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Recumbent folds: key structural elements in orogenic belts

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Abstract

This review has two main parts. The first of them presents existing ideas and data related to recumbent folds, reviewing aspects such as the physical conditions of the development of these folds, the strain inside the folded layers, the kinematic mechanisms of their formation, the role of gravitational forces, the tectonic context of their development and the structures associated with them. In the second part, the above ideas are discussed and possible mechanisms for the development of these folds are presented. It is proposed that initial perturbations of the layers are essential to give rise to the asymmetry of recumbent folds. These perturbations may be non-planarities of the layering or may be linked to the existence of a core or basement of competent rock that hinders the normal propagation of the deformation. This could explain why many large recumbent folds have a root zone.

Deformation with an important component of simple shear is a general condition for the formation of recumbent folds. In areas with very low grade metamorphism, competent layers often play an active role during the deformation and undergo buckling with the development of an overturned fold limb, which can be stretched and thinned to finally produce a pair of recumbent folds separated by a thrust. In areas with low or medium metamorphism, buckling under a simple shear regime is probably the most important mechanism for producing large folds with gentle or moderately dipping axial surfaces; subsequent kinematic amplification by coaxial strain components with vertical maximum shortening is important for the formation of recumbent folds. These components involve a sub-horizontal stretching that can cause a problem of strain compatibility and give rise to a

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basal thrust.In areas deformed under high *P* and *T* conditions, recumbent folds can develop by flow perturbations and kinematic amplification of folds; this is probably a common mechanism in ductile shear zones.

Keywords: folding; geometry; strain; kinematics; lineations; orogens

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1. Introduction

The term *recumbent fold* was defined over a century ago and folds fitting this definition have been recognized on a wide variety of scales and in varied geological contexts. According to Geikie (1905, p. 137), "*Recumbent fold is the name given to a flexure, the axial plane of which approaches horizontality*". A similar definition was given by Leith (1913, p. 105) and Ries and Watson (1914, p. 154), and this is consistent with current usage by most authors. Fleuty's (1964) scheme for the classification of folds based on the attitude of the axial plane proposes an angle of 10° for the maximum dip of the axial plane of a recumbent fold and this convention has been widely accepted (e.g., Ramsay, 1967, p. 358-359; Dennis, 1972, p. 144; Ramsay and Huber, 1987, p. 322; Price and Cosgrove, 1990, p. 235; Ghosh 1993, p. 225; van der Pluijm and Marshak, 2004; p. 244). Although this classification scheme is based on geometry rather than on the mode of origin of the folds concerned, it is evident that the formation of recumbent folds requires a different set of physical conditions than those that lead to folds with upright axial planes. In this paper we attempt to identify these specific conditions and the tectonic regime in which these conditions pertain.

Large recumbent folds were recognized in the 19th century in the Alps (Escher von der Linth, 1841; Gerlach, 1869; Renevier, 1877; Heim, 1878) and were named "recumbent-fold nappes" (Bertrand 1884, 1887), "fold nappes" (Haug, 1900), and "first-order nappes" (Termier, 1906). Nowadays the term "fold nappe" is most commonly used, but it has a certain ambiguity. Most authors agree that a large recumbent fold is a fold nappe, but a terminological problem arises when the lower limb of the fold is cut by a thrust. Thus, for example, the Morcles Nappe (Helvetic zone of the Swiss Alps) is considered by Ramsay and Huber (1987), Price and Cosgrove (1990) and Epard and Escher (1996) to be a fold nappe, whereas it is referred to by Twiss and Moores (1992) as a thrust nappe. In agreement with most authors, and in line with Dennis et al. (1981), we prefer to define a fold nappe as a large recumbent fold whose lower limb may be cut by a thrust. Another point of contention is the length of the overturned limb, i.e. the length of the stratigraphic inversion necessary for the structure to be considered a nappe. Some authors suggest a minimum inversion of 5 km (e. g., Dennis et al., 1981; France, 1987), whereas others consider the minimum to be 10 km (e.g., Ramsay and Huber, 1987, p. 521). In any case, this decision is arbitrary and the change from 5 to 10 km is unlikely to involve a qualitative change in the kinematics of the structure.

It is interesting to consider further the common association of a large recumbent fold and a thrust cutting the lower limb. Explanation of this association may help to understand the kinematics of recumbent folds. Another additional characteristic of recumbent folds is that they are usually either close folds or isoclinal folds, and in some cases are more tightened than the non-recumbent folds found in other areas of the same orogen (e.g., Sanderson, 1979; Bastida et al., 2010).

The development of near-isoclinal large recumbent folds facing towards the foreland is common in the early phases of deformation in orogenic belts. Upright or inclined folds often appear superposed on these folds. In many cases, this superposition gives rise to interference patterns of the Type 3 of Ramsay (1967). This change in the style of folding with time must be associated with a change in the local stress regime. Consideration of the reasons for this change is key to our understanding of the kinematics and mechanics of orogenic belts.

In addition to the large recumbent folds with their corresponding parasitic folds, outcrop-scale recumbent folds are found in ductile shear zones developed commonly in the hinterland of orogens (e.g. Aller and Bastida, 1993; Yassaghi et al., 2000; Williams and Jiang, 2005). In these cases, near-isoclinal anticline-syncline fold pairs are common. Many of these folds have curved hinges and are sometimes sheath folds.

As will be seen in the examples described below, references to recumbent folds are countless and these folds have been described in a wide variety of regions: Archean cratons or terranes, greenstone belts, Proterozoic orogens, the Caledonian belt, the Northern Appalachian Mountains (Acadian deformation), the European Variscan belt, the Ural Mountains, the Moroccan Paleozoic massifs, the Central and Southern Appalachian Mountains, the Canadian Rocky Mountains, the Alps, the Himalayas, the Axial Zone of Pyrenees and the Betic Cordillera.

Recumbent folds usually develop in the internal zones or the internal/external transitional zones of orogenic belts, i.e., in a compressional tectonic regime (e.g. Hatcher, 1981; Ramsay, 1981; Dietrich and Casey, 1989; Simancas et al., 2004; Fernández et al. 2007; Ryan and Dewey, 2011). Nevertheless, they can also develop in extensional regimes (e.g., Platt, 1982; Froitzheim, 1992; Vissers et al., 1995; Orozco et al., 1998; Harris et al., 2002). Recumbent folds can occur as sedimentary structures, associated with slumps or overturned cross-bedding (e.g., McKee et al., 1962; 1971; Allen and Banks, 1972; Hendry and Stauffer, 1975; Fitches and Maltman, 1978; Doe and Dott, 1980; Plint, 1983; Owen 1985, 1987, 1996; Farrell and Eaton, 1987; McClay, 1987; Paim, 1995; Nigro and Renda, 2004; Nichols, 2009; Pye and Tsoar, 2009; Alsop and Marco, 2011, 2012, 2013). They can occur also in glaciers of ice or salt and in the sediments associated with ice glaciers, such as push moraines or deformation tills (e.g., Hudleston, 1976, 1977, 1983; Talbot, 1979, 1998; Talbot and Aftabi, 2004; Sans et al., 1996; Hambrey and Lawson, 2000; van der Wateren, 2002; Hooke, 2005; Hudec and Jackson, 2006, 2007; Evans, 2007; Leseman et al., 2010), or in impact structures (Price, 2001).

The fact that recumbent folds are often large and subsequently re-deformed structures makes it difficult to establish their original geometry. The isoclinal character of these large folds, together sometimes with the scarcity of outcrops and the monotony of the lithology, can even make them difficult to recognize. This complicates their kinematic and mechanical analysis. Furthermore, the wide variety of geological settings in which recumbent folds occur and their relation to other structures poses a number of additional problems.

The aim of this paper is to review the different theories for the development of recumbent folds and to consider how the analysis of the bulk strain undergone by the folded rocks can help to understand the mechanisms of formation of recumbent folds. We also seek to explain the observed relationship between recumbent folds and the other structures they are often associated with. Finally, we try to analyze the broader significance of recumbent folds in the context of orogenic belts. The analysis will be focused on the recumbent folds generated by tectonic deformation. Recumbent sheath folds have a peculiar geometry and pose specific problems that are not considered here.

2. Existing ideas and data relating to recumbent folds

Many papers make mention of recumbent folds and a variety of theories have been proposed to explain their development. For their systematization we consider in this section the descriptions and analyses made about the physical conditions, strain state, kinematics, driving forces (specifically the role of the gravitational forces), tectonic regime, and structural associations relating to recumbent folds.

2.1 Physical conditions of the development of recumbent folds

Recumbent folds can develop in a notable variety of physical conditions. However, there are certain specific conditions under which they are more common. Holland and Lambert (1969) asserted that the onset of metamorphic reactions marks the upper boundary of a soft zone within the orogenic crust. According to these authors, the formation of micas and amphiboles from sedimentary minerals is commonly associated with the development of widespread tectonic foliation and of similar (class 2) recumbent folds.

Fyson (1971) remarked on the common temporal change in orogenic belts from isoclinal recumbent folds to more open upright or steeply inclined folds. He asserted that during this change the metamorphic grade can increase or decrease and therefore the influence of the temperature on the development of folds is not easy to evaluate. Fyson emphasized the influence of lithostatic pressure, a high value of which generates horizontal extension and deformation. According to this idea he emphasizes, in agreement with the interpretations by de Sitter and Zwart (1960) of the Variscan deformation of the Pyrenees, the existence of structural levels (superstructure and infrastructure), with recumbent folds restricted to the infrastructure (Fig. 1). Mattauer (1973, pp.181-192) developed this concept introducing a middle structural level characterized by the dominance of parallel folds, between the upper structural level dominated by faults and a lower structural level with flow, non-parallel folds and widespread tectonic foliation. Subsequently, this structural model was only used occasionally by a few authors (Zwart, 1979, 1981; Murphy, 1987; Carreras and Capella, 1994) until the revival of this concept (Culshaw et al., 2006) based on the development of orogenic thermal-mechanical numerical models describing channel flow in the middle-lower crust (Beaumont et al., 2001; Godin et al., 2006). These models illustrate the contrasting styles of deformation in infrastructure and superstructure (Beaumont et al., 2006). One of these models (hot fold nappes) simulates a mid-lower crust that is forcibly expelled outward over a lower crustal indentor to create fold nappes that are inserted into the mid crust. This model is driven tectonically by, for example, collision with a strong crustal indentor. Williams and Jiang (2005) and Williams et al. (2006) explain the development of recumbent folds in mid- to-deep crustal levels (infrastructure) due to modification of earlier structures by large magnitude simple-shear, which in some cases can be related with channel flow processes. These structures contrast with the upright folds developed in the upper levels (superstructure). The existence of structural levels in the context of channel flow with recumbent folds in the infrastructure has also been described by Kuiper et al. (2006) in Shuswap Complex of the Canadian Cordillera.

An example of crustal division into infrastructure and superstructure not linked to channel flow processes has been described by Carr and Simony (2006) in the Omineca Belt (Canadian Cordillera). The suprastructure is dominated by moderately inclined to upright folds, with axial-planar cleavage and sharply defined ductile thrusts in metamorphic rocks in greenschist or subgreenschist facies, whereas the infrastructure is characterized by tight recumbent folds, rootless folds, boudinage and strongly foliated migmatite, paragneisses and orthogneisses.

A spatial variation of the metamorphic grade through a fold nappe is common. For example, Dietrich and Casey (1989) and Casey and Dietrich (1997) describe a vertical and horizontal gradient of metamorphism in the Helvetic nappes (Fig. 2), so that the metamorphic grade increases from the upper nappe (Wildhorn) to the lower nappe (Morcles), and, for any specific nappe, from the front to the root zone. The Kübler index of the illite shows that these variations range from diagenetic to epizonal conditions and they correlate with an increase in strain and in the intensity of the tectonic foliation associated with the folds (Dietrich and Casey, 1989; Goy-Eggenberger and Kübler, 1990; Crespo-Blanc et al., 1995). In the north-eastern extension of these nappes, three units can be distinguished from bottom to top: Doldenhorn (lateral equivalent to the Morcles Nappe), Gellihorn and Wildhorn nappes. The ductile deformation is more intense in the Doldenhorn nappe, where large recumbent folds were developed (Herwegh and Pfiffner, 2005). The metamorphism of this area was studied by Burkhard (1988), using the Kübler index of illite and clay minerals, and Herwegh and Pfiffner (2005), using calcite-graphite and calcite-dolomite thermometry in carbonate rocks. The latter authors established the structural and metamorphic evolution of the area (Fig. 3), and they concluded that the maximum temperatures (from 300 to 380ºC) were reached in the Doldenhorn Nappe during the development of the recumbent folds.

Another example of well-established physical conditions during the formation of large recumbent folds is the Mondoñedo nappe unit (Westasturian-Leonese Zone, NW Spain). This unit consists of a stack of kilometer-scale recumbent folds with an associated axial plane foliation (Fig. 4). During the development of the large recumbent folds a medium *P/T* ratio metamorphism increased downwards and towards the root zone from greenschist facies to amphibolite facies (Bastida et al., 1986; Arenas and Martínez-Catalán, 2003) with syn-tectonic crystallization of biotite, almandine and staurolite in the highest grade zone. The metamorphism continued after the development of recumbent folds with a decrease of the *P/T* ratio and replacement of staurolite by andalusite. As a consequence of the rock softening produced by this change, a ductile kilometer-scale shear zone developed in the basal part of the unit with formation of mylonites and small-scale recumbent folds with the geometry of sheath folds in some cases. The evolution continued with the development of a basal thrust and general retrograde metamorphic conditions.

In addition to the cases cited above, there are many examples of fold nappes or small recumbent folds developed under greenschist facies conditions. For example, recrystallization in the biotite and garnet zones during development of the Tay Nappe (Grampian Highlands, Scotland) was suggested by Roberts (1974). The pelitic rocks of the Kishorn Nappe (northwest Scotland) contain chlorite and white mica, and the structure was developed under low greenschist facies conditions (Coward and Whalley, 1979; Potts, 1983, and references therein). The Courel kilometer-scale recumbent fold (Iberian Massif, NW Spain) also developed under greenschist facies conditions (chlorite zone) (Barrera et al. 1989). Michard et al. (2008) described small recumbent folds developed in Cambrian-Ordovician metagreywackes and metapelites of the Sehoul Block (Moroccan Palaeozoic massifs) with associated recrystallization in greenschist facies. Similarly, Chalouan et al. (2008) describe small recumbent folds developed in the Ketama Mesozoic rocks (Morocco) under low greenschist facies conditions, at temperatures between 200 and 300ºC and pressures close to 300 MPa.

Descriptions also exist of recumbent folds formed under lower metamorphic grades. An example is the recumbent folding of the Pardailhan Nappe (Montagne Noire, French Massif Central), which developed in a mainly anchizonal area, with an associated axial plane cleavage (Echtler, 1990).

Recumbent folds also formed in many areas under metamorphic conditions higher than those of the greenschist facies. Classical examples of recumbent folds developed in high grade rocks are described by Wenk (1956) and Haller (1956a and b, 1957) in the fjords of eastern Greenland and by Michot (1956, 1957) in the Caledonides of Norway. A summary of these studies can be found in de Sitter (1958). Another example is found in the Cabo Ortegal Complex (NW Spain) (Marcos, 1998; Marcos et al., 2002). This is an allochthonous unit in the Variscan belt with mafic and ultramafic rocks where large near-isoclinal recumbent folds formed in amphibolite facies conditions have been described. These authors suggest that folds were basically developed during the exhumation of the Cabo Ortegal Nappe from higher grade metamorphic conditions. Windley and Garde (2009) described another good example in the Archean craton of West Greenland, where kilometer-scale isoclinal recumbent folds deformed orthogneisses and mafic and ultramafic rocks in granulite facies conditions. In a different tectonic setting, Orozco et al. (2004) described kilometer-scale recumbent folds in the Alpujárrides nappes (Betic Cordillera, SE Spain) developed under an extensional regime and high grade metamorphic conditions, at temperatures between 600 and 700ºC and decreasing pressure.

