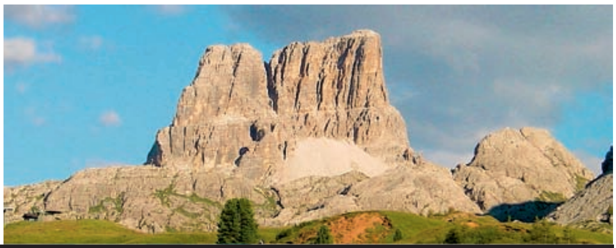

1. Elements of rheology

The kinds of structures that develop in rocks during deformation depend on: 1) the orientation and intensity of the forces applied to the rocks; 2) the physical conditions (mainly temperature, pressure, occurrence and pressure of fluids) under which the rocks are deformed; 3) the mechanical properties of rocks, which are strongly dependent on the physical conditions. At relatively low temperature and pressure and for high intensity of the applied forces, rocks generally undergo brittle deformation, i.e., rocks are characterised by loss of cohesion along discrete surfaces (either faults or fractures) and deformation is accommodated by the mechanism of fracturing (e.g., cataclastic flow). At higher temperature and pressure, rocks normally undergo ductile deformation, i.e., a solid state deformation in which no cohesion loss occurs both at the crystalline scale and at larger scales (no evidence of brittle fracturing). The deformation is accommodated by intracrystalline plastic mechanisms (e.g., from low to high temperature: pressure solution, intracrystalline deformation, twinning, recovery, recrystallization, solid-state diffusion creep, superplasticity). Ductile deformation is generally associated to the generation of folds, stretching and thinning of layers, generation of planar (foliations) and linear (lineations) tectonic fabrics. It is however stressed that not necessarily folding of a sedimentary succession may be accommodated by plastic deformation mechanisms. More frequently, folding in foreland fold-and-thrust belts is accommodated both at meso- and micro-scales by brittle fracturing. In this case the term brittle-ductile deformation is adopted to indicate that geometries of structures may be similar to those occurring in ductile regimes but deformation is accommodated by brittle deformation mechanisms.

The strength of rocks in the brittle regime (brittle yield stress, expressed in terms of differential stress, σ_1-σ_3) is described by the Sibson's law (SIBSON, 1981). This law, based on the Coulomb-Navier frictional criterion, predicts a linear increase with depth of the yield stress:

$$\sigma_1-\sigma_3=\beta\rho \boldsymbol{g}\boldsymbol{z}(1-\lambda) \qquad (1)$$

where β is a parameter depending on the fault type, ρ is the density, $\boldsymbol{g}$ is the gravity acceleration, $\boldsymbol{z}$ is the depth and λ is the pore fluid pressure ratio (i.e., the ratio between fluid pressure and lithostatic load). It is immediately clear that brittle rock strength is quite independent on rock type (with the exception of particluar lithologies such as shales and evaporites, characterised by very low strengths) and increases linearly with depth. It is however dependent on the type (extensional, strike slip or compressional) of the structural setting and on fluid pressure.

The ductile behaviour of rocks is described by various power law creep relations, such as, for example:

$$\sigma_1 - \sigma_3 = \left(\frac{\dot{\varepsilon}}{A}\right)^{1/n} \exp\left(\frac{Q}{nRT}\right) \qquad (2)$$

where $\dot{\varepsilon}$ is the effective strain rate, A is the generalised viscosity coefficient, n is the stress power law exponent, Q is the activation energy, R is the gas constant and T is the absolute temperature (e.g., CARTER & TSENN, 1987). A, n and Q are typically lithology dependent and are determined by laboratory experiments on rock or mineral samples. The ductile strength of rocks diminishes exponentially with depth and is strongly lithology, strain rate and temperature dependent (fig. 1).

