Nature of thrust zones in deep water sand-shale sequences: outcrop examples from the Champsaur Sandstones of SE France.

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Compressional structures are commonly found in deep water turbidite successions that contain important hydrocarbon reservoirs. Here we present a field study of deformation in the Champsaur Sandstones of the French Alps to illustrate the geometry of natural thrust zones in weakly lithified turbidites. The evolution of these zones results from a combination of faulting and folding. A general model for thrust zone evolution is developed where, even though there is a tendency to produce localised, through-going slip surfaces, folding can occur at various stages. Segments of dismembered fold limbs carried up thrust ramps, potentially increase the fault zone permeability and fluid communication through the stratigraphic pile. The propensity for folding is increased by the presence of thin-bedded sand-shale successions. While imaging these detailed architectures may lie beyond the reach of seismic methods in the sub-surface, it may be possible to predict the nature of thrust zones from knowledge of the sand-shale stratigraphy, and *vice versa*.

Key words: Turbidites, hydrocarbons, thrust tectonics.

Introduction

Sandy turbidites found in deep water fold and thrust belts and many convergent plate boundaries are increasingly exploited as hydrocarbon reservoirs. The performance of such reservoirs is dependent, in no small measure, on the internal 3D connectedness of the composite sandstone body. In deformed rocks, this attribute is a function not only of the primary stratigraphic architecture but also the deformation style. The purposes of this contribution are (1) to provide descriptions of compressional tectonics recorded in outcrops of turbidites from the Alpine foreland basin of SE France and (2) to assess the possible control upon structural style imposed by the architecture of the host sediments. While seismic data may resolve the presence of faults within a section, much of the detail of fault architecture may lie beyond the reach of seismic methods. It is hoped, therefore, that field analogue studies can be used to inform the prediction of fault zone architecture from sub-surface data-sets.

Historically, the analysis of contractional structural styles developed in well-layered sedimentary sequences has been approached from different perspectives. In the thrust tectonics literature, the presence of mechanically weak horizons has long been thought to promote the development of thrust flats (e.g. Elliott & Johnson, 1980). Consequently a succession of turbidites with alternating sand-rich and sand-poor intervals might be expected to have developed rather complex thrust trajectories made up of ramps through the sand-rich sediment packages and flats localised in the shaly parts. Displacement on these profiles might be expected to have given rise to complex fault-bend fold styles. In addition, the presence of many potential thrust flats increases the likelihood of single layer duplex structures having effectively thickened-up individual sand-rich bed packages.

In contrast to the thrust perspective, the presence of a layered rheology promotes buckle folding during layer-parallel contraction (see Price & Cosgrove, 1990, for extensive review). The geometry of folds is strongly influenced by the stacking characteristics of competent and incompetent layers. Rapid alternations in this property, generating a strong mechanical anisotropy, tends to create kink or chevron-type folds when compressed (Biot, 1961) – indeed the classic geometrical study of these structures (Ramsay, 1974) describes folds of this type from turbidite successions. If the competent units are more widely separated from each other then a broader range of fold styles with greater hinge curvature can be created. In general, increasing the thickness of incompetent units will allow the competent beds to fold with less interference and therefore fold wavelengths will be much less than for the thin-layered, anisotropic case. This permits greater disharmony in folding. In these situations the key controls on fold wavelength are the viscosity contrast between the layers and the matrix together with the thickness of the competent layer.

Many deformed terrains contain both buckle folds and thrusts. Commonly the sites of thrusting can be shown to have been pre-conditioned by earlier buckling so that thrusts break through fold structures (e.g. Williams & Chapman, 1983). Thus in these cases the thrust ramp initiates within a material that has previously experienced layer-parallel shortening. The final deformational geometry, after buckling followed by displacement up the thrust ramp, is a composite structure with features common both to fault bend folds and the precursor buckle (e.g. Cooper & Trayner, 1986). This structural style is relatively common in alternating carbonate-shale sequences (Butler, 1992a, b) where the key issue is how the propensity for strain localisation varies through a multilayer. Efficient localisation promotes simple thrust geometries while inefficient localisation is reflected in greater buckling. These styles can interact through the sedimentary pile. When the whole thrust zone eventually links through, the forelimb areas of the precursor buckle folds can be carried as horses within which bedding is steeply dipping. In short, precursor buckling tends to generate thrust zones with more complex internal architectures.

Setting

The Champsaur Sandstones are a sequence of late Eocene- early Oligocene turbidites that form part of the sedimentary fill of an early Alpine foredeep basin (e.g. Sinclair, 1997; Figure 1). In common with much of the foredeep fill, the sandstones form the youngest part of a deepening-upwards sedimentary triptych that started with a transgressive shallow-water carbonate (the Nummulitic limestone) followed by deeper water marls before the onset of turbidite sand deposition. For the most part this c. 650m thick turbidite succession is volcaniclastic and shows broadly east to north-easterly dispersal. In its type area the Tertiary sequence oversteps from south to north a substrate of Jurassic marls and limestone down to pre-Triassic crystalline basement (the Pelvoux massif). This substrate confined the turbidites during deposition by forming a broadly south to south-easterly-dipping palaeoslope (Waibel, 1990). The nature of the pre-Tertiary deformation is controversial (see Ford, 1996, for discussion) but need not concern us here.

The Champsaur turbidite succession shows distinct variations in the amount of sandstone up section. In general the sand-rich beds form packages up to about 80m thick with km-scale lateral continuity. This large-scale layering is disrupted at several levels in the succession by erosional channels (e.g. Waibel, 1990; McCaffrey *et al.,* 2002) with significant amounts of sand body amalgamation. The channels do not form continuous layers but can be traced as sinuous ribbons (McCaffrey *et al.*, 2002). Interspersed between these sand-rich packages are turbidites with significantly lower sand content. In general the bedding in these units is laterally extensive. Overall then, the Champsaur Sandstones show significant layering, both on the scale of individual beds and on the scale of bed-sets.

Thrusts and folds in the Champsaur region were first described in detail by Lory (1901-2) and then by Gignoux & Moret (1937). Ford (1996) presents a model for the kinematic evolution of the area upon which subsequent work by Bürgisser (1998) has built. The area is covered by a 1:50,000 geological map (Debelmas *et al.*, 1980). Our study builds on these foundations and utilises field observations together with oblique air photography coverage acquired during on-going investigations of the sedimentary architectures and facies (e.g. McCaffrey *et al.*, 2002).

The Tertiary sequence in Champsaur is weakly metamorphosed to prehnite-pumpelliite facies, generating a spotty appearance to the sandstones. The metamorphism was caused by tectonic burial beneath far-travelled Alpine thrust sheets (the Embrunnais-Ubaye nappes). However, the structures described in this paper predate the metamorphism – the spotty metamorphic fabric is undeformed. Furthermore the structures in the sandstones have developed without a penetrative cleavage, in marked contrast to the fold structures formed further east in the region that formed beneath the major thrust sheet.

The central part of the Champsaur outcrop preserves turbidites with remarkably little tectonic disruption, considering their position within the Alpine orogen (Figure 2). This low strain zone is bounded by two zones of structural complexity. These are described in turn, starting in the west.

Queyrel fold-thrust zone.

In the west, the Champsaur outcrop terminates on a high ridge culminating in the Pic Queyrel (2435m) with 600m of vertical relief around the peak. Individual sand bodies were correlated around Pic Queyrel using a combination of air photographs, draped onto a digital elevation model and visualised using ArcView. This digital mapping enhanced field studies and additional distant views acquired by helicopter and from adjacent hills (Figure 3). Results are presented as three serial sections and a structural map (Figure 4).

