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Arcuate fold and thrust patterns in the Subalpine Chains of Southeast France

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Abstract—The arcuate pattern of folds and thrusts is examined in the Subalpine Chains of Southeast France. Five segments are recognised; (1) a NNE–SSW trending zone between Grenoble and Die, (2) an E–W zone between Die and Sisteron, and (3); (4), & (5) the Arcs de Digne, Castellane, Nice, trending NNW–SSE, E–W, and N–S, respectively. Folds of three ages are known: pre-Senonian (recognised in Dévoluy), Paleocene to early Eocene and late Oligocene to late Miocene. The last phase was accompanied by low-grade metamorphism and cleavage development. Thrusts were formed during the last two phases.

Examination of the age of folding and thrusting, directions of tectonic transport and the continuity of structures reveals that segments (1), (2) & (3) are not continuous. Segment (2) has a northward sense of tectonic transport and is thought to be of pre-Senonian and Paleocene age. It is crossed in Dévoluy by the late Oligocene structures which are continuous from segment (3) into segment (1). Segments (3), (4) & (5) are continuous. Their curvature is thought to be primary and to have resulted from gravity gliding.

A section through the Digne Thrust Sheet is examined in terms of models of thrusting proposed by Elliott and by Chapple. It is found that the compressional component in Chapple's model was completely dominant in overcoming the resistance to sliding during the Miocene phase, the glacier spreading component having a negligible effect. Other thrusts, in particular those of the Arc de Castellane, do not appear to fit the model. It is thought that gravity gliding may have been the mechanism of thrusting in these areas.

Resumé—La virgation des plis et des chevauchements dans les chaînes subalpines du sud-est de la France a été étudiée. Cinq segments ont été définis; (1) un segment NNE–SSW entre Grenoble et Die, (2) un segment E–W entre Die et Sisteron et, (3), (4) & (5), les arcs de Digne, Castellane et Nice, respectivement NNW–SSE, E–W et N–S. Trois générations de plis sont connus, une phase pré-Sénonien definie en Dévoluy, une phase Eocène précoce et une phase tarde Oligocène à tarde Miocène. Cette dernière phase est accompagnée d'un métamorphism de faible degré et du développement de schistosité. Les chevauchements a sont produits pendant les deux dernières phases.

L'examen de l'âge des plissements et des chevauchements, des directions de transport tectonique et de la continuité des structures montre que les segments (1), (2) et (3) ne sont pas continus. Le segment (2) d'âge pré-Sénonien et Paleocène probable, a une direction de transport tectonique vers le nord. Cela est recoupé en Dévoluy par la phase d'âge tarde Oligocène, dont les structures sont continues entre les segments (3) à (1). Les segments (3), (4) et (5) sont continus, leur courbure est interpretée comme étant primaire et resultant d'un écoulement par gravité.

Une coupe à travers l'écaille de Digne est étudiée et comparée aux modèles de chevauchure proposés par Elliott et par Chapple. Il se montre que la composante de compression du modèle de Chapple était dominante et depassait de loin la résistance au glissement pendant la phase Miocène, la composante d'extension de type glacier ayant un effet négligeable. D'autres chevauchements ne correspondent pas au modèle, en particulier ceux de l'arc de Castellane. Pour ceux ci l'emplacement par gravité pourrait être le mécanisme de chevauchure.

INTRODUCTION

THE SUBALPINE Chains of Southeast France provide a fascinating area for the study of arcuate fold and thrust belts and for the study of thrust tectonics at high structural levels. The area of this study extends from Grenoble south around the outcrop of the External Zone to the Mediterranean Coast between Nice and Antibes (Fig. 1). Along the length of the External Zone within this area the trends of folds and thrusts change from NNE–SSW between Grenoble and Die to E–W between Die and Sisteron, NNW–SSE in the Arc de Digne, E–W in the Arc de Castellane and N–S in the Arc de Nice (Fig. 1).

Debelmas & Lemoine (1970) provide an excellent review of the geology of the Western Alps and Goguel (1963) discusses some of the specific problems of the Subalpine Chains. Recent structural maps and interpretations of the area have been compiled by Debelmas (1974), Vialon (1974) and Graham (1978). Debelmas & Kerckhove (1973) and Lemoine (1973) discuss the thrust tectonics of the Western Alps and include specific examples from this area. Debelmas (1970) presents an excursion guide to the Savoie and Dauphine Alps, which provides a useful collection of panoramas and cross-sections.

The aims of this study are twofold.

(a) To examine and inerpret the arcuate pattern of folds and thrusts.

(b) To reassess the relative roles of gravity gliding and crustal compression in the thrust tectonics of the area in the light of recent work on thrusting (Elliott 1976a, b, Chapple 1978).

STRATIGRAPHY

The total thickness of the Mesozoic and Tertiary cover of the Subalpine Chains varies from about 2 to 7 km. There are significant stratigraphic variations across and along the length of the Chains (Figs. 2 and 3); these