Recumbent folds developed under conditions of middle- or high-grade metamorphism in relation with channel flow processes in the middle-lower crust have been described in several orogenic belts; some relevant examples are considered below. In the Appalachian Inner Piedmont, Hatcher and Merschat (2006) described an imbricate stack of fold nappes formed in dominantly sillimanite-zone conditions. In this area, transpressive subduction to > 15 km depth initiated partial melting forcing escape from the collision zone in an along-strike orogenic channel. In the Shuswap Complex (southern Canadian Cordillera), Kuiper et al. (2006) described four structural domains with different attitudes and vergence of folds. In the three lower domains, recumbent folds formed under amphibolite facies conditions and exhibit vergence changes consistent with the model of channel flow deformation described by Williams et al. (2006). In the Greater Himalayan Sequence of the Western Himalaya, Searle et al. (2007) described a recumbently folded isograd pattern affecting a sequence of migmatites, leucogranite sheets and metamorphic rocks with up to kyanite or sillimanite + K-feldspar grade. This deformation is associated with a channel flow episode that produced extrusion of a mid-crustal layer of Indian Plate rocks during the Miocene. This episode postdates formation of major fold nappes during the late Eocene in condition of peak kyanite grade metamorphism.

A particular case of recumbent folds are those developed in extremely ductile rocks, such as ice or salt. These folds are formed under surficial conditions, even at temperatures below 0ºC. Both rock types show a non-linear viscous rheological behavior and flow under their own weight near to the Earth´s surface. Irregularities in the flow can give rise to tight or isoclinal recumbent folds of sub-similar geometry at different scales (e.g. Hudleston, 1976, 1977, 1983; Talbot, 1979, 1998; Hambrey and Lawson, 2000; Talbot and Aftabi, 2004; Hooke, 2005; Hudec and Jackson, 2006, 2007; Evans, 2007). These folds sometimes have curved hinges. The best examples of recumbent folds in salt are found in the Zagros

Mountains of Iran (Fig. 5) (e.g., Talbot, 1979, 1998; Talbot et al., 2000; Talbot and Aftabi, 2004). The recumbent folds developed in glacier ice and salt are not part of the internal structure of orogenic belts, but their present-day formation enables them to be used as analogues of recumbent folds in orogens (Hudleston, 1977; Talbot and Aftabi, 2004).

2.2. Strain distribution within recumbent folds

Knowledge of the strain state in folded layers is very important for helping unravel folding kinematics. Unfortunately, the common lack of adequate strain indicators often makes analysis difficult. Hence, the strain estimates have been mainly made in large recumbent folds developed in very low- or low-grade metamorphic conditions, where the strain markers are better preserved.

In the western Helvetic nappes Ramsay (1981), Siddans (1983), Dietrich (1989) and Casey and Dietrich (1997) estimated the total strain using several types of strain markers (ooids, fossils, pebbles, pressure shadows, etc.) and found a great variation in the obtained values (Fig. 2). In general, the X/Z axial ratio of the strain ellipsoid increases downwards and towards the root zone in the stack of the nappes. In the Morcles Nappe, Ramsay (1981) observed intense strain in the lower limb, with $X/Z > 100$ in some cases, whereas Siddans (1983) obtained lower values. The upper limb shows very low values of strain, with $X/Z < 4$ in most cases. The lower limb is cut by a thrust and its footwall appears nearly undeformed (Ramsay, 1981) or moderately deformed (Siddans, 1983; Casey and Dietrich, 1997). In agreement with the strain distribution, a well-developed slaty cleavage appears in the lower limb, whereas in the upper limb the cleavage has a weak development. Assuming constant volume, Casey and Dietrich (1997) observe cleavage when the maximum principal stretch (major semi-axis of the strain ellipsoid), $\sqrt{\lambda_1} > 1.5$.

Azor et al. (1994a) made strain measurements using R_f/ϕ (Ramsay, 1967; Dunnet, 1967; Ramsay and Huber, 1983; Lisle, 1985) and Fry (Fry, 1979) methods in the upper Precambrian and Ordovician rocks deformed by the Hornachos recumbent fold (Central Iberian Zone, Spain) (Fig. 6). The *X*-axis of the strain ellipsoid is parallel to the hinges of the recumbent folds and coincides approximately with the stretching lineation observed on the foliation surfaces associated with these folds. The measured strain ratios are slightly higher on the inverted limb of the recumbent fold (Fig. 6).

In the kilometer-scale Courel recumbent fold of the Iberian Massif, NW Spain (Fig. 7), Fernández et al. (2007) analysed the kinematic mechanisms of nine parasitic folds, located at or near the hinge zone of the major fold, using the program *FoldModeler*, which is a two-dimensional numerical simulator of folds formed by a controlled superposition of folding mechanisms (Bobillo-Ares et al., 2004). The modeled finite *Y/Z* ratios obtained on the fold profile ranges between 1.9 and 2.8 in the outer arc of the hinge zone, and in general between 3 and 6.3 in the hinge zone's inner arc. In order to obtain the amount of flattening suffered by the major fold, the method developed by Srivastava & Shah (2006) was applied; the obtained value for the flattening *Y/Z* ratio was of about 5. Microstructural methods based on the shape of the grains and centre-to-centre techniques were also used and they gave surprisingly low values. Fernández et al. (2007) assert that these differences could be due to the existence of deformation mechanisms that are not registered by these methods.

The stretching lineation (*X*-direction) has variable orientations with respect to the recumbent fold hinge, but it is noteworthy that in many cases this lineation is approximately parallel to the corresponding fold hinge. For example this is so in the case described by Williams and Jiang (2005) in recumbent folds of the Monashee Complex (Canadian Cordillera) developed in a regime of high strain by simple shear that produced a rotation of the fold hinges towards the stretching direction. Another case is described by Azor et al.

(1994a and b) and Martínez-Poyatos et al. (1995, 1998) in recumbent folds of the southern part of the Central-Iberian Zone (Iberian Massif, Spain). In this case, the authors conclude that the stretching lineation represents the transport direction in a transpressive context with an important component of simple shear. Fernández et al. (2007) have also found stretching lineation parallel to the fold hinge in the Courel recumbent syncline (Iberian Massif, NW Spain), but they consider that this is incompatible with a simple shear regime and interpret it to be the result of a dominant coaxial flattening with sub-vertical maximum shortening that occurred during the advanced stages of the recumbent folding.

2.3. Kinematics of recumbent folds: simple shear versus coaxial deformation

Simple shear deformation has been invoked in many ways for the formation of recumbent folds. From a historical perspective, the first author to describe the development of a fold nappe with a basal thrust was Heim (1878). According to him these folds formed as a result of a sub-horizontal shearing acting on the overturned limb of a fold that finally breaks forming the thrust (Fig. 8). This general model of evolution, proposed initially to explain the development of Alpine fold nappes, has continued to be used long after to explain the development of fold nappes elsewhere; e.g., Potts (1983) in the case of fold nappes at Kishorn, Scotland, and at Mellene, Norway.

Theoretical and experimental analyses made by several authors indicate that recumbent folds can be produced by buckling in a regime of simple shear (Ghosh, 1966; Manz and Wickham, 1978; Sanderson, 1979; Ez, 2000; Carreras et al., 2005; Llorens et al., 2013). This mechanism requires a near-horizontal shear plane and a layering dipping in the top shear direction (see Fig. 20 below). As the initial dip angle decreases, a greater amount of simple shear, and consequently more shortening, is required to form recumbent folds. As simple shear progresses, buckling can be followed by a passive amplification due to strain superposition (kinematic amplification) (Carreras et al., 2005; Aller et al. 2010).

Ramsay et al. (1983) proposed a simple shear origin for large recumbent folds in the Helvetic nappes; these authors suggested that the folds formed in the frontal culmination wall of the nappe, where the layering had reached a substantial dip before the development of the recumbent folds (Fig. 9). Hence, the folds were formed by buckling and kinematic amplification in a simple shear regime. Subsequently, Rowan & Kligfield (1992) proposed that the folds of the Wildhorn Nappe (Helvetic nappes) formed by simple shear in horizontal layers between two gently dipping thrusts. Deformation in a simple shear regime has been also invoked to explain outcrop-scale recumbent folds, often related to ductile shear zones or thrusts (Sanderson, 1982; Rattey and Sanderson, 1982; Williams and Jiang, 2005; Bastida et al., 2010).

In the case of recumbent folds developed in salt, Talbot (1998) and Talbot and Aftabi (2004) propose a mechanism for salt extrusions in the Zagros Mountains in which they are formed by an advance of the frontal part of the salt glacier like a caterpillar tank track. The top of the salt body moves faster than the bottom surface, which attaches to the bedrock. Hence, the top rolls over the advancing flow front. This mechanism was simulated by experiments producing the lateral compression of a circular body of a Newtonian-viscous polymer (polydimethylsiloxane) with passive coloured markers layers (Talbot and Aftabi, 2004).

Hudleston (1976, 1977, 1983) studied the strain, flow and development of recumbent folds in the Barnes Ice Cap (Baffin Island, Canada). He concluded that the ice flows essentially by shearing and that the recumbent folds in glacier ice can form by superposition of homogeneous simple shear on an initially undulatory layering. This author also suggested that recumbent folds in orogenic belts may form in an analogous manner. In fact, a similar mechanism has been suggested by Williams and Jiang (2005) for shear zones developed in high grade metamorphic conditions.

Ez (2000) and Aller et al. (2010) remark that the folds obtained by the mechanism proposed by Hudleston (1977) require an amount of superposed simple shear strain, γ, greater than 5 to produce tight or isoclinal recumbent folds similar to those common in orogenic belts. This can be achieved in ductile shear zones developed in the middle to lower crust, but is more difficult at orogenic-scale in low-grade metamorphic zones, where these structures also occur. Several authors have proposed that a coaxial strain with maximum shortening in a vertical direction could be very important in the development of recumbent folds. According to Ez (2000) and Aller et al. (2010), coaxial deformation produces rotation of an inclined axial surface towards a recumbent attitude with less strain than simple shear; therefore, coaxial deformation requires less work than simple shear to produce recumbent folds. Hence, Ez (2000) asserts that a probable cause for the development of almost recumbent isoclinal folds may be a coaxial strain with vertical maximum shortening of inclined folds with an overturned limb. Fernández et al. (2007) suggest that the kilometer-scale recumbent fold of the Courel (NW Spain) was a result of the following evolution (Fig. 10): (a) development of a supratenuous rollover anticline, related to a growth fault; (b) tectonic uplift of the southern part of the section giving rise to an overall dip of the layers towards the foreland; (c) minor layer shortening, buckling and rotation in a simple shear regime with a shear direction foreland directed and gently plunging towards the hinterland; (d) flattening with the maximum shortening direction steeply inclined, and involving an important area decrease in the profile plane and stretching in the axial direction, and (e) superposition of gentle folding and rotation of the recumbent folds to the present position.

To explain the geometry, strain pattern and metamorphism of the Helvetic nappes (Fig. 2), Dietrich and Casey (1989) proposed that these nappes are the result of the superposition of simple shear and pure shear. This strain model is produced by a tectonic extrusion from the root zone. The pure shear part of the deformation is of variable magnitude; it decreases towards the external zone of the orogenic belt, in a direction perpendicular to the fold hinge, and it involves a positive vertical extension in the frontal part of the nappes and a vertical shortening in the internal part. The nappe is thus comparable with a concentric fold with a neutral surface. The simple shear is heterogeneous vertically, although constant from the internal to the external zones. At the level of maximum simple shear, a thrust appears in the external part of the nappe and ductile deformation in the internal part. Subsequently, the same authors (Casey and Dietrich, 1997) gave a different explanation for the development of the Helvetic nappes. Specifically, they propose the following explanation for the development of the Morcles and Diablerets nappes (Fig. 11): a) a first stage of layer-parallel shortening and buckling with development of gentle folds in an overall context of pure shear; b) amplification of the folds as a result of the development of a shear zone by heterogeneous simple shear with a shear direction (direction of the relative displacement of the top of the shear zone) oblique to the middle surface of the folds (the shear zone includes the rocks that will become the Morcles and Diablerets nappes); c) thinning of the shear zone and development of the basal thrust of the Diablerets Nappe; d) continuation of the development of the shear zone affecting the rocks of the Morcles Nappe, and e) development of its basal thrust.

Numerical and experimental models producing recumbent folds with complex strain distributions have been developed by several authors (e.g. Bucher, 1956; Brun and Merle, 1988; Vacas-Peña and Martínez Catalán, 2004; Beaumont et al., 2006). In these models, the existence in a deep level of an obstacle that hinders the propagation of the deformation enables the formation of recumbent folds.

2.4. The role of gravitational forces

In buckling theory gravitational forces are often neglected. This assumption might be reasonable when describing the mechanics of small folds, but not so in the case of large folds where gravitational forces are always important. De Sitter (1954) asserted that gravitational forces always influence the development of structures. Gravity can contribute to the development of recumbent folds enabling gravitational sliding or gravitational spreading. De Sitter (1958) states that gravitational sliding was active in the case of many folds formed in shallow depths, as for example in slump structures. According to this author, a recumbent fold can be considered to be due to gravity when there is a sloping plane along which the slipping has occurred. Shackleton (1969) asserted that many deformations, including most large recumbent folds, are probably the result of gravitational sliding and they do not involve shortening of the crust.

Lateral gravitational spreading involves ductile flow and lateral expansion of material under the vertical stress due to its own weight. It has also been invoked in the development of recumbent folds in glacier ice or salt under shallow conditions by some authors (e.g., Hudleston, 1976, 1977, 1983, Talbot, 1979, 1981; Cooper, 1981; Hudec and Jackson, 2006, 2007).

Gravitational sliding and lateral gravitational spreading can be difficult to distinguish and it is possible that both work together in many cases, so that the term "gravitational flow" has commonly a very general meaning. Cummins and Shackleton (1955) explained the Tay Nappe (Grampian Highlands of Scotland) as the result of "deep-reached gravitational flow", and Read and Farqhuar (1956) consider that the recumbent folds of the Banff Nappe (NE Scotland) were due to flow toward the lower part of the limb of a bulge.

In some cases, authors have considered that gravitational flow that gives rise to the recumbent attitude of the folds is produced at an advanced stage of evolution of the folding. For example, Echtler (1990) concluded that the Pardailhan Nappe (Montagne Noire, France) evolved firstly by development of overturned folds over a thrust under the main action of tectonic stresses, then a part of the structure was uplifted, and finally the folds were rotated along extensional faults under the main action of gravitational forces (Fig. 12). Likewise, Aerden and Malavielle (1999) described the development of a fold nappe in the Montagne Noire by gravitational collapse from initially upright folds.

Ez (2000) proposed vertical compression of overturned folds or bucking of vertical layers as common causes for the development of recumbent folds. In order to explain the kinematics of the kilometer-scale recumbent fold of the Courel (NW Spain), Fernández et al. (2007) suggest the operation of sub-vertical shortening due to gravity at advanced stages of its evolution, as a consequence of the tectonic superposition due to folding.

Several authors have investigated the influence of gravity on the development of recumbent folds by physical experiments or by the finite element method. Bucher (1956, 1962) experimentally produced recumbent folds in models with layers of several types of stitching wax subjected to compression. In a series of experiments these folds developed in the warmer (and therefore weaker) part of the model. In other experiments, the stiffer part of the models was replaced by a wooden strut in the basal part of the model. This strut operated as a buttress that facilitated development of a recumbent anticline with a greatly thinned lower limb (see Fig. 25b below). According to this author, the recumbent character of the folds was produced by flattening of the material under its own weight. Similar results were obtained by Vacas Peña and Martínez-Catalán (2004) using the finite element method with viscous layers and a rigid block in a part of the base of the model (Fig. 13). These authors remark that the inclusion of gravity forces in the model was a necessary requirement for the development of recumbent folds.