The differential yield stress required to produce failure is given by the lesser of the brittle and ductile yield stresses (GOETZE & EVANS, 1979). In the upper crust, at shallow levels, brittle strength is lower than ductile strength and fracturing occurs. At deeper depths ductile strength is lower and plastic deformation occurs. The depth at which brittle and ductile strengths are equal is called brittle-ductile (or alternatively brittle-plastic) transition depth and is the border between the brittle and ductile deformation fields. Such depth is not constant since it varies

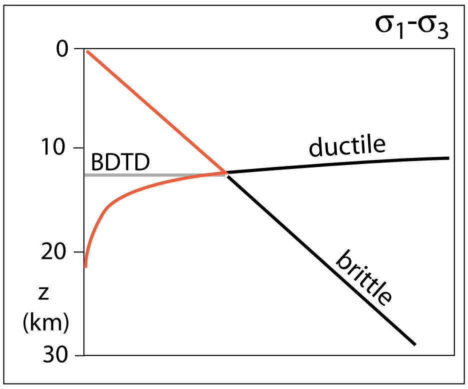

Fig. 1 - Schematic representation of the strength of rocks (yield stress, expressed in terms of differential stress, σ_1-σ_3) in the brittle and ductile regime. Brittle strength is linearly increasing with depth whereas ductile strength diminishes exponentially with depth. The depth at which the two curves cross each other is named brittle-ductile transition depth (BDTD). Above such depth, rocks will undergo brittle deformation, whereas below they will deform ductilely. The strength of rocks with depth will follow the red line.

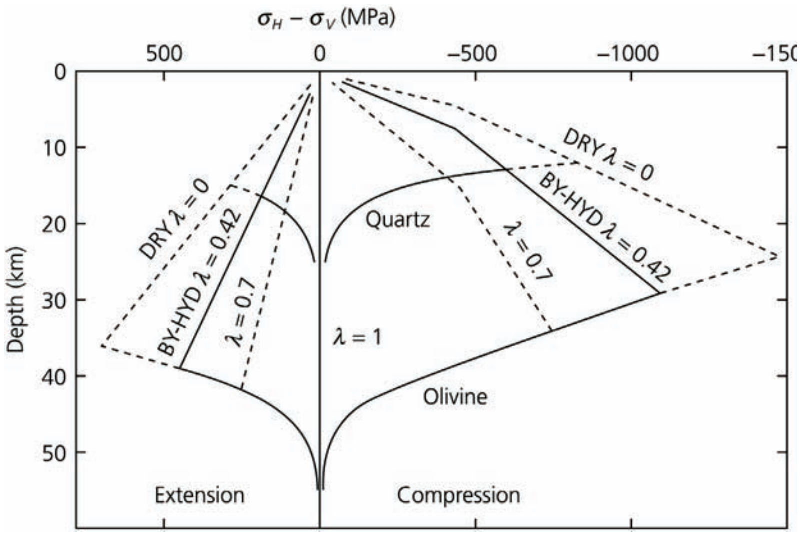

Fig. 2 - The effects of fluid pressure on the depth of brittle to ductile transition are shown. An increase of fluid pressure (increasing λ) induces a lowering of the brittle yield stress and consequently a deepening of the brittle-ductile transition. Brittle strength curves are provided for extensional and compressional settings. Notice also that, for a fixed depth, the ductile strength is larger for olivine than for quartz. Quartz (that is the main mineralogical component of upper crustal crystalline rocks) is therefore characterised by shallower brittle-ductile transitios than olivine (that is the main component of the lithospheric mantle). (After BRACE & KOHLSTEDT, 1980).