North side section

On the north side of the outcrop (Figures 3a and 4a), the Tertiary succession is folded into a large, NW-facing, inclined syncline with a steep SE limb. This structure is part of a large fold pair with the axial trace of the complementary anticline running close to the summit of Pic Queyrel. The forelimb of this composite structure contains minor thrusts at outcrop that effectively displace rocks from the forelimb towards the syncline. These structures are picked out by the turbidite sandstones. The underlying Nummulitic limestone is deformed into a fold pair with an amplitude of less than 30m, with no evidence of thrusting at this level. Higher in the succession thrusts on this section show low displacements with throws of less than the thicknesses of the main sand units (Figure 4a). Nevertheless the importance of thrusting in the disruption of bed continuity appears to increase up section. At this site these thrusts have juxtaposed sand panels. Kinematically the thrusts transfer displacement from the main forelimb area of the fold pair into the lower limb of the syncline. This type of behaviour has been described from other competent-incompetent multi-layers (e.g. Butler, 1992b). These thrusts are presumed to develop relatively late in the evolution of the structure, perhaps in response to mechanical hardening as the larger-scale fold tightens up. The syncline does indeed show progressive tightening and increasing curvature up section (Figure 4a).

South face section

The structure on the southern side of the Pic Queyrel is dramatically different (Figures 3b and 4b). Here a thrust has carried the Nummulitic limestone up into the turbidites. A dramatic asymmetric fold pair in these turbidites crops out ahead of the hanging-wall ramp in the limestones. The forelimb of this fold pair is cut by minor thrusts that bound a small wedge of sand. Otherwise the hanging-wall to this thrust is remarkably undeformed, with sub-horizontal planar beds cropping out. These geometries are suggestive of tip-line folding, with the thrust that carries the Nummulitic limestone terminating up-section a few metres beyond the hanging-wall cut-off of the limestone. However, closer analysis of the southern section reveals problems with a simple thrust tip interpretation. The horizontal separation between hanging-wall and footwall cut-offs for the Nummulitic limestone against this thrust exceeds 350m. This slip cannot be accommodated with a tip line just a few tens metres ahead of the hanging-wall cut-off line. Furthermore, the limestone has been elevated above its position in the footwall by c. 200m by being carried onto a section of turbidites. These cut-offs define a footwall ramp (X of Figure 4b). Consequently the turbidites must be offset by an additional thrust segment – and a hanging-wall ramp to this must lie to the NW of the outcrops (Y on Figure 4b). The bulk of the displacement must transfer onto this segment that presumably runs along a low net-to-gross horizon. Thus the whole thrust profile forms a staircase of ramps and flats through the turbidites (Figure 4b). Moreover, this structure has been folded so that the thrust locally dips to the NW. This fold structure is presumed to pass through to the fold pair in the Nummulitic limestone on the northern slopes of Queyrel.

The thrust segment that carries the hanging-wall ramp of Nummulitic limestone represents a minor splay from the inferred larger structure. However, this segment appears to post-date the fold in its footwall, the structure that folds the main thrust (Figure 4b). Consequently a model for the structural evolution is proposed (Figure 5) which has broadly simultaneous folding and thrusting. As folding continues it reorients part of the thrust above, which is therefore abandoned with part of the hanging-wall ramp. New thrust segments break through in the hanging-wall (Figure 5b).

Linking the sections

The west face of Pic Queyrel (Figures 3c and 4c) provides outcrop linkage between the two profiles. Here a thrust cuts up section through the turbidites with an inclined anticline carried in its hanging-wall. In the footwall there is a prominent sand package above a thick shaly unit that collectively show little evidence of folding (Fig. 3c). However, the thrust itself is decorated by a thin tract of sand, up to 10 m across. This forms a thin, fault bound horse which has significance for correlating sand bodies across the thrust zone. At first sight it might appear that the sand above the thick shale in the footwall correlates directly with the sand that has been carried onto it in the hanging-wall. With this correlation the throw on the thrust zone would be rather small, approximately equal to the thickness of the main sand body (Figure 6a). However, the presence of the sand horse continuing down dip below the footwall sand body, juxtaposed across the underlying shale (Figure 6b), strongly suggests that this sand horse and the sand body in the hanging-wall were derived from a stratigraphically deeper unit (Figure 6c). Consequently the thrust zone must accommodate a throw equal to or greater than the thickness of the sand-shale couplet in the footwall.

3D geometry

Given the above description the next task is to relate the outcrops to a 3D structural model. The serial sections for Figure 4 are constructed 500m apart indicating that lateral changes in structural style can be abrupt. The intermediate section (Figure 4c) shows a combination of thrusting and folding: consequently the displacement on the thrust is less than seen to the south. Presumably the shortening is taken up by folding in this section. The test however lies in the shortening. To examine these variations we measured change in bed-length for a single level within the turbidites, 180m above the Nummulitic limestone, a level that is the best defined by our study. The northern section (Figure 4a) shows 200m shortening at this level while in the southern section (Figure 4b) this rises to 670m. Both of these values carry uncertainities of about 20%. Nevertheless the lateral variation is significant, given that the two sections are just 1 km apart. Note that this is a far more rapid gradient in shortening that might be expected for shallow thrust systems – Elliott's (1976) "bow-and-arrow" rule for example predicts gradients of about 10% (i.e. section 1km apart should show no more than a 100m difference in shortening). The implications of 3D geometry proposed here is evident on a structural map (Figure 4d) which shows footwall and hanging-wall cut-off lines (for top limestone) that converge rapidly.

So are there problems in our estimation of shortening for the cross-sections? The displacements recorded by the southern section (Figure 4b) are conservative. The footwall cut-off of the Nummulitic limestone used in the section could be sited further back as it has been eroded from the section line. The shortening in the northern section is unlikely to exceed greatly that estimated here because there is no mappable offset of the Nummulitic limestone along the northern outcrop edge of the Tertiary in Champsaur. One solution might be to invoke bed-parallel slip in the turbidites of several hundred metres in addition to folding in the northern section. While theoretically possible no such detachments have been found. This model would also require a SW-dipping lateral ramp within the turbidite sand succession of Queyrel yet there are no indications of such a structure. Consequently we infer that there is a significant along-strike gradient in shortening across Queyrel.

The Palastre Thrust Zone

The eastern margin of the relatively undeformed tract of Champsaur turbidites is defined by the Palastre thrust zone. This structure was amongst the first described from this part of the Alps (Lory, 1901-2) and it is clearly visible from the open ground to the south (Figure 7a). All previous descriptions (e.g. Bürgisser, 1998) note that the displacement is accommodated not on a single discrete fault surface but in a thrust zone within which turbidite sandstones show varying degrees of deformation. The Palastre Thrust Zone directly underlies a major tract of folding and thrusting that carries the pre-Tertiary substrate (chiefly Jurassic limestones and shales) up and across the Champsaur sandstones. Deformation within this over-riding tract is complex, with panels of steep and locally overturned sediments. With respect to the underlying Palastre Thrust Zone, these over-riding structures may be considered as a tectonic allochthon. Outliers of the allochthon lie above the Champsaur sandstones on the hilltops of Soleil Boeuf, Pic Sud de Venasque and Le Pouveyret (Figure 1). Consequently the Palastre Thrust Zone represents a sub-thrust structure with respect to the allochthon and it is in this context that we have framed the following discussion. The structural evolution of the allochton itself will not be considered further here (but see Bürgisser, 1998).

