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Notes

Stress and deformation in subduction zones: insight from the record of exhumed metamorphic rocks

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Abstract: High pressure (HP) and ultrahigh (UHP) metamorphic rocks are exhumed from subduction zones at high rates on the order of plate velocity (cm/year). Their structural and microstructural record provides insight into conditions and physical state along the plate interface in subduction zones to depths of >100 km. Amazingly, many identified (U)HP metamorphic rocks appear not to be significantly deformed at (U)HP conditions, despite their history within a high strain rate mega-shearzone. Other (U)HP metamorphic rocks seem to be deformed exclusively by dissolution–precipitation creep. Indications of deformation by dislocation creep are lacking, apart from omphacite in some eclogites. Available flow laws for dislocation creep (extrapolated to low natural strain rates, which is equivalent to no deformation on the time scales of subduction and exhumation, i.e., 1 to 10 Ma) pose an upper bound to the magnitude of stress as a function of temperature along the trajectory followed by the rock. Although the record of exhumed (U)HP metamorphic rocks may only be representative of specific types or evolutionary stages of subduction zones, for such cases it implies: (1) strongly localized deformation; (2) predominance of dissolution–precipitation creep and fluid-assisted granular flow in the shear zones, suggesting Newtonian behaviour; (3) low magnitude of differential stress; which (4) is on the order of the stress drop inferred for earthquakes; and (5) negligible shear heating. These findings are easily reconciled with exhumation by forced flow in a low viscosity subduction channel prior to collision, implying effective decoupling between the plates.

An unexpected wealth of high-pressure (HP) and previously unknown ultrahigh-pressure (UHP) metamorphic rocks has been identified within the last two decades (Harley & Carswell 1995; Schreyer 1995; Coleman & Wang 1995; Ernst & Liou 1999; Liou 1999; Ernst 1999). The characteristic low temperature/pressure ratios require that these rocks have been exhumed from subduction zones. Thus their record provides valuable insight into the physical state at depth and the trajectories followed by these rocks, which can be integrated into the results of geophysical field studies, laboratory experiments, and numerical simulations (e.g. Peacock 1996; Hacker & Peacock 1995; Ernst & Peacock 1996; Hacker 1996) to address fundamental questions concerning subduction zones. These questions comprise the degree of localization of deformation in interplate shear zones, the magnitude of shear stress, the nature and position of subduction zone seismicity, the interaction between deformation and phase transformations, the role of devolatilization, the pore fluid pressure and transport properties and their variation in space and time, and finally the typical particle trajectories followed by rocks during progressive subduction and exhumation. The latter can only be recorded to a limited extent by the present structure of mountain belts, due to the general lack of complete information. The purpose of the present paper is to highlight the significance of the record of (U)HP metamorphic rocks, beyond that of occurrence of uncommon mineral assemblages or derived very high pressures, and to emphasize the opportunities presented by the available information. As most of the peculiar features have not received appropriate attention by researchers so far, it is hoped that this paper will stimulate more investigations focussed on the objective assessment of the (micro)structural record of exhumed rocks, and its interpretation in terms of subduction zone rheology and the general geodynamics of convergent plate boundaries.

The record of metamorphic rocks exhumed from subduction zones

Pressure–temperature paths

Pressure–temperature (P–T) paths are available for a great number of HP- and UHP-metamorphic terranes. Most reveal significant cooling during decompression (e.g. Harley & Carswell 1995; Carswell & Zhang 1999; Ernst

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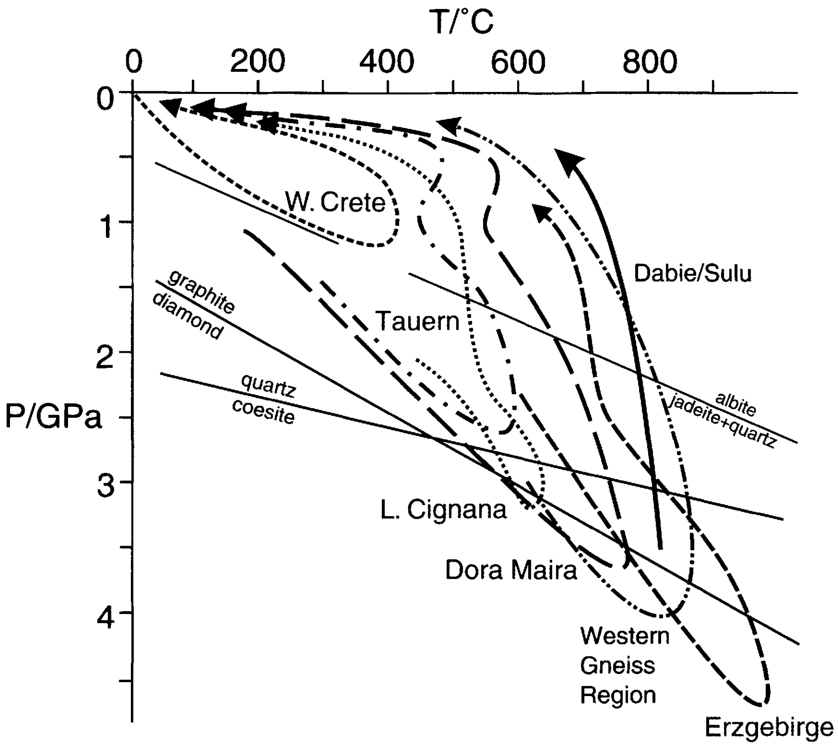


Fig. 1. Selected P–T paths of HP and UHP metamorphic rocks: Erzgebirge (Massonne 1999), Dora Maira (Chopin & Schertl 1999), Lago Cignana (Reinecke 1998), Tauern (Stöckhert *et al.* 1997), Western Gneiss Region, Norway (Terry *et al.* 2000*a*, *b*), Dabie/Sulu, China (Wallis *et al.* 1999), Crete (Küster & Stöckhert 1997). Note that the shape of P–T paths is always subject to large uncertainties and that a (U)HP metamorphic terrain cannot be characterized by a single loop. Nevertheless, the inferred general form is similar for most

1999), although a number of exceptions occur with a significant portion showing near-isothermal decompression (e.g. Carswell *et al.* 2000) or a more complex shape with significant changes in slope (Reinecke 1998). Some examples are displayed in Fig. 1. In general, most P–T paths are presented without consideration of error bars, which can be very large for thermobarometry. Apart from the uncertainties in the position of phase equilibria and the formulation of the usually applied geothermobarometers, it is very difficult to assess local equilibrium within a phase assemblage in a natural rock, and in fact equilibrium may not be attained at all in many (U)HP metamorphic mineral assemblages. Furthermore, a subduction-related high pressure metamorphic complex is neither characterized by a single type of P–T loop, nor is a single P–T trajectory sufficient to derive or to test tectonic models (Gerya *et al.* 2001). Notwithstanding these restrictions, the available information appears to be sufficiently robust and the points made in this contribution are not seriously affected by the uncertainties.

Time constraints

The results of isotopic dating with a variety of methods indicate that exhumation of continental UHP metamorphic rocks of the Dora Maira Massif (Gebauer *et al.* 1997; Rubatto & Hermann 2001), and UHP metamorphic oceanic slices in the Piemonte Zone (Amato *et al.* 1999), both tectonic units of the western Alps, took place at rates on the order of plate velocity. Very high rates of burial and heating are also indicated by preservation of steep compositional gradients in zoned minerals and around inclusions (Perchuk *et al.* 1999; Perchuk & Philippot 1997, 2000). As the geochronological data can only provide average rates, and the time brackets may cover late stages of slower cooling and decompression after reaching normal crustal depth, the actual exhumation rates may be even higher than the minimum rates suggested by these studies. For older UHP metamorphic terranes, such as the Triassic Dabie Sulu belt in Central China (e.g. Hacker *et al.* 1997, 2000), the Paleozoic Kokchetav massif in Kazachstan (e.g. Kaneko *et al.* 2000; Maruyama & Parkinson 2000), the Caledonian Western Gneiss Region in Norway (e.g. Terry *et al.* 2000*b*), and the Variscan Erzgebirge in Germany (e.g. Massonne 1999), larger age uncertainties preclude precise estimates on the exhumation rates, although the available data appear to reconcile with exhumation to normal crustal depths within less than 10 Ma.

The record of deformation at (U)HP conditions

Apart from some structural and microstructural observations reported from the Dora Maira Massif (e.g. Wheeler 1991; Henry *et al.* 1993; Michard *et al.* 1993, 1997), the UHP-metamorphic belt in China (e.g. Wallis *et al.* 1997, 1999; Hacker *et al.* 2000; Faure *et al.* 1999), and the Kokchetav massif in Kazachstan (e.g. Yamamoto *et al.* 2000), the deformation-related fabrics of UHP-metamorphic rocks have received much less attention when compared to the numerous studies on phase relations. In particular, a quantitative assessment of the deformation mechanisms and the inferred boundary conditions of deformation at great depth, and of the changing state of stress along the particle trajectory (e.g. van der Klauw *et al.* 1997), is not available for UHP metamorphic units. For less deeply buried HP-metamorphic rocks there are numerous (micro)structural studies, but sufficiently detailed documentation and objective interpretation is similarly scarce. In part, this may be a consequence of the rather inconspicuous (micro)structural record, and in particular of the apparent absence of high-strain shear zones commonly observed in other crustal units. In fact, the record of (U)HP metamorphic rocks appears to be rather systematic and intriguing for the given tectonic setting. Most (U)HP-metamorphic rocks, where they are not affected by the common late-stage overprint at normal crustal depth under greenschist to amphibolite facies metamorphic conditions, reveal a common picture: many are not notably deformed at HP or UHP-conditions at all, or the recorded deformation took place by poorly specified deformation mechanisms other than dislocation creep.

For instance, for the coesite-bearing pyrope quartzites (Chopin *et al.* 1991; Schertl *et al.* 1991; Compagnoni *et al.* 1995) of the Dora Maira Massif, western Alps, Michard *et al.* (1993, 1995) describe a preferred orientation of minerals overgrown by the large pyrope poikiloblasts. Also, these authors present evidence of strain shadows to the large pyrope crystals, implying a weak deformation at UHP metamorphic conditions. Apart from the presumably inherited compositional layering and related preferred orientation of minerals grown during UHP metamophism, there is no further record of pervasive deformation in the pyrope quartzites.

A particulary impressive rock body in the Dora Maira Massif is the UHP metamorphic Brossasco granite (Biino & Compagnoni 1992; Bruno *et al.* 2001). This km-sized granite body, intersected by late greenschist-facies shear zones, has remained undistorted during burial to >100 km depth and subsequent exhumation, with the original magmatic fabric perfectly preserved (Fig. 2). In view of the ease with which granite bodies elsewhere acquire a gneissic fabric, the undistorted UHP metamorphic Brossasco granite is taken to represent a stress gauge that indicates that the maximum differential stress as a function of temperature has remained too low to drive any significant deformation along its trajectory in the subduction zone (Stöckhert & Renner 1998; Renner *et al.* 2001). Similar undeformed UHP metamorphic granites (Hirajima *et al.* 1993; Wallis *et al.* 1999) and gabbros (Zhang & Liou 1997) have been reported from the Dabie-Sulu UHP belt in China. In general, UHP metamorphic rocks appear either weakly deformed at depth, with a foliation and mineral preferred orientation pattern inherited from an earlier stage and overgrown by the UHP mineral assemblage, or essentially undeformed, as in rocks with preserved primary magmatic fabrics.

