE dipping thrust planes is dominantly dip-slip, these planes corresponding to the frontal thrust ramps. Planes whose orientations are intermediate between these two trends are generally planes of oblique-slip, and interpreted as oblique thrust ramps.

Some of these basement thrusts correspond to reactivated Jurassic extension faults, and anomalous slip-directions may result. This is due to the problems encountered in the dip-slip reactivation of steeply dipping faults (see Chapter 3). Nevertheless, the overall consistency of a NW-SE kinematic indicator trend accords well with the WNW to NW thrust transport direction obtained using other methods.

ii) Southern Bourg d'Oisans area: A study of the thrust transport direction using the methods listed above, yields different results to the NW transport direction obtained for the northern sector.

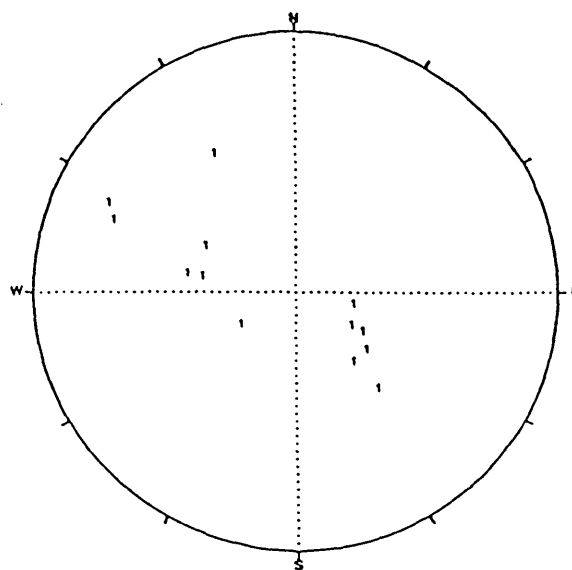
An analysis of the fault cut-off trends (Figure 2.1.46), is itself insufficient to determine accurately the direction of thrust transport because of the 180° range of possible movement directions from a single cut-off trend. As no obvious lateral ramp structures were identified, it

a. Bourg d'Oisans (N)



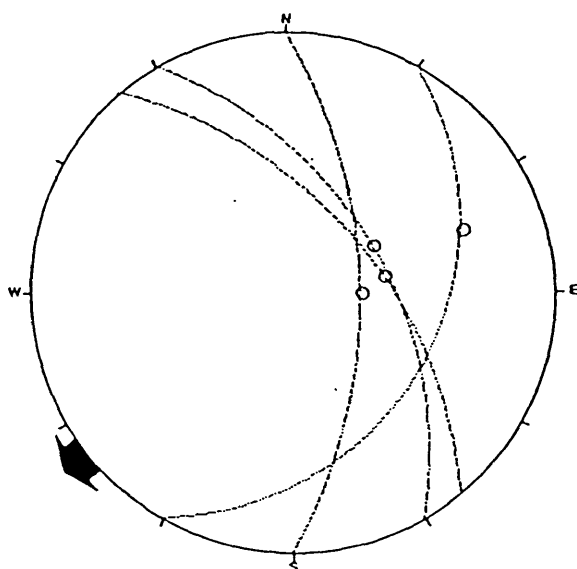
Codes	n°	%
1.	0	0.00
2.	0	0.00
3.	0	0.00
4.	0	0.00
5.	0	0.00
6.	0	0.00
7.	0	0.00
8 - FAULT PLANE	13	50.00
9 - SLICKENSIDES	13	50.00
Total number of data	26	

b. Slickensides B. d'O. (N)



Codes	n°	%
1 - slickensides	13	100.00
2.	0	0.00
3.	0	0.00
4.	0	0.00
5.	0	0.00
6.	0	0.00
7.	0	0.00
8.	0	0.00
9.	0	0.00
Total number of data	13	

c. Bourg d'Oisans (S)



Codes	n°	%
1	0	0.00
2.	0	0.00
3.	0	0.00
4	0	0.00
5.	0	0.00
6.	0	0.00
7.	0	0.00
8 - FAULT PLANE	4	50.00
9 - SLICKENSIDES	4	50.00
Total number of data	8	

Figure 2.1.45:

- Slip-lineation and fault plane data for the northern Bourg d'Oisans area (Wulff net).
- Equal area (Schmidt net) plot of slip-lineations for the northern Bourg d'Oisans area reveals NW-SE trend.
- Slip-lineation and fault plane data for the southern Bourg d'Oisans ' area (Wulff net).

Arrows signify thrust transport direction, see text for explanation.

was necessary to combine information on the thrust geometry with a study of the kinematic indicator trends to determine the thrust transport direction. The slip-lineation data recorded from basement thrust planes are all consistent with a SW thrust transport direction. These data are presented stereographically in Figure 2.1.45 c, and show a thrust transport direction towards 235°. The difference in thrust transport direction between northern and southern areas, becomes most apparent when comparing rose diagrams of the slip-lineation data (Figure 2.1.47), or histogram plots of slip-direction against frequency for all the Bourg d'Oisans area (Figure 2.1.48).

Having demonstrated the existence of two thrust transport directions in the Bourg d'Oisans area, it is more difficult to determine what relationship exists between them. The problem is further confounded by the out-of-sequence, break-back geometry of the Signal de Lauvitel thrust, which slices through an earlier SW-directed thrust structure. Two possible solutions are:

1. The two transport directions reflect movements on two different age thrust systems.
2. The two directions are of the same age and accommodated on the same thrust system.

The critical relationship is that between the Lac du Vallon faults which accommodate NW-directed thrust displacements, and the thrust structures in the southern Bourg d'Oisans area, which accommodate SW-directed displacements. Unfortunately, it is impossible to locate the southward continuation of the Lac du Vallon faults in the Rochail basement, and therefore difficult to determine the relationship between these two directions. However, movement on the Le Neyrarel thrust is towards the SW, and this thrust appears to slice through the Lac Du Vallon faults which accommodate NW-directed movements. This relationship would suggest that the SW-directed structures are later, and cut through earlier NW-directed structures.

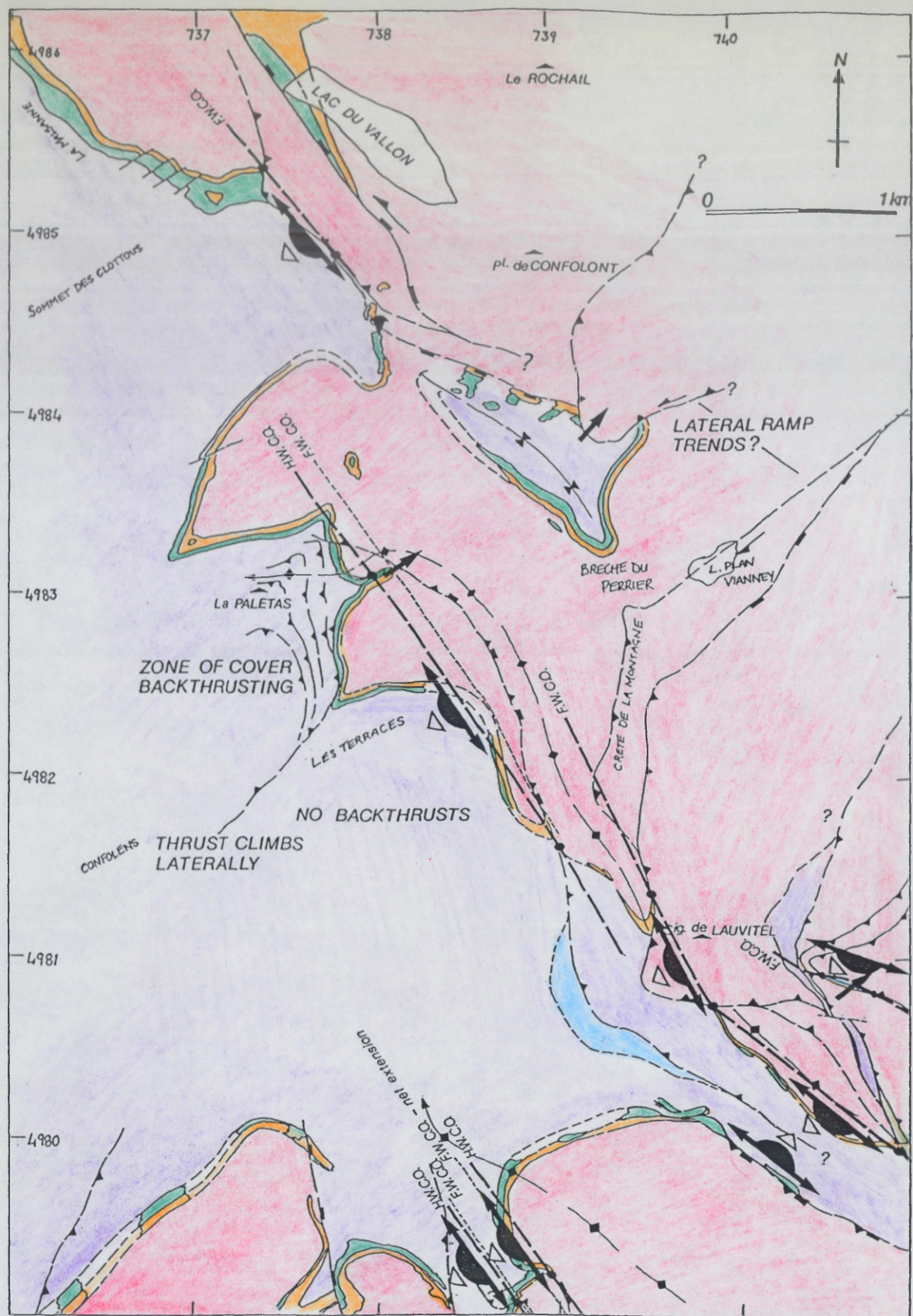


Figure 2.1.46: Simplified structural map of the southern Bourg d'Oisans area showing main faults, and hanging-wall (HWCO) and foot-wall (FWCO) cut-off trends for the basement-Trias unconformity on basement-cover thrust faults. Location and legend indicated in Figs. 1.7 & 1.8. Based on mapping by the author.

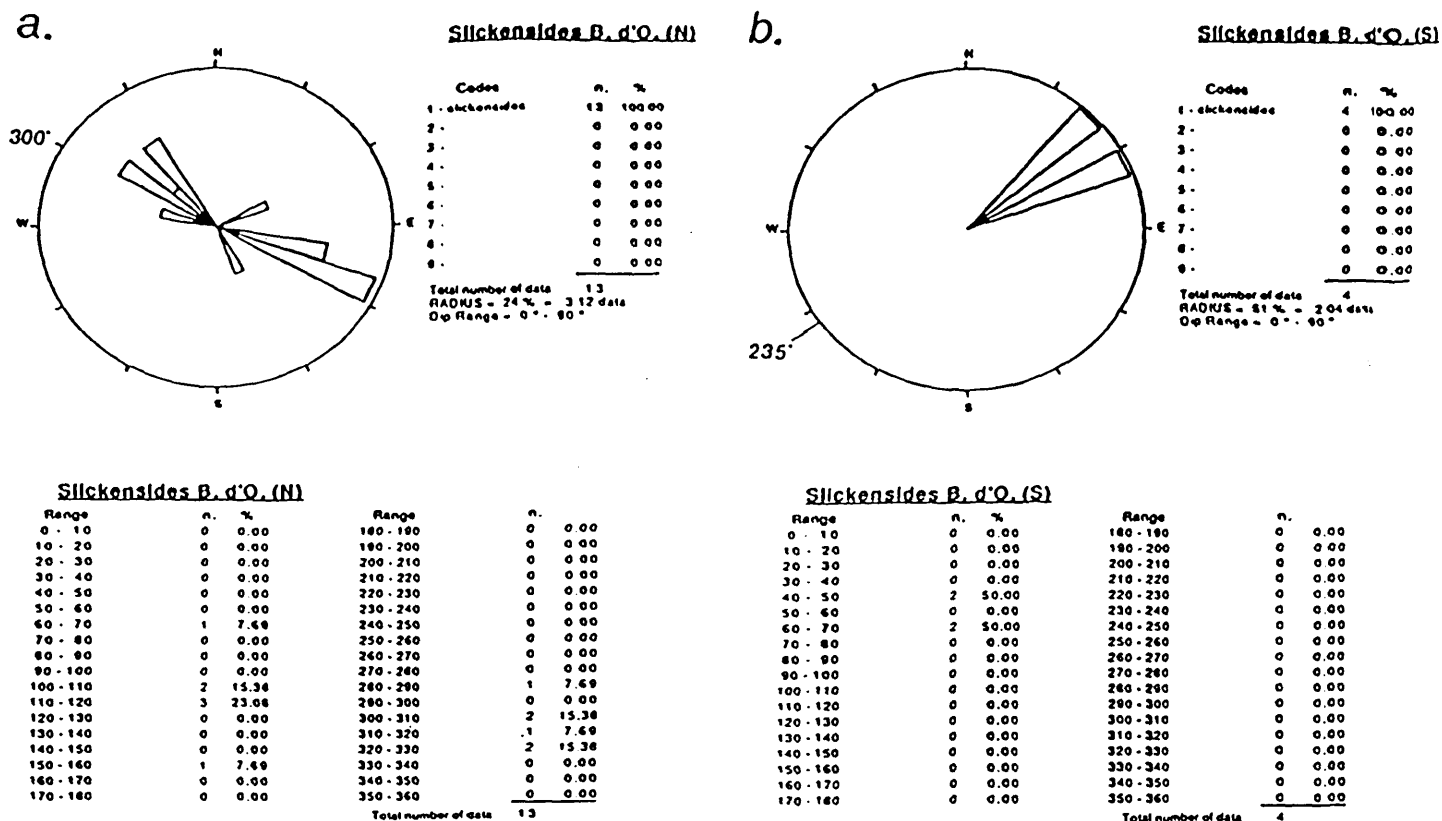


Figure 2.1.47: Rose diagrams of slip-lineation trends for a) the northern Bourg d'Oisans area, and b) the southern Bourg d'Oisans area, revealing the NW and SW thrust transport directions.

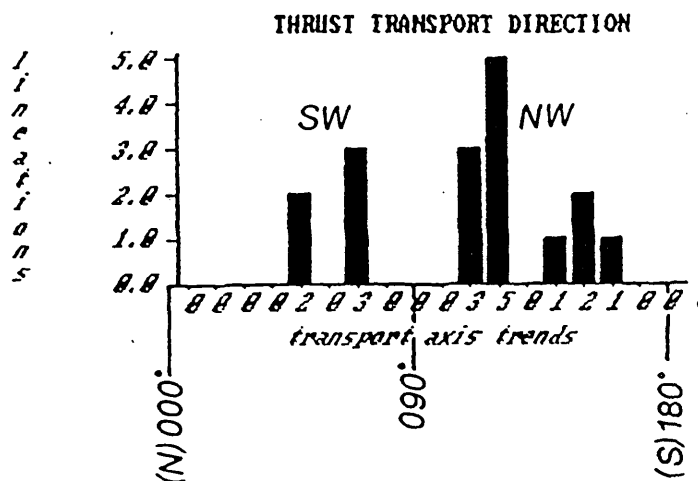


Figure 2.1.48: Histogram of slip-lineation data for the Bourg d'Oisans area. The two transport trends, NW and SW are clearly evident.

However, it is also plausible that the two transport directions belong to the same thrust system and were operative simultaneously. To understand the significance of these two transport directions, it is necessary to consider the continuity of structures and thrust relationships, in adjacent areas.

2.1.3.6 La Mure half-graben:

The stratigraphy of the La Mure half-graben closely resembles that of the Bourg d'Oisans half-graben (see Section 2.1.2).

The La Mure syn-rift sediments are largely unaffected by the effects of Alpine compression, giving support to the interpretation that the Ornon fault acted as a buttress during Alpine thrusting.

The syn-rift sediments are deformed by downward facing, disharmonic fold structures, well exposed along the D217 road from Valbonnais to La Mure, and which are assigned a gravitational origin. The fold axes trend N160, and associated detachment surfaces are mostly bedding-parallel, detaching down the half-graben slope. Internal sedimentary discordances, indicative of syn-sedimentary tectonic activity, are characterized by the progressive eastward-directed onlap of younger sediments onto older sediments which have been tilted towards the W. This suggests that tilting was controlled by extensional movements on the La Mure fault. These earlier gravitational structures are refolded by later folds which may be of Alpine age. These later folds are fairly open, and upright to slightly westward-verging, with a weak axial planar cleavage that overprints the gravity structures. The orientation of these later folds is similar to that of the early gravity folds, with the folds plunging 20° towards 190°. The pre-rift Triassic dolomites dip 28-30°W and are exposed on the W-facing slopes of Le Coiro. The La Mure half-graben is itself re-oriented by deeper level movements on the external Belledonne thrust.

2.1.4 Interpretative model for the extensional-compressional evolution of the Bourg d'Oisans and La Mure half-grabens

2.1.4.1 Hypothetical extensional reconstruction:

An examination of the extensional basin fault geometry which prevailed during late Triassic to early Cretaceous times in the Bourg d'Oisans area will be attempted. To determine the original normal fault geometry, and unravel the timing of the extension, a consideration of the following is necessary:

- 1) The present contractional geometry; including structural styles and the mechanical response of pre-existing extensional faults during the Alpine compression.
- 2) The isostatically induced vertical movements (subsidence and uplift events), governed by extensional fault movements and a mechanical thinning of the crust.
- 3) The corresponding sedimentary response; facies distributions (deep and shallow water sediment relationships), intra-basinal unconformities, detrital input and sediment supply rates, tilting and onlap relationships, olistolith generation and the sealing of faults by later sedimentation.
- 4) The kinematics and dynamics of extensional faulting and basin formation.

Some of the above mentioned aspects are covered in greater detail in Chapter 4.

Figure 2.1.49, is a schematic reconstruction of the Bourg d'Oisans half-graben basin geometry after removing the effects of the Alpine shortening. The distribution of the different sediment types was used to determine the history, and to a certain extent, the geometrical expression of different extensional events.

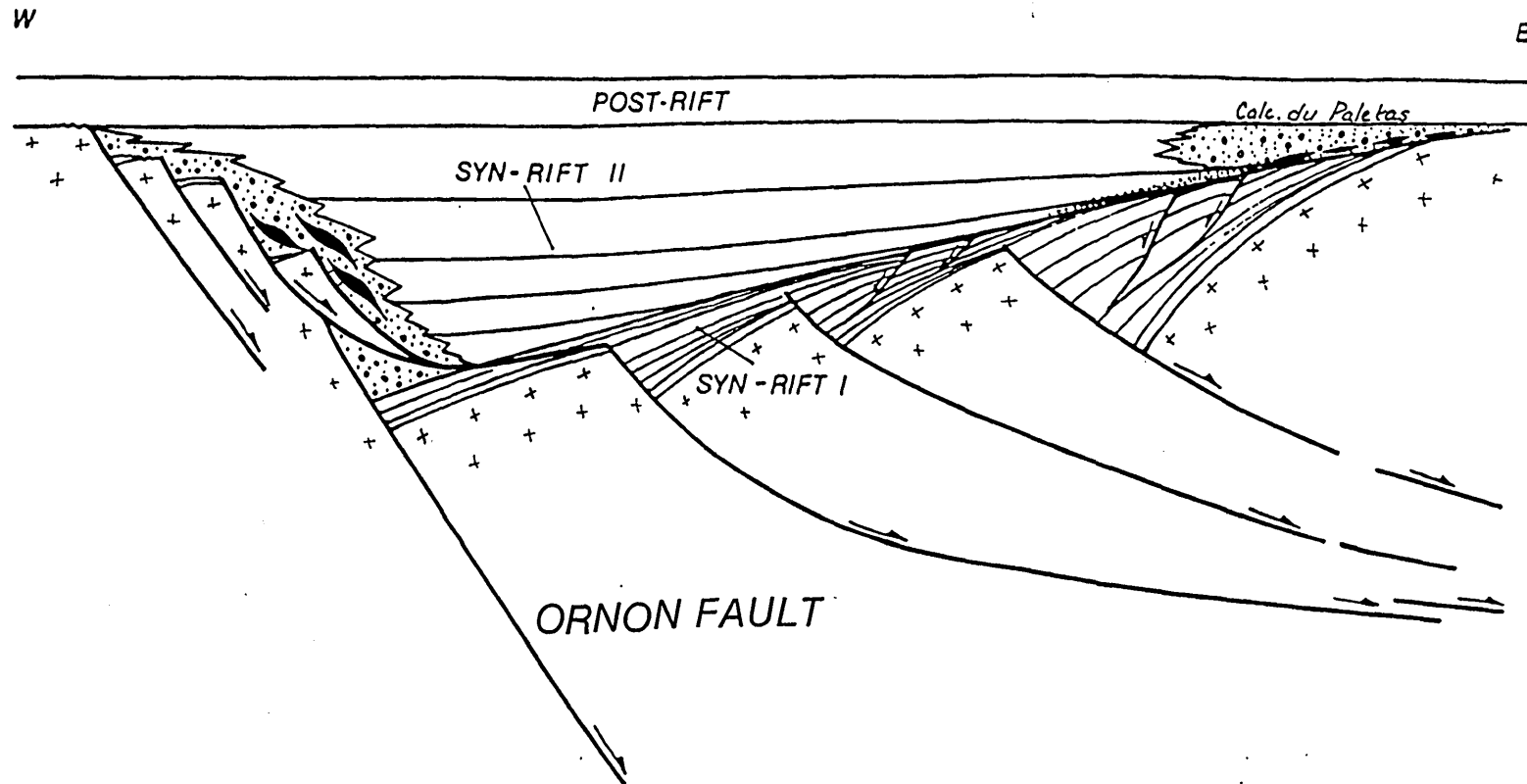


Figure 2.1.49: Schematic reconstruction of the Jurassic normal fault geometry and stratigraphy, for the Bourg d'Oisans half-graben. The Ornon fault is drawn as a steep breaching fault while the geometry of the early gently dipping faults remains speculative. They may be planar or listric in geometry.

The main feature of note on this reconstructed section, is the two contrastingly different geometry of normal fault. This is interpreted to be a direct consequence of two different ages of extensional faulting, an interpretation that is substantiated by sedimentological and local isostasy considerations across this section line. The faults in the eastern part of the section, with lower dip-angles and an inferred shallow depth to detachment, are demonstrably of late Triassic to early Liassic age. The period of maximum subsidence and hence fault activity, associated with the larger and steeper Ornon, and La Mure (not shown) faults, occurs much later than this.

1. Late Triassic to early Liassic extension: Faults identified as belonging to this category include the Armentier half-grabens whose syn-rift infill is predominantly of Triassic age, with some evidence of Lower Liassic growth (Hettangian).

In the southern Bourg d'Oisans area, the La Paletas half-graben which suffered a reactivation and inversion during the Alpine compression, shows little in the way of Triassic growth and the syn-rift infill comprises Lower Liassic age sediments (Hettangian and Sinemurian). A restoration of the present contractional geometry shows that the half-graben sediments are sealed by post-Sinemurian age sediments. Other extensional faults of the same age, may include the Lac du Vallon faults, although the available evidence cannot confirm this. The necessary basement thrusts, inferred to resolve space problems in the lower levels of the constructed cross-sections (Section 2.1.3.3) and explain the steep hanging-wall dips to the Ornon fault, may equally correspond to older extensional half-graben faults reworked as thrust faults during the Alpine compression. These faults would have dipped gently to the SE and be of late Triassic to early Liassic age.

It is impossible to say conclusively whether any sedimentary growth of this age occurred on the steeper Ornon and La Mure faults, the critical lower syn-rift

sediments being located in the present subsurface. It is possible that these faults were initiated much, or slightly later, so that they may, or may not, record Lower Liassic growth on their hanging-walls.

2. Late Liassic to early Dogger extension: Subsidence related to extensional movements on steeply dipping, as opposed to the gently dipping high level faults, is invoked to explain the rapid deepening recorded by deep water pelagic sediments in the Bourg d'Oisans and La Mure half-grabens. Both structural and sedimentological arguments provide evidence that this increased subsidence was accommodated by extensional movements on the Ornon and La Mure faults which at the time possessed steep Mesozoic dips.

i) The large subsidence anomaly: The sedimentary record quite clearly records a rapid and sudden deepening around the Toarcian-Aalenian boundary (Barfety, 1985). The N-S trend of deep and shallow water facies in the Bourg d'Oisans and La Mure half-graben basins, clearly relates this sudden deepening to extensional movements on the Ornon and La Mure faults. Because we are dealing with a sediment starved basin, this sudden rapid deepening resulted in the lowering of the basin floor to beneath the CCD; marked by a transition from calcareous to non-calcareous sedimentation.

With this rapid subsidence the existing basin topography became sub-merged, and the earlier gently dipping extensional faults were sealed by the progressive eastward directed onlap of deeper water pelagic shales. This reflects the broad wavelength (and high amplitude) subsidence effects related to extension on the Ornon and other similar age faults.

- A number of sedimentological and tectonic features observed in the hanging-wall sediments of these larger half-grabens, unequivocally demonstrate that this dramatically increased rate of vertical subsidence is of tectonic origin. In the hanging-wall sediments of the

Ornon fault, a number of minor structures are generated in response to the increased hanging-wall dips developed during extension and fault block rotation. Tilting and onlap relationships, Carixian age olistoliths (Barfety & Gidon, 1983), sudden increases in detrital sediment supply and calci-turbidites all reflect foot-wall uplift, increased basinward palaeoslope dips and active extensional faulting. Gravity collapse structures also reflect these increasing hanging-wall dips.

Barfety & Gidon (1983) demonstrate a Toarcian age for the olistoliths located along the base of the Ornon fault scarp. The similar age of the transition from shallow water to deeper water sedimentation and olistolith formation, is not considered coincidental, and both can be attributed to active extensional fault movements. Seismic aftershock may have provided the necessary triggering mechanism for both the gravitational detachments initiated down the hanging-wall dip-slope of the half-graben and the formation of olistoliths derived from the foot-wall fault scarp.

These later faults are interpreted as belonging to a later breaching family of extensional faults (see Chapter 4). Unfortunately, insufficient data restricts a determination of the relative ages of fault movement between the Ornon and La Mure faults, thereby constraining the direction of fault propagation and the collapse sequence of this later extensional fault set.

ii) Foot-wall uplift: From the stratigraphy described by Barfety & Gidon (1983) and from field observations, the apex of the Ornon fault block was subjected to an important vertical uplift and erosion. This is interpreted to be the result of local isostatic effects (Jackson and Mckenzie, 1983; Barr et al., 1985; Barr, 1987), and is more commonly referred to as foot-wall uplift.

From Section 2.1.2, the Calcaire du Petit Renaud and Calcaire du Paletas shallow water clastic limestones which range in age from Lower Callovian to Upper Oxfordian, rest

unconformably on the pre-rift basement (see Figure 2.1.49). This uplift and erosion is attributed to foot-wall uplift of the Bourg d'Oisans fault block. The adjacent fault which bounds this block is tentatively interpreted to be the La Muzelle fault, or is now obscured by Alpine basement thrusting along the western Rochail. The earlier faults rotated to gentle dips during extension are also affected by this same episode of foot-wall uplift and erosion, supporting the younger age interpretation of the more widely spaced, and steeper dipping extensional faults.

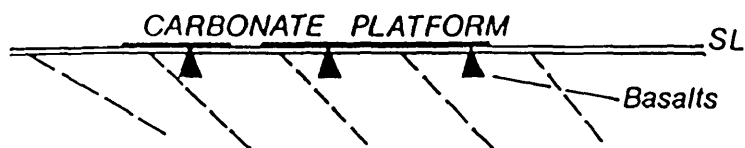
From these relationships, the uplifted foot-walls of the breaching extensional fault system of Toarcian-Aalenian age, belong to two, of three categories described by Barr (1987):

- 1) The foot-walls were uplifted above sea-level, eroded and then subsided below sea-level.
- 2) The foot-walls were uplifted above sea-level and remained above sea-level throughout the syn-rift stretching phase, and only subsided during the thermal re-equilibration of the margin.

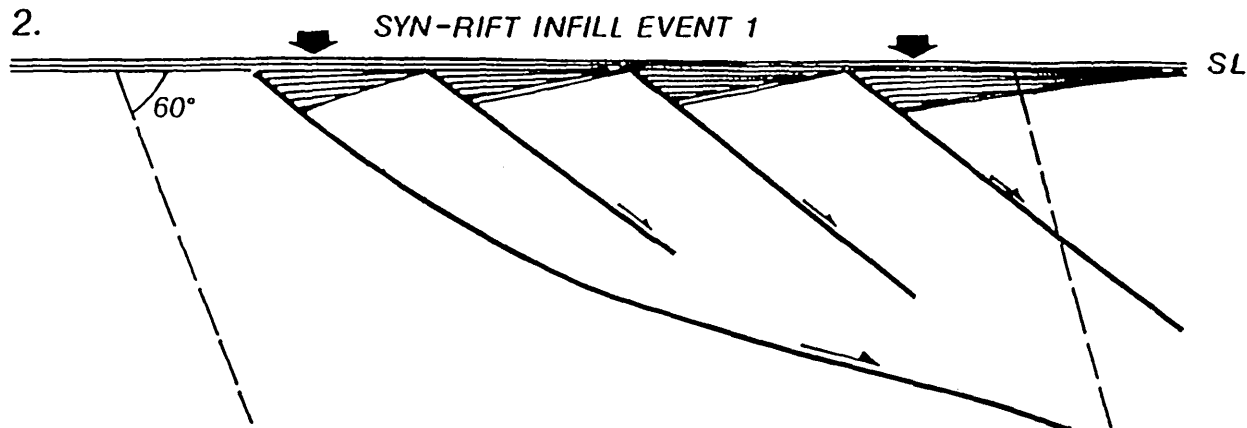
As the post-rift sediments have now been largely removed by erosion from the critical area, it is impossible to choose between these two possibilities. However should the age dating of the Calcaire du Petit Renaud and Calcaire du Paletas prove correct, the second of the two possibilities appears to be the most likely. The degree and nature of the foot-wall uplifts can have important implications for the initial fault geometry, the depth to detachment, the amount of crustal stretching and the original spacing between faults (Barr, 1987).

The evolution of the two fault systems, interpreted to belong to two different extensional pulses operative during the syn-rift stretching phase, is depicted schematically in Figure 2.1.50. A more complete discussion of the history of syn-rift extension and the geometry of syn-rift faulting is

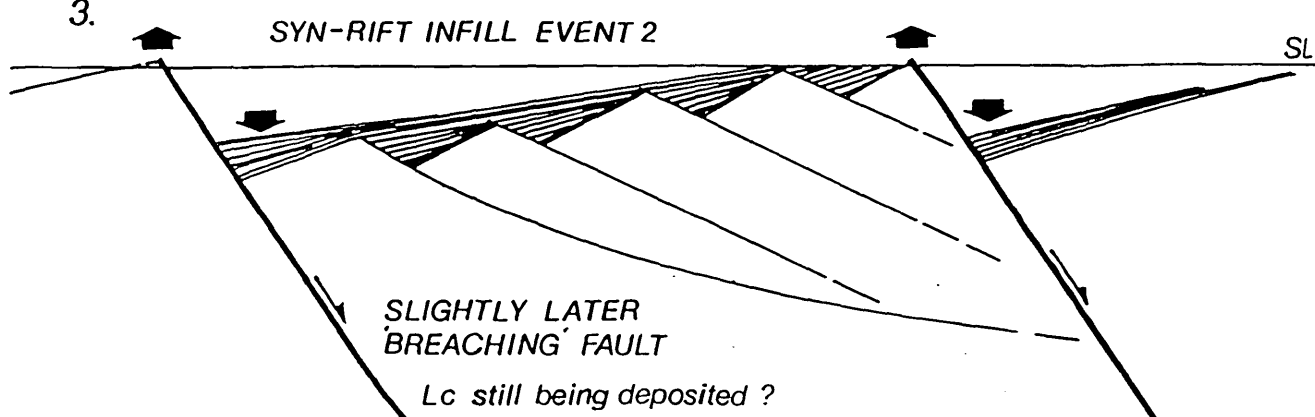
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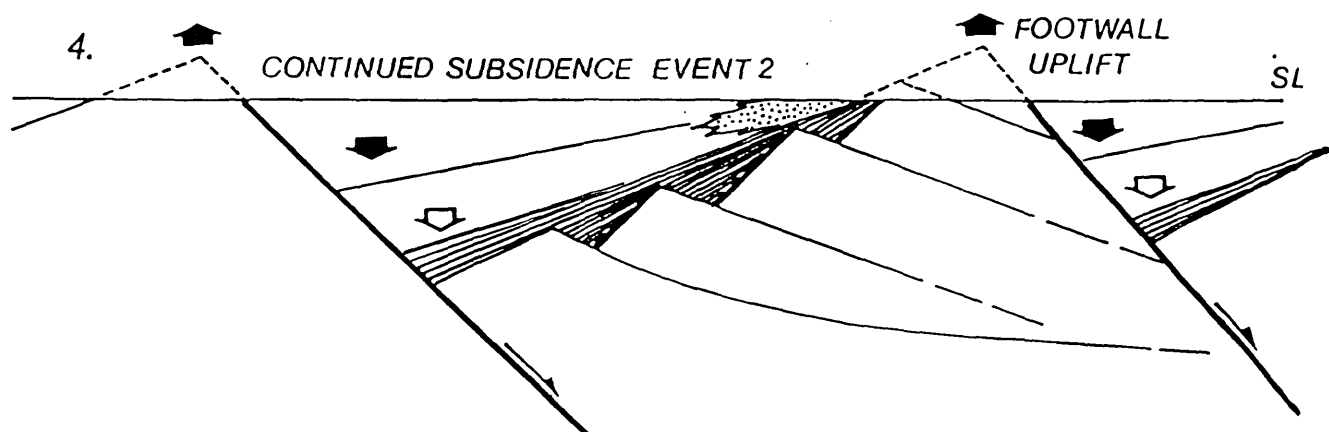
2.



3.



4.



0 1km
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Figure 2.1.50: Suggested structural and stratigraphic evolution of the Bourg d'Oisans half-graben. 1) and 2), late Triassic to early Liassic fault system initiated, 3) and 4), late Liassic steep breaching faults become active and a larger half-graben basin develops.

given in Chapter 4.

3. Extension direction:

The N-S alignment of the deep and shallow water ^e sediments adjacent to the Ornon and La Mure faults and the occurrence of NW -SE trending transfer faults, such as the La Sue fault and the Lac du Vallon faults, suggests that the dominant Jurassic extension direction was NW-SE, parallel to the strike of these transfer faults. Information concerning the late Triassic to early Liassic extension direction is more limited, although the arrangement of the faults identified as belonging to this phase suggests that the extension direction was also NW-SE.

2.1.4.2 Models for the compressional evolution:

Three separate models can be proposed to explain the observed shortening of the syn- and post-rift cover sequence, these are:

1. Gravity sliding model.
2. Basement fault reactivation model (excess bedlength problem).
3. Thin-skinned rearward compression and buttressing model.

The plausibility of each of the above models will be assessed.

1. Gravity sliding model:

During lithospheric stretching, if the upper crustal extension is accommodated on a series of planar faults (otherwise known as the "domino model"; described by Morton & Black 1975; Le Pichon & Sibuet 1981; Wernicke & Burchfiel 1982; Jackson & McKenzie 1983), a direct and inevitable consequence is the rotation of both fault planes and the blocks they bound, thereby maintaining strain

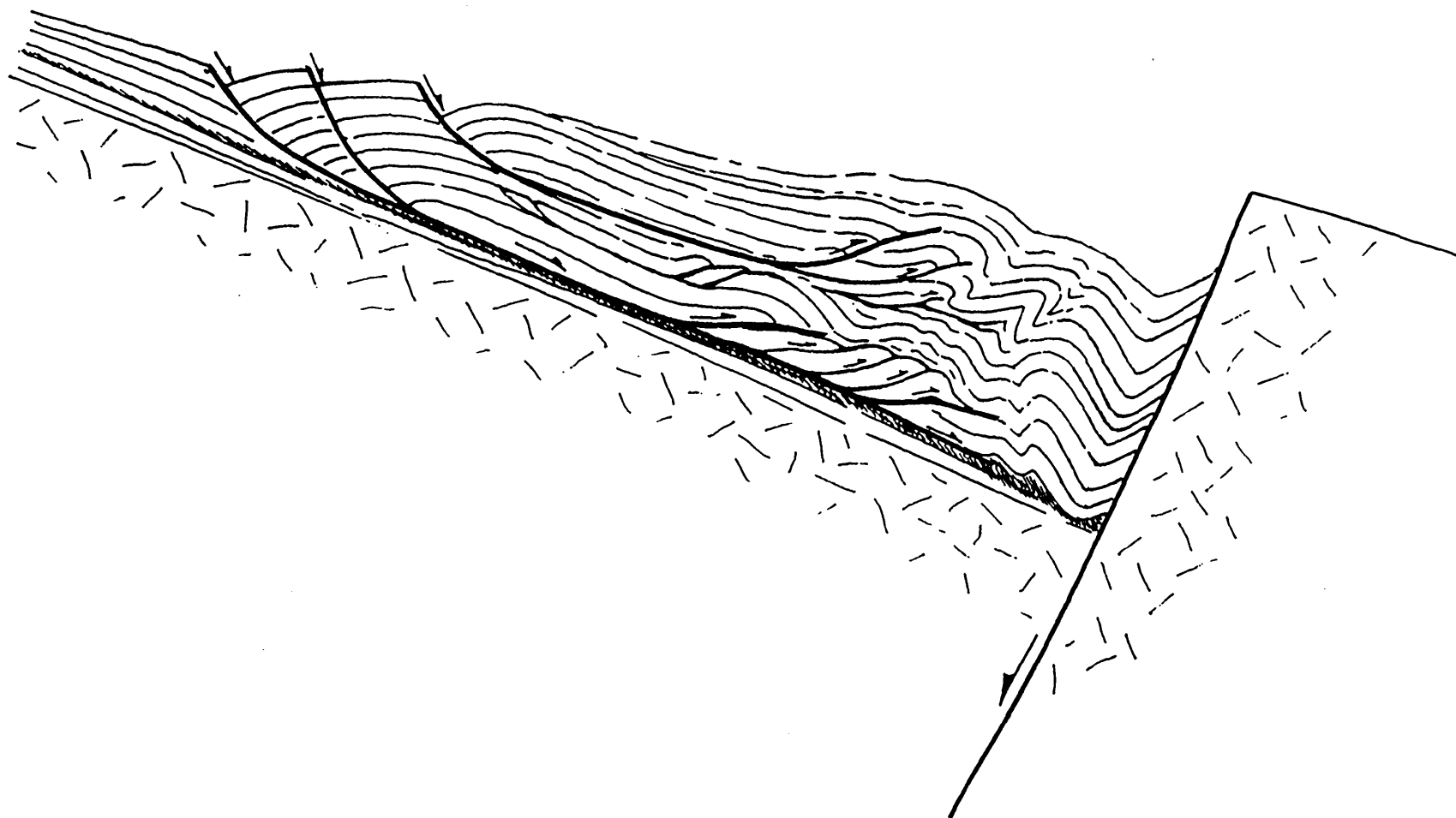


Figure 2.1.51: Schematic diagram to show gravity sliding of the syn-rift sequence down the hanging-wall half-graben dip slope. Extension in the upper part of the half graben must balance the contraction adjacent to the normal fault scarp.

compatability across the stretched zone (Jackson, 1987).

Rotational fault blocks are characterized by a progressive tilting of the hanging-wall basement-cover interface such that with increasing values of β (as defined by Mckenzie, 1978), the hanging-wall dip angle becomes more severe. Should this basinward dip attain a critical slope angle, failure may occur, initiating a gravity sliding of the syn-rift sedimentary sequence down the half-graben hanging-wall slope (Figure 2.1.5.1). This type of model may explain some of the gravity structures and shortening observed in the Ornon and La Mure sedimentary basins.

i) Mathematical description:

The mechanics of gravity sliding down a half-graben slope can be analysed mathematically, by adapting some of the arguments used to model the motion of thrust sheets and thrust accretionary wedges (e.g. Hubbert & Rubey, 1959; Kehle, 1970; Chapple 1978; Davis et al., 1983).

Hubbert & Rubey (1959) used the reduced friction theory, whereby abnormally high pore fluid pressure reduce the effective load stress (σ_n), and having no effect on the tangential shear stress (τ_b), aid the gravity sliding of a rigid mass down a gently inclined plane.

Assumming that failure along the basal plane is controlled by the Mohr Coulomb criterion:

$$\tau = C_0 + \mu \sigma_n \quad (1)$$

where τ is the shear strength of the rock, σ_n is the normal stress acting on the slide plane, C_0 the cohesion of the slide plane, μ the coefficient of internal friction ($\mu = \tan \theta$), this can be modified to take into account the effects of pore fluid pressures by using the criterion of Terzaghi (1923) (see also Hubbert & Rubey, 1959; Hobbs et al., 1976; Mugnier & Vialon, 1986) and

equation (1) becomes:

$$\tau = C_0 + \mu \sigma_n^l \quad (2)$$

where σ_n^l is the effective normal stress ($\sigma_n - P$ fluid). σ_n^l can also be represented by:

$$\sigma_n^l = (1 - \lambda) \sigma_n \quad (3)$$

$$\text{where the ratio } \lambda = \frac{\text{hydrostatic pressure}}{\text{lithostatic pressure}} = \frac{\rho_w g H}{\rho g H} \quad (4)$$

The ratio λ , for average rock types and burial depths generally lies between 0.3 and 0.5, however, Hubbert & Rubey (1959), document λ values of 0.9 from exploratory wells. Substituting (3) into equation (2) gives:

$$\tau = C_0 + \mu (1 - \lambda) \sigma_n \quad (5)$$

Following on, from Figure 2.1.52, the rigid mass induces a tangential shear stress (τ_b) on the basal plane:

$$\tau_b = -\rho g H \sin \theta \quad (6)$$

If the gravitational forces induce a shear stress along this basal surface that is equal to or greater than the shear strength under the natural conditions of the rock (Elliott, 1976 b), then failure and sliding will occur, that is if $\tau_b = \tau$, or $\tau_b > \tau$. In this event:

$$-\rho g H \sin \theta > K_b \quad (7)$$

where the value K_b is the yield strength of the basal layer. In the model developed here K_b may either correspond to a plastic limit (ductile deformation) or a rupture limit (brittle deformation) (Figure 2.1.53). This relationship may also be expressed in the form:

$$-\rho g H \sin \theta = C_0 + (1 - \lambda) \sigma_n \quad (8)$$

where, following Hubbert & Rubey (1959), the value of C_0 is taken to be zero.

ii) Modification of the model:

The unmodified application of the reduced friction sliding theory to the gravity detachment of syn-rift sediments down the half-graben slope, is unrealistic because the mechanical details of the half-graben sedimentary wedge do not conform to the mechanical rigid mass envisaged by Hubbert & Rubey (1959). Instead the typical half-graben sedimentary wedge, which thickens towards the basin bounding fault and thins onto the hanging-wall dip slope, is characterized by a number of mechanically weak bedding planes. This effect is often amplified by the alternating competent-incompetent bedding arrangement (limestone-shale sequences, or deltaic clastic-shale sequences) which is often characteristic of the cyclic sedimentation patterns found in developing half-graben basins. Furthermore, if the sedimentation is rapid, as is often the case, this condition will favour the development of overpressured horizons, not only at the base, but also within the half-graben sedimentary sequence increasing the potential for bedding-parallel detachment.

Therefore, in addition to the existence of a mechanically weak basal layer, parts of sedimentary cycles or individual bedding planes may constitute important mechanical discontinuities along which the potential for sliding exists. In the modified model, it is necessary to examine this bedding-parallel slip deformation, and each bedding plane is modelled as a mechanical discontinuity which dips basinward and defines gently tapering elongate sedimentary wedges which thicken towards the extensional boundary fault (Figure 2.1.53). Each bedding plane discontinuity becomes increasingly more inclined with depth - a consequence of the cumulative subsidence and rotation suffered by each bed once deposited, thereby increasing the potential for detachment in the lower, deeper levels of the half-graben basin. By taking into account these bedding

plane discontinuities it is possible to model the general plastic behaviour of the sedimentary prism independent of any penetrative internal deformation.

For each bedding plane discontinuity it is assumed that sliding is governed by the Mohr coulomb criterion for failure, and like Hubbert & Rubey (1959), the cohesive strength contribution is considered to be zero (see also Davis et al., 1983) and is therefore neglected. This may or may not be a valid assumption, and indeed Hsu (1969) claims that it is not; later disputed in a reply by Hubbert et al. (1969). The stresses operative along each discontinuity plane within the sedimentary wedge can be calculated from:

$$\tau_b = -\rho g H \sin \theta_x \quad (9)$$

where the θ_x value varies with depth i.e. for each bedding discontinuity the θ_x value will increase, as will the H value and the σ_n value. σ_n can be calculated from:

$$\tau = C_0 + \mu \sigma_n^I \quad (10)$$

and assuming C_0 to be zero, then:

$$\sigma_n^I = \tau / \mu \quad (11)$$

Substituting for σ_n^I in equation (11) gives:

$$(1-\lambda) \sigma_n = \tau / \mu \quad (12)$$

and:

$$\sigma_n = \tau / \mu (1-\lambda) \quad (13)$$

where the coefficient of friction (μ) characteristic of each sliding plane relates the values of the normal and tangential shear stresses during sliding so that:

$$\tau = \tan \theta_x \sigma_n \quad (14)$$

RIGID SLIDING MODEL - No internal deformation

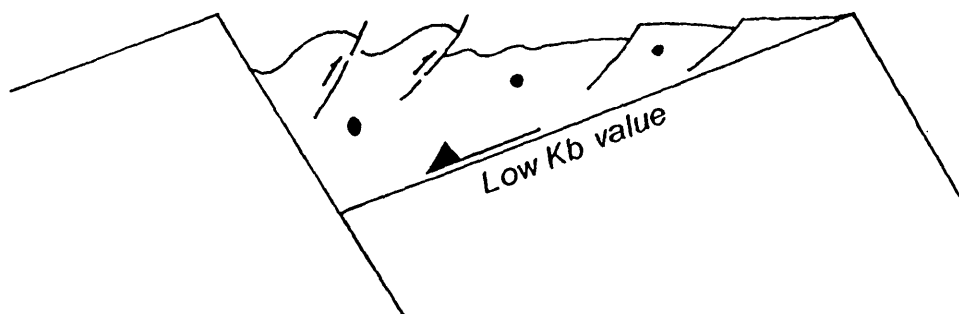


Figure 2.1.54: Gravity sliding down a half-graben slope along a weak basal layer with minimal internal deformation of the half-graben infill.

DISTRIBUTED SHEAR MODEL - Displacement by internal deformation

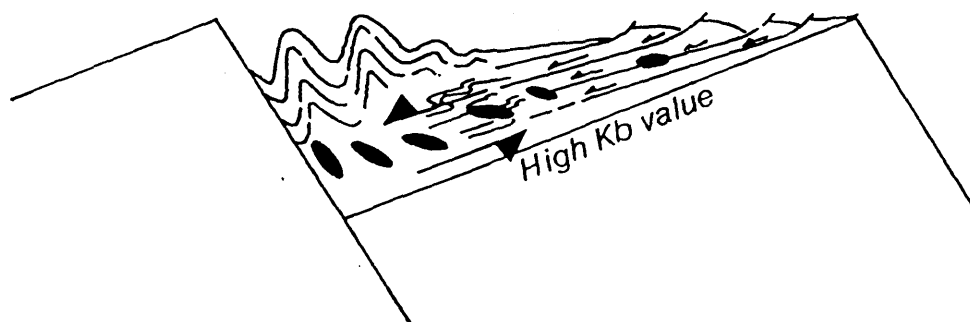


Figure 2.1.55: Gravity sliding down a half-graben slope where the shear deformation is distributed across potential detachment horizons within the half-graben sequence because a weak basal layer is missing.

If the tangential shear stress exceeds this value sliding may commence along the bedding planes.

From this simple model it may be possible to relate the degree of internal deformation caused by gravity sliding down the half-graben slope, to the local extension factor (β) across the basin which controls the magnitude of the hanging-wall dip, for different mechanical parameters from natural half-graben settings.

In the situation where the mechanical properties of the basal plane favour detachment at low hanging-wall tilt values, i.e the basal yield strength value (K_b) is low due to the existence of a perfectly plastic material such as salt, or the sediments at the base of the syn-rift sequence are abnormally overpressured, detachment and the majority of displacement will be accommodated along this weak basal layer, and bedding-parallel decoupling at higher levels within the syn-rift sequence will be kept to a minimum (Figure 2.1.54). However, should this basal layer have a high yield strength value (K_b), gravity sliding may only be initiated at larger stretch values (β), and the majority of the shear deformation will be more distributed, accommodated by numerous zones of bedding-parallel slip and decoupling throughout the syn-rift sedimentary sequence (Figure 2.1.55).

iii) Burial depth and overpressuring effects:

Because of the tapering nature of the sedimentary wedge it is important to consider the increasing depth of burial effect towards the deeper parts of the half-graben basin. An increase in burial depth will result in a proportional increase in the normal (σ_n) and tangential shear (τ_b) stresses. This coupled with increasing dip values in the deeper parts of the basin (θ_x), will increase the potential for sliding in the deeper parts of the basin (Figure 2.1.56). This will be especially true if the sedimentary formations at lower levels are overpressured, thereby reducing the value of the effective

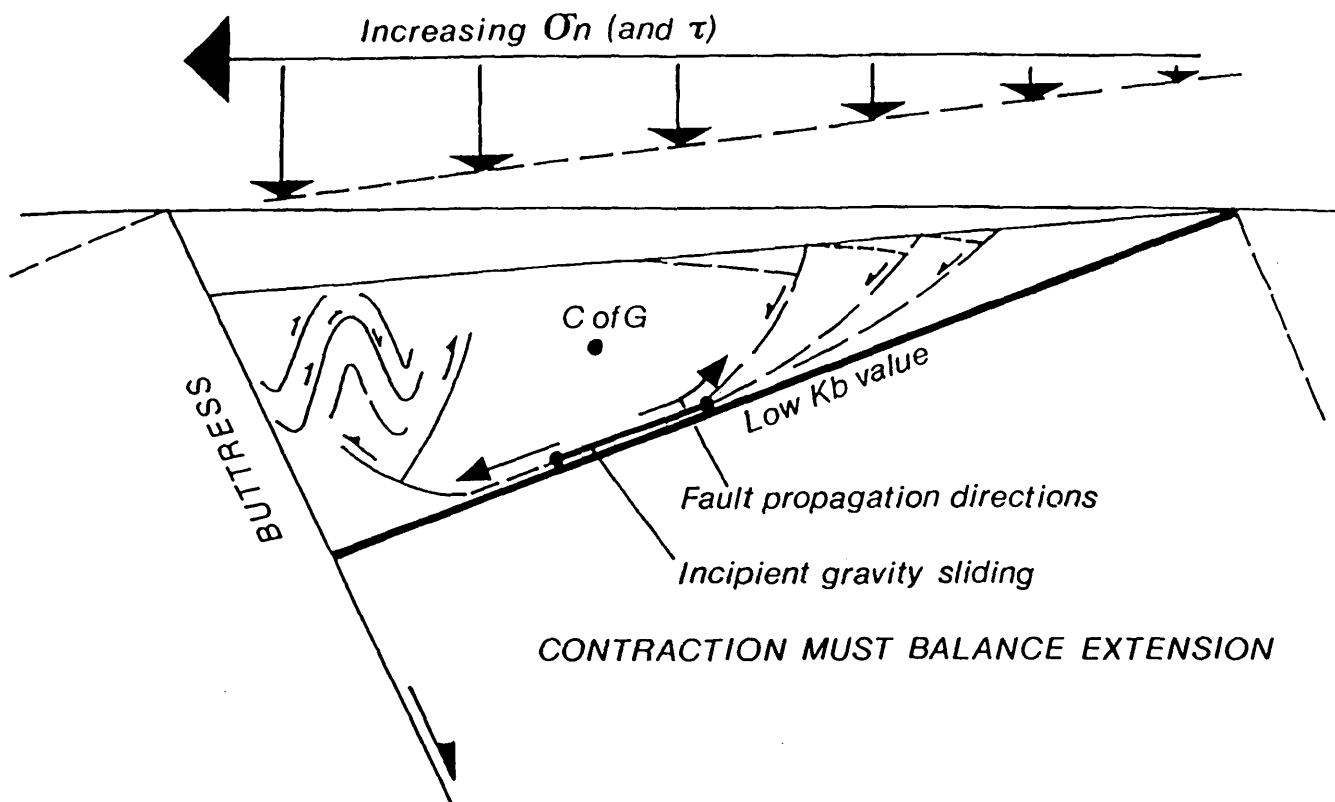


Figure 2.1.56: The general geometry of gravity sliding down a half-graben slope. See text for details.

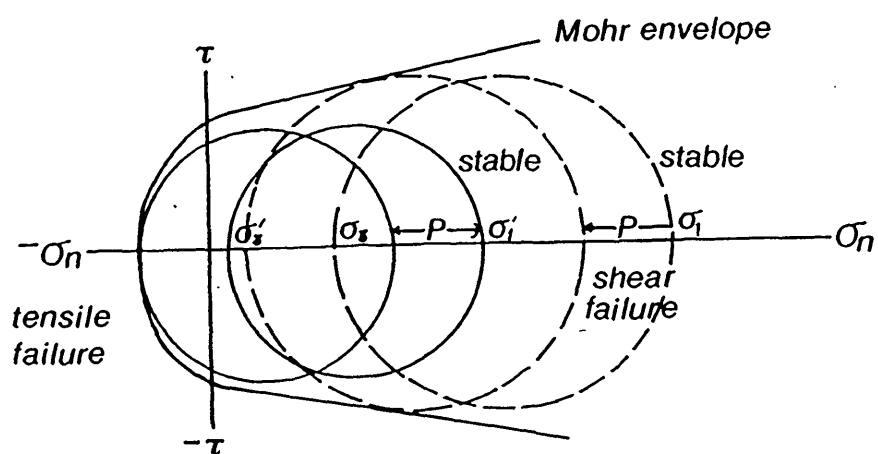


Figure 2.1.57: The effect of an increase in pore fluid pressure is to move the Mohr circle along the σ_n axis by a distance P towards instability and failure (the Mohr envelope is shown to follow the Griffith criteria for failure).

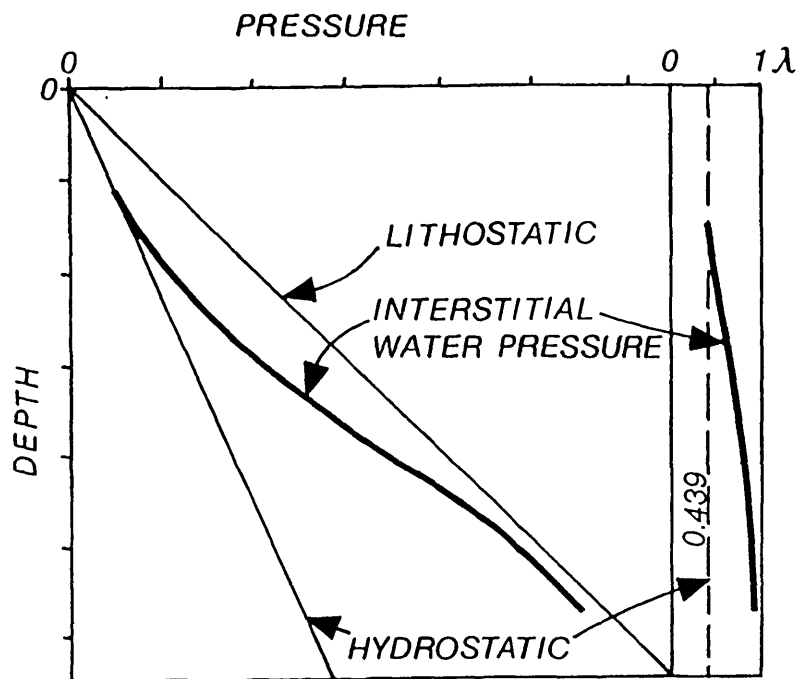


Figure 2.1.58: Variations in pore fluid pressure with depth. λ = pore fluid pressure/lithostatic pressure (after Davies et al., 1983).

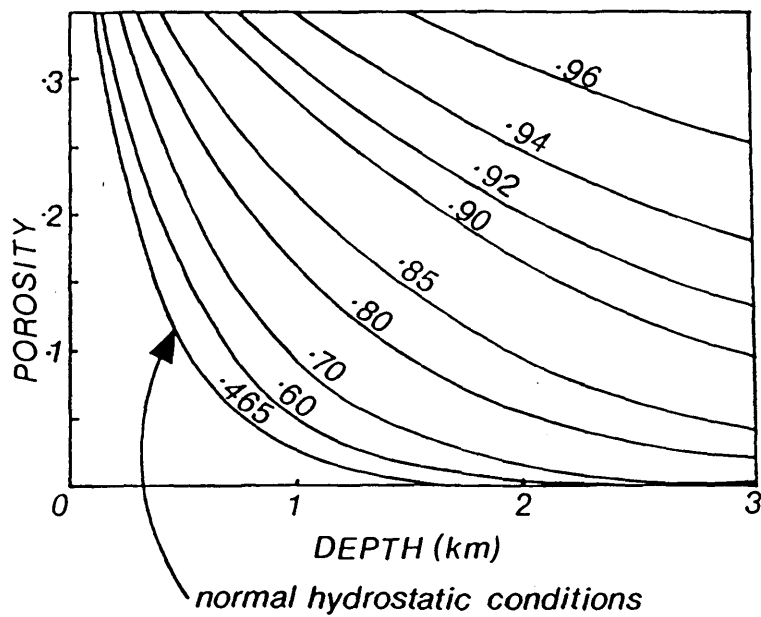


Figure 2.1.59: The relationship between porosity and depth of burial for different values of λ .

normal stress but having no complimentary effect on the value of the tangential shear stress (i.e reducing the resistance to shearing). On the Mohr diagram the effect of an increase in the pore fluid pressure is to move the Mohr circle along the σ_n axis by a distance P towards instability and failure (Figure 2.1.57). The increase in lithostatic pressure and hydrostatic pressure with depth, may also be accompanied by an increase in λ (λ = hydrostatic pressure / lithostatic pressure). This relationship is shown in Figure 2.1.58, while Figure 2.1.59 shows the effect of different λ values on porosity for increasing depths of burial. Abnormal fluid pressures may occur where there are restrictions to fluid flow due to the presence of impervious horizons such as evaporites and compacted shales, and it is well known that an elevation of pore fluid pressures can have a triggering effect on fault movements and sliding, as shown by the Rangely oilfield earthquake experiments (Raleigh et al., 1982). This leads to the notion of potential detachment horizons within a sedimentary sequence (Gretener, 1981). If the potential detachment level corresponds to an evaporitic horizon, the increasing temperature-depth profile will have a similar weakening effect on the shear strength of the potential detachment horizon. From these considerations, and from Figure 2.1.56, it appears that the incipient stages of sliding will be initiated in the deeper, lower sections of the half-graben.

iv) Buttreassing - the end effects:

The all important boundary forces induced by the rigid effect of the basement fault scarp and so far neglected in this analysis, are fundamentally important because they prohibit the forward motion of the detached gravity slide. Thus before gravity sliding can commence, and during gravity sliding, these down dip boundary forces must be overcome. This effect, otherwise known as buttressing (Kehle, 1970), will ultimately control the velocity of the deformation, that is the rate of advance of the gravity slide and the degree of internal deformation. This is

largely limited by the rate at which shortening and folding at the front of the gravity slide, adjacent to the buttress, provides space for the syn-rift material to occupy downslope. Ultimately, the amount of extension in the upper parts of the half-graben should balance the contraction adjacent to the buttress.

v) Kinematics of sliding after initiation:

Once sliding is initiated, the mechanical parameters controlling the dynamics of the deformation will change. For example the centre of gravity of the gravity slide will alter its position, thus removing the potential for further sliding. Also, immediately after initiation of sliding, the coefficient of kinetic friction will be much lower than the coefficient of static friction, facilitating sliding movements in its early stages. However, once the extensional faulting in the upper part of the gravity slide reaches the free surface, a dramatic reduction in the pore fluid pressures will slow down the gravity slide. Thus with the prohibition to forward movement exerted by the down dip fault scarp buttress and the progressive and perhaps rapid dissipation of pore fluids once sliding is initiated, the amount of internal deformation and displacement which can occur by this mechanism is considered somewhat limited.

vi) Triggering mechanism:

In reality the initiation of gravity sliding although strongly influenced by, is not entirely dependant on the critical slope angle being reached, but related to a more violent source of energy, such as the seismicity associated with earthquakes caused by slip on the basin bounding fault. Once sliding is initiated, continued slip will be sustained until the stress is dissipated.

vii) Distributed shear theory:

The deformation along the weak basal layer and potential decoupling horizons in the sedimentary cover may

not correspond to sliding along discrete kinematic discontinuities, but rather the resulting deformation may be distributed across a number of low viscosity zones, which deform by simple rectilinear shear when subjected to loading (Kehle, 1970). With this mechanism several decoupling horizons may be operative simultaneously. Kehle (1970) argues that his model of fluid like shear requires less work to move a thrust mass, than the reduced frictional sliding theory. However this should not invalidate equation (8) (Kehle, 1970; see also Davis et al., 1983).

As should be evident, the phenomena of gravity sliding is very much centred around the respective roles of gravity, surface forces and the mechanical characteristics of the syn-rift sediments. Adapting the approach employed by Elliott (1976; see also Goguel, 1983) it should be possible to calculate the the amount of energy expended during the gravitational emplacement of the rock mass in the Earths gravitational field.

In summary, this analytical model is based upon a number of simplistic assumptions whose validity may be questionable. However, in its present form this model still has many useful applications, especially to half-graben basins in present day extensional settings where the relative extension factor (β) and the hanging-wall dip angles can be determined from seismic data, and the mechanical parameters from well data (variations of rock type and strength, λ values etc.). Large scale examples of gravity sliding have been documented from shelf-edge regions off the South African coast; Dingle (1977) describes collapse structures from the shelf-edge slope of the Orange basin using seismic data.

In the Alpine examples cited here, it is suggested that this mechanism can explain the gravity structures described from the Ornon and La Mure sedimentary basins and may account for some of the shortening and cleavage development of the syn-rift sediments adjacent to the rigid

fault scarps. However, because the later Alpine displacement and deformation is coaxial with respect to the early gravity deformation, it is impossible to determine the extent and amount of shortening that is associated with early gravity sliding down the half graben slope; the later Alpine structures will only serve to amplify any existing sub-vertical cleavage and tighten buckle folds developed adjacent to the basement fault scarp.

For mechanical reasons it is also suggested that this mechanism cannot accommodate large magnitudes of shortening, and some other mechanism must be sought to explain the observed shortening and strain affecting the syn-rift sediments of the Bourg d'Oisans area. This leads us to consider the alternative hypotheses.

2. Basement fault reactivation model (excess bedlength problem):

If at depth the geometry of the Ornon fault plane is listric, during Alpine compression a reverse or oblique reactivation would result in geometrically necessary back-thrusting, upright folding and cleavage development before the early extension on the Ornon fault is cancelled. This is depicted schematically in Figure 2.1.60, and is a direct consequence of the bed lengths at higher levels in a syn-rift growth sequence (L1), being longer than the pre-extension marker (L0) (Figure 2.1.61, after Gibbs, 1987; see also Chapter 3).

Similar structures may also result for different mechanical reasons. For example, with listric normal faults the gently dipping part or flat, during subsequent compression will reactivate with relative ease while a necessary decrease in displacement occurs at the foot of the ramp, or where the listric fault profile increases its dip (see Chapter 3). Should the ramp orientation become too steep, the displacements along the fault may be reduced to zero, and continued compression is accommodated by a dominantly pure shear deformation of the hanging-wall. This

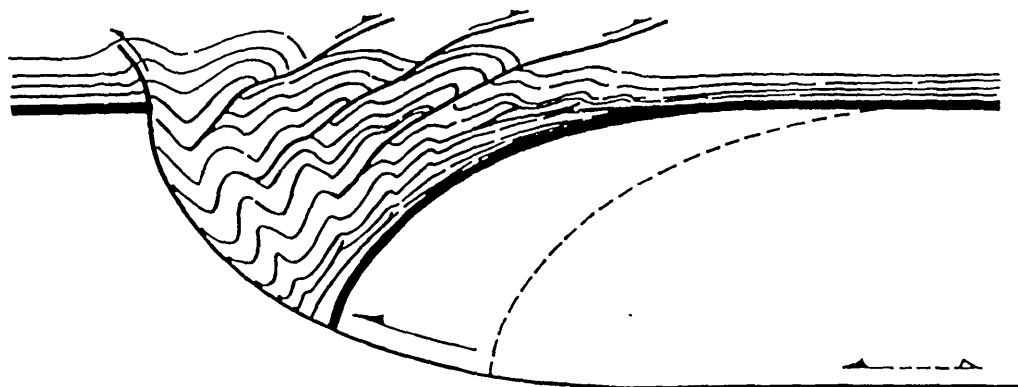


Figure 2.1.60: Intense back-thrusting and penetrative deformation of the half-graben infill during the reverse reactivation of a listric normal fault profile.

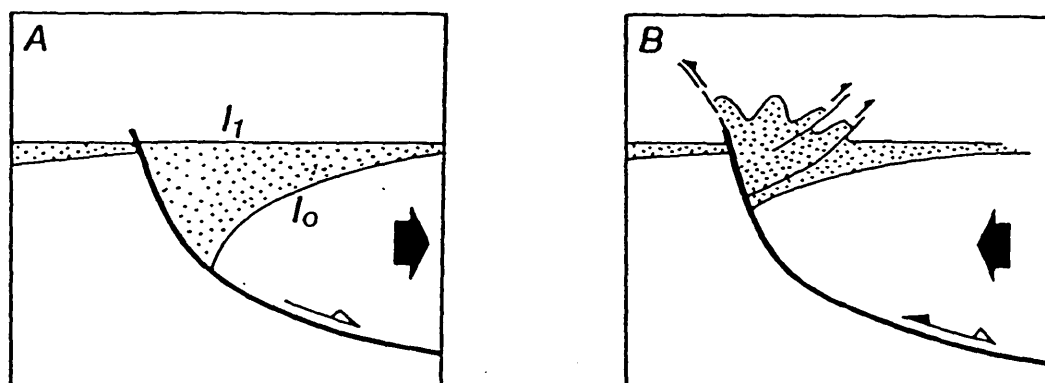


Figure 2.1.61: Folding and back-thrusting of the higher levels of the half-graben infill has been shown to be due to longer bedlengths in the upper parts of the half-graben sequence (after Gibbs, 1987). A. Extension, B. Inversion.

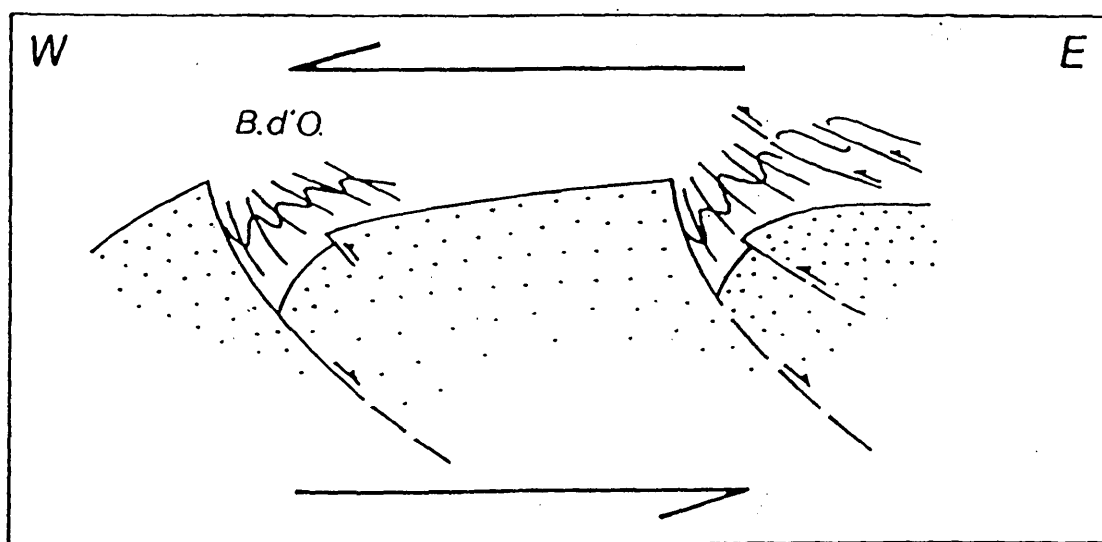


Figure 2.1.62 Illustrates the model proposed by Gratier & Vialon (1980), where the initial half-graben geometries are modified in an overall regime of progressive simple shear.

results in an exaggerated hanging-wall fold in the basement and upright folding and cleavage development in the sedimentary cover. This model is hinted at, although not fully implied, by Gratier & Vialon (1980) who attribute the overall geometry of the Bourg d'Oisans half-graben to a modification of the extensional structures described above, in a regime of progressive simple shear. This model was later reiterated by Tricart & Lemoine (1986) to explain the steep hanging-wall dips to the Ornon fault.

Although this model can explain some of the compressional structures observed to the E of the Bourg d'Oisans area, for the Bourg d'Oisans half graben an alternative explanation of the observed structure is preferred. The main arguments behind this are proposed in the next section. The absence of shear fabrics consistent with a reverse or oblique reactivation along the Ornon fault supports this third and last hypothesis.

3. Thin-skinned rearward compression and buttressing model:

It is proposed that a substantial component of the displacement and internal deformation described from the Bourg d'Oisans syn-rift sedimentary sequence results from a thin-skinned decoupling of the Jurassic cover from its underlying Triassic and pre-Triassic basement, such that the hanging-wall has moved towards the WNW-NW, structurally downwards but stratigraphically remaining at the same level in the half-graben. To the E this cover shortening is clearly associated with the thrust stacking of the Rochail basement, the thrusts of which are both newly formed, and old extensional faults reactivated as thrusts during the Alpine compression. Upon climbing into the overlying syn-rift cover sequence these basement thrusts either branch onto or breach the overlying cover detachment. The displacements from these thrusts which transfer along the base of the syn-rift sequence must die out to the E of the Ornon fault.

i) Buttrressing effect of the Ornon fault:

As in the gravity sliding model, the buttressing effect of the basement fault scarp - that is the prohibition of foreland directed cover thrust displacements, dictates the style and intensity of deformation affecting the syn-rift sediments. However, unlike the gravity sliding model, a horizontal rearward compression provides the driving force, such that the deformation rate and degree of internal deformation are controlled by the rate of advance of the basement thrust sheets to the E, and the corresponding volume reduction occurring within the half-graben sedimentary prism.

These W to NW directed displacements of the cover sequence are therefore accommodated by:

1. Discrete slip along a basal syn-rift detachment.
2. The accumulation of large strains in the form of ductile folding and penetrative cleavage development.

The resistance due to buttressing is characterized by a distinctive structural style and a progressive increase in strain intensity approaching the Ornon fault. Immediately adjacent to the Ornon fault the cover shortening is expressed in the form of tight upright folds with a penetrative slaty cleavage and pronounced down-dip mineral stretching lineation, the latter recording important horizontal contraction and vertical extension of the cover adjacent to the fault scarp (see Figure 2.1.65). These folds and continued slip along the cleavage discontinuity planes, must accommodate some of the thin-skinned displacement of the sedimentary cover, which became blocked against the mechanically rigid basement fault scarp. This resulted in a dominantly pure shear deformation adjacent to the buttress, where the cover sediments remained locally autochthonous (Tricart & Lemoine, 1986). The general axial trend of these folds is tangential to the basement fault scarp and they probably

tightened earlier folds related to Jurassic gravity sliding.

Away from the fault these tight upright folds are more open and verge towards the E, forming a complex back-fold-thrust system which accommodates the remainder of the foreland directed displacements within the closed system of the Jurassic half-graben.

ii) Interaction between "buttressing" and "wedge" tectonics:

One problem for which no adequate explanation has yet been provided, is the present steep dips of the Triassic pre-rift sequence in the hanging-wall to the Ornon fault. Dips in excess of 40-45° were recorded along the Paletas and Malsanne sections.

Assuming a planar fault geometry for the Ornon and La Mure faults, the degree of hanging-wall dip after extension (β) is a simple function of the amount of displacement and rotation occurring on a system of planar rotational faults during extension (see Figure 2.1.63). The relative extension (β) is expressed by the simple relationship:

$$\beta = \frac{L1}{L0} = \frac{\sin \phi}{\sin(\phi_i - \theta)} \quad (15)$$

(after Mckenzie, 1978; see also Le Pichon & Sibuet, 1981; Jackson & Mckenzie, 1983), or alternatively, following Wernicke & Burchfiel (1982), the % extension can be represented by:

$$(L1 - L0)100 = \left[\frac{\sin(\phi + \theta)}{\sin \phi} - 1 \right] 100 \quad (16)$$

or,

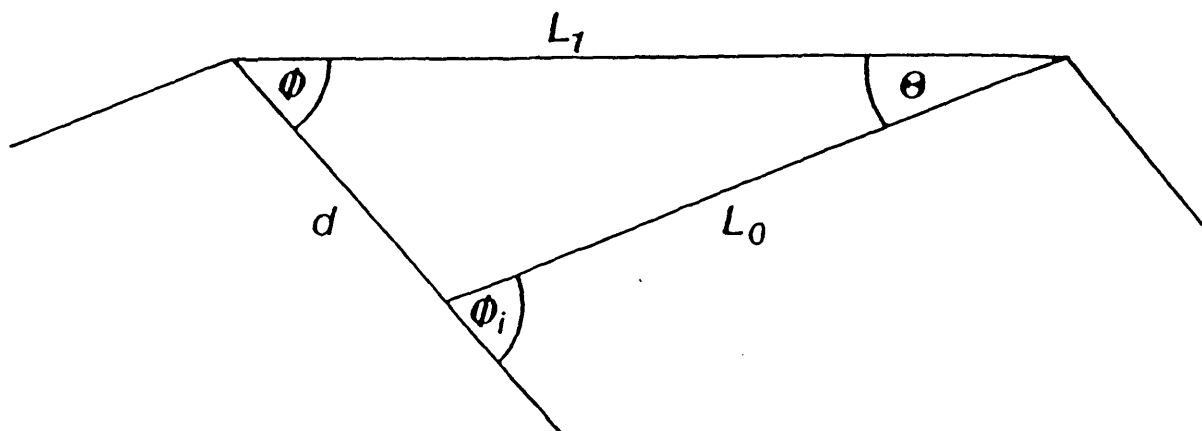
$$\left[\frac{\sin \phi_i}{\sin(\phi_i - \theta)} - 1 \right] 100 \quad (17)$$

Assuming that the Ornon fault was initiated at a typical

Andersonian angle of dip ($\phi_i = 60^\circ$), and that the dip preserved in the hanging-wall of the Ornon fault corresponds to the original Jurassic tilt angle of the hanging-wall fault block ($\theta = 40^\circ$), from simple trigonometry the corresponding dip of the Ornon fault plane with the horizontal (ϕ), should be in the order of 20° . This is clearly not the case, and although it is possible that the Ornon fault suffered a back rotation and steepening during subsequent Alpine compression, this seems unlikely as both the foot-wall and hanging-wall blocks are obliged to suffer the same passive angular rotation, thereby preserving the value of (ϕ_i); unless the foot-wall block were massively strained internally. Furthermore, assuming the above dimensions are the original Jurassic ones, from equation (15), this would give an extension value (β) of 2.5 or 150 % extension (see Figure 2.1.64). This estimate is considered too high for the Dauphinois basin given its position on the European passive margin, relative to the Ligurian Tethys (see Chapter 4). An alternative explanation must be sought for these steep hanging-wall dips.

Following Gratier & Vialon (1980), Tricart & Lemoine (1986) attribute these steep hanging-wall dips to a folding of the basement-cover interface, incurred during the attempted reactivation of the main Ornon fault. However, the findings here show that:

1. The shortening of the syn-rift sediments in the Bourg d'Oisans half-graben results from a buttressing of important horizontal displacements, whereby the Jurassic cover is decoupled from its underlying basement to the E.
2. The steep hanging-wall dips to the Ornon fault do not derive from the folding of the basement-cover interface during the attempted reactivation of the Ornon fault, but record the steep dips of the fore-limbs of basement anticlines which formed passively over foot-wall thrust ramps in the basement. These basement thrusts are responsible for the decoupling of the Jurassic cover



$$\beta = \frac{L_1}{L_0} = \frac{\sin \phi_i}{\sin(\phi_i - \theta)}$$

Figure 2.1.63: Geometrical relationship between the amount of extension (β), the initial fault dip (ϕ_i) and the amount of bed rotation.

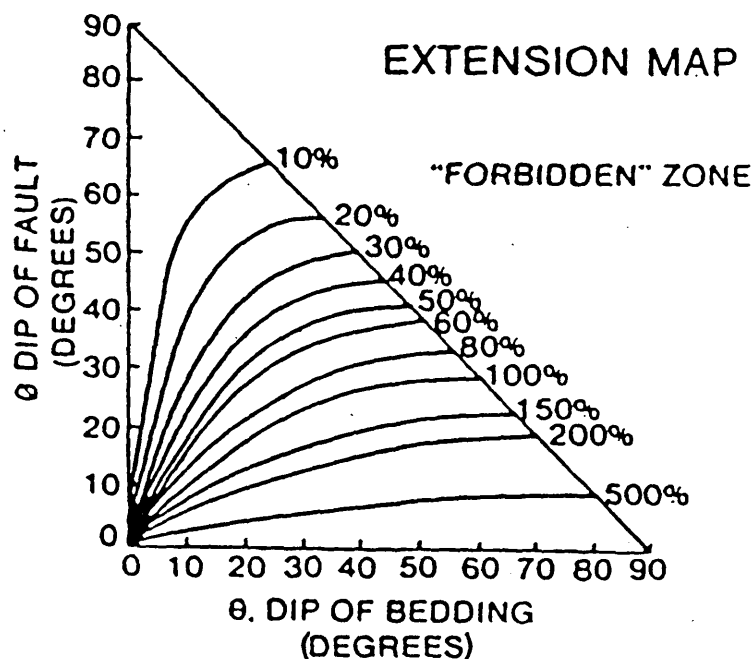


Figure 2.1.64: Graphical representation of the amount of fault dip (ϕ), bedding dip (θ), and % extension (from Wernicke & Burchfiel, 1982).

sequence.

With the horizontal detachment of the sedimentary cover and the prohibiting effect of the Ornon buttress, excess bed length rapidly accumulates in the hanging-wall of the cover detachment. This excess bedlength is accommodated by:

1. Ductile folding and penetrative internal deformation of the cover sequence.
2. Back-thrusting along the upper surface of the newly emplaced basement thrust wedges.

This last mechanism is sometimes referred to as "tectonic wedging" (Price, 1986). The phenomena of tectonic wedging has been described experimentally under conditions of rigid-plastic behaviour, where intersecting conjugate shear zones delineate mechanical wedges which force the outside layers to part and delaminate (Paterson, 1958; Heard, 1960; Price, 1986).

In the field the diagnostic feature of tectonic wedging is a reversal in the vergence or displacement direction, between the base and the upper surface of the wedge. In the Bourg d'Oisans example the wedge comprises mechanically rigid crystalline basement, which is detached and driven into the weaker surrounding sedimentary cover. The basement thrust wedge is bound above and below by zones of reverse faulting or shear. The westward-directed displacements from these basement thrusts sole out along the base of the syn-rift cover sequence and a number of eastward-verging back-folds and thrusts splay from this sole detachment. The locus of the back-thrusting is therefore controlled by the mechanical difference between the rigid basement and weaker sedimentary cover, as the former forces the overlying strata up and over the driving thrust wedge. The zone of back-thrust shear is characterized by either a very weak back-thrust fabric, as observed along in the Oulles-Huez section (Figure 2.1.15),

or a more penetrative fabric with several successive phases of back-thrusting and folding with complex overprinting relationships, as occurs in the southern sector of the Bourg d'Oisans half-graben (Figure 2.1.37). These variations in style are obviously dependant on the magnitude of the fore-directed displacements entering the confines of the half-graben basin from basement levels. In such a system it is necessary to balance the displacements such that:

displacements entering = folding + internal strain + back-thrust displacements.

The mechanics and problems of displacement compatibility are essentially the same as those described for a passive-roof duplex (Banks & Warburton, 1986). The geometric and kinematic components of this system are shown in Figure 2.1.65.

iii) Ornon fault as a stress riser:

In regions of cover detachment early normal faults will disrupt a simple layered stratigraphy, causing the localisation and positioning of thrust ramps, as demonstrated from field and seismic studies in the Idaho-Wyoming thrust belt and from 2-D photoelastic modelling (Wiltschko & Eastman, 1983). With steep basement faults a zone of strong stress concentration occurs immediately in front of the basement fault. The thrust ramp may be generated from this zone, such that the displacements bypass the obstacle to continued forward propagation, removing the need for the build-up of large displacements and the generation of complex back-thrust systems. This relationship, with the cover thrust displacements ramping over the basement fault scarp, is the observed along the Oulles-Mizoen section (Figure 2.1.15), where the thrust ramp slices through incipient stages of a back-fold system, developing in response to the buttressing effect of the Ornon fault.

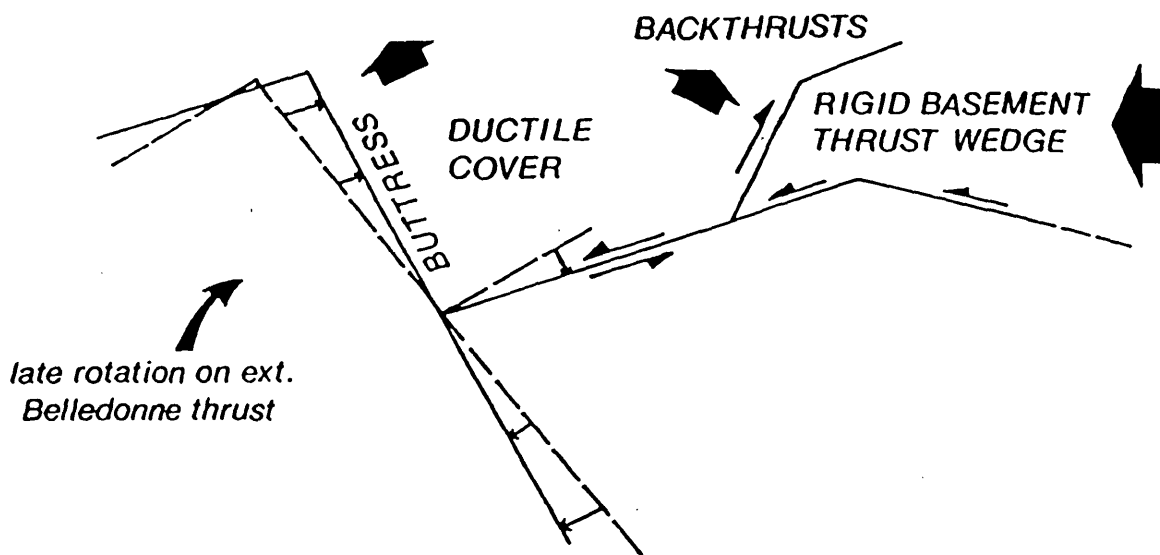


Figure 2.1.65: Geometrical and kinematic components of tectonic wedging and buttressing, with particular reference to the Bourg d'Oisans half-graben example.

2.1.4.3 Sequence of basement-cover thrusting:

Generally, the sequence of thrusting appears to have been directed towards the more external parts of the thrust belt. The Jurassic cover sequence is interpreted to have detached in the foot-wall to a basement thrust sheet which, because of the buttressing effect of the Eastern Grandes Rousses fault, is interpreted to be the Internal Grandes Rousses thrust sited to the W of this buttress. However, some thrust displacements from the Ultra-Dauphinois thrust system to the E (Beach, 1981) may have ramped at the site of this early extensional fault scarp. Continued shortening at basement levels is accommodated by basement thrusts which either branch onto or breach the cover detachment. This basement-cover thrust package on the western margin of the Rochail and Pelvoux massifs is deformed by later out-of-sequence break-back thrusts, indicating that this is not a simple foreland propagating sequence.

2.1.4.4 Structural styles along the Bourg d'Oisans half-graben:

A comparison of the four cross-sections constructed across the Bourg d'Oisans area demonstrates the profound variations in structural style along strike.

1. Oulles - Mizoen cross-section:

The increase in shear strain intensity towards its base, shows that the Mesozoic cover is decoupled from its underlying basement. The shortening associated with this is pinned at its western margin by the Ornon fault scarp buttress. This is characterized by an increase in strain intensity towards the Ornon fault.

Excluding shortening from the intensely strained Oulles section, a shortening value of 26 % was obtained from folding along on the Cote Alamele section. This value is significantly less than those obtained for the same structures along strike to the S where back-thrusting

intensifies.

Although shortening increases approaching the Ornon fault, reasons for the low shortening value obtained from the Cote Alamele section are two-fold:

1. The buttress role of the Ornon fault scarp is partially negated because of the thrust ramp which takes most of the cover shortening out from the restricted confines of the half-graben.
2. The amount of basement shortening in the foot-wall to this cover thrust ramp (i.e. to its W) is negligible compared with the basement shortening to the S.