Brun and Merle (1988) made experiments with a viscous slab (silicone putty) flowing under its own weight over an inclined ramp with basal perturbations consisting of transversal plasticine ridges. As a result, isoclinal cylindrical recumbent folds were obtained. The overturned limbs appeared greatly thinned and the geometry of folds was nearly similar (Class 2).

2.5. Tectonic context of recumbent folds: compressional regime versus extensional regime

Large recumbent folds occur in most orogenic belts, and they are associated usually with compressive tectonic regimes. However, several authors have described development of recumbent folds under extensional conditions.

Froitzheim (1992) described recumbent folds formed during syn-orogenic extension in the Australpine nappes (Switzerland). According to this author, the folds formed by vertical shortening and horizontal stretching. This coaxial deformation could be combined with noncoaxial deformation developed by simple shear in a shear zone dipping in the same directional sense as the layers, but dipping less than the latter. Harris et al. (2002) have described the development of recumbent folds by simple shear in extensional ductile shear zones similar to those described above. No significant differences in the geometry of the folded layers are observed between recumbent folds formed in extensional shear zones and those formed in compressional shear zones, the attitude of the shear zone being the main difference (Fig. 14).

In the Betic Cordillera, Orozco et al. (1998, 2004) found a close relation between recumbent folds and extensional faults and concluded that the folds developed during an extensional event. In other cases, an extensional event has given rise to the recumbent attitude after the development of the folds, by rotation during the displacement along normal faults. This is also the case in the Pardailhan Nappe (Montagne Noire, France; Fig. 12) (Echtler, 1990).

2.6. Structural associations related to recumbent folds

Recumbent folds usually present several types of associated structures that have great relevance to understanding their kinematics. These folds are usually isoclinal or have very low interlimb angles. Hence, tectonic foliation generally appears associated with them. However, there are cases in which this foliation is not present or is poorly developed. An example appears in the Cabo Ortegal Complex (Variscan belt of NW Spain) (Marcos, 1998; Marcos et al., 2002), where recumbent folds, developed in the amphibolite facies conditions, deformed a mylonitic foliation whilst a foliation associated with the folds is not observed in the field, except for a weak preferred orientation of hornblende (in amphibolites) or phyllosilicates (in gneisses) in some hinge zones.

The association of large recumbent folds and thrusts is so typical that it commonly appears in the definition of fold-nappe (for example, Dennis et al., 1981). Classical examples of this association are the Helvetic nappes, in which a thrust cuts the overturned limb of the recumbent folds. Many similar examples can be found in other areas. More complex associations of recumbent folds cut by thrusts have been described in fault-propagation folds developed in the foreland of orogenic belts (e.g. Storti and Salvini, 1996).

A very common structural association consists of large recumbent folds deformed by upright folds whose hinges are usually parallel or nearly parallel to those of the recumbent folds. This association can appear in several forms. Often, the recumbent folds are deformed in a large antiform or dome. This is, for example, the case of the Helvetic nappes of Morcles, Diablerets and Wildhorn (Ramsay, 1981) (Fig. 15). Another good example is the Pardailhan Nappe (Montagne Noire, France), which is a large recumbent anticline deformed by a major antiform in the axial zone of this part of the Variscan belt (Harris et al., 1983; Echtler, 1990). A cross section through the Ossa-Morena Zone (Iberian Massif, Spain) also exhibits a good example of large recumbent folds deformed by a large antiform in whose core an imbricate thrust system appears (Simancas et al., 2003, 2004) (Fig. 16). More examples can be found in the Caledonides of Ireland and Scotland (e. g. the Connemara antiform and the Tay Nappe in NW Mayo). In some cases, the large upright fold that deforms the large recumbent fold is a synform. This is, for example, the case in the Cabo Ortegal Complex (NW Spain) (Fig. 17).

Another common form of structure consists of a pair of large upright folds (antiform and synform) deforming recumbent folds. A good example of this situation is the Mondoñedo Nappe (NW Spain) (Fig. 4). Here the antiform has been interpreted as a culmination due to an antiformal stack of thrusts sheets, a stacking of folds, a ramp associated with a later out-of-sequence thrust, or a combination of these structures (Pérez-Estaún et al., 1991).

Cases in which superimposed folding consisting of several upright folds deform recumbent folds are also common. Examples have been described in the Central Iberian Zone (Iberian Massif, Spain) (Azor et al., 1994a; Martínez Poyatos et al., 1995, 1998), in the Montagne Noire (Aerden and Malavieille, 1999), and in the Alborán domain (Betic Cordilleras, Spain) (Orozco et al., 2004).

3. Analysis and discussion

The wide variety of geological settings in which recumbent folds occur within the orogenic belts suggests that the mechanics of these structures is controlled by a multiplicity of factors, so that it is not possible to explain the development of all types of recumbent folds present in rocks using a single model. However, it is possible to analyze and discuss the different ideas on this subject and establish some constraints on the development of these folds.

3.1. A starting point: Heim's model

The simple model described by Heim (1878) attempts to explain the development of a thrust along the overturned limb of a recumbent anticline (Fig. 7). We can consider three stages in this model: a) development of the initial folds (an anticlinal-synclinal couple); b) shifting apart of the two hinges with stretching and thinning of the overturned limb; and c) failure and development of a thrust. The model does not detail the mechanisms or the strain distribution associated with these stages and it was developed to explain the kinematics of great scale structures, such the Alpine fold nappes. The evolution described by the model has been proposed to explain some specific structures and the question that arises concerns its general validity when applied to other examples. The increase of strain, development of cleavage and metamorphism towards the basal part of the Helvetic fold nappes found by Ramsay (1981), Siddans (1983) and Casey and Dietrich (1997) agrees with the Heim's model, but the increase of these features towards the root zone does not agree with this model. Hence, the models to explain the development of these nappes presented by Ramsay et al. (1983), Dietrich and Casey (1989) and Casey and Dietrich (1997) differ from the Heim's model.

The geometrical implications of Heim's model are difficult to analyze in large recumbent folds; it is easier to check them in small-scale folds. Fig. 18 shows two pairs of small recumbent folds developed in metapsammites and phyllites under metamorphic conditions of the biotite zone. The overturned limb of the lower pair of folds is cut by a thrust and the general geometry is comparable with that resulting in the final stage of the Heim's model. In contrast, the upper pair of folds in Fig. 18 shows the overturned limb thickened with respect to the normal limbs. This thickening is not observed in the evolutionary stages of Fig. 8.

Fig. 19 shows a near-isoclinal recumbent anticline-syncline pair developed in a ductile shear zone under metamorphic conditions of the staurolite zone. These folds have a thickened overturned limb and therefore their geometry is different to that found in the evolutionary stages of Heim's model.

Observation of minor folds like those in the above mentioned examples suggests that the development of the evolutonary stages shown in Fig. 8 depends on the lithological characteristics of the multilayer and the physical conditions of the deformation. In sedimentary multilayers with marked competence contrast under conditions of the diagenesismetamorphism transition or low grade metamorphism and affected by a bulk non-coaxial deformation regime (e.g., simple or sub-simple shear), the competent layers can exhibit active behavior and undergo buckling with development of an overturned limb. Later, this limb can suffer a nearly homogeneous stretching and thinning until eventual failure in the context of the mentioned non-coaxial deformation. In some of these cases, like the fold nappes at Kishorn, Scotland, and Mellene (Norway), Heim's model may approximate geological reality (Potts, 1983).

In the case of a metasedimentary multilayer folded under high temperature conditions, the rock ductility is high and competence contrasts are low. In such circumstances the folding may initiate by a flow perturbation, so that the effect of buckling is limited and the folding grows by kinematic amplification in the context of a non-coaxial deformation. In this case, the overturned limb can appear thickened or thinned depending on the geometry of the initial perturbation and the final orientation of the limb in the shortening or stretching fields respectively of the finite strain ellipse in the fold profile. If, for example, the recumbent folds developed in the context of progressive simple shear parallel to bedding in the normal limbs and the initial irregularity is gentle, the forelimb will pass firstly through a stage of thickening. When this limb becomes overturned it enters the field of incremental thinning, although it will remain thickened until it enters the field of the finite thinning (Ramsay, 1980). If the temperature is high and the behavior is very ductile, the development of a thrust along the overturned limb can be very difficult. As we will see below, the non-coaxial deformation involved in the development of most recumbent folds probably has simple shear and pure shear components. The latter are not considered in Heim's model.

3.2. Whole-rock strain and recumbent folds

With the exception of the Helvetic fold nappes, the information obtained from strain estimates in recumbent folds is not sufficient to infer the kinematic mechanisms responsible for developing these folds. Some methods used to measure the strain give very low values of the strain ratio in the fold profile plane. This is due to these methods not taking into account some deformation mechanisms of the rocks, such as grain boundary sliding. Hence, in order to infer these mechanisms it is only possible to use the information provided by the fold geometry and the associated minor structures. This information must be checked with analogue or numerical models in order to shed light on the mechanics of recumbent folds.

Simple shear plays a very important role in the development of recumbent folds. It is the simplest strain mode that is able to explain the facing of recumbent folds and does not produce strain incompatibility with adjacent undeformed areas. In addition, superposition of a coaxial strain with a vertical direction of maximum shortening or a combination of simple shear and coaxial strain can favor the development of recumbent folds, because it enables the recumbent attitude to be achieved with less strain than simple shear as the only mechanism (Fernández et al., 2007; Aller et al., 2010). The coaxial strain component must increase with time during the folding process, since its contribution in the first stages would prevent the development of buckling except in cases with high initial dip of the multilayer. Thus, simple shear probably dominates in the first stages of the recumbent folding, whereas the coaxial part would become more important in the later stages as a consequence of the increase of the lithostatic stress due to the tectonic superposition associated with the folding development.

Assuming a simple shear regime on layers dipping in the top shear direction or in a direction opposite to it, with an obliquity between the layering and the shear plane so that the vertex of this obliquity angle points opposite to the sense of shear, we infer that the development of recumbent folds in layers with a specific rheology mainly depends on three factors: a) the initial obliquity angle (θ_0) (Fig. 20), b) the plunge of the shear direction, and c) the amount of simple shear (γ). Fig. 21 shows the variation of shortening of the layer with the

γ value for different angles θ_0. The curves have been obtained using the equation (Malvern, 1969, p. 164):

$$\lambda_{\mathbf{N}} = N^{T}CN, \qquad (1)$$

where λ_N is the quadratic elongation for the considered initial unit vector represented by the matrix N, N^T is the corresponding transpose, and C is the matrix of the Green deformation tensor for simple shear. The curves have a maximum whose value increases as θ_0 decreases. Likewise, as θ_0 decreases, the γ value of the maximum increases. On the other hand, the layers rotate progressively as γ increases, so that this maximum shortening is reached when the envelope surfaces of the folds becomes perpendicular to the shear direction. In this position, if shortening gives rise to folding, recumbent folds will develop in competent layers when the shear direction is sub-horizontal. When θ_0 is low, the shortening corresponding to this position and the γ value necessary to reach it are high (for instance, if θ_0 = 10º, shortening ≈ 83% and γ ≈ 5.7). It seems to be difficult to achieve these values in the development of large recumbent folds in low-grade metamorphic areas, where these structures are common, although they are possible in the case of ductile shear zones or high grade metamorphic rocks (Williams and Jiang, 2005). If θ_0 is much higher, the value of γ needed to reach the perpendicular orientation is lower, but the maximum shortening has low values, and the development of tight recumbent folds is improbable (for instance, if θ_0 = 40º, shortening ≈ 36% and γ ≈ 1,3). In addition, high values of θ_0 are geologically unrealistic in most cases. Nevertheless, in agreement with the results of Llorens et al. (2012), overturned folds can be generated with shortening >50% and final obliquity angle (θ, dip of the median surface of the folds; Fig. 21) ≥40º. These folds involve lower γ values than those developed when θ = 90º, and they can be later tightened and converted into recumbent folds by flattening with vertical maximum shortening. From these considerations, a simple shear regime with θ_0 values between 15° and 20º could give rise to tight or close overturned or recumbent folds with γ values between 1 and 3.5, interval that can be common in the development of large recumbent folds.

For γ values higher than those corresponding to the maximum of the curves in Fig. 21, an interval of incremental stretching begins and the folds undergo an unfolding episode, in which a local change in the asymmetry of the folds could occur, with normal limbs longer than the overturned limbs. This situation is rare in nature. On the other hand, a plunge of the shear direction opposite to the direction of the relative displacements of the top of the shear zone gives rise to a piling up of rocks and a tectonic superposition during folding that, if not balanced by rapid erosion, would produce an increase of the lithostatic stress.

A question arises when natural recumbent folds are compared with those produced or simulated by simple shear of competent layers using analogue modelling or numerical methods. The latter develop by buckling and rotation and are nearly symmetric (Ghosh, 1966, Manz and Wickham, 1978, Carreras et al. 2005; Llorens et al. 2012), whereas the former are usually asymmetric folds. The characteristics of the initial irregularities of the layers could explain this difference. Differently to the models, in the natural multilayers where large recumbent folds developed, buckling probably takes place by amplification of a few initial irregularities with appreciable initial amplitude or by the influence of other perturbations prior to folding. In the case of recumbent folds originated in ductile shear zones under high P and T conditions, initial irregularities would give rise to flow perturbations whose amplification is a geometrical effect of a strain superposition (kinematic amplification), as suggested by Hudleston (1977) for recumbent folds formed in glacier ice. In these cases, the buckling plays at most a secondary role.

There are many ways to model geometrically the late evolution of recumbent folds in which the coaxial part of the strain with vertical maximum shortening dominates over the simple shear component. Two possibilities are: a) superposition of pure shear on the simple shear, or b) superposition of a combination of simple shear and simultaneous pure shear with progressive increase of the components of pure shear (sub-simple shear). In addition it is also possible to include a volume change in the coaxial part of the strain. Amongst such models many variants are obviously possible. All these possibilities give rise to flattening of the overturned or recumbent folds previously formed during the stage of simple shear. The flattening also involves a rotation of the median surface and the axial plane, the latter decreasing its dip. Hence, this stage enhances the recumbent character of the folds. A history like that described above successfully explains the geometry of the kilometer-scale Courel recumbent fold (Variscan Belt, NW Spain; Fig. 10) (Fernández et al., 2007).

Let us consider now the 3D case, that is to say, the possibility of the development of recumbent folds by simple shear in layers with any orientation with respect to the shear direction. According with Ramsay (1980) and Treagus and Treagus (1981), the orientation of the hinges of the folds depends on the obliquity between the layer to be folded and the shear direction, so that the axial direction must coincide with the direction of the major axis of the whole-rock finite strain ellipse of the folded layer. To obtain geometrical and strain data for any deformed plane in a context of simple shear, we have used the analysis made by Ramberg (1976) and Ramberg and Ghosh (1977). Measuring the initial dip direction α_0 of the layer from 0° to 360º clockwise from the shear direction (Fig. 22), assumed horizontal, and considering as an approximation that the axial plane is perpendicular to the median surface of the folds, we have analyzed the possibility of development of recumbent folds in layers with dip direction α_0 of 0°, 30°, 60° and 90º (for the symmetry of the simple shear model the results are equally valid for 330°, 300° and 270º respectively) and initial dips θ_0 from 0 to 30º (Fig. 23). When the dip direction is 30º, the results are comparable to the 2D case (this corresponds to the 3D case with a dip direction of 0º), in the sense that recumbent folds are possible. Based on the same argument described above for the 2D case, the most favorable dips for the formation of these folds are those of 15° to 20º. The values of γ required for the development of recumbent folds for a dip direction of 30º are somewhat higher than those required for a dip direction of 0º. For a dip direction of 60º nearly recumbent folds are only possible for a layer with an initial dip of 30º, and for a dip direction of 90º recumbent folds are not possible by simple shear for initial dips ≤ 30º. Similarly to the 2D case, superposition of a coaxial strain with vertical direction of maximum shortening on overturned or recumbent folds generated by simple shear would decrease the dip of the axial plane and would increase the recumbent character of the folds.

