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Alpine tectonics and diapiric structures in the Pre-Betic zone of southeast Spain *

F. MOSELEY, J. C. CUTTELL, E. W. LANGE, D. STEVENS and J. R. WARBRICK

Department of Geological Sciences, University of Birmingham, Birmingham, B15 2TT, England

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Abstract—The Pre-Betic is the most northerly of the Alpine zones forming the Betic Cordilleras of southern Spain. It consists of strongly folded and faulted Mesozoic and Tertiary rocks, the oldest of which are ferruginous and gypsiferous Triassic mudstones, followed by a predominantly carbonate facies of Cretaceous, Palaeogene and Miocene age. Although this sequence is interrupted by a number of minor unconformities, the major structures were formed during the middle or late Miocene. The highly incompetent Triassic rocks are the most strongly deformed, and form diapiric intrusions discordant to regional structural trends in the younger rocks. The latter are essentially of two facies; massive competent limestones which are deformed by relatively simple folds of large wavelength, and highly incompetent marl–limestone interbeds with complex disharmonic folds and crush belts. Faults include low-angle and high-angle thrusts, gravity slides and wrench faults. The regional tectonic strike is ENE to NE, but the diapiric intrusions mostly follow WNW and N directions. These intrusions have pushed the younger rocks aside, the result being polyphase structures of several trends.

Less intense post-Miocene tectonics are mostly associated with continued diapirism and have resulted in the folding and tilting of the late Miocene to Quaternary clastic sediments.

INTRODUCTION

THE PRE-BETIC is the most northerly structural zone of the complex Alpine fold belt of the Betic Cordilleras, which extend from Cape Nao in southeast Spain to Gibraltar in the southwest, a distance of 600 km. The fold belt is bounded to the north by the Hercynian Meseta and is subdivided from south to north into the three major zones; (a) the Betic (*sensu stricto*), (b) the Sub-Betic and (c) the Pre-Betic (Fig. 1). The Betic zone is composed of a great variety of nappes which range from the metamorphic nappes of the Nevado-Filabride Complex, through the variable metamorphic rocks of the Alpujaridde Nappes, to the non-metamorphic Carboniferous and Mesozoic rocks of the Betic of Malaga Nappe. To the north of the Betic zone there is the non-metamorphic Sub-Beltic zone made up of a series of Mesozoic carbonate nappes, and north of this again lies the largely autochthonous Pre-Betic zone of complex folds in Mesozoic and Tertiary rocks.

The general geology of the Pre-Betic zone has been described by Darder (1945), Rios *et al.* (1958, 1961), Moseley (1968, 1973) and Champetier (1972). It mainly consists of Triassic ferruginous and gypsiferous mudstones followed by largely carbonate sequences of Cretaceous and Tertiary age. The culmination of the Alpine orogeny occurred midway through the Miocene. Post-orogenic deposits include late Miocene gypsiferous marls and Pliocene to Quaternary fanglomerates and calcretes.

The area described here consists of 600 km^2 of steep mountainous terrain with many peaks and ridges exceeding 1000 m. Exposure is consequently good, with 200–500 m vertical cliffs commonplace, but access can be difficult because of extensive tomillares (garigue). One of us (F.M.) has worked in this region for many years, the others surveyed small critical areas during undergraduate thesis work.

OUTLINE OF STRATIGRAPHY

The oldest strata in the region are of Triassic age, the most common rocks being red ferruginous shales and mudstones with grey banding and a high proportion of gypsum, the latter forming beds up to 20 m thick. Gypsum is also a major component of the tectonic breccia which is a common feature of these rocks. Other Triassic rocks include massively-bedded limestones and dolomites, which form rafts up to 1 km long, isolated and tectonically emplaced in the mudstones during diapiric uplift (see below). There are also basic igneous rocks which, like the massive limestones, occur as blocks tectonically isolated by diapiric uplift.

Jurassic rocks are not present in the area but Cretaceous limestones and marls are important. Massive limestone, especially of Turonian and Cenomanian age, forms high precipices and is highly competent, constrasting strongly with the marl and marl–limestone interbeds (see below). Palaeocene, Eocene and Oligocene strata are of similar lithologies to those of the Cretaceous and are equally important, with competent massive limestones and incompetent marls having important influences on structural style. A small but significant unconformity separates the Eocene from the Oligocene, and locally (for example northeast of the area shown in Fig. 1) Oligocene strata rest directly on upper Cretaceous limestones.

There are extensive Miocene outcrops, but the true stratigraphical junction between the Oligocene and Miocene has not yet been determined in this region. For

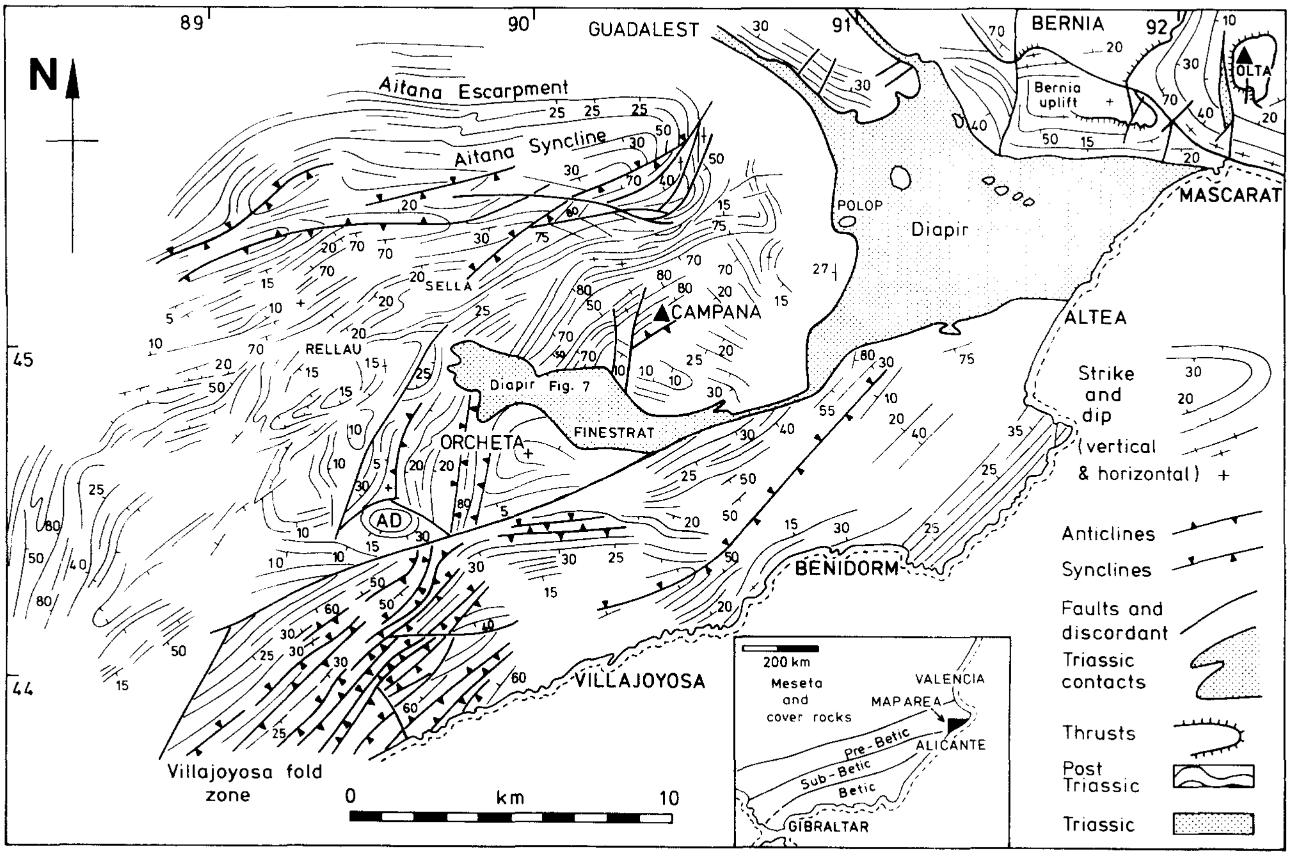

Fig. 1. Structural map of the southern Costa Blanca. The blank areas are alluvium, fanglomerate and unexposed areas of terrace cultivation. AD, Amadorio dome.

mapping purposes however it is convenient to take it at a strong facies change between massive limestones and a sequence of glauconite marls, sandstones and sandy limestones. The latter are followed by a thick sequence of marls, sandstones and conglomerates exposed east of the area depicted in Fig. 1. Their deposition preceded the main phase of the Alpine orogeny, most of the structures described below belonging to this episode. It is not possible to state a precise date for the earth movements because the Miocene sequence is not known in sufficient detail, but Cabanas *et al.* (1973) have indicated that strata as young as Langhian are involved in the main folding near Javea, a few kilometres to the east. This compares with late Burdigalian folding in the Sub-Betic, and a conformable sequence as young as late Langhian in the Pre-Betic west of Cartagena, 200 km southwest of the area described here (Hermes 1977a, b). Stratigraphic and palaeontological work now in progress should make it possible to date the event more precisely in the near future.

