

Consequences of progressive eclogitization on crustal exhumation, a mechanical study

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SUMMARY

The very low water content of the granulitic unit of Holsnøy, in the Bergen Arcs, Norway, caused its partial metastable preservation throughout Caledonian burial and exhumation, leading to the observed mixture of completely eclogitized and uneclogitized rocks. The eclogitization of the granulite-facies protolith led to a density increase $\sim 10\%$, but also to a large rheological weakening that resulted in the localization of all deep ductile deformation in the eclogite fraction.

We address the consequences on subduction dynamics of the concomitant evolution of large-scale density and rheology during progressive eclogitization of crust by comparing the behaviour of buried crust with various properties in the channel-flow model. Large-scale buoyancy and density are integrated in a single adimensional parameter, the ‘exhumation number’ α , which describes the capacity of the crust to exhume.

For a crust whose eclogitization causes a very large viscosity decrease (‘burial resistant’ end-member), α goes through a maximum in the partially eclogitized zone, that is, there is a given proportion of eclogite for which the crust is strongly weakened but still significantly lighter than surrounding mantle. For this partially eclogitized crust with α max, the maximum possible downward flow is very low, this zone acts as a bottleneck and all incoming crust in excess is forced backwards and starts exhuming. This return flow zone is bounded downwards in the partially eclogitized zone by α max, while its upper limit propagates upwards at a constant rate.

The curve $\alpha(\text{eclogite fraction})$ controls the maximum proportion of eclogite of the crust that can be exhumed. We also demonstrate that large exhumation rates reached by certain UHP units are incompatible with any stationary flow regime, whatever the nature of subducted crust. The additional presence of a weak layer on top of the channel favours burial for narrow channels and exhumation for large ones, but does not qualitatively change the dynamics of the subduction channel.

This study shows that deep crustal circulation in subduction zones and exhumation from large depths are controlled by the evolution of large-scale crustal properties with progressive eclogitization.

Key words: crustal deformation, eclogitization, exhumation, metamorphism, rheology, subduction zone.

1 INTRODUCTION

The presence in former or present collision zones of coesite (Chopin 1984; Smith 1984) or microdiamonds (Dobrzhenetskaya *et al.* 1995) in crustal rocks now at the surface evidences their former burial to large depths (>100 km) and subsequent exhumation. The geodynamical context that allows to reach such depths is subduction: crust attached to the dense subducting lithosphere is dragged down into the mantle, and at some point starts its way up. The second part of

this evolution is problematic: which mechanism does trigger movement reversal, what are the conditions of exhumation? The main purpose of this study is to explore the consequences of metamorphic reactions and phase changes on this process.

Large-scale analogue (Chemenda *et al.* 1995, 1996) as well as numerical modelling (Burov *et al.* 2001) of subduction zones highlight the major forces at stake: vertical movements at large depths of subducted crust are prescribed by the balance between buoyancy of buried crust and downpulling drag by lithospheric mantle. In the

second type of model, exhumation of HP rocks is driven by crustal convective instability, while in the first type it results from the formation of one crustal-scale thrust. In both cases exhumation starts when buoyancy overcomes downward pull. Additionally, in corner-flow models (Shreve & Cloos 1986), the fixed wedge shape of the subduction channel induces a forced return flow of crust independent of density contrasts.

The effects of metamorphic transformations on the behaviour of the subduction channel have seldom been explored although variations in material properties and particularly in density and rheology are likely to bring about large modifications of geometry and dynamics of subduction zones. Eclogite-facies metamorphic reactions affecting the crust as it is buried cause a well-defined density increase (e.g. Austrheim 1987; Bousquet *et al.* 1997) and rheological changes that are still not well constrained. The ongoing controversy over the dominant deformation of omphacite, one of the main eclogite-facies minerals, either dislocation creep (Bascou *et al.* 2002) or diffusion creep (Mauler *et al.* 2001), illustrates the lack of data on mechanical behaviour of high-pressure minerals. As a result, when accounted for, crustal eclogitization reactions are only integrated in numerical models through the density variation (Burov *et al.* 2001; Doin & Henry 2001).

Another complexity arising when handling metamorphic reactions is that they are neither instantaneous (with respect to P–T changes) nor spatially homogeneous, as evidenced by the preservation of successive parageneses in a single rock. A large fraction of the anhydrous Precambrian granulitic unit of the Bergen Arc underwent Caledonian burial and exhumation without any sign of reworking (Austrheim 1987; Austrheim & Griffin 1985), while the rest was in average strongly eclogitized. As a result, the kilometre-scale rheology of partially eclogitized units, which is the property to be considered when studying large-scale movements, depends not only on the small-scale rheology of un- and eclogitized materials, but also on the geometries of un- and transformed volumes.

The first objective of this paper is to qualitatively estimate eclogite-facies rocks rheology at all scales, from completely eclogitized samples up to kilometre-sized partially transformed units, using both existing literature data and our own field observations. The second objective is to analyse the consequences of such partial metamorphic transformations on deep circulation in subduction channels. The framework of this mechanical study is the channel-flow model (England & Holland 1979; Shreve & Cloos 1986) that enables to understand the interplay between buoyancy and viscosity variations in the flow of progressively eclogitizing rocks.

2 GEOLOGICAL SETTINGS

2.1 The Lindås Nappe, a Caledonian high-pressure unit

The Lindås Nappe belongs to the Bergen Arcs, a series of arcuate thrust sheets centred around Bergen, Norway. The protolith is considered as a portion of lower continental crust and contains abundant mafic material (Kolderup & Kolderup 1940).

The northwestern part of Holsnøy (Fig. 1) belongs to The Lindås Nappe and is constituted of gabbroic to anorthositic rocks. The geology and metamorphic history of the area is well constrained by numerous prior studies (Austrheim 1987, 1990a,b, 1994; Austrheim & Engvik 1995; Austrheim & Griffin 1985; Austrheim & Mørk 1988; Boundy *et al.* 1992). The protolith was pervasively recrystallized and deformed under granulite-facies conditions during the Grenvillian orogenesis (~900 Ma Austrheim & Griffin 1985; Bingen *et al.*

2001; Cohen *et al.* 1988). It was then involved in the Caledonian orogenesis and experienced successive eclogite-, amphibolite- and greenschist-facies metamorphism around 400–460 Ma (Austrheim 1990b; Bingen *et al.* 2004, 1998; Boundy *et al.* 1997b). Maximum burial conditions, recorded by the eclogite-facies metamorphism, were estimated by thermobarometric studies as ~15–20 kbars and 700°–800°C (Austrheim & Griffin 1985; Boundy *et al.* 1992; Jamtveit *et al.* 1990; Matthey *et al.* 1994; Raimbourg *et al.* accepted). Due to the very low water content of the granulite-facies protolith, only the fraction of it that was provided external fluids experienced Caledonian metamorphic reactions; as a result the transformation of northwestern Holsnøy is highly heterogeneous and a large proportion of the granulite-facies unit was metastably preserved throughout Caledonian history (Austrheim 1987; Austrheim & Engvik 1995; Austrheim & Griffin 1985).

2.2 Role of eclogitization in Holsnøy exhumation

The age of eclogite-facies metamorphism on Holsnøy (460–420 Ma), albeit controversial, is by large older than the onset of Caledonian large-scale extension driven by plate divergence (~400–405 Ma Fossen 1992, 2000; Roberts 2003). Both burial and exhumation of Holsnøy occurred, therefore, during convergence and collision. The general structure of the Caledonian orogen is interpreted as resulting from the subduction of Baltica under Laurentia (Griffin *et al.* 1985; Krogh 1977). Continental subduction is also the only widely accepted framework where crustal rocks reach depths >100 km, like coesite-bearing units in the nearby Western Gneiss Region (WGR) (Carswell *et al.* 2001; Smith 1984; Wain 1997). Within the geometrical framework of this subduction, we interpreted the eclogite shear zones studied on Holsnøy as large thrusts that enabled the mechanical decoupling within the subducting slab between dense mantle and light crustal slivers (Fig. 1b) that started their buoyant ascent from then on (Raimbourg *et al.* 2005). Eclogitization reactions and the consequent viscosity drop are, therefore, responsible for the initiation of exhumation.

3 RHEOLOGICAL CHANGES ASSOCIATED TO ECOLOGITIZATION

3.1 Sample scale rheology: experimental and natural data

The Holsnøy eclogite paragenesis is composed of omphacite, garnet, kyanite and minor phengite, $\pm$ rutile, $\pm$ quartz, $\pm$ amphibole (Boundy *et al.* 1992), and the volumetric dominant mineral is omphacite. Whereas the metastable granulite-facies fraction was not deformed during Caledonian subduction, most eclogitized zones show pervasive deformation, attesting to the much higher ductility of eclogite than granulite (Austrheim & Griffin 1985). This rheological softening is the most obvious in narrow transition zones between granulite and eclogite, where the partially eclogitized granulitic foliation is bent to merge into eclogitic foliation (Fig. 2).

The unambiguous localization of deformation in omphacite-dominated fraction (eclogite) and respective undeformability of plagioclase-dominated fraction (granulite) do not nevertheless match experimentally derived rheological laws. Diopside is 10 times stronger than anorthite under similar thermodynamic conditions (Dimanov *et al.* 2003), what is confirmed by the very strong behaviour of dry calcic clinopyroxene aggregates (Bystricky & Mackwell 2001). The yet much weaker rheology of natural omphacite seen on Holsnøy is also documented in the Western Alps

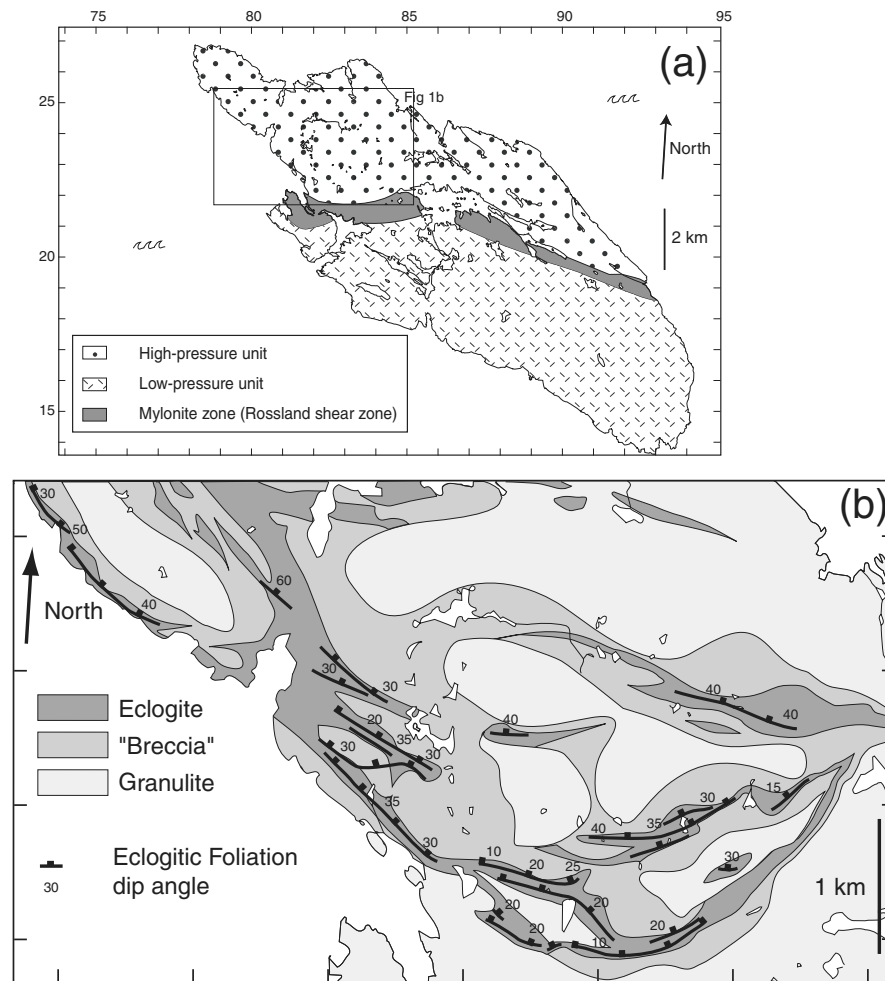


Figure 1. (a) Geological map of Holsnøy modified from (Birtel *et al.* 1998, Boundy *et al.* 1997b, Kühn 2002, Ragnhildsveit & Helliksen 1997). Location of Fig. 1(b). Coordinates correspond to UTM grid, zone 32: 22 = 67 22 000, 81 = 2 81 000. (b) Northwestern Holsnøy from (Austrheim *et al.* 1996; Boundy *et al.* 1992, 1997a; Raimbourg *et al.* 2005), with heterogeneous distribution of eclogitic overprint on granulite. Eclogite, Breccia and Granulite corresponds respectively to degrees of eclogitic metamorphism >80, 40–80, <40 % . Structural data are given for the large-scale eclogitic shear zones.

(Piepenbreier & Stöckhert 2001). The presence of Na in the high-pressure pyroxene may significantly soften it (Jin *et al.* 2001), but these experiments were carried out at very high temperatures (1500 K) and their extrapolation to real deformation conditions may be difficult. Furthermore, beside discrepancies in creep rates of omphacite, even deformation mechanisms are debated: Anisotropic omphacite crystallographic preferred orientations in natural samples, studied by electronic backscatter diffraction, have been interpreted as resulting either from diffusion creep (Mauler *et al.* 2001) or dislocation creep (Bascou *et al.* 2002) or from the contribution of both processes (Bystricky & Mackwell 2001; Godard & Van Roermund 1995). As a result of this variance in natural and experimental data, and given the very strong strain partitioning on Holsnøy, we shall consider in the following that eclogite is much weaker than granulite, without quantifying the rheological difference.

3.2 Large-scale rheology of the mixture eclogite-granulite

The large-scale rheology of the partially eclogitized unit of Holsnøy depends not only on the hand-scale samples rheology, but also on

the geometry of un- and eclogitized volumes. As eclogitization is controlled by fluid input, the physical processes enabling fluid flow at large distances determine the propagation of eclogite zones into the granulite.

As can be seen from the distribution of eclogite domains (Fig. 1b), eclogite zones have an highly anisotropic shape, as they are constituted of 10–50 m wide but kilometre long shear zones containing scarce granulite-facies blocks (Fig. 3). Outside of these shear zone, that is, in most of the unit, the granulite is mostly preserved from eclogitization, which is confined in few narrow bands (Fig. 3). Such a pattern implies that the fluid penetration into the granulite, controlling its eclogitization, was also strongly anisotropic.

