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# Eclogitisation of dry and impermeable granulite by fluid flow with reaction-induced porosity: insights from hydro-chemical modelling

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## Abstract

Eclogitisation is a major metamorphic process of continental subduction zones, where transformation of dry lower crustal rocks into eclogites seem to correlate with seismogenic events. Eclogitisation can occur at high pressure during hydration of granulite, but the physical processes controlling the hydration of dry, impermeable granulite remain poorly understood. Here, we present a new fully coupled hydro-chemical model of a non-deforming porous rock which undergoes metamorphic reactions in response to fluid pressure variations. Conservation equations for total and solid mass are solved, and fluid and solid densities are calculated with look-up tables computed from models relying on equilibrium thermodynamics. Our model shows that a fluid pressure pulse generates a pressure gradient that causes densification when the pressure required for eclogitisation is reached. The reaction generates porosity and subsequent porous fluid flow into the initially non-porous impermeable granulite. This process lasts as long as the pressure pulse is maintained, but high pressure within eclogite can persist for a longer time. The hydration front propagates tens of centimetres into the granulite in the order of weeks to months. We show that propagation of a hydration-reaction front is effectively a diffusive process, with diffusivity in the order of  $10^{-9} \text{ m}^2 \cdot \text{s}^{-1}$  for eclogitisation as in Holsnøy, Norway. Reactive hydration of impermeable granulite is possible because its solid density is smaller than that of eclogite. We discuss the application of our model for eclogitisation and also for other reactions for which hydration of impermeable rock is possible.

*Keywords:* Eclogitization, hydration front, reaction-induced porosity, hydro-chemical modelling

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## 1. Introduction

The physical properties of the continental crust, such as density, viscosity and permeability, critically affect fluid migration, orogenic cycles and seismicity, and are primarily controlled by the metamorphic state of the lower crust (Austrheim, 1998; Jackson et al., 2004). Eclogitisation of dry and mechanically strong crustal rocks involves a large increase in solid density, transient rheological weakening, and fluid flow (e.g. Austrheim, 1987; Bras et al., 2021), which play a major role for the stress state of the crust, seismicity and long-term lithosphere dynamics (Labrousse et al., 2010; Jamtveit et al., 2018a; Yamato et al., 2019; Moulas et al., 2022). For example, the strong feedbacks between metamorphism, deformation, and fluid transport in the lower crust (John and Schenk, 2006; Austrheim, 2013; Jamtveit et al., 2019) are crucial for determining the physical mechanisms that initiate earthquakes (Miller, 2013; Menegon et al., 2021). Hence, studying eclogitisation of dry crustal rocks is crucial for our understanding of the properties and dynamics of the deep crust (Jamtveit et al., 2019).

The interplay between metamorphism, deformation, and fluid flow can be inferred through geophysical investigations in active orogens (e.g. in the Himalaya, Hetényi et al., 2007), high pressure laboratory rock deformation experiments (Incel et al., 2017; Shi et al., 2018) and mathematical modelling based on the concepts of continuum mechanics and thermodynamics (Malvoisin et al., 2020; Schmalholz et al., 2020; Moulas et al., 2022; Yamato et al., 2022). However, investigating this interplay with direct petrological and structural observations from exhumed crustal rocks in the field is essential to understand how metamorphic reactions control the properties of the deep crust (Jackson et al., 2004; Labrousse et al., 2010; Baisset et al., 2023).

The island of Holsnøy, in southwestern Norway, constitutes an ideal natural laboratory for investigating the interplay between metamorphism, deformation and fluid flow in the lower crust (Austrheim, 1987, 2013; Putnis et al., 2017). A large part of the island is made of anorthositic granulites, which were formed at ca. 930 Ma during the Grenvillian orogeny (Bingen et al., 2001). During the Caledonian orogeny, the granulites were subducted during convergence and collision between Baltica and Laurentia (Andersen et al., 1991). The high-pressure event occurring during the collision stage caused the partial eclogitisation of the granulitic lower crust, associated with fluid infiltration and shear deformation, and was presumably initiated by fracturing associated with earthquakes (Austrheim, 1987). Peak pressure ( $P$ ) and temperature ( $T$ ) conditions in Holsnøy were 670-690 °C and 21-22 kbar (Bhowany et al., 2018). The relation between eclogite and granulite is exceptionally well preserved, making Holsnøy an ideal analogue for the metamorphic state of the lower crust being deeply buried during orogenic events (Jackson et al., 2004; Labrousse et al., 2010).

In Holsnøy, eclogitisation is spatially heterogeneous at the outcrop scale and is localised in eclogite shear zones crosscutting the granulite protolith, and in eclogite fingers overprinting the granulite without associated deformation (Zertani et al., 2019; Kaatz et al., 2021; Putnis et al., 2021). Our study concerns both eclogite shear zones and eclogite fingers (see examples in figure 1), which will

37 both be referred to as "eclogitisation zones" in the rest of this study.

38 Several authors (e.g. [Austrheim, 1987](#); [Zertani et al., 2022](#); [Kaatz et al., 2022](#)) argued that eclogite  
39 and granulite coexist in close spatial proximity in eclogitisation zones because granulite is metastable:  
40 granulite experienced eclogite facies  $P - T$  conditions, but remained largely untransformed because  
41 it lacked fluids to trigger eclogitisation. Only regions infiltrated by fluids transformed into eclogite.  
42 However, recent studies suggested that the coexistence of granulite and eclogite due to local pressure  
43 variations may be a more consistent interpretation of the field data ([Putnis et al., 2021](#) and references  
44 therein). Here, we do not aim to demonstrate which interpretation is more feasible, because our study  
45 is compatible with both interpretations.

46 Several studies suggest that eclogitisation zones widen over time ([Jamtveit et al., 2000](#); [Jolivet  
47 et al., 2005](#); [Bras et al., 2021](#); [Kaatz et al., 2021](#)), but existing numerical models do not explain  
48 this widening with consistent physics, coupling metamorphic reaction and fluid flow. For example,  
49 [Jamtveit et al. \(2000\)](#) explain the development of static eclogitisation zones with a model coupling  
50 solid state diffusion and generation of a network of fractures in an elastic model, without porous fluid  
51 flow. [Bras et al. \(2021\)](#) and [Kaatz et al. \(2021\)](#) model the widening of eclogitisation zones with a  
52 diffusion equation which mimics the effects of porous fluid flow and/or solid state diffusion of hydrogen  
53 and oxygen. [Moulas et al. \(2022\)](#) present a purely mechanical model for metamorphism of the lower  
54 crust, without any hydro-chemical aspect, to study the long-term (order of millions of years) impact  
55 of such metamorphism on crustal dynamics. [Malvoisin et al. \(2020\)](#) consider porous fluid flow with  
56 a fully coupled hydro-chemical-mechanical model and show that hydration of dry granulite can occur  
57 in the order of weeks. However, they focus on the influence of the reaction on the generation of  
58 seismicity but not on the mechanism of widening of eclogitisation zones. Therefore, although several  
59 models already describe the eclogitisation of granulite, it is still not clear how to predict the evolution  
60 of eclogitisation zones based on theoretical models which couple reactions and porous fluid flow.

61 Here, we investigate the time evolution of eclogitisation driven by porous fluid flow following  
62 Darcy's law. We apply a simple but physically consistent one dimensional (1D) hydro-chemical (HC)  
63 model involving metamorphic reactions, based on the concepts of continuum mechanics and local  
64 equilibrium thermodynamics, similar to models presented by [Malvoisin et al. \(2015\)](#), [Plümper et al.  
65 \(2017\)](#), and [Schmalholz et al. \(2020\)](#). Pressure variations modify the rock density by solid compress-  
66 ibility and by replacement of low-density granulite-facies minerals with high density eclogite-facies  
67 minerals. Density changes that decrease the rock volume generate porosity, and hence permeability if  
68 it is dependent on porosity. Permeability associated with fluid pressure gradients leads to fluid flow  
69 and reaction front propagation. Our aims are to (1) quantify the widening of the eclogitisation zones  
70 and (2) explain the mechanism of hydration of a zero-porosity, and thus impermeable, granulite by  
71 porous fluid flow.

72 From a technical point of view, modelling fluid flow into a zero-porosity region is indeed challenging.

73 The zero porosity limit is typically met in studies involving two-phase flow (e.g. [Scott and Stevenson,](#)  
74 [1986](#)). Although the set of governing two-phase equations reduce to single-phase equations in the zero  
75 porosity limit (e.g. [Katz et al., 2007](#)), accounting for transient variations of the relative proportion  
76 of porous and non-porous domains in a single simulation can be challenging. Several studies get  
77 around this problem by having a minimal background porosity in the dry phases (e.g. [Räss et al.,](#)  
78 [2019](#); [Schmalholz et al., 2020](#)), which allows to apply two-phase flow equations to the whole domain.  
79 Other authors have addressed this problem by including a specific treatment of these equations (e.g.  
80 [Arbogast et al., 2017](#)). In this paper we also address this issue in the context of the hydration of a  
81 zero-porosity region.

82 With our model, we show that fluid flow induced hydration and subsequent eclogitisation of a dry  
83 and impermeable rock is possible. We explain the physical mechanisms by which the eclogitisation  
84 zones can widen into an initially zero-porosity region, and provide a numerical tool allowing to simulate  
85 this process.

## 86 2. Methods

87 We consider a 1D hydro-chemical (HC) model, similar to that of [Plümper et al. \(2017\)](#), where a non-  
88 deforming porous rock undergoes metamorphic reactions in response to fluid pressure variations. The  
89 model is based on local thermodynamic equilibrium and mass conservation. Metamorphic reactions  
90 lead to changes in rock density and exchange of H<sub>2</sub>O between the fluid and the solid, hence modifying  
91 the porosity of the system. Since we use a porosity-dependant, Carman-Kozeny type permeability  
92 ([Costa, 2006](#)), changes in porosity also modify the permeability. Reaction-induced permeability leads  
93 to fluid flow controlled by fluid pressure gradients.