We have assumed above that the fold hinge has the direction of the major axis of the finite whole-rock strain ellipse in the plane of the folded layer (Ramsay, 1980; Treagus and Treagus 1981). However, another possibility is that the fold hinge forms in the direction of the whole-rock strain ellipse in the plane of the layer at a certain moment in the folding history and then rotates passively in agreement with the subsequent strain (Flinn, 1962; Sanderson, 1973; Ramsay, 1979, Jiang, 1999). In the second case, the long axis of the strain ellipse on the layer can make an angle τ with the fold axis (Fig. 22). When the initial dip direction of the layer is 0º, τ is always 0º, since the direction of the major axis of the strain ellipse on the layer does not change during folding, and it coincides with the strike of the layer and is perpendicular to the shear direction. When the initial dip direction of the layer is not 0º, assuming that fold hinges are fixed in the first increment of the deformation and then they rotate passively with the deformation, the angle τ is generally less than 6º in a context of simple shear (Fig. 24). This value is probably a maximum, since in the first stages of the deformation a layer-parallel shortening occurs before the buckling and hinge migration is probably a common mechanism in the first stages of buckling.

The existence of vertical maximum shortening involves the dominant influence of the lithostatic stress, and the increase of this shortening during folding may be related to the increase of the tectonic superposition associated with folding. The role of gravitational forces in the development of recumbent folds, as inferred from the strain models, is confirmed by analogue and finite elements models (Bucher, 1956, 1962; Brun and Merle, 1988; Vacas Peña and Martínez Catalán, 2004). The flattening of recumbent folds due to lithostatic stress explains why these folds are in many cases nearly isoclinal, and sometimes more tightened than the non-recumbent folds of adjacent areas.

Due to the complexity of geological reality, simple numerical models cannot fully explain peculiarities of many specific recumbent folds. The models used here do not consider the complex kinematic mechanisms arising from the anisotropies and rheological heterogeneities of the rocks like those considered by Bucher (1956) in his experimental models, nor the existence of fractures that can favor the development of recumbent folds. An example of this complexity is found in the Helvetic fold nappes. Each of these nappes consists of a set of folds with axial surfaces converging towards the core of the fold nappe. Hence, complex models have been proposed to explain these fold nappes (e.g., Dietrich and Casey, 1989; Casey and Dietrich, 1997). Specifically, Dietrich and Casey (1989) propose an increase of the pure shear part of the deformation towards the internal zone of the orogenic belt.

The existence of a root zone is common in large recumbent folds. Clear examples of this structure are found in the Helvetic Nappes (Ramsay, 1981), the nappes of the west Central Highlands (Roberts and Treagus, 1979; Thomas, 1979), and the recumbent folds of the Appalachian Inner Piedmont (Hatcher, 1981) (Fig. 25). The geometry of these structures suggests that a tectonic extrusion is involved in the development of the recumbent folds. In these cases, it is probable that the existence of a perturbation prior to folding, comparable to a wall buttress, hindered the normal propagation of the deformation and induced a climb up of the rocks to overcome the obstacle, involving the development of a recumbent fold to accommodate the ductile deformation to the geometry of the obstacle. The formation of recumbent folds in relation to obstacles in the bedrock has been observed in salt glaciers in the Zagros Mountains of Iran (Fig. 5) (Talbot, 1979, 1981, 1998). The control exerted by obstacles on the formation of recumbent folds is suggested by the results of experimental analysis and finite element method (Bucher, 1956, 1962; Brun and Merle, 1988; Vacas Peña and Martínez Catalán, 2004; Beaumont et al. 2006) in which the influence of gravitational forces is necessary. Comparison of a result of Bucher's experiments with a large recumbent fold of Appalachian Piedmont is very illustrative (Fig. 25). The state of strain in these folds cannot be explained by a superposition of simple kinematic mechanisms of folding.

Sometimes the folds acquire their recumbent attitude as a result of deformation occurred after their formation, by ductile deformation or by sliding along a fault. The outcrop-scale chevron folds of Millook Haven (SW England) are an example of this change in attitude. The development of these folds has been interpreted as due to a deformation in a regime of simple shear (Sanderson, 1979; Bastida et al. 2007). Comparing the geometry of these folds with folds simulated with numerical methods, the later authors suggest that they could have formed by an event of buckling plus rotation induced in a general regime of simple shear, followed by a kinematic amplification in the same regime. However this scenario would have given rise to non-recumbent overturned folds and a tilting is necessary to explain the recumbent attitude. This tilt was probably due to sliding along later extensional faults (Shackleton et al., 1982). A similar example, although on a larger scale, is the Pardailhan Nappe (Montagne Noir, France; Fig. 11) (Echtler, 1990).

Flattening involving vertical shortening and horizontal stretching can create a problem of strain compatibility at the boundaries with the upper and lower adjacent areas where the stretching disappears. Upwards the vertical shortening and the horizontal stretching can decrease progressively as the lithostatic and differential stresses decrease. This general upward decrease of the stresses would give rise to a corresponding decrease of the horizontal stretching or a maintaining of the stretching with a change to brittle deformation and development of normal faults, as suggested in the general models of Platt (1986) and Fossen and Tikoff (1997). The progressive change is more difficult downward and it is more probable that a discontinuity, such as a thrust or a ductile shear zone, is kinematically necessary to solve the strain incompatibility (e.g., Sanderson, 1982; Ramsay and Huber, 1987, pp. 612-613). In this case, the location of the lower boundary of the recumbent folds, where the thrust or the shear zone developed, will be commonly associated with a lithological and/or structural discontinuity, as for example the roof of a massive competent basement.

3.3. Stretching lineations in recumbent folds

When cleavage is developed in the context of simple shear, a stretching or mineral lineation approximately parallel to the *X*-direction can occur. If the initial dip direction of the folded layer is 0º relative to the shear direction, an angle of 90º between the stretching lineation and the fold axis could be expected. An example of this situation is found in the recumbent folds of the Mondoñedo Nappe (Variscan belt, NW Spain) (Matte, 1968).

Stretching lineations parallel to the axis of recumbent folds are very common (e.g. Harris et al., 1983; Mattauer et al., 1983; Hossack and Cooper, 1986; Escuder Viruete, 1998; Azor et al., 1994a and b; Martínez-Poyatos et al. 1995, 1998; Fernández et al., 2007; Morales et al., 2011). This parallelism is also common in non-recumbent folds (e.g. Grujic and Mancktelow, 1995 and references therein).

Both in recumbent folds and in non-recumbent folds, many authors have interpreted the stretching lineation as indicative of the movement direction (or shear direction) of the rocks in a context of simple shear (e.g. Hanmer and Passchier, 1991; Menegon et al., 2008). When this lineation is parallel to the hinges of the folds, it is interpreted that the folds were formed with their hinges in a direction which depends on the initial orientation of the layers and then they rotated with the deformation towards the *X*-axis (stretching lineation), which if the simple shear is intense ($\gamma > 10$ after Grujic and Mancktelow, 1995) is very close to the shear direction. This interpretation is probably applicable to small-scale recumbent folds developed in ductile shear zones, where high strains are involved, but not to large recumbent folds developed in areas with low metamorphism, where the strain measured is lower, with *X/Z* values usually < 5. These cases cannot be explained solely by deformation in a simple shear regime. In agreement with the kinematic mechanisms described above, the development of large recumbent folds requires, in addition to a deformation in a regime of simple shear, a superposed coaxial strain with sub-vertical maximum shortening. This coaxial component controls the strain ellipsoid shape and eventually dominates its orientation (Teyssier and Tikoff, 1999; Czeck and Hudleston, 2003). The development of a stretching lineation parallel to the fold axis is probably associated with the homogeneous strain superposed on the folds and not with the previous history of buckling. Whereas the buckling is produced during a general simple shear regime, the homogeneous deformation is mainly associated with a flattening during the coaxial strain. If this flattening has the *X*-direction parallel to the fold axis, it is possible that the stretching lineation forms parallel to this axis. This situation can be favored by a particular structural location of the area, for example in the extrados of an orocline (Ries and Shackleton, 1976). This is the case in the Courel kilometer-scale recumbent fold (Fernández et al., 2007), located in the extrados of the Iberian orocline (Variscan belt, NW Spain).

If the shear strain associated with the simple shear is high and it produces a stretching lineation, the superposition of a coaxial strain with vertical maximum shortening and *X*-direction parallel to the fold axis can give rise to a sudden switching of up to 90º in the *X*-direction at a specific moment in the development of the deformation (cf. Teyssier and Tikoff, 1999). One might think that this evolution may lead in some cases to two stretching lineations. Although this is not common, a case of coexistence of two stretching lineations has been described in the Morcles fold nappe (Helvetic nappes) (Casey and Dieterich, 1997). Lineations oblique to the axes of kilometer-scale recumbent folds are not common. Nevertheless, they can occur if the dip direction of the layers to be folded is initially oblique or perpendicular to the shear direction. More complex situations with lineations oblique to the fold axes or to the apical direction of sheath folds have been proposed by Jiang and Williams (1999) and Kuiper et al. (2007) as a result of triclinic deformation in transpressive shear zones.

3.4. Large recumbent folds: the role of tectonic and lithostatic stress

Recumbent folds are structures typical of orogenic belts, and therefore they usually develop in a compressional tectonic regime. However, they can also develop in areas with local extensional deformation within orogens. Although the development of a recumbent fold is obviously a progressive process, we can distinguish in general two stages in its evolution:

1) In the first stage the tectonic stress predominates over the lithostatic stress, and the buckling or flow perturbation of rocks takes place dominantly in a regime of simple shear. We can distinguish three main cases:

a) The shear direction plunges gently towards the hinterland in a direction opposite to the relative displacements on the top of the shear zone. It involves tectonic superposition of rocks and tangential shortening. This seems to be the usual situation in orogenic belts.

b) The shear direction is horizontal. In this case, the horizontal direction does not undergo finite longitudinal strain and neither tectonic superposition nor denudation is produced.

c) The shear direction plunges towards the foreland in the same direction of the relative displacements. It involves tangential stretching, which can lead to tectonic denudation. This can occur locally in orogenic belts.

2) The second stage is a consequence of the tectonic superposition (case ‘a’ above), which increases the contribution of the lithostatic stress to the development of recumbent folds. It involves flattening with vertical maximum shortening and horizontal extension, and enhances the recumbent character of the folds. Since it occurs in the compressive context of an orogenic belt and involves a strain incompatibility with the substrate, the extensional deformation of this stage must be accommodated by compressional structures, such as a basal ductile shear zone or a thrust, which usually translate the body of folded rocks towards the foreland. Therefore, the last stages of the development of many large recumbent folds involve a ductile vertical shortening and horizontal stretching driven by lithostatic stress and a rock translation driven by tectonic stress. This evolutionary model offers an explanation for the common association of large recumbent folds and thrusts.

4. Conclusions

Recumbent folds are key structures in the development of orogenic belts. They occur in a variety of geological settings and have different sizes and geometry. Hence, it is not surprising that they form by several different mechanisms giving rise to different strain patterns in the folded layers. In all cases, a common factor controlling the location of these folds must be the existence of initial perturbations in the multilayer that also explain the asymmetry that these folds usually have. These can be deviations from a planar form of the layers, the existence of faults or the presence of an obstacle (e.g. existence of a core or basement of competent massive rock) that hinders the propagation of the deformation. Consequently, the development of a large recumbent fold at the orogenic scale is frequently due to the existence of a mechanical contrast between the involved multilayer and the adjacent rocks. This situation can explain the existence of a root zone in many large recumbent folds, since the folding multilayer must climb up to continue the propagation of the deformation.

In areas deformed under conditions of very low grade of metamorphism, the competent layers can behave in a mechanically active manner during folding and undergo buckling with development of an overturned limb and subsequent stretching, thinning and failure of this limb to give rise to a fold nappe with a basal thrust (Heim's model).

In areas deformed under high *T* and *P* conditions, the rocks can undergo high ductile deformation and recumbent folds can develop by flow perturbations, with little or no intervention of buckling. In these cases, an important mechanism is the kinematic amplification associated with nearly homogeneous strain with components of simple shear and coaxial strain; the latter mainly operates in the last stages of the folding. This can be the formation mechanism for recumbent folds common in ductile shear zones.

In areas with low-grade metamorphism, where large recumbent folds are common, buckling under a simple shear regime, with a sub-horizontal or gently inclined shear direction, is probably important in the first stages of folding. In these cases, the progressive decrease in the dip of the axial plane to produce recumbent folds requires in a first stage a null or low initial angle between the shear direction and the dip direction of the layers and a low obliquity angle between these and the shear direction. Subsequently, a kinematic amplification with components of simple shear and coaxial strain, involving a vertical maximum shortening, is required. The latter components are important in the last stages of folding, when the lithostatic stress associated with the tectonic superposition is high. Sub-horizontal stretching due to the coaxial components can generate problems of strain compatibility in the basal boundary of the deformed rocks and give rise to a basal thrust producing an important displacement towards the foreland.

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FIGURE CAPTIONS

Figure 1. Model of crustal evolution developed by Fyson (1971) in which two structural levels have been distinguished (infrastructure and suprastructure). In the stage (a) recumbent folds develop in the infrastructure, whereas in a subsequent stage (b) upright folds develop in the suprastructure and in the upper part of the infrastructure. In this the upright folds have lower amplitude and are superimposed on recumbent folds.

Figure 2. Synthetic cross section through the western Helvetic nappes (after Casey and Dietrich, 1997; modified from Dietrich and Casey, 1989) showing strain ellipses (*XZ* sections of the strain ellipsoid) taken from the above authors and Ramsay and Huber (1983).

Figure 3. Tectonothermal evolution of the Helvetic nappes of Doldenhorn (Do), Gellihorn and Wildhorn (Wi). The left column shows evolutive geological section of the nappes, and the right column shows the corresponding thermal zonation. Aa, Aar massif. After Herwegh and Pfiffner (2005).

Figure 4. Cross section through the northern part of the Mondoñedo nappe (Westastur-leonese Zone, Variscan belt of NW Spain) (after Bastida et al. 1986). VF = Viveiro fault.

Figure 5. Idealized section showing the fold trains developed in the northern salt glacier of Kuh-e-Namak (southern Iran) and associated with irregularities in the bedrock (after Talbot, 1979).

Figure 6. Structural schemes of the Hornachos recumbent fold (southern border of the Central Iberian Zone, Variscan belt of SW Spain) showing the ratios *X/Y* (left) and *X/Z* (right) of the strain ellipsoid at several points of the structure (after Azor et al. 1994a).

Figure 7. Cross section through the kilometer-scale recumbent fold of the Courel (Variscan belt of NW Spain) (After Fernández et al., 2007).

Figure 8. Heim's model to illustrate the development of a fold nappe (after Heim, 1878; in Leith, 1913).

Figure 9. Cross sections to explain the development of the western Helvetic nappes (after Ramsay et al., 1983).

Figure 10. Evolutionary model to explain the development of the kilometer-scale recumbent fold of the Courel (Variscan belt of NW Spain) (After Fernández et al., 2007).

Figure 11. Evolutionary model for the Morcles and Diablerets nappes (western Helvetic nappes) (after Casey and Dietrich, 1997). For explanation, see text.

Figure 12. Idealized cross sections to explain the development of the Pardailhan nappe (Montagne Noire, Variscan belt of the French Massif Central) (after Echtler, 1990).

Figure 13. Finite element model of folding of a multilayer showing formation of recumbent folds in the lower layers near to the edge of a rigid block. The recumbent folds were a result of the local shear induced by the block and the effect of gravity. After Vacas-Peña and Martínez Catalán (2004).

