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Interfacial friction-induced pressure and implications for the formation and preservation of intergranular coesite in metamorphic rocks

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The finding of intergranular coesite and coesite pseudomorphs has been taken as critical evidence to support that the formation of coesite was related to regional ultrahigh pressure (UHP) metamorphism rather than the pressure-vessel effect. Previous laboratory deformation experiments found that coesite occurs in practically undeformed strain-forbidden zones immediately adjacent to the piston–sample interfaces while the mean stress applied to the specimen is remarkably lower than the quartz–coesite equilibrium boundary. The formation and preservation of intergranular coesite in eclogites from UHP metamorphic terranes and the occurrence of coesite in strain-forbidden zones within experimentally deformed rocks can be more satisfactorily explained by the theory of interfacial friction-induced pressure, which is a well-known in metal forging. During certain episodes of fast tectonic deformation under high transient differential stresses, the interfacial friction can induce very significant deformation pressures in thin layers of weak materials (e.g., SiO_2) between large garnet crystals that are refractory and mechanically strong, thus act as excellent anvils. The local deformation pressure, which deviates from the lithostatic value, is likely an intrinsic property of highly constrained flows of weak materials between strong walls at least on a microstructural scale.

1. Introduction

Ultrahigh pressure (UHP) is defined as a type of metamorphism that occurs at very high lithostatic pressures within the eclogite facies but above the stability field of quartz (e.g., Liou et al., 1998; Carswell and Compagnoni, 2003). The quartz–coesite transition boundary is thus regarded as the lower pressure limit of UHP metamorphism. It is an accepted tenet of geology that the occurrence of coesite relics or pseudomorphs after coesite (former coesite) in eclogite, paragneiss and marble is a critical indicator for UHP metamorphism of supracrustal rocks that were subducted and exhumed from depths greater than ~100 km during continental–continental collision (e.g., Chopin, 1984, 2003; Schreyer, 1995; O'Brien and Ziemann, 2008). The relationship between lithostatic pressure (P_L) and depth (z) is inferred from Pascal's law: $P_L = \rho gz$, where ρ is the average density of the overlying rocks from surface to depth z, and g is the acceleration of gravity. However, laboratory deformation experiments (e.g., Hobbs, 1968; Green, 1972; Hirth and Tullis, 1994), which showed that coesite could form under differential stress conditions below its stability field determined under hydrostatic pressures (e.g., Kitahara and Kennedy, 1964), casts doubt on this paradigm with important implications for understanding the mechanics of naturally deforming rocks (Fig. 1). In the geodynamic Earth, solid polymineralic rocks do not act as an incompressible fluid for which Pascal's law is valid (e.g., Mancktelow, 1993, 1995, 2008), and deform under nonhydrostatic stresses (i.e., $\sigma_1 \geq \sigma_2 \geq \sigma_3$, where σ_1, σ_2 and σ_3 are the maximum, intermediate and minimum stresses, respectively) on either a regional scale or a microstructural scale (e.g., Turcotte and Schubert, 1982; Stöckhert, 2002). The finding of pressure-induced incipient amorphization of α-quartz and transition to coesite in an eclogite from Antarctica (Palmeri et al., 2009) and eclogite-facies pseudodotachylytes in the Bergen arcs of western Norway (Austrheim and Boundy, 1994) suggests that the state of stresses at some sites during certain episodes in continental collision zones can extremely deviate from the hydrostatic state ($\sigma_1 = \sigma_2 = \sigma_3$).

Hirth and Tullis (1994) experimentally deformed cylindrical specimens (d_0 = 4.8–6.3 mm, h_0 = 10 − 14 mm) of Heavitree quartzite using a modified Griggs-type apparatus. Coesite occurs mainly in two cone-shaped domains immediately adjacent to the pistons, where practically no crystal plastic deformation occurs

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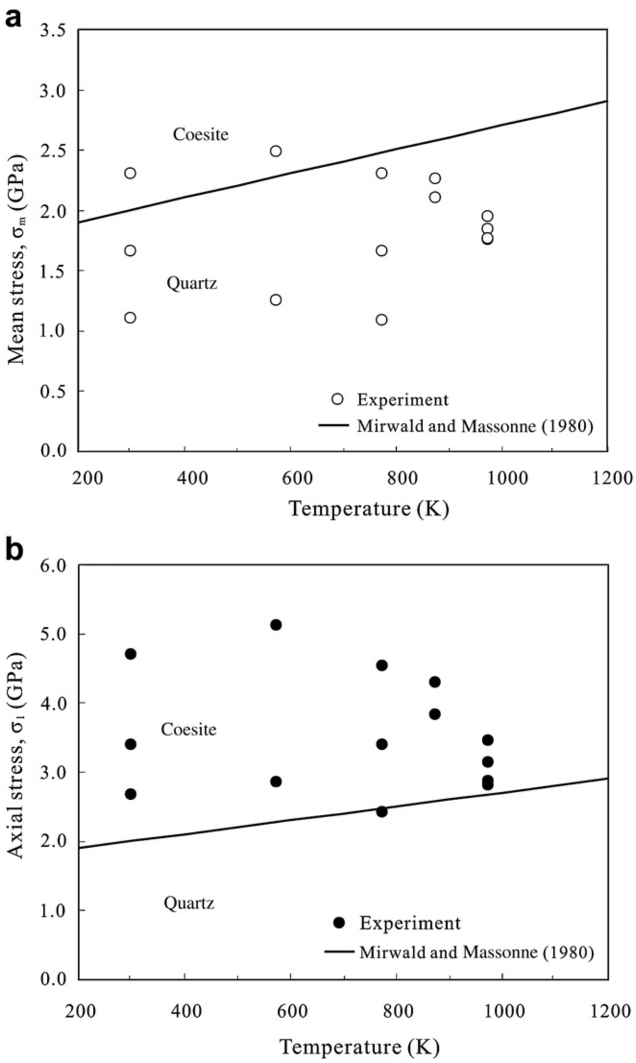