### 94 2.1. Governing system of equations

95 In our model, the volatile species, assumed here to be pure H<sub>2</sub>O, are either structurally bound to  
96 the solid, having a particular solid mass fraction  $X_{\text{H}_2\text{O}}$ , or constitute the fluid filling the porosity. We  
97 assume that dissolution of solid species, such as Si, Al, Na and Ca, in the fluid is negligible. Following  
98 [Malvoisin et al. \(2015\)](#); [Plümper et al. \(2017\)](#); [Schmalholz et al. \(2020\)](#), solving this system requires  
99 (1) conservation of total mass per unit volume of the system ( $\rho_T$ , i.e. sum of fluid mass and solid  
100 mass) and (2) conservation of the mass per unit volume of solid species ( $\rho_T^s$ ). All variables used in  
101 this study are listed in table 1.

#### 102 2.1.1. Conservation of total mass

103 We consider a porous medium, with porosity  $\phi$ , which consists of a solid phase with density  $\rho_s$  and  
104 a fluid phase with density  $\rho_f$  stored in the pores. Per unit volume, the total mass  $\rho_T$  of the porous  
105 medium is:

$$\rho_T = \rho_f \phi + \rho_s (1 - \phi). \quad (1)$$

106 In a general case, the one-dimensional conservation equation of a given quantity A, expressing the  
 107 variation of quantity A (per unit volume) over time, is:

$$\frac{\partial A}{\partial t} = -\frac{\partial q_A}{\partial x}. \quad (2)$$

108 In absence of diffusive flux, the flux of A ( $q_A$ ) is purely advective and can be expressed as:

$$q_A = Av_A, \quad (3)$$

109 which is simply the amount of A multiplied by the velocity of A. Using equations (1) and (3), the  
 110 advective flux of total mass  $\rho_T$  can be written as:

$$q_{\rho_T} = \rho_f \phi v_f + \rho_s (1 - \phi) v_s, \quad (4)$$

111 where  $v_f$  and  $v_s$  are respectively the velocities of the fluid and the solid. Using equations (2) and (4),  
 112 the conservation of total mass  $\rho_T$  is then:

$$\frac{\partial \rho_T}{\partial t} = -\frac{\partial}{\partial x} (\rho_f \phi v_f + \rho_s (1 - \phi) v_s). \quad (5)$$

113 In the absence of solid deformation ( $v_s = 0$ ), the conservation equation of total mass simplifies to:

$$\frac{\partial \rho_T}{\partial t} = -\frac{\partial}{\partial x} (\rho_f \phi v_f). \quad (6)$$

114 The force balance for the fluid in the absence of gravity and solid velocity can be expressed using  
 115 Darcy's law:

$$\phi v_f = -\frac{k_0 \phi^3}{\eta_f} \left( \frac{\partial P_f}{\partial x} \right), \quad (7)$$

116 where  $k_0$  is the permeability coefficient in a porosity-dependant Carman-Kozeny type permeability  
 117 expression,  $\eta_f$  is the fluid viscosity, and  $P_f$  is the fluid pressure. Substituting equation (7) into equation  
 118 (6) yields the total mass conservation equation:

$$\frac{\partial \rho_T}{\partial t} = \frac{\partial}{\partial x} \left[ \rho_f \frac{k_0 \phi^3}{\eta_f} \left( \frac{\partial P_f}{\partial x} \right) \right]. \quad (8)$$

### 119 2.1.2. Total solid mass

120 We assume that solid components, such as Si, Al, Na and Ca, never enter the fluid phase, whereas  
 121 the volatile component H<sub>2</sub>O can be part of the solid phase in the form of hydrous minerals.  $X_{\text{H}_2\text{O}}$   
 122 is the mass fraction of volatile components structurally bound in the solid. The total solid mass (without  
 123 structurally bound H<sub>2</sub>O) of the system per unit volume,  $\rho_T^s$ , is:

$$\rho_T^s = \rho_s (1 - \phi) (1 - X_{\text{H}_2\text{O}}). \quad (9)$$

124 Solid components remain in the solid, so in the absence of solid velocity there is never any advective  
 125 flux of solid mass (equation 3), and thus there is no time evolution of  $\rho_T^s$ :

$$\frac{\partial \rho_T^s}{\partial t} = 0. \quad (10)$$

126 Hence, equation (9) can be used to express the solid mass conservation equation if  $\rho_T^s$  is a known,  
 127 constant parameter of the model.

## 128 2.2. Precomputed thermodynamic data

### 129 2.2.1. Thermodynamic model

130 The equations described above require prior knowledge of the densities of fluid, eclogite and gran-  
131 ulite at any pressure, as well as the amount of fluid structurally bound to the solid  $X_{\text{H}_2\text{O}}$ . We compute  
132 these values by minimisation of Gibbs free-energy using the software *Perple\_X* (Connolly, 2005), with  
133 the internally consistent thermodynamic data set of Holland and Powell (1998).

134 To study the eclogitisation of granulite in Holsnøy, we consider the model chemical system  $\text{Na}_2\text{O}-$   
135  $\text{CaO}-\text{SiO}_2-\text{Al}_2\text{O}_3-\text{H}_2\text{O}$ . We do not consider FeO and MgO because the main reactions of granulite to  
136 eclogite are destabilisation of the anorthite and albite components of plagioclase (e.g. Bras et al., 2021):  
137 with increasing pressure and in the presence of water, anorthite transforms into zoisite, kyanite and  
138 quartz, and albite transforms into jadeite and quartz. Our modelled granulite is therefore composed  
139 of only plagioclase, and our modelled eclogite of jadeite, zoisite, kyanite, quartz. We use a system  
140 composition of 57.64 wt%  $\text{SiO}_2$ , 26.92 wt%  $\text{Al}_2\text{O}_3$ , 8.76 wt% CaO and 6.68 wt%  $\text{Na}_2\text{O}$ , without  
141 FeO or MgO, corresponding to a representative composition of plagioclase in Holsnøy granulite (Bras  
142 et al., 2021). We use the solid solution for plagioclase of Newton et al. (1980) (Pl(h) in *Perple\_X*).  
143 We use a constant bulk composition and variable water content, as there is no major variation of  
144 composition between granulite and eclogite, besides the introduction of water (Centrella, 2019; Bras  
145 et al., 2021). Rock properties are computed at a constant temperature of 700°C, consistent with  
146 temperature estimates in both granulite and eclogite by Bhowany et al. (2018).

147 The aim of the thermodynamic calculation is to compute bulk rock densities as a function of pres-  
148 sure in order to use them in our model, and not to predict mineral assemblages. We can therefore  
149 simplify the thermodynamic database by only including minerals that are present in substantial pro-  
150 portions in Holsnøy granulites and eclogites. This simplification produces the pseudosection displayed  
151 in figure 2A. With this simplified set-up, dry granulite is modelled with pure plagioclase and no wa-  
152 ter at low pressure, and hydrous eclogite is modelled with the products of plagioclase destabilisation  
153 (jadeite + kyanite + zoisite + quartz) and water at high pressure (figure 2A, yellow line). Due to  
154 the aforementioned simplifications the modelled data is not identical to field data, but has the same  
155 general behaviour: a jump in density and amount of fluid stored in minerals between granulite and  
156 eclogite.

157 When solid-fluid interactions are considered, it is the fluid pressure that determines which phases  
158 are stable in a local equilibrium domain, rather than the mean stress (e.g Dahlen, 1992). Indeed,  
159 Llana-Fúnez et al. (2012) experimentally demonstrated that metamorphic reactions are controlled by  
160 fluid pressure rather than by confining pressure. Fluid pressure is therefore an appropriate macroscopic  
161 proxy of the pressure rocks are in equilibrium with, but in the absence of porosity there is no fluid  
162 pressure. Hence, throughout this study we use the general term "pressure" ( $P$ ) to designate both fluid  
163 pressure in porous rocks and mean stress in rocks without porosity.

The solid density  $\rho_s$ , fluid density  $\rho_f$ , and the amount of fluid structurally bound in the solid  $X_{\text{H}_2\text{O}}$  at each pressure are precomputed with the thermodynamic model described above, for use as look-up tables during simulations:

$$\rho_s = \rho_s(P), \quad (11a)$$

$$\rho_f = \rho_f(P), \quad (11b)$$

$$X_{\text{H}_2\text{O}} = X_{\text{H}_2\text{O}}(P). \quad (11c)$$

165 Additionally, an inverse look-up table can be computed using these data. At any given pressure, the  
 166 values of  $\rho_s$ ,  $\rho_f$  and  $X_{\text{H}_2\text{O}}$  are thermodynamically constrained (equation 11). Since the value of  $\rho_T^s$  is  
 167 a known parameter of the model,  $\phi$  can be computed at any pressure as well, using equation (9).  $\phi$ ,  
 168  $\rho_s$  and  $\rho_f$  can then be used to compute  $\rho_T$  with equation (1), ending up with a total mass look-up  
 169 table ( $\rho_T = \rho_T(P)$ , figure 2B).  $\rho_T$  must monotonously increase with pressure, which is a consequence  
 170 of thermodynamic stability (e.g. Landau and Lifshitz, 1980). Hence, for every given pressure there is  
 171 a uniquely corresponding  $\rho_T$ . This uniqueness makes the inverse statement true: for every given  $\rho_T$   
 172 there is one uniquely corresponding pressure. We therefore have an inverse look-up table of the form:

$$P = P(\rho_T). \quad (12)$$

173 This inverse look-up table approach allows our model to account for both porous and non-porous  
 174 regions. There is thus no need to assume an artificially low porosity in the non-porous domain.