Figure 14. Progressive development of recumbent folds in an extensional ductile shear zone formed by simple shear for different values of the shear strain (γ) (after Harris et al. 2002; modified from Ramsay et al., 1983).

Figure 15. Block diagram showing the 3D-geometry of the western Helvetic nappes (after Ramsay, 1981). The recumbent folds are folded by a large anticlastic antiform (terminology from Lisle and Toimil, 2007).

Figure 16. Geological interpretation of a migrated crustal seismic image through the Ossa Morena Zone (Iberian Variscan belt of SW Spain) (after Simancas et al., 2003). LC, lower crust; IRB, Iberseis Reflective Body (band of high seismic reflectivity); UP, upper

Proterozoic; p-CP (white with dark grey lines), pre-Carboniferous Paleozoic; LC, unconformable Lower Carboniferous.

Figure 17. Composite **c**ross section (approximately E – W) through Cabo Ortegal complex (Variscan belt of NW Spain) showing recumbent folds folded by a large synform (after Marcos et al., 2002).

Figure 18. Two pairs of small recumbent folds developed in a multilayer of Cambrian metapsammites and phyllites (Benquerencia beach, Lugo; Variscan belt of NW Spain). The folds look eastward (left of the photograph) and were tilted by a subsequent deformation. The upper pair has a thickened overturned limb whereas the lower pair has the overturned limb cut by a thrust.

Figure 19. A pair of small recumbent folds developed in Cambrian rocks of the basal shear zone of the Mondoñedo nappe (Punta das Cabras, Lugo; Variscan belt of NW Spain). The overturned limb (or short limb) is thicker than the normal limbs.

Figure 20. 2D schemes of the deformation by simple shear of a plane. (a) Initial configuration with obliquity angle θ_0. (b) Deformed configuration after a shear strain γ with final obliquity angle θ of the median surface of the folds; MS, median surface.

Figure 21. Shortening by simple shear of layers dipping in the shear direction or in a direction opposite to it as a function of the shear strain for different values of the initial obliquity angle (θ_0).

Figure 22. Scheme of the deformation by simple shear of a plane with dip direction α_0 (measured clockwise from the shear direction) and dip θ_0. The bulk finite strain ellipse of the deformed plane (median surface of the folds) and the possible angle τ between the major axis of the ellipse and the axial direction of the folds are shown. $\tau = 0$ when the axial direction is controlled by the finite strain, and $\tau \neq 0$ when the axial direction is fixed for a certain

increment of the progressive deformation and then rotates passively. AP, axial plane; MS median surface.

Figure 23. Dip of the axial plane (β) in simple shear, with horizontal shear direction, as a function of the shear strain for layers with different dip directions and initial dips between 10 and 30º. Great numbers on the curves indicate dip direction/dip and the small numbers on the curves indicate shortening percent of the layer. Dashed lines join points of equal shortening. The shaded area corresponds to the field of recumbent folds.

Figure 24. Variation of the angle τ between the whole-rock maximum stretch direction of the deformed layer and the fold axis as a function of the shear strain γ, for conditions of simple shear deformation, dip directions of 30º, 60º and 90º and dips of 10º and 30º (the numbers on the curves indicate dip direction/dip). It is assumed that the fold axis forms in the direction of the whole-rock strain ellipse in the beginning of the deformation of the layer and then it rotates passively in agreement with the subsequent strain.

Figure 25. Comparison of a large recumbent fold in the Appalachian Inner Piedmont (a) with an experimental recumbent fold produced in a multilayer with layers of stitching wax (b). (a) after Hatcher (1972, 1981), (b) after Bucher (1956).

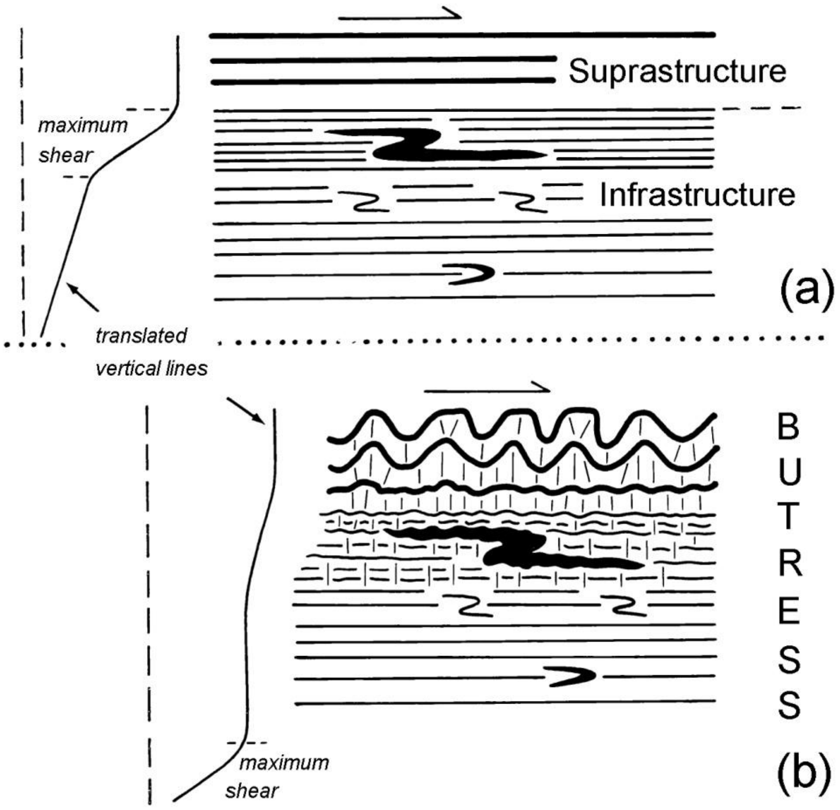

Suprastructure
maximum
shear
Infrastructure
(a)
translated
vertical lines
B
U
T
R
E
S
S
maximum
shear
(b)

Figure 1

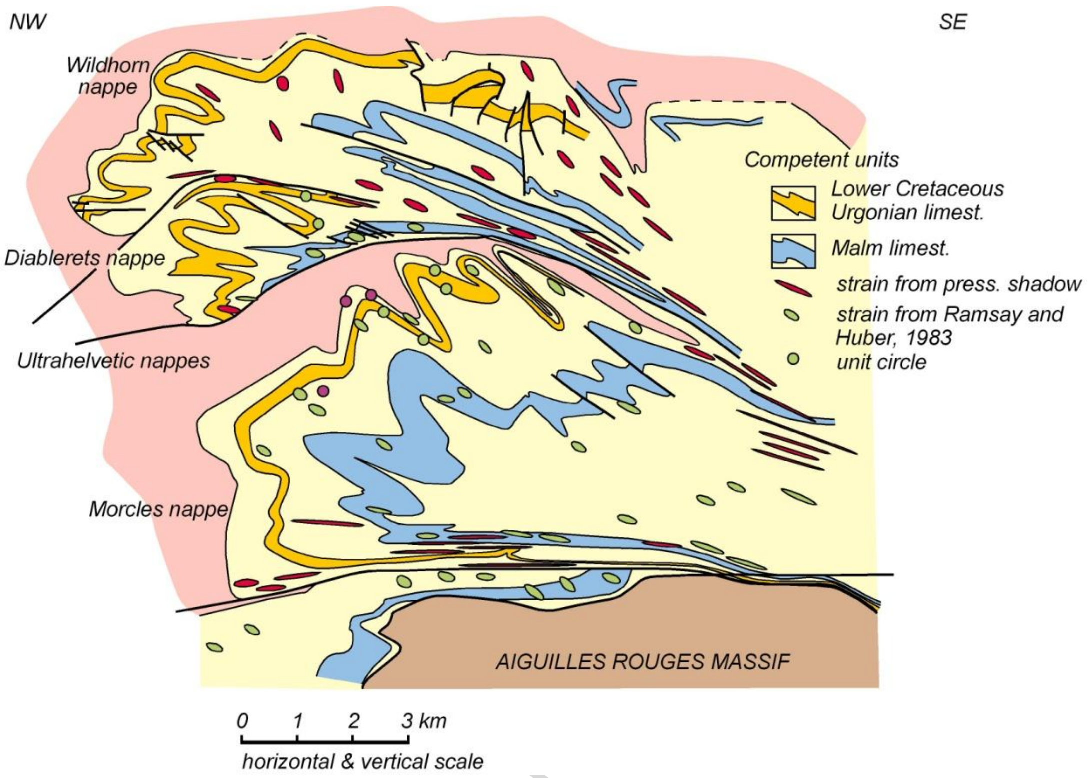

NW
SE
Wildhorn nappe
Diablerets nappe
Ultrahelvetic nappes
Morcles nappe
AIGUILLES ROUGES MASSIF
Competent units
Lower Cretaceous Urgonian limest.
Malm limest.
strain from press. shadow
strain from Ramsay and Huber, 1983
unit circle
0 1 2 3 km
horizontal & vertical scale

Figure 2

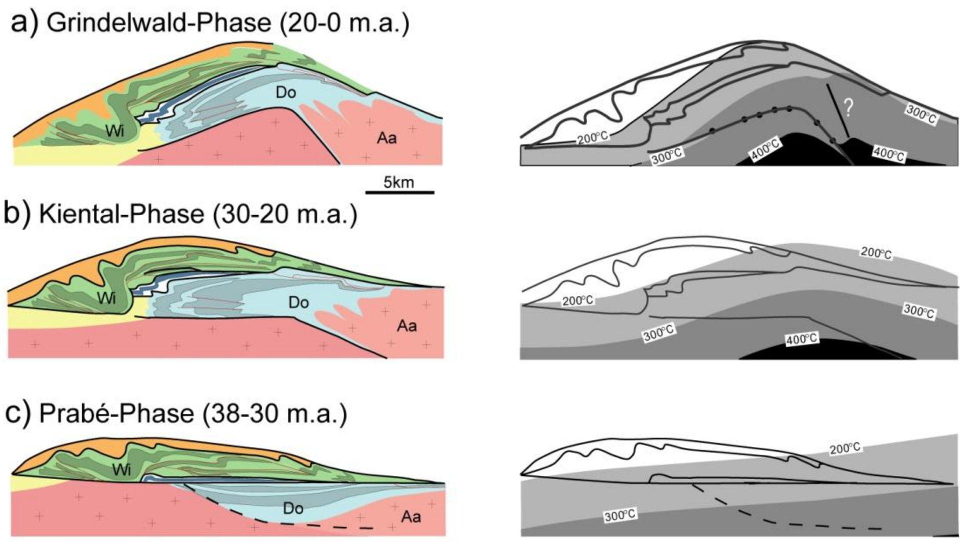

a) Grindelwald-Phase (20-0 m.a.)
Do
Wi
Aa
5km
200°C
300°C
400°C
300°C
400°C
?
b) Kiental-Phase (30-20 m.a.)
Wi
Do
Aa
200°C
200°C
300°C
300°C
400°C
c) Prabé-Phase (38-30 m.a.)
Wi
Do
Aa
200°C
300°C

Figure 3

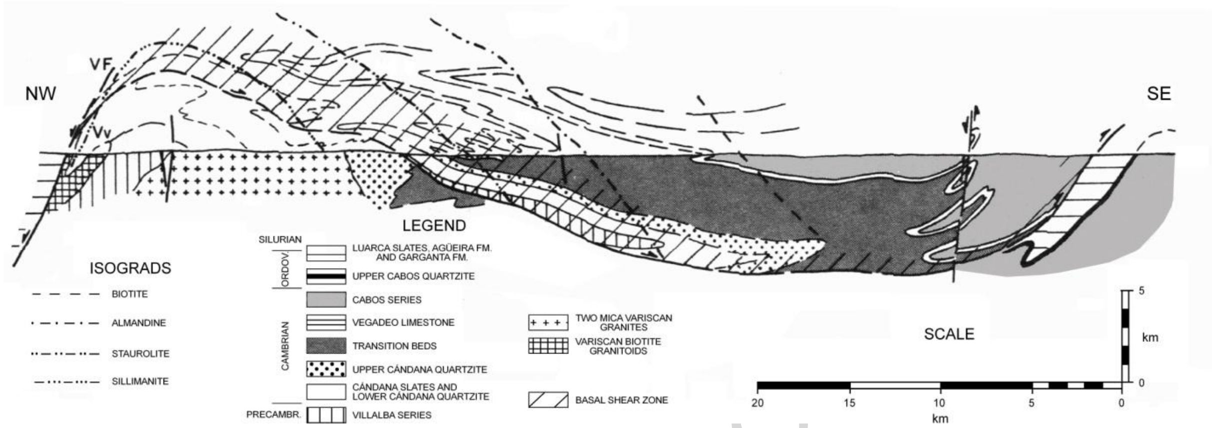

NW
VF
Vv
SE
ISOGRADS
BIOTITE
ALMANDINE
STAUROLITE
SILLIMANITE
LEGEND
SILURIAN
ORDOV.
CAMBRIAN
PRECAMBR.
LUARCA SLATES, AGÜEIRA FM. AND GARGANTA FM.
UPPER CABOS QUARTZITE
CABOS SERIES
VEGADEO LIMESTONE
TRANSITION BEDS
UPPER CÁNDANA QUARTZITE
CÁNDANA SLATES AND LOWER CÁNDANA QUARTZITE
VILLALBA SERIES
TWO MICA VARISCAN GRANITES
VARISCAN BIOTITE GRANITOIDS
BASAL SHEAR ZONE
SCALE
5
km
0
20
15
10
5
0
km

Figure 4

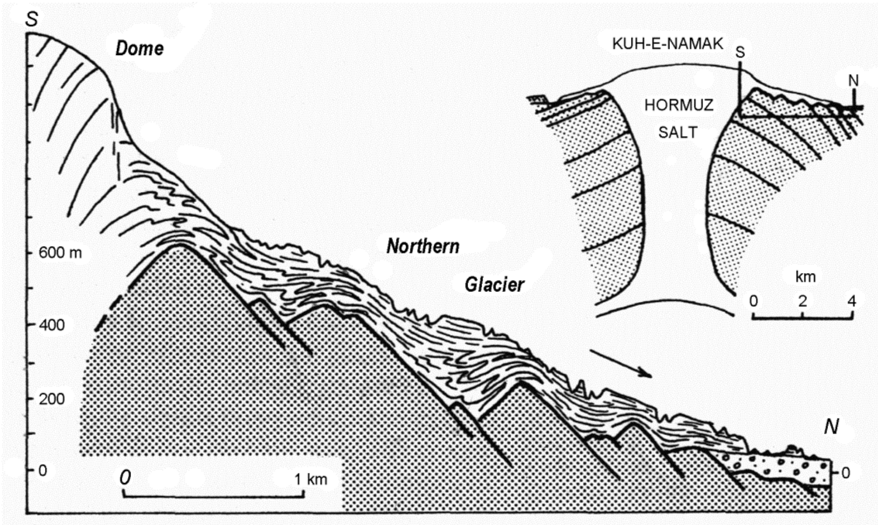

S
Dome
600 m
400
200
0
0
1 km
Northern
Glacier
N
0
KUH-E-NAMAK
S
N
HORMUZ
SALT
km
0
2
4