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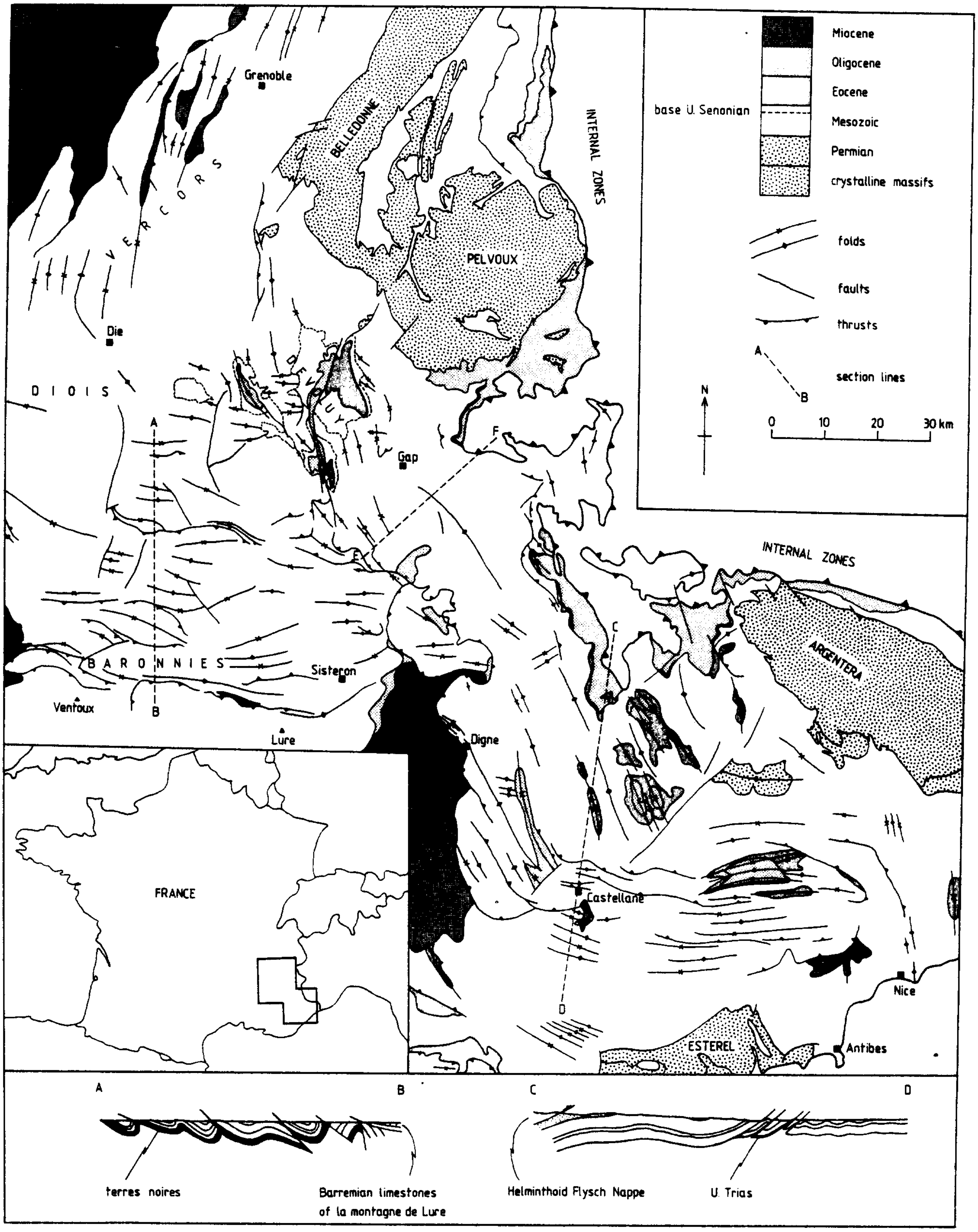


Fig. 1. Structural map of the Subalpine Chains of Southeast France. After La Carte des Alpes Occidentales, 1:250,000, B.R.G.M., J. Goguel (1936), A. Faure-Muret (1955), Flandrin (1966) and Graham (1978). Section C–D after Lemoine (1973).

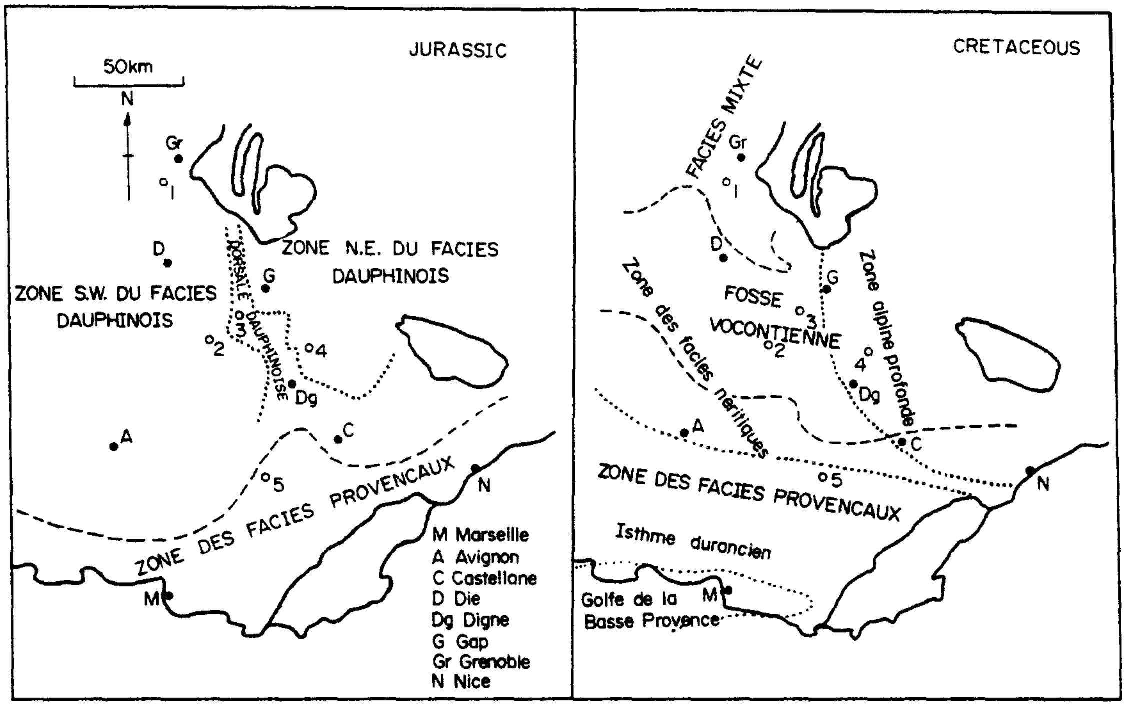


Fig. 2. Sketch maps showing the distribution of Mesozoic facies in Southeast France. Based on data from Gignoux & Moret (1944), Gignoux (1955), Artru (1967), Gariel (1961), Mouterde (1961), Riche *et al.* (1961), Debelmas (1970) and Debelmas & Lemoine (1970). External crystalline massifs are outlined by solid lines. Open circles refer to numbered sections in Fig. 3.