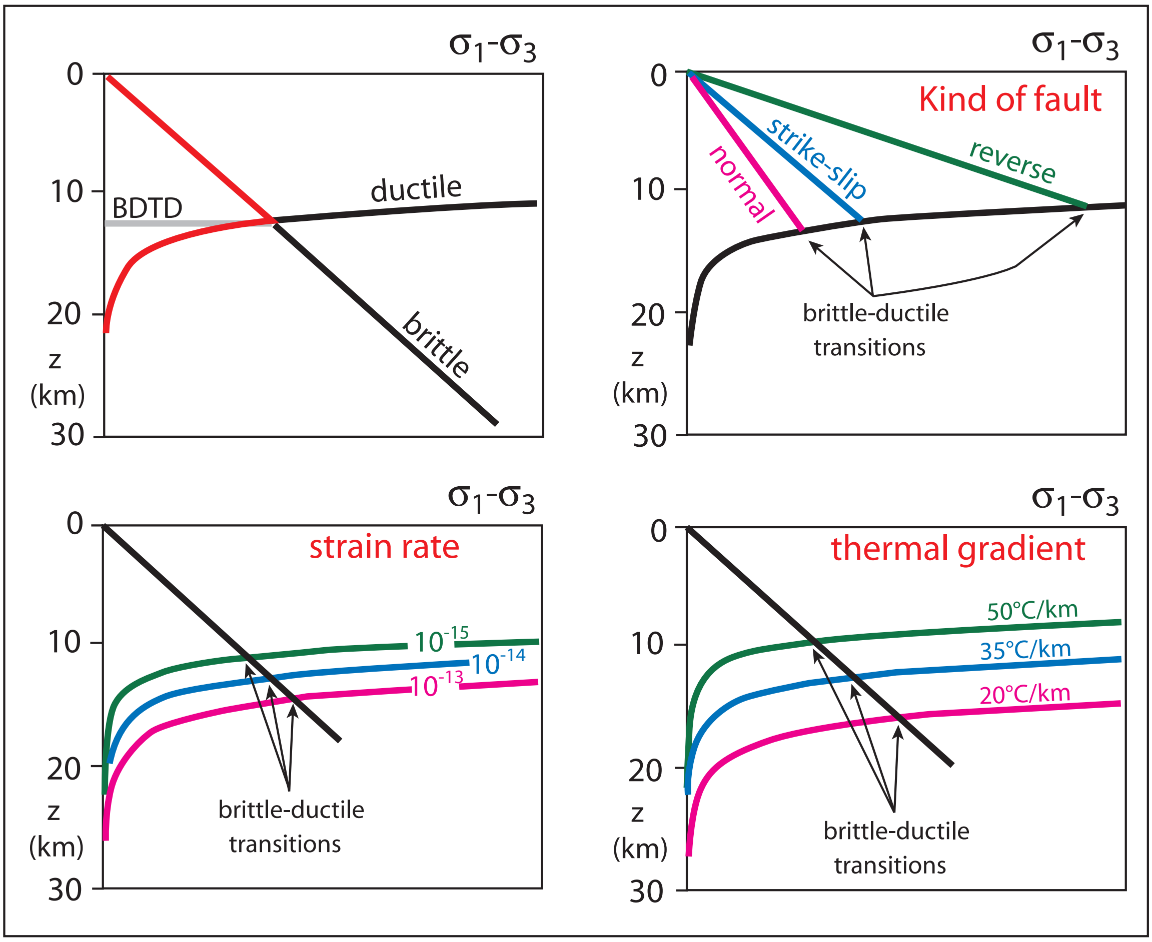

Fig. 3 - The brittle-ductile transition depth (BDTD) is strongly dependent on the kind of structural regime active in a given region, on the strain rate and on the thermal gradient. In this figure, the sensitivity of the BDTD on these parameters is schematically shown. In all the panels the lithology is kept constant. Areas characterised by thrust faulting are characterised by larger brittle strength with respect to areas characterised by strike-slip and normal faulting. As a consequence, keeping constant the other controlling parameters, the brittle-ductile transition depth in contractional areas will be shallower than in strike-slip (intermediate depth) and extensional (deeper depth) areas. Strain rate strongly controls the ductile strength. Typical strain rates in deforming areas vary between 10^{-15} s^{-1} (slow deformation) and 10^{-13} s^{-1} (fast deformation). Increasing strain rates induce higher ductile rock strengths. This is mainly due to the fact that, for fast deformations, migration of lattice defects, recrystallization and other plastic deformation mechanisms are not capable to keep the pace with deformation. Keeping constant the other controlling parameters, areas characterised by faster deformations will be characterised by deeper brittle-ductile transition depths. The thermal gradient of an area is controlled by the regional mantle heat flow, by the crustal radiogenic heat production (that is in turn controlled by crustal lithologies) and by local effects (e.g., magmatic intrusions, heat production generated along major regional faults). An area characterised by larger thermal gradient will be characterised, at a fixed depth, by higher temperatures with respect to areas characterised by low thermal gradients. Since ductile strength is inversely proportional to temperature, at the same depth areas characterised by high gradients will also be characterised by low ductile strength. Therefore, keeping constant the other controlling parameters, larger thermal gradients will be associated to shallower brittle-ductile transition depths.