Figure 5

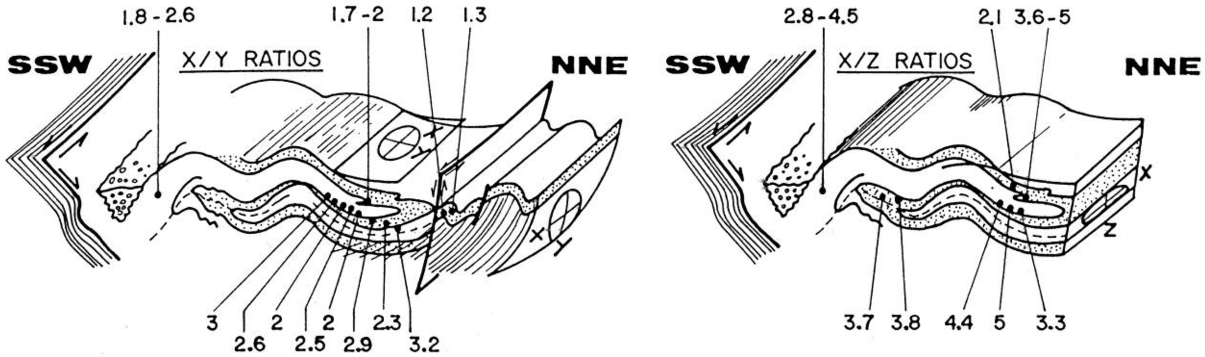

1.8 - 2.6
SSW
X/Y RATIOS
1.7 - 2
1.2
1.3
NNE
3
2.6
2
2.5
2
2.9
2.3
3.2
2.8 - 4.5
SSW
X/Z RATIOS
2.1
3.6 - 5
NNE
X
Z
3.7
3.8
4.4
5
3.3

Figure 6

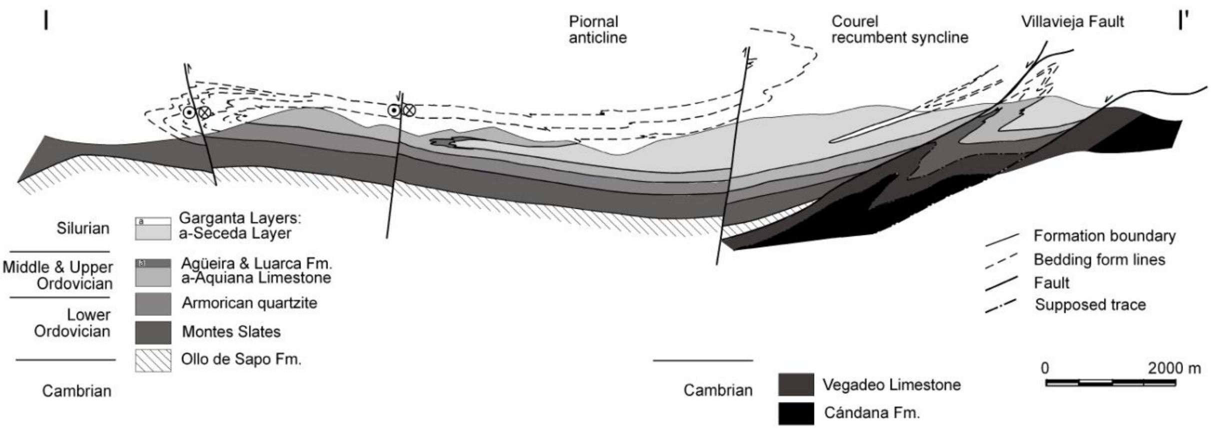

I
Piornal
anticline
Courel
recumbent syncline
Villavieja Fault
I'
Silurian
Garganta Layers:
a-Seceda Layer
Middle & Upper
Ordovician
Agüeira & Luarca Fm.
a-Aquiana Limestone
Lower
Ordovician
Armorican quartzite
Montes Slates
Ollo de Sapo Fm.
Cambrian
Formation boundary
Bedding form lines
Fault
Supposed trace
0
2000 m
Cambrian
Vegadeo Limestone
Cándana Fm.

Figure 7

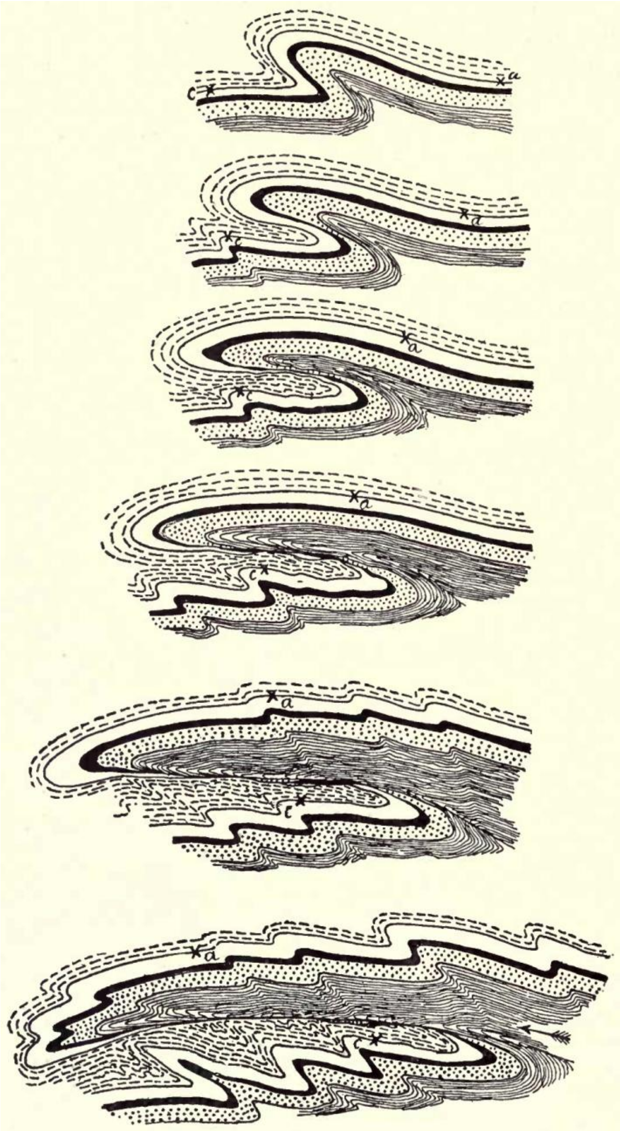


Figure 8

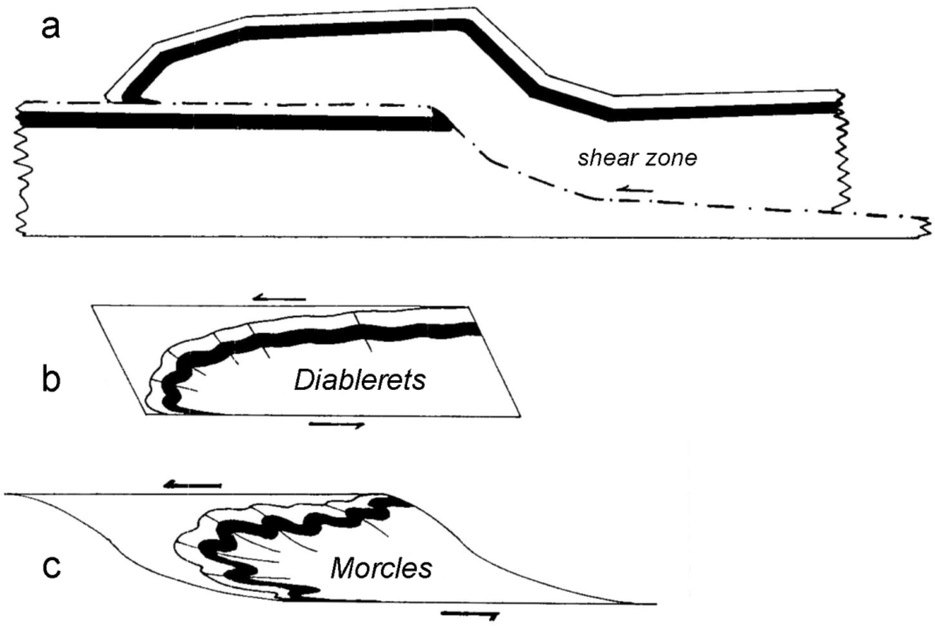


Figure 9

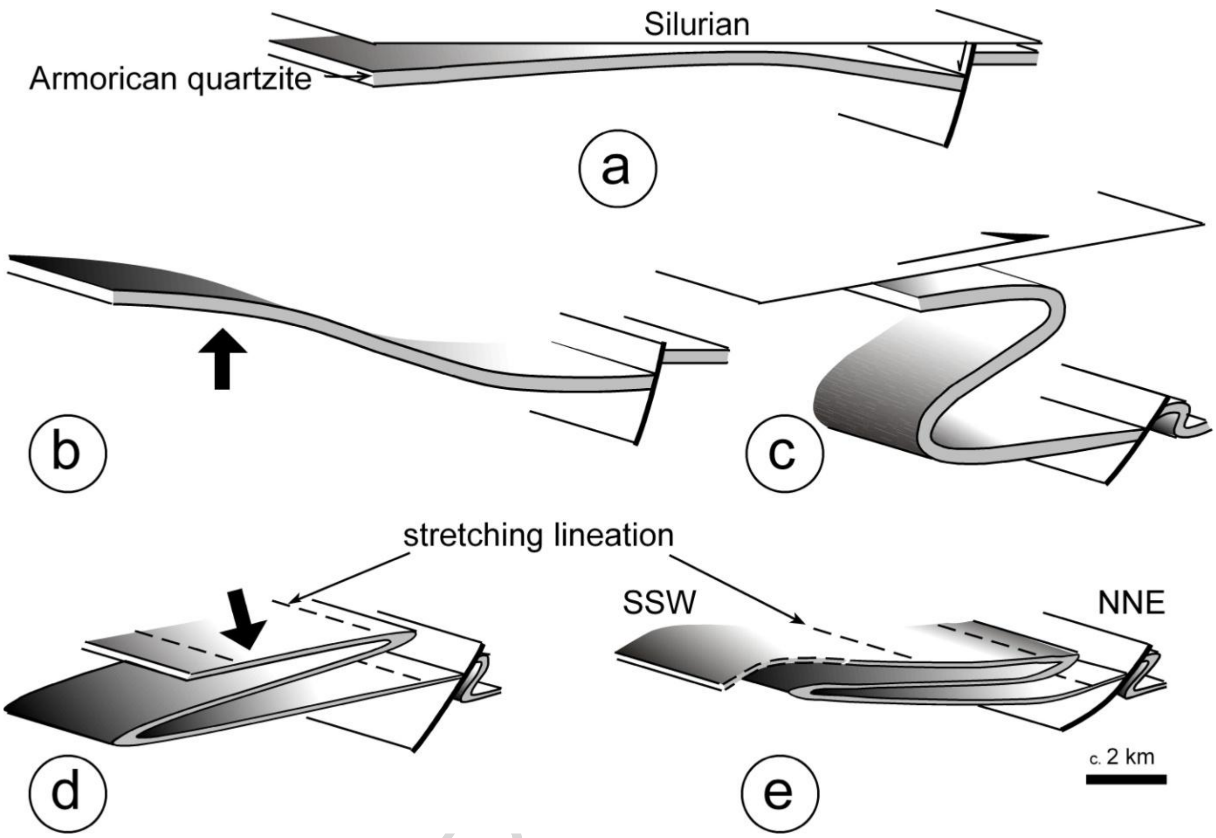

Silurian
Armorican quartzite
a
b
c
stretching lineation
SSW
NNE
d
e
c. 2 km

Figure 10

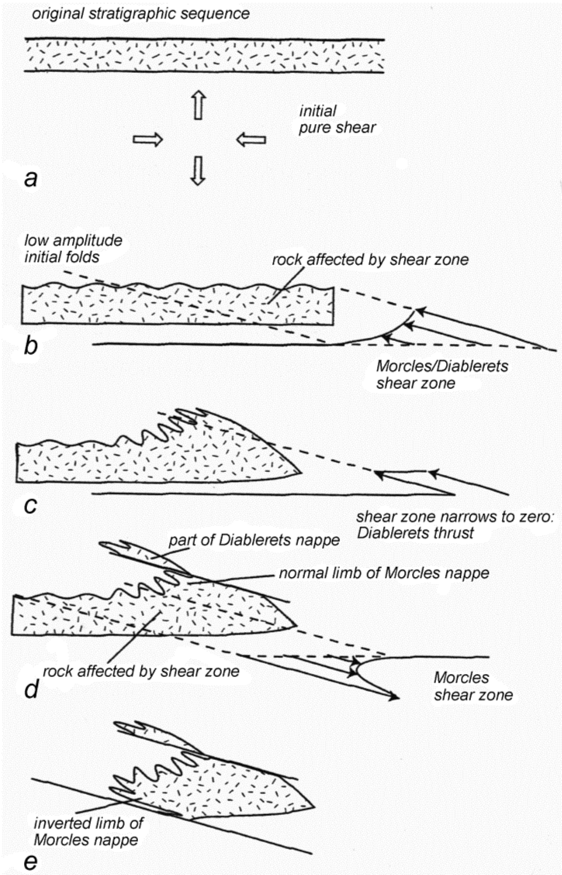

original stratigraphic sequence
initial
pure shear
a
low amplitude
initial folds
rock affected by shear zone
b
Morcles/Diablerets
shear zone
c
shear zone narrows to zero:
Diablerets thrust
part of Diablerets nappe
normal limb of Morcles nappe
rock affected by shear zone
d
Morcles
shear zone
inverted limb of
Morcles nappe
e