Fig. 1. Plots of the mean stress (a) and the axial stress σ_1 (b) versus temperature for experimentally deformed samples in which coesite was observed in strain-forbidden zones (Hirth and Tullis, 1994). The $\alpha -$ quartz $-$coesite phase boundary from Mirwald and Massonne (1980). The differential stress on each specimen of Green (1972) and Hobbs (1968) was not known because the pistons punched into the specimen and the resolution in stress measurement was poor with large uncertainty in the solid-medium Griggs-type deformation apparatus of 1960–1970's.

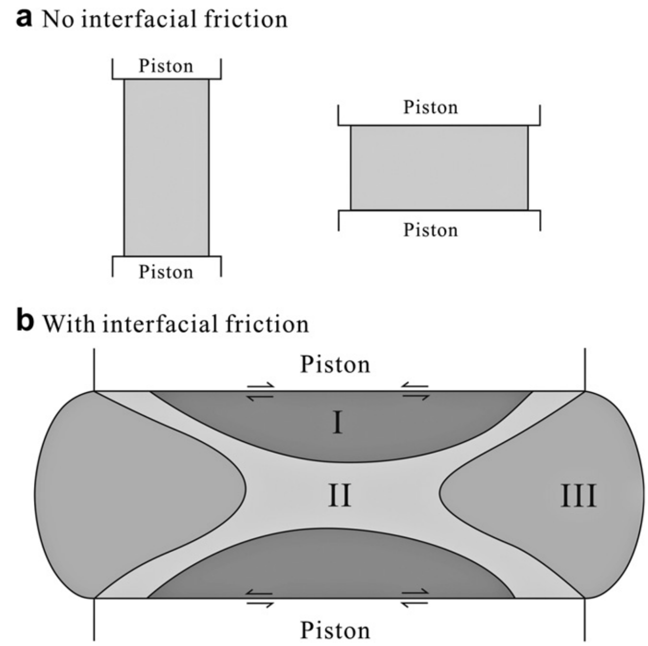

Fig. 2. Schematic illustration of homogeneous (a: without interfacial friction) and inhomogeneous (b: with interfacial friction) strain distributions in a cylindrical rock sample deformed between two rigid anvils. A heterogeneously deformed sample can be divided into three zones: "strain-forbidden zone" (I), intensive strain zone (II) and intrusion zone (III).

(Fig. 2b). These poorly deformed domains can be called the strain-forbidden zones or strain shadows. Occasionally coesite grains are also found along grain boundaries perpendicular to σ_1 in the areas away from the piston–sample interfaces or some fault zones where stresses were concentrated. Coesite makes up ~5% of the material in the strain-forbidden zones, but <1% of the overall sample. In the diagram of mean stress $\sigma_m = (\sigma_1+\sigma_2+\sigma_3)/3$ versus temperature (Fig. 1a), most of the deformed samples containing of coesite are within the stability field of $\alpha -$ quartz. In the diagram of axial stress σ_1 versus temperature (Fig. 1b), however, these coesite-bearing samples all lie within the stability field of coesite. Hirth and Tullis (1994) proposed a hypothesis that the magnitude and orientation of the maximum principal stress (σ_1) rather than the confining pressure ($\sigma_2 = \sigma_3$) or the mean stress play a crucial role in controlling the quartz–coesite transition.