A simplified sequence of events is as follows:

1. The cover sequence became detached in the foot-wall to a basement thrust to the E, as yet not identified. Minor buckling of the half-graben sediments adjacent to the Ornon fault occurred, and an incipient back-fold-thrust system developed.
2. With continued rearward compression a thrust ramp-shear zone develops some distance back from the fault scarp buttress. A convergence of fold facing directions or thrust triangle zone results.
3. Continued shortening at basement levels caused the development of a minor back-thrust cleavage and folds adjacent to the basement fault scarp became tightened.

2. Grand Renaud cross-section:

The southward continuations of the Cote Alamele folds are tighter. This may reflect an increase in basement shortening in the subsurface, which also accounts for the increasing southward elevation of the cover structures. As yet, no back-thrust system is developed, though the cover folds are upright to eastward-verging.

In the eastern part of the section, the basement-cover succession is repeated forming an anticlinal stack of basement-cover thrust sheets at the western edge of the Rochail. Some of these basement thrusts breach an earlier detachment of the cover, ramp through Lower Liassic limestones and flatten along the Terres Noires easy slip horizon so that the post-rift sequence is detached.

3. La Malsanne - Le Neyrarel cross-section:

The southern equivalent of the Cote Alamele structures is represented by a complex back-fold-thrust system which accommodates an increased shortening of the cover. The increased elevation of the pre-Triassic basement and steep foreland dips suggests the presence of important basement thrusting in the subsurface. A direct correlation can be made between the increased shortening documented in the sedimentary cover and the inferred basement thrusting at deeper levels in the section. The accumulation of displacements from the basement thrusting should balance the increased shortening sedimentary cover.

4. Le Perier - La Paletas cross-section:

The continuation of the Malsanne back-fold-thrust system towards the S, is marked by an intensification of back-thrusting, with several episodes superimposed. From simple line length estimates, the cover shortening exceeds 69 %. The map area occupied by the cover sediments between the Ornon fault and the Rochail basement also becomes constricted. This increased intensity of back-thrusting and shortening can be directly correlated with increased shortening of the basement in the hanging-wall to the Ornon fault.

In summary, it is concluded that the progressive along strike changes in structural style documented from the Bourg d'Oisans area are related to the pre-existing Jurassic fault geometry, which imposes severe constraints on the way in which horizontal displacements entering into

the half-graben can be accommodated.

Because of numerous problems surrounding the use of conventional section balancing techniques in this area, only local percentage shortening estimates were made for comparative purposes. Some of the problems encountered include an incomplete knowledge of a rapidly varying stratigraphic template, the intense and distributed nature of the cover deformation due to buttressing, complex tectonic overprinting relationships and the absence of suitable stratigraphic markers in the syn-rift sequence from which reliable estimates of the shortening can be obtained.

2.1.4.5 Structural evidence concerning the Mesozoic fault dips:

In Section 2.1.4.1, it is argued that the Triassic to Mid-Jurassic rifting is governed by two fault systems of different age; an early closely spaced system of faults rotated to gentle SE dips during extension, and a later widely spaced set which deform the earlier faults and possessed substantially steeper Mesozoic dips.

A number of structural arguments concerning the way in which these pre-existing extensional faults behave during Alpine compression, support this interpretation.

Considering the mechanics of fault reactivation (Chapter 3, Section 3.3), it is the faults that have suffered large angular rotations during extension that will be the easiest to reactivate when the maximum compressive stress is sub-horizontal, as occurs in a compressional regime (Jaeger & Cook, 1979; Sibson, 1985; Gillcrist et al., 1987). Faults with steeper dips and strike orientations perpendicular to the compression direction, will reactivate only should the coefficient of sliding friction be low. Furthermore, during the development of a thrust wedge, as is envisaged at the western margin of the Rochail massif, the maximum compressive stress will be

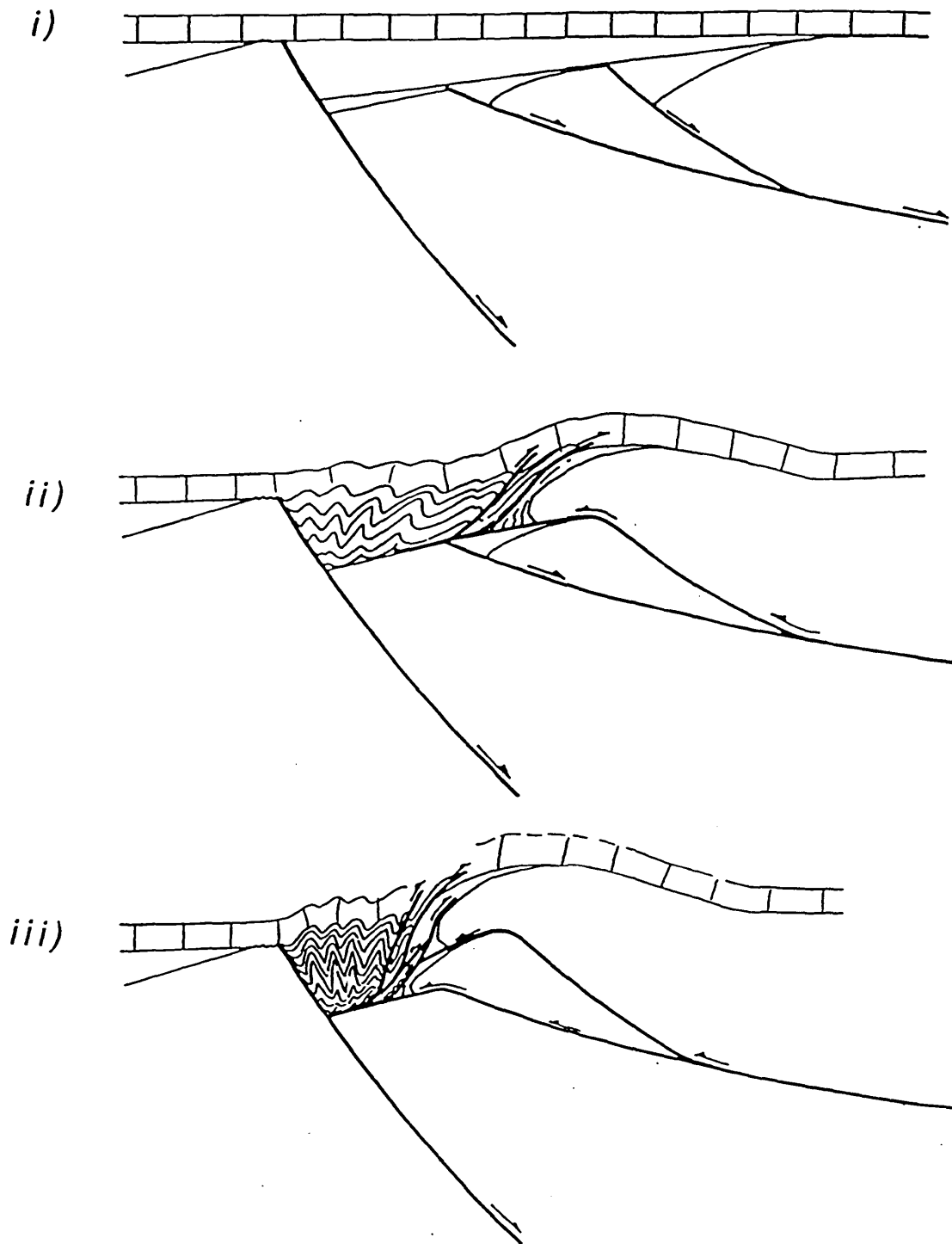


Figure 2.1.66: Schematic compressional evolution of the Bourg d'Oisans half-graben; i) Early Cretaceous basin configuration prior to Alpine compression, ii) thrust stacking of the gently dipping extensional fault blocks, and decoupling of the syn-rift cover sequence, iii) with increased shortening the buttressing effect of the Ornon fault results tight folding and back-thrusting of the syn-rift sequence.

inclined at an angle (α) to the horizontal (Chapple, 1978) rendering the reactivation of the fault almost impossible. Thus faults with low dips to the maximum compressive stress will reactivate, while steeper dipping faults buttress the contractional displacements. From the data presented, it is clear that the Ornon fault served to buttress the Alpine contractional thrust displacements while faults in the hanging-wall to the Ornon fault were reactivated. This supports the interpretation proposed in Section 2.1.4.1.

2.1.4.6 Physical and chemical conditions of cover deformation:

From microprobe analyses, Gratier (1979) noted a concentration of insoluble elements along the S1 cleavage planes. Assuming this increase was due only to the solution of the most mobile elements (calcite and quartz), he calculated the amount of contraction accommodated by pressure solution processes perpendicular to the cleavage plane. Shortening across the cleavage was estimated to be around 17 %, with an elongation perpendicular to this contraction. In any cover shortening estimate this value obtained by Gratier (1979) must be included.

From fluid inclusions, Bernard et al. (1977) estimated the pressure-temperature conditions of cover deformation in the Bourg d'Oisans half-graben, showing:

1. The thermal effect of the basement, with temperatures of 335-370°C recorded for Alpine age veins in the Rochail basement, and 350-250°C for the lower parts of the sedimentary cover with a marked decrease in formation temperature away from the basement.

2. Estimated pressures ranging from 2.0 to 2.5 Kb. Assuming the fluid pressure (FP) equals the lithostatic, then 2.0 Kb is equivalent to 7.5 to 9.5 km of overburden. As the known sedimentary thickness up to the Upper Cretaceous does not exceed 2.5 km, this seems

unlikely. Alternatively, the fluid pressures may be greater than the lithostatic i.e. $\lambda > 1.0$, a condition which would lead to hydraulic fracturing. The pressure recorded by the fluid inclusion would be slightly less than the transitory overpressure but higher than the lithostatic.

These findings have important implications for the gravity sliding hypothesis suggested earlier, increasing the plausibility of Liassic age detachment and sliding down the half-graben slope. The dissipation, or lowering of fluid pressures from this phase may also partly explain the penetrative nature of the cover strains induced by later Alpine age thrust detachments.

2.2 Northern margin of the Pelvoux massif

2.2.1 Introduction

The area covered under this section is indicated on Figure 2.2.1. Understanding the structure of this area is critical as it marks the junction between different basement massifs; the Grandes Rousses massif passes southwards into the Rochail massif, whilst both these massifs are divided from the Pelvoux massif by a thin strip of N-S trending Mesozoic cover sediments.

As in the Bourg d'Oisans area, the structure is dominated by a N-S trending structural depression filled with deep water Mesozoic sediments which separate the uplifted Pelvoux basement from the Grandes Rousses-Rochail basement (Figure 2.2.1). The structures uplifting the Pelvoux massif plunge northwards under the Mesozoic cover of the Plateau d'Emparis (Figure 2.2.1), where the narrow sediment filled depression broadens out into a large expanse of Mesozoic cover sediments, shortened and telescoped by thrusts of the Ultra-Dauphinois thrust system (Beach, 1981a).

The western limit of this sedimentary depression is bounded by steeply E-dipping normal faults, interpreted as preserved Jurassic extensional structures. The interpretation presented here differs from that proposed by Barfety & Gidon (1981), and Rudkiewicz & Vibert (1982), in that two important Jurassic age normal faults, the accident de l'Herpie-Pied Moutet and the Eastern Grandes Rousses fault, are interpreted to have controlled the original geometry of this deep water Jurassic basin.

2.2.2 Evidence for extensional faulting - Mesozoic stratigraphy

Sedimentary facies relationships defined by Barfety & Gidon (1981), are cited as evidence for active Jurassic extensional tectonics, contemporaneous with sedimentation.

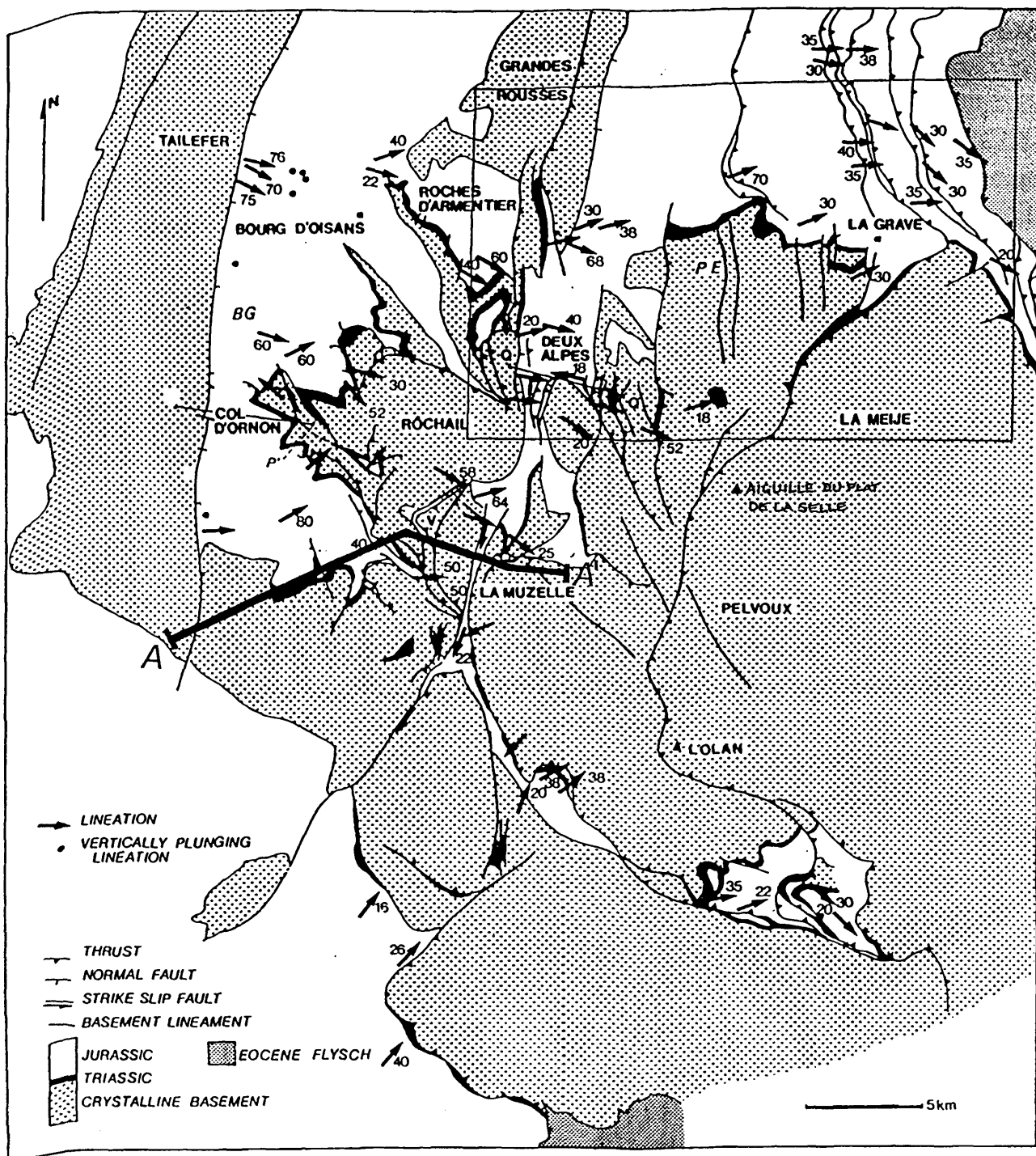


Figure 2.2.1: Structural map of the Pelvoux massif showing the location of the area covered in Section 2.2 (rectangle). The cross-section line AA' for Figure 2.3.7 is shown.

As mentioned above, the structure of the area is dominated by a N-S trending depression of deep water Jurassic sediments. These deep water sediments range in age from Toarcian to Aalenian (Barfety & Gidon, 1981), and possibly up to Bajocian-Bathonian (Dogger) (BRGM, La Grave sheet, 1976). The basement-cover contacts on the borders of this depression are interpreted as steeply dipping Jurassic age normal faults; the Eastern Grandes Rousses fault, and l'Herpie-Pied Moutet fault (see Rudkiewicz & Vibert, 1982; and Section 2.1.3.2). Structural observations support this early normal fault interpretation.

The intensity of the Alpine cover deformation precludes any accurate determination of the original stratigraphic thickness of these deep water formations. Extensional movements on the Eastern Grandes Rousses and Pied Moutet faults provided the necessary subsidence for the deep water sedimentation.

Towards the E, these Toarcian to Aalenian deep water carbonates and shales progressively onlap older extensional faults and shallow water syn-rift sediments exposed on the western margin of the Pelvoux massif. The findings of Barfety & Gidon (1981), from the western border of the Pelvoux, are summarized below. The boundary between Carixian and Aalenian sediments i.e. the Lower-Upper Liassic boundary, is constrained palaeontologically (Barfety & Gidon, 1981).

On the western slopes of Les Perons (G.R.2740,49865) (Figure 2.2.3, and Figure 2.2.2), Upper Liassic sediments of Toarcian age are in direct contact with the basement and Lower Liassic breccias are interstratified with these Toarcian sediments. On the lower slopes of the Grande Cote (G.R.2750,49862), blocks of crystalline basement and Triassic dolomites rest in a Lower Liassic matrix, which to the E passes in to an undisturbed bedded sequence of Lower Lias, the thickness of which does not exceed 150 m (Barfety & Gidon, 1981). In places the Lower Liassic sediments are completely missing and Toarcian shales rest unconformably

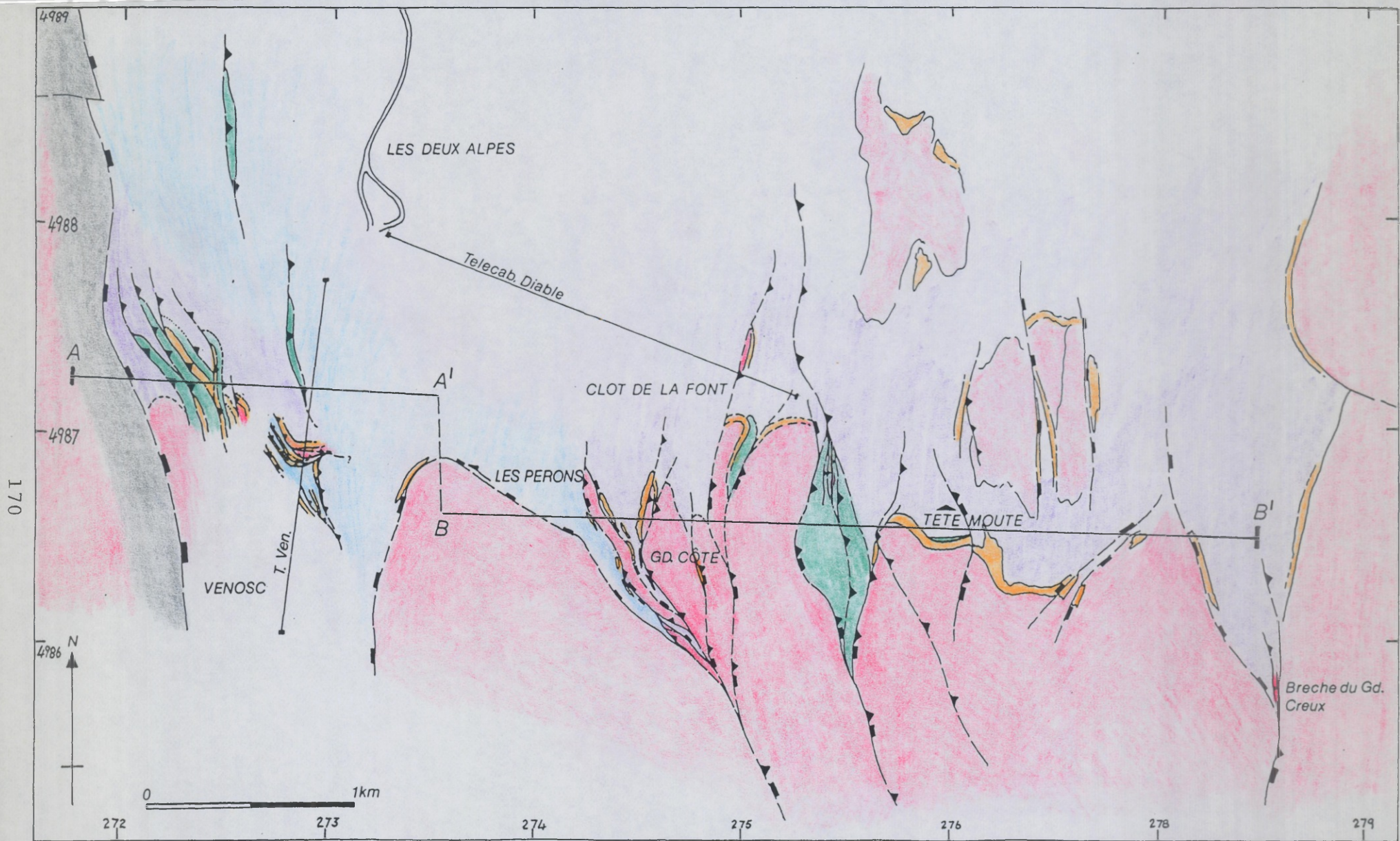


Figure 2.2.2: Structural map of the Deux Alpes area, northern margin of the Pelvoux massif. AA'-Venosc section line (Fig. 2.2.6), AB'-Deux Alpes cross-section (Fig. 2.2.3). For the location and legend refer to Figs. 1.7 & 1.8 (Ch. 1). Based on mapping by the author.

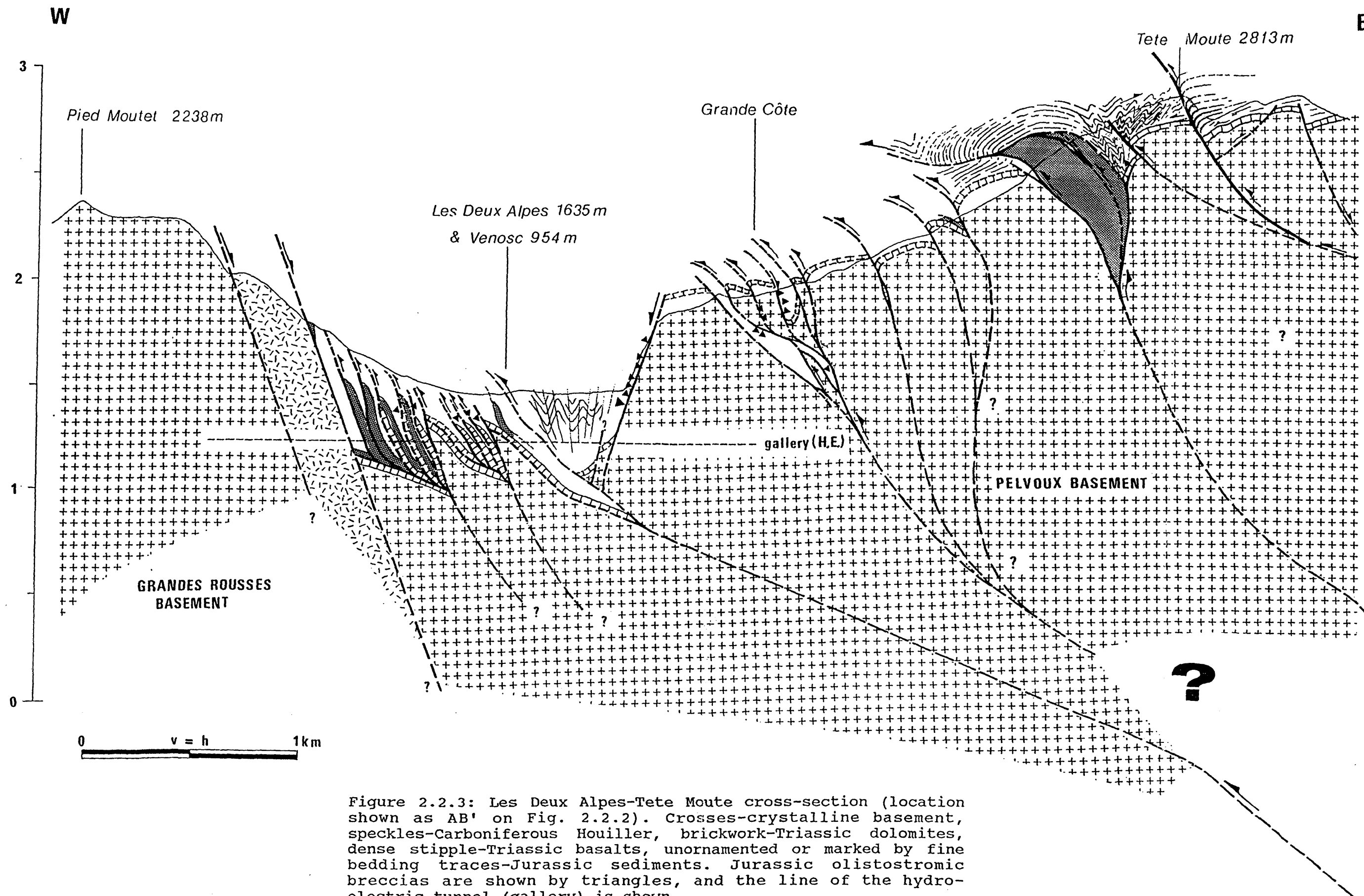


Figure 2.2.3: Les Deux Alpes-Tete Moute cross-section (location shown as AB' on Fig. 2.2.2). Crosses-crystalline basement, speckles-Carboniferous Houiller, brickwork-Triassic dolomites, dense stipple-Triassic basalts, unornamented or marked by fine bedding traces-Jurassic sediments. Jurassic olistostromic breccias are shown by triangles, and the line of the hydro-electric tunnel (gallery) is shown.

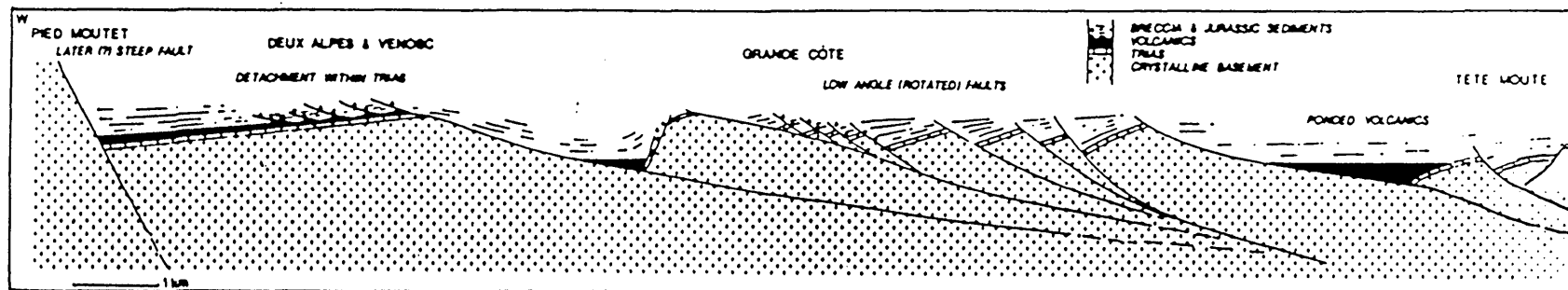


Figure 2.2.4: Suggested cross-section through the Deux Alpes area before the compressional tectonics, showing the low dip of the normal faults. Ornamentation as for Fig. 2.2.3.

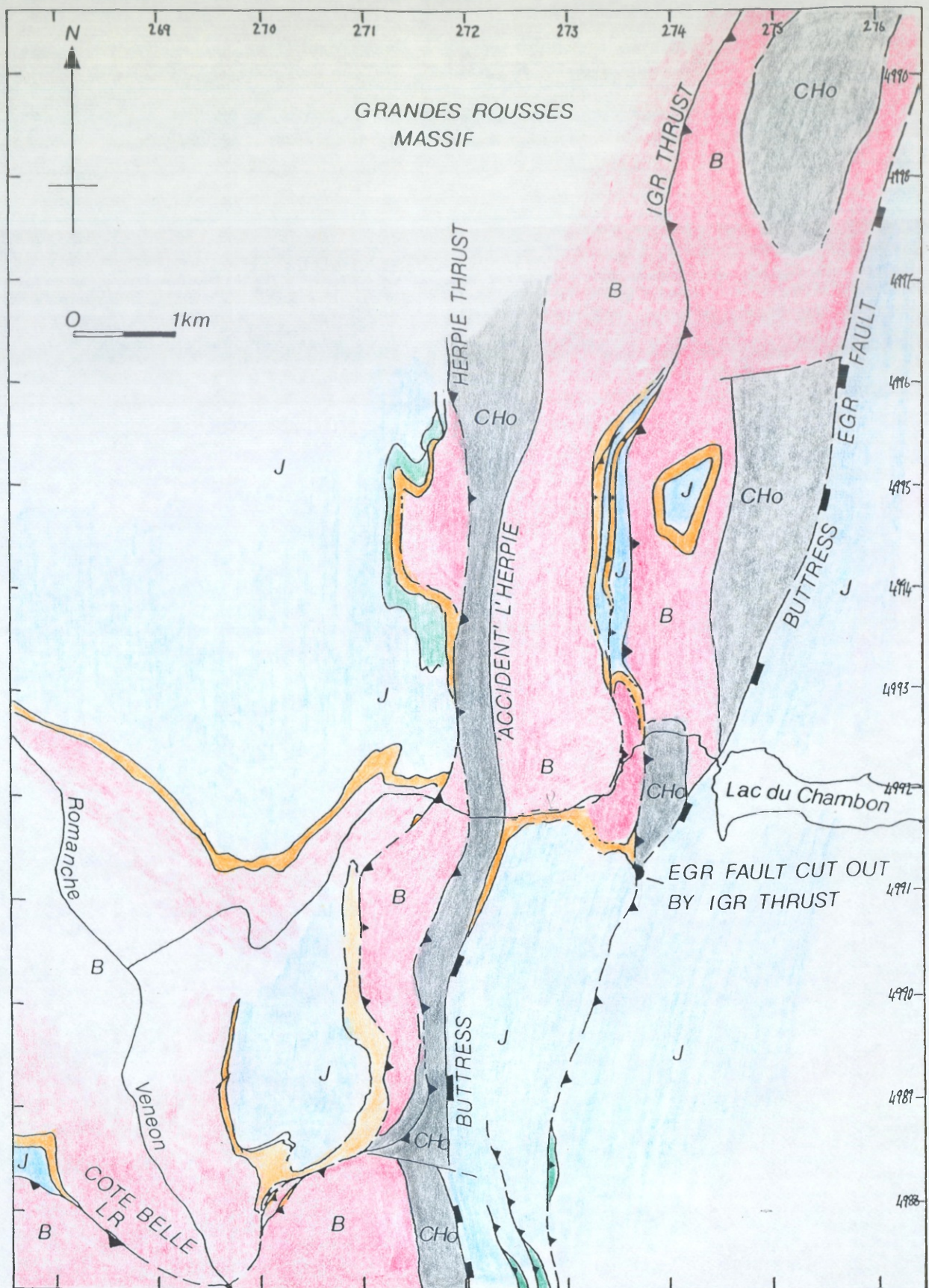


Figure 2.2.5: Simplified structural map of the Grandes Rousses and northern Rochail massifs. Locality and main legend indicated in Figs. 1.7 & 1.8. B-crystalline basement, CHO-Carboniferous Houiller, J-Jurassic, IGR-Internal Grandes Rousses thrust, EGR-Eastern Grandes Rousses fault. Based on mapping by the author and the BRGM 1:50 000 Vizille (1972), and La Grave (1976) maps.

upon crystalline basement. On the upper slopes of the Tete Moute (G.R.2762,49864), alkali basalts reflect further this Lower Liassic extension. To the immediate E of these basalts exists a reduced Lower Liassic sequence.

A hypothetical extensional reconstruction across Les Deux Alpes-Tete Moute depicting the Jurassic facies relationships, is shown in Figure 2.2.4.

2.2.3 Structure

2.2.3.1 Les Deux Alpes-Tete Moute cross-section:

Figure 2.2.3 is a section along a naturally exposed E-W trending cliff along the northern margin of the Pelvoux massif. The location of the section line is shown on Figure 2.2.2.

i) Western slopes of Pied Moutet:

As discussed in Section 2.1, a zone of basement-cover thrust imbrication which branches with the Herpie thrust in the Grandes Rousses massif and the Cote Belle lateral ramp to the S, can explain the elevation of the Pied Moutet basement at Le Sapey (G.R.7425,49890). The importance of the thrust displacements associated with the Cote Belle - Villard-Notre-Dame thrust system, and the need to conserve displacements along strike, has important implications for the local deep structure. Assuming the N-S trending Pied Moutet-La Muzelle fault is of Jurassic age, as displacements associated with the Cote Belle-VND thrust system are not observed to transfer laterally towards the SE i.e. the trace of the Pied Moutet-La Muzelle fault remains continuous and is not offset by the Cote Belle lateral ramp (Figure 2.2.1 & 2.2.5), these thrust displacements must continue to the W of the Jurassic fault. This relationship suggests that at depth, the Pied Moutet fault is sliced through by basement thrusts of the VND-Cote Belle system.

ii) Pied Moutet fault:

The Pied Moutet fault is continuous with the Herpie fault to the N, and the La Muzelle fault to the S (Figure 2.2.5). It apparently formed along and is parallel to, a N-S trending band of Carboniferous Houiller sediments. The exposed fault plane dips 66° E, and juxtaposes basement against mostly Liassic age sediments. An extensional deformation fabric, and small ($< 1\text{m}$) lenses of sheared Triassic spillites and dolomites are observed along the fault plane. It is impossible to determine whether these lenses were introduced tectonically into the fault zone, or are syn-sedimentary and later deformed during Alpine compression. The Liassic shales adjacent to the fault are intensely deformed and are characterized by a penetrative slaty cleavage that obliterates the original bedding trace. A down-dip stretching lineation, manifest by stretched objects, locally records extension values in excess of 500 %, and is observed on the cleavage surface. This fabric is reminiscent of the pure shear fabric recorded adjacent to the Ornon fault.

iii) Venosc imbricates:

To the immediate E of the Pied Moutet fault, the Triassic and Jurassic cover sediments have suffered a significant shortening by thrust imbrication (Figure 2.2.2). A composite E-W section is shown in Figure 2.2.6. The subsurface part of this section was constructed using data from a hydro-electric tunnel (see Letourner, 1943), which confirms that the thrust repetitions observed at the surface can be continued downwards to deeper levels. Basement is also involved in the thrust imbrication. The displacements associated with this thrust imbrication are transferred along a roof thrust in the Upper Liassic and Middle Jurassic shales. Above this roof thrust the Middle Jurassic shales shorten by folding and internal deformation; probably a direct result of buttressing against the Pied Moutet fault. The poor outcrop and the documentation of micro-breccias and other indicators of

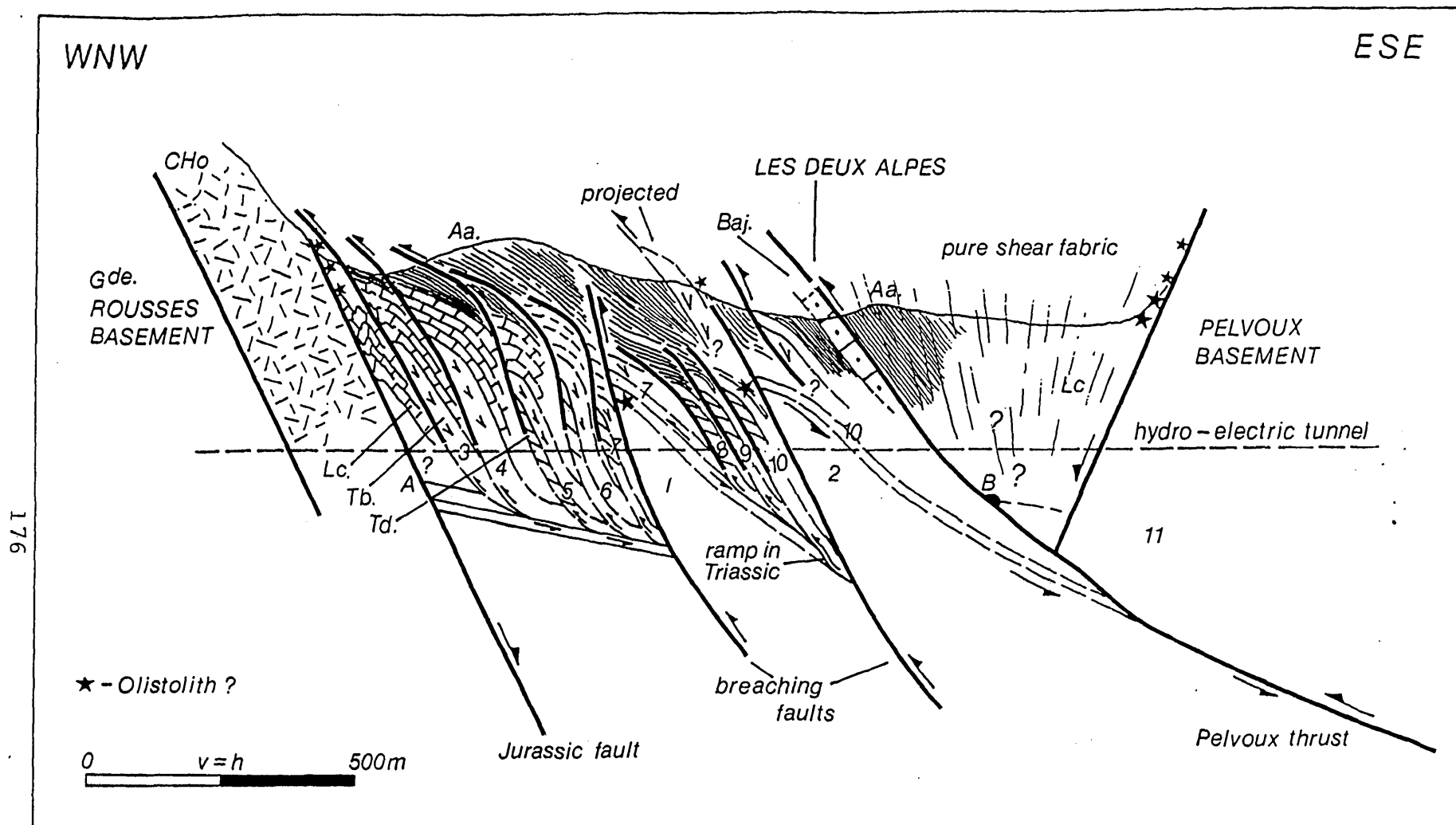
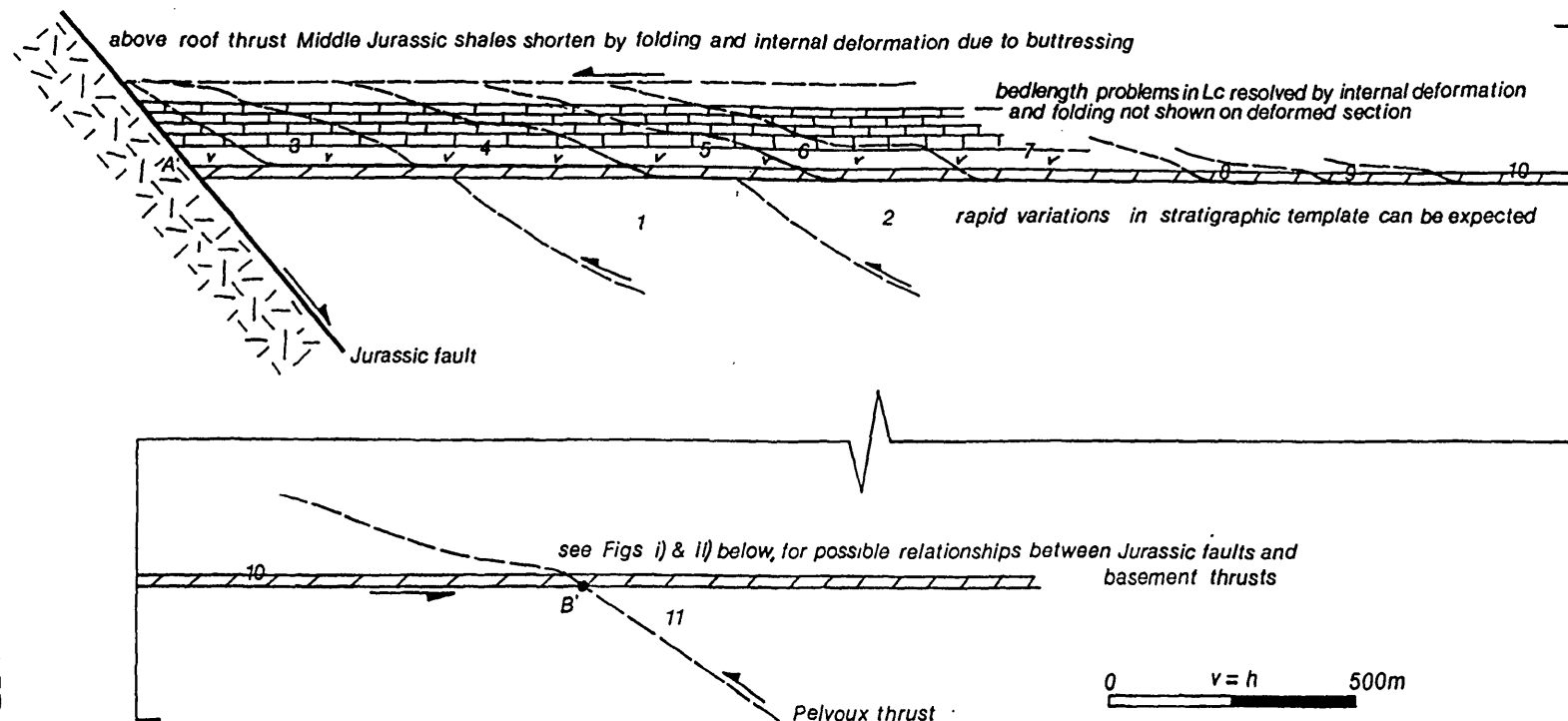


Figure 2.2.6: Structural cross-section across the Venosc imbricates (Location shown as AA' on Fig. 2.2.2), line length balanced for the Triassic horizons. Unornamented-crystalline basement, CHO-Carboniferous Houiller, Td-Triassic dolomites, Tb-Triassic basalts, Lc-Lias calcaire, Aa-Aalenian shales, Baj-Bajocian carbonates.



$$\% \text{ SHORTENING} = L_0 - L_1 / L_0 \times 100 = 3.7 - 1.1 / 3.7 \times 100 = 70.3 \% \text{ min. slip on Pelvoux thrust} = 2.6 \text{ km}$$

Figure 2.2.7: Restored cross-section for the Venosc imbricates (see Fig. 2.2.6). Legend as for 2.2.6.

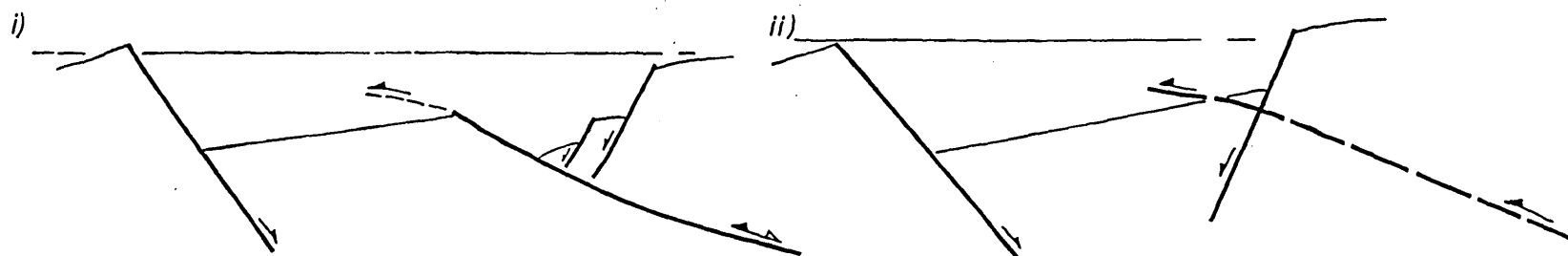


Figure 2.2.8: Possible geometries of the Venosc graben prior to Alpine compression. See text for explanation.

syn-sedimentary faulting (Barfety, 1985), questions the validity of the balanced solution proposed (Figure 2.2.7). In reality, it is difficult to differentiate between large sedimentary olistoliths, of the type described from the Ornon fault, which are passively caught up in later Alpine thrusting and poorly exposed thrust imbricates. It is highly probable that both are present.

The eastern margin of the sedimentary depression is bordered by a steep W-dipping normal fault. Parallel to the fault plane is a large block of Triassic dolomites, interpreted to be a Jurassic age olistolith. Smaller syn-sedimentary breccias are also observed along the fault plane.

The Venosc structure may therefore be considered as representing the remnants of an original Jurassic graben, the sediments of which are protected from any Alpine deformation associated with cover detachments further E. This necessitates a substantial overthrusting of the Pelvoux basement towards the WNW to explain the shortening and imbrication of the Venosc cover sediments. From the balanced restoration proposed in Figure 2.2.7, a minimum slip of 2.6 km is interpreted to have occurred on the Pelvoux thrust. Because of the poor nature of the outcrop and the large amounts of internal strain in the cover sediments, this slip estimate probably constitutes a gross underestimate. The basement thrust which brings up the Pelvoux massif may have originally been a low-angle Jurassic age detachment fault which was reactivated during Alpine compression (Figure 2.2.8.i) (Gillcrist et al, 1987), or the thrust displacements may branch and link with the Internal Grandes Rousses thrust, against which the Eastern Grandes Rousses early normal fault is cut out. The latter implies that the thrust bringing up the Pelvoux basement developed independantly of any pre-existing structure (see Figure 2.2.8.ii). The cover imbrication observed at the Col de Ruinee, and the Venosc imbricates, may be related in that both were generated in the foot-wall to the same basement overthrust. The Pied Moutet fault

served to block this thrust translation of the cover, possibly causing it to imbricate and ensuring a pure shear-upright deformation regime adjacent to the fault scarp.

iv) Grande Cote:

As discussed in Section 2.2.2, Barfety & Gidon (1981) provide evidence for Liassic age extensional tectonics contemporaneous with sedimentation, the early normal fault contacts largely being recognised from syn-sedimentary relationships, such as onlaps, olistoliths and fault scarp breccias.

On the southern slopes of Les Perons, Toarcian age sediments rest directly against a steeply dipping basement fault scarp, itself lined with micro-breccias of syn-sedimentary origin. The Alpine cleavage in the cover sediments dips steeply towards the NE (68° E/160) and is aligned parallel to the basement fault plane. To the E, towards the Grande Cote, the cleavage is partially wrapped around an indenting basement block which is thrust westwards over a foot-wall of cover sediments.

On the upper slopes of the Grande Cote are a number of steeply dipping reverse basement faults. Some of these can be traced northwards to the Clot de la Font where they flatten out as thrusts in the overlying sedimentary cover. In the foot-walls of these basement faults are Liassic age syn-sedimentary fault breccias and deca-metric blocks of crystalline basement and Triassic dolomites, interpreted to be Liassic age olistoliths (Plate 2.2.1). Some Liassic age faults still preserve a net extension, with fault scarp breccias on their hanging-wall. Such a structure occurs on the Clot de la Font and is now a syncline tightened in the foot-wall to a basement thrust (Figure 2.2.3). Others have suffered a complete reactivation and inversion.

To the S on the lower slopes of the Grande Cote, all of these Liassic normal faults and Alpine basement thrusts branch onto a single detachment surface in the basement

(Figures 2.2.2 & 2.2.3). This fault broadens out into a broad shear zone, the kinematic indicators from which are all consistent with a reverse thrusting movement. Where these thrust displacements climb into the sedimentary cover, the Le Perons normal fault profile is imbricated (Figure 2.2.9). This geometry suggests that at depth the original fault may be re-used, while at higher structural levels, new failure planes develop and the original fault profile is imbricated.

Overall, the evidence presented suggests that a number of extensional imbricates are restacked during compression and then transported over a foot-wall succession of cover sediments on a frontal basement thrust which accommodates most of the compressional displacements.

v) Tete Moute:

On the lower slopes of the Tete Moute, what is demonstrably a thrust, with a sense of displacement of hanging-wall towards the W, emplaces younger Triassic alkali basalts over older crystalline basement. The lower parts of the basalts are intensely sheared and are characterized by a steeply E-dipping, penetrative Alpine cleavage. At higher levels the basalts are repeated and thickened by thrusting, whilst the uppermost basalts are unstrained and preserve thermal contraction joints. In the overlying sedimentary cover, what are interpreted as the syn-rift sediments of a Liassic age half graben, are folded into a tight W-verging anticline with an overturned lower limb which is cut out against an E-dipping thrust fault (Plate 2.2.2). This thrust is continuous with the thrust fault at the base of the basalts. The remainder of the half-graben sediments to the E are deformed by tight upright folds with a sub-vertical cleavage. These folds progressively assume an eastward-vergence developing into a back-fold system which can be traced to the N. It is surmised that the basalts accumulated in the lower part of a half-graben (Figure 2.2.4) and with subsequent compression, were shortened in the foot-wall to a basement

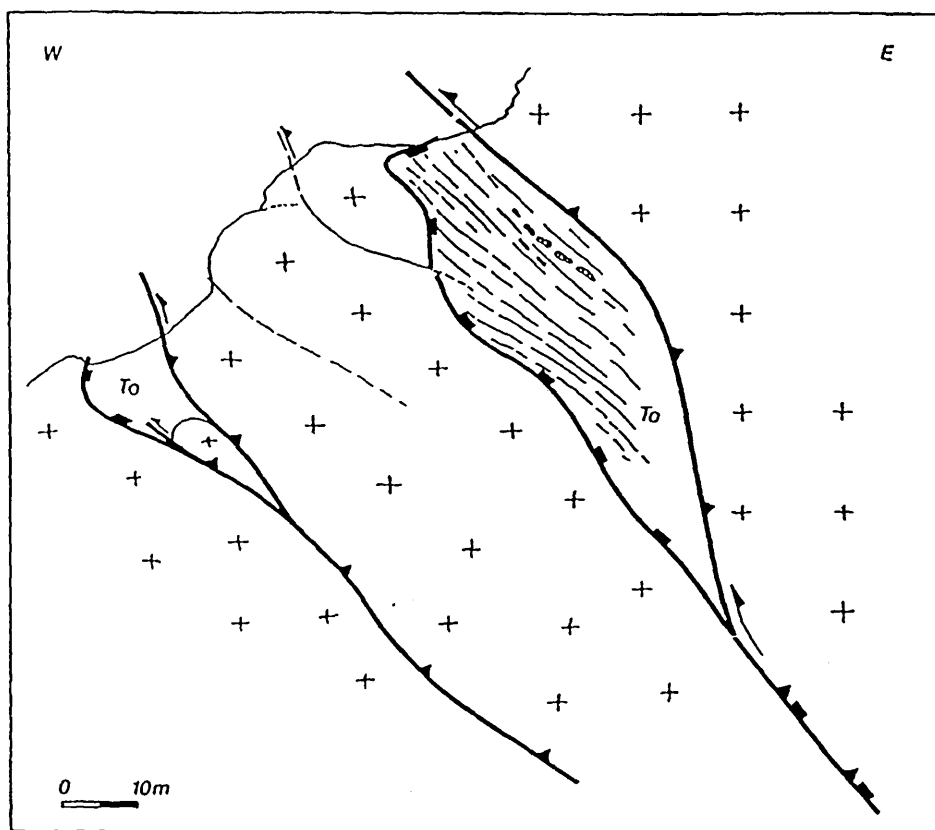


Figure 2.2.9: Field sketch of the imbricated Les Perons early normal fault profile. Crosses-basement, To-Toarcian, bedding trace and syn-sedimentary breccias also indicated. See text for details.

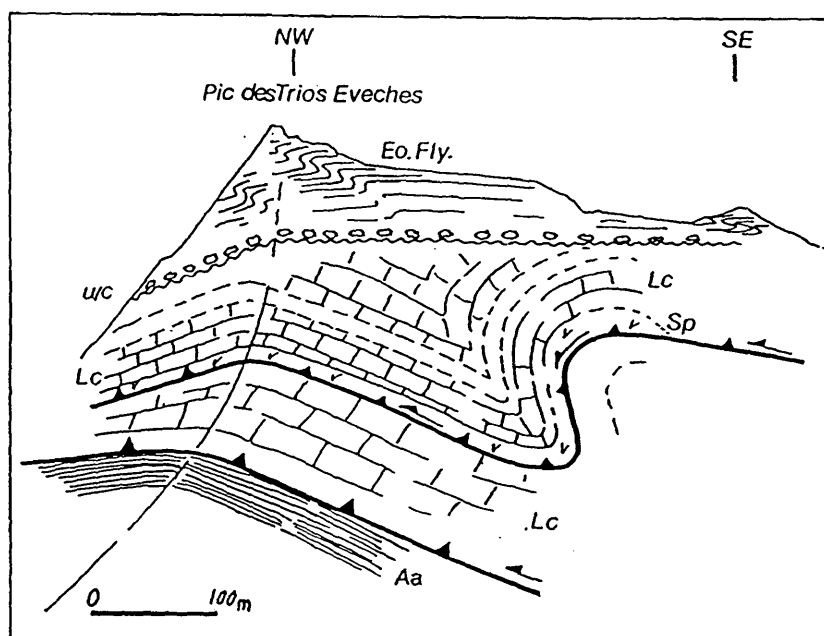


Figure 2.2.12: Field sketch of the Pic des Trios Eveches (see Fig. 2.2.11 for location) showing thrust structures which apparently pre-date the overlying Eocene flysch sequence. Eo.Fly-Eocene flysch, Aa-Aalenian shales, Lc-Lias calcaire, Sp-Triassic spillites.

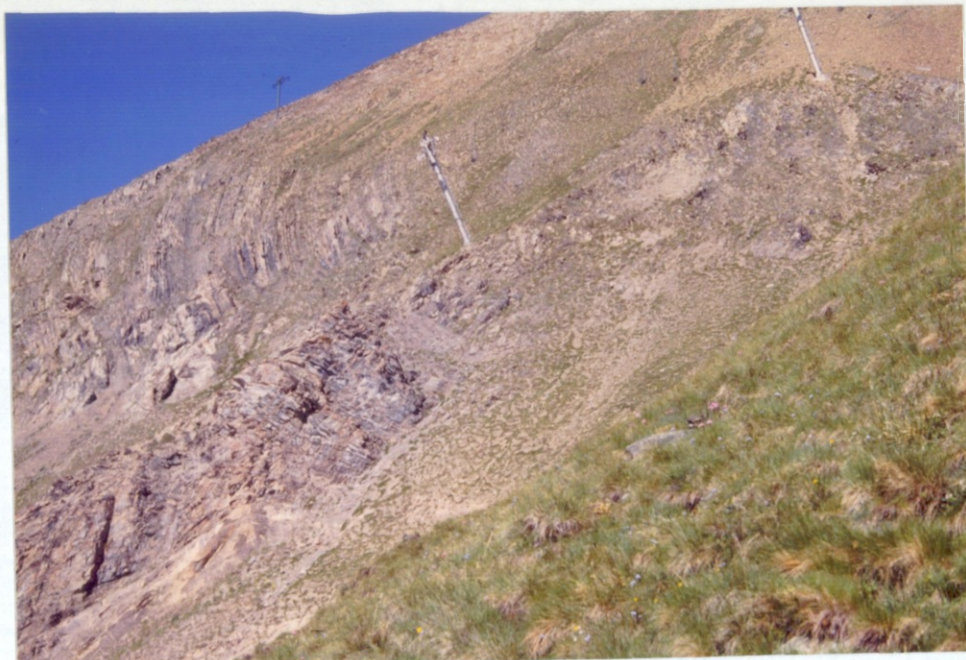


Plate 2.2.2: Hanging-wall fold and thrust fault in Liassic sediments exposed on the Tete Moute. At a lower structural level the same fault is in net extension, emplacing a hanging-wall of Triassic basalts over a foot-wall of basement. See text for details. Looking NE.

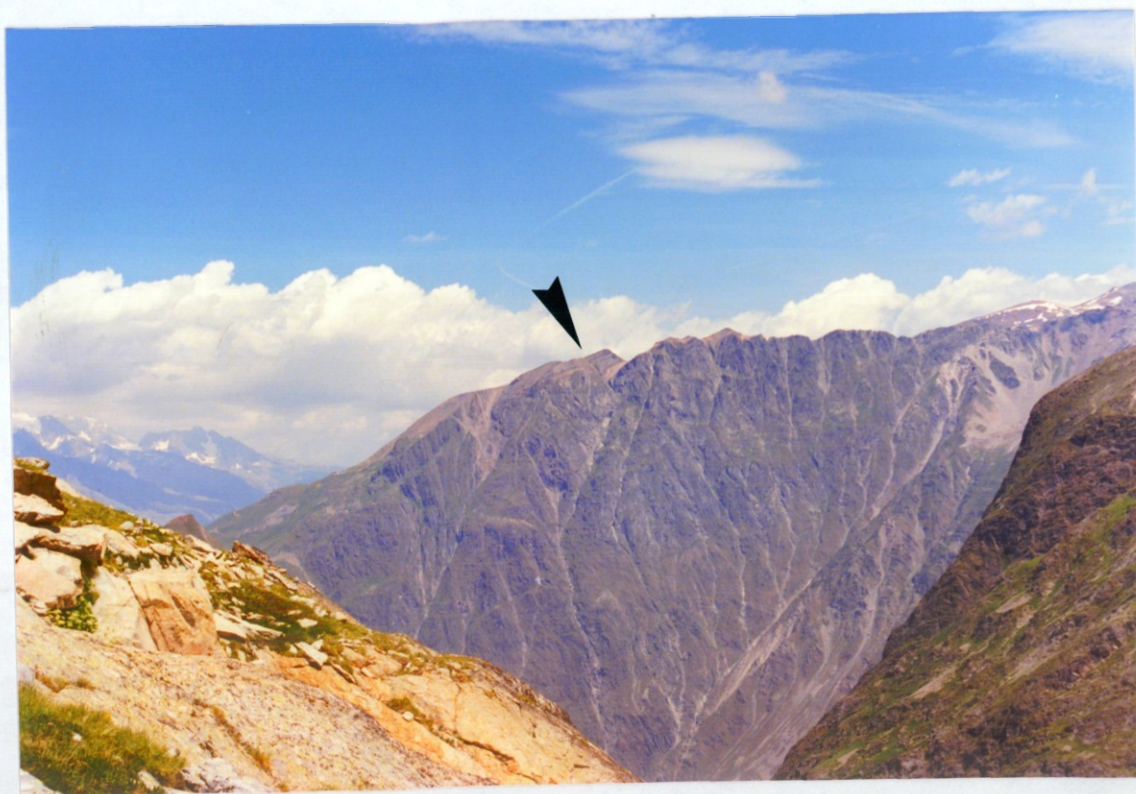


Plate 2.2.3: Panoramic of the Deux-Alpes - Tete Moute cliff section showing the inverted half-graben of the Tete Moute (arrow). Looking N, see text for details.

thrust sheet which reactivated the old extensional fault plane and carried the Liassic half-graben in its hanging-wall. The hanging-wall was then cut by a later break-back thrust.

The Tete Moute summit provides a classic example of an inverted Liassic half graben (Plate 2.2.3). The fault is still in net extension, preserving minor Liassic growth in its hanging-wall. The comparative foot-wall sequence of Triassic and Liassic age is dramatically reduced. In response to the Alpine compression a sub-vertical fault developed, along which minor reverse movement occurred. The hangingwall sediments adjacent to this fault are folded into a gentle monocline.

This cross-section is representative of the complex array of structures mapped along the northern margin of the Pelvoux massif, from Les Deux Alpes to the glacier Mont-de-Lans and the La Meije thrust.

2.2.3.2 Hypothetical extensional reconstruction:

i) Structural evidence for the early extensional fault geometry:

The deformed and restored sections presented in Figures 2.2.3 & 2.2.4, differ from those given by Barfety & Gidon (1981) in that the structural inversion is considered to be more complex with several levels of detachment accommodating significant amounts of thrust shortening; they consider that the present day structural configuration results from a minor Alpine tightening of the pre-existing extensional structures.

In the eastern part of the section it is suggested that most of the pre-existing normal faults dipped toward the E at low to moderate angles, presumably a result of rotation during extension; a presupposition supported by the apparent relative ease with which these faults reactivated as thrusts during Alpine compression. The

initial amount of extension is difficult to judge as Alpine thrusting has modified the original cut-off angles and steepened the dip of the normal faults.

More specifically, the basement thrusts of the Grande Cote are interpreted to have originally been part of a normal fan system (Gibbs, 1984), detaching on to a low-angle fault in the basement which was subsequently reactivated during Alpine compression. The contractional displacements derived from this thrust reactivation presumably transfer along bedding-parallel surfaces unaffected by earlier extensional faulting, in Middle Jurassic shales or post-rift sediments now removed by erosion. The final contractional geometry resembles that of a compressional duplex with its sole fault lying in basement and its roof thrust in the undisturbed post-rift cover sequence. In the absence of detailed stratigraphic observations, such a geometry could be easily misinterpreted within a thin-skinned thrust context.

The structural style prevailing in the Venosc graben (with 70 % shortening), suggests that the Pied Moutet and Eastern Grandes Rousses faults possessed significantly steeper Mesozoic dips ($50-60^{\circ}$?), acting as buttresses during Alpine shortening and are possibly part of a slightly later breaching system to the earlier gently dipping normal faults. Much of the shortening in the Venosc graben probably reflects the resistance to continued forward thrusting exerted by the steeply dipping Pied Moutet fault.

In summary, a large Alpine shortening was accommodated by the thrust reactivation of the more gently dipping normal faults and the formation of new basement thrusts (the Pelvoux thrust), while the steeply dipping Jurassic normal faults served to buttress the contractional displacements climbing out from the basement.

ii) Stratigraphic evidence for the extensional fault geometry:

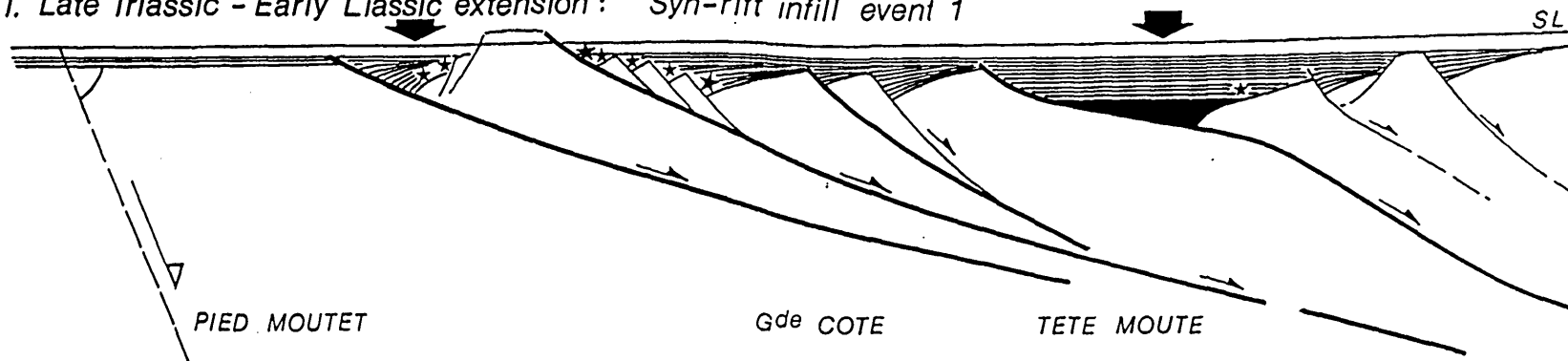
The low-angle, gently dipping faults, later reactivated as Alpine thrusts, are dominantly of Lower Liassic age with a late Triassic-early Liassic initiation age. The half-graben sediments are characterized by Lower Liassic syn-sedimentary fault breccias which are dated as younger than Carixian (Barfety & Gidon, 1981). The magnitude of subsidence associated with this extension is interpreted to be in the order of 200-300 m, deduced from sedimentary thicknesses, and the shallow water character of the half graben sediments.

Deeper water sedimentation, and hence the period of maximum subsidence, commenced in the Toarcian and apparently continued throughout the greater part of the Middle Jurassic. No movements of this age are documented for the earlier gently dipping faults, which are progressively sealed by the eastward-directed onlap of deeper water sediments. This relative deepening apparently increased towards the Pied Moutet fault. To the E Toarcian and Aalenian sediments rest directly on crystalline basement. This may reflect localised foot-wall uplift during the early extensional phase, or be related to the uplift and erosion of the early fault system caused by the broader isostatic effects associated with later extension.

The steeper faults, Pied Moutet and Eastern Grandes Rousses, are interpreted to belong to the same, slightly later breaching family to which the Ornon and La Mure faults belong (Section 2.1). The evolution of the fault systems related to these two different pulses of syn-rift extension, are depicted schematically in Figure 2.2.10.

No direct evidence of the Mesozoic extension direction (i.e. the documentation of possible transfer faults) was observed. However, N-S facies trends, and the presupposition that the later Alpine shortening was coaxial with respect to the Mesozoic extension (Chapter 3),

1. Late Triassic - Early Liassic extension : Syn-rift infill event 1



2. Late Liassic - Mid. Jurassic extension : Syn-rift infill event 2

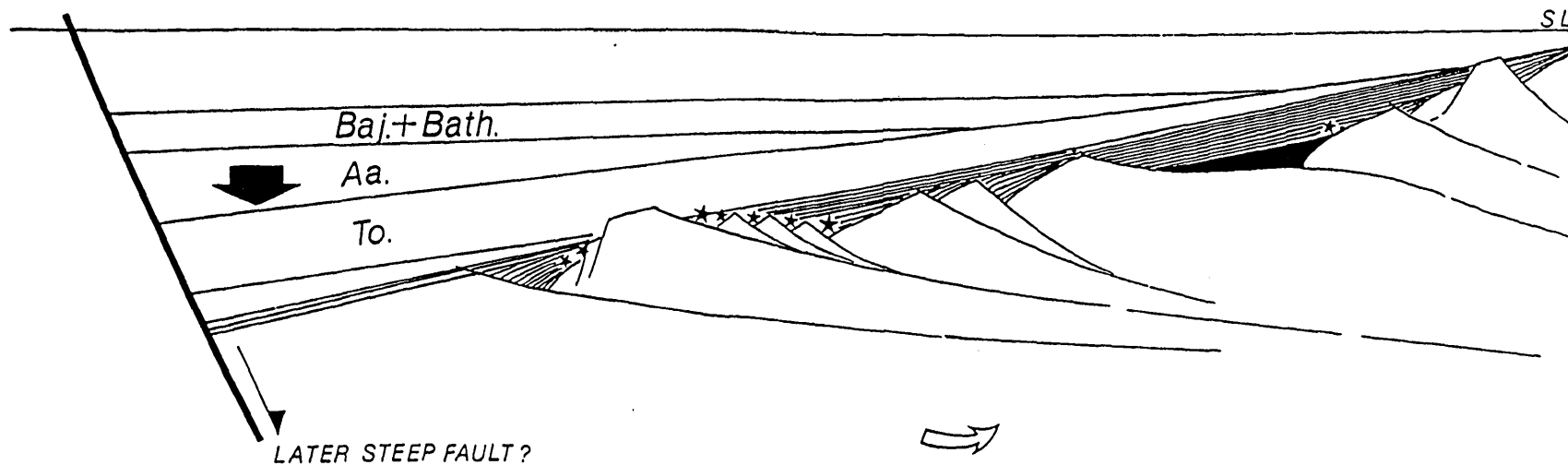


Figure 2.2.10: Suggested evolution of the Jurassic normal fault system across the Deux Alpes section line. To-Toarcian, Aa-Aalenian, Baj & Bath-Bajocian & Bathonian. See text for details.

suggests that the extension direction is the same as proposed for the Bourg d'Oisans area (Section 2.1.4.1).

2.2.3.3 Continuation of the Deux Alpes section:

Continuing the Deux Alpes section eastwards, the basement-Triassic unconformity is offset by a number of minor normal faults of indeterminate age. They may be of similar age to structures observed on the Plateau d'Emparis to the N, some of which are clearly Liassic in age, whilst others post-date a penetrative, flat lying Alpine deformation fabric in the sedimentary cover. The next important basement thrust is observed at the Breche du Creux (Figure 2.2.2), where from map relationships this thrust apparently branches with a Liassic age extensional fault to the S, suggesting thrust reactivation of the Liassic fault occurred during the Alpine compression. In the foot-wall to the this basement thrust, is a small (< 10 m) fault bound slice of basement, completely enveloped in the surrounding cover sediments and which may have been plucked from the foot-wall of the extensional fault scarp during reactivation. This thrust structure continues towards the Plateau d'Emparis where it becomes difficult to trace in the overlying Mesozoic cover sediments.

In the hanging-wall of this thrust fault, at Le Jandri (G.R.2800,49870), horizontally bedded Triassic and Liassic sediments are exposed. The continuation of this section line further to the E is restricted by the glacier Mont-de-Lans which extends right up to the La Meije thrust.

2.2.3.4 La Meije thrust:

The La Meije thrust has long been identified as a thrust fault (Vernet, 1961). Controversial ages and different directions of movement have been proposed by several authors. These include an "early Alpine" phase, of late Cretaceous or Palaeocene age (Barbier, 1963; Gidon, 1979; Bravard & Gidon, 1979), with a NNW direction of overthrusting (Bartoli et al, 1974), and SW directed post-

Eocene thrusting (Beach, 1981a; 1981b). A similar controversy surrounds the age and the direction of thrusting of the Combeynot massif, to the E of the La Meije thrust (e.g. Bravard & Gidon, 1979; Mathews, 1984).

The La Meije thrust is exposed to the S of the village of La Grave (Figure 2.2.11), directly beneath La Meije and Le Rateau glaciers (Plate 2.2.4). The Upper Liassic and Middle Jurassic shales in its foot-wall are penetratively deformed by tight asymmetric folds which verge towards the W. A slaty cleavage, axial planar to these folds dips 40-45° to the NE, striking 130, and contains a mineral stretching lineation indicative of a thrust transport direction to towards the W to SW. Fibres from pressures shadows indicate a strong anti-clockwise rotational component to the cover deformation.

The La Grave BRGM 1:50 000 geological map (1976) shows the La Meije thrust as a normal stratigraphic contact, with a reversed (overturned) Triassic and Jurassic sequence. However, based on our own observations (A. Pecher and the author), this interpretation is considered incorrect. The reasons for this are two fold:

1. The La Meije thrust plane emplacing basement over Upper Liassic sediments is clearly exposed. The fault is marked by a strongly developed mylonite, and several hanging-wall thrust splays deform the La Meije granite. The La Meije granite is also deformed by Alpine age shear zones, characterized by a ductile foliation/cleavage which becomes sheared parallel to localized zones of high strain.

2. Where exposed, the Triassic dolomites along the base of this thrust contact are strongly sheared and in places mylonitised.

These relationships suggest that either the reversed stratigraphic sequence is a foot-wall syncline to the La Meije basement overthrust, in which case the displacements

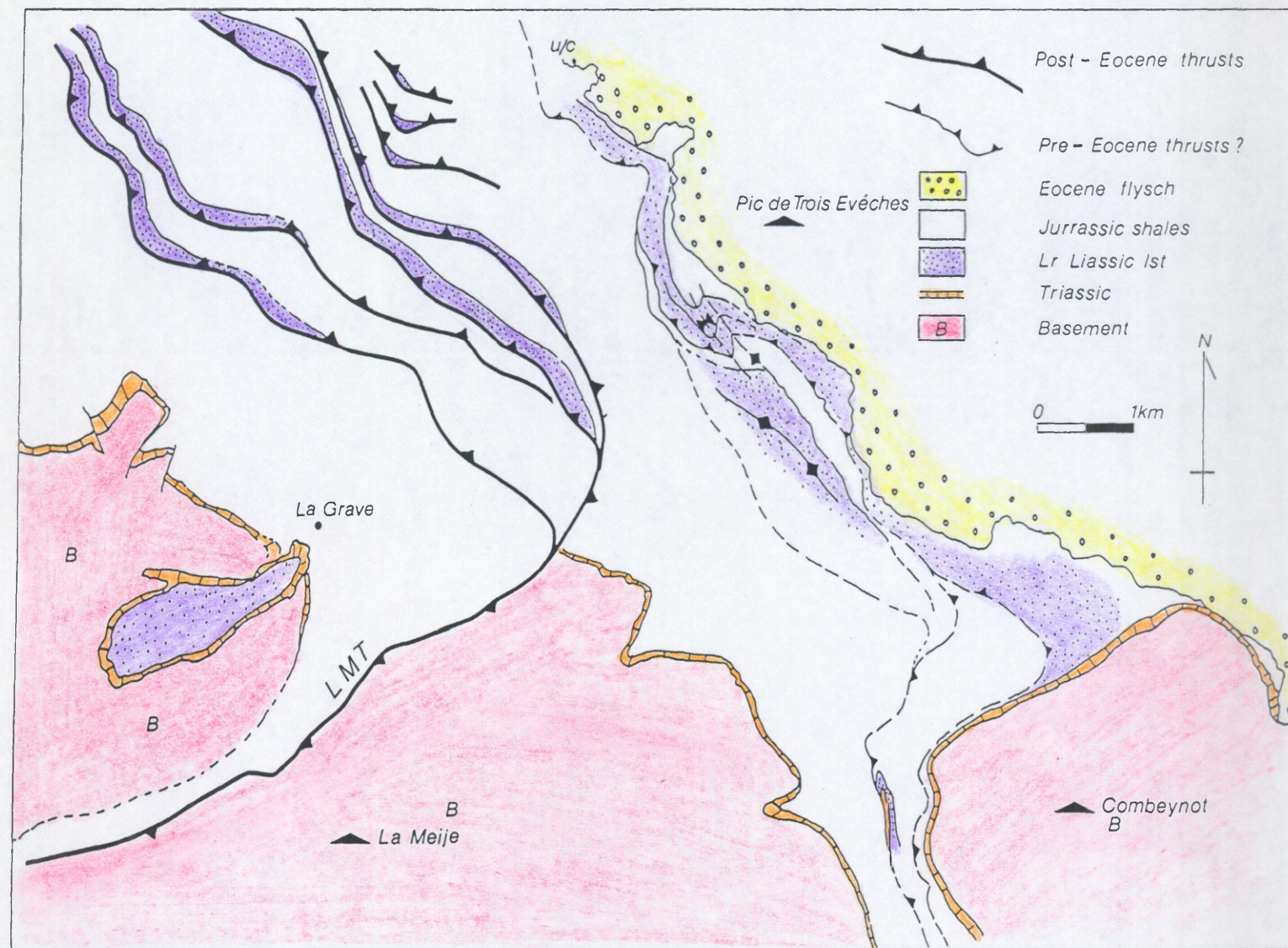


Figure 2.2.11: Simplified structural map of the La Grave area, based on the BRGM 1:50 000 La Grave sheet (1976). Location is indicated on Fig. 1.7.

associated with this fault are not as large as envisaged by some authors (Beach, 1981a & b; Butler et al., 1986), or the Triassic dolomites correspond to a "far travelled horse" (Hossack, 1978) plucked from the foot-wall during thrusting. In either case, the stratigraphic relations in the footwall to the La Meije thrust need redefining. Indeed the lithological appearance of what is designated as Lower Liassic sediments (BRGM La Grave sheet, 1976) in the immediate foot-wall to the La Meije thrust, bears a closer resemblance to the Middle Jurassic black shales.

The thrust plane itself dips between 34-40° to the SE, and the orientation of linear fabrics from the fault zone suggest that the thrust movement direction was towards the W.

In the immediate foot-wall sediments of the La Meije thrust, irregular shaped blocks of crystalline basement are completely enveloped in the surrounding cover sediments (Plate 2.2.5). These can best be interpreted as sedimentary olistoliths. However considering the large thrust displacements associated with the La Meije thrust, the original position of the fault scarp from which they were derived is now obscured.

2.2.3.5 Timing of the thrusting:

Upon reaching the overlying cover sequence, the La Meije thrust apparently flattens, and splays into a number of cover thrusts which shorten and telescope the Ultra-Dauphinois sedimentary cover (Figure 2.2.11). The precise age of these cover thrusts and the La Meije basement thrust is not clear, however, they may post-date the Eocene flysch deposition and explain the eastward tilting of this Eocene flysch sequence to the E (Figure 2.2.11). No pre-Eocene thrust movements (Bartoli et al., 1974) were identified.

The age of the Combeynot thrust to the E, may pre-date the deposition of the Eocene flysch (Bravard & Gidon, 1979). The basal Eocene unconformity rests directly on



Plate 2.2.4: Shows the La Meije thrust (arrow) emplacing a hanging-wall of basement over a foot-wall of Lower and Mid-Jurassic cover sediments. Looking SE.



Plate 2.2.5: Basement olistolith enveloped in Upper Liassic sediments, lying in the immediate foot-wall to the La Meije thrust. Geologist (A. Pecher) for scale.

Jurassic age sediments and even erodes down into the Combeynot basement massif (Figure 2.2.11). This suggests that substantial uplift, possibly related to thrust movements on the Combeynot thrust, pre-dates the deposition of the Eocene flysch sequence. Indeed cover thrusts which emerge from the Combeynot basement thrust can be traced to the NW and dated by the Eocene flysch. Reconnaissance work by the author reveals that on the SW facing slopes of the Pic de Trois Eveches, a thrust flat which emplaces Triassic spillites and Lower Liassic sediments over a Lower Liassic foot-wall sequence, is folded by movements on deeper thrusts, and therefore apparently pre-dates the deposition of the Eocene flysch which rests unconformably across this structure (Figures 2.2.11 & 2.2.12). Later thrust related slip may have exploited the base of the flysch (Beach, 1981; Mathews, 1984), but the lower level thrusts affecting the Jurassic sediments seem to be cross-cut by the unconformity. Further work is required to verify this hypothesis.

2.2.3.6 Thrust transport directions:

In the absence of identifiable lateral ramp structures, the thrust movement direction for the structures along the northern margin of the Pelvoux massif was determined using cut-off trends measured in the field, combined with a study of the kinematic indicators. The slip-direction indicators are presented in map form (Figure 2.2.13), stereographically, and in rose diagrams to illustrate the regional transport trend.

1. Le Sapey imbricates: These correspond to the basement-cover imbricates on the western slopes of the Pied Moutet. The slip indicators recorded from these faults indicate a W-direction of thrust transport (Figure 2.1.14 a & b). This direction differs from that obtained for the VND-Cote Belle thrusts with which they are interpreted to branch (Section 2.1).

2. Venosc imbricates: From the zone of cover thrust

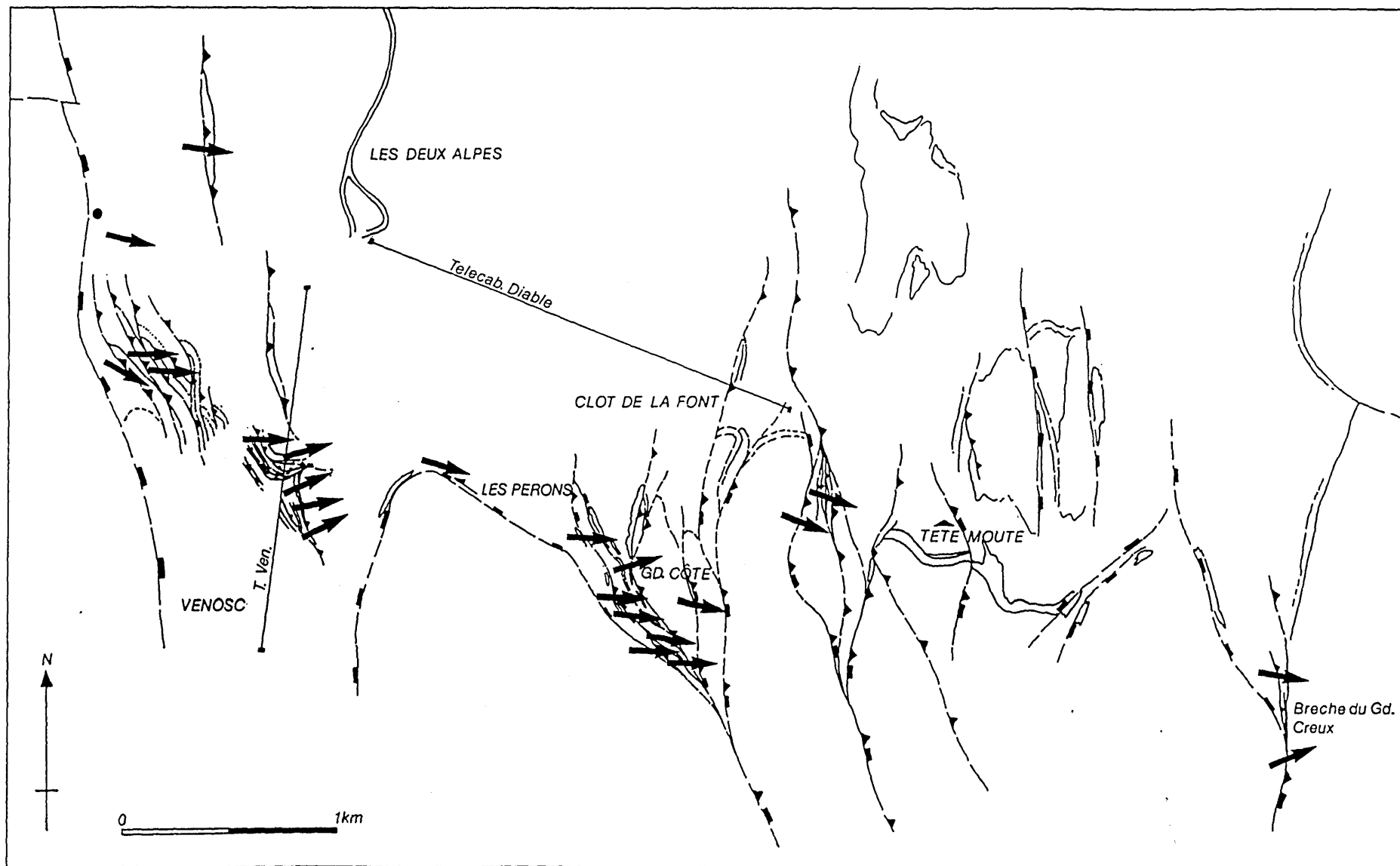


Figure 2.2.13: Map showing the slip-lineation trends from thrust faults, and mineral stretching directions in the cover, for the Deux Alpes area. The map area is the same as that covered by Fig. 2.2.2.

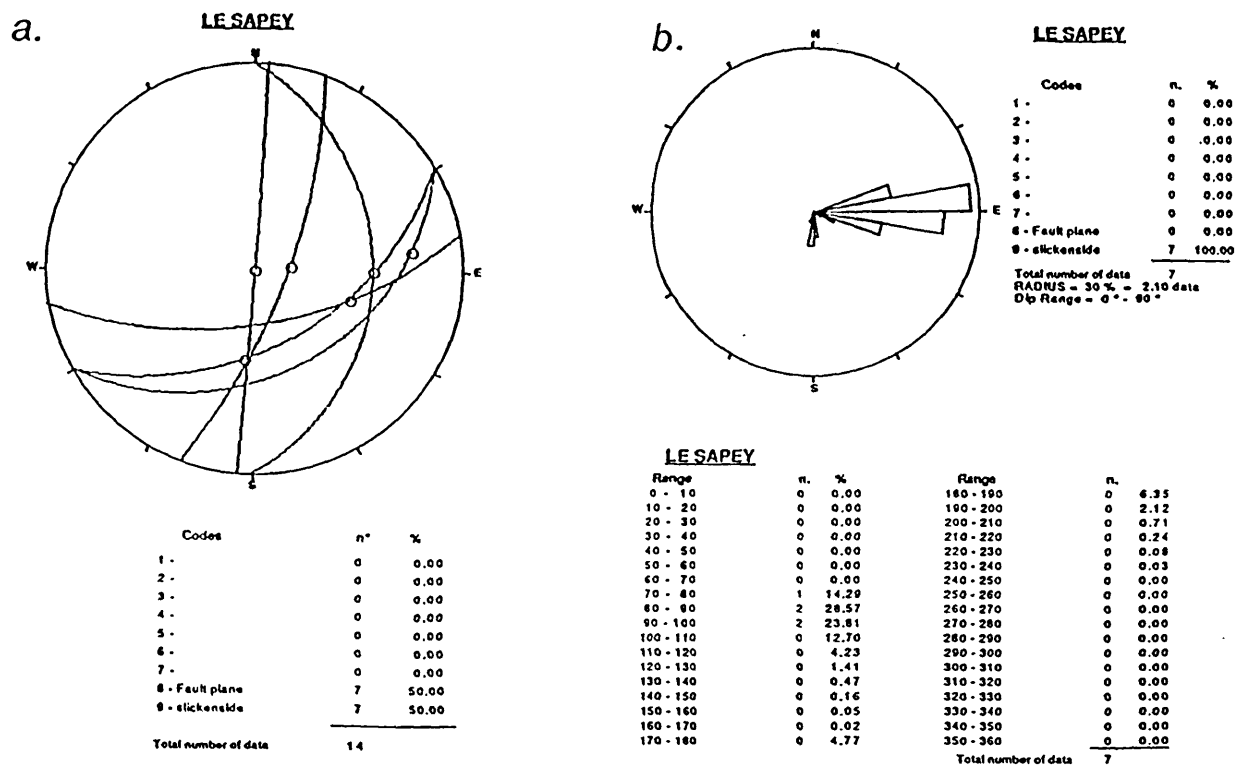


Figure 2.2.14: Equal angle (Wulf net) plot of slip-lination and fault plane data (a), and rose diagram (b) of the slip-lination trends for Le Sahey (to the W of Pied Moutet).