In contrast to the quartz- or coesite-rich rocks described above, eclogites derived from precursors of basaltic composition frequently reveal a marked foliation and a pronounced shape (SPO) and crystallographic (CPO) preferred orientation of omphacite, even in rocks that underwent HP metamorphism at rather low temperatures (e.g. Buatier *et al.* 1991; Philippot & van Roermund 1992; Godard & van Roermund 1995; Piepenbreier & Stöckhert 2001). The CPO is attributed to deformation by dislocation creep and is consistent with the established glide systems in clinopyroxene (e.g. Mauler *et al.* 2000). However, at least in some cases, the combined SPO and CPO of omphacite in eclogites could also be the result of oriented nucleation and anisotropic growth. Also, many eclogites lack an obvious SPO or CPO of omphacite and

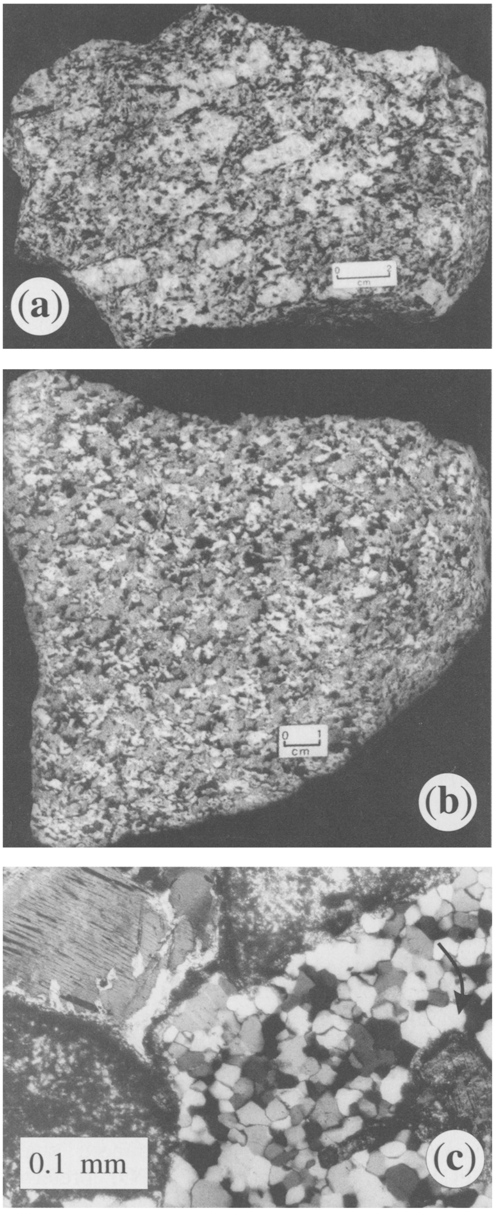


Fig. 2. Structure and microfabric of the UHP metamorphic Brossasco granite from the Dora Maira massif, western Alps. Undistorted magmatic fabric (**a**, **b**) and microstructure (**c**) revealing fine-grained quartz aggregates with foam structure. This quartz has grown at the expense of coesite, that in turn had formed from magmatic quartz during UHP metamorphism. Biotite rimmed by coronitic garnet, plagioclase decomposed to fine-grained aggregates of breakdown phases.

even primary magmatic structures (e.g. undistorted pillows, Bearth 1959) are preserved in places. Finally, it should be stressed that the occasional record of deformation of omphacite by dislocation creep should not be mistaken as an indication that this deformation regime is relevant for the subduction shear zone in general.

Microstructures and deformation mechanisms

Evidence for deformation by dissolution precipitation creep is mainly restricted to low to intermediate grade HP metamorphic rocks. An unequivocal microstructural record for this deformation mechanism has not been described from higher grade (U)HP metamorphic rocks. On the other hand, evidence for significant deformation by dislocation creep is lacking. In particular microstructures of major constituents indicating dynamic recrystallization or recovery and a pronounced CPO (apart from the forementioned omphacite fabrics in some eclogites) are rarely observed. Three representative examples (dealt with elsewhere in more detail) are given below.

(1) Delicate fossils in carbonate rocks of Crete, Greece, buried to 35 km depth, were not distorted when overgrown by coarse-grained metamorphic aragonite (Fig. 3a). Also, they remained undistorted during folding and boudinage at depth, as syn-HP-deformation was essentially by dissolution–precipitation creep not affecting the interior of large grains (Stöckhert *et al.* 1999). The grain-size sensitivity of dissolution–precipitation creep is reflected by boudinage of the coarse-grained layers within homogeneously deformed fine-grained aragonite marbles, revealing a microstructure characteristic for dissolution–precipitation creep (Fig. 3b) and lacking a CPO (Wachmann & Stöckhert, unpublished data). In siliciclastic rocks of the same series, progressive deformation at HP metamorphic conditions ($c. T = 400 \pm 50\,°C$, $P = 1 \pm 0.3\,GPa$) was essentially by dissolution–precipitation creep with enhanced dissolution at mica–quartz interfaces in poly-phase phyllites (Fig. 3c), and stress concentration in pure (nearly single phase) quartzites and quartz veins deforming by dislocation creep (Schwarz & Stöckhert 1996; Stöckhert *et al.* 1999).

(2) Garnet poikiloblasts (Fig. 4a) in metasediments from the Tauern Window, eastern Alps, crystallized during HP metamorphism at the maximum depth of burial of *c.* 70 km ($T \approx 600\,°C$ at $P \approx 2.5\,GPa$) have overgrown

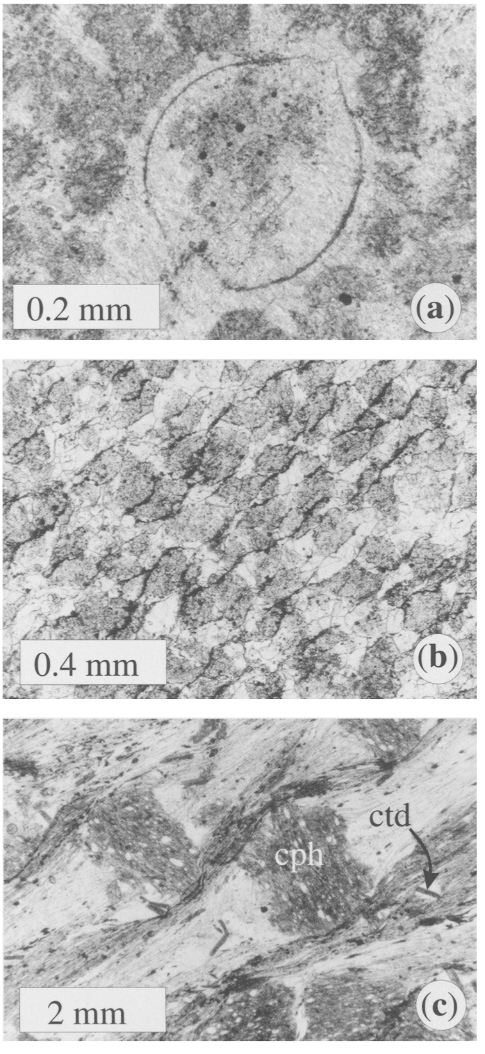


Fig. 3. Microstructures of low-temperature HP metamorphic rocks from Crete (see Stöckhert *et al.* 1999 for details). (**a**) Coarse-grained aragonite marble boudin with undistorted fossil shell overgrown by metamorphic aragonite, precluding pervasive deformation of this carbonate rock during burial to >25 km depth (crossed polarizors). (**b**) Fine-grained aragonite marble homogeneously deformed by dissolution–precipitation creep in the stability field of aragonite, with unsoluble opaque particles enriched at interfaces normal to shortening direction and clear overgrowth at interfaces normal to the stretching direction (crossed polarizors). (**c**) Phyllite deformed by dissolution–precipitation creep. Mg-carpholite (cph) poikiloblasts have overgrown a foliation defined by mica and truncated clastic quartz grains, with strain shadows and strain caps at carpholite and chloritoid (ctd) revealing ongoing deformation by dissolution–precipitation creep. There is no evidence for crystal plastic deformation of quartz.

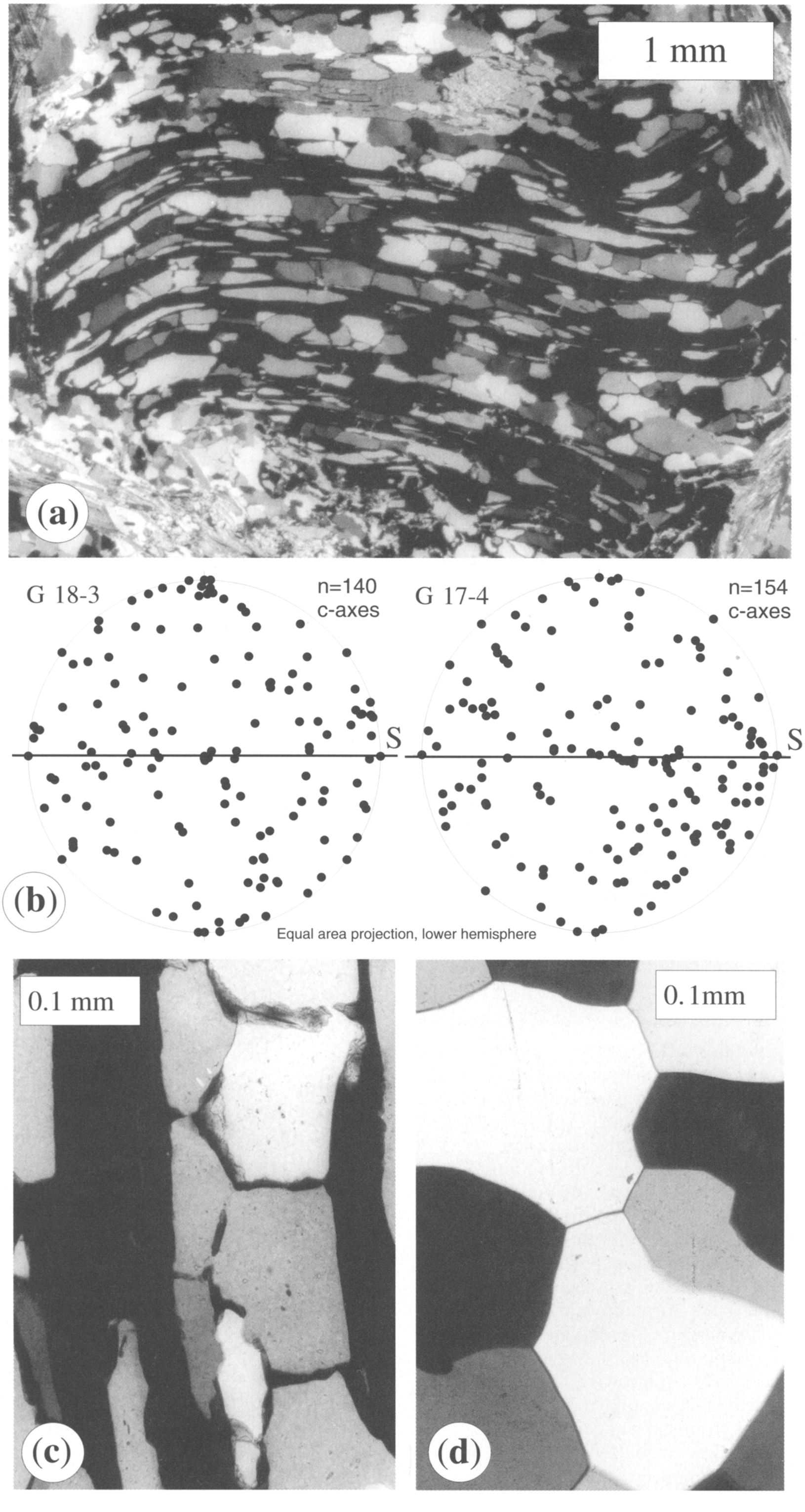
quartz aggregates with a typical foam microstructure (Fig. 4c; Stöckhert *et al.* 1997) and random crystallographic orientation (Fig. 4b; Eismann & Stöckhert, unpublished data). The lack of a quartz CPO as preserved inside the garnet in the well-foliated schist indicates that dislocation creep was not activated during progressive burial to 70 km depth. Also, the foam structure controlled by interfacial free energy shows that stress was too low to drive deformation by dislocation creep at the stage when the rock became detached from the downgoing plate and started to return towards the surface.