The fluid input through pristine granulite was achieved at scales >m by transport into a network of macroscopic fractures (Fig. 4) filled with hydrous eclogite-facies minerals (Austrheim & Griffin 1985; Austrheim & Mørk 1988). The origins of these fractures are not known, they may have been inherited from a tectonic phase prior to eclogitization. Alternatively they may have formed in relation to eclogite-facies metamorphism. In the first hypothesis, the geometry of eclogitized domains passively resulted from the distribution of these inherited fractures, therefore, necessarily

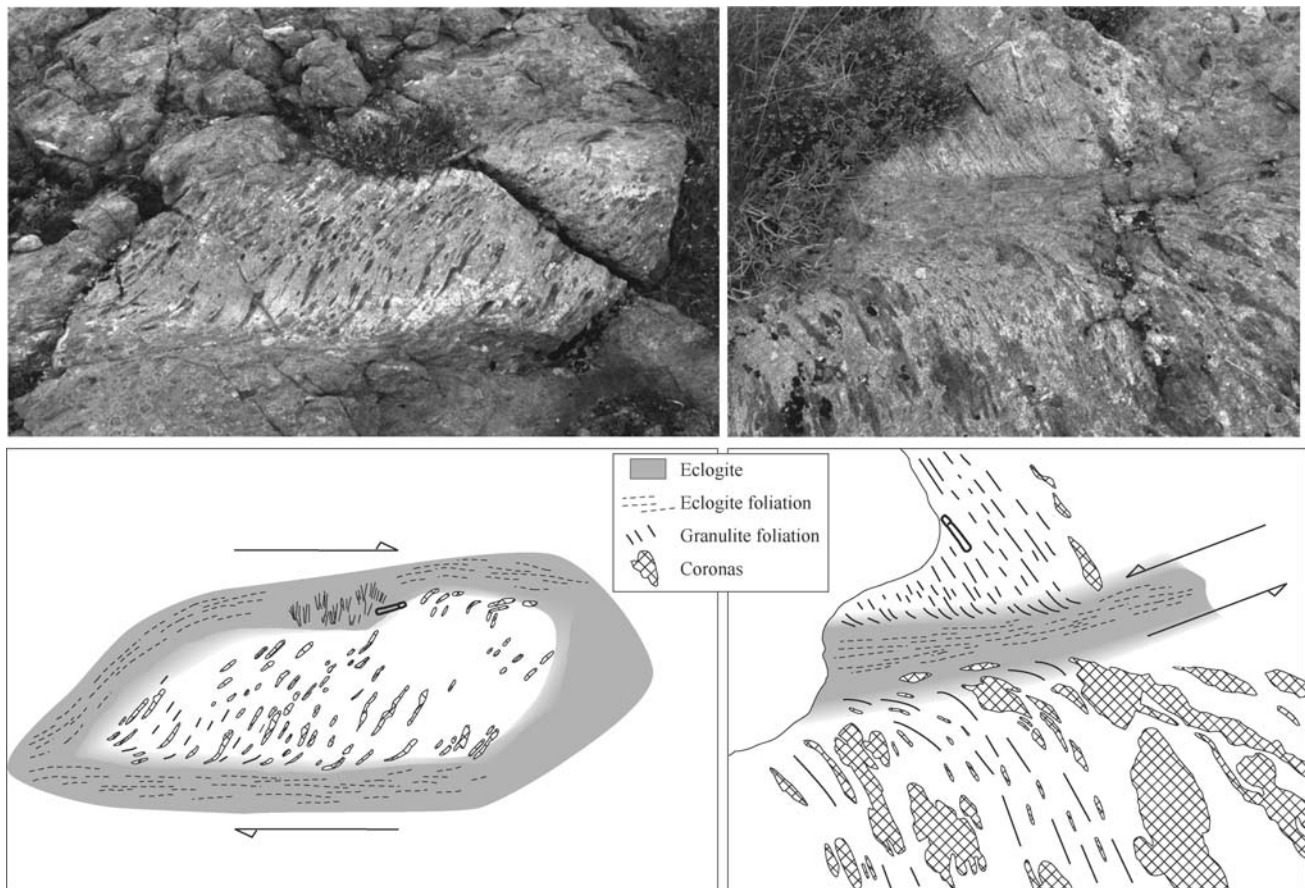


Figure 2. Localization of deep deformation and rheological differences between eclogite and granulite. The granulitic foliation, which is undeformed in the core of the granulite, bends to merge into the eclogitic foliation. This foliation deflection, either on the granulite boudin rim (left picture), or on the border of a narrow eclogitic shear zone (right picture), is used to deduce the sense of shear.

strongly anisotropic itself. In the hypothesis where fracturing occurred during eclogite-facies metamorphism, several processes can account for the anisotropic geometry of the fractured domain that isolates eclogite from granulite domain. The study of ductile deformation showed that consistent large-scale stresses were applied to the whole unit (Raimbourg *et al.* 2005). Such shear stresses on a rheologically contrasted material (weak eclogite+strong granulite) result in stress concentration in particular directions around the eclogite domains, leading to anisotropic fracturing and further anisotropic eclogite domain expansion. In addition, if the eclogite domain is itself slightly anisotropic, the metamorphism-related volume changes within the eclogite domain lead to stress concentration and fracturing with such a geometry that it will tend to increase the anisotropy of eclogite domain. Therefore, there is no real need to invoke inherited fractures to explain the anisotropic geometry of observed eclogite domains, as stress concentrations due both to metamorphism-related rheological contrast and volume change can account for anisotropic fracturing leading to anisotropic eclogite domains.

The precise description of the physical processes enabling the propagation of eclogitized domains would be necessary to quantify the evolution of the average rheology of the whole unit at all stages during eclogitization. In the final state, as the very weak shear zones are continuous throughout the unit, the average rheology of the unit is close to eclogite rheology, even if the proportion of eclogite is still much lower than granulite.

4 EXHUMATION OF PARTIALLY ECLOGITIZED CRUST

The physical processes controlling the propagation of eclogitized zones consequently determine the evolution of large-scale rheology during progressive eclogitization. Parallel to these complex rheological variations, the density of a partially eclogitized unit varies proportionally to the fractions of protolith and eclogite. Eclogitization results in a density increase hindering the capacity to exhumate. On the other hand, in Holsnøy granulites, the large rheological softening induced by eclogitization enabled mechanical decoupling from the subducting lithospheric mantle and subsequent exhumation (Raimbourg *et al.* 2005). The ‘channel flow’ model (England & Holland 1979; Shreve & Cloos 1986) is a powerful analytical framework to study this equivocal effect of eclogitization on exhumation. Its simplicity enables to study the interplay between buoyancy and viscous drag, two major forces controlling burial and exhumation.

4.1 The subduction channel model

4.1.1 General geometry and boundary conditions

The subduction channel model has been defined to analyse the movements of sediments on top of a subducting plate (England & Holland 1979; Shreve & Cloos 1986). We apply here the model not to the

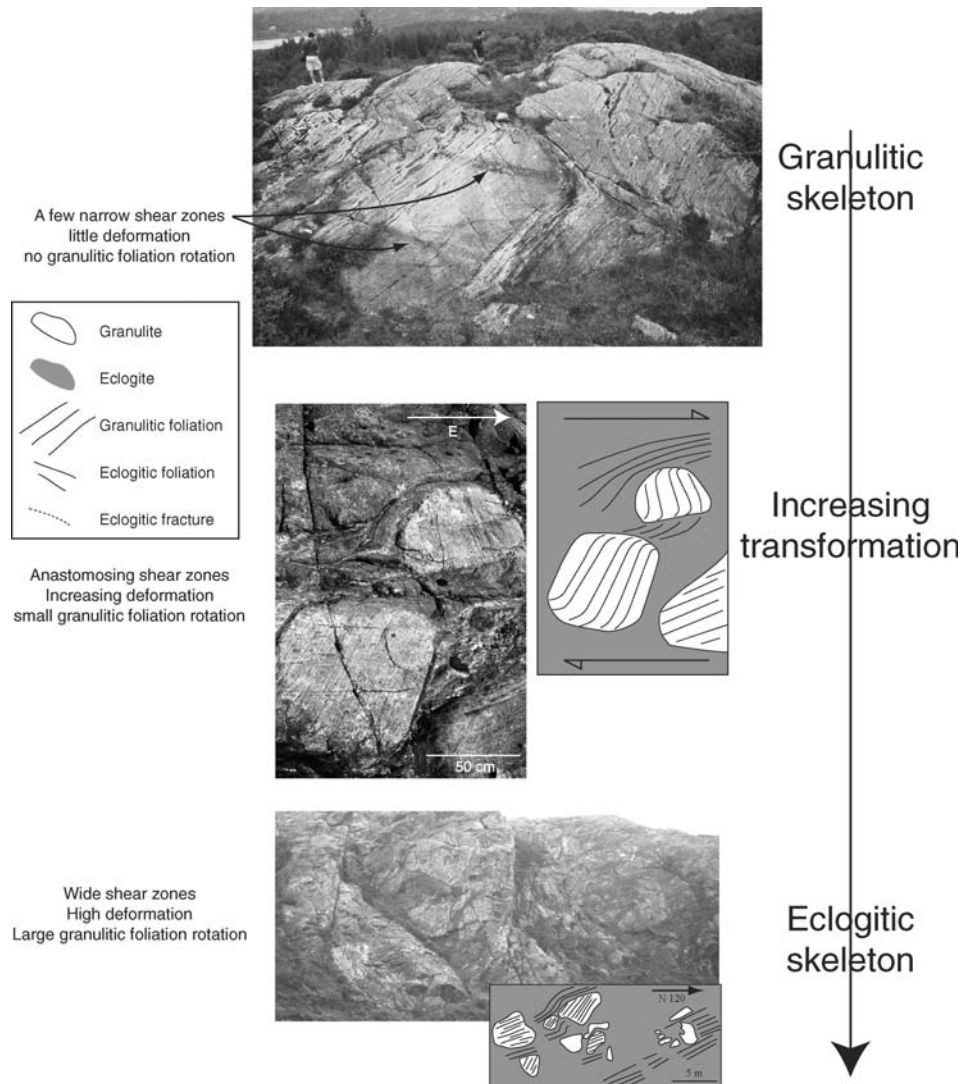


Figure 3. The extent of eclogitization is highly variable throughout the granulitic unit. In little transformed areas, eclogite is present only in scarce and narrow shear zones cutting through a rigid granulitic skeleton. In the most transformed areas, granulite is preserved only in rare boudins embedded in an eclogitic skeleton. The rheology difference between strong granulite and weak eclogite results in the localization of all the deformation in the eclogite fraction. The disruption of the granulitic skeleton with eclogitization progress is marked by the increase in the misorientation of adjacent granulite volumes. In the most transformed zones, the independent rigid rotation of granulite boudins results in the random orientation of their foliations.

sediments but to the crust overlying the subducting slab and we ignore the thin layer of sediments, in a first approach. Accretionary prisms, which can be as deep as 30–40 km (e.g. Bousquet *et al.* 1997; Chi *et al.* 2003; Rubio *et al.* 2000) are not described by the channel flow model, whose upper vertical limit is, therefore, below such structures. Its lower limit is not constrained, as a certain amount of lower crust within the subducting slab is assumed to disappear into the mantle (Jolivet *et al.* 2003; Le Pichon *et al.* 1988).

The bottom boundary of the channel, in the y direction (Fig. 5), is the subducting lithospheric mantle; the roof is either the overriding mantle wedge or the thickened crustal orogen, depending on the depth. The velocity is prescribed as the plate velocity U at the bottom, and null velocity at the top. Such velocity conditions imply that all shear deformation induced by the velocity difference between subducting plate and overriding mantle is accommodated by the crust within the channel. It is justified if the mantle on top of the channel is much stronger than the crust within.

The channel width gradient $\frac{\partial h(x)}{\partial x}$ and the transverse pressure gradient $\frac{\partial P(x,y)}{\partial y}$ are assumed to be small, material properties to be constant in y direction. As a result the velocity $\vec{u}$ is everywhere parallel to $\vec{x}$ —that is, the flow is laminar. The pressure within the channel at a given depth is fixed as being equal to the mantle lithostatic pressure, $P = \rho_{\text{mantle}} gz$, outside of the channel.

4.1.2 'Pressure-' versus 'geometry-' imposed boundary conditions

The pressure profile defined above (P in the channel equal to P out of the channel) is not imposed by the model, it is one possible set of boundary conditions that we have chosen to build up our model (referred to hereafter as 'pressure-imposed' model). Alternative choice has been preferred in Mancktelow (1995), where the geometry of the channel, that is, channel width variations, have been fixed *a priori* ('geometry-imposed'). As a result of the variable width of the

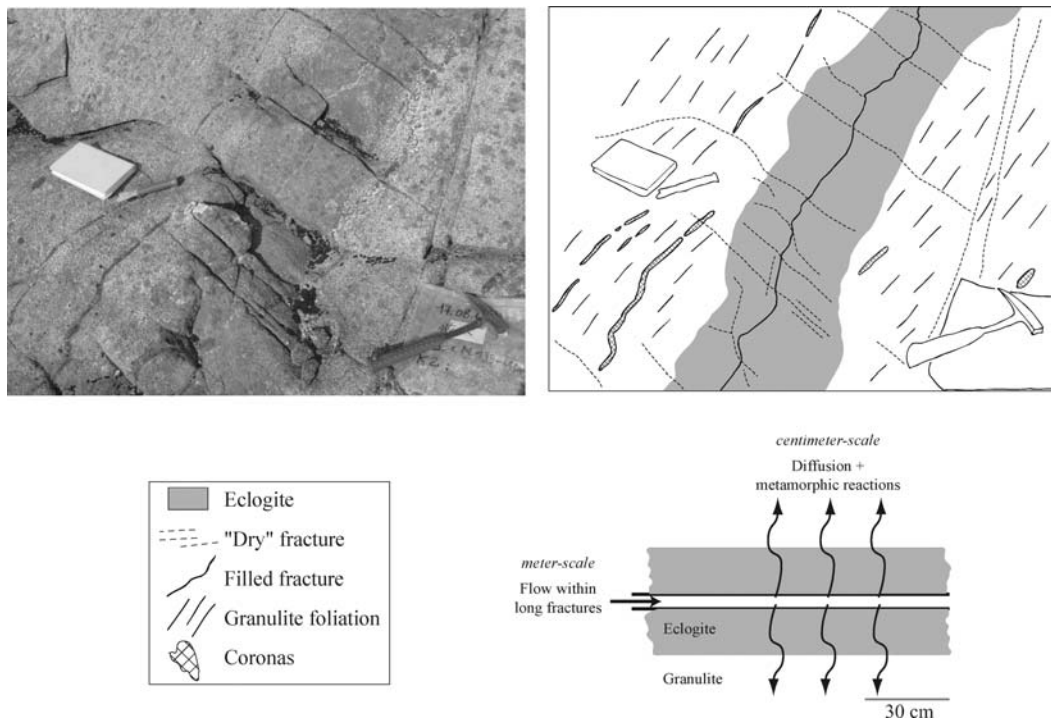


Figure 4. Mechanism for propagation of eclogitized zones into granulite. In the highly anhydrous granulite, fluid input is necessary to trigger eclogitization, and the sharp front between light granulite and dark eclogite coincides with the hydration front. Fluid input is achieved by the combination of channelized flow within metre-scale fractures and subsequent diffusion from the fracture into host rock. Note the presence of 'dry' fractures (as opposed to eclogitic fractures, filled with hydrous eclogitic minerals) crosscutting the eclogite volume and its boundary, whose origin is unclear.