Post-orogenic younger Miocene rocks are not widespread, but they are well exposed near Finestrat (Fig. 2), where there are plant-bearing gypsiferous grey marls, and some sandstones and conglomerates, possibly of Messinian age. Post Miocene folding affecting these rocks and Pliocene to Quaternary fanglomerates and calcretes is attributed to continued rise of the diapiric intrusion.

STRUCTURE

The unusually wide structural variety in this region is a result of the impact of Alpine tectonics on lithologies of vastly different competence. The massive limestone formations were the most resistant to deformation and now form large folds, commonly with wavelengths of several kilometres. They contrast with the incompetent marls and marl–limestone interbeds which form small-scale folds with wavelengths ranging from 400 m to less than 1 m. The greatest deformation of all is associated with diapiric intrusions; in places the Triassic gypsiferous mudstones are squeezed into most complex forms, and the Mesozoic and Tertiary rocks which overlie them and are adjacent to the diapirs are also strongly deformed. A confusing additional complication is introduced by syn-depositional sedimentary structures, especially in the Eocene mudstone–limestone sequence, which closely resemble some of the more complex tectonic structures.

Sedimentary structures

The sedimentary structures of particular interest to this study are those in the Eocene pebbly mudstone–nodular limestone facies, which could so easily be interpreted as an additional tectonic element. Detailed study of these sequences, however, leaves no doubt that they are of

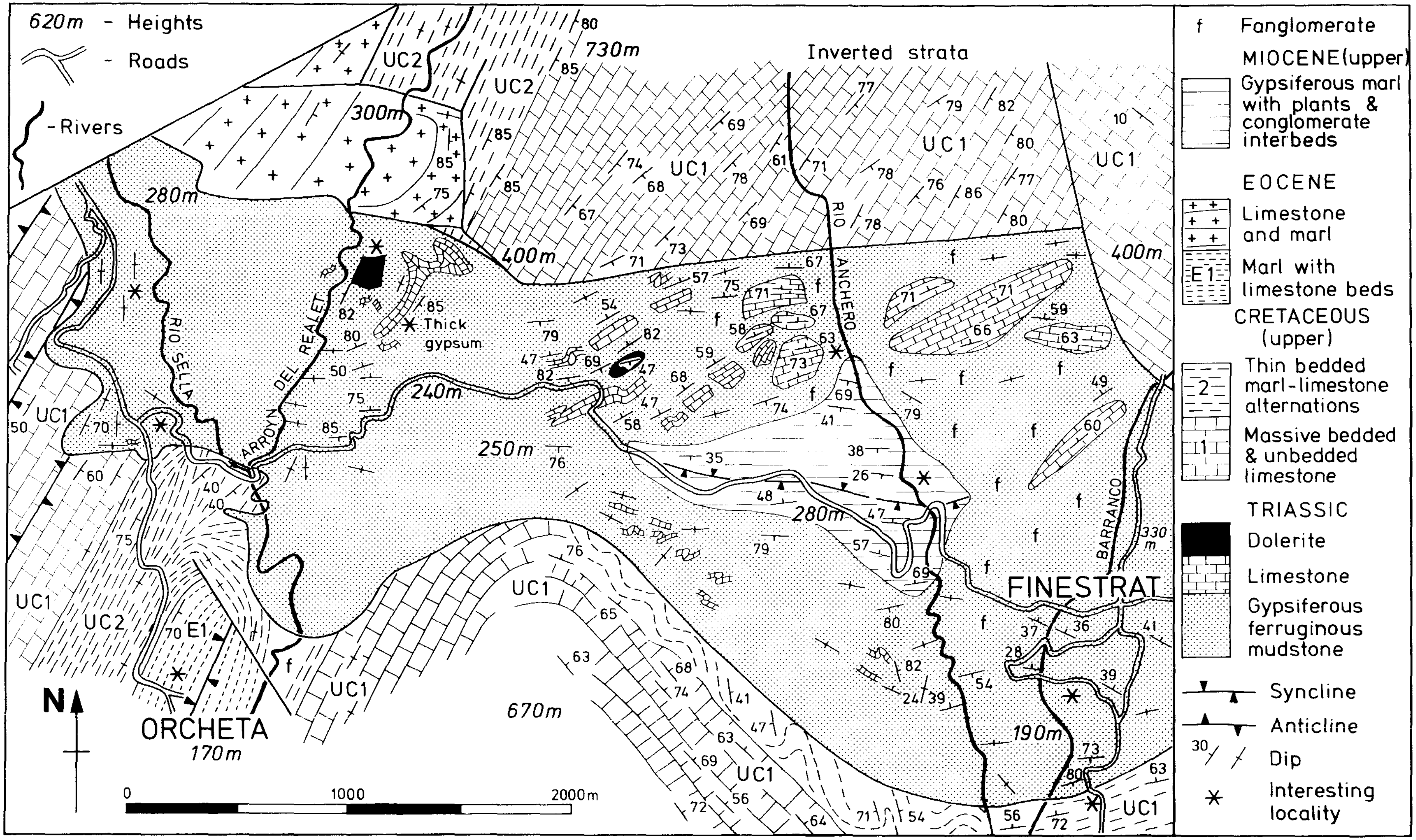

Fig. 2. The Finestrat diapir and surrounding area. Note that the structures within the diapir mostly strike E–W. There is an important syncline in late Miocene strata and a 30–40° tilting of post-Miocene gravels in the same area (not shown on this map) indicating the late stage of the final movements. Note that the strike of the Cretaceous rocks is nearly at right angles to the diapir (location 900 459, Fig. 1).

sedimentary origin, and they are interpreted as submarine debris flows down gentle slopes, perhaps triggered by earthquakes. There is no problem with the pebbly mudstones which were clearly formed by sedimentary processes. They contain mudstone fragments with contorted bedding which resulted from movement whilst still in an unconsolidated state, and fragments of limestone already lithified at the time of inclusion (Moseley 1981, fig. 94). The matrix is Nummulitic mudstone or marl. The mudstones generally grade upwards into breccia which, without the details of field relations could easily be mistaken for tectonic breccias such as those described later. The breccias in turn grade into highly-folded nodular limestones. Figure 3 illustrates the character of the breccia and the folds which are interpreted as syn-depositional slumps of the nodular limestone. Figure 4 shows the steep dips recorded during a routine survey of those areas with slump folds. These dips are anomalous when compared with the comparatively simple tectonic structure in the more competent nodular limestone outcrops and they probably contribute to the differences in fold orientation indicated by bedding and plunge in Fig. 5.1. As stated above, the field relations (non-coaxial folds, folds only in the more competent nodular limestone, brecciation at the base of the limestones and the absence of comparable structures in the incompetent marls) leave no doubt that the folds are of sedimentary origin; nevertheless they are similar in style and magnitude to tectonic structures in some of the regions of more severe deformation.

Structures in Cretaceous to 'Middle' Miocene rocks.

Major structures with a regional northeast Alpine trend. The major NE-trending structures shown in Fig. 1 are the Benidorm syncline, Campana anticline and the Aitana syncline. All were formed in massive Cretaceous or Eocene limestones, but 10 km east of the area depicted in Fig. 1 there is the broad Benisa syncline developed in Miocene rocks, currently under investigation.

The Aitana syncline (Fig. 6) is the best exposed of these structures and provides a good example of a large amplitude fold in competent limestone. It is an impressive structure when seen from the surrounding peaks, and has a rigid framework of massive Eocene limestones which exceed 300 m in thickness. The structure is large but basically simple, being essentially an ENE asymmetric concentric fold with virtually no plunge, although towards the west it is split into two by a complementary anticline (Fig. 1). The northern limb presents a continuous north-facing escarpment along the Aitana skyline, and there are deeply-eroded dip slopes dipping south at 30° into the Barranco del Arc, which descends to Sella, and very nearly coincides with the axis of the main branch of the syncline. The southern limb is much steeper, in fact much of it is vertical, and complicated by important longitudinal faults which slice diagonally across the outcrops to isolate a number of precipitous sharp ridges which rise to about 1000 m. The syncline is, however, terminated in the east by an abrupt increase in plunge from zero to vertical, and the Eocene limestone outcrop passes round the nose of the fold as a series of impressive crags of vertical limestone to form the Sierra Benimantell (Figs. 6 and 7). This sudden change in plunge can probably be attributed to the proximity to the east of an important outcrop of diapiric Trias which extends from Altea to Guadalest (Fig. 1). To the southwest, the southern limb of the syncline is also unusual. One fault-bounded ridge swings through a wide arc as a vertical 700 m-high knife edge, eventually to assume a SSW trend before it is abruptly truncated by the Trias diapir 2 km north-northeast of Orcheta (900 455, Figs. 1 and 2).