Figure 11

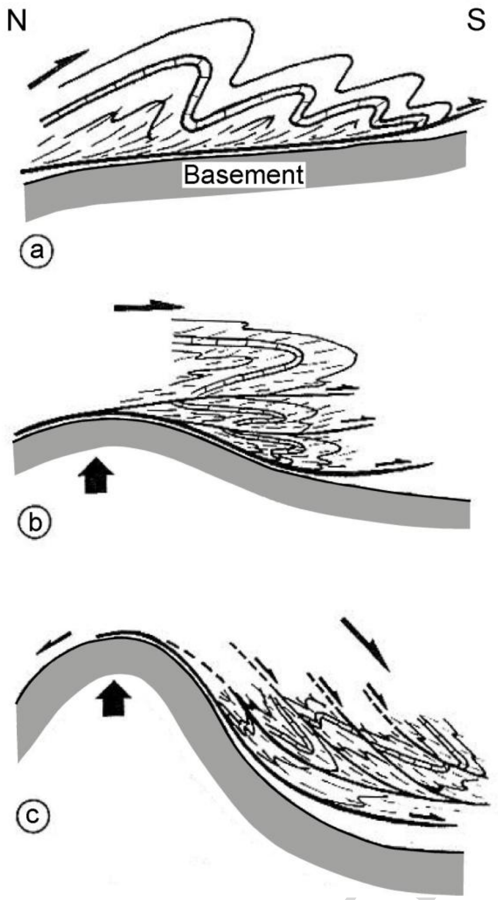


Figure 12

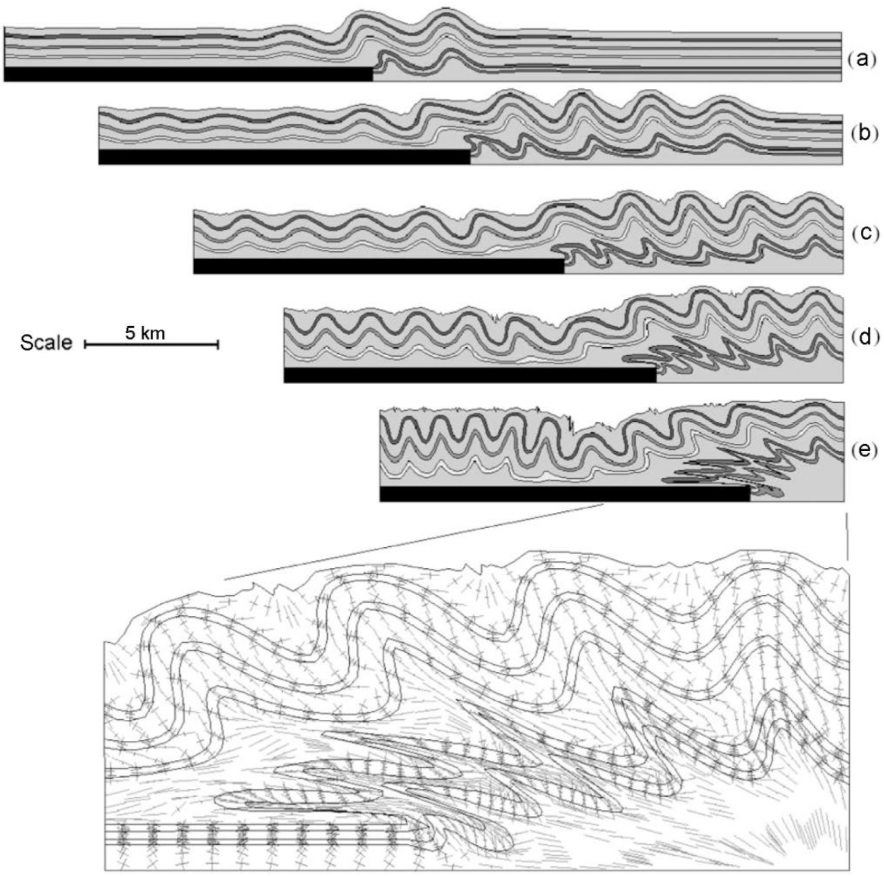


Figure 13

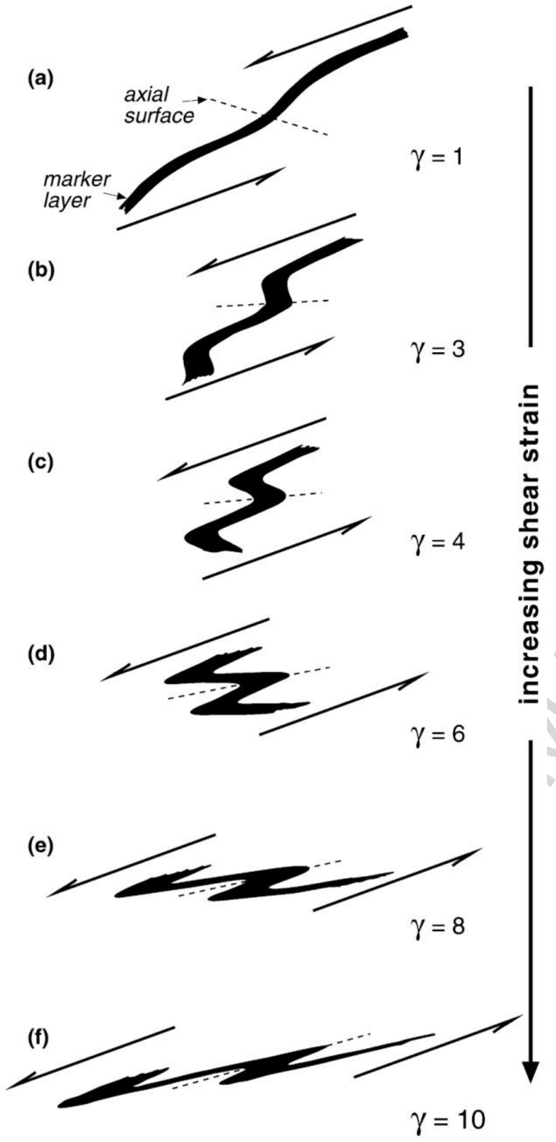

(a)
axial
surface
marker
layer
γ = 1
(b)
γ = 3
(c)
γ = 4
(d)
γ = 6
(e)
γ = 8
(f)
γ = 10
increasing shear strain

Figure 14

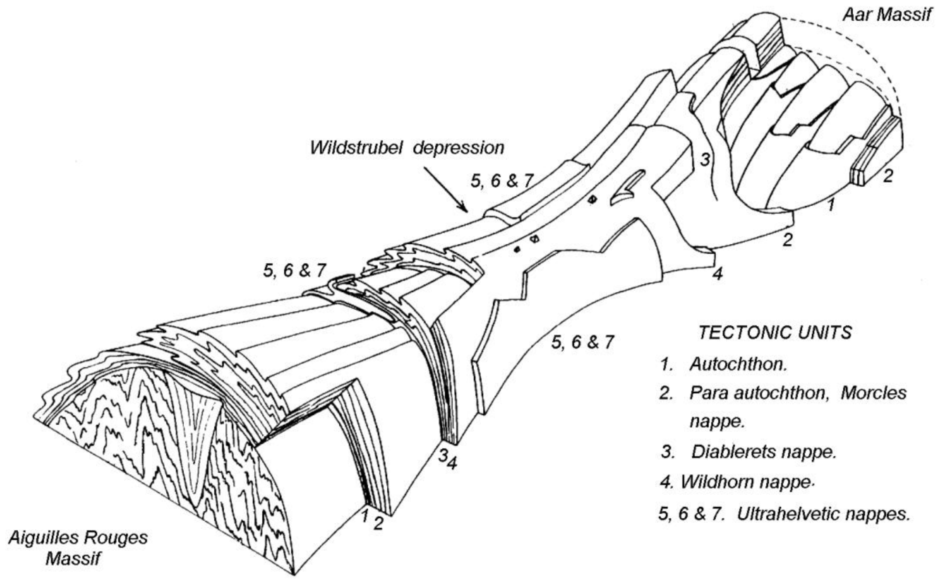

Aar Massif
Wildstrubel depression
5, 6 & 7
3
2
1
2
5, 6 & 7
4
TECTONIC UNITS
1. Autochthon.
2. Para autochthon, Morcles nappe.
3. Diablerets nappe.
4. Wildhorn nappe.
5, 6 & 7. Ultrahelvetic nappes.
5, 6 & 7
3
4
1
2
Aiguilles Rouges Massif

Figure 15

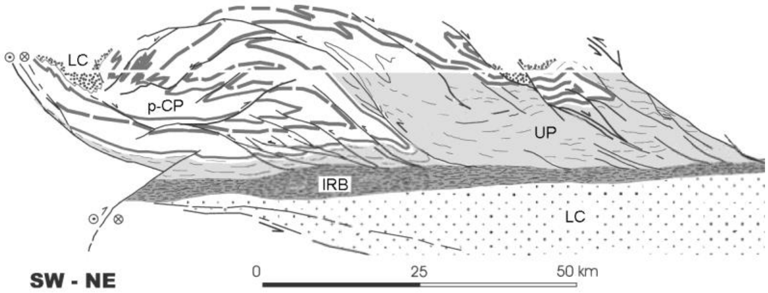


Figure 16

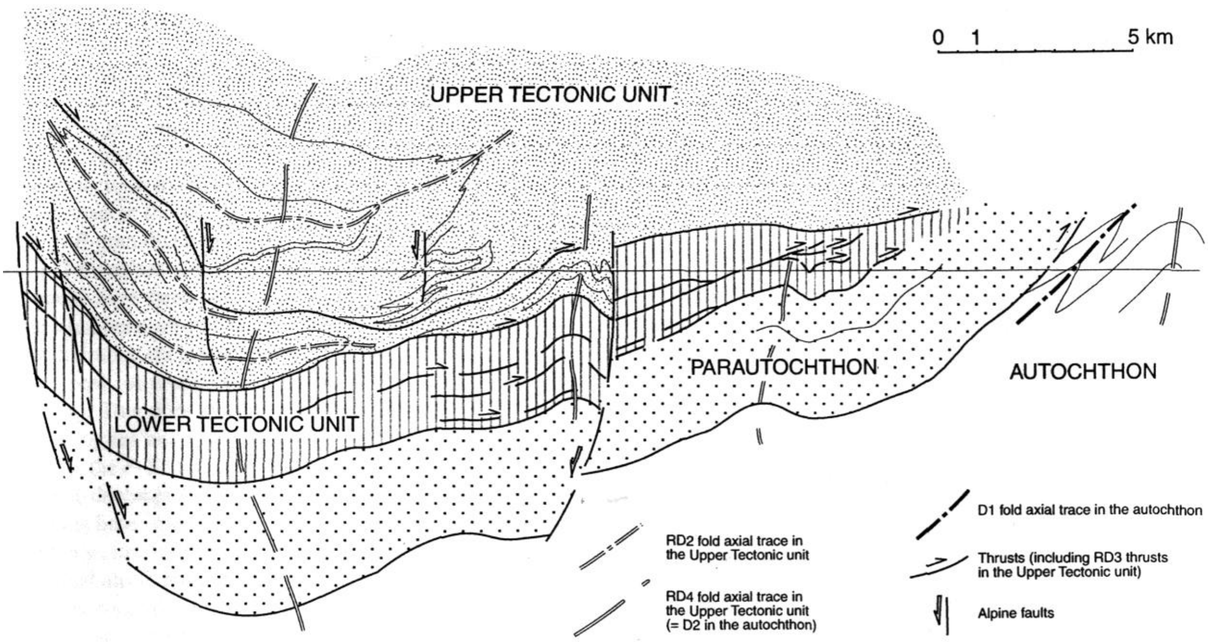

0 1 5 km
UPPER TECTONIC UNIT
LOWER TECTONIC UNIT
PARAUTOCHTHON
AUTOCHTHON
RD2 fold axial trace in
the Upper Tectonic unit
RD4 fold axial trace in
the Upper Tectonic unit
(= D2 in the autochthon)
D1 fold axial trace in the autochthon
Thrusts (including RD3 thrusts
in the Upper Tectonic unit)
Alpine faults

Figure 17

Figure 18

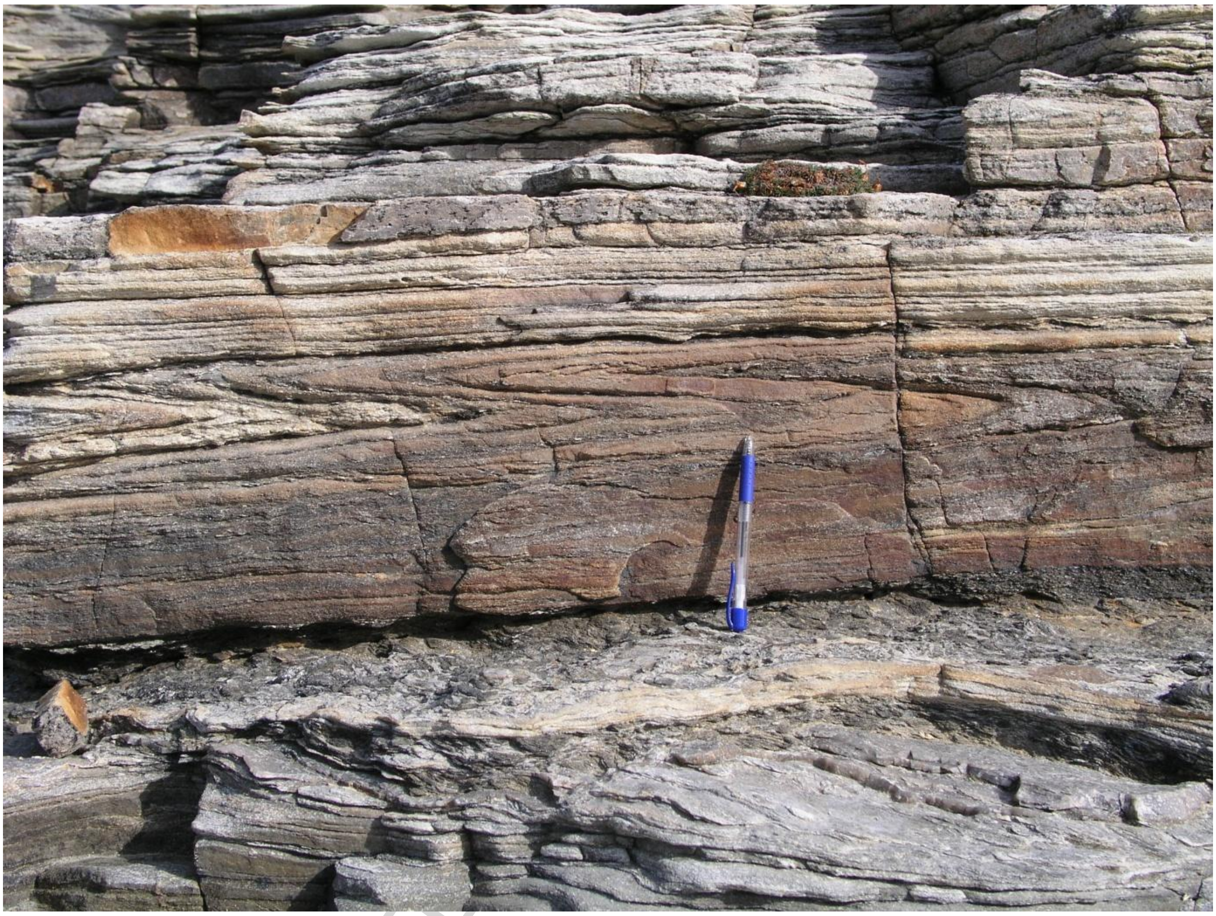


Figure 19

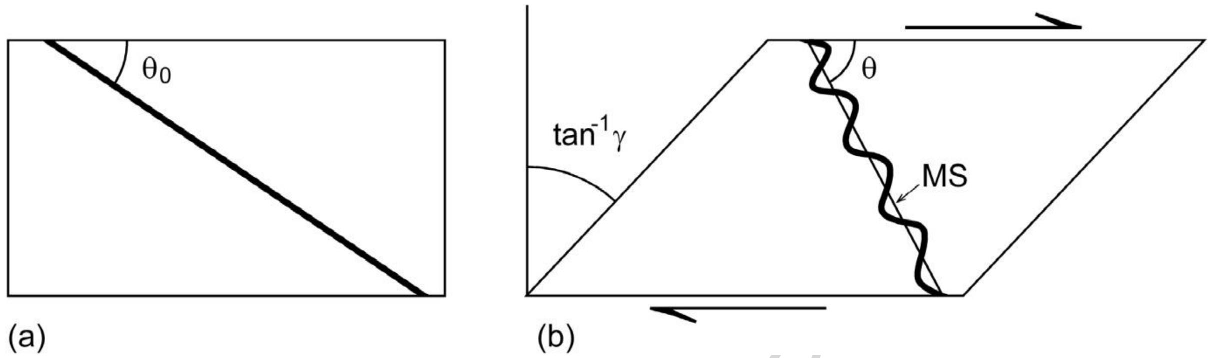


Figure 20

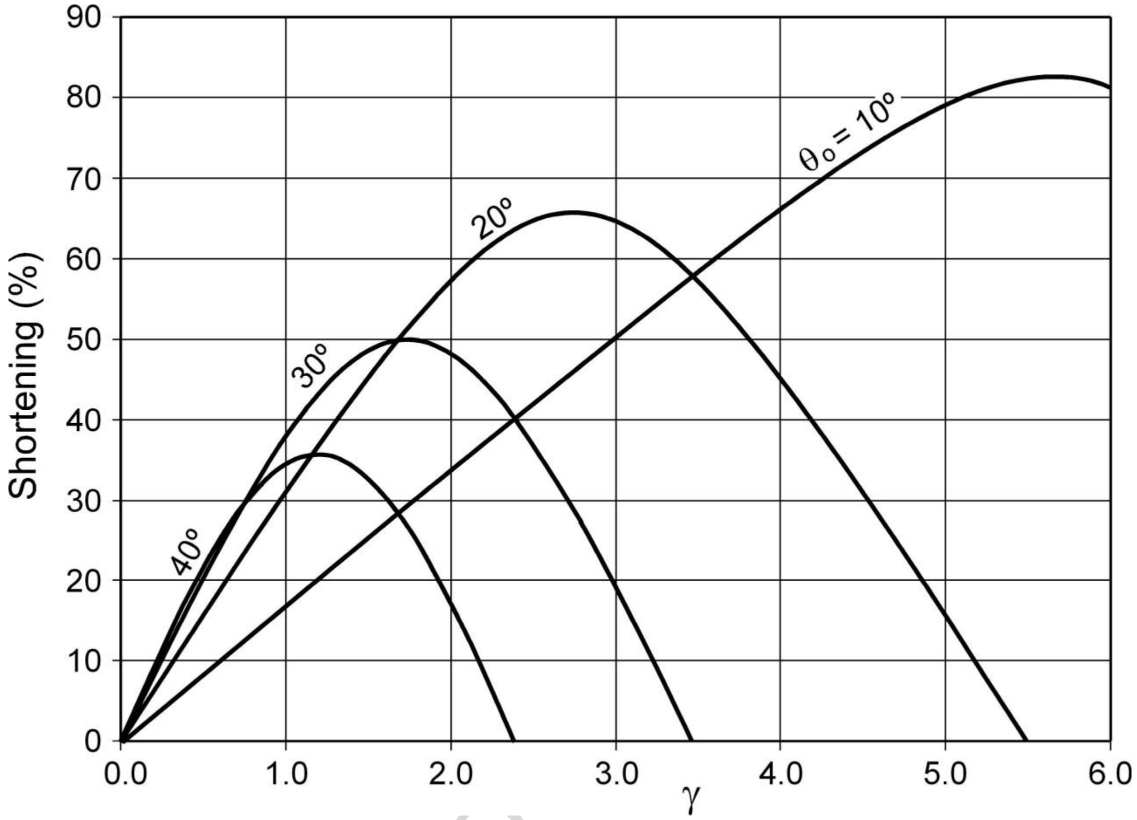

Shortening (%)
γ
θo= 10°
20°
30°
40°
0
10
20
30
40
50
60
70
80
90
0.0
1.0
2.0
3.0
4.0
5.0
6.0