Displacement on the Palastre Thrust Zone can be estimated from the offset of the basal marls that underlie the turbidite sandstones. The dip-separation between the hanging-wall and footwall cut-offs of this unit is about 1000m. Beneath the Palastre summit the thrust exhibits a large-scale footwall ramp that cuts through 450 m of turbidites. The displacement carries the basal marl that stratigraphically underlies the turbidite sandstone most of the way up the ramp. Further up the footwall ramp the thrust climbs onto an important low net-to-gross horizon. At this level the dip of the thrust decreases so that it refracts into the next sandstone unit at a lower angle (Figure 2). In this regard the Palastre Thrust Zone has a fault bend upon it. At the inflection point the sandstone package in the footwall to the thrust, directly underlying the main shaly horizon, has been folded and slices of the fold have been carried onto the overlying shaly unit (Figure 7b). This slice of sandstone truncates folds in the shaly unit below. The complex is then cut by steeper thrusts of a few metres displacement. The effect of this is to increase sandstone linkage across the shale and to reduce the curvature on the thrust.

Displacements of 1000m up the Palastre Thrust Zone should be expected to have deformed the hanging-wall strata into a fault-bend fold. The amplitude of the predicted, broad anticline should approximate the throw on the thrust that carries it. This prediction can be investigated by looking at the continuity of main turbidite units in the hanging-wall to the Palastre Thrust Zone and the overlying allochthon. The key feature is that the base of the allochthon can be traced from above the footwall ramp on the Palastre thrust to the outliers that lie ahead of the ramp. The regional section (Figure 2) shows that the base of the allochthon is not folded in the manner predicted above. The turbidite sandstones between the allochthon and the footwall ramp on the Palastre Thrust Zone must be truncated by the thrust at the base of the allochthon. The immediate footwall to the allochthon contains a tract of sand-rich turbidites that are detached and which have variable dips. This zone is interpreted as being the dismembered upper part of the hanging-wall to the Palastre Thrust Zone. These observations are consistent with at least some of the movement history on the base of the allochthon being after much of the movement on the Palastre Thrust Zone.

The continuity of beds in the footwall Palastre Thrust Zone can be investigated directly in the Roranche valley ahead of its footwall ramp. Here there is no evidence for bed-scale disruption of the sandstones – associated deformation is only found within a width of about 100m of this thrust zone. While erosive sedimentary surfaces within the sand bodies demonstrate that there has been no tectonic slip within these units, shaly horizons show evidence for bed-confined shear. The basal marls to the turbidites show extensive development of an anastamosing weak cleavage and s-c type fabrics (in the sense of Hanmer & Passchier, 1991) that indicate a shear sense of top-WNW (Figure 8). The effect of this deformation is to transfer slip into the footwall to the Palastre Thrust Zone which might be expected to link into contractional structures within the overlying turbidite sandstones. Similar fabrics are found in the upper shale-rich unit, the horizon at which the Palastre Thrust Zone ramp refracts.

Apparently isolated deformation exists at sites within the turbidite sandstones ahead of the Palastre Thrust Zone. One such example crops out on the southern flank of the Pointe Nord de Venasque where an isolated thrust dips to the west (Figure 9). As such it represents a back-thrust with respect to the overall tectonic vergence recorded in the Champsaur. The minor thrust shows a throw of just a few metres and transects a major sandstone package. In detail this thrust consists of soft-linked segments separated by a zone of kinking. The beds adjacent to the fault segments show minor folding. Similar backthrusts exist elsewhere in this same sandstone package (e.g. on the north side of the Co de Venasque). Collectively they represent minor attempts to thicken this unit by imbrication and presumably link kinematically to the Palastre Thrust Zone via the inferred weak detachment zones in the shaly units. This behaviour conforms to the thrust model for multilayers where the weak horizons tend to form thrust "flats" (detachments) between which the more competent units can thicken up by intra-formation thrusting.

A general model for deformation in the turbidite succession

The Champsaur area described above displays a range of different structural styles developed within the same sediment pile. These range from relatively discrete, near planar thrust zone segments through broader zones of thrusting to larger-scale folds with overturned dip-panels. However, on the kilometre scale, these deformations are relatively confined and enclose volumes of turbidites with essentially no tectonic deformation. Deformation is therefore heterogeneous and localised. We now develop a model for the evolution of a generalised thrust zone in the Champsaur (Figure 10).

We assume that the initial tectonic loading condition is layer parallel shortening. In this situation rheological multi-layers have a tendency to buckle. Variations in the alternating thicknesses of weak and strong layers in the Champsaur example are represented by the variations in sand content within the turbidite stack. Note that this is not equal to the individual bed thickness. It seems more probable that the sand-rich bed-set packages, up to about 80m thick, behave as coherent mechanical layers. The amplitude/wavelength characteristics of buckles are strongly controlled by the amount of shortening, the viscosity contrast between layers and the layer thickness (e.g. Biot, 1961). However, as Ramberg (1961) shows, the different layers will interact mechanically via a zone of contact strain equal to about one fold wavelength. Consequently, variations in the multi-layer character – in effect the variations in bulk sand content through the stratigraphic pile – can engender variations in structural style. In experimental studies of buckle folds where competent layers of differing thickness are separated by thick incompetent layers, this complexity is represented by disharmony in folding through the multilayer (e.g. Ramberg, 1964). However, when the incompetent and competent layers are relatively thin the resultant anisotropy tends to generate larger wavelength, harmonic folding. Note that the differing fold styles have different mechanisms (e.g. Price & Cosgrove, 1990). Folding of strongly anisotropic materials is accommodated principally by flexural slip on bedding planes while single layer folding is generally accommodated by limb rotation together with tangential longitudinal strain in the hinge zones. These differences might be reflected in different distributions of associated damage within beds (e.g. fracture or porosity-occluding solution fabrics). However, further discussion of these issues lies beyond the scope of this paper.

In all cases folding eventually ceases and displacement on thrusts achieve the shortening. However, the timing of thrusting relative to fold amplification is likely to vary from layer to layer. Consequently we envisage the deformation zone to evolve from a fold on to a through-going thrust ramp via a period where thrust segments operate in series with folds linked kinematically through the multi-layer. Note that as different wavelengths of fold develop in multi-layers, eventually order is imposed by the thicker competent units which define the dominant wavelength for further amplification (e.g. Ramberg, 1964). Consequently thrusts formed in thin units early in the deformation may become folded (with a cessation of slip upon them) as a larger scale of folding is imposed upon them. Note that folds grow along their hinge lines so that this imposition of folding overprinting thrusts could happen into the plane of a section. Perhaps this provides an explanation of the cycling between folding and thrusting inferred for the Queyrel area (Figure 5).

The later that thrusts form in the fold amplification of a particular layer, the greater the propensity for rotated bedding and the subsequent development of complex thrust zone architectures, decorated with minor horses of steep to overturned beds. Applying Ramberg's (1961) notion of contact strain, we predict that folding is likely to progress further when competent layers are widely separated from each other by thick incompetent layers. In this context, folding should be expected to be more important adjacent to the thick, low net-to-gross units. In the case of the Palastre Thrust Zone, such deformation is seen at the top of the steep segment of the footwall ramp, at the base of the upper, thick shaly unit.

Classical buckle models (e.g. Biot, 1961; Ramberg, 1961) predict that folds forming in this way should form in long trains. Aspects of this prediction are discussed by Casey *et al*. (in press), elsewhere in this volume. However, in the Champsaur area the deformation zones are not obviously parts of longer fold trains but rather are isolated. This suggests some form of strain softening - perhaps enhanced by the formation of weak fault zones that become progressively more linked to define a thrust ramp. In this way the far-field layer-parallel shortening is resolved into an inclined zone that effectively accommodates thrust-sense simple shear – a distributed ramp. So far our discussions have been confined to a two dimensional view. However structures can vary in style in three dimensions, as is evident at Pic Queyrel. This example has a fold passing laterally into a more thrust-dominated structural style. Conventionally this relationship might appear to represent a tip zone to the thrust (e.g. Williams & Chapman, 1983) with the implication that increasing deformation should see the thrust break through the fold. Therefore any particular site should see a history of folding evolving into thrusting. However, the Queyrel example behaves differently. The fold structure deforms thrusts, and thrust splays are inferred to show a break back sequence. This implies alternations in periods of folding and thrusting in single sites – rather than the simple linear progression from folding to thrusting.

