Sustainable densification of the deep crust

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ABSTRACT

The densification of the lower crust in collision and subduction zones plays a key role in shaping the Earth by modifying the buoyancy forces acting at convergent boundaries. It takes place through mineralogical reactions, which are kinetically favored by the presence of fluids. Earthquakes may generate faults serving as fluid pathways but the influence of reactions on the generation of seismicity at depth is still poorly constrained. Here we present new petrological data and numerical models to show that, in the presence of fluids, densification reactions can occur very fast, in the order of weeks, and consume fluids injected during an earthquake, which leads to porosity formation and fluid pressure drop by several hundreds of megapascals. This generates a mechanically highly unstable system subject to collapse and further seismic wave emission during aftershocks. This mechanism creates new pathways for subsequently arriving fluids, and thus provides a route for self-sustained densification of the lower crust.

INTRODUCTION

Mineralogical reactions modify the density and the rheology of the entire lithosphere with important implications for geodynamics in collision and subduction zones (Austrheim, 1987; Hetényi et al., 2007). The presence of fluids at grain boundaries increases element transport and reaction rates by several orders of magnitude (Rubie, 1986; Farver and Yund, 1992).

Permeability estimates of the dry crust (10^{-20} - 10^{-16} m²; Ingebritsen, 2012) preclude pervasive flow as a fluid transport mechanism, thus requiring the formation of preferential fluid pathways for metamorphic reactions to proceed (Plümper et al., 2017; Omlin et al., 2017). Seismic events generate a high-permeability damage zone able to transport fluids and induce reactions in the seismogenic zone (Caine et al., 1996; Reches and Dewers, 2005; typically 0-30 km depth on continents). Earthquakes are also recorded at deep-crustal and intermediate-depth levels in collision and subduction zones (Frohlich, 2006; Halpaap et al., 2019). Thermodynamic calculations predict metamorphic reactions at pressure-temperature conditions consistent with the locations where these earthquakes are recorded (Hacker, 2003; Hetényi et al., 2007; Nakajima et al., 2013), and some laboratory experiments of deformation lead to acoustic emissions due to fracturing during densification reactions (Shi et al., 2018; Incel et al., 2019). There is thus a feedback loop between earthquakes and metamorphic reactions at convergent margins but the underlying mineralogical and mechanical processes of this loop are still enigmatic.

Rocks exposed in the Bergen Arcs (Norway) provide detailed insights into this problem. Their mineralogical assemblages experienced Grenvillian granulite to eclogite metamorphic facies transformation in the root of the Caledonian orogeny (Austrheim, 1987) at 2.0-2.2 GPa and 670-750 °C (Raimbourg et al., 2007; Bhowany et al., 2018), and record frictional melting

(pseudotachylite) during seismic slip (Austrheim and Boundy, 1994). Previous studies reported coseismic damage in the vicinity of pseudotachylites (Austrheim et al., 1996, 2017; Putnis et al., 2017; Petley-Ragan et al., 2018). The eclogite facies minerals form in and along the pseudotachylites, suggesting that the reaction occurs immediately after the earthquake (Pollok et al., 2008; Austrheim et al., 2017). Failure may rely on stress-pulses linked to earthquakes in the seismogenic zone (Jamtveit et al., 2018), or on weakening by shear heating (Braeck and Podladchikov, 2007). However, reactions can also be the cause for the onset of deformation rather than a response as suggested by cracking due to density change associated with reaction (Kirby et al., 1996; Malvoisin et al., 2017). Here we acquire constraints on the interplay between reaction and deformation in the orogenic roots to predict the couplings between lower crust's densification, fluid flow at depth and the related seismicity.

REACTION IN A CLOSED SYSTEM

A granulite from Holsnøy Island (Bergen Arcs; 60°35.561'N, 5°4.374'E) is locally eclogitized and cross-cut by a completely recrystallized, ~ 5 mm wide pseudotachylite (Figure 1). The wall rock to the pseudotachylite contains two different generations of garnet.

Garnets G1 are up to 5 mm wide with compositions in major elements ($X_{\text{Prp}} = 0.50$; $X_{\text{Grs}} = 0.17$) similar to the single generation of garnets found in the unreacted granulite at ~1 m away from the pseudotachylite, suggesting crystallization in granulite facies conditions. They are fragmented by micro-cracks parallel to the pseudotachylite and aligned inclusions of eclogitic minerals (Figure 1B), indicating coseismic deformation as already proposed by Austrheim et al. (2017). The dominant phase in the matrix is fine-grained plagioclase with a jigsaw puzzle texture interpreted as a result of cataclasis during slip along the pseudotachylite (Putnis et al., 2017; Petley-Ragan et al., 2018).