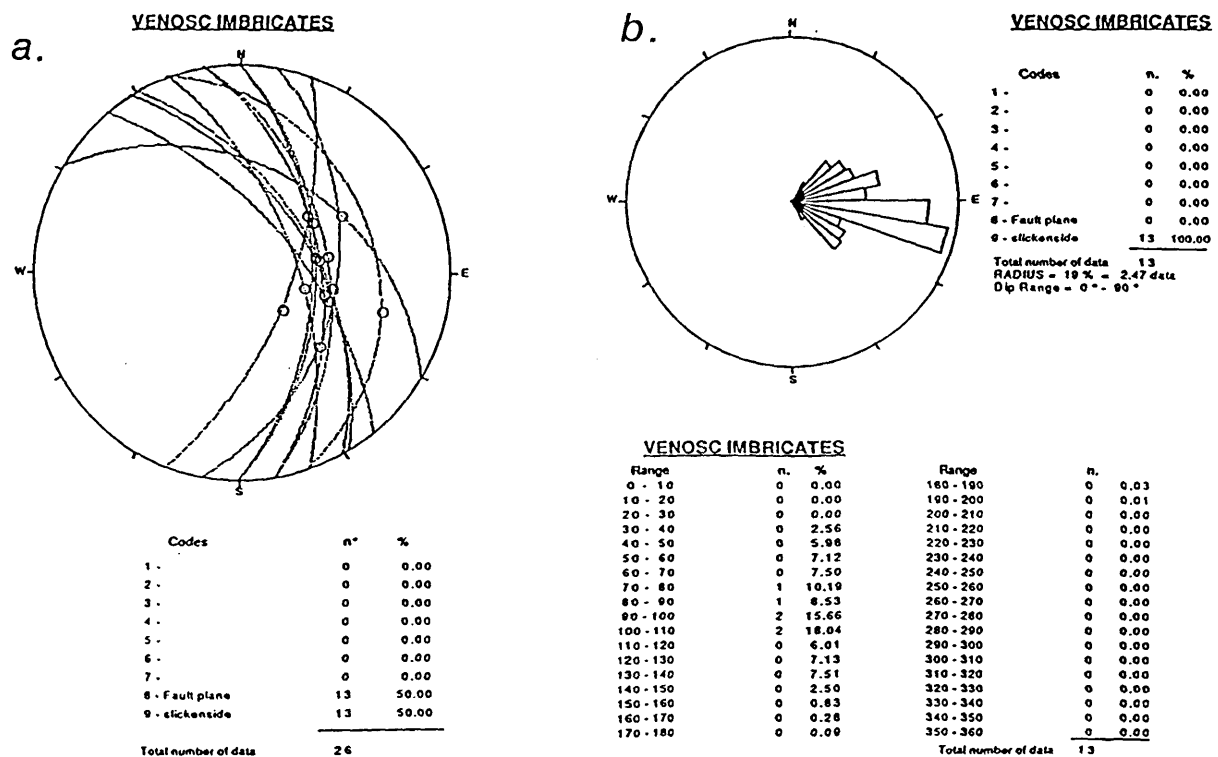


Figure 2.2.15: Equal angle (Wulf net) plot of slip-lination and fault plane data (a), and rose diagram (b) of the slip-lination trends for the Venosc area.

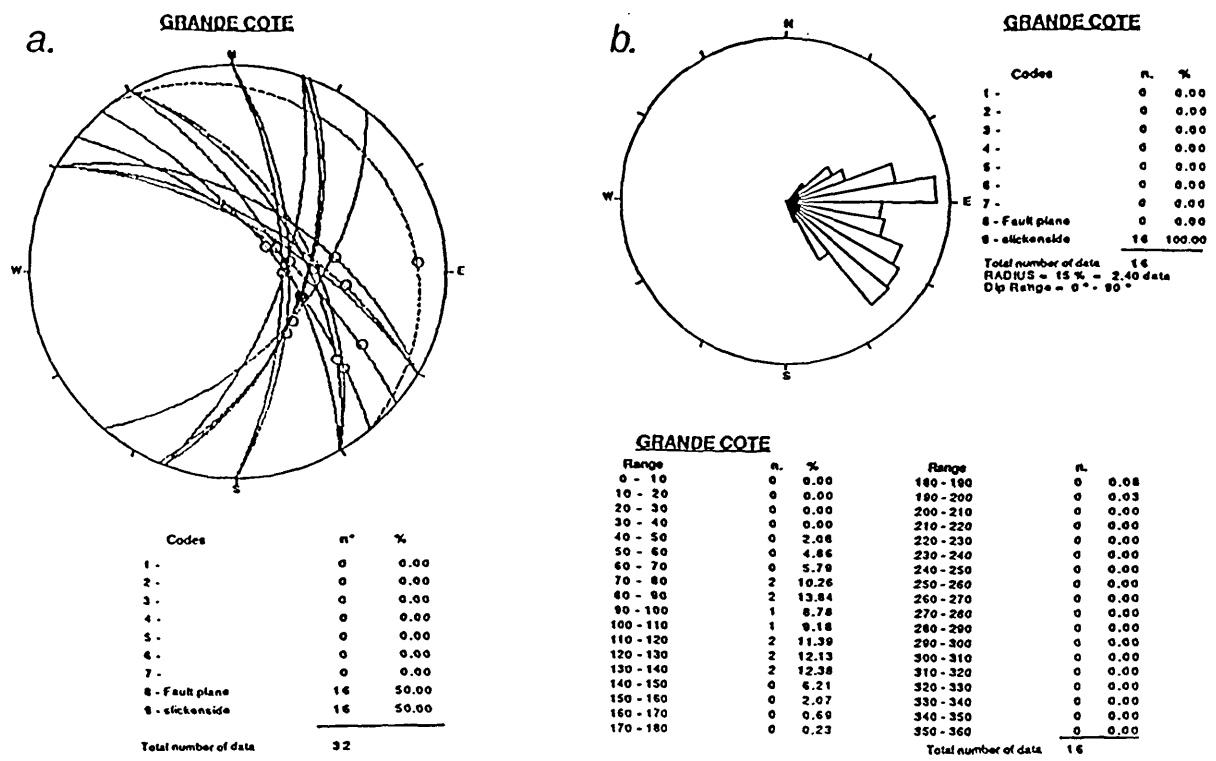


Figure 2.2.16: Equal angle (Wulff net) plot of slip-lineation and fault plane data (a), and rose diagram (b) of slip-lineation trends for the Grandes Cotes.

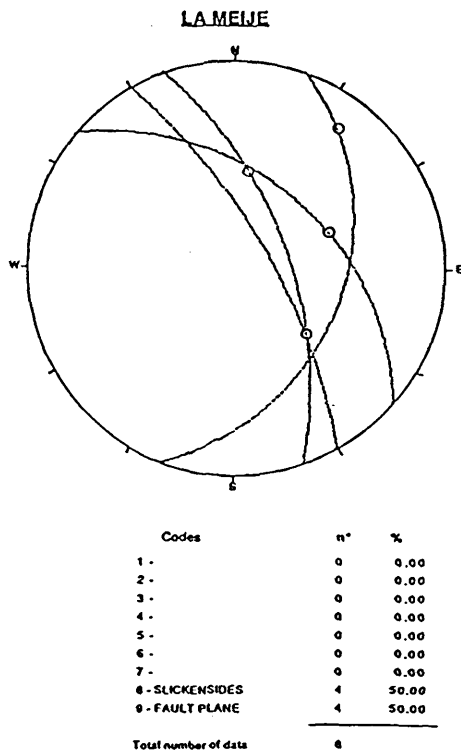


Figure 2.2.17: Equal angle (Wulff net) plot of slip-lineation and fault plane data from the La Meije thrust and its foot-wall.

imbrication, the slip-lineation direction varies in plunge direction from 110° to 070°, a substantial 40° transport range. The mean direction of thrust transport appears to be towards the W (Figure 2.1.15 a & b).

3. Grande Cote - Tete Moute: Local variations in slip-direction are recorded, although the general movement direction appears to have been towards the W to WNW (Figure 2.2.16 a & b). The wide ranging slip-direction recorded for individual fault planes may reflect the reactivation of a pre-existing fault system.

4. The La Meije - Combeynot thrusts: Data from the La Meije thrust and the immediate foot-wall indicate that the direction of overthrusting was towards the W (Figure 2.2.17). This is consistent with the cut-off trends for the La Meije fault, but disagrees with the SW movement interpretation proposed by Beach (1981a & b), and the early Alpine NNW interpretation proposed by Bartoli et al. (1974).

Mathews (1984) interprets the movement direction for the Combeynot thrust to be towards the WNW, basing this interpretation on stretching data recorded from the foot-wall. Some of this movement may be assigned to an early, pre-Eocene, Alpine phase of movement.

5. Ultra-Dauphinois cover thrust detachment: The inferred tectonic movement direction of the Ultra-Dauphinois thrust system is largely taken from the work of Beach (1981a & b; 1982) who recorded finite extension directions using stretched belemnites throughout thrust pile. In the lower part of the Jurassic cover sequence, towards the Triassic boundary, the extension directions are strongly suggestive of a SW movement direction. However, at higher structural levels in the thrust sequence, this direction changes to become NW-SE. These data are consistent with the field observations made by the author. Its significance will be discussed in Chapter 5.

2.3 Central Pelvoux area

2.3.1 Vallon Lanchatra:

The structure of the Lanchatra area is shown in Figure 2.3.1. The structures shown are largely continuous with those described from the northern margin of the Pelvoux massif (Section 2.2). On the S side of the Veneon valley at La Coche (G.R.2743,49835) (Figure 2.3.1), a steeply E-dipping fault juxtaposes Liassic shales against crystalline basement. A hanging-wall of Lower Lias, Trias and basement is folded into a tight N-S trending anticline, the vertical to overturned W-facing limb of which, is cut out against the basement fault plane. The hanging-wall basement anticline is deformed by a penetrative axial planar cleavage (Plate 2.3.1) which is clearly Alpine in age; the cleavage affects the overlying Triassic and Liassic sediments. This structure is continuous to the N and S. To the N it links with or is the same structure as the inverted half-graben described from the Tete Moute (Section 2.2.3.1). To the S this anticline can be traced to the head of the Lanchatra valley (Figure 2.3.1). The overall geometry is interpreted as a recompressed Jurassic age extensional structure which has remained in net extension. The intense hanging-wall strain is a common trait of such structures (Gillcrist et al, 1987; see also Chapter 3) and reflects the problems of reactivating a non-planar fault profile.

The continuation of this Jurassic age structure along the W slopes of La Coche is uncertain and it is apparently cut-out against a basement thrust fault (Figure 2.3.1). This thrust continues northwards into the Venosc area, where it may branch with the thrust uplifting the western margin of the Pelvoux massif (Section 2.2). Towards the S, the La Coche fault may have originally branched with what is interpreted to be a Jurassic age extensional structure, exposed along the eastern flanks of the Pied de Barry (Figure 2.3.1). Adjacent to this fault surface is a chaotic syn-sedimentary breccia, comprising decametric sized

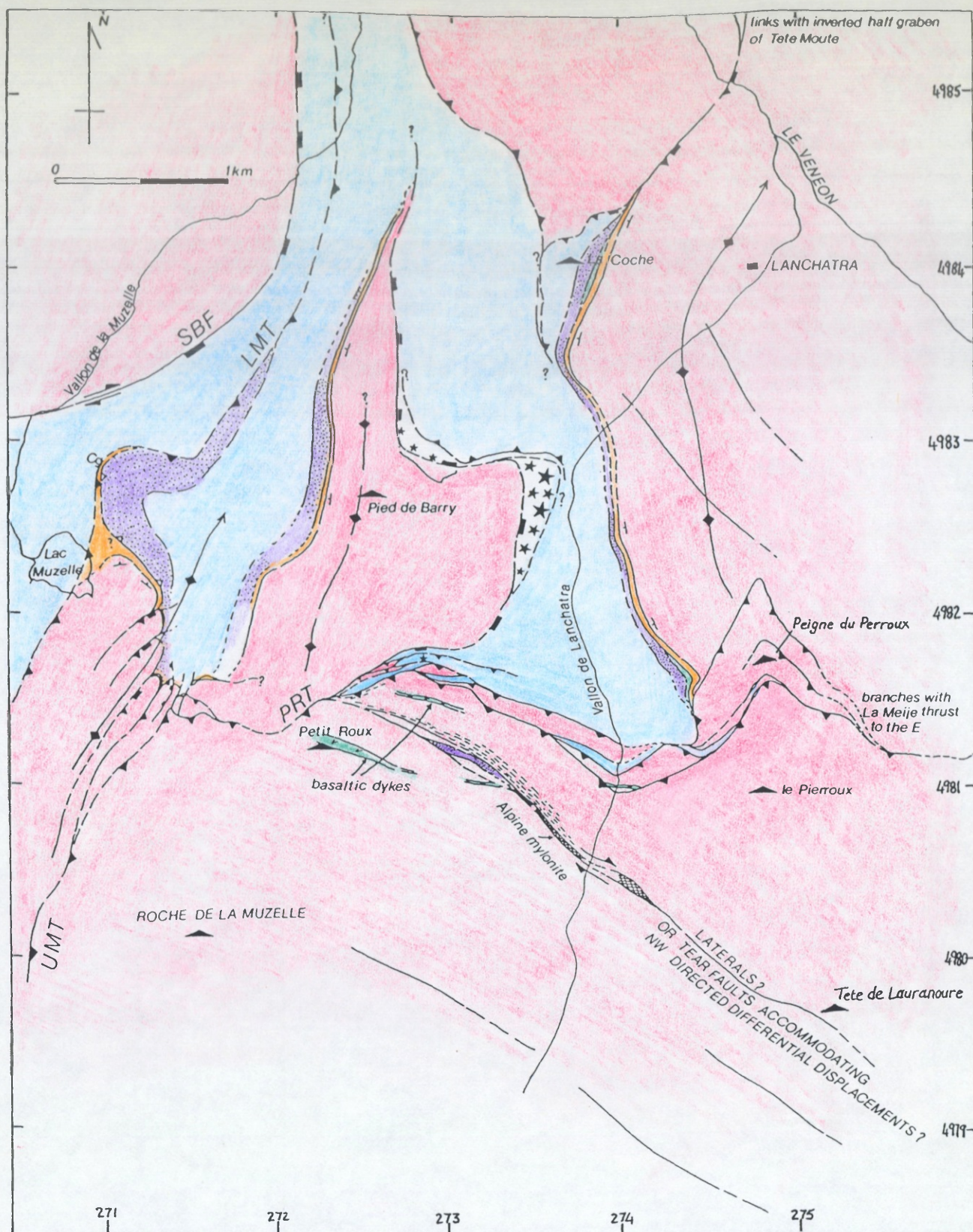


Figure 2.3.1: Simplified structural map of the Lanchatra area. The location and legend are indicated in Figs. 1.7 & 1.8. UMT-Upper Muzelle thrust, LMT-Lower Muzelle thrust, SFB-Ser Barbier fault, PRT-Petit Roux thrust. Based on mapping by the author.



Plate 2.3.1: The La Coche basement anticline with a penetrative axial planar Alpine cleavage. This is interpreted as an inversion structure related to the reactivation of the Le Barry normal fault. See text for details. Looking N with the Deux Alpes section in the background.



Plate 2.3.2: Deformed syn-sedimentary fault breccia in the foot-wall to the Petit Roux basement thrust. The deformed breccia comprises deformed basement clasts and Upper Liassic sediments. R1 shear bands give the sense of movement. The camera lens cap is 5.5 cm in diameter. See text for locality.

blocks of Triassic basalt, dolomite, and crystalline basement, very similar in character to that described from the foot of the Ornon fault scarp (Section 2.1.3.1). This syn-sedimentary breccia is overlapped by Upper Liassic shales with a notable absence of Lower Liassic age sediments. On the W face of the Pied du Barry, a reduced but normal succession of Triassic and Liassic sediments dip steeply to the W. This perhaps corresponds to the now folded, foot-wall dip slope of the Jurassic age extensional structure exposed on the eastern slopes of the Pied du Barry. Given the fault's present moderate dip (58°E), and evidence for back-steepening by deeper thrust structures in its foot-wall, the original attitude of this fault was probably gently dipping, as has been suggested for the Les Deux Alpes structures. Parallel to the fault plane, in the Liassic shales, is a thrust detachment corresponding to a zone of high strain characterized by shear bands consistent with a thrust sense motion. These cover thrust displacements are interpreted to derive from the partial thrust reactivation of this extensional fault plane, which also accounts for the tight hanging-wall anticline and upright pure shear cleavage in the Liassic cover sediments.

At the head of the Lanchatra valley, an E-W trending basement-cover thrust contact (Vernet, 1961) terminates the southern margin of the Lanchatra syncline (Figure 2.3.1). From map relationships, these basement-cover thrusts appear unaffected by the structurally lower N-S trending anticlines of La Coche and the Pied du Barry (Figure 2.3.1), implying an out-of-sequence break-back thrust geometry. The principal thrust, emplacing granite over Liassic cover sediments, crops out on the lower slopes of Le Petit Roux (Figure 2.3.1). This same thrust fault can be traced westwards along the lower slopes of La Muzelle as a continuous gently dipping feature. At le Petit Roux, in the foot-wall to this fault, the basement-cover contact is extensively imbricated (Figure 2.3.2). Barfety & Gidon (1981) interpret these basement thrust imbricates to be syn-sedimentary olistoliths, an interpretation not followed here, although smaller irregular shaped blocks of basement

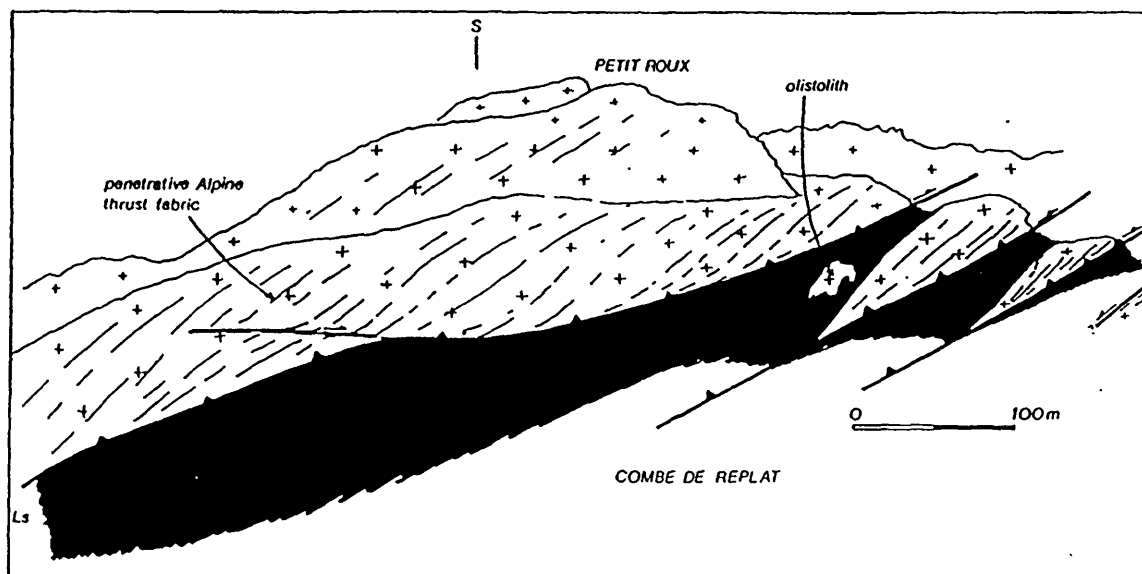


Figure 2.3.2: Field sketch of basement-cover thrust imbricates in the foot-wall to the Petit Roux thrust. Crosses-crystalline basement, Ls (black)-Lias schisteux.

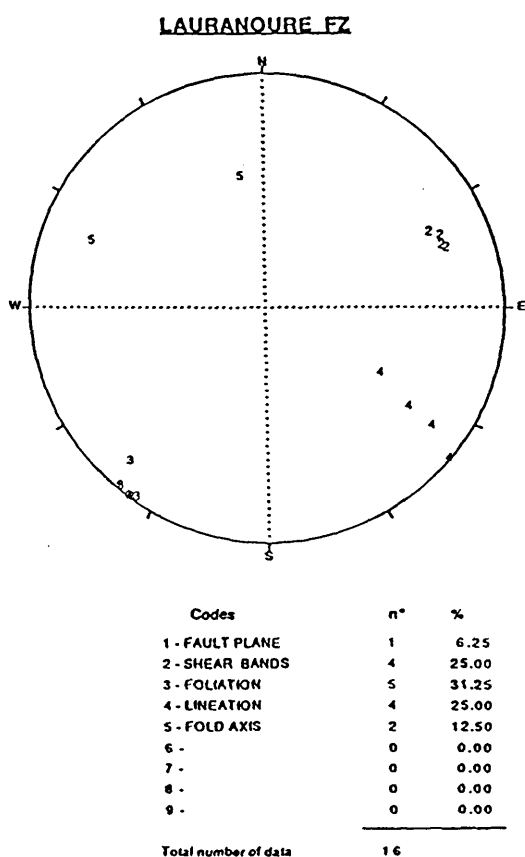


Figure 2.3.3: Equal area (Schmidt net) plot of structures in the Lauranoure fault zone, planes are plotted as poles.

documented from the zone of basement-cover imbrication are possibly sedimentary olistoliths (Figure 2.3.2).

The larger elongate slices of crystalline basement are clearly basement-cover thrust slices, their lower surfaces corresponding to discrete fault breaks, which upon passing into basement, broaden out into wider shear zones. The thrust sheets themselves are deformed by a penetrative Alpine cleavage, often deformed by Alpine age extensional shear bands (Figure 2.3.2). At the base of the lowermost thrust imbricate, where it cuts into the basement of the Pied du Barry, is a syn-sedimentary breccia of basement clasts and Liassic shales which is penetratively deformed by subsequent thrusting (Plate 2.3.2). The orientation of the mineral stretching direction and slickenside lineations recorded from these thrust surfaces, indicate the thrust transport direction was towards the NW. This is consistent with the direction obtained from shear bands and other kinematic indicators.

The exact origin of these basement thrust slices remains uncertain; they may have been derived locally from a foot-wall imbrication of the Pied du Barry fault scarp, implying that the main thrust movements are accommodated on the upper Petit Roux fault and that the imbricates are not far travelled. However, given that this zone of basement-cover thrusting corresponds to a zone of dominantly strike-slip movement, these basement imbricates could equally have been transported for a considerable distance.

To the E, the basement forming the Peigne du Pierroux ridge (G.R.2756,49807) is also extensively imbricated, and represents the eastward continuation the Petit Roux thrusts. Sheared lenses of Liassic cover sediments were documented along the thrust faults which are predominantly within the basement.

E of Peigne du Pierroux, at the level of the Veneon valley, the continuation of this basement thrust surface is difficult to locate - it must either branch with, or be the same fault as the La Meije thrust which was mapped along

the lower slopes of the Aiguille du Plat de la Selle, to the E (see Figure 2.2.1) (see also Bartoli et al., 1974).

Within the La Muzelle-Petit Roux thrust sheet, a number of vertical NW-SE trending faults deform the Laurantou granite (Figure 2.3.1). The most important of these was mapped from the Tete de Laurantou (G.R.2750,49790) across to the Petit Roux, where cover sediments are incorporated within the fault zone, confirming its Alpine age. The width of deformed zone within the Laurantou granite attains a thickness of 300m. Bartoli et al. 1974, describe this fabric as a blastomylonite, which is deformed by more discrete brittle cataclastic zones. Shear bands, and the sense of curvature of the main shear fabric are consistent with a dextral motion, confirmed by the orientation of the cleavage, shear folds and a horizontal lineation recorded from the Liassic cover of le Petit Roux. The orientations of the structural data recorded from this fault zone are shown stereographically in Figure 2.3.3.

These vertical faults are interpreted as tear faults within the La Muzelle-Petit Roux sheet which may accommodate a differential NW directed displacement. They may however, have been initiated as Variscan fault zones, as they show the same NW-SE trend as most of the Variscan structures in Western Europe.

2.3.2 Summary

The La Meije-Muzelle thrust faults divide this region up into two distinctive structural domains (Figure 2.3.4). A structurally lower northern domain, comprises Jurassic age extensional faults which have been restacked during Alpine compression. This northern domain is characterized by a strong cataclastic deformation of the basement, in which an Alpine age foliation is well developed (Bartoli et al., 1974). In contrast however, the structurally higher southern domain is characterized by the comparatively undeformed La Meije, Etages, and Turbat-Laurantou granites

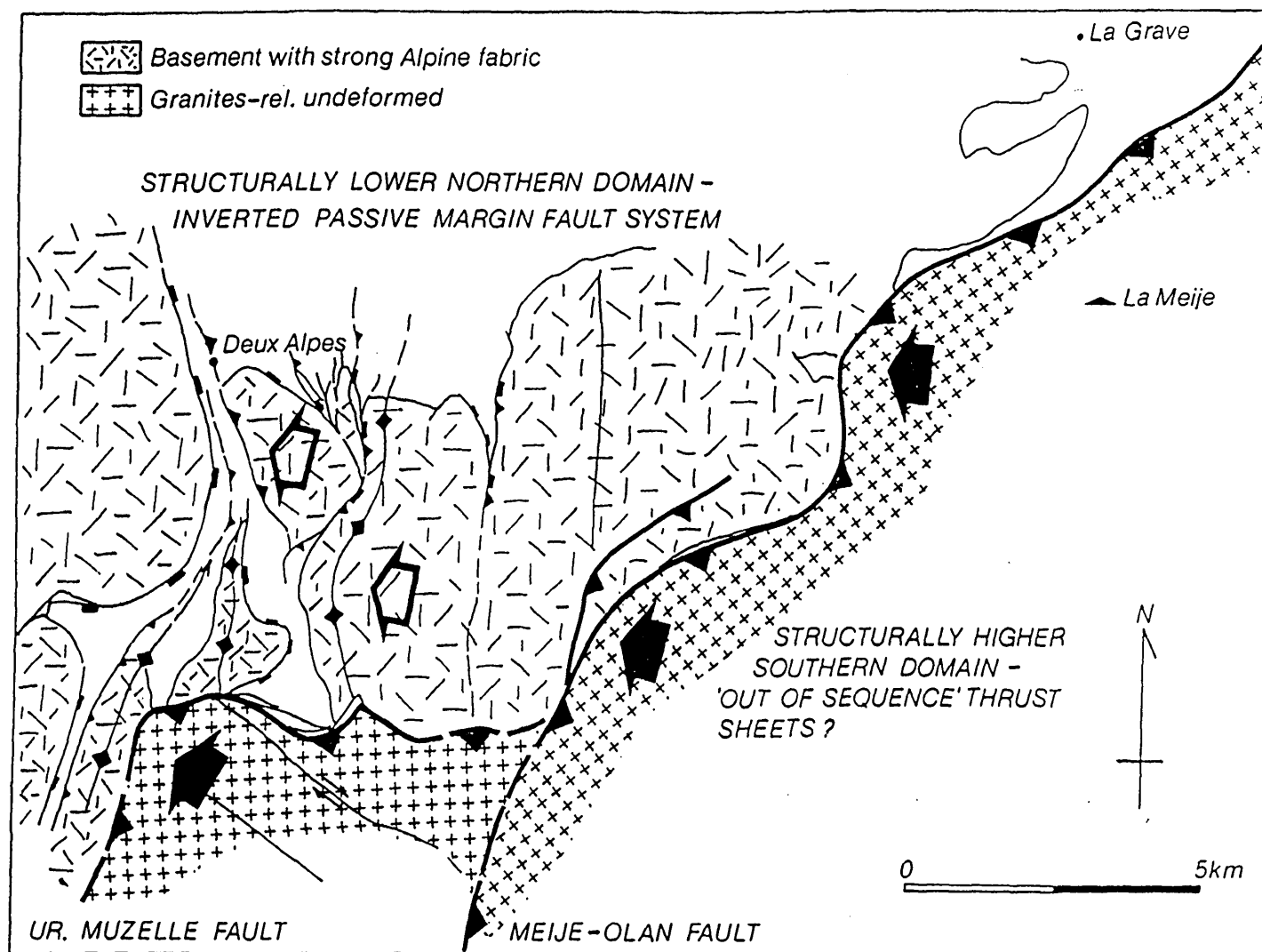


Figure 2.3.4: Simplified map of the northern and central parts of the Pelvoux massif shows the La Meije break-back thrust system which appears to truncate older inversion structures, dividing the area into two distinctive structural domains (modified after Bartoli et al., 1974). The arrows indicate the direction of Alpine thrusting.

which have been carried passively in the hanging-wall to the La Meije and Muzelle basement thrusts. This contrast in structural style between these two domains may reflect the sequence of deformation and thrusting.

The intensity of the Alpine cleavage in the basement of the northern domain can be assigned to the attempted reactivation of the passive margin fault system. The map relations seem to suggest that the La Meije thrust and subsequent foot-wall deformation was largely break-back and out-of-sequence, thrusting undeformed granitic masses over the already stacked extensional fault system. Unfortunately there have been no accurate means of dating the inversion structures, described from the northern margin of the Pelvoux, other than from geometrical relationships, and the break-back nature of the La Meije thrust can only be inferred.

2.3.2 La Muzelle:

This section includes the Vallon de la Muzelle, Col du Vallon and Col de la Muzelle areas (Figures 2.3.1 & 2.3.5). The location of these different map areas is indicated in Chapter 1 on Figure 1.7.

1. Ser Barbier: The N-S trending Pied Moutet Jurassic age normal fault described in Section 2.2 is continuous with the Ser Barbier fault exposed in this area (Figure 2.2.1), which juxtaposes deep water Liassic shales against the Rochail basement. During the Alpine compression this fault was active as a dextral strike-slip fault, the sense of shear being determined from shear fabrics observed along and adjacent to the fault plane.

2. Col du Vallon: At the Col du Vallon (Figure 2.3.5), the Ser Barbier fault intersects a NNE-SSW trending normal fault which juxtaposes Upper Liassic shales in the E against Carboniferous basement. Barfety & Gidon (1981) document the presence of syn-sedimentary olistoliths comprising Carboniferous Houiller and crystalline



Figure 2.3.5: Simplified structural map of the Vallon Valsenestre area (location and legend indicated in Figs. 1.7 & 1.8). LMT-Lower Muzelle thrust, CDLMF-Col de la Muzelle fault, BLT-Breche de Lauvitel thrust, CVT-Cabane du Vallon thrust, SLT-Signal de Lauvitel thrust, LPA-La Paletas anticline. Based on mapping by the author.

basement, enveloped in the Upper Liassic shales. Both of these faults, interpreted to have a Jurassic age of initiation, are sliced through by later Alpine thrusts. On the S-ridge of the Aiguille de Venosc (G.R.7440,49835) a S-dipping Alpine thrust fault branches from the Ser Barbier fault and can be followed along to the W-face of the Tete de la Muraillette at an altitude of 2350 m (Figure 2.3.5). This fault intersects the Muraillette mountain as a gently dipping planar feature which overthrusts a small sliver of Triassic dolomites along the N-ridge of the Muraillette. Another small slice of Triassic dolomites marks the trace of the thrust fault along the W-face.

On the NE-face of the Muraillette, the Rochail basement is uplifted along a WNW-ESE trending, steeply S-dipping fault. In the immediate foot-wall to this fault, thrust imbricates of Carboniferous Houiller and Upper Liassic shales were mapped; this interpretation differs from the sedimentary olistolith interpretation proposed by Barfety & Gidon (1981). The eastward continuation of this fault towards the Lac Muzelle (Figure 2.3.5) is not clear, suggesting that this fault and an adjacent fault may have been Jurassic age transfer faults to the Pied Moutet and Col de la Muzelle early normal faults; however further evidence is required to substantiate this hypothesis. Stretching lineation data from this fault and the foot-wall imbricates record a tectonic movement direction towards the WNW (Figure 2.3.17). From map relationships this fault appears to be truncated by the gently dipping thrust mapped on the W-face of the Muraillette, which from along strike relationships to the S, is interpreted to belong to a family of slightly later SW to WSW (?) directed thrusts (See Vallon Valsenestre, Section 2.3.3).

3. La Muzelle: To the E of the Lac Muzelle, the Pelvoux basement is uplifted on the Muzelle thrust (Lower Muzelle fault, see Figures 2.3.1 & 2.3.5). The basement in the hanging-wall of this thrust is folded into a N-S trending anticline, which plunges N-wards under the Mesozoic cover sediments. The continuation of this thrust in the

sedimentary cover is delineated by the sporadic occurrence of Triassic cargneules, and a hanging-wall sequence of Lower Liassic limestones and Upper Liassic shales, which are thrust over Upper Liassic shales in its foot-wall. The thrust displacements pass N-wards into the Venosc "graben", now greatly shortened and constricted by thrust displacements which have climbed out of the basement. To the S, the orientation of this thrust fault, emplacing basement over Upper Liassic shales, steepens dramatically.

The Col de la Muzelle is characterized by a narrow pinched in zone of cover sediments limited by the lower Muzelle thrust on its eastern side and a steep N-S trending, E-dipping fault on its W-side. Adjacent to this fault, termed here the Col de la Muzelle fault, there is a notable absence of Lower Liassic sediments with Mid-Jurassic shales juxtaposed against the Rochail basement, and a Jurassic age interpretation is proposed. The Alpine cleavage in the cover sediments is extremely penetrative and upright suggesting that this fault may have blocked some of the Alpine contractional displacements. Horizontal striae observed along the fault plane, and shear folds are indicative of a dextral strike-slip movement along the fault. This may well correspond to a late reactivation.

The hanging-wall anticline of the lower Muzelle thrust is deformed by what are interpreted to be a number of Liassic age extensional faults restacked during the Alpine compression. Some of these faults still preserve a net extension. As was proposed for the northern margin of the Pelvoux massif, it may be inferred that these Liassic age faults initially possessed gentle E-dips.

2.3.3 Vallon Valsenestre:

The Vallon Valsenestre area is shown in Figure 2.3.5. The present structure is the result of interaction between inversion (the reactivation of older normal faults) and later discrete thrusting which develops independantly of pre-existing structure.

1. High level break-back thrust system:

The slightly later, gently dipping thrust fault described from the Col du Vallon area, splays into several basement thrust imbricates towards the S. These intersect the E-W trending topographic ridge of the Signal de Lauvitel-Breche de Valsenestre (Figure 2.3.5) and can be traced southwards to the E of the Valsenestre village by small cover "pinches" lying along the thrust planes. This zone of overthrusting delineates the eastern margin of the Lac Labarre "syncline" (Barfety & Gidon, 1984), or more precisely, the eastern limit of the Bourg d'Oisans sedimentary cover in this area.

The Signal de Lauvitel thrust is a gently dipping planar feature which post-dates a large amplitude NW-SE trending basement fold, the La Paletas anticline (Section 2.1.3.2). From its map geometry the Signal de Lauvitel thrust "oversteps" this structurally lower anticline and cuts back sharply towards the E to branch with the imbricate thrust system exposed on the Breche de Lauvitel-Breche de Valsenestre section (Figures 2.3.5 & 2.3.7). To the SE, the surface trace of these thrust faults come together and branch S of the Vallon Valsenestre where they once again truncate the SE continuation of the La Paletas anticline.

To the E, two important thrust faults overthrust Mesozoic cover sediments along the Breche de Lauvitel (G.R.7408,49808). In the hanging-wall of the Signal de Lauvitel thrust and the foot-wall to a thrust exposed on the W face of the Breche de Lauvitel, the Calcaire du Paletas formation (Barfety & Gidon, 1984) rests directly on the Signal de Lauvitel basement, with Triassic and Liassic formations missing. This contact is sedimentary and not tectonic, apart from some minor, later thrust slip. The bedding orientation of the sedimentary cover is parallel to that of the basement-cover contact as is the main Alpine cleavage. This absence of Triassic and Liassic age

sediments is interpreted to be the result of foot-wall uplift with corresponding erosion of the reduced sedimentary sequence deposited on the apex of the Ornon fault block (Section 2.1.2). The Calcaire du Paletas formation is overthrust to the E by a thrust fault on the W-face of the Breche de Lauvitel, called here the Cabane du Vallon fault. The geometry of this fault is shown in Figure 2.3.7. Deformed Triassic spillites and some cagnueles outcrop in the foot-wall to the fault and the hanging-wall is collapsed down towards the E by a minor normal fault. Stretching lineations recorded from the fault plane are consistent with a WSW direction of overthrusting (Figure 2.3.17).

Farther to the E is another important basement-cover thrust, the Breche de Lauvitel thrust. The hanging-wall of this eastern fault is deformed by a number of brittle Alpine imbricate thrust faults. Along an E-W profile, this thrust cuts upwards through the stratigraphy towards the W, emplacing crystalline basement over a foot-wall of Triassic dolomites and Lower Liassic sediments (Figure 2.3.6). However, this same thrust fault proceeds to cut stratigraphically downwards towards the WNW, limiting the thrust transport direction towards the WSW to SW, unless the fault is cutting through a pre-existing high structure. In the foot-wall of this fault, the orientation of shear bands (Plate 2.3.3), mineral stretching lineations and striations recorded from along the fault plane show the thrust transport to be towards the WSW (Figure 2.3.17), suggesting this structure corresponds to an oblique frontal cut-off. It must be emphasised that cut-off trends may be misrepresentational of the true thrust transport direction, giving more information on the trend, or orientation of pre-existing compressional or extensional structures. This fact and the absence of a complete stratigraphy, precludes any accurate determination of the displacement.

To the E a vertical to steeply E-dipping normal fault, which branches onto the Breche de Lauvitel thrust, reactivates this fault into extension. Indeed most of these

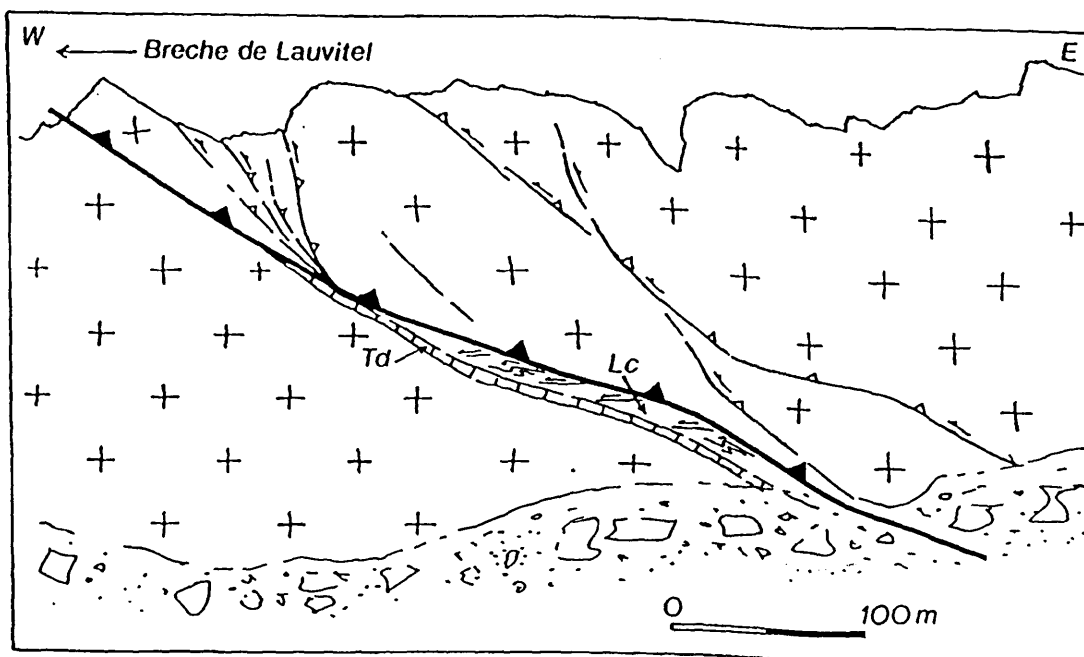


Figure 2.3.6: Field sketch showing hanging-wall deformation to the Breche de Lauvitel thrust (see Plate 2.3.3). Crosses-crystalline basement, Td-Triassic dolomites, Lc-Lias calcaire.

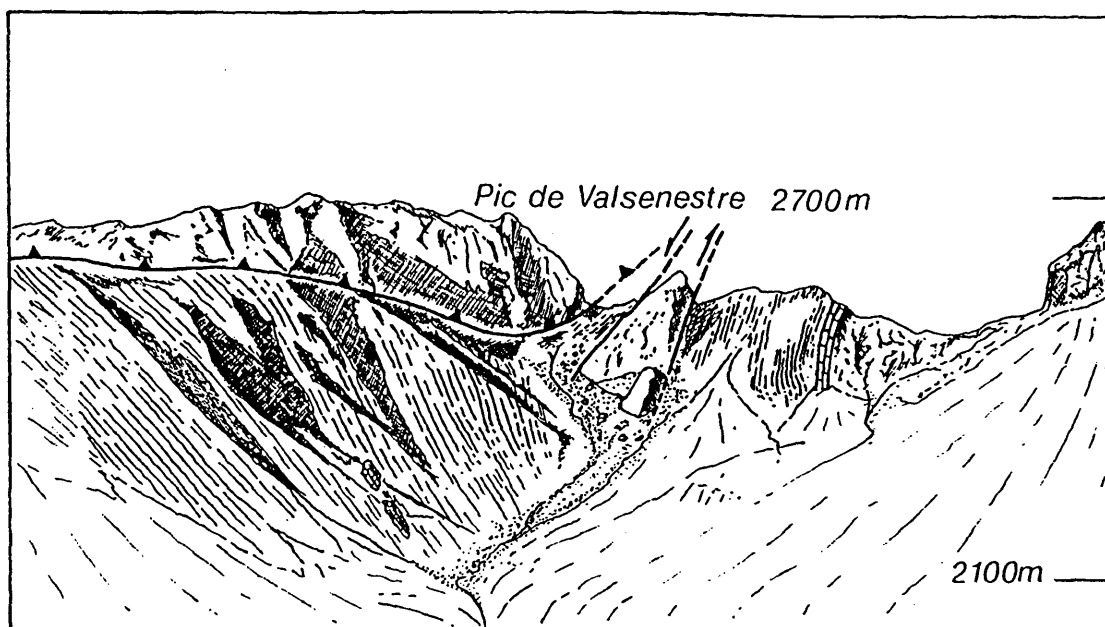


Figure 2.3.10: Line drawing of foot-wall short cut/strike-slip imbricate along the Col de la Muzelle fault (G.R.7421,49755). The Pic de Valsenestre thrust is also shown.



Plate 2.3.3: Hanging-wall deformation of the Breche de Lauvitel basement thrust. See Fig. 2.3.6 for location and details.



Plate 2.4.1: Shows the Lower and Upper Aiguille de Morges thrusts (arrows) repeating Triassic basalts and the Liassic cover sequence. Looking S.

SW

NE E

W

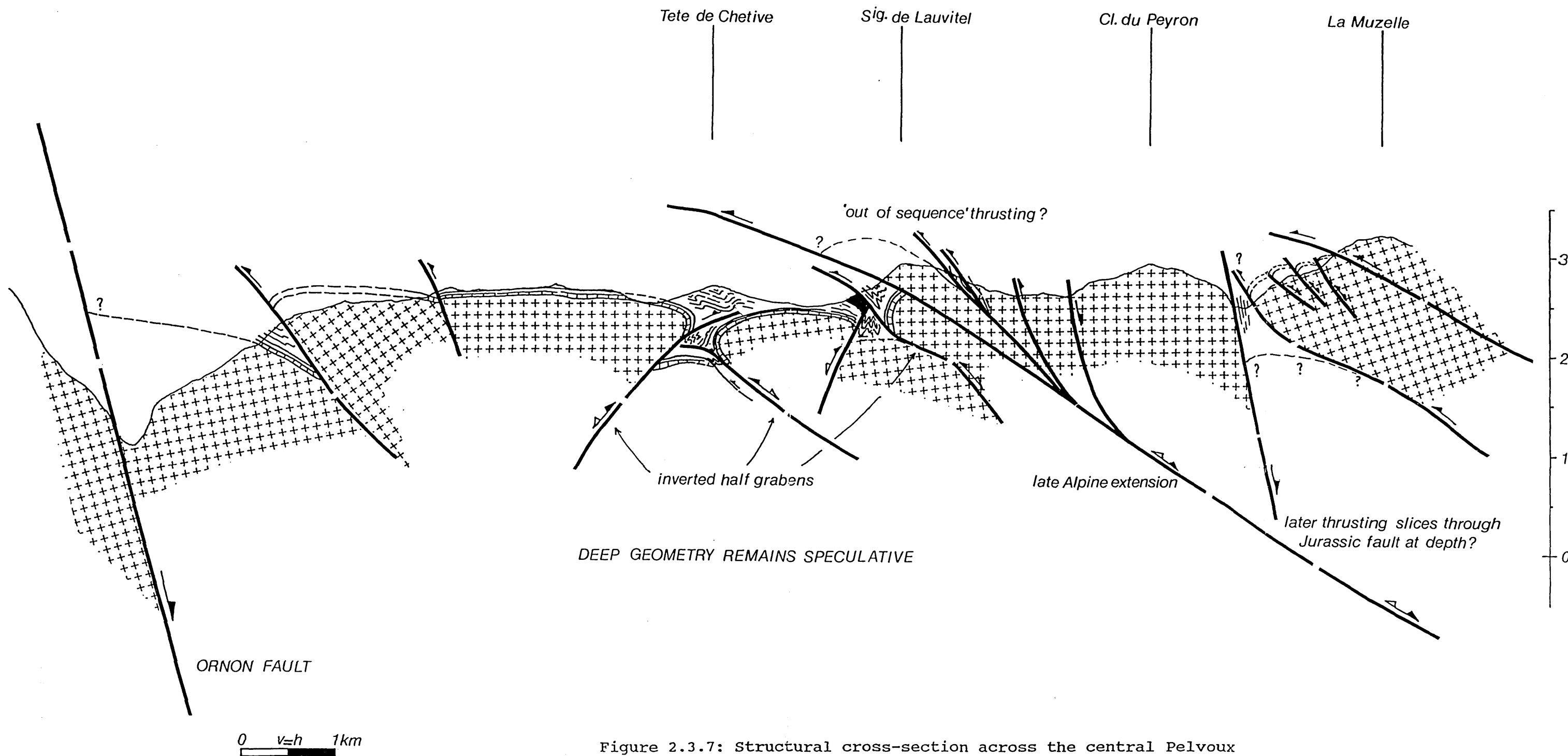


Figure 2.3.7: Structural cross-section across the central Pelvoux area (location of section line shown as AA' on Figure 2.2.1). It shows the principal thrusts, reactivated Jurassic normal faults, and late Alpine extension. Movements occur out of the section plane and the section is not balanced. Crosses-crystalline basement, brickwork-Triassic dolomites and basalts, Jurassic cover is indicated by fine bedding traces, and the Terres Noires is shown in black.

higher level basement thrusts have been reworked by post-Alpine extension.

Both the Cabane du Vallon fault and the Breche de Lauvitel fault can be traced across the Vallon Valsenestre towards the S where they may branch with thrusts along the western slopes of the Pic de Valsenestre (Figure 2.3.8).

2. Lac Labarre "syncline":

A band of Mesozoic cover sediments parallels the NW-SE trend of the La Paletas anticline. This has been termed the Lac Labarre syncline by Barfety & Gidon (1984) and represents the constricted SE continuation of the Bourg d'Oisans half-graben in this area (Figure 2.3.5). The western margin of the sedimentary cover is limited by a NW-SE trending, steep reverse fault, which dips to the SW. Along this fault, at le Pisse Roux (G.R.7410,49794), sedimentary olistoliths of Triassic dolomite were mapped (Figure 2.3.5). In the foot-wall to this steep reverse fault, the cover sediments of Lias calcaire and Calcaire du Paletas are deformed into NE-verging folds (back-folds) which to the E progressively become upright and eventually W-verging, generating a minor fold-facing confrontation. The W-verging structures are generated in the foot-wall to the SE continuation of the Cabane du Vallon fault. Towards Lac Labarre (Figure 2.3.5) this back-folding and thrusting become less evident and the cover sediments are deformed by tight upright chevron-type folds, which trend NW-SE (N140) and contain a penetrative upright axial planar cleavage. At the Lac Labarre, Liassic sediments overthrust Middle Jurassic and Oxfordian black shales along a gently dipping thrust fault which dips 30° to the NE. A penetrative thrust-related cleavage is present in both the foot-wall and hanging-wall (45° E dip) and the stretching lineation direction is consistent with a SW direction of overthrusting. These cover thrust displacements are interpreted to emerge from the La Paletas basement thrust located in the subsurface. The W-ward continuation of this section is shown in Figure 2.3.7. At the Combe Guyon

(G.R.7395,49792), to the SE of the Tete de Chetives (G.R.7389,49795), a basement anticline of large wavelength plunges NW under the Mesozoic cover sediments. In the foot-wall to this is a minor thrust imbricate which generates a compressional hanging-wall fold in the Triassic dolomites. To the SE of this, directly S of the Tete de Chetives summit, is a large wavelength NE-verging back-fold and thrust which affects the Lias, Trias, and basement. The overlying Liassic cover sediments of the Tete de Chetives are deformed by a number of bedding-parallel thrusts and W-verging folds. The section line (Figure 2.3.7) was continued towards the Ornon fault where a number of basement thrust structures uplift the basement terminating the southward continuation of the Bourg d'Oisans half-graben. These most probably correspond to reactivated Jurassic age faults.

A feature of note concerning the Lac Labarre structures, is the arcuate trend and continuity of both the La Paletas anticline and the cover folds, which continue into the Bourg d'Oisans area. It is possible that this swing in structural trend was inherited from the Jurassic age extensional faults in this area.

2.3.4 Pic de Valsenestre:

The structures described in Sections 2.2.2 and 2.2.3 are continuous with those of this area (Figures 2.3.1, 2.3.5, & 2.3.8). The location of the map areas is shown in Figure 1.7 (Chapter 1).

1. SE continuation of the Col de la Muzelle early normal fault: The N-S trending strip of cover sediments described in Section 2.2.2, towards the S divides into two branches. The western branch parallels the Col de la Muzelle early normal fault which passes to the W of the Pic de Valsenestre (Figure 2.3.9) and eventually links with the Aspres-les-Corps lineament further S (Section 2.5). The Jurassic age interpretation is substantiated by the presence of two isolated sedimentary olistoliths, enclosed

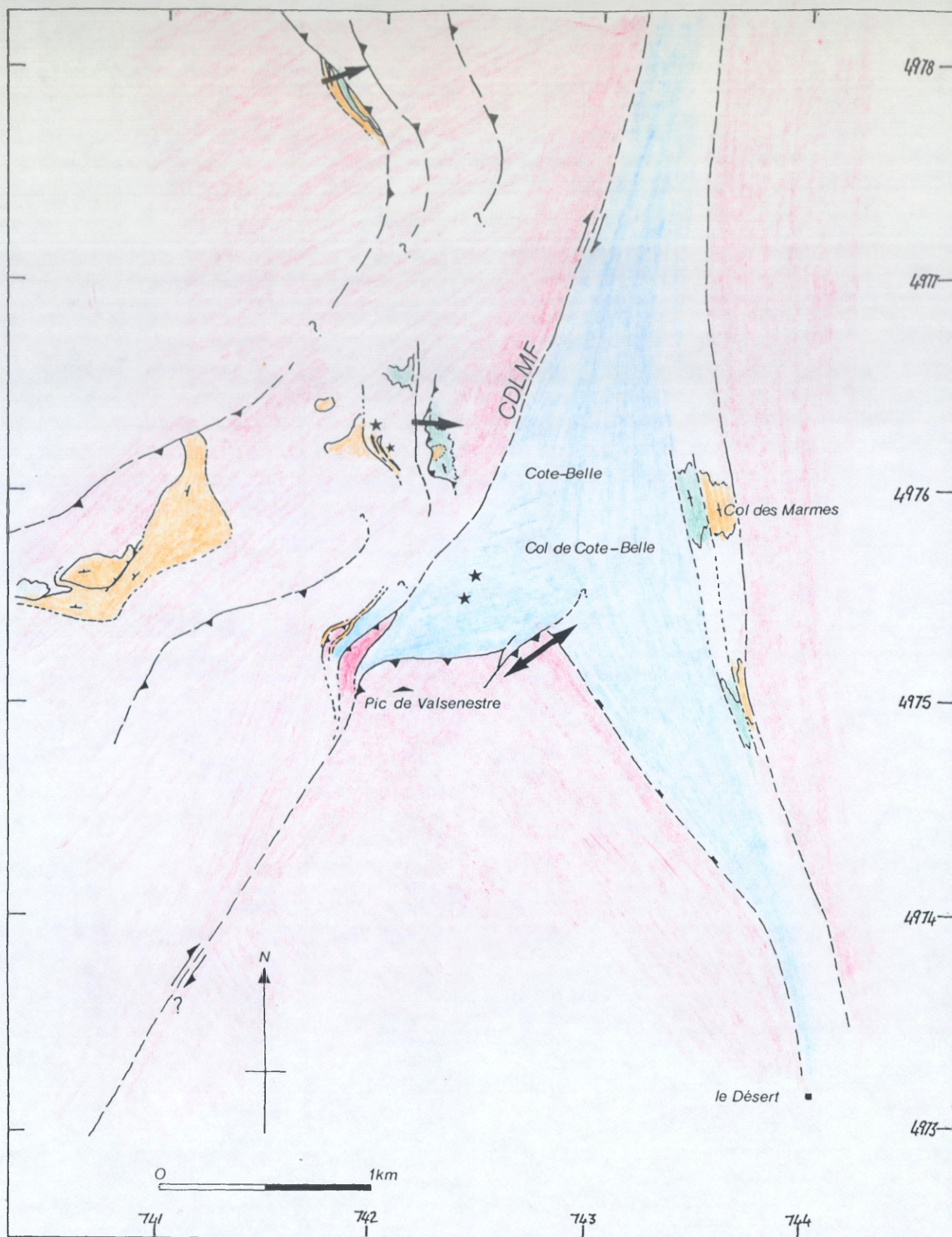


Figure 2.3.8: Simplified structural map of the Pic de Valsenestre area (location and legend indicated in Figs. 1.7 & 1.8). Stars show Jurassic olistoliths. Based on mapping by the author.

in cover sediments on the N-face of the Pic de Valsenestre. Where the La Muzelle fault intersects this N-face, a sizeable fault-bound sliver of crystalline basement is completely enveloped in the surrounding sedimentary cover (Figure 2.3.10). The lower surface corresponds to a thrust fault which suffered a SW-direction of oblique reverse-slip. The upper surface is interpreted to be the La Muzelle early normal fault plane, which was subsequently imbricated during these strike-slip to reverse thrust movements. This interpretation differs slightly from the one proposed by Gillcrist et al. (1987), which is now in need of modification. The foot-wall of the La Muzelle fault is also imbricated, however, the steep walled SW-face of the Pic de Valsenestre restricts both the La Muzelle fault and basement imbricates to be followed towards the SW. A. Pecher (pers. comm. 1985) has mapped lenses of deformed cover sediments along the SW continuation of this fault zone.

A number of basement thrusts were mapped to the W of this fault and can presumably be correlated with the basement-cover imbrication in the Vallon Valsenestre (Figure 2.3.8 & 2.3.9).

2. Cover deformation of the Cote-Belle: As at the Col de la Muzelle, the cover sediments are characterized by a penetrative upright deformation fabric which obliterates the original bedding trace. The main cleavage is sub-parallel to the trend of the La Muzelle fault scarp exposed towards the W. In the Middle Jurassic shales, this fabric is locally folded by tight asymmetric shear folds (< 0.5 m scale) which also fold an earlier lineation. Such shear folds are characteristic of the deformation affecting the Middle Jurassic shales. In this narrow strip of sedimentary cover, both the cleavage and small scale shear folds are deformed by later shear bands. These shear bands are widespread and of consistent orientation (030-040° strike and dip 30-45° SE) and sense of shear is always dextral. The orientation of these structures is depicted stereographically in Figure 2.3.12. They are interpreted as

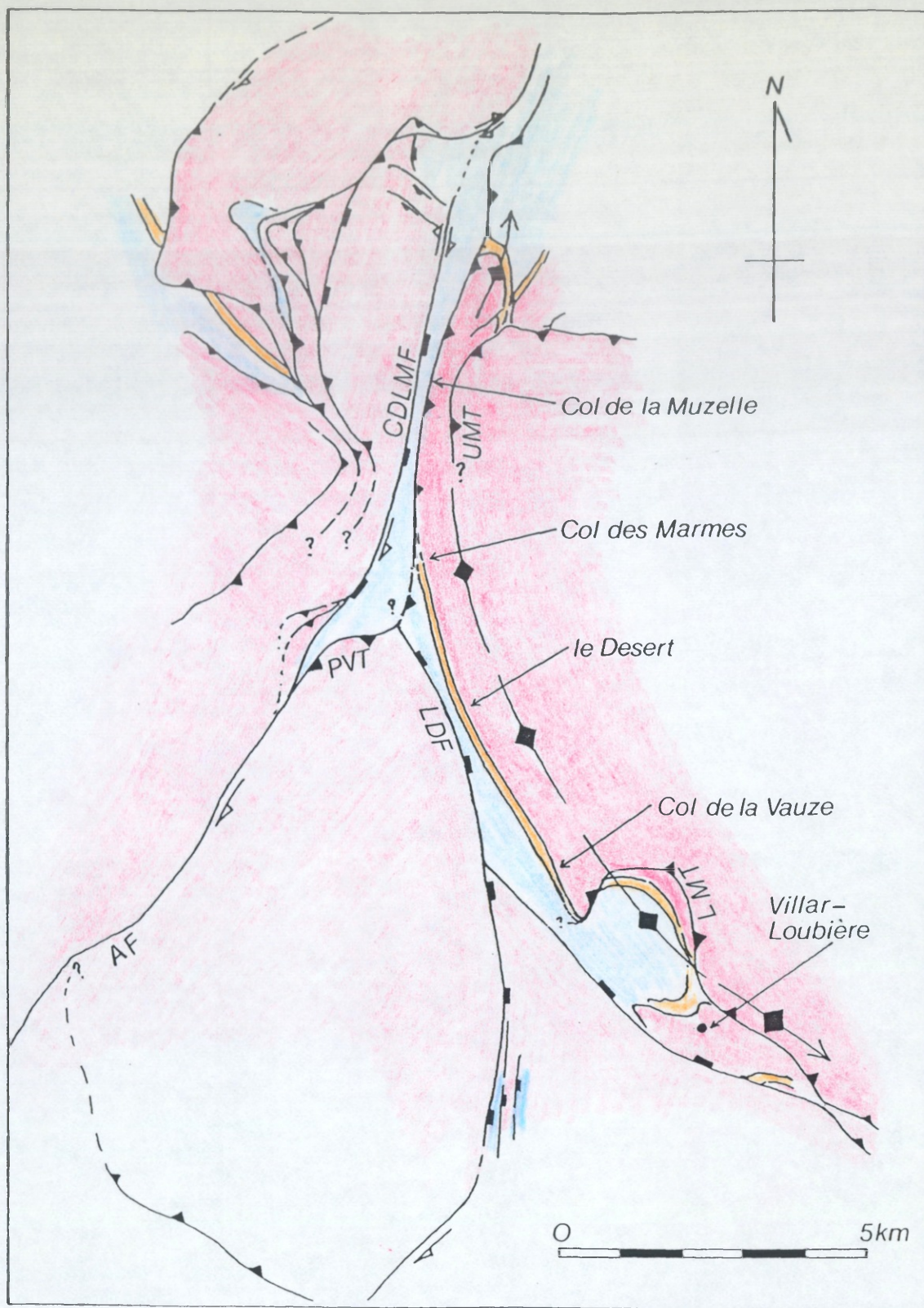


Figure 2.3.9: Simplified structural map of the central Pelvoux area (location and legend indicated on Figs. 1.7 & 1.8). UMT-Upper Muzelle thrust, LMT-Lower Muzelle thrust, CDLMF-Col de la Muzelle fault, LDF-Le Desert fault, PVT-Pic de Valsenestre thrust, AF-Aspres fault. Based on mapping by the author.

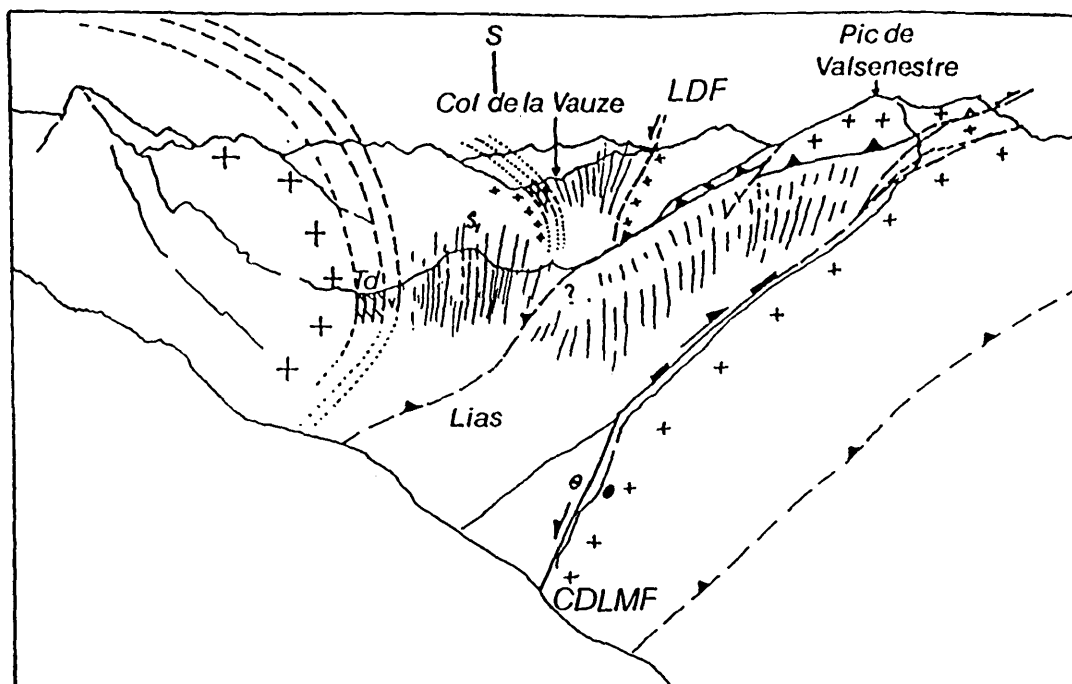


Figure 2.3.11: Panoramic sketch across the Col des Marmes, Pic de Valsenestre, with Col de la Vauze in the background. Shows the La Muzelle anticline and the Col de la Muzelle fault (CDLMF). Crosses-basement, Td-Triassic dolomites and spillites, LDF-Le Desert fault.

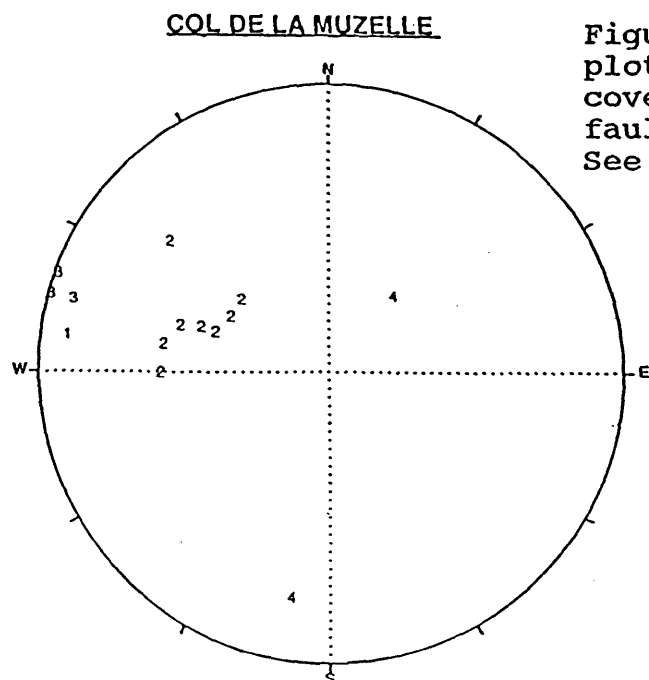


Figure 2.3.12: Equal area (Schmidt net) plot of structures along and in the cover adjacent to the Col de La Muzelle fault. Planar features plotted as poles. See text for details.

Codes	n°	%
1 - FAULT PLANE	1	7.14
2 - SHEAR BANDS	8	57.14
3 - CLEAVAGE	3	21.43
4 - LINEATION	2	14.29
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 -	0	0.00
9 -	0	0.00

Total number of data

14

accommodating a late dextral reverse-strike-slip movement of the cover along the La Muzelle fault. At the Col de la Cote-Belle to the E (G.R.7438,49760), only a single penetrative upright vertical cleavage with a pronounced down-dip mineral stretching lineation is observed.

3. Col des Marmes: At the Col des Marmes (Figure 2.3.9) a normal stratigraphic sequence of Lower Liassic and Triassic basalts and dolomites, rest with a normal contact on crystalline basement. The bedding is vertical and strikes N-S. This is interpreted to be the forelimb of a large basement anticline which to the N cuts out against the lower Muzelle thrust fault (see Section 2.3.2), rotated eastwards by later foot-wall imbrication. The pure shear deformation fabric recorded in the cover sediments supports the Jurassic age interpretation for the Col de la Muzelle fault.

4. Pic de Valsenestre thrust: The Valsenestre thrust corresponds to a gently S-dipping planar feature (Figure 2.3.11). Shear fabrics and lineations recorded from the exposed fault plane are consistent with a SW direction of thrust movement. Three models may be proposed to explain its geometry:

i) To the N this fault branches with, or is the same fault as the lower Muzelle thrust, and slices through an earlier Jurassic fault which may be interpreted as a transfer fault to the Col de la Muzelle fault (Figure 2.3.9). This model is consistent with the apparent "late" geometry of the Valsenestre thrust which truncates the upright cleavage of the Cote-Belle. Both the Col de la Muzelle fault and a possible NW-SW trending transfer fault, may have blocked some of the cover thrust displacements associated with the emplacement of the Lower Muzelle thrust sheet. An ensuing foot-wall collapse sequence resulted in the Valsenestre thrust slicing through the steep Jurassic transfer fault. One problem with this model is the whereabouts of the frontal part of this thrust. It may be present W of the Pic de Valsenestre, forming a basement-basement thrust contact,

making it difficult to locate.

ii) From map relationships alone, the Valsenestre thrust may correspond to a frontal thrust ramp generated during an early N to NNW phase of crustal shortening (see Bartoli et al., 1974; Gidon, 1979), which reactivated the existing Jurassic fault pattern. In this case, both the Col de la Muzelle fault and its Jurassic age transfer fault, would have been reactivated as oblique frontal ramps during such a compression. This model however, is not compatible with the apparent late age of the Valsenestre thrust, which post-dates an earlier penetrative upright cleavage.

iii) The map pattern of the Valsenestre thrust and the steep E-dipping fault which passes into the Le Desert area (Figure 2.3.9), may be interpreted within a folded thrust context. This implies that large displacements are associated with the Valsenestre thrust, which would have to be equated with a higher level basement thrust in the more internal Pelvoux. Subsequent movement on the lower level Muzelle thrust would fold and dome up the overlying Valsenestre thrust sheet to give the present map geometry. However, no evidence along the Le Desert fault was found to support such an interpretation.

The first solution proposed is considered to be the most satisfactory solution, although clearly further work is required.

2.3.5 Eastern branch of the Muzelle sedimentary cover:

The Le Desert fault, apparently cut out by the Valsenestre thrust, can be traced south-eastwards in the central Pelvoux towards the Valgaudmar-Aiguille de Morges area (Section 2.4), controlling the geometry of the eastern branch of the Muzelle cover.

At Le Desert and the Col de la Vauze (Figure 2.3.9), this fault dips steeply to the E and the cover sediments adjacent to this fault are deformed by a penetrative

upright Alpine cleavage which trends sub-parallel to the basement fault and also dips steeply toward the E. A pronounced down-dip mineral stretching lineation is present along the cleavage plane. The La Muzelle anticline documented in the previous section is also continuous through the central Pelvoux area and a normal basement-Trias-Lias sequence forms the eastern boundary of the eastern cover branch. This corresponds to the steeply dipping forelimb of the La Muzelle anticline.

At the Col de la Vauze, the Le Desert basement fault branches in two: the eastern branch which continues its SE trend into the Aiguille de Morges area and a western branch passes into the Vaulgaudmar-Severaisse valley to the SW (Figure 2.3.9). Shear sense indicators associated with the western branch are consistent with a dextral Alpine movement.

S of the Col de la Vauze, erosion breaches the La Muzelle anticline producing a semi-window which reveals the foot-wall Triassic and Liassic cut-offs (Figure 2.3.9) of the La Muzelle thrust. The La Muzelle thrust plane is a gently dipping planar feature which defines a semi-circular trace around the semi-window. Where exposed along a profile section, the La Muzelle thrust dips gently toward the SW, implying folding or upward doming by deeper structures. Stretching data recorded from the fault plane and its immediate foot-wall are consistent with an overthrust direction towards the WSW.

Also S of the Col de la Vauze, adjacent to the steep basement fault in the W, lower Liassic sediments are deformed by tight upright chevron-type folds which plunge 35-40° towards the SE, reflecting the surface expression of a deeper hidden structure. The fold axial planes and cleavage are upright and trend sub-parallel to the basement fault scarp, suggesting the latter exerts an important control over the prevalent structural style. Toward the E these folds become more open, and are progressively modified by a SW directed simple shear. Upright axial

planes at lower structural levels, become sheared over towards the WSW at higher structural levels; a response to the emplacement of the Muzelle basement sheet (Figure 2.3.13). The folding of the Liassic cover sediments also implies local decoupling of the cover from its basement, which presumably occurred in the foot-wall to the La Muzelle thrust during the early stages of movement.

The continuation of the La Muzelle thrust into the overlying cover is impossible to trace. This may be due to the intense strains accommodated adjacent to the pre-existing basement fault, or may simply reflect the poor nature of the exposure.

2.3.6 Olan fault:

The Olan fault corresponds to a major thrust fault juxtaposing basement against basement for the greater part of its length, making it difficult to follow through the "Haute Montagne" of the central Pelvoux. To the N it may branch with, or be the same fault as, the La Meije thrust in the northern Pelvoux.

In the southern central Pelvoux, the Olan fault is exposed to the W of the Olan summit (Figure 2.3.15), where it emplaces a hanging-wall of amphibolites over a foot-wall of granites and migmatites.

The as yet unpublished BRGM 1:50 000 St. Christophe geological map shows the Olan fault to be continuous with the Rif du Sap fault, a major basement-cover thrust in the Aiguille de Morges thrust complex (Section 2.4). On the N side of the Severaisse valley the Rif du Sap fault emplaces basement over a foot-wall of poorly exposed Triassic dolomites and basement. The bedding in the dolomites parallels the fault plane, and the thrust plane dips 66 NE, striking 140. A mineral stretching lineation plunges towards 050-060, the latter implying a SW direction of overthrusting. However Mugnier (1984, unpublished data), made systematic analyses of mineral slickensides along the

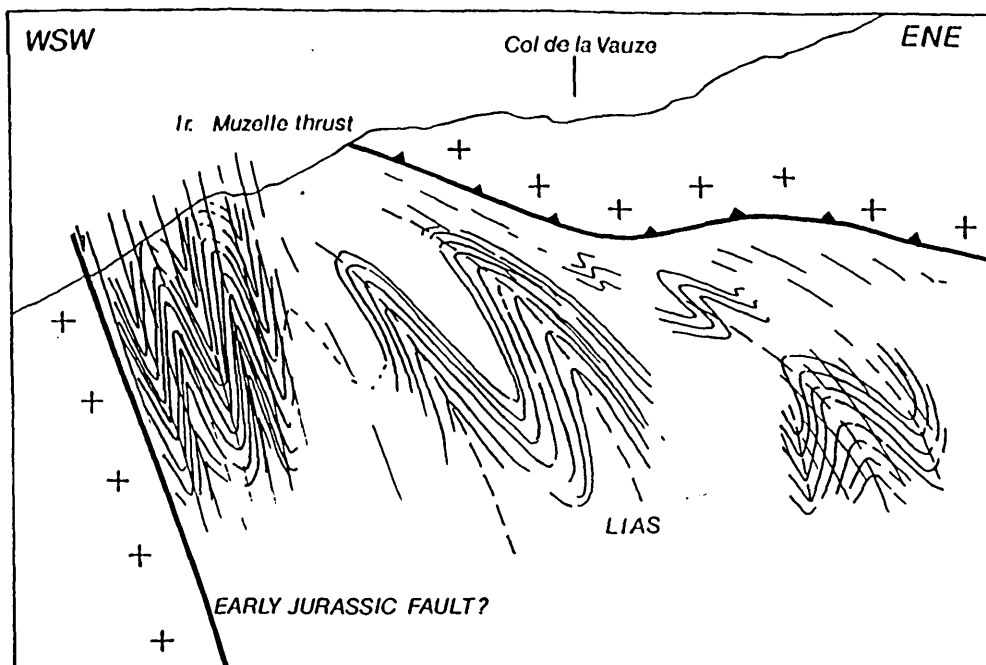


Figure 2.3.13: Sketch section across the Col de la Vauze showing the buttressing effect of the Le Desert fault, and the Lower Muzelle thrust.

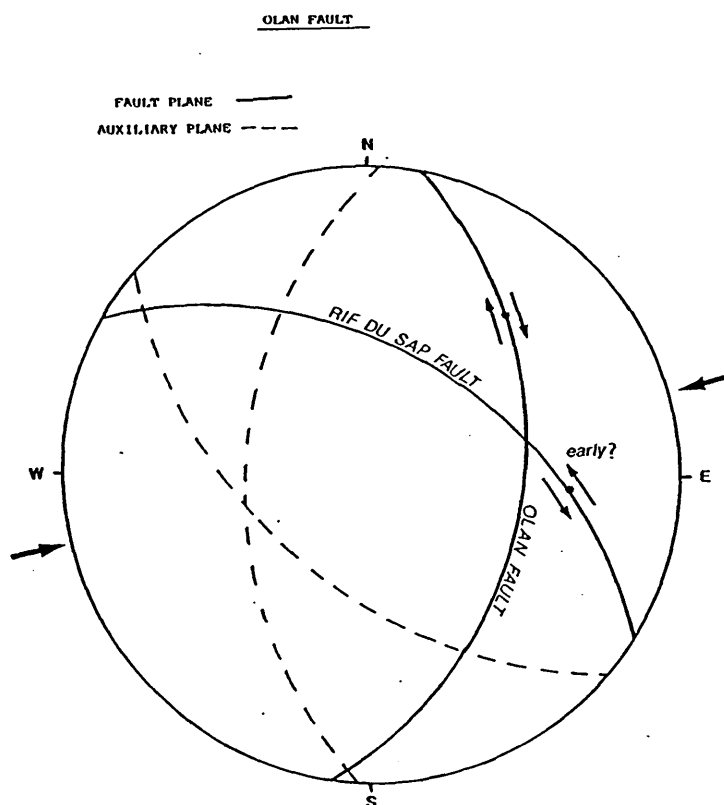


Figure 2.3.14: Equal angle (Wulf net) plot of the slip-lineation data from the Olan and Rif du Sap faults (Mugnier, unpublished data). See text for details.

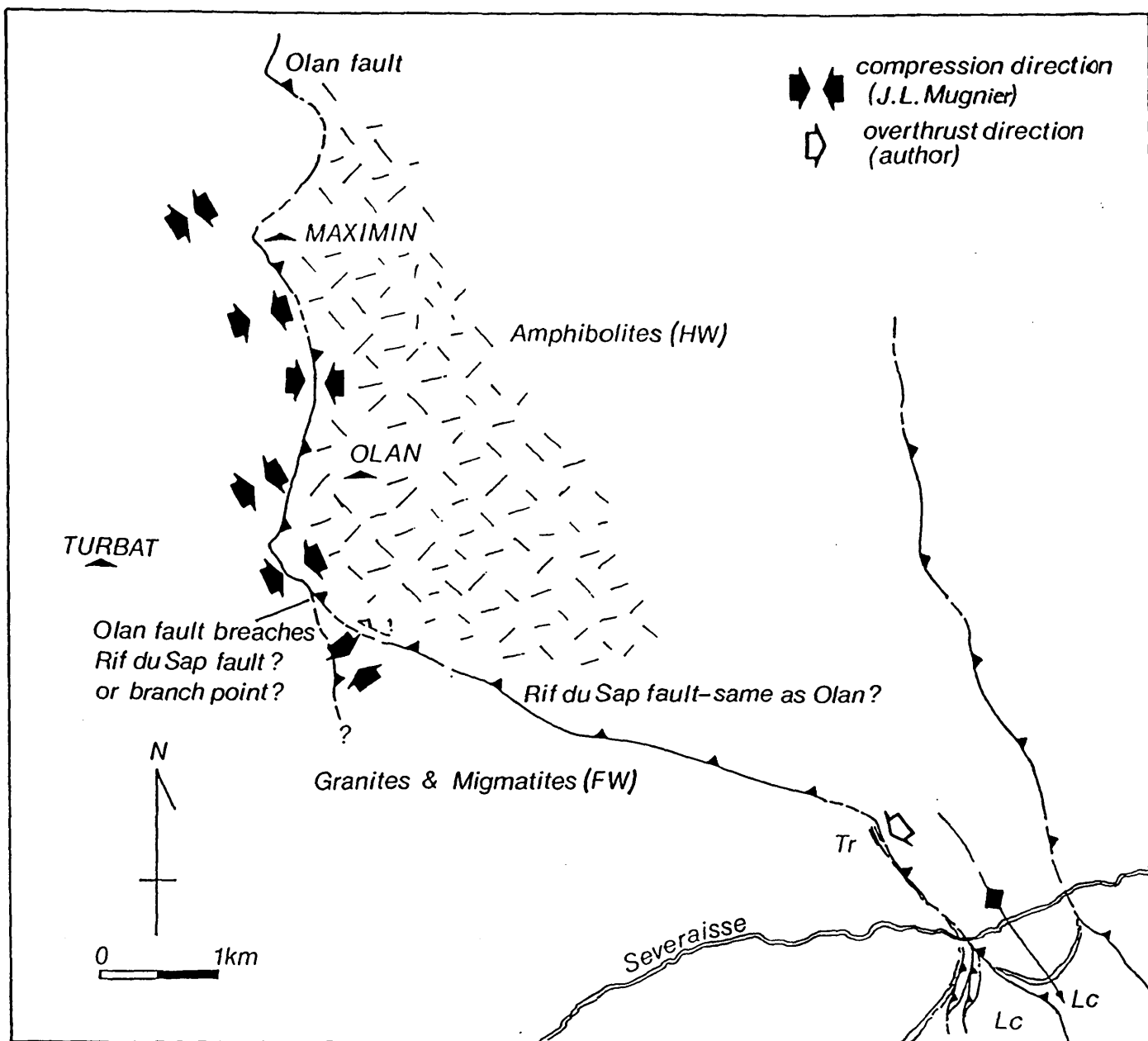


Figure 2.3.15: Sketch map of the Olan and Rif du Sap faults showing the compression directions obtained by Mugnier (unpublished data) and the overthrust direction obtained by the author.

fault planes and using the method of Angelier (1983), and Angelier & Goguel (1978), recorded different directions of compression for the Olan and Rif du Sap faults. For the Rif du Sap fault he recorded an early (?) more ductile mineral stretching lineation reconcilable with a NW direction of overthrusting, whereas for the Olan fault, later (?) more brittle fibre slickensides consistently give a WSW-ENE direction of compression (Figures 2.3.14 & 2.3.15). From this he suggested that the Olan and Rif du Sap faults are of two different ages, the Olan fault being later, carrying the Rif du Sap fault passively in its hanging-wall. Two major problems arise from this interpretation:

1. The southward continuation of the Olan thrust fault - no such fault of appreciable size and displacement can be traced to the S.
2. The mineral stretching lineation consistent with a SW direction of overthrusting recorded from the Rif du Sap fault along the N side of the Severaisse valley.

It is suggested here that the Rif du Sap fault and the Olan fault are of the same age and are possibly the same fault; if not the Olan-Rif du Sap intersection constitutes a thrust "branch" and not a thrust "breach" relationship. However the two contrasting directions of mineral stretching direction may suggest that these faults have suffered two distinctive phases (NW and WSW) of thrust movement.

2.3.7 Thrust transport directions - a summary:

The directions and relative ages of thrusting were determined using a combination of geometrical relationships, structural style, and a study of the kinematics using slip indicators exposed on the fault surfaces.

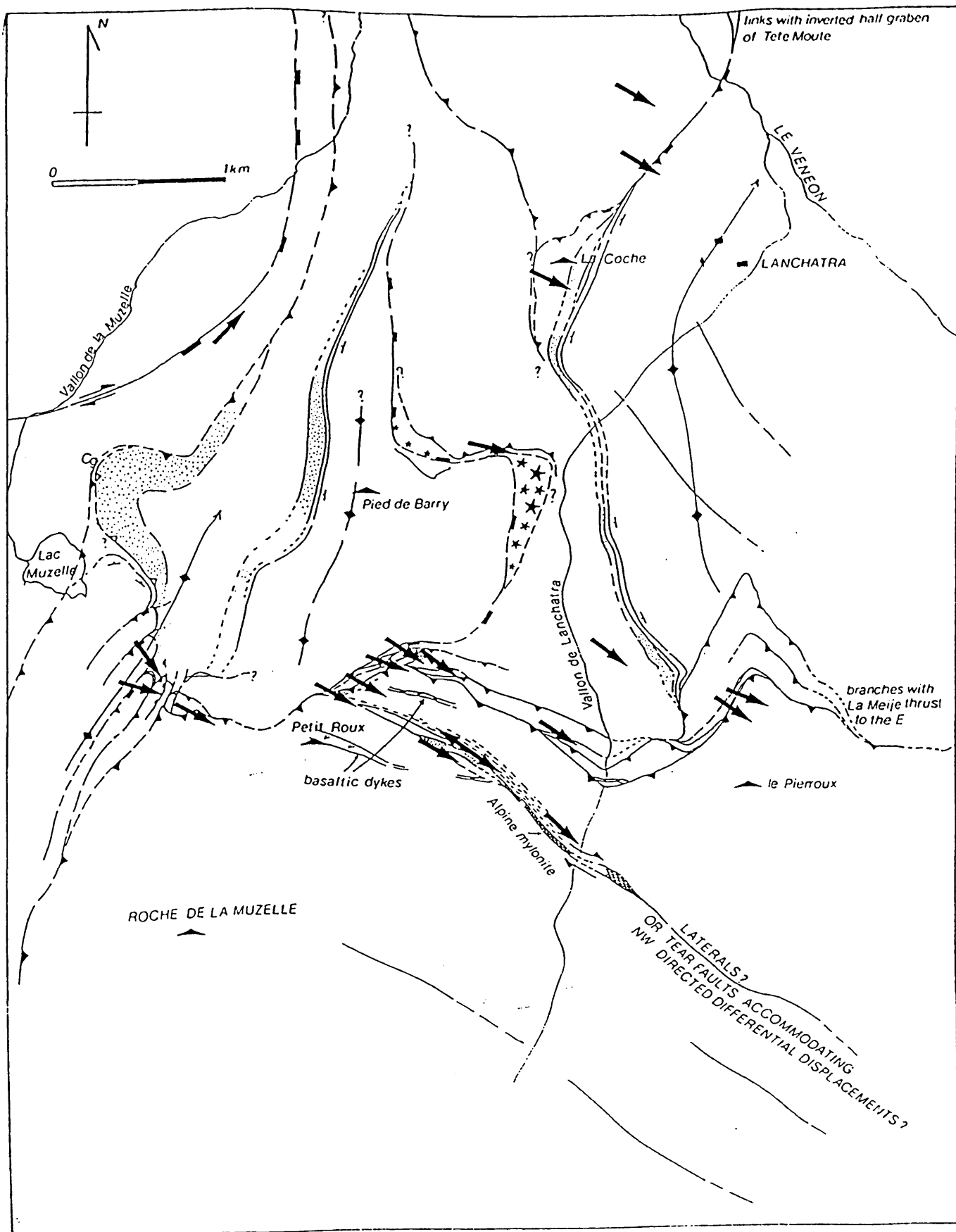


Figure 2.3.16: Slip-lineation data for the Lanchatra area (location and legend are the same as for Fig. 2.3.1). See text for details.