Discussion

The trivial conclusion of our study is that structural styles in the Champsaur Sandstones are complex and no single end-member model for deformation (ideal buckling, simple fault-bend folding) explains the range of structures present. By implication an expectation of complexity should be assumed for other settings involving compressional deformation of complex multi-layers such as turbidites. The obvious questions to ask are whether these variations are important in understanding the geometry and history of hydrocarbon reservoirs and if so what strategies might be adopted for reducing risk in structural interpretations of such systems.

In answer to the first question, understanding the geometry of compressional deformation could be important. The development of thrusts preceded by buckles can develop fault zones of finite width and internal complexity within which the possibility of sand-sand juxtapositions are greatly enhanced. Therefore we might predict that the vertical permeability in these types of systems is potentially much greater than might be predicted using thrust tectonic models with very discrete fault surfaces.

Although the disruption of stratigraphic units may be sufficient for the general nature of deformation similar to that of our Champsaur analogue to be detected on seismic reflection data-sets in the subsurface, the details of fault zone architecture lie beyond the reach of seismic methods. Indications of the complexity might be deduced from wells that cross the fault zones. While core offers the best chance of identifying complex fault zone architecture, less direct information such as from dip-meter and image-logging data could well differentiate between discrete thrusts and the broader thrust zones with their tendency for increased sandstone juxtapositions.

An alternative approach is to model structural evolution in a specific site calibrated by the local stratigraphy and any available structural data. As buckle folding is highly sensitive to the apparent viscosity contrasts between competent (high sand content) and incompetent (high shale content) units and the relative thickness of these layers, gaining understanding on the nature of the stratigraphic layering and on the rate of deformation (from growth strata) should aid prediction. Mechanical solutions are likely to be more powerfully predictive than simple empirical approaches. This seems a potential fruitful avenue for further investigation. While the approach may not yield unique solutions it may at least provide a basis for assigning risks on reservoir models using different fault zone architectures and their impact on reservoir models.

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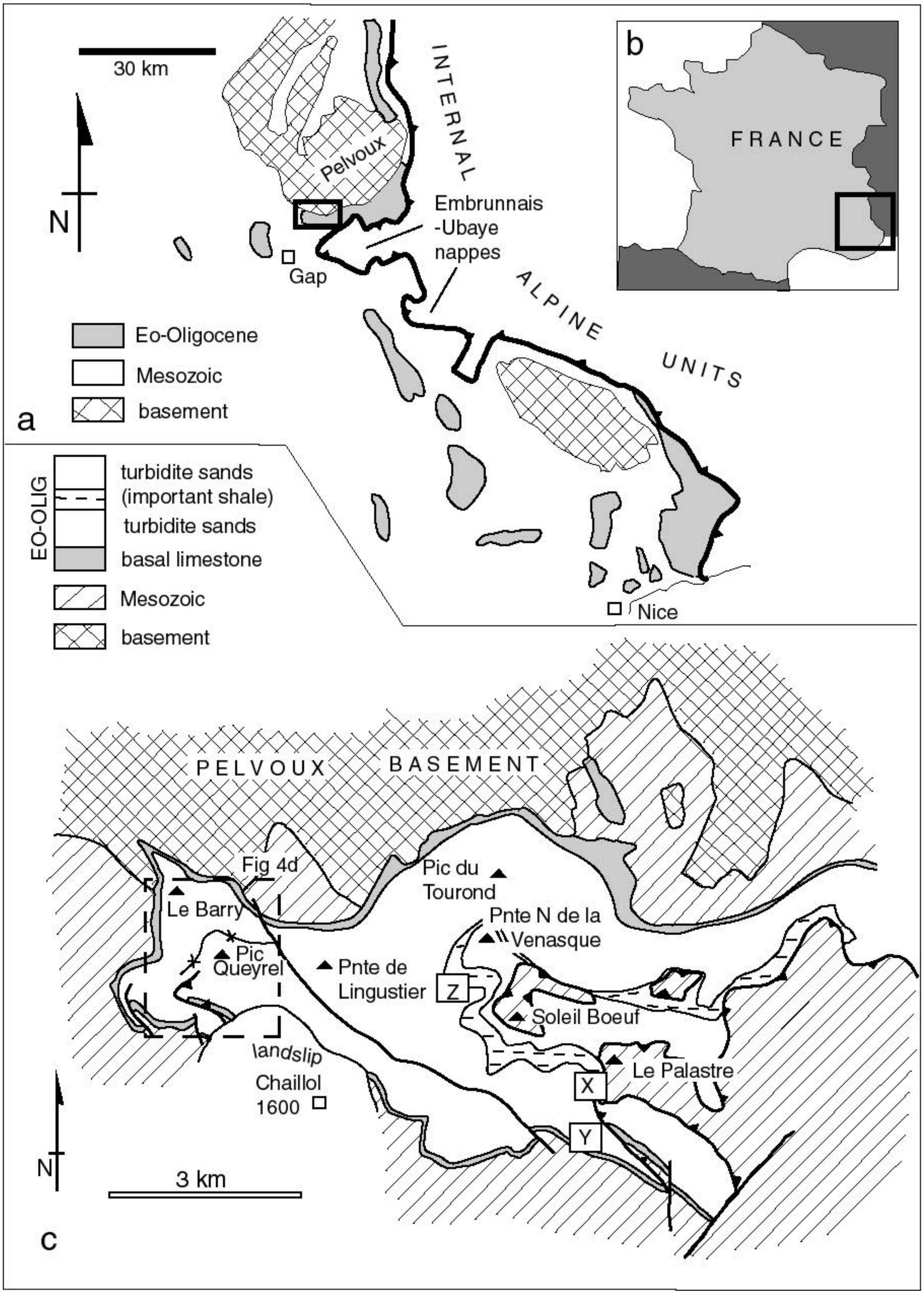
Figures

Figure 1 (a) Simplified geological sketch map of the SW Alps (located in SE France, Figure 1b) showing the outcrops of Eocene-Oligocene turbidites (Gres d'Annot and Gres du Champsaur). (c) Simplified geological sketch map of the Champsaur outcrops (Chaillol hills, boxed area on Fig. 1a). X, Y and Z refer to the locations of Figures 7b, 8 and 9 respectively. Chaillol 1600 is a small ski-station.

Figure 1



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-Ubaye
nappes
Gap
ALPINE
UNITS
Eo-Oligocene
Mesozoic
basement
a
b
FRANCE
Nice
EO-OLIG
turbidite sands
(important shale)
turbidite sands
basal limestone
Mesozoic
basement
PELVOUX
BASEMENT
Fig 4d
Le Barry
Pic
Queyrel
Pnte de
Lingustier
Pic du
Tourond
Pnte N de la
Venasque
Z
Soleil Boeuf
landslip
Chaillol
1600
X
Le Palastre
Y
N
3 km
c

Figure 2 Cross-section through the Champsaur outcrop (Chaillol hills). The section line involves some projection (place names and key shown on Figures 1c). X, Y and Z refer to the locations of Figures 7b, 8 and 9 respectively. The inverted basement-Mesozoic contact below Le Barry testifies to the complex deformation that predated the deposition of the Tertiary succession here. The key is as Figure 1c, although shale horizons are omitted for clarity.

