

Fluids and rock deformation

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Virtually all metamorphic rocks have interacted with fluids at some time during their history. Some polymetamorphic rocks have distinct textures and mineralogies associated with a later (retrogressive) fluid interaction. Metasomatism can only occur when fluids are out of chemical, and sometimes thermal, equilibrium with the adjacent rocks. Many fluids may never leave a chemical signature in the rocks through which they have passed (THOMPSON, 1983; WOOD and WALTHER, 1986; FERRY, 1986).

Minute amounts of volatile species are almost always present in the disconnected porosity of grain junctions. This fluids is not in contact with all mineral surfaces (WATSON and BRENNAN, 1987). Metamorphic fluids are released by devolatilisation as the rocks P-T path passes through the appropriate metamorphic reactions. How the fluids are able to interact with the adjacent rocks depends upon the flow pathways that are available, or that the fluids can make for themselves through chemical and mechanical responses of the minerals (see FYFE et al., 1978; THOMPSON, 1987).

The rate of metamorphic fluid production

The rate of fluid production is directly proportional to the rate of heat supply (e.g. WALTHER and Orville, 1982). The relations

are not simple to evaluate because of the generally endothermic nature of most devolatilisation reactions occurring at different times in the metamorphic pile (RIDLEY, 1986). Rapid devolatilisation only occurs near thermal perturbations (THOMPSON, 1987). Thus most of the spectacular features of metamorphic rocks, such as vein systems or shear zones, are catastrophic phenomena and indicate fluid-rock interaction on short timescales, occurring in response to transient mechanical or thermal instabilities (e.g. faults, shear zones or magma bodies). The largest part of a metamorphic pile will experience slow thermal and related mechanical responses. In these environments, rock deformation and fluid production keep pace with the mineral reactions.

Fluids and the rheological properties of rocks

Rock mechanics experiments have shown that under certain conditions of P, T, σ , $\dot{\epsilon}$ etc., if devolatilised fluid is confined, rock samples can fracture even when they are undergoing long-term ductile behaviour (see original references in the summaries by KIRBY, 1983, 1985; MURRELL, 1985; BRODIE and RUTTER, 1985). Under other conditions, even trace amounts of fluid are able to hydrolytically weaken crystals so that they exhibit quasi-plastic instead of brittle behaviour (GRIGGS-

see KIRBY, 1984, 1985). The chemical mechanisms of hydrolytic weakening are poorly understood but are thought to involve $\text{Na}^+\text{-H}^+$ or $\text{Si}^{4+}\text{-4H}^+$ exchanges coupled with vacancy migration (AINES et al., 1984; PATERSON, 1986). As hydrolytic weakening is very mineral specific, the deformational behaviour of multiminerals composites (rocks) in the presence of fluids is poorly understood, although it is usually assumed that bulk deformation is controlled by the rheological response of the weakest phase (SCHMID, 1982).

Deformation and equilibrium domains in polymetamorphic rocks

Prograde regional metamorphic terrains frequently show approaches to textural and mineralogical equilibrium. In many polymetamorphic terrains, however, the degree of such equilibrium varies on quite small scale. Pods of higher-grade rocks are frequently surrounded by finer-grained lower-grade «retrograded» assemblages (KOONS and RUBIE, 1983; RIDLEY and DIXON, 1984). Because of transiently increased porosity during cataclastic deformation and mineral reactions, access of fluids may be temporarily enhanced. Stable isotope studies of minerals from local deformation regimes in metamorphic rocks have revealed both domains of extremely restricted fluid motion (FRÜH-GREEN, 1985a, 1985b), localised channel flow (RUMBLE et al., 1982), and massive channelised flow (KERRICH et al., 1984).

Fluid release and deformation mechanisms

Metamorphic reactions frequently produce fine grain-sized products which may locally enhance ductility and localise deformation (RUBIE, 1983). The volume changes associated with most mineral reactions are such that transient local porosity increases might occur and any fluid released by reaction will remain at the reaction site until fluid pressure locally exceeds the least principal stress and induces microcracking even in a nominally ductile regime.

In the shallower, more brittle, part of the crust rapid release of fluid from a deep crustal reservoir, or from devolatilisation reactions, can cause massive fracturing and initiation of hydrothermal systems (SIBSON, 1981, 1986; NORTON, 1984). In deep crustal levels, massive fluid influx can change the deformation mechanism to cataclasis (RUTTER, 1983; BRODIE and RUTTER, 1985) and initiate shear zones (WHITE and KNIPE, 1978; RUTTER and BRODIE, 1985). Subsequent deformation behaviour is controlled by the rate of propagation of fractures (SEGALL, 1984) or the shear-zones (RUTTER and BRODIE, 1985) compared to the rate at which fluid can percolate through the less permeable porous rock-matrix between the high-flow fluid channelways (ETHERIDGE et al., 1984; SEGALL, 1984; THOMPSON, 1987).

Localisation of deformation

Processes that localise deformation include strain/fabric softening, rotation of local strain anisotropies, devolatilisation reactions which promote brittle failure by causing an increase in pore fluid pressure, or at a rheological boundary where a change of deformation mechanism occurs in a gradient of temperature and mean stress (POIRIER, 1980; CASEY et al., 1987). Plots of rheological experiments for single phase materials in stress-temperature space (e.g. KIRBY, 1985) show well the temperatures at which common minerals become weak and thus more easily deformed. Assuming that the bulk strength of rocks is controlled by the rheological responses of the weaker phase, the rheological behaviour of individual rocks (HANDY and WALKER, 1987), or indeed the rheological stratification of the whole lithosphere can be investigated (e.g. RANALLI and MURPHY, 1987; SONDER and ENGLAND, 1986; ENGLAND, 1987).

Within the lithosphere primary detachment caused by the above localisation mechanisms can occur at a specific crustal depth (BUTLER et al., 1987) determined principally by rheological properties, temperature and mean stress gradients, and deformational boundary conditions (CASEY et al., 1987).

After the first detachment and, for

example, the emplacement of a thrust sheet, the subsequent deformational behaviour is strongly controlled by the evolution of the perturbed thermal regime (THOMPSON et al., 1987).

Rheological properties and deformation regimes

Because the rheological properties of individual minerals varies with both temperature and pressure, the weaker component of a rock is not necessarily the same in all metamorphic facies. STÜNITZ (1987) has shown that amphibole is weaker in blueschist to greenschist facies transitions in metabasic rocks but that plagioclase is weaker in amphibolite to greenschist facies transitions. Likewise changes in deformation mechanism are likely when rocks are reacting to produce melts of fluids. For deformation to be successful, a favorable combination of rheological and thermal parameters is required to occur within relatively short geological time intervals ($< \approx 10$ Ma).

Escape of fluids and melts from deforming rocks

The segregation of volatile species to form a fluid, or of melt to form a magma, are much discussed but still poorly understood phenomena. Particularly underinvestigated is the role of shear in fluid or melt localisation and the feedback system involving deformation. Shear can promote mineral reactions, change the rate of deformation or even the mechanism, and localise the weaker fluid or melt phase into layers.

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