### 175 2.3. Solving procedure

176 All partial differential equations are approximated with a finite-difference discretisation (e.g. Gerya,  
 177 2019). Our HC model consists of a closed system of 6 unknowns ( $\rho_s$ ,  $\rho_f$ ,  $X_{\text{H}_2\text{O}}$ ,  $\rho_T$ ,  $\phi$ ,  $P$ ) and 6 equa-  
 178 tions (8, 9, 11a, 11b, 11c, 12). We use a pseudo-transient method (e.g. Räss et al., 2022) to iteratively  
 179 solve these 6 equations until a converged solution is found, proceeding as follows:

- 180 1. Use the forward look-up tables (equation 11) to compute  $\rho_s$ ,  $\rho_f$  and  $X_{\text{H}_2\text{O}}$  as a function of  $P$ ,
- 181 2. Rearrange equation (9) to solve for  $\phi$  as a function of  $\rho_s$  and  $X_{\text{H}_2\text{O}}$ :

$$\phi = 1 - \frac{\rho_T^s}{\rho_s(1 - X_{\text{H}_2\text{O}})}, \quad (13)$$

- 182 3. Explicitly solve equation (8) to compute  $\rho_T$  as a function of  $\phi$ ,  $P$  and  $\rho_f$ :

$$\rho_T^{\text{new}} = \rho_T^{\text{old}} + \frac{\partial}{\partial x} \left[ \rho_f \frac{k\phi^3}{\eta_f} \left( \frac{\partial P}{\partial x} \right) \right] \times dt, \quad (14)$$

- 183 4. Use the inverse look-up table (equation 12) to compute  $P$  as a function of  $\rho_T$ .

184 In practice, at each time step we iterate this set of equations until the difference in calculated pressure  
 185 from one iteration to another is smaller than a chosen convergence tolerance.

## 186 2.4. Model configuration

187 The parameters of our reference model are chosen to be close to natural values reported in the  
188 literature. Unless stated otherwise, all simulations displayed in this study use the parameters described  
189 below and shown in table 1.

### 190 2.4.1. Starting conditions

191 A common explanation of the development of eclogitisation zones (e.g. [Austrheim, 1987](#); [Zertani  
192 et al., 2019](#); [Kaatz et al., 2021](#)) can be summarised as follows: (1) Lower crustal earthquakes cause  
193 stress pulses and propagate fractures in the brittle granulite, which act as a pathway for fluids. (2)  
194 Fluid availability triggers eclogitisation. (3) Infiltration of fluid into the granulite further promotes  
195 eclogitisation. Our model reproduces this configuration (figure 3): when the simulation starts, the  
196 entire domain is made up of granulite, with pressure set at  $P_{bg} = 15$  kbar, close to the pressure of  
197 seismic faulting reported by [Bhowany et al. \(2018\)](#). In the centre of the domain a pulse of high fluid  
198 pressure is applied to a narrow band ( $w_0 = 1$  cm), mimicking the initial crack that allows for fluid flow  
199 into the granulite host-rock. Introducing the shorthand notation  $\Delta P_r = P_r - P_{bg}$ , with  $P_r$  the reaction  
200 pressure, that expresses the increase in pressure needed for the granulite to cross the reaction, it is clear  
201 that the pressure pulse must be greater than  $\Delta P_r$  to trigger eclogitisation. The pressure of the fluid  
202 pulse is set at  $\Delta P_f = 10$  kbar, consistent with estimates of possible overpressure in Holsnøy ([Centrella,  
203 2019](#); [Jamtveit et al., 2018b](#); [Moulas et al., 2022](#)), and fluid pressure pulses in other geological contexts  
204 (e.g. [Vrijmoed et al., 2009](#); [Viète et al., 2018](#)). A study of the effect of  $\Delta P_f$  is presented in the result  
205 section. To assert the robustness of our approach, the initial pressure perturbation is applied as a step  
206 function:  $P = 25$  kbar in the central crack and 15 kbar in the background.

### 207 2.4.2. Pressure pulse duration

208 Episodic high fluid pressure pulses in eclogite-facies rocks have been linked to dehydration reactions  
209 ([Taetz et al., 2018](#); [Hoover et al., 2022](#)). Indeed, dehydration of adjacent units of the subducting plate  
210 can cause accumulation of pore fluids, resulting in periodic fluid pressure pulses that may cause brittle  
211 failure ([Audet et al., 2009](#); [Broadwell et al., 2019](#)). [Viète et al. \(2018\)](#) showed that metamorphic rocks  
212 can record such fluid overpressures events.

213 The scenario we model is consistent with these observations. A fluid pressure pulse is applied in  
214 the central crack during a period  $t_p = 50$  days, that can be interpreted as a single short pulse, or a  
215 series of very short-lived pressure pulses. This is consistent with the study of [Malvoisin et al. \(2020\)](#),  
216 who conclude that eclogitisation of dry granulite promotes further earthquake generation, which can  
217 self sustain the presence of high pressure pulses. It is also in agreement with the study of [Putnis  
218 \(2021\)](#) who argue that the transition in texture and composition from an untransformed granulite to  
219 an eclogite finger suggest a series of fluid pulses that advance the eclogitisation front into the granulite.  
220 Sporadic fluid pressure pulses can also be related to seismic aftershocks ([Jamtveit et al., 2018a](#)).

221 During the 50 days of the fluid pulse event, the model is an open system and the pressure in  
222 the initial crack is always  $P_{\text{bg}} + \Delta P_f = 25$  kbar. This period is thereafter referred to as the open  
223 system period. After the first 50 days, the system closes and the pressure is free to drop in the initial  
224 crack. This period is referred to as the closed system period and lasts another 50 days, after which the  
225 simulation ends. [Kaatz et al. \(2021\)](#) likewise considered both open and closed systems, but in terms  
226 of amount of available fluid instead of fluid pressure.

### 227 *2.4.3. Total solid mass*

228 To assert the robustness of our approach, we set the background granulite porosity to 0 by using  
229 equation (9): given that  $X_{\text{H}_2\text{O}} = 0$  in granulite, we set  $\rho_T^s = \rho_s$  to ensure that  $\phi = 0$ . In other  
230 words, the solid mass per unit volume  $\rho_T^s$  must be equal to the density of granulite at 15 kbar, i.e.  
231  $2685 \text{ kg} \cdot \text{m}^{-3}$  (figure 2B). In the central crack, where the total pressure is higher than the reaction  
232 pressure, the granulite of density  $2685 \text{ kg} \cdot \text{m}^{-3}$  is replaced by eclogite, with a density of  $3228 \text{ kg} \cdot \text{m}^{-3}$ .  
233 As the total solid mass is conserved (equation 9), the porosity in the eclogite increases from 0 to 16%  
234 (figure 2C).

### 235 *2.4.4. Permeability and fluid viscosity*

236 In a Carman-Kozeny type relationship, the permeability is equal to  $k_0\phi^3$ , where  $k_0$  is a permeability  
237 coefficient and  $\phi$  the porosity. In our model, eclogite porosity is around 0.16 (figure 2C). Using a  
238 permeability coefficient  $k_0 = 10^{-19} \text{ m}^2$  yields an eclogite permeability of  $4 \times 10^{-22} \text{ m}^2$ , consistent  
239 with estimates of crustal permeability by [Shmonov et al. \(2003\)](#) or [Mindaleva et al. \(2020\)](#). Granulite  
240 permeability is 0 because its porosity is 0. We use a dynamic fluid viscosity  $\eta_f = 10^{-3} \text{ Pa} \cdot \text{s}$ , consistent  
241 with [Audéat and Keppler \(2004\)](#); [Hack and Thompson \(2011\)](#).

### 242 *2.5. Validation of the model: resolution test*

243 Resolution tests in both space and time were performed and are presented in supplementary figures  
244 S1 and S2. The results of tests for spatial resolution (S1) show convergence with increasing resolution,  
245 indicating that the employed finite differences discretisation is appropriate for HC modelling. We use  
246 400 numerical grid points for the discretisation in space. One simulation can be computed in a few  
247 minutes on a standard laptop. Changing the time resolution (Courant–Friedrichs–Lewy condition,  
248 CFL) does not affect the predicted eclogite width (S2A) and has very little effects on the modelled  
249 pressure (S2B).

## 250 **3. Results**

### 251 *3.1. Introductory examples*

252 Figure 4 illustrates the behaviour of the two end-member cases described previously: open system  
253 with a fluid pulse continuous in time, and closed system with an instantaneous fluid pulse only during

Table 1: Model variables, and parameters used in the reference model. The equation determining the value of the model unknowns are indicated.

Symbol	Meaning	Unit	Value
$P$	Pressure	Pa	Eq. 12
$\phi$	Porosity	-	Eq. 9, 13
$\rho_T$	Total mass per unit volume	$\text{kg} \cdot \text{m}^{-3}$	Eq. 1, 14
$\rho_s$	Solid density	$\text{kg} \cdot \text{m}^{-3}$	Eq. 11a
$\rho_f$	Fluid density	$\text{kg} \cdot \text{m}^{-3}$	Eq. 11b
$X_{\text{H}_2\text{O}}$	H <sub>2</sub> O mass fraction	-	Eq. 11c
$X_s$	$1 - X_{\text{H}_2\text{O}}$	-	
$v_f$	Fluid velocity	$\text{m} \cdot \text{s}^{-1}$	
$\beta$	Solid compressibility	$\text{Pa}^{-1}$	
$T$	Temperature	$^{\circ} \text{C}$	700
$P_{\text{bg}}$	Background pressure	kbar	15
$\Delta P_f$	Pulse fluid pressure	kbar	10
$P_r$	Reaction pressure	kbar	19.4
$\Delta P_r$	$P_r - P_{\text{bg}}$	kbar	4.4
$w_0$	Width of initial crack	m	0.01
$t_p$	Duration of fluid pulse	days	50
$\rho_T^s$	Total solid mass per unit volume	$\text{kg} \cdot \text{m}^{-3}$	2685
$k_0$	Permeability coefficient	$\text{m}^2$	$10^{-19}$
$\eta_f$	Fluid viscosity	$\text{Pa} \cdot \text{s}$	$10^{-3}$
$n$	Porosity exponent	-	3

254 the first numerical time step. The separate study of these two cases is useful to understand the  
 255 following HC models that include both behaviours.

