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Normal faults, layering and elastic properties of rocks

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ABSTRACT

We study mesoscale normal faults cutting alternating limestone and clay-rich layers in several localities in the South-Eastern Mesozoic sedimentary basin (France). The displacement gradients, defined as the displacement variation per unit length along fault profile, and the mean fault dips are correlated to the structural and petrophysical properties of the host rock, including the carbonate content, the stiffness, the layering pattern and the maximum burial depth. Analysis of the fault dips indicates that the faults propagate (downward or upward) from an initial fracture from one unit to another rather than through connections of fractures that nucleated in different units. The fault dips are consistent with shear, mixed mode and tensile failures in limestone units. They are consistent with shear failure or are abnormally low in the clay-rich units. Among the studied attributes, in limestone units the failure mode is related to the contrast of the Young's modulus between the limestone and the clay-rich layers. A high contrast promotes tensile failure, whereas a low contrast promotes mixed mode or shear failures. In clay-rich layers, the dip is related to the layering pattern and abnormally low dips are promoted in thin clay-rich layers surrounded by thick limestone units. Specific displacement gradients characterize each lithology of the layered section. It ranges from 0.06 to 0.2 in the clay-rich units and it increases with Young's modulus. The analyses of the fault dips and the displacement gradients have implications in term of local stress. Both analyses converge and they can be related to a variation in stress controlled by the variation in stiffness through the layering.

1. Introduction

In sedimentary basins, the faults generally grow in heterogeneous layered sections with different lithologies. Field work, seismic data and modeling show that the layering highly controls the fault structure (Ferrill and Morris, 2008; Lansigu and Bouroullec, 2004; Larsen and Gudmundsson, 2010; Peacock and Sanderson, 1992; Reyer et al., 2012; Roche et al., 2012a), the fault growth (Childs et al., 1996; Ferrill and Morris, 2003; Larsen et al., 2010; Roche et al., 2012b, 2013; Schöpfer et al., 2006), as well as the displacement accumulation (Gross et al., 1997; Gudmundsson, 2004; Larsen and Gudmundsson, 2010; Muraoka and Kamata, 1983; Roche et al., 2012b; Wibberley et al., 2007; Wilkins and Gross, 2002). This paper deals with the variations in fault dips and displacement gradients related to the layering. These parameters may give insight into the rock properties and the state of stress within the rocks. Likewise, they are key parameters to assess the length of an isolated fault. This is fundamental to discuss the moment magnitude of earthquakes in seismic fields, the cap rock integrity, the leak-off pressure processes, the fracture-induced permeability, and the containment capability of a host rock in oil and gas reservoir fields.

The fault dip changes with the lithology in layered sections. The dip is commonly lower in compliant units such as clay-rich units than in stiff units such as limestone or sandstone units (Davison, 1987; Ferrill and Morris, 2003; Peacock and Sanderson, 1992; Roche et al., 2012a; Schöpfer et al., 2007; Van der Zee et al., 2008). Such refraction may have various origins. First, this may result from a late differential compaction within the layered section (Davison, 1987; Mandl, 1988; Wang, 1995). This process is limited for faults that grew at important depth (i.e. <500 m) because the compaction decreases exponentially with depth (Giles, 1997). In addition, the refraction is also likely weak for fault initiated at shallow depth (i.e. <500 m) because the fault dips should decrease in both compliant and stiff units due to the compaction, and the faults should display a low dip (i.e. <60°) in the stiff layers (Davison, 1987). Then, the refraction may result from the fault growth within the layered section. In that case, the shallow dip in the compliant layer is formed by the connection of two steeply dipping fractures located in the surrounding stiff units. The fractures that connect may be either unrelated, for instance they are part of different joint or fault

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sets, or they are related together and comprised in a shear zone (Childs et al., 1996; Larsen and Gudmundsson, 2010; Schöpfer et al., 2006; Walsh et al., 2003). In both cases, the change in dip must be somehow related to stress deflection in the inferred along-dip echelon (Larsen and Gudmundsson, 2010; Peacock and Sanderson, 1994), and a relationship should arise between the refraction and the configuration of the fracture sets or the shear zone. Finally, the refraction may result from the continuous propagation of the fault through the layered section (i.e. without connection of pre-existing fractures) (Roche et al., 2012a,b; Schöpfer et al., 2006). The changes in dip are then associated with a stress-refraction (Bradshaw and Zoback, 1988), or with a variation in effective stress state. This may be related to the change in mechanical properties through the layered section including the contrast in strength (i.e. friction and/or cohesion) (Ferrill and Morris, 2003; Mandl, 1988), the contrast in stiffness (i.e. Young's modulus) (Roche et al., 2013) or a fluid overpressure in one unit due to the contrast in permeability.