Figure 21

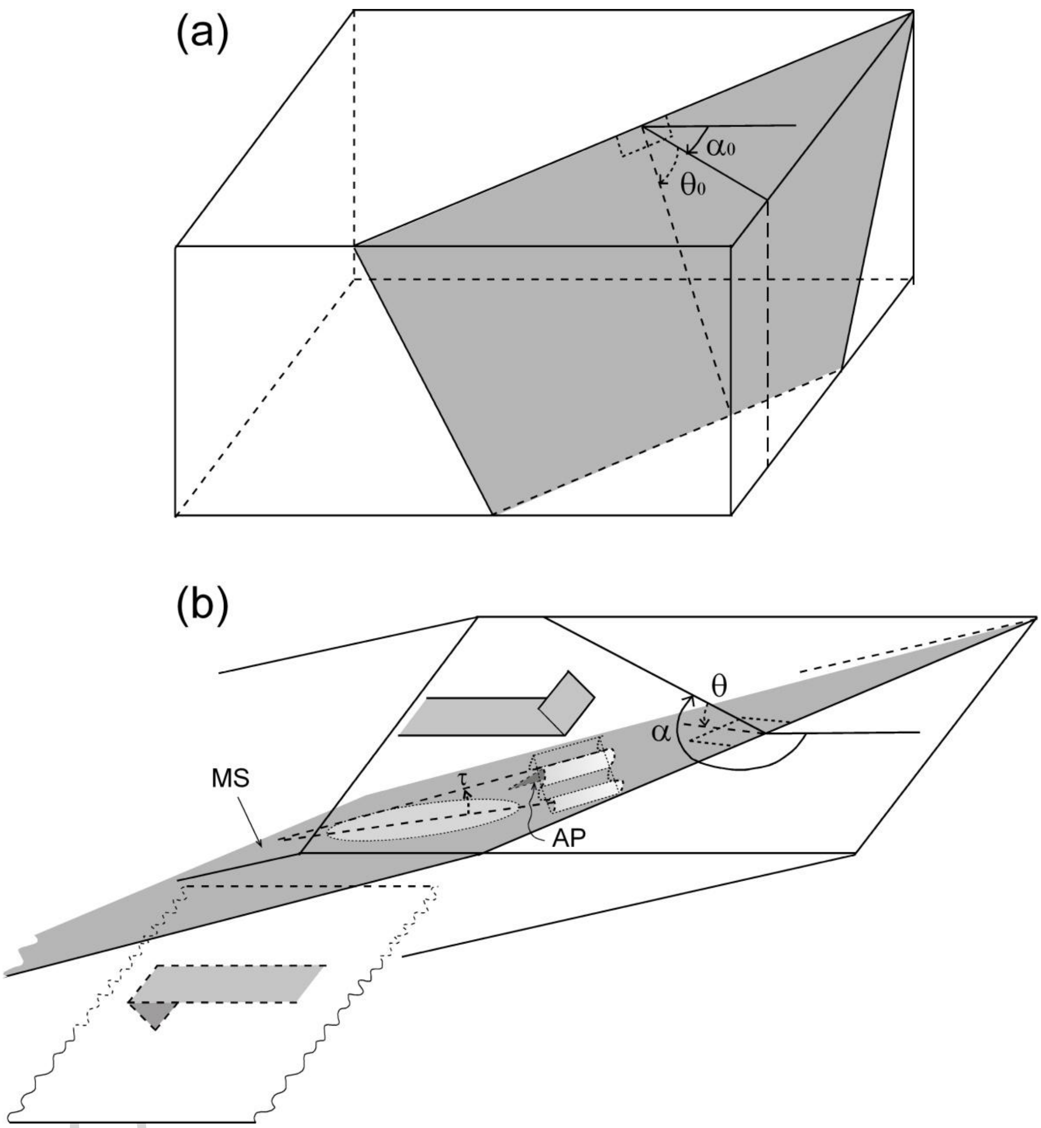


Figure 22

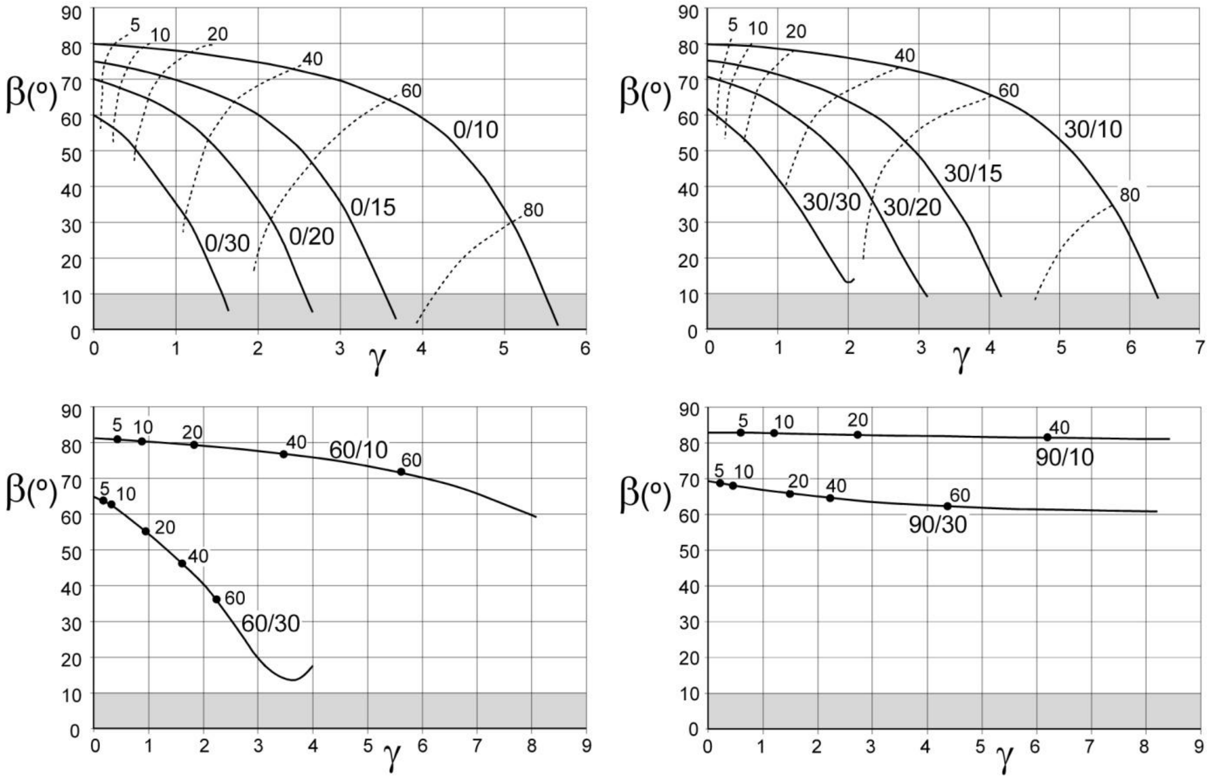


Figure 23

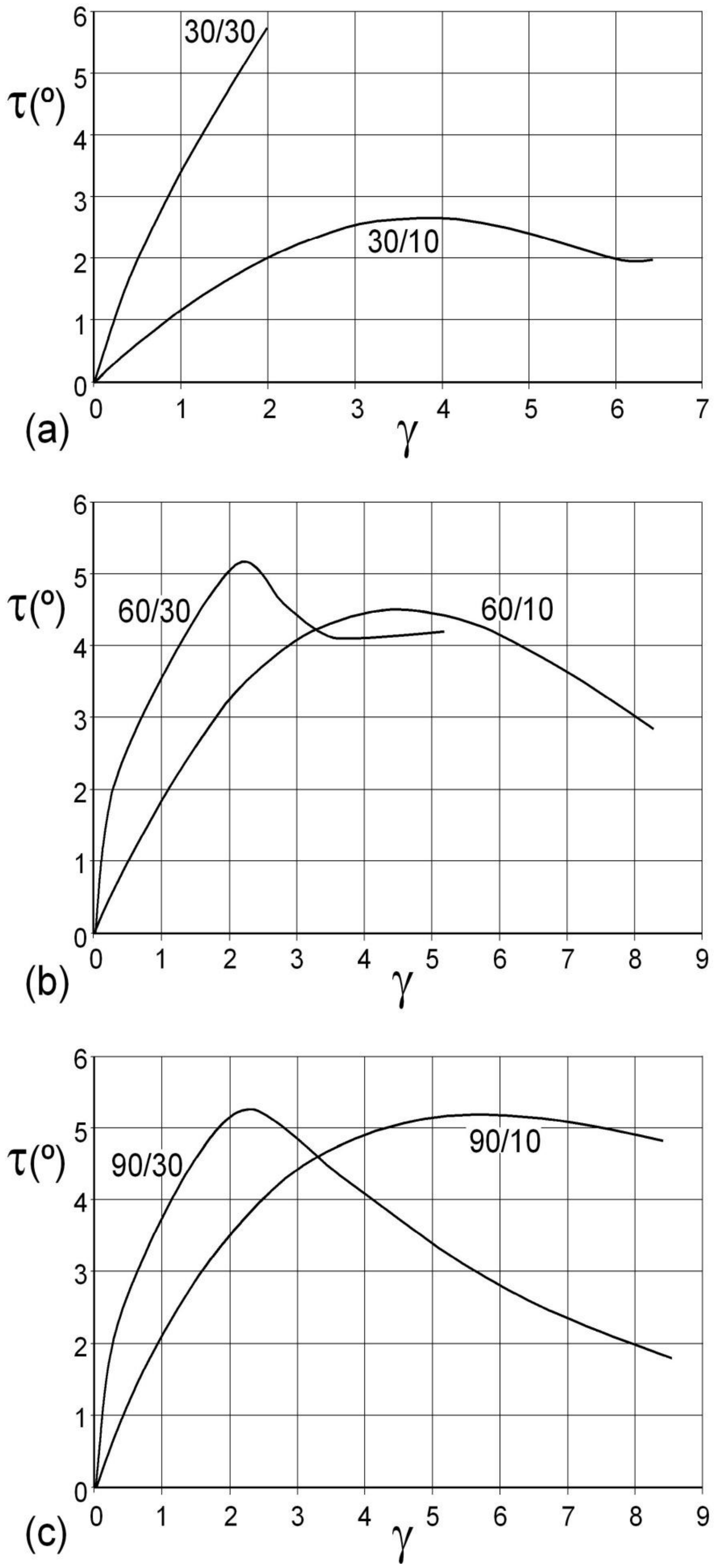


Figure 24

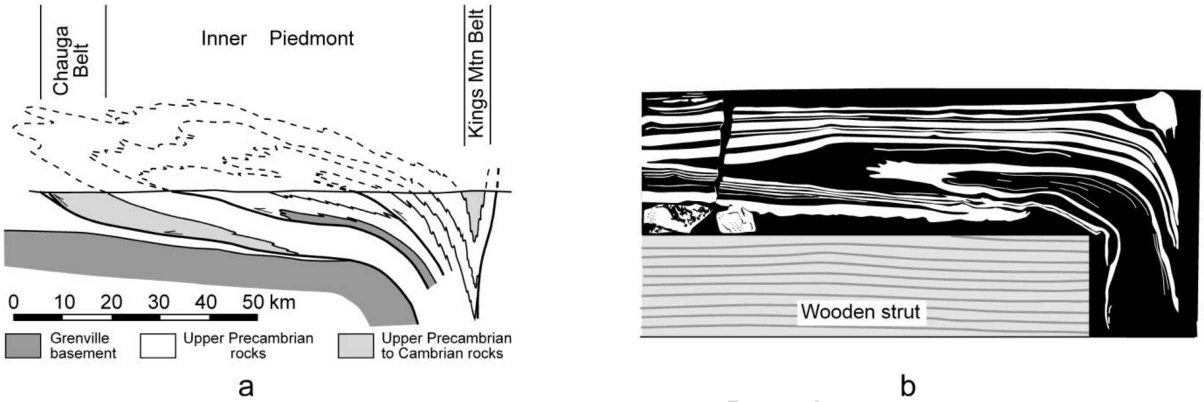
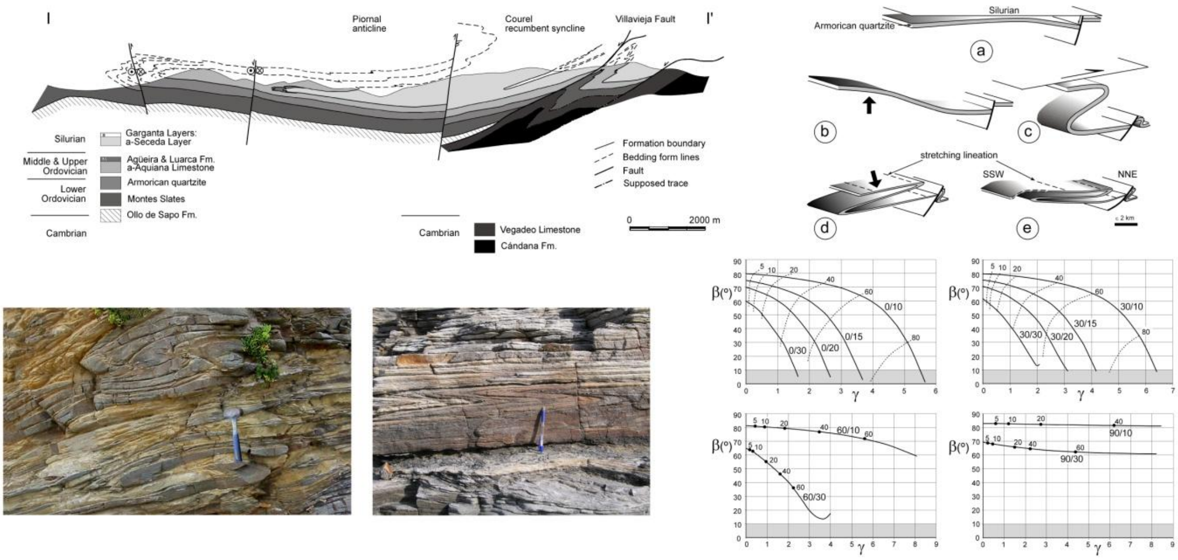

Chauga Belt
Inner Piedmont
Kings Mtn Belt
0 10 20 30 40 50 km
Grenville basement
Upper Precambrian rocks
Upper Precambrian to Cambrian rocks
a
Wooden strut
b

Figure 25



Graphical abstract