Jurassic—dotted lines refer to Lias and Middle Jurassic facies, broken lines to Upper Jurassic facies. Cretaceous—broken lines and capital letters refer to Lower Cretaceous facies, dotted lines and lower case lettering to Upper Cretaceous facies.

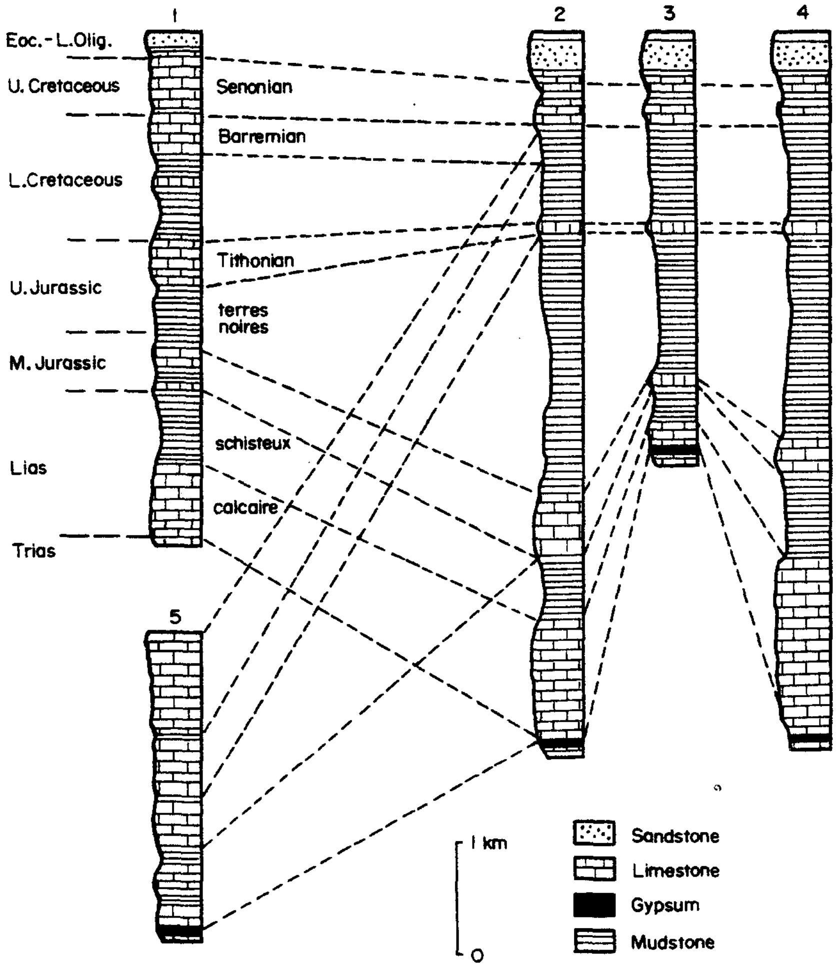


Fig. 3. Representative stratigraphic sections for the numbered sites shown in Fig. 2. Based on same data sources as Fig. 2 plus Subieta (1977).

variations to some extent controlling the tectonics. The controls are the occurrence of incompetent marly or evaporite-bearing horizons which provide décollement levels, and massive, competent limestone formations which behaved as relatively rigid blocks and now control the topography.

In the north of the area, in Vercors, the Trias succession lacks the evaporite-bearing formation and the Lias and Dogger appear to have remained stuck to their crystalline basement, the Belledonne Massif. The dominant décollement horizon is the 'terres noires' of Oxfordian age. Above this, limestones of Tithonian and Barremian age appear to have controlled folding. In the southern Subalpine Chains evaporite-bearing dedolomitised limestones, the cargneules, appear in the Upper Triassic succession. Below these, the basal Trias quartzite and local Permian redbeds have remained stuck to their crystalline basement, the Argentera and Esterel Massifs. Sedimentation during Jurassic times appears to have been controlled by a residual Hercynian relief and by active faults. A series of basins in which clastic formations, notably the 'terres noires' and Lias schisteux, attain their maximum thickness, extended obliquely to the subsequent Alpine Arc (Figs. 2 and 3). During Early Cretaceous times an E–W trough, the Vocontien Basin, evolved (Fig. 2) in which sedimentation was more uniformly marly (Fig. 3). In particular, the Urgonian facies of the Barremian is missing in Baronnies and the Tithonian limestones become thin. In these regions it is the Upper Cretaceous limestones that appear to have controlled folding. Further south still, in the Arcs de Castellane and Nice, Upper Jurassic limestones thicken and a pure limestone of Barremian age appears in the Montagne de Lure and Mont Ventoux (Fig. 3), although this thins towards Nice. The Upper Cretaceous limestones thin to zero on the Durancian Isthmus (Fig. 2). It was only in late Cretaceous and Tertiary times that sedimentary facies followed trends parallel to the Alpine Arc. In eastern parts of the External Zone the Nummulitique, of Eocene and Oligocene age, is represented by a marine flysch facies, for example the Grès d'Annot and the Grès de Champsaur. Locally they are strongly transgressive onto the 'terres noires' in the Gap Basin and the Pelvoux Massif to the north. Westwards the Oligocene and Miocene occur only as continental facies in restricted basins.

