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Comparison of paleostress magnitudes from calcite twins with contemporary stress magnitudes and frictional sliding criteria in the continental crust: Mechanical implications

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Abstract

The evolution with depth of paleo-differential stress magnitudes (mostly derived from calcite twinning paleopiezometry) in various tectonic settings is compared to the modern differential stress/depth gradients deduced from in situ measurements. Despite dispersion, both independent sets of stress data support to a first-order that the strength of the continental crust down to the brittle-ductile transition is generally controlled by frictional sliding on well-oriented pre-existing faults with frictional coefficients of 0.6–0.9 under hydrostatic fluid pressure. Some ductile mechanisms may, however, relieve stress and keep stress level beyond the frictional yield, as for instance in the detached cover of forelands. The main conclusion is that despite inherent differences, contemporary stress and paleostress data can be combined to bring useful information on the strength and mechanical behaviour of the upper continental crust over times scales of several tens of Ma.

Keywords: Calcite twins; Paleostress; Contemporary stress; Differential stress magnitudes; Friction; Strength; Continental crust

1. Introduction

The concept of stress is of primary importance when dealing with the mechanics of materials. Although the validity of applying concepts of continuum mechanics to natural rock masses that are neither continuous nor purely solid may be questioned, in practice the idealisation of reality by the definition of an equivalent continuum has proved very powerful for many mechanical purposes. The mechanical behaviour of rocks is in many instances very satisfactorily explained and predicted by referring to stresses, which is in turn a good justification of the wide use of this concept in tectonics studies.

The motivation to characterise the distribution of stresses in the crust arises from applied geological purposes such as geological hazards, engineering activities and resource exploration, but also from fundamental geological purposes, such as understanding the mechanical behaviour of geological materials and deciphering various tectonic mechanisms, from those related to plate motions at a large scale to those causing jointing and faulting or even microstructures at a smaller scale.

Many techniques of measurements of contemporary stresses were improved for applied geological purposes and for a better understanding of current seismicity, active fault kinematics and crustal strength [e.g., Harper and Szymanski (1991), Cornet (1993), Engelder (1993), Amadei and Stephansson (1997) and more recently Ljunggren et al. (2003) and Zoback et al. (2003). See also Zoback and Zoback (1989) and Zoback (1992) for a discussion of a quality ranking and rationale for such ranking for a variety of contemporary stress indicators]. These methods determine contemporary stress components by different techniques that rely upon different bases of measurement: fluid pressure for hydraulic fracturing, i.e., a quantity directly related to stress; geometry of finite deformation and/or reloading strains for borehole techniques; relief strains, evolution of relief strains, reloading strains, finite deformation state for techniques on cores; and seismic radiation for earthquakes. Theoretically, methods

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using hydraulic fractures, boreholes or rock cores all have the potential for the complete determination of the stress tensor from a set of measurements at a point location but in practice this only applies when rock coring is undertaken.

Coevally, methods of paleostress reconstructions based on mechanical interpretation of various structural or petrographic elements in natural rocks have been set out in order to decipher the past tectonic evolution. However, Earth scientists characterising contemporary stresses using in situ measurements or earthquake focal mechanisms and those improving and applying methods of paleostress reconstructions seem not to speak the same language and not to deal with similar mechanical concepts. Despite few attempts at including results of paleostress reconstructions within young rock formations in the compilation of the present-day global tectonic stress pattern (Zoback et al., 1989), contemporary stress and paleostress studies are generally carried out separately and results have never been combined or even critically compared, especially in terms of stress magnitudes, an important topic in Earth Sciences.

The KTB deep drill hole program has led to characterisation of the orientations and magnitudes of contemporary stresses to depth of 8 km (Brudy et al., 1997) and renewed considerations on the strength of the upper continental crust in intraplate settings have been proposed (Townend and Zoback, 2000; Zoback and Townend, 2001), emphasising the importance of hydrostatic fluid pressure in keeping the strength of the crust high. It is thus timely to try to compare these results with the increasing number of results on paleostress magnitudes gained in various settings worldwide. It can be expected that the combination of both independent approaches could be fruitful for the understanding of the rheology of the upper crust, provided that the concepts are similar or at least comparable and that the underlying mechanics is the same.

This paper aims at comparing currently available paleostress data with contemporary stress magnitudes in the continental crust. For this purpose, the most commonly used paleopiezometry techniques based on development of mechanical twins in calcite are first reviewed; then the pattern and the geological meaning of available estimates of paleodifferential stress magnitudes are discussed and compared to the stress-depth relationships derived from in situ stress measurements. The ultimate purpose is to demonstrate that both independent sets of data can be reconciled and combined to create a powerful tool for understanding crustal state of stress and rheology of the continental crust over time scales of several tens of Ma.

2. Concept of stress tensor vs. paleostress tensor

Defining a stress tensor in a rock material requires theoretically that an elementary representative volume (ERV) may be identified, which is the smallest volume for which there is equivalence between the continuum material and the real rock. The ERV is the physical representation of the mathematical point. The ERV is sufficiently small for its mechanical properties to be considered constant—a direct consequence is that there is no significant body force nor any resultant moment acting on this volume—but the ERV should, however, be sufficiently large to avoid inhomogneities found at the grain scale. A rock mass is assimilated to an equivalent continuum material and this equivalence holds for volume larger than the ERV.

Because rocks are heterogeneous, the concept of stress refers to mean forces per unit area for rock materials. The local stress tensor is defined (1) at any point in a rock mass by the components of the mean surface traction supported by the faces of the smallest cube which surrounds completely the ERV, and (2) at a given, instantaneous time. The complete determination of the local stress tensor requires the appraisal of six independent variables which correspond to the eigenvalues and the eigenvectors of the tensor (referred to as principal stress components σ_1, σ_2 and σ_3 and principal stress directions).

The determination of stress always involves the direct measurement of some intermediate quantity and the implicit or explicit application of some constitutive relation that relates the intermediate quantity to the stress. For example, the determination of stress commonly involves the measurement of a displacement in an elastic material and the determination of the stress from the equations of elasticity, either by calibration or by direct calculation. In nature, because the geological history of rock masses is usually complex hence never known precisely, because the constitutive equations describing the mechanical behaviour of rocks remain fairly approximate, and because the detailed structure of a rock mass cannot be determined exactly, it is impossible to evaluate by straight computation the natural stress field. For this reason, means have been developed to try to measure or to reconstruct it. The purpose of most of these measurements/reconstructions is to determine the local (contemporary or paleo-) stress tensor, or some of its components, irrespectively of the various mechanisms which may be its cause.

Whereas in situ stress measurements deal with the true concept of local stress tensor, the basic concept in paleostress reconstructions is that of a mean stress tensor. This means that the paleostress tensor considered is averaged over several thousands or even several millions years (the duration of a tectonic event) and over a given rock volume. Among the hypotheses is that the stress field is homogeneous throughout the volume of interest from which measurements are taken (especially, gradients such as gravitational stress are neglected). In practice, determining a mean stress tensor is equivalent to determining the mean stresses applied at the boundaries of the rock volume considered. For instance, the analysis of fault slip data leads to a stress tensor which is averaged on time and space; the analysis of calcite twins (Section 2) yields also a mean stress tensor averaged over the time span of a tectonic event, but the small size of samples required for performing the study makes the determined tensor closer to the theoretical stress tensor, i.e., defined at a point.