(3) The classical pyrope quartzites of the Dora Maira Massif, western Alps reveal a coarse-grained (0.3–0.5 mm) foam structure of quartz (Fig. 4d). The quartz formed at the expense of coesite in an early stage of exhumation (Chopin *et al.* 1991) does not reveal a CPO (Lämmerhirt & Stöckhert, unpublished data). The preservation of a conspicuous palisade-type SPO pattern of elongated quartz grains (e.g. Michard *et al.* 1995) related to the coesite–quartz phase transformation, the undisturbed interfacial free energy controlled foam structure, and the lack of CPO preclude any post-UHP deformation by dislocation creep during exhumation.

For these examples, the microstructural record implies that the levels of differential stress (as a function of temperature) required to drive deformation of calcite/aragonite or quartz by dislocation creep were generally not reached at HP-conditions, and during most of the burial and early exhumation history. Likewise, the undistorted UHP metamorphic granites indicate that the same holds true for the stage of UHP metamorphism in the stability field of coesite.

Experimental constraints on flow strength

The ultimate flow strength of rocks as a function of temperature and strain rate is limited by dislocation creep (e.g. Ranalli 1995; Evans and Kohlstedt 1995). Thus, for rocks not deformed at (U)HP conditions at all, or exclusively deformed by mechanisms like dissolution–precipitation creep without activation of crystal plastic processes, an upper bound to differential stress at (U)HP conditions can be obtained by extrapolation of flow laws for dislocation creep to a strain rate of $10^{-15}\,s^{-1}$, which is equivalent to no notable deformation on the time scales of subduction and exhumation (1 to 10 Ma). As a first approximation, the ultimate strength of many polyphase crustal rocks buried in subduction zones and undergoing (U)HP metamorphism is controlled by the predominating


1 mm
(a)
G 18-3
n=140
c-axes
S
G 17-4
n=154
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(b)
Equal area projection, lower hemisphere
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(c)
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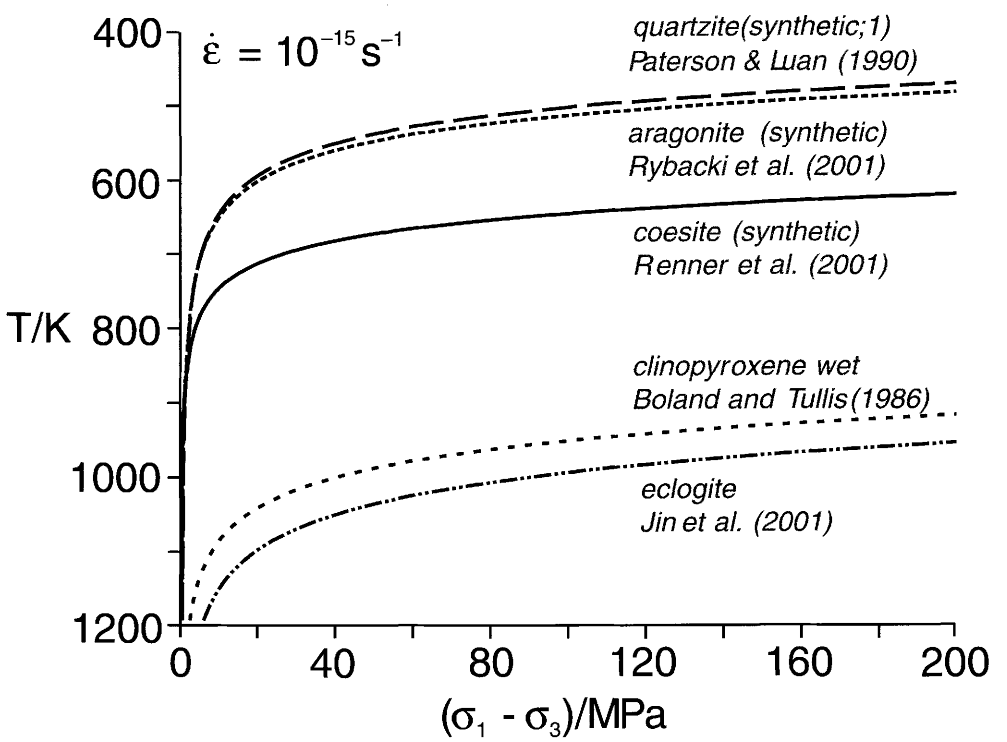


Fig. 5. Strength as a function of temperature for rock-forming minerals relevant at HP and UHP metamorphic conditions, based on experimental flow laws for dislocation creep (power law, $d\varepsilon/dt = A\exp(-Q/RT)\sigma^n$, with $d\varepsilon/dt$ denoting the strain rate in s^{-1}, A is the pre-exponential factor given as $MPa^{-n}\,s^{-1}$, Q is the activation energy in kJ/mole, n is the stress exponent, T is the absolute temperature in K, and R is the gas constant). The parameters for quartz are from Paterson & Luan (1990; $Q = 135$ kJ/mol, $n = 3.1$, $\log A = -7.19$ [$MPa^{-n}n^{-1}$]), for aragonite from Rybacki *et al.* (in prep.; $Q = 246$ kJ/mol, $n = 5.25$, $\log A = 0.52$ [$MPa^{-n}n^{-1}$]), for coesite from Renner *et al.* (2001; $Q = 257$ kJ/mol, $n = 3.1$, $\log A = -3.16$ [$MPa^{-n}n^{-1}$]), for clinopyroxene from Boland & Tullis (1986; $Q = 490$ kJ/mol, $n = 3.3$, $\log A = 5.17$ [$MPa^{-n}n^{-1}$]), and for eclogite from Jin *et al.* (2001; $Q = 480$ kJ/mol, $n = 3.43$, $\log A = 3.3$ [$MPa^{-n}n^{-1}$]. For discussion of the experimental techniques, data reduction, and uncertainties the reader is referred to the original papers. The strain rate of $10^{-15}\,s^{-1}$ is equivalent to no significant deformation on the time scales inherent to burial and exhumation of (U)HP metamorphic crustal slices, i.e. 1 to 10 Ma.

constituent, i.e. omphacite for basaltic composition, quartz or coesite for granitic composition, and aragonite for carbonate rocks.

An experimental flow law for dislocation creep of quartz has been provided by, among others, Paterson & Luan (1990), for coesite by Renner *et al.* (2001), for aragonite by Rybacki *et al.* (in prep.), and for omphacite with additional garnet, i.e. eclogite, by Jin *et al.* (2001). The graphical representation of these flow laws extrapolated to a geological strain rate of $10^{-15}\,s^{-1}$ is depicted in Fig. 5. Unfortunately, the crucial test of the reliability of these flow laws and their extrapolation (i.e. the comparison with natural

Fig. 4. Interfacial free energy controlled quartz microstructure in HP metamorphic rocks. (**a**) Poikiloblastic garnet in micaschist from the Eclogite Zone of the Tauern Window, eastern Alps (see Stöckhert *et al.* 1997 for details). (**b**) Two examples for the absence of a CPO (*c*-axes) of quartz overgrown by poikiloblastic garnet at HP metamorphism, as shown in (**a**) and (**c**), indicating that dislocation creep of quartz was not activated in the foliated schist during burial to about 70 km (Eismann & Stöckhert, unpublished data). (**c**) Closer view showing interfacial free energy controlled quartz microstructure (foam structure) overgrown and preserved by garnet shown in a) during HP metamorphism at *c.* 600 °C and 2.5 GPa. (**d**) Foam structure of quartz that has formed at the expense of coesite in UHP metamorphic pyrope quartzite (Chopin *et al.* 1991, Schertl *et al.* 1991) from the Dora Maira massif, western Alps. There is no indication of crystal plastic deformation of quartz during exhumation.

microstructures in rocks with well-constrained metamorphic conditions, e.g. Stöckhert *et al.* 1999; Hirth *et al.* 2001) is impossible in the case of coesite and aragonite for two reasons. First, these phases are readily transformed to quartz and calcite, respectively, upon exhumation (Mosenfelder & Bohlen 1997; Liu & Yund 1993) and syn-(U)HP microstructures are erased. Second, based on observations on natural (U)HP metamorphic rocks, it remains doubtful whether stress levels sufficient for deformation by dislocation creep are ever typically reached at natural (U)HP conditions, and consequently it is doubtful whether the microstructures required for comparison are even formed in nature. In contrast to coesite and aragonite, omphacite microstructures have been reported from some eclogites, reflecting deformation by dislocation creep, even at temperatures below 500 °C (Piepenbreier & Stöckhert 2001). Deformation by dislocation creep at such low temperatures is in conflict with the extrapolation of experimental data for diopside (e.g. Boland & Tullis 1986), that invariably predict an unrealistically high flow strength for $T < 700\,^{\circ}C$ at low geological strain rates (Fig. 5). Even the eclogite flow law proposed by Jin *et al.* (2001) does not allow deformation at reasonable stress levels at temperatures below 700 °C (Fig. 5). Based on the natural record it is therefore proposed that sodic pyroxenes may have a significantly lower flow strength compared to diopside. This prediction is supported by new experimental results on synthetic jadeite aggregates (Orzol *et al.* 2001), and awaits further verification and extension to the omphacite solid solution series. As pointed out by Jin *et al.* (2001) and Piepenbreier & Stöckhert (2001), the flow strength of eclogite is of particular interest, as metamorphosed basaltic oceanic crust is expected to form a continuous layer between both lithospheric plates in subduction zones, with the ultimate strength of eclogite thus posing an upper bound to interplate shear stress.

Discussion

Localization of deformation

Rocks that remained weakly deformed or undeformed at (U)HP-conditions indicate a high degree of localization of deformation within the subduction zone. Assuming a simple situation without return flow, a cumulative thickness of interplate shear zone(s) of 2 km, and a convergence rate of only 1 cm/year, strain rates would be on the order of $10^{-13}\,s^{-1}$. Higher convergence rates and smaller cumulative thickness of the shear zones, as suggested by the huge volumes of undeformed or weakly deformed (U)HP metamorphic rocks, require correspondingly higher strain rates, maybe even of the order of $10^{-11}\,s^{-1}$, which are well above the rates usually considered as relevant for long-term tectonic processes. Also, more complicated kinematic patterns, including return flow in a subduction channel as a feasible mechanism for rapid exhumation of (U)HP metamorphic rocks (see below), imply higher strain rates.