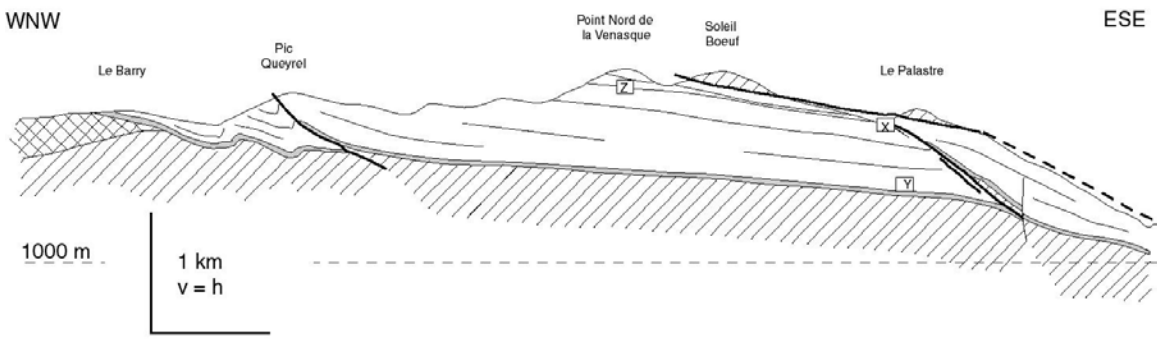


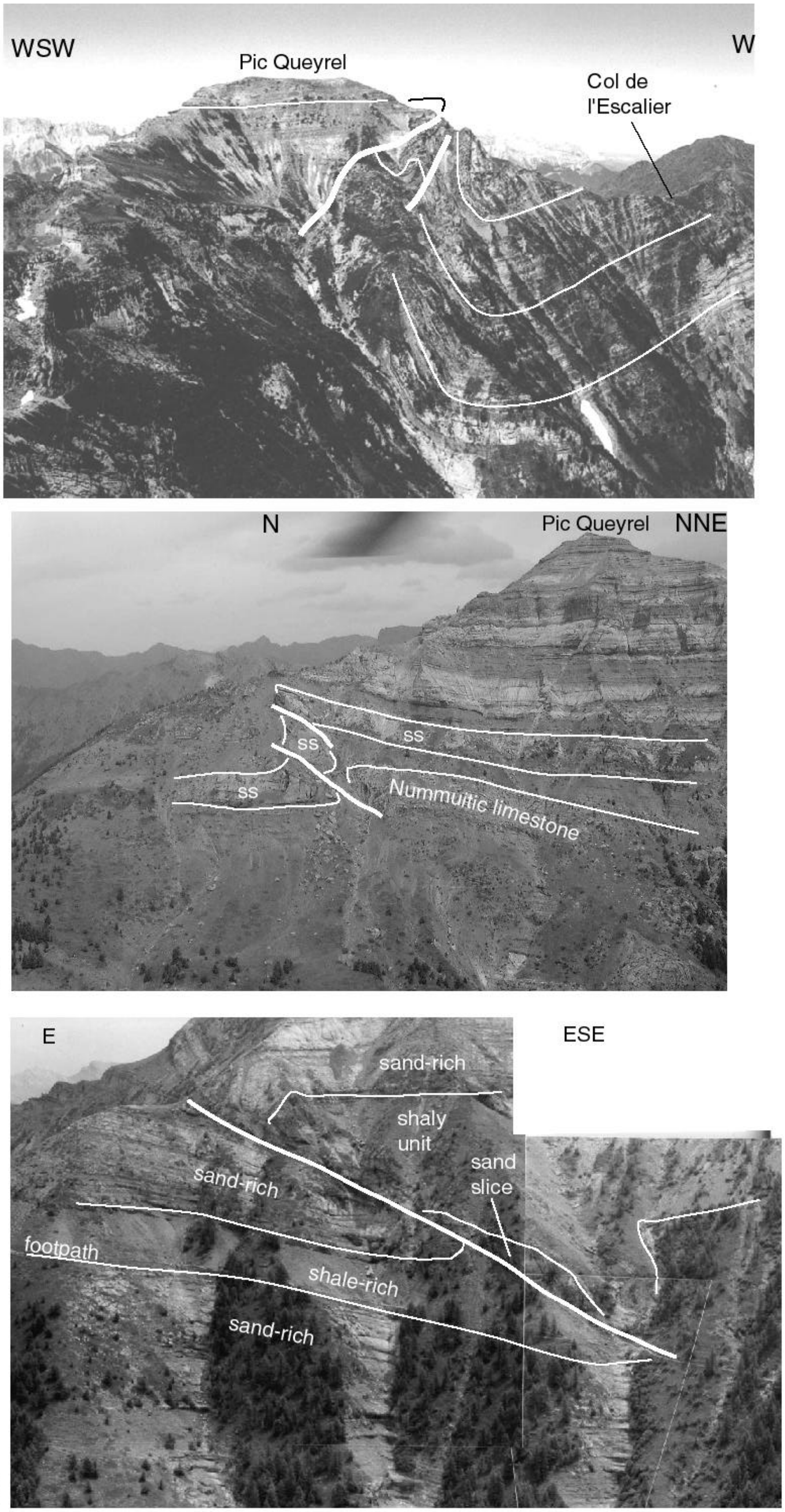
Figure 2

Figure 3 Photographs of structural geometries in the Pic Queyrel (2435m) area. (a) Looking SW onto the NE face of Pic Queyrel (taken from near the Pointe de Lingustier) showing the NW-facing synform picked out by sandbodies in the turbidites. The visible section is c. 600m high (see Figure 4a). (b) The south face of Pic Queyrel (photograph taken from helicopter, looking north), showing the light-coloured Nummulitic limestones carried up into the Champsaur sandstones (see Figure 4b). (c) West side of Pic Queyrel (photograph taken from helicopter), looking highly obliquely to the tectonic transport direction (see Figure 4c).

a)

b)

c)



WSW
Pic Queyrel
W
Col de
l'Escalier
N
Pic Queyrel
NNE
ss
ss
ss
Nummulitic limestone
E
ESE
sand-rich
shaly
unit
sand
slice
sand-rich
footpath
shale-rich
sand-rich

Fig. 3

Figure 4 Structural geometry of the Pic Queyrel area. A-A', B-B' and C-C' are serial sections along the lines shown on Figure 4d. The key for the cross-sections (a-c) is as in Figures 1c and 2. In the map the Nummulitic limestone is shown by diagonal ruling, while the turbidite sandstones and shales are unornamented except for a prominent shale unit (shaded).