Major structures of other trends. In some areas these structures are related to diapiric uplift, as in the case of those associated with the WNW trending Bernia line (Moseley 1973). In other areas the reasons for anomalous trends are not immediately apparent, but underlying diapirism, where the diapirs have not yet been revealed at the surface by erosion, seems to be the most likely cause. In addition to the Bernia structure, there are others with N–S trends which become dominant towards the west of the area shown in Fig. 1. The most important of these are the Amadorio dome and anticline and the Orcheta syncline, all in upper Cretaceous rocks (Figs. 2 and 8).

The Amadorio dome is a faulted structure picked out by a prominent limestone member, with underlying marl eroded to form a saucer-shaped valley. The dips radiate outwards in all directions, possibly reflecting interference between NE–SW and N–S structures. The dome is continued northwards by the Amadorio anticline in massive Cretaceous limestones continuously exposed in the 400 m deep antecedent Amadorio gorge. The Orcheta syncline, parallel to the anticline, brings in the highest Cretaceous (marls and limestones) and lower Eocene strata (marls) which are incompetent and deformed by minor structures (Fig. 9). The eastern limb of the syncline is vertical, and to the north both syncline and anticline are abruptly truncated by the Finestrat diapir (Fig. 3).

Minor structures. Minor structures are almost always restricted to the incompetent lithologies of marl and marl–limestone interbeds, with the more complex folds also associated with regions of diapric uplift. Lithological control is extremely important and disharmony between the thicker limestone members and marl-dominated sequences is commonplace.

The Villajoysa fold belt typifies minor folds formed outside the influence of the diapirs. These folds are in Eocene marls, pebbly mudstones and nodular limestones (Figs. 4 and 8), and have wavelengths ranging from 400 m to less than 10 m. The complexity varies from open folds with limb dips of about 30°, mostly developed in the thicker and more competent limestone members, to tight to isoclinal folds in the marl–dominated sequences. The latter frequently display faulted and brecciated axial planes. The Eocene rocks of the Villajoysa fold belt also exhibit the sedimentary structures described above, and care is required to differentiate between these and tectonic structures. The former are rarely coaxial with each other or with the tectonic structures, and they only occur in the

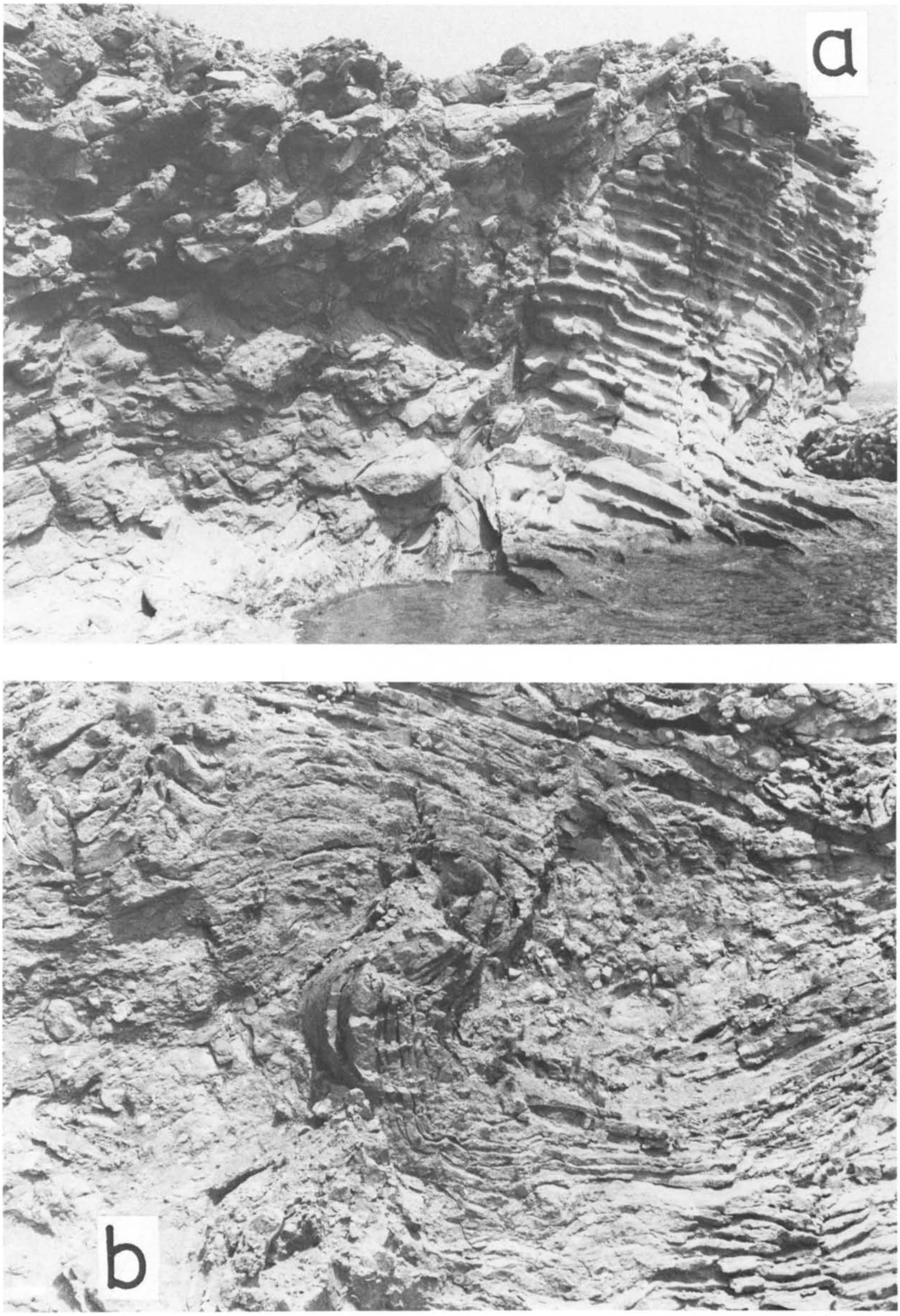

Fig. 3. (a) 15 m high cliff section near Charco (see Fig. 4) exposing Eocene limestone breccia. The bedded nodular limestone on the right is one boulder in the breccia, which is interpreted as a submarine debris flow. It grades upwards from pebbly mudstones. (b) Bedded nodular limestone above the breccia shown in (a), which has been involved in slump folding. The amplitudes of the folds are about 3 m and the hinge zones are brecciated.