Garnets G2 are $< 30 \mu\text{m}$ wide, and Ca-richer and Mg-poorer ($X_{\text{Prp}} = 0.30$; $X_{\text{Grs}} = 0.38$) than garnets G1. They are part of the eclogitic assemblage consisting of omphacitic pyroxene ($X_{\text{Jad}} = 0.27\text{-}0.39$), kyanite, quartz and carbonate in patches surrounding garnets G1, and zoisite in the plagioclase-rich matrix. The eclogitic patches are surrounded by a symplectic rim composed of an intergrowth of plagioclase and diopsidic pyroxene (Figure 1A).

To constrain the fluid input phase, we measured the O-H absorption of nominally anhydrous minerals with Fourier transform infrared spectroscopy (see Supplemental information).

Concentrations are given as H_2O with uncertainties corresponding to the standard deviation estimated on n measurements for garnet and to the propagation of the uncertainty on the absorption coefficient of Katayama et al. (2006) for clinopyroxene. At less than 1 cm from the pseudotachylite, clinopyroxene contains $210 \pm 37 \mu\text{g/g H}_2\text{O}$ ($n = 34$) and garnet G1 contains $178 \pm 32 \mu\text{g/g H}_2\text{O}$ ($n = 25$) (sample BA7). Garnets G2 are too small to be analyzed. Garnets G1 are composed of different types of domains: i) inclusion-rich domains with bulk H_2O contents above $300 \mu\text{g/g}$ probably formed by fragmentation followed by local recrystallization and fluid inclusions trapping as described in Austrheim et al. (2017; Supplementary Figure S1), indicating the presence of a free aqueous fluid; ii) inclusion-free domains containing $150 \mu\text{g/g H}_2\text{O}$ in their core and less than $50 \mu\text{g/g}$ at their rim (Figure 2A and B). These latter H_2O concentrations are higher than the minimum concentration of $7 \mu\text{g/g H}_2\text{O}$ measured in garnets found in a fresh granulite at ~ 50 cm from the pseudotachylite (sample BA3), and having major element composition similar to garnets G1 (Supplemental table S1). This suggests H_2O incorporation by diffusion but the zoning is opposite to the one expected for diffusion from micro-cracks into garnet G1. H uptake and loss by diffusion in garnets G1 is likely governed by a proton-polaron reaction (Reynes et al., 2018), whereby charge balance is maintained by conversion of Fe^{3+} to

Fe^{2+} during RedOx reactions in the matrix. For the metamorphic temperature of 750°C , experimentally determined diffusivities for this type of defect range from $D = 10^{-15}$ - $10^{-13} \text{ m}^2.\text{s}^{-1}$ (Blanchard and Ingrin, 2004; Kurka et al., 2005; Reynes et al., 2018). For such diffusivities, the timescale for diffusion and reaction are similar for eclogitization conditions. We therefore modelled water zoning in garnet with a 2D numerical model considering both H_2O diffusion in garnet and hydration reaction in the plagioclase matrix (see Supplemental Materials). The model is able to reproduce the measured H_2O zoning in garnet if we consider the system is closed after fluid injection (Figure 2 and Supplementary Figure S3). The free water initially present in the matrix both reacts to form hydrous minerals and diffuses into garnet up to the point where the consumption of H_2O by the hydration of the matrix leads to a lowering of $a_{\text{H}_2\text{O}}$. Garnet then releases its stored H leading to a reversal of the H_2O diffusion profile (Figure 2C to E). This produces a higher amount of reaction products around the garnet (Supplementary Figure S4), similar to the eclogitization patches observed in the natural samples. This reversal of the diffusion profile occurs for the whole range of diffusivities measured experimentally ($10^{-15} \text{ m}^2.\text{s}^{-1} < D < 10^{-13} \text{ m}^2.\text{s}^{-1}$) and for reaction rates up to two orders of magnitude smaller than the one measured experimentally for plagioclase hydration (Schramke et al., 1987; Supplementary Figure S2).

We found in the natural sample, at the transition from eclogite to symplectite (Figure 1B), $\sim 10 \mu\text{m}$ wide amphibole inclusions in omphacitic pyroxene (Supplementary Figure S5). Chlorine and potassium increase from 0 to $\sim 3 \text{ wt.}\%$ and 0 to $\sim 4 \text{ wt.}\%$ (Supplementary Figure S5), respectively, from the clinopyroxene core to its rim in contact with the symplectite. The uptakes of H_2O by hydration reaction and CO_2 by carbonate formation lead to an increase in soluble element activity in the fluid inducing the progressive enrichment in Cl and K in the reaction

products (Kullerød, 1996). This desiccation process also requires reaction in a closed system after a fluid pulse. Such a desiccation process was already reported in the context of salt precipitation during lower crust hydration (Markl and Bucher, 1998).