DEFORMATION PHASES—THE AGES OF FOLDS AND THRUSTS

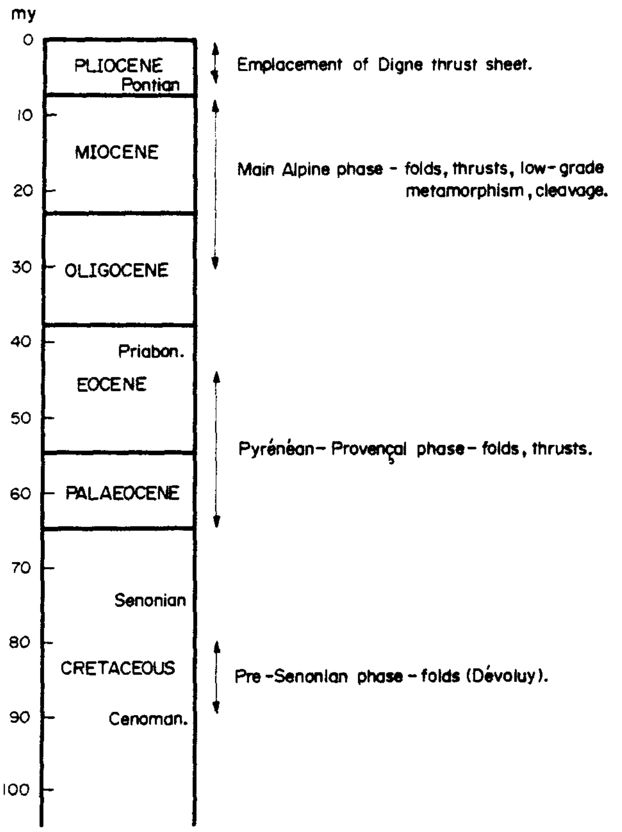
Three periods of fold-producing deformation can be recognised in the Subalpine Chains by conventional stratigraphic criteria, that is the ages of the strata involved in the folds and the ages of the strata that overstep eroded folds. They are (Table 1):

(1) *Pre-Senonian folds.* In Dévoluy rocks of up to Cenomanian age are involved in locally tight, upright, generally E–W trending folds that are overlain unconformably by Upper Senonian limestones (Flandrin 1966, Debelmas 1970, fig. 84). Although usually described as trending E–W, some folds of this age trend NNW–SSE (Fig. 1). No thrusts of this age are known to the author, though presumably, like the later folds, they result from décollement along a weak layer above the basement.

(2) *Pre-Nummulitique folds and thrusts.* About 12 km west of Gap, Upper Senonian limestones are involved in E–W trending folds that are overlain unconformably by nummulitic limestones of Priabonian age. E–W trending folds of this age are also found in Baronnies (Flandrin 1966) and in the Arc de Digne (Graham 1978). Some of the E–W trending thrusts in Baronnies are also of this age (Flandrin 1966). These pre-Nummulitique folds are the 'plis anciens' of Goguel (1963) and were described as belonging to the Pyrénéan–Provençal phase by Lemoine (1972).

(3) *Late Oligocene – post-Miocene folds and thrusts.* Various authors have suggested a late- or post-Miocene age for the main Alpine phase of deformation in this area (Goguel 1963, Gidon 1964, Debelmas 1974). The evidence for this is that Miocene rocks are involved in the folds (e.g. Debelmas 1970, fig. 62). Elsewhere, Miocene rocks unconformably overlie folds that involve Oligocene rocks (Fig. 1). In the Baronnies area both situations are known (Flandrin 1966). It may be the case that some of the Miocene molasse was deposited in basins that developed over growing synclines of this age. The regional cleavage and low-grade metamorphism in

Table 1. The timing of events in the Subalpine Chains of Southeast France



the eastern parts of the Subalpine Chains are also associated with deformation of this age. Like Graham (1978), this author considers the bulk of main Alpine phase deformation to be of pre- or Miocene age. Note, however, that Debelmas (1970, fig. 4, 1974) considers it to be of Pliocene–Quaternary age. Throughout the area Miocene rocks are overthrust by older rocks (Fig. 1). Final emplaement of the Parpaillon Nappe and its basal slices, of Internal Zones origin, in Embrunais and Ubaye also occurred in Miocene times (Kerckhove 1969).

Later events include the final emplacement of the Digne Thrust Sheet over a known distance of 7 km, dated on stratigraphic grounds as Pliocene (Gigot *et al.* 1974), the flexing of the thrust plane and the Tertiary Valensole Basin, and the rejuvenation of pre-existing faults, which continues to produce modern seismicity in a N–S compressive regime (Philip & Tapponnier 1976).

THE ARCUATE FOLD AND THRUST PATTERN

The overall pattern of folds and thrusts in the Subalpine Chains is shown in Fig. 1 (see also Debelmas 1974, Vialon 1974, Graham 1978). It is immediately apparent that individual folds are of limited axial extent. Most can only be followed for about 10 km along their axial traces. The structural map of Debelmas (1974) is similar in this respect, although Graham (1978) shows longer axial traces, and Vialon (1974) shows only trends, not individual folds. This point becomes of importance in discussion of the continuity of structures around the arcs. The arcuate pattern between Grenoble and Nice falls into five segments.

(1) *Grenoble – Die.* Folds have a fairly uniform NNE–SSW orientation with a few exceptions trending N–S or NNW–SSE. Where the ages of folds and thrusts can be demonstrated they belong to the main Alpine phase. The sense of vergence of fold axial planes and the displacement directions on thrusts (Debelmas 1970, fig. 61) indicate a tectonic transport direction from the E towards the W.