considerably from region to region. This is mainly due to the variation of the factors controlling brittle (structural setting, fluid pressure) and ductile (temperature gradient, lithology, strain rate) strengths (fig. 2, 3). It has to be stressed that the brittle-ductile transition does not occur sharply and at the same depth in a given region. This is mainly due to the fact that the factors controlling the brittle-ductile transition depth may vary even along short distances (for example the fluid pressure may be controlled by laterally discontinuous acquifers) and the mineralogical content of rocks may also vary significantly. Moreover, the different minerals forming one single lithology (e.g., phyllosilicates, quartz and feldspars in an upper crustal

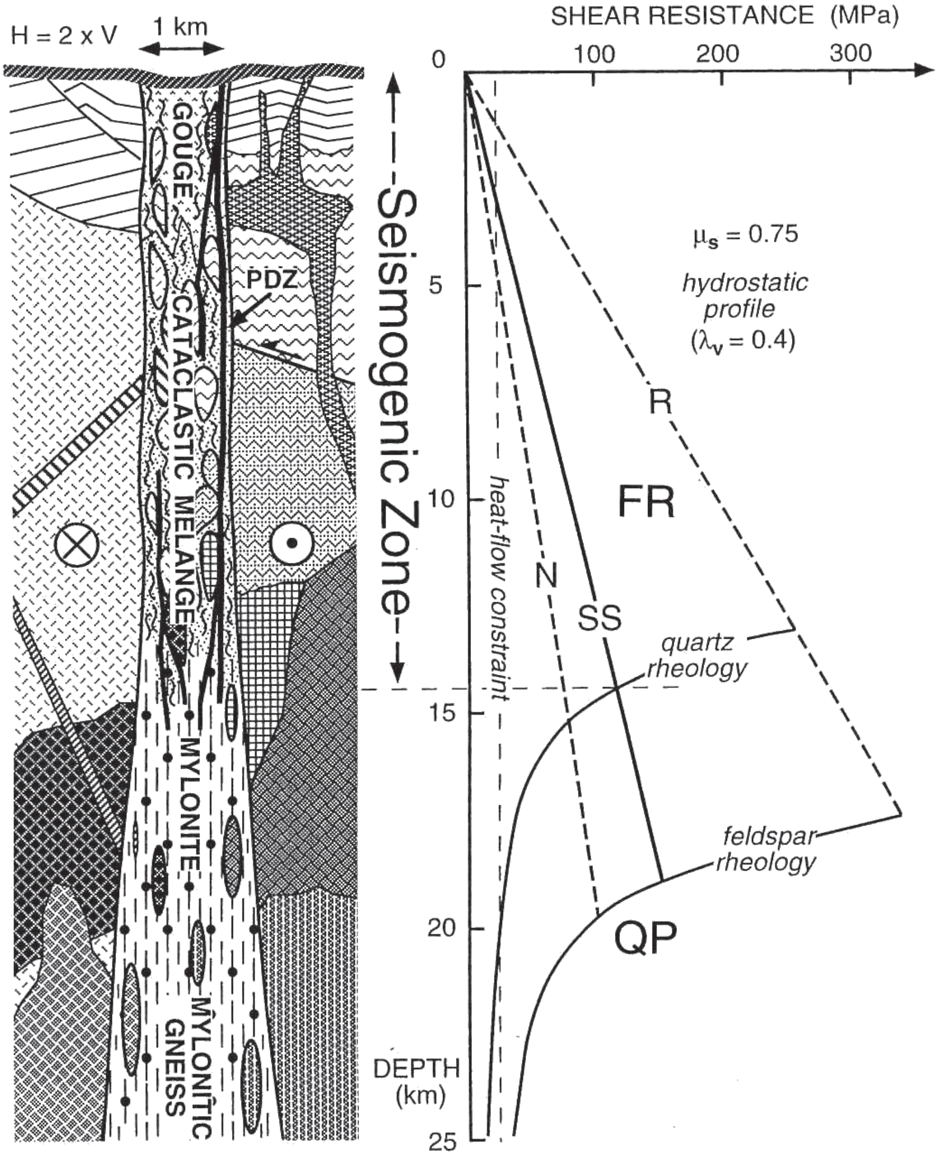

Fig. 4 - Fault rocks associated to brittle (breccias, fault gouge and foliated/non-foliated cataclasites) and ductile (mylonites) structural settings (left panel). Notice that the transition from brittle to ductile behavior is rather ditributed. This is mainly due to the fact that the brittle ductile transition for different rock forming minerals occurs at different depths. In the right panel the intersection between the brittle strength curves (for normal faults, N, for strike slip, SS and for reverse faults, R) and the ductile strength curves for quartz and feldspar is shown. Quartz and feldspar are typical mineralogical components of upper crustal rocks, such as gneisses. In this example, at depths of about 15-18 km, quartz may deform plastically whereas feldspar may be still characterised by fracturing (after Sibson, 2001).

gneiss) may be characterised by different brittle-ductile transition depths.

The active deformation mechanism will control the kinds of rocks generated along fault zones (fig. 4). A major fault zone crossing the upper crust from the surface down to depths of 15-20 km will be characterised by brittle (most probably seismogenic) behavior at shallow depths and ductile (aseismic) behavior below the brittle ductile transition. Typical fault lithologies associated to brittle regimes are breccias, gouge and cataclasites (that can be either foliated or characterised by absence of internal organization). Ductile portions of faults are associated to mylonites.