### 256 3.1.1. Open system

257 Figure 4A shows the behaviour of an open system. The high pressure perturbation in the centre of  
 258 the domain is maintained, mimicking a continuous fluid supply, or a succession of fluid pressure pulses  
 259 very close in time. This results in a continuous influx of mass into the system, since  $\rho_T = \rho_T(P)$  (figure  
 260 2B). The rest of the configuration is identical to that presented in figure 3, except that the pressure  
 261 perturbation is applied at only one numerical grid point. The continuous high pressure perturbation  
 262 creates a pressure gradient between the crack and the granulite, leading to a fast hydration due to  
 263 high Darcy flow. The eclogitisation zone expands rapidly from the crack, and is around 30 cm wide  
 264 after 100 days.

### 265 3.1.2. Closed system

266 Figure 4B shows the behaviour of a closed system. Here, there is no additional supply of fluid  
267 pressure after the initial pressure perturbation, so the total mass present in the system is conserved.  
268 In a few days, the pressure profile evolves from the initial pressure perturbation to a plateau at the  
269 reaction pressure  $P_r$ . Afterwards, this plateau widens, although much more slowly than in the open  
270 system case, and persists long after the initial pressure perturbation has decayed.

### 271 3.2. Reference model

272 The time evolution of our reference model is presented in figure 5, with a total simulation time of  
273 100 days. The system is open during the first half of the simulation, with a behaviour similar to figure  
274 4A, and is closed during the second half, with a behaviour similar to figure 4B. Granulite instanta-  
275 neously transforms into eclogite when the pressure leaves the granulite stability field. Because of the  
276 densification reaction, the eclogitisation zone is characterised by a high reaction-induced porosity. The  
277 eclogitisation zone spreads rapidly during the open system phase, from its initial width of 1 cm to 23  
278 cm in 50 days (figure 5C). During the closed system stage the eclogite width increases only slightly,  
279 from 23 to 26 cm. Most of the widening occurs while the system is open, i.e. when a fluid pressure  
280 perturbation is active. At the end of the simulation the system is composed of a central eclogite region  
281 with  $P = 19.4$  kbar and  $\phi = 0.16$ , with a sharp transition to the granulite region with  $P = 15$  kbar  
282 and  $\phi = 0$ .

### 283 3.3. Parametric study

284 Simulations with varying duration of the pulse ( $t_p$ ) and varying magnitude of the fluid pressure  
285 pulse ( $\Delta P_f$ ) were performed in order to determine how they affect the width of the eclogitisation  
286 zone (figure 6). Each simulation lasts for twice the duration of the fluid pulse: if  $t_p = 50$  days, the  
287 simulation stops after 100 days. This allows us to compare the width of the eclogitisation zones in the  
288 same closed-system conditions. Every other parameter is the same as in the reference model (table 1,  
289 results in figure 5). Figure 6A shows that  $t_p$  and  $\Delta P_f$  both contribute to increasing the width of the  
290 eclogitisation zone. Since eclogitisation occurs when  $P$  reaches the reaction pressure  $P_r$ , any model  
291 with a pressure pulse lower than  $\Delta P_r = 4.4$  kbar yields an eclogite width of 0.

292 Using the data displayed in figure 6A, we derived a formula for the width of an eclogitisation zone  
293  $w$  as a function of  $t_p$ ,  $\Delta P_f$ , the width of the initial crack  $w_0$  and the pressure increase required to  
294 eclogitise the granulite  $\Delta P_r$ :

$$w \approx \alpha \sqrt{t_p (\Delta P_f - \Delta P_r)} + w_0, \quad (15)$$

295 where  $\alpha$  is a constant determined empirically (with unit  $\text{m} \cdot \text{Pa}^{-0.5} \cdot \text{s}^{-0.5}$ ), and with  $t_p$  in seconds,  $\Delta P_f$   
296 and  $\Delta P_r$  in Pascals,  $w$  and  $w_0$  in metres. For a simulation time that is twice the duration of the  
297 open system,  $\alpha = 4.76 \times 10^{-9} \text{ m} \cdot \text{Pa}^{-0.5} \cdot \text{s}^{-0.5}$ . This formula predicts  $w$  with little deviations from the

298 expected value (fig 6B,  $r^2 = 0.997$ ), and is shown with the dotted line on figure 6C. For a simulation  
 299 that only lasts until the system closes,  $w$  is predicted by equation (15) with a constant  $\alpha = 4.47 \times 10^{-9}$   
 300  $\text{m} \cdot \text{Pa}^{-0.5} \cdot \text{s}^{-0.5}$ , and is shown with the dashed line on figure 6C. For a given  $\Delta P_f$ , the time evolution  
 301 of a model follows closely the widths predicted by equation (15).

302 Equation (15) shows that  $w$  has a square root dependency to the duration of the pulse  $t_p$  and to  
 303 the pulse pressure overstep, relative to the reaction pressure,  $\Delta P_f - \Delta P_r$ . For our reference model,  
 304 with parameters  $t_p = 4.3 \times 10^6$  s (50 days),  $\Delta P_f = 10^9$  Pa,  $\Delta P_r = 4.4 \times 10^8$  Pa, and  $w_0 = 0.01$   
 305 m, equation (15) yields  $w = 0.24$  m, very close to the modelled value (figure 5). With parameters  
 306  $t_p = 1.58 \times 10^8$  s (5 years),  $\Delta P_f = 5 \times 10^8$  Pa,  $\Delta P_r = 4.4 \times 10^8$  Pa, and  $w_0 = 0.01$  m, equation (15)  
 307 yields  $w = 0.47$  m.

## 308 4. Discussion

### 309 4.1. General features of the HC model

#### 310 4.1.1. Time evolution of the system

311 Our results show that a very sharp pressure/porosity reaction front can propagate into a rock that  
 312 has zero porosity, and hence, zero permeability. When the system is open and is driven by a high  
 313 pressure gradient, eclogite width increases as a function of  $\sqrt{t}$ , which is typical of a diffusion process  
 314 (e.g. Turcotte and Schubert, 2002), although the model does not explicitly solve a diffusion equation.  
 315  $\Delta P_f$  is as important as  $t_p$  in the widening of eclogitisation zones: the pressure gradient determines  
 316 how fast fluid can flow into the dry granulite.

317 When the system closes, the impermeability of granulite causes the fluid flow to be extremely slow,  
 318 preventing fluid pressure diffusion. As a result, the high pressure plateau persists long after the end  
 319 of the short-lived fluid pressure pulse (figures 4b, 5a), preserving the eclogitisation zone without any  
 320 sustained pressure perturbation.

321 In our reference model (figure 5), a 25 cm wide eclogitisation zone can form in 100 days. This  
 322 result is in agreement with that of Beinlich et al. (2020), who show that a reaction front can advance  
 323 as fast as 10 cm per year, and with that of Taetz et al. (2018), who show that the duration of fluid-  
 324 rock interaction in an eclogite mélange is 1 to 4 months. Our time estimate is also close to that of  
 325 the hydro-mechanical-chemical model of Malvoisin et al. (2020), who conclude that eclogitisation of  
 326 granulite can occur on the order of weeks.

#### 327 4.1.2. Contribution of compressibility and eclogitisation to front propagation

328 Two processes contribute to an increase in solid density, and thus an increase in porosity: compress-  
 329 ibility (increase in density of a given phase with increasing pressure) and eclogitisation (replacement  
 330 of low density minerals by high density minerals). We compared the importance of these two processes  
 331 by switching them on and off in different simulations (figure 7). The results show that both effects

332 contribute to the widening of the hydration zone. Eclogitisation generates most of the porosity by  
333 a significant increase in solid density, creating a stronger fluid flow and therefore a faster hydration  
334 front. Compressibility creates only little porosity, and therefore limited hydration.

#### 335 *4.1.3. Simplifications of the 1D HC model*

336 **Permeability.** We consider a homogeneous, isotropic-permeability rock, whereas metamorphic  
337 rocks can have anisotropic permeabilities that are 1 to 2 orders of magnitude higher along the meta-  
338 morphic fabric than in the normal direction (e.g. [Acosta and Violay, 2020](#)). This anisotropy cannot  
339 be included in our 1D model, but a 2D model would allow to model preferential growth of eclogite  
340 fingers in the direction of the metamorphic fabric of the granulite ([Putnis et al., 2021](#)).

341 **Deformation.** The aim of the present paper is to show, with a simple hydro-chemical model, that  
342 a high pressure perturbation, here caused by a fluid pulse, can propagate and drive a reactive porous  
343 fluid flow into an impermeable rock. Coupling shear deformation with hydro-chemical processes is out  
344 of the scope of this study, and an extension of our work would include viscous deformation in order  
345 to simulate the growth of shear zones. During deformation, the weak shear zones could exhibit larger  
346 pressures than the strong, dry host-rock, if they have a suitable orientation relative to the background  
347 stress field ([Moulas et al., 2014](#); [Jamtveit et al., 2018b](#)). Stress caused by shear deformation could self  
348 sustain the high pressure anomaly in the shear zone and promote hydration front propagation and  
349 shear zone widening.

#### 350 *4.2. Applications to Holsnøy case study*

##### 351 *4.2.1. High pressure sustainability in eclogite*

352 In Holsnøy, eclogitisation occurs in (1) eclogite shear zones, that undergo deformation during  
353 eclogitisation (e.g. [Austrheim, 1987](#); [Bras et al., 2021](#)), and (2) eclogite fingers that undergo static  
354 eclogitisation without associated macroscopic deformation (e.g. [Jamtveit et al., 2000](#); [Zertani et al.,](#)  
355 [2019](#); [Baïssset et al., 2023](#)). In our model we attribute overpressure only to a short-lived fluid pressure  
356 pulse, but additional causes of overpressure may be at play, in both eclogite fingers and shear zones.