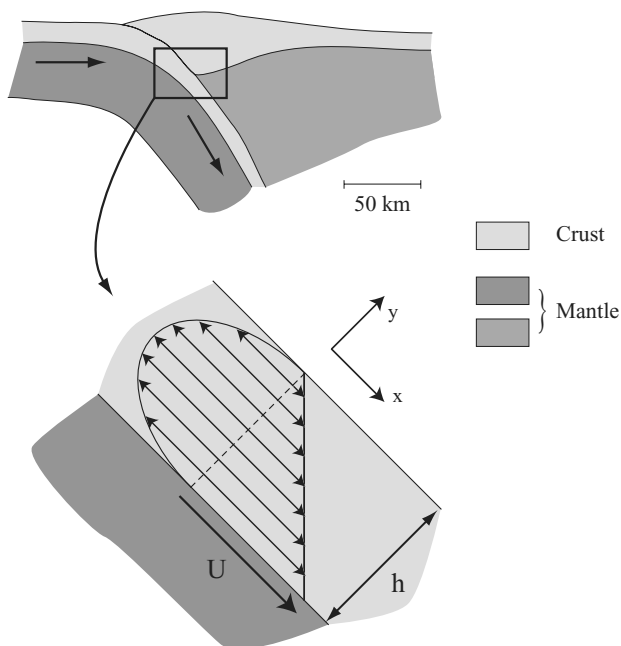


Figure 5. Subduction channel model. The material within the channel is the crust belonging to the subducting slab. The channel is bounded at the bottom by the lithospheric mantle, at the top by the crust or the mantle of the overriding plate, depending on the depth and the subduction zone considered. The bottom velocity is the plate convergence rate and the top velocity is zero. The velocity field within the channel is the sum of a linear field, related to basal dragging, and a parabolic field related to buoyancy and viscosity of buried crust.

channel, large over- and underpressures (difference for a given depth to the mantle lithostatic pressure), of the order of 1 GPa, develop in the material flowing within the channel.

In order to determine the domain of validity of these two approaches, we develop in (Raimbourg and Kimura 2006) a model coupling flow within the channel and deformation of the surrounding mantle triggering channel width variations. The coupling is achieved by an additional constitutive equation relating the deformation of the channel geometry $\frac{\partial h}{\partial t}$ to the amplitude of the non-lithostatic pressure within the channel, with the generic form

$$\frac{\partial h}{\partial t} = L \frac{(P_{\text{channel}} - P_{\text{mantle}})}{\eta_{\text{mantle}}},$$

where L is the length of the channel and η_{mantle} the mantle viscosity. Adding such an equation, $P(x, t)$ or $h(x, t)$ are no longer prescribed but are free to vary.

Without getting into too much detail, the conclusions of this study of coupled crust and mantle deformation are:

(i) We can define two distinct domains of behaviour, depending on whether the deformation of the channel walls is neglectable or not (i.e. 'geometry-imposed' or not). The boundary between these two domains is defined by $\lambda = \frac{h_0^3}{L^3} \frac{\eta_{\text{mantle}}}{\eta_{\text{crust}}} = 1$, where h_0 is the average channel width. For $\frac{h_0}{L} \approx 0.5 \cdot 10^{-1}$ (e.g. if $h_0 = 5$ km and $L = 100$ km), in the range $1 < \frac{\eta_{\text{mantle}}}{\eta_{\text{crust}}} < 10^4$ (i.e. in $\lambda \ll 1$ domain), the deformation of the geometry of the channel cannot be neglected, in contradiction with the assumption underlying 'geometry-imposed' models. In contrast, for much stronger mantle ($\frac{\eta_{\text{mantle}}}{\eta_{\text{crust}}} > 10^4$, that is, $\lambda \gg 1$), the channel can be considered as rigid.

(ii) Within the regime $\lambda \ll 1$, the channel walls deformation inhibits any significant over and underpressures to develop. As a

result, the assumption that everywhere in the channel $P_{\text{channel}} = P_{\text{mantle}} = \rho_{\text{mantle}}gz$, used in ‘pressure-imposed’ models such as the one developed here, corresponds to the regime $\lambda \ll 1$.

(iii) Using the parameters used by Mancktelow (1995) yields $\lambda \ll 1$, in contradiction with the assumption the author made that the channel is rigid! In other words, the settings of his study is such that channel walls deformation cannot be neglected, resulting in much lower overpressures than calculated by the author.

4.1.3 Velocity profiles

The flow within the channel is governed by Stokes equations with neglectable acceleration

$$\frac{\partial \tau_{xy}}{\partial y} = \frac{\partial \phi}{\partial x},$$

where $\phi = P - \rho_{\text{crust}}gz$ is the hydraulic potential, τ_{xy} the shear stress, P the pressure, ρ_{crust} the crustal density, g the vertical component of acceleration within the gravity field and z the depth. As discussed above, the pressure at any depth in the channel is approximated as being equal to the lithostatic pressure outside of the channel $P = \rho_{\text{mantle}}gz$. As a result, the hydraulic potential gradient is proportional to the density difference between crust and mantle:

$$\begin{aligned} \frac{\partial \phi}{\partial x} &= (\rho_{\text{mantle}} - \rho_{\text{crust}})g \sin \beta \\ &= \Delta \rho g \sin \beta, \text{ where } \beta \text{ is the dip of subduction} \end{aligned}$$

Crustal constitutive equations under a large range of P–T conditions, relating stress and strain rate, are the last ingredient to derive the flow field from the stress balance equations. Monomineralic flow laws are often used to approximate rock rheologies; as stated above, omphacite rheology is unfortunately not well constrained. In particular the deformation mechanism, either by diffusion or dislocation creep, corresponding respectively to Newtonian or non-Newtonian rheologies, highly influences the flow pattern within the channel (Gerya & Stöckhert 2002a). Resulting exhumation rates, as well as natural evidences described in Stöckhert (2002), Stöckhert *et al.* (1997) and Stöckhert *et al.* (1999), led Gerya *et al.* (2002b) to prefer a Newtonian rheology for high-pressure crust in their numerical model. For the same reasons and for the sake of simplicity we use a Newtonian rheology defined by

$$\tau_{xy} = \eta \frac{\partial u}{\partial y},$$

where η is the viscosity of the crust.

For a given channel width h , the velocity profile is given by

$$\bar{u} = -\frac{\alpha}{2} \bar{h}^2 (\bar{y} - \bar{y}^2) + (1 - \bar{y}),$$

where $\bar{u} = \frac{u}{U}$, $\bar{y} = \frac{y}{h}$ ($0 < \bar{y} < 1$) are adimensional variables, $\bar{h} = \frac{h}{h^*}$ and $\alpha = \frac{\partial \phi}{\partial x} \frac{h^{*2}}{\eta U}$ are adimensional parameters and U the plate velocity. The width h^* is defined from the problem dimensions as $h^* = \sqrt{\frac{2(\eta)U}{(\frac{\partial \phi}{\partial x})}}$, where $\langle \eta \rangle$ and $\langle \frac{\partial \phi}{\partial x} \rangle$ are average crust viscosity and hydraulic potential gradient, respectively. It corresponds for this average crust to the limit between circulation pattern with or without exhumation (see next section).

The velocity field is the sum of two terms (Fig. 5): one parabolic field generated by the hydraulic potential gradient and one linear

field resulting from the velocity of the lower boundary. The latter is always directed downwards and represents the dragging by the subducting lithospheric mantle, while the former has a sign given by the density difference between crust and mantle, which varies according to the crustal phase changes, and represents the buoyancy-driven movements, which can be directed either downwards or upwards.

4.1.4 Effect of α value on flow pattern

The parameter α integrates all relevant material properties and thus completely determines the flow pattern within the channel. In particular, density and viscosity variations associated to eclogitization control the value of α ,

$$\alpha = \frac{\frac{\partial \phi}{\partial x} h^{*2}}{\eta U}$$

through $\frac{\partial \phi}{\partial x}$ at the numerator and η at the denominator, respectively. α is named hereafter the ‘exhumation number’. The sign of α depends on the value of $\frac{\partial \phi}{\partial x}$, thus on crust density.

For a crust heavier than the mantle, the exhumation number is negative, as $\frac{\partial \phi}{\partial x} < 0$. Both buoyancy and lithospheric dragging driven velocity fields (parabolic and linear flows in Fig. 5, respectively) are directed downwards and no exhumation occurs in this case. Exhumation—that is, upward-directed velocities—can occur only when the crust is lighter than the mantle, for $\alpha > 0$, under certain conditions. This behaviour highly contrasts with ‘geometry-imposed’ models, where exhumation can occur as a result of forced flow in narrowing channel, irrespective of density contrast.

In the case $\alpha > 0$, the possibility of exhumation is controlled by the relative contribution of the buoyancy-driven flow, directed upwards, and the lithospheric dragging driven flow, directed downwards. The intensity of the former varies as α and $\bar{h}$ (Fig. 6): for large α and/or $\bar{h}$, the buoyancy-driven flow increases sufficiently for exhumation to occur near the top (in the y -direction) of the channel. This can be seen through the evolution of flow pattern with $\bar{h}$ variations, for a given positive exhumation number α : the range of channel widths $\bar{h}$ can be divided in two distinct sets: for $\bar{h} < \sqrt{\frac{2}{\alpha}}$, everywhere across the channel velocities are directed downwards, while for $\bar{h} > \sqrt{\frac{2}{\alpha}}$ the channel is divided into a bottom subducting zone and an top exhuming zone with reverse velocities. Conversely, for a fixed channel width $\bar{h}$, circulation patterns with exhumation are characterized by large exhumation numbers α , whereas small α ’s imply regimes without exhumation. The physical interpretation of these variations is quite simple in this case: large α result either from large hydraulic gradients $\frac{\partial \phi}{\partial x}$ or from low crust viscosity η . Large hydraulic gradients imply low density of the crust—that is, large positive buoyancy of the crust—which favours upward flow. Low crust viscosity decreases the effect of the boundary conditions (velocities imposed on top and bottom of the channel) on the circulation pattern and consequently enhances the upward buoyancy-driven flow.

The effect of density and rheology variations on flow patterns and exhumation are, therefore, faithfully integrated into the exhumation number and can be summarized as follows: *the larger the value of the exhumation number α , the greater the material potential to exhume/resistance to burial*. This property underlies all the conclusions drawn hereafter.

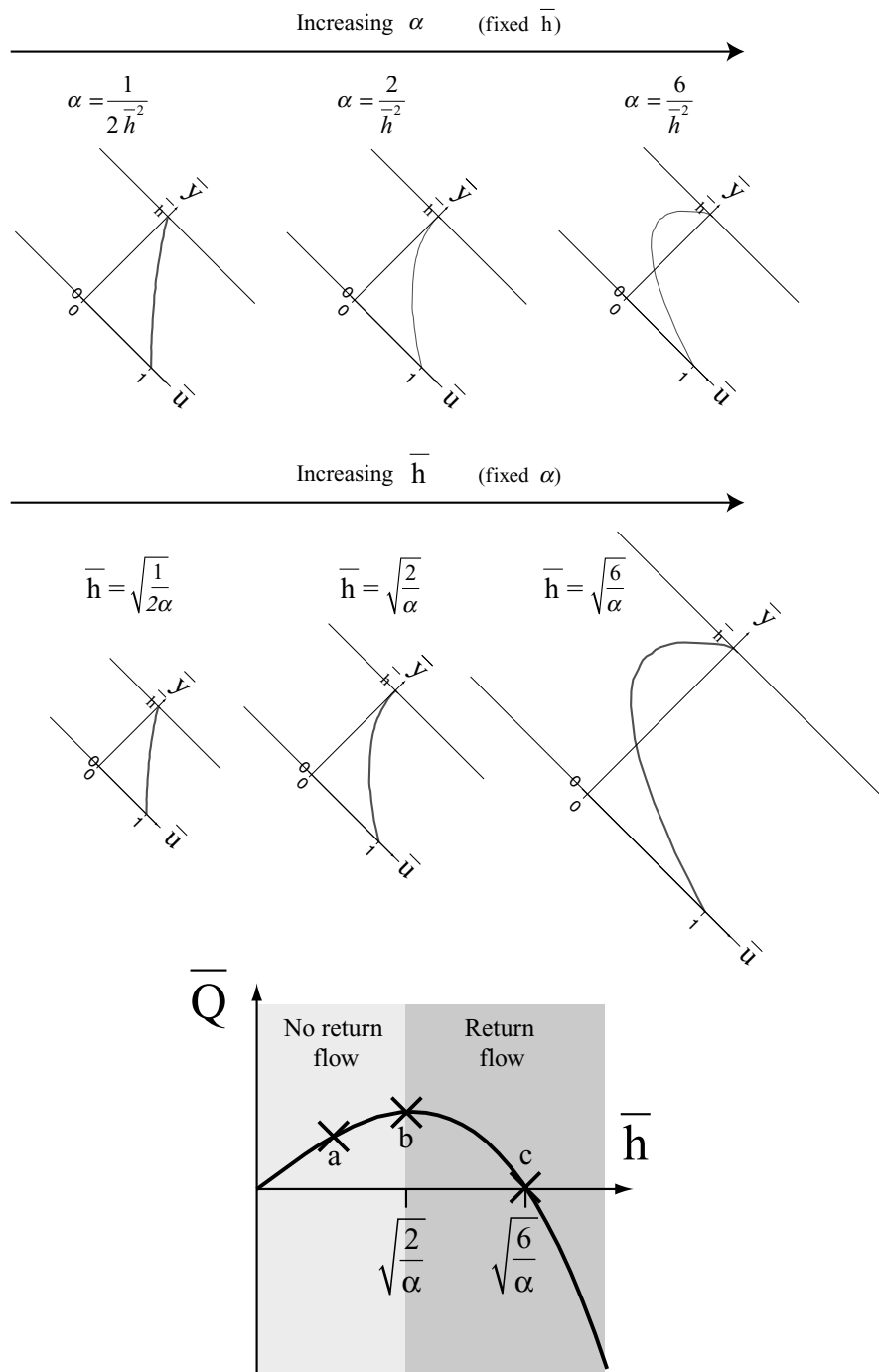


Figure 6. Top: Velocity fields for various positive α , for a fixed channel width $\bar{h}$. The increase in α results in the decrease in the downward flow and eventually in the formation of a return flow in the upper portion of the channel. Middle: Velocity fields for various channel widths $\bar{h}$, for fixed positive α . For $\bar{h} \leq \sqrt{\frac{2}{\alpha}}$, everywhere in the channel velocities are directed downwards, while for $\bar{h} \geq \sqrt{\frac{2}{\alpha}}$ in the upper section of the channel the crust is moving upwards. Bottom: $\bar{Q}(\bar{h})$ curve for a positive α . $\bar{h} = \sqrt{\frac{2}{\alpha}}$ separates regimes with and without return flow, $\bar{h} = \sqrt{\frac{6}{\alpha}}$ separates positive and negative net flow domains. The particular channel width $\bar{h} = \sqrt{\frac{2}{\alpha}}$ corresponds also to the maximum net flow $\bar{Q}_{\max} = \frac{1}{3}\sqrt{\frac{2}{\alpha}}$.