Another difference between contemporary stress determinations and paleostress reconstructions is that in situ stress measurements provide ‘‘snapshots’’ of ambient crustal stresses while paleostress tensors reflect crustal stresses at the particular time of tectonic deformation (internal rock deformation/ fault zone reactivation). One has therefore to draw attention on the actual meaning of a paleostress estimate (especially in terms of magnitude) which is based on deformation that may have occurred over many millions of years. In the following for instance, in the absence of dynamical or fluid-enhanced recrystallisation, paleopiezometry based on calcite twinning will be considered to provide estimates of the peak differential stress ($\sigma_1 - \sigma_3$) attained during a particular event of the tectonic history of the rock mass.

3. Determination of the local paleostress tensor components based on analysis of calcite twin data: principal stress orientations and relative stress magnitudes

Twinning of minerals depends on the magnitude of the shear stress which has been applied to them. It has been proposed to make use of this property for evaluating the stresses which have been supported by a rock during its history (e.g., Tullis, 1980). This may provide relevant information concerning the past stress field and the present stress field for recent deposits. Calcite is the most sensitive mineral for twinning and the most likely to provide useful tectonic information, especially in foreland settings where the outcropping formations are mainly sedimentary rocks.

Since the pioneering work of Turner (1953), several methods of stress analysis have been developed on the basis of calcite twin data (Jamison and Spang, 1976; Laurent et al., 1981; Laurent, 1984; Laurent et al., 1990; Etchecopar, 1984; Pfiffner and Burkhard, 1987; Nemcok et al., 1999). These methods are based on the widespread occurrence of *e*-twinning in calcite aggregates deformed at low pressure and temperature (Fig. 1). *E*-twinning occurs with a change of form of part of the host crystal by an approximation to simple shear in a particular sense and direction along specific crystallographically defined *e* planes {01−12}, in such a way that the resulting twinned portion of the crystal bears a mirrored crystallographic orientation to the untwinned portion across the twin plane.

Methods of strain/stress analysis based on calcite twins share the fundamental assumption that the measured twins formed in a homogeneous stress field and were not passively rotated after formation. These methods are best applied to very small strains (Fig. 1) that can be approximated by coaxial conditions (Burkhard, 1993); in this case, the orientation of small twinning strain can be reliably correlated with paleostress orientation. In addition, the *e*-twinning in calcite is not thermally activated and is not sensitive to either strain rate or confining pressure; Spiers (1979) has further shown that deformation is distributed inhomogeneously between grains while the stress is much more homogeneous at the scale of the aggregate. Calcite twinning therefore fulfils most requirements for paleopiezometry.

In contrast to Turner's (1953) method which only yields σ_1 and σ_3 orientations and to the technique of Jamison and Spang (1976) which only provides values of ($\sigma_1 - \sigma_3$) without any information on stress orientations and relative stress magnitudes (see Section 4.1.1), the computerised inversion of calcite twin data (stress inversion technique, SIT, summarised hereinafter) provides 5 parameters among the 6 of the complete stress tensor. It is to date the only technique which allows simultaneous calculation of principal stress orientations and differential stress magnitudes from a set of twin data, and which therefore allows to relate unambiguously differential stress magnitudes to a given stress orientation and stress regime.

The SIT assumes homogeneous state of stress at the grain scale and constant critical resolved shear stress (CRSS) for twinning τ_a. The inversion process is very similar to that used for fault slip data (Etchecopar, 1984), since twin gliding along the twinning direction within the twin plane is geometrically comparable to slip along a slickenside lineation within a fault plane. But the inversion process additionally takes into account both the twin planes oriented so that the resolved shear stress τ_s (the component of the shear stress along the twinning direction) was greater than τ_a (i.e., effectively twinned planes), and the twin planes oriented so that τ_s was lower than τ_a (i.e., untwinned planes), a major difference with inversion of fault slip data (Lacombe and Laurent, 1996) recently re-highlighted by Fry (2001). The inverse problem thus consists of finding the stress tensor that best fits the distribution of measured twinned and untwinned planes. This tensor must theoretically meet the major requirement that all the twinned planes consistent with it should sustain a resolved shear stress τ_s larger than that exerted on all the untwinned planes.

The inversion of slip (gliding) data along twin planes leads only to four parameters of the complete stress tensor T. These four parameters, which define the reduced stress tensor T′, are the orientations of the three principal stress axes and the stress ellipsoid shape ratio Φ [$\Phi = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$]. T′ is such that the complete stress tensor T is a function of T′: $\mathrm{T} = k\mathrm{T}' + l\mathbf{I}$, where k and l are scalars ($k = (\sigma_1 - \sigma_3) > 0$; $l = \sigma_3$) and $\mathbf{I}$ the unit matrix. The stress tensor solution is consequently searched as a reduced stress tensor T′ and, in addition, the maximum differential stress ($\sigma_1 - \sigma_3$) is scaled to 1 (Etchecopar, 1984). For this normalised tensor the resolved shear stress τ_s acting along any twin plane therefore varies between −0.5 and 0.5.

The first step of the inversion consists of obtaining a solution by applying a number of random tensors. For each tensor the stress components are calculated for all the twinned and the untwinned planes. However, because the resolved shear stress τ_s exerted on some untwinned planes may in practice be greater than that exerted along some twinned planes compatible with the tensor, the second step of the process consists of minimising the function, *f*, ideally equal to 0, defined as:

$$f = \sum_{j=1}^{N} \left(\tau_{sj} - \tau_{a'} \right)$$

where $\tau_{a'}$ is the smallest resolved shear stress applied on the twinned planes compatible with the tensor, and τ_{sj} are the resolved shear stresses applied on the N untwinned planes j such that $\tau_{sj} > \tau_{a'}$ (for more details, see Etchecopar, 1984). The $\tau_{a'}$ value is deduced from the inversion and corresponds to the CRSS for the normalised tensor used for calculation. The

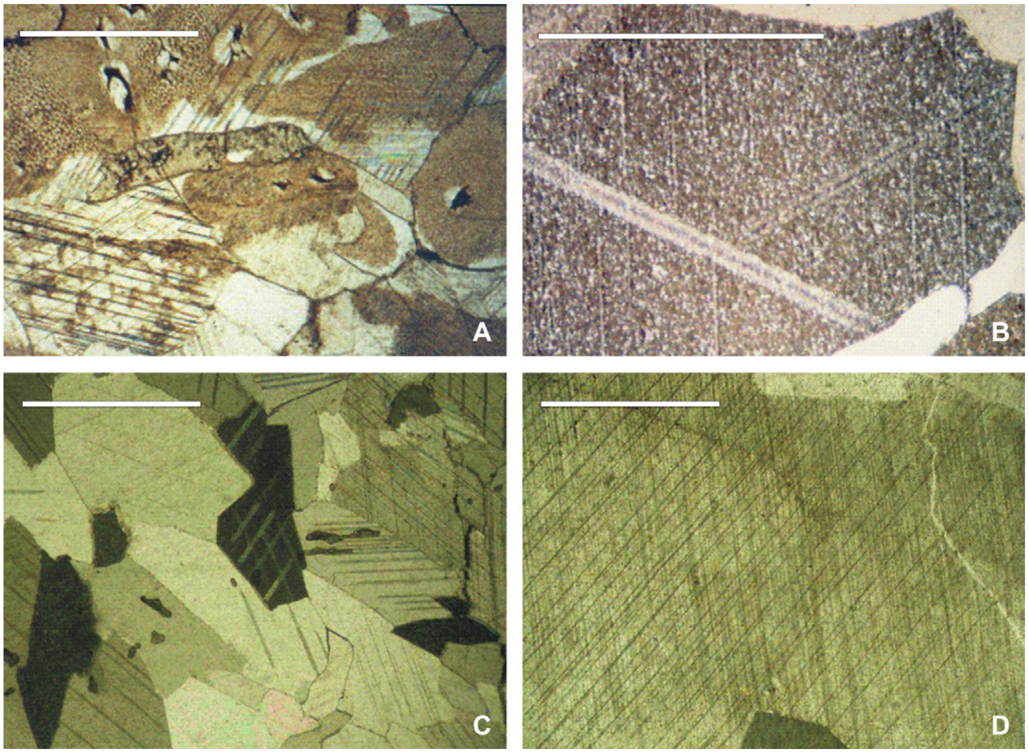

Fig. 1. Example of calcite twin material commonly used for inversion and paleostress estimates. (A) Twinned grains within coarse-grained limestone from southern France. (B) Close-up of a calcite grain twinned on its 3 potential e-twin planes. (C, D) Twinned grains within coarse-grained limestones from Iran displaying low twin density (C) and high twin density (D) (courtesy of Khalid Amrouch). Internal strain is achieved under a thin-twin regime. Scale bar is about 200−300 μm.