Green (1972) experimentally deformed Dover flint at temperatures of 450–900 °C and at confining pressures of 1.0–2.0 GPa using a solid-medium Griggs-type apparatus and found that coesite occurred only in the region between pistons (where two strain-forbidden zones are overlapped) rather than in the extrusion zone (Domain 3, Fig. 2b). Coesite occurred only in the samples shortened >50% at a strain rate of 10^{-4} s^{-1} while all other samples shortened <50% at strain ratios of 10^{-7}–10^{-5} s^{-1} were free of coesite. Coesite appeared in a sample (C460) shortened to 60% at 750 °C and a confining pressure of only 2.0 GPa, which is much lower than the quartz–coesite boundary (2.87 GPa at 750 °C) determined hydrostatically by Kitahara and Kennedy (1964). The explanation offered by Green (1972) is that the presence of high dislocation density displaced the quartz–coesite transition boundary to lower pressures. The viability of the interpretation has been ruled out by the experimental observation that the dislocation density has no effect on the position of the equilibrium phase boundary between quartz and coesite (Ingrin and Liebermann, 1989).

Hobbs (1968) reported that coesite occurred in single crystals of quartz shortened to 50% at a confining pressure of 1.5 GPa and the maximum differential stress ($\sigma_{1-}\sigma_3$) of ~0.9 GPa at 900 °C. The mean stress is 1.8 GPa which is 1.2 GPa below the quartz–coesite boundary at the same temperature as determined hydrostatically by Kitahara and Kennedy (1964). Hobbs (1968) postulated that the stored strain energy is responsible for the nucleation and growth of coesite outside of the hydrostatic stability field.

A well-known metal-forging theory on interfacial friction-induced pressure (e.g., Unksov, 1961; Harris, 1983; Dieter, 1986) has been applied to various geological interpretations. Jamieson (1963) used the theory to emphasize that the pressure, which is attained as the result of confinement within a stronger pressure vessel with a geometrical arrangement intensifying the pressure, can be 3–9 times the normal compressive strength of the vessel-constructed material. Mancktelow (1993, 1995, 2008) used the concept first to explain the development of isolated eclogites in metamorphic terrains and then to deal with the tectonic overpressure within a flow channel between two rigid walls on a regional scale. Ji et al. (2000) applied the theory to explain why the flow strength of quartz–plagioclase layered composites increases remarkably with decreasing the thickness of the layers, and the thin-layered composites are significantly stronger than particulate counterparts with the same composition.

Here we use the concept of interfacial friction-induced pressure (e.g., Unksov, 1961; Harris, 1983; Dieter, 1986) to provide an alternative interpretation for the formation of coesite in the samples deformed experimentally at differential stress conditions below its stability field under hydrostatic pressures (e.g., Hobbs, 1968; Green, 1972; Hirth and Tullis, 1994). Because the effect of interfacial friction-induced pressure is a corollary of highly constrained flow of weak materials between strong walls, we will also examine the potential implications of the model for the formation and preservation of intergranular coesite between mechanically strong minerals in UHP metamorphic rocks.

2. Interfacial friction-induced pressure

If there is no friction between a cylindrical specimen and anvils, this specimen can be shortened homogeneously and its cylindrical shape can be retained in the regime of plastic deformation (Fig. 2a). The flow strength of the sample material (σ_0) is equal to the applied differential stress: $\sigma_0 = \sigma_1 - \sigma_3$. However, there is almost always friction between the specimen and anvils. The frictional forces oppose the lateral flow of the material, which increases the diameter of the sample during compressive deformation. Shear stresses caused by the friction forces at the contact planes between sample and anvils are distributed unevenly along the height of the specimen: maximum at the ends and zero halfway along the height of the specimen. The inner material of the sample is supported by the exterior, producing an extra deformation pressure in the specimen (e.g., Dieter, 1986). As a result, an initially cylindrical sample becomes barrel shaped after compressive deformation because the material at specimen mid-height can flow out undisturbed while the ends of the specimen near the contact with the anvils are practically undeformed (Fig. 2b). Thus, a deformed specimen generally displays three domains with distinct strain characteristics (Fig. 2b): Domain I, adjacent to the ends, which is a cone-shaped strain-forbidden zone where practically no crystal plastic deformation occurs. Domain II is a most intensely deformed zone where the material flows from the center to the periphery. Domain III is an intrusion zone where the deformation is intermediate between Domains I and II (Unksov, 1961; Dieter, 1986; Ji et al., 2000).

The heterogeneous deformation of the specimen (Fig. 2b), resulted from the interfacial friction, requires a higher axial stress (σ_1^*) and greater total energy expenditure than for a homogeneous deformation without friction (Fig. 1a). $\sigma_1^* > \sigma_1$, where σ_1 is the axial stress to deform the same material if no friction occurred between the specimen and anvils. With increasing the diameter-to-thickness ratio (d/h, Fig. 3), two strain-forbidden zones approach and overlap, causing a significant increase in σ_1^* for a given incremental deformation. For a fixed diameter, a shorter specimen requires a greater differential stress ($\sigma_1^* - \sigma_3$) to produce the same quantity of compressive strain (Hawkes and Mellor, 1970; Goretta and Routbort, 1987; Odom and Adams, 1994) because of relatively larger volume fraction of the strain-forbidden zones. A sheet specimen with an extremely large d/h value becomes almost undeformable under axial compression. In uniaxial creep tests of minerals, for example, a thin film of platinum is often placed between sample and alumina pistons to avoid potential chemical reaction between them. It is generally found that the platinum films are not deformed after the test, indicating that the interfacial friction inhibits efficiently the lateral flow of the platinum (Ji et al., 2000). Thus, it appears risky to compare the flow strengths of samples without taking into consideration their aspect-ratios (d/h) at which the flow strengths are measured. The true flow strength in compression without friction (σ_0) can be indirectly obtained by plotting ($\sigma_1^* - \sigma_3$) versus d/h for several different sets of data and extrapolating the curve to $d/h = 0$ (Dieter, 1986).