4.1.5 Channel width variations in steady state regime

The adimensional net flow across a given section is given by the velocity integral

$$\begin{aligned}\bar{Q} &= \frac{Q}{Uh^*} = \frac{1}{Uh^*} \int_0^h u dy = \int_0^h \frac{u}{U} \frac{h}{h^*} d\left(\frac{y}{h}\right) \\ \bar{Q} &= \int_0^1 \bar{u}\bar{h} d\bar{y} \\ \bar{Q} &= \frac{\bar{h}}{2} - \frac{\alpha\bar{h}^3}{12}\end{aligned}$$

Under steady-state regime the mass conservation requires that $\bar{Q}$ is constant all along the channel, that is,

$$\forall x, \quad \bar{Q}(x) = cst \Rightarrow \frac{\bar{h}(x)}{2} - \frac{\alpha(x)\bar{h}(x)^3}{12} = cst$$

The variations in channel width $\bar{h}(\bar{x})$ are, therefore, completely constrained by the exhumation number $\alpha(x)$ of the material filling the channel (Fig. 7). In our approach to the subduction channel, the mantle deforms and adapts passively to the flow properties within the channel; it is, therefore, the material *inside* the channel that controls the channel width and its time and space variations. This point of view is diametrically opposite to Mancktelow (1995) approach, where the width of the channel is fixed *a priori* by the rigid mantle that bounds it.

4.1.6 Variations of net flow with $\bar{h}$ for fixed α

If $\alpha > 0$, the parametric $\bar{Q}(\bar{h})$ curve, which depends on α , has a maximum flow $\bar{Q}_{\max}$ for $\bar{h} = \sqrt{\frac{2}{\alpha}}$ (Fig. 6 bottom):

$$\bar{Q}_{\max} = \frac{1}{3} \sqrt{\frac{2}{\alpha}}$$

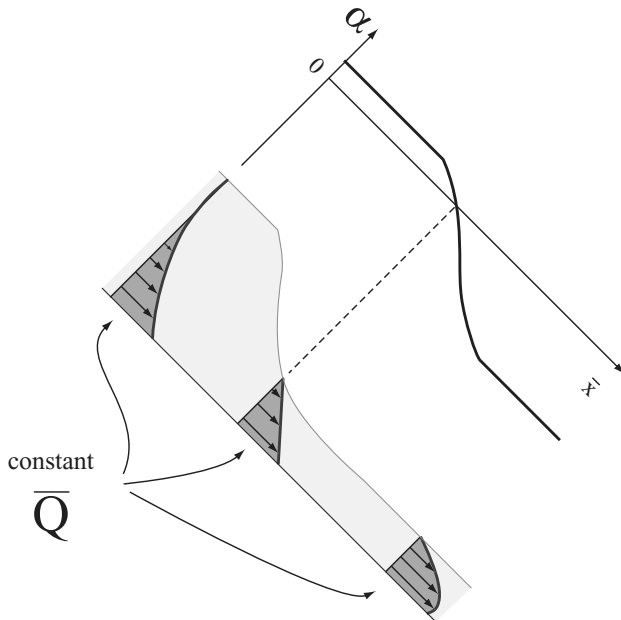


Figure 7. The morphology of the channel, that is, the $\bar{h}(\bar{x})$ profile, is completely determined by the $\alpha(\bar{x})$ profile under steady state conditions (constant $\bar{Q}$ over channel length). In the completely eclogitized zone, α is largely negative, resulting in a narrow channel. For the same net flow, in shallower domains where the exhumation number is higher, the channel is wider.

As explained above, for a fixed positive α , the buoyancy-driven flow, directed upwards, is favoured by large channel widths. The channel width $\bar{h} = \sqrt{\frac{2}{\alpha}}$ corresponds to the start of the return flow but also to the maximum of net flow $\bar{Q}$. For larger $\bar{h}$'s, the upward buoyancy-driven flow rapidly increases and the net flow $\bar{Q}$ decreases constantly and eventually changes sign (negative $\bar{Q}$ means that in average over the channel section more material is moving upwards than downwards, which is achieved when the upward buoyancy-driven flow overtakes the downward dragging driven flow).

4.1.7 Eclogitization modelling in the subduction channel

To analyse the role of eclogitization in deep crustal movements we have designed a model decomposed in three domains: The shallow domain, made of uneclogitized crust (named hereafter '0', for 'eclogite fraction equal to 0'), the deep one where the crust is completely eclogitized ('1', for 'eclogite fraction equal to 1') and an intermediate domain where eclogitization is variably completed, from no phase transformation at the boundary with the uneclogitized domain '0', to complete transformation at the boundary with eclogitized domain '1'. The exhumation number is constant within homogeneous domains '0' and '1', and varies with x in the intermediate domain.

4.2 Variations in the exhumation number related to eclogitization—end-member study

The structural analysis of the kinematics of eclogite-facies deformation on Holsnøy has shown that it recorded the first steps of exhumation. Indeed, the formation of mechanically weak large eclogite-facies shear zones, enabling decoupling between crust and lithospheric mantle within the slab, actually triggers the initiation of exhumation (Raimbourg *et al.* 2005). In the case of Holsnøy granulite, the influence of eclogitization-related density increase is much smaller than the viscosity decrease: the whole unit is *in average* not much eclogitized, as only the large shear zones are highly transformed, hence the density increase is relatively low. On the contrary, the *average* rheology of the whole unit is much decreased, as weak eclogite shear zones are continuous throughout the unit. Therefore, the difference in the concomitant evolution of *average* density and rheology during eclogitization on Holsnøy strongly promoted exhumation. Such a crust is named hereafter *R* type material, as resistant to burial.

On the contrary, let us consider a very different crust, where eclogitization still induces density increase but no rheological weakening. For this material, eclogitization unambiguously promotes burial. Such a material is named hereafter *C* type material, as compliant to burial.

Both density increase and a mechanical softening affect α , respectively through the hydraulic potential gradient and the viscosity. In the following we analyse how the qualitative differences between *C* and *R* type material end-members influence α profiles in the partially eclogitized domain. The definition of *C* and *R* type α profiles is then used in the subduction channel model to understand how and to which extent the *C* or *R* nature of buried crust, in other words the characteristics of eclogitization-related material changes, influences exhumation.

4.2.1 Burial resistant end-member (Holsnøy granulitic unit): *R* type crust

4.2.1.1 Density variations

The eclogite density, measured on the samples we collected (Table 1), is relatively variable around 3.3–3.35 g cm⁻³, comparable

Table 1. Measured density of granulite- and eclogite-facies samples.

Sample	Density	Type
H30a	3.34	eclo
H30b	3.32	eclo
H26	3.22	eclo
K02	3.22	eclo
G03	3.37	eclo
H01	3.29	eclo
N03	3.27	eclo
H14	3.40	eclo
H14	3.39	eclo
M02b	2.76	gran
A04	2.98	gran
G01	2.96	gran
M02a	2.75	gran
N01	3.10	gran

to mantle densities, while the granulite density, also variable, is around 2.8–2.9 g cm⁻³, both values in agreement with Austrheim (1987). The density of partially eclogitized granulite is a linear function of the eclogitized fraction f_e

$$d(f_e) = d_0(1 - f_e) + d_1 f_e,$$

where d_0 and d_1 are densities of uneclogitized and completely eclogitized granulite, respectively ($f_e = 0$ and $f_e = 1$).

4.2.1.2 Rheology variations

The mechanical behaviour of a mixture of two Newtonian materials is enclosed between the Voigt and Reuss bounds, which estimate overall properties using uniform strain or stress, respectively (Nemat-Nasser & Hori 1993). If η_0 and η_1 are granulite and eclogite viscosities, the viscosity of partially eclogitized granulite is comprised between

$$\eta^{\text{Voigt}}(f_e) = (1 - f_e)\eta_0 + f_e\eta_1 \quad (\text{uniform strain - Voigt estimate})$$

and

$$\eta^{\text{Reuss}}(f_e) = \left(\frac{1 - f_e}{\eta_0} + \frac{f_e}{\eta_1} \right)^{-1} \quad (\text{uniform stress - Reuss estimate})$$

The Voigt estimate rheology is stronger than Reuss estimate—that is, $\eta^{\text{Voigt}} > \eta^{\text{Reuss}}$. The variation with eclogite fraction f_e of α calculated with Reuss and Voigt estimates for rheology is shown on Fig. 8, with three different eclogite densities: inferior, equal or superior to mantle density, resulting in positive, null or negative α in the eclogite domain, and $\frac{\eta_0}{\eta_1} = 100$. For eclogite as dense as the mantle as in Holsnøy unit (central diagram in Fig. 8), α is constantly decreasing for the Voigt estimate, while it shows a maximum for intermediate f_e (close to 0.5 if $\eta_0 \gg \eta_1$) for the Reuss estimate.

The real large-scale rheology (5 × 5 km) of the mixture of un- and eclogitized crust depends on eclogite and granulite rheology and on the geometry of un- and eclogitized fractions. This large-scale rheology is highly anisotropic due to the asymmetrical shape of eclogitized volumes. Nevertheless we are not interested in determining the whole set of parameters describing the rheology of the partially eclogitized unit, but rather in the mechanical response to particular boundary conditions, namely non-coaxial shear stress, thought to correspond to high-pressure deformation in the subduction channel (Jolivet *et al.* 2005; Raimbourg *et al.* 2005).

The rheology of the granulite with little eclogite volume, corresponding to the inception of eclogitization, can be roughly approximated by the moduli of randomly distributed diluted spherical inclusions of eclogite within a granulite matrix (François *et al.* 1995,

page 159, Nemat-Nasser & Hori 1993, page 209). Under prescribed macrostrain, and with the assumption that $\eta_0 \gg \eta_1$, the overall viscosity η reduces to

$$\eta(f_e) \approx \left(1 - \frac{15}{9} f_e \right) \eta_0$$

As the strong phase, granulite, is the coherent and stress-supporting phase, the overall rheology is strong and close to Voigt estimate

$$\eta^{\text{Voigt}} = (1 - f_e)\eta_0 + f_e\eta_1 \approx (1 - f_e)\eta_0$$

As the eclogite fraction increases, it reaches the critical rheological fraction (CRF). Near this point, a continuous shear zone of weak eclogite concentrates a large proportion of the deformation, under the given shear boundary conditions, resulting in a drastic change in rheology (Fig. 9). For higher f_e , the rheology is controlled by the weak phase, and the α of partially eclogitized unit evolves near the α -curve built with Reuss rheology.

The value of the CRF for an material constituted of 2 phases with contrasted rheologies has been estimated ~20–25% of weak phase, on the basis of experiments on solid-melt (Vigneresse *et al.* 1996), anhydrite–halite assemblages (Ross *et al.* 1987) or theoretical arguments (Takeda 1998). These studies nevertheless consider isotropic geometry and organization of the two phases. In the case of Holsnøy, the very asymmetric propagation of eclogitized domains, leading to the formation of narrow and long shear zones cutting through relatively untransformed granulite, eventually resulted in a large strength drop when the shear zones connected for an average low proportion of eclogite. Therefore, in this case of very anisotropic eclogite domains, the rheological transition (CRF) occurred for a relatively low proportion of eclogite, probably <20 %.

4.2.1.3 α^R profile

Calculating the exact evolution of α^R for the Holsnøy granulitic unit is not possible, as neither eclogite rheology nor the geometry of eclogite domains are precisely known. Nevertheless, using the authorized domain defined by Reuss and Voigt bounds and the considerations above, we can qualitatively draw the α^R profile (Fig. 9 bottom) in the partially eclogitized domain. The key feature of this curve is that α^R goes through a maximum for a certain eclogite proportion, and this property, much contrasting with the C type crust defined below, will be described hereafter as determining to a great extent the circulation pattern and the possibility of exhumation.

4.2.2 Burial compliant end-member: C type crust

We now consider, as an alternative candidate to eclogitization, a lower continental crust that reaches the eclogite-facies through blueschist- or amphibolite-facies metamorphism and is relatively rich in fluids.

4.2.2.1 Density variations

The density of a lower crust with an andesitic composition after complete eclogitization, equal to 3.37 (Bousquet *et al.* 1997; Goffé *et al.* 2003), is of the order of mantle density. The previous density of the crust, equilibrated on the prograde path in the blueschist- or albite–epidote–amphibolite–facies, is 3.13 or 2.98–3.06, respectively.

4.2.2.2 Rheology variations

Many field occurrences of high-pressure rocks evidence the high strength of eclogite-facies assemblages compared to other metamorphic assemblages. For example, within the Drøsdal body in the WGR, Norway (Foreman *et al.* 2005) or within Holsnøy

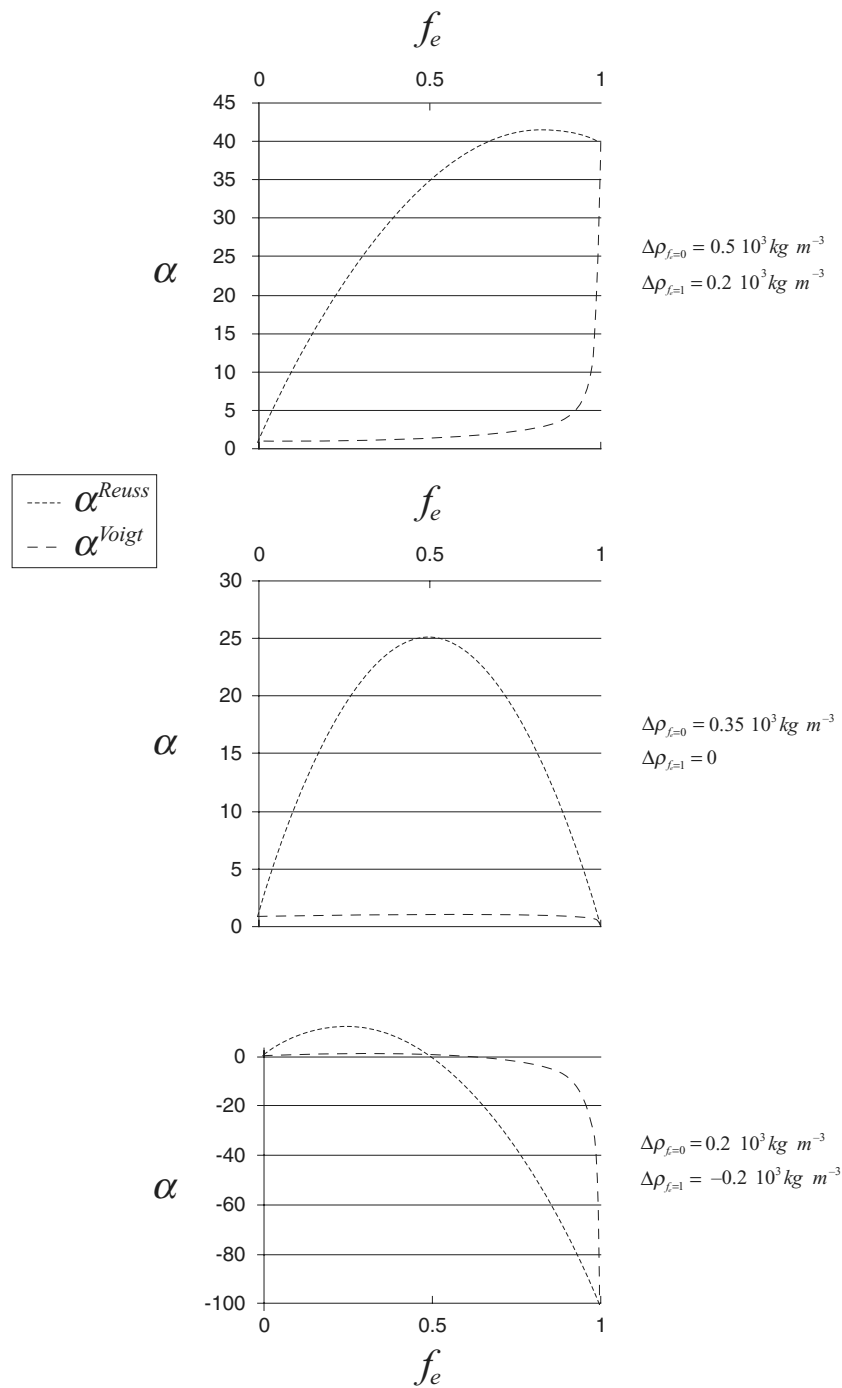


Figure 8. α variations versus eclogitization extent (proportion of eclogite in the granulite unit) for three different rocks. $\Delta \rho$, positive when crust is lighter than mantle and conversely, changes as crust is partially eclogitized. While the density changes are proportional to eclogite fraction, the rheology variations are much more difficult to constrain. The effective viscosity and resulting α of the mixture must nevertheless lie between Reuss and Voigt end-members, corresponding to prescribed macrostress ($\underline{\sigma}(x) = \underline{\sigma}^0$) and macrostrain ($\underline{\varepsilon}(x) = \underline{\varepsilon}^0$), respectively. All curves were computed for a weak eclogite with $\frac{\eta_0}{\eta_n} = 100$.