In any case, the postulated shear zones are obviously difficult to identify in the field. Indeed, the present author is not aware of any descriptions and detailed microstructural analyses of high-strain shear zones unequivocally developed at UHP conditions in the literature. There may be several reasons for this shortcoming. First, it is possible that the shear zones developed during progressive burial at HP and UHP metamorphic conditions are obliterated by progressive metamorphism. Second, they may be continuously active during burial and exhumation, or repeatedly reactivated, with localization possibly controlled by the bulk composition of a weak rock type. In this case the microstructures developed at higher metamorphic grade would become replaced by those developed later at lower grade during exhumation, with the high grade fabrics being erased. Third, it appears possible that the (micro)structural record of high strain deformation at (U)HP conditions is unspecific, due to specific deformation mechanisms predominating at HP and in particular at UHP metamorphic conditions. In view of the apparent lack of strongly deformed rocks in crustal slices rapidly exhumed from a deep-reaching mega-shearzone, this third possibility is particularly attractive and deserves careful evaluation.

Upper bound on flow stress

Undeformed rocks and/or the unequivocal microstructural record of rocks deformed exclusively by dissolution–precipitation creep prove that stresses (compared to temperatures) were too low to drive dislocation creep along the entire trajectory. As such, these rocks serve as a kind of 'flight recorder', a stress gauge that would have recorded episodes of sufficient stress by corresponding permanent strain and the respective microfabrics indicative of deformation by dislocation creep. Taking a strain rate on the order of $10^{-15}\,s^{-1}$ as equivalent to no notable deformation on the time scales of subduction and exhumation (1 to 10 Ma), an upper bound to stress as a function of temperature is imposed by extrapolation (subject to the uncertainties outlined above) of the flow laws for dislocation creep of quartz/coesite or aragonite to this strain rate (Fig. 5).

As a first approximation, the kinematic framework of a subduction zone suggests a layered structure (e.g. Ji & Zhao 1994; Kirby *et al.* 1996). Also, the isotherms are closely spaced and subparallel to the subduction shear zone (e.g. Peacock 1996). Evidently, interplate deformation should then be localized in the weakest layer of this sandwich structure, the strength of which in turn limits the average shear stress and the degree of interplate coupling. If this is true, the preservation and exhumation of undeformed rocks, combined with the bounds imposed by extrapolation of the experimental flow laws, i.e. the 'flight recorder concept' outlined above, indicates that the interplate shear zones must be very weak, with a very low effective viscosity of the respective materials. The same holds true for rocks deformed exclusively by mechanisms other than dislocation creep. Evidently, this conclusion is rather robust, as the eventually identified microstructural record of deformation by dislocation creep may simply reflect a temporary local stress concentration, related to spatial heterogeneity of the crust (e.g. Schwarz & Stöckhert 1996), or to instantaneous stress redistribution during seismic events (e.g. Taylor *et al.* 1996). In contrast, the failure of the exhumed stress gauge to indicate any deformation by dislocation creep along its entire particle path is clearly more significant, as it precludes attainment of the required stress level (as a function of temperature) throughout the entire burial and exhumation history.

For continental material, with a significant volumetric proportion of quartz respectively coesite, the flow laws depicted in Fig. 5 pose an upper bound to differential stresses of <10 MPa at the typical temperatures of HP-metamorphism (using the quartz flow law of Paterson & Luan, 1990), and barely a few MPa at the typical temperatures of UHP-metamorphism, using the flow law for coesite (Renner *et al.* 2001). Likewise, the flow law for aragonite (Rybacki *et al.*, in prep.) indicates that in the case of the aragonite marbles exposed on Crete, which remained undeformed or exclusively deformed by dissolution–precipitation creep, differential stress never exceeded a few MPa at the peak temperatures of $400 \pm 50\,^{\circ}C$.

Clearly the extrapolation of experimental flow laws of rock-forming minerals to typical geological strain rates is subject to considerable uncertainty (e.g. Paterson 1987). As the quoted uncertainties in the experimental data depend on the mode of correction and data reduction, which in turn depend on the understanding of the processes in the experimental sample assembly, the reader is referred to the details given in the original papers. In the case of quartz, comparison of the predictions of experimental flow laws with both the microstructural and thermometric record of exhumed rocks on one hand (Stöckhert *et al.* 1999; Hirth *et al.* 2001), and with the depth distribution of intracontinental earthquakes (Chen & Molnar 1983; Meissner & Strehlau 1982) on the other hand, supports the validity of the extrapolation. Furthermore, the flow law for coesite displayed in Fig. 5 represents an upper bound to strength (Renner *et al.* 2001). The upper bounds on the magnitude of stress in subduction zones proposed in this paper are based on these flow laws for quartz and coesite, respectively. Notwithstanding the remaining uncertainty, it should be born in mind that unequivocal microstructural evidence of deformation in the dislocation creep regime is common in metamorphic rocks from a broad variety of tectonic settings. The typical absence of such microstructures in HP and UHP metamorphic rocks indicates comparatively low stress levels at the given temperatures – independent of the problems inherent in the extrapolation of experimental flow laws.

Comparison with information on stress in present day subduction zones

The magnitude of shear stress in subduction zones has been inferred from seismological information (e.g. Choy & Boatwright 1995), from heat flow data (e.g. von Herzen *et al.* 2001; Hyndman & Wang 1993), and theoretical models (e.g. Molnar & England 1990; Wang & He 1999). The typical values proposed by these studies span two orders of magnitude, ranging from close to zero (e.g. Wang & He 1999; Wang 2000) to several tens of MPa (e.g. von Herzen *et al.* 2001), or even 100 MPa (e.g. Molnar & England 1990). Clearly, the stress field can be highly heterogeneous on various length scales, and thus a 'typical value' for a subduction zone may be a spurious number. However, the fact that rocks have been buried and exhumed without being subject to sufficient stress to drive deformation of their major constituents by dislocation creep at any time is consistent with those results indicating a very low shear stress, at least at elevated temperatures beyond the seismogenic zone (Tichelaar & Ruff 1993; Ruff & Tichelaar 1996).

A seismological quantity derived from radiation spectra of earthquakes (e.g. Scholz 1990; Stacey 1992) is the stress drop. Although the physical significance of the calculated values is not entirely clear and the quantitative result is model-dependent (Beresnev 2001), stress drop is generally accepted to range between about 1–10 MPa, which is the same order as the upper bound to differential stress in subduction zones posed by the microstructural record of exhumed (U)HP metamorphic rocks combined with experimental flow laws. In this case, with information so far restricted to the depth range $<c.$ 150 km, the stress drop for subduction zone earthquakes should be close to total and the stored energy should be almost quantitatively radiated as seismic waves, with a very high efficiency (e.g. Stacey 1992). Negligible remaining stress would be consistent with lubrication by a fluid and effective decoupling along the fault plane due to zero effective pressure.

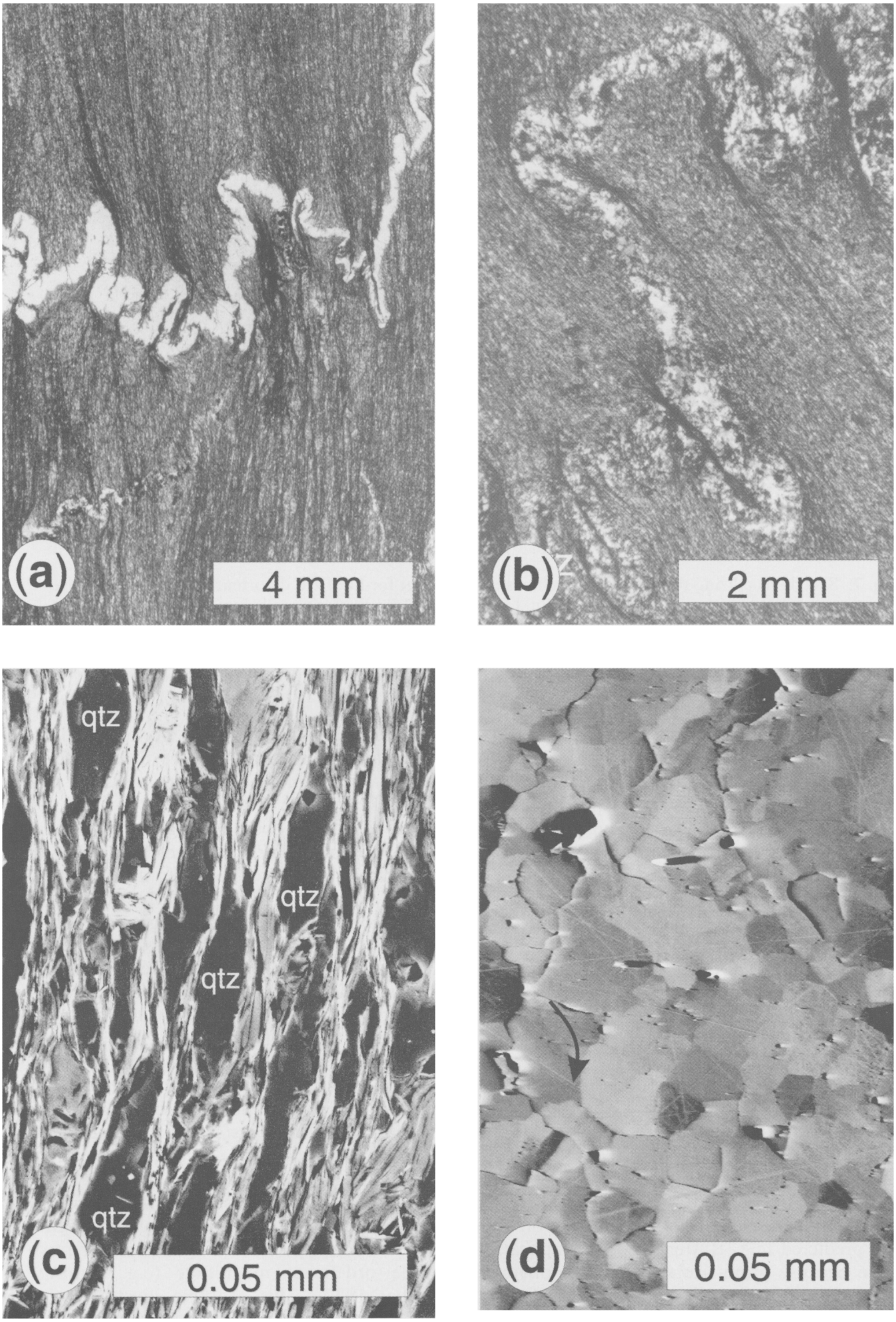
Suspected deformation mechanisms

At present, predictions of the potential deformation mechanisms in the low viscosity shear zones are speculative. At any rate, volatiles are carried down in subduction zones in huge amounts and liberated as a fluid phase during dehydration reactions. High pressures strongly enhance the solubility of rock-forming minerals in aqueous solutions (e.g. Manning 1994; Shen & Keppler 1997) resulting in an extraordinarily high concentration of solutes at conditions of HP and UHP metamorphism (e.g. Scambelluri *et al.* 1998; Scambelluri & Philippot 2001). In view of the continuous fluid supply, likely deformation mechanisms (apart from the possibility of brittle failure at very low effective pressure $\sigma_{ii}/3 - P_{fluid}$) could be dissolution–precipitation creep and grain boundary sliding with grain shape accomodation by dissolution and precipitation in the presence of an interconnected aqueous fluid or a hydrous melt (Fig. 6). Constitutive equations to predict the behaviour of materials undergoing deformation by fluid or melt assisted granular flow, with a melt fraction below the critical value (e.g. Rosenberg 2001) for flow to be concentrated in the melt itself, have been proposed by Paterson (1995, 2001). In this model, strain rate is a function of effective grain or aggregate size d (decreasing proportional to $1/d^2$) and linearly dependent on stress (Newtonian behaviour), controlled by either viscosity of the flowing melt, rate of diffusion within the melt, or rate of reaction at the solid–melt interfaces (Paterson 2001). Experimental studies on partially molten mantle rocks in the diffusion creep regime have shown that the strain rate is increased by a factor of 25 in the presence of 7% of melt (Hirth & Kohlstedt 1995).