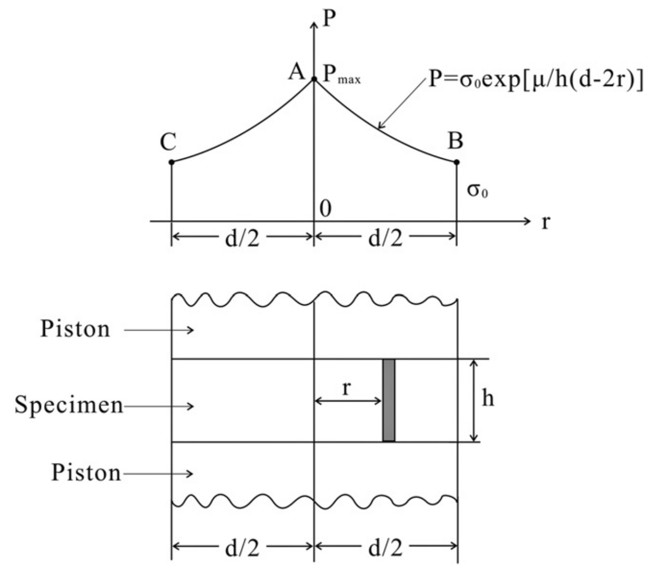

Fig. 3. Schematic illustration of the deformation pressure (P) varying with position along an interface of a cylindrical specimen (diameter d and height h) that is compressed between a pair of parallel anvils. If small pressure-measuring devices were placed at intervals on the interface during the axial compression, the indications on the pressure-gauge dials would show the variations as in this figure. Coulomb friction is assumed.

The frictional shear stresses lead to an extra deformation pressure (p) in the material immediately adjacent the rigid pistons (Dieter, 1986), which increases smoothly from the equivalent of the true flow stress of the material in compression without friction (σ_0) at the edges to a peak value (P_{max}) at the center, as illustrated by experimental measurements (Unksov, 1961; Jamieson, 1963; Wu and Bassett, 1993). A mathematical expression for the variation of deformation pressure with position at the interface was given in Unksov (1961):

$$p = \sigma_0 \exp\left[\frac{\mu}{h}(d - 2r)\right] \quad (1)$$

where μ is the friction coefficient of the interface, and r is the distance from a given point in the disk to the center of the disk (Fig. 3). This equation, which is valid whatever confining pressure is applied (e.g., 0.1 MPa for metal forging at room pressure, and several hundreds to several thousands of MPa for laboratory rock deformation experiments), has been used in Jamieson (1963), Mancktelow (1993, 1995, 2008) and Ji et al. (2000) for analyzing the possible occurrence of overpressure in geological processes on various scales. As shown in Fig. 3, the deformation pressure is distributed symmetrically about the axial line and has a maximum value (P_{max}) equal to $\sigma_0\exp(\mu d/h)$ at the center of the disk and a minimum value equal to σ_0 at the edges of the disk. The results for a series of specimens with a fixed diameter ($d = 10$ mm) and various d/h ratios (Fig. 4) were calculated assuming $\mu = 0.45$, 0.60 and 0.75. For most rocks at confining pressures above and below 200 MPa, $\mu = 0.60$ and 0.85, respectively (e.g., Byerlee, 1978). The P/σ_0 value increases with increasing the d/h ratio and μ.

The average deformation pressure ($\overline{p}$) of the disk to compression (i.e., the mean height of the friction hill) is given by Dieter (1986) as

$$\overline{p} = 2\sigma_0\left(\mu\frac{d}{h}\right)^{-2}\left[\exp\left(\mu\frac{d}{h}\right) - \mu\frac{d}{h} - 1\right] \quad (2)$$

For a given value of d/h ratio, the $\overline{p}/\sigma_0$ value increases remarkably with μ. For a given μ, increasing the d/h value easily leads to a significant increase in the average pressure.