Isolated non-restricted faults are faults growing without interaction with other faults, showing no physical contact between their tips and a restrictor, such as, lithological interfaces or other faults, and without an abnormally steep near-tip displacement gradient. For these faults the distribution of the fault displacement, observed in cross section, is also controlled by the layering. The shape of the displacement profile and the relationship between the maximum displacement (*Dmax*) and the length (*L*) (i.e. *Dmax–L*) display correlations with the layering. This results from the dependence of the displacement gradients to the medium dissected by the fault (Ferrill and Morris, 2008; Gross et al., 1997; Gudmundsson, 2004; Muraoka and Kamata, 1983; Roche et al., 2012b; Wibberley et al., 2007; Wilkins and Gross, 2002). When faults are modeled as ideal crack surfaces in an elastic homogeneous medium, the *Dmax/L* ratio depends on the rock properties (Bürgmann et al., 1994; Cowie and Scholz, 1992; Gudmundsson, 1987a,b; Gudmundsson, 2004; Gudmundsson et al., 2013; Martel and Pollard, 1989; Schultz and Fossen, 2002; Walsh and Watterson, 1987; Welch et al., 2009). For a circular normal fault, the displacement is linked to the length of the fracture *L* with the following equation:

$$Dmax = \frac{2\Delta\tau(1+\nu)}{E}L, \tag{1}$$

where E and ν are Young's modulus and Poisson's ratio. $\Delta\tau$ is the driving shear stress and depends on the remotely applied stress, the pore pressure and the strength of the rocks. Using the static modulii in this relationship is more suitable for the long term displacement whereas the dynamic moduli are more suitable for the slip during co-seismic rupture (Gudmundsson, 2011). In this paper, we assume that the displacement gradients that are the displacement variation per unit of length along a fault profile are linked to the properties of the rocks along the fault following the same relationship. The displacement gradient is therefore expected to be inversely and directly proportional to Young's modulus and shear stress acting on the fault plane, respectively (Fig. 1). However such theoretical relationships have not been illustrated by field data yet.

The current study attempts to compare the fault dip and the displacement gradient in each rock type of layered sections with the structural and petrophysical properties of the layering such as the thickness, the carbonate content, the maximum burial depth and the stiffness of the layer.

2. Geological setting and data acquisition

2.1. Geological setting

We study faults in seven localities named Trescléoux, St-Didier, Vogüé, Flaviac, and three others referred to as Hauterivian sites. The sites are located in the South-Eastern Mesozoic sedimentary basin of France, which was part of the North European continental margin of the Tethys Ocean. They are either in the Ardèche margin that is the

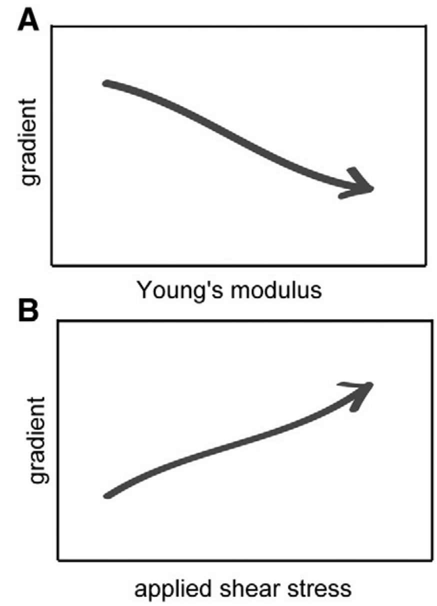

Fig. 1. Schematic evolution of the displacement gradient with Young's modulus (A), and with the applied shear stress (B), assuming a "crack model" in an elastic medium (Gross et al., 1997; Gudmundsson, 2011).

Western boundary of the South-Eastern basin, or in the Vocontian trough that is the deepest part of the basin (Fig. 2). The faults cut layered sections of two lithologies called here the limestone and clay-rich units. The layers are a few decimeters high and tilted 12° to 66° (Fig. 3). The slickensides observed on the fault planes indicate that the faults were formed as normal faults, which is also consistent with the normal offsets of the sedimentary interfaces (Figs. 2, 3). The mineralogy and mechanical properties are almost constant within a same unit. These sedimentary systems were deposited in the Late Jurassic and Early Cretaceous (Early Oxfordian, Late Oxfordian, and Hauterivian stages; Table 1) over the course of the Mesozoic sedimentary infilling over the Hercynian crystalline basement, after the Late-Triassic/Early-Jurassic rifting (Bill et al., 2011; Lemoine and De Graciansky, 1988; Lemoine et al., 1986), and continuously until the Late Cretaceous (Baudrimont and Dubois, 1977). The basin was inverted during two main phases of compression in the Late Eocene (Pyrenean compression) and Late Miocene (Alpine compression) (Champion et al., 2000; Choukroune et al., 1973; Dumont et al., 2012; Sibuet et al., 2004). Complex structures were developed during these compressions. It includes E–W and N–S folding, double-verging E–W-trending thrust faults mainly located in the southern part of the Basin, westward N–S to NNW–SSW trending thrust faults and a highly complex set of oblique strike-slip faults (Fig. 2). N–S striking graben structures opened up under an E–W distension phase between the two orogeneses in the Oligocene (Arthaud et al., 1977; Bergerat, 1987; Roure et al., 1992). Though it is difficult to ascertain the age of the faulting in the study sites, the normal fault formation is likely associated with this Oligocene distension. The directions of the minimum principal stress (σ_3) responsible for the faults are quite similar in all the sites (i.e. E–W to SE–NW) and in agreement with the Oligocene stretching (Fig. 2), and the faults formed before the bed tilting. The only exception is the almost N–S direction of σ_3 at Vogüé site, which may suggest that the faults took place there before the Oligocene.

2.2. Field data collected on faults

The outcropping portions of the normal faults are 50 to 680 cm long and their maximum vertical displacements (*Dmax*) range from 1 cm to 1 m. Dips of the fault planes and offsets of the sedimentary markers are measured systematically all along the faults and in each layer. We build 10°-interval frequency histograms using the fault dip data before the