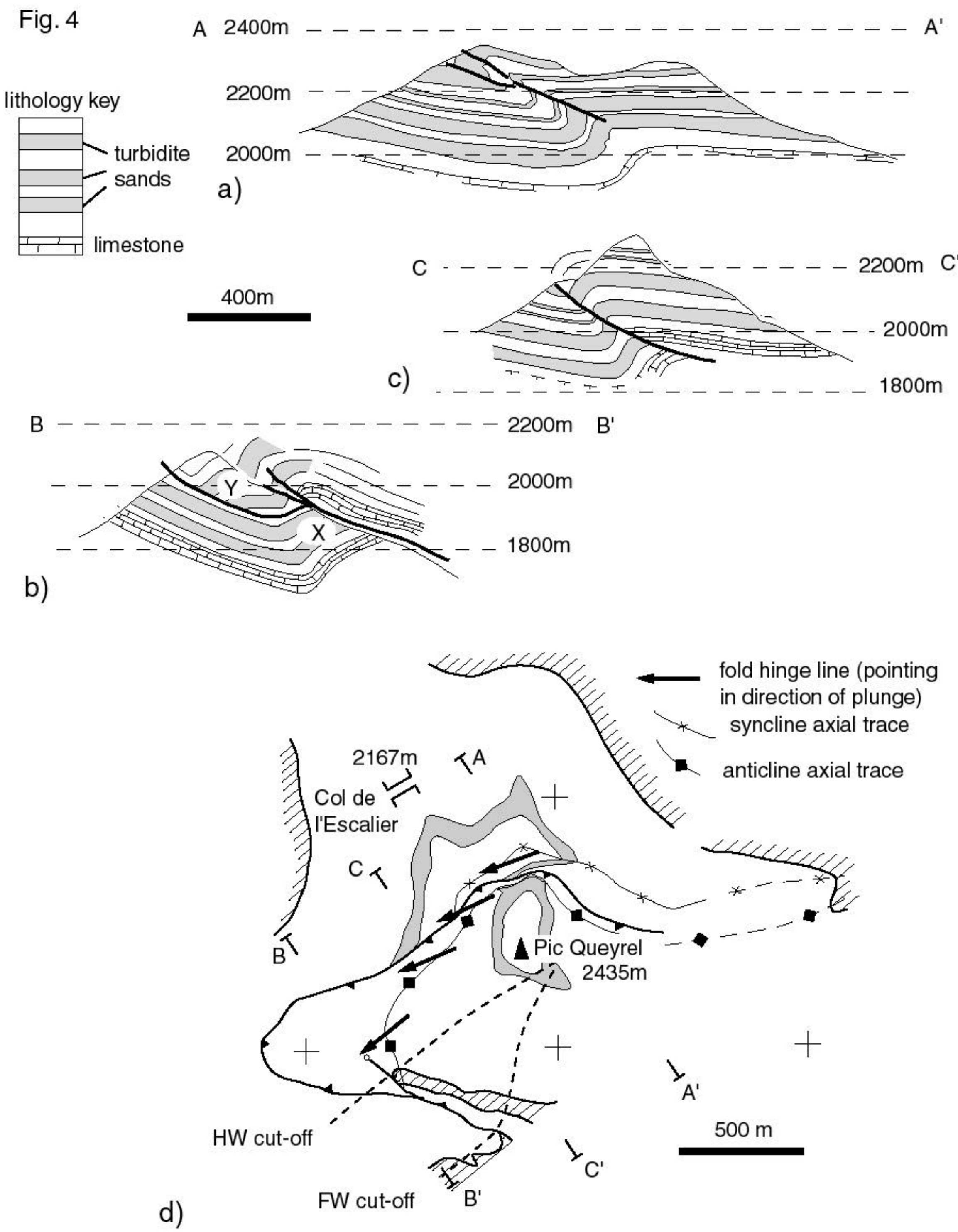


Figure 5 An model for the structural evolution (a-b in time) of the section through the southern Queyrel area (Figure 4b). A thrust ramp has formed, climbing up through the Nummulitic limestones and about 200m of turbidites. However, the footwall to this structure is about to fold (approximate position of axial surfaces are indicated). After the fold pair has amplified (Figure 5b), any renewed slip up the thrust ramp generates a new thrust segment (short-cut) in the hanging-wall, carrying part of the back-limb of the fold pair across the forelimb.

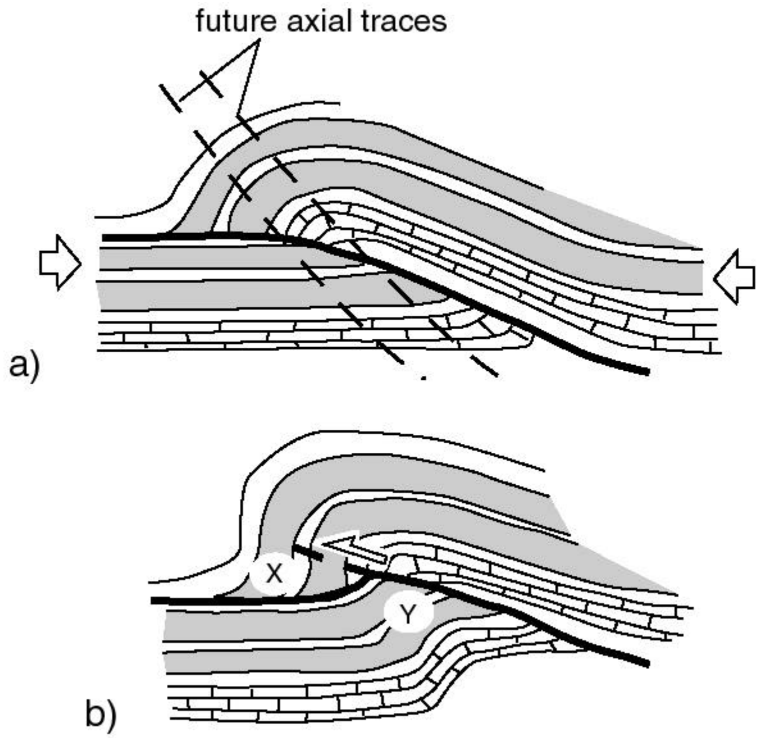


Fig. 5

Figure 6 The implication of sand slices for the correlation of sand bodies across the thrust zone as seen on the western side of Queyrel. (a) shows an interpretation with rather low displacement (juxtaposed sand units in hanging-wall and footwall correlate). In this situation, sandstone slices can exist between the leading hanging-wall cut-off and training footwall cut-off of the sand unit (x) but there can be no sand slices hindward of the trailing footwall cut-off (y) until the next (underlying) sand unit is encountered. In outcrop (represented by boxed area on Figure 6b) the presence of sandstone slices hindward of the trailing footwall cut-off must be derived from a deeper (underlying) sand unit. This is illustrated on the schematic partial restoration (Figure 6c). Thus the juxtaposed sand units do not correlate.