optimisation process leads to the reduced stress tensor solution that includes the largest number of twinned planes and simultaneously corresponds to the smallest value of f. The orientations of the three principal stresses σ_1, σ_2, and σ_3 are calculated, as well as the value of the Φ ratio that indicates the magnitude of σ_2 relative to σ_1 and σ_3. Numerous studies have demonstrated the potential of the twin inversion technique to derive regionally significant stress patterns even in polyphase tectonics settings (e.g., Lacombe et al., 1990, 1994; Rocher et al., 1996, 2000, 2004; and references therein).

4. How can absolute paleostress magnitudes be determined using calcite twins?

The study of absolute stress magnitudes in the crust is an important topic in Earth sciences, but to date our knowledge of the actual stress level sustained by the continental crust remains poor. Data on contemporary stress magnitudes in the crust are few (see Section 6), and inferring paleostress magnitudes from the development of natural geological structures is inherently difficult.

Paleopiezometry basically relies upon establishing a close relationship between the state of stress and the development of a conspicuous element in the rock itself and calibrating it experimentally. The first type of approach combines the inversion of fault slip data with rock mechanics data and/or determination of fluid pressures prevailing during brittle deformation. For instance, Bergerat et al. (1985), Reches et al. (1992) and Angelier (1989) used friction and/or failure envelopes to put additional constraints on the stress tensors derived from inversion of fault slip data. Lespinasse and Cathelineau (1995) then André et al. (2002) combined inversion of fault slip data and fluid inclusions data to derive principal stress magnitudes.

The alternative approach is based on the study of paleopiezometers such as dislocation density in calcite (e.g., Pfiffner, 1982), dynamic recrystallisation of calcite and quartz (e.g., Twiss, 1977; Weathers et al., 1979; Kohlstedt and Weathers, 1980), and mechanical twinning in calcite and dolomite (e.g., Jamison and Spang, 1976; Rowe and Rutter, 1990; Lacombe and Laurent, 1992, 1996). Because a great part (more than the half) of reported paleostress estimates are based on calcite twinning paleopiezometry (including the stress inversion technique) (Table 1), I summarise hereafter how differential stress magnitudes can be determined using calcite twinning, and further how combination with rock mechanics data may help constrain the complete stress tensor.

4.1. Quantifying differential stress magnitudes using calcite twins

4.1.1. Summary of the principles of previous estimates of differential stress magnitudes based on the techniques of Jamison and Spang (1976) and Rowe and Rutter (1990)

The basis of the widely used method of Jamison and Spang (1976) is that in a sample without any preferred crystallographic orientation, the relative percentages of grains twinned

Table 1
Characteristics of paleo-differential stress/depth data reported in Fig. 3A

Differential stress magnitude/Depth data	Location	Stress regime	(Paleo-) piezometric technique used	Value of fluid pressure	References
A	Causses, France	Strike-slip	Microfaults	Unknown	Rispoli and Vasseur, 1983
B1	Zagros, Iran (prefolding)	Reverse/Strike-slip	Calcite twins (SIT)	Unknown	Amrouch et al., 2005
B2	Zagros, Iran (postfolding)	Reverse/Strike-slip	Calcite twins (SIT)	Unknown	Amrouch et al., 2005
C	SW Taiwan Foothills	Strike-slip	Calcite twins (SIT)	Unknown	Lacombe, 2001
D	Appalachian Plateau, USA	Strike-slip	Dislocation density in calcite	Unknown	Engelder, 1982
E	Germany	Strike-slip/Reverse	Microfaults/Frictional criterion	0.4–1.4 MPa	Bergerat et al., 1985
F	Appalachian Plateau, USA	Strike-slip	Residual stresses measured *in situ*	Unknown	Engelder and Geiser, 1984
G	W Taiwan Foothills (prefolding)	Reverse/Strike-slip	Calcite twins (SIT)	Unknown	Lacombe, 2001
H	Burgundy, France	Strike-slip	Calcite twins (SIT)	Unknown	Lacombe and Laurent, 1992
I	Quercy, France	Strike-slip/Reverse	Calcite twins (SIT)	Unknown	Tourneret and Laurent, 1990
J	Provence, France	Strike-slip/Reverse	Calcite twins (SIT)	Unknown	Lacombe et al., 1991
K	W Taiwan Foothills postfolding	Strike-slip/Reverse	Calcite twins (SIT)	Unknown	Lacombe, 2001
L	Morocco	Strike-slip	Microfaults/Frictional criterion	Unknown	Petit, 1976
M	Paris Basin, France	Strike-slip	Calcite twins (SIT)	Unknown	Lacombe et al., 1994
N1	Subalpine Chain, France	Reverse	Calcite twins (JS)	Unknown	Ferrill, 1998
N2			Calcite twins (RR)		
O	Cantabrian Zone, Spain	Reverse	Calcite twins (RR)	Unknown	Rowe and Rutter, 1990
P	Moine Thrust, Scotland	Reverse	Quartz recrystallised grain size	Unknown	Christie, 1963; Twiss, 1977
Q	Massif Central, France	Strike-slip	Microfaults/Frictional criterion	50 ± 10 MPa ~hydrostatic	Lespinasse and Cathelineau, 1995
R	Pyrenees, Spain	Reverse	Calcite twins (JS)	Unknown	Holl and Anastasio, 1995
S	McConnel Thrust Rockies, Canada	Reverse	Calcite twins (JS)	Unknown	Jamison and Spang, 1976
T	Glarus Thrust, Alps, Switzerland	Reverse	Dislocation density in calcite	Unknown	Pfiffner, 1982
U	Infrahelvetic complex, Alps Switzerland	Reverse	Dislocation density in calcite	Unknown	Pfiffner, 1982
V	Shear Zone, Colorado, USA	Strike-slip?	Quartz recrystallised grain size—dislocation density in quartz	Unknown	Kohlstedt and Weathers, 1980
W	Pioneer Landing Fault Zone, USA	Reverse	Dolomite twins (JS/RR)	Unknown	Newman, 1994
X	Shear Zone, Wyoming, USA	Strike-slip?	Dislocation density in quartz	Unknown	Weathers et al., 1979
Y1, Y2	Lorraine, France	Strike-slip	Calcite twins (SIT)	Unknown	Rocher et al., 2004

on 0, 1, 2 or 3 twin plane(s) depend on the applied $(\sigma_1 - \sigma_3)$ value. Since this relationship has been experimentally calibrated, knowing these relative percentages in a sample, and under the hypothesis of a constant critical shear stress value for twinning, the order of magnitude of $(\sigma_1 - \sigma_3)$ can be estimated. Among the limitations of this method is that it does not take into account the grain size dependence of twinning, does not check the mutual compatibility of measured twin systems and does not allow to relate the differential stress estimates to a given stress regime since principal stress orientations are not determined.