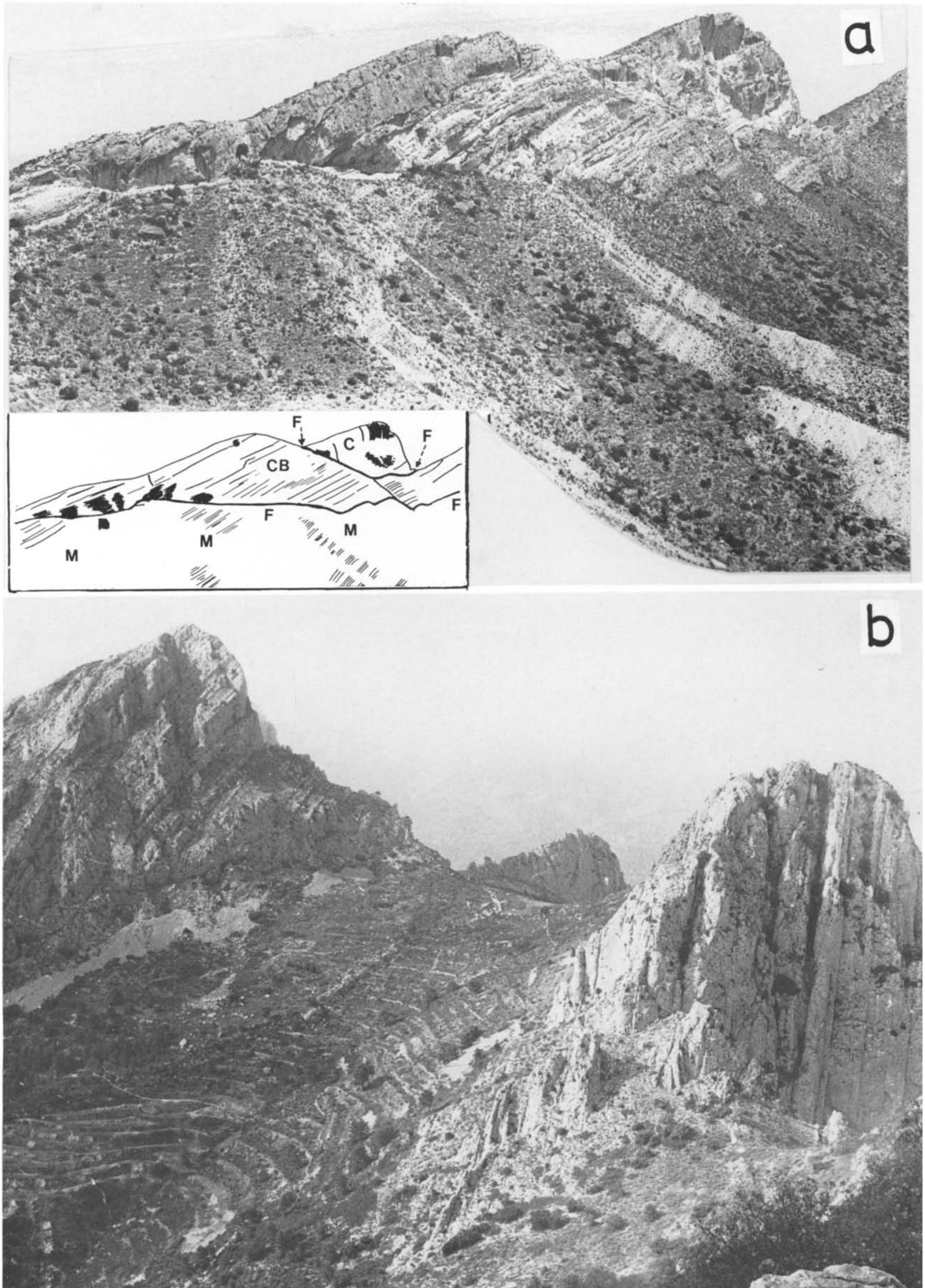

Fig. 7. (a) View of the faulted northwest margin of the Amadorio dome from the south-southeast showing Upper Cretaceous marls and limestones. M, marl; CB, bedded limestone; C, massive limestone; F, faults (see Fig. 8). (b) The Sierra de Benimantell forming the eastern termination of the Aitana Syncline. The peaks (Eocene limestone) are in the hinge zone of the syncline which here steepens to a vertical plunge. Spot height 1036 m (see Fig. 6) is on the right of the photograph which was taken looking north from spot height 1050 m.

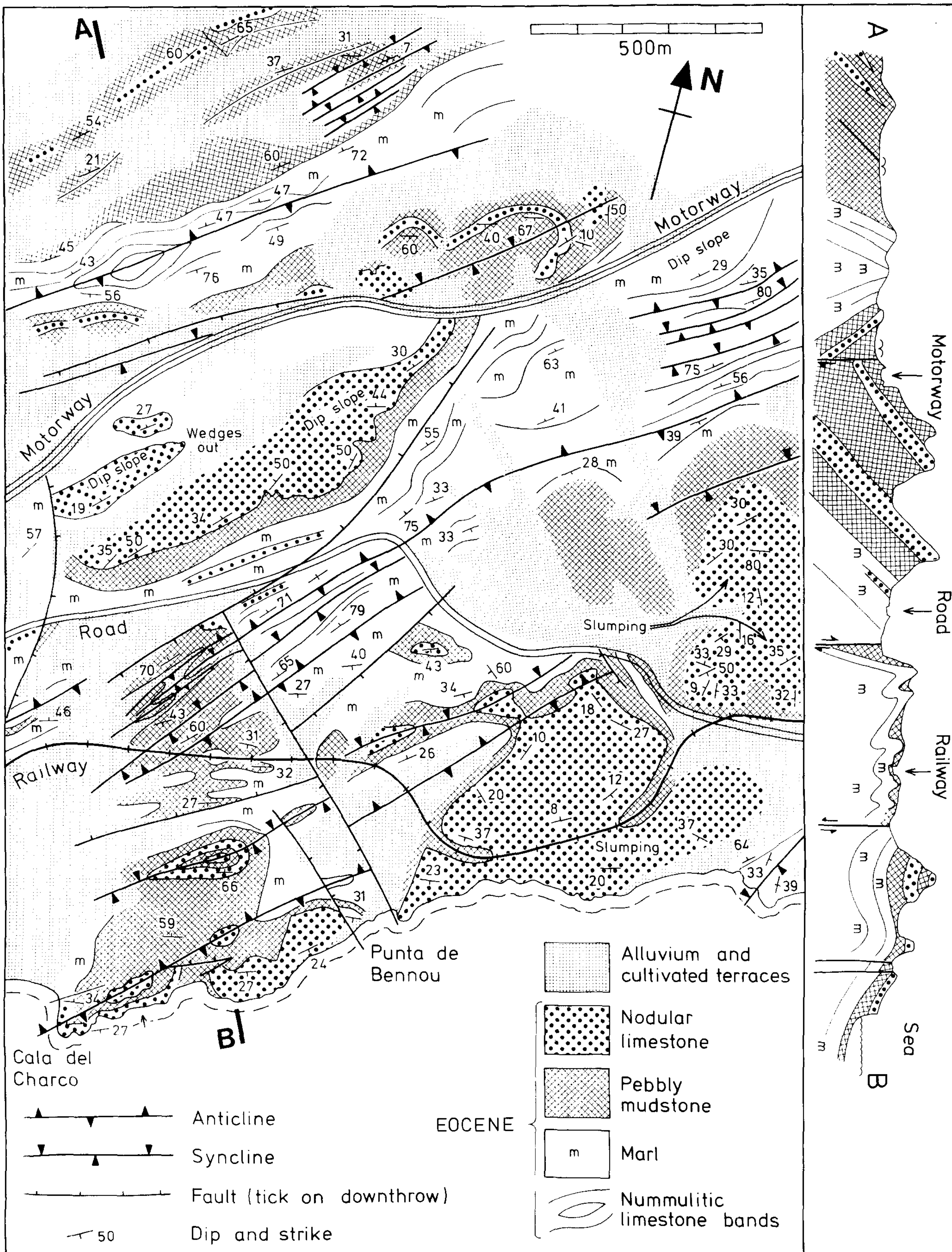

Fig. 4. Detailed map of the southern part of the Villajoyosa fold belt. The folds in the competent nodular limestone are open, with the steep variable dips a result of slumping and not tectonics (Fig. 3). The folds in the less competent marls are tighter (Location 897 436, Fig. 1).