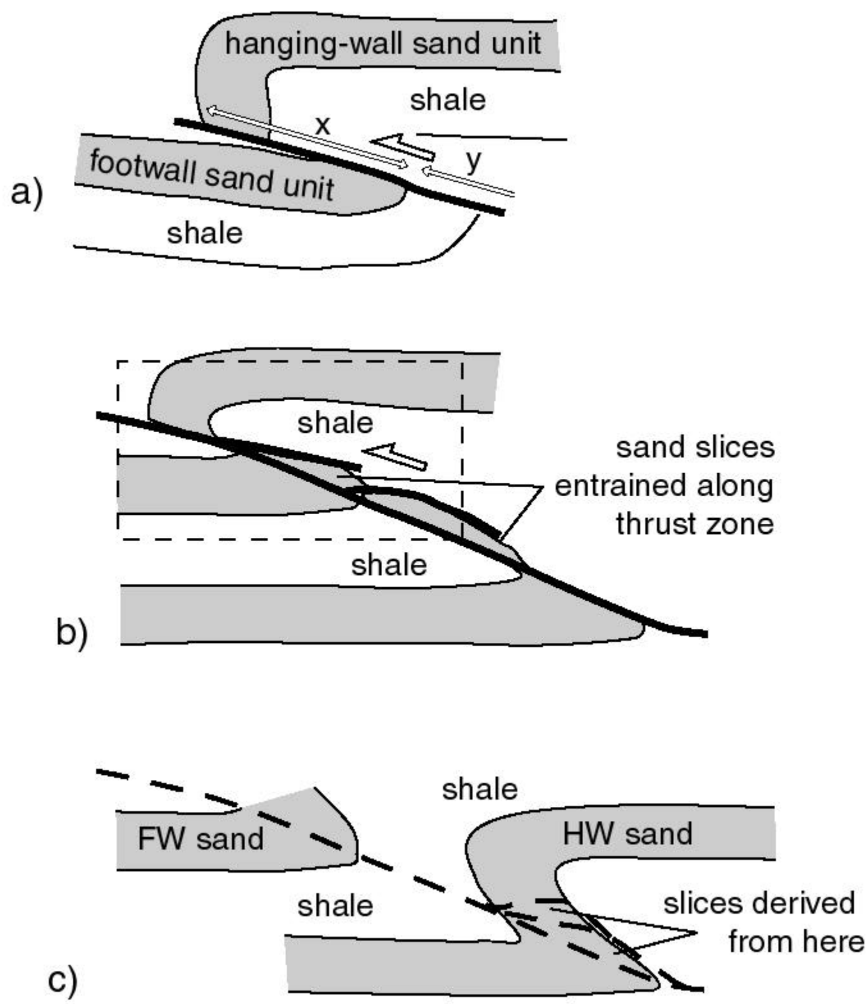
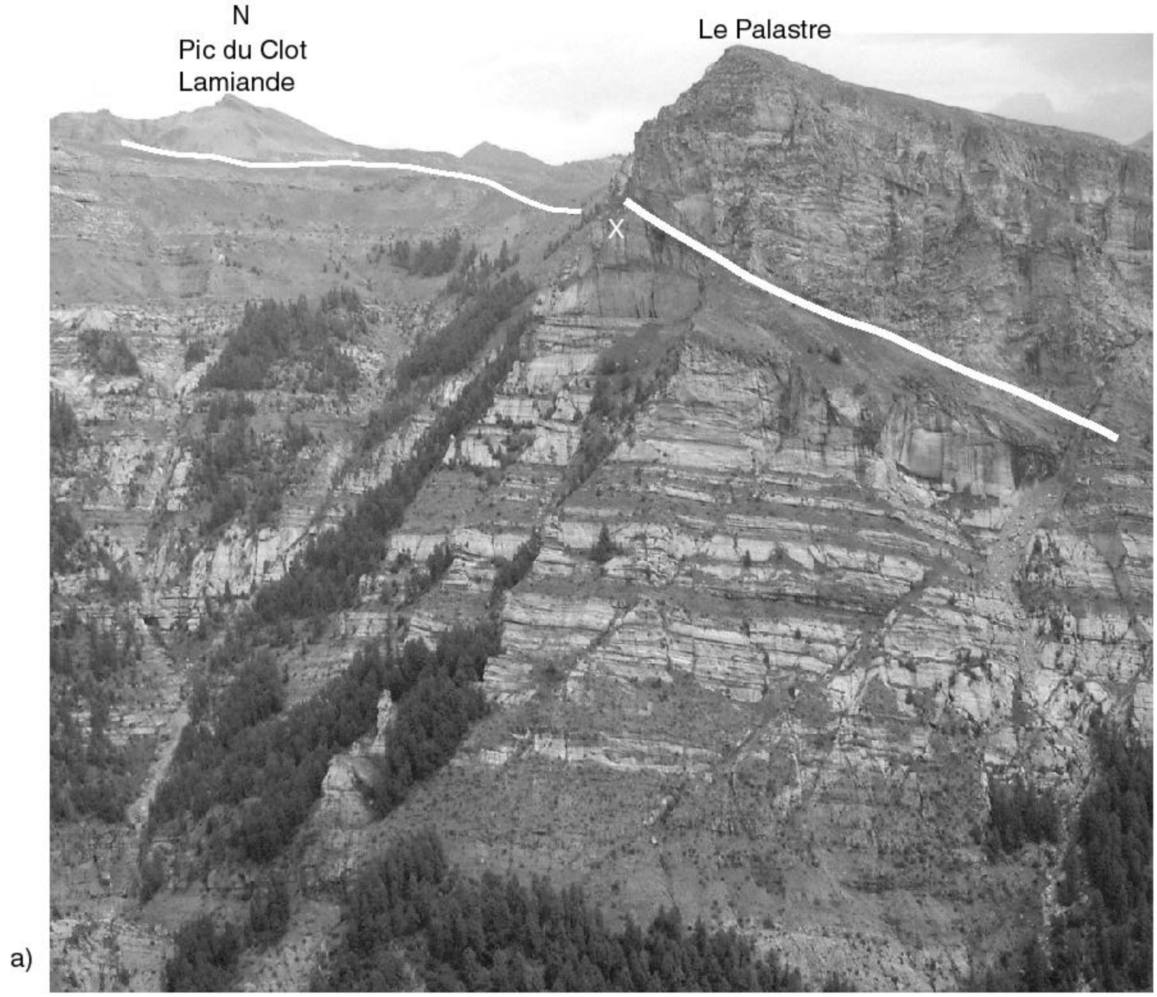
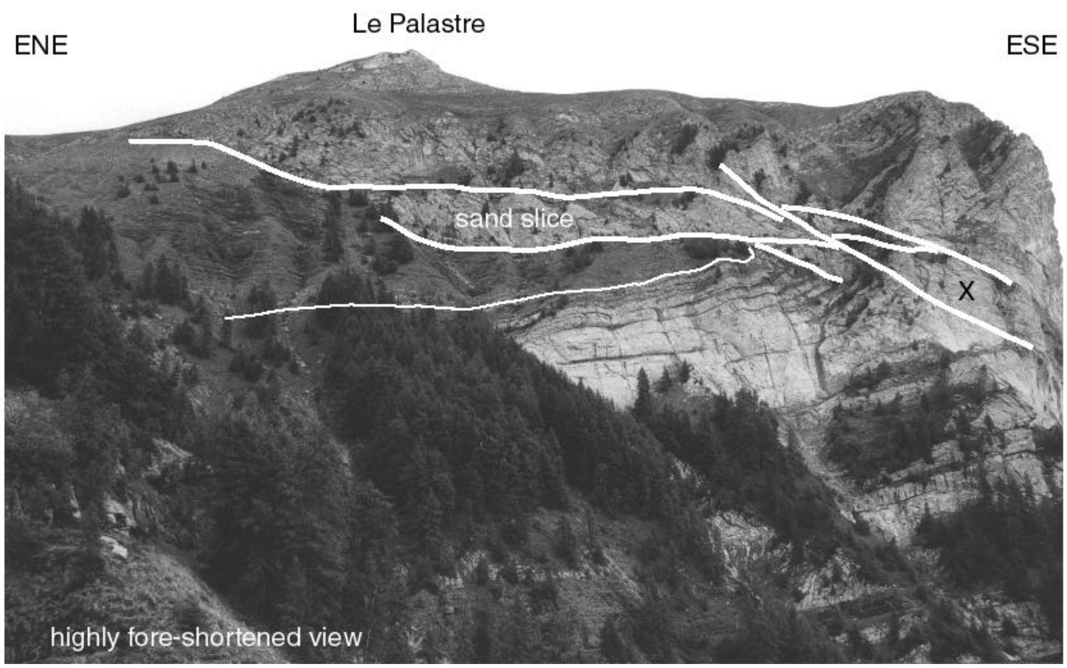


Fig. 6

Figure 7 The Palastre Thrust Zone (Figure 2). (a) photograph of the south face (taken from helicopter). (b) Detail of the horse of sandstone that crops out towards the top of the footwall ramp in the Palastre Thrust Zone.


N
Pic du Clot
Lamiande
Le Palastre
X

a)


Le Palastre
ENE
ESE
sand slice
X
highly fore-shortened view

b)

Figure 8   Anastamosing weak pressure solution cleavage in the basal marls at the foot of the Palastre section. The geometry of these foliations is akin to s-c fabrics and is indicative of top to WNW layer parallel shear.