The method of Rowe and Rutter (1990) is based on the sensitivity of the twinning incidence, twin volume fraction and twin density to differential stress; in turn, estimates of these parameters are used to infer differential stress magnitudes. The first two criteria are found to be largely dependent on grain size, since twinning is easier in large grains. The twin density (number of twins per mm) is poorly grain-size dependent. The use of this last criterion provides reasonable results of $(\sigma_1 - \sigma_3)$ when applied at high temperature, but leads to overestimates of differential stresses when applied at a low-temperature twinning deformation (see discussion in Ferrill, 1998). As a result, the Rowe and Rutter paleopiezometer best applies to twinning deformation at high temperatures (between 200 and 800 °C), and at large strains (7–30%), and is not appropriate for evaluating differential stress magnitudes from calcite twinning analysis in outer parts of orogens. In addition, this method shares the same limitation than the Jamison and Spang method in not checking the mutual consistency of twin sets from which differential stress values are derived and not allowing to relate differential stress estimates to a given state of stress.

4.1.2. Quantifying differential stress magnitudes using inversion of calcite twin data

The four parameters defining the reduced stress tensor T′ (orientation of principal stress axes and Φ ratio) being derived from inversion of calcite twin data (section 2), quantifying the deviatoric stress tensor T_D defined as:

$$T_D = T - [(\sigma_1 + \sigma_2 + \sigma_3)/3]\mathbf{I}$$

requires the determination of a fifth parameter of the complete stress tensor: the scalar k ($= \sigma_1 - \sigma_3$). This determination relies on the existence of a constant CRSS for twinning τ_a and on the accurate estimate of $\tau_{a'}$ that corresponds to the normalised value of the CRSS for the reduced stress tensor used for calculation.

4.1.2.1. Assumption of a constant Critical Resolved Shear Stress for twinning. Differential stress estimates using the twin inversion technique are based on a constant CRSS τ_a for twinning. This assumption is also shared by the technique of Jamison and Spang (1976). Using this technique, a constant CRSS of 10 MPa is generally adopted (e.g., Craddock et al., 1993, 2002; Craddock and Van Der Pluijm, 1999; Gonzales-Casado and Garcia-Cuevas, 1999). Based on the analysis of experimentally deformed samples, Lacombe and Laurent (1996) and Laurent et al. (2000) have demonstrated the reliability of the twin inversion technique, but also emphasised that the CRSS is sensitive to strain hardening. Their results suggest that the CRSS for twinning can be considered constant for samples displaying a nearly homogeneous grain size. For samples displaying a mean grain size of ∼200–300 μm and deformed between 0 and 100 °C at 2–2.5% strain, it equals 10 MPa; for the same samples deformed at nearly 1–1.5% strain, the CRSS rather equals 5 MPa (Lacombe, 2001). For samples displaying different grain sizes, a crystal size-CRSS relationship such as that proposed by Rocher et al. (2004) can be further adopted to improve differential stress estimates.

4.1.2.2. Determination of differential stress magnitudes. Under the assumption of a constant CRSS for twinning τ_a, the differential stress magnitudes can be determined as follows (Etchecopar, 1984; Lacombe and Laurent, 1996; Laurent et al., 2000):

$$(\sigma_1 - \sigma_3) = \tau_a/\tau_{a'}$$

and

$$(\sigma_2 - \sigma_3) = \Phi(\sigma_1 - \sigma_3)$$

where $\tau_{a'}$ is the smallest resolved shear stress applied on the twinned planes accounted for by the stress tensor and therefore the normalised value of the CRSS when $(\sigma_1 - \sigma_3)$ is scaled to 1. As for other techniques of paleostress estimation based on calcite twin analysis (e.g., Jamison and Spang, 1976; Rowe and Rutter, 1990), the paleopiezometric technique used herein yields the maximum differential stress $(\sigma_1 - \sigma_3)$ related to a given paleostress orientation, because the differential stresses are computed by taking into account the maximum percentage of twinned planes consistent with the tensor and therefore the smallest $\tau_{a'}$ value. In the absence of recrystallisation, the meaning of such differential stress estimates is therefore that of the peak stresses sustained by rocks during a given episode of their tectonic history.

In order to prevent bias due for instance to local record of stress concentrations, which may lead to overestimated stresses which may be not representative of the far-field stress of interest, several samples are usually collected for a given locality and one usually retains the weighted mean of the differential stress values as the most reliable estimate. At all localities, limestone samples are collected away from major fault zones where the stress field is known to be very inhomogeneous in both orientation and magnitude. The calcite twin-based paleostress estimates therefore meet the assumptions of stress homogeneity and low-finite strain and are likely representative of stresses at larger scale.

4.2. Quantifying absolute stress magnitudes and estimating the depth at which peak differential stresses recorded by calcite twinning prevailed

Knowing the deviatoric stress tensor, a single parameter is missing in defining the complete stress tensor. This sixth parameter corresponds to the isotropic component of the tensor that cannot be determined using calcite twinning only since twinning does not depend on isotropic stress. In order to assess the actual magnitudes of σ_1, σ_2 and σ_3, it is thus necessary to fix directly one of them, or at least to determine a third additional relationship between them.

The method used herein to determine the complete stress tensor relies upon combination of analyses of calcite twins and rock mechanics data (Lacombe and Laurent, 1992). For a given tectonic event and a given site, it consists of finding the values of σ_1, σ_2, and σ_3 required for consistency between newly formed faulting, frictional sliding along pre-existing planes, and calcite twinning.

The differential stress value $(\sigma_1 - \sigma_3)$ determined from calcite twins fixes the scale (i.e., the diameter) of the Mohr circle associated with the tensor. To completely describe the stress regime (orientation and magnitude), the isotropic component of the tensor is missing. The determination of this parameter, which corresponds to the position of the Mohr circle along the normal stress axis in the Mohr diagram, can be done in the following way (Fig. 2).

4.2.1. Failure

If subsets of newly formed faults and twins provide similar reduced stress tensors in a given site (description of stress inversion applied to fault slip data is out of the scope of this paper, see for instance Angelier, 1989), it can be reasonably inferred that they formed during the same tectonic event, and therefore they may be thought to have developed