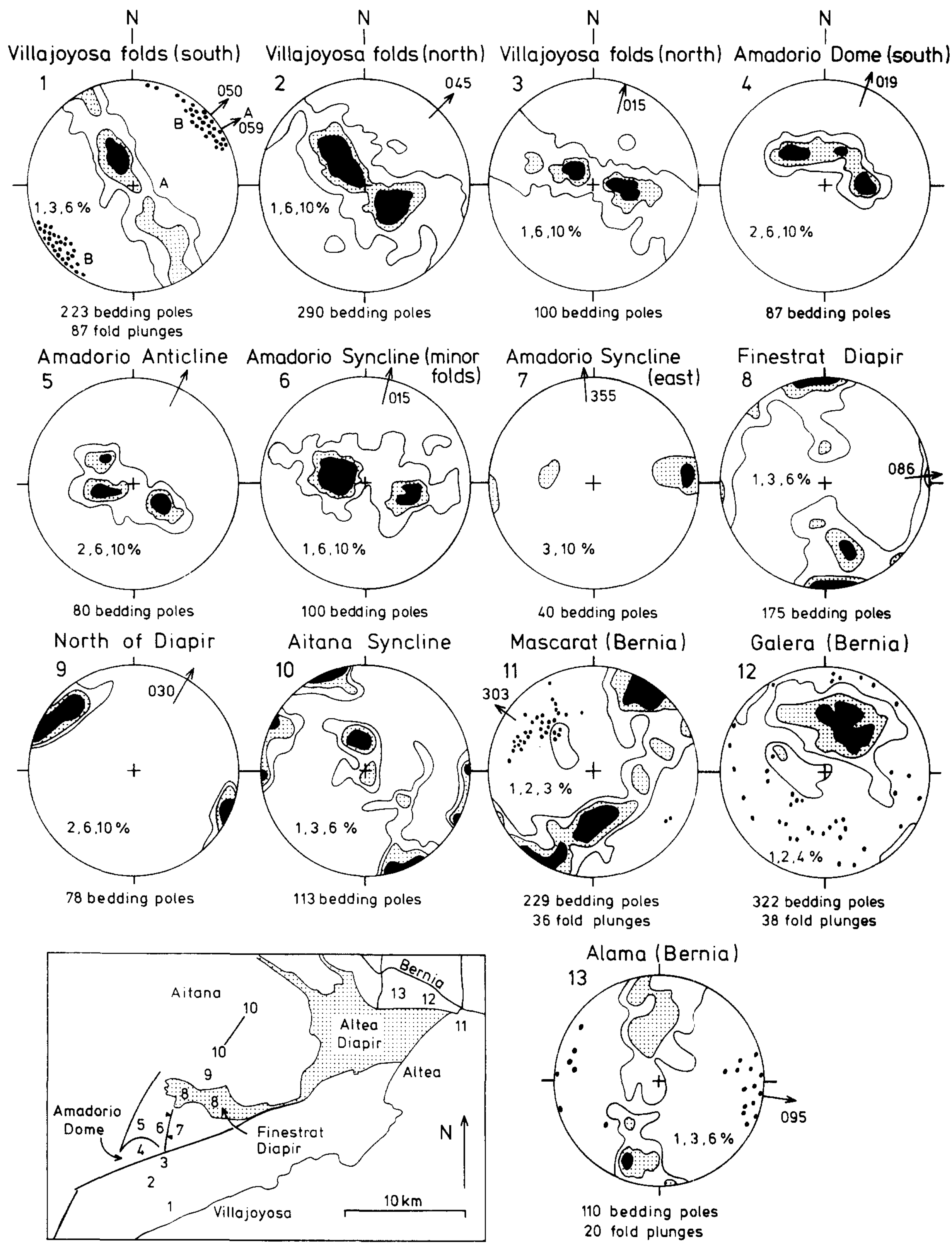

Fig. 5. Contoured equal-area stereographic projections for the areas shown on the inset map. Diagrams 1–7 show the change from a NE–SW to a N–S axial trend. Notice that in (1) the bedding poles (A) and fold plunges (B) suggest different structural trends, but this is probably because many bedding dips were taken in pebbly mudstone and nodular limestone formations subject to strong slumping. (8) illustrates the cross-cutting E–W trend of the largerly Triassic Finestrat diapir. The scattered maxima of (10) result from the swing in strike seen on Fig. 1. (11), (12) and (13) illustrate the WNW–ESE axial trend of the Bernia uplift. The scattered plunges of (12) are explained by diapiric rise, resulting in gravity sliding (see text and Moseley 1973).

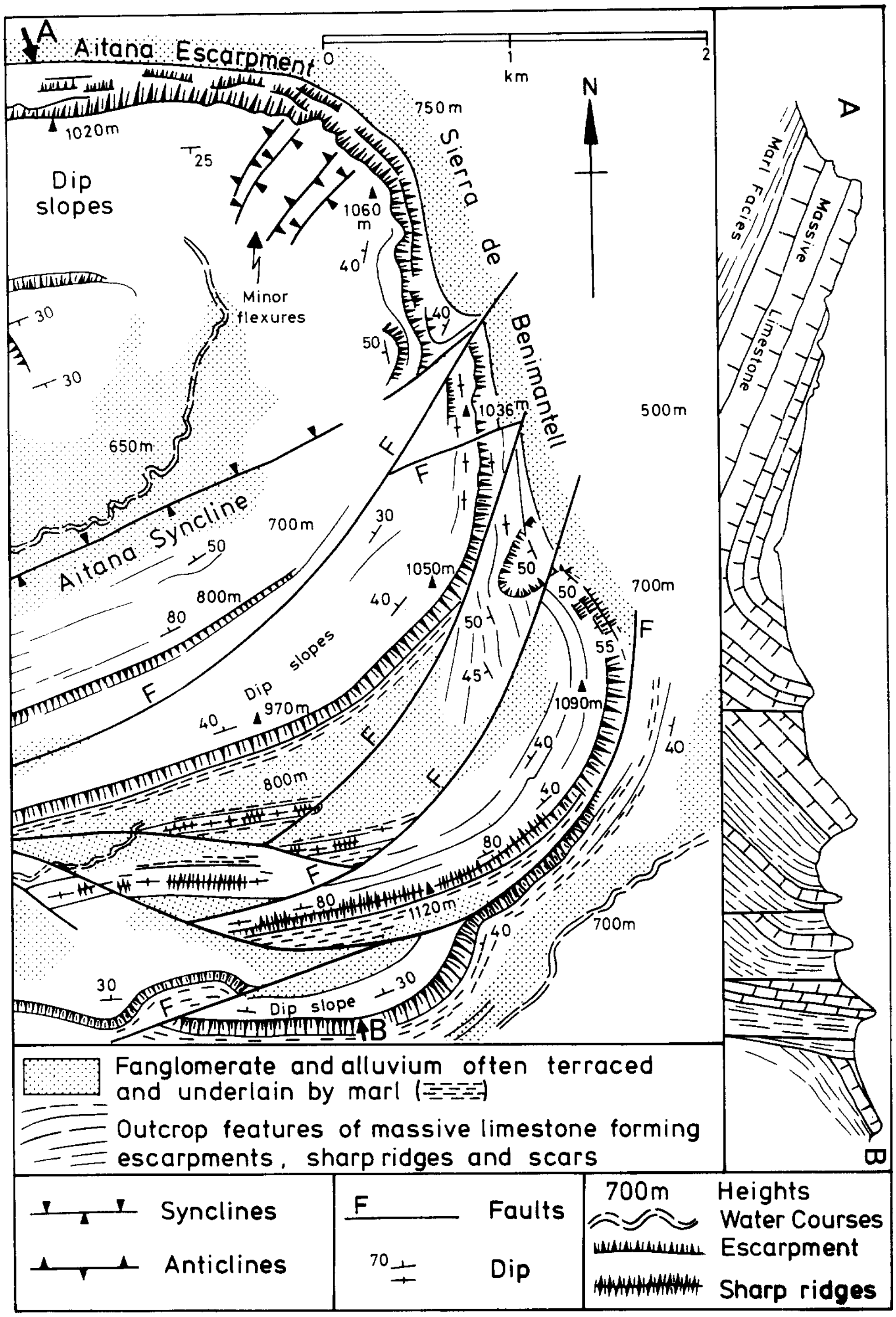

Fig. 6. Structural map and cross-section of the eastern termination of the Aitana syncline showing the rapid increase in plunge from sub-horizontal to near vertical. The rocks involved are of Eocene age (located at grid reference 905 456, Fig. 1).

more competent nodular limestone beds. Another feature of this region is that the fold trend swings from NE in the south to NNE in the north (Fig. 8). The reasons for the latter trend are almost certainly the same as the reasons for the N–S trend of the Amadorio dome and anticline and the Orcheta syncline, and will be referred to in the discussion below.

The folds in Cretaceous marl–limestone facies differ in style from those just described, mostly because of the greater degree of disharmony they display. There is a rapid passage from massive upper Cretaceous limestone into this facies, and there are also some thick limestone members within marl–limestone sequences which have resulted in ideal conditions for the formation of disharmonic and décollement structures. One region where they occur is on the eastern limb of the Amadorio anticline (Fig. 9a) where uniform 30° dip-slopes in massive lime stones are followed by a great variety of minor folds in the marl–limestone facies. These folds become increasingly complex near to the Finestrat diapir where recumbent