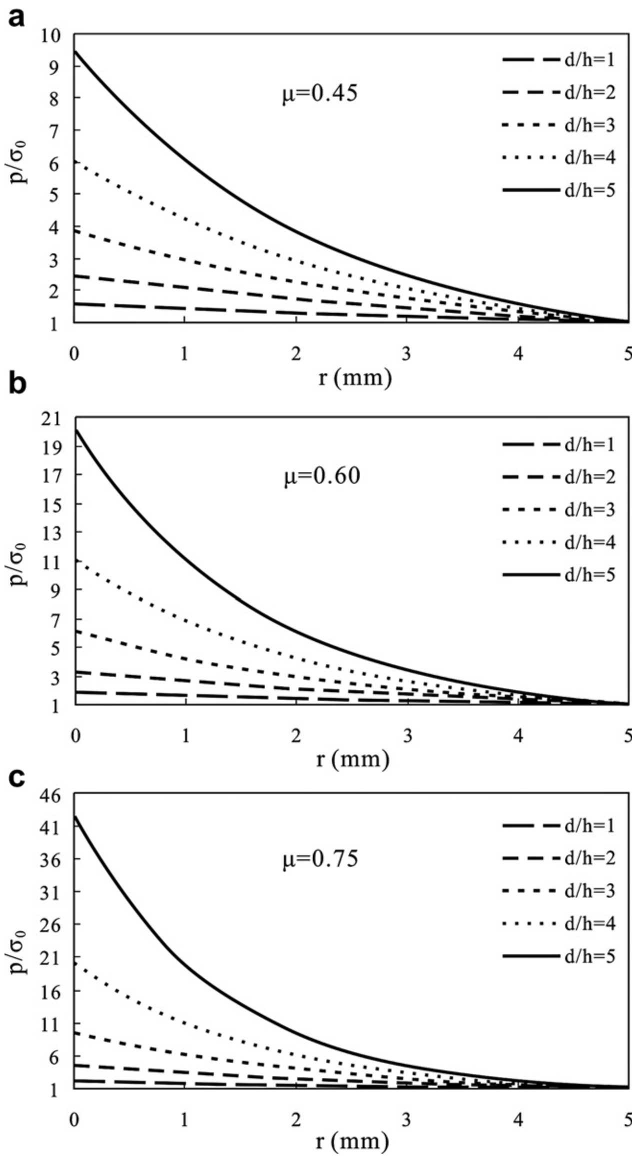

Fig. 4. Deformation pressure curves in terms of the sample aspect-ratio (d/h) and friction coefficient (μ=0.45, 0.60 and 0.75 for a, b, and c, respectively). The sample diameter $d = 10$ mm.

Equations 1 and 2 are based on an assumption that the shear stresses are proportional to the normal stresses over the whole of the interface (i.e., the Coulomb friction). In practice, the interfacial shear stress cannot exceed the shear strength of the interface. If the shear strength of the interface is constant and equal to $m\tau_0$, where τ_0 is the shear strength of the sample, and m is the interface sliding factor, then according to Dieter (1986),

$$p = \sigma_0 + \frac{m\tau_0(d-2r)}{h} \tag{3}$$

where m is defined as

$$m = \frac{d-d_s}{d} \tag{4}$$

and d_s is the total length of the sliding interface,

$$d_s = \frac{h}{\mu}\ln\frac{1}{2\mu} \tag{5}$$

when $\mu < 0.50$, interface sliding takes place over only a length of $d_s/2$ from each edge of the disk while no sliding occurs in the central part of the interface. As long as $\mu \geq 0.50$, $d_s = 0$, interface sliding is impossible ($m = 1$) and thus perfect sticking friction will occur. This condition is particularly suited for the rocks in the deep crust and upper mantle. Then, the maximum deformation pressure under the condition of sticking friction:

$$p_{\max} = \sigma_0 + \frac{d}{h}\tau_0 \tag{6}$$

To the first approximation, the shear strength of the sample (τ_0) can be estimated from the following relation based on the von Mises criterion (e.g., Dieter, 1986; Jaeger et al., 2008):

$$\tau_0 = \sigma_0/\sqrt{3} \tag{7}$$

The von Mises criterion is part of a plasticity theory that applies best to ductile materials such as metals, ceramics and rocks that are power-law flow materials and elastic prior to yield. The von Mises criterion implies that the onset of yield for the ductile materials does not depend on the hydrostatic component of the stress tensor. There are the following differences between the von Mises and Tresca (or the maximum-shear-stress) criterions: (1) The Tresca criterion does not take into consideration the intermediate principal stress and suffers from the major difficulty that it is necessary to know in advance which are the maximum and minimum principal stresses; and (2) The von Mises criterion is less conservative than the Tresca criterion because that the Tresca yield surface is circumscribed by von Mises one. Therefore, the Tresca criterion predicts plastic yield already for stress states that are still elastic according to the von Mises criterion. The principle of interfacial friction-induced pressure described above is always valid whenever the von Mises criterion or the Tresca criterion is used. Equations (1)–(6) have been proved by experimental measurements (Unksov, 1961).