357 **Eclogite fingers.** Consistent with our work, [Putnis \(2021\)](#) inferred a pressure gradient from  
358 the host-rock granulite to an eclogite finger, deducing that the eclogite finger experienced a higher  
359 pressure. It was also shown ([Jamtveit et al., 2000](#); [Baïssset et al., 2023](#)) that static eclogitisation of  
360 granulite causes perturbation in the local stress field, and could accelerate the propagation of fluids.  
361 [Yamato et al. \(2022\)](#) recently proposed that eclogitisation-induced densification of reacting granulite  
362 can generate sufficient shear stress to fail the rocks, and self-propagates eclogitisation in a finger-like  
363 structure. Reaction-induced densification therefore appears to be a likely mechanism that can sustain  
364 high pressures where eclogitisation takes place.

365 **Shear zones.** Sustained high pressure in eclogite shear zones can be achieved by weakening-  
366 induced overpressure ([Jamtveit et al., 2018b](#)), and numerous studies showed that eclogitisation me-

367 mechanically weakens the granulite (Jolivet et al., 2005; Bras et al., 2021; Bâisset et al., 2023). This  
368 weakening can have crucial effects: Mancktelow (1993) showed that in deforming rocks, weak and  
369 strong layers can experience pressure differences of several kbar. Numerical simulations and analytical  
370 solutions show that the presence of a weak shear zone within a strong stressed lower crust can sig-  
371 nificantly increase the pressure inside the shear zone, potentially more than 10 kbar (Schmalholz and  
372 Podladchikov, 2013; Moulas et al., 2014; Jamtveit et al., 2018b; Moulas et al., 2022). The relevance of  
373 weakening induced pressure variations has also been suggested based on field observations (e.g. Luisier  
374 et al., 2019). Weakening-induced pressure increase in eclogite shear zones is therefore probably signif-  
375 icant (Jamtveit et al., 2018b) and could self-sustain overpressure in the eclogite, efficiently promoting  
376 fluid flow and further eclogitisation of the dry lower crust.

#### 377 4.2.2. Metastability

378 As stated in the introduction, there is currently a debate on the pressure state of granulite in  
379 Holsnøy: the juxtaposition of eclogite and granulite regions can be either interpreted as (1) metasta-  
380 bility of granulite at eclogite-facies conditions at 21 kbar (Bhowany et al., 2018) or (2) local pressure  
381 variations between granulite and eclogite (see discussion in Putnis et al., 2021). Here, we show that  
382 the local pressure hypothesis is consistent with widening of eclogitisation zones. However, we can also  
383 test the metastable hypothesis with two modifications to our numerical code:

- 384 1. Set the background pressure to 21 kbar, and
- 385 2. Change the reaction pressure from the value predicted by thermodynamic modelling (19.4 kbar)  
386 to 21.01 kbar, just above the background pressure.

387 By doing so, granulite is stable up to 21 kbar, at eclogite pressure conditions. Any little increase in  
388 pressure caused by fluid flow triggers the transition from granulite to eclogite, satisfyingly reproducing  
389 the metastable behaviour where the reaction is triggered by a fluid influx. We here use a pressure  
390 of 21 kbar, close to the peak pressure estimated by Bhowany et al. (2018), but any pressure greater  
391 than  $P_f$  could be used. The general behaviour of the system is the same as in the stable scenario (see  
392 supplementary figure S3), the only difference being a wider eclogitisation zone in the metastable case  
393 (32 cm) compared to the stable case (24 cm). Although metastable granulite does not thermodynam-  
394 ically require a large pressure increase to transform into eclogite, a pressure gradient is still necessary  
395 to drive the fluid into the dry granulite by Darcy flow.

#### 396 4.3. Reaction front propagation controlled by density of solid components

##### 397 4.3.1. Theoretical considerations

398 We showed that widening of high pressure, hydrated zones is possible in the case of the granulite  
399 to eclogite reaction. In a general case, propagation of a hydration front in a dry rock is only possible  
400 if the metamorphic reaction induces porosity, otherwise the fluid could not flow and no hydration

401 would occur. In our hydro-chemical model without solid deformation, porosity is constrained by rock  
 402 properties, which implies that changes in thermodynamic properties caused by a metamorphic reaction  
 403 are enough to predict whether it generates porosity. Since we consider that solid mass is only stored  
 404 in the solid phase and not in the fluid phase, equation (9) can be used to describe the transformation  
 405 of a protolith (rock a) with solid density  $\rho_s^a$ , mass fraction of volatile components in the solid  $X_{\text{H}_2\text{O}}^a$   
 406 and no porosity, into a transformed rock b with  $\rho_s^b$ ,  $X_{\text{H}_2\text{O}}^b$  and reaction-induced porosity  $\phi_r$ :

$$\rho_T^s = \rho_s^a(1 - X_{\text{H}_2\text{O}}^a) = \rho_s^b(1 - X_{\text{H}_2\text{O}}^b)(1 - \phi_r). \quad (16)$$

407 Introducing the notation for the mass fraction of solid components in the solid  $X_s = 1 - X_{\text{H}_2\text{O}}$ ,  
 408 equation (16) is used to compute the porosity generated by the reaction from rock a to rock b:

$$\phi_r = 1 - \frac{\rho_s^a X_s^a}{\rho_s^b X_s^b}. \quad (17)$$

409 The reaction generates porosity when  $\phi_r$  is greater than 0, which is true only if:

$$\rho_s^b X_s^b > \rho_s^a X_s^a, \quad (18)$$

410 i.e. if the metamorphic reaction increases  $\rho_s X_s$ . Equation (18) can be rearranged as:

$$\frac{X_s^b}{X_s^a} > \frac{1}{\rho_s^b/\rho_s^a}. \quad (19)$$

411 Figure 8 shows the porosity created by the metamorphic reaction that transforms rock a into rock b  
 412 with increasing pressure, excluding any solid deformation. No porosity is created in the white field  
 413 because reactions in this field decrease the amount of solid mass stored in the solid ( $\rho_s^b X_s^b < \rho_s^a X_s^a$ ).  
 414 As expected from equation (19), the boundary between the two regions follows an inverse relationship.  
 415 On figure 8, reactions are displayed on a prograde path: rock a is stable at low pressure and rock b at  
 416 high pressure. Only the prograde reactions in the top right field of the diagram can propagate into a  
 417 rock with no porosity, through a positive pressure pulse. If reactions in the white field (e.g. periclase  
 418  $\rightarrow$  brucite) were instead displayed on a retrograde path (brucite  $\rightarrow$  periclase), they would appear in  
 419 the colour field, meaning that these reverse reactions can propagate a low pressure perturbation into  
 420 a no porosity, high pressure region. This is for instance the case in the study of [Schmalholz et al.](#)  
 421 (2020), who show with numerical model similar to ours that a front of low pressure, dense periclase  
 422 can propagates into a high pressure, low density brucite region.

423 Two situations are hence capable of generating a reaction/pressure front that propagates into a  
 424 zero-porosity region. If the high pressure rock has a higher  $\rho_s X_s$  than the low pressure rock (colour  
 425 field in figure 8), a sharp high pressure, high density front can propagate into a low pressure, low-  
 426 density rock. Conversely, if the high pressure rock has a lower  $\rho_s X_s$  than the low pressure rock (white  
 427 field), a sharp low pressure, high density front can propagate into a high pressure, low-density rock.  
 428 Such situations were also discussed by [Malvoisin et al.](#) (2015), although they described these systems

429 in terms of negative and positive Clapeyron slopes instead of, respectively, increase and decrease of  
430  $\rho_s X_s$ . They showed that a sharp pressure/porosity front propagates only for negative Clapeyron slope  
431 and positive pressure pulses, and for positive Clapeyron slopes and negative pressure pulses.

#### 432 4.3.2. Applications

433 Even if  $X_s$  and  $\rho_s$  changes contribute in exactly the same manner to reaction-induced porosity, in  
434 natural metamorphic reaction changes in density are generally much greater than changes in  $X_s$ . For  
435 instance the data of [Centrella \(2019\)](#) shows that eclogitisation of granulite increases density by 7%,  
436 but decreases  $X_s$  by only 0.5%. Therefore, even if in a general case an increase in  $\rho_s X_s$  is required,  
437 in most natural cases an increase in  $\rho_s$  (a decrease in solid volume) should be sufficient to explain  
438 reaction-induced porosity in a high pressure front.

439 Consistent with our work, [Putnis and John \(2010\)](#) showed that increased porosity enhances fluid  
440 transport, mineral replacement and propagation of a reaction front. Our results suggest that in  
441 purely rigid, non-deforming rocks, only reactions which decrease the solid volume (i.e. increase solid  
442 density) would be able to efficiently propagate through generation of porosity. [Hövelmann et al. \(2012\)](#)  
443 experimentally showed that carbonation reaction of peridotite fills the pre-existing porosity because  
444 the reaction increases the rock volume, which ultimately self-limits the reaction.

445 However there is evidence that other mechanisms can generate porosity at the reactive interface,  
446 even when the reaction lowers the solid density. [Jamtveit et al. \(2008\)](#) showed that stress induced  
447 by volume increase can generate extensive (micro-)fracturing, which creates a connected network that  
448 allows fluid transport. In a locality nearby Holsnøy, [Moore et al. \(2020\)](#) described sharp amphiboliti-  
449 sation fronts into granulite, with enhanced permeability/porosity in the less dense amphibolite. They  
450 showed evidence of abundant mass redistribution, which our model does not account for, resulting in  
451 an overall mass loss during fluid rock interaction. This can explain generation of porosity even though  
452 amphibolitisation increases the rock volume.

453 Shear zones in Monte Mucrone (Italian Alps) studied by [Früh-Green \(1994\)](#) do not show evidence  
454 of widening. Their chemical analysis shows that in the meta-diorite protolith and the eclogite shear  
455 zone,  $X_s$  and  $\rho_s$  are almost identical ( $X_s^2/X_s^1 = 0.994$ ,  $\rho_s^2/\rho_s^1 = 1.007$ ). This indicates that the  
456 reaction, close to [1, 1] coordinates on figure 8, generate neither porosity nor fracturing induced by  
457 volume increase, and thus no hydration front that propagates into the protolith.