(2) *Die – Sisteron.* A detailed structural map and discussion of this area are given by Flandrin (1966). Folds have an approximately uniform E–W trend except in Dévoluy, where NNW–SSE folds belonging to the pre-Senonian and main Alpine phases are found. The E–W folds in Dévoluy and probably in Diois are of pre-Senonian age, while those in Baronnies are of the Pyrénéan–Provençal or main Alpine phase. Most of the folds and thrusts cannot be dated stratigraphically, but the main Alpine phase folds appear to be reactivated folds of the Pyrénéan–Provençal phase. Many E–W thrusts involve Miocene rocks but their geometry is so closely related to folds of the Pyrénéan–Provençal phase that it seems likely that they originated during that phase, and were reactivated during the main Alpine phase (Flandrin 1966). Some can be seen to be of the Pyrénéan–Provençal phase. The sense of displacements on thrusts and the direction of vergence (Fig. 1, section A–B) indicate a tectonic transport from S to N. Geophysical work in Diois and Baronnies (Flandrin & Weber 1966) suggests that the basement is not directly involved in the E–W folds, though there is a possibility that the E–W thrust forming the northern boundary of La Montagne de Lure may be situated over a basement fault. Like Debelmas (1974), but unlike Graham (1978), the present author believes that individual folds do not pass continuously from segment 1 into this segment. One, near Die, appears to have the correct curvature, but dies out before attaining an E–W orientation. Northwest of Gap outcrop patterns suggest intersecting N–S and E–W trending folds, but their ages are unknown. Further northwest N–S trending main Alpine phase folds are clearly superposed on E–W trending pre-Senonian folds in Dévoluy. Towards the east this segment passes below the Digne Thrust Sheet.

(3) *Arc de Digne.* Pyrénéan–Provençal phase folds trend WNW–ESE southeast of Digne and are crossed by main Alpine phase folds trending NNW–SSE. Thrusts are parallel to the main Alpine phase folds. The Digne Thrust itself probably originated during the main Alpine phase (see later section). Displacement directions on thrusts and fold vergence (Debelmas 1970, fig. 107) indicate tectonic transport from NE to SW. Northwards, this segment passes into the Digne Thrust Sheet, where E–W trending folds of the Pyrénéan–Provençal phase are known, and NE–SW trending folds of unknown age are also found (Fig. 1).

(4) *Arc de Castellane.* Folds and thrusts trend generally E–W. Those that involve the Nummulitique are of the main Alpine phase. Of the others some can be demonstrated to be of the Pyrénéan–Provençal phase, but the age of most is uncertain. Displacement directions on thrusts and fold vergence (Fig. 1, section C–D) indicate tectonic transport from N to S. Like Debelmas (1974), but unlike Graham (1978), the present author cannot trace individual folds from segment 3 into this segment, although some thrusts link the segments (Fig. 1).

(5) *Arc de Nice.* Folds and thrusts trend N–S or NNW–SSE. Where their age can be demonstrated they are of the main Alpine phase. Only thrusts can be followed around from segment 4 into this segment, where tectonic transport directions are from the E towards the W.

Continuity of structures around the whole arc from Grenoble to Nice is not observed. In particular, the E–W zone between Die and Sisteron has an anomalous tectonic transport direction relative to the rest of the arc, and it contains a high proportion of folds that are of pre-Senonian age in northern parts and of the Pyrénéan–Provençal phase in southern parts. These folds are clearly crossed by N–S trending main Alpine phase folds in Dévoluy.

INTERPRETATION OF THE ARCUATE FOLD AND THRUST PATTERN

Goguel (1963) and Gidon (1965) suggest that the arcuate fold and thrust pattern in the Subalpine Chains may be due to the response of the cover rocks to sinistral displacements along NE–SW transcurrent faults in the basement. This idea has been considerably developed by Viadon (1974) and modified by Graham (1978). Palaeomagnetic data on Permian rocks from the Argentera Massif, dôme de Barrot and Esterel Massif (Bogdanoff & Schott 1977) preclude the idea of subsequent bending of an originally straight N–S belt; and, as Graham (1978) notes, ductile shear zones in the cover rocks cannot be reasonably invoked to explain the two E–W segments.

Vialon (1974), in a far reaching discussion, draws attention to the late-Hercynian fault systems trending N 50° E (failles cévenoles) and N 170° E. He notes the role of the N 50° E system in controlling Mesozoic sedimentation and suggests that their reactivation as sinistral transcurrent faults, during N–S crustal compression, produced the arcuate fold pattern seen in the cover rocks. In particular, it is suggested that such faults extend from Nimes to the SE border of the Pelvoux Massif and through the Arc de Castellane (Vialon 1974, fig. 9). This hypothesis is further developed by Boudon *et al.* (1976), who interpret the evolution of the Western Alps in terms of persistent N–S compression acting on an assemblage of basement blocks. Shearing along the margins of these blocks is thought to have reoriented the pre-Senonian and Pyrénéan–Provençal phase folds and caused the main Alpine phase folds.

Graham (1978) reports the results of model simulations using wood blocks to represent the basement and wet tissue-paper to represent the cover rocks. He finds that sinistral displacement along two transcurrent faults, the Diois and Provençal Faults (Graham 1978, fig. 9), must have taken place during the main Alpine phase at the same time as the cover rocks were being compressed and folded. Estimated figures for the amounts of displacement are 15 and 10 km, respectively.

Entirely independent evidence for large-scale sinistral transcurrent faults through this area during Late Cretaceous times is presented by Debelmas (1975) and by Trümpy (1976). They use stratigraphic arguments involving the geographic distribution of the Helminthoid Flysch. Trümpy (1976) suggests that such a fault zone may be the eastern extension of the North Pyrenean Fault.

There are, however, a number of arguments against this basement transcurrent fault hypothesis. Trümpy (1976) points out the problem of reconciling such fault movement contemporaneous with the high-pressure eoalpine metamorphism (100–60 Ma, Frey *et al.* 1974), probably as a result of subduction at the European Plate margin (Desmons 1977, Tapponnier 1977). Graham's (1978) argument encounters difficulties in that the eastern extension of the Provençal Fault does not match up with the change in foliation orientation in the Argentera Massif (Graham 1978, fig. 11 and the elusive fig. 12). There is also a mechanical argument that makes the hypothesis seem intuitively wrong. Folding and thrusting in the cover rocks of the Subalpine Chains is permitted by décollement along the evaporite-bearing Upper Trias, the Lias schisteux or the 'terres noires' (Goguel 1952). This means that these horizons were mechanically weak, so that they could not transmit large shear stresses from the formations above to those below. The formations above can thus slide and/or fold independently of those below. The basement transcurrent fault hypothesis requires that the opposite be the case, that is sufficiently large shear stresses can be transmitted upwards through the décollement horizons to bend the formations above. This seems intuitively impossible. An alternative explanation is thus sought for the origin of the arcuate pattern.