Combining Equations (6) and (7), the following equation can be obtained:

$$p_{\max} = \sigma_0\left(1 + \frac{\sqrt{3}}{3}\frac{d}{h}\right) = [1 + 0.577(d/h)]\sigma_0 \tag{8}$$

Clearly, the friction-induced pressure in the strain-forbidden zones immediately near the rigid walls becomes particularly important at large values of d/h. Furthermore, with increasing compressive strain, the specimen is progressively shortened, d/h increases, and the strain-forbidden zones approach and finally overlap, causing an increase in $P_{\max.}$ The maximum deformation pressure can be easily 5–6 times higher than σ_0 as long as the d/h value is sufficiently large. The above principle about the interfacial friction-induced pressure, which describes an intrinsic property of highly constrained flow of weak materials between strong walls, is well-known in mechanical metallurgy.

3. Discussion

3.1. Interpretation of the experimental data

The experimental results that coesite formed under differential stress conditions below its stability field under hydrostatic pressures (e.g., Hobbs, 1968; Green, 1972; Hirth and Tullis, 1994) can be more satisfactorily interpreted according to the theory of interfacial friction-induced pressure.

The first point concerns why coesite occurs only in the region between pistons (where two strain-forbidden zones are overlapped in highly shortened samples) rather than in the extrusion zone (Domain 3 in Fig. 2). In the strain-forbidden zones, the local differential stress is lower than the yield strength of the material, and therefore little crystal plastic deformation occurs in the zones. In the material directly adjacent to the pistons, the average deformation pressure induced by interfacial friction equals to σ_1 while the maximum deformation pressure in the central part even exceeds σ_1 in magnitude. Hence coesite nucleates and crystallizes in the strain-forbidden zones with the local deformation pressures $\geq\sigma_1$.

The second point is why coesite occurs only in the samples shortened >50% at a strain rate of 10^{-4} s^{-1} while all other samples shortened <50% at strain ratios of 10^{-7}–10^{-5} s^{-1} were free of coesite. For example, coesite appeared in a sample (C460) shortened to 60% ($d/h = 3.7$) at 750 °C and a confining pressure of only 2.0 GPa (Green, 1972). As shown by Equations (8) and Fig. 4, the maximum deformation pressure at the interface between the sample and the anvil increases with either d/h or σ_0. The σ_0 value of a sample deformed at a strain rate of 10^{-4} s^{-1} is certainly much higher than that at 10^{-7}–10^{-5} s^{-1}. For a given σ_0, the d/h ratio should be above certain critical value so that the maximum deformation pressure in the strain-forbidden zones becomes higher than the critical pressure for the quartz–coesite transition.

3.2. Potential implications for the formation of intergranular coesite

Large-scale tectonic overpressure (ΔP) has been invoked to be responsible for HP/UHP metamorphism (e.g., Rutland, 1965; Brace et al., 1970; Mancktelow, 1993, 1995; 2008; Petrini and Podladchikov, 2000). $\Delta P = \sigma_m - P_L$, where σ_m is the mean stress equal to $(\sigma_1 + \sigma_2 + \sigma_3)/3$, and P_L is the lithostatic pressure $\Delta P > 0$ in a compressive regime as the regional σ_1 is horizontal while $\Delta P < 0$ in a extensional regime as the regional σ_1 is vertical. The maximum tectonic overpressure (ΔP_{max}) occurs when $\sigma_1 = \sigma_2 > \sigma_3$ in a compressive regime: $\Delta P_{max} = 2\sigma_d/3$ (e.g., Brace et al., 1970), where the differential stress or flow strength $\sigma_d = \sigma_1 - \sigma_3$. Clearly, the maximum tectonic overpressure is limited by the flow strength of dominant rocks, which exceeds hardly 0.2 GPa at natural strain rates (10^{-13}–10^{-15} s^{-1}) on geological time scales. For this reason, some authors (e.g., Brace and Kohlstedt, 1980; Burov et al., 2001; Renner et al., 2001; Green, 2005) argued the effect of large-scale tectonic overpressure on the quartz–coesite transition and UHP metamorphism is marginal in regional scale. Recently the 2D numerical modeling conducted by Li et al. (2010) suggested that the magnitude of regional tectonic overpressure in the bottom corner of a wedge-like confined flow channel can be ~0.3 GPa, which is about 10% of the lithostatic pressure.

However, whether coesite is formed by a small-scale local overpressure or a regional lithostatic pressure is still inconclusive. As coesite or coesite pseudomorph occurs predominantly as inclusions in refractory, impermeable and mechanically strong minerals such as garnet and zircon, consideration of the pressure-vessel effect (e.g., Gillet et al., 1984; van der Molen and van Roermund, 1986; Nishiyama, 1998; Zhang, 1998; Perrillat et al., 2003) leads to a ratiocination that the inclusion pressures can be tremendously higher than the lithostatic pressure on the host as long as the latter has not relaxed by plastic deformation or not fractured to allow the relaxation of the inclusion pressures. Raman spectroscopy and synchrotron XRD data indicated that present-day, internal pressures on monocrystalline coesite and polycrystalline quartz inclusions within unfractured host minerals such as garnet, zircon and diamond are still as high as 1.9–2.3 GPa (Parkinson and Katayama, 1999; Parkinson, 2000), 2.4 GPa (Ye et al., 2001) and 3.6 GPa (Sobolev et al., 2000). The pressures at room temperature correspond to those close to or above the quartz–coesite transformation boundary, indicating that garnet, zircon and diamond are indeed excellent pressure vessels in which plastic relaxation is impossible at temperatures lower than ~800 °C (e.g., Mosenfelder et al., 2005; Wang and Ji, 1999; Li et al., 2006).