Microstructures indicative of such a deformation mechanism operative at high temperatures in the presence of a supercritical fluid at (U)HP metamorphic conditions have not been identified in natural rocks so far, and, unfortunately, the respective microstructural record cannot be expected to be unequivocal. Any synkinematic fabrics are likely to be obliterated by grain growth and mineral reactions when deformation is accompanied or followed by annealing.

For the pressure and temperature realm of UHP metamorphism, the recent experimental identification of supercritical fluids (Shen & Keppler 1997; Bureau & Keppler 1999) in aqueous silicate systems has opened new perspectives. At these conditions, there is no coexistence between a low density/low viscosity fluid and a high density/high viscosity silicate melt. Supercritical fluids with intermediate properties are formed. Inclusions of these fluids, now crystallized to yield a complex mineral assemblage including diamond, have been identified in garnet from UHP metamorphic gneisses from the Erzgebirge, Germany (Stöckhert *et al.* 2001). Experimental investigations on the distribution of such fluids at typical UHP metamorphic conditions ($T = 700\,^{\circ}C$ at $P = 3.5$ GPa, melt fraction about 10%) have shown (Mönicke *et al.* 2001) that the fluid-filled interstices are characterized by a high ratio γ_{ss}/γ_{sl} (with γ_{ss} denoting the interfacial free energy between crystals, and γ_{sl} that between the fluid and a crystal) and by predominantly rational (low-Miller index) interfaces, resulting in a highly irregular and complex shape of the

Fig. 6. Example from Franciscan HP–LT metamorphic siliciclastic series (Pacheco Pass), visualizing the contrast in effective viscosity between single phase material (folded quartz vein) undergoing deformation by dislocation creep and polyphase material (phyllite) undergoing deformation by dissolution–precipitation creep. **(a)** and **(b)** are optical micrographs, **(c)** and **(d)** are SEM images of the phyllite and the quartz vein, respectively. Note that, as a first approximation, the grain size is identical in both materials. The fold shape suggests a viscosity contrast of two to three orders of magnitude (e.g. Ramsay & Huber 1987).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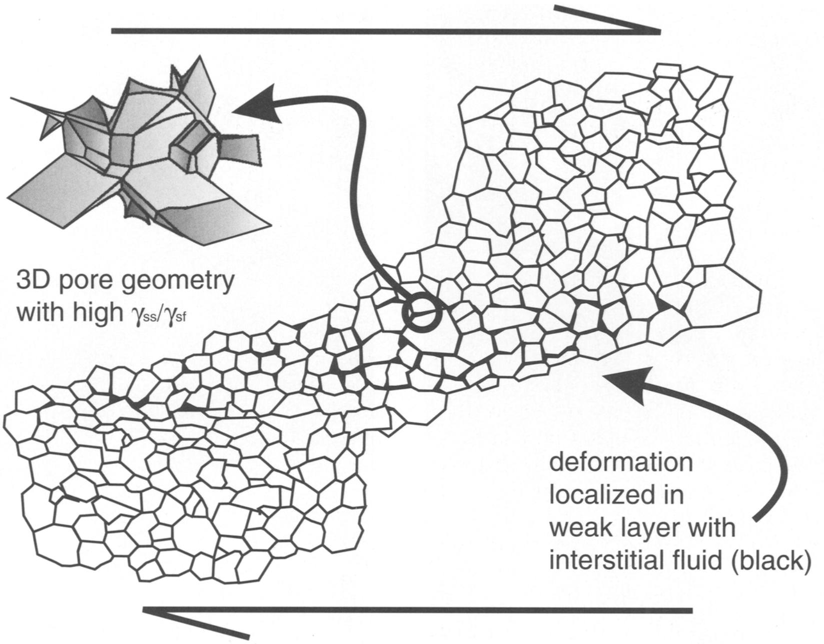

Fig. 7. Sketch visualizing the proposed mechanism to account for localized deformation and low effective viscosity at UHP metamorphic conditions: fluid assisted granular flow, controlled by the quantity and distribution of (supercritical) fluid in different rock types. The characteristic shape of the fluid-filled interstitials is based on experimental studies at UHP metamorphic conditions (Mönicke *et al.* 2001). Stress becomes concentrated at the edges of the wedge-shaped offshoots.

fluid-filled interstices with a low volume/interface area ratio (Laporte & Provost 2000). Owing to their shape, the interstices are interconnected in three dimensions, but the solid portions of the rock are polycrystalline and polyphase aggregates, as considered in the model proposed by Paterson (2001). The further exploration of the composition, reaction kinetics, density, viscosity and pore geometry of these supercritical fluids for a variety of host rocks at (U)HP conditions will be an important task for the near future, as the rheology of deeply subducted crust and in particular the localization of deformation into narrow shear zones possibly related to specific rock compositions (Fig. 7) may crucially depend on the properties and distribution of the interstitial fluids.

For both of the interrelated deformation mechanisms inferred above, i.e. dissolution–precipitation creep and fluid assisted granular flow, grain-size-sensitive behaviour and a linear (Newtonian) dependence of strain rate on stress are predicted (e.g. Rutter 1983). Complications arise because grain-size sensitivity is obviously not sufficient to describe the influence of microstructure on dissolution–precipitation creep. Instead, the portion and nature of the interfaces between unlike minerals controls viscosity (Schwarz & Stöckhert 1996; Stöckhert *et al.* 1999), as also indicated by experimental studies (e.g. Hickman & Evans 1995; Bos *et al.* 1999). This is illustrated here using an example of a very low grade HP-metamorphic phyllite from the Pacheco Pass area in the Franciscan Mélange (e.g. Ernst 1993), where an early formed quartz vein became folded during progressive deformation of the phyllitic matrix (Fig. 6). The microstructure of the monophase quartz vein indicates deformation by dislocation creep, with a recrystallized grain size of *c.* 0.005–0.01 mm. In contrast, the phyllitic matrix indicates deformation exclusively by dissolution–precipitation creep, with clastic quartz grains truncated at their interphase boundary with mica flakes, but not deformed by crystal plastic processes and not recrystallized. The diameter of the flattened clastic quartz grains in the phyllite compares to the recrystallized grain size in the quartz vein. The shape of the folded vein suggests a high contrast in effective viscosity, which may be two to three orders of magnitude (e.g. Ramsay & Huber 1987). This example illustrates that a polyphase rock capable of undergoing deformation by dissolution–precipitation creep has a much lower strength compared to a single-phase quartz vein, notwithstanding the similar grain size. Constitutive equations relevant to natural rock deformation and crustal properties must take this effect into account (e.g. Wheeler 1992), with a purely grain-size sensitive term being insufficient. Furthermore, significant stress concentration can occur in specific layers, the microstructural record of which will therefore not reflect the typical average stress level.

At present, the effective viscosities of polyphase rocks undergoing deformation by dissolution–precipitation creep and fluid-assisted granular flow are poorly constrained by experimental results. From the viscosity contrast with material that was deformed by dislocation creep and for which experimental flow laws are available, effective viscosities on the order of 10^{19} Pa s or below have been suggested for low grade HP-metamorphic phyllites on Crete deformed by dissolution–precipitation creep (Stöckhert *et al.* 1999). This order compares well to the viscosities proposed by Shreve and Cloos (1986) and Cloos and Shreve (1988*a*, *b*) for sedimentary material in subduction channels.

For a given grain size d, increasing temperature will decrease the effective viscosity. However, grain size in metamorphic rocks generally increases with increasing temperature, and the increase in viscosity of a material undergoing deformation by dissolution–precipitation creep or fluid-assisted granular flow is proportional to d^2 or d^3. As such, the effect of increasing temperature on viscosity may be counterbalanced to some extent by increasing grain size, and viscosity changes with depth may be moderate.

Implications for the kinematic pattern and tectonic models

In combination with the P–T paths, the high exhumation rates pose upper bounds on the size of the exhumed bodies and thus stimulate questions concerning the kinematics and driving forces for exhumation, and in particular the inherent rheological aspects. There are a number of concurrent exhumation models (see Platt 1993 for a review). The shape of the P–T paths in combination with the constraints on the relevant time-scale discussed above, i.e. exhumation rates corresponding to plate velocity, and are best reconciled with the corner flow model (e.g. Cloos & Shreve 1988*a*, *b*). This concept is also consistent with the limited size of many (U)HP metamorphic rock bodies and their intimate association with metamorphic rocks revealing markedly different P–T conditions or P–T paths (e.g. Chopin *et al.* 1991; Henry *et al.* 1993; Eide 1995; Dong *et al.* 1998; Cong *et al.* 1999; Chopin & Schertl 1999; Terry *et al.* 2000*a*). One possible exception is the very extensive UHP metamorphic area in the Dabie/Sulu region in China which may constitute a coherent continental slice (Hacker *et al.* 2000; Ye *et al.* 2000*a*, *b*), contrasting with the small slices identified elsewhere. Chemenda *et al.* (1995) have proposed a model for the exhumation of coherent continental crust driven by buoyancy. However, as the ultimate strength of continental crust at UHP metamorphic conditions should be controlled by coesite, Stöckhert and Renner (1998) and Renner *et al.* (2001) have questioned this concept, because at typical UHP metamorphic conditions the crust should be too weak to withstand the buoyancy forces.

Exhumation of small and internally undeformed (U)HP metamorphic slices by forced flow in a subduction channel narrowing with depth requires a low viscosity matrix, which has not so far been unequivocally identified. Possible reasons for this have been considered above in the discussion of the nature of the shear zones, which in their entity can be considered as equivalent to what is referred to as matrix here.

It is possible that – at least to some extent – the low viscosity matrix is represented by serpentinite derived from the progressively hydrating hanging wall of the subduction channel (Peacock 1993; Peacock & Hyndman 1999; Gerya *et al.* 2001). Serpentinite and peridotite are intimately associated with most occurrences of smaller (U)HP metamorphic slices in mélange type associations (e.g. Little *et al.* 1993; Blake *et al.* 1995; Parkinson 1996; Guillot *et al.* 2000) and serpentine mud diapirism rising from up to 25 km depth with incorporated blueschist facies blocks has been observed in the active Mariana forearc (Fryer *et al.* 1999).