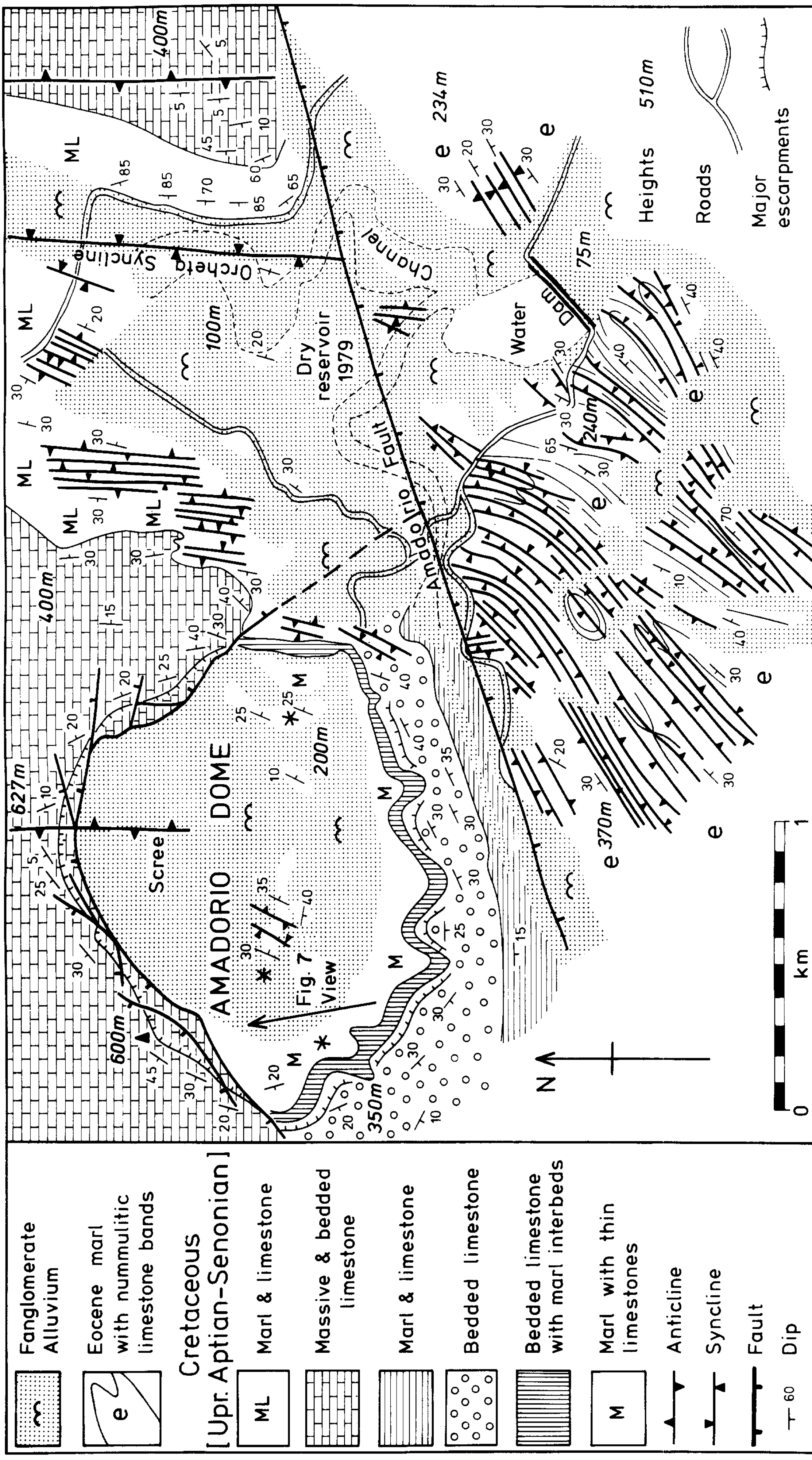

Fig. 8. Map of the northern part of the Villajoyosa fold belt and the Amadorio dome showing contrasting structural styles and orientations. Good fossil localities are shown by asterisks (location 895 445, Fig. 1).

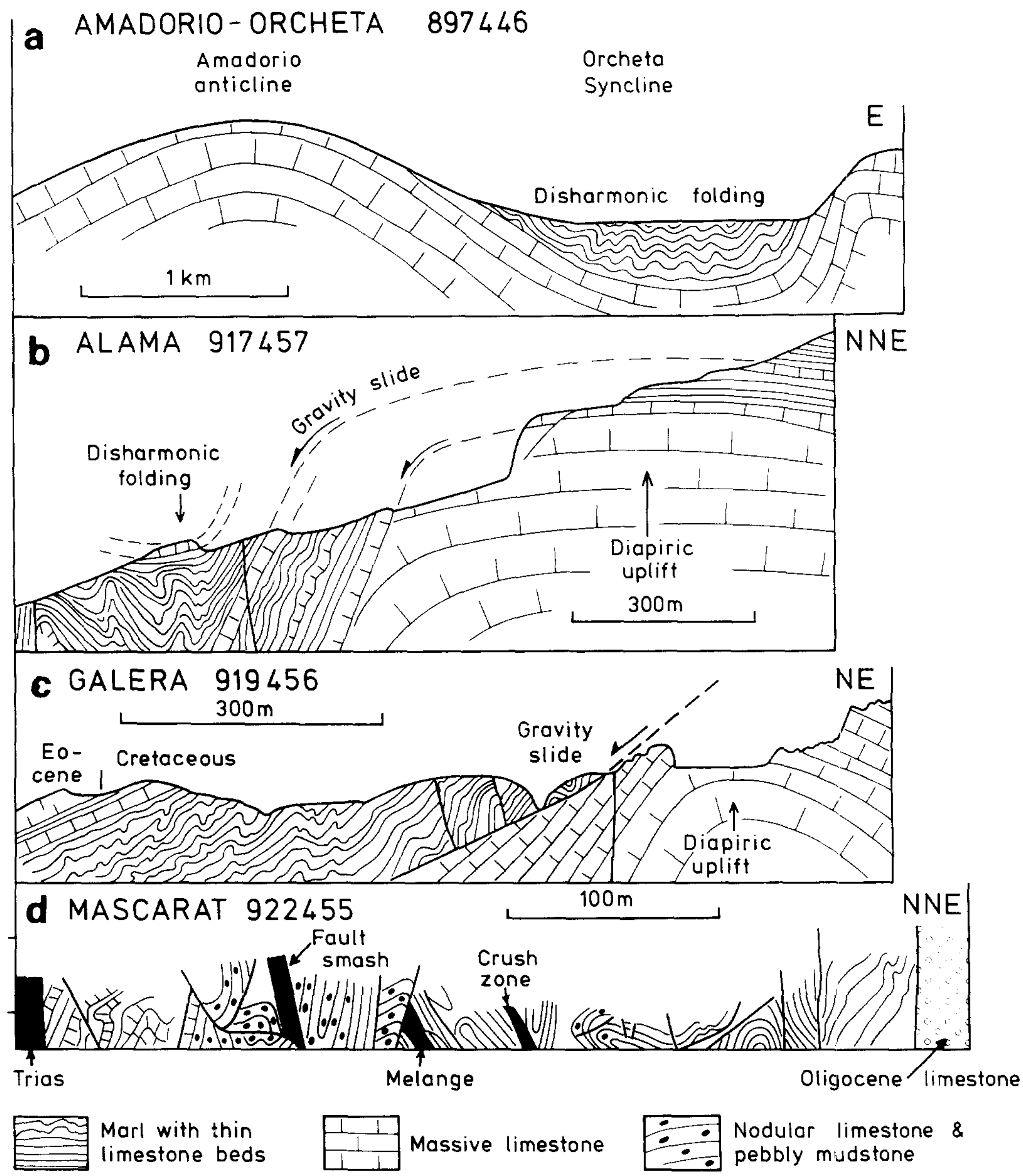

Fig. 9. (a) Cross-section showing an Upper Cretaceous sequence of competent and incompetent strata, the latter folded disharmonically (see Fig. 8). (b) & (c). Diapiric uplift forming the Galera anticline in Upper Cretaceous limestones, and inducing gravity slides on the southern limb. Disharmonic folding occurs in the incompetent marl–limestone interbeds. (d) Complex tectonic zone in Eocene rocks of the Mascarat shore.

isoclines and refolded folds are exposed. Comparable complex folds occur along the Bernia line, again associated with diapiric uplift (Figs. 9b & c and Moseley 1973, figs. 8, 9 and 10). In all cases there is a direct relation between lithology and fold complexity with the thicker limestones deformed into simple structures of concentric type (e.g. the Galera anticline, Moseley 1973, fig. 7 and plate 3), and the thin-bedded limestone and marl units deformed into smaller-scale, more complex folds. In some parts of the Bernia region (Alama) all these structures are coaxial but in others (Galera) there is clear indication of polyphase deformation (Moseley 1973, fig. 7). This aspect is considered further in the discussion below.

The most severe deformation to be seen in this region is in the Eocene marl–limestone facies of the 'Bernia line', continuously exposed along the Mascarat shore (Fig. 9d and Moseley 1973, figs. 4, 5 and 6). There is a wide range of competence in these rocks which resemble those of the Villajoysa fold belt. They include nodular limestones, pebbly mudstones, marl–limestone interbeds and marls but deformation has proceeded further than in the Villajoyosa fold belt. Complex tectonic structures are superimposed on sedimentary structures of the type described above (Fig. 3), and it is with some difficulty that one is distinguishable from the other. Fortunately however, the tectonic folds are strongly coaxial with WNW trends parallel to the Bernia line (Fig. 5.11) and there seems little doubt that most of the tight folds and shears of Fig. 9d are of tectonic origin. The more competent nodular limestones form upright, overturned or recumbent folds of moderate tightness. More severe structures occur in the marl–limestone interbeds in which fold hinges in the limestone beds become isolated, or crushed (Moseley 1973, figs. 5 and 6, 1981, fig. 95). In extreme cases the limestone beds and the marls form a tectonic mélange which resembles the sedimentary boulder beds previously described. Orientations of fold plunges and linear elements in the mélange are always ENE; parallel to the Bernia line (Fig. 5.11). The latter is believed to be related to the intrusion of the Altea diapir (see below).