A consensus has been reached about that the overpressure, which is the difference between the inclusion pressure and the lithostatic pressure on the host, is produced by thermomechanical (e.g., thermal expansion, compressibility, and elastic anisotropy) mismatch between the inclusion and the host during changes in temperature and/or pressure. However, conflicting results have suggested different depths at which coesite inclusions formed. The elastic models of Gillet et al. (1984) and van der Molen and van Roermund (1986) suggested that "coesite included in garnet cannot have formed during progressive metamorphism from quartz at lithostatic pressures below the coesite stability" because "the negative volume change during the transformation of quartz into coesite prevents growth from a quartz inclusion in the elastic garnet, and the differential elasticity of quartz and garnet cannot alone produce the internal pressure required to stabilize coesite". These authors envisaged that "garnet grew around coesite and not around quartz". Implicit in these models is the assumption that the coesite inclusions were captured by their host minerals under lithostatic UHP conditions and the role of inclusion overpressure is only to inhibit the volume increase necessary for coesite to transform to quartz and to allow the preservation of coesite during the fast exhumation of UHP metamorphic rocks. A fundamental question is whether the local overpressure within a quartz inclusion captured by garnet and zircon below the coesite stability field, produced by the differential expansion between the inclusion and the host, is high enough to allow the transformation of quartz to coesite during any P–T evolution. As the inclusions are much smaller than their host, the negative volume change during the transformation of quartz to coesite may be readily accommodated by the elastic and/or plastic deformation of the host. The thermomechanical modeling of Wu and Chi (2003) suggested that quartz enclosed in garnet could be transformed to coesite at a depth of about 60 km, corresponding to a pressure far below the stability field of coesite under hydrostatic conditions. The authors proposed that the occurrence of coesite inclusions in mechanically strong minerals such as garnet and zircon, which resulted from high overpressures in the inclusions, is not a reliable indicator for regional UHP metamorphism of supracrustal rocks or deep continental subduction (>100 km). All the elastic models described above have not been tested by experiments yet and are still a matter of debate.

The finding of intergranular coesite in eclogitic mylonites from the Sulu metamorphic belt, China (e.g., Zhang and Liou, 1996, 1997; Ye et al., 1996; Zhao et al., 2003) has been regarded as the most convincing evidence to support that coesite was formed by the regional UHP metamorphism rather than the local overpressure induced by the pressure-vessel effect. The eclogitic mylonites are highly deformed with a strong planar fabric defined by a compositional banding with alternating garnet- and omphacite-rich layers (e.g., Ji et al., 2003). Most of the intergranular coesite and coesite pseudomorphs are found along the interface between garnet and garnet or between garnet and garnet-dominated polymineralic aggregate in the garnet-rich layers. Later-formed ($T < 350$ °C) radial extensional fractures from the relict intergranular coesite rimmed with palisade or mosaic quartz polycrystalline aggregate occur in both garnet and omphacite. The tensile fractures, which allowed the relaxation of the pressure that the surrounding strong phases applied on the intergranular coesite and the infiltration of fluids, and thus the coesite to quartz transition, could be initiated by the differential expansion of the coesite inclusion and host garnet. The eclogitic mylonites constitute ductile shear zones wrapping around weakly deformed, lens-shaped boudins consisting of partially eclogitized metagabbro. The internal parts of the boudins, which were extremely dry and inaccessible to fluid, are thus less deformed and only partially eclogitized (e.g., Zhang and Liou, 1997).

Garnet is also an excellent anvil material because it is mechanically strong (e.g., Wang and Ji, 1999; Li et al., 2006), and extremely stable over a wide P–T interval (e.g., Godard and van Roermund, 1995). During tectonic deformation, the interfacial friction could induce significantly strong deformation pressures in the thin layers of quartz between garnet and garnet or between garnet and garnet-dominant polymineralic aggregate in the eclogites. The interfacial deformation pressure can be several times higher than the flow stress ($\sigma_0 = \sigma_{1-}\sigma_3$) of the bulk rock, depending on the relative length with respect to the thickness (d/h) of the thin layer. Under the condition of sticking friction, which is particularly suited for the rocks in the deep crust and upper mantle, the maximum deformation pressure can be approximately determined according to $P_{\max} = [1 + 0.577(d/h)]\sigma_0$. For $\sigma_0 = 0.1$ GPa (e.g., Molnar and England, 1990), $P_{\max} = 0.27$ GPa if d/h = 3, $P_{\max} = 0.39$ GPa if d/h = 5, and $P_{\max} = 0.677$ GPa if d/h = 10. In the case of $P_{\max} = 0.677$ GPa, intergranular coesite may occur along the garnet–garnet interfaces at a lithostatic pressure of 2.1 GPa, equivalent to a depth of ~70 km in a continental collision mountain root. If this inference withstands further scrutiny, the interfacial friction-induced pressure would be an alternative interpretation for the formation of intergranular coesite in eclogitic mylonites.