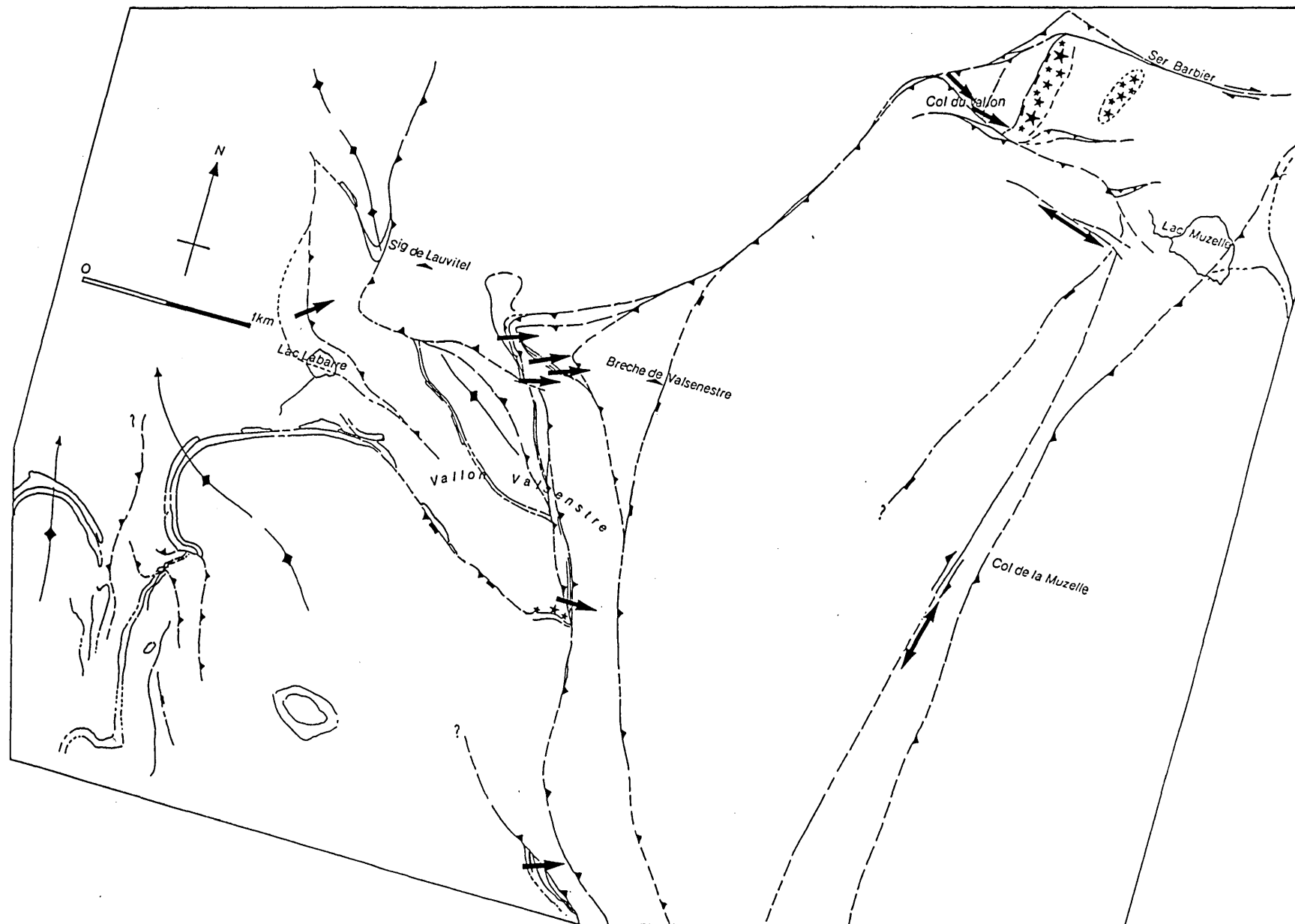


Figure 2.3.17: Slip-lineation data for the Vallon Valsenestre area (location and legend same as for Fig. 2.3.5).

1. Vallon Lanchatra:

The mean movement direction in the northern part of the area was towards the WNW (Figures 2.3.16, & 2.3.18 a & 2.3.19). This proposed movement direction is compatible with that determined from lateral ramp orientations and NW-SE trending tear faults.

It has been previously (Section 2.3.1) argued that the sequence of thrusting does not conform to a simple foreland-directed piggyback system, but rather records "pulses" of shortening of different ages. An early phase, appears to reactivate the pre-existing Mesozoic extensional fault system and these structures are post-dated by structurally higher and later basement thrusts, whose development appears to be uninfluenced by pre-existing structure. The general tectonic movement direction for both these events was evidently towards the WNW.

2. Vallon Valsenestre:

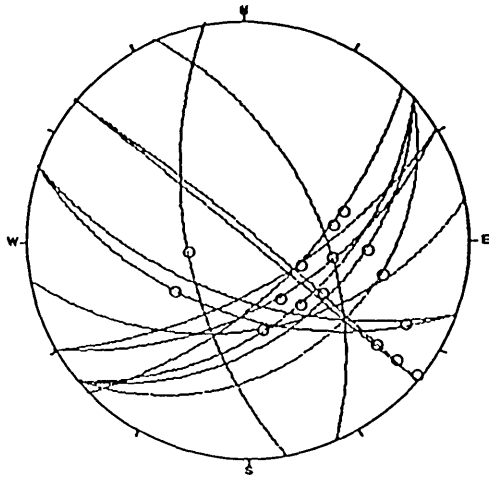
In the Vallon Valsenestre area, a WSW direction of tectonic transport appears to dominate (Figures 2.3.17). A high level SW-directed basement thrust sequence appears to post-date earlier inversion structures, which also have a WSW direction of tectonic transport. At the Col du Vallon, the basement thrusts also post-date WNW directed inversion structures. Some pre-existing NE-SW oriented faults appear to facilitate the squeezing of movements out towards the SW (i.e. the Col de la Muzelle and Ser Barbier faults).

3. Pic de Valsenestre and southern areas:

A WSW direction of tectonic transport dominates in this area (Figure 2.3.18 c). Some of the structures, such as the La Muzelle anticline, which is overthrust towards the WSW, are laterally continuous into the northern area dominated by WNW-directed thrust movements.

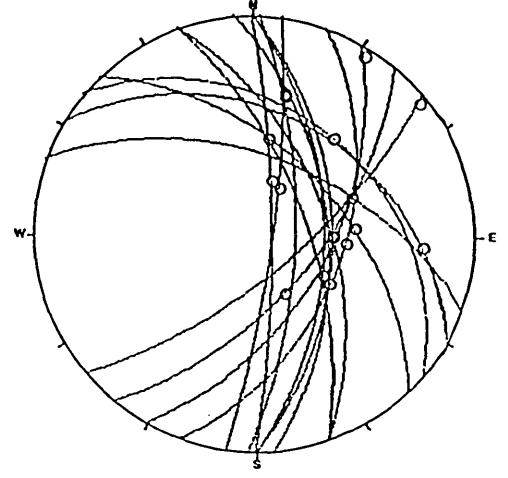
A discussion concerning the significance of these

a. LANCHATRA



Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - Fault plane	16	50.00
9 - slickenside	16	50.00
Total number of data	32	

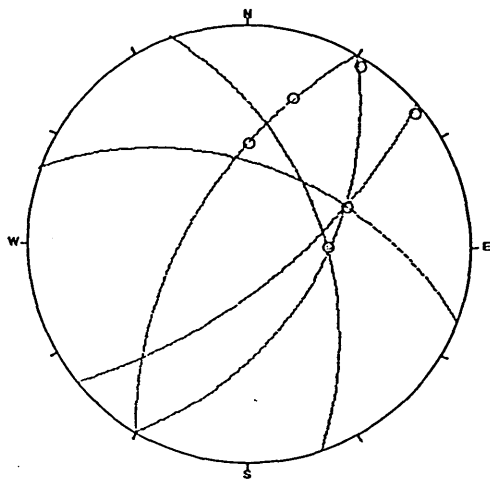
b. VALLON VALSENESTRE



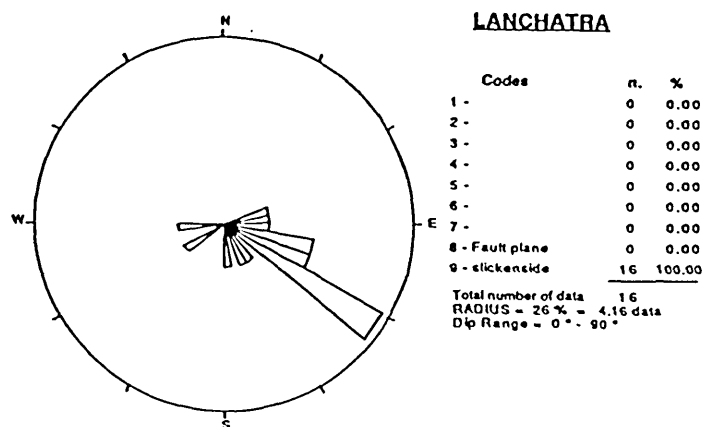
Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - SLICKENSIDES	17	50.00
9 - FAULT PLANE	17	50.00
Total number of data	34	

Figure 2.3.18: Equal angle (Schmidt net) plots of slip-lineation and fault plane data for the Lanchatra (a), Vallon Valsenestre (b), and Pic de Valsenestre (c) areas. See text for details.

c. PIC DE VALSENESTRE



Codes	n°	%
1 -	0	0.00
2 -	0	0.00
3 -	0	0.00
4 -	0	0.00
5 -	0	0.00
6 -	0	0.00
7 -	0	0.00
8 - SLICKENSIDES	6	50.00
9 - FAULT PLANE	6	50.00
Total number of data	12	



LANCHATRA

Range	n.	%	Range	n.
0 - 10	0	0.00	180 - 190	0
10 - 20	0	0.00	190 - 200	0
20 - 30	0	0.00	200 - 210	0
30 - 40	0	0.00	210 - 220	0
40 - 50	0	0.00	220 - 230	0
50 - 60	0	0.00	230 - 240	1
60 - 70	0	0.00	240 - 250	0
70 - 80	1	6.25	250 - 260	0
80 - 90	1	6.25	260 - 270	1
90 - 100	1	6.25	270 - 280	0
100 - 110	2	12.50	280 - 290	0
110 - 120	2	12.50	290 - 300	0
120 - 130	4	25.00	300 - 310	0
130 - 140	0	0.00	310 - 320	0
140 - 150	1	6.25	320 - 330	0
150 - 160	1	6.25	330 - 340	0
160 - 170	0	0.00	340 - 350	0
170 - 180	1	6.25	350 - 360	0

Total number of data 16

Figure 2.3.19: Rose diagram of the slip-lination plunge directions for the Lanchatraz area only. See text for details.

thrust transport directions and the timing of inversion and thrusting will be given in Chapter 5.

2.4 Aiguille de Morges area

The Aiguille de Morges area comprises a quadrangle of Triassic and Jurassic cover sediments surrounded by basement (Figure 2.4.1). It records where major basement thrusts have climbed into the overlying cover sequence which is substantially shortened and telescoped as a result.

2.4.1 Southern boundary:

The structural map pattern is that of a complex linked thrust system where thrusts splay from a common floor thrust and repeat the cover stratigraphy extensively. This floor thrust is best observed along the southern boundary of the cover thrust complex where it parallels the basement-cover contact (basement-Triassic unconformity). This mapped southern section is unique in that the basement-cover contact and overlying cover thrusts have been re-oriented and tilted towards the NNE, presumably by deeper thrust structures in the Chaillol-Clapouse basement, and the present map geometry presents a moderately inclined section. As no mapping was undertaken S of the Aiguille de Morges area, the precise geometry of these deeper level basement structures, responsible for the upward doming of the Chaillol-Clapouse basement and tilting of higher level thrusts remains undisclosed.

Along this southern boundary the cover sequence overlying the floor thrust is characterized by minor omissions of Triassic stratigraphy, such that Triassic basalts or Liassic limestones and shales rest directly on the basement. The thin Triassic carbonate platform sequence documented from other areas of the Pelvoux and adjacent massifs, is largely replaced by cargneules occurring as isolated tectonic lenses; presumably strongly tectonised by movement along the basal slip-surface. Given the character of the Triassic basic volcanism, where the rift basalts were not extruded as a continuous extensive sheet onto a

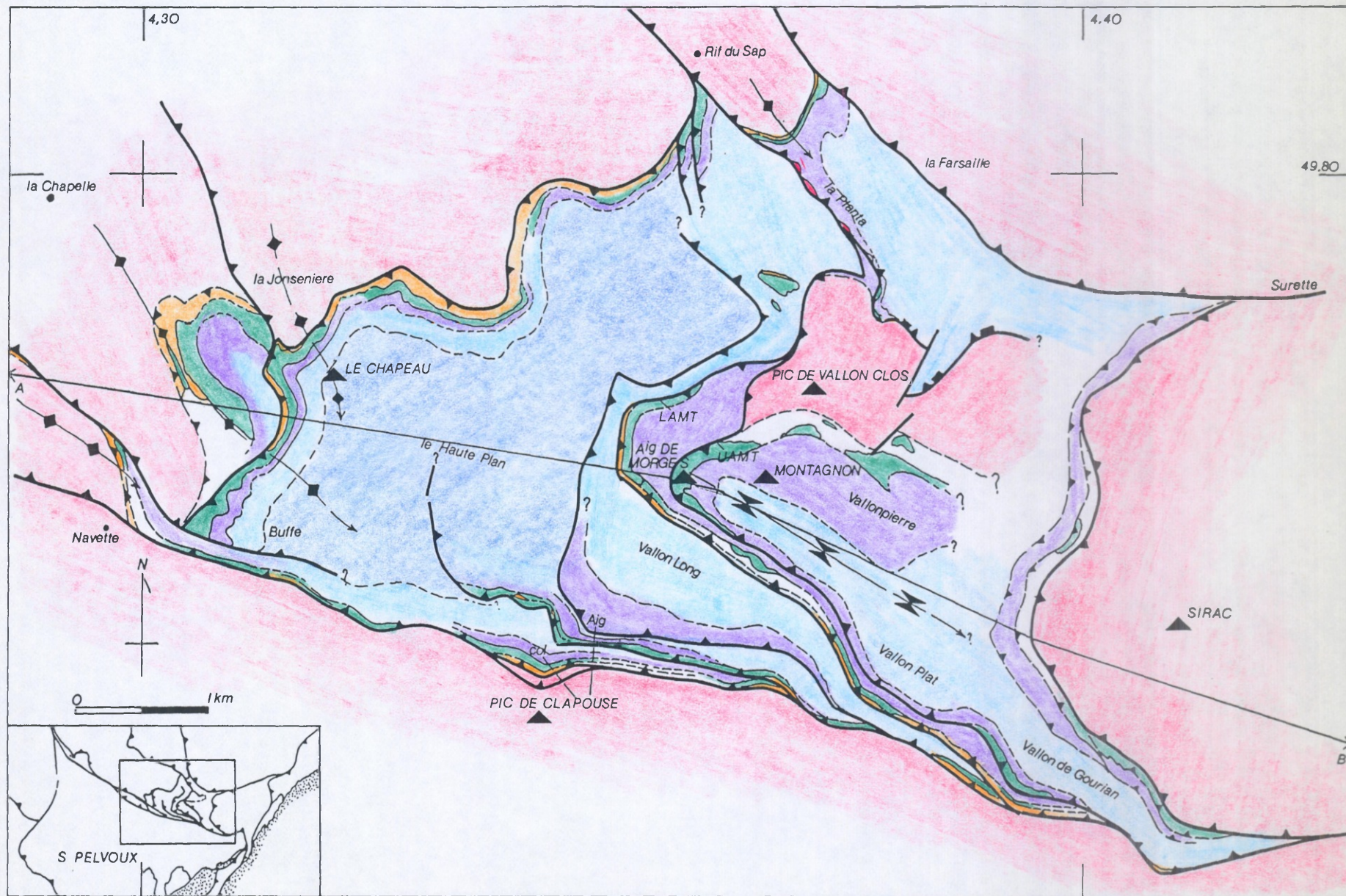


Figure 2.4.1: Structural map of the Aiguille de Morges area (location and legend indicated on Figs. 1.7 & 1.8). A-B denotes the section line shown in Figure 2.4.2, LAMT-Lower Aiguille de Morges thrust, UAMT-Upper Aiguille de Morges thrust. Based on mapping by the author.

shallow carbonate platform and the tectonised nature of the Triassic cargneules, this omission of stratigraphy is easily accounted for.

In the Vallons de Gourian, Plat and Long (Figure 2.4.1), a number of cover thrusts splay from the main floor thrust, and follow a ramp-flat staircase trajectory. The principal flat detachment, or floor thrust, exploits the Triassic cargneule horizon such that the competent basalt horizons which are unusually thick (upto 150 m), invariably appear in the hanging-walls of the long flat cover detachments, and cargneule is associated with each cover thrust fault. These cover thrusts consistently ramp from the floor thrust, through the Triassic and Lower Liassic sequence (informally known as the Lias calcaire), and flatten along the Lower-Upper Liassic boundary or within the Upper Liassic shales (informally known as the Lias schisteux). Most of these thrusts are interpreted to ultimately ramp into the Middle Jurassic shale sequence, as observed at the Col de Clapouse, where a thrust carrying Triassic and Liassic sediments in its hanging-wall ramps into the Middle Jurassic shales of le Haute Plan, where the resulting deformation is both penetrative and distributed, and the trace of the thrust fault impossible to follow. It is conceivable that the roof thrust to this lower level thrust imbrication is in the Middle Jurassic shales or Oxfordian black shales (Terres Noires). This thrust repetition of the Triassic and Liassic stratigraphy, well displayed on the Aiguille de Morges peak, is shown in Plate 2.4.1.

The cover thrust sheets are characterized by a flat-lying penetrative cleavage and both cleavage and long flat thrust detachments dip to the NE, reflecting a superimposed NNE tilting. The present day orientation and geometry of these cover thrust sheets suggests that the sequence of thrust development was piggyback and directed towards the foreland.

Generally, NE-SW foot-wall and hanging-wall cut-off

trends could have been derived from either frontal ramps in a NW directed thrust system, lateral ramps in a SW system, or oblique ramps to some intermediate direction of thrust transport.

2.4.2 Sirac thrust:

In the Vallon Vallette E of the Vallon Gouirian, the basement-cover detachment from which the cover thrusts splay, branches with a major basement overthrust - the Sirac thrust (Vernet, 1961). The Sirac basement sheet is interpreted to have travelled along the basement-Triassic unconformity, a distance that is equal to the total displacement associated with the cover thrusting recorded in its foot-wall. Both the Sirac thrust sheet and basement-Triassic unconformity have been tilted to the NNE, suggesting a piggyback sequence of thrust development. Two thrust slivers of Triassic basalts and Lower Liassic limestones in the immediate foot-wall to the Sirac thrust, are interpreted as far travelled tectonic horses, plucked from the foot-wall during the initial emplacement and transported during subsequent horizontal movement of the Sirac basement sheet. Towards the ESE the hanging-wall cover sequence of the Sirac basement crops out and the thickness of the Sirac basement sheet diminishes southwards. Farther S at the Col de Meollion, an imbrication of the Jurassic cover in the foot-wall to the Sirac thrust post-dates the deposition of the Eocene Gres du Champsaur flysch. However, some pre-Eocene uplift of the Sirac basement must have occurred, because the Eocene flysch rests unconformably upon basement. The age and timing of basement thrusting will be assessed in Section 2.4.8. In the N, the trace of the Sirac basement thrust follows the Surette valley (Figure 2.4.1) to link with a zone of basement-cover imbrication (Vernet, 1961), and the Alfroide basement-cover imbricates in the eastern Pelvoux (Gidon, 1979).

2.4.3 Vallonpierre synform and Pic de Vallon Clos thrust:

The Vallonpierre synform trends N120 and plunges gently to the ESE (Figure 2.4.1), this plunge being the result of an eastward back-steepening induced by thrust movements on the Lower and Upper Aiguille de Morges thrusts. The synform is asymmetric in cross-section and faces toward the SSW, the S facing limb being steeper than the N facing one. On the W slope of the Pic de Vallon Clos, the Pic de Vallon Clos thrust is a gently dipping planar feature emplacing a hanging-wall of basement over a foot-wall of Jurassic sediments. The thrust continues into the overlying sedimentary cover as the Upper Aiguille de Morges thrust (UAMT), a hanging-wall flat which parallels the base of the Triassic basalts emplacing them over a foot-wall of Lower Liassic limestones and shales. The Vallonpierre syncline is interpreted as the hanging-wall adaptation to the Pic de Vallon Clos basement thrust such that the steeper northern limb corresponds to the forelimb of a hanging-wall ramp anticline carried over a foot-wall flat, or more gently dipping thrust ramp. The trend of the Vallonpierre syncline reflects the WNW-ESE trend of this hanging-wall ramp in the subsurface, i.e. the hanging-wall basement cut-off line. In the subsurface, the Pic de Vallon Clos thrust is interpreted to branch with and transfer its displacement onto, the Sirac thrust.

The trend of this hanging-wall thrust ramp may be interpreted as a frontal ramp in a SW transport system, a lateral ramp in a NW transport system, or an oblique thrust ramp to some intermediate thrust transport direction.

The Lower Aiguille de Morges thrust (LAMT) (Figure 2.4.1) also branches with the Pic de Vallon Clos thrust at le Chateau, to the immediate S of the Pic de Vallon Clos. Here the Triassic basalts in the hanging-wall of the LAMT cut out against the fault. For a SW direction of thrusting, this thrust cuts down the hanging-wall stratigraphy in the direction of thrust transport. However, a W or NW direction of thrusting is compatible with this cut-off trend.

2.4.4 Rif du Sap - Olan fault:

The Pic du Vallon Clos thrust can be traced to the N as a gently dipping planar feature emplacing basement over a foot-wall of Upper Liassic shales, and a Lower and Upper Liassic sequence over Upper Liassic shales along the Planta ridge (Figure 2.4.1). Basement reappears in the hanging-wall of this thrust at the Rif du Sap, and it is clear that the Pic de Vallon Clos thrust is the same fault as the Rif du Sap - Olan thrust (see Section 2.3.6).

Where the Rif du Sap - Olan thrust climbs from basement into the overlying cover, the hanging-wall basement and its cover are folded into a tight anticline which plunges steeply to the SE. The Rif du Sap - Olan fault at the level of the Rif du Sap village corresponds to a steeply dipping feature (60° - 70°), becoming less steep with altitude (45°). This downward steepening of dip possibly reflects back-steepening by deeper level thrusts, as does the steep SE plunge of the Rif du Sap anticline. Removing the effects of later rotation, the Rif du Sap fold would have probably have faced towards the SW or W. From this limited information it is difficult to commend a definitive direction of thrust transport.

Along the Planta ridge the trace of the Olan - Rif du Sap fault is delineated by fault-bound tectonic lenses of basement completely enveloped by the surrounding cover sediments (Plate 2.4.2). Associated with the Lower Liassic sediments in the hanging-wall are more irregular shaped blocks of crystalline basement, interpreted to be syn-sedimentary in origin (sedimentary olistoliths).

The NE side of the Pic de Vallon Clos basement is delineated by a steeply dipping normal fault contact which juxtaposes Upper Liassic shales against basement. The deformation regime in the cover sediments adjacent to the basement fault is locally upright and penetrative, suggesting that such a fault may have existed prior to the main compressional deformation. This basement fault is cut



Plate 2.4.2: Shows tectonic horses of basement completely enveloped in Liassic cover sediments on the La Planta ridge. Ruck-sack in the foreground gives the scale. See text for locality and details.



Plate 2.5.1: Tight chevron-type folds deforming Lower Liassic sediments in the immediate foot-wall of the Chaillol basement thrust. See text for locality and details. Looking SSE.

out against the Pic de Vallon Clos basement thrust, and shear fabrics adjacent to the fault plane suggest also that it has suffered some reverse slip. Using the limited evidence cited above, this basement normal fault is interpreted to be a Jurassic age fault, and the small basement lenses are interpreted to have been derived from the pre-existing basement fault scarp during its attempted reactivation. Equally, the Pic de Vallon Clos thrust may be interpreted as a foot-wall shortcut structure (see Chapter 3), although the large amount of displacement associated with thrusts in its foot-wall supports an interpretation where in the subcrop, the Pic de Vallon Clos thrust slices through the pre-existing normal fault.

Farther to the E another basement thrust, the Farsaille thrust (Figure 2.4.1), overrides Upper Liassic shales in its foot-wall. The present-day steep dip of this fault suggests a piggyback thrust sequence interpretation.

2.4.5 Le Chapeau - Navette:

In the W of the Aiguille de Morges area, a set of complex thrust relations exist. On the Crete de la Cote Grasse, which forms the W face of Le Chapeau (Figure 2.4.1), an imbricate thrust splay from the main floor thrust resulting in a simple repetition of the Triassic and Liassic cover sequence. The exposed thrust geometry corresponds to a hanging-wall flat thrust over a foot-wall flat. In the N at la Jonseniere, a minor basement thrust is interpreted to branch with the main floor thrust, reactivating the cover thrust ramp such that the la Jonseniere basement overthrusts a foot-wall of Triassic basalts and dolomites. To the S, the branch line of the Crete de la Cote Grasse thrust with the floor thrust at the base of the Triassic sequence also crops out, however, both floor thrust and higher level cover imbricate are folded and thrust by a structurally lower fault. This thrust fault climbs out of the basement to the N of Navette at la Lauziere and breaches the Aiguille de Morges floor thrust and the Crete de la Cote Grasse imbricate as it continues

to ramp into the overlying sedimentary cover. The la Lauziere thrust is itself back-steepened by another basement thrust which flattens along the basement-Triassic unconformity, to the immediate N of Navette. To the E of Navette the floor thrust to the cover thrust system is itself re-oriented by deeper structures, dipping steeply to the NE. The foot-wall bedding cut-off trends associated with the Crete de la Cote Grasse thrust, where it ramps from the floor thrust, can be interpreted in the context of either a SW or NW transport system, or any intermediate direction.

The re-orientation of higher level structures by movement on lower level thrust sheets explains the steeply eastward-plunging Buffe folds (60° towards 100°) which affect the Triassic and Liassic hanging-wall cover sequence of the Crete de la Cote Grasse thrust. The initial orientation of the Buffe folds prior to the movement on deeper level structures is uncertain. They may have possessed an initial NW-SE trend and faced towards the SW. Folds with such an orientation could be interpreted in the context of a thrust transport direction towards anywhere between NW and SW.

2.4.6 E-W cross-section:

An E-W cross-section constructed approximately parallel to the thrust movement direction (see Section 2.4.7), is shown in Figure 2.4.2. This section illustrates the importance of cover shortening in the foot-wall to the Sirac thrust.

The overall geometry is interpreted as a thrust duplex, with the floor thrust exploiting the basement-cover interface and the roof thrust, the Mid-Jurassic shales or Oxfordian black shales (Terres Noires). The intervening thrust faults which transfer displacement between the floor and roof thrusts follow a complex ramp-flat staircase trajectory. The breaching and consequent folding of the floor thrust by basement thrusts is well documented at

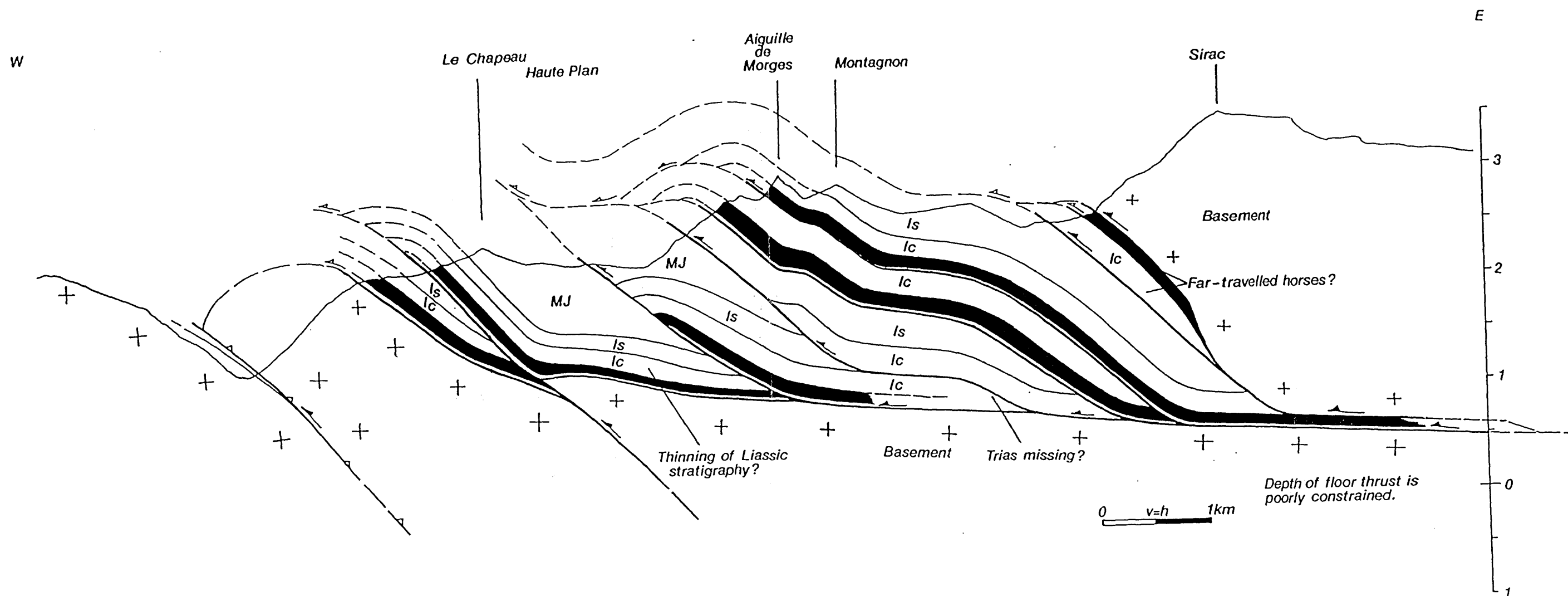


Figure 2.4.2: Structural cross-section across the Aiguille de Morges area (location shown as A-B on Fig. 2.4.1). Crosses-crystalline basement, black-Triassic basalts, lc-Lias calcaire, ls-Lias schisteux, MJ-Middle Jurassic shales.

Navette.

The Jurassic stratigraphy is poorly constrained and therefore simplified. The complexity is largely due to the penetrative character of the cover deformation. Although stratigraphic facies changes indicative of syn-sedimentary faulting were not recognised, more detailed sedimentological and stratigraphic studies may provide additional information concerning the geometry of the pre-contractional template.

2.4.7 Thrust transport directions:

It is difficult to conclude a definitive direction of thrust transport for the Aiguille de Morges area, largely because of conflicting evidence from adjacent areas into which the Aiguille de Morges thrust structures are continuous. The best estimate of thrust transport direction is arrived at by analyzing the geometry of thrusting (cut-off trends, lateral ramp trends etc.) and kinematic indicator trends (slickenside lineations, aligned fibres, and mineral stretching directions).

In the Aiguille de Morges area, stratigraphic cut-off trends, whether measured directly in the field or obtained from maps, serve only to discriminate against certain transport directions, rather than reveal the true direction. This is because of the 180° transport range possible from a single cut-off trend, so that it is necessary to identify both frontal and lateral cut-off trends before the true transport direction can be identified.

A critical cut-off trend occurs in the hanging-wall of the Lower Aiguille de Morges thrust, beneath the Pic de Vallon Clos (Figure 2.4.4). For a SW transport direction interpretation the LAMT thrust would have to cut down the hanging-wall stratigraphy in the direction of thrusting. This is the only cut-off direction that cannot be reconciled with a SW or WSW thrust transport direction.

Similarly, the cut-off trends associated with the basement hanging-wall anticlines at Navette and Rif du Sap, are difficult to reconcile with a NW direction of thrust transport, and fit more easily into a WSW directed thrust system, although re-orientation by deeper level structures renders the use of such folds and their hanging-wall cut-offs as indicators of thrust transport direction as ambiguous.

An assessment of the kinematic indicators reveals a consistent W-WNW trend, although minor trend variations are evident (Figure 2.4.3)

From the above, a SW thrust transport direction interpretation can be excluded at least for the higher level thrust sheets, and a NW direction of thrusting can be excluded for the Navette and Rif du Sap basement anticlines. A W to WNW thrust transport direction is proposed for the structurally higher thrust sheets, although for the lower level basement thrusts this direction is not so apparent. If a WSW direction of thrust transport is inferred for these lower level basement sheets, this would imply a progressive change in transport direction towards the foreland, and necessitate a minor anti-clockwise rotation of the higher level thrust sheets. Such a WSW transport interpretation is compatible with the kinematic indicators and the geometry of the basement thrust structures in the Navette area and the continuation of the Navette basement structures into the Villar Loubiere area where they branch with the lower Muzelle thrust. A W to WNW transport direction interpretation for the Rif du Sap - Olan fault is also in conflict with the kinematics suggested for the Olan fault (Section 2.3), where a component of WSW directed thrusting may be important. It is feasible that the kinematics of this area are quite complex and cannot be interpreted in the context of a single thrust transport direction.

Unfortunately, no obvious lateral ramp structures were identified, other than for the continuation of the

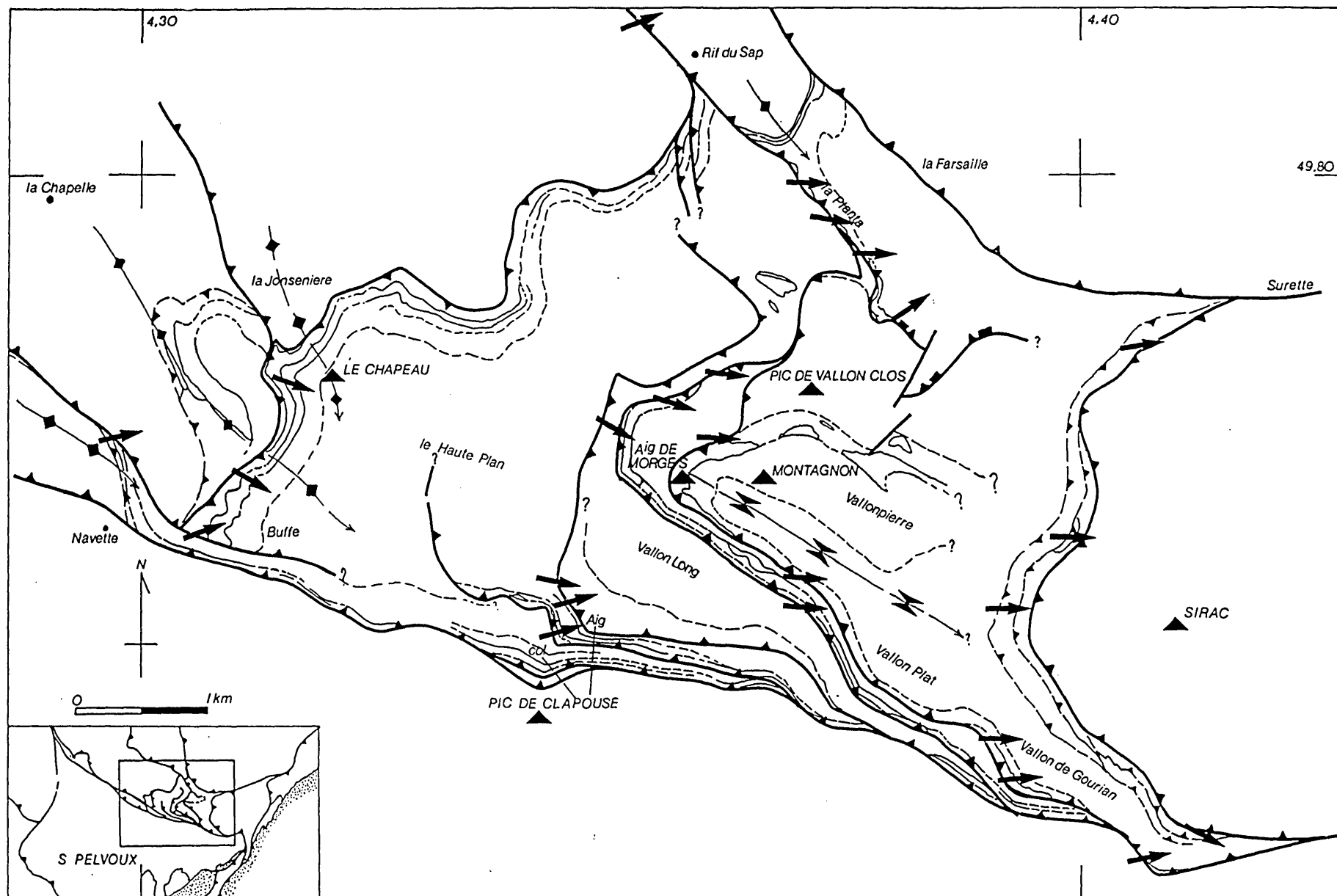


Figure 2.4.3: Slip-lineation data and mineral stretching lineation trends for the Aiguille de Morges area (location and legend as for Fig. 2.4.1). See text for details.

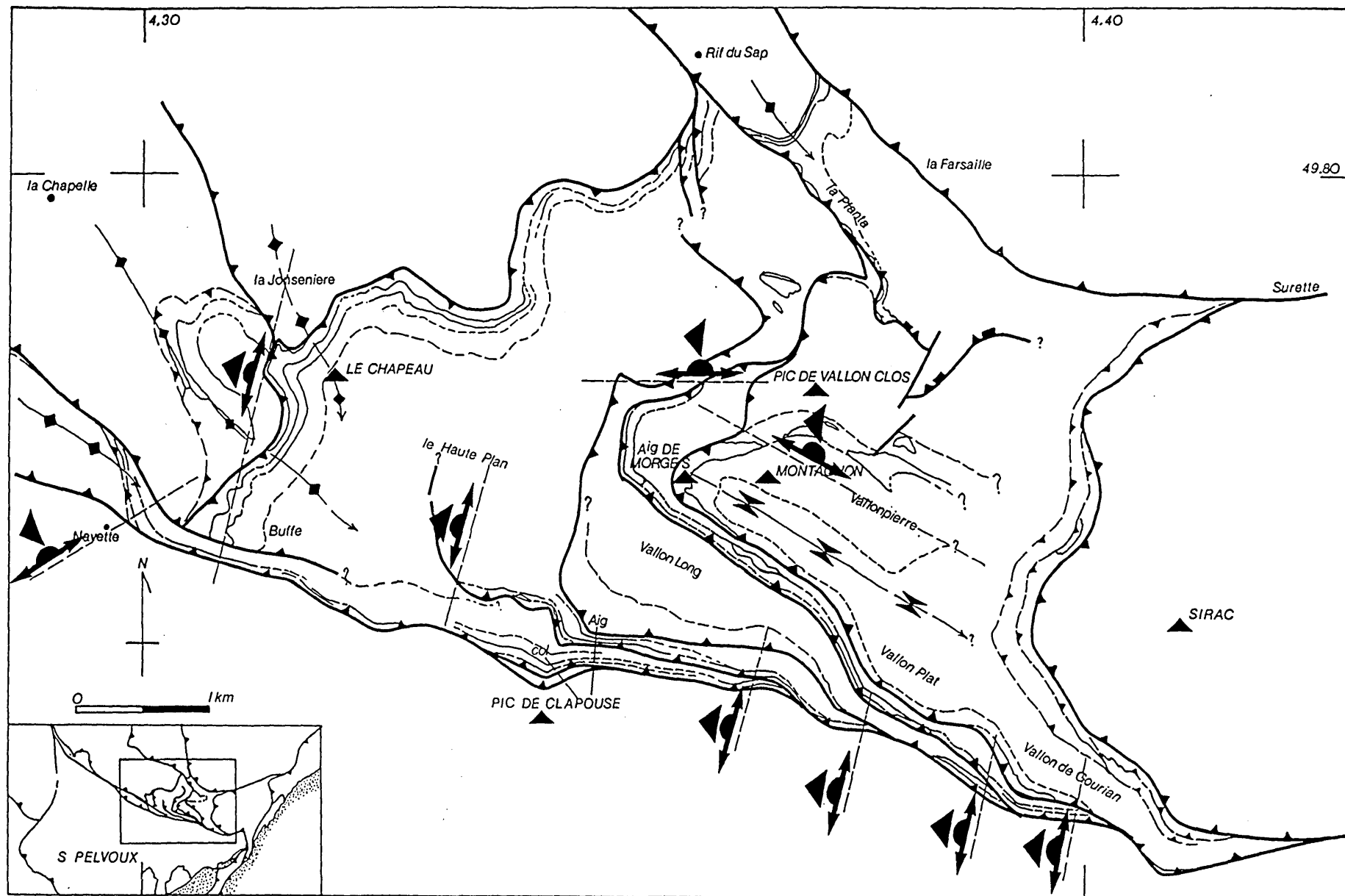


Figure 2.4.4: Fault cut-off trends for the Triassic in the Aiguille de Morges area (location and legend as for Fig. 2.4.1). Apex of triangles denotes the direction of thrust transport.

Farsaille and Sirac basement thrusts along the Surette valley, which may correspond to a lateral structure. Bijou-Duval (1975) documents a contrast in basement types along either side of this dislocation; the distinctive Sirac migmatites being juxtaposed against anatexites of the Farsaille sheet. From these relationships he inferred a substantial dextral displacement along the northern boundary of the Sirac sheet. Even if this structure is a lateral ramp to the Sirac basement sheet, its orientation may not be representative of the true transport direction, especially if it corresponds to a reactivated Jurassic structure. Care must be exercised when interpreting such structures to be indicative of the true thrust transport direction.

2.4.8 Timing and sequence of thrusting:

In the general absence of syn-orogenic sediments, the precise age of thrusting documented from the Aiguille de Morges area is difficult to assess. However, the southward continuation of these structures and their relationship with the Eocene (Priabonian age) Gres du Champsaur flysch permits an age to be suggested. In the Meollion valley to the S (Figure 2.4.5), the continuation of the Sirac thrust and the Aiguille de Morges cover thrust equivalents, breach the Eocene flysch unconformity thereby advocating a post-Eocene age interpretation. A section across the Col du Meollion (Chambers, 1987; see also BRGM 1:50 000 Orcieres sheet, 1980) shows the Eocene unconformity which directly overlies the Clapouse-Chaillol basement, to be overthrust by Jurassic shales which in turn are themselves overthrust by the Sirac basement, carrying Eocene flysch on its hanging-wall. These post-Eocene thrusts continue along the length of the Meollion valley, imbricating the Jurassic and overlying Gres du Champsaur flysch sequence. The displacements associated with the Sirac - Aiguille de Morges thrust system may ultimately roof into the vast expanse of Oxfordian black shales (Terres Noires) on which the city of Gap is centred, and which can be traced southwards beyond the Lac du Serre Poncon in the hanging-

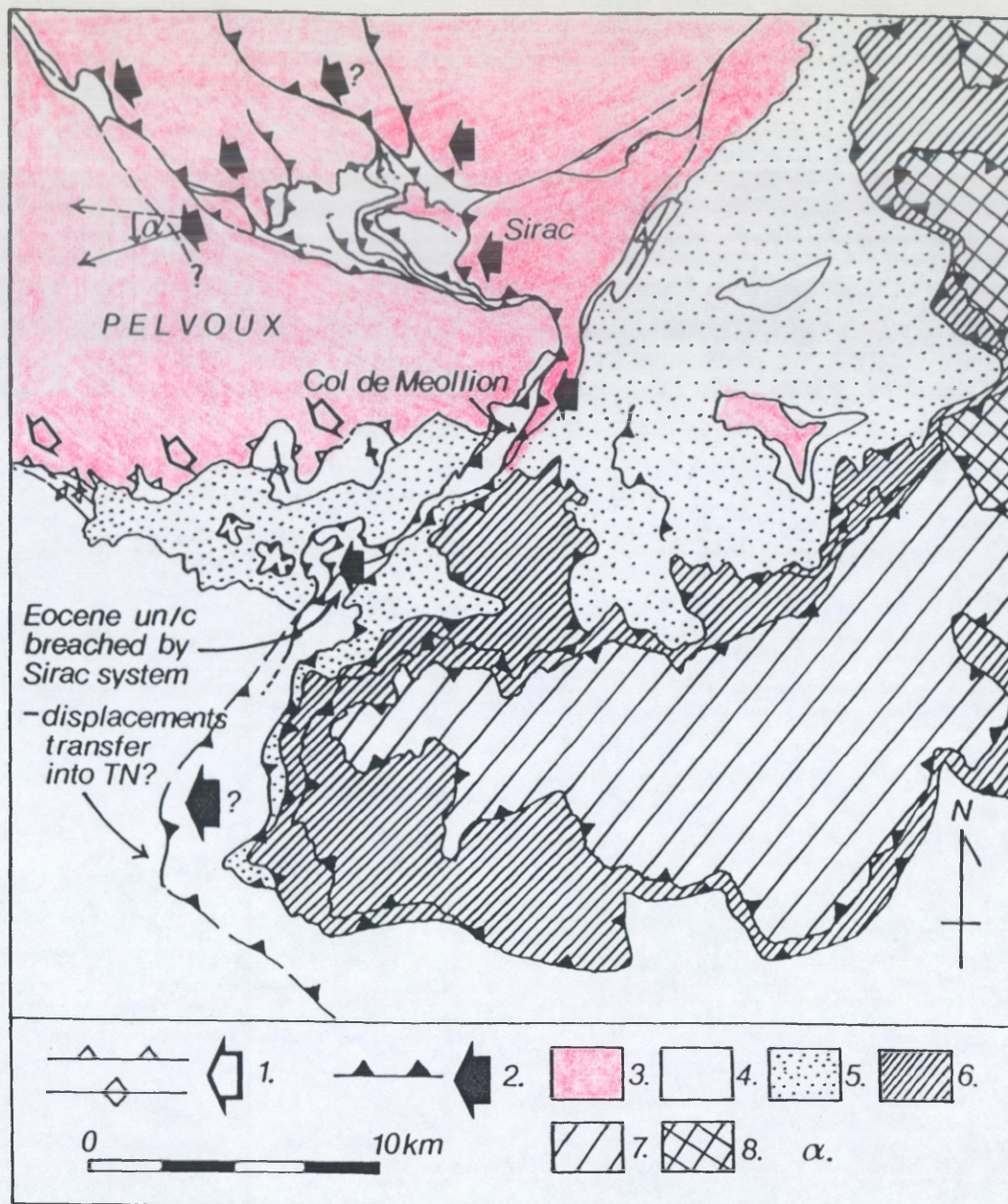


Figure 2.4.5: Simplified structural map of the southern Pelvoux showing the distribution of the different age structures (see text for details). 1. Folds and thrusts pre-dating the Nummulitique flysch (no. 5), 2. thrusts post-dating the Nummulitique flysch, arrow denotes direction of thrusting, 3. crystalline basement, 4. Jurassic cover, 5. Gres du Champsaur (Nummulitique) flysch, 6. & 7. Embrunais-Ubaye nappes, 8. Briançonnais nappes, - rotation of thrust transport direction. Based on mapping by the author and the BRGM Orcieres 1:50 000 geological map (1980).

wall of the Digne thrust system (BRGM 1: 250 000 Gap sheet, 1980). Alternatively the Sirac displacements may roof onto the Frontal Pennine thrust which emplaces the internal Alpine units over the external ones and therefore structurally overlies the Sirac thrust system.

Although a post-Eocene age interpretation is proposed for the Sirac-Aiguille de Morges thrust system, the overlap of the Eocene Gres du Champsaur flysch onto the Sirac and Clapouse-Chaillol basement indicates a pre-Eocene deformation event during which the Pelvoux basement and its sedimentary cover suffered uplift and erosion.

Assuming a piggyback thrust sequence for the Aiguille de Morges area (see previous sections), it is possible to date some of the compressional structures in the central Pelvoux, from their continuity into this area. For example, the Lower Muzelle fault exposed in the Villar-Loubiere area is continuous with the basement structures exposed at Navette. Assuming a post-Eocene age for the Navette basement thrusts, the same age of formation can be assigned to the La Muzelle structure, and the majority of basement thrusting throughout the central Pelvoux - inference.

The Aiguille de Morges thrust complex presents a key area for understanding the continuity of thrusts and displacements through the central Pelvoux and indeed around the Western Alpine arc. This aspect will be examined further in Chapter 5.

2.5 SW margin of the Pelvoux massif

The principal structural elements making up the SW margin of the Pelvoux massif are shown in Figure 2.5.1.

2.5.1 Chaillol thrust:

S of the Severaisse-Valgaudemard valley (Figure 2.5.1), the SW limit of the Pelvoux basement is fault controlled. The Chaillol basement overthrusts a foot-wall sequence of Triassic and Jurassic sediments along a moderately NE dipping thrust plane - termed here the Chaillol thrust. This foot-wall cover sequence is overturned (60° NE/160), and is cut out against the basement thrust plane. A foot-wall syncline interpretation, concurrent with the high bedding cut-off angle and overturning of the foot-wall sequence is proposed. The basement thrust plane crops out at an altitude of 1470 m, to the N of the settlement of Mallarua (G.R. 7424,49600), dipping 44° NE and striking 120° . A planar thrust fabric, more or less sub-parallel to this, affects both the hanging-wall basement and foot-wall basalts. A mineral stretching lineation recorded from the fault plane fabric plunges towards 050° , and is supportive of a SW thrust transport interpretation. The well exposed foot-wall Liassic limestones and shales are deformed by tight SW-verging chevron-type folds (Plate 2.5.1), with an average axial trend towards 148° , and a penetrative axial planar cleavage (50° NE/150). The formation of these folds is interpreted to be contemporaneous with the SW overthrusting of the Chaillol basement. A minor thrust imbrication of the overturned foot-wall sequence is also evident along the lower slopes, towards les Costes (Figure 2.5.1).

A SW transport interpretation is further substantiated from the nature of the northward termination of the Chaillol basement sheet: it is terminated against a NE-SW trending thrust fault interpreted to be a lateral ramp. This is best exposed SW of St. Jaques-en-Valgaudemard (G.R. 7423,49620) (Figure 2.5.1), where the Chaillol basement

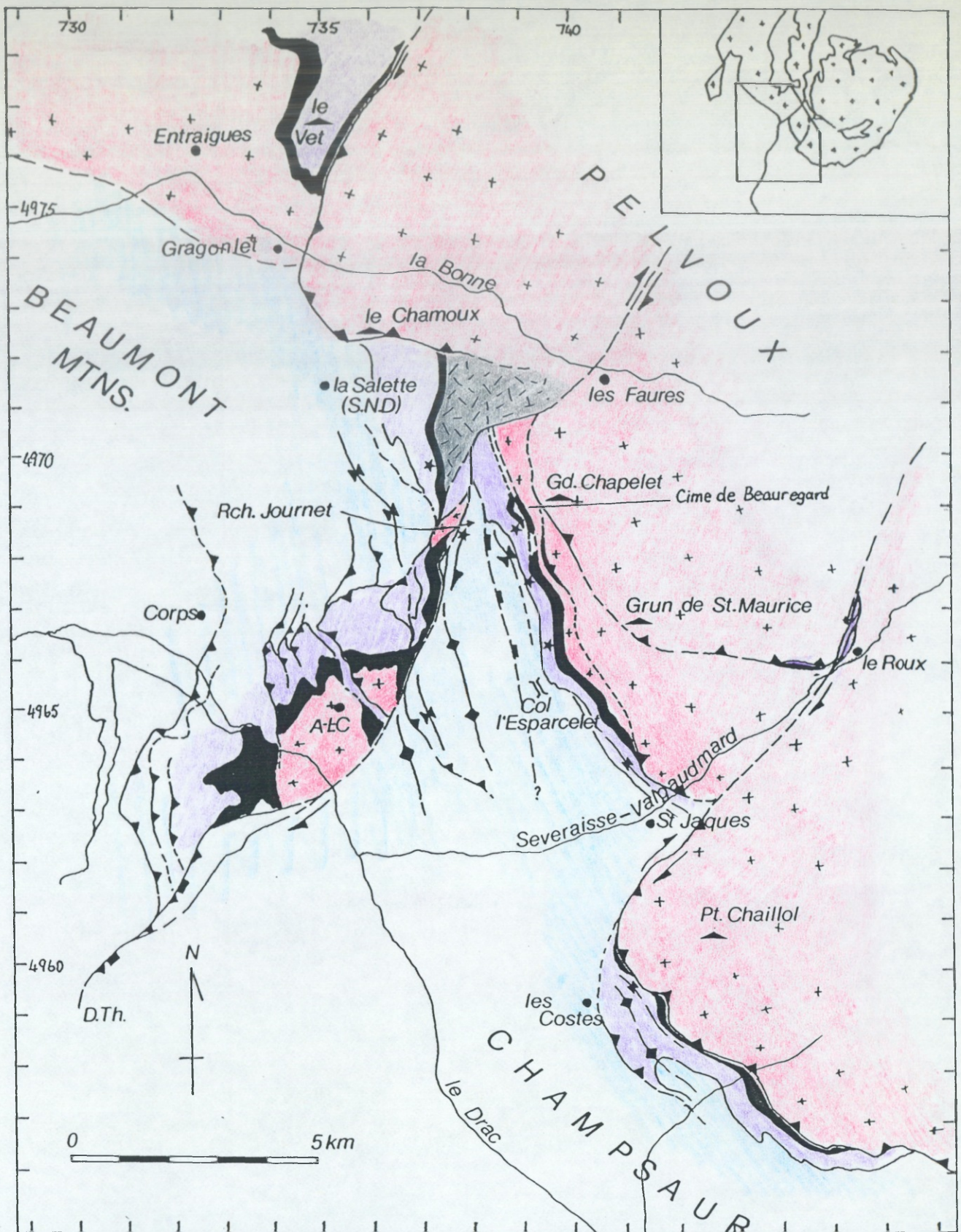


Figure 2.5.1: Structural map of the SW Pelvoux area (location and legend indicated in Figs. 1.7 & 1.8). Based on mapping by the author. Black - Triassic.

overthrusts Mid-Jurassic shales. The thrust plane dips 35-40°SE, striking 020°, and fibrous slickensides, shear bands, and asymmetric shear folds in its foot-wall are consistent with a SW direction of thrust movement. The geometries of the foot-wall cut-off trends for the overturned Triassic and Liassic provide additional evidence that this structure is a lateral ramp.

The Chaillol lateral ramp follows the Severaisse-Valgaudemard valley and may link with a zone of basement-cover imbrication exposed at le Roux (Figure 2.5.1), which ultimately branches with structures in the central Pelvoux (Figure 2.3.9). Because of poor exposure, the origin of these basement-cover imbricates is uncertain. They may be of extensional origin (i.e. sedimentary olistoliths or foot-wall collapse imbricates) later deformed within a strike-slip zone, or newly formed. It is conceivable that like other Alpine structures documented from the central Pelvoux, the Chaillol lateral ramp was partly inherited from a pre-existing Jurassic structure.

A newly identified gently NE-dipping basement thrust which overthrusts Triassic and Lower Liassic sediments towards the SW, was mapped to the WNW of the village of St. Maurice-en-Valgaudemard, and is termed here the St. Maurice thrust. This may branch with the Chaillol lateral ramp in the Severaisse-Valgaudemard valley.

The Chaillol thrust and associated foot-wall deformation can be dated from its relationship with the Eocene to Oligocene Gres du Champsaur flysch exposed on the southern margin of the Pelvoux massif. To the E of les Infournas, the BRGM 1:50 000 Orcieres sheet (1980) shows the Chaillol thrust to be overlapped by the Gres du Champsaur flysch sequence. This is confirmed by the work of Chambers (1987), who demonstrated that the eastward continuation of the Chaillol thrust was truncated and transgressed by the basal Eocene flysch unconformity and from fold and fault geometries and kinematic indicators, proposed a direction of thrust transport towards the SW or

SSW. In view of the new data presented here, a SW directed, pre-Eocene interpretation is proposed for the Chaillol thrust. This disagrees with the interpretation proposed by Gidon (1979) who postulated a SE thrust transport direction for the Chaillol basement on southern margin of the Pelvoux, correlating this thrusting with the pre-Senonian deformation documented from the Devoluy. The expression of pre-Senonian shortening in the Pelvoux massif, if any, will be considered in Chapter 5. A post-Eocene SSW regional tilting of the Eocene flysch is also evident, observable from the Route Napoleon and confirmed by the work of Chambers (1987). The SW dip in the order of 30°, is interpreted to signal post-Eocene thrusting at a deeper unexposed level of the section. This may be related to movement on what is interpreted to be the lateral continuation of the external Belledonne thrust to the NW.

2.5.2 Col l'Esparcelet:

N of the Severaisse-Valgaudemard valley (Figure 2.5.1), by the village of Moussuq (G.R. 7420,49640), the Pelvoux basement overthrusts a foot-wall of overturned Liassic sediments, with the Triassic presumably located in the unexposed foot-wall. The basement-cover thrust geometry closely resembles that described S of the Severaisse-Valgaudemard valley and a similar foot-wall syncline interpretation may be proposed. The overturned Jurassic sequence is likewise deformed by tight SW verging chevron type folds with a NW-SE axial trend. To the N, erosion exposes a deeper section and the thrust trace cuts back into the basement, such that the overturned basement-Triassic unconformity is exposed.

Further N towards the Col l'Esparcelet, a complication arises affecting the interpretation of this overturned sequence. In the Ravine de Lavey (G.R.7400,49670), S of the Col, Lemoine et al. (1985) document evidence in favour of Liassic age extensional faulting. From this ravine they describe an isolated sedimentary olistolith of Triassic basalts and dolomites, which on all mappable sides is

juxtaposed against Hettangian and Sinemurian limestones. In the absence of obvious fault contacts, and from other observations, I found no reason to contradict this interpretation; indeed it is difficult to propose any alternative.

To the N of the Col l'Esparcelet (Figure 2.5.1) a detailed structural section of the overturned sequence was logged (G.R.7397,49688). The results from this log are summarized in Figure 2.5.2. Similar relationships to those described by Lemoine et al (1985) S of the Col, were documented. An irregular shaped body of bedded Triassic basalts and dolomites is enveloped by Lower Liassic limestones and syn-sedimentary micro-breccias at the lateral margins of the basalt body were observed. An interpretation involving a sedimentary olistolith is preferred. The overturned basement-cover contact is folded into a SW-verging anticline, the backlimb of which is sliced through by a gently NE-dipping thrust bringing up the Pelvoux basement (Figure 2.5.3). Kinematic indicators suggest that the motion on this thrust was towards the SW. To the N this fault cuts back into the basement and presumably branches with a structurally higher, gently dipping basement thrust, which outcrops on the SW face of the Grand Chapelet (Figure 2.5.4).

On the SW margin of the Pelvoux, at the Cime de Beauregard (Figure 2.5.1), the orientation of the basement-cover contact remains vertical to overturned. In the overturned Liassic sequence are undeformed Triassic basalts and dolomites which are bedding-parallel and stratigraphically out of place (Figure 2.5.4). Unfortunately the lower contact is not exposed making the interpretation of this structure difficult. In view of the other olistoliths documented to the S, this structure could be interpreted in the context of an enormous sedimentary olistolith. However it is unlike any other olistolith described from Pelvoux and adjacent basement massifs (this Chapter), with no evidence of syn-sedimentary deformation in the surrounding sediments and an unusual bedding-parallel geometry. Indeed

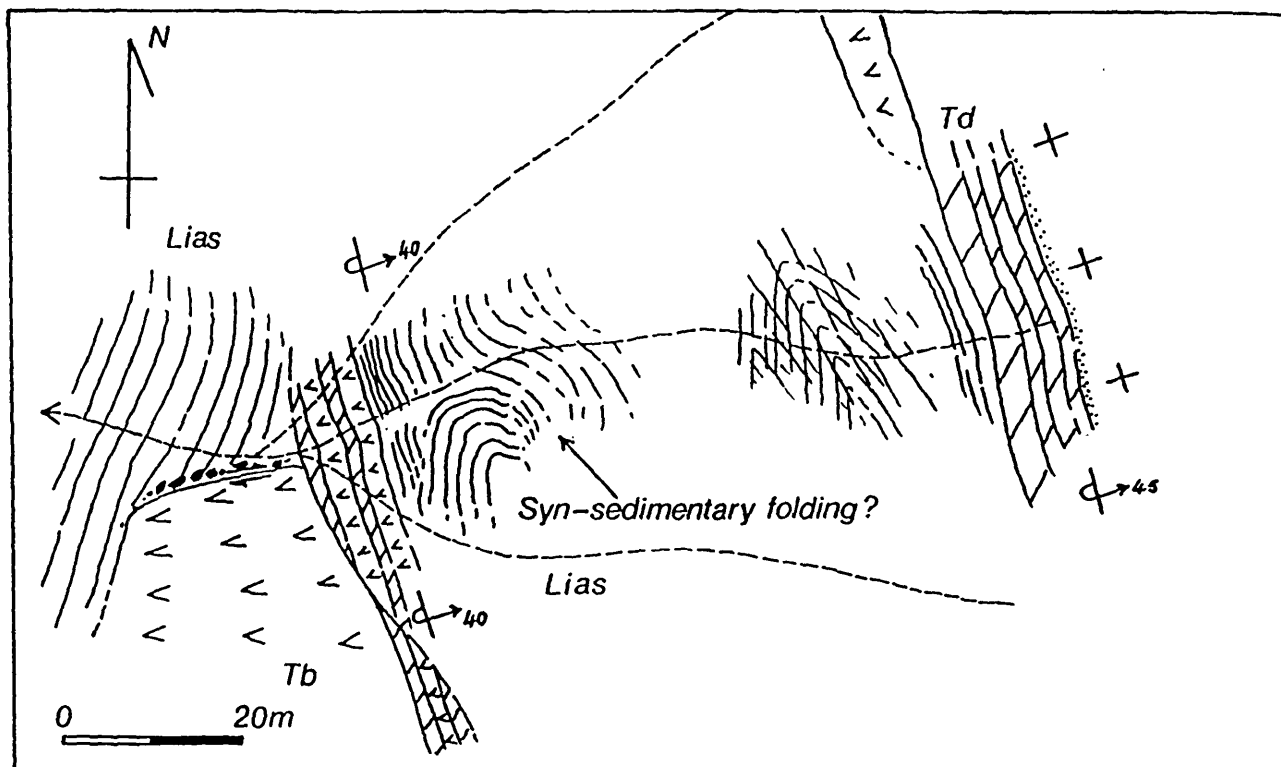


Figure 2.5.2: Structural and stratigraphic log of the SW Pelvoux margin, N of Col l'Esparcelet (G.R.7397,49688), showing olistolith in Lower Liassic sediments. Crosses-basement, Td-Triassic dolomites, V-basalts.

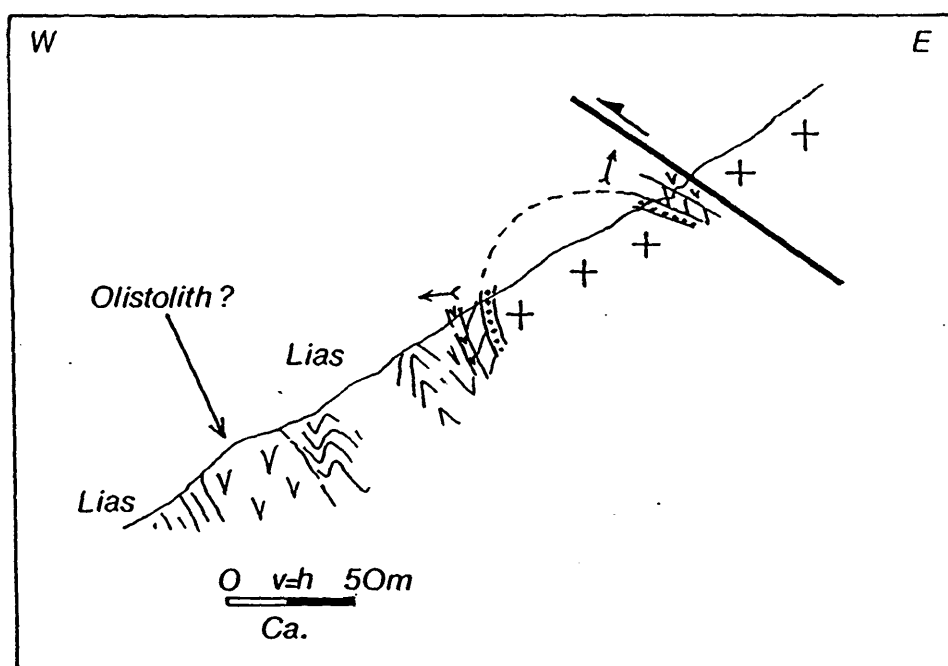


Figure 2.5.3: Schematic cross-section across the SW Pelvoux margin (Col l'Esparcelet, G.R.7397,49688). See text for details. Key is the same as for Fig. 2.5.2.

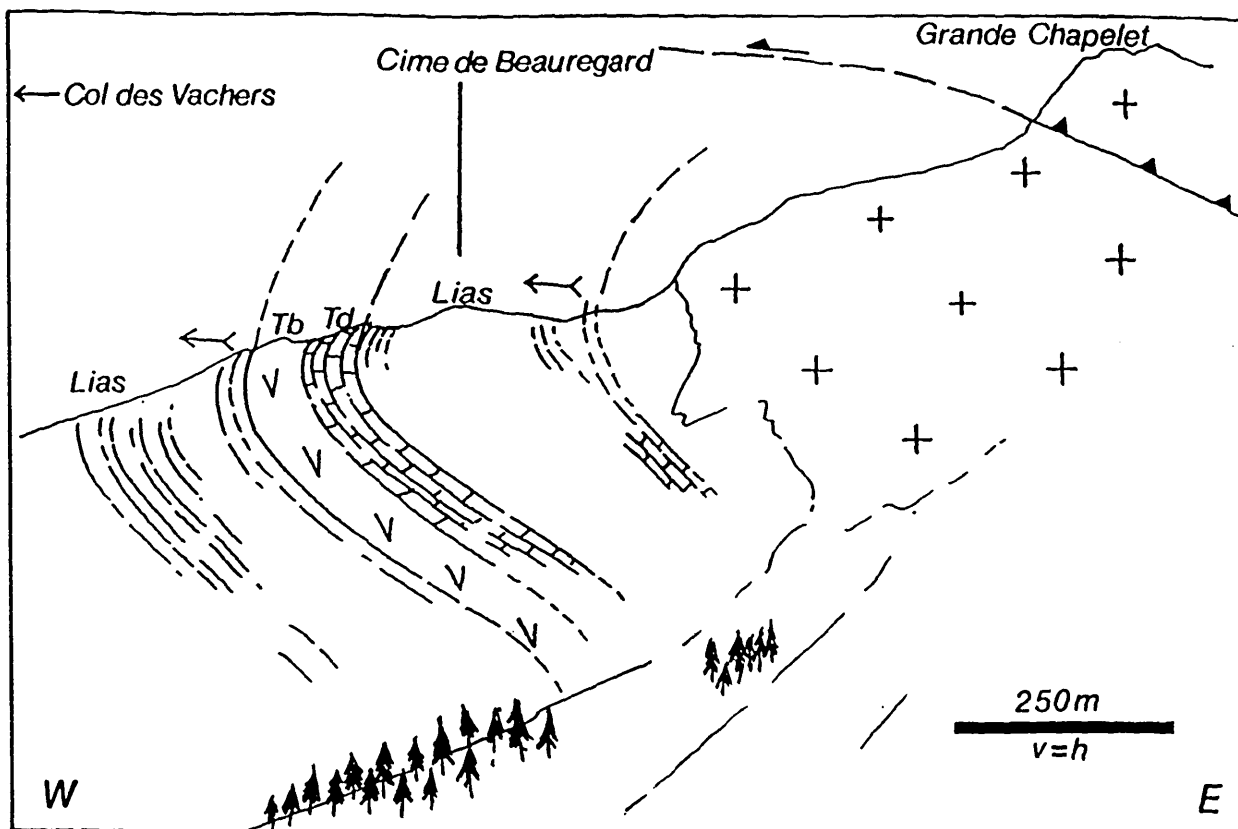


Figure 2.5.4: Field sketch across the SW Pelvoux margin at the Cime de Beauregard. Crosses-crystalline basement, Td-Triassic dolomites, Tb-Triassic basalts. See text for details.

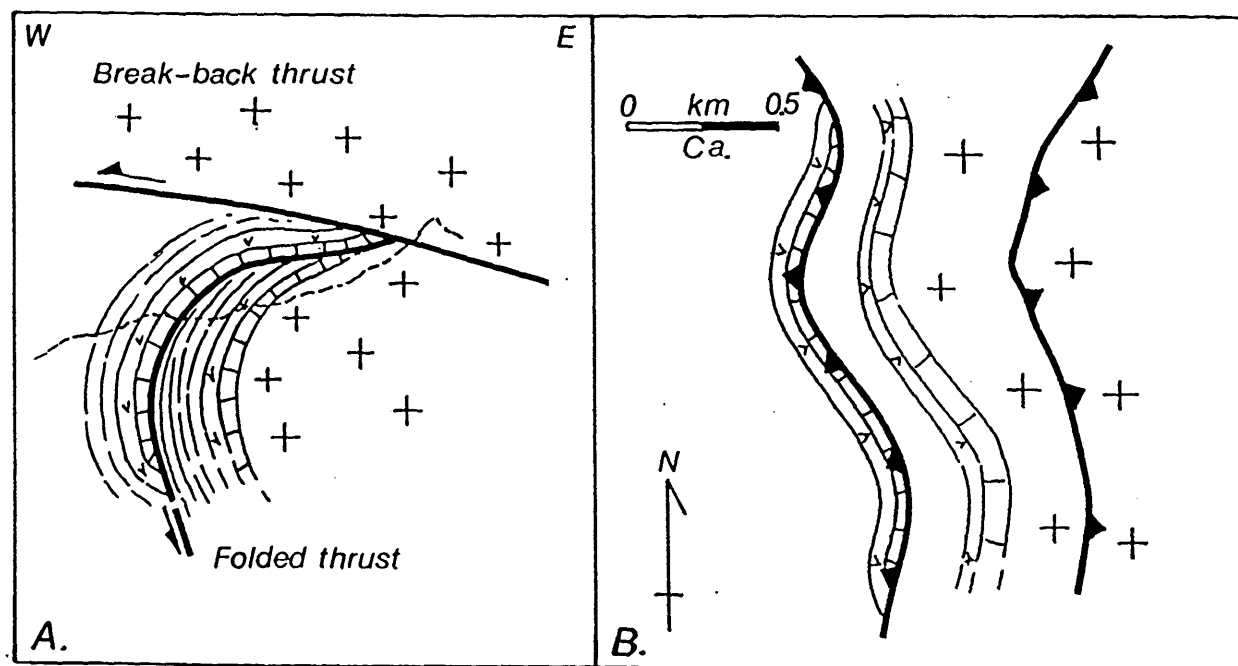


Figure 2.5.5: Alternative thrust interpretation for the structure at the Cime de Beauregard. Key as for Fig. 2.5.4. See text for details.

it is difficult to envisage how and from where it came if a sedimentary olistolith interpretation is proposed. A folded thrust interpretation could equally suffice to explain the structure. A hanging-wall flat/foot-wall flat thrust geometry, later folded during the uplift of the Pelvoux basement (Figure 2.5.5 B), could account for the bedding-parallel relationship and absence of internal deformation in the basalts and dolomites. The pinching out laterally of the Triassic sediments could equally be explained by hanging-wall lateral ramps in a ramp-flat staircase trajectory (Figure 2.5.5 B). This folded thrust interpretation requires a detachment of the sedimentary cover from its underlying basement prior to the folding and uplift of the Pelvoux basement. Unfortunately, no evidence supporting this is documented elsewhere along the SW margin of the massif.

2.5.3 Aspres-les-Corps fault:

The basement fold and thrusts of the SW Pelvoux are terminated laterally against a NE-SW trending, steeply dipping to vertical fault exposed at the Tete du Gris (G.R. 7385,49715), which to the NE juxtaposes Carboniferous Houiller against crystalline basement, and to the SW Carboniferous Houiller against the Jurassic cover sequence. To the NE this fault can be traced across the Entraigues-Bonne valley, where it passes to the W of les Faures and into the Valsenestre area where it may continue as the Col de la Muzelle fault. To the SW it corresponds to the Aspres-les-Corps fault (Gidon et al, 1976; Gidon & Aprahamian, 1981). At the SW margin of the Pelvoux, the NW side of this fault shows a normal basement-cover stratigraphic contact which dips 45°NW. This steep NW dip may reflect deeper thrusting, explaining the cover shortening to the W. The structure previously described is cut-out by a NW-SE trending basement thrust exposed at le Chamoux (G.R. 7361,49725), to the NE of Sanctuaire de Notre Dame de la Salette (Figure 2.5.1), which uplifts the Pelvoux basement on its SW margin. This thrust can be traced northwards across the Entraigues-Bonne valley by the

village of Gragonlet, and from its changed trend to NE-SW, is interpreted as a lateral structure to the SW directed Chamoux thrust. Gidon and Aprahamian (1981) interpret the Chamoux fault to be a Jurassic normal fault which initially dipped to the W, and was sheared over to the SW during subsequent Alpine deformation. However this interpretation is not consistent with observed foot-wall deformation affecting the Liassic cover, or the three-dimensional geometry of the Chamoux thrust.

At the Rochers de Journet (G.R. 7382,49692) the contact relationship of the Aspres-les Corps fault with the adjacent sedimentary cover is well exposed. The fault surface is vertical to marginally overturned towards the E (Figure 2.5.6). An apparent thickening of Bajocian carbonates towards the basement fault and minor normal faulting apparently sealed by overlying Bajocian shales (Plate 2.5.2) may provide evidence that this fault was active throughout the Jurassic and is therefore an original Jurassic structure. However, like the many identified Jurassic structures in this area, the original Jurassic contact has been modified by later compression. The cover sediments of le Cucuron (Figure, 2.5.6) display little in the way of shortening or internal deformation. However a marked intensification of cleavage and minor folding occurs adjacent to the basement fault. This vertical cleavage strikes N150 and is asymptotic in plan view. The sense of curvature suggests that this fault suffered a dextral strike-slip reactivation, presumably during Alpine compression. This dextral motion is consistent with the movement recorded along the Col de la Muzelle fault, with which this structure is interpreted to be continuous. The presence of a number of olistolith type bodies mapped in sediments adjacent to this fault supports the postulate of a Jurassic movement history.

Further S at the village of Aspres-les-Corps, a lozenge shaped body of basement and Triassic basalts crop-out adjacent to the Aspres-les-Corps fault (Figure 2.5.8). The basement is itself imbricated by Alpine thrusts. One



Plate 2.5.2: Mid-Jurassic sedimentary growth against the Aspres-
les-Corps fault. See Fig. 2.5.6 for locality and details.

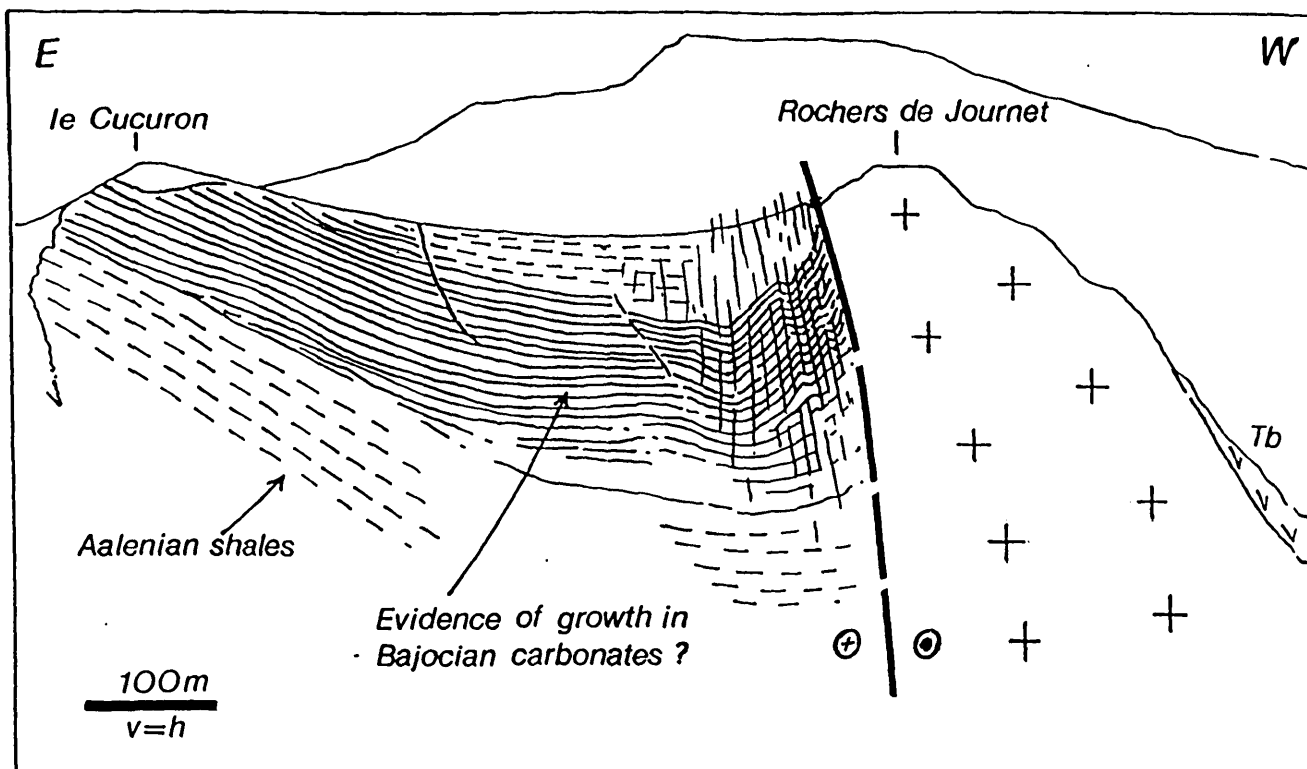


Figure 2.5.6: Field sketch across the Aspres-les-Corps fault at the Rochers de Journet (Fig. 2.5.1) showing possible Mid-Jurassic growth. See also Plate 2.5.2.

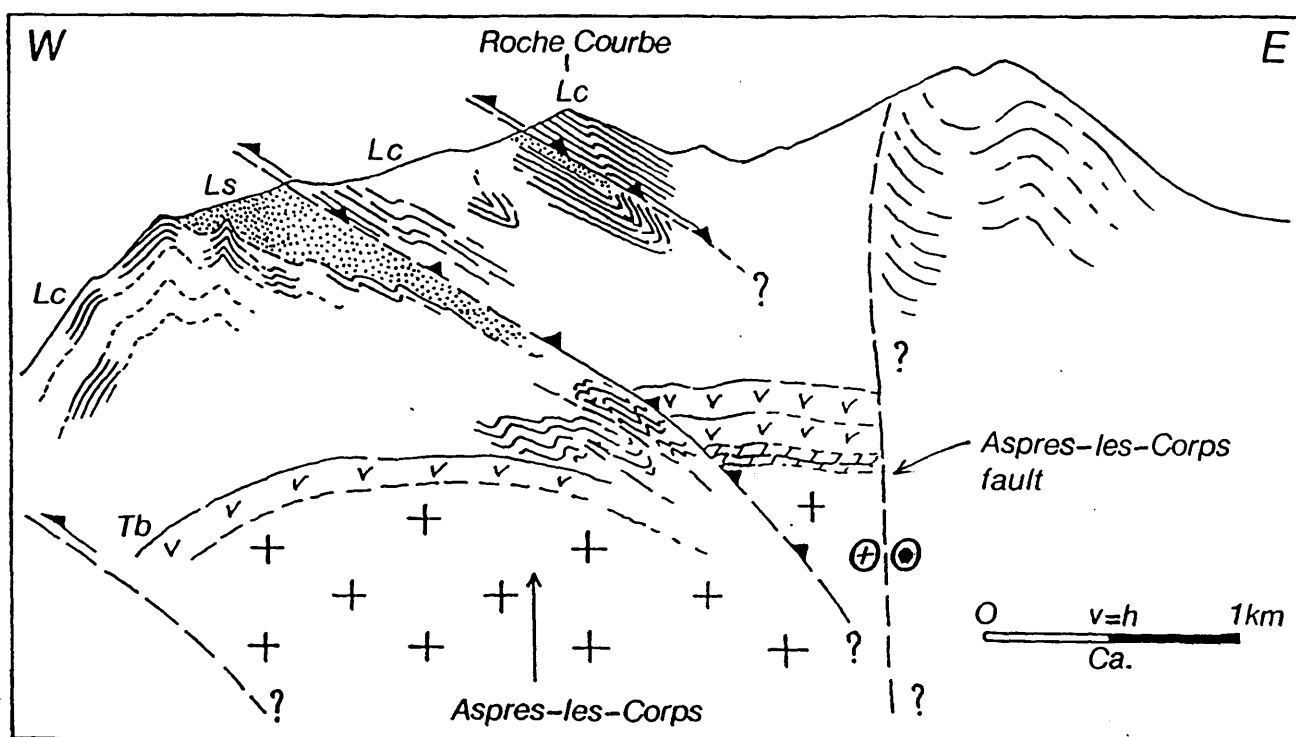


Figure 2.5.7: Sketch section across the Aspres-les-Corps structure at Aspres. Crosses-basement, brickwork-Triassic dolomites, Tb-Triassic basalts, Lc-Lias calcaire, Ls-Lias schisteux. See text for details.

such thrust is well exposed both to the NE and SE of Aspres-les-Corps. On the E side of the Ravine Serre (G.R. 7360,49660) basement and basalts overthrust a Lower Liassic sequence and both thrust fault and accompanying penetrative foot-wall deformation are exposed. Tight foot-wall folds trend NW-SE (N140°-N145°), and are strongly asymmetric towards the SW. From the systematic analysis of cut-off trends and mineral stretching lineations from minor foot-wall thrusts, a SW direction of thrust transport is suggested. The average trend of the mineral stretching lineations is towards 040°, substantiating this claim. This basement thrust can be traced into the overlying cover where it emplaces Lower Lias over Upper Lias. Indeed most of the Liassic stratigraphy overlying the Triassic basalts is extensively imbricated (Figures 2.5.7). Gidon et al. (1976) show the NW contact of the basalts against the Lower Lias to be a thrust contact. However in the area of Aspres-les-Corps the exposure is too poor to be definitive about the exact nature of this contact, although it appears to be a normal stratigraphic contact. Furthermore at Beaufin, to the S, the contact between the Triassic basalts and overlying Liassic sequence is clearly exposed and conforms to a normal stratigraphic contact dipping steeply (80°) towards the WNW. This steep dip is interpreted to result from basement thrusting deeper in the section, which is apparent to the W, emplacing a Lower Liassic sequence over Mid-Jurassic shales. This same contact may well, in part, correspond to the Devoluy thrust (Figure 2.5.8), or at least branch with it in the subsurface.

The Aspres-les-Corps lineament has featured in many accounts concerning the Alpine geology of this area. Goguel (1963) attributed the present alignment of structures along the Aspres-les-Corps fault to an important dextral strike-slip displacement which could be traced from the Eastern Grandes Rousses fault in the N, through the central Pelvoux via the Col de la Muzelle, and into the Aspres-les-Corps area. Vernet (1974) referred to the basement slices along the Aspres-les-Corps faults as "ecailles flottantes", translatable as "floating tectonic

wedges", and suggested that they are the result of a "tectonique d'ejection verticale" by an "axe anticlinal profonde". In the most comprehensive account to date, Gidon et al. (1976) proposed a complex and long-lived tectonic history commencing with a Jurassic age of initiation, substantiated by stratigraphic data (Gidon and Aprahamian, 1981), a sinistral strike-slip movement during the Priabonian ("phase Nummulitique") reflecting N-S compression, and a late Neogene phase of dextral movement associated with the formation of the Devoluy thrust.

From this work, the conclusive last phase of movement on this fault is deemed to be a dextral strike-slip movement. However, the complex anastomosing fault pattern (Figure 2.5.8) and thrust imbrication of the Aspres-les-Corps basement may be interpreted in one of two ways:

1. In the subsurface the pre-existing Aspres-les-Corps lineament is sliced through and truncated by a low-angle basement thrust(s), which to the SW becomes the Devoluy thrust.
2. The present structural geometry has arisen from the strike-slip reactivation of a pre-existing non-planar fault profile (in map plan), and is analogous to the structural styles and geometries described from areas of strike-slip tectonics.