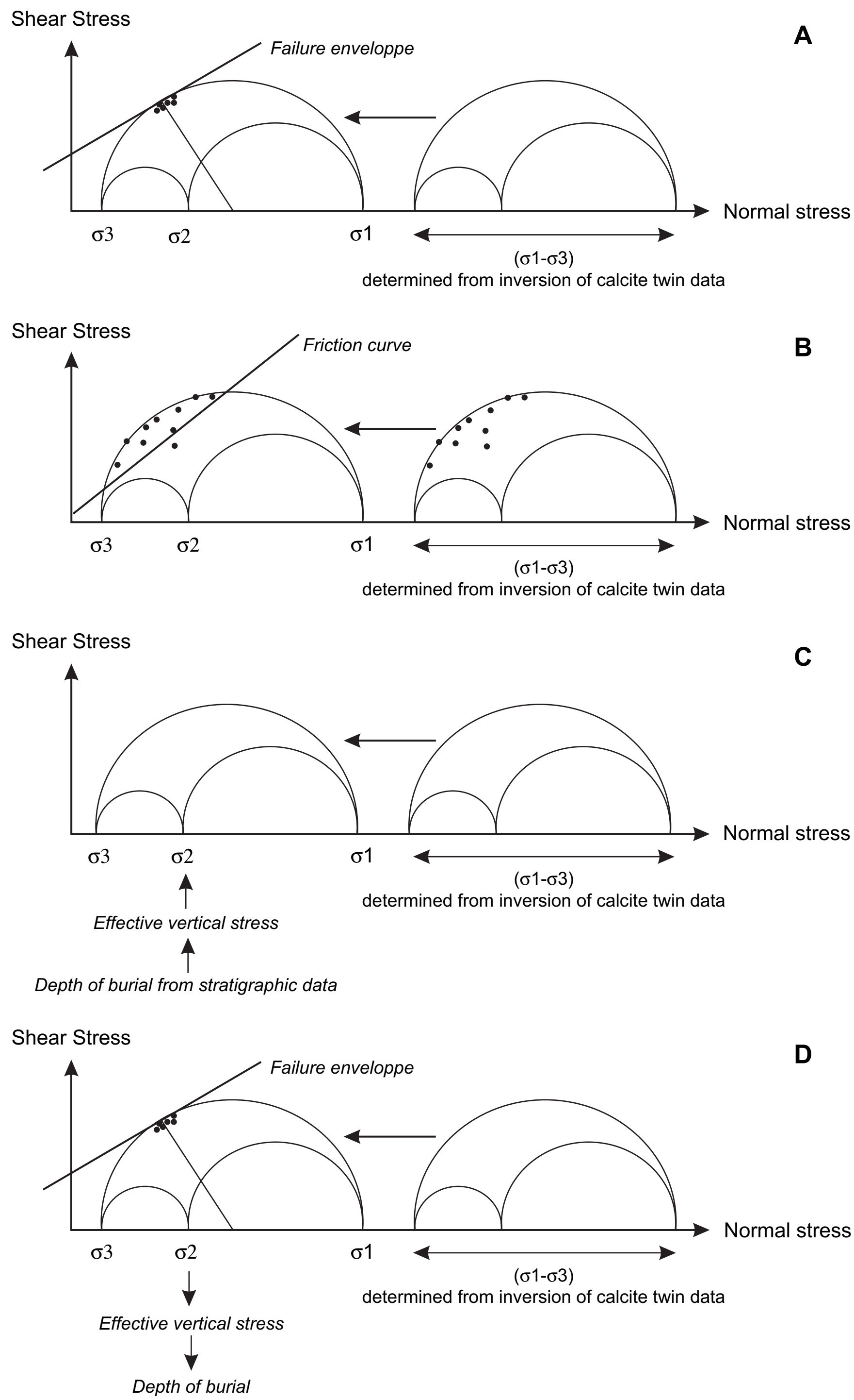

Fig. 2. Principle of determination of principal stress magnitudes based on combination of inversion of calcite twin data with rock mechanics data. (A) Calcite twins, newly formed faults and fit with the fresh failure envelope. (B) Calcite twins, reactivated inherited faults and fit with friction curve. (C) Calcite twins Paleodepth estimate from stratigraphic data and determination of the magnitude of the vertical principal stress. (D) The use of above-mentioned failure or friction criterion fixes the magnitude of the vertical principal stress, which allows estimate of paleodepth of deformation. Hydrostatic fluid pressure is assumed.

"contemporaneously" in terms of geologic time. As calcite twins record the "peak" stress reached during the entire history of the rock mass, the maximum recorded differential stress should correspond to the stress which induced rock failure, because fresh failure releases stress. Consequently, if new faults developed, fitting the (σ_3,σ_1) Mohr circle determined from calcite twins with the intrinsic failure (or crack development) curve yields values of principal stresses that prevailed just before rupture (Fig. 2A).

4.2.2. Friction

Because all points that represent reactivated pre-existing planes should lie above the friction curve (Byerlee, 1978) in the Mohr diagram, fitting the (σ_3,σ_1) Mohr circle determined from calcite twins with the friction line, so that inherited faults are located above the friction line, allows determination of the principal stress values. This requirement imposes a constraint on the position of the Mohr circle along the normal stress axis and yields principal stress magnitudes (Fig. 2B). Note that this may apparently contradict Byerlee's law since each individual reactivated pre-existing plane should lie on (and not above) the friction curve since a fault plane cannot sustain a shear stress/normal stress ratio greater than that required for frictional sliding on it. It is simply a consequence of considering a mean stress tensor, as previously discussed in Section 1.

4.2.3. Evaluation of the vertical principal stress

Most of paleostress reconstructions based on calcite twins (and fault slips: e.g., Lacombe et al., 1990; Lisle et al., 2006) yield a highly plunging principal stress axis, close to vertical (within the usual 10° range of uncertainties), provided that passively rotated stress axes due to folding are interpreted after backtilting to their prefolding attitude. The verticality of one principal stress is also indicated by widespread joints in foreland environments which are very close to vertical, a condition that is most likely if one of the principal stresses is vertical (Engelder and Geiser, 1980). Considering a vertical principal stress is also common in most in situ measurements of contemporary stresses. This assumption was assessed by McGarr and Gay (1978), who concluded that it is basically valid although slight departures from this rule are common. Most of the data reported by these authors were obtained in mines "often in regions of complex geology" and thus they expected that orientations of stress in other areas might conform more closely to this assumption. In their compilations of different types of in situ stress indicators in of North America, Zoback and Zoback (1980) made similar arguments which were later extended to the global compilation of in situ stress indicators (Zoback et al., 1989; Zoback, 1992). Further indications for this assumption to be valid are given by the analysis of focal plane solutions (e.g., Angelier, 2002) at various sites which show that one of the principal stresses is often within a few degrees of vertical. Brudy et al. (1997) also recently concluded that the assumption was valid at the KTB site. Cornet (1993), however, points out that the vertical direction is not a principal direction in the vicinity of the ground surface in mountainous area or, more generally, anywhere the ground surface is not horizontal. The question then arises of determining the depth up to which this non-verticality of one of the principal stresses may be observed. When the vertical direction is not a principal direction, assuming that one principal component of the local stress tensor is equal to the weight of overburden is of course wrong.

As mentioned before, paleostress reconstructions deal with stresses associated with tectonic deformation of rocks at depth, hence generally of larger magnitudes and of longer duration (the duration of a "geological tectonic event") than "snapshots" of ambient stresses provided by in situ measurements. These paleostresses are consequently "averaged" over several millions years, so they are controlled to the first order by the vertical (gravity) and the horizontal (tectonic forces) directions and do not reflect local and/or temporal sources of stress perturbation as for instance the evolving-with-time topography can be. As a result, it is generally correct to equal the magnitude of the vertical principal stress to the overburden load, ($\rho_{gz} - P_f$) where ρ is the average density of the overburden. g is the acceleration of gravity, z is the depth, and P_f is the fluid pressure. Paleodepth of overburden can be evaluated using stratigraphic data in favourable settings (Fig. 2C).The actual pore fluid pressure at the time of deformation (and the porosity as well) being unfortunately often out of reach, hydrostatic conditions are usually adopted as the most realistic conditions of fluid pressure. Results obtained in deep boreholes suggest that this assumption is reasonable and justified in most cases (see Section 6).

As the inversion of calcite twin data provides directly differential stress magnitudes, the determination of the principal stress magnitudes only requires one piece of information among the above criteria. If the failure or the friction criterion is available, the magnitude of the vertical principal stress is fixed. If the burial thickness is unknown, this estimate of the value of the (effective) vertical stress can be used to estimate the weight of overburden, and therefore the depth at the time of deformation (Fig. 2D). This estimate can be compared with stratigraphic information (if available) in order to check for consistency. Such a procedure has been applied in Taiwan to derive Plio-Quaternary principal stress magnitudes in the Western Foothills (Lacombe, 2001).