Three scenarios for subduction zones with a low viscosity subduction channel are displayed in Fig. 8. Figure 8a illustrates the simple situation with continuous subduction and no return flow, which is probably irrelevant when considering the record of rapidly exhumed HP and UHP metamorphic rocks. In this case, exhumation of HP and UHP rocks could only be achieved after collision. Figure 8b introduces return flow in a narrowing subduction channel, with exhumed material forming an orogenic wedge, following the principle proposed by Cloos (1982) for the Franciscan complex in California. Figure 8c shows a scenario derived from the model shown in Fig. 8b, but introduces an important modification. Here, forced flow in the low viscosity

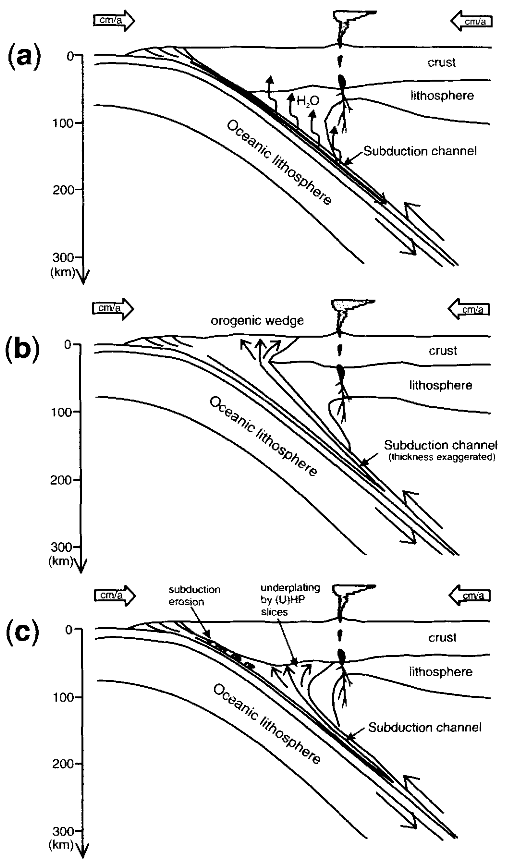

Fig. 8. Three scenarios for subduction zones, based on the subduction channel concept (not to scale). (**a**) Subduction flux greatly exceeds return flux; no exhumation of (U)HP metamorphic rocks driven by forced flow. (**b**) Subduction channel with exhumation by forced flow into an accretionary prism. (c) Subduction channel with forced flow and exhumation to a deep crustal level, from which the rocks may be further exhumed after continental collision at a later stage, and thus after a prolonged history with P, T depending on their successive position in the underplated volume. In contrast to model (**b**), model (**c**) allows subduction erosion to take place concomitantly with exhumation. Also, it allows long-term storage of former (U)HP metamorphic material at a deep crustal level, with the record of (U)HP metamorphism being selectively erased, depending on rock type and presumably availability of fluids.

subduction channel (controlled by rheology and thus by realization of the boundary conditions required to activate the specific deformation mechanisms envisaged above) is restricted to the P–T realm of HP and UHP metamorphism. Material delivered by return flow is stored at the base of the crust of the upper plate, eventually losing its (U)HP memory with time, but with a fair chance of becoming exhumed after collision. This scenario reconciles with the evidence that UHP metamorphism seems to happen prior to collision (e.g. Eide & Liou 2000), and not as a consequence of collision. Note that the scenario in Fig. 8c takes into account the possibility of subduction erosion (von Huene & Scholl 1991), with material removed from the front of the upper plate carried down into the subduction channel. This implies that UHP metamorphic continental crust is not necessarily part of the downgoing plate, but can be derived from an active continental margin during continuous subduction of oceanic lithosphere. This scenario appears to be consistent with petrological, geochronological and (micro)structural information, and may also reconcile with the depth distribution of subduction thrust earthquakes (e.g. Ruff & Tichelaar 1996; Peacock & Hyndman 1999).

Are these conditions valid for subduction zones in general?

All conclusions drawn here are subject to the general validity of the record of the exhumed (U)HP-metamorphic rocks. In fact, it may be strongly biased, since exhumation may be highly selective and restricted to specific types of subduction zones or episodes of reorganization. It is possible that exhumation during plate convergence is facilitated at retreating margins undergoing subduction roll back (e.g. Royden 1993). Notably, exhumed high-pressure metamorphic rocks are merely absent along the active continental margin of the Andes. There, corresponding to the scenario illustrated in Fig. 8c, HP and UHP metamorphic rocks may be stored at deeper crustal levels of the upper plate (and eventually losing their memory) while subduction still continues. These rocks may reach the surface only after collision and underplating by continental crust, as in most part of the Alps.

Notwithstanding the uncertainties about their general validity, the features discussed in this paper indicate that, at least under certain conditions, extensive volumes of rock can be buried at convergent plate boundaries to depths well exceeding 100 km, and then return to normal crustal levels within a time span of 1 to 10 Ma, without undergoing any significant deformation. This means that they were not subject, at any time during their burial and exhumation history, to the level of differential stress (as a function of temperature) required for deformation by dislocation creep, or in cases even for any deformation by whatsoever mechanism.

Summary and conclusions

Amazingly, we can learn a lot about stress and deformation in subduction zones from undeformed rocks, or from rocks that underwent deformation by dissolution–precipitation creep or fluid-assisted granular flow, a mechanism for which appropriate constitutive equations and experimental calibrations allowing application to natural conditions are not yet available. Following the principle of a 'flight recorder', exhumed rocks are considered as a stress gauge that would provide an appropriate signal, if subject to a sufficient stress (as a function of temperature, and for appropriate time span) to cause a permanent deformation. Although deformation by dislocation creep seems to be only exceptional in subduction zones, the experimental flow laws for this deformation regime can be applied to provide an upper bound to stress.

If the microstructural record of exhumed (U)HP metamorphic rocks is of general significance for subduction zones, and if we accept the validity of experimental flow laws and extrapolation of laboratory data as feasible, a tentative summary and account of implications of these findings looks as follows.

(1) The entire absence of deformation observed in some (U)HP metamorphic rocks, and the absence of deformation during exhumation prior to reaching a normal crustal depth in many, is much more significant than any record of deformation identified so far.

(2) Deformation must be highly localized, although the microstructural record of the shear zones appears to be incomplete, either because it is not specific and hence not recognized, or obliterated by later annealing and progressive reactions.

(3) A well-recorded deformation mechanism in lower grade HP metamorphic rocks is dissolution–precipitation creep. At greater depth, supercritical fluids at partially or completely wetted grain and solid phase boundaries are suspected to facilitate fluid assisted granular flow and thus to control crustal strength.

(4) Both deformation mechanisms imply Newtonian behaviour, with the effect of increasing temperature on viscosity counterbalanced to some extent by increasing grain size.

(5) As a large-scale model, corner flow in a narrowing subduction channel appears feasible for these low viscosities, allowing exhumation of internally undeformed small slices with plate velocity and cooling during decompression.

(6) Interplate shear stress is expected to be very low in subduction zones (a few MPa or less) and thus of the same order as the stress drop inferred for seismic events.

(7) The previous point implies that the stress drop for earthquakes may be total rather than partial, and that the efficiency of seismic radiation is high.

(8) Finally, any significant contribution of shear heating to the thermal budget of subduction zones is unlikely.

In the present context there appear to be two research goals still outstanding. The first is to develop a better understanding of dissolution–precipitation creep and fluid-assisted granular flow in complex polyphase materials. This includes derivation of constitutive equations and experimental calibration, which are needed as input for more realistic numerical simulations, and a more complete and objective assessment of deformation at depth in (U)HP terranes. In the case of evidence for deformation by crystal plastic processes, the possibilities of local stress concentration related to spatial heterogeneity, and of short-term deformation related to rapid synseismic loading should be carefully explored. The second goal is to improve the integration of geophysical field studies with laboratory experiments, numerical simulations and, last but not least, the record of natural rocks, as here (in contrast to Hutton's famous statement) the past may be the key to the present.

This paper summarizes some basic concepts and ideas developed over the last decade, with continuous benefit from discussions with many colleagues, in particular J. Renner, E. Rybacki, R. Wirth, G. Dresen, W. Schreyer, H.-J. Massonne, H.-P. Harjes, to name just a few. The organizers of the DRT 2001 meeting at Nordwijkerhout are thanked for the invitation to present these results and thoughts as a keynote. S. Thomson is thanked for correcting the English. The reviews by Harry Green and by an anonymous referee, who urged me to add an additional note of caution concerning the extrapolation of experimental flow laws, are gratefully acknowledged. Financial support has been provided by the Deutsche Forschungsgemeinschaft within the scope of the Research Group 'High Pressure Metamorphism in Nature and Experiment' and within the Collaborative Research Center SFB 526 'Rheology of the Earth – from the Upper Crust into the Subduction Zone'.

References

Amato, J. M., Johnson C. M., Baumgartner, L. P. & Beard, B. L. 1999. Rapid exhumation of the Zermatt-Saas ophiolite deduced from high-precision Sm-Nd and Rb-Sr geochronology. *Earth and Planetary Science Letters*, **171**, 425–438.

BEARTH, P. 1959. Über Ekogite, Glaukophanschiefer und metamorphe Pillowlaven. *Schweizer Mineralogisch-Petrologische Mitteilungen*, **39**, 267–286.

BERESNEV, I. A. 2001. What we can and cannot learn about earthquake sources from the spectra of seismic waves. *Bulletin of the Seismological Society of America*, **91**, 397–400.

BIINO, G. & COMPAGNONI, R. 1992. Very-high pressure metamorphism of the Brossasco coronite metagranite, southern Dora-Maira massif, western Alps. *Schweizerische Mineralogische und Petrographische Mitteilungen*, **72**, 347–363.

BLAKE JR., M. C., MOORE, D. E. & JAYKO A. S. 1995. The role of serpentinite melanges in the unroofing of UHPM rocks: an example from the Western Alps of Italy. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 182–205.

BOLAND, J. N. & TULLIS, T. E. 1986. Deformation behavior of wet and dry clinopyroxenite in the brittle to ductile transition region. *Geophysical Monograph*, **36**, 35–49.

BOS, B., PEACH, C. J. & SPIERS, C. J. 1999. Frictional-viscous flow of simulated fault gouge caused by the combined effects of phyllosilicates and pressure solution. *Tectonophysics*, **327**, 173–194.

BRUNO, M., COMPAGNONI, R. & RUBBO, M. 2001. The ultra-high pressure coronitic and pseudomorphous reactions in a metagranodiorite from the Brossasco-Isasca Unit, Dora-Maira Massif, western Italian Alps: a petrographic study and equilibrium thermodynamic modelling. *Journal of Metamorphic Geology*, **19**, 33–43.

BUATIER, M., VAN ROERMUND, H., DRURY, M. R. & LARDEAUX, J. M. 1991. Deformation and recrystallization mechanisms in naturally deformed omphacites from the Sesia-Lanzo zone; geophysical consequences. *Tectonophysics*, **195**, 11–27.

BUREAU, H. & KEPPLER, H. 1999. Complete miscibility between silicate melts and hydrous fluids in the upper mantle: experimental evidence and geochemical implications. *Earth and Planetary Science Letters*, **165**, 187–196.

CARSWELL, D. A. & ZHANG, R. Y. 1999. Petrographic characteristics and metamorphic evolution of ultrahigh-pressure eclogites in plate-collision belts. *International Geology Review*, **41**, 781–798.

CARSWELL, D. A., WILSON, R. N. & ZHAI, M. 2000. Metamorphic evolution, mineral chemistry and thermobarometry of schists and orthogneisses hosting ultra-high pressure eclogites in the Dabieshan of Central China. *Lithos*, **52**, 121–155.

CHEMENDA, A. I., MATTAUER, M., MALAVIEILLE, J. & BOKUN, A. N. 1995. A mechanism for syn-collisional rock exhumation and associated normal faulting: results from physical modelling. *Earth Planetary Science Letters*, **132**, 225–232.

CHEN, W. P. & MOLNAR, P. 1983. Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. *Journal of Geophysical Research*, **88**, 4183–4214.

CHOPIN, C. & SCHERTL, H. P. 1999. The UHP Unit in the Dora-Maira-Massif, Western Alps. *International Geology Review*, **41**, 765–780.

CHOPIN, C., HENRY, C. & MICHARD, A. 1991. Geology and petrology of the coesite-bearing terrain, Dora Maira massif, western Alps. *European Journal of Mineralogy*, **3**, 263–291.

CHOY, G. L. & BOATWRIGHT, J. L. 1995. Global patterns of radiated seismic energy and apparent stress. *Journal of Geophysical Research*, **100**, 18205–18228.