The minerals constituting the main lithologies forming the lithosphere are characterised by radically different ductile strength (fig. 5). As a consequence, the lithological layering of the lithosphere (sedimentary covers, upper crustal crystalline basement, lower crust, lithospheric mantle) will determine a rheological layering within the lithosphere. The rheological layering of the lithosphere varies according to the type of lithosphere (continental vs. oceanic), to the thickness of the various layers and to the parameters controlling brittle and ductile strength (fig. 6). Typical lithospheric rheological profiles are characterised by the alternance of brittle and ductile levels and have a christmas-tree (or saw-tooth) shape. The occurrence of ductile low strength levels between stronger brittle layers is a major factor controlling the deep geometry of extensional (e.g., rift zones) and compressional areas (e.g., mountain belts). Weak ductile layers, that may be located both in the sedimentary covers and in the crustal crystalline rocks, will, in fact, behave as lithospheric scale detachment levels.

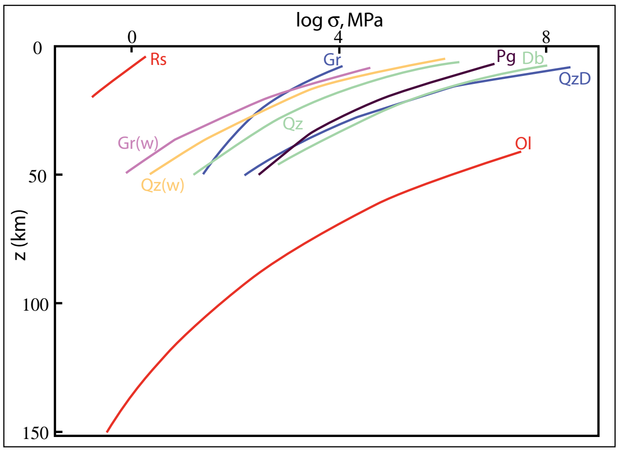

Fig. 5 - Ductile strength of different rocks and rock forming minerals calculated for a cold continental geothermal gradient. Rs, rock salt; Gr(w), wet (i.e., hydrated) granite; Gr, granite; Qz, quartz; Qz(w), wet quartz; Pg, plagioclase; Db, diabase; QzD, quartz-diorite; Ol, olivine. Calcite and limestones have strength intermediate between salt and quartz. Quartz, plagioclase and olivine are the main constituents of upper crust, lower crust and lithospheric mantle. Notice that the ductile strength of the lithospere increases with depth. The minerals of the crust have lower strenght than the minerals of the mantle (after RANALLI & MURPHY, 1987).