(Austrheim & Griffin 1985) the eclogitic foliation is crosscut by late amphibolite-facies shear zones, attesting of the lower strength of the lower-grade assemblage. The possibility of deformation localization and shear zone formation implies that the material within the deformed zone is weaker than surrounding material, therefore, demonstrating in the examples given above the higher strength of

host eclogite with respect to lower-grade assemblages formed within shear bands.

Additionally to the field examples, flow laws derived from laboratory experiments allow to some extent to compare the rheology of different rocks. If we consider for simplification that eclogite is governed by omphacite rheology, the flow laws for clinopyroxene by

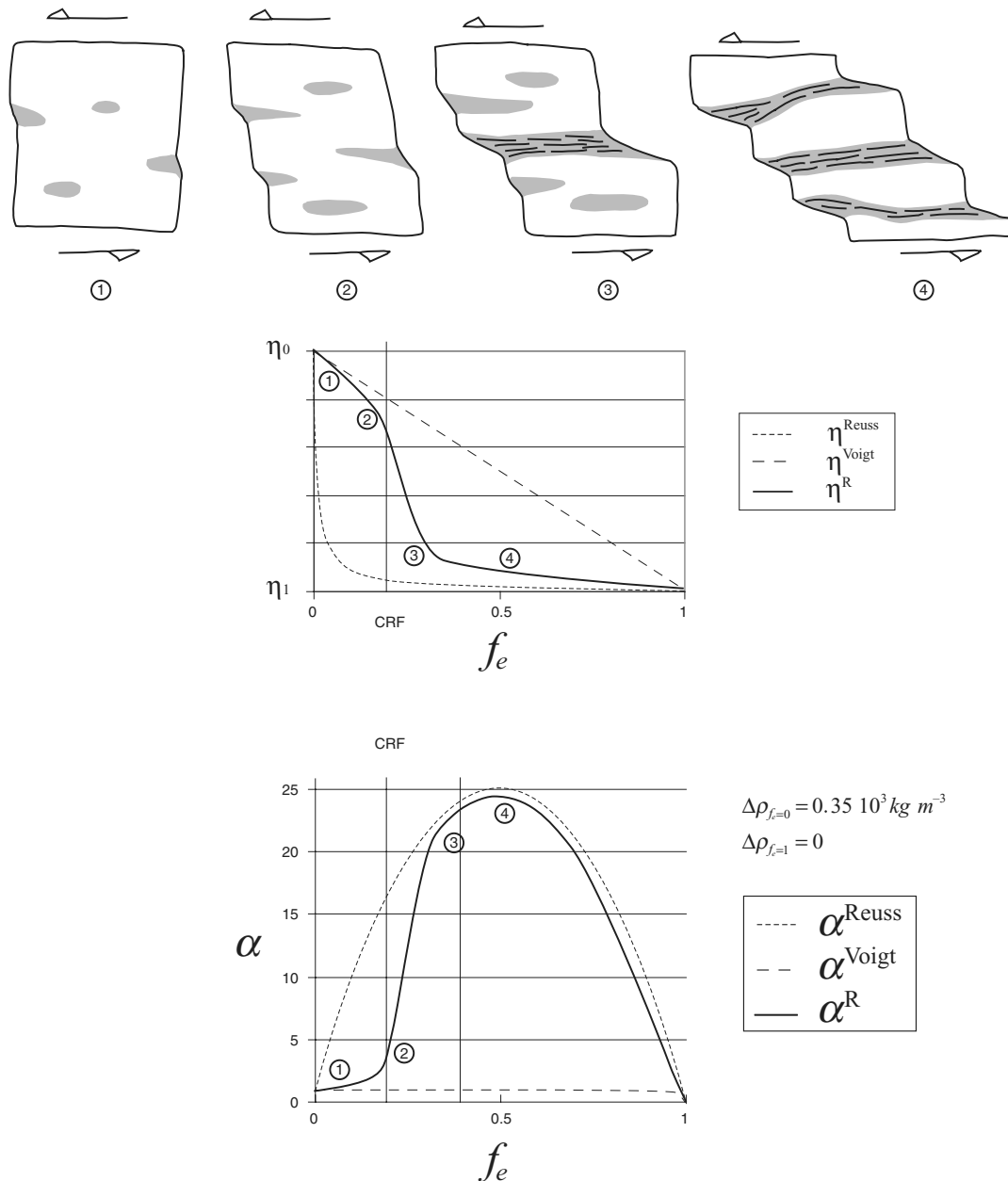


Figure 9. Qualitative evolution of effective viscosity and parameter α for the eclogitizing granulitic unit. For a low proportion of weak eclogitic material (1–2), the stress supporting matrix is the strong granulite and the effective viscosity under the prescribed shear is close to the Voigt estimate. For an eclogite proportion larger than the CRF, an eclogite volume cuts through the granulitic unit and strongly decreases the effective viscosity of the unit (3–4), as all the strain localizes into the eclogite fraction. The resulting α -curve (solid line) shows a maximum near Reuss estimate maximum for $f_e \approx 0.5$. The Reuss and Voigt estimates were computed with $\frac{\eta_0}{\eta_1} = 100$.

Bystricky & Mackwell (2001) and Dimanov *et al.* (2003) evidence the very high strength of eclogite, which deforms at much smaller rates than rocks controlled by anorthite (Rybacki & Dresen 2000).

Nevertheless it is an uneasy task to compare the mechanical properties of metamorphic facies, as rheology varies according to many parameters, such as grain size, mineral composition or fluid abundance. Considering eclogite-facies paragenesis, an increase in the jadeitic fraction of the eclogite-controlling pyroxene leads to rheological softening (Jin *et al.* 2001). Additionally, an increase in fluid content strongly weakens a given rock, as evidenced by the

strength difference between dry and wet anorthosite (Rybacki & Dresen 2000).

The fluid parameter may be considered as the main reason why anhydrous granulite-facies rocks on Holsnøy were stronger than their eclogite-facies equivalent, on the contrary to the prediction of the flow laws presented above. The quasi-absence of deformation in equivalent anhydrous granulite-facies bodies in the neighbouring WGR (Krabbendam *et al.* 2000) can be tentatively explained by their similar strong rheology. Given this strong behaviour of dehydrated granulite-facies rocks, eclogitization leads to softening.

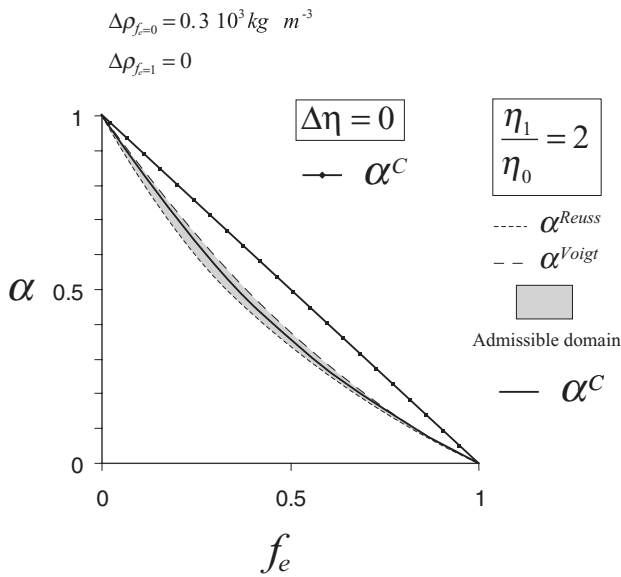


Figure 10. α variations versus eclogitization extent for a lower continental crust previously equilibrated under amphibolite facies. For $\Delta\eta > 0$ (eclogite stronger than uneclogitized crust), α -curves are concave upwards, whereas for $\Delta\eta = 0$ it is a straight line. As a result, when eclogitization induces no weakening but on the contrary either no or some hardening (C type crust), the exhumation number is constantly decreasing in the partially eclogitized zone.

On the contrary to it, lower crustal rocks not dehydrated by intense granulite-facies metamorphism are likely to undergo mechanical hardening during eclogitization.

We will, therefore, consider as a characteristic C type material such an hydrated crust whose prograde eclogitization induces rheological changes ranging from no to little hardening.

4.2.2.3 α^C variations

When eclogitized crust is stronger than uneclogitized crust, Reuss and Voigt curves are concave upwards, when both rocks have the same rheology the α^C -curve is a straight line (Fig. 10).

As a result, for an hydrated lower continental crust whose eclogitization leads to no to little hardening, the exhumation number α^C is maximum for $f_e = 0$, in other words the exhumation number of partially eclogitized crust is lower than uneclogitized crust. This material defines the compliant to burial (C type) crust that we use in the following.

4.3 Consequences of variations in the exhumation number related to eclogitization upon steady state flow regime

In the following we analyse the variations in the circulation pattern when the nature of buried material changes, using the two end-members that we have defined above. Constant and rapid variations in nature and properties of buried crust are likely, as continental crust is laterally heterogeneous. As a result, steady state circulation, corresponding to homogeneous buried crust, may not be reached. Nevertheless, considering simple situations with steady state is the first necessary step which enables (i) to fully understand the basic processes governing channel flow variations and (ii) to determine the general trends of circulation variations when studying more general situations where the nature of buried material changes constantly. We consider in the following the scenario where:

(a) First a given material of type C is buried for a time sufficiently long so that steady state flow is achieved. For this material, α changes

in the zone of eclogitization, but the flow $\bar{Q}(x)$ is constant all along the channel.

(b) Then the nature of subducted material changes for a given type R material, which is buried for a long time.

We will not consider the rapid variations in flow and channel geometry at the transition time when type R material is substituted for type C material in the subduction channel, but rather the state sometime after this transition, when a new flow pattern is controlled by the type R material.

4.3.1 Variations of the exhumation number parallel to the subduction channel

For both end-members C and R type materials, α varies along burial and eclogitization progress. In addition, as C and R type materials are different, their exhumation number in the uneclogitized domain are likely to differ ($\alpha_0^R \neq \alpha_0^C$). Nevertheless, we are not interested in the change in flow pattern when the uneclogitized properties of buried crust vary, but rather when the behaviour during eclogitization varies. In other words, subducting either a very soft and light crust or a strong and dense crust obviously affects the flow within the channel, irrespective of the way these two contrasted crust may be progressively eclogitized at depth, but this is not the point we are trying to analyse. We will rather consider two materials which have similar behaviour in the uneclogitized domain ($\alpha_0^R = \alpha_0^C$), but which evolve very differently during eclogitization, with C type material being affected by little or no hardening, in contrast to strongly softened R type materials. Doing so, we study only the effect of contrasted eclogitization behaviour on the flow pattern within the channel.

As a result we use the two end-members described above, with the condition that they have a similar behaviour in the uneclogitized domain ($\alpha_0^R = \alpha_0^C$). While in the domain of partial eclogitization the exhumation number of C type material decreases ($\alpha_0^C \geq \alpha^C(f_e) \geq \alpha_1^C$), it goes through a maximum for R type material ($f_e^{\max}$, $\alpha^R(f_e^{\max}) = \max(\alpha)$ (Fig. 11 top). There is, therefore, a peak in α profile within the partially eclogitized domain of R type crust, which does not exist for C type crust.

4.3.2 Maximum net flow and necessary return flow

When $\alpha > 0$, the net-flow curve $\bar{Q}(\bar{h})$ is convex upwards and shows a maximum flow $\bar{Q}_{\max}$ for $\bar{h} = \sqrt{\frac{2}{\alpha}}$ (Fig. 6 bottom left):

$$\bar{Q}_{\max} = \frac{1}{3} \sqrt{\frac{2}{\alpha}}$$

The larger α , the smaller this maximum flow $\bar{Q}_{\max}$: This property is key to analyse the evolution from the subduction of low to high exhumation number material. We consider here steady state regimes, where mass conservation equation requires that flow is constant everywhere,

$$\forall x, \quad \bar{Q}(x) = \frac{\bar{h}(x)}{2} - \frac{\alpha(x)\bar{h}(x)^3}{12} = cst.$$

In a first phase (Fig. 11 left), only C type crust is subducted; the flow $\bar{Q}_{eq}^C$ is near the maximum flow in the superficial '0' domain; in the deep domain '1' where the exhumation number is lower, the same flow $\bar{Q}_{eq}^C$ is reached with a narrower channel ($\bar{h}_1 < \bar{h}_0$, see the left $\bar{Q}(\bar{h})$ curve on Fig. 11). In the intermediate domain, as f_e increases continuously between '0' and '1' domains, α^C decreases