Petrological, geochemical and geochronological studies of UHP metamorphic rocks revealed that subduction and exhumation of continental slabs are “fast-in” and “fast-out” tectonic processes taking place in cold subduction zones with a temperature increase by 5–10° C/km from the Earth’s surface (e.g., Schreyer, 1995; Liou et al., 1998). A fast strain rate is a necessity because the tectonic deformation is associated with the long distance transport of UHP rocks en route to the mantle and back to the surface. The effect of interfacial friction-induced pressure certainly becomes more pronounced at faster strain rates and lower temperatures because the differential stress (σ_0) increases with increasing strain rate and decreasing temperature. The local differential stress in HP/UHP metamorphic rocks, estimated from microstructures (e.g., jadeite’s deformation twins, and recrystallized grain size) is on the order of some hundreds of MPa (e.g., Lenze et al., 2005).

Furthermore, experiments of Hirth and Tullis (1994) on quartz–coesite transition together with those of Vaughan et al. (1984) on olivine–spinel transformation suggest that under nonhydrostatic stresses, stability of a phase does not depends on the mean stress (σ_m) but the local normal stress across a special interface. In the experimentally deformed quartzite (Hirth and Tullis, 1994), nucleation and growth of coesite occur in the grain boundaries perpendicular to σ_1 and particularly in strain-forbidden zones where the interfacial friction-induced pressures locally equal or exceed σ_1. In nature, it should be also true that nucleation and growth of coesite occur in the grain boundaries perpendicular to σ_1 and in the regions immediately adjacent to the interfaces with rigid or strong minerals such as garnet.

Unlike the coesite inclusions within rigid minerals such as garnet and zircon, intergranular coesite grains are extremely rare in the UHP rocks exhumed to the Earth’s surface for mainly two reasons: (1) absence of the pressure-vessel effect, and (2) easy access of fluids along the grain boundaries. Previous experiments (e.g., Boyd and England, 1960; Mosenfelder and Bohlen, 1997) demonstrated that the matrix coesite may transform to quartz within hours at temperatures as low as 450 °C provided fluid is available. Thus, intergranular coesite is extremely difficult to preserve from retrograde recrystallization, annealing and fluid infiltration during exhumation. Most of intergranular coesite in UHP metamorphic rocks might have been transformed to quartz whose microstructures would be replaced by those developed later at lower grade deformation during exhumation, and consequently there is no preserved evidence that intergranular coesite existed. The successful preservation of rare intergranular coesite (e.g., Liou and Zhang, 1996; Ye et al., 1996; Zhao et al., 2003), which has persisted metastably in the exhumed eclogitic mylonites from the Sulu metamorphic belt (China), results from the lack of fluid (Liou and Zhang, 1996; Mosenfelder et al., 2005), low temperature (<450 °C), and nonhydrostatic deformation environment during rapid exhumation. The nonhydrostatic deformation environment allowed the interfacial friction-induced overpressure to favor the preservation of intergranular coesite between strong minerals.

4. Concluding remarks

The interfacial friction-induced pressure, which is a well-known mechanical theory in metal forging, provides a viable explanation for the experimental observations that coesite forms in practically undeformed “strain-forbidden zones” directly adjacent to the piston–sample interfaces while the average stress applied to the specimen is significantly lower than the quartz–coesite equilibrium boundary (e.g., Hobbs, 1968; Green, 1972; Hirth and Tullis, 1994). The interfacial friction-induced pressure can be several hundreds of MPa higher than the mean stress exerted on the specimen. Furthermore, the effect of interfacial friction-induced pressure becomes more pronounced at faster strain rates and lower temperatures because the differential stress increases with increasing strain rate and decreasing temperature. If the experimental results can be extrapolated to the natural orogenic belts, the interfacial friction-induced pressure seems to play an important role in the formation and preservation of intergranular coesite in eclogites from the Sulu UHP metamorphic terrane. This recalls that a local pressure deviation from the lithostatic value due to the presence of differential stress and deformation may be an intrinsic property of highly constrained flows of weak materials between strong walls on either regional scale (Rutland, 1965; Mancktelow, 1993, 1995; 2008; Petrini and Podladchikov, 2000; Li et al., 2010) or microstructural scale (Unksov, 1961; Jamieson, 1963; Dieter, 1986; Ji et al., 2000; Lenze et al., 2005).

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