458 In general, retrogression of high grade rocks replaces dense and anhydrous minerals (garnet, py-  
459roxene) by low density and/or hydrous minerals (amphiboles, mica, chlorite, plagioclase). Low grade  
460 rocks therefore generally have a lower rock density and a higher water content, i.e. a lower  $\rho_X$  than  
461 higher grade rocks (e.g. [Hacker et al., 2003](#)). Our work implies that retrogression must be supported  
462 by some other porosity-generating mechanisms, such as mass transport or brittle deformation caused  
463 by volume increase. We suggest that in absence of such mechanisms, retrogression may be limited

464 and high grade rocks may be preserved during their exhumation.

#### 465 4.4. Hydraulic diffusion

##### 466 4.4.1. Hydraulic diffusion during eclogitisation

467 [Kaatz et al. \(2022\)](#) showed that H<sub>2</sub>O content variations across eclogite shear zones are affected  
 468 by solid state diffusion of hydrogen, but their work does not show if solid state diffusion plays a  
 469 significant role in the widening of the eclogitisation zones. [Putnis and John \(2010\)](#) showed that  
 470 mineral replacement during metamorphic reactions occur by fluid transport associated with reaction-  
 471 induced porosity, rather than solid state diffusion. For this reason, our model includes Darcy flow  
 472 instead of solid state fluid diffusion of hydrogen and oxygen atoms into the crystalline lattice, like  
 473 some papers have done in the past to model eclogitisation ([Bras et al., 2021](#)).

474 Nevertheless, figure 6 shows that the front propagation distance is proportional to the square root  
 475 of time, implying that fluid transport by reactive Darcy flow is a diffusive transport, termed hydraulic  
 476 diffusion (e.g. [Connolly, 1997](#)). By analogy with a general diffusion law, a hydraulic diffusivity  
 477 constant  $D_H$  can be used to relate the characteristic length of diffusion  $L$  and the characteristic time  
 478 of diffusion  $\tau$ :

$$L = \sqrt{D_H \tau}. \quad (20)$$

479 Using this relationship, a hydraulic diffusivity can be determined for the granulite-eclogite case when a  
 480 fluid pulse is active, using our empirically derived equation (15) with  $\alpha = 4.47 \times 10^{-9} \text{ m} \cdot \text{Pa}^{-0.5} \cdot \text{s}^{-0.5}$ .  
 481 The characteristic length of diffusion is the distance travelled by the hydration front, in our case  
 482  $L = \frac{w-w_0}{2}$ . The characteristic time is the duration of the fluid pressure pulse ( $\tau = t_p$ ). Reformulating  
 483 equation (15) with these parameters yields

$$L = \sqrt{\left(\frac{4.47 \times 10^{-9}}{2}\right)^2 (\Delta P_f - \Delta P_r) \tau}. \quad (21)$$

484 The ratio of permeability coefficient to fluid viscosity  $k_0/\eta_f$  has a first order control on the timing of  
 485 the hydration process. Tests for varying simulation times  $t$  and  $k_0/\eta_f$  indeed show that the results of  
 486 our model are not only a function of  $t$ , but rather  $t \times k_0/\eta_f$  (supplementary figure S4). As an example,  
 487 in a simulation where  $k_0/\eta_f$  is 10 times greater than our reference model, the results displayed on  
 488 figure 5 appear after 10 days instead of 100. We can therefore introduce the  $k_0/\eta_f$  ratio in equation  
 489 (21):

$$L = \sqrt{0.05 \frac{k_0}{\eta_f} (\Delta P_f - \Delta P_r) \tau}. \quad (22)$$

490 Substituting equation (22) into equation (20) yields an equation for  $D_H$ :

$$D_H \approx 0.05 \frac{k_0}{\eta_f} (\Delta P_f - \Delta P_r). \quad (23)$$

491 In our reference model where  $k_0/\eta_f = 10^{-16} \text{ m}^3 \cdot \text{s} \cdot \text{kg}^{-1}$  and  $\Delta P_f - \Delta P_r = 5.6 \times 10^8 \text{ Pa}$ , equation  
 492 23 yields a hydraulic diffusivity in the order of  $10^{-9} \text{ m}^2 \cdot \text{s}^{-1}$ . This is orders of magnitude higher than

493 solid state diffusivity of hydrogen or oxygen atoms in rock forming minerals, that range from  $10^{-11}$   
 494 to  $10^{-18} \text{ m}^2\cdot\text{s}^{-1}$ , in most cases  $10^{-13}$  to  $10^{-15} \text{ m}^2\cdot\text{s}^{-1}$ , at temperatures relevant for this study (Ingrin  
 495 and Blanchard, 2006; Farver, 2010). Our work therefore shows that the propagation of a reacting  
 496 hydration front by porous flow is a diffusive process that is potentially orders of magnitudes faster  
 497 than solid-state diffusion.

#### 498 4.4.2. General hydraulic diffusivity

499 Equation (23) applies to our granulite to eclogite model. In a general case, the width of the  
 500 hydration zone must also be controlled by the reaction-induced porosity  $\phi_r$ . Tests with varying  $\phi_r$   
 501 (figure 9) reveal that the width of the hydration zone is proportional to  $\phi_r$ , therefore:

$$L = \frac{\phi_r}{\phi_r^e} \sqrt{0.05 \frac{k}{\eta_f} (\Delta P_f - \Delta P_r) \tau}, \quad (24)$$

502 where  $\phi_r^e = 0.1625$  is the porosity induced by eclogitisation of granulite (figure 2C). Additionally,  
 503 since  $\phi_r$  is thermodynamically constrained by equation (17), we can substitute  $\phi_r$  for properties of the  
 504 protolith ( $\rho_s^a, X_s^a$ ) and of the transformed rock ( $\rho_s^b, X_s^b$ ):

$$L = \left(1 - \frac{\rho_s^a X_s^a}{\rho_s^b X_s^b}\right) \sqrt{1.9 \frac{k}{\eta_f} (\Delta P_f - \Delta P_r) \tau}. \quad (25)$$

505 Substituting equation (25) into equation (20) yields a general equation for the hydraulic diffusivity  
 506  $D_H$  that can be estimated for any reaction:

$$D_H = 1.9 \left(1 - \frac{\rho_s^a X_s^a}{\rho_s^b X_s^b}\right)^2 \frac{k}{\eta_f} (\Delta P_f - \Delta P_r). \quad (26)$$

## 507 5. Conclusion

508 We present a physically consistent hydro-chemical model of eclogitisation of impermeable granulite,  
 509 that can explain the widening of eclogite shear zones and eclogite fingers in Holsnøy. Our work shows  
 510 that hydration of a zero-porosity rock is possible only if density changes controlled by pressure changes  
 511 are taken into account: solid compressibility or metamorphic reactions. Fluid overpressure will drive  
 512 a reactive, porous fluid flow into an impermeable rock, as long as the high-pressure rock is denser than  
 513 its impermeable protolith. Our model can simulate any reaction inducing density changes, as well as  
 514 metastable scenarios.

515 In our model without mechanics, overpressure is only caused by a fluid pressure pulse, and future  
 516 work should take solid deformation into account. Deformation could enhance the overpressure and  
 517 subsequent reactive fluid flow, as long as the high pressure rock is mechanically weaker than its  
 518 protolith.

519 Local pressure perturbations of several kbars, caused by fluid pulses or weakening-induced over-  
 520 pressure, can induce reaction front propagation at a rate of tens of centimetres per year, as long as

521 the pressure perturbation is present. The rate at which the reaction front propagates is typical of a  
522 diffusive process. This complex and coupled hydro-chemical process can hence be approximated by a  
523 diffusion law, where the diffusivity can be calculated for any reaction from the pressure gradients and  
524 rock properties.

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## 532 7. Data Availability

533 All numerical results have been generated with a self-developed MATLAB code, which is available  
534 on the platform Zenodo under: <https://doi.org/10.5281/zenodo.7966767>.

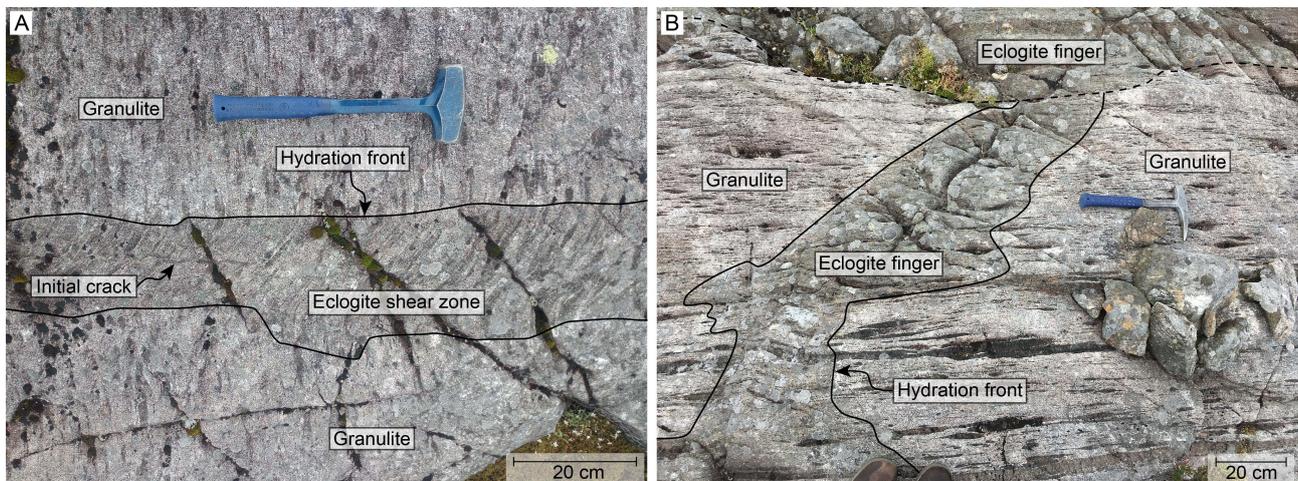


Figure 1: Photographs from Holsnøy (western Norway), showing a halo of eclogitisation overprinting the granulite. (A) Eclogite shear zone (eclogitisation associated with deformation); 60°35'11"N, 5°07' 34"E. The shear zone develops around a narrow central crack. (B) Eclogite finger (no deformation); 60°35'27"N, 5°00'44"E.