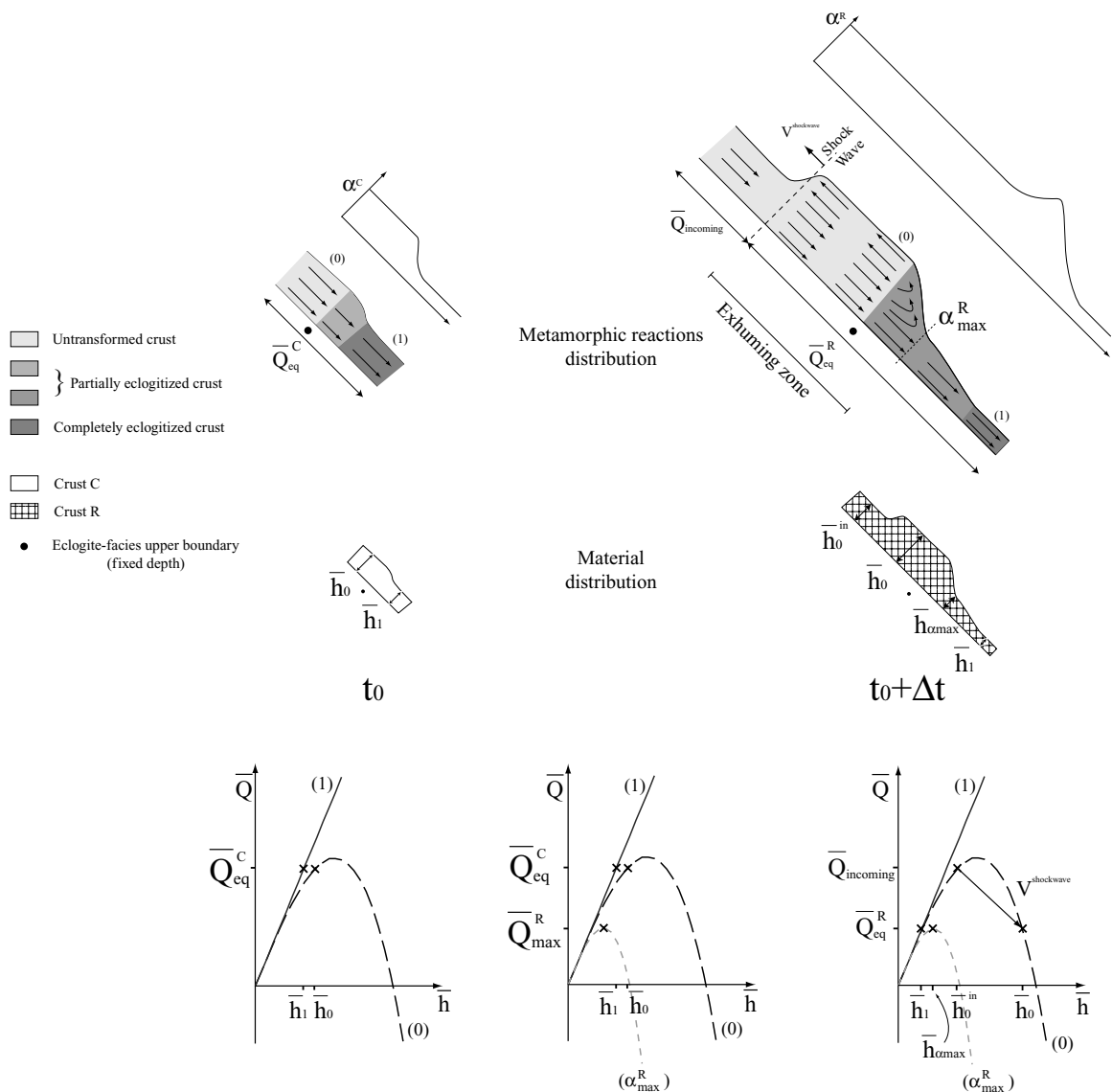


Figure 11. Consequences on the flow within the subduction channel of the material properties of partially eclogitized crust. On the left, for the C type crust, the α profile shows a constant decrease in the partially eclogitized domain. The corresponding $\bar{Q}(\bar{h})$ curves are intermediate between uneclogitized (0) and completely eclogitized (1), and at steady state flow $\bar{Q}_{eq}^C$, the width $\bar{h}$ smoothly decreases from $\bar{h}_0$ to $\bar{h}_1$ within the partially eclogitized zone. In this configuration there is no return flow anywhere. For the subsequently subducted R type crust, such as the granulite on Holsnøy, the α evolution in the partially eclogitized domain shows a maximum for a given transformation extent f_e . The maximum net flow $\bar{Q}_{max}^R$ for this maximum α is incompatible with the incoming flow $\bar{Q}_{eq}^C$ from domain (0), determined by the previous steady state flow of C type crust (middle diagram). In the final stable regime, the equilibrium flow $\bar{Q}_{eq}^R$ is determined by the maximum flow at the bottleneck (i.e. at the maximum of α), and upstream of the bottleneck the large channel width enables material exhumation. The match between $\bar{Q}_{eq}^R$ and incoming flow $\bar{Q}_{in}^{incoming}$ in the domain (0) is made by a shock wave, moving upwards with a velocity $V_{shockwave}$ prescribed by line between the two points $(\bar{h}_0^{in}, \bar{Q}_{in}^{incoming})$ and $(\bar{h}_0^R, \bar{Q}_{eq}^R)$, located above and below the shock wave, respectively (see bottom right diagram).

from α_0^C to α_1^C , and $\bar{h}$ decreases from $\bar{h}_0$ to $\bar{h}_1$. The variations in $\bar{h}(\bar{x})$ reflect the variations in exhumation number along metamorphic transformation of the subducting crust—that is, the $\bar{h}(\bar{x})$ profile is determined by the $\alpha(\bar{x})$ profile.

In a second phase, when all the C type material has been subducted, the behaviour of R type material (see α -profiles on top of Fig. 11) in the partially eclogitized domain, characterized by small maximum net flow, prevents large quantity of crust to be buried. The maximum net flow $\bar{Q}_{max}^R$ through the partially eclogitized domain, determined at the point where α^R is maximum, is no longer com-

patible with the net flow $\bar{Q}_{in}^{incoming}$ ($\bar{Q}_{in}^{incoming} = \bar{Q}_{eq}^C$) arriving from domain '0' (Fig. 11, middle $\bar{Q}(\bar{h})$ curve). The partially eclogitized domain of R type crust act, therefore, as a bottleneck for the whole subducting system, which must adapt through $\bar{h}$ evolution, not only within the intermediate domain but also in down- and upstream domains. This evolution implies to force all exceeding material arriving at the bottleneck, where the net flow is very small, to reverse velocity and start exhumation.

After transient evolution, the long-term solutions—where $\bar{Q}(x) = \bar{Q}_{max}^R$ is constant everywhere but at very shallow

Velocity Field

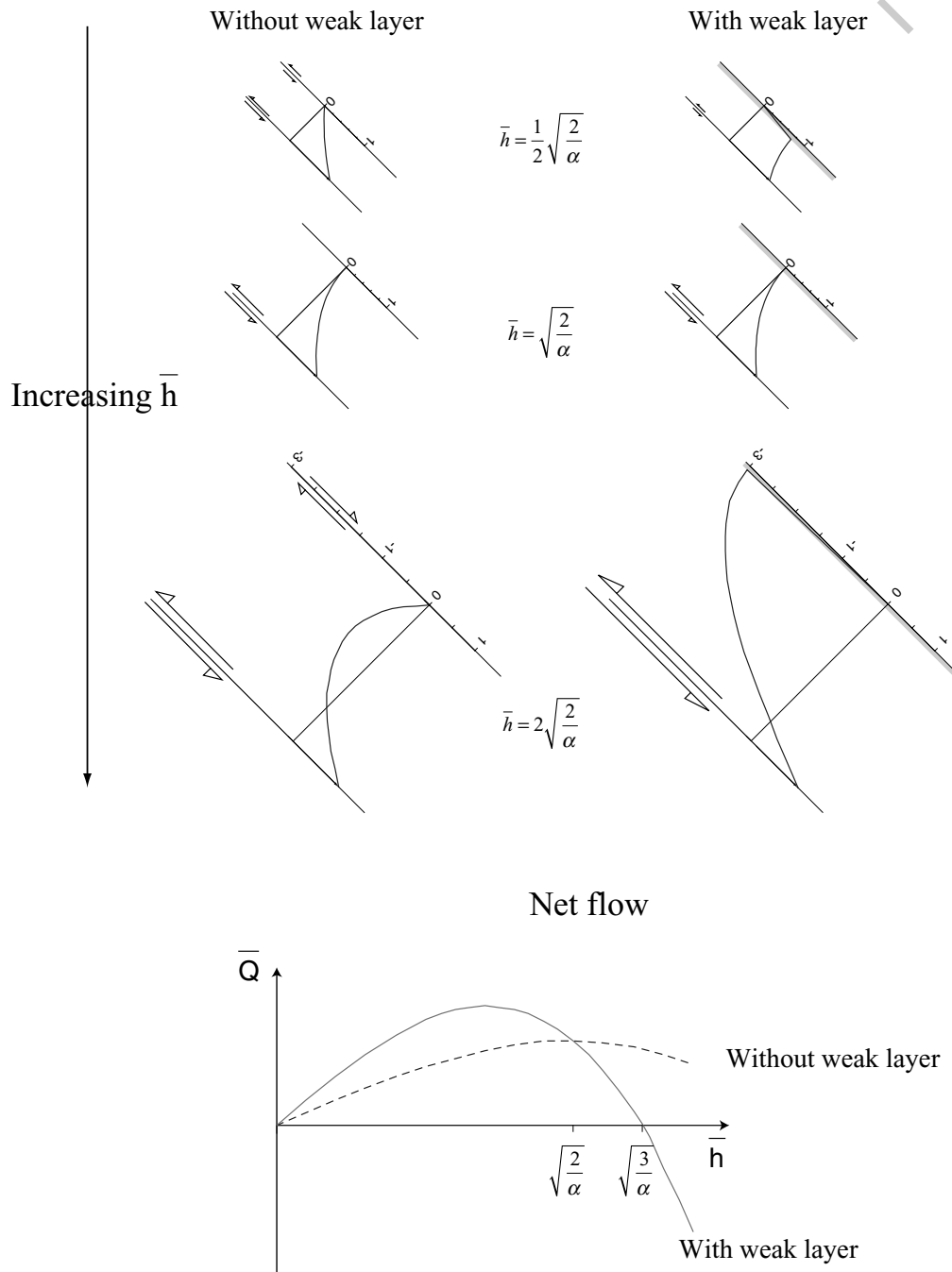


Figure 12. Consequences on the flow pattern and shear stress distribution of the presence of a weak layer on top of the subduction channel. The velocity fields for varying channel widths $\bar{h}$'s are shown with or without the weak layer. The shear stresses at the base and top of the channel are represented by arrows at scale, when no arrow stress is zero. For $\bar{h} \leq \sqrt{\frac{2}{\alpha}}$, shear stresses exerted by the hanging wall at the top of the channel without weak layer are directed upwards, and the net flow without weak layer is smaller than with it (thus average burial rates are smaller), and conversely for $\bar{h} \geq \sqrt{\frac{2}{\alpha}}$.

levels- can be described qualitatively (Fig. 11 right $\bar{Q}(\bar{h})$ curve): within the partially eclogitized domain, the bottleneck allows the maximum possible flow $\bar{Q}_{\max}^R$ ($\bar{Q}_{eq}^R = \bar{Q}_{\max}^R$). Possible solutions within '0' domain are twofold for $\bar{Q}_{eq}^R$; as too much material was

supplied from domain '0' to the bottleneck, the appropriate solution is the larger $\bar{h}$, for which the flow enables the return towards the surface of exceeding material supplied at the bottleneck region ($\bar{h}_0 > \sqrt{\frac{2}{\alpha_0^R}}$). The channel widths variations along $\bar{x}$ (see the particular

values $\bar{h}_0$, $\bar{h}\alpha_{\max}$ and $\bar{h}_1$ on the right $\bar{Q}(\bar{h})$ curve) enable the flow to be constant in and around the partially eclogitized zone of R type crust.

4.3.3 Shock wave and exhuming zone

The variations in flow pattern due to the change from C type to R type material subduction do not occur simultaneously everywhere in the channel, on the contrary they start from the partially eclogitized zone and propagate upwards and downwards. In other words, far upstream (and downstream) from the partially eclogitized zone, the flow has not yet been affected by any change while in the partially eclogitized zone the circulation has been completely modified. As a result, far upstream, the net flow is still $\bar{Q}_{\text{incoming}}$, while in the partially eclogitized zone it has changed to the new equilibrium value $\bar{Q}_{\text{eq}}^R$. The match between these two zones is achieved by a discontinuity in $\bar{h}$ and $\bar{Q}$ (named hereafter ‘shock wave’, see the circulation pattern Fig. 11 top right). The movements of this shock wave is prescribed by mass conservation: at the point where the flow $\bar{Q}$ is discontinuous, the balance of in- and outcoming material determines the instantaneous displacement of the discontinuity

$$\bar{Q}(x^+) - \bar{Q}(x^-) = (\bar{h}^+ - \bar{h}^-)\bar{V}^{\text{shockwave}},$$

where $+$ and $-$ refer to up- and downstream points in the vicinity of the shock wave. In our case, $\bar{Q}_{\text{incoming}} > \bar{Q}_{\text{eq}}^R$, therefore, exceeding incoming material is stocked through the widening of the channel from $\bar{h}_{\text{upstream}} = \bar{h}_0^{\text{in}}$ to $\bar{h}_{\text{downstream}} = \bar{h}_0$ at the rate $\bar{V}^{\text{shockwave}}$ in the x -direction (Fig. 11 bottom right). The domain where $\bar{Q} = \bar{Q}_{\text{eq}}^R$ extends, therefore, as the shock wave propagates upwards. The shock wave displacement can be interpreted as the propagation of the variations in flow pattern, due to the change from C to R type material, into the whole domain.

This shock wave may be better understood through the analogy with kinematic waves in automobile traffic arriving at a junction with a traffic light (Lighthill *et al.* 1955a,b), where the channel width $\bar{h}$ variable is replaced by the car concentration c (and the $\bar{Q}(\bar{h})$ relation by a similar $\bar{Q}(c)$ relation): when at sometime the traffic is stopped as the light gets red (equivalent to the set-up of the bottleneck in our model, when high- α material, letting through little flow $\bar{Q}$, is subducted), a traffic jam forms (null net flow $\bar{Q}(c)$),

thus null car velocity, maximum car concentration c , equivalent to our exhuming zone) and propagates backwards. The shock wave corresponds to the back limit of the traffic jam, where fast arriving cars must almost stop and are incorporated into the growing traffic jam. This discontinuity between a zone of stopped cars and a zone of fast driving cars propagates backwards, as cars are constantly provided to the traffic jam zone.

The movement of the shock wave, described by the equation above, controls the rate at which the ‘exhuming zone’, that is, the domain where material velocities are directed upwards, expands upwards. The ‘exhuming zone’ expansion rate is different from material exhumation rates, that are recorded in some natural samples. The condition for exhumation to actually occur, that is from our point view for rocks to reach the surface, is that the ‘exhuming zone’ grows enough to reach shallow levels; this condition is controlled by the ‘exhuming zone’ expansion rate.

4.3.4 Maximum eclogitization

The maximum of α within the partially eclogitized domain marks the deeper boundary of the return flow zone, which means that crust more eclogitized ($f_e > f_e(\alpha_{\max})$) carries on downwards and is never exhumed. The concomitant evolution of large-scale rheology and density, through exhumation number α evolution, determines precisely the maximum transformation of exhuming crust.

4.3.5 Exhumation rates

Let us consider material exhumation rates within the ‘exhuming zone’ in the long-term solution after R type material has been introduced into the system.