REACTION-INDUCED WEAKENING

We study the impact of fluid consumption by reaction on the evolution of fluid pressure and porosity with a model coupling reaction, fluid flow and viscoelastic deformation (Malvoisin et al., 2015; Omlin et al., 2017). The 1 x 1 m-sized model starts with a granulite reacting in eclogite facies conditions towards thermodynamic equilibrium at 750°C. The timescale of reaction depends on a first-order kinetic law incorporating the effect of the distance from the equilibrium and the dependency of reactive surface area on porosity. The initial fluid pressure is 2.0 GPa but it evolves as a result of eclogitization and viscous compaction. The above scenario of earthquake-induced fluid pulse followed by reaction translates here as an initial high-porosity zone in the center of the model representing the damage zone surrounding the pseudotachylite (up to 0.5 % of porosity) and no fluid flux at the boundaries (Supplementary Movie S1). The results discussed in the following are obtained with $r_0 = 5.10^{-6} \text{ s}^{-1}$, a permeability of $k = 10^{-19} \text{ m}^2$ and a viscosity of $\eta_\phi = 10^{17} \text{ Pa.s}$ (Schramke et al., 1987; Bürgmann and Dresen, 2008; Ingebritsen, 2012).

The formation of a ~ 18 % denser assemblage leads to a fluid pressure decrease from 2.0 to 1.5 GPa in the numerical model. In the meantime, porosity increases by one order of magnitude (Figure 3A and B) and the new mineralogical assemblage evolves from eclogite to amphibolite facies in the center of the model (Figure 3C). Such a sequence of mineralogical assemblages reproduces the observation in the eclogitic patches at Holsnøy, with eclogitic garnets in the core of the patches and amphibole further out, followed by plagioclase-bearing symplectite at the rim

(Supplementary Figure S5). At larger scale, newly formed Ca-rich plagioclase preferentially occurs in the pseudotachylite and at its rim (Supplementary Figure S6). Thermodynamic calculations indicate that Ca content in plagioclase increases when pressure decreases (Supplementary Figure S6B). This suggests a larger fluid pressure decrease in the highly permeable zones near the pseudotachylite as predicted by the numerical model (Figure 3A). After one week, the model indicates a decrease in porosity to its initial value due to viscous deformation favored by high effective pressure and porosity (Yarushina and Podladchikov, 2015). This estimate of reaction duration is consistent with the results of Taetz et al. (2018) even though it strongly depends on r_0 which is poorly constrained.

A sensitivity analysis reveals that k has no effect on the results due to the no fluid-flux boundary condition and that the model results discussed above are valid for $r_0 > 2 \cdot 10^{-6} \text{ s}^{-1}$ at $\eta_\phi = 10^{17} \text{ Pa.s}$ and $r_0 > 2 \cdot 10^{-6} \text{ s}^{-1}$ at $\eta_\phi = 10^{18} \text{ Pa.s}$ (Supplementary Figure S9). Low η_ϕ values favor limited fluid pressure and porosity variations if r_0 is at least three times smaller than the experimental estimates from Schramke et al. (1987). Together with the uncertainty on the kinetic data, a low fluid availability is the main candidate for a lowering of r_0 . However, the observations of hydrous mineral precipitation and water diffusion demonstrate the presence of an aqueous fluid during eclogitization. Moreover, a low fluid availability implies a higher viscosity (10^{19} Pa.s for dry plagioclase at 700°C ; Bürgmann and Dresen, 2008) for which r_0 has to be more than 100 times smaller than the experimental estimates to produce limited fluid pressure and porosity variations. Alternatively, plastic and brittle deformations, not modelled here, may contribute to compaction.

In nature, the loading is not isotropic as assumed here. Under differential stresses, compaction is associated with strain localization (Wong and Baud, 2012). At high pressure, the connection of

densification sites (anticracks) forms shear instabilities (transformational faulting, Kirby et al., 1996) which have been observed in experiments (Incel et al., 2019) and in natural settings (Nakajima et al., 2013). The pore collapse modelled here is likely to generate other earthquakes (aftershocks) possibly connecting fluid reservoirs to not yet reacted granulites and preparing pathways for subsequent fluid injection phases. This mechanism is probably not the only one responsible for the eclogitisation of ~ 40 % of the Holsnøy Island (Austrheim et al., 1996). Nevertheless, it self-sustains and continuously induces fluid-assisted seismicity as long as there is a fluid source available.