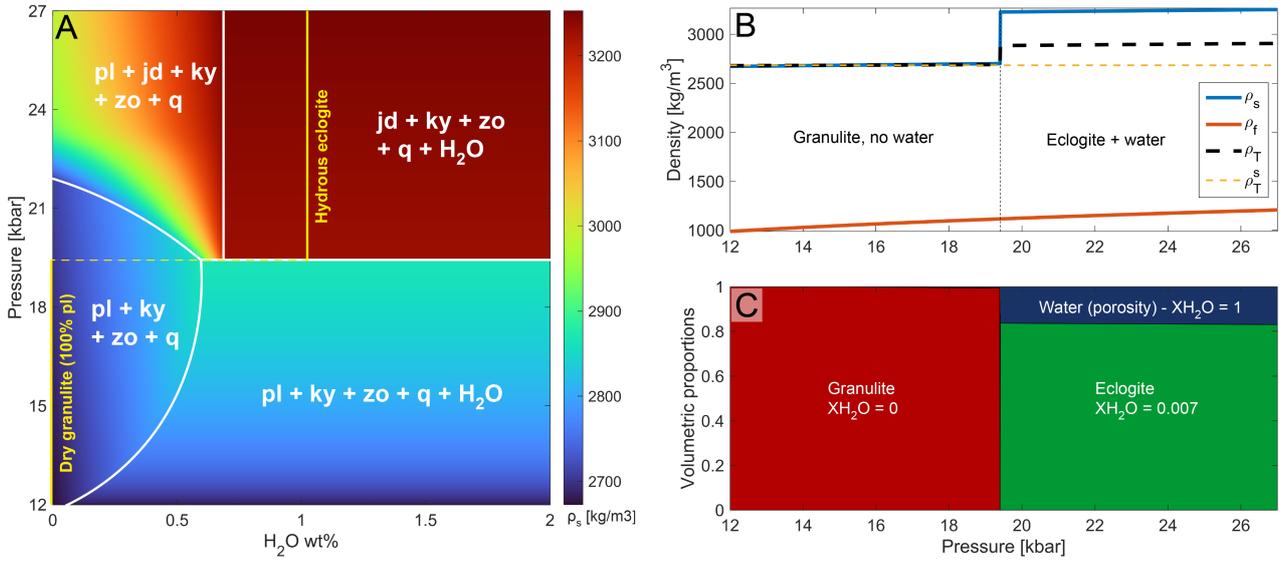


Figure 2: Thermodynamic data used in the model. (A) *Perple\_X* pseudosection used to mimic the stable mineralogical assemblages resulting from destabilisation of anorthosite, as a function of water content and pressure. The yellow line shows the  $H_2O$  wt% conditions at which we retrieve solid density data: dry conditions for granulite at low pressure, and 1 wt% (saturated in  $H_2O$ ) for eclogite at high pressure. Abbreviations after Holland and Powell (1998): pl: plagioclase, ky: kyanite, zo: zoisite, q: quartz, jd: jadeite. (B) Corresponding 1D look-up tables for solid ( $\rho_s$ ) and fluid ( $\rho_f$ ) density as a function of pressure, and total density ( $\rho_T$ ) look-up table calculated from equation (12). The total solid mass ( $\rho_T^s$ ) is a constant value of  $2685 \text{ kg} \cdot \text{m}^{-3}$ . (C) Phase composition of the system. At 15 kbar, granulite has zero porosity.

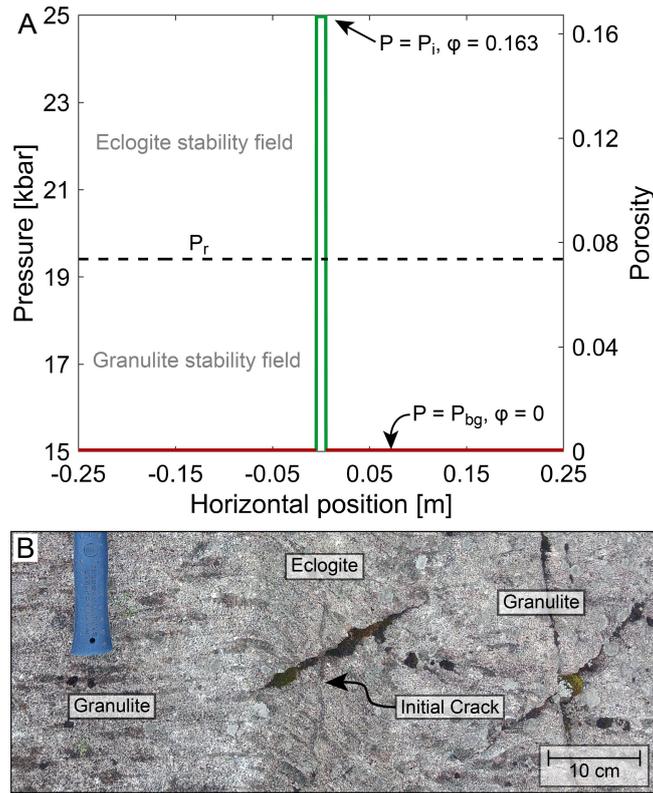


Figure 3: Model configuration. (A) Initial profiles of pressure and porosity. All the domain is made up of granulite with no porosity initially placed at 15 kbar ( $P_{bg}$ , red line), except the central narrow crack that is submitted to a high fluid pressure pulse. This initial pressure  $P_i$ , higher than the reaction pressure  $P_r$  (dotted line), is enough to trigger eclogitisation and generate porosity (green line). (B) Detail of the shear zone from photograph 1A (top view) with features comparable our configuration: a central narrow crack cutting through the granulite, from which the eclogitisation front propagates.

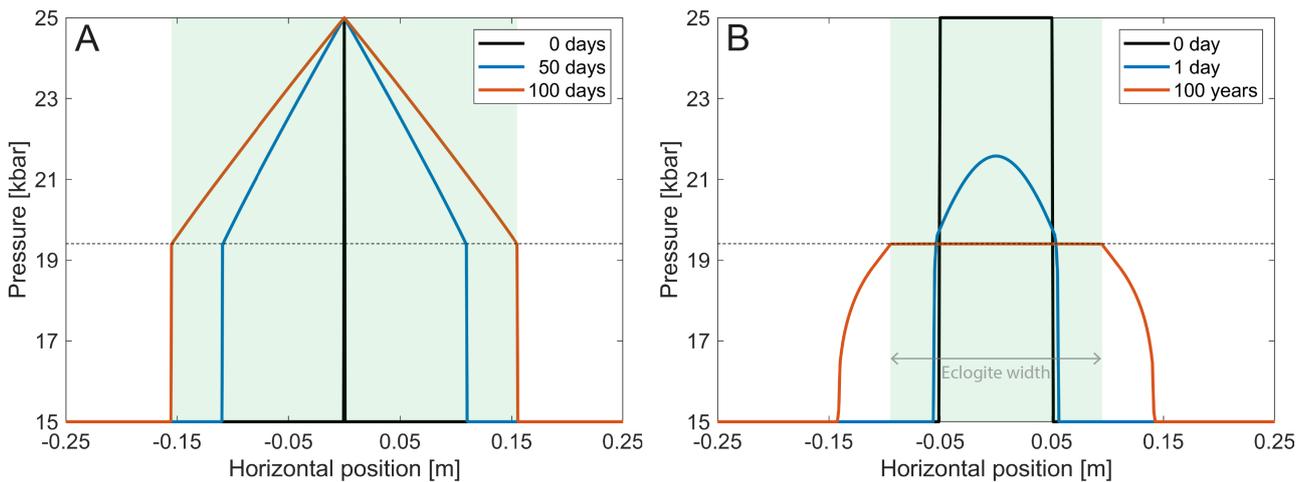


Figure 4: Introductory examples to the hydro-chemical model. Light green areas represent the eclogite width after 100 years. A: Open system. In the centre of the domain, the pressure is always fixed at 25 kbar as a boundary condition. This corresponds to a fluid pulse that is continuously supplied by the central crack. B: Closed system. There is no fluid supply except the initial step-like pressure profile that decays rapidly, and then slowly diffuses into the granulite. Notice the difference in time scales between A and B.

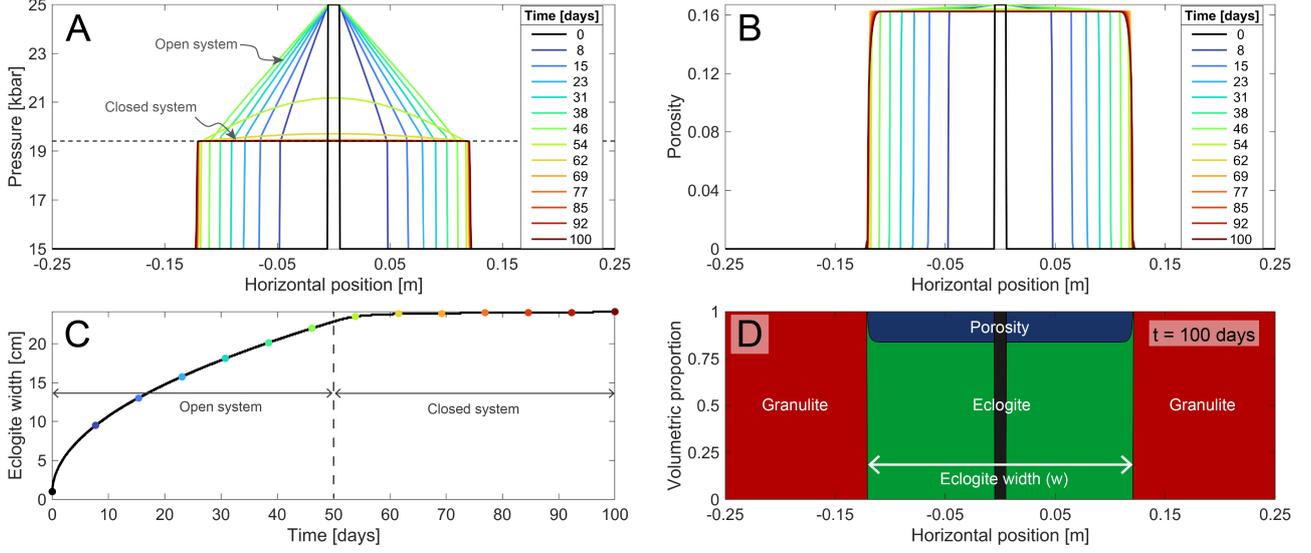


Figure 5: Result of a simulation after 100 days. The fluid pulse lasts 50 days with a behaviour similar to the model of figure 4A, and then the fluid supply ends, rapidly shifting to the behaviour shown in figure 4B. (A) Evolution of the pressure and (B) evolution of the porosity with time. (C) Evolution of the width of the eclogitisation zone over time. The colour dots correspond to profiles in A and B. (D) Phase composition at the end of the simulation (100 days). The narrow black region shows the region where the fluid pulse is applied, of width  $w_0 = 1$  cm. It is already eclogitised at the start of the simulation.