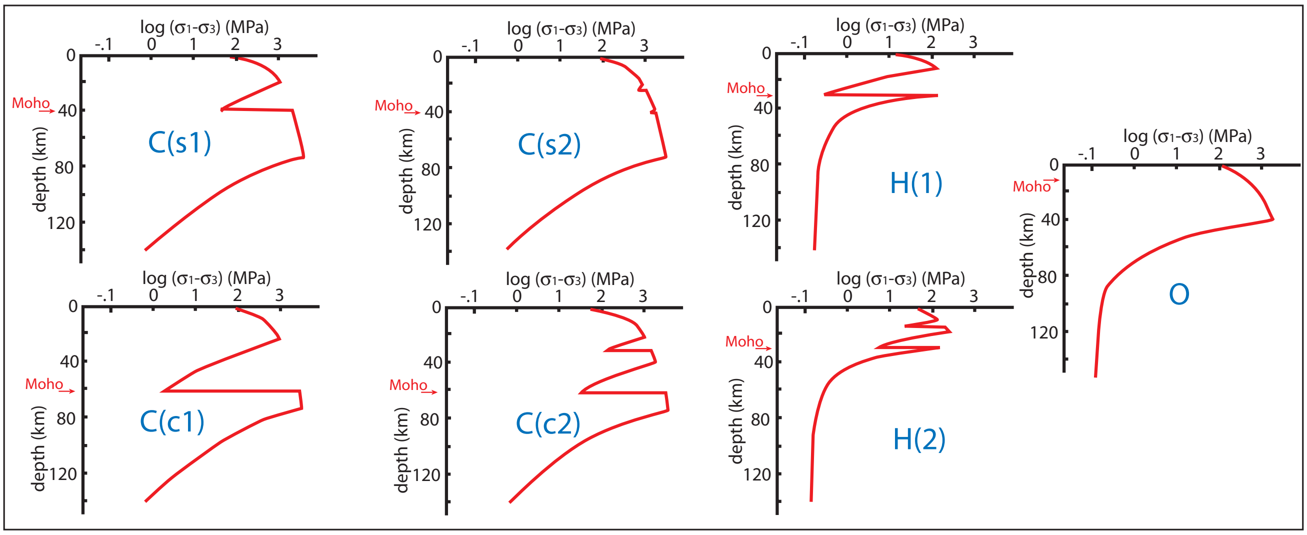

Fig. 6 - Lithospheric scale rheological profiles are characterised by the alternance of brittle and ductile structural levels. Ductile layers may act as major detachment levels along deforming plate margins (e.g., rift zones and collisional belts). The profiles were calculated for different geodynamic setting: C(s1), precambrian shields with 150 km thick lithosphere and unlayered (i.e., quartz.granitic) 40 km thick crust; C(s2), precambrian shields with 150 km thick lithosphere and layered (i.e., quartz-granitic over intermediate basic) 40 km thick crust; C(c1) continental convergence zones with 150 km thick lithosphere and unlayered 60 km thick crust; C(c2) continental convergence zones with 150 km thick lithosphere and layered 60 km thick crust; H(1), continental extensional zones with 50 km thick lithosphere and unlayered 30 km thick crust; H(2), continental extensional zones with 50 km thick lithosphere and layered 30 km thick crust; O, mature oceanic lithosphere, 75 km thick and with a 10 km thick basic crust. The lithospheric rheological profile is even more complicated if one considers sedimentary cover as well. As discussed in the previous figures, evaporites and limestones are characterised by ductile strengths lower than crystalline rocks. Additional brittle-ductile transitions may therefore occur within the sediment succession (after RANALLI & MURPHY, 1987).