5. Evolution with depth of paleo-differential stress magnitudes

5.1. On the difficulty of establishing a paleostress/paleodepth relationship

Collecting data on contemporary stress and paleostress magnitudes with depth is fundamentally different. In drill holes, contemporary stresses are determined directly at a given depth, or at least in a narrow depth interval. In contrast, paleopiezometers are generally sampled and analysed after they have reached the surface, i.e., after exhumation from an unknown depth, and establishing a differential stress versus depth relationship for paleostresses requires independent determination of differential stress and depth. In addition, in case of polyphase tectonism, deciphering the stress/depth evolution requires to unambiguously relate a differential stress value to both a depth of deformation and a tectonic event. As mentioned in Section 3, inversion of calcite twin data provides an efficient tool to perform such estimates in favourable cases.

Paleodepth estimates are usually derived from stratigraphic/sedimentological studies in forelands or even fold-thrust belts, from paleothermometry coupled with considerations on paleothermal gradient, recrystallisation accompanying ductile deformation or fabric development, or from metamorphism. Uncertainties on the stress versus depth relationship of interest are partly due to uncertainties on depth estimates.

Second, depending on the geological setting studied (forelands, fold-thrust belts or fault zones in metamorphic belts), the paleopiezometric methods used are different (e.g., analysis of dislocation density in calcite, dynamic recrystallisation of calcite and quartz or mechanical twinning in calcite: Table 1). The poor knowledge we have on the paleostress levels sustained by natural rocks is partially linked to the fact that the paleopiezometric techniques do not share the same limitations, each of them having particular conditions of application; this leads to multiple sources of methodological uncertainties which are superimposed to the variability of natural phenomena. This is illustrated by a method dependence of estimates (e.g., Ferrill, 1998), which additionally group according to regional versus fault zones studies (e.g., Newman, 1994).

5.2. To what extent do paleostress magnitudes derived from paleopiezometry in the sedimentary cover reflect the (frictional) stress gradient in the deeper crystalline basement?

Many actively deforming forelands are earthquake deficient. This is particularly true in the upper 2–4 km in the sedimentary cover where folds develop mainly by mechanisms including diffusion-mass transfer and other ductile mechanisms such as calcite twinning. In this depth range (the upper-tier of the ductile flow regime in the sense of Engelder, 1993), the so-called brittle crust creeps by ductile mechanisms. These mechanisms relieve stresses, so the differential stress level is kept beyond the frictional yield that likely prevails at greater depth in the basement (Section 6). This is particularly true in many settings where the cover is detached from basement.

The way differential stresses evolve with depth when crossing a weak decollement within or at the base of the cover, and more generally when crossing a rheological contrast induced by a local structural or lithological inhomogeneity also requires attention. Cornet and Burlet (1992) reported for instance such a contemporary stress field discontinuity at the Echassieres site (France), related to a fault zone displaying breccia and a high kaolinite content.

Since the stress/depth gradient above and below the decollement can possibly be different, the amount of coupling between cover and basement may at least partially control the level of differential stresses sustained by cover rocks. If coupling is low (e.g., efficient, weak decollement), paleostress magnitudes derived from paleopiezometry in the cover may not reliably reflect deeper crustal stresses in that the stress gradient can change when crossing the decollement. When basement/cover coupling is more efficient, as in stable forelands of orogens and plate interiors despite the common widespread occurrence of weak shale beds in basins, the stress gradient derived from cover paleopiezometry can be extended with more confidence to the crust.

5.3. Evolution with depth of paleo-differential stress magnitudes

The paleo-differential stress magnitude data reported in Fig. 3A correspond either to reverse stress regimes or to strike-slip stress regimes (Table 1), sometimes to mixed reverse/strike-slip stress regimes (subhorizontal σ_1 axis and low Φ values). The plot of these paleo-differential stress data against depth suggests a trend of increasing differential stresses with depth. In an attempt at interpreting this stress-depth relationship in a more precise way, friction curves were built and reported in Fig. 3A. Because the frictional strength of a faulted rock mass depends on pore pressure (Hubbert and Rubey, 1959), estimates of the frictional strength of the brittle crust depend on the pore pressure at depth (Brace and Kohlstedt, 1980). Hence, for given stress and pore pressure regimes, one can estimate the maximum differential stress expected on the basis of the frictional-failure equilibrium hypothesis, and compare this with observations. For a favourably oriented pre-existing cohesionless fault plane, the condition of reactivation, which therefore applies to a critically stressed crust, can be written as follows (Jaeger and Cook, 1969):

$$(\sigma_1 - P_f)/(\sigma_3 - P_f) = [(\mu^2+1)^{0.5}+\mu]^2$$

This equation can be used to predict differential stress as a function of depth in a crust in frictional equilibrium. Because most available paleostress magnitude data come from compressional settings (i.e., with the largest principal stress σ_1 being horizontal, so that states of stress are of strike-slip (σ_2 vertical) or reverse-type (σ_3 vertical) regimes: Anderson, 1951), the following equations for strike-slip and reverse-type regimes have been considered:

$$\sigma_1 - \sigma_3 = 2\rho_{gz}(\lambda - 1)(1 - [(\mu^2+1)^{0.5}+\mu]^2) / (1 + [(\mu^2+1)^{0.5}+\mu]^2)$$

and

$$\sigma_1 - \sigma_3 = \rho_{gz}(\lambda - 1)(1 - [(\mu^2+1)^{0.5}+\mu]^2)$$

with $\lambda = P_f/\rho_{gz}$. λ is close to 0.4 for hydrostatic conditions, close to 0.9 for nearly lithostatic conditions and equals 0 for dry conditions. The curves corresponding to strike-slip (SS) and reverse faulting (C) regimes for values of λ of 0.38 (hydrostatic) and 0 (dry) and for values of the friction coefficient μ of 0.6 and 0.9 have been drawn accordingly on Fig. 3A.

Different stress–depth evolutionary trends can be distinguished. A large set of paleostress data more or less fit the theoretical curves of frictional-failure equilibrium corresponding to strike-slip (SS) and reverse faulting (C) regimes under nearly hydrostatic fluid pressure ($\lambda \sim 0.38$) and friction coefficient μ between 0.6 and 1 (I and II on Fig. 3A). These data therefore support a critically stressed crust under hydrostatic conditions. Some data however reflect stress levels down to ~7 km depth beyond the frictional limit (III on Fig. 3A), which suggests possible occurrence of some creep in the crust which therefore relieves stresses and keeps the differential stress level beyond the frictional yield (upper-tier of the ductile flow regime: Engelder, 1993). Below 10 km, some stress data lie also under the frictional limit, and presumably reflect stresses at or just below the brittle-ductile transition (IV).