As described above, continuity of structures around the three northern segments is not seen and the directions of tectonic transport in the two E–W segments are opposite to each other. It is thus not necessary for the whole arcuate structure to be explained by a single mechanism. The E–W segment between Die and Sisteron seems to be anomalous in both its tectonic transport direction and the age of its folding and thrusting, which is thought to belong mainly to the pre-Senonian and Pyrénéan–Provençal phases, with minor reactivation during the main Alpine phase. South of Baronnies, pure white limestones of Barremian age appear in the Montagne de Lure and Mont Ventoux, south of which they are undeformed. Further south still, between the Durance river and the Mediterranean coast around Marseille, the E–W fold and thrust belt of Provence is found. The folds here are of the Pyrénéan–Provençal phase (Goguel 1963) and the direction of tectonic transport is also towards the north (Debelmas 1974). Like Flandrin (1966), it is suggested that the region from Diois and Dévoluy south to the Mediterranean coast forms a single tectonic unit with a relatively undeformed central area related to the presence of the Tithonian and Barremian limestone facies. The fold and thrust belt was presumably caused by either a northwards compression or northwards gravitational gliding during the Pyrénéan–Provençal phase (Fig. 4).

Intersecting N–S and E–W folds are exposed northwest of Gap and in Dévoluy. The N–S outcrops of Oligocene rocks in Dévoluy occupy the cores of main Alpine phase synclines (Debelmas 1970, fig. 87). These are clearly superposed on the earlier E–W folds. The N–S to NNE–SSW trending folds of Vercors are also of main Alpine phase age where their age can be demonstrated with certainty. Recent work on the pattern of main Alpine phase cleavage found in the cover of the Belledonne, Grandes Rousses and Pelvoux Massifs, dôme de Remollon, demi-fenêtre d'Embrun and the Digne Thrust Sheet (Gratier *et al.* 1973, Siddans 1977, Subieta 1978) indicates the continuity of the main Alpine phase deformation from the Digne Thrust Sheet northwest into Vercors, cutting across the earlier E–W fold and thrust belt. Tectonic transport directions during the main Alpine phase, in both the Digne Thrust Sheet and Vercors, are from the E or NE towards the W or SW. These northern parts of the arcuate Subalpine Chains are thus thought to result from intersecting fold and thrust belts of different ages (Fig. 4). Parallelism of main

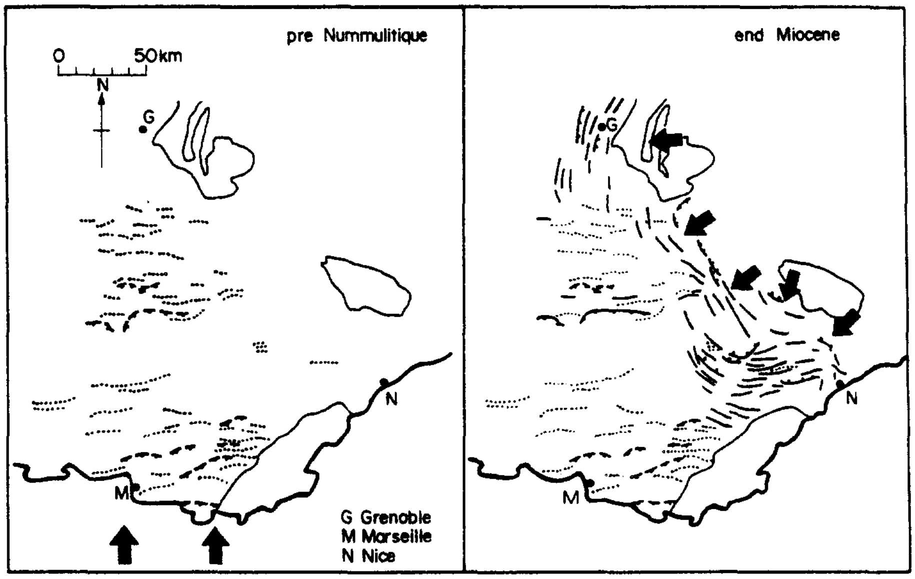


Fig. 4. Structural sketch maps showing the development of the Southern Subalpine Chains.
Dotted lines—axial traces of Pyrénéan–Provençal phase folds.
Broken lines—traces of Pyrénéan–Provençal phase thrusts.
Solid lines—axial traces of main Alpine phase folds and thrusts.

Alpine phase and Pyrénéan–Provençal phase structures in Baronnies is attributed to reactivation of the earlier structures by the later phase.

Continuity of structures is seen to some extent in the Arcs de Digne, Castellane and Nice. Although early E–W folds are known, main Alpine phase folds and thrusts occur in all segments. The direction of tectonic transport in all segments is consistently outwards, that is away from the Internal Zones, indicating compression from, or gravity gliding off, the more internal parts. The curvature of these segments is thus thought to be primary, possibly resulting from gravity gliding and spreading off the internal parts of the late-Oligocene – post-Miocene Alpine Chain which were topographically high (Fig. 4). This would require either two relatively high spots, situated approximately at the sites of the present northwest and southeast ends of the Argentera Massif, or a single low spot in the present site of the central Argentera Massif. Two lines of evidence support the latter hypothesis. These are the continuity of main Alpine phase structures from the Arc de Digne northwest into Dévoluy, and the observations of Bouma & Stanley (1964) on palaeocurrent directions in the Grès d'Annot, which are from the south, indicating that the Argentera Massif was not exposed as a topographic high during the Nummulitique. Evidence for low ground around the Arcs de Digne, Castellane and Nice during Miocene times is seen in the distribution of Miocene basins.

THRUSTING IN THE SUBALPINE CHAINS

"We must remember that the mathematical tool is only applied to a model, not the geological body or phenomenon. The chosen model has to be as simple as possible to permit calculation and this simplicity is fascinating" (Lemoine 1973).