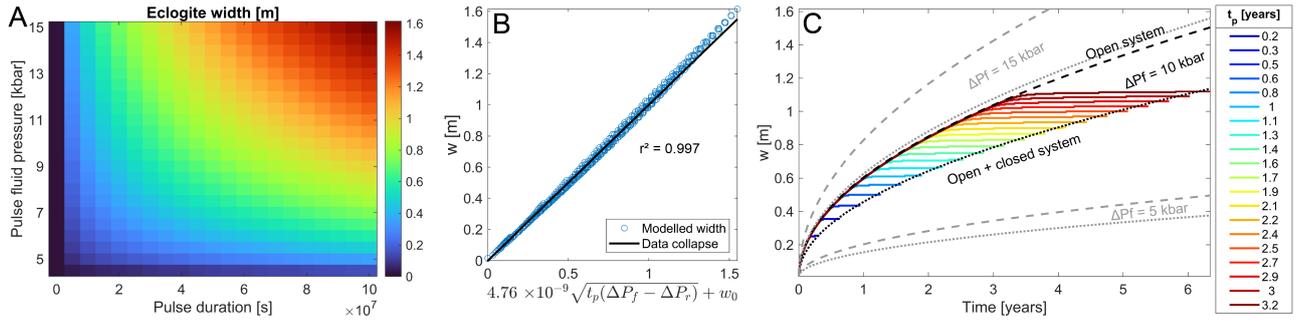


Figure 6: Parametric study. (A) Eclogitisation zone width  $w$  as a function of the duration of the fluid pulse  $t_p$  and of the fluid pressure of the pulse  $\Delta P_f$ . The duration of the simulation is  $2 \times t_p$  for all simulations. (B) Data collapse for the eclogite width at the end of the simulation (open + closed system). See text for details and explanation of the parameters. (C) Time evolution of the eclogite width  $w$  for different  $t_p$  and  $\Delta P_f$ . The colour lines show the modelled time evolution for different  $t_p$  and  $\Delta P_f = 10$  kbar. The dotted lines show the eclogite width time evolution predicted in 6B and equation (15) with  $\alpha = 4.76 \times 10^{-9}$ , for the indicated  $\Delta P_f$ . The dashed lines show the eclogite width time evolution at the end of the open system predicted by equation (15) with  $\alpha = 4.47 \times 10^{-9}$ . The predictive lines aligns closely to the modelled colour lines. Colour lines for the other  $\Delta P_f$  are not plotted to reduce cluttering on the diagram, but they would align with the dotted and dashed lines in a similar fashion.

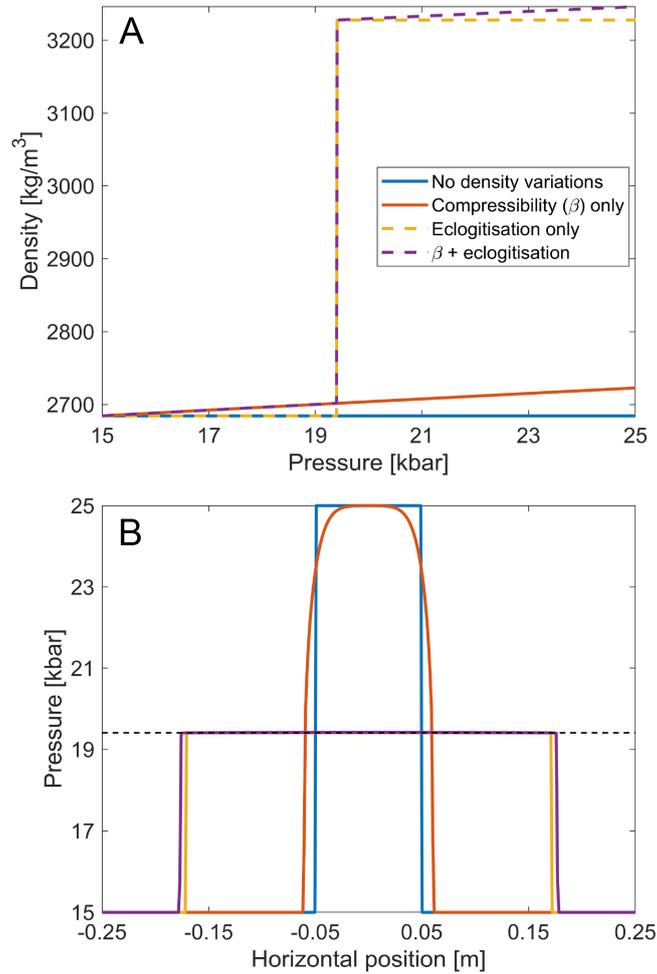


Figure 7: Effect of density changes on the widening of an eclogitisation zone. (A) Solid density as a function of pressure for 4 different scenarios. Blue: the solid is incompressible and non-reactive; its density never increases. Orange: the solid is non-reactive; its density only increases due to compressibility. Yellow: the solid is incompressible, its density only increases due to transformation of granulite into eclogite. Purple: the solid density increases due to compressibility and transformation of granulite into eclogite, as used throughout this study. (B) Pressure profiles after 100 days. The blue profile is identical to the initial pressure profile. Thickness of the initial perturbation (compared to figures 3, 5) was increased for clearer visualisation.

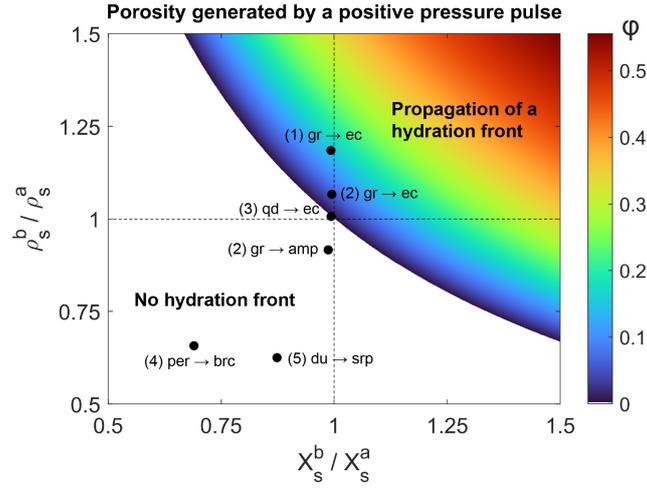


Figure 8: Porosity generated by the transformation of a low pressure rock a (with solid density  $\rho_s^a$  and proportion of non volatile component  $X_s^a$ ) into a high pressure rock b (with  $\rho_s^b$  and  $X_s^b$ ). Porosity is created only when the reaction increases the solid mass stored in the solid, i.e.  $X_s^b \rho_s^b > X_s^a \rho_s^a$ . Reactions in the white field do not generate porosity. Abbreviations: gr = granulite; ec = eclogite; amp = amphibolite; qd = quartz-diorite; per = periclase; brc = brucite; du = dunite; srp = serpentinite; atg = antigorite. Source of  $\rho_s$  and  $X_s$  data: (1) This work, (2) Centrella (2019); (3) Früh-Green (1994); (4) Schmalholz et al. (2020); (5) Malvoisin et al. (2021).

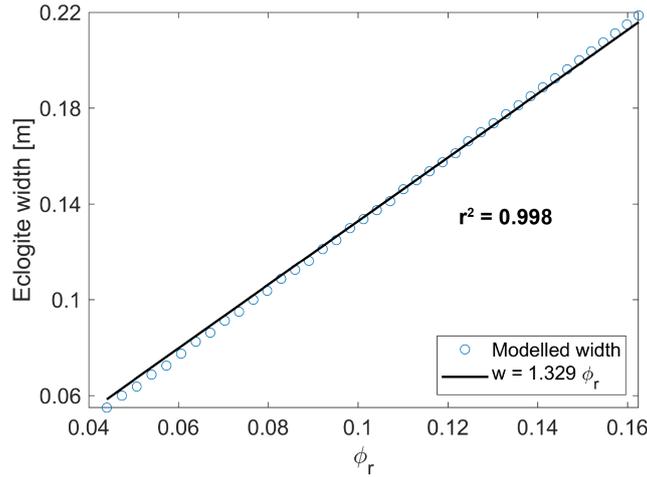


Figure 9: Linear fit of hydration zone width as a function of reaction-induced porosity  $\phi_r$ . The width of the hydration zone increases linearly with  $\phi_r$ . The parameters of the simulations are the same as in the reference model, except  $w_0 = 0$  m. The simulation ends when the system closes, after 50 days.  $\phi_r$  is set by modifying the densities of granulite and eclogite in the look-up table.

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