Diapiric structures

Interpretation of these structures as diapiric is based on field relations. The boundaries between the Triassic rocks of the diapirs and the younger rocks are either tectonic (clearly defined high-angle faults), or piercement contacts (reminiscent of contacts between plutonic intrusions and country rocks). Supporting evidence comes from internal structures within the Triassic outcrops, particularly the mélange of large blocks of limestone and basic igneous rock set in a matrix of gypsiferous ferruginous mudstone, itself generally brecciated and highly deformed.

Diapir-country rock contacts. As indicated above there are two different forms of contact. First there are piercement contacts. The field relations show that in some areas the junction is highly irregular, for example north of Orcheta (Fig. 2). In this region exposures of Triassic, Cretaceous and Eocene rocks are excellent, and there is a sinuous but high-angle contact between the Triassic and younger rocks. This contact is discordant to structures in all the rocks, and we can see no alternative to interpreting it as an intrusive contact comparable with those between plutonic intrusions and country rocks. Secondly there are many areas where the boundaries between Triassic and younger rocks are major faults. In these situations there are usually one or two major fault planes with numerous subparallel shears. The faults are nearly always high-angle, and have N–S or WNW trends, although there are some exceptions. The fault zones are commonly more than 100 m wide. Good examples are to be seen in the roadside section south of Finestrat (Fig. 2), where there are numerous shear planes in Cretaceous marly limestone, and also at several localities between Altea and Olta. In this latter region the faults are vertical, and in several places exhibit horizontal slickenside lineations. Between Mascarat and Altea la Vieja the faults trend WNW and mainly separate Triassic sediments from massive Eocene limestones, but in the east there is a 100 m-wide wedge of Miocene marl between the Trias and the Eocene, indicating the late age of the final movements (Moseley 1973, fig. 12). Further east near Olta (Figs. 1 and 10), there is a N–S faulted strip of Triassic rocks, bounded on both sides by Miocene marls (Moseley 1973, plate 1). The trends of these faults are mostly WNW (ranging to E–W) and N–S, and are considered to be fundamental wrench faults related to NNW–SSE Alpine compression, and to have controlled the orientations of the diapiric intrusions (see discussion below).

Internal structures within the diapirs. The most impressive features within the diapirs are the large-scale mélange structures, in which disoriented blocks up to 1 km long are set in a matrix of gypsiferous, ferruginous mudstone. The Finestrat diapir (Fig. 2) shows this characteristic well, with blocks of massive-bedded limestones isolated from each other, and aligned approximately parallel to the trend of the diapir. There is also a large block of dolerite which shows a normal igneous-sedimentary contact on one side, but is sharply truncated elsewhere; it appears to have been a sill, another section in a sill occurs near Olta (Moseley 1973, p. 11). This latter sill is 50 m thick and ranges from gabbro in the centre to basalt at the margin. The Altea diapir also contains blocks of limestone, some of Triassic and some of Cretaceous age, and blocks of basic igneous rock, such as the basalt and agglomerate of Cap Negret between Altea and Mascarat.

The greater part of the Trias is made up of ferruginous-gypsiferous mudstone. This material is exceedingly incompetent, and has suffered extremely complex and chaotic deformation. Much of the rock has become gypsiferous breccia, but this passes into bedded, strongly folded rock especially where there are thin siltstone and dolomite beds in the mudstone (Moseley 1973, fig. 13). Systematic survey, however, reveals some order in the chaos, and in the case of the Finestrat diapir the dips are high angle and the overall strike is subparallel to the long axis of the diapir (Fig. 5.8).

Faults

Faults have already been referred to where they form integral parts of other structures. To appreciate their overall significance however it is necessary to review the total fault pattern.

High-angle faults. Most of these faults have WNW–ESE or N–S trends and are believed to have originated as wrench faults. Certainly their trends fit well with the theoretical trends for dextral and sinistral wrench faults related to a regional ENE–WSW strike. In detail however, there are complications, and analysis of all the structures makes it seem likely that horizontal, oblique and vertical movements have occurred at different times. Exposed fault planes display slickenside lineations parallel to all these senses, but because of the complex tectonic history, and the effects of diapirism, there are few localities where strike–slip movement can be conclusively demonstrated by displacements of marker horizons or other structures.

The Mascarat–Bernia fault complex is one in which it is undesirable to describe high-angle faults separately from thrusts since some of them are composite (Fig. 10). Perhaps the most impressive of the faults are those forming the 100 m-vertical cliffs which intersect at 110° in the Cala de Mascarat. One of the fault planes trends WNW and extends towards Bernia. It is displaced 110 m dextrally by the other, a N–S fault which, further north, brings in the wedge of Trias west of Olta. On theoretical

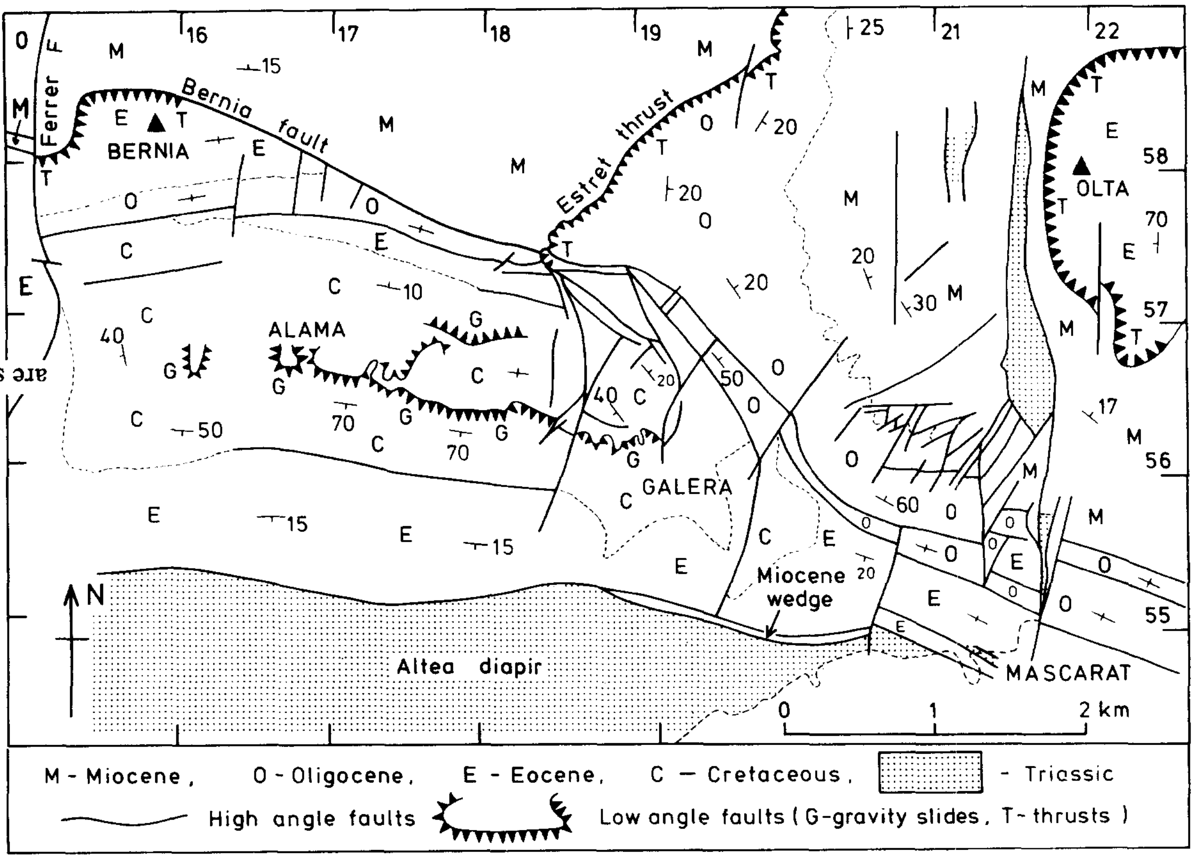

Fig. 10. Map of the fault pattern in the Bernia region (917 457, Fig. 1).

grounds a dextral displacement here is difficult to explain. The WNW fault continues for about 8 km and then bends into the Estret thrust, which brings Oligocene limestone on top of Miocene marl (Fig. 10). The WNW line is continued by the 200–300 m-high northern precipice of the Sierra Bernia, where Oligocene and Eocene limestone is brought against soft Miocene rocks. It then bends into the Bernia thrust, which in turn is crossed by the important N–S Ferrer fault. The latter extends to the north of the area shown in Fig. 1 to run along the axis of the Sierra de Ferrer, a N–S belt of vertical Cretaceous and Oligocene limestones (Rios *et al.* 1961, Moseley 1981, fig. 99). Of the other faults in the Mascarat–Altea region, the most important is that forming the northern margin of the Altea diapir. In the east there are two parallel faults with a 100 m wide strip of Miocene marl between Triassic mudstone and Eocene limestone (Fig. 10). Further west and northwest its outcrop is lost in badly-exposed ground, but it is probably continued by the Trias wedge extending to Guadalest (Fig. 1).