The flow pattern within the exhuming zone is constrained by the channel width $\bar{h}_0$, which itself is dependent on the maximum flow within the bottleneck $\bar{Q}_{\text{max}}^R$ (Fig. 13). As exhumation rates increase with $\bar{h}_0$, the highest exhumation rates are achieved when $\bar{Q}_{\text{max}}^R$ is minimum. Physically, this means that the more the bottleneck resists material burial, the larger the return flow upstream of the bottleneck. Very small $\bar{Q}_{\text{max}}^R$ are reached for very large maximum α in the bottleneck (e.g. corresponding to very weak eclogite). In the case most favourable to exhumation, α_{max}^R is so high that almost no material can flow downwards through the partially eclogitized zone ($\bar{Q}_{\text{max}}^R = 0$), and maximum and average exhumation rates are,

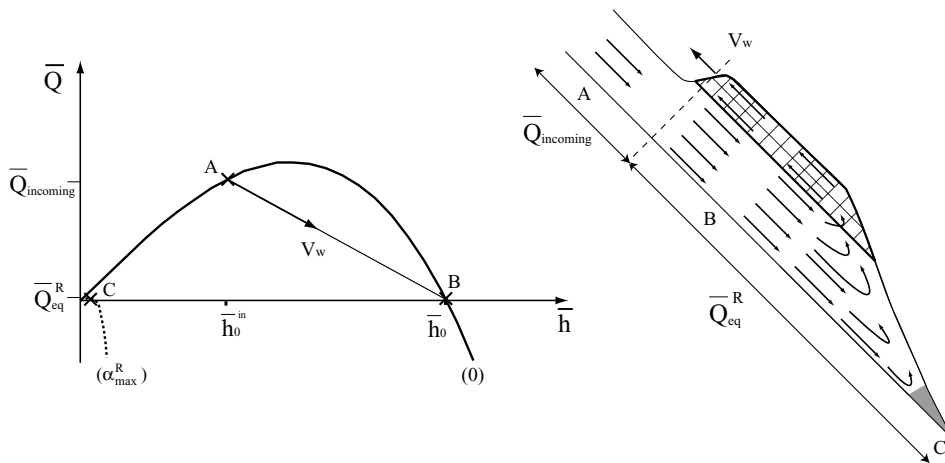


Figure 13. In steady-state subduction ($\bar{Q} \geq 0$), the most favourable situation for exhumation occurs when α is very large in the partially eclogitized zone, preventing almost all crust to flow through it ($\bar{Q} \simeq 0$), resulting in the non-eclogitized domain upstream of it in a large channel whose upper portion is exhuming. In this case, we computed the maximum and average exhumation velocity in the exhuming zone (hatched zone). Letters show the correspondence between locations along the channel and flow parameters ($\bar{Q}$, $\bar{h}$).

respectively:

$$\bar{u}^{\max} = -\frac{1}{3} \Rightarrow u^{\max} = -\frac{1}{3}U$$

$$\langle \bar{u}^{ex} \rangle = \frac{\bar{Q}^{ex}}{\bar{h}^{ex}} \approx -0.22 \Rightarrow \langle u^{ex} \rangle = -0.22U$$

where the average is computed over the portion of the channel where velocities are negative, that is, for $1 \geq \bar{y} \geq \frac{2}{\alpha \bar{h}^2}$ (this portion exists as $\bar{h} \geq \sqrt{\frac{2}{\alpha}}$).

These velocities must be understood as an upper bound to exhumation rates, corresponding to stationary burial of material most favourable to exhumation.

4.4 Effect of a weak layer on top of the channel

One of the ideas underlying the subduction channel model is that all deformation due to the relative movement between subducting and overlying plates is partitioned in a layer at the interface between both plates. In the above analysis we have supposed that this deforming layer was the crust on top of the subducting slab. Alternatively, the deforming layer is not necessarily constituted of subducting crust only, but can include a weak layer on top of it. Metasedimentary units can play this role for shallow levels (Shreve & Cloos 1986), but their occurrences in HP and UHP context (e.g. UHP metasediments in Zermatt-Saas (Lapen *et al.* 2003; Reinecke 1998), or in the Dabie-Sulu Region (Compagnoni & Rolfo 1999)) are rather scarce and it is likely that only a small fraction reach the high-pressure field. For deeper levels, the possible presence of a wedge of weak and light serpentinized mantle overlying the subducting crust may greatly enhance its exhumation (Gerya *et al.* 2002b; Guillot *et al.* 2000; Schwartz *et al.* 2001).

4.4.1 Model description

The presence of a continuous layer of weak material ‘lubricating’ the top of the subduction channel is not incompatible with the model developed above, but the null velocity condition must be replaced by a mixed condition associating stress and velocity on top $\tau_{\text{top}} = \eta_{\text{weak}} \frac{u_{\text{top}}}{\delta h_{\text{weak}}}$, where η_{weak} and δh_{weak} are the viscosity and the width of the weak material layer, respectively. The serpentinite used in Gerya *et al.* (2002b), for $T = 700^\circ\text{C}$ and $\sigma = 100$ MPa, deforms between 10^3 and 10^5 times faster than unhydrated mantle (rheology from Ranalli 1995), but its rheology is not well constrained by laboratory experiments.

For a given strain rate, rheological weakening results in stress decreasing, and the maximum stress applied under geologically realistic strain rates are all the lower as the material is weak. This consideration led us to model the weak layer not by a very unconstrained flow law, but by prescribing null stress within it

$$\tau_{\text{top}} = 0.$$

For a given channel width h , the velocity profile is given by

$$\bar{u} = -\frac{\alpha}{2} \bar{h}^2 (2\bar{y} - \bar{y}^2) + 1,$$

where $\bar{u} = \frac{u}{U}$, $\bar{y} = \frac{y}{h}$ ($0 < \bar{y} < 1$), $\bar{h} = \frac{h}{h^*}$ and $\alpha = \frac{\partial \sigma}{\partial \dot{\epsilon}} \frac{h^{*2}}{\eta U}$ are the same variables as above.

In this model the very weak serpentinized layer is sufficiently thin so that we do not consider its internal deformation; instead it is used to mimic a free-slip surface on the upper border of the

subduction channel. Nevertheless, a larger layer of serpentinite on top of the channel, which is lighter than unhydrated mantle, should also experience channel-like flow. The circulation within the resulting channel, mixing crust and serpentinized mantle, should be accounted by more sophisticated 2-D models such as in Gerya *et al.* (2002b).

4.4.2 Results

The velocity field and boundary stresses for various channel widths are shown in Fig. 12, with or without weak layer on top. The effect of the weak layer changes as channel width increases: For a narrow channel (small $\bar{h}$), there is less deformation in the crust, which is buried at a velocity close to the plate velocity U , and the downward flow is thus larger. For larger channel, when buoyancy forces overtake the viscous drag, the largest upward velocities are attained on the upper boundary, and the return flow is larger. Thus, if a weak layer enhances burial for narrow channels, it enhances exhumation for large channels. The transition occurs for $\bar{h} = \sqrt{\frac{2}{\alpha}}$: the net flow without weak layer is larger than the one with for $\bar{h} > \sqrt{\frac{2}{\alpha}}$, and conversely (Fig. 12 bottom). This transition corresponds to a sign change in the stress at the channel top in the model without weak layer: for $\bar{h} < \sqrt{\frac{2}{\alpha}}$, the upward shear stress applied by the upper boundary restrains crust burial, while for $\bar{h} > \sqrt{\frac{2}{\alpha}}$ the upper shear stress is directed downwards and restrains crust exhumation. The weak layer, which tends to decrease much the absolute value of the stress exerted by the upper boundary on the crust within the channel, has reverse effects, depending on the action of the upper boundary.

Maximum channel width for *stable* subduction regime—that is, for $\bar{Q} \geq 0$ —occurs in the model with weak layer for $\bar{h} = \sqrt{\frac{3}{\alpha}}$ —that is, for a narrower channel than without weak layer. The maximum velocity in this configuration, reached next to the weak layer, is

$$\bar{u}^{\max} = -\frac{1}{2} \Rightarrow u^{\max} = -\frac{1}{2}U,$$

while average exhumation velocity is

$$\langle \bar{u}^{ex} \rangle = \frac{\bar{Q}^{ex}}{\bar{h}^{ex}} \approx -\frac{1}{3} \Rightarrow \langle u^{ex} \rangle = -\frac{1}{3}U$$

If the introduction of a weak layer changes the velocity profiles, it does not affect the qualitative relations between $\bar{h}$, $\bar{Q}$ and α , or the order of exhumation rates in steady state subduction regime. The conclusions on the effect of eclogitization, based on the comparison of two materials affected by metamorphic reactions in a very different fashion are, therefore, still valid whether a weak layer is present or not.

5 DISCUSSION

5.1 Model assumptions

The simplicity of the channel-flow models enables to gather all relevant material properties into a single adimensional parameter α , the exhumation number, and to describe the flow pattern with unidimensional variables $\bar{Q}(\bar{x})$ (or $\bar{h}(\bar{x})$). The assumptions underlying the model are nevertheless very strong and their relevance must be assessed.

5.1.1 Unidimensional velocities

All velocities in the channel-flow model are parallel to the subduction plane. This assumption requires that the $\vec{x}$ variations of the channel width are really smooth. The shape of $\alpha(x)$ profiles (not adimensional $\bar{x}$), which determine $\frac{\partial \alpha}{\partial x}$ gradients and consequently width gradients $\frac{\partial h}{\partial x}$, depends first on material properties involved in α , but also on the actual $\vec{x}$ -size of the partially eclogitized zone. This size depends on the rates at which metamorphic reactions progress in time and propagate in space. Metastable preservation of shallow parageneses into high-pressure fields can be very efficient: igneous texture and mineralogy was largely preserved in metagranodiorite from the Brossasco-Isasca unit in the Dora-Maira massif in spite of its burial up to 24 kbars (Bruno *et al.* 2001), while the eclogitization of anhydrous granulite-facies rocks on Holsnøy or in the Flatraket Complex (Wain *et al.* 2001) was restricted to shear zones crosscutting virtually untransformed host granulite. These examples evidence that as for rheology, the rate of metamorphic transformation of a kilometre-sized unit (large-scale kinetics) depend not only on mineral-scale reaction kinetics, but also on larger-scale physical processes such as fluid transport and deformation. Furthermore the size of the partially eclogitized domain is the result of the competition between reaction kinetics and burial rates, both being highly variable in different geological settings.

All these uncertainties preclude to assess precisely the actual length of the partially eclogitized zone, and to estimate whether channel width variations are sufficiently smooth for the circulation to be unidimensional, as described in the model. Nevertheless, within the deep circulation (>40 km) the $\vec{y}$ -directed circulation is of second-order with respect to $\vec{x}$ -directed one, as attested by petrological data as well as orogen-scale numerical modelling. The shape of numerous high- and ultrahigh-pressure P–T paths shows burial and exhumation along the same cold gradient (e.g. in Dora-Maira massif, Chopin & Schertl 1999; Schertl *et al.* 1991; Simon *et al.* 1997 or in Kokchetav massif, Parkinson 2000). These gradients prescribe movements near the subduction plane, which are in agreement with results from numerical modelling showing velocities in the high-pressure domain mainly parallel to subduction (Burov *et al.* 2001; Gerya *et al.* 2002b).

5.1.2 Boundary conditions

5.1.2.1 Upper boundary velocity

If the velocity condition at the bottom of the subduction channel ($u = U_{\text{convergence}}$) is rather natural (plate velocity), kinematic boundary conditions on top are more difficult to prescribe.

As already discussed when setting up the model, three classes of boundary conditions can be distinguished, on the basis of the relative viscosity of material in- and outside the channel, that is, crust and mantle:

(a) $\eta_{\text{mantle}} \gg \frac{L^3}{h_0^3} \eta_{\text{crust}}$ (geometry imposed): the mantle is so much more viscous that it does not deform at the timescales relevant for the flow within the channel. As a result, the velocity on the upper boundary is null, and the geometry of the channel is imposed *a priori*. The model developed in Mancktelow (1995), based on the assumption of rigid channel walls, does nevertheless not satisfy these conditions.

(b) $\eta_{\text{mantle}} \ll \frac{L^3}{h_0^3} \eta_{\text{crust}}$ and $\eta_{\text{mantle}} \gg \eta_{\text{crust}}$ (e.g. if $h_0 = 5$ km and $L = 100$ km, $10^4 \gg \frac{\eta_{\text{mantle}}}{\eta_{\text{crust}}} \gg 1$) (lithostatic pressure imposed): the mantle is slightly more rigid than the crust, such that it does not deform significantly under the action of shear stresses exerted

by the flow within the channel; therefore, the velocity on top of the channel is null. Nevertheless it is not rigid enough to sustain large variations to the lithostatic pressure that trigger rapid movements from the boundary of the channel in the direction perpendicular to x . As a result, the pressure within the channel is always close to mantle lithostatic values.

(c) $\eta_{\text{mantle}} \leq \eta_{\text{crust}}$: If the mantle is not more rigid than the crust, then it is necessary involved in the flow in the x direction. In this case two approaches can be considered: either one can define a new channel of deforming material (crust at the bottom plus soft mantle on top) between two boundaries of rigid mantle, and apply the null velocity condition on top of this channel. Alternatively, one can consider only the flow within the crust, by applying a condition on top of the crustal layer that accounts for the deformation in the mantle on top of the channel (we have followed this method in the preceding section). Whatever approach is preferred, the use of the channel flow model still requires that the flow is unidimensional in deforming zone of crust and mantle. This implies that the geometry of the soft mantle on top of the channel is a layer parallel to the channel rather than a wedge with a large angle.

5.1.2.2 Material loss or gain

In the channel flow model designed by Shreve & Cloos (1986), underplating of sediments was accounted for by a loss term in the mass conservation equation. Material loss occurred when shear stresses applied on the boundary exceeded a threshold. This criterion does not seem to be relevant, because the effect of shear stresses is ambiguous, as high shear stresses on the border may imply either channel loss or gain, depending on the material scraped off. It should be also noted that under prescribed velocity conditions, the intensity of shear stresses are absolutely not indicative of the movements within the channel: as $\bar{h}$ increases, upward net flow increases, as well as downward drag by the subducting slab (Fig. 12 left).

Another possible criterion for channel material gain or loss is the evolution of the pressure within the channel. For a prescribed geometry of the channel with variable width, variations to the lithostatic pressure along $\vec{x}$ are likely to give rise to exchange of material between channel infill and walls, with mantle being injected in the channel at underpressures and conversely. Nevertheless, under the conditions stated above on the rheological contrast between crust and mantle, tectonic over- and underpressures have low amplitude and do not result in large exchanges between in- and out-of-the-channel domains.

5.1.3 Stability of steady-state solutions

While the solutions presented above describe some equilibrium flow pattern, one can wonder whether such solutions are stable. In other words, if small perturbations of these steady-state solutions, such as small variations in the channel geometry, are present, do these perturbations amplify (unstable equilibrium) or do they vanish (stable equilibrium).

To show that the steady-state solutions are stable, let us consider the very simple case where α is constant and positive (i.e. crust lighter than the mantle), and where the channel has the same width h_0 over the whole channel length but for a small portion where it smoothly narrows down to $h_0 - \Delta h$ (Fig. 14).