Lemoine (1973) provides an excellent review of ideas on the thrust tectonics of the Western Alps. Notable contributions were made by Goguel (1948) in his remarkable memoir, in which he noted the lubricating effect that could be provided by plastic deformation of a weak basal layer; for example, a sheet 4 km thick can slide down a slope of 2° under the action of gravity if it has a weak basal layer of yield stress about 40 bars. Such a situation exists in many Subalpine thrust sheets where the evaporite-bearing Upper Trias is a notable décollement horizon. Bordet (1949) suggested that folds in the frontal parts of thrust sheets (e.g. Fig. 1, section C–D) may be produced by shortening in the locked frontal parts of gravity-slipped sheets. Goguel (1949), however, observed that localization of folds may be due to thickness variations in the sedimentary succession. He also noted that the Subalpine thrust sheets may have been pushed by the overriding Pennine Nappes. Lemoine (1973), after consideration of the slope problems, kinetics and possible mechanical models of gravity gliding, concluded that gravity gliding may have played a role during folding of the Subalpine Chains, but it is very likely that this role was secondary and occurred mainly at the end of the folding.

Two recent models of thrusting (Elliott 1976a, b, Chapple 1978) may provide further insight into the mechanics of these thrust sheets. The model of Chapple (1978) appears to include some of the characteristics of some Subalpine thrust sheets, in that a weak basal layer is involved, the thrust sheet is prism-shaped and deformed internally at its back end. In practice it is difficult to find thrust sheets whose geometry and post-emplacement history are sufficiently well defined to apply the model. The Digne Thrust Sheet, however, is well exposed in the dôme de Remollon and the demi-fenêtre d'Embrun. A vertical cross-section is shown in Fig. 5, it was drawn approximately in the direction of tectonic transport as indicated by fold geometry and the 'bow-and-arrow' rule (Elliott 1976b), and after restoration of post-thrusting extension faults and diapiric gypsiferous masses. Displacements in the trace of the present-day topography indicate the positions of these faults. No attempt has been made to balance the cross-section because there are primary thickness variations in Jurassic formations, and the penetrative deformation associated with cleavage development is heterogeneous and not plane strain, so that there are area changes along the line of this section. The level of the basement is only known for central parts of the section where it crops out in the dôme de Remollon. It is assumed that some sort of basement horst existed along the line of the Dorsale dauphinoise (Fig. 2). This is indicated at the southwest end of the section where it is suggested that its presence may have provided the ramp above which folds developed in the thrust sheet. In northeast parts of the section the stratigraphy is exposed in the demi-fenêtre d'Embrun only down to the Lias calcaire, though the gypsiferous Upper Trias crops out in the Barcelonnette area to the southeast. Below the gypsums, lower parts of the Trias succession probably thicken towards the east. The basal layer of the thrust sheet is the gypsiferous Upper Trias formation.

The thrust sheet was deformed together with an overlying slice of allochthonous rocks, of Subbriançonnais Zone origin, during the main Alpine phase (Debelmas & Kerckhove 1973) prior to emplacement over Miocene rocks (Fig. 1). The increase in main Alpine phase deformation towards the back end of the sheet was outlined by Siddans (1977). The strain data of that study, in the Lias schisteux and 'terres noires', has been projected onto the section plane and the resulting ellipses are shown in Fig. 5. From a study of illite crystallinity in the Digne area, in the Rhaetic above and below the thrust plane, Dunoyer de Segonzac & Abbas (1976) were able to demonstrate that the thrusting produced a metamorphic inversion, the crystallinity in the hanging-wall being about 4.0 compared with about 7.5 in the foot-wall, suggesting a displacement of some 20 km. Also, near Digne a displacement of at least 7 km along a Pliocene erosion surface can be demonstrated as post-Pontain (Gigot *et al.* 1974), giving a minimum average displacement rate of about 1 mm/y. The thrust sheet is thus interpreted as having had a two-stage evolution, the post-Pontian movements and an earlier displacement associated with the main Alpine phase pre-dating the arrival of the Parpaillon Nappe and its basal slices.

Consider first the early displacements. The geometry of the sheet at the time of thrusting is uncertain, the uncertainty arising from the extent of erosion prior to the emplacement of the Embrunais nappes and the curvature of the sheet. In the line of section the base of the Embrunais nappes forms an approximately horizontal trace, which can be reasonably continued along the length of section indicated by a solid line. The observed down-cutting of the base of the Embrunais nappes clearly suggests that the upwards curvature of the sheet was either primary or due to flexing after displacement but before the arrival of the Embrunais nappes. The pattern of the Eocene overstep suggests it was primary. The geometry of the sheet is best considered under les Parias. The backslope of the lower surface of the sheet (cf. angle θ, Chapple 1978, fig. 2a) is taken as the tangent drawn to the thrust plane dipping 2.5° NE (Fig. 5). The forward slope of the upper surface of the sheet (cf. angle α, Chapple 1978, fig. 2a) is taken to be parallel to the inferred Miocene erosion surface along which the Embrunais nappes were emplaced, that is a horizontal surface. The minimum thickness of the sheet (cf. h_o, Chapple 1978, fig 2a) under les Parias is 3.5 km.

A value of 500 bars is taken as the yield stress for the material of the thrust sheet above the basal layer. This figure is chosen for the reasons discussed by Chapple (1978) combined with the lack of sandstones and presence of mudstones in this particular thrust sheet. Putting these values into Chapple's (1978) equation 35, the yield stress of the weak basal layer evaluates at 43.6 bars, which is about the same as that established by Goguel (1948) for the gypsiferous Upper Trias. With zero surface slope the glacier-sliding term becomes zero, indicating that the compressional term was completely dominant in overcoming the resistance to sliding of the weak basal layer. The physical meaning of this is that, if the thrust sheet was originally a uniform sheet dipping 2.5° NE, no displacement would have occurred until compression had thickened the back end of the sheet so that the surface slope was zero. According to the Chapple model the sheet would have stopped moving when the compressive stress was dissipated.