CLOOS, M. 1982. Flow melanges: numerical modeling and geologic constraints on their origin in the Franciscan subduction complex, California. *Geological Society of America Bulletin*, **93**, 330–345.

CLOOS, M. & SHREVE, R. L. 1988*a*. Subduction-channel model of prism accretion, melange formation, sediment subduction, and subduction erosion at convergence plate margins: 1. background and description. *Pure and Apllied Geophysics*, **128**, 455–500.

CLOOS, M. & SHREVE, R. L. 1988*b*. Subduction-channel model of prism accretion, melange formation, sediment subduction, subduction erosion at convergence plate margins: 2. implications and discussion. *Pure and Applied Geophysics*, **128**, 501–545.

COLEMAN, R. G. & WANG, X. 1995. Overview of the geology and tectonics of UHPM. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 1–32.

COMPAGNONI, R., HIRAJIMA, T. & CHOPIN, C. 1995. Ultra-high-pressure metamorphic rocks in the Western Alps. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 206–243.

CONG, B., WANG, Q. & ZHAI, M. 1999. New data regarding hotly debated topics concerning UHP metamorphism of the Dabie-Sulu belt, East-Central China. *International Geology Review*, **41**, 827–835.

DONG, S., CHEN, J. . & HUANG, D. 1998. Differential exhumation of tectonic units and ultahigh-pressure metamorphic rocks in the Dabie Mountains, China. *The Island Arc*, **7**, 174–183.

EIDE, E. A. 1995. A model for the tectonic history of HP and UHPM regions in east central China. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 391–426.

EIDE, E. A. & LIOU, J. G. 2000. High-pressure blueschists and eclogites in Hońgan: a framework for addressing the evolution of high- and ultra-high-pressure rocks in central China. *Lithos*, **52**, 1–22.

ERNST, W. G. 1993. Metamorphism of Franciscan tectonostratigraphic assemblage, Pacheco Pass area, east-central Diablo Range, California. Coast Ranges. *Geological Society of America Bulletin*, **105**, 618–636.

ERNST, W. G. 1999. Metamorphism, partial preservation, and exhumation of ultrahigh-pressure belts. *The Island Arc*, **8**, 125–153.

ERNST, W. G. & LIOU, J. G. 1999. Overview of UHP metamorphism and tectonics in well-studied collisional orogens. *International Geology Review*, **41**, 477–493.

ERNST, W. G. & PEACOCK, S. M. 1996. A thermotectonic model for preservation of ultrahigh-pressure phases in metamorphosed continetal crust. *In*: BEBOUT, G. E., SCHOLL, D. W., KIRBY, S. H. & PLATT, J. P. (eds) *Subduction Top to Bottom*. American Geophysical Union, Washington, 171–178.

EVANS, B. & KOHLSTEDT, D. L. 1995. Rheology of rocks. *In*: AHRENS, T. J. (ed.) *Rock Physics & Phase Relations – A Handbook of Physical Constants*. American Geophysical Union, Washington, 148–165.

FAURE, M., LIN, W., SHU, L., SUN, Y. & SCHÄRER, U. 1999. Tectonics of the Dabieshan (eastern China) and possible exhumation mechanism of ultra high-pressure rocks. *Terra Nova*, **11**, 251–258.

FRYER, P., WHEAT, C. G. & MOTTL, M. J. 1999. Mariana blueschist mud volcanism: Implications for conditions within the subduction zone. *Geology*, **27**, 103–106.

GEBAUER, D., SCHERTL, H.-P., BRIX, M. & SCHEYER, W. 1997. 35 Ma old ultrahigh-pressure metamorphism and evidence for very rapid exhumation in the Dora Maira Massif, Western Alps. *Lithos*, **41**, 5–24.

GERYA, T. V., STÖCKHERT, B., MARESCH, W. V. & PERCHUK, A. L. 2001. Exhumation dynamics of high-pressure rocks in a hydrated mantle wedge: constraints from 2D numerical experiments. *Beihefte zum European Journal of Mineralogy*, **13**, 64.

GODARD, G. & VAN ROERMUND, H. 1995. Deformation-induced clinopyroxene fabeics from eclogites. *Journal of Structural Geology*, **17**, 1425–1443.

GUILLOT, S., HATTORI, K. H. & DE SIGOYER, J. 2000. Mantle wedge serpentinization and exhumation of eclogites: insights from eastern Ladakh, northwest Himalaya. *Geology*, **28**, 199–201.

HACKER, B. 1996. Eclogite formation and the rheology, buoyancy, seismicity and H_2O content of oceanic crust. *In*: BEBOUT, G. E., SCHOLL, D. W., KIRBY, S. H. & PLATT, J. P. (eds) *Subduction Top to Bottom*. American Geophysical Union, Washington, 337–346.

HACKER, B. R. & PEACOCK S. M. 1995. Creation, preservation, and exhumation of UHPM. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 159–181.

HACKER, B. R., RATSCHBACHER, L., WEBB, L. & SHUWEN, D. 1997. What brought them up? Exhumation of the Dabie Shan ultrahigh-pressure rocks. *Geology*, **23**, 743–746.

HACKER, B. R., RATSCHBACHER, L. *ET AL*. 2000. Exhumaton of ultrahigh-pressure continental crust in east central China: Late Triassic–Early Jurassic tectonic unroofing. *Journal of Geophysical Research*, **105**, 13339–13364.

HARLEY, S. L. & CARSWELL, D. A. 1995. Ultradeep crustal metamorphism: a prospective view. *Journal of Geophysical Research*, **100**, 8367–8380.

HENRY, C., MICHARD, A. & CHOPIN, C. 1993. Geometry and structural evolution of ultra-high-pressure and high-pressure rocks from the Dora Maira massif, western Alps, Italy. *Journal of Structural Geology*, **15**, 965–981.

HICKMAN, S. & EVANS, B. 1995. Kinetics of pressure solution at halite–silica interfaces and intergranular clay films. *Journal of Geophysical Research*, **100**, 13113–13132.

HIRAJIMA, T., WALLIS, S. R., ZHAI, M. & YE, K. 1993. Eclogitized metagranitoid from the Su-Lu ultra-high pressure province, eastern China. *Proceedings of the Japan Academy*, **69**, Series B, 249–254.

HIRTH, G. & KOHLSTEDT, D. L. 1995. Experimental constraints on the dynamics of the partially molten upper mantle: deformation in the diffusion creep regime. *Journal of Geophysical Research*, **100**, 1981–2001.

HIRTH, G., TEYSSIER, C. & DUNLAP, W. J. 2001. An evaluation of quartzite flow laws based on comparisons between experimentally and naturally deformed rocks. *International Journal of Earth Sciences*, **90**, 77–87.

HYNDMAN, R. D. & WANG, K. 1993. Thermal constraints on the zone of major thrust earthquake failure: the Cascadia subduction zone. *Journal of Geophysical Research*, **98**, 2039–2060.

JI, S. & ZHAO, P. 1994. Layered rheological structure of subducting oceanic lithosphere. *Earth and Planetary Science Letters*, **124**, 75–94.

JIN, Z. M., ZHANG, J., GREEN, H. W. & JIN, S. 2001. Eclogite rheology: implications for subducted lithosphere. *Geology*, **29**, 667–670.

KANEKO, Y., MARUYAMA, S., *ET AL*. 2000. Geology of the Kikchetav UHP-HP metamorphic belt, Northern Kazakhstan. *The Island Arc*, **9**, 264–283.

KIRBY, S., ENGDAHL, E. R. & DENLINGER, R. 1996. Intermediate-depth intraslab earthquakes and arc volcanism as physical expressions of crustal and uppermost mantle metamorphism in subducted slabs. *In*: BEBOUT, G. E., SCHOLL, D. W., KIRBY, S. H. & PLATT, J. P. (eds) *Subduction Top to Bottom*. American Geophysical Union, Washington, 195–214.

KÜSTER, M. & STÖCKHERT, B. 1997. Density changes of fluid inclusions in high-pressure low-temperature metamorphic rocks from Crete: A thermobarometric approach based on the creep strength of the host minerals. *Lithos*, **41**, 151–167.

LAPORTE, D. & PROVOST, A. 2000. Equilibrium geometry of a fluid phase in a polycrystalline aggregate with anisotropic surface energies: dry grain boundaries. *Journal of Geophysical Research*, **105**, 25937–25953.

LIOU, J. G. 1999. Petrotectonic summary of less intensively studied UHP regions. *International Geology Review*, **41**, 571–586.

LITTLE, T. A., HOLCOMBE, R. J. & SLIWA, R. 1993. Structural evidence für extensional exhumation of Blueschist-bearing serpentinite matrix melange, New England Orogen, Southeast Queensland, Australia. *Tectonics*, **12**, 536–549.