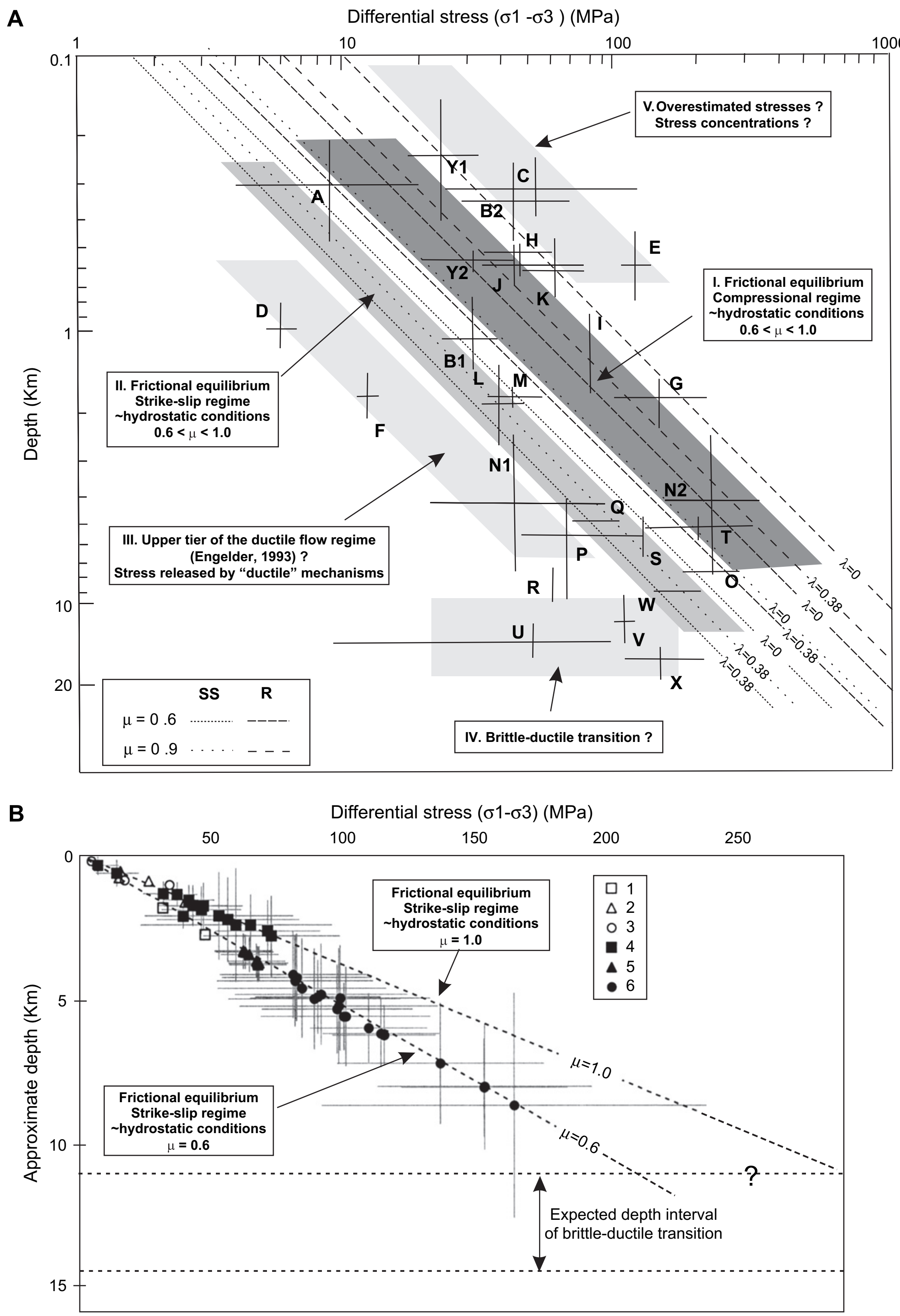

Fig. 3. (A) Differential stress values derived from paleopiezometry versus depth relationship (log-log). Modified and completed after Davis and Engelder, 1985; Engelder, 1993; Ferrill, 1998; Lacombe, 2001. For details concerning stress/depth data, refer to Table 1. Calcite twin paleopiezometry: RR, Rowe and Rutter technique; JS, Jamison and Spang technique; SIT, Stress Inversion technique (Etchecopar, 1984. completed by Lacombe and Laurent, 1992 for the determination of the complete stress tensor). Dashed lines illustrate stress–depth relationships predicted using Coulomb frictional-failure theory for coefficients of friction μ of 0.6 and 0.9, pore pressures of 0 ($\lambda = 0$) and hydrostatic ($\lambda = 0.38$) and various tectonic regimes (SS, strike-slip; R, reverse). (B) Differential stress versus depth relationship derived from in situ stress measurements (after Townend and Zoback, 2000; Zoback and Townend, 2001). 1, Fenton Hill; 2, Cornwall; 3, Dixie Valley; 4, Cajon Pass; 5, Siljan; 6, KTB. Site Dixie Valley reveals a normal stress regime; all other sites correspond to a strike-slip stress regime. Dashed lines illustrate stress–depth relationships predicted using Coulomb frictional-failure theory for coefficients of friction of 0.6 and 1.0.

Finally, some differential stress data lie above the frictional limit for the compressional regime (V) and cannot be easily interpreted. It can be suspected that these data reflect overestimated stresses: this is the case for the estimates by Bergerat et al. (1985) (datum E on Fig. 3A).

6. Comparison with the evolution with depth of contemporary differential stress magnitudes. Implications for the strength and behaviour of the continental crust

6.1. What we know about the present-day differential stress magnitudes at depth

The most widely used model of crustal stress and strength is based on the assumption that the lithosphere is in failure equilibrium with the state of stress in the upper crust controlled by its frictional strength (Byerlee, 1978). Evidence supporting that intraplate continental crust is in (or at least close to) a state of failure equilibrium are derived from studies of seismicity and contemporary stresses. They have been discussed by Townend and Zoback (2000); they are: (1) the widespread occurrence of seismicity induced by either reservoir impoundment (Simpson et al., 1988; Roeloffs, 1996) or fluid injection (e.g., Pine et al., 1983; Zoback and Harjes, 1997), (2) earthquakes triggered by other earthquakes (Stein, 1999), and (3) in situ stress measurements in deep wells and boreholes. Data on in situ stress magnitude measurements come mainly from intraplate regions where the state of stress is generally of strike-slip or compressional type: United States, South Africa (McGarr and Gay, 1978); Scandinavia, central Europe (Rummel et al., 1986); Monticello, South Carolina (Zoback and Hickman, 1982); Cornwall, England (Pine et al., 1983); Yucca Mountain, Nevada (Stock et al., 1985); Fenton Hill, New Mexico (Barton et al., 1988); Cajon Pass, California (Zoback and Healy, 1992; Vernik and Zoback, 1992); KTB, Germany (Brudy et al., 1997; Zoback and Harjes, 1997).

Application of Coulomb faulting theory with laboratory-derived coefficients of friction (e.g., Byerlee, 1978) allows prediction of critical stress levels in reverse, strike-slip, and normal faulting environments as a function of depth and pore pressure (e.g., Hubbert and Rubey, 1959; Sibson, 1974; Jaeger and Cook, 1969; Brace and Kohlstedt, 1980). Fig. 3B shows the data compiled by Townend and Zoback (2000) together with the theoretical curves for a critically stressed crust under hydrostatic conditions. Most of these in situ stress measurements are consistent with Coulomb frictional-failure theory incorporating laboratory-derived frictional coefficients, μ, of 0.6—1.0 and hydrostatic fluid pressure for a strike-slip stress regime. The crust's brittle strength is quite high (hundreds of mega-pascals) under conditions of hydrostatic pore pressure. Note, however, that the stress/depth gradient depends explicitly on the stress configuration, i.e., normal, strike-slip or reverse stress regime.

However, a complete stress profile taken from a three-well experiment on the Appalachian Plateau (Evans et al., 1989) shows stresses well under the frictional limit. This suggests that ductile mechanisms are possibly currently acting in some settings, that they relieve stress and keep the crustal state of stress beyond the frictional limit. As a result, friction likely governs in numerous settings, but it is probably not a general rule for the entire upper crust, especially its uppermost part (the detached cover).