The brittle basement thrusting and ductile deformation of the cover can possibly be explained in the context of a transpressional "restraining bend" in the pre-existing Jurassic fault (Figure 2.5.8). On the S side of the fault a set of en echelon folds whose fold axes and axial cleavage asymptote into the shear direction, are characteristic of a transpressional restraining fault bend and associated with the change in strike orientation of the Aspres-les-Corps fault. Likewise the brittle thrusting and deformation of the rigid basement may also reflect the problem of reactivating a non-planar fault profile, and the compressional deformation may overcome the irregularity in

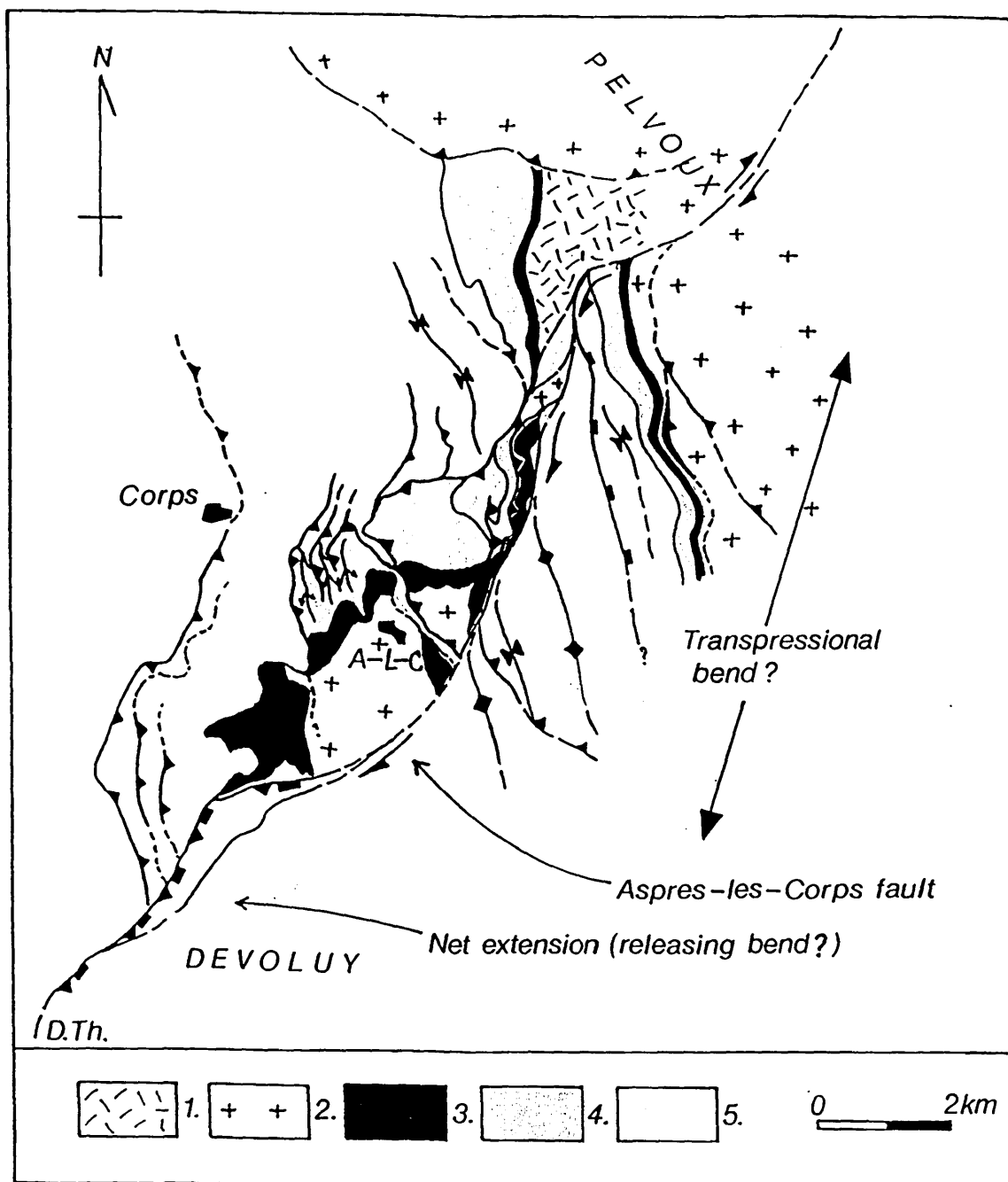


Figure 2.5.8: Structural map of Aspres-les-Corps and surrounding area showing suggested strike-slip geometry. 1. Carboniferous Houiller, 2. crystalline basement, 3. Triassic basalts and dolomites, 4. Lias calcaire, 5. Lias schisteux. Based on mapping by the author.

the pre-existing fault by propagating thrusts into the foot-wall. In cross-section, the deeper geometry may resemble that of a flower structure with the thrusts at the surface steepening downwards and branching off the pre-existing fault at depth. The transport direction recorded by these thrust imbricates may therefore only record a local direction of transport reflecting the local stress distributions around the non-planar segment of the fault. The local net extension observed on the Devoluy thrust (Figure 2.5.8) along its northern margin can also be explained if the Devoluy thrust reactivates the pre-existing Aspres-les-Corps structure at depth.

2.6 Subalpine Chains: Vercors and Devoluy

2.6.1 Introduction

From N to S the Subalpine Chains curve in an anti-clockwise fashion around the external basement massifs of the Pelvoux and Belledonne (Figure 2.6.1). A certain amount of controversy surrounds the continuity of fold and thrust structures from the Vercors into the Devoluy Subalpine Chains via the E-W trending structures of the Diois (for reviews see Graham, 1978; Siddans, 1979). This aspect and the evolution of arcuate fold and thrust patterns in the external French Alps will be addressed in Chapter 5.

The external basement massifs can be considered as late structural culminations (see Chapter 5) which expose deeper level basement-cover thrust and inversion geometries. The adjacent Subalpine Chains however, comprise structures formed at a much higher structural level, explaining the marked contrast in structural style between these two distinctive structural domains.

Despite the absence of field studies in the Subalpine Chains during the present study, two section lines based on published data are considered across the Vercors and Devoluy, to illustrate different interpretations and possible relations between the Pelvoux and Belledonne basement massifs and the adjacent Vercors and Devoluy Subalpine Chains.

2.6.2 Vercors cross-section

Two tentative solutions are proposed for a single cross-section line across the Vercors massif. The first solution (Figure 2.6.3 A & B) assumes a model of thin-skinned thrusting, whilst the second (Figure 2.6.4 A & B) examines the validity of a model which invokes basement involvement at depth, either as newly formed basement thrusts, or through the reactivation of pre-existing Jurassic structures. The viability and constraints for each

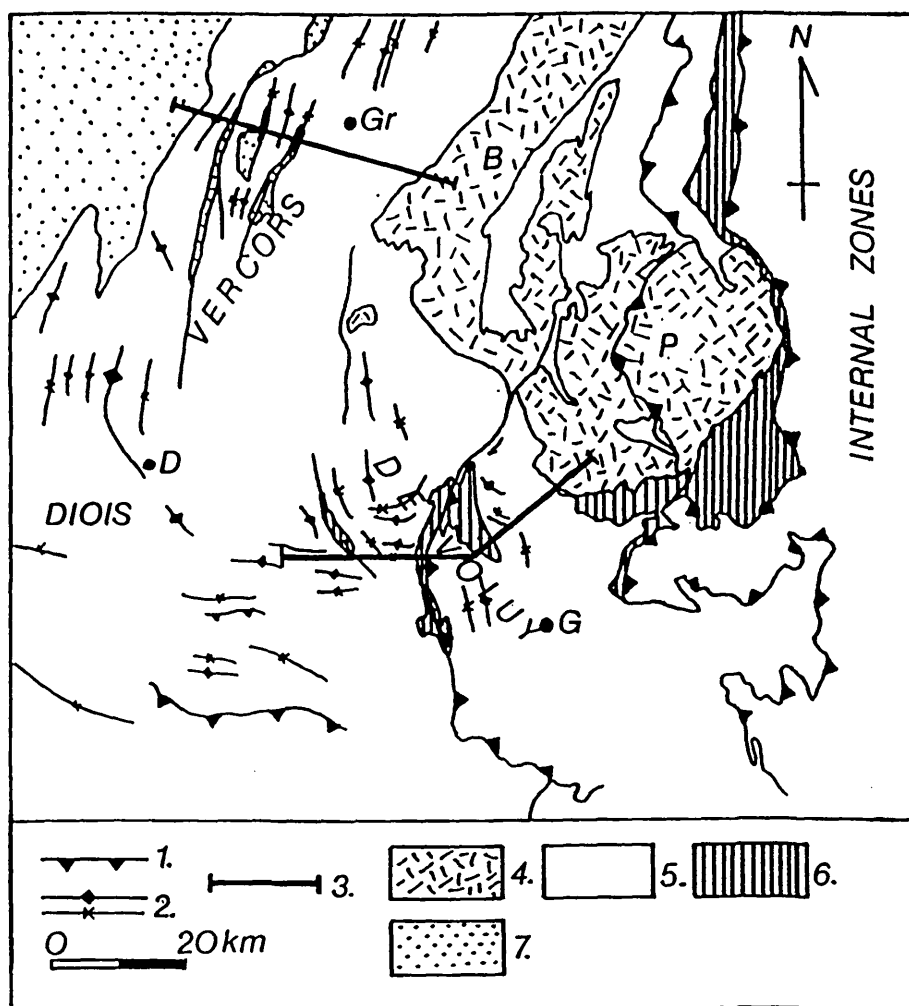


Figure 2.6.1: Simplified structural map showing the location of the Vercors and Devoluy Subalpine Chains with respect to the Pelvoux massif. 1. thrusts, 2. folds, 3. section lines, 4. basement, 5. Mesozoic cover, 6. Eocene-Oligocene, 7. Miocene. See text for references and details.

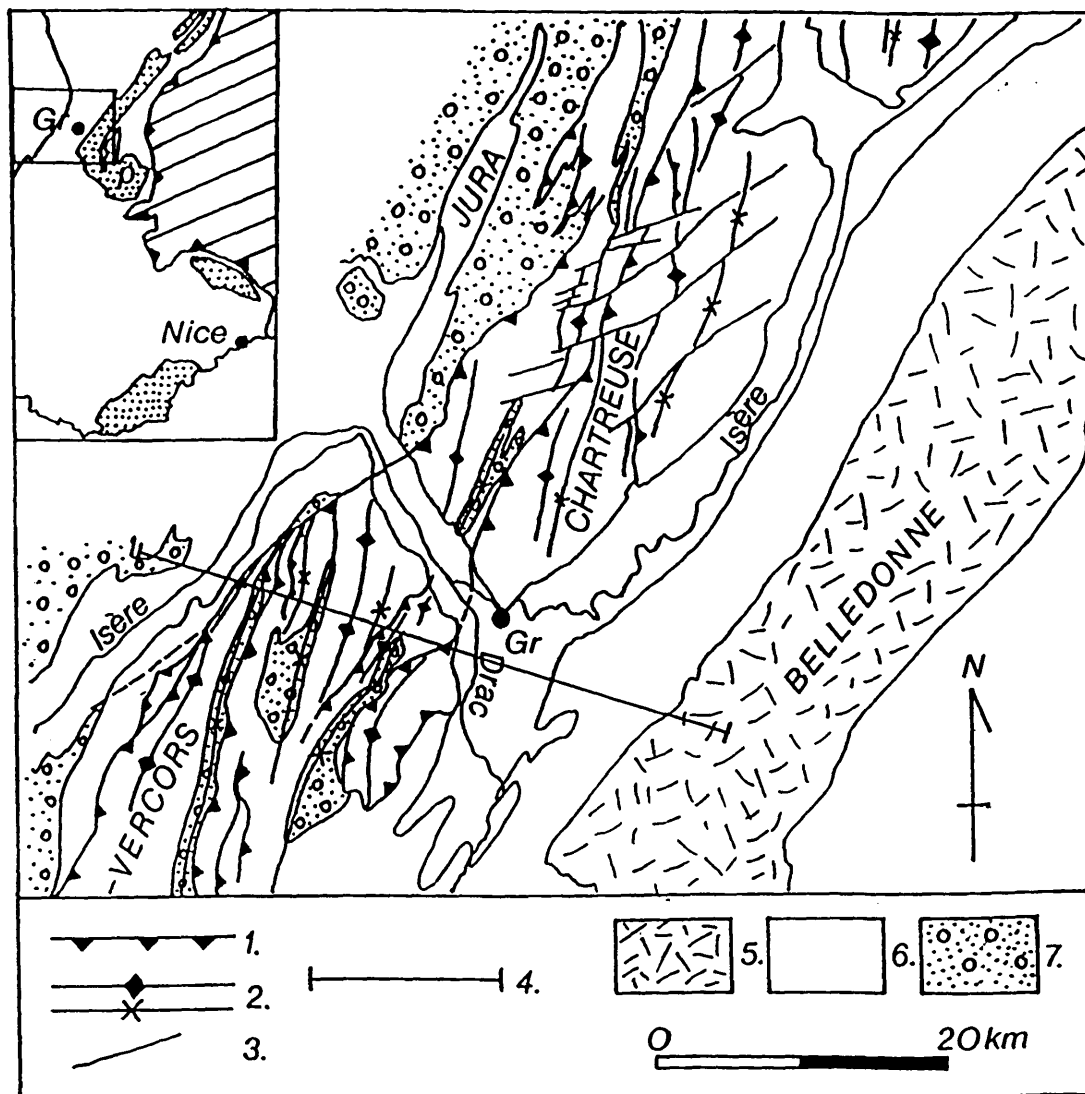


Figure 2.6.2: Simplified structural map of the Vercors and Chartreuse Subalpine Chains in the Grenoble region, 1. thrusts, 2. folds, 3. strike-slip faults, 4. Vercors section line, 5. basement, 6. Mesozoic cover, 7. Miocene molasse, Gr-Grenoble. Based on the BRGM 1:50 000 Grenoble map sheet (1978).

model will be discussed.

The location of this cross-section line is shown in Figures 2.6.1 & 2.6.2, and is parallel to the designated thrust transport direction for this part of the Subalpine Chains (Mugnier, in press). The surface structure is taken from the BRGM 1:50 000 Grenoble map sheet (1978) and a cross-section by Gidon & Arnaud (1978). The depth to the pre-Triassic crystalline basement is from Menard (1979; 1980) who collated seismic and borehole data to produce a map of the pre-Triassic basement in SE France. The step shown in the pre-Triassic basement is also determined geophysically and is known as the Isere fault (see also Siddans, 1982). Menard (1979; 1980) was the first to propose that the Belledonne basement was detached at its base and overthrusts the Subalpine cover, a hypothesis later confirmed by the deep crustal reflection profile ECORS-CROP (Bayer et al., 1987). Since this thesis is not concerned with the detailed timing of thrusting in the Vercors, the age of the deformation can be considered broadly as Miocene, as the main thrusts override Miocene molasse in their foot-walls. The potential detachment levels in the Subalpine cover sequence probably comprise i) the basement-cover interface, especially if Triassic evaporitic levels exist beneath the Subalpine Chains, and as suggested by Goguel (1952) to be more important in the Subalpine Chains, ii) the "Lias schisteux" and iii) the "Terres Noires", which have already been shown to be important detachment horizons in the Pelvoux and Rochemaure (this Chapter), where the easy slip Triassic gypsum is largely missing.

One of the main geological problems of the Vercors plateau is its present elevation above the "regional". The reference levels chosen to determine this elevation are the post-rift Tithonian and Urgonian limestones, unaffected by Tethyan extension. The "regional" levels of these reference horizons are given by well data in the undeformed foreland (Figure 2.6.5), assuming that the foreland reference levels have not been altered by Oligocene normal

faulting associated with the Rhone-Bas Dauphine. Along the chosen section line these reference levels are elevated well above their foreland regionals without evidence of major thrusting exposed at the surface. This effect, which becomes more pronounced in the central and southern Vercors, requires some sort of explanation.

2.6.2.1 Thin-skinned thrust model:

One solution developed using a thin-skinned thrust model, assumes that this excess elevation is accomplished entirely by detachment and a horizontal shortening of the Subalpine cover in response to the westward overthrusting of the Belledonne basement (Figure 2.6.3 A & B).

The exposed surface thrust geometry allows for a complete repetition of the post-rift sequence, with a level of decoupling along the Oxfordian black shales. Irregularities in this overriding sheet are interpreted as imbricate thrusts which developed in a piggyback fashion prior to its emplacement and a longer wavelength fold related to no obvious surface structure, reflects a hanging-wall ramp anticline in the subsurface which folds the overlying detachment, with the latter now acting as a roof thrust. A lower level thrust detachment is shown to exploit the basement-cover interface, however it may equally follow the Oxfordian black shales such that no detachment occurs along the basement-cover interface.

This model infers that the undisturbed foreland stratigraphy is continuous underneath the present mountain front and the subsurface offset of the basement-cover interface - the Isere fault, is considered as a pre-existing Jurassic structure which controls the thickening of the pre-Oxfordian lithologies. In this model the Isere fault has no direct influence over the development of the thrust structures which make up the Vercors Subalpine thrust belt.

At the mountain front, the build up of horizontal

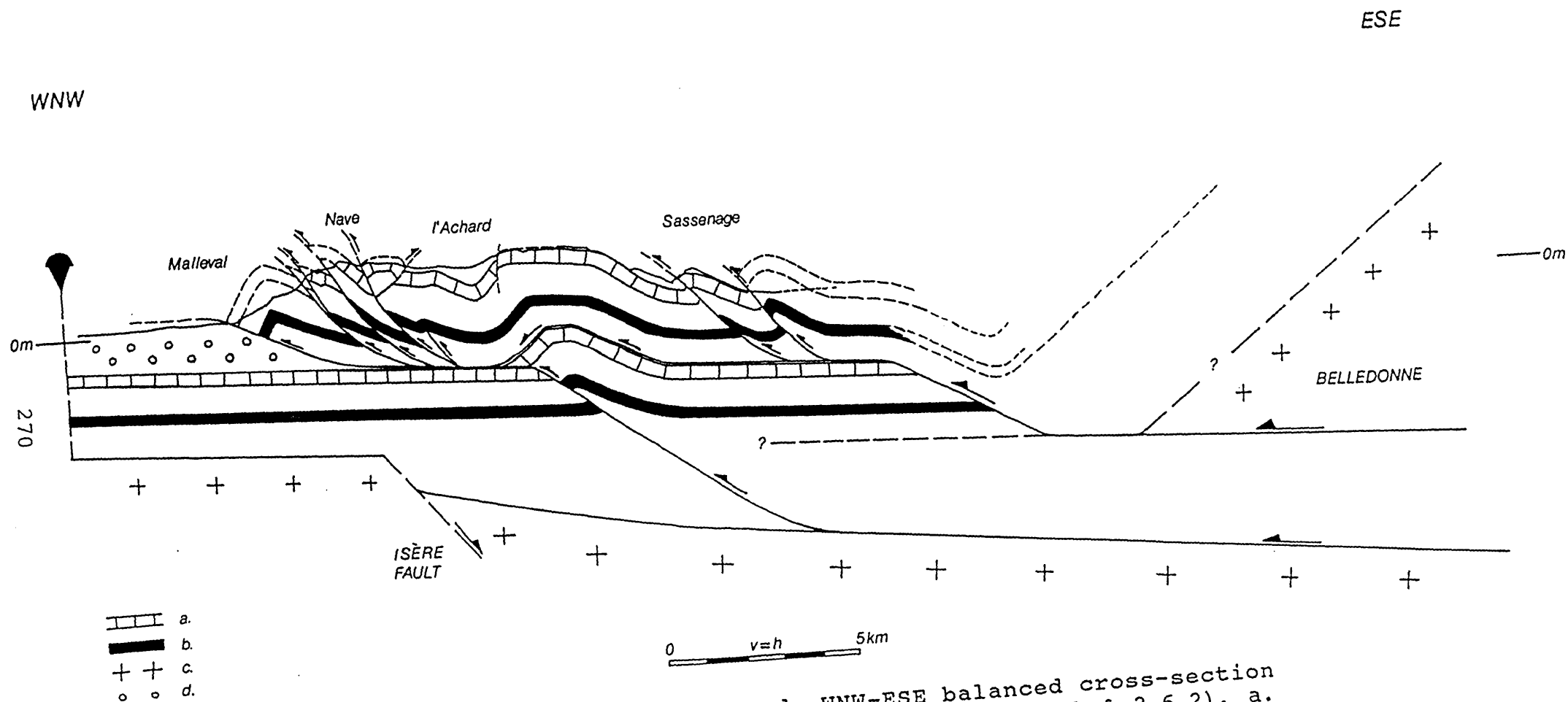


Figure 2.6.3 A: Thin-skinned, WNW-ESE balanced cross-section across the Vercors (location shown on Figs. 2.6.1 & 2.6.2). a. Urgonian, b. Tithonian, c. basement, d. Miocene molasse. See text for discussion.

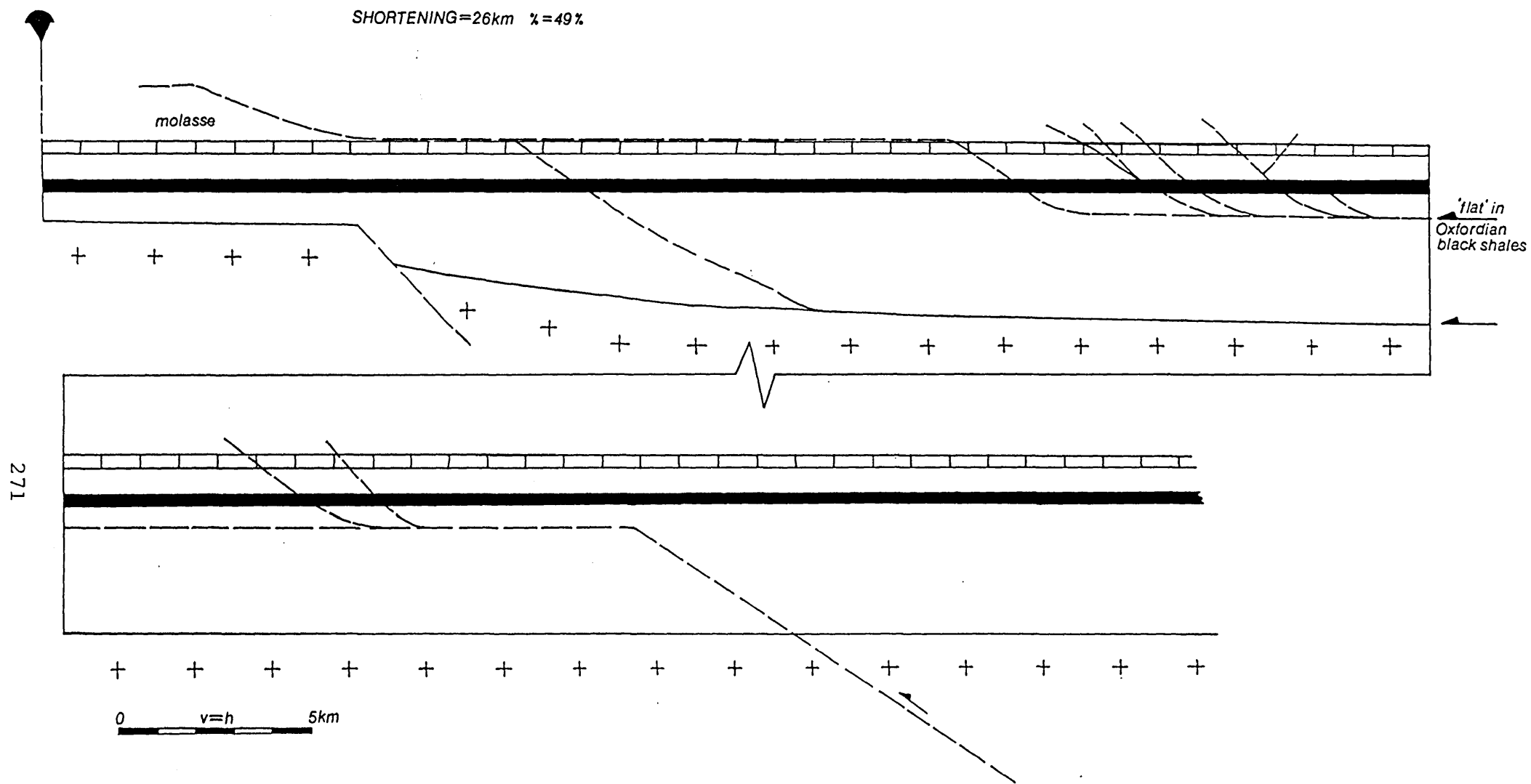


Figure 2.6.3 B: Restored cross-section across the Vercors. Legend is the same as for Fig. 2.6.3 A.

WNW

ESE

BEAUREGARD-BARRET

St. LATTIER

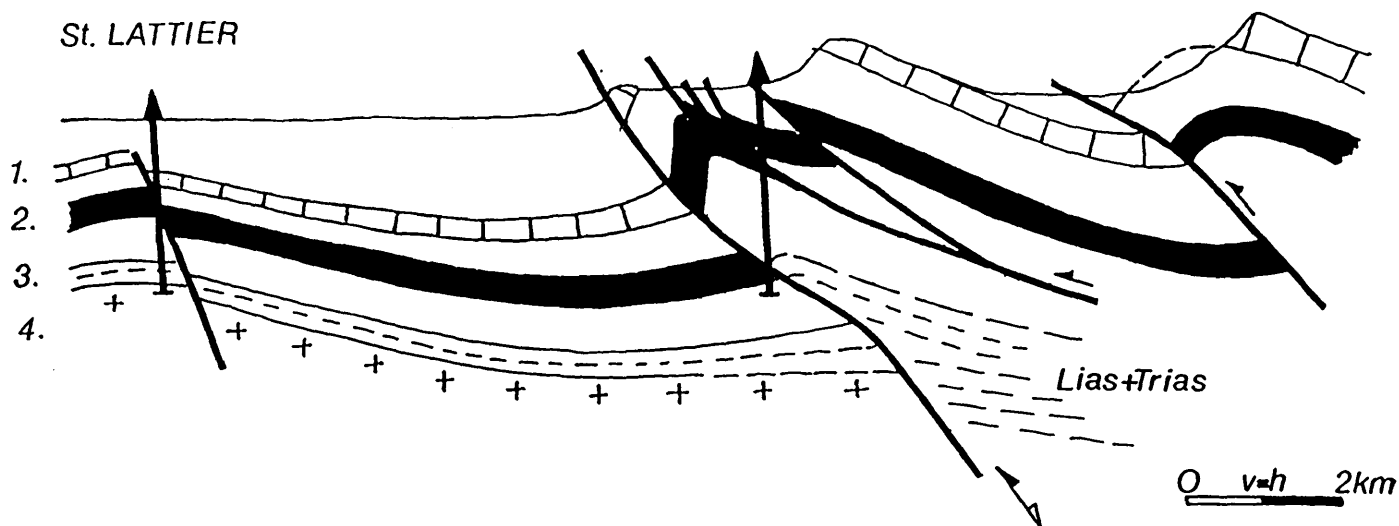


Figure 2.6.4: The Vercors mountain front structure, modified after societe de petrole de Valence (1966) and Mugnier (unpublished data). 1. Urgonian, 2. Tithonian, 3. Lias & Trias, 4. basement. See text for details.

WNW

ESE

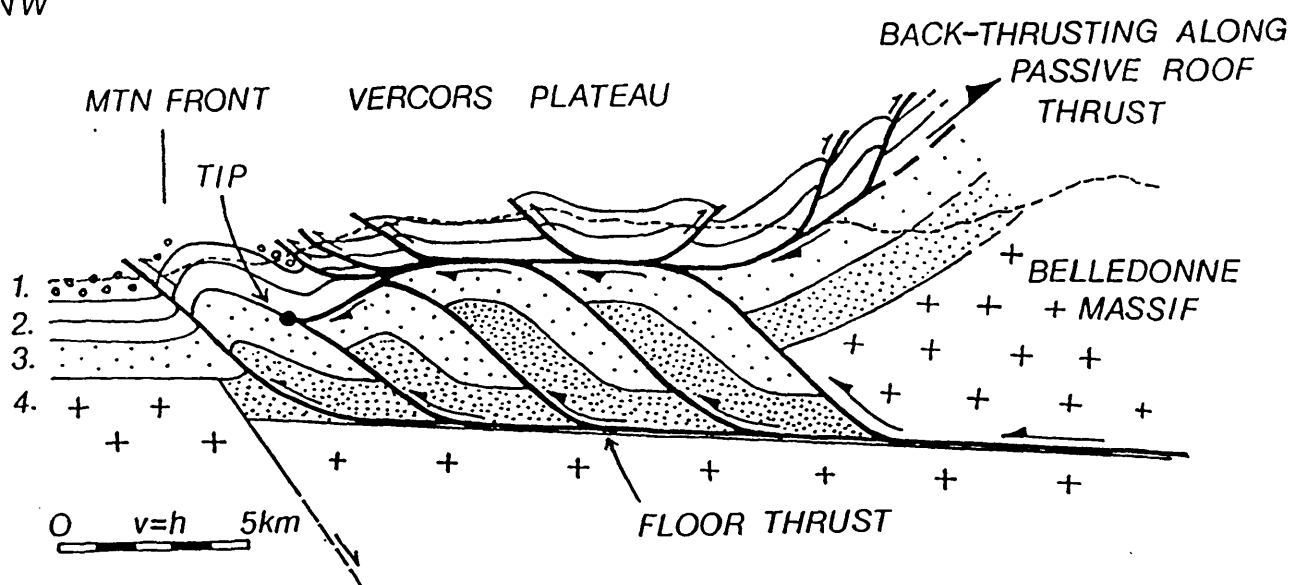


Figure 2.6.5: Thin-skinned passive-roof thrust model to explain the structure and elevation of the Vercors plateau. 1. Miocene molasse, 2. Cretaceous, 3. Upper Jurassic (light stipple) & Mid Jurassic to Lower Jurassic (dense stipple), 4. basement. See text for details.

displacements responsible for this regional elevation, may have passed into foreland basin sediments subsequently removed by erosion, or the hanging-wall was transported a considerable distance along an old Miocene land-surface. Either way, a substantial horizontal displacement is required at the present mountain front. The displacement at the mountain front is however constrained by well data to the S of this section line (Figure 2.6.4), and a maximum displacement of 2 kms is recorded. In a thin-skinned thrust model the only other way to lose the large build of displacement necessary to explain the observed regional elevation, is by back-thrusting along a bedding-parallel flat with no stratigraphic separation. In this case a deep thrust duplex, shortening only pre-Oxfordian sediments, transfers its displacement onto an Oxfordian roof thrust on which the operative sense of shear is towards the hinterland, i.e. similar to a passive-roof duplex. This is shown schematically in Figure 2.6.5. These back-thrust displacements should be apparent to the W of the Belledonne basement anticline, along a hanging-wall flat/foot-wall flat thrust contact in the Oxfordian black shales of the Isere valley. Rock types will make them difficult to recognise. The frontal thrust must however breach the overlying roof thrust, or if the roof thrust has not propagated this far out onto the foreland, breach the Oxfordian decoupling horizon to explain the present mountain front structure.

2.6.2.2 Thick-skinned inversion model:

The second solution involves less horizontal shortening of the Subalpine cover, and an additional thrust fault in the basement (Figure 2.6.6 A & B). Again two levels of detachment are envisaged in the Subalpine cover, one along the Oxfordian black shales and the other along the basement-cover interface. As the amount of displacement on the frontal thrust is restricted to 2 kms and thrusts branching from a deeper floor thrust cannot easily explain the consistent regional elevation of the chosen reference horizons, uplift of the underlying basement on a steep

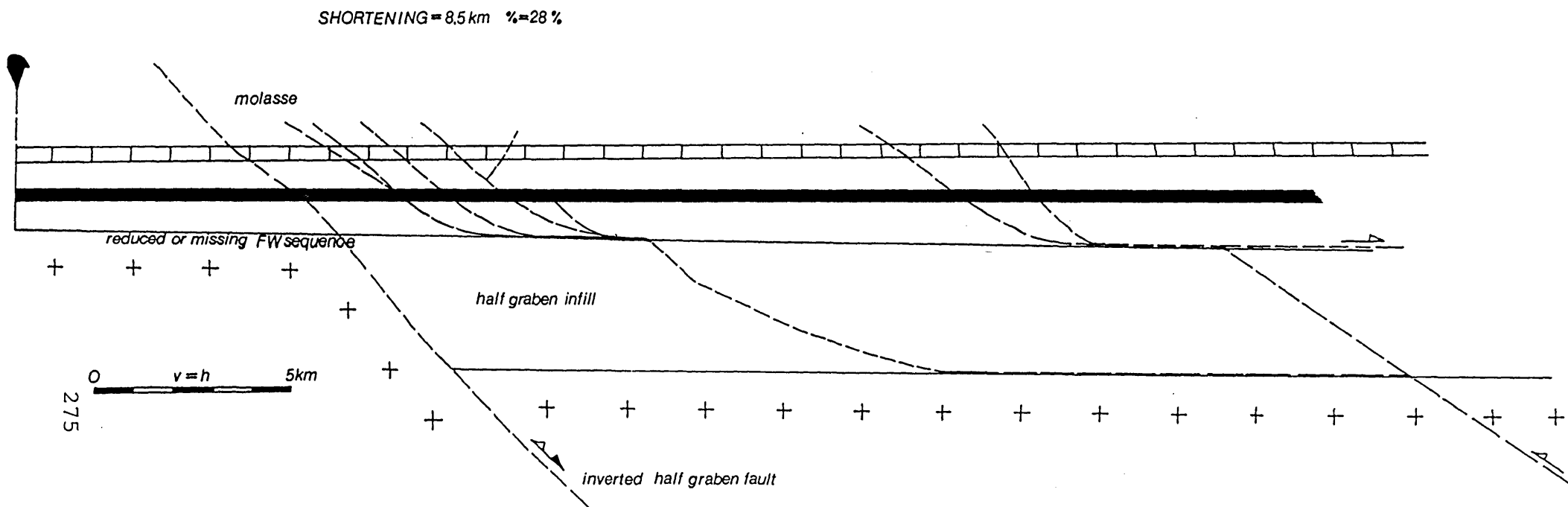


Figure 2.6.6 B: Restored version of the section in Fig. 2.6.6 A across the Vercors. Legend is the same as for Fig. 2.6.6 A.

reverse fault provides another alternative. The relative positioning of the Isere fault, the present mountain front structure and the vertical elevation of the Vercors may not be coincidental. If the Isere fault is interpreted as a pre-existing Jurassic structure, the effect of a reverse reactivation during Alpine compression would be to cancel out a part of the original extension at the level of the basement, and elevate the post-rift sequence of the Vercors plateau. Thus the present mountain front structure would be continuous with the reactivated Isere fault at depth. At the same time the minor thrusting of the Vercors plateau could be attributed to a thin-skinned deformation induced by the emplacement of the external Belledonne basement. Any additional shortening required to explain the present elevation of the Belledonne basement could be accommodated by back-thrusting in the manner described above. This last phenomenon together with isostatic adjustment and uplift of the mountains may well explain the present vertical elevation of the Vercors plateau in the absence of any substantial cover shortening.

2.6.3 Devoluy cross-section

As for the Vercors, two tentative cross-sections are presented for a single section line across the Devoluy Subalpine Chain. The first section (Figure 2.6.8 A & B) examines the evolution of the Devoluy in the context of a thin-skinned thrust model where the cover is decoupled from its underlying basement and the second (Figure 2.6.9 A & B), in the context of a thick-skinned inversion model where the structural development is strongly influenced by pre-existing structure inherited from a Tethyan rifting event. This section line also provides new evidence concerning the timing of the Alpine compression. However this aspect will only be touched upon and is dealt with in more detail in Chapter 5.

The basic structure of the Devoluy area is depicted in Figure 2.6.7. In addition to the main Alpine structure, the area is characterized by pre-Senonian folds which trend E-W

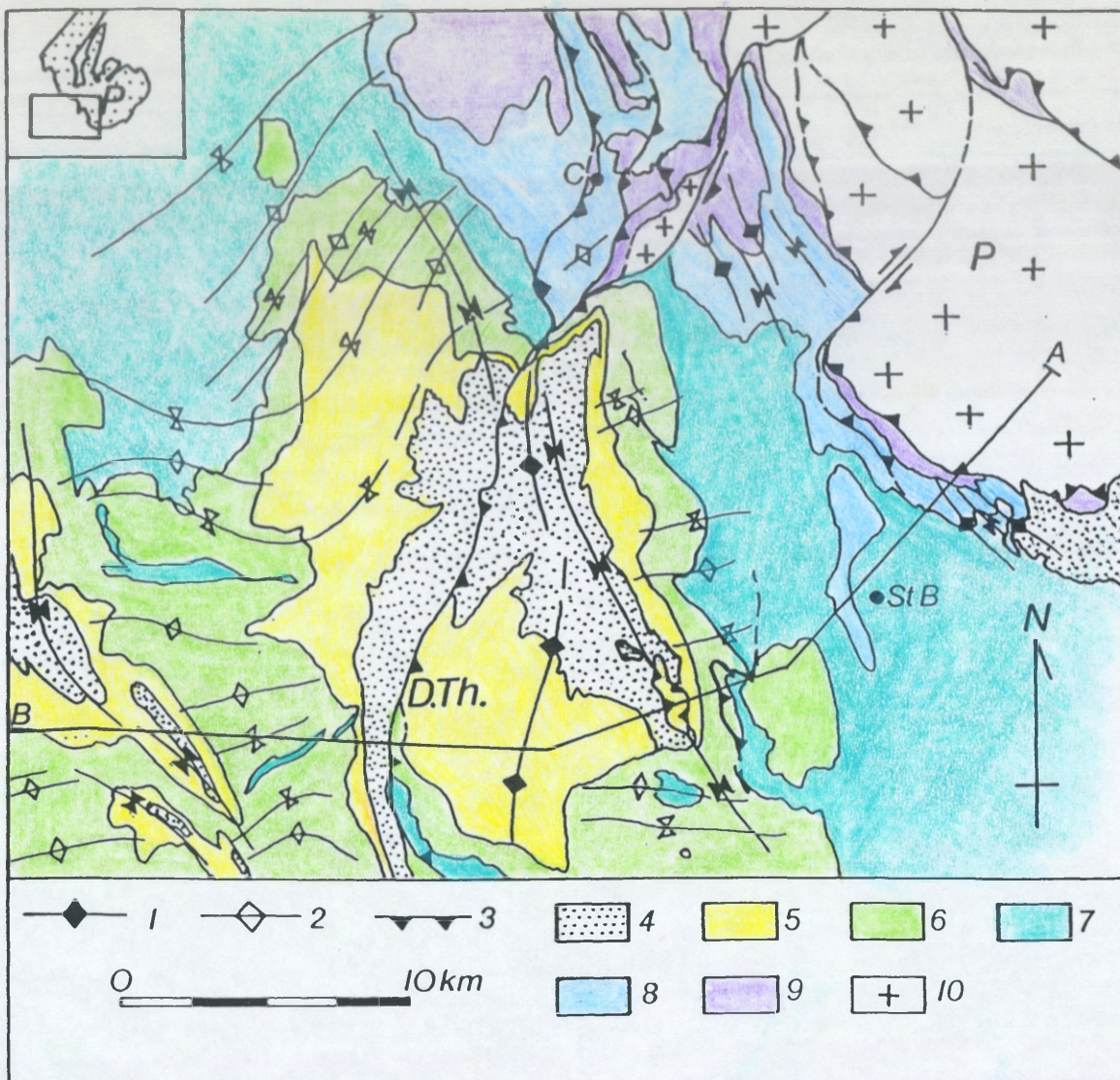


Figure 2.6.7: Simplified structural map of the Devoluy Subalpine Chain and the SW Pelvoux. 1. Oligo-Miocene folds, 2. Pre-Senonian folds, 3. thrusts, 4. Tertiary, 5. Upper Cretaceous 6. Lower Cretaceous 7. Terres Noires, 8. Dogger, 9. Lias & Trias, 10. basement. Adapted from the the BRGM 1:50 000 St Bonnet sheet (1980).

to NE-SW. The age of the most recent Alpine thrusting event is taken to be late Oligocene to early Miocene with the Devoluy thrust overriding the Oligocene Gres de St. Disdier molasse. Evidence for an early Eocene Alpine event also exists. The Oligo-Miocene deformation may be marginally older than thrusting in the frontal parts of the Vercors.

The section location is shown on Figures 2.6.1 & 2.6.7 and is parallel to the designated transport direction for this area: WSW for the SW margin of the Pelvoux, whilst the thrust transport direction for the Devoluy, where only reconnaissance field work has been carried out, is more uncertain.

The eastern, or Pelvoux part of the section was constructed using data collected by the author, and the Devoluy part of the section is taken from the BRGM 1:50 000 St. Bonnet sheet and a surface section by Gidon (1978) on this map sheet. As regards subsurface constraints, the depth to the pre-Triassic crystalline basement was adopted from Menard's (1979; 1980) compilation, although this does not provide information on the geometry of the basement-cover contact.

2.6.3.1 Thin-skinned thrust model:

The principal decoupling horizon for this thin-skinned construction (Figure 2.6.8 A & B) is interpreted to be the basement-cover interface - presumably a Triassic evaporite level, with some local decoupling along the Oxfordian black shales to explain the broad box-type anticline-synclinal pair at the eastern end of the section. For mechanical reasons, the Oxfordian black shales do not generally constitute an effective decoupling horizon in the Devoluy. This is largely because of a pre-Senonian deformation so that during the Oligo-Miocene Alpine compression, hinges of pre-Senonian folds tend to buckle, and the development of long flat thrust detachments are not mechanically favoured. However, should this pre-Senonian event be controlled by a decoupling of the sedimentary

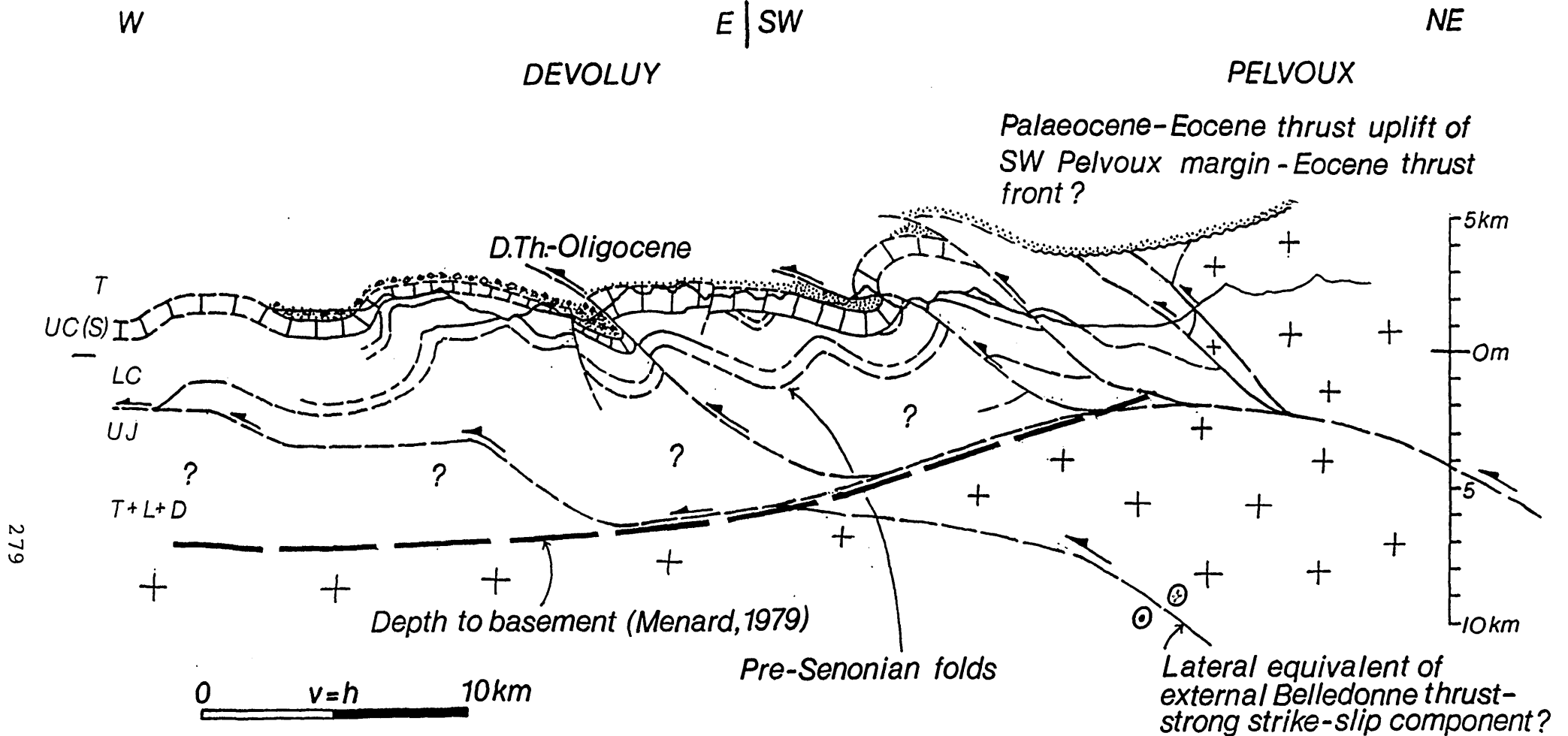
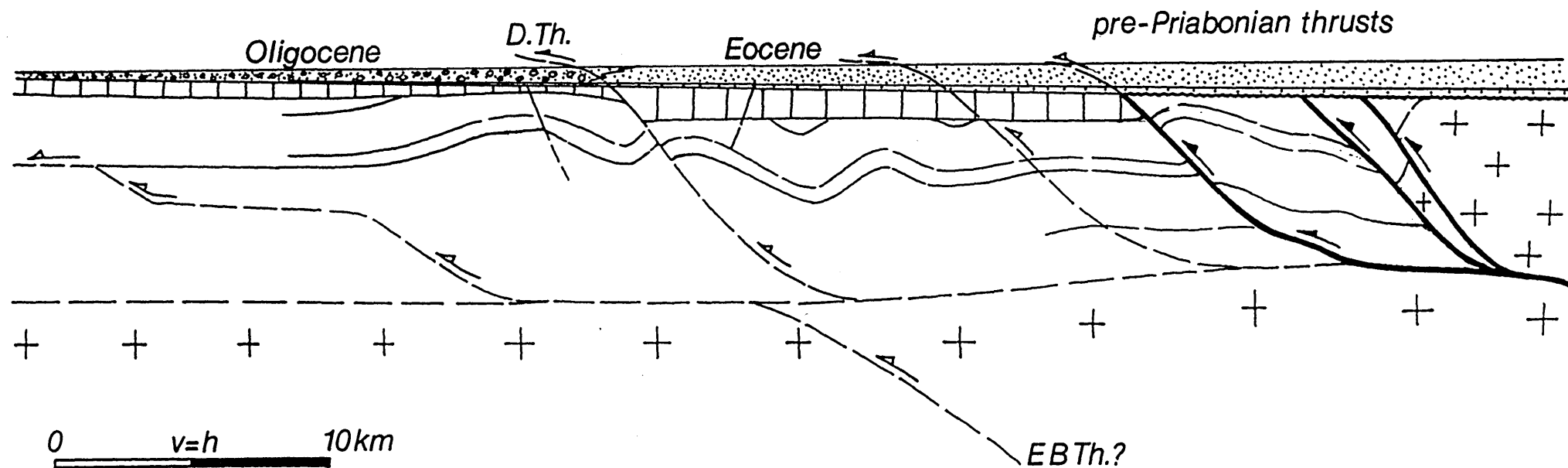


Figure 2.6.8 A: Thin-skinned cross-section across the Devoluy (location shown on Figs. 2.6.1 & 2.6.7). Crosses-basement, T+L+D-Trias, Lias & Dogger, UJ (TN)-Upper Jurassic (Terres Noires), LC-Lower Cretaceous, UC (S)-Upper Cretaceous (Senonian), T-Tertiary. See text for discussion.



NB. Lower part of section not balanced because of pre-Senonian deformation.

Figure 2.6.8 B: Partial restoration of the cross-section in Figure 2.6.8 A, removing the main Oligo-Miocene deformation. The legend is the same.

cover from its underlying basement along Triassic evaporites, this basement-cover contact may still be re-utilized as an effective decoupling interface during any subsequent deformation, even if the compression direction happens to be different.

This pre-Senonian deformation causes innumerable problems when attempting to balance sections parallel to the Oligo-Miocene transport direction. This is because this section line is not parallel to the tectonic transport direction of the pre-Senonian deformation and the lower parts of the section cannot be balanced using a simple two-dimensional model. We have an imperfect knowledge of the stratigraphic thickness, with Jurassic and Cretaceous thickness changes along this section. For the above reasons, only the Oligo-Miocene Alpine deformation is restored. The Senonian limestones have been used as a reference level for this and also aid in the prediction of the subsurface thrust geometry. The lower parts of the section are only area balanced and may have substantial errors.

The main points concerning this section are:

1. Any regional decoupling is along the basement-cover interface, for the mechanical reasons previously stated.
2. The subsurface continuation of the external Belledonne thrust must have some expression in this area. It may be partly responsible for the elevation of the basement-cover interface SW of the Pelvoux, as shown by the map of Menard (1979; 1980). The movement direction associated with this thrust need not be the same as for the higher level structures - towards the WSW, but may be towards the WNW, as in the Vercors. If the external Belledonne thrust continues along the basement-cover interface then the higher level structures will be transported "en mass" in the hanging-wall of this fault towards the WNW, causing additional problems for balanced cross-section constructions. The exact hanging-wall geometry of this

basement thrust naturally remains speculative.

3. Two different ages of Alpine thrusting (additional to the pre-Senonian deformation) are documented along this section line. This comment is also pertinent to the thick-skinned interpretation (Figure 2.6.9 A & B).

i) The timing of the deformation along the SW margin of the Pelvoux is constrained by unconformable relations between the Pelvoux basement and the Eocene to Oligocene Gres du Champsaur flysch (Sections 2.4 & 2.5). From this it can be proposed that the Alpine deformation commenced prior to the deposition of the of the "Calcaire Nummulitique" flysch series (Priabonian). This event, which involved basement, appears to have generated a substantial topographic relief which was subsequently eroded and subsided during the initiation of the Alpine foredeep, to be overlapped by foredeep flysch sediments derived from the growing Alpine mountain belt in the E. The thrust transport direction for this event was towards the WSW to SW (see also Chambers, 1987) making it difficult to reconcile with the N-S pre-Senonian Devoluy compression.

ii) Important uplift of the Pelvoux basement took place much later during the Oligo-Miocene event. The effect of this on the SW margin of the Pelvoux was to inflict a regional SW tilt to the Eocene unconformity (Section 2.5), reflecting deeper basement thrusting. This Oligo-Miocene deformation may have reactivated a pre-existing thrust ramp remaining from the early Eocene phase. These later thrust displacements are interpreted to transfer along the basement-cover interface, eventually emerging to the WSW in an Oligocene to early Miocene foreland basin. Continued uplift followed with the piggyback development of the external Belledonne thrust, which is also interpreted to flatten along the basement-cover contact. A more comprehensive account of the timing will be presented in Chapter 5.

2.6.3.2 Thick-skinned inversion model:

One of the attributes of the thin-skinned thrust model presented above is that it postulates a tentative explanatory relationship between the WNW-directed thrusting of the external Belledonne and Vercors, and the WSW deformation of the southern Pelvoux and higher structural levels of the Devoluy. Unfortunately, the details of the model are not consistent with observations concerning the NE continuation of the Devoluy thrust into the Pelvoux massif along the Aspres-les-Corps fault (Section 2.5).

In section 2.5, evidence is presented supporting Jurassic movements on the Aspres-les-Corps fault and its NE continuation, the Col de la Muzelle fault. The last compressional movement along this fault was a dextral strike-slip movement which is linked with the formation of the Devoluy thrust. The Aspres-les-Corps fault is continuous with the Devoluy thrust (Figure 2.6.7) and it is suggested, acted as preformed lateral ramp. At the level of the Devoluy thrust, and to the NW of the Aspres-les-Corps fault, compression and thrusting clearly dominate. However at Beaufin and Aspres-les-Corps the fault remains in net extension i.e. not all the Jurassic extension is cancelled out by the Oligo-Miocene contraction. Gidon and Pairis (1976) provide independent evidence that the Devoluy/Aspres-les-Corps fault was a site of repeated tectonic activity from the Middle Jurassic onwards. They describe a reduced Senonian sequence (50 m) in the foot-wall to the Devoluy thrust and a corresponding hanging-wall sequence with a thickness exceeding 200 m. They argue that throughout the Senonian the "Devoluy fault" was an active half-graben growth fault (Figure 4, Gidon & Pairis, 1976; see also cross-sections), and further proposed Priabonian age strike-slip movements along the Devoluy fault. These lines of evidence together suggest that the present Devoluy thrust was first initiated as a Jurassic normal fault (the Aspres-les-Corps fault), which was later to accommodate Senonian extension, late Eocene movement, and finally a reversal of this extension during Oligo-

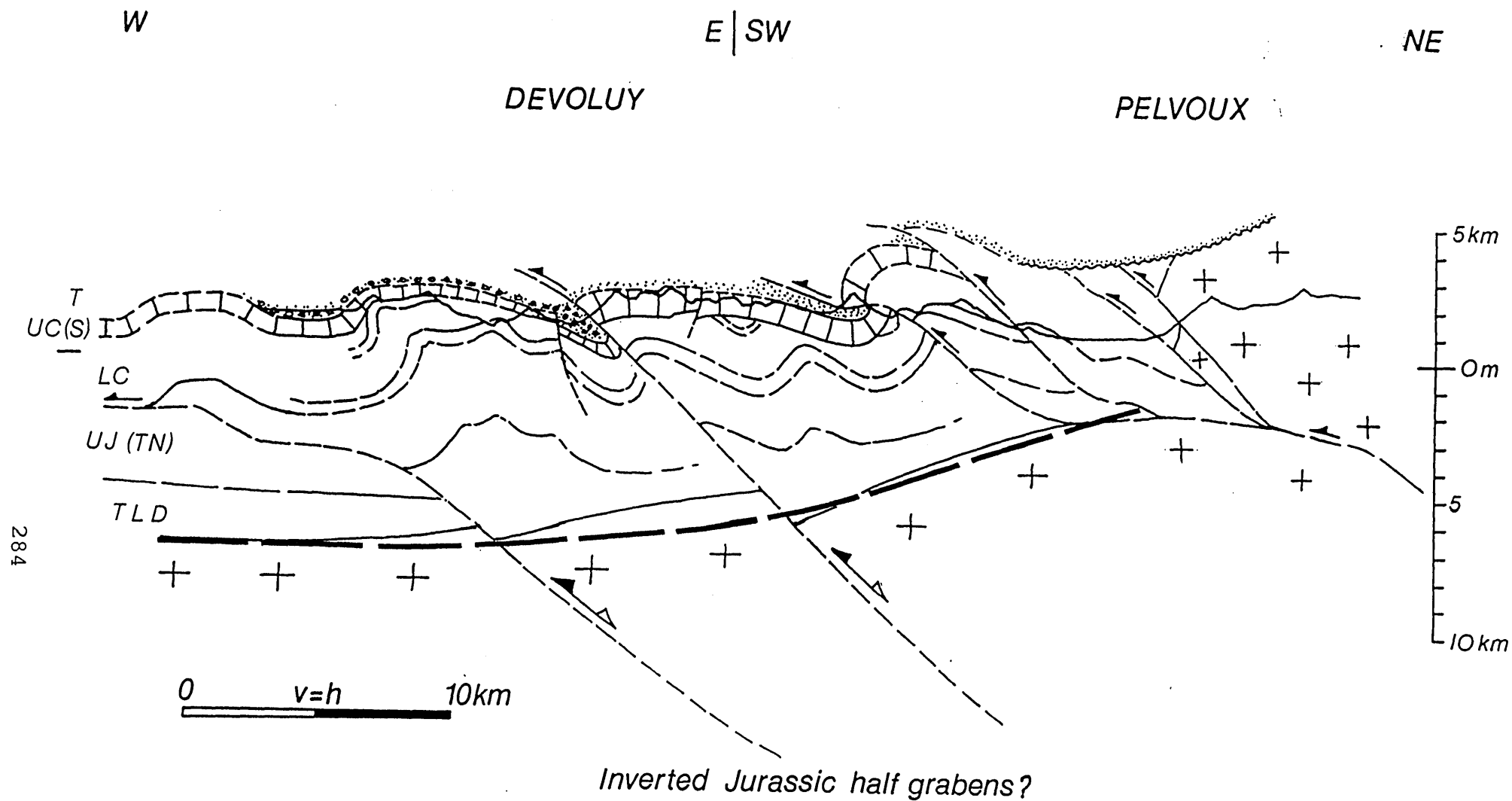


Figure 2.6.9 A: Alternative thick-skinned inversion model for the Devoluy cross-section. Legend is the same as for Fig. 2.6.8 A.

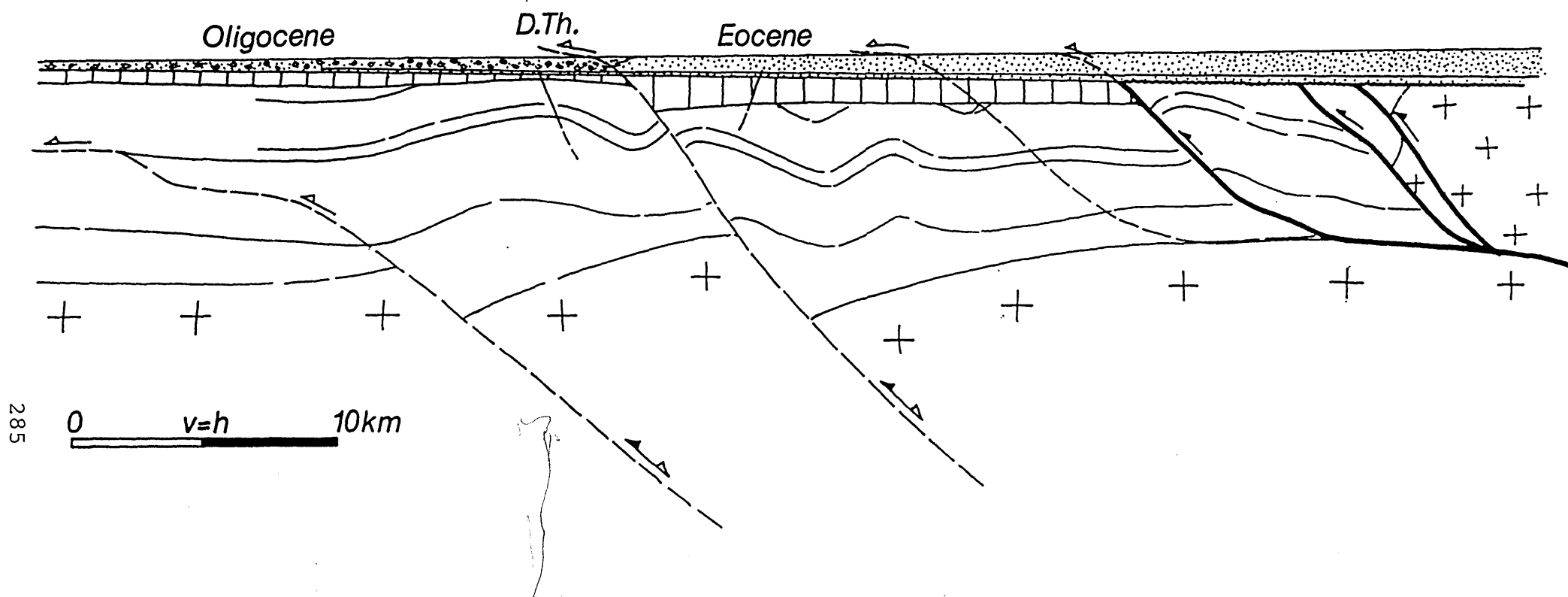


Figure 2.6.9 B: Partial restoration of section 2.6.9 A, removing the effects of the Oligo-Miocene deformation. The legend is the same.

Miocene Alpine compression, resulting in the formation of the Devoluy thrust. Any cross-section constructed through the Devoluy must take this evidence into account, so that the deep geometry of the Devoluy thrust involves basement and may reactivate a pre-existing Jurassic normal fault.

The existence of this inferred Jurassic inversion structure challenges the postulate of a thin-skinned basement-cover detachment under the Devoluy. Two solutions concerning the westward continuation of this section are possible:

1. No thin-skinned detachment of the Devoluy cover from its basement exists, and the cover shortening recorded along this section line reflects the inversion of pre-existing Jurassic or Cretaceous normal faults which offset the basement-cover interface. A shortening of the basement level would cancel any pre-existing extension restoring the basement-cover interface back to something approaching a horizontal surface. The half-graben infills would be thrust out, and the uniform post-rift sequence must shorten to balance the cancelling of the early extension at a deeper level. Contractional displacements from inverted faults in the basement may interact with potential decoupling horizons in the post-rift cover sequence, generating what appear to be thin-skinned fold and thrust structures at a higher structural level (see Chapter 3). However the line lengths from such a section will not balance to a layer-cake stratigraphy. Given the importance of this phenomenon in the adjacent basement massifs and that it is mechanically intuitive that pre-existing crustal fractures should fail under compression should the correct conditions prevail (see Chapter 3), a thick-skinned inversion model may explain the observed cover deformation equally as well as any thin-skinned thrust model.

2. Some thin-skinned thrust detachment may occur at the SW margin of the Pelvoux in response to an Oligo-Miocene thrusting of the Pelvoux basement. This thin-skinned thrusting did not propagate beyond the present position of

the Devoluy thrust, which all the evidence suggests, is a reactivated Jurassic structure. However thin-skinned deformation may still be present to the W of the Devoluy fault in response to the external Belledonne basement thrust, if this continues to exploit the basement-cover interface or a higher level decoupling horizon. As mentioned previously this also accounts for the elevation of the basement SW of the Pelvoux massif.

From regional considerations, it is suggested that this last alternative is the most likely option as the model incorporates both the reactivation of pre-existing Jurassic structures, for which there is evidence and accounts for the lateral continuation of the external Belledonne thrust into this area.

Finally, in view of the fact that a full regional study of the Vercors, Diois and Devoluy Subalpine Chains was not undertaken, some interpretational error may be present in the sections presented. This should not however detract from the elementary problems exposed through the attempted construction of balanced cross-sections in these areas. Not all of these problems encountered can be resolved using a thin-skinned thrust model and the existence of older inherited basement structures may also play a fundamental role. Unfortunately, all too often there is insufficient information concerning the geometry of the deeper structures making it difficult to choose between selected models.

PART II:

Theory and the development of regional models

Chapter 3: Structural inversion

3.1 Introduction

The presence and importance of early, syn-sedimentary faults is now being recognised in many orogenic belts and it has now been suggested that much of the intra-continental deformation within the crust is accommodated by the reactivation of pre-existing structures (Mckenzie, 1972; Sykes, 1978; Jackson, 1980; Sibson, 1985; Etheridge, 1986). This is especially true in regions where compression and uplift affect earlier basins or oceanic margins, leading to a tectonic inversion (Winslow, 1981; Dewey, 1982; Ziegler, 1983; Harding, 1983; 1985; Bally, 1984; Beach, 1985; Etheridge, 1986). Extensional intra-cratonic basins and continental margins seem to be especially prone to failure and reactivation by compressional structures, when the stress system changes from one of dominantly extension to one of compression (Le Pichon & Sibuet, 1981). In middle to upper crustal levels, this failure is often accommodated by a reversal of displacement along the pre-existing extensional faults.

Some of the better documented examples of basin inversion and fault reactivation have been described from commercial seismic data, from areas which are still below sea-level or of only low topographic relief, for example from the North Sea basins (Ziegler, 1983), or South Sumatra basin (Harding, 1983). These do not involve large amounts of crustal shortening. However within collisional mountain belts, where the crustal shortening and uplift are more extreme, the reactivation of a pre-existing extensional fault system and the progressive development of a foreland fold-thrust system has rarely been documented (e.g. see Beach, 1981a; Davies, 1982). This research, carried out in the external French Alps, emphasises the importance of inversion and the reactivation of a pre-existing extensional fault system in controlling the fundamental structural style of the mountain belt.

3.2 Types of inversion

The term inversion is restricted to describe regions which have experienced a reversal in uplift or subsidence rates, that is areas which have changed from being ones of subsidence to ones of uplift or vice versa. Following Harding (1985), an area which has changed from uplift to subsidence can be considered to have suffered a positive inversion, and an area which has changed from uplift to subsidence has suffered a negative inversion. In the Alpine thrust belt we are mainly concerned with positive inversion, and in particular that related to the compression and uplift of sedimentary basins and the reversal of movement on normal faults. Gillcrist et al. (1987) proposed a basic subdivision of inversion into weakly and strongly inverted basins (Figure 3.1). In this work, I restrict myself to a consideration of basin inversion in strongly inverted regions, such as occurs in the French Alps. For further reference to, and examples of, weakly inverted basins, the reader is referred to Gillcrist et al. (1987) and the works referenced therein. There are several other mechanisms of inversion which are discussed briefly below.

1. Uplift due to flexure:

Uplift can be generated by the flexural bending of a lithospheric plate as a result of localised loading by sedimentation (Beaumont, 1978) or by the accumulation of a thrust mass (Beaumont, 1981; Karner & Watts, 1983; Royden & Karner, 1984). Such flexural basins have a slight peripheral bulge at their margin, where there may be localized erosion if lifted above sea-level (Figure 3.2 a).

2. Uplift along low-angle strike-slip faults:

Changes in the strike of a fault, or offsets in the fault may cause the local development of basins or zones of uplift (Crowell, 1974; Reading, 1980). This is attributed

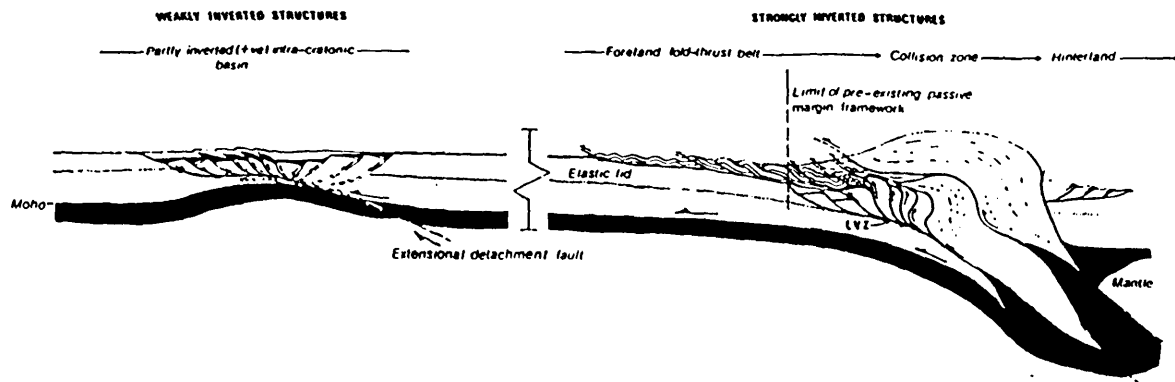


Figure 3.1: A schematic cross-section through weakly and strongly inverted basins, the strongly inverted basins lying near the original plate margin.

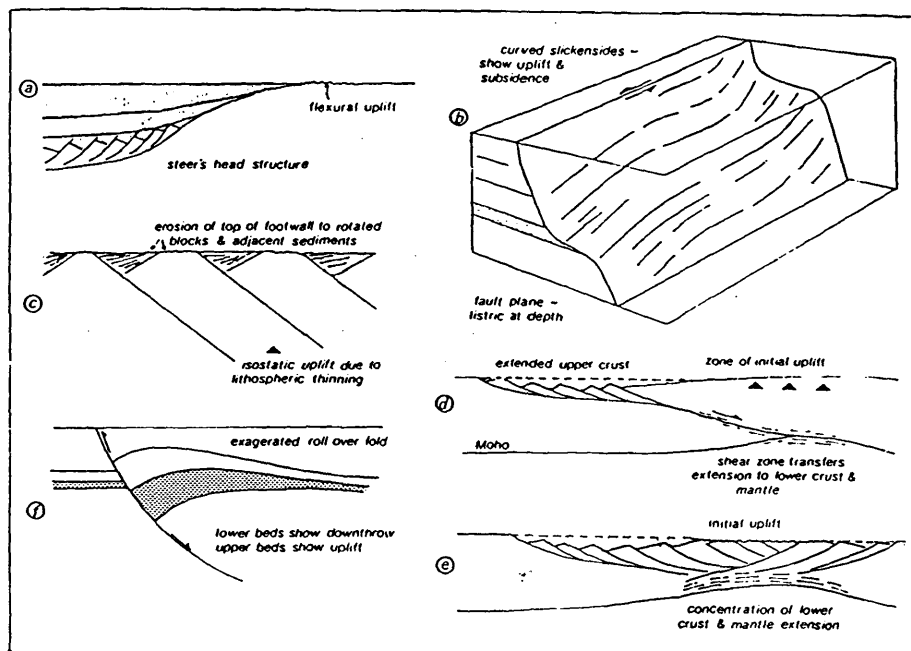


Figure 3.2: Some of the main uplift mechanisms for sedimentary basins: a) flexural uplift at the margin of a subsidence basin, b) localized uplift and subsidence along a listric strike-slip fault, whose axis of curvature is not horizontal, c) foot-wall uplift of rotated blocks, d) isostatic uplift from thinning of the dense lower crust and mantle, leaving the less dense upper crust intact, following the model of Wernicke (1981; 1985), e) heterogeneous thinning of the crust leading to uplift above a concentrated zone of lower crust and mantle, f) uplift by the compressive reactivation of earlier faults.

to variations in the sense of curvature or offset along one wall of the fault, or the existence of ramp-flat structures which are not purely concentric in three-dimensions (Figure 3.2 b). Continued strike-slip movements may cause existing basins to invert, and the topographically higher zones to become depressed.

3. Isostatic uplift of sedimentary basins:

Jackson & McKenzie (1983) argue that the localised foot-wall uplift of a large normal fault, is due to isostatic effects arising from the removal of denser material from the hanging-wall (Figure 3.2 c). Larger scale examples of isostatic uplift may occur from the heterogeneous thinning of the crust and lithospheric mantle. The McKenzie model of lithospheric thinning (McKenzie, 1978; Le Pichon et al., 1982) assumes no uplift during the initial stretching and following thermal subsidence phases, whereas the Wernicke model of crustal extension on a low-angle shear zone (Wernicke, 1981; 1985) assumes an initial subsidence beneath thinned upper crust, offset from a zone of initial uplift above a thinned mantle (Figure 3.2 d).

These two models are two extreme end members of a general model of heterogeneous lithospheric stretching, where the stretching factor (β) may vary along any vertical section through the lithosphere (Coward, 1986). Thus regions which show subsidence during an early phase of stretching, may be subsequently uplifted if the later stretching is locally confined to the lower crust and lithospheric mantle (Figure 3.2 e). This is similar to the depth dependant model of stretching (Beaumont et al., 1982) which predicts uplift as a result of increased extension in the lower lithosphere at the edge of a basin, with respect to that in the upper lithosphere. Some of the phases of uplift, evidenced in the now telescoped Tethyan passive margin stratigraphy, may simply reflect localised isostatic uplift which follows such a heterogeneous crustal stretching (see Chapter 4, and Gillcrist et al., 1987).

4. From slight changes in plate vectors:

Strike-slip belts which bound sedimentary basins or act as compartmental faults to heterogeneous extension across the basin, may become zones of transpression and uplift as the plate movements change, due to the complex kinematics of plate motions on a spherical Earth (Dewey, 1975).

5. Major stress reversals in the earth's crust:

Many inverted basins develop from major changes in stress magnitudes, so that directions of extension reverse to become directions of compression. The Alpine system and its NW foreland show excellent examples of this. From the Bathonian to the late Cretaceous, extension in the mid-Atlantic led to phases of NW-SE extension across the North Atlantic province and Europe. However during the late Cretaceous, opening of the Labrador sea produced a change in relative plate motion between Europe and Africa, with the result that NW Europe was placed under compression. During the opening of the North Atlantic from the earliest Eocene, the relative plate motions were such that Europe and Africa were convergent and Europe was affected by a NW-SE compressive stresses, a situation which has continued until this present day, as shown by *insitu* stress measurements (Griener, 1975; Griener & Lohr, 1979; Froidevaux et al., 1979; Baumann, 1981) and recent seismic activity (Dunning, 1977; Pavoni, 1980; Dorel et al., 1983). This is the main mechanism of structural inversion that we observe during the formation of the Alpine thrust belt.

3.3 Controls on the geometry of inversion

The Alpine age inversion structures documented in part I of the thesis show that a wide range of geometries are associated with fault reactivation, and not just simple uplift along the earlier normal faults. This section discusses the origins of some of these geometries and more generally, why particular basins might invert and others

remain inert.

1. Capability of faults to reactivate:

In considering the reactivation of earlier normal faults in the upper brittle levels of the earth's crust during positive inversion, we can use the Navier-Coulomb criteria for brittle failure, effectively represented by a Mohr diagram as in Figure 3.3 a. In the compressive field, the failure criterion can be expressed by the linear relationship:

$$\tau = C_0 + \mu \sigma_n \quad (18)$$

where τ is the shear strength of the rock, σ_n the normal stress acting on the fault plane, C_0 the cohesion of the fault and μ the coefficient of internal friction ($\mu = \tan \phi$). If a fault plane already exists in the rock, then the cohesion on this plane may be reduced to zero and the coefficient of internal friction will change:

$$\tau' = \mu' \sigma_n \quad (19)$$

where μ' represents the coefficient of frictional sliding on the plane. From the Mohr diagram (Figure 3.3 a), a fracture plane oriented between β_1 and β_2 will be reactivated before a new fracture develops in the originally intact rock. As the cohesion and sliding friction decrease, then the range orientation ($\beta_1 - \beta_2$) of potential fault reactivation increases. Similarly for a fixed confining stress (σ_3), C_0 and μ , then the necessary stress difference ($\sigma_1 - \sigma_3$) varies with β :

$$\sigma_1 - \sigma_3 = \frac{2 C_0 + 2 \mu \sigma_3}{(1 - \mu \cos \beta) \sin 2 \beta} \quad (20)$$

where β is the angle between the normal to the original fault plane and the maximum compressive stress (σ_1) (from Jaeger & Cook, 1979). The fault planes which will fail first will be those whose orientation is closest to the

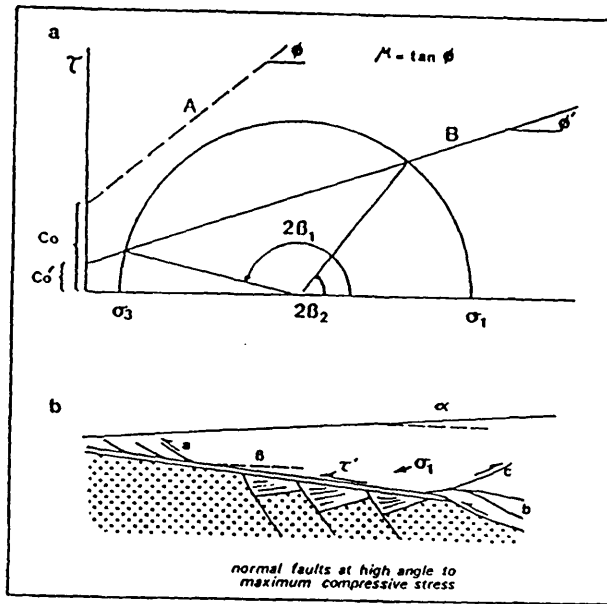


Figure 3.3: a) The Mohr circle and Mohr envelope for brittle failure, assuming Coulomb failure criteria. A - the envelope for wall rock, with a cohesion of c_0 and angle of internal friction ϕ , B - the failure envelope for a pre-existing fracture, with lower cohesion c_0' and angle of sliding friction ϕ' . See text for discussion.

b) A typical thrust wedge, with basal slope β , and surface slope α . See text for discussion.

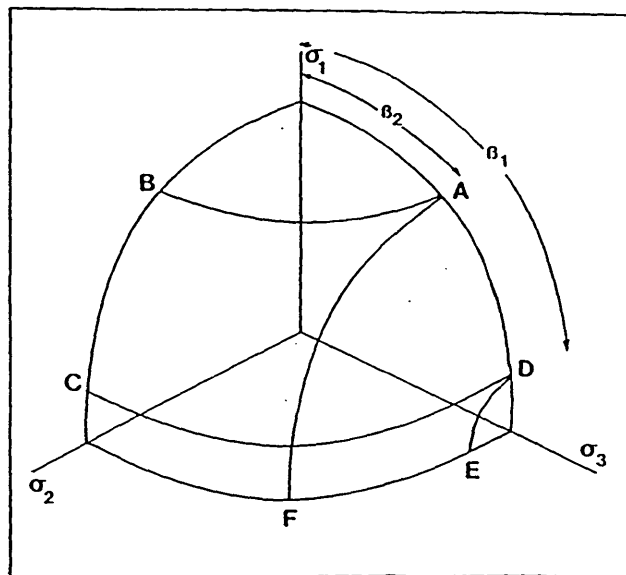


Figure 3.4: Conditions for failure in three dimensions - see text for discussion (after Jaeger & Cook, 1979).

failure plane for isotropic rock with the same mechanical properties as the fault plane. If no planes have the right orientation, the stress difference ($\sigma_1 - \sigma_3$) would need to reach a value:

$$\frac{C_0 + \sigma_3 \tan \phi}{\sin \phi - \tan \phi + \tan \phi \cos \phi} \quad (21)$$

for which a new fracture plane would develop in the originally intact rock.

It will be normal faults which had been rotated to gentle dips during extension which will fail preferentially during inversion. For a fault whose dip changed abruptly, the flat part could fail during inversion, while other fold and fault structures are created at the foot of the ramp. Assuming horizontal compression, then normal faults with 60° dips will only fail if the coefficient of sliding friction is low. During the development of a thrust wedge, the maximum compressive stress will be inclined to the horizontal (Chapple, 1978), (Figure 3.3 b), and hence will be at an even greater angle to the steeply dipping normal faults.

Sibson (1985) concluded that sliding on pre-existing faults was facilitated by the presence of abnormally high pore fluid pressures, reducing the effective stress ($\sigma_n - P$ fluid) towards zero. As was demonstrated by the Denver and Rangely earthquake studies (Evans, 1966; Raleigh et al., 1979), artificially elevated pore fluid pressures causes the reactivation of faults. However, in some cases the cohesive strength of the rock may exceed that of the surrounding intact rock. Surprisingly little work has been done on attempting to quantify the material strength properties of fault gouges. Donath & Cranwell (1981) propose that in the brittle regime, any pre-existing fault is likely to have a lower cohesive strength than the surrounding intact rock. However, experimental work by Logan et al. (1981) showed that montmorillonite free gouge zones were capable of sustaining high shear stresses and

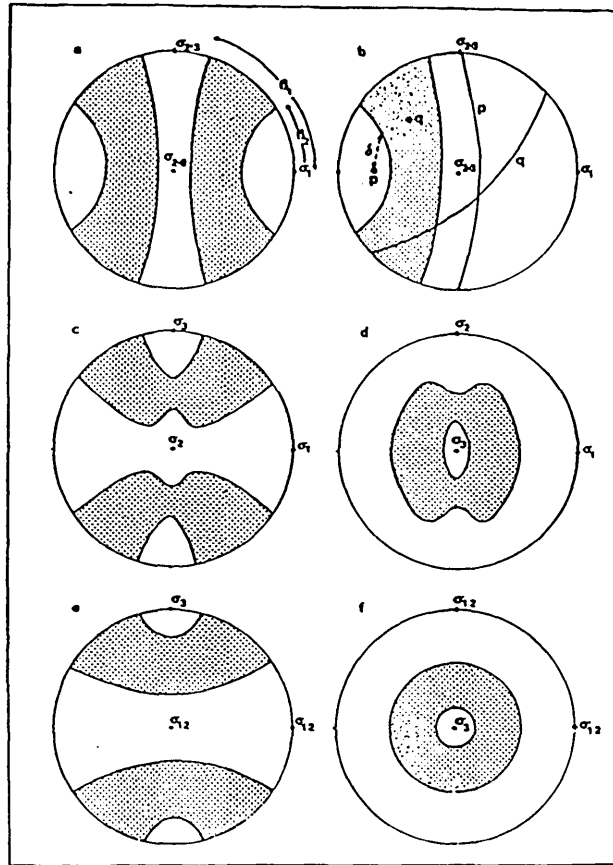


Figure 3.5: Stereographic projections to show regions where poles to pre-existing fractures must lie (shaded) for failure to take place. a) for stress state $\sigma_2 = \sigma_3$, b) for the same stress state as a, but showing how faults or fractures may be reactivated if the maximum compressive stress is oblique (by angle δ) to the plane, c) for stress state $\sigma_1 > \sigma_2 > \sigma_3$ (σ_3 horizontal), d) for stress state $\sigma_1 > \sigma_2 > \sigma_3$ (σ_3 vertical), e) for stress state $\sigma_1 = \sigma_2$ (σ_3 horizontal), f) for stress state $\sigma_1 = \sigma_2$ (σ_3 vertical). After Jaeger & Cook (1979).

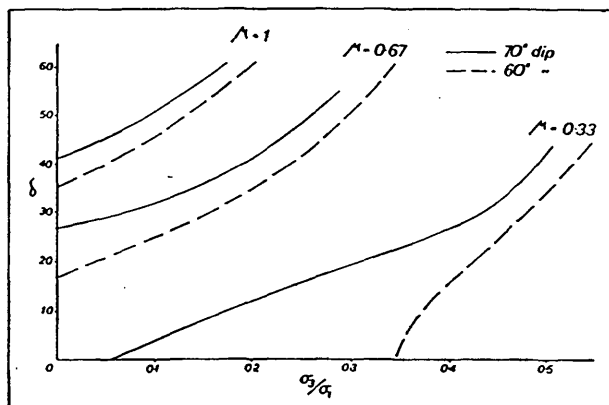


Figure 3.6: Graph showing how the angle δ must vary for different dips of the pre-existing fault, different coefficients of sliding friction and different stress ratios σ_1 / σ_3 . The stress state is of $\sigma_1 > \sigma_2 = \sigma_3$.

concluded that it was not only the earlier fault plane which failed during later deformation. Fluids along the fault plane may result in mobility and deposition of quartz and/or calcite in the dilational fractures, especially if the active fault zone is extensional. This may have a similar strain hardening effect (Etheridge, 1986).

The above discussion has considered failure in two dimensions, where the fault planes are considered to be parallel to the intermediate compressive stress. In 3-D the fault planes which are not parallel to the intermediate stress may also satisfy the failure criterion; fault reactivation is partly dependant on the magnitude of the intermediate compressive stress. Jaeger & Cook (1979) consider two extreme 3-D stress configurations, where:

$$\sigma_1 = \sigma_2 > \sigma_3 \quad \&, \quad \sigma_1 > \sigma_2 = \sigma_3 \quad (22) \quad \& \quad (23)$$

Thus in Figure 3.4, the regions in which a normal fault plane must fall, for sliding to occur, will be ABCD for $\sigma_1 > \sigma_2 = \sigma_3$, and ADEF for $\sigma_1 = \sigma_2 > \sigma_3$. These are represented in stereographic form in Figure 3.5.

Thus a fault plane dipping at 60° or more, may not be reactivated by horizontal compression, if the maximum compressive stress is perpendicular to the fault plane, but if it is oblique, by an angle δ then failure can occur. The angle δ depends on the type of 3-D stress (that is the $\sigma_1 : \sigma_2 : \sigma_3$ ratio), the coefficient of sliding friction, the stress difference ($\sigma_1 - \sigma_3$) as well as the dip of the fault. Figure 3.6 shows the relationship between the necessary angle δ for failure and the stress ratio σ_1 / σ_3 , for one particular stress configuration.

Steeply dipping faults, therefore, may be reactivated by oblique slip movements. In the more extreme cases of inversion, as occurs in the French Alps, where the normal faults had strikes perpendicular to the Alpine compression (see Chapter 4), steeply dipping faults such as the Ornon fault and Pied Moutet faults (Chapter 2) blocked Alpine

movements and developed tight fold and thrust systems on their hanging-walls.

The siting and geometry of high level brittle fault reactivation may be controlled by movement on ductile faults deeper in the crust. The mechanical properties of deeper mylonite zones are controlled by a series of creep flow laws (White, 1976; Etheridge, 1986) and fine grained mylonites are inherently weaker than the surrounding intact rock. Deep seismic studies, such as across the British offshore basins, shows that early fault zones are preferentially reworked at depth (Cheadle et al., 1987).

Major transfer or transform faults may control the orientation of later inversion structures, by concentrating the thrust movements in a particular direction. Less work will be involved if crustal shortening occurred parallel to the trends of large early faults rather than across them. This may account for the apparent parallelism of fault movements in NW Europe from the Variscan to the Alpine.

2. Strength of the crust:

A possible strength profile through the crust and upper mantle is shown in Figure 3.7 b, assuming simple rheologies dependant on particular minerals, such as wet quartz in the crust and olivine in the mantle. The upper part of the crust will behave in a brittle fashion, the strength increasing with confining pressure and depth. In the lower part of the crust, however, ductile deformation will dominate, the ductile transition occurring at depths between 15-20 km. If more realistic crustal compositions and variations are assumed, where the local strength may be controlled by a variety of minerals: dry quartz, feldspar, pyroxenes, then the shape of the strength/depth curve becomes more complex. Low strength regions occur at the base of the crust, due to the contrast between quartz and olivine rheology and other low strength regions are predicted in the lower crust, assuming major compositional

and lithological boundaries (Kuznir & Park, 1987). The lithosphere strength is critically controlled by the crustal thickness and the rate of or time since, stretching. If the lithosphere is stretched rapidly and homogenously (see McKenzie model, 1978), the strength curve will be modified as shown in Figure 3.7 d, the total strength of the stretched lithosphere being less than that of the unstretched lithosphere. With time, the lithosphere should cool back to its original geothermal gradient and the strength curve will modify as shown in Figure 3.7 e, the total strength of the stretched lithosphere now being greater than that of the unstretched lithosphere. Similarly a slow extension rate will lead to a strengthening of the lithosphere (Kuznir & Park, 1987).

If extension follows the Wernicke model (Wernicke, 1981; 1985), where the stretching in the upper crust is transferred laterally to the lower crust and mantle, then there will be relatively little effect on the crustal strength curve beneath the faulted basin (Figure 3.7 f and g). In the region where the lithospheric mantle is thinned, there will be an increase in the geothermal gradient and hence a decrease in strength, immediately following stretching, but with time this will revert to almost its normal strength profile.