From the strain data shown in Fig. 5 the reciprocal quadratic elongations parallel to the thrust trace in this line of section have been calculated. The method of Hossack (1978) can then be used to evaluate the bulk shortening in the thrust sheet associated with the main Alpine phase deformation. This is also shown in Fig. 5. Between la Saulce and les Parias, now a distance of 29 km apart, the square root of the bulk reciprocal quadratic elongation is 1.25, indicating a shortening of 20%. The part of the thrust sheet under les Parias has thus been displaced about 7.5 km further to the SW than that at la Saulce.

Subsequent emplacement of the Embrunais nappes onto the back end of the thrust sheet must have effectively increased the forwards surface slope and thickness of the combined sheet. A tangent drawn to the present-day topography dips 3.5° SW and gives a thickness of 4.5 km (cf. Elliott 1976a, fig. 1). This is right at the lower end of the range of thicknesses that Elliott's model applies to, but it is a minimum thickness. The density of Lias schisteux and 'terres noires' samples in this area is about 2.7 (Siddans 1977). Substituting these values into equation 4 of Elliott (1976a) gives a mean basal shear stress of about 70 bars which, according to the above argument and Goguel (1948), is considerably in excess

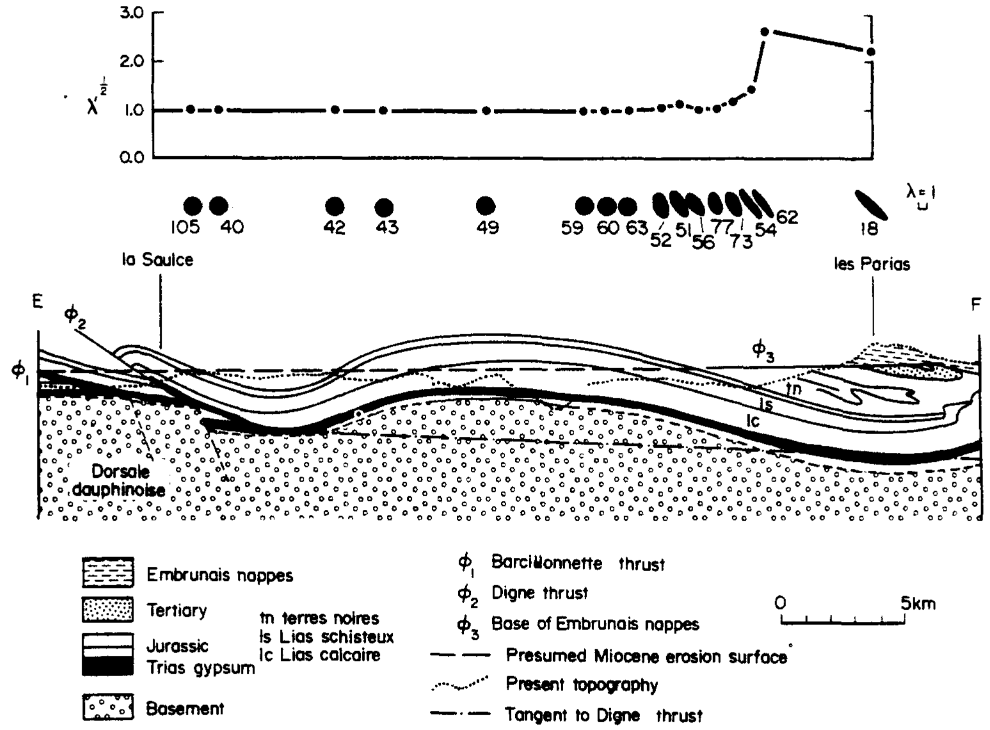


Fig. 5. Vertical section along line E–F (Fig. 1) after restoration of post-thrusting extension faults and diapirism. In part after La Carte Géologique de la France, 1:80,000 sheet 200 (Gap). Strain data from Siddans (1977), to which numbered strain analyses refer, scale bar λ = 1, is length of radius of sphere of same volume as strain ellipsoid. Upper diagram shows variation of the square root of the reciprocal quadratic elongation parallel to the trace of the thrust plane along the section.

of the yield stress for the Upper Trias gypsums. It thus seems possible that the piggy-back effect of emplacement of the Embrunais nappes onto the back end of the Digne Thrust Sheet produced the observed post-Pontian displacement of the sheet. In this case it is appropriate to ask why it stopped moving. The answer is probably because of the three-dimensional shape of the sheet. At some stage the thrust plane probably stopped propogating laterally, so that the sheet became pinned by its ends.

It appears that for the Digne Thrust Sheet emplacement may have been achieved along the lines suggested by Goguel (1949) and Lemoine (1973). Displacement was associated with the main Alpine phase and related to compression at the back end of the sheet at the time of the approach of the Embrunais nappes, and was followed by gravity induced sliding after the emplacement of the Embrunais nappes had provided a forward sloping surface. Application of the Elliott and Chapple models to this sheet shows that such a history is mechanically possible, though not necessarily the true history of emplacement. Other thrust sheets in the Subalpine Chains do not appear to have the characteristics of the models. In particular, internal deformation seems to be totally lacking in the thrust sheets of the southern Arc de Digne and the Arc de Castellane. Perhaps gravity gliding as envisaged by Goguel (1948) and Fallot & Faure-Muret (1949) occurred. A 3° slope implies an altitude difference of only 2 km along a slope 40 km long, less than the maximum present-day difference in altitude between the Argentera and Esterel Massifs. However, the present-day topography along this line could hardly be described as a smooth slope, it resembles more the 'nappe-traps' of Lemoine (1973). Alternatively, as discussed earlier, a forward surface slope of 3.5° provides more than an adequate mean basal shear stress, according to the Elliott model, for a sheet 4.5 km thick with the gypsiferous Upper Trias at its base.

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