In such a configuration, the flow is not constant throughout the channel and the mass conservation can be expressed as:

$$\frac{\partial \bar{Q}}{\partial \bar{x}} + \frac{\partial \bar{h}}{\partial \bar{t}} = 0$$

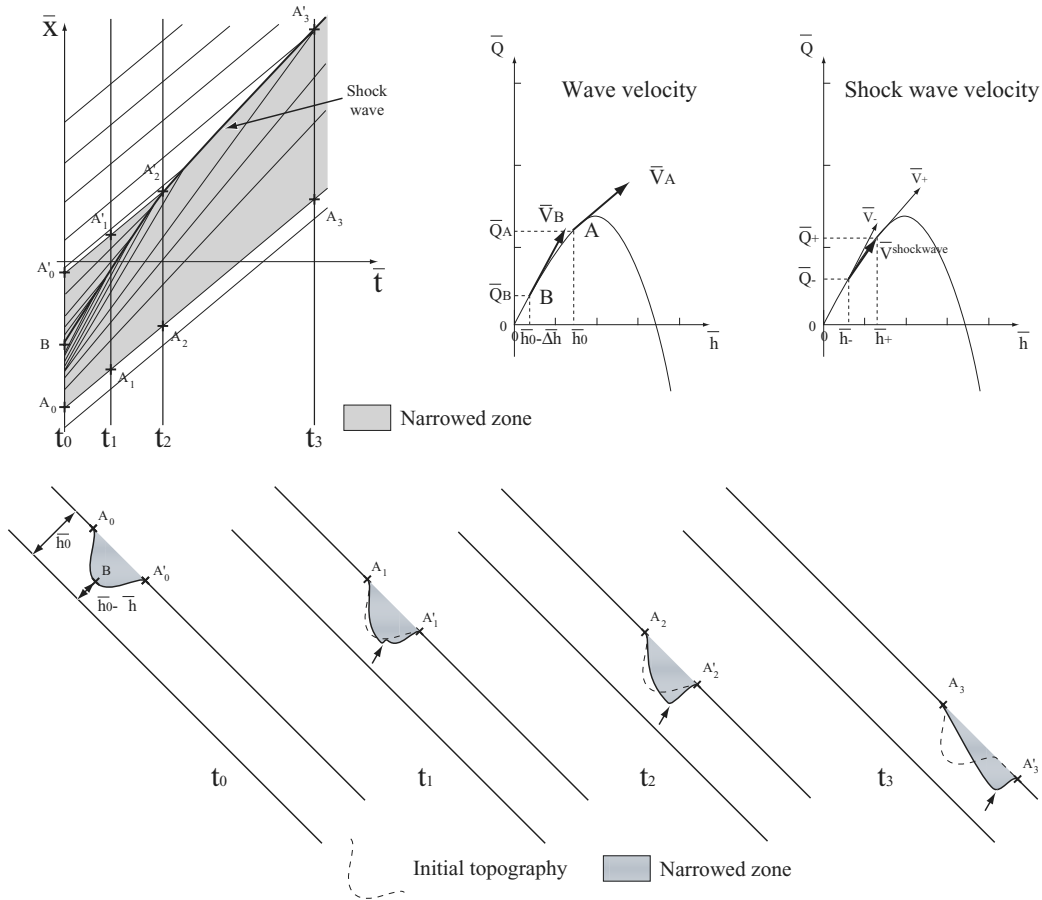


Figure 14. Evolution of the channel geometry with initial width perturbation. Top left: space–time diagram showing the trajectories of the kinematic waves. At t_0 , within the narrowed zone ($A_0 \rightarrow A'_0$), waves velocities first increase ($A_0 \rightarrow B$) then decrease ($B \rightarrow A'_0$). The convergence of kinematic waves leads to the formation of a discontinuity ('shock wave') also propagating downwards (downward = higher x). From $\bar{t}_0$ to $\bar{t}_2$, the narrowed zone (light gray) keeps its original length and moves downwards at a velocity equal to $\bar{V}_A$. From $\bar{t}_2$ on, it expands as the downward limit starts moving faster. Top middle: $\bar{Q}(\bar{h})$ curve showing channel properties outside of the narrowed zone ($\bar{Q}_A, \bar{h}_0, \bar{V}_A$) and for the narrowest channel ($\bar{Q}_B, \bar{h}_0 - \Delta\bar{h}, \bar{V}_B$). Top right: Shock wave velocity is prescribed as the chord joining $(\bar{Q}, \bar{h})^-$ to $(\bar{Q}, \bar{h})^+$; it is, therefore, included in the range $\bar{V}^- < \bar{V}^{\text{shockwave}} < \bar{V}^+$. Bottom: Snapshots of channel geometry at the given times of the space–time diagram, with letters enabling the correspondence. The grey surface ('missing material') is constant over time. The initial narrowing profile (translated to coincide) is shown for comparison with evolving topography. The arrow shows the location of the shock wave, first within the narrowed zone, then at its downward limit. After long time, the narrowed zone gets very long and the topography tends to $\bar{h}_0$.

It can be rewritten as

$$\frac{\partial \bar{Q}}{\partial \bar{x}} + \frac{1}{c} \frac{\partial \bar{Q}}{\partial \bar{t}} = 0, \quad (1)$$

where

$$c = \frac{\partial \bar{Q}}{\partial \bar{h}} \bigg|_{\bar{h}(\bar{x})}. \quad (2)$$

Eqs (1) and (2) define kinematic waves motions, described in detail in Lighthill *et al.* (1955a,b), which enable a simple and graphic analysis of the evolution of the system (such an approach, enabling to study transient evolution, is also described in detail in Raimbourg (submitted)). At $t = 0$, everywhere in the channel, we associate to the flow $\bar{Q}(\bar{x})$ and the width $\bar{h}(\bar{x})$ a wave which 'carries' the given flow $\bar{Q}(\bar{x})$ and which moves with a velocity given by eq. (2). From there on, the evolution of the channel can be studied through the motion of these waves carrying a constant $\bar{Q}(\bar{x})$ within a space–time diagram (Fig. 14).

At $t = 0$, outside of the narrowed portion of the channel, wave velocity is constant and equal to $\bar{V}_A = \frac{\partial \bar{Q}}{\partial \bar{h}} \bigg|_{\bar{h}=\bar{h}_0}$ (point A), while

within it, the wave velocity is increased up to $\bar{V}_{\text{wave}}^{\text{max}} = \frac{\partial \bar{Q}}{\partial \bar{h}} \bigg|_{\bar{h}=\bar{h}_0-\Delta\bar{h}}$ (point B). As a result, at the point of convergence between these waves carrying different $\bar{Q}(\bar{x})$ and travelling at different speed, a discontinuity ('shock wave') is formed, which propagates forwards with a velocity prescribed by mass conservation equation

$$\bar{Q}(x^+) - \bar{Q}(x^-) = (\bar{h}^+ - \bar{h}^-) \bar{V}^{\text{shockwave}},$$

where + and – refer to up- and downstream points in the vicinity of the shock wave (Fig. 14 right). At any time, the velocity of the shock wave is intermediate between the velocity of the waves merging into it ($\bar{V}^- < \bar{V}^{\text{shockwave}} < \bar{V}^+$).

The space–time diagram enables to draw snapshots of the geometry of the channel at characteristic times (Fig. 14 bottom). For the sake of mass conservation, the volume of 'missing material', that is, the difference between the actual channel profile and a channel with constant width $\bar{h}_0$ (grey surface in Fig. 14 bottom), must remain constant.

From t_0 to t_2 , the length of the narrowed zone does not change, but its geometry changes. The shock wave, which moves downwards

faster than the deeper limit of the narrowed zone (points A'_0 , A'_1 and A'_2), gets closer to it. At the same time, the 'jump' at the shock wave, that is, the discontinuity between x^+ and x^- , increases. The minimal width of the channel (located at x^- , the upper limit of the shock wave), which was $h_0 - \Delta h$ at $t = 0$, slowly increases ($\bar{h}_{\min}(t_i) > \bar{h}_{\min}(t_j)$ if $t_i > t_j$), as the most rapid waves (corresponding to the narrowest channel) are progressively absorbed into the shock wave.

From t_2 on, the narrowed zone grows larger than its initial length, as the shock wave reaches and overtakes the deeper limit of the narrowed zone. Concomitantly, the discontinuity between x^+ and x^- decreases, and the minimum width of the channel increases. For very large times, the narrowed zone gets very long, as a result of which everywhere within it the width tends to h_0 . The shock wave, located at the deeper limit, tends also to disappear, as the property of the merging wave in x^- converge to the properties of the wave in x^+ ($(\bar{Q}, \bar{h})^- \rightarrow (\bar{Q}, \bar{h})^+ = (\bar{Q}_A, \bar{h}_0)$).

From t_0 to t_2 , the narrowed zone moves downwards with a velocity $\bar{V}_A = \frac{\partial \bar{Q}}{\partial \bar{h}} \bigg|_{\bar{h}=\bar{h}_0} = \frac{1}{2} - \frac{a\bar{h}^2}{4}$, which should not be confused with the average material velocity $\langle \bar{V} \rangle_{\text{material}} = \frac{\partial \bar{Q}}{\partial \bar{h}} \bigg|_{\bar{h}=\bar{h}_0} = \frac{1}{2} - \frac{a\bar{h}^2}{12}$. From t_2 on, the downward limit of the narrowed zone (point A'_3) moves downwards faster, but the upward limit (points A_0 to A_3) keeps the constant velocity $\bar{V}_A$.

In conclusion, within the modelling framework of the 1-D subduction channel, perturbations to steady-state geometry tends to decrease in amplitude at the same time as they are swept downwards with a velocity slightly different from material velocity. It is nevertheless possible that particular channel topographies, with dense mantle on top of light crust along a sudden narrowing step in channel width, should trigger local unstable instabilities such as Rayleigh–Taylor ones, but such 2-D phenomena cannot be accounted for by our model.

5.2 Crust involved in the channel flow

Before reaching the subduction channel, the crust on top of the subducting lithospheric plate is involved in shallow circulation. Depending on the internal coherence of the crust and the crust–sediments coupling, a variable upper fraction of this crust is scraped off and starts exhuming, while the rest reaches the deep channel-flow circulation.

On the other hand, the most abundant UHP rocks now at the surface are felsic rocks (Ernst 2001), indicative of upper crustal levels. This abundance of felsic rocks to exhume has certainly much to do with their greater buoyancy (e.g. petrogenetic grid in Bousquet *et al.* 1997). An alternative reason lies in the upper crust structural position in the channel-flow: as the bottleneck created by partially eclogitized crust triggers return flow, incoming crust is divided between a lower portion that keeps on downwards and the rest that is forced back. For real heterogeneous materials, this corresponds to the formation of large shear zones decoupling slivers on top of subducting crust, enabling their exhumation. The crustal depth reached by such shear zones, or equivalently the fraction of incoming crust involved in return flow, depends on the capacity of the bottleneck to let material through. The unusual exhumation of large tracts of lower crust such as on Holsnøy can be explained in our model by a very high α , which highly decreases the capacity of the bottleneck and forces to exhume almost the totality of incoming crust, including lower levels.

5.3 Heterogeneities modelling

The channel flow model assumes for sake of simplicity that subducted crust is made of a single homogeneous material, and the resulting velocity field is rather smooth. The study of high-pressure provinces shows on the contrary that high-pressure deformation is highly heterogeneous at all scales. Heterogeneities whose size is largely inferior to the model's smallest dimension—that is, channel width—are accounted for by averaging local material properties and using effective parameters. The conclusions drawn from this study are to be understood in the same 'effective' meaning: the largest eclogite fraction of exhumed material, corresponding to the maximum of α , is an average over the channel width. Effective properties, averaged over a few kilometres, can explain the exhumation of completely eclogitized kilometre-sized mafic bodies embedded in a buoyant and little eclogitized felsic matrix in the WGR, such as Drösdal body (Foreman *et al.* 2005), or the concomitant exhumation of light and weak serpentine with eclogitic rocks (De Sigoyer *et al.* 2004; Guillot *et al.* 2001, 2000).

5.4 Petrochronological constraints on subduction models

5.4.1 Exhumation rates

Exhumation of many high- and ultrahigh-pressure terranes proceeded in 2 phases, with a rapid exhumation with little or no cooling from peak burial, followed by a much slower return to the surface (Duchêne *et al.* 1997). The rates of this deep exhumation, which we studied here through the channel flow model, are highly variable (see for example the review of exhumation rates in the WGR by Labrousse *et al.* 2004), but there is a growing number of evidences for very high rates (Perchuk & Philippot 2000), of the order of plate velocities (3.4 cm yr⁻¹, Dora Maira, Rubatto & Hermann 2001; 1–2.6 cm yr⁻¹, Zermatt-Saas, Amato *et al.* 1999; 1.5–3 cm yr⁻¹, Kokchetav, Hacker *et al.* 2003; 1.7–5.1 cm yr⁻¹, eastern Papua New Guinea, Baldwin *et al.* 2004). Such high rates are incompatible with steady state subduction—that is, with positive or null net flow— even considering the material most favourable to exhumation (with extremely large α_{max}^R). It is, therefore, likely that these rates occurred in transient evolution of the system when channel geometry and velocities changed rapidly according to variations in the nature of buried material. Alternatively, focusing of exhumation flow on top of the channel, due to across-the-channel gradients of temperature, could explain such high rates, but our 1-D model does not allow us to consider such bidimensional configurations. It must nevertheless be noted that: (i) some of these high rates are controversial (for example in the UHP unit of Kokchetav, exhumation rates estimated by Katayama *et al.* (2001) are 0.5 cm yr⁻¹, compared to 1.5–3 cm yr⁻¹ (Hacker *et al.* 2003)) and (ii) surface records are not necessary faithful to the deep processes, as rapidly exhuming terranes are much more likely to preserve HP and UHP relics without complete retrograde overprinting; therefore, rapidly exhuming units may be much more abundant on the surface than they are actually at depth.

5.4.2 P–T paths

The cold gradients leading to peak metamorphic conditions of most HP and UHP terranes is a strong argument in favour of burial in subduction context. P–T evolution on the retrograde path is on the other hand much more variable, from hair-pin like P–T paths with

high cooling rates during exhumation (see the review by Chopin 2003), to much more open clockwise P–T paths (Carswell & Zhang 1999). Most HP and UHP units in western Norway show a deep exhumation with little cooling from peak conditions down to 600–650°C ~10 kbars (Labrousse *et al.* 2004). The thermal field in the subduction zone may vary a lot with respect to the nature of buried material and convergence rates (Goffé *et al.* 2003; Peacock & Wang 1999). Moreover, the dynamics of the subduction channel strongly affect the P–T evolution of involved crust. Cooling rates during uplift are the result of the balance between heat lost to adjacent units and exhuming rates. As argued by Ernst (1999) and Ernst (2001), the thinner the metamorphic unit, the more efficient the heat loss; therefore, a large channel width, resulting in a large width of its exhuming fraction, reduces heat transfer to footwall and hanging wall and favours near-adiabatic decompression (see for example thermal field evolution with increasing wedge width in Gerya *et al.* 2002b). Concomitantly, within a large channel exhumation rates are high, also impeding heat transfer at depth. In the subduction channel model, exhumation rates and cooling during decompression are strongly related, both directly and indirectly through channel width.

Although viscous deformation has been considered as another main factor controlling temperature field through shear heating (Burg & Gerya 2005), this hypothesis is difficult to assess on natural data. It must be noted that whatever model is considered to explain the deep organization of subduction zones, crustal units with the highest exhumation rates are by definition little deformed (velocity gradients—that is, deformation rates—are null at velocity extrema) and little affected by shear heating. As these rapidly exhuming units are the most refractory to late overprinting, both deformation and thermal conditions deduced from them may not be faithful to actual deep thermal and kinematic framework.

6 CONCLUSION

To study the structure and dynamics of subduction zones, large-scale properties of rocks are necessary. As metamorphic reactions are neither instantaneous nor ubiquitous, most metamorphic rocks consist of juxtaposed volumes which equilibrated under different conditions. When studying the effect of eclogitization on deep crustal circulation, it is, therefore, not sufficient to consider only end-members, that is, no and complete transformation, we must also consider the whole range of rocks whose eclogite proportion varies continuously between 0 and 100 per cent.

The consequences for exhumation of progressive eclogitization and related evolution of physical properties