The crustal anisotropy and decrease in grain size on deep level shear zones may aid later reactivation, as has been discussed above. Thus early mylonites may reactivate as the deep floor thrust to a collisional system. However, if the mylonites were annealed or uplifted to shallow crustal levels, they could become stronger than the surrounding rock.

Thus if compression follows closely after a phase of extension, the lithosphere beneath basins formed by homogeneous extension of crust and mantle will be weaker than that beneath basins formed by heterogeneous extension. Their strength will depend upon the original heat flow, the amount of stretching, the rate of stretching and the

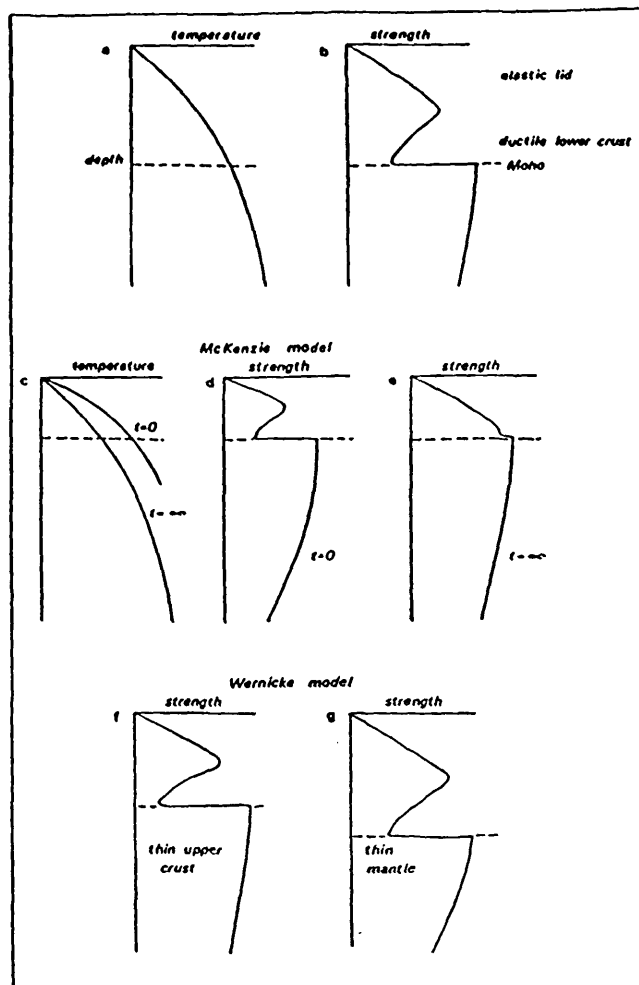


Figure 3.7: Simplified temperature-depth and strength-depth profiles through the upper lithosphere for the initial state. (a and b) for the McKenzie model of homogeneous stretching (c, d and e, for initial subsidence, time $t=0$ and after thermal equilibration at time $t=\infty$), for the Wernicke model of heterogeneous stretching at time $t=0$ (f) and $t=\infty$ (g).

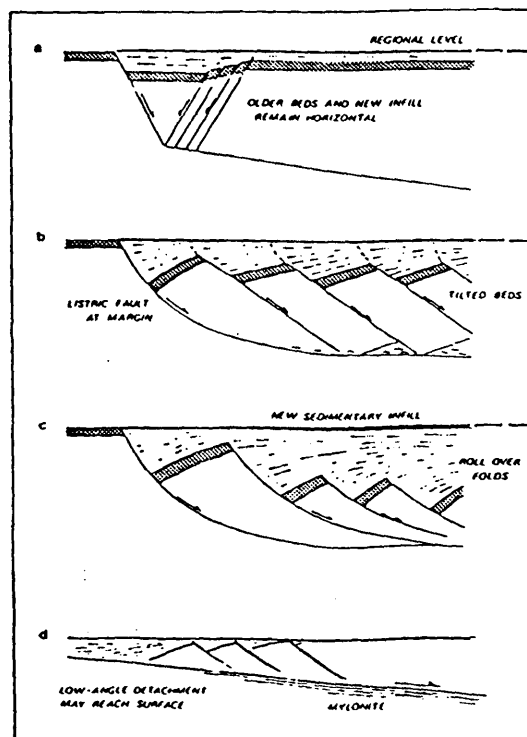


Figure 3.8: The main types of normal faults - see text for discussion.

time elapsed since stretching. Thus some basins may invert early in the compressional history, but inversion may cease as the basin thickens and the lithosphere cools. The basins may invert sequentially depending on how their strength increases with time, until all become as strong as the adjacent unstretched lithosphere.

3. Geometry of the original normal faults:

Some of the structures developed during inversion result from the accommodation of extra bedlength during thrust reactivation of the normal faults, while others from the blocking of thrust movements. Figure 3.8 summarizes the main types of extensional fault systems. The faults may be straight at typical Andersonian (Anderson, 1951) angles of dip and the beds and fault planes may suffer no rotation during extension (Figure 3.8 a). The faults obviously have to change shape or decouple at depth, generally by slip along a more ductile bedding plane. The beds drop down as a triangular keystone block and as extension continues, numerous antithetic and synthetic faults develop to fill in the extra space. Jackson (1987) describes such faults using seismic data from the Aegean.

During inversion it may be possible to reactivate the the low-angle detachment and the steeper faults may be squeezed up, depending on their orientation relative to the maximum compression direction and the frictional properties of the faults. Pop-ups and box-shaped forced folds may be produced in the cover.

A second type of extensional fault system produces planar faults above a decoupling zone and the faults and intervening blocks rotate to very low dips during extension, similar to a collapsing pack of dominoes (Figure 3.8 b). During inversion the rigid fault blocks may either slide over each other as individual thrusts, or all slide together with a back-rotation of the rigid blocks and associated steepening of dips, decreasing the potential for the failure of individual fault planes. Gently dipping

faults produced by a high degree of crustal extension, may fail as individual thrusts, stacking up the rigid blocks of basement in a foot-wall propagating sequence (Figure 3.9). Here the amount of crustal shortening may exceed that of the initial extension value and the inverted structures may be difficult to distinguish from a thrust stack affecting previously undeformed foreland, unless any syn-tectonic sedimentary breccias or sedimentary onlaps and unconformities can be distinguished.

During back-rotation, the sediments in the half-grabens will be uplifted and shortened and they will be either folded or thrust out of the half-grabens (Figure 3.10). An imbricate stack may be produced, where the thrusts shallow upwards to join a roof thrust, but steepen downwards in to the earlier normal faults. The resulting sections may appear unbalanced, if a layer cake stratigraphy were to be assumed. High level beds may be considerably shortened by imbrication or folding, while lower beds appear relatively undeformed (Figure 3.10). Jackson (1980) invoked this mechanism of inversion to explain the deep structure of the Zagros region of Iran. Here the cover sequence shows simple folding while the seismic activity shows activity on steeply dipping faults at depth. Indeed the whole of the Arabian foreland shows evidence of structural inversion and fault reactivation, caused by late Miocene continental collision in the Taurus overthrust belt to the N (Lovelock, 1984).

Listric extensional faults (Figure 3.12 a) curve to flatten downwards and produce a roll-over anticline on their hanging-walls (Gibbs, 1984). Inversion may initially accentuate the roll-over fold, but eventually will push out the sedimentary wedge on a steep, shallowing upwards thrust (Figure 3.12 b and c). The reactivation may be more intense on the the gentler dipping parts of the listric fault, displacement decreasing towards the steeper dipping parts of the fault. Where inversion is weak, a gentle compressional anticline will form. Where reactivation is more intense, continued shortening may be accommodated by

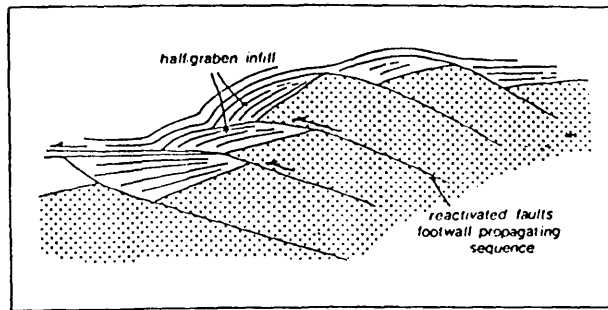


Figure 3.9: Inversion of the rotated domino type faults by foot-wall propagation of thrust reacti vation.

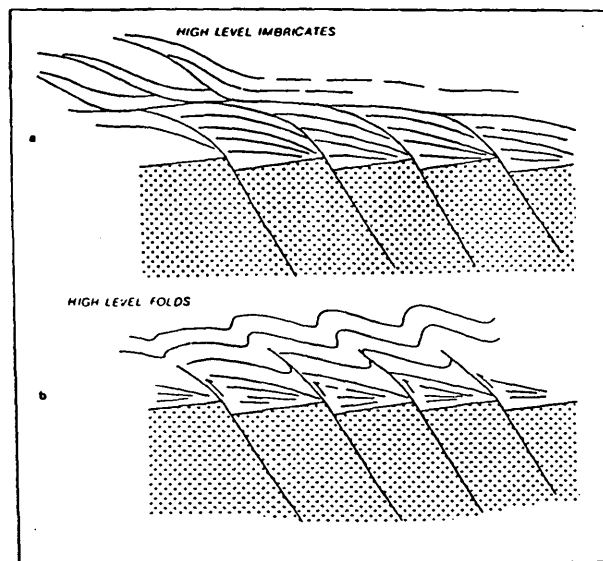


Figure 3.10: Inversion of the rotated domino faults by back-rotation. Note the extra apparent shortening in the cover, by thrust imbrication (a) or folding (b).

a pure shear straining of the hanging-wall of the extensional fault, leading to the exaggeration of the roll-over anticline, and upright folding with steep penetrative cleavage development in the syn-rift sediments adjacent to the fault scarp. Variations in sedimentary thickness, attributed to the differential growth of a normal fault, affect the final inversion geometry and the amount of stratigraphic separation. Reverse stratigraphic separation occurs at higher structural levels, while the lower levels maintain the normal fault offsets (Figure 3.2 f).

4. Amount of shortening relative to original extension:

As mentioned earlier, on a broad scale it is possible to distinguish between weakly and strongly inverted basins (Figure 3.1), where the strongly inverted basins may ultimately develop into a mountain belt and generally, but not always, occur at the sites of continental collision. The French Alps provide excellent examples of strong inversion associated with collision, whilst the Pyrennees have formed a thickened crust and mountain belt from an intra-cratonic basin (Souquet et al., 1980; Bally, 1984).

During plate collision the thinned continental margin will be the first to deform, the thrusts and folds gradually developing away from the plate margin (Dewey et al., 1986). The type of structures which develop will depend on the time elapsed since extension. Old passive margins will have a strong crust and the deformation will propagate into the underlying plate by gradual accretion of thrust slabs to the underthrusting plate. The unfaulted sedimentary carapace, developed during the thermal subsidence of the passive margin will be accreted to the overthrust slab as a thin-skinned fold and thrust belt. The lower syn-rifting sediments will become involved in the thrust tectonics, the original normal faults acting as stress risers and probably controlling the position of the imbricate thrust ramps. The thinned basement will deform by reactivating the older extensional faults or by developing

new fault surfaces, depending on the orientation of the faults relative to the maximum compression direction.

During the deformation of a younger passive margin, however, there may be a different sequence of structures. The thinned crust would still be weak and could fail by reactivating the extensional faults early in the collision history, before the thin-skinned deformation of the sedimentary carapace.

The French Alps show both aspects of continental margin inversion. The ages and the different types of inversion are discussed in Chapter 5.

5. Linkage of inversion tectonics:

Just as extensional structures need to be linked in 3-D, so do compressional structures and tectonic inversion. Thus inversion cannot die out laterally without developing local strains. Whilst the inversion structures and displacements must link within the Alpine thrust belt, so they must link on a larger cratonic scale such that the tectonic inversions which occurred in the N Sea and the English Channel must link with deformation and displacement occurring in the Alpine system. Similarly the tectonic inversions occurring in the Palmyras on the southern Arabian margin, (Lovelock, 1984) must link with the crustal deformation and displacement associated with continental collision to the N.

Bally (1984) and Ziegler (1982) have suggested that the displacements link at depth by detachments at the base of the crust, or within the crust (see also Figure 3.1). Presumably the depth of the detachment zone depends on crustal thickness and geothermal gradient; some tectonic inversion may be linked by detachment in the lower lithosphere. Old basins, which are stronger than the adjacent crust, may form buttresses to any lower crustal or mid-lithospheric detachments, localising the compressional deformation to the adjacent previously undeformed plate.

3.4 Summary of inversion geometries from the Pelvoux:

The inversion geometries encountered in the Pelvoux and adjacent basement massifs, described in Chapter 2, are summarized in Figures 3.11 a and b.

As should be apparent from Chapter 2, the occurrence of tight asymmetric hanging-wall folds deforming basement, and related to the reactivation of pre-existing extensional faults, are common in the studied area. In some cases, at the pre-rift and basement stratigraphic level, the inverted extensional faults still remain in net extension after compression e.g. the La Coche anticline (Section 2.3). If the deformation is intense, or the amount of shortening to be accommodated on the old extensional fault system is large, there may be failure of either the hanging-wall or foot-wall rock, leading to the development of foot-wall short cut geometries (Knipe, 1985), (Figure 3.11 a (1)), or hanging-wall break-back thrust sequences which slice through the inversion anticline e.g. Signal de Lauvitel, Section 2.1 and 2.3 (see also Figure 3.12 a (2), and Gillcrist et al., 1987, for discussion concerning the mechanics).

Where secondary antithetic or synthetic extensional faults affect the hanging-wall or foot-wall basement of a pre-existing extensional fault, controlling the sedimentary prism geometry within the larger half-graben, after contraction and inversion a number of characteristic geometries will result (Figure 3.11 a). Antithetic faults will reactivate to produce basement "pop-up" structures (Figure 3.11 a (3)). Alternatively, thrust reactivation of minor synthetic faults which transfer contractional displacements into the confines of the larger half-graben sedimentary prism, will result in complex cover deformation patterns (see Bourg d'Oisans example, Section 2.1). Such a system of listric extensional imbricates restacked during compression, will produce a finite structure resembling that of a compressional thrust duplex as the contractional displacements transfer onto the base of the post-rift

sedimentary succession (Figure 3.11 a (4.ii)). Unless any syn-sedimentary fault breccias or other features indicative of active extensional tectonics during sedimentation can be defined, the final contractional geometry may be difficult to distinguish from that of a thin-skinned thrust stack involving basement and a non-tectonised foreland shelf sequence.

Other typical inversion geometries reflect the initial geometry of extension prior to contraction. Examples exist where the hanging-wall of the old extensional fault is completely thrust out over foot-wall sediments which are tightly "pinched in" beneath the basement thrust fault e.g. Les Deux Alpes, Section 2.2 (Figure 3.11 a (5)). The original extension on the old normal fault is completely cancelled during inversion. Where this compressional geometry occurs, it is suggested that the extensional fault possessed an initial steep dip and was planar, such that during compression a high angle reverse fault developed. Alternatively, contractional displacements from an inverted extensional fault may transfer along the pre-rift/syn-rift stratigraphic boundary, so that the hanging-wall basement of the old extensional fault and its half-graben infill are passively carried along a foot-wall flat which exploits the pre-rift/syn-rift contact (Figure 3.11 a (6)). The final geometry of the contractional structure resembles that of a basement thrust ramp anticline overlying a foot-wall flat in a simple thin-skinned thrust system which deforms a layer cake stratigraphy. However the anomalous stratigraphic cut-offs of the half-graben infill against the fault plane, its internal deformation and the overturned hanging-wall basement fold reveal its otherwise hidden inversion history. One such example, the La Paletas half graben, is presented from the Bourg d'Oisans area in Chapter 2.

The remaining diagrams in Figure 3.11 b illustrate the excess bedlength problems encountered during the attempted reactivation of listric normal faults, and the interaction between pre-existing extensional fault scarps

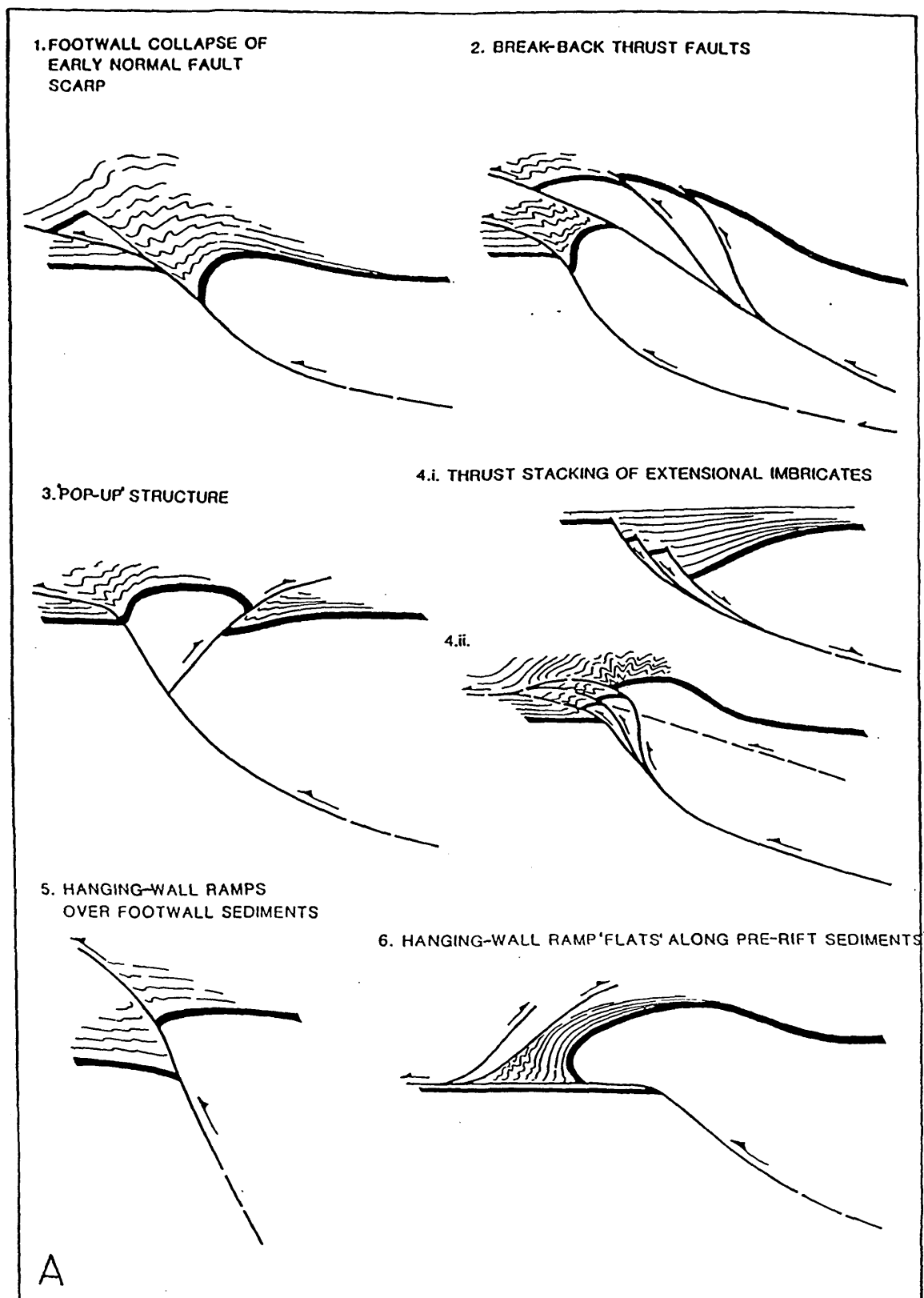
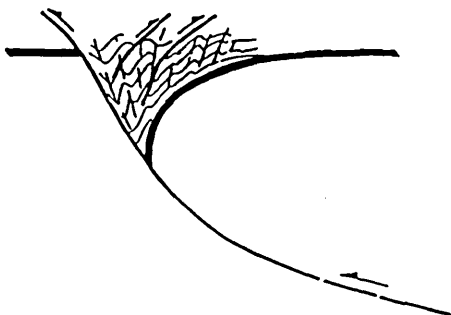
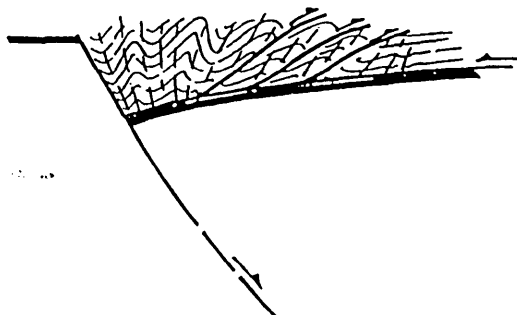


Figure 3.11 A & B: Series of characteristic inversion geometries documented from the Pelvoux area. See text for discussion.

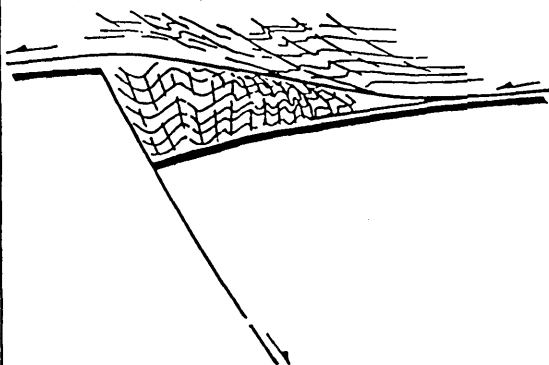
7. DEEP LEVEL REACTIVATION OF A LISTRIC NORMAL FAULT PROFILE WITH COVER BACKTHRUSTING



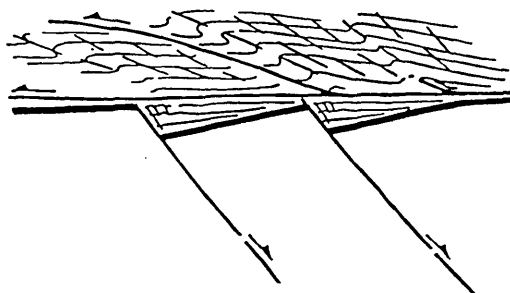
8. NORMAL FAULT ACTS AS A STRESS RISER
- Shortening taken out of the 1/2 graben by backthrusting



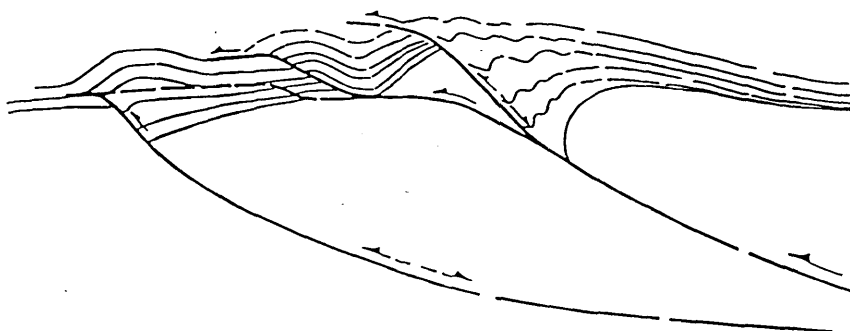
9. NORMAL FAULT ACTS AS A STRESS RISER
- Shortening taken out of the 1/2 graben by 'out of sequence' thrusts



10. COVER DETACHMENT OVERSTEPS SYN-SEDIMENTARY FILL



11. THIN SKINNED THRUSTS CUTTING DOWN SECTION - TRUNCATION OF EARLIER INVERSION STRUCTURES



B

Figure 3.11 A & B

and thrust detachments, within and of, the sedimentary cover. In regions of cover detachment, the early normal faults will disrupt the simple layered stratigraphy and act as effective stress risers (Wiltchko & Eastman, 1983). With steep basement faults, a zone of strong stress concentration occurs. This may cause a flat detachment to ramp (Figure 3.11 b (9)), as observed along the Oulles-Mizoen section in the northern sector of the Bourg d'Oisans half-graben. Alternatively the normal fault scarp may act as a buttress. This results in the generation of complex back-fold-thrust systems in the hanging-wall of the extensional fault. Such a back-thrust system may be difficult to differentiate from back-thrusting developed in the higher levels of a half-graben infill, to accommodate excess bedlengths during the reactivation of the deeper basement fault (Figure 3.11 b (7)). Where there is a strong lithostratigraphic control over the development of cover detachments, a cover thrust detachment may completely "overstep" early half-graben basins, leaving the syn-rift infill undeformed (Figure 3.11 b (10)). This phenomena can be observed along the Rochers d'Armentier (Section 2.1).

The final illustration (Figure 3.11 b (11)), depicts the complex interaction between the thin-skinned thrusting of the sedimentary cover, and the reactivation of older basement structures. Thin-skinned thrust faults in the sedimentary cover, which emerge from reactivated extensional structures in the basement, may cut down the stratigraphic sequence, truncating earlier long wavelength-low amplitude inversion anticlines.

3.4 Thin-skinned thrust verses thick-skinned inversion models

The phenomena of inversion tectonics, convincingly demonstrated with examples from the external basement massifs of the French Alps and the inversion and uplift of major fault blocks upto 1,300 km from the present Alpine deformation front (Ziegler, 1982), suggests that a zone of lithospheric decoupling must exist between the Alpine

orogen and its adjacent foreland, to link with an intra-crustal, crust-mantle or deeper lithospheric mantle decoupling beneath the Alpine foreland (Bally, 1984).

The inferred existence of such an intra-crustal detachment increases the problem of differentiating between "thick-skinned" inversion and "thin-skinned" thrust models to explain the observed cover deformation in the more external parts of thrust belts: the Subalpine Chain equivalents.

In a thin-skinned thrust model, a common basic assumption is the decoupling of the cover sequence from its underlying basement, which presents a smooth uninterrupted profile, unaffected by earlier tectonic events. However in the absence of seismic data, or good well control, the inferred existence of an entirely passive basement may not be a valid assumption. Even where deep geophysical data (refraction and reflection seismics) reveal a normal crustal thickness and magnetic or gravity profiles suggest a smooth surface to the basement-cover interface, this may not be sufficient for the implementation of a thin-skinned thrust model. Where the crust has been previously stretched and thinned by extensional faulting, pre-existing faults inherited from this extensional phase may reactivate during compression, returning the crust back to its normal thickness, and readjusting the basement-cover contact back to a horizontal and apparently uninterrupted profile. During this phase of reactivation, the overlying sediments unaffected by extension (i.e. post-rift sequence) must shorten by folding and thrusting to accommodate the basement shortening. The final surface thrust belt geometry will closely resemble that of a thin-skinned Rocky Mountain style thrust belt, yet balanced cross-sections constructed using conventional methods will yield incorrect solutions.

Jackson (1980) describes this thick-skinned style of deformation using earthquake focal depths to demonstrate reverse faulting in the basement of the southern Zagros mountains of Iran. At the surface this deep deformation is

decoupled from the cover deformation by salt, and the overlying cover deforms by folding.

The two cross-sections constructed across the Subalpine Chains and presented in Chapter 2, epitomize the problem of selecting a thin-skinned thrust or thick-skinned inversion model, where in the latter, shortening is accommodated by the thrust reactivation of earlier normal faults.

3.6 Implications for balanced cross-section techniques

Inversion tectonics involving the reactivation of pre-existing extensional faults, which controlled sedimentation patterns, has a number of important implications concerning the use of the generally accepted conventional balanced section techniques.

Palinspastic reconstructions through compressional fold and thrust belts using balanced section techniques assume an adequate knowledge of the pre-deformational stratigraphic template (e.g. Bally, 1966; Dahlstrom, 1969; Elliott & Johnson, 1980). However there are obvious problems in template prediction and construction if the region has been deformed by an earlier phase of extensional tectonics, contemporaneous with sedimentation. Such an extensional template will clearly be interactive with later thrusting and compression, the pre-existing extensional structures having a profound effect on the positioning of later thrust structures and the structural style. In such areas, the iterations usually required to construct a balanced section may no longer be valid, and modifications of the technique may be required.

The extensional reconstruction proposed must be compatible with the observed compressional geometry, i.e. the interactive nature of the pre-thrusting template provides a further test of the proposed model. However, restorations will only predict the approximate geometry of the initial template. The precise original geometry of

extension is difficult to determine because of:

1. Frequently strain modified cut-off relationships.
2. Re-orientation, and the 2-D rotations of faults and the blocks they bound during inversion.
3. The 3-D aspect of the rotation, that is movements out of the plane of the section - this poses the greatest problem.

Mechanical considerations and structural style may provide valuable information concerning the relative initial orientations of early extensional faults.

With particular reference to the Alpine thrust belt, a two-fold zonation of the external parts can be proposed:

1. An outer foreland zone, characterized by a simple foreland fold-thrust belt which telescopes a stratigraphic template approximating to a uniform gently tapering wedge i.e. the post-rift sequence which corresponds to an almost layer-cake sequence, or a platform sequence beyond the basin margin i.e. the Jura. Balanced section techniques can generally be applied with confidence in this zone, although the possibility of inversion at depth cannot be excluded.
2. An inner zone characterized by broad basement uplifts, controlled by the complex interaction of fault reactivation (inversion) and the formation of new crustal faults. The unadulterated application of balanced section techniques in this zone (e.g. Butler, 1983; Mathews, 1984) is questionable.

In terms of original basin/passive margin setting, the outer zone may correspond to the flexural margin of the basin, comprising uniform post-rift sediments resting directly on basement or a pre-rift sequence if the rift shoulders suffered uplift and erosion from early thermal and later flexural effects (peripheral bulge) during the syn-rift extension phase (Royden & Keen, 1980; Beaumont et

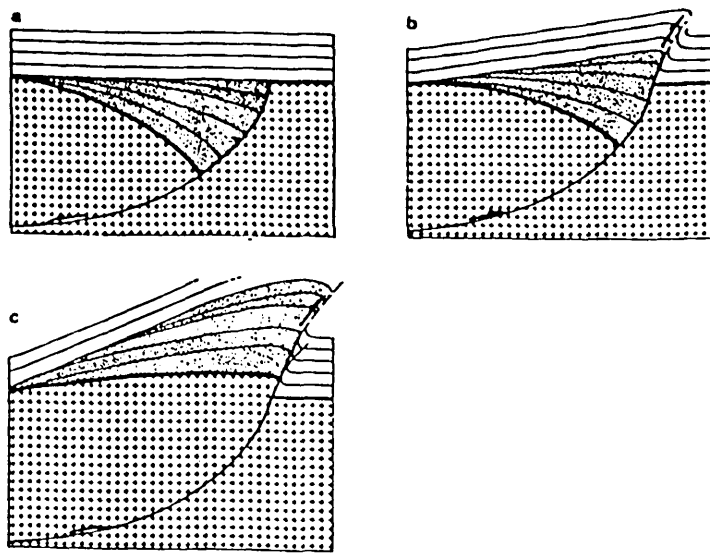


Figure 3.12: Inversion of listric faults, producing a forced fold in the uplifted cover.

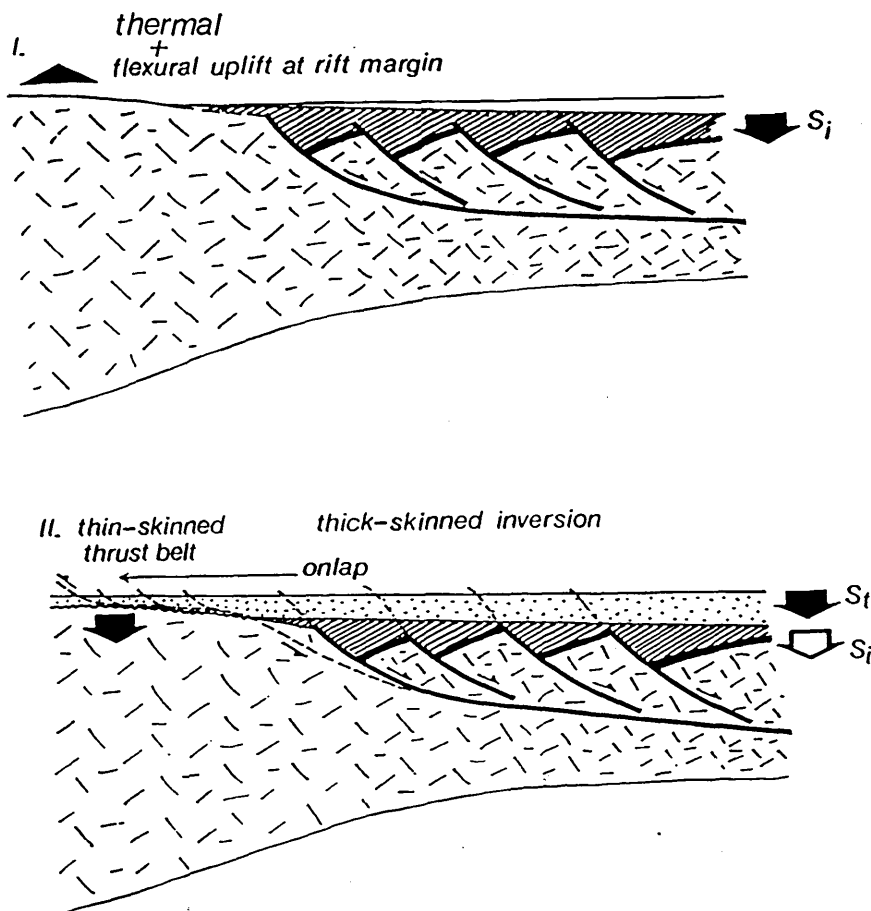


Figure 3.13: i) Thermal and flexural uplift at the rift margin during the early stages of stretching, and ii) during compression and inversion, a thin-skinned fold and thrust belt will develop beyond the basin margin, while fault reactivation and basement thrusting is more important in the area of pre-existing extension. S_i =initial isostatic subsidence, S_t =thermal subsidence.

al., 1982), (Figure 3.13 a). The inner zone corresponds to the zone where the syn-rift stretching is more pronounced and the subsidence is fault controlled. This zone corresponds to the part of the passive margin fault system which is subsequently reworked during compression (Figure 3.13 b).

Chapter 4: Model of Tethyan basin development

4.1 Introduction

The stratigraphy of the Western Alps can be considered in the context of a starved passive margin sequence (Graciansky et al., 1979; Lemoine et al., 1981; 1986; Tricart & Lemoine, 1986; Lemoine & Trumpy, 1987), a concept based on comparisons made with modern Atlantic type margins (Montadert et al., 1979).

In this Chapter, the tectonic subsidence history of the European Tethyan margin is analysed and a thermo-mechanical model given for crustal stretching, based on the results of this subsidence analysis. Comparisons are made with other stretching models. The basic data for this subsidence analysis comprises well data obtained from the University of Grenoble, and previously published accounts of the Tethyan stratigraphy across the European margin, referenced in the following sections.

4.2 Tethyan basin stratigraphy

As described in Chapter 1, the European margin of the Western Alps is divisible into several tectonic domains, each possessing a distinctive Triassic and Jurassic stratigraphy (Lemoine et al., 1986). Each domain corresponds to part of a Mesozoic age basin, generated during the main Tethyan rifting. With the subsequent formation of a passive margin and the onset of the thermal subsidence phase (see McKenzie, 1978), these sub-basins lose their individual identity.

In the French Alps, the Mesozoic stratigraphy correlated across the domains of the Cevennes-Ardeche, Dauphinois, Briançonnais and Piemont, is extremely variable and summarized in Figure 4.1. The Triassic and Jurassic stratigraphy of each of these domains will be discussed.

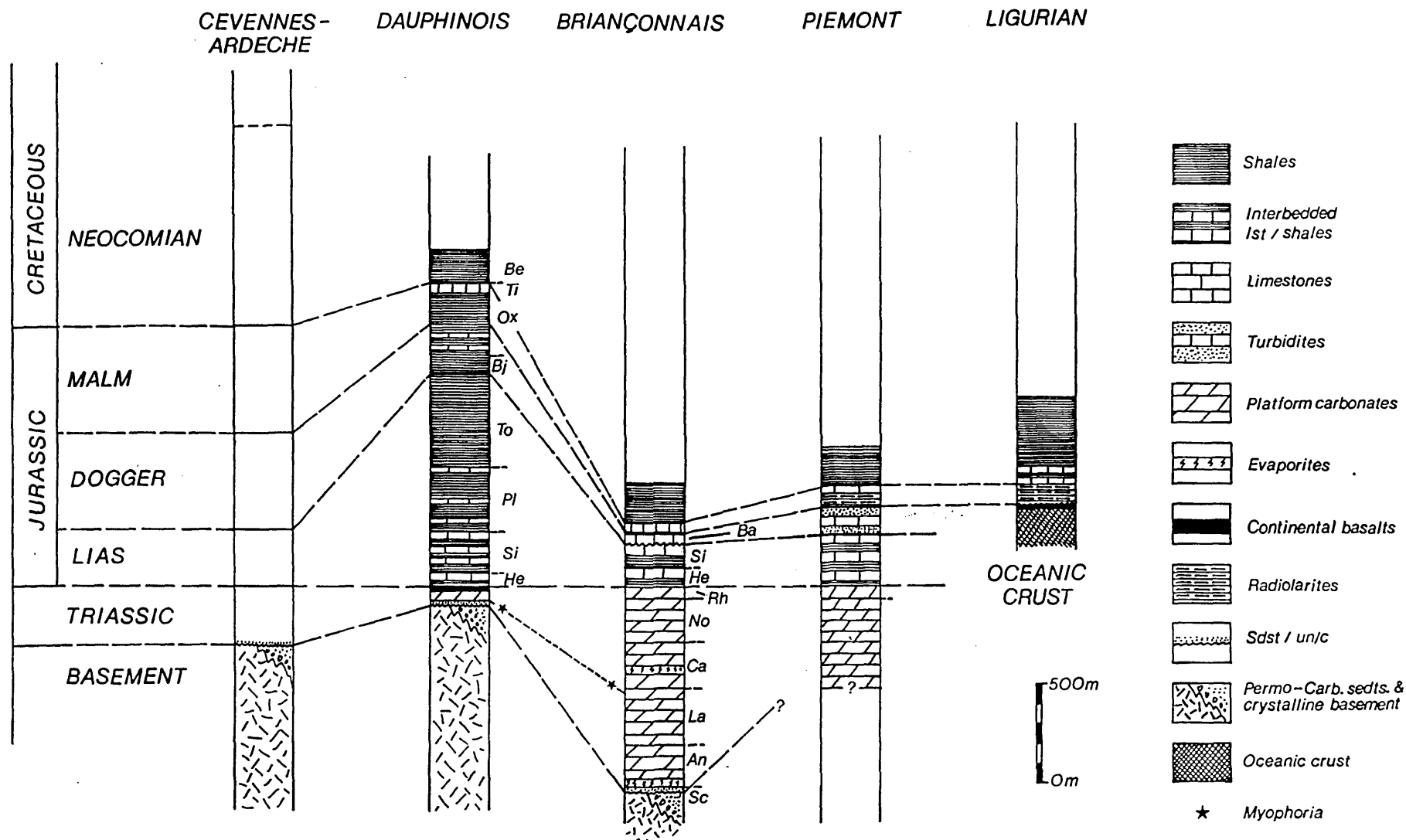


Figure 4.1: Mesozoic stratigraphy of the different domains across the European Tethyan margin (see Fig. 4.2). Sc-Scythian, An-Anisian, La-Ladinian, Ca-Carnian, No-Norian, Rh-Rhaetian, He-Hettangian, Si-Sinemurian, Pl-Pliensbachian, To-Toarcian, Baj-Bajocian, Ba-Bathonian, Ox-Oxfordian, Ti-Tithonian, Be-Berriasian. Based on published accounts of the stratigraphy and well data (for details and references, see text).

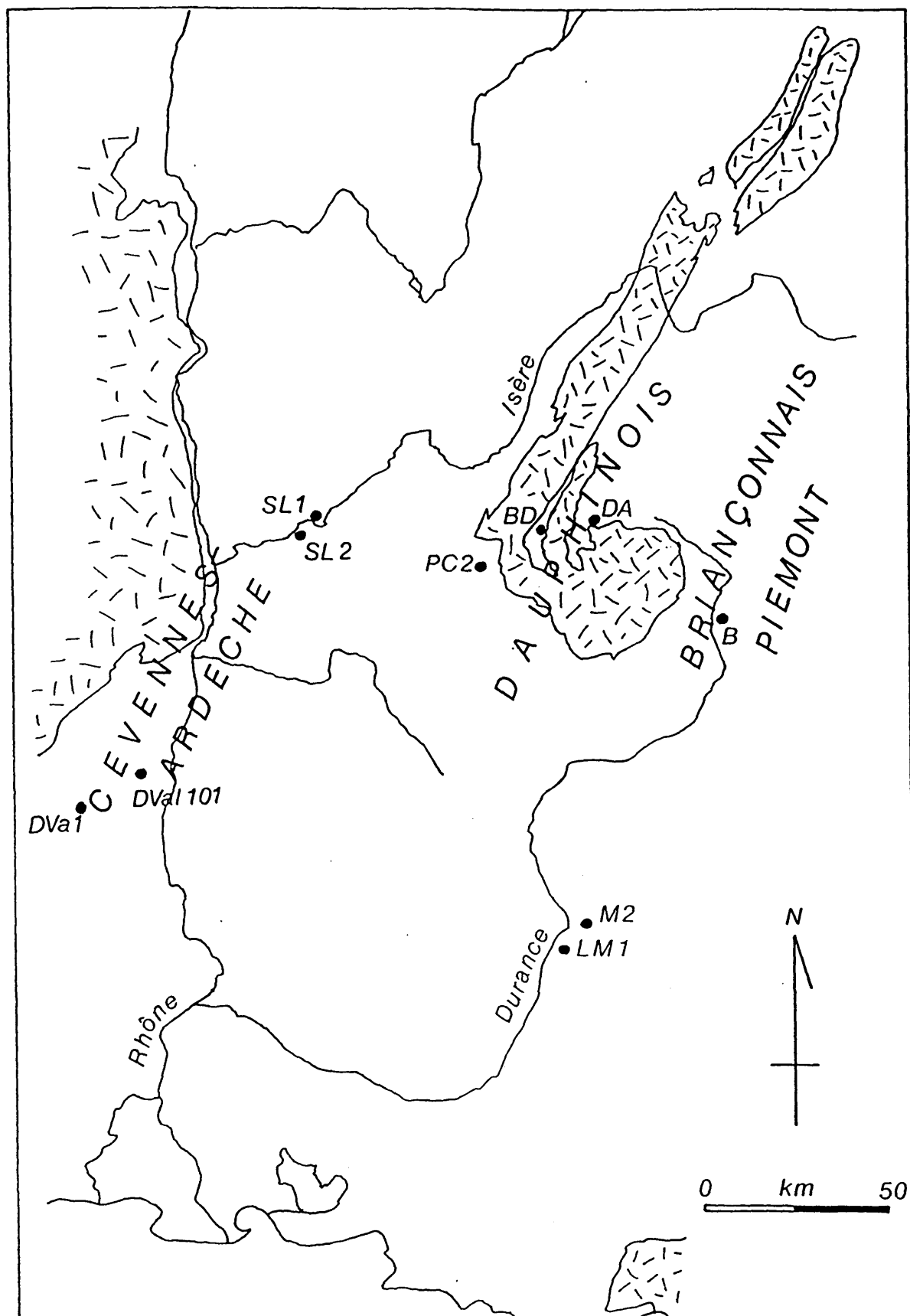


Figure 4.2: Map of the French Alps and SE France showing the location of the wells and stratigraphic sections used in the subsidence-history analysis. PC-Pierre Chatel, BD-Bourg d'Oisans, DA-Deux Alpes, B-Briançonnais.

1. Cevennes-Ardecche domain:

Well data concerning lithologies for this section are limited, and only information on the formation tops is available from Dva 1 and Dval 101 (Figure 4.2). It is assumed that the sediments are largely equivalent to those outcropping in other areas of the French Alps. The relevant sedimentary thickness data recovered from Dva 1 are:

pre-rift crystalline basement; 252 m Trias; 240 m Lias; 490 m Dogger; 555m Malm; 969 m Neocomian.

The thicknesses obtained from nearby Dval 101 are consistent with these.

2. Dauphinois domain:

TRIASSIC: Overlying the Pelvoux pre-rift basement, the Triassic sequence corresponds to a markedly reduced series of supra-tidal dolomites, with a strong continental influence at their base; a basal sandstone passes upwards into 15-30 m of bedded dolomites, which towards the top of the series, are intercalated with continental alkaline basalts with individual flows up to 50 m in thickness. At the base of the dolomites, a *Myophoria* bearing fossil horizon correlated with a similar horizon the Briançonnais domain (Figure 4.1), marks a late Ladinian-early Carnian transgression (Lemoine et al., 1986) and the dolomite-basalt series is thus interpreted to be of Carnian-Norian age. Dumont (in press) documents a Rhaetian hiatus. Little evidence for Triassic faulting exists, other than the Armentier half-grabens (Section 2.1).

JURASSIC: i) **LIAS** - Hettangian shallow water limestones and shales herald the beginning of a Jurassic eustatic cycle (Dumont, in press). The Hettangian to Carixian (lower Pliensbachian), predominantly limestone-shale sequence is characterized by numerous facies and thickness variations across short horizontal distances, local fault scarp breccias and olistoliths (Barfety & Gidon, 1981). The

Hettangian, Sinemurian and Carixian sediments locally attain thicknesses of up to 300 m. The Domerian (upper Pliensbachian) is characterized by deeper water shale sedimentation and broader facies variations develop across wider areas; shallow water shoals comprising predominantly crinoid debris, located over uplifted foot-walls to faults, pass basinward into deeper water hemipelagic shales. The Toarcian sees the transition to almost entirely deep water shale sedimentation and the change from calcareous to non-calcareous sedimentation (lowering of basin floor to beneath the CCD). Calci-turbidites were still generated from the uplifted foot-walls of faults and pass basinward into the deeper water shales. Sedimentary thicknesses are variable ranging from 100's of metres to 0 m, depending on relative position with respect to the hanging-wall dip slope. Large basement olistoliths of this age are also documented adjacent to basement fault scarps.

ii) **DOGGER:** During the Aalenian, deep water shales were deposited, whilst the Bajocian sees a return to calcareous sedimentation, characterized by shales interbedded with carbonates (Barferty & Gidon, 1984). A strong differentiation of paleogeography is evident with nodular and condensed sequences deposited over uplifted foot-walls and detrital input where the foot-walls are uplifted above sea-level. Thicknesses range from 100's of metres to 0 m. The Callovian is characterized by the deposition of deeper water shales.

iii) **MALM:** Both the Callovian and the Oxfordian are characterized by the deposition of deep water black shales (Terres Noires). These sediments seal the pre-existing submarine topography and consequently thickness variations range from 100's of metres to 0 m. The Kimmeridgian pelagic shales and competent Tithonian limestones, preserve a more uniform thickness across the Dauphinois domain (150 m). These are overlain by Neocomian pelagic shales.

3. Briançonnais domain:

The Briançonnais domain presently comprises several stacked thrust sheets (Tricart, 1984).

TRIASSIC: Detailed sedimentological work by Baud & Megard-Galli (1975) and Megard-Galli & Baud (1977) demonstrate that shallow marine carbonates overlies Scythian tidal marine orthoquartzites (50-200 m), which rest unconformably on pre-rift crystalline basement and Permo-Carboniferous deposits. Four main stages can be distinguished (see Lemoine et al., 1986):

1. 5-10 m of Early Anisian evaporites, characteristic of a coastal sabkha environment.
2. 300-400 m of Middle Triassic platform carbonates, characterized by 3 transgressive cycles.
3. Thick Carnian evaporites, marking the beginning of a new transgressive cycle which can be correlated across to the Pelvoux in the Dauphinois domain.
4. 250-300m of Norian-Rhaetian platform carbonates, divided into two Formations (Lemoine et al., 1986): the Hauptdolomit Formation - 200-250 m of dolomitized carbonates and the Rhaetian Formation - 50-100 m of micritic limestones, black shales and dolomitic mudstones. The latter corresponds to a wide-spread Rhaetian transgression across much of western Europe (Lemoine et al., 1986), but which is absent from the Pelvoux.

JURASSIC: In each Briançonnais thrust sheet, with the exception of the Peyre Haute nappe, early to Middle Jurassic sediments are missing. The presence of Hettangian and Sinemurian limestones unconformably overlain by late Jurassic platform and then deeper water limestones (Lemoine et al., 1986), suggests that this absence partly results from uplift and erosion rather than non-deposition. Extensive kharstification (Lemoine et al., 1985), which

marks the contact between Triassic carbonates, and Bathonian limestones, gives evidence of sub-aerial exposure. Callovian to Oxfordian non-calcareous deposition suggests a lowering of the deposition level to beneath the CCD (Lemoine et al., 1986). The Sub-Briançonnais domain to the W, is characterized by the "nappes de Breche" (Lemoine et al., 1986) which comprise detritus and breccias derived from the adjacent uplifted Briançonnais.

4. Piemont domain:

In the Piemontais domain, Lower Liassic limestones and shales are present. The Dogger is however characterized by a syn-rift flysch, comprising largely of dolomitic and siliceous clasts, derived from the uplifted Briançonnais domain. The Malm is characterized by deep water pelagic limestones (Lemoine et al., 1986).

5. Ligurian domain:

In the Queyras area, overthrust remnants of oceanic crust have been documented (Tricart & Lemoine, 1986, Lemoine et al., 1986). The sediments (Schistés Lustrés) deposited on this oceanic crust have been dated from microfossil discoveries (De Wever & Caby, 1981). The sequence consists of Oxfordian radiolarian cherts, Malm-Berriasian pelagic limestones, Neocomian limestones and shales, Middle Cretaceous black shales and Senonian calcschists (Lemoine et al., 1986).

4.3 Subsidence-history

The quantitative procedure developed to calculate tectonic subsidence-history (Sleep, 1971; Sclater & Christie, 1980) and modified for fully lithified strata (Bond & Kominz, 1984), was not applied to the sediments of the French Alps for the following reasons:

1. The syn-rift sediment type, with a high shale percentage, proved unsuitable for analysis (see Bond &

Kominz, 1984). This was mainly due to widespread deformation and penetrative cleavage formation, which may have substantially altered the original stratigraphic thickness (see also Hossack, 1978), making conventional decompaction corrections problematic. Indeed, Beach (1982), from a study of extended belemnites, estimates that a flattening strain may have reduced the thickness of some lower level shale horizons in the Ultra-Dauphinois zone by up to 85 %.

2. Although a good structural control has been established for the deformation occurring in the Dauphinois domain, the distributed and penetrative nature of the strain results in numerous intra-cover detachment zones and thrusts, of indeterminate displacement. An incomplete control on local thickness and facies variations and uncertain correlations across faults introduces errors when determining original stratigraphic thicknesses. This factor may be most significant in the Briançonnais domain, where, although the syn-rift sediments are suited to a subsidence analysis, they are detached at their base. A complete understanding of the thrust geometry is a necessary pre-requisite before stratigraphic thickness columns can be chosen for a tectonic subsidence analysis.

3. Difficulties are encountered in the determination of water depths, essential to the construction of water-filled basin curves for a sediment starved margin. Only the shallow water carbonate banks, where the effects of water depth are minimal, could be used in a tectonic subsidence analysis.

4. Such an undertaking is very time consuming.

i) Subsidence curves:

If only the timing and patterns of subsidence are being investigated and the subsidence curves are not being analysed in a quantitative fashion to determine rates of subsidence or crustal extension (e.g Wood & Barton, 1984),

then an analysis of cumulative sediment thickness curves uncorrected for decompaction could suffice. Decompacted curves possess the same general slope trends, but will show an overall increase in slope angle.

The subsidence-history curves in Figure 4.3 plot time along the horizontal abscissa and cumulative sediment thickness along the vertical ordinate axis and assume that the space for sediment accumulation is controlled by the subsidence of the basement. The geologic time scale of Harland et al. (1982) was used.

ii) Limitations of the curves:

Because of the uncertainties surrounding the determination of water depths or palaeobathymetry, which is influenced by subsidence rate, sedimentation rate and eustatic sea-level changes, the cumulative thickness curves for a sediment starved basin will underestimate the amount of subsidence, ignoring extra subsidence caused by sediment and water loading. However in this analysis, water depths and their relation to tectonic subsidence events and eustatic sea-level changes, are assessed in a qualitative fashion by identifying shallowing or deepening trends.

A semi-quantitative and qualitative analysis of the subsidence-history across the European Tethyan margin, using cumulative sediment thickness curves and information concerning lithology types and depositional environment follows.

1. TRIASSIC SUBSIDENCE-HISTORY:

i) Cevennes-Ardeche: Well data from this area indicates steady subsidence and the deposition of Triassic sediments. Information concerning lithology types is limited.

ii) Dauphinois: Subsidence curves constructed using data from wells in the Vercors area, on the Cevennes-Dauphinois border (wells SL 1 & SL 2, Figure 4.3), show little in the

way of subsidence, with thick anhydrite-dolomite sequences (59 m and 118 m) and a reduced Rhaetian sequence (28 m). In the Bourg d'Oisans area and northern Pelvoux (Figures 4.2 & 4.3), minor subsidence commenced in the Late-Ladinian with reduced subsidence and possible uplift in the Rhaetian, characterized by continental alkaline basalts and a Rhaetian hiatus.

ii) Briançonnais: Subsidence commenced in the Anisian and continued at a steady rate up until the Rhaetian, when an apparent slowing down occurred (Figure 4.3). Lemoine et al. (1986) provide evidence for extensional faulting during sedimentation and suggest that nearly 1000 m of platform carbonates are divided in two by an aborted Carnian rift event.

iv) Piemont: The Triassic subsidence-history is similar to that documented from the Briançonnais domain.

2. JURASSIC SUBSIDENCE-HISTORY:

i) Cevennes-Ardeche: Significant thickness variations recorded for the Lias in adjacent wells suggests important fault controlled subsidence (DVA 1, 240 m; DVal 101, a colossal 2028 m). However the Lias covers a time range of 25 my, and the overall subsidence may have occurred over a much shorter time interval, such as 5 my. A substantial amount of subsidence can be inferred prior to the Aalenian. This substantial and rapid subsidence continued throughout the Dogger, Malm, and Neocomian. However, the degree of stratigraphic control is insufficient to reconstruct a detailed Jurassic subsidence-history.

ii) Dauphinois: Data from two wells in the Vercors (Figures 4.2 & 4.3) show an increased rate of subsidence during the Hettangian which continued at a more gentle rate through the Sinemurian and increased again at the Toarcian-Aalenian boundary. The subsidence rate then eased up at the Oxfordian-Kimmeridgian boundary. In the Pelvoux and Bourg d'Oisans area, a progressive increase in the subsidence

rate occurred during the Hettangian; the curve continues to steepen through the Sinemurian and the maximum period of subsidence commenced at the Domerian-Toarcian boundary. The rate of subsidence then slowed down from the Callovian onwards.

iii) Briançonnais: After a slowing down of subsidence during the Rhaetian, increased rates of subsidence are again resumed during the Hettangian and Sinemurian: the same subsidence phase documented from the Dauphinois domain. It is suggested that the rapid Toarcian-Aalenian deepening event documented from the Dauphinois and Cevennes areas did not affect the Briançonnais and that during this period, the Briançonnais was undergoing uplift, emergence and erosion. However, the possibility that this deepening event did occur and that the sediments were subsequently uplifted and removed by a post-Aalenian to pre-Bathonian erosion cannot be entirely excluded. From Bathonian times onwards, the Briançonnais subsided rapidly, in places attaining depths below the CCD during the Callovian (Lemoine et al., 1986).

iv) Piemont: The Hettangian and Sinemurian subsidence events are recorded, yet the uplift of the Briançonnais is not. The rate of subsidence during this uplift period is difficult to determine, although syn-rift flysch sediments were derived from the adjacent Briançonnais high. Rapid subsidence to beneath the CCD during the Callovian can be correlated with the formation of oceanic lithosphere in the Ligurian domain to the E.

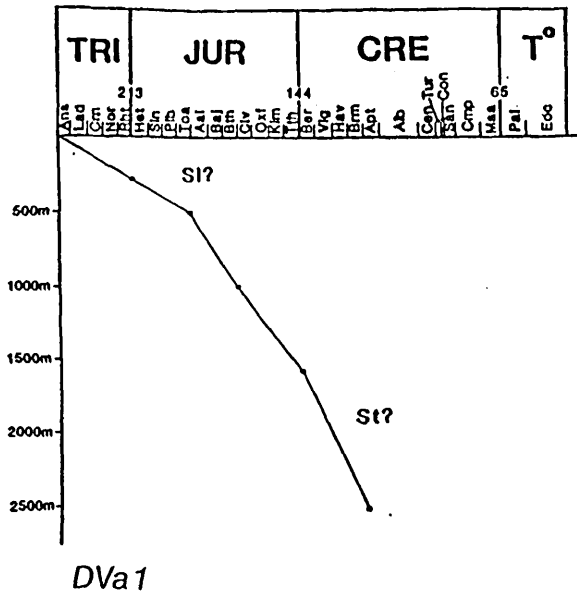
v) Ligurian: The deposition of deep water radiolarian cherts during the Callovian and later pelagic shales marks the beginning of sea-floor spreading in this domain.

Summary of the main subsidence events:

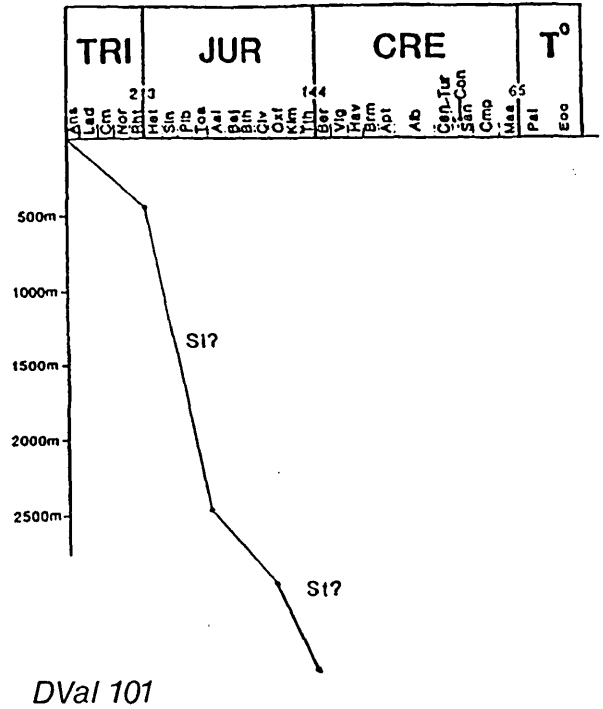
Three tectonically controlled subsidence events are identified. Because of the poor stratigraphic control, the timing intervals are very approximate. In reality, each

1. CEVENNES-ARDECHE

CEVENNES-ARDECHE BORDER

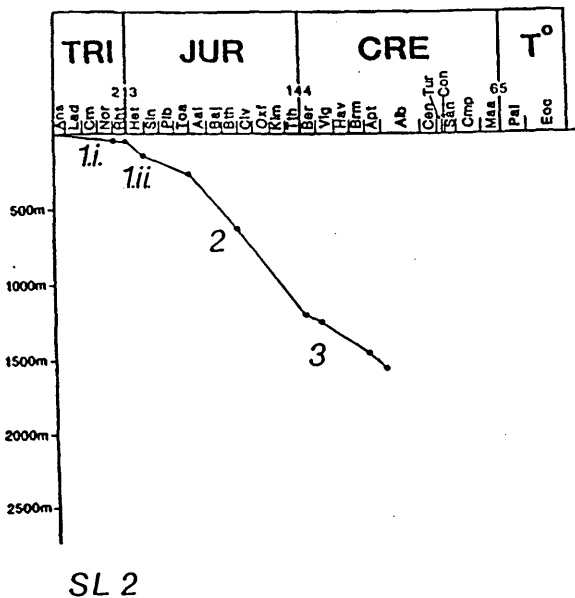


CEVENNES-ARDECHE BORDER



2. DAUPHINOIS

VERCORS UNSLIPPED FORELAND



VERCORS UNSLIPPED FORELAND

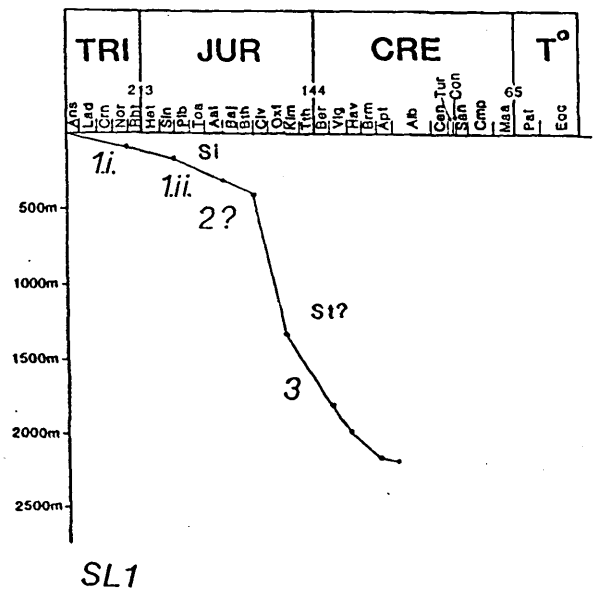
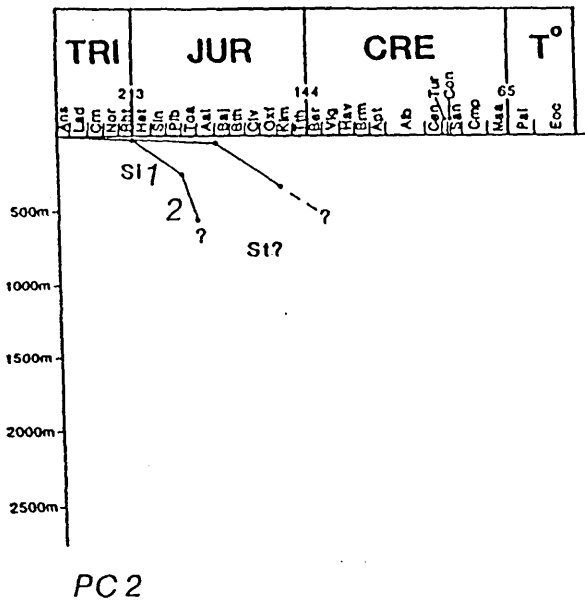


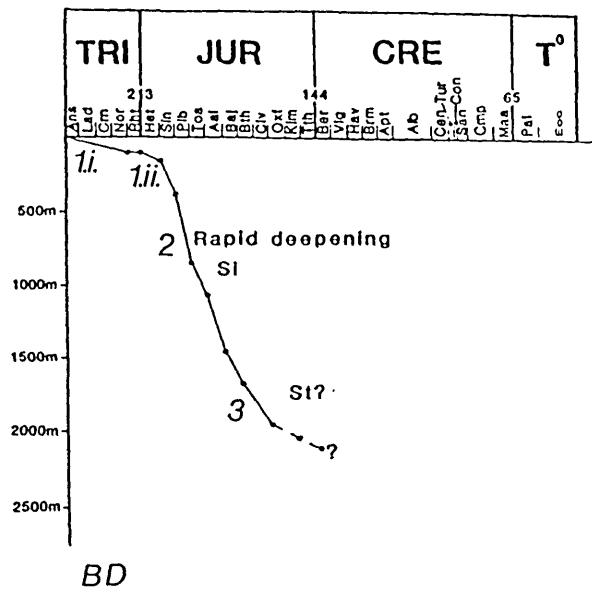
Figure 4.3: Cumulative sediment thickness curves, uncorrected for decompaction, constructed for the different domains across the European Tethyan margin (for locations see Fig. 4.2). Black circles show the number of data points used to construct each curve, Si-initial subsidence, St-thermal subsidence, 1.i, 1.ii, 2 and 3, are tentatively correlated with the main subsidence events (see text for discussion).

2. DAUPHINOIS (CONT.)

PIERRE CHATEL-LA MURE



DAUPHINOIS-BOURG D'OISANS



DAUPHINOIS-PELVOUX

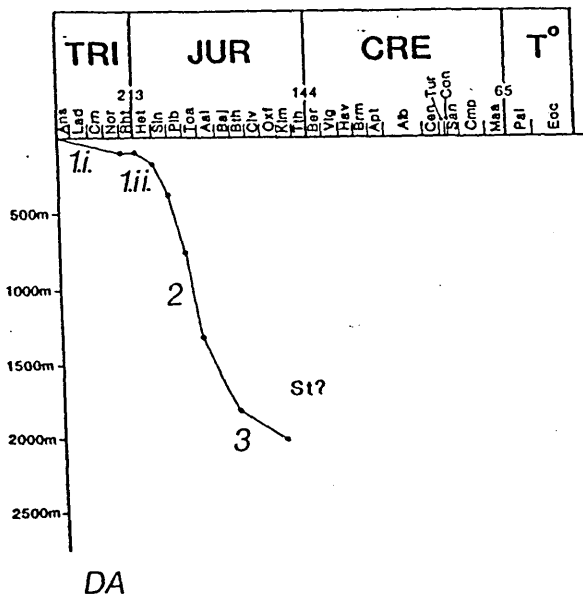
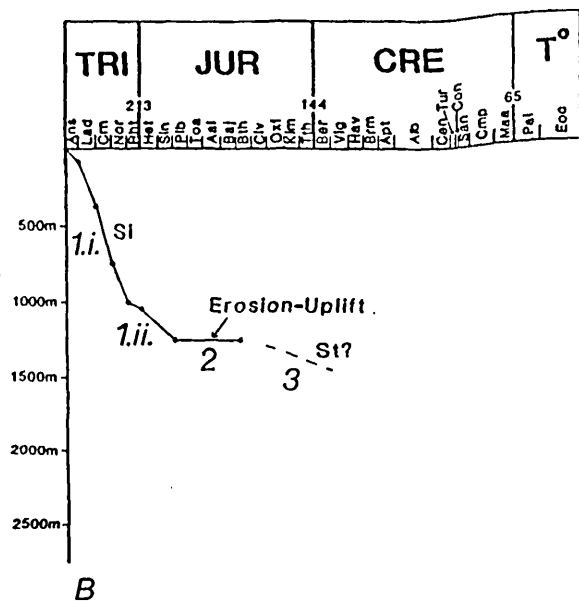


Figure 4.3

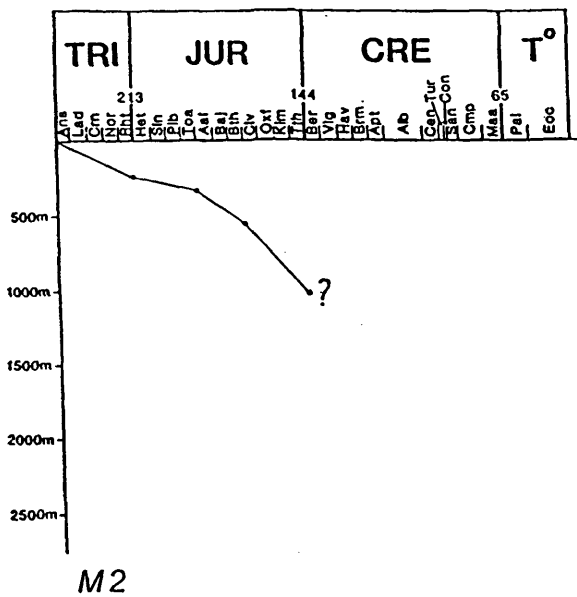
3. BRIANÇONNAIS

BRIANÇONNAIS DOMAIN



4. PROVENCE PLATFORM

MIRABEAU-PROVENCE PLATFORM



LES MEES-PROVENCE PLATFORM

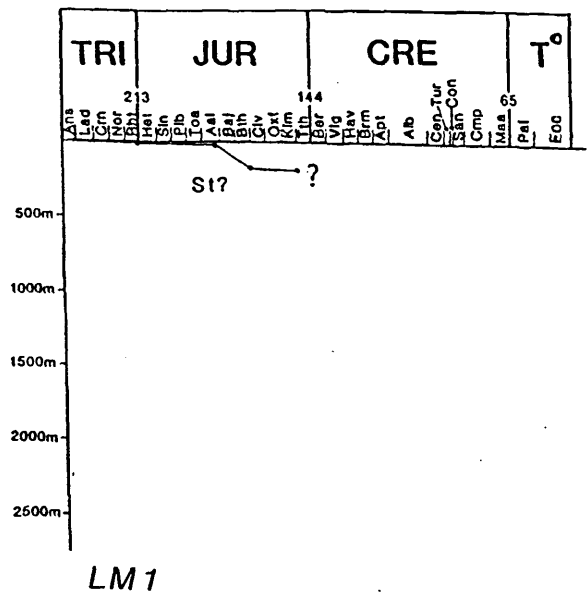


Figure 4.3

event probably corresponds to a rapid subsidence accommodated over a short time interval, separated by longer periods of controlled, regular subsidence.

EVENT 1: Divided into two sub-events:

i) Initiated in the Anisian in the Briançonnais and Piemontais (24 my duration ?) where it is well expressed and Carnian in the Dauphinois (12 my duration ?), where the amount of subsidence is minor.

The Rhaetian corresponds to a period of reduced subsidence in the Briançonnais and uplift in the Dauphinois (Pelvoux). This period divides EVENT 1 into i) and ii).

ii) Renewed subsidence occurred during the Hettangian and Sinemurian and is well expressed across the margin (13 my duration ?).

EVENT 2: Dramatically increased rates of subsidence, relative to EVENT 1, occurred during the Toarcian and Toarcian-Aalenian boundary. This was most pronounced in the Dauphinois zone and in the Cevennes. EVENT 2 corresponds to a negative subsidence (uplift) in the Briançonnais. The two phenomena are considered to be related.

EVENT 3: There was rapid subsidence to bring sedimentation beneath the CCD during the Callovian. This was well expressed in the Briançonnais and Piemontais, but poorly expressed in the Dauphinois.

Deepening trends in the short-term eustatic sea-level curve for the Jurassic (Hallam, 1981) (Figure 4.5), coincide with subsidence events 1 ii), 2 and 3. The amount of subsidence recorded is clearly too much to be solely related to sea-level changes, however these sea-level trends will enhance any deepening effects induced by the tectonic subsidence events. The different subsidence events recorded across the European margin are summarized in Figure 4.4.

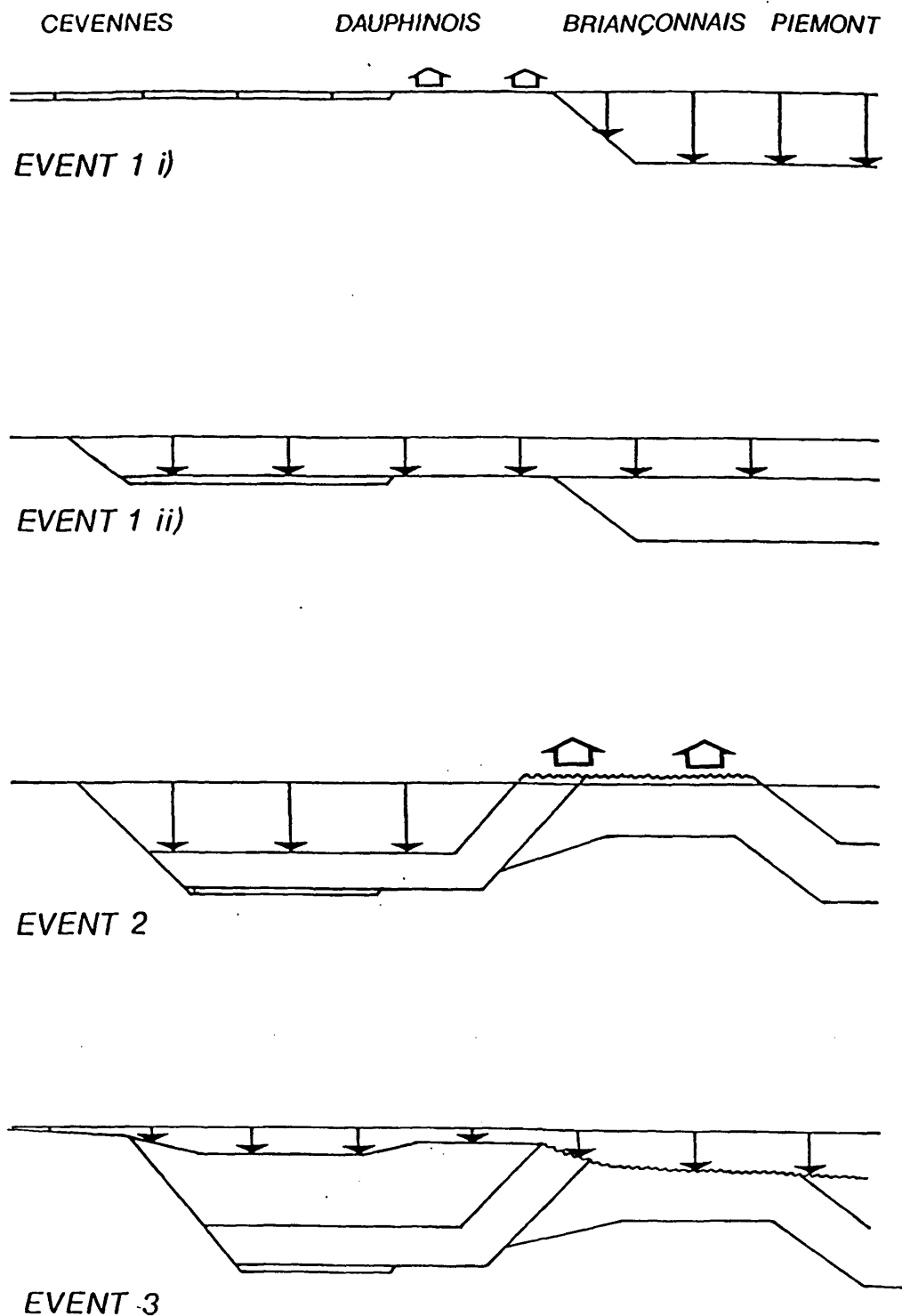


Figure 4.4: Simplified diagram showing contrasts in vertical subsidence movements (negative and positive) between the different domains of the European Tethyn margin, for each of the main events identified. See text for discussion.

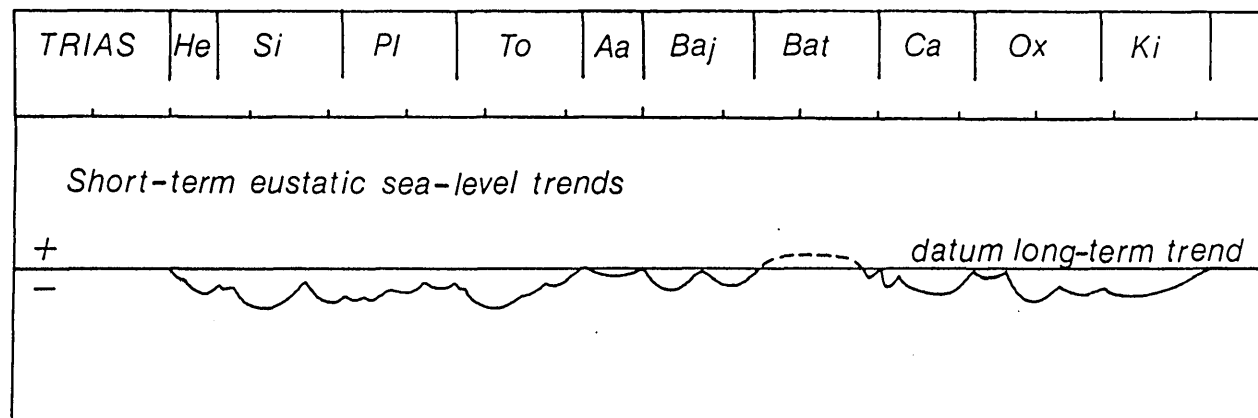


Figure 4.5: Short-term variations from the long-term trend of eustatic sea-level for the Jurassic (After Hallam, 1978; 1981; Morton, 1987). See text for details.

4.4 Geometry of normal faulting

It has been argued previously in Chapter 2, that in the Dauphinois area, two normal fault systems of different age and distinctive geometry can be detected.

From both structural and stratigraphic considerations, a late Triassic-early Liassic system of normal faults were apparently rotated to gentle dips and consequently reactivated during the Alpine compression. From considerations of isostasy, the relatively low initial subsidence associated with this phase of crustal extension (EVENT 1. ii) suggests that only the upper part of the crust was thinned, supporting the interpretation of a high level shallow crustal detachment, at least in the Dauphinois.

The period of maximum subsidence (EVENT 2) appears to be associated with a larger, more widely spaced set of steeply dipping normal faults. From the large isostatic subsidence associated with these faults, this episode of faulting apparently thinned a greater portion of the European crust in the Dauphinois domain, suggesting a deeper sole detachment.

From the general normal fault reactivation patterns (Chapter 5), it is argued that there was more reactivation and stacking of the passive margin faults in the eastern part of the Pelvoux, where the faults had been rotated into gentle dips above a shallower extensional detachment. The steeper faults, more ubiquitous to the W, apparently blocked a large part of the Alpine contractional displacement. Locally there appears to have been some interference between the two fault systems and the steeper faults such as the Ornon and Eastern Grandes Rousses faults, breach the earlier detachment (Chapter 2). Figure 4.6 is an interpretative cross-section through the external basement massifs at the latitude of the Pelvoux and Figure 4.7 is a simplified section to show the suggested extensional fault geometry at the end of the Mesozoic. One

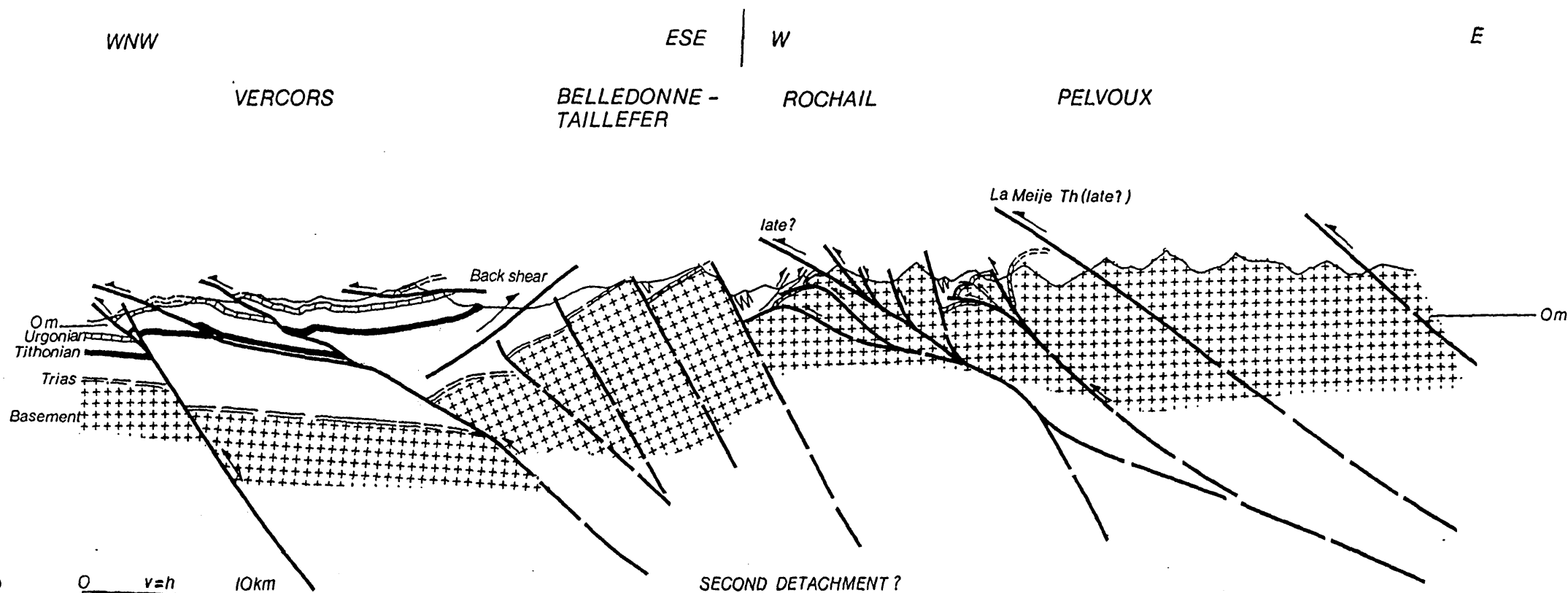


Figure 4.6: Suggested surface structure across the Pelvoux and the Vercors (after Gillcrist et al., 1987). As discussed in Section 2.6, different thrust geometries may be envisaged for the Vercors and Belledonne parts of the cross-section. Some of the thrust sheets have moved in a SW direction, out of the plane of section. The location of the section is shown as AA' on Fig. 1.2.

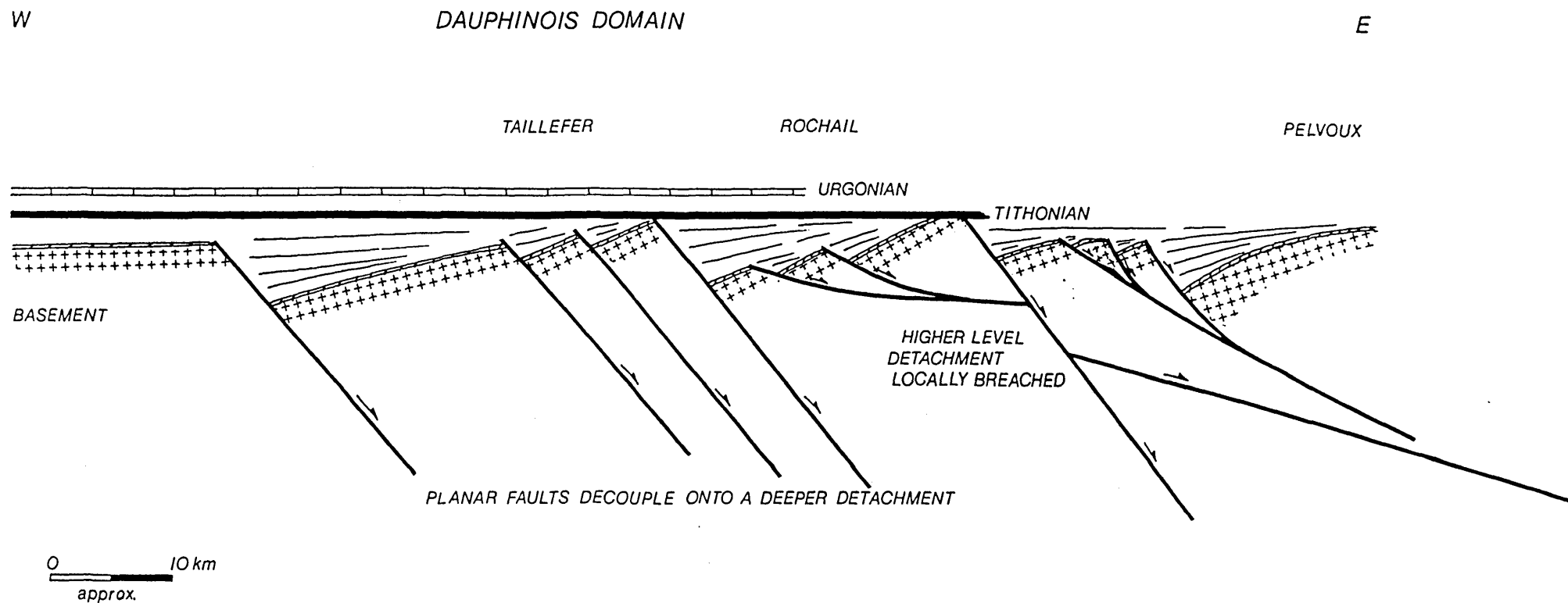


Figure 4.7: Suggested restoration to show the geometry of Jurassic faulting after removal of the effects of the Alpine compression. See text for discussion.

important question to ask is what was the deeper crustal expression of this upper crustal extension in the Dauphinois domain further to the SE.

1. The development of a second fault system: mechanical arguments

Jackson (1987), assuming that planar crustal faults initiate at a typical Andersonian angle of 60° (Anderson, 1959), proposed that a relative extension of 1.7β is the maximum extension that can be accommodated by a single generation of faults, if they are to be seismically active. With a linear stress-strain relationship, this lock-up value is clearly related to the frictional properties of the fault surfaces and the geometry of faulting. Extension beyond this critical factor may be accommodated by:

1. Continued fault activity on the same faults by aseismic creep.
2. A second generation of "breaching" faults may become active, cutting the first generation faults, which have been rotated to dips in the order of 30° .
3. Extension on new faults, which possibly exploit a pre-existing mechanical anisotropy in the crust, continues in an adjacent area - cratonward propagation of the rift system ?.

The situation most suited to the European margin is 3), with 2) locally developed where there is interference between the later and earlier fault systems.

The thinning of the crust above the brittle-ductile transition will cause it to migrate downwards in the crust, where the transition is pressure-temperature dependant for a constant stretching rate (Sibson, 1977). In this case the switching of a sole fault to a deeper sub-horizontal shear could be mechanically favoured (Coward & Gibbs, 1987).

4.5 Non-uniform stretching model

Over the last 10 years, a multitude of lithospheric stretching models which involve either homogeneous stretching of the entire lithosphere, displacement on a lithospheric-scale shear zone, or some combination of the above two end members, have been developed. The three main types of model are depicted in Figure 4.8.