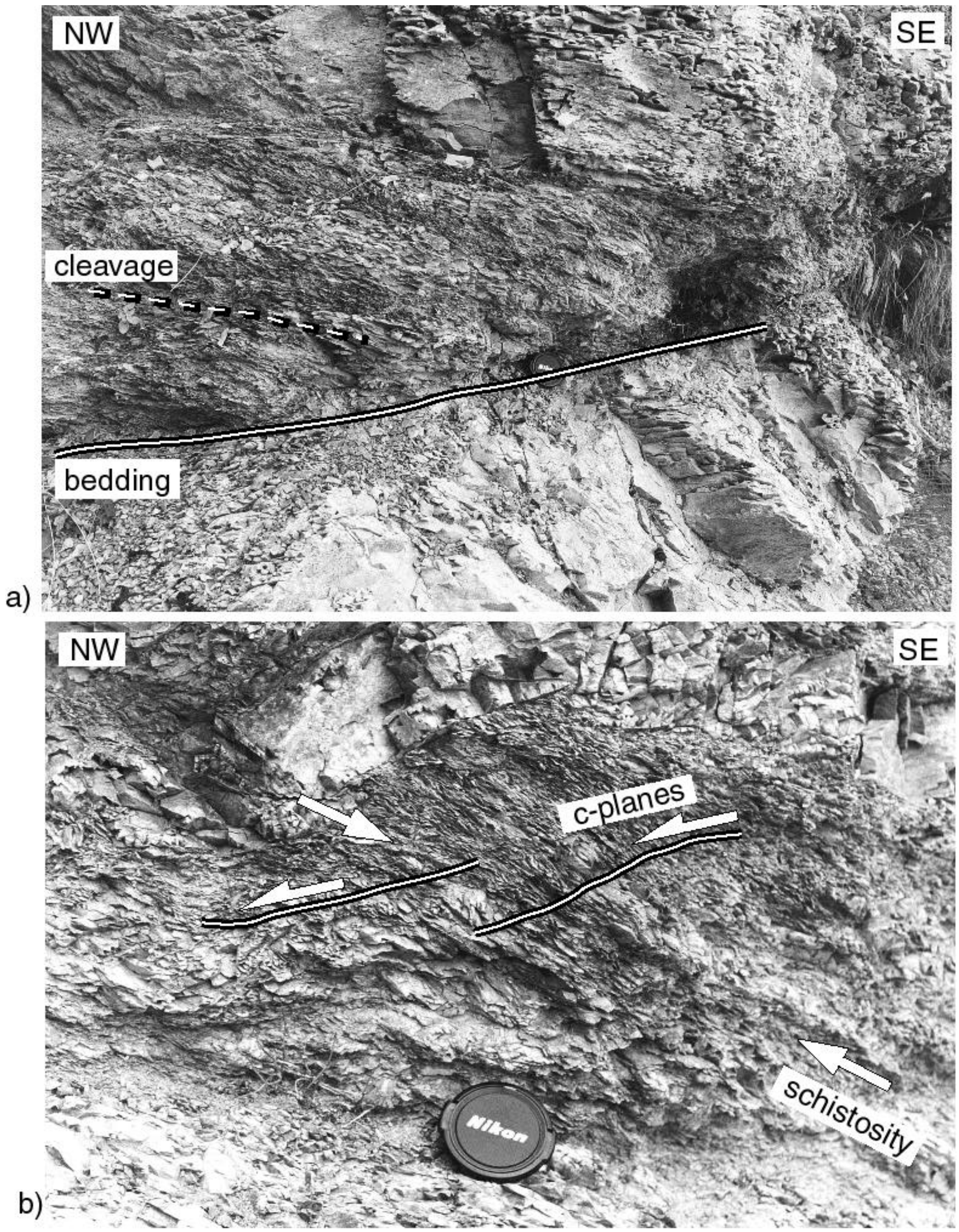
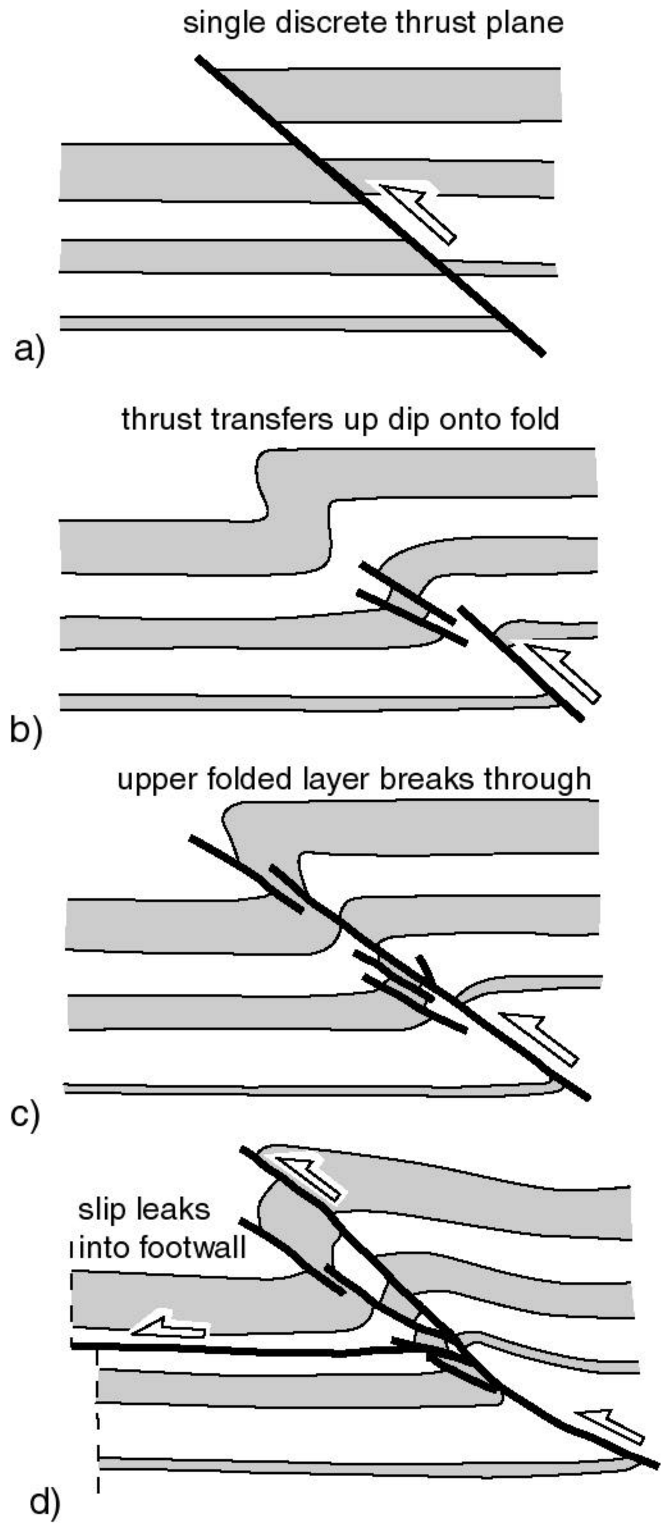


Fig. 8

Figure 9 Spatially isolated, west-dipping thrust exposed on the south ridge of Pointe Nord de la Venasque.

Figure 10 General models for the evolution of thrust zones through multilayer sandstones and shales based on the Champsaur study. a) shows an idealised thrust geometry where displacement is localised onto a single narrow thrust surface. In this geometry, the propensity for sand juxtaposition is simply predicable on the basis of knowing the vertical sequencing of sandstone and shale together with the thrust zone displacement. This idealised geometry is not believed to be applicable to our case study. The remaining diagrams show the idealised evolution (b-d in time) of a thrust zone based on the Champsaur outcrops. At an early stage layers can accommodate shortening through varying degrees of thrusting and buckle folding. With increasing bulk strain there is increased probability of thrust segments linking up. Further slip can be transferred out into adjacent parts of the multilayer. As shown here, slip is transferred into the footwall. Composite thrust zones such as this, involving buckle folding and localised fault slip greatly increase the propensity for sand-on-sand juxtapositions compared with the simple thrust model (Figure 10a). There is also the increased chance for isolated, shale-enveloped sandstone slices will be developed along the thrust zone.

Fig. 10



single discrete thrust plane
a)
thrust transfers up dip onto fold
b)
upper folded layer breaks through
c)
slip leaks
into footwall
d)