NE to ENE high-angle faults, sub parallel to the regional strike form another important set. The only important one in this region however, is the Amadorio fault (Figs. 1 and 8) which forms part of the southern margin of the Finestrat and Altea diapirs; other faults in this category isolate knife-edge ridges on the southern limb of the Aitana syncline (Fig. 6).

Low-angle faults. These structures include the composite thrust–wrench faults already referred to, the gravity slides of the Bernia region (Moseley 1973) and near-horizontal thrust sheets apparently unrelated to other faults.

The composite thrust–wrench faults appear to have formed in relation to a NW–SE compressive stress, with one set curving into the other in a manner comparable to many other wrench fault sets described in the literature (e.g. De Sitter 1964, Soper & Moseley 1978). Other structures in the Bernia region were interpreted as gravity slides by Moseley (1973), because it was believed that the Bernia uplift was a response to the diapiric rise of the underlying Trias rather than lateral compression. Major structures such as the Galera anticline were interpreted in this way and it was concluded that the low-angle faults on the southern flank of the Bernia uplift could be explained most easily by a rising anticline which induced gravity sliding towards the south. Much of the complex folding in Cretaceous rocks already referred to is attributed to this cause (Figs. 9 and 10). The most impressive of the low-angle faults are the thrust sheets of Eocene limestone resting on Miocene rocks. Most of these structures are to the north of the area shown in Fig. 1 (Rios *et al.* 1961), and were not investigated in detail during this survey. The Olta thrust on the eastern margin of the area shown in Fig. 1 extends further east to the Ifach peninsula.

DISCUSSION

The Mesozoic and Tertiary rocks of the Costa Blanca exhibit a great variety of fold trends. In most situations one fold direction swings gradually into another, but there are areas where trend changes are abrupt and the possibility of polyphase tectonics has to be considered. The predominant fold trend throughout the Costa Blanca is WSW–ENE but two other trends; WNW–ESE and N–S are also common. The last two trends are also the directions followed by most of the elongate Trias diapirs and they are also the trends of the majority of the high-angle faults. The latter may have originated as complementary sets of wrench faults in a Pre-Triassic 'basement' which activated disharmonic and diapiric movements in the gypsiferous Trias, with associated folding in the overlying cover rocks. Thus most of the folds with trends diverging from the regional WSW–ENE direction probably formed during the late Miocene rather than at separate earlier periods. Fig. 1 shows a number of areas where such structures are well displayed. The WNW-trending Bernia–Mascaret uplift is particularly important, as are the N–S folds around Orcheta and the WNW-cross-cutting Finestrat diapir. Other structures of similar type in areas adjacent to that covered by this survey are: (1) the WNW-trending Mongo syncline 20 km to the northeast of Bernia (Moseley 1968), (2) the Sierra del Ferrer, which extends for 8 km immediately north of the Sierra Bernia, and consists of vertical N–S striking Cretaceous, Oligocene and Miocene rocks together with vertical strike faults (Rios *et al.* 1961, Moseley 1981, fig. 99) (the Ferrer fault shown in Fig. 10 is the southern extension of this zone) and (3) N–S folds in Cretaceous limestones 10 km west of the area shown in Fig. 1. The latter N–S folds are bounded by a large Triassic outcrop (Verdu & Gonzales 1978).

The Bernia line is a complex WNW fold-fault structure within which many of the major and minor fold hinges trend WNW and plunge gently (Fig. 5.11), but just to the north and east of the line there are abrupt changes in trend both to NE–SW and N–S. The former trend is exposed west and north of Olta (920 458) where the strike of the Oligocene limestones swings through 90° within a distance of 7 km (the NE–SW strike occurs just to the northeast of the area shown in Fig. 1 and is the regional strike of the Pre-Betic zone). Immediately west of Olta (921 457) there is a fault-bounded N–S strip of Trias, the fault extending to Mascarat, where it is clearly exposed as a wrench fault (Moseley 1973, fig. 10). Further west in the Sierra de Benimantell (Figs. 1, 6 and 7), the N–S belt of vertical Eocene rocks which terminate the Aitana syncline may have been pushed up by the rising western margin of the Altea diapir. Further to the southwest, between Campana and Finestrat (900 450), the steeply-dipping southern limb of the Aitana syncline swings steadily into a N–S strike. This structure is truncated by the Finestrat diapir, and although the same trend continues south of the diapir the structures are quite different (Figs. 1 & 2). The conclusion is that the structures are displaced along a WNW–ESE strike-slip fault, and that the Finestrat diapir has been emplaced along the fault zone. A little further south, the Amadorio dome (Figs. 1 and 2) could be explained by intersecting NE–SW and N–S trends, and it is possible that underlying diapiric Trias could be responsible for the latter trend.

These observations show that the diversity of fold trend in the Pre-Betic zone can be explained by postulating a NW–SE maximum compression giving rise to NE–SW or ENE–WSW folds and complementary WNW–ESE and N–S wrench faults, nearly all of which formed during the same tectonic episode. There are also important post-Miocene structures which are associated with the diapiric uplift. The best example is seen within the Finestrat diapir (Fig. 2), where late Miocene plant-bearing marls have been folded into an E–W syncline with limb dips reaching 60°, and post-Miocene bedded sands and gravels with no component of depositional dip have been tilted to 40°. Similar steep dips in Quaternary gravels are seen on the flanks of the Sierra Bernia. This suggests that, whilst the main phase of tectonic activity was late Miocene, movements have continued to the present day, especially those related to the continued rise of the diapirs.

Finally it is desirable to comment on emplacement mechanisms for the diapirs. There is an extensive literature dealing with diapir formation, and it is clear that explanations differ widely from region to region according to the nature of the diapiric materials, the overlying rock and the tectonic forces involved. The reader is directed to Braunstein & O'Brien (1968) for a review and descriptions of a variety of diapiric environments, and these and other articles make it apparent that each case has to be examined in relation to local field conditions rather than to a predetermined set of rules. However, one has to decide on an acceptable definition for a diapir, because at the one extreme it is restricted to structures formed by purely halokinetic processes (Burns *et al.* 1977); that is buoyant low density rocks, rising through, and folding the overlying strata, with no regional tectonic forces involved; whereas at the other extreme décollement structures formed during orogeny have been described as diapiric. Both of these extreme views seem to be unrealistic, the first because there are few diapirs not triggered by some form of tectonic activity, and the second because disharmony is such a common tectonic feature, for example in the Upper Cretaceous rocks described above. Few would disagree with the suggestion that true diapirs should consist of low density rocks with an intrusive relationship to surrounding rocks, but we believe that to rule out involvement in regional tectonics makes the definition too restrictive.

Classical diapirs of the halokinetic type are to be found in some regions, for example north Germany (Sanneman 1968). Diapirs in northern Spain which have similar stratigraphical relationships to those described here have been attributed to thickness variations in the Cretaceous overburden (Brinkmann & Lögters 1968). This is believed to have controlled their distribution and exerted the pressure necessary to initiate movement of the saliferous beds. A different type of diapirism is that associated with transform faults where volume changes from peridotite to low-density serpentinite have been suggested, leading to intrusion of serpentinite–gabbro–basalt–chert–limestone megabreccias (Bonatti 1978, Moseley & Abbots 1979). On the other hand Burns *et al.* (1977) have recently disagreed with former diapiric interpretations for breccia-cored domes in the Flinders Ranges of South Australia, claiming that there is no density contrast or intrusive relationship between the megabreccias (believed to be diapiric by Dalgarno & Johnson 1968) and the surrounding rocks. Burns *et al.* (1977) suggest that these structures were formed at a décollement surface and resulted from folding. The examples just referred to by no means cover the whole field of diapirism, and to emphasise what has already been stated, it is the field relations of each area which are important, and no investigation should rely too heavily on conclusions from other regions.

The diapirs of the Costa Blanca are piercement structures as illustrated in Fig. 2, but they are also tectonically controlled. The region possesses most of the characteristics necessary for the formation of diapirs. The gypsiferous Triassic mudstones are plastic and incompetent, and would be expected to flow following the application of a tectonic force. They are also of low density with a weight differential when compared with the predominatly limestone overburden, and the resulting buoyancy would favour diapiric rise. Although much of the overburden consists of massive limestone of low ductility, it has been folded and fractured and these structures, especially the faults, are likely to have given the initial instability which triggered the diapiric movements, and provided the pathways along which flowage took place. There is ample evidence, already presented, that the movement of the diapirs has been controlled in this way.

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