At first appearance the extensional evolution of the European Tethyan margin corresponds quite well to a simple uniform stretching model (McKenzie, 1978) (Figure 4.8 A), with a prolonged 62-74 my duration of stretching; the 62 or 74 my ages depend on whether or not the pre-Carnian rift phase (Lemoine et al., 1986) is considered to be part of the Tethyan rifting. This was followed by a thermal subsidence phase. However, the McKenzie model of uniform crustal stretching only predicts an instantaneous positive or negative subsidence following a phase of crustal stretching. This is dependant upon the initial crust-lithosphere thickness ratio; a critical Cz/Lz ratio of 0.138 exists, above which subsidence occurs, and below which uplift will occur. Such a model cannot explain the pattern of contrasting differential subsidence observed across the European margin and in particular, the post-Sinemurian/pre-Bathonian inversion and uplift of the Briançonnais domain. In considering the effects of Alpine compression, the restored Briançonnais stratigraphy suggests this inversion and uplift is of regional extent and not caused by local isostatic effects during faulting (i.e. foot-wall uplift).

It is proposed that a model of heterogeneous crustal stretching (Wernicke, 1981; 1985; Coward, 1986), or a non-uniform stretching model (Royden & Keen, 1980; Beaumont et al., 1982) (Figure 4.8 C), which allow for locally different values of stretching between the upper crust and the lower crust and mantle lithosphere, can best explain these patterns of differential subsidence and uplift.

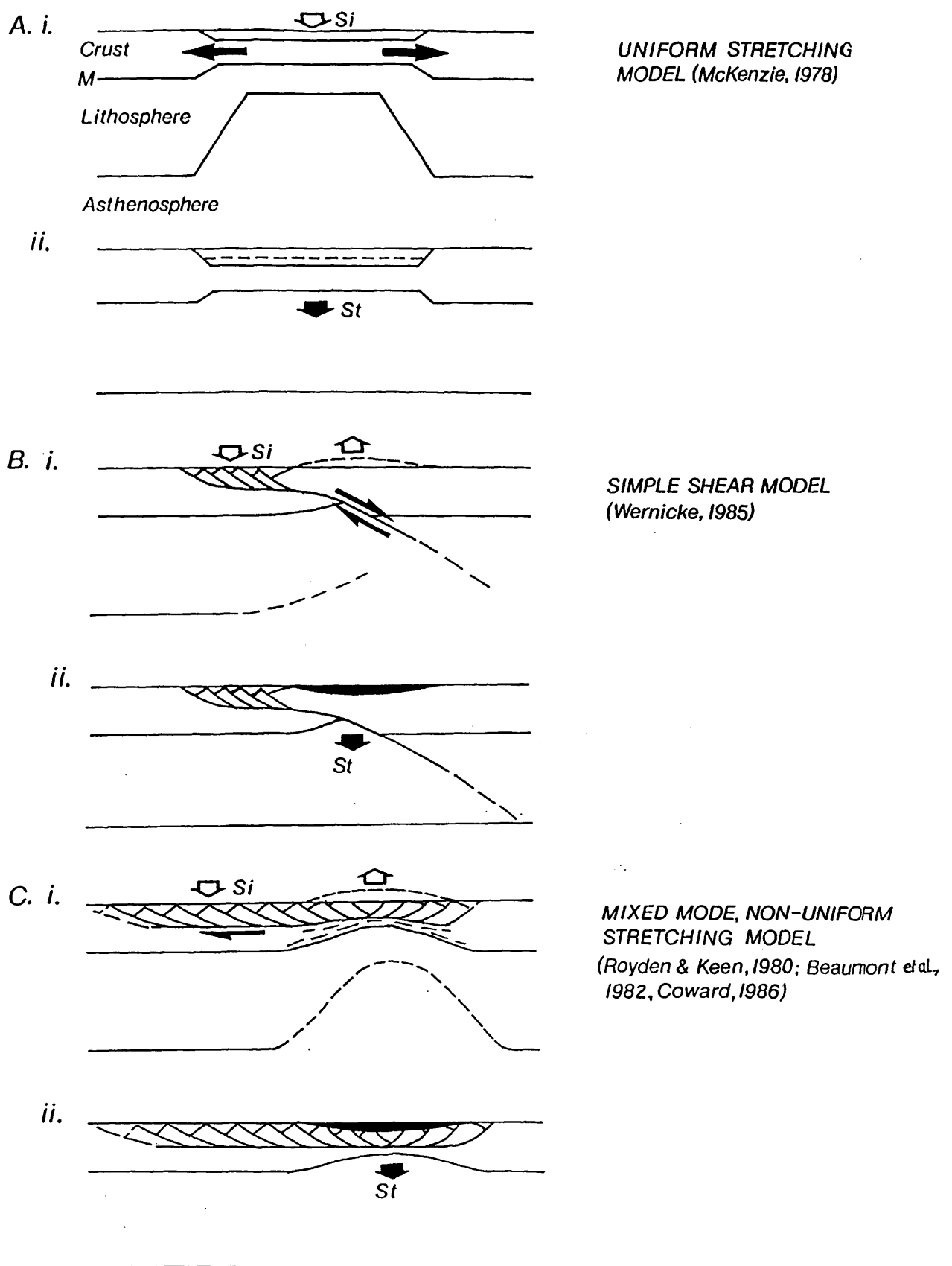


Figure 4.8: Examples of the main different models of lithospheric stretching and basin evolution (A,B,C). i) initial stretch phase (Si), ii) thermal subsidence phase (St). Arrows indicate sense of subsidence (i.e. negative or positive).

i) Stretching model:

Coward (1986) describes a typical sedimentary basin with an outer zone where the upper crust only is stretched and an inner zone, where the upper crust is stretched by a factor β , and the lower crust is stretched by the factor β , plus an extra factor β' required to balance the upper crustal extension at the basin margin. A zone of initial subsidence develops where only the upper crust is thinned, whereas in the part of the basin where the whole of the lithosphere is thinned, the addition of the extra factor β' in the lower crust and mantle lithosphere may cause uplift. This effect is enhanced if the lower crust is not considered as a homogeneous granitic mass of density 2.8 g cm^{-3} , but increases in density towards its base due to an increase in the mafic and ultra-mafic content. The subsequent phase of thermal subsidence affects the areas where largely lower crust and lithospheric mantle have been thinned (Figure 4.8 C).

ii) Application of the model:

The tectonic evolution of the European margin throughout the Mesozoic is modelled on the existence of two crustal detachments (Figures 4.9, 4.10 & 4.11). However upon passing into middle and lower crustal levels, these faults may equally become zones of broad ductile shear and the crust may deform by a combination of simple shear and pure shear stretching, or the deformation may concentrate along zones of more intense stretching or simple shear. Any interpretation concerning the deep crustal geometry remains speculative.

Subsidence EVENT 1 i) is interpreted to result from a thinning of the whole lithosphere in the Briançonnais and Piemont basins, generating an instantaneous isostatic subsidence (Figure 4.9). In the Dauphinois basin this subsidence event is poorly recorded, with minor uplift and the extrusion of continental alkaline basalts. The basalts suggest that the asthenosphere underlying the Briançonnais

RIFT INITIATION: LATE TRIASSIC - EARLY LIASSIC

HETEROGENEOUS STRETCHING MODEL FOR THE EVOLUTION OF TETHYS

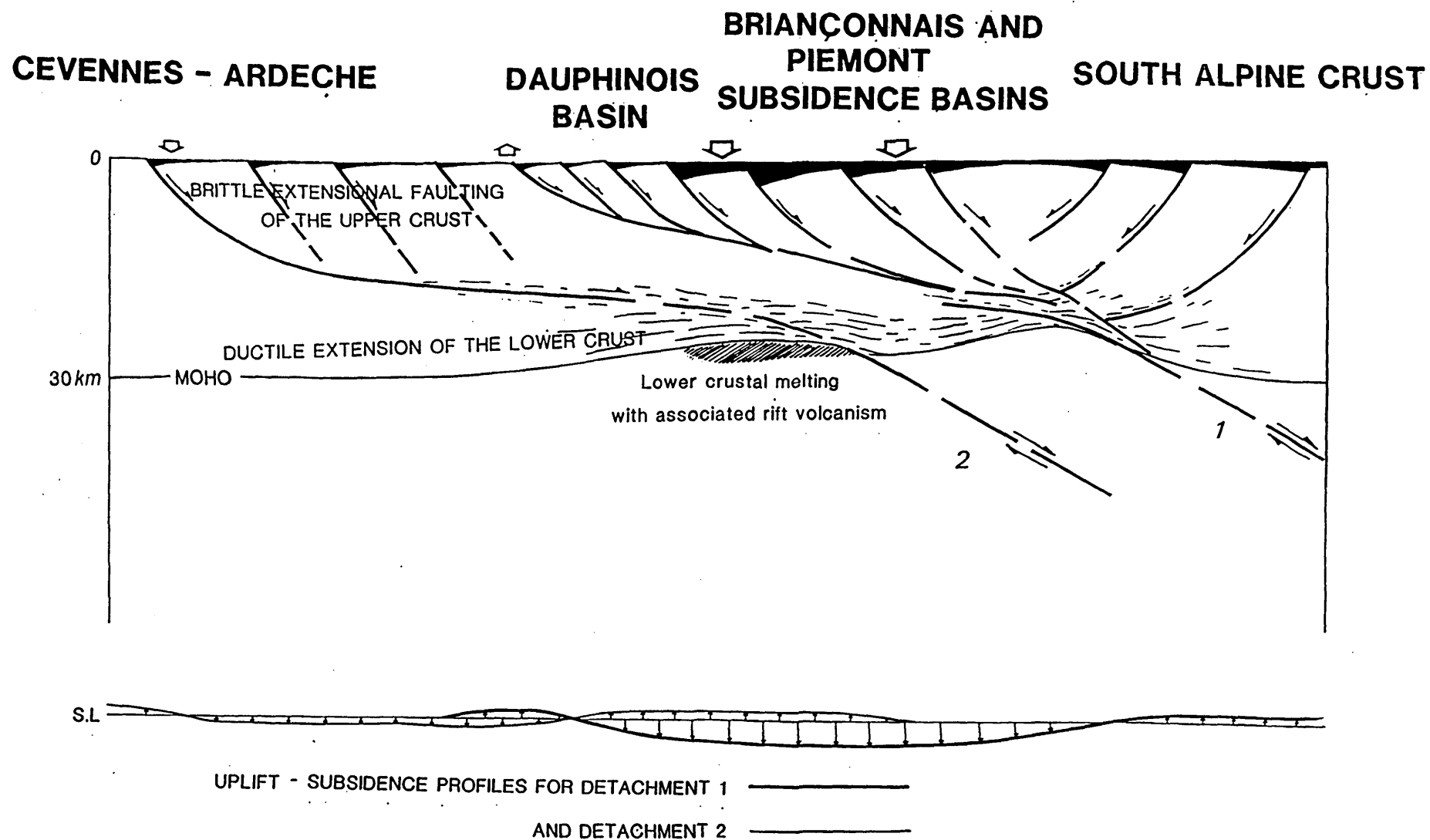


Figure 4.9: Model to show the suggested extensional fault geometry and differential thinning of the lower crust during the late Triassic to early Liassic. See text for discussion.

reached a level high enough to initiate lower crustal melting and the passage of these basic volcanics to the then basin margin (Pelvoux-Argentera axis) may reflect an inherent asymmetry in the extensional detachment system. The reduced sequence in the Dauphinois is interpreted to reflect the temporary shoulder of the developing rift, where the subsidence-uplift is controlled by thermal and flexural mechanisms (Beaumont et al., 1982).

The extension appears to have propagated away from the area of initiation (EVENT 1. ii) and upper crustal extension affects the Dauphinois domain. It is suggested that the amount of upper crustal extension associated with this is relatively high and that the faults of the Dauphinois were rotated into low dips.

Renewed extension was accommodated on a second detachment (No. 2, Figure 4.10) and the majority of the upper crustal stretching is transferred to the Dauphinois and Cevennes-Ardeche basins to the W (EVENT 2). The upper and middle crustal levels were thinned by a factor β , while the lower crust suffered little if any extension. To the E the lower crust and possibly lithospheric mantle beneath the already thinned upper crust of the Brianconnais, thinned by an extra factor β' , to balance the upper crustal extension in the Dauphinois and Cevennes-Ardeche basins. The extra thinning of dense lower crust, coupled with thermal heating resulted in the uplift and emergence of the Brianconnais basin, whilst the thinning of only the upper crustal levels in the Dauphinois and Cevennes basins, resulted in an isostatic subsidence.

Continued extension was accommodated towards the Piemont basin where continental break-up occurred and sea-floor spreading was initiated, forming the Ligurian ocean (Figure 4.11). This corresponded to EVENT 3 which signified the end of rifting and the beginning of thermal contraction of the European passive margin. In a McKenzie model of lithospheric stretching, the subsidence curves associated with the thermal contraction phase should should assume an

MID-JURASSIC EXTENSIONAL RIFT PHASE

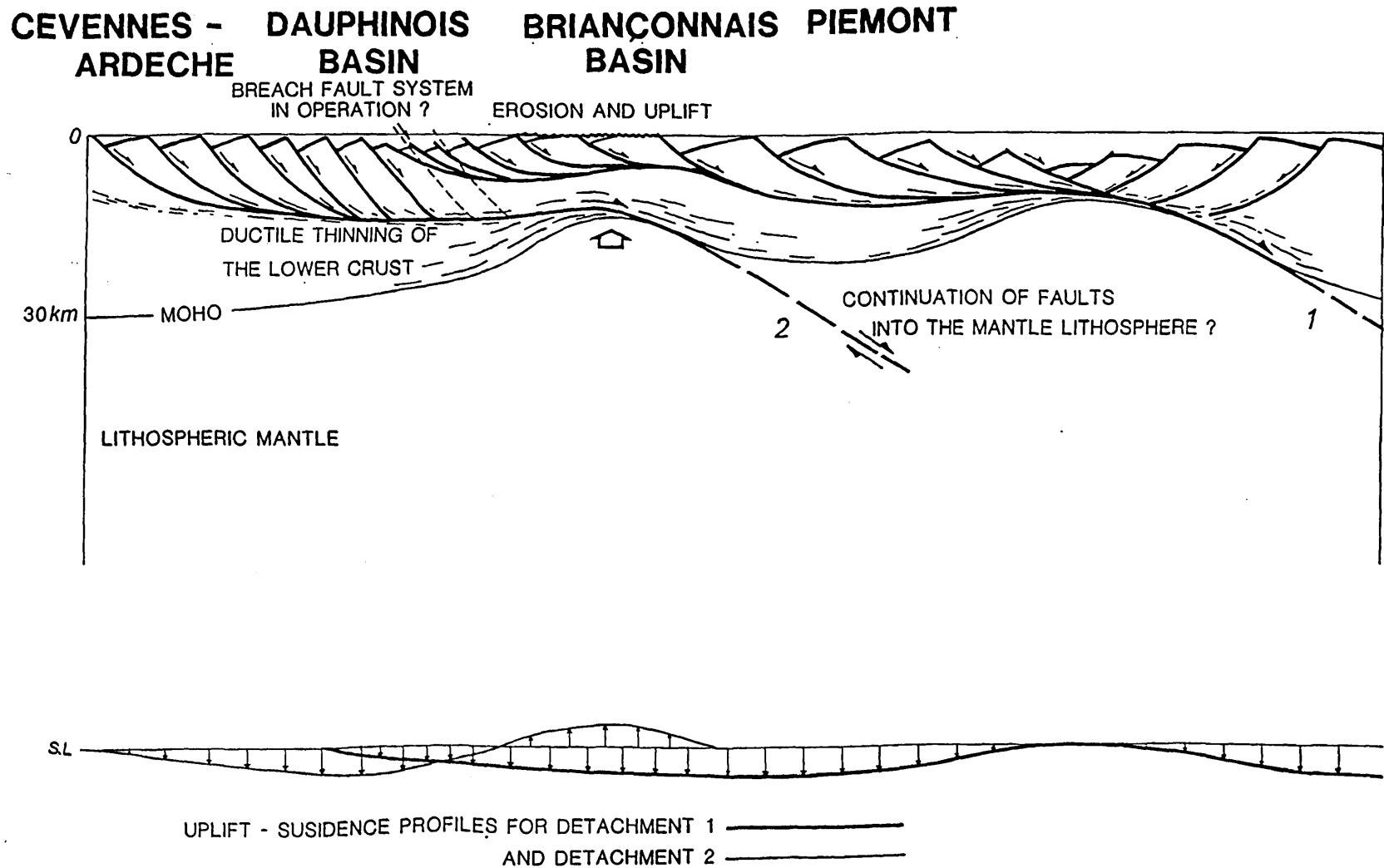


Figure 4.10: Model to show the suggested geometry of crustal faulting and thinning for the early to middle Jurassic. Note the uplift and erosion of the Briançonnais. See text for discussion.

LATE JURASSIC THERMAL SUBSIDENCE PHASE

INHERENT ASYMMETRY OF OPPOSING PASSIVE MARGINS ?

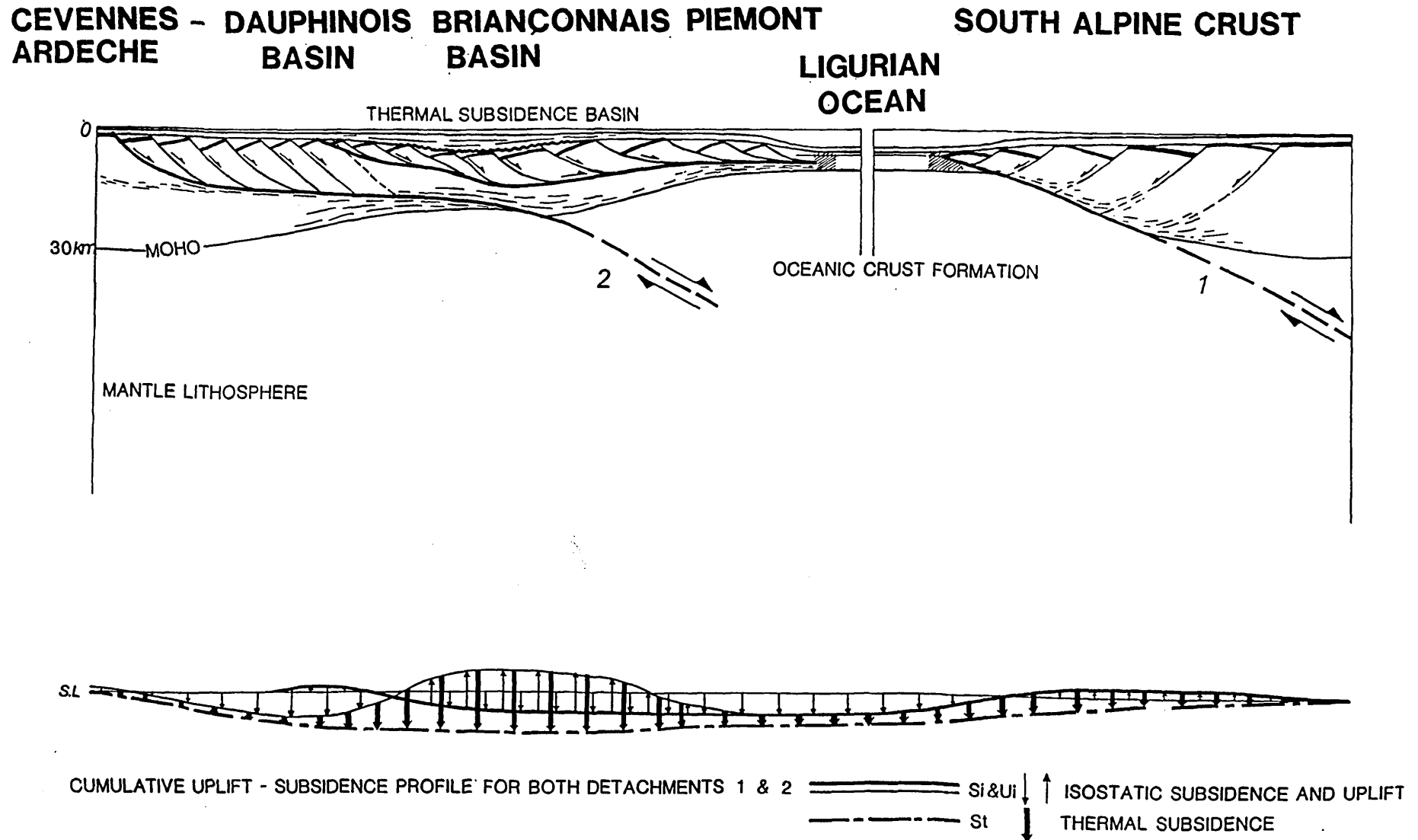


Figure 4.11: Model to show the late Jurassic to early Cretaceous thermal subsidence of the European Tethyan margin. Note the maximum subsidence occurs in the Briançonnais and Piemont domains. See text for explanation.

exponential form. This should hold true for any non-uniform stretching model due to the relatively high rates of heat loss sustained during the early stages of cooling. However unlike that of the McKenzie model, the form of this subsidence curve will also vary from area to area depending on the relative amounts of upper crustal thinning and superimposed lower crustal and lithospheric mantle thinning with associated heating. Areas such as the Briançonnais, where the whole crust has been thinned substantially can be expected to have subsided fairly rapidly during this phase. Similarly, a large thermal subsidence can be expected in the Piemont domain where the largest amount of lithospheric thinning and continental break-up occurred. Areas such as the Dauphinois and Cevennes-Ardeche can be expected to have subsided less rapidly and by smaller amounts. This model fits well with the observed data. The syn-rift and post-rift history of the European Tethyan margin is summarized in Figure 4.12.

Lemoine et al. (1986) pointed out that the end of rifting was not synchronous over all the European passive margin. If the Tethyan oceanic crust were formed in a series of rhombohedral pull-aparts, different ages for the formation of oceanic crust could be expected along the strike of the Tethyan rift system. Furthermore, it appears that the pattern of thermal subsidence across the European margin may have been disrupted by a Mid-Cretaceous extensional event. This has been termed the Atlantic phase (Graciansky, pers.comm., 1988) and relates to the opening of the N Atlantic. This extension which is well expressed in the Vocontian basin (Ricou & Siddans, 1986) and the Diois and Baronnies (Gerlier & Mascle, 1987), is interpreted to be the eastward continuation of crustal thinning associated with the formation of the Parentis basin and the Aptian-Albian Bay of Biscay opening (Pinet et al., 1987). However, increased early Cretaceous subsidence in the Valais and Helvetic Alpine domains to the N is difficult to reconcile with the extension responsible for the formation of the Parentis basin. Hsu (1982) explains this increased subsidence by a down bending of the

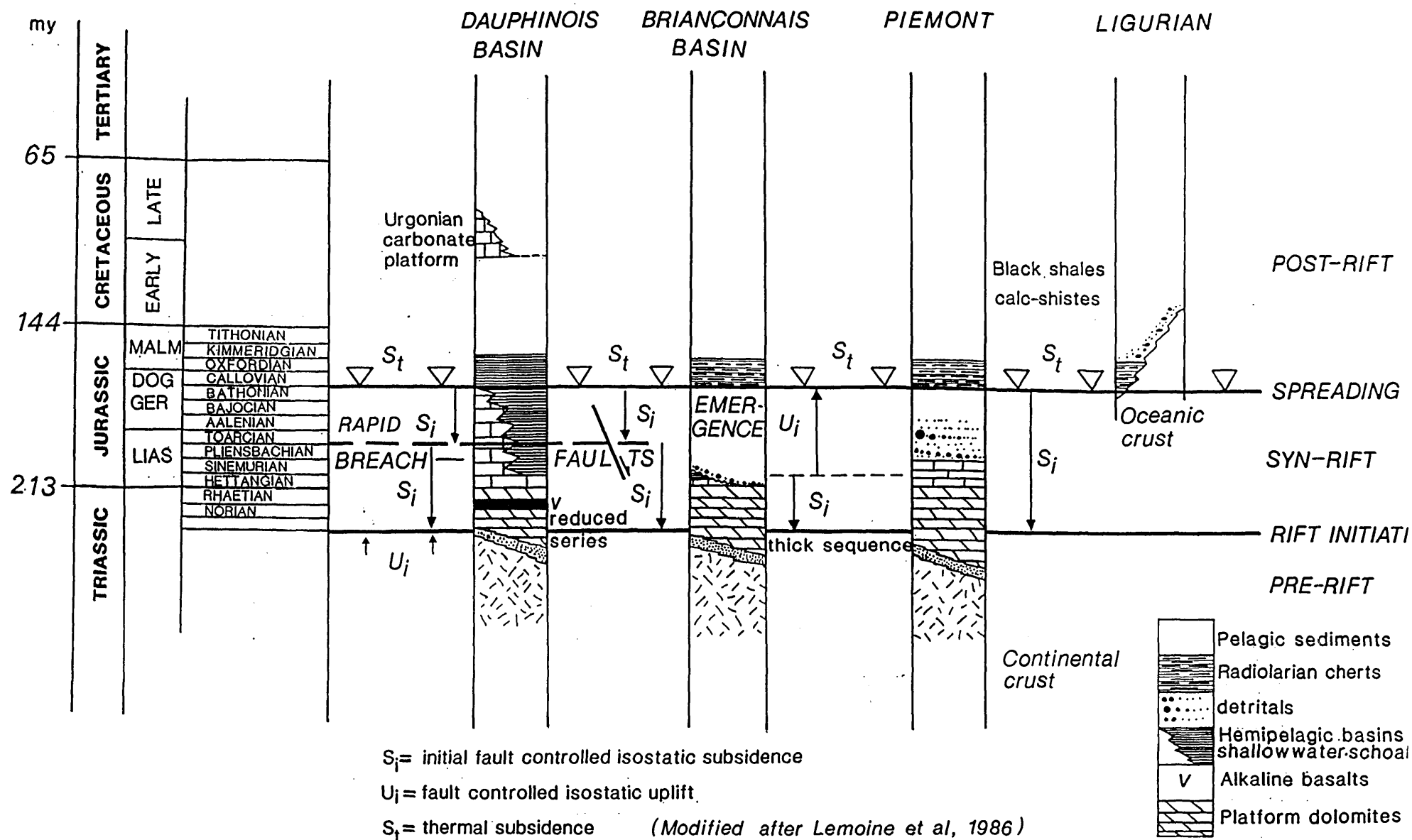


Figure 4.12: Simplified diagram to show the main extensional tectonic events and the accompanying sedimentary response across the European Tethyan margin. See text for discussion.

lithosphere, when the European passive margin changed to an active compressional one. Although alternative explanations exist, a detailed account of the effects and causes of an early Cretaceous extensional event is considered beyond the scope of this thesis.

4.6 Alternative hypotheses

The proposed model of Jurassic stretching is compatible with the subsidence-history of the European Tethyan margin and the geometry of upper crustal normal faulting inferred from the mechanical behaviour of these faults during the Alpine compression. However alternative stretching models may also explain the observed subsidence patterns and in particular, the subsidence anomaly of the Briançonnais uplift. Lemoine et al. (1986) propose a single westward dipping lithospheric-scale shear zone (see Wernicke, 1985) to explain the exposure of upper mantle rocks and the Briançonnais uplift. Equally, a model involving the non-conservation of volume and mass (gabbro-eclogite phase transformation), which involved the assimilation of lower crustal rocks by hotter mantle (Pinet et al, 1987), may explain the Briançonnais uplift. Which ever model is proposed, it appears that the uplift of the Briançonnais can be related to some deep seated thermal event, although the causes of this remain speculative.

In summary, the complex pattern of subsidence can be interpreted to be the result of a long and complex heterogeneous stretching process, rather than a single and simple homogeneous event.

4.7 Variscan inheritance

The positioning of some of the Tethyan normal faults may have been dictated by pre-existing Variscan structure. The geometry of faulting and fault kinematics affecting the European margin throughout the Mesozoic and Tertiary appears to have been strongly influenced by a number of fundamental basement faults, which may correspond to

older, late Variscan thrusts. Two orthogonal fault trends of N50 (Cevennes trend) and N140 transect the European margin (see Figure 1.2). These fundamental fault trends appear to have controlled much of the tectonic development of Central and SE France and in NW Europe may explain the apparent parallelism of movements from late Variscan time onwards (Zeigler, 1982; Beach, 1987). Stephanian movements on the inherited fault pattern resulted in a number of late Carboniferous to Permian age basins. The presence of what are now substantially shortened and telescoped Carboniferous and Permian age sediments further to the E in the present day Alpine thrust belt and on the Apulian margin of present day Italy, suggests that an extensive late Carboniferous basin system may have covered a large part of Western Europe, prior to Alpine contraction. Late Carboniferous extension associated with this basin forming event probably thinned the European crust substantially after Variscan orogenic thickening, prior to the onset of Mesozoic Tethyan extension. Any extensional model proposed must therefore take into consideration the possible effects of older structures on the final geometry of faulting.

A model of inversion, where the principal extensional detachment faults which controlled the extensional evolution of the European Tethyan margin are reactivated during Alpine compression, is presented in Chapter 6.

Chapter 5: Tectonic transport directions, timing and the Western Alpine arc

5.1 Introduction

The present position of the Pelvoux massif has for some time been central to a fundamental problem in the external French Alps; that is the relationship between NNE-SSW trending thrust structures with a WNW transport direction N of the Pelvoux and NW-SE trending thrust structures with a SW transport direction to its S. The structures of the Pelvoux mark the critical transition from one system to the other, and are therefore coincidental with an arcuate bend in the external Western Alpine thrust belt (Figure 5.1).

To date there has been no definitive work resolving the thrust geometry, timing and directions of tectonic transport in the Pelvoux and adjacent basement massifs, linking individual structures, or sets of structures around the Alpine arc. The aims of this Chapter are thus three-fold:

1. To unravel the timing and kinematics of thrusting in the Pelvoux and adjacent basement massifs.
2. To examine the continuity of structures and displacements through the Pelvoux massif and hence about the Western Alpine arc.
3. To uncover the origins and controls which governed the formation of the Western Alpine arc.

5.2 Tectonic movement directions in the Pelvoux and adjacent massifs

The geological criteria used to constrain the relative timing and thrust transport directions are discussed in the relevant sections of Chapter 2. This section will attempt to summarize the various directions and assess the

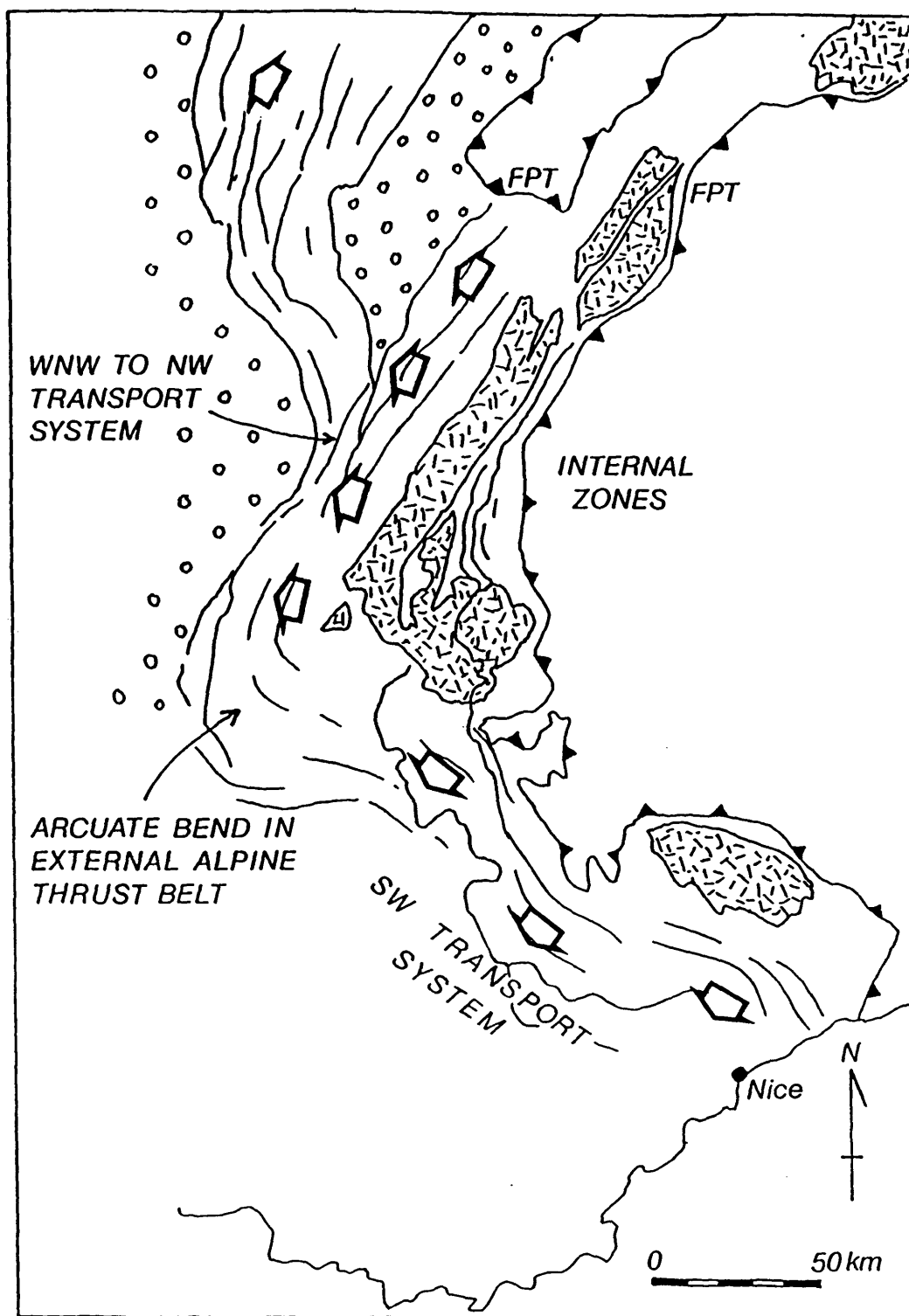


Figure 5.1: Simplified map of the Western Alpine arc showing the trend of the main structures and the two directions of thrust transport (SW and NW) in the external zones. The crystalline basement massifs and the molasse foreland basin sediments are shown.

displacement patterns within a regional tectonic framework. The movement directions recorded for individual basement thrust sheets, and packages of basement thrust sheets, are summarized in Figure 5.2.

5.2.1 Thrust transport directions in the basement:

In the northern Rochail, the Grandes Rousses and along the northern margin of the Pelvoux (Sections 2.1 & 2.2), the dominant direction of thrust transport is towards the WNW. This occurs irrespective of whether the basement thrusts evolved by reactivating pre-existing Jurassic extensional structures, or developed independantly, although there is evidence to suggest that steeply dipping pre-existing faults such as the Lac du Vallon fault (Section 2.1) may have influenced the direction of tectonic movement. Some basement thrusts such as those documented from the Deux Alpes section (Section 2.2) reactivate pre-existing passive margin faults, where as others such as the Cote Belle-Rochail thrusts (Section 2.1) are interpreted to slice through pre-existing Jurassic structures at depth. On the basis of geometrical arguments presented in Sections 2.2 and 2.3, the La Meije thrust is inferred to have an out-of-sequence or break-back geometry with respect to inversion structures in its foot-wall. The movement of the La Meije thrust sheet is towards the W and not the SW as suggested by Beach (1981). This westward direction of thrusting differs slightly from the direction recorded for the lower level structures, perhaps supporting the break-back interpretation.

In the southern Bourg d'Oisans and Rochail area a complication in the direction and sequence of thrust development is encountered. A WSW direction of thrust transport dominates and locally, a distinctive late thrust sequence can be demonstrated. Earlier SW directed basement thrusts decouple along the basement-cover interface of the Bourg d'Oisans half-graben (i.e. La Paletas, Section 2.1). These earlier structures, where observed, conform to reactivated Jurassic structures and are themselves

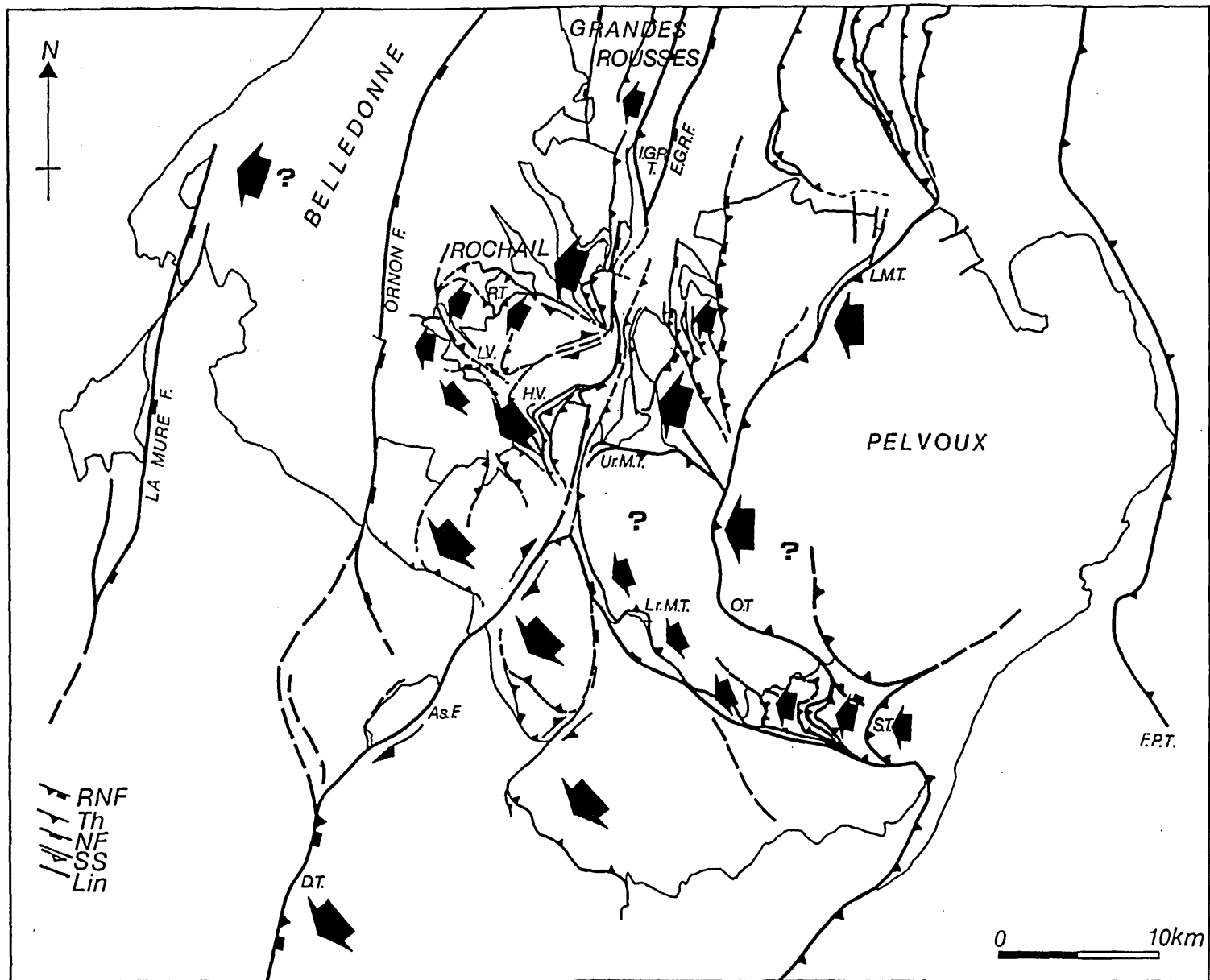


Figure 5.2: The distribution and directions of thrusting recorded for basement thrust sheets in the Pelvoux and adjacent massifs. Arrows indicate the direction of overthrusting. Legend, RNF-reactivated normal fault, Th-thrust, NF-normal fault, SS-Strike-slip fault.

truncated or sliced through by higher level basement thrusts belonging to a later WSW directed system, well exposed in the Haute Vallon Valsenestre and the Signal de Lauvitel (Section 2.3). These WSW directed structures of the higher level thrust system can be traced north-eastwards into the Lauvitel (Section 2.1 and 2.3) and Les Deux Alpes area, where their continuity is lost. This suggests that locally, dextral strike-slip movements may occur on the Eastern Grandes Rousses fault, however no evidence was found to support this. The age relationship between the WNW directed thrusting described from the N and the high level WSW directed thrusting is uncertain. They may be of similar age with the WSW thrusting being restricted to the southern area. On the basis of structural styles it could be argued that the inversion structures of this area pre-date higher level basement thrusts which slice through the earlier inversion structures. The swing in structural trend in the southern Bourg d'Oisans area (Section 2.3) may be inherited, reflecting an initial trend change of the pre-existing Jurassic structures.

In the Lanchatra area of the central Pelvoux (Section 2.3), it was argued that higher level thrusts divide the area into two distinctive structural domains: a structurally lower domain characterized by inversion geometries and a structurally higher domain of undeformed granites which may have been emplaced as part of the same break-back thrust sequence to which the La Meije thrust is attributed. The thrust transport direction proposed for this area is towards the WNW.

It is the central Pelvoux (Section 2.3) where the most complex set of relationships exist. This is interpreted to be partly due to the presence of pre-existing Jurassic faults. One such fault, the Col de la Muzelle fault, forms a continuous structure through the Pelvoux, linking with the Pied-Moutet fault to the N (Section 2.2) and the Aspres-les-Corps fault to the SW (Section 2.5). A dextral strike-slip movement is associated with this fault and as no such movement was documented along the Pied-Moutet or

Eastern Grandes Rousses faults to the N, this pre-existing fault appears to have been responsible for the partitioning and channelling of displacements towards the WSW. The relationship of this fault with the higher level WSW directed thrusts of the Vallon Valsenestre is uncertain - at depth they may slice through this fault or branch off it.

Another critical structure which is continuous through the central Pelvoux is the Lower Muzelle thrust and its associated hanging-wall anticline. Where this thrust has climbed into the overlying cover sequence towards Les Deux Alpes (Section 2.3) its thrust transport direction is difficult to determine. In the La Muzelle area, from scant evidence, a WNW transport interpretation was proposed. However, a WSW transport interpretation may equally suffice and is consistent with the thrust transport direction obtained farther S at Villar-Loubiere (Section 2.3) and the Aiguille de Morges area (Navette, Section 2.4). The arcuate trend of the anticlinal trace (Figure 5.2) may equally reflect or mimic the pre-existing extensional fault trend, as could the variation in thrust transport direction. Although there is evidence to suggest that the Col de La Muzelle fault corresponds to a Jurassic structure and its eastern branch, the le Desert fault may be interpreted as a transfer fault to the Col de la Muzelle fault (Figure 5.2 & 5.9), there is no direct evidence to suggest that the Lower Muzelle thrust itself is a reactivated Jurassic extensional structure.

The other principal structure which is continuous through the Pelvoux massif is the La Meije-Olan thrust. A westward direction of thrust transport is attributed to this fault although anomalous directions are documented from the Olan region (Section 2.3). To the S in the Aiguille de Morges area (Section 2.4), this structure branches with the Sirac basement thrust which is interpreted to have a westward direction of tectonic transport. Higher level thrust structures to the E of the Meije-Olan thrust were not studied.

In the Aiguille de Morges area (Section 2.4), a study of the kinematic indicators and thrust geometries imply that a progressive change in thrust transport direction may occur at deeper levels in the structural pile. For example the movement direction of the Olan and Sirac thrusts is towards the W. At lower structural levels, represented by the basement thrust sheets of Navette, a WSW direction of thrust transport prevails. This direction is confirmed from the continuity of these structures with the Lower Muzelle thrust. If a progressive anti-clockwise rotation of the transport direction does occur, the initial direction of emplacement of the La Meije-Olan thrust sheet may have been towards the WNW, and kinematic indicators associated with this WNW emplacement would be subsequently rotated anti-clockwise about a vertical axis giving their present E-W orientation.

On the SW margin of the Pelvoux all of the structures mapped form part of a WSW to SW transport system, with no evidence for WNW directed movements (Section 2.5). Pre-existing NE-SW trending Jurassic normal faults impose important constraints on the direction of tectonic transport.

5.2.2 Thrust transport direction of the sedimentary cover:

The displacement direction of the detached cover in the area studied is much more difficult to ascertain. This is largely due to the presence of pre-existing Jurassic normal faults (see Chapter 2) which displace the basement-cover interface, off-setting potential decoupling horizons within the syn-rift sequence. These faults can therefore be considered as heterogeneities which impede the forward propagation of the fold-thrust system. The resulting strain patterns, geometries and structural styles affecting the cover sediments, to a large extent, reflect the geometry of these pre-existing extensional structures. As described in Chapters 2 and 3, locally upright deformation regimes occur adjacent to pre-existing normal fault scarps, so that cover strain patterns reflect the spatial distribution of pre-

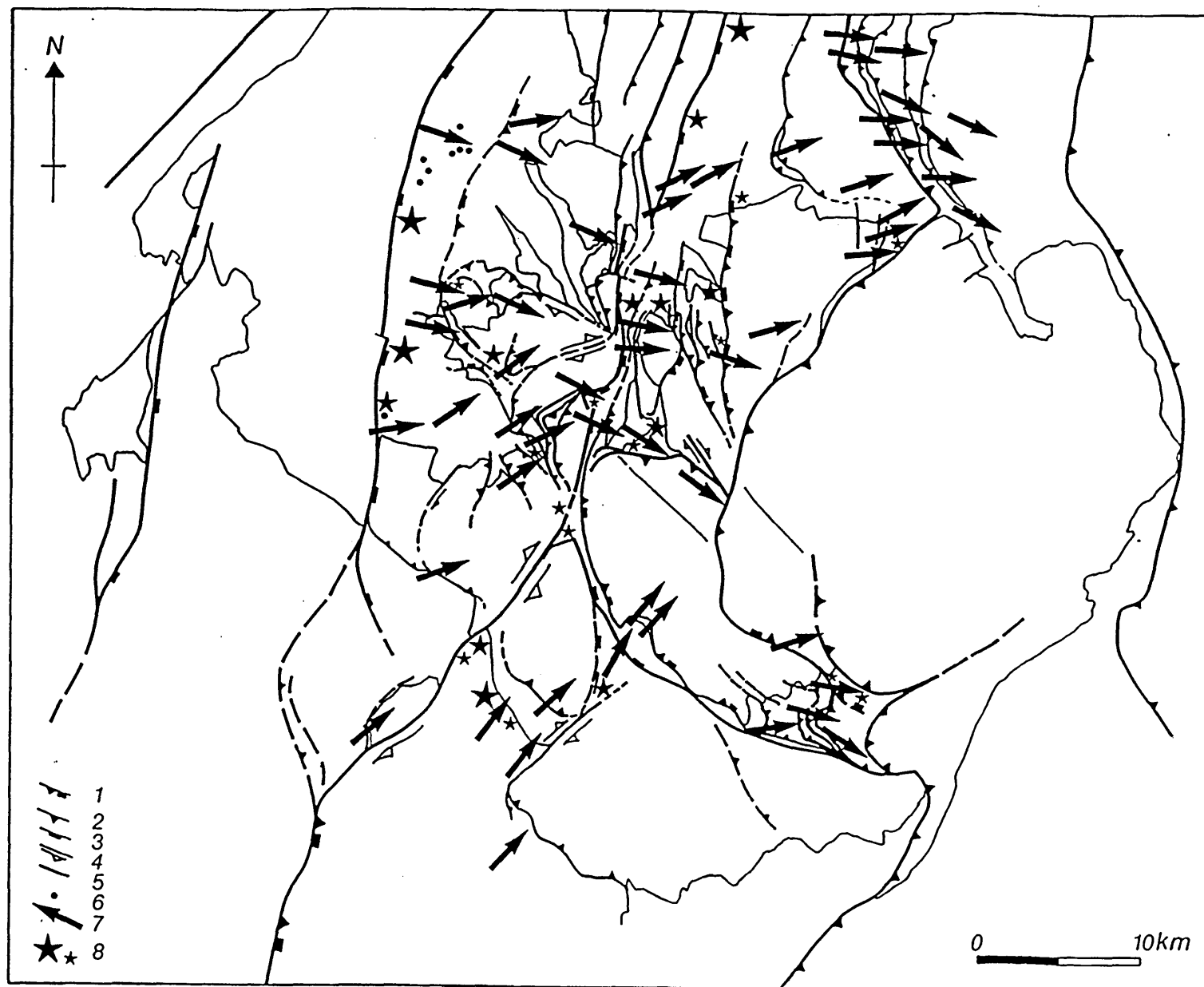


Figure 5.3: Structural map of the Pelvoux showing the distribution of slip-lineations from basement-cover thrusts and mineral stretching lineations in the sedimentary cover. Legend, 1. reactivated normal fault, 2. thrust, 3. normal fault, 4. strike-slip fault, 5. fault, 6. vertically plunging mineral lineation, 7. lineation plunge direction, 8. olistoliths.

existing extensional faults rather than that expected of simple foreland fold-thrust system.

However, along the northern margin of the Pelvoux where the cover is decoupled from its basement and forms the Ultra-Dauphinois thrust system (Beach, 1981a & b), the orientation of the finite stretching direction throughout the sediment/thrust sheet pile is variable. Close to the basement-cover interface the finite stretching direction orientation is ENE-WSW. This changes to a WNW-ESE orientation at higher structural levels. This change in is shown in Figure 5.3.

Along the southern margin of the Pelvoux, Chambers (1987) documents fold and ramp orientations and kinematic indicator orientations that are consistent with a NW direction of tectonic transport for the Gres du Champsaur flysch sequence. The regional significance of these cover transport trends will be discussed later.

5.3 Timing of the deformation

5.3.1 The Pelvoux and adjacent massifs:

In the absence of syn-tectonic foredeep sediments and syn-tectonic unconformities, it is difficult to date conclusively many of the thrust structures in the Pelvoux and adjacent massifs. However, at the southern and north-eastern margins of the Pelvoux movements on some of the thrust structures can be dated with respect to the deposition of Eocene to Oligocene foredeep flysch sediments - the Aiguille d'Arves flysch to the NE, the Gres du Champsaur flysch to the S and the Gres de St. Disdier in the Devoluy. By extrapolation, either from direct continuity with dated structures, or on the basis of structural styles, it is possible to make certain inferences concerning the timing of the deformation in the Pelvoux.

The stratigraphy of the foreland basin flysch sequence

can be divided into five distinctive stratigraphic units. These are in ascending order, a basal continental facies, a transgressive shelf carbonate (Calcaires Nummulitique), Globigerina bearing marls (Marnés à Globigerinés), a turbiditic sandstone sequence (Grés du Champsaur) and a tectonised sedimentary melange (Schistés à Blocs). The Calcaires Nummulitique is dated as Priabonian (Debelmas, 1975). The timing of the deformation affecting the Pelvoux massif can thus be broadly be divided into pre- and post-Nummulitique events.

1. Pre-Nummulitique deformation:

Along the southern and eastern margins of the Pelvoux massif, the Calcaires Nummulitique can be observed to rest unconformably on Jurassic limestones and crystalline basement. Chambers (1987) demonstrated that the basal Eocene conglomerate constitutes a plane of no tectonic slip and therefore corresponds to a preserved sedimentary contact. This relationship is testimony to a pre-Priabonian uplift and erosion of the Pelvoux basement.

Thrusts at the SW margin of the Pelvoux were shown to pre-date the Nummulitique transgression (Sections 2.5 and 2.6). Previous workers correlate this pre-Nummulitique uplift and erosion of the Pelvoux basement with pre-Senonian deformation occurring in the Devoluy (Barbier, 1948; Goguel, 1963; Gidon, 1979; Siddans, 1979). The WSW tectonic transport direction recorded for pre-Nummulitique basement thrusting in the SW Pelvoux (Section 2.5) is difficult to reconcile with the N-S pre-Senonian compression of the Devoluy. However, the possibility that these two events are one in the same, cannot be ruled out. If they are not, the pre-Senonian deformation of the Devoluy may still have some as yet unrecognised expression in the Pelvoux.

Along the NE margin of the Pelvoux, evidence in support of a pre-Nummulitique event involving the Combeynot basement has previously been discussed in Section 2.2.

Bravard & Gidon (1979), and Gidon (1979), assume the direction of pre-Nummulitique transport was towards the N or NNW, correlating this with the "Arvinche" deformation (Barbier, 1948). However the geometry of thrusting is difficult to reconcile with a northerly direction of thrust transport. The main pre-Nummulitique transport direction may have been towards the WNW, and locally towards the SW in the southern Pelvoux. The reason for two different directions of tectonic transport may become evident as the text proceeds.

From these relationships it is tentatively suggested that the Alpine deformation of this area commenced prior to the transgression of the Calcaire Nummulitique (Priabonian - 42 ma) and corresponds to an Early Tertiary Alpine event which involved the thrusting and subaerial exposure of basement. This is consistent with and may also be correlated with the pre-Upper Lutetian (45 ma) Alpine deformation documented by Elliott et al. (1985), to the S in the Haute Provence and Grés d'Annot sector of the Alpine foreland basin.

Clearly future work is required to define further the geometry and timing of this Early Tertiary Alpine event. This could be accomplished from combined structural and sedimentological studies of the Eocene-Oligocene flysch series to determine the effect of pre-existing structures on flysch sedimentation patterns in the foreland basin.

2. Post-Nummulitique deformation:

The main Alpine deformation in this area is considered to be late Oligocene to early Miocene in age and therefore post-dates the deposition of the Grés du Champsaur and Aiguilles d'Arves flysch.

Along the northern margin of the Pelvoux, the cover shortening associated with the Ultra-Dauphinois thrust system is interpreted to have developed in response to the W or WNW directed overthrusting of the La Meije basement.

The stacking of these cover thrust sheets may well be responsible for the eastward steepening of the basal unconformity of the Aiguille d'Arves flysch. Indeed to the N, Parish (1986) has demonstrated that the cover thrusts of the Ultra-Dauphinois thrust system post-date the deposition of the Eocene flysch. The displacements associated with the Ultra-Dauphinois thrust system and the La Meije thrust can be traced through the central Pelvoux via the Olan thrust into the Aiguille de Morges area, where they are transferred onto the Sirac thrust by courtesy of a branching between the Olan and Sirac thrusts. To the S of the Pelvoux, the Sirac thrust and cover thrusts in its foot-wall breach the Priabonian unconformity, advocating a post-Eocene age interpretation. The thrust structures described from the central Pelvoux such as the Lower and Upper Muzelle thrust faults are interpreted to be part of the same break-back thrust sequence to which the Meije-Olan-Sirac thrusts belong and therefore post-date the deposition of the Eocene flysch. It is also possible to demonstrate that the dextral strike-slip movements along the Aspres-les-Corps fault, which is interpreted to be a lateral structure to the Devoluy thrust, are post-Oligocene age because the Devoluy thrust overrides Oligocene molasse in its foot-wall (Section 2.6).

The age of structures in the Rochail and Valsenestre area is more difficult to determine in the absence of syn-tectonic sediments. However it could be argued on the basis of structural styles that the higher level break-back thrusts are the same age as the Meije-Olan-Sirac system and the structures which they truncate, which tend to display inversion type geometries, are pre-Nummulitique in age. However this cannot be confirmed and these inversion structures could equally belong to a complex sequence of post-Nummulitique inversion and thrusting. The inferred distribution of post and pre-Nummulitique thrusts and inversion structures and those of indeterminate age, is shown in Figure 5.4.

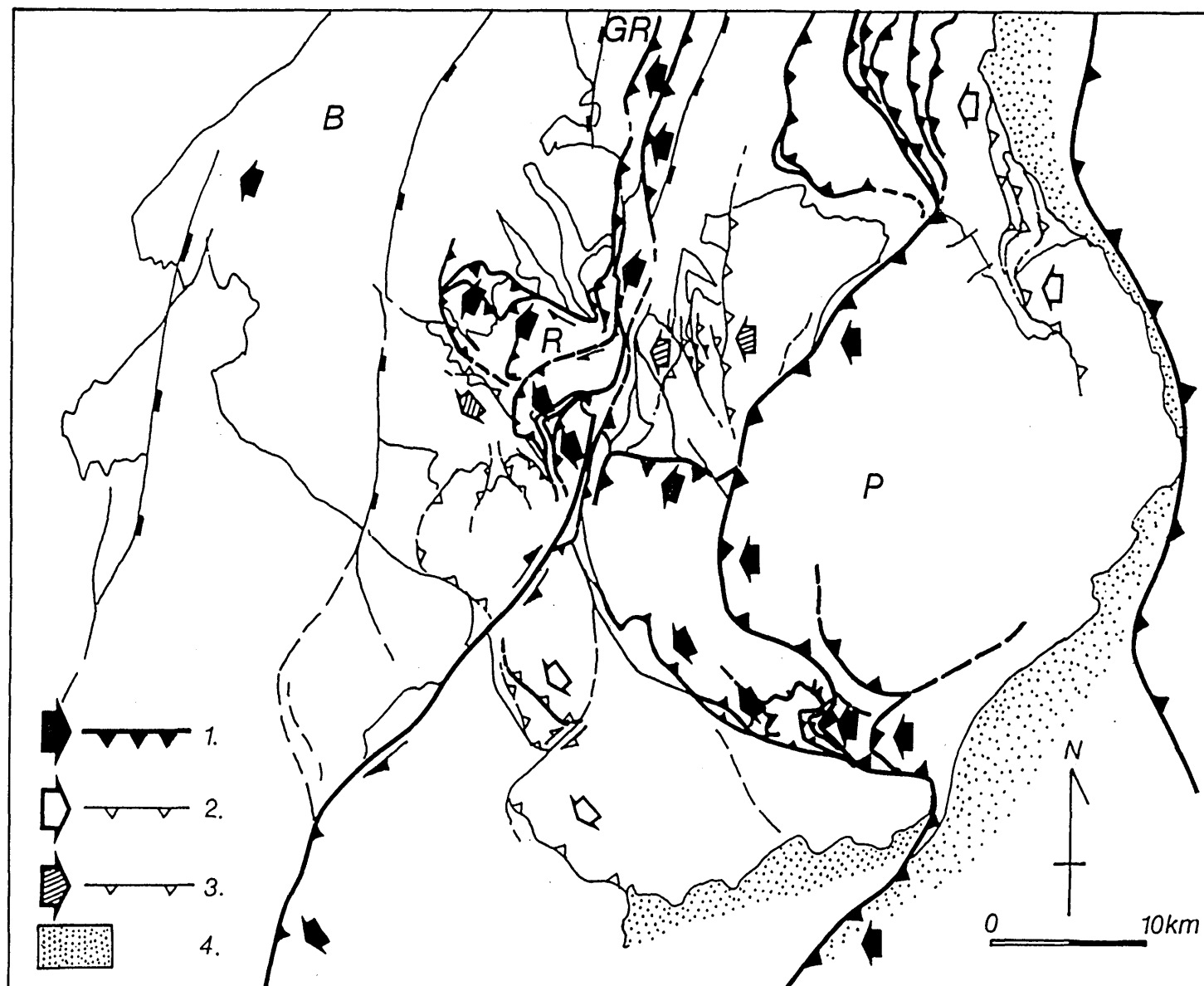


Figure 5.4: Shows the distribution and direction of thrust transport for pre- (2) and post- (1) Nummulitique structures, and structures of unknown age (3), Eocene-Oligocene flysch (4), P-Pelvoux, B-Belledonne, R-Rochail, G.R-Grandes Rousses.

5.3.2 The Subalpine Chains:

Both the origin of its arcuate shape, and the age of the structures in the Subalpine Chains of SE France, have over the years been the subject of intense debate. The following account is therefore largely based on previously published work.

The timing and emplacement direction of the Belledonne basement massif was not addressed in previous sections, largely because these aspects are constrained by the deformation in the adjacent Subalpine Chains.

In the Subalpine Chains of SE France, three principal ages of structure have been documented (Siddans, 1979). These are:

1. In the Devoluy Cenomanian limestones are folded and unconformably overlain by Upper Senonian limestones. These folds correspond to the "ante Senonian" folds of Lory (1984), Barbier (1963), and Gidon (1979) and are commonly referred to as pre-Senonian folds (Siddans, 1979).
2. In the Baronnies (Flandrin, 1966) and the Arc de Digne (Graham, 1978), a set of E-W trending structures fold Upper Senonian limestones and pre-date the deposition of the Calcaire Nummulitique. These correspond to the "plis anciens" of Goguel (1963) and were attributed to a Pyrenean-Provençal phase of northward directed compression (Lemoine, 1972).
3. Late Oligocene-Miocene folds and thrusts (the main Alpine phase) in the Vercors and Devoluy are dated by molassic sediments (Section 2.6). Siddans (1979) suggests that the fold and thrust structures of the Vercors are continuous with those of the Devoluy.

Whether the pre-Nummulitique structures of the Pelvoux form an integral part of an E-W trending Pyrenean-Provençal thrust belt (see Barbier, 1948; Goguel, 1963; and

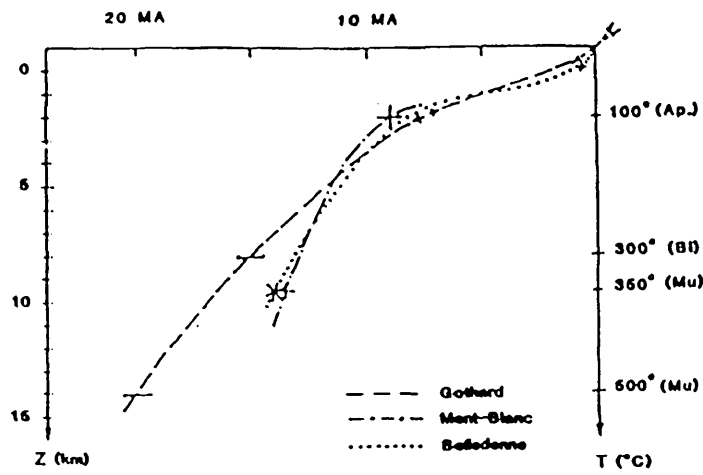


Figure 5.5: Radiometric data on the cooling and uplift of the external crystalline massifs during the end Oligocene and Miocene. After compilation by Mugnier & Menard (1986).

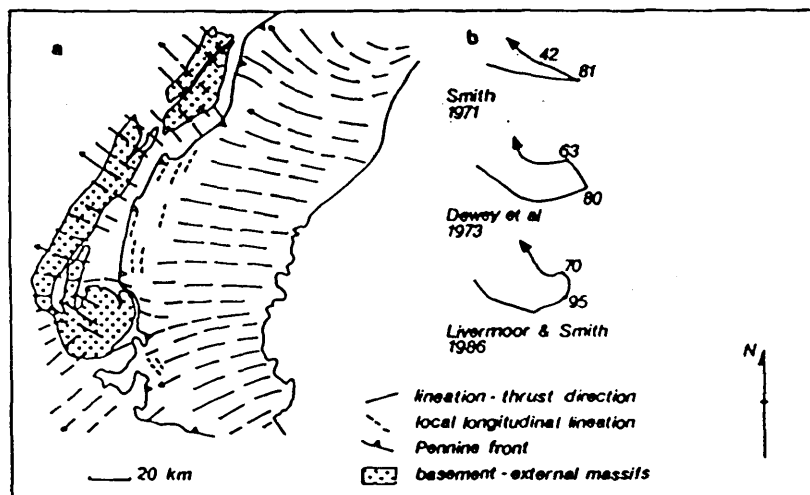


Figure 5.11: a) Lineation directions in the Western Alps, after Malavielle et al. (1984). b) Relative motions of Africa with respect to Europe. See text for discussion.

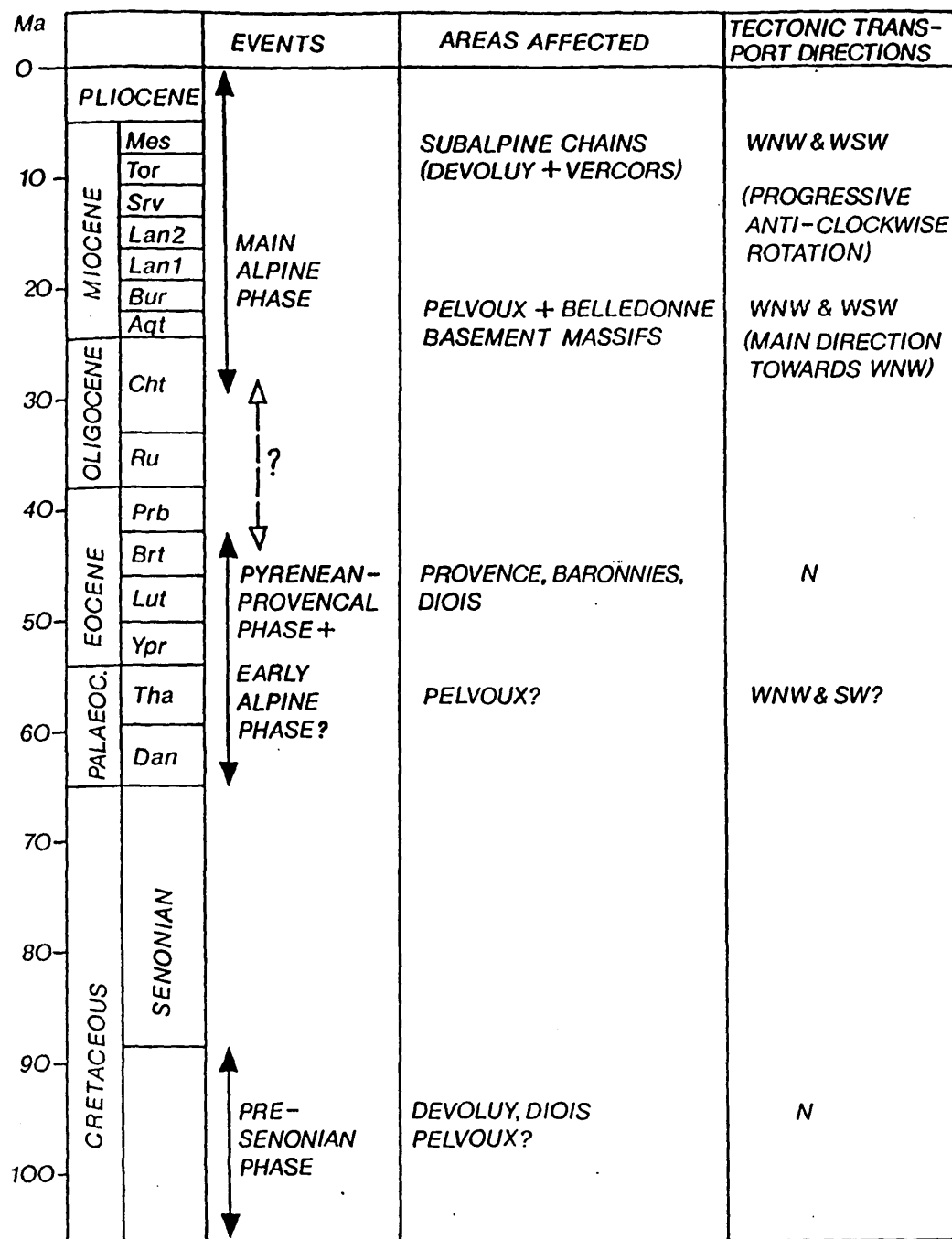


Figure 5.6: Timing of the main tectonic events in the Pelvoux and adjacent Subalpine Chains. See text for discussion.

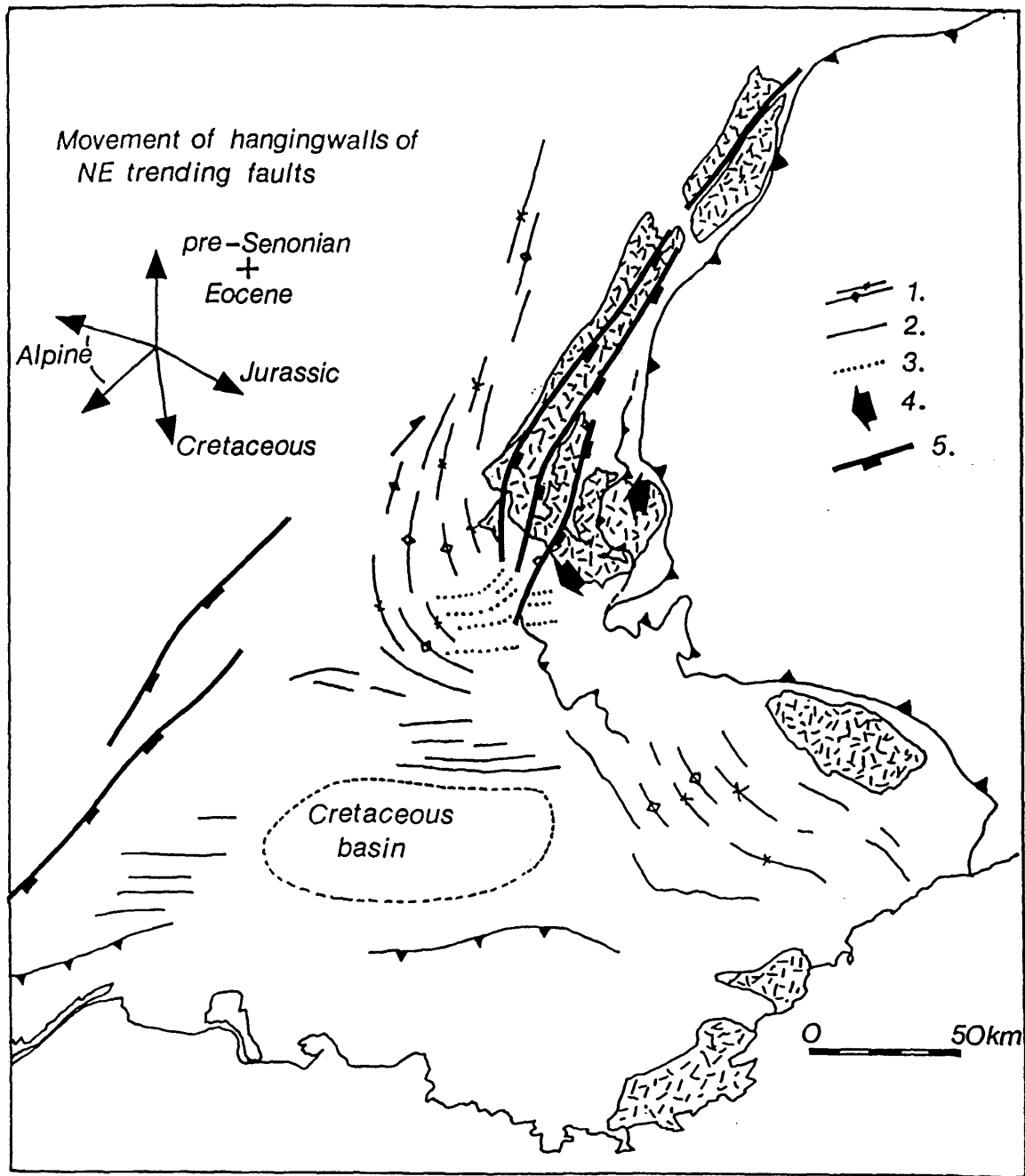


Figure 5.7: Shows the tectonic transport directions for the main tectonic events in SE France and the external French Alps, 1. Miocene-Pliocene fold-thrusts, 2. Eocene fold-thrusts, 3. Pre-Senonian folds, 4. pre-Nummulitique deformation in the Pelvoux, 5. Jurassic extension.

Siddans, 1979), or reflect a superimposed belt of deformation related to early continental collision, is difficult to ascertain. If we cannot invoke a separate pre-Nummulitique deformation of the Pelvoux, given the direction of tectonic transport recorded, the postulate of a syntaxial bend in the Pyrenean-Provencal thrust belt must be considered.

Menard (1979) and Thouvenot & Perrier (1980), on the basis on geophysical data, proposed that the external Belledonne basement was detached along a crustal scale thrust and emplaced towards the NW to WNW. Assuming that the emplacement of the massif induced a decoupling of the Subalpine cover in its foot-wall, this thrust can be broadly dated as late Oligocene to early Miocene. This age interpretation is further refined by radiometric dating of the uplift of the external basement massifs (Schaer & Jeanricard, 1974; Mugnier & Menard, 1986). From curves which show the cooling and uplift history of the external basement massifs (Figure 5.5), assuming that this uplift is related to the movement of the Belledonne basement up and over a crustal ramp, thrusting should have been initiated before 15-20 ma as uplift commenced before this date (Mugnier & Menard, 1986).

The timing of tectonic events for the external French Alps in SE France and the associated tectonic transport directions, are summarized in Figures 5.6 & 5.7.

5.4 Continuity of structures and displacements through the Pelvoux massif and around the Western Alpine arc

From the work outlined in this thesis it is possible to trace, with some degree of certainty, the continuity of individual thrust structures, or assemblages of thrusts through the Pelvoux massif. The aim of this section is to briefly outline these continuous relationships, revealing the missing link between the WNW and WSW/SW thrust systems of the Western Alpine arc. The displacement trajectory paths of the deformation will be discussed in Section 5.6.

5.4.1 La Meije-Olan-Sirac thrusts:

The branching arrangement of these thrusts creates a continuous linked system through the Pelvoux and is responsible for the transference of large post-Nummulitique displacements from the Ultra-Dauphinois thrust system in the N, to the Provence thrust system in the S. Along the southern margin of the Pelvoux massif the southward continuation of the Sirac system is observed in the Meollion valley. The thrust displacements from this may ultimately transfer to a roof thrust which exploits the vast thickness of Oxfordian black shales exposed in the hanging-wall of the Digne thrust system (Figure 5.8). The W-direction of thrust transport for the Sirac, Olan and La Meije thrusts may still be operative along this roof thrust, locally contradicting the SW transport interpretation proposed by workers in Haute Provence. However, a progressive anti-clockwise rotation of the tectonic transport direction to the WSW has already been documented for lower level basement structures in the southern Pelvoux, and this is compatible with the SW direction recorded for the southern Subalpine Chains.

5.4.2 Aspres-les-Corps fault - Devoluy thrust:

While the Aspres-les-Corps fault is continuous through the Pelvoux massif, the post-Nummulitique SW displacements recorded along its southern part where it passes into the Devoluy thrust, are not. The pre-existing NE-SW trending Aspres-les-Corps fault appears to have directed displacements towards the WSW, during the main W to WNW directed Alpine compression. Indeed pre-existing faults of this orientation may have partly caused the rotation and WSW transport direction documented for the southern Pelvoux (see later).

5.4.3 The external Belledonne thrust:

Assuming the external Belledonne basement thrust to be partially responsible for the elevation of the Pelvoux and

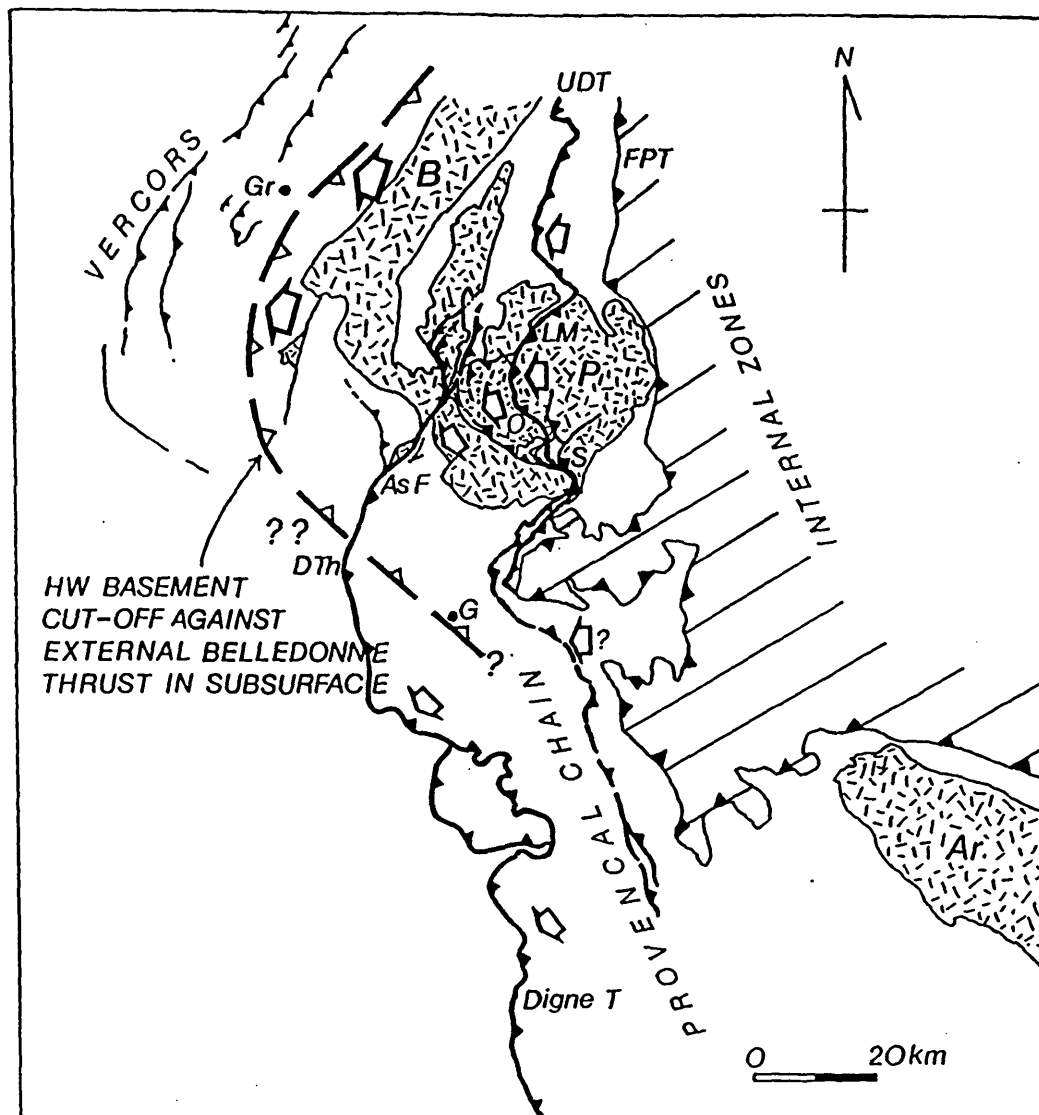


Figure 5.8: Continuity of Oligo-Miocene structures through the Pelvoux massif and around the Alpine arc. Arrows indicate direction of Alpine thrusting, Ar-Argentera, P-Pelvoux, B-Belledonne, S-Sirac, O-Olan, LM-La Meije, AsF-Aspres fault, DTh-Devoluy thrust, Gr-Grenoble, G-Gap. See text for details.

Belledonne massifs, its subsurface geometry and contribution to the arcuate shape of the Western Alpine thrust belt is difficult to determine, largely because it is a geophysically inferred fault (Menard, 1979, 1980; see also Thouvenot & Perrier, 1980), although recently imaged on the ECORS-CROP (Bayer et al., 1987) deep crustal seismic reflection profile. The deep geometry of this thrust can however be guessed at from the continuity of cover fold and thrust structures in the adjacent Subalpine Chains (Figure 5.8). Gratier et al. (1973) and Siddans (1979) demonstrate the continuity of main Alpine phase deformation (cleavage, thrusts and folds) from the Digne thrust sheet NW into the Vercors, showing them to cut across earlier E-W trending structures of the Diois, or partially reactivate them. This suggests that the external Belledonne basement thrust is also continuous around the Alpine arc. However, its relationship to the SW directed basement thrusting on the southern margin of the Pelvoux remains uncertain. A number displacement path trajectories are possible for the Pelvoux and Belledonne basement massifs (see later), but a detailed three-dimensional restoration (map and section) of the adjacent Subalpine Chains is necessary before these can be accurately constrained.

5.5 Origin of the Western Alpine arc

5.5.1 Arcs and oroclines

In many mountain belts, segments of the belt are curved or bent in plan view. The Western alpine thrust belt is similar and major fold axes and thrusts are curved forming tangents to a nearly 90° arc (Figure 5.1).

Cary (1955) distinguished two types of bend: primary bends (nonrotational arcs) which are initiated with their present curved form and secondary bends, or oroclines (rotational arcs) where a later strain is superimposed on a previously linear belt. There should be a progressive variation in magnetic declination associated with oroclines. However, as pointed out by Marshak (1988), in

Carey's terminology an orocline does not include bends that developed coevally with the structural grain of the belt, and yet may show palaeomagnetic evidence for rotation. Following the terminology of Marshak (1988), for nonrotational arcs, segments of the belt do not change strike during its development, and for oroclines, segments of the belt do change strike during or after its development.

Compressional deformation along non-linear continental margins, indentors (Tapponier, 1977) and the compressional reactivation of an inherited right angle bend in a pre-existing fault geometry, will all lead to the formation of primary nonrotational arcs. Interaction between a thin-skinned fold and thrust belt and obstacles on the foreland (the stratigraphic pinch out of a glide horizon, or pre-existing basement fault scarps of limited lateral extent) may lead to the formation of oroclines (Marshak, 1988). In the Western French Alps it is proposed that the main arc is of a primary nonrotational origin and that secondary oroclines affect the thin-skinned parts of the thrust belt i.e. the Subalpine Chains and the Jura. The displacement path trajectories are necessarily complex to accommodate apparent changes in tectonic transport direction around the arc. This includes the development of a strong anti-clockwise rotational component to the deformation, which may explain late tangential extension expressed as dextral strike-slip movements along pre-existing basement faults in the foreland.

5.5.2 Other models:

Argand (1916; 1924), one of the pioneers of Alpine tectonics, was the first to address the difficult problem of tectonic transport directions and the origin of the Western Alpine arc. He proposed a plastic flowage of the internal zones, radiating outwards from an indenting Adriatic promontary, generating a radially outward verging fold and thrust pattern. This model was later contested by Goguel (1963) who recognised that the restoration of

radially diverging thrust nappes along the proposed displacement trajectory paths generates serious space problems, locating all the nappes simultaneously in the Po Plain! (see also Ricou & Siddans, 1986). Such a model also requires large amounts of arc parallel extension, which to the best of my knowledge, has not been documented in the Western Alps. Goguel's alternative model was that for most of its early history, the Western Alps was a sinistral strike-slip zone accommodating a northward displacement of Africa (Adriatic Promontary) relative to Europe, a model later reiterated by Ricou & Siddans (1986). Goguel (1963) suggested a gravity mechanism to explain the general westward emplacement of higher level thrust nappes in the Western Alps and Ricou & Siddans (1986) invoke a late "hypercollision". Attempts to discredit this model, or more precisely the model proposed by Ricou & Siddans (1986), are largely based on studies of small scale kinematic movement indicators in the Western Alps (Malavielle et al., 1984; Choukroune et al., 1986), and the elusive evidence for early N-directed sinistral strike-slip displacements (see Butler et al., 1986). However, pre-Senonian structures and "Pyrenean" early Eocene structures indicative of early N-S compression are in evidence throughout most of SE France (Goguel, 1963; Siddans, 1979). Other advocates of a single northward movement of Adria relative to Europe include Bouden et al. (1976), who also consider the Western Alps to be a sinistral strike-slip zone with a general N-S compression operating on a pre-existing assemblage of crustal faults and Tapponier (1977), who considered the Western Alpine system in the context of a rigid-plastic indentation of Eurasia by peninsula's of the African continent, where northward collision was accommodated by "lateral expulsion" in the Western Alps.

Other authors, in attempting to explain the arcuate form of the Western Alps, attach more importance to later NW or W-directed movements of Adria with respect to Europe. Laubscher (1971 and recently updated, 1988), proposed a "push-pull" arc subduction model to explain the origin and timing of the Western Alpine and Apenninic arcs. He

considered that the Western Alp arc developed in response to the westward movement of an Adriatic indenter (the "Insubric plate") bound to the N by the dextrally transpressive Insubric fault and to the S by a zone of sinistral strike-slip, loosely defined as the Ligurian fault (Laubscher, 1971).

Away from plate scale models, Siddans (1979) examined arcuate fold and thrust patterns in the Subalpine Chains of SE France, concluding that a right angle bend in the Subalpine Chains is the result of two intersecting fold-thrust belts of different ages. This interpretation is partly true, in that pre-existing fold and thrust structures in the Baronnies and Devoluy influenced the development of later Alpine structures and distort the arcuate geometry. However they may also conceal or mask the continuity of these Alpine structures around the main Alpine bend, as defined by the present distribution of the external basement massifs. So structures from one segment to another in the Alpine arc may in fact be continuous yet obscured because of near parallelism or by the reactivation of earlier structures. Graham (1978) also argued for the continuity of structures around other Subalpine arcs to the SE and demonstrated that continuity patterns may be obscured where the near parallelism of old and new structures occurs.

Vialon (1974) emphasises the importance of Late Variscan orthogonal basement fault trends on the European foreland and Graham (1978) later developed this model to explain the Arc de Digne and the Arc de Castellane, invoking sinistral shearing on NE-SW trending vertical basement faults, although Siddans (1979) discusses arguments against this mechanism. The model to be presented here retains the importance aspect of these pre-existing fault trends, and attempts to explain the variations of tectonic transport direction in the Pelvoux and adjacent basement massifs and on the larger scale, the origin of the Western Alpine arc, by a combination of the reactivation of a pre-existing passive margin fault pattern and a

progressive change in compression direction from SE-NW to E-W. This change in compression direction may be associated with the rotation of an Adriatic indenter accommodated by late dextral strike-slip movements on the Insubric fault.