LIU, M. & YUND, R. A. 1993. Transformation kinetics of polycrystalline aragonite to calcite – new experimental data, modeling, and implications. *Contributions to Mineralogy and Petrology*, **114**, 465–478.
MANNING, C. E. 1994. The solubility of quartz in H_2O in the lower crust and upper mantle. *Geochimica et Cosmochimica Acta*, **58**, 4831–4839.
MARUYAMA, S. & PARKINSON, C. D. 2000. Overview of the geology, petrology and tectonic framework of the high-pressure–ultrahigh-pressure metamorphic belt of the Kokchetav Massif, Kazakhstan. *The Island Arc*, **9**, 439–455.
MASSONNE, H.-J. 1999. A new occurrence of microdiamonds in quartzofeldspathic rocks of the Saxonian Erzgebirge, Germany, and their metamorphic evolution. *Proceedings of the VIIth International Kimberlite Conference: Capetown*, **2**, 533–539.
MAULER, A., BYSTRICKY, M., KUNZE, K. & MACKWELL, S. 2000. Microstructures and lattice preferred orientations in experimentally deformed clinopyroxene aggregates. *Journal of Structural Geology*, **22**, 1633–1648.
MEISSNER, R. & STREHLAU, J. 1982. Limits of stresses in continental crusts and their relation to the depth-frequency distribution of shallow earthquakes. *Tectonics*, **1**, 73–89.
MICHARD, A., HENRY, C. & CHOPIN, C. 1995. Structures in UHPM rocks: a case study from the Alps. *In*: COLEMAN, R. G. & WANG, X. (eds) *Ultrahigh Pressure Metamorphism*. Cambridge University Press, Cambridge, 132–158.
MICHARD, D., CHOPIN, C. & HENRY, C. 1993. Compression versus extension in the exhumation of the Dora-Maira coesite-bearing unit, Western Alps, Italy. *Tectonophysics*, **221**, 173–193.
MOLNAR, P. & ENGLAND, P. 1990. Temperatures, heat flux, and frictional stress near major thrust faults. *Journal of Geophysical Research*, **95**, 4833–4856.
MÖNICKE, A., BURCHARD, M., DUYSTER, J., MARESCH, W. V., RÖLLER, K. & STÖCKHERT, B. 2001. Experimental studies on fluid distribution and properties at ultra-high pressure metamorphic conditions: Implications for the flow strength of deeply subducted crust. *Beihefte zum European Journal of Mineralogy*, **13**, 126.
MOSENFELDER, J. L. & BOHLEN, S. R. 1997. Kinetics of the coesite to quartz transformation. *Earth and Planetary Science Letters*, **153**, 133–147.
ORZOL, J., STÖCKHERT, B. & RUMMEL, F. 2001. Experimental deformation of synthetic polycrystalline jadeite aggregates. *EOS*, **82**, F1145.
PARKINSON, C. D. 1996. The origin and significance of metamorphosed tectonic blocks in mélanges. *Terra Nova*, **8**, 312–323.
PATERSON, M. S. 1995. A theory for granular flow accomodated by material transfer via an intergranular fluid. *Tectonophysics*, **245**, 135–151.
PATERSON, M. S. 1987. Problems in the extrapolation of laboratory rheological data. *Tectonophysics*, **133**, 33–43.
PATERSON, M. S. 2001. A granular flow theory for the defomation of partially molten rock. *Tectonophysics*, **335**, 51–61.
PATERSON, M. S. & LUAN 1990. Quartzite rheology under geological conditions. *In*: KNIPE, R. J. & RUTTER, E. H. (eds) *Deformation Mechanisms, Rheology and Tectonics*. Geological Society, London, Special Publications, **54**, 299–307.
PEACOCK, S. M. 1993. Large-scale hydration of the lithosphere above subducting slabs. *Chemical Geology*, **108**, 49–59.
PEACOCK, S. M. 1996. Thermal and petrologic structure of subduction zones (overview). *In*: BEBOUT, G. E., SCHOLL, D. W., KIRBY, S. H. & PLATT, J. P. (eds) *Subduction Top to Bottom*. American Geophysical Union, Washington, 119–134.
PEACOCK, S. M. & HYNDMAN, R. D. 1999. Hydrous minerals in the mantle wedge and the maximum depth of subduction thrust earthquakes. *Geophysical Research Letters*, **26**, 2517–2520.
PERCHUK, A. & PHILIPPOT, P. 1997. Rapid cooling and exhumation of eclogitic rocks from the Great Caucasus, Russia. *Journal of Metamorphic Geology*, **15**, 299–310.
PERCHUK, A. L. & PHILIPPOT, P. 2000. Geospeedometry and time scales of high-pressure metamorphism. *International Geology Review*, **42**, 207–223.
PERCHUK, A., PHILIPPOT, P., ERDMER, P. & FIALIN, M. 1999. Rates of thermal equilibration at the onset of subduction deduced from diffusion modeling of eclogitic garnets, Yukon-Tanana terrane, Canada. *Geology*, **27**, 531–534.
PHILIPPOT, P. & VAN ROERMUND, H. L. M. 1992. Deformation processes in eclogitic rocks: evidence for the rheological delamination of the oceanic crust in deeper levels of subduction zones. *Journal of Structural Geology*, **14**, 1059–1077.
PIEPENBREIER, D. & STÖCKHERT, B. 2001. Plastic flow of omphacite in eclogites at temperatures below 500 °C – implications for interplate coupling in subduction zones. *International Journal of Earth Sciences*, **90**, 197–210.
PLATT, J. P. 1993. Exhumation of high-pressure rocks: a review of concepts and processes. *Terra Nova*, **5**, 119–133.
RAMSAY, J. G. & HUBER, M. I. 1987. *The Techniques of Modern Structural Geology Volume 2: Folds and Fractures*. Academic Press, London.
RANALLI, G. 1995. *Rheology of the Earth*. Chapman & Hall, London.
REINECKE, T. 1998. Prograde high- to ultrahigh-pressure metamorphism and exhumation of oceanic sediments at Lago di Cignana, Zermatt-Saas Zone, Western Alps. *Lithos*, **42**, 147–189.
RENNER, J., STÖCKHERT, B., ZERBIAN, A., RÖLLER, K. & RUMMEL, F. 2001. An experimental study into the rheology of synthetic polycrystalline coesite aggregates. *Journal of Geophysical Research*, **106**, 19411–19429.
ROSENBERG, C. L. 2001. Deformation of partially molten granite: a review and comparison of experimental and natural case studies. *International Journal of Earth Sciences*, **90**, 60–76.
ROYDEN L. H. 1993. The tectonic expression slab pull at continental convergent boundaries. *Tectonics*, **12**, 303–325.

RUBATTO, D. & HERMANN, J. 2001. Exhumation as fast as subduction. *Geology*, **29**, 3–6.
RUFF, L. J. & TICHELAAR, B. W. 1996. What controls the seismogenic plate interface in subduction zones? *In*: BEBOUT, G. E., SCHOLL, D. W., KIRBY, S. H. & PLATT, J. P. (eds) *Subduction Top to Bottom*. American Geophysical Union, Washington, 105–112.
RUTTER, E. H. 1983. Pressure solution in nature, theory and experiment. *Journal of the Geological Society London*, **140**, 725–740.
SCAMBELLURI, M. & PHILIPPOT, P. 2001. Deep fluids in subduction zones. *Lithos*, **55**, 213–227.
SCAMBELLURI, M., PENNACCHIONI, G., & PHILIPPOT, P. 1998. Salt-rich aqueous fluids formed during eclogitization of metabasites in the Alpine continental crust (Austroalpine Mt. Emilius unit, Italian western Alps). *Lithos*, **43**, 151–167.
SCHERTL, H.-P., SCHREYER, W. & CHOPIN, C. 1991. The pyrope-coesite rocks and their country rocks at Parigi, Dora Maira massif, western Alps: detailed petrography, mineral chemistry, and P-T path. *Contributions to Mineralogy and Petrology*, **108**, 1–21.
SCHOLZ, C. H. 1990. *The Mechanics of Earthquakes and Faulting*. Cambridge University Press, Cambridge.
SCHREYER, W. 1995. Ultradeep metamorphic rocks: the retrospective viewpoint. *Journal of Geophysical Research*, **100**, 8353–8366.
SCHWARZ, S. & STÖCKHERT, B. 1996. Pressure solution in siliciclastic HP-LT metamorphic rocks – constraints on the state of stress in deep levels of accretionary complexes. *Tectonophysics*, **255**, 203–209.
SHEN, A. H. & KEPPLER, H. 1997. Direct observation of complete miscibility in the albite-H_2O system. *Nature*, **385**, 710–712.
SHREVE, R. L. & CLOOS, M. 1986. Dynamics of sediment subduction, melange formation, and prism accretion. *Journal of Geophysical Research*, **91**, 10229–10245.
STACEY, F. D. 1992. *Physics of the Earth* (third edition). Brookfield Press, Kenmore.
STÖCKHERT, B. & RENNER, J. 1998. Rheology of crustal rocks at ultrahigh pressure – when continents collide. *In*: HACKER, B. R. & LIOU, J. G. (eds) *Geodynamics and Geochemistry of Ultrahigh-Pressure Rocks*. Kluwer, Amsterdam, 57–95.
STÖCKHERT, B., BRIX, M. R., KLEINSCHRODT, R., HURFORD, A. J. & WIRTH, R. 1999. Thermochronometry and microstructures of quartz – a comparison with experimental flow laws and predictions on the temperature of the brittle–plastic transition. *Journal of Structural Geology*, **21**, 351–369.
STÖCKHERT, B., DUYSTER, J., TREPMANN, C. & MASSONNE, H.-J. 2001. Microdiamond daughter crystals precipitated from supercritical COH + silicate fluids included in garnet, Erzgebirge, Germany. *Geology*, **29**, 391–394.
STÖCKHERT, B., MASSONNE, H. & NOWLAN, E. U. 1997. Low differential stress during high-pressure metamorphism: the microstructural record of a metapelite from the Eclogite Zone, Tauern Window, Eastern Alps. *Lithos*, **41**, 103–118.
STÖCKHERT, B., WACHMANN, M., KÜSTER, M. & BIMMERMANN, S. 1999. Low effective viscosity during high pressure metamorphism due to dissolution precipitation creep: the record of HP-LT metamorphic carbonates and siliciclastic rocks from Crete. *Tectonophysics*, **303**, 299–319.
TAYLOR, M. A. J., ZHENG, G., RICE, J. R., STUART, W. D. & DMOWSKA, R. 1996. Cyclic stressing and seismicity at strongly coupled subduction zone. *Journal of Geophysical Research*, **101**, 8363–8381.
TERRY, M. P., ROBINSON, P., HAMILTON, M. A. & JERCINOVIC, M. J. 2000*b*. Monazite geochronology of UHP and HP metamorphism, deformation, and exhumation, Nordoyane, Western Gneiss Region, Norway. *American Mineralogist*, **85**, 1651–1664.
TERRY, M. P., ROBINSON, P. & KROGH RAVNA, E. J. 2000*a*. Kyanite eclogite thermobarometry and evidence for thrusting of UHP over HP metamorphic rocks, Nordoyane, Western Gneiss Region, Norway. *American Mineralogist*, **85**, 1637–1650.
TICHELAAR, B. W. & RUFF, L. J. 1993. Depth of seismic coupling along subduction zone. *Journal of Geophysical Research*, **98**, 2017–2037.
VAN DER KLAUW, S. N, REINECKE, T. & STÖCKHERT, B. 1997. Exhumation of ultrahigh-pressure metamorphic oceanic crust from Lago di Cignana, Piemontese zone, western Alps: the structural record in metabasites. *Lithos*, **41**, 79–102.
VON HERZEN, R., RUPPEL, C., MOLNAR, P., NETTLES, M., NAGIHARA, S. & EKSTRÖM, G. 2001. A constraint on the shear stress at the Pacific-Australian plate boundary from heat flow and seismicity at the Kermadec forearc. *Journal of Geophysical Research*, **106**, 6817–6833.
VON HUENE, R. & SCHOLL, D. W. 1991. Observations at convergent margins concerning sediment sub duction, subduction erosion, and the growth of continental crust. *Reviews of Geophysics*, **29**, 279–316.
WALLIS, S., ENAMI, M. & BANNO, S. 1999. The Sulu UHP Terrane: A Review of the Petrology and Structural Geology. *International Geology Review*, **41**, 906–920.
WALLIS, S. R., ISHIWATARI, A., *ET AL.* 1997. Occurrence and field relationships of ultrahigh-pressure metagranitoid and coesite eclogite in Su-Lu terrane, eastern China. *Journal of the Geological Society*, **154**, 45–54.
WANG, K. 2000. Stress-strain 'paradox', plate coupling, and forearc seismicity at the Cascadia and Nankai subduction zones. *Tectonophysics*, **319**, 321–338.
WANG, K. L. & HE, J. 1999. Mechanics of low-stress forearcs: Nankai and Cascadia. *Journal of Geophysical Research*, **104**, 15191–15205.
WHEELER, J. 1991. Structural evolution of a subducted continental sliver: the northern Dora Maira massif, Italian Alps. *Journal of the Geological Society*, **148**, 1101–1113.
WHEELER, J. 1992. Importance of pressure solution and coble creep in the deformation of polymineralic rock. *Journal of Geophysical Research*, **97**, 4579–4586.

YAMAMOTO, H., ISHIKAWA, M., ANMA, R. & KANKEO, Y. 2000. Kinematic analysis of ultrahigh-pressure–high-pressure metamorphic rocks in the Chaglinka-Kulet area of the Kokchetav Massif, Kazakhstan. *The Island Arc*, **9**, 304–316.

YE, K., CONG, B. & YE, D. 2000*a*. The possible subduction of continental material to depths greater than 200 km. *Nature*, **407**, 734–736.

YE, K., YAO, Y., KATYAMA, I., CONG, B., WANG, Q. & MARUYAMA, S. 2000*b*. Large areal extent of ultrahigh-pressure metamorphism in the Sulu ultrahigh-pressure terrane of East China: new implications from coesite and omphacite inclusions in zircon of granitic gneiss. *Lithos*, **52**, 157–164.

ZHANG, R. Y. & LIOU, J. G. 1997. Partial transformation of gabbro to coesite-bearing eclogite from Yangkou, the Sulu terrane, eastern China. *Journal of Metamorphic Geology*, **15**, 183–202.