A second arising point is that most data reported in Fig. 3B were collected in intraplate settings where a strike-slip type state of stress prevails. Therefore, collecting new stress data is crucial to confirm the general validity of both the applicability of laboratory derived strength parameters to the crust Kohlstedt et al., 1995 and the assumption that the crust is in a state of failure equilibrium in many different settings. Moreover, because evidence comes only from in situ measurements which are essentially shallower than 8 km, outstanding questions about crustal stress magnitudes are whether the findings at this maximum depth reached by boreholes can be extended to midcrustal depths approaching or reaching the brittle-ductile transition.

6.2. Comparison of the evolution with depth of paleo- and contemporary differential stress magnitudes, and discussion

Compared to contemporary stress magnitudes reported on Fig. 3B (Townend and Zoback, 2000), estimates of paleostress magnitudes reported in Fig. 3A show a large dispersion, although an increase of differential stresses with depth is clearly depicted. There are many reasons to explain this dispersion. It is mainly due to methodological uncertainties and to the variety of sampled natural settings.

In contrast to most in situ measurements carried out at depths shallower than 8 km in intraplate setting where a strike-slip type state of stress prevails, paleostress estimates are of mixed origins, combining magnitudes from strike-slip and compressional (reverse) regimes (Table 1); in addition, the usual collection of many paleostress magnitude data not only away, but also close to, or within, orogens (and therefore close to fault zones/past plate boundaries) may increase the dispersion. Finally, the hypothesis of hydrostatic pore pressure usually adopted for paleodepth estimate may not be valid in some particular compression zones where it may be closer to lithostatic than to hydrostatic.

However, it is reasonable to consider that, at the present-day state of our knowledge and with the available set of data, most contemporary stress and paleostress data support a first-order frictional behaviour of the upper continental crust. The mechanical strength of the brittle crust, away from (but also possibly close to) plate boundaries, is in numerous settings governed by frictional sliding on pre-existing favourably oriented faults. The critically stressed upper continental crust is therefore able to sustain differential stresses as large as 150—200 MPa, so its strength makes it able to transmit a significant part of orogenic stresses from the plate boundary across the far foreland (Townend and Zoback, 2000; Lacombe, 2001; Zoback and Townend, 2001).

Theoretical strength profiles for the crust basically separate the crust into an upper brittle and lower ductile regime. Whereas the strength in the brittle part is controlled by the frictional strength of pre-existing, favourably oriented faults, and varies with the tectonic regime (Sibson, 1974; see above), the strength of the lower ductile part is described by flow laws for appropriate rock types, temperatures, and strain rates. The transition between these zones is usually depicted as the intersection of the two strength envelopes, but the transition presumably occurs over an interval of several kilometres as the rheology of the material gradually changes from brittle to ductile (e.g., Scholz, 1990). Commonly, the 350 °C isotherm is thought to be the onset of plasticity in rocks of gneissic and granitic composition. For a temperature gradient of 27–30°/km, this temperature corresponds to a depth of 11–13 km. There is no present-day stress information of stress magnitudes in the 10–15 km depth interval, although this would provide a very useful complement to available stress data. Estimates of regional paleostress magnitudes corresponding to different depths of deformation collected at various places worldwide, especially including fault zone studies in metamorphic belts where depth of deformation presumably has approached or reached the brittle–ductile transition, such as those reported on Fig. 3A, are therefore a way to extend our knowledge of stress/depth relationship at depth.

Models as well as simple extrapolation of the stress/depth gradient from shallower depths suggest that stresses at mid-crustal depth should be of the order of hundreds of MPa for near-hydrostatic pore pressure. Despite inhomogeneities of stress magnitudes which may expectedly arise in fault zones due to the high finite strain of the rocks (grain-size reduction, rotational deformation at the grain scale or dynamic or fluid-enhanced recrystallisation may influence paleopiezometry) and the record of local effects such as stress concentrations, the reported (unfortunately few) paleostress estimates at the brittle-ductile transition values appear to be lower than those expected by simply extrapolating frictional gradients (Fig. 3A). Data are too few to unambiguously argue whether these lower stress values reflect a lower strength related to the onset of the crustal ductile behaviour at the place and at the depth where deformation within the fault zones occurred, or if an stress-depth gradient characteristic of the upper crust can be extrapolated with confidence below 11 km. A fluctuating brittle-ductile transition (350–450 °C?) must be defined in order to take into account the variable geotherm and therefore the settings where stress estimates in fault zones have been performed. Taking into account such a fluctuating BDT, and although further studies are needed, one can safely consider that to a first order, available paleostress and stress magnitudes, even at mid-crustal depths, are more or less in agreement with laboratory-derived friction laws.

As a result, at the present-day of our knowledge, independent contemporary stress and paleostress data support that upper crustal stresses are/were generally probably close to, and in all cases limited by, frictional faulting equilibrium. In settings where friction clearly governs stress, this implies uniform stress differences at a given depth and a given stress regime, regardless of the ''intensity'' of deformation, the style of deformation being probably simply a function of the strain rate.

One has, however, to keep in mind that some data nevertheless indicate paleostress and contemporary stress conditions beyond the frictional limit, suggesting that stresses may be locally relieved by ductile mechanisms (upper tier of the ductile flow regime: Engelder, 1993), probably in the uppermost cover part of the crust. These ductile mechanisms, such as diffusion-mass transfer, may prevent the stress level to reach the frictional limit, lowering the seismogenic potential of the (detached) uppermost sedimentary cover.

Another argument that may support that crustal stresses are not always at the frictional yield comes from the works by Craddock et al. (1993) and Van der Pluijm et al. (1997). These authors have documented a fall-off in differential stresses with distance from the orogenic front in forelands of various orogens. Because no information about the depth at which paleo-differential stresses were evaluated is available, there is no way to normalise each individual stress datum by the depth at which twin deformation occurred, so this fall-off may well be an artefact of the different depths at which paleostresses were determined. However, this decay of differential stresses is observed within the so-called undeformed foreland away from the deformation front; so one can infer that because no significant orogen-related deformation and therefore no large lateral variations of erosion and exhumation have occurred in the foreland, rocks presently at the surface and from which measurements are taken were likely deformed at nearly similar depths. The absence of orogen-related large deformation in the foreland indicate that far-field orogenic differential stresses presumably did not reach the stress conditions required for frictional faulting. If this decreasing trend in differential magnitudes away from the deformation front does occur and has a geological meaning in terms of stress attenuation as stated by the authors, it indicates that lateral stress gradients may occur in the upper crust; only ductile flow mechanisms characteristic of the upper-tier of the ductile flow regime are able to permit such stress gradients below the frictional stress limit.

7. Conclusion

Quantitative estimates of crustal stresses and strength are central to many problems of lithospheric mechanics. This paper presents a first attempt at comparing and combining paleostress magnitude data with contemporary stress data.

Concepts and techniques underlying determinations of contemporary stresses and paleostresses are inherently different, and both types of stress data do not have strictly the same geological meaning: contemporary stresses measured in situ reflect local, instantaneous ambient crustal stresses, while reconstructed paleostresses reflect ancient crustal stresses at the particular time of tectonic deformation, averaged over the duration of a tectonic event and over a given rock volume. Although to this respect contemporary stresses and paleostresses are not directly comparable, their analyses however rely on the same mechanics, and they constitute complementary stress data sets.

Combination of both types of stress data provides new constraints on the differential stress gradients with depth, which are to date still poorly known. Combining contemporary and paleostress data allows us to extend our stress/depth database in various settings, i.e., away horizontally from drill holes, and vertically by obtaining information on stress magnitudes at depth more or less continuously down to the brittle-ductile transition. Finally, such a combination of stress data therefore brings useful information on the strength and mechanical behaviour of the upper continental crust over times scales of several tens of Ma, and should be taken into account in future modelling.

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