5.5.3 Early basin configuration and the Western Alpine arc:

Throughout this thesis (mainly Chapters 2 & 4), structural and stratigraphic evidence are presented for i) a Jurassic extensional event and ii) that Jurassic extension faults were reactivated as contraction faults during the Tertiary Alpine compression. It follows that pre-existing Jurassic fault patterns must have some bearing on the overall fold and thrust trends developed. In consequence to this, the pattern of Jurassic faulting for both the Pelvoux and adjacent basement massifs and for SE France as a whole will be examined. It is suggested that the main right-angle bend in the Alpine thrust belt was inherited from the pattern of Jurassic faulting such that the overall thrust belt trend strongly reflects the European passive margin fault geometry.

5.5.3.1 Jurassic faults in the Pelvoux:

From the identification and mapping of preserved Jurassic faults and Jurassic extension faults reactivated as contraction faults during Tertiary Alpine compression (Figure 5.2), a very speculative reconstruction of the pattern of Jurassic faulting for the Pelvoux and adjacent massifs can be proposed (Figure 5.9).

Only the principal Jurassic fault trends are depicted in Figure 5.9; these usually correspond to the steeper and later "breaching" faults (see Chapters 2 & 4). The pattern of normal faulting of the earlier (Lower Liassic) gently dipping faults is more difficult to reconstruct because the degree of reactivation is more severe and only the obvious or general trend is displayed.

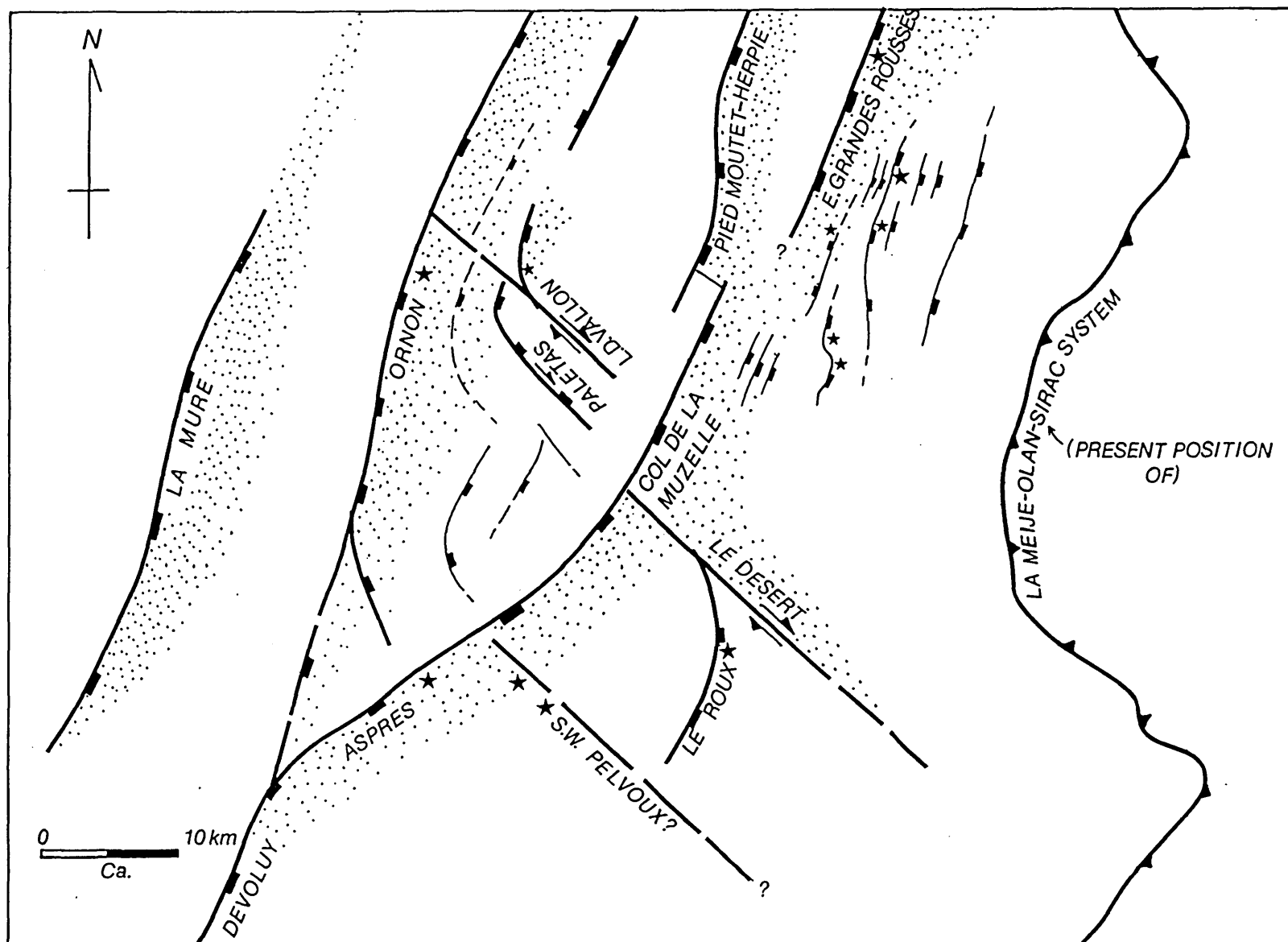


Figure 5.9: Suggested reconstruction for the pattern of Jurassic faulting in the Pelvoux area. Stipple indicates deeper water basins and stars olistoliths. See text for discussion.

On both the deformed map (Figure 5.2) and suggested reconstruction, two dominant fault sets, or trends of faults are evident. A NNE-SSW to NE-SW trend which includes the La Mure and Ornon faults and a not so evident NW-SE trend which includes the Lac du Vallon faults, the le Desert fault and possibly the SW margin of the Pelvoux, although the presence of a pre-existing extensional fault here is debatable.

The congruency or parallelism of deep and shallow water facies trends with the NE-SW fault trends (Barfety & Gidon, 1984; Barfety, 1985), suggests that the general strike of these faults is perpendicular to the principal extension direction. The less obvious NW-SE trending faults are interpreted as transfer faults which formed parallel to the Jurassic extension direction.

Having defined a pattern of Jurassic faulting that in terms of the exact pre-contractional geometry is highly speculative, although the general positioning and trend of the structures is probably quite accurate, it is interesting to compare this with the pattern of compressional reactivation and thrusting and the documented variations in tectonic transport direction. It must be borne in mind that this pre-existing passive margin fault system may have suffered more than one episode of compression and thrusting (re. timing, Section 5.2).

The main feature is that WSW or SW tectonic transport is not apparent northward of a NW-SE trending line coincident with the present position of the Lac du Vallon-le Desert transfer faults. For the basement deformation pre-dating the higher level break-back system of thrusts (a definitive age cannot be given; it may be pre- or post-Nummulitique), this NW-SE pre-existing trend is significant and the early inversion structures of the La Paletas and Haute Valsenestre may mimic the pre-existing fault trend. An explanation for the causes of the SW transport direction will be given later. For the later break-back thrust system, dated as post-Nummulitique, the control of pre-existing Jurassic faults on thrust geometry and transport

direction is not so evident. Except that is for the suitably oriented sections of the steeply dipping NE-SW trending faults such as the Aspres-les-Corps fault. The orientation of this fault is such that during the Alpine contraction, rather than reactivate as a basement thrust, an oblique dextral strike-slip reactivation occurs, locally deflecting displacements towards the SW. This may be interpreted as squeezing out against a pre-existing buttress.

5.5.3.2 Pattern of Jurassic faulting across SE France:

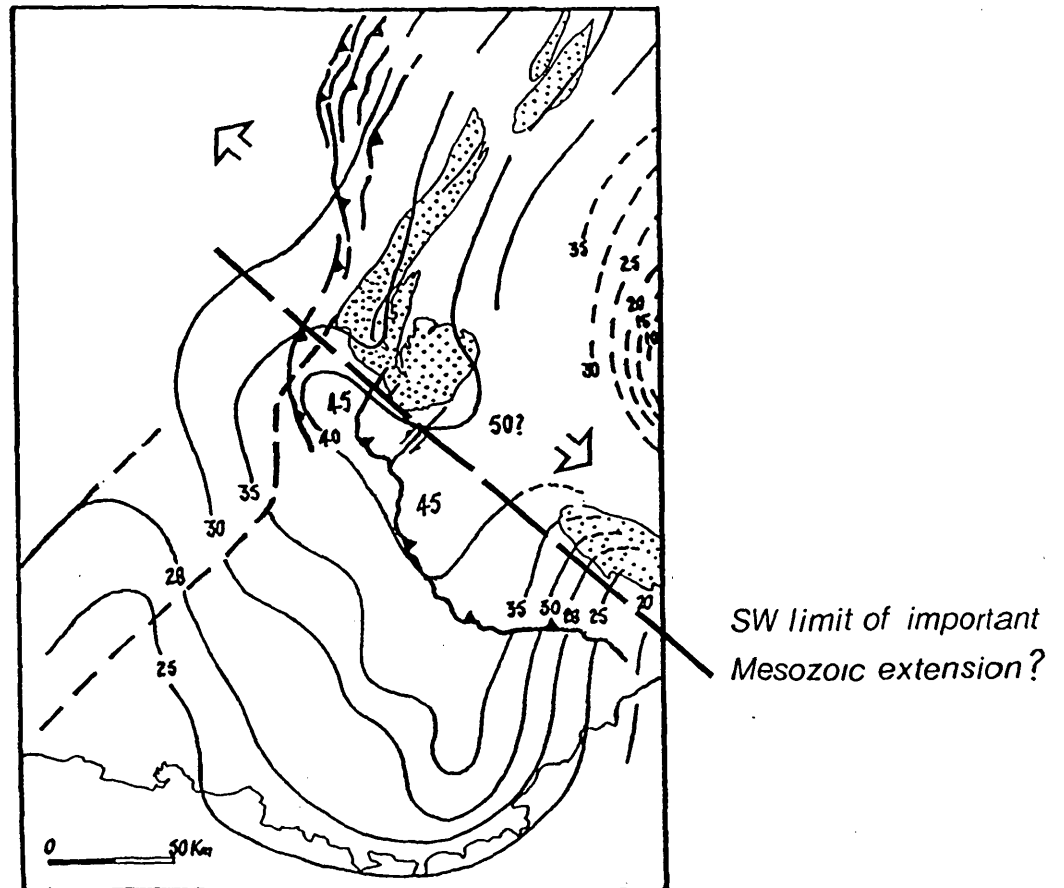
The main tectonic elements of SE France display the same structural trends as those across much of Western Europe (Beach, 1987) and are the same as the Jurassic fault trends documented from the Pelvoux and adjacent massifs. Beach (1988) suggests that the NW-SE trending faults are a reactivated set of late Variscan oblique-slip faults. This Variscan inheritance has long been established (Mattauer, 1973; Vialon, 1974; Arthaud & Matte, 1975). Vialon (1974) proposed that two orthogonal faults sets, the NE-SW "Cevennes" trend and a NW-SE trending set, controlled much of the Mesozoic and Tertiary tectonic development of SE France and it has been suggested in Chapter 4, that these Variscan fault trends may have controlled the fault geometry of the European Tethyan margin. The recognition of these Variscan trends, mainly the NW-SE trend defined by mylonite zones in Pelvoux and Argentera basement massifs and palaeomagnetic data on Permian rocks from the Argentera massif, Dome de Barrot and Esterel massif (Bogdanoff & Schott, 1977; see also Siddans, 1979), precludes the idea of subsequent bending of an originally linear Alpine belt.

As was suggested for the Pelvoux, the early Jurassic syn-rift facies trends show a remarkable parallelism with the NE-SW trending faults (Barfety, 1985) of the Dauphinois domain, exposing the probable Jurassic extension direction. Lemoine et al. (1986) suggest a southern boundary to the Dauphinois basin at the Provence platform. However the geometry of this proposed basin margin is poorly

understood. It may correspond to a complex transfer zone which accommodates a progressive decrease of extension in a direction parallel to the Pelvoux-Argentera axis.

Circumstantial evidence suggests that the Pelvoux-Argentera axis may correspond to an initial transfer fault trend in the Jurassic fault system on the European passive margin. Perrier & Vialon (1980) mapped the depth of the Moho (crustal thickness map) in SE France using Bouguer gravity maps. An anomalously thick crust (40-45 km) trends NW-SE underneath the Devoluy and southern Subalpine Chains. The crustal thickness underneath the adjacent Pelvoux massif is estimated to be between 35 and 40 km. At first appearance this relationship seems unusual. With evidence of crustal thrusting and presumably associated crustal thickening in both the Pelvoux and Belledonne basement, one would expect the crustal thickness here to be greater than underneath the Devoluy. This may be due to a poor interpretation of the gravity data, or alternatively, it may reflect original crustal thickness variations after the Jurassic extension. If the crust of the Pelvoux had been thinned more than the crust underneath the adjacent southern Subalpine Chains, a differential extension between the two domains requires the presence a complex transfer zone possibly parallel to the Pelvoux-Argentera axis. Subsequent crustal thickening may result in a larger crustal thickness underneath the Subalpine chains than is present underneath the Pelvoux, even if the Pelvoux has suffered more in the way of crustal shortening. From this it may be suggested that the Pelvoux-Argentera axis corresponds to a Jurassic age transfer zone defining the SW limit of important Mesozoic extension, or separating a zone of crustal thinning from an adjacent zone of little or no crustal thinning (Figure 5.10).

From the foregoing discussion it is evident that the pre-existing extensional fault trends have influenced to a greater or lesser extent, the Alpine fold and thrust trends and direction of tectonic transport. However it cannot explain fully the progressive change in transport



DEPTH TO MOHO CONTOURED -
dashed lines delineate the Ivrea body
(after Perrier & Vialon 1980)

Figure 5.10: Possible Jurassic transfer fault trend along the Pelvoux-Argentera axis. See text for discussion.

direction, from a WNW direction to a SW direction around this part of the Alpine arc. This is attributed to the interaction between larger scale rotational movements of an Adriatic indenter and the partial reactivation of pre-existing faults on the European foreland (see Section 5.6 below).

5.6 Passive margin fault geometry, indentation and rotation - a model

A model which takes into account the pre-existing passive margin fault geometry and the interactions of indentation and rotation, is proposed to explain the Western Alpine arc and its apparent divergent tectonic transport directions.

5.6.1 Simple rectilinear NW-displacement model:

In the northern part of the external Alps the dominant Oligocene to Miocene displacement direction is towards the NW (Butler et al., 1986). This direction is consistent with the relative plate motions of Africa and its Adriatic promontory relative to Europe, which between 70 ma and 80 ma changed from a northward direction of closure to a NW direction (Smith, 1971; Dewey et al., 1973; Livermore & Smith, 1986; see Figure 5.11).

In any collisional model the crust of the European foreland should not be considered as a rigid undeformed mass, but as a heterogeneous body, characterized by a complex linked network of pre-existing discontinuities inherited from earlier geological event(s), e.g. Tethyan stretching. Given the correct stress conditions each discontinuity may be capable of reactivation as reverse, oblique or strike-slip faults before new crustal faults are generated. However, this fragile condition may only be transitory and after a certain degree of reactivation the potential for further reactivation may decrease dramatically, after which the crust fails by developing new faults. This may have occurred in the Pelvoux after the

pre-Nummulitique deformation and inversion of the pre-existing fault system. Not all of the pre-existing faults need reactivate during this phase.

In Section 5.5.3.2 it was argued that the Pelvoux-Argentera axis corresponds to a Jurassic transfer trend. This interpretation may resolve the difficulties encountered by Graham (1978), when attempting to trace the eastward continuation of an inferred NE-SW basement fault (the Provence fault) in the southern Subalpine Chains (Figure 5.12). From the mechanical arguments presented in Chapter 3, less work is involved if the pre-existing extension and subsequent compression directions are coaxial, so that pre-existing lateral structures simply reverse their sense of displacement. In a simple rectilinear NW-displacement model, the Pelvoux-Argentera axis would thus be operative as a preformed sinistral lateral ramp.

In such a system potential exists for SW tectonic transport, even though the dominant displacement direction is towards the NW. This potential is very much dependant on the pattern of Jurassic faulting. From Chapter 3, reactivation is largely a function of the 3-D stress configuration ($\sigma_1: \sigma_2: \sigma_3$ ratio), the coefficient of sliding friction, the stress difference ($\sigma_1 - \sigma_3$), as well as fault dip and the angle δ between the maximum compressive stress and the strike of the fault. Steeply dipping faults, slightly oblique to the main compression direction, may therefore reactivate in an oblique or strike-slip fashion. In the case of the Western Alps, if the maximum direction of compression is slightly oblique to Pelvoux-Argentera trend, then a zone of transpression will result, locally generating a WSW direction of tectonic transport. This WSW direction will form an angle δ with the NE-SW trending pre-existing faults which are too steep to reactivate as thrusts and therefore reactivate as dextral oblique or strike-slip faults. The net effect is a tectonic transport towards the SW. If the Pelvoux-Argentera Jurassic transfer zone has a complex fault

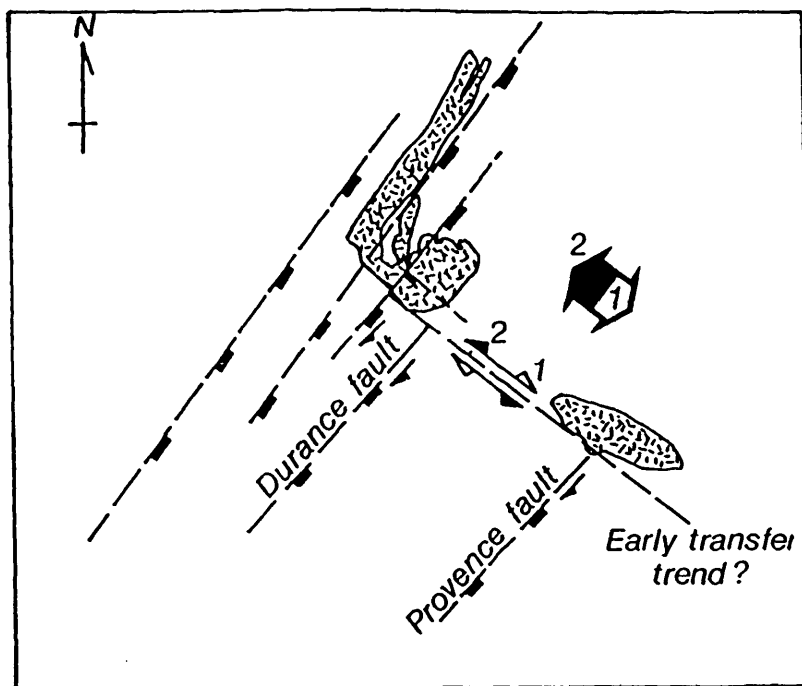


Figure 5.12: Shows a possible relationship between the Pelvoux and Argentera massifs and inherited NE-SW trending basement faults in SE France. 1. Jurassic extension direction, 2. dominant Alpine thrust transport direction. See text for details.

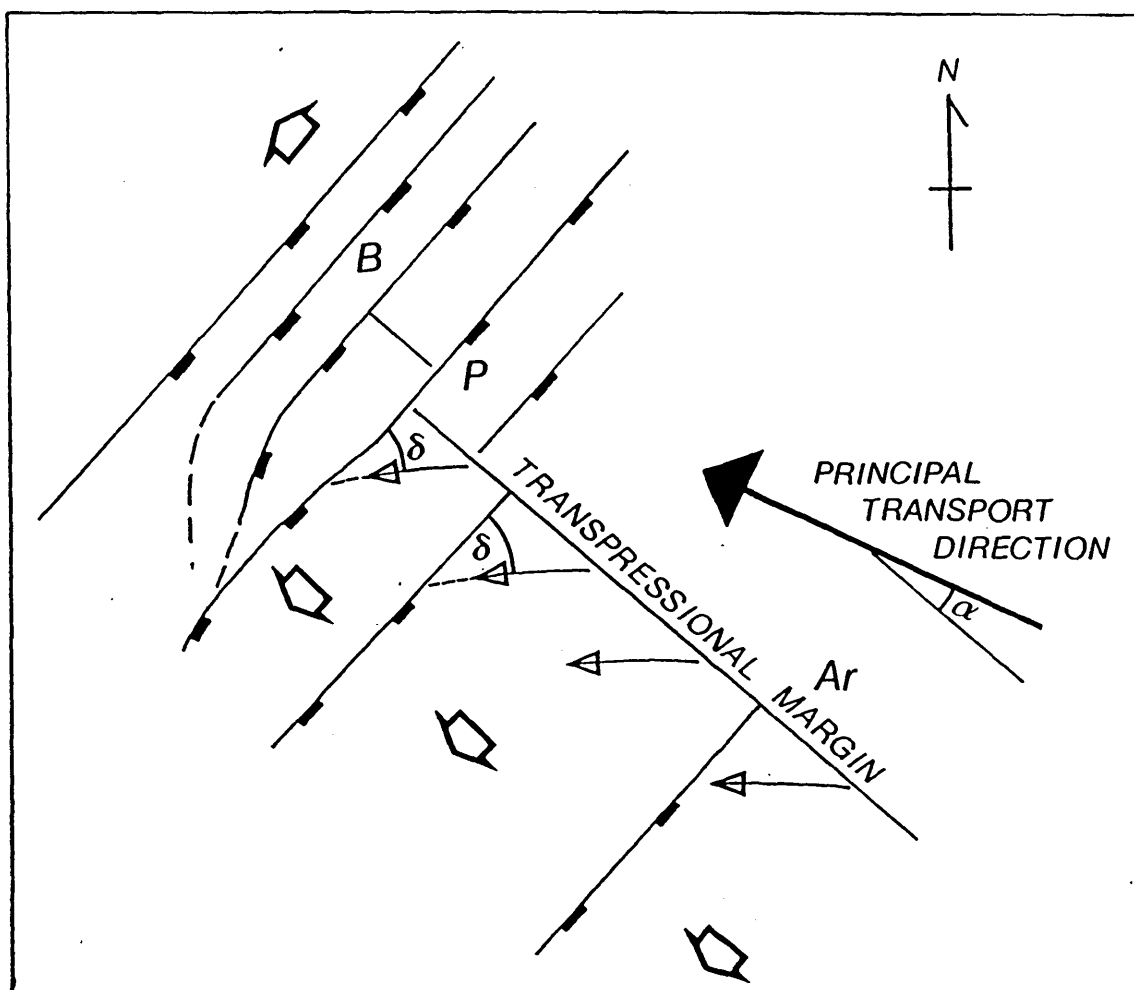


Figure 5.13: Shows the possible effect of the inherited passive margin fault geometry on directions of thrust transport during Alpine compression. B-Belledonne, P-Pelvoux, Ar-Argentera. See text for discussion.

geometry, a pinning effect may result, introducing an anti-clockwise rotational aspect to the deformation. The above aspects are illustrated schematically in Figure 5.13.

5.6.2 Interaction of indentation and rotation:

Although a combination of the pre-existing passive margin fault geometry and a simple rectilinear NW-displacement model, is to some extent able to explain the divergent transport directions around the external Alpine arc, such a model is not consistent with all the geological data. Mineral stretching lineation maps (Malavielle, 1984; Choukroune et al., 1986) show a dominantly E-W stretching for the Internal Alps, and a divergent fan pattern for external zones (see Figure 5.11). Merle et al. (1984), and Choukroune et al. (1986) recognise an early NW Cretaceous displacement for the Embrunais - Ubaye Helminthoid flysch nappes and document a progressive rotation of the transport direction, to a W direction and eventually SW direction during the Oligocene (Figure 5.15 C). Choukroune et al. (1986) suggest that early collision in the Western Alps is characterized by a NW direction of overthrusting, and that a progressive anti-clockwise rotation of the transport direction occurs with time. W of the Belledonne massif, in the Vercors and Chartreuse Subalpine massifs, extension tangential to the Alpine arc is accommodated by dextral strike-slip faulting which offsets Miocene age thrusts and folds. The position of these cover faults is coincidental with the NE extension of the Isere basement fault (Section 2.6) and they are interpreted as riedel shears formed in the cover in response to dextral strike-slip movements along this fault (Figure 5.15).

It therefore seems that the progressive change in tectonic transport direction is not only restricted to the Pelvoux massif, and on the scale of the Western Alpine arc, there are significant variations of the displacement direction in both time and space. This change in tectonic transport direction appears at its most acute during the final stages of collision, when in the northern segment of

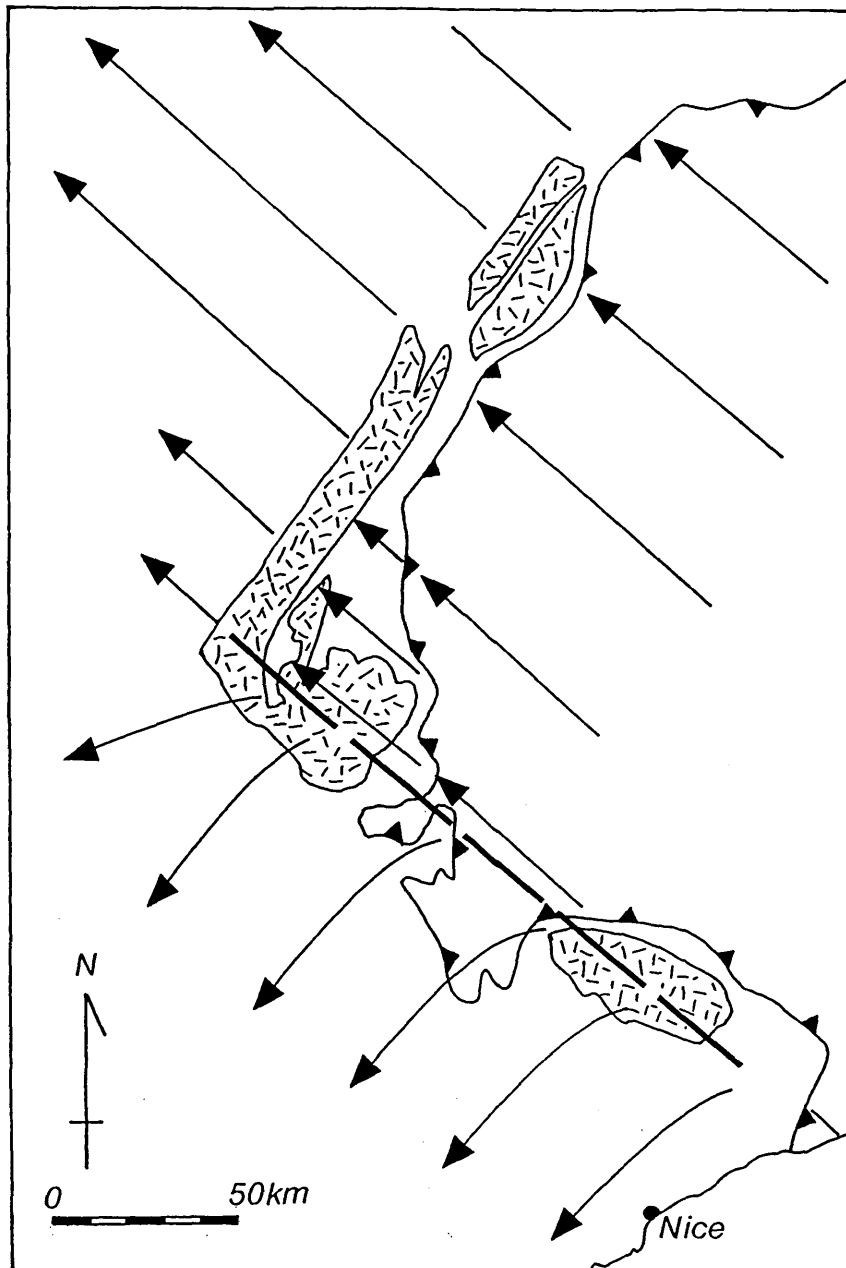


Figure 5.14: Main Alpine phase displacement trajectory paths for a simple rectilinear NW-displacement model for the French Alps. See text for discussion.

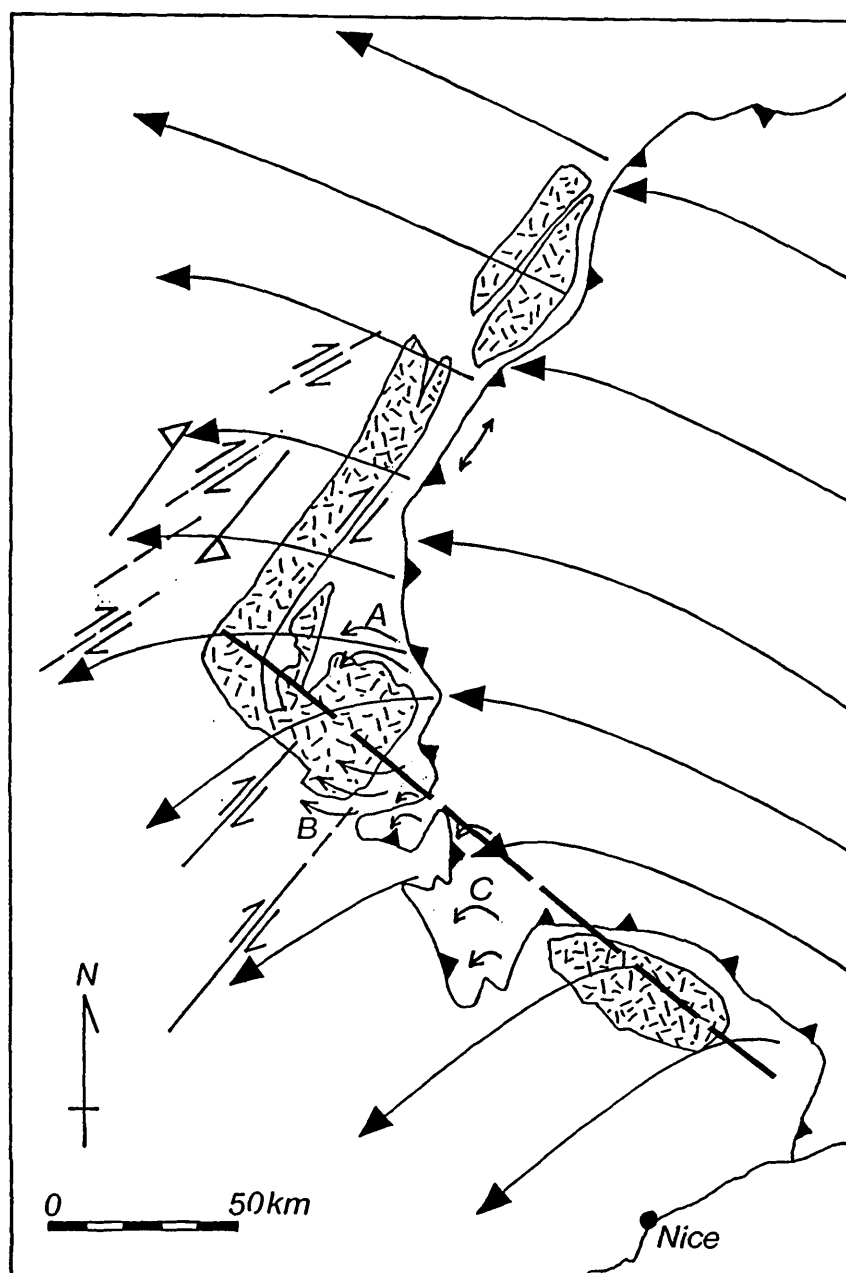


Figure 5.15: Displacement trajectory paths for the main Alpine phase deformation showing a progressive regional anti-clockwise rotation of transport direction with time, and SW-directed displacements along the Pelvoux-Argentera. A. Local rotations along the northern Pelvoux (Beach, 1982; this study), B. Local rotation along the southern Pelvoux (Tricart, 1982; Chambers, 1987), C. Local rotations in the Embrun-Ubaye nappes (Merle et al., 1984; Choukroune et al., 1986). Late dextral strike-slip reactivation of pre-existing NE-SW trending faults is also shown.

the chain which had previously been dominated by NW to WNW directed displacements, pre-existing NE-SW basement faults and possibly pre-existing thrusts which span the length of the arc (Frontal Pennine thrust ?) reactivate as dextral strike-slip faults (Figure 5.15). A simple rectilinear NW-displacement model is clearly too simplistic. This anti-clockwise rotational component may reflect a larger scale plate rotation of an Adriatic indenter, bound to the N by the dextrally transpressive Insubric-Tonale line (Laubscher, 1971, 1988). As the displacements along the Insubric fault lock up, the dextral movements may then be transferred outwards onto the European foreland, reactivating pre-existing structures. The extra Tertiary rotation of Italy due to the opening of the Balearic basins, estimated to be 17° by Lowrie (1986), will also contribute to the progressive change in thrust transport direction documented from the Western Alpine foreland.

In summary it is proposed that in the Pelvoux district, the principal direction of tectonic displacement was towards the NW, and that with time an increasing anti-clockwise rotational component was introduced. The cause of this rotation may have been a local pinning along the transpressional southern margin (Pelvoux-Argentera axis). With time the rotation became more pronounced, reactivating early thrust structures and basement faults out on the Alpine foreland. That this same stress field is still operative today, is demonstrated by the present seismicity of the Alpine foreland (Frechet, 1978; Menard et al., 1988, Vialon et al., in press). The displacement trajectory paths of the two models are shown in Figures 5.14 & 5.15. On Figure 5.15, the two local directions of cover transport (A & B) at the northern and southern margins of the Pelvoux massif are interpreted to result from a local pinning and deflection of the displacement direction around the borders of the Pelovux massif, which at the time of thrusting formed a resistant body due to previous uplift (see also Tricart, 1982).

Chapter 6: Crustal structure

6.1 Introduction

This short chapter intends to examine some of the more speculative problems concerning the deep structure of the Alps, and serves to combine available geophysical data (e.g. ECORS-CROP profile, Bayer et al., 1987) with surface-derived information presented in other parts of the thesis. Aspects to be considered include the effect of Jurassic faulting and crustal thinning on the present crustal geometry; it has previously been suggested from the study of surface exposures that pre-existing Jurassic extensional structures controlled many fundamental aspects of the Alpine compressional deformation.

6.2 Geophysical data

In the Swiss Alps, Hsu (1979) showed on the evidence of seismic velocity inversions in the European Alpine crust, that the external basement massifs overthrust one another as a result of Alpine compression. In the French Alps, Menard (1979; 1980) from seismic velocity profiles and prospective well data, compiled a map of the depth to the pre-Triassic crystalline basement. This shows a NNE-SSW zone of thickened Mesozoic cover parallel to, and to the immediate W of the Belledonne basement massif. This was interpreted to have formed in direct response to the WNW overthrusting of the Belledonne basement. Indeed, the steeply westward dipping basement-cover contact of what can be considered as the crystalline front in the external French Alps, can be readily interpreted as a homoclinal fault-bend fold (Suppe, 1983), suggesting the presence of a crustal foot-wall ramp in the subsurface somewhere to the ESE (see Section 2.6).

Thouvenot & Perrier (1980) used a velocity analysis of the seismic data provided by the ALP75 experiment (Figure 6.1) to document the presence of an intra-crustal low-velocity zone (LVZ) or velocity reversal, underlying the

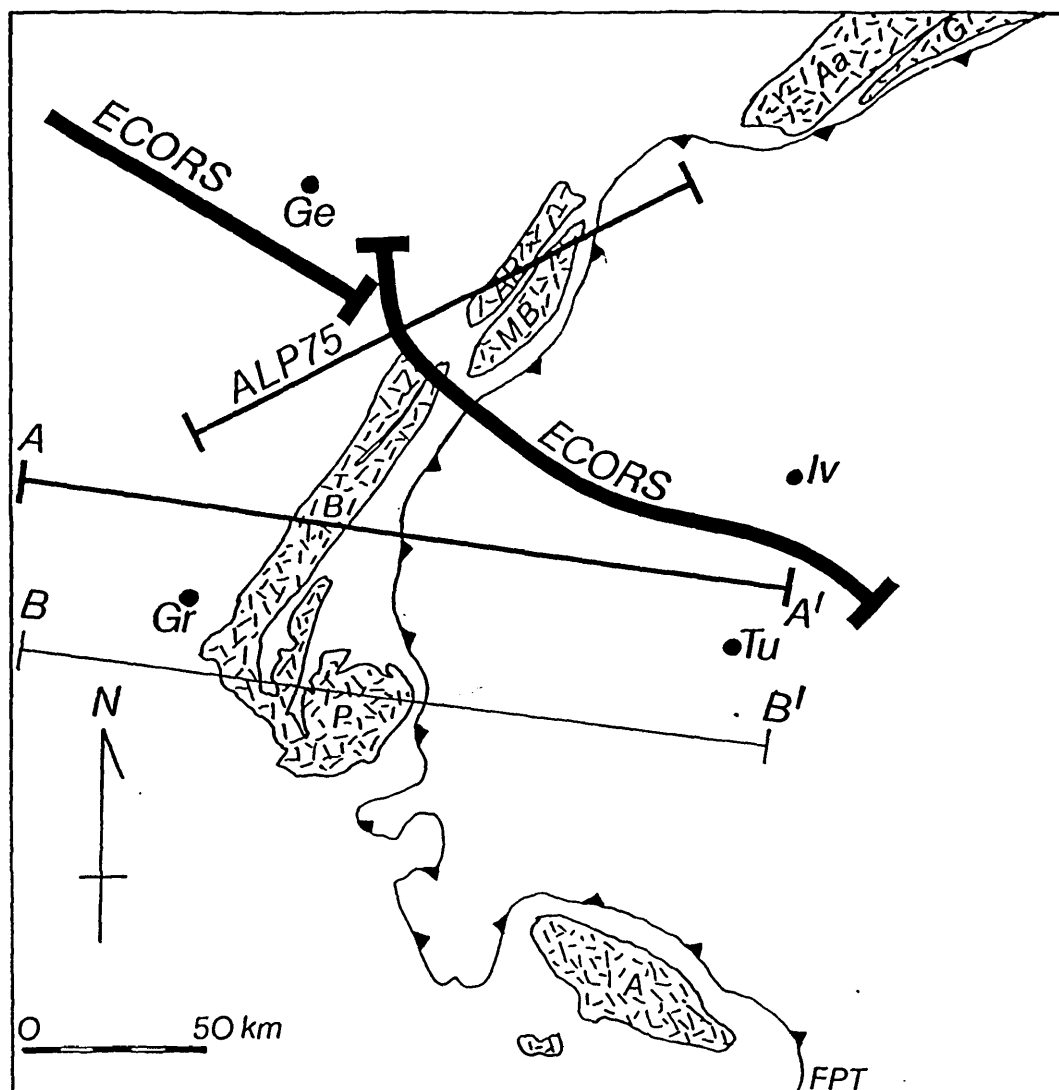


Figure 6.1: Location of the seismic profiles mentioned in the text. Section line AA' corresponds to Fig. 6.2 and section line BB' corresponds to Fig. 6.5. Gr-Grenoble, Ge-Geneva, Tu-Turin, Iv-Ivrea, G-Gotthard, Aa-Aar, AR-Aiguille Rouges, MB-Mont Blanc, B-Belledonne, P-Pelvoux, A-Argentera.

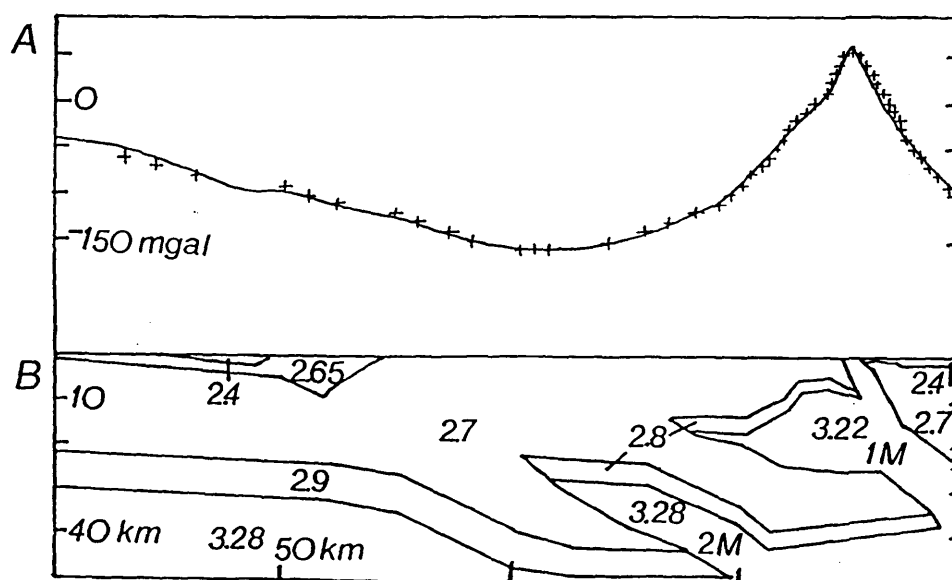


Figure 6.3: Structure of the Western Alps modelled from gravity. A. Full line: calculated anomaly, crosses: observed anomaly, B. gravity model with corresponding densities. Taken from Thouvenot & Menard (1984).

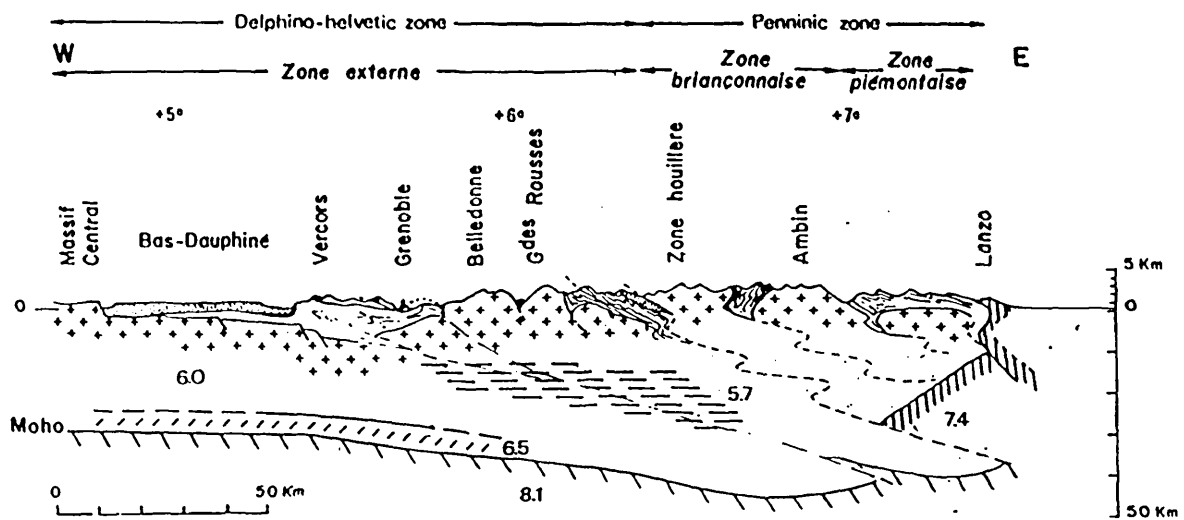


Figure 6.2: Interpretative cross-section across the Western Alps by Perrier & Vialon (1980) (location shown as AA' on Fig. 6.1). See text for details.

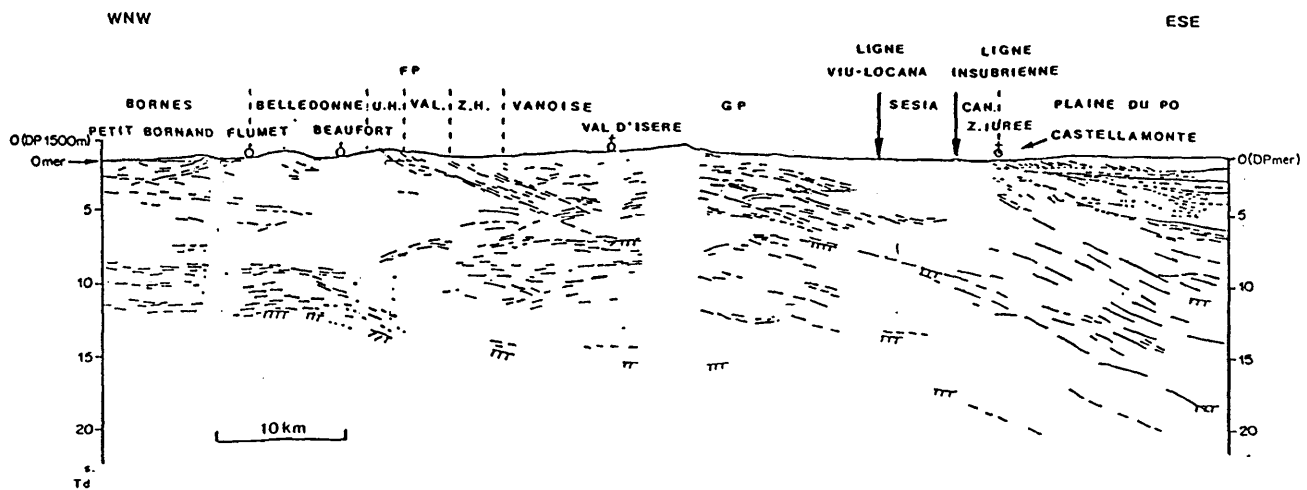


Figure 6.7: Provisional line drawing of the ECORS-CROP profile (from Bayer et al., 1987).

external crystalline massifs at a depth of 10-15 km, and which was absent from the crust under the adjacent Subalpine Chains. They suggested that this LVZ corresponds to a zone of higher plasticity, or intra-crustal decoupling, linking with the thrust ramp of the Belledonne massif. This constitutes the sole thrust of the Western Alps (see Butler, 1983; 1986).

From the combined analysis of seismic-velocity profiles and gravimetric data, Perrier & Vialon (1980) presented an interpretative cross-section of the crustal structure of the Western Alps, from the Po Plain to the Massif Central, at the latitude of Grenoble (see Figure 6.2). The gravity data reveals a large positive Bouguer anomaly (20 to 60 mgals) below the Ivrea zone suggesting the presence of a high density (3.2 g cm^3) mass. This positive Bouguer anomaly diminishes rapidly to the W forming a broad negative anomaly which reaches a maximum (-140 to -160 mgals) under the Brianconnais, suggesting a crustal thickness of 45 to 50 km compared with a 29 km thick crust under the Rhone Valley. Perrier & Vialon (1980) interpret this geophysically defined Ivrea body (see Breckhemer, 1968) as a mantle thrust slice which is continuous with the Frontal Pennine thrust to the W (Figure 6.2). The Belledonne thrust is also shown to pass downwards across the lower crust to join with another recorded offset of the Moho at a lower level than the Ivrea mantle slice.

Menard & Thouvenot (1984) proposed an alternative cross-section to that of Perrier & Vialon (1980). From a re-interpretation of the gravity and arrival times from a 1958 seismic experiment (Closs & Labrousse, 1963), they defined a second important mantle slice (M2 - Figure 6.3) to the W of the main Ivrea body, whose "birds head" shape was first defined by Berckhemer (1968). The shape of this second mantle slice is modelled from the available seismic and gravity data (Figure 6.3). The existence of two mantle thrust slices (M1 & M2) i.e. the flaking of the mantle lithosphere, forms their basic hypothesis, and like the

interpretation of Perrier & Vialon (1980), the thrust surfaces controlling their geometry correspond to the Frontal Pennine and Belledonne thrusts respectively. Menard & Thouvenot (1984) also contemplate to which crust the Ivrea mantle slice belongs i.e. European, South Alpine or Tethyan. They conclude that it is difficult for the Ivrea mantle slice (M1) to have originally been a part of the South Alpine lithosphere (contrary to the suggestions of Schmid et al. (in press - see later) and suggest that it was a part of either the Tethyan or European lithosphere. They also show that the mantle lithosphere beneath the Alps displays abnormal seismic velocities ($<8 \text{ kms}^{-1}$) which are reminiscent of the seismic velocities documented from areas of crustal extension or rifts. This may be a reflection of earlier Tethyan lithospheric extension. The interpretation proposed by Menard & Thouvenot (1984) is important, as it forms the basis of a crustal extension-inversion model presented in Section 6.4.

During the closing months of this work, the preliminary interpretations of a deep seismic reflection profile (ECORS-CROP) across the Western Alps were published (Bayer et al., 1987). The implications of this newly acquired data for the interpretations presented in this work are discussed in Section 6.5.

6.3 Other interpretations

The geophysical interpretations of the deep structure of the Western Alps discussed in the foregoing opened a number of unresolved problems which have been subsequently readdressed by a number of authors (e.g. Beach, 1981; Butler, 1983; 1986).

Caby et al. (1978) postulate the loss of European crust by subduction in the internal Alps. More recently, Butler (1983; 1986) addresses anew this problem of insufficient or "missing" continental crust beneath the Western Alps. On the basis of thin-skinned balanced section constructions across the Alps, Butler (op. cit.) infers

that large amounts of upper crustal shortening are incurred on a high level crustal detachment, generating the requirement for a middle to lower crustal wedge to bring about a crustal balance. Butler (1986) allows this by the subduction of large volumes of bouyant lower to middle European crust, with a granulite-eclogite phase transformation, the resulting density increase providing the necessary hidden load requirement for lithospheric flexure in the Alpine foreland and the Po Basin (Karner & Watts, 1983; Royden & Karner, 1984). Such an analysis remains useful, but it does bring into question the validity of "thin-skinned" interpretations, which fail to incorporate effects arising as a result of pre-existing Mesozoic extension. The requirement of a middle to lower crustal wedge of such enormous dimensions may simply be an artifact of the assumptions and balancing method used.

Other workers addressing the problem of insufficient crust under the Western Alps (e.g. Helwig, 1976; D'Argenio & Alvarez, 1980; Beach, 1981c), attach greater importance to a pre-existing extensional phase which may have thinned the European crust, reducing its thickness substantially prior to the onset of Tertiary Alpine compression. Helwig (1976) considered the evolution of the Eastern Alps in the context of a Wilson cycle, where previously thinned continental crust is subsequently re-thickened back to, or greater than, its original normal thickness by the reactivation of pre-existing normal faults during compression (see also D'Argenio & Alvarez, 1980, who proposed a similar restacking model for the Apennines). Beach (1981c) also argued that if the whole crust were thinned by normal faulting, it could be restacked to form a deep level crustal duplex, giving the present crustal thickness beneath the Alps. These models eliminate the requirement of a lower crustal wedge and the subduction of large volumes of continental crust that the palinspastic reconstructions based on thin-skinned thrust models demand (e.g. Butler, 1983; 1986). They also eradicate the serious space problems which arise under the Po Plain when similar N-S crustal subduction profiles are proposed by other

authors for the Eastern and Southern Alps (e.g. Roeder, in press). The deep structure of the Alps is clearly a three-dimensional problem.

If the European crust were substantially thinned, as can be expected at the ocean-continent boundary, this may facilitate the process of crustal subduction (McKenzie, 1969; Le Pichon et al., 1982) and the model proposed by Butler (1986) may still prove valid. Furthermore, should both the extensional and later compressional deformation be largely accommodated along the same mid-crustal decoupling zone, such that it is not uniform in any vertical section through the crust, a decoupling during compression similar to that suggested by Butler (1986), with the exception that the shortening reactivates the pre-existing extensional detachment, may still call for the subduction of unshortened and possibly previously thinned continental crust.

Models alternative to those proposed in Section 6.2, may explain the present subsurface configuration of the Ivrea body. Handy (1986), and Schmid et al. (in press) suggest that the Ivrea body is a mantle slice of South Alpine affinity that is later cut through by the Insubric fault, which is considered to be a late intra-crustal fault accommodating back-thrusting followed by dextral strike-slip movements (Laubscher, 1971). Schmid et al. (in press) therefore locate the main collisional suture beneath the Ivrea body such that both are deformed by back-thrusting and back-folding of the later Insubric phase. However, should the Ivrea body be mantle lithosphere of European or Tethyan affinity, it could be expected to act as a rigid stress guide during collision, thereby providing one explanation of the back-thrust deformation in the internal zones of the Western Alps. This could be considered in the context of "tectonic wedging" (Price, 1986), induced by the emplacement of a mechanically rigid Ivrea mantle wedge.

6.5 Crustal extension-inversion model

Prior to the publication of the ECORS-CROP deep seismic reflection profile interpretations (Bayer et al., 1987), a simplified model of Tethyan extension, followed by Tertiary Alpine compression and inversion of the European Tethyan margin by the reactivation of pre-existing passive margin faults was developed.

On the basis of structural and stratigraphic evidence, two normal fault systems of slightly differing age and distinctive geometry have been postulated in the Pelvoux region of the Western Alps (see Chapters 2 and 4). In Gillcrist et al. (1987), from the extrapolation of surface-derived data, it was suggested that several levels of crustal detachment controlled the extensional and compressional evolution of European Tethyan margin. For example, in the Pelvoux region two detachment levels were considered to have been operative, a shallow (early) detachment which climbed to the surface at what is now the Rochail massif (see also Chapter 2) and a deeper level detachment, which climbed to the surface somewhere to the W of the Belledonne massif. It was suggested that these detachments developed during the Mesozoic extension and were reworked by Alpine thrusting.

This working hypothesis, of two principal detachment faults, needs to be justified. It is supported by:

- 1) Structural and stratigraphic observations concerning the geometry of both early normal and later thrust faulting (see Chapter 2).
- 2) The subsidence patterns, or geodynamic history of the European Tethyan margin (see Chapter 4).
- 3) The geophysical observations of Menard & Thouvenot (1984) (this Chapter).

Nevertheless, any inference concerning the deep structural geometry remains at best speculative, especially

considering our current knowledge concerning the deformation kinematics and physical processes affecting middle to lower crustal levels in both extensional and compressional terrains.

Menard & Thouvenot (1984) also postulate two principal detachment faults in their interpretative cross-section of the Western Alps. The lowest of these corresponds to the lower detachment of Gillcrist et al. (1987) and shows a ramp-flat trajectory which incorporates a slice of mantle lithosphere (M2) at a depth of 30 km. The higher detachment corresponds to the detachment under the Pelvoux region and to the E carries the geophysically determined Ivrea body on its hangingwall.

It is possible that the main detachment faults of Menard & Thouvenot (1984) represent older extensional faults which governed the evolution of the European Tethyan margin. Brodie & Rutter (1987), disagree with the suggestions of some theoretical stretching models (e.g. McKenzie, 1978) which assume that extension of the lower crust is accommodated by a pervasive pure shear ductile deformation and postulate that high-temperature faulting of lower crustal rocks may also form an important mechanism of lower crustal extension. They base this assumption on observations of lower crustal rocks from the Ivrea zone. Such a mode of failure of the lower crust is implicit in the simple shear stretching models recently proposed (e.g. Wernicke, 1985; 1986; Beach, 1986). However, Brodie & Rutter (1987) also state that it is impossible to determine whether bulk pure shear stretching is accommodated on conjugate sets of low-angle high-temperature faults, or if a dominant single shear sense prevails. It is theoretically possible that both bulk pure shear deformation, and zones of localized simple shear faulting are operative (see Coward, 1986) and that the low-angle detachment faults of Wernicke (1985) may show a ramp-flat modification (see Beach, 1986). As high-temperature fault zones in the lower crust and mantle lithosphere may concentrate strain, low-angle ramp-flat detachment faults developed during a period

of crustal extension, may constitute zones of inherent weakness that are susceptible to failure and reactivation during compression. In this way it may be envisaged that the detachment faults of Menard & Thouvenot (1984) correspond to earlier Jurassic faults. From the inversion geometries documented at the surface, the higher level detachment may correspond to the necessary decoupling horizon for the earlier normal faults (see Chapters 2 and 4), whilst the deeper decoupling horizon corresponds the extensional detachment onto which the later breaching faults decouple (i.e the Ornon and La Mure faults).

On the section depicting the late Jurassic to early Cretaceous extensional configuration (Figure 6.4), albeit an extremely speculative reconstruction, a thrust reactivation of the second detachment, which is proposed to explain the early to Mid-Jurassic uplift of the Briançonnais, will result in the formation of the second mantle slice (M2) forecast by Menard & Thouvenot (1984). The present elevation of the basement rocks of the Pelvoux massif also requires explaining. The combined effects of restacking pre-existing extensional fault blocks by a thrust decoupling along the higher level extensional detachment, together with passive uplift of the Pelvoux on a deeper thrust ramp in the lower detachment surface, may provide a necessary explanation (Figure 6.5). A conceptual model which involves the reactivation of pre-existing Mesozoic detachment faults during Tertiary Alpine compression is presented in Figures 6.5 & 6.6.

An additional point of interest concerning the detachment model proposed for the extensional evolution of the European Tethyan margin, is the form of the Pogallo fault suggested by Handy (1986) and Schmid et al. (in press). They interpret this to be a Triassic to Jurassic age deep crustal normal fault which separates deeper parts of the South Alpine crust from more intermediate crustal levels and was subsequently rotated to its present steep attitude by the emplacement of the Ivrea mantle body during Alpine orogenesis. After removal of the effects of

CEVENNES - DAUPHINOIS BRIANÇONNAIS PIEMONT SOUTH ALPINE CRUST
ARDECHE BASIN BASIN LIGURIAN OCEAN

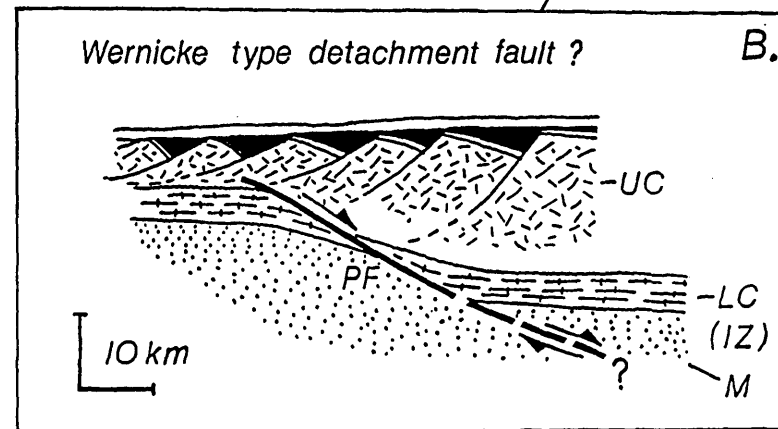
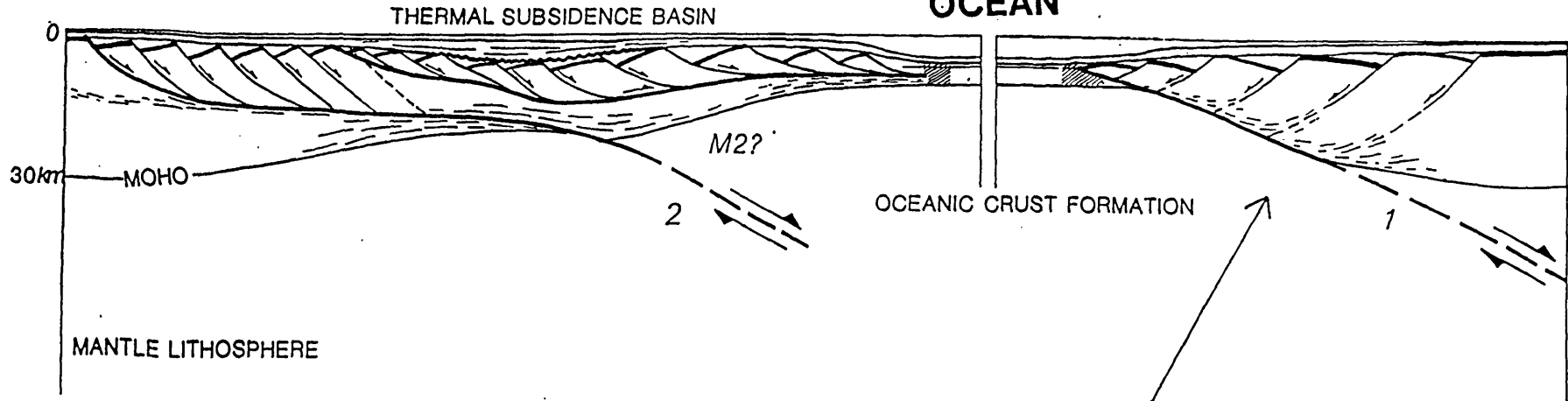


Figure 6.4: A. Hypothetical extensional geometry at the end of the Jurassic, B. the Pogallo Fault (PF) (see Handy, 1986; Schmid et al., in press) which may correspond to a Wernicke-type detachment fault in the lower South Alpine crust. UC-upper crust, LC-lower crust, IZ-Ivrea zone, M-mantle. Figure B modified after Handy (1986). See text for details.

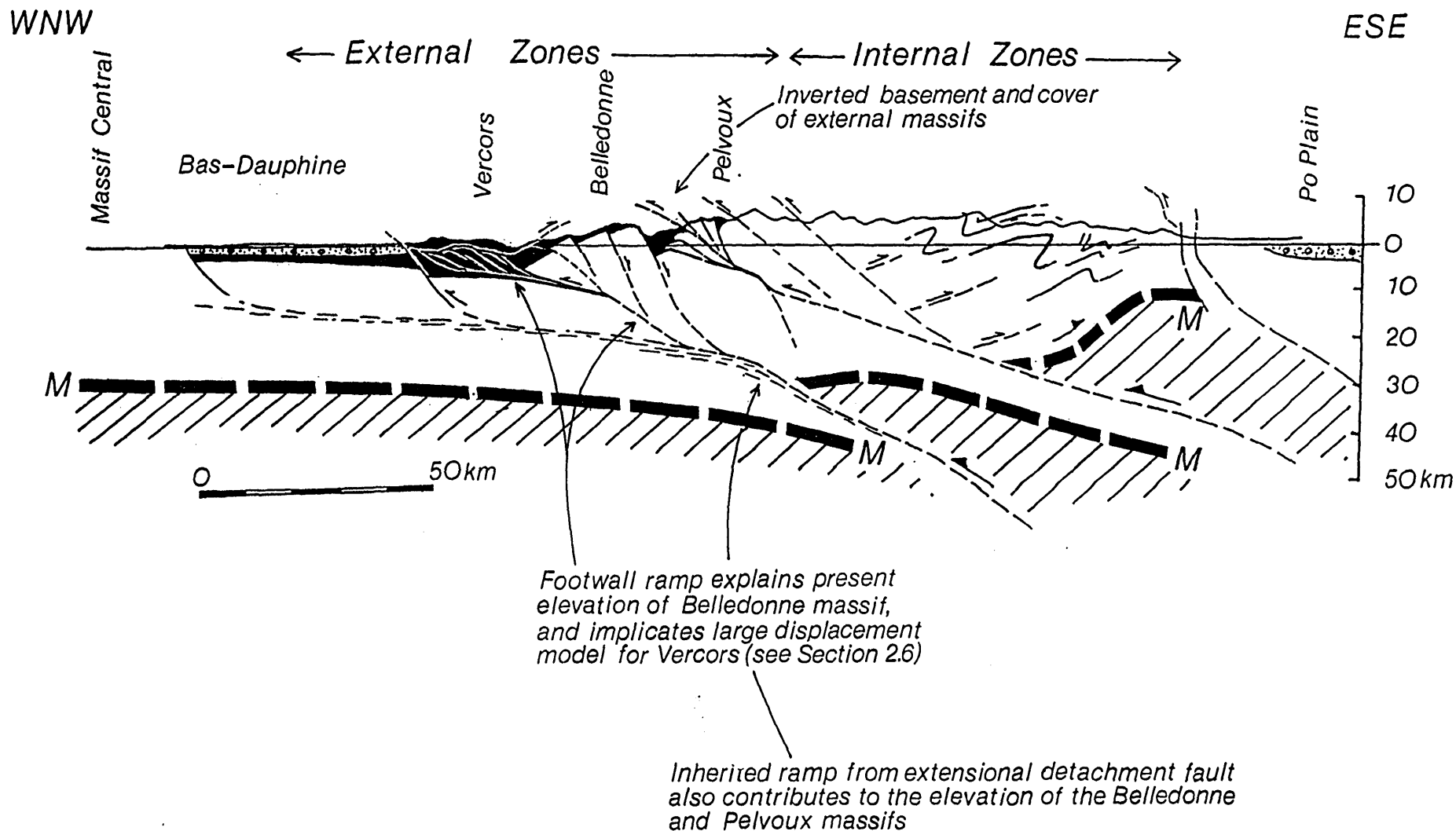


Figure 6.5: Interpretative cross-section across the Western Alps at the latitude of the Pelvoux (BB' on Fig. 6.1). Based on the gravity profile of Thouvenot & Menard (1984) and surface data presented in this thesis. See text for details.

WNW

EXTERNAL PART OF THE EUROPEAN ALPINE-TETHYAN EXTENSIONAL MARGIN

ESE

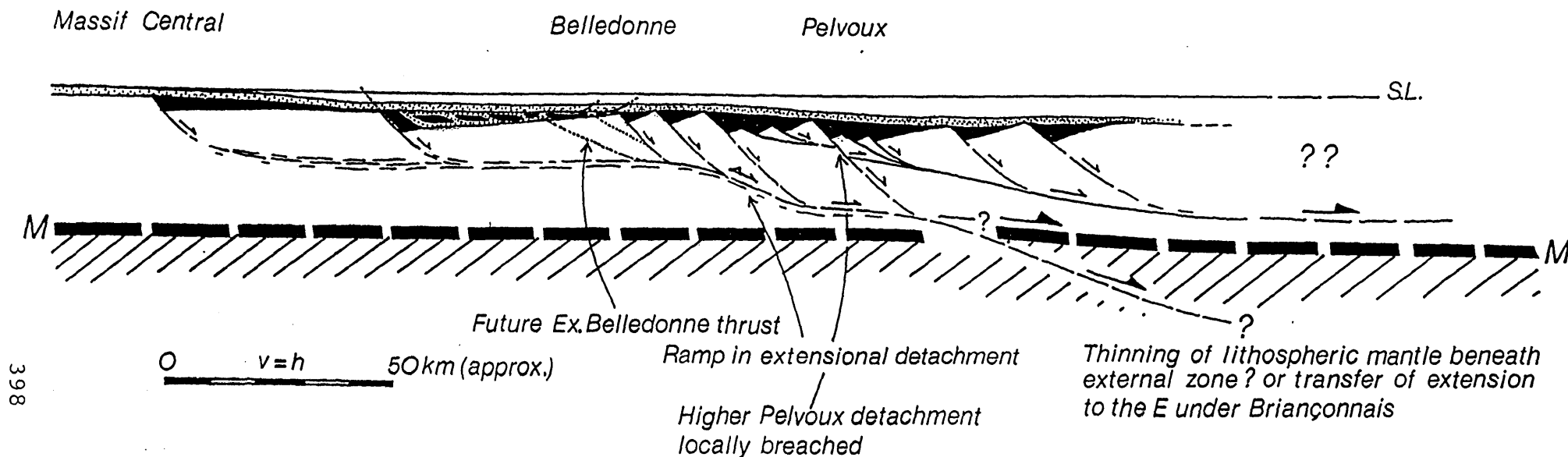


Figure 6.6: Suggested extensional geometry for the external parts of the European Tethyan margin prior to Alpine contraction and inversion (restored from section in Fig. 6.5). See text for details.

Alpine deformation they suggest that the Pogallo fault was a low-angle extensional fault which originally dipped gently to the ESE. This dip sense is opposite to that inferred for the majority of upper crustal normal faults affecting the Southern Alpine margin (Bernoulli, 1964, Kalin & Trumphy, 1977). In view of this evidence, it is tempting to interpret the Pogallo fault as part of a simple shear fault system (Wernicke, 1985; or with a modified crustal geometry e.g. see Beach, 1986; Coward, 1986) on which continental break-up occurred (Figure 6.4 A & B), especially if the Ivrea body is taken to be of European or Tethyan affinity. However, given the low extent of exposure of deep crust, it may also be part of a system of high-temperature, low-angle, conjugate fault sets which accommodate a bulk pure shear deformation of the lower crust.

6.5 The ECORS-CROP profile - a modification of the model

From an examination of the surface-derived data (Chapter 2) and the new findings of the ECORS-CROP deep seismic reflection profile, the crustal extension-inversion model presented in Section 6.4 may be oversimplified and can be modified.

The level of erosion and exposure in the external basement massifs of the French Alps is such that it provides an insight into the mechanism of upper crustal shortening and thickening that is operative: the basement massifs can be considered as late structural windows disclosing a number of critical structural and sedimentological relationships.

It has been postulated in Chapters 2 and 5 that a number of break-back, basement-involved thrusts, form part of a late out-of-sequence thrust system in the central and western Pelvoux. These later thrust surfaces are discordant with respect to an older progressive strain, generating the discontinuous thrust surface map patterns so characteristic of out-of-sequence thrusting (Boyer & Elliott, 1982). The

discordant older structures in their foot-walls are often inversion structures, formed by the compressional reactivation of earlier extension faults.

A line drawing of the ECORS-CROP reflection profile is reproduced from Bayer et al. (1987) (Figure 6.7). It lies a considerable distance to the N of the Pelvoux area, too far to extrapolate the field data collated and presented in this thesis, and a re-interpretation of this profile is not attempted here. However, the important points outlined by Bayer et al. (1987) are:

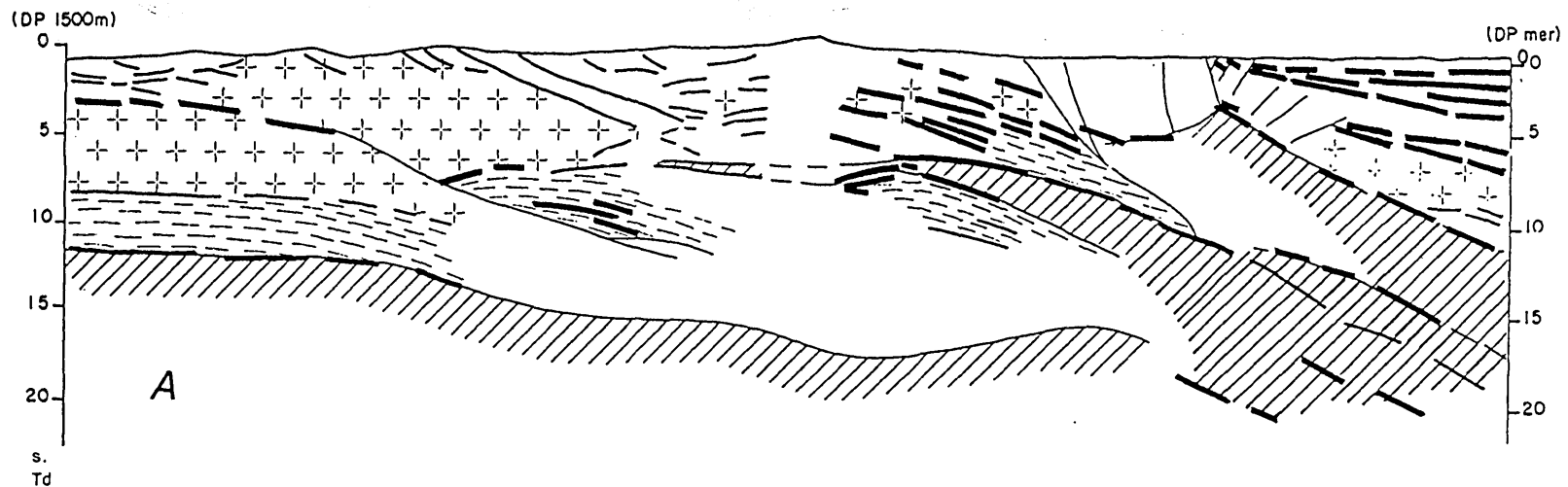
1. A strongly reflective layered lower crust dips gently eastwards under the Belledonne massif. This is cut out by the deeper continuation of the Belledonne thrust which is either a crustal thrust, or a lithospheric thrust as suggested by Menard & Thouvenot (1984).
2. In the upper parts of the profile, faults such as the Frontal Pennine thrust (FP) appear as strong continuous reflectors with an apparent 25° eastward dip. These faults may branch or merge from a thrust interpreted to involve the mantle lithosphere at 6 seconds which overlies a foot-wall of reflective lower crust. Such a detachment could be compatible with the higher level detachment inferred in Section 6.4.
3. To the E of the FPT, W dipping reflectors may be interpreted as back-thrusts on the W flanks of a large domal structure culminating in the Grand Paradis (tectonic wedging ?).
4. A zone of seismic transparency to 8 seconds over the Ivrea zone, suggests the presence of steep structures.

The gently dipping, continuous nature of seismic events interpreted to represent the Frontal Pennine thrust and structures in its foot-wall, is consistent with the out-of-sequence break-back character of the thrusts documented from the Pelvoux. These are interpreted to

branch from the the same basal detachment as the FPT. The form of these events on the ECORS-CROP line is similar to that of the Wind River thrust imaged on the COCORP study (Smithson et al., 1978) which corresponds to a gently dipping, continuous break through the upper parts of the Earths' crust. Combining this seismic information on the deep geometry with surface observations from the Pelvoux area, it is possible that the main Alpine crustal shortening and overthrusting is pre-dated by the inversion of the European Tethyan margin, accommodated by a reactivation of the passive margin fault system. However, rather than be considered as two distinctive and separate deformational events, such a sequence may be progressive and geodynamically continuous, reflecting the inherent weakness of pre-existing discontinuities in the European plate. With early Alpine compression and the onset of continent-continent collision, pre-existing normal faults out on the European foreland may reactivate. With continued plate convergence there may be a time when the compressive deformation can no longer be accommodated by the reactivation of pre-existing extensional faults i.e. the amount of contraction may greatly exceed the amount of original extension. At this point the European crust will deform by renewed failure initiating the overthrusting of new basement slabs, giving what appears as the out-of-sequence thrusts.

Intra-crustal deformation of the European plate will be accentuated by the difficulties encountered in subducting bouyant European continental crust. This is depicted in the two preliminary interpretations presented by Bayer et al. (1987), reproduced in Figure 6.8 A & B. The first of these interpretations (Figure 6.8 A) infers a thrust duplex of lower crustal rocks, the thrusts transferring displacements onto a main lithospheric roof thrust from which the FPT branches. The second model involves mantle lithosphere in the crustal imbrication (Figure 6.8 B).

What ever the final interpretation, one inescapable



WNW

ESE

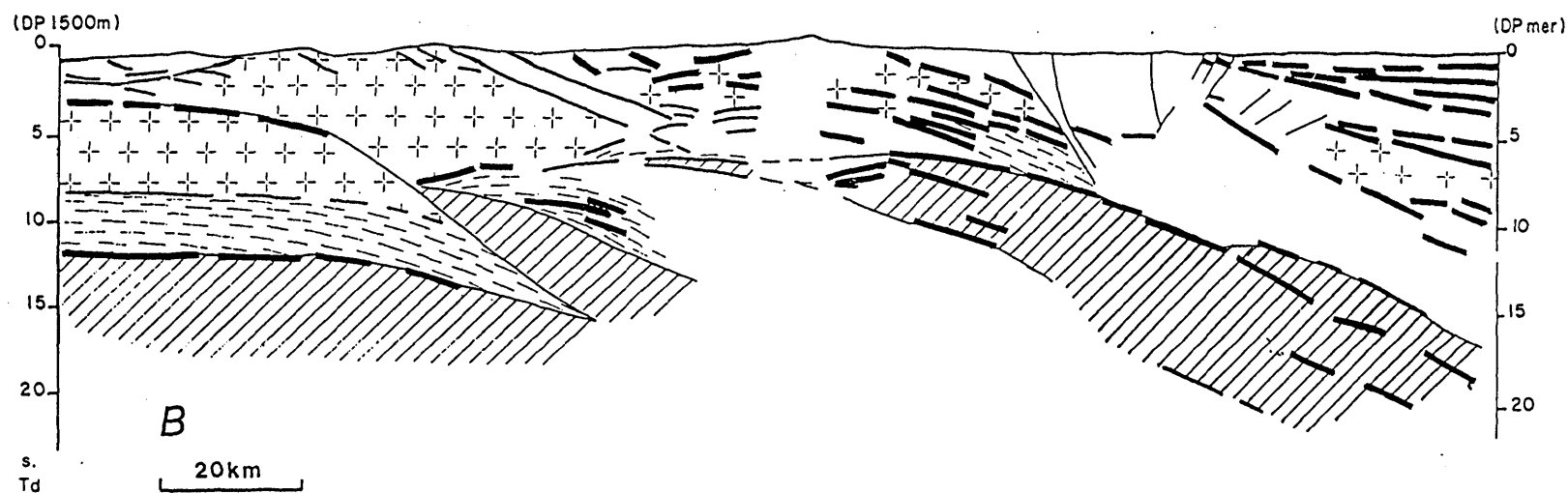


Figure 6.8: Two interpretations of the ECORS-CROP profile. A. Thrust imbrication of lower crustal rocks, B. involvement of mantle lithosphere in the thrusting. Crosses-upper crust, dashes-lower crust, diagonal hatching-mantle. Taken from Bayer et al. (1987).

consequence, is that convergence between the Italian and European plates appears to have largely been accommodated by the intra-plate deformation of Europe; both through a reworking of the pre-existing passive margin fault system and the development of new crustal thrusts. Some of these thrusts, such as the Belledonne thrust, generate foothills type thrust belts (the Subalpine Chains) in their footwalls as the contractional displacement sole out at the base of the Mesozoic cover sequence. The sequence of collapse and thrusting may not be as simple as is envisaged here and further work is required to date movements on individual structures. What is clear, is that the effects of early faulting and crustal extension may be substantial, as is the reactivation of these structures during collision. What remains is for these effects to be quantified. Estimates of crustal thinning and realistic estimates of displacement magnitude, are still required for input into crustal models constrained by seismic reflection data. These models can then be tested against models of plate convergence (see Roeder, in press).

Chapter 7: Conclusions

The conclusions of this thesis can be grouped and presented under several headings. Generally, they concern the Mesozoic to Tertiary structural and stratigraphic history of the study area and on a regional level, the geodynamic and tectonic development of a part of the European Tethyan margin and the external Western Alpine thrust belt.

1. Mesozoic extension and Tethyan passive margin development

i) During the Mesozoic, what now constitutes the Western Alpine thrust belt and its adjacent foreland in SE France, evolved as a rifted continental margin to the Ligurian Tethyan ocean. From combined structural and stratigraphic studies, the following history of rifting can be proposed:

Rifting was initiated during the Anisian (243 ma) in the Briançonnais and propagated westwards into the Dauphinois during the Carnian (231 ma). Subsidence was minor, but increased during the Hettangian and Sinemurian (213 ma). This corresponds to the first main subsidence event. A second subsidence event occurred at the Toarcian-Aalenian boundary (188 ma). These two events are interpreted to be the isostatic response to a mechanical thinning of the European crust by normal faulting. A final subsidence event commenced during the Callovian (169 ma) and was coeval with the onset of sea-floor spreading, marking the onset of the main thermal contraction phase.

ii) In the Pelvoux area, two normal fault systems of different age and geometry are correlated with the first two subsidence events. Upper crustal extension associated with the first subsidence event was accommodated on a system of normal faults subsequently rotated to more gentle SE-dips. These faults and their basal detachment are locally breached by a later set of steeply dipping normal faults related to the second main subsidence event. A NW-SE

extension direction is postulated for both of these events.

iii) A pattern of differential subsidence across the European margin is interpreted to reflect a long and complex heterogeneous stretching process, rather than a simple and single homogeneous event. A model of heterogeneous lithospheric stretching is consistent with the timing, geometry and spatial distribution of Mesozoic normal faults and can explain the negative subsidence anomaly (uplift) of the Briançonnais and rapid subsidence of the Dauphinois. The evolution of the European margin is speculatively modelled on two SE-dipping extensional detachment faults. These relate the Briançonnais subsidence anomaly to some deep-seated thermal event, reflecting the kinematics of lithospheric stretching.

iv) The effects of a Cretaceous extensional event were superimposed onto the thermally subsiding European margin. This is interpreted to reflect the formation of the Parentis basin and the Albian-Aptian opening of the Bay of Biscay. However, the local geometry and timing of this extensional event remain ill-defined.

v) The geometry of Tethyan normal faulting appears to have been influenced by and may partially reactivate, a pre-existing late Variscan fault pattern on the European foreland.

2. Structural inversion and the influence of pre-existing passive margin faults on Alpine thrusting

i) The Pelvoux area of the French Alps provides a classic example of structural inversion related to continental collision. A full range of inversion structures are present.

ii) The amount of Alpine contraction greatly exceeds the amount of earlier Tethyan extension and thrust detachment in the Pelvoux basement and decoupling of its syn- and post-rift cover sequence is important. The Oxfordian Terres

Noires easy slip horizon appears to have acted as a roof thrust to thrust faults which climbed out from the basement.

iii) Mesozoic passive margin faults offset potential stratigraphic decoupling horizons introducing heterogeneities into the pre-contractinal stratigraphic template. They strongly influenced the geometry and structural style of later Alpine thrust structures in the study area.

iv) The structural style is strongly influenced by the dip of the early normal fault. Gently dipping normal faults reactivate as thrusts during Alpine compression, inducing a decoupling of the syn- and post-rift cover sequence. Steeply dipping normal faults act as stress risers disturbing the regional stress trajectories and locally may cause:

a) A pinning of cover thrust displacements at the foot of the normal fault scarp i.e. buttressing. Penetrative and locally upright cover strains and complex back-fold-thrust systems result.

b) The formation of out-of-sequence thrust ramps in the hanging-wall of the pre-existing fault. These take thrust displacements out of the confines of the half-graben and transfer them along the base of the post-rift sequence (Oxfordian black shales). This then acted as a roof thrust.

v) Pre-existing normal faults controlled the position and orientation of folds and thrusts and fold trends often parallel the early basin bounding faults. Thrusts and tight hanging-wall folds in the study area are often curved in plan view and mimic the Jurassic normal fault geometry.

vi) On the basis of structural style, a two-fold zonation of the external parts of the Alpine thrust belt may be proposed:

a) An outer foreland zone characterized by Rocky Mountain foothills type thrusting: this corresponds to the area beyond the basin margin where sedimentation was largely controlled by a thermal mechanism of subsidence i.e. steers head configuration. The Jura and possibly the Subalpine Chains are equivalent to this zone.

b) An inner zone where the reactivation of pre-existing passive margin faults (structural inversion) is important. Crystalline basement massifs are formed as the extensional fault blocks are restacked. The structural style is not of a Rocky Mountain foothills type and balanced cross-sections constructed using thin-skinned principles may be misleading. Care must be taken not to force balanced layer-cake type sections and more attention given to the nature of the pre-contractional template i.e. the early basin history. The Pelvoux and other external basement massifs correspond to such a zone.

4. Timing and tectonic transport directions

i) The timing of the Alpine compression is complex and in the Pelvoux area, the deformation sequence is not a simple foreland-directed piggyback thrust sequence. Several directions of thrust transport were also recorded.

ii) The earliest compressional deformation is dated stratigraphically as pre-Senonian, and is best observed in the Devoluy. The tectonic transport direction appears to have been towards the N or NNW. The extent or importance of this deformation in the neighbouring Pelvoux massif is not clear.

iii) Alpine deformation affecting the Pelvoux basement and its Mesozoic cover can be divided into pre- and post-Nummulitique (Eocene flysch) events. The tectonic transport direction for the pre-Nummulitique deformation, appears to have been towards the WNW in the eastern Pelvoux and locally towards the SW in the southern Pelvoux, although the precise geometry and kinematic details of this event

remain poorly defined. Both directions are difficult to reconcile with the pre-Senonian deformation in the Devoluy. The main phase of Alpine shortening and important uplift is post-Nummulitique and of late Oligocene to Miocene age. In the Pelvoux this comprises a break-back, basement-involving thrust sequence. Structures which the break-back thrusts post-date in the central Pelvoux, cannot be dated conclusively. They may be pre-Nummulitique, or form part of a complex sequence of post-Nummulitique basement thrusting and inversion. A progressive anti-clockwise rotation of the post-Nummulitique thrust transport direction, from a WNW-direction in the northern Pelvoux, to a SW-direction in the southern part of the Pelvoux is evident. The SW-directed displacements locally reactivate pre-existing NE-SW trending Jurassic faults. Major post-Nummulitique thrusts are continuous through the central Pelvoux and can thus be traced around the Western Alpine arc, linking structures of the WNW and SW directed thrust systems.

iv) The Western Alpine arc is considered to be of a primary non-rotational origin reflecting the earlier passive margin fault geometry. The variations in thrust transport direction are attributed to, a) the local pinning of thrust displacements along the SW margin of the Pelvoux massif which may correspond to a pre-existing transfer fault trend (Pelvoux-Argentera axis), and b) plate rotations in the Mediterranean related to the opening of the Balaeric basins.

6. Deep crustal structure

Interpretations of the recent ECORS-CROP seismic reflection profile across the Western Alps, when combined with the surface-derived information presented in this thesis, provides some insight as to the extent and mechanism of crustal shortening. Before the release of the ECORS-CROP profile, the deep structure of the Western Alps was modelled on the reactivation of two pre-existing extensional detachment faults. However, the ECORS-CROP profile has shown that this may be too simplistic and

additional shortening and thickening of the European crust by thrusting during Alpine orogenesis must also be taken into consideration. Nevertheless, given the importance of inversion tectonics documented from the surface studies, the effect of Tethyan extension and thinning of the European crust must clearly have some bearing on the geometry of crustal deformation. The quantification of these extensional effects, and the incorporation of this data into models of the crustal structure is clearly something towards which future research should be directed.

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