A current analogy to lower-crustal metamorphic earthquakes could be the cluster of events beneath southernmost Tibet (Monsalve et al., 2006) where eclogitization occurs (Hetényi et al., 2007). The range of magnitudes spans from 1 to 4, and the largest event is not purely double-couple but includes an isotropic component (Alvizuri and Hetényi, 2019). On Holsnøy Island, pseudotachylites are dispersed along a roughly linear structure extending over 5 km, with a thickness of at least 150 m. Faulting-related displacement on individual ruptures can locally reach 1.7 m (Jamtveit et al., 2018) but is generally smaller, around 0.1 m on average. If rupture occurred simultaneously along this structure (5 km long, 150 m wide, 10 cm average slip and assuming a shear modulus of 50 GPa), it would yield a Mw 4.3 earthquake, comparable to the largest Himalayan event. This is a maximum estimate for Holsnøy since the pseudotachylites are not aligned along a single fault plane (Austrheim, 2013), nevertheless much smaller events would also compare to the Himalayan cluster with detections down to magnitude 1.

Alternatively, rupture on a non-planar fault could have also occurred in Holsnøy.

CONCLUSION

We show here that earthquakes may lead to a fluid pulse followed by hydrothermal reactions (Figure 4A and B) similar to seismic pumping occurring at shallow depth (Sibson et al., 1975). Reactions subsequent to seismicity progressively consume the fluid in a closed system (desiccation), leading to porosity generation and fluid pressure decrease by up to several hundreds of megapascals (Figure 4C). This significantly weakens the lower crust mechanically, leading to aftershocks that generate new fracture networks by a process similar to transformational faulting (Figure 4D). These networks may be isolated and the reactions shut down as in the sample studied here. Alternatively, they may be connected to a (subsequent) fluid source allowing for the reaction to continue. The reactions lead to extremely short residence time for free fluids in the deep crust, which is otherwise non-reactive over the majority of geological time (Figure 4E).

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AUTHOR CONTRIBUTIONS

B.M., H.A. and G.H. designed the study. B.M. and H.A. collected the samples. B.M., J.R. and J.H. acquired and processed Fourier Transform Infrared data. B.M. and L.P.B. acquired and processed X-ray microtomography data. B.M. and Y.Y.P. developed the numerical models. All authors were involved in interpretation of results and writing the manuscript.

FIGURE CAPTIONS

Figure 1. Mineralogical distribution in sample HKCB_2018_BA7. A: Phase identification based on X-ray mapping with the microprobe; B: 3D garnet distribution (in red) determined with X-ray microtomography (vertical view of the core); Grt: garnet; Pst: pseudotachylite.

Figure 2. H₂O distribution in granulitic garnets found in the vicinity of the pseudotachylite. A and B: H₂O concentration in garnet measured by Fourier infrared spectroscopy. The grains surrounded with a black ellipse display a higher water concentration in their core than at their rim. C to E: prediction of the time evolution of H₂O concentration with a 2D diffusion-reaction model. The system is composed of both garnet grains and a matrix undergoing hydration in a closed system (Neumann boundary conditions; no fluid flux at the boundaries). First, H diffuses from the matrix into the garnets (C). After approximately 1.5 days, H₂O concentration becomes higher in the garnets than in the matrix (D) due to fluid consumption in the matrix with the formation of hydrous phases. Garnets then release their H by diffusion into the matrix (E) leading to hydrous phases formation around garnets. The scale bar is 200 μm .

Figure 3. Numerical modelling of the couplings between reaction, deformation and fluid flow during desiccation (eclogitization reaction in a closed system). Temperature is fixed to 750 °C during the simulation and the initial fluid pressure is 2.0 GPa in the entire model. A: Fluid pressure. The dashed lines delimit the stability fields of several minerals. (+) refers to a mineral stable at a pressure above the line. (-) refers to a mineral stable at a pressure below the line. B: Porosity. C: Mineral modes. In A and B, the evolution with time of the parameters is displayed at

four points aligned along the initial Gaussian distribution in porosity: in the center of the model (red), at 8 cm from the center (blue), at 15 cm from the center (green) and at 25 cm from the center (black). These points are also displayed in Supplementary Movie S1.

Figure 4. Sequence of processes associated with reaction-induced aftershocks in the lower crust.

The table below the sketches gives the evolution of the mechanical, hydrological and petrological conditions as well as the duration of each process.

Figure 1

Figure 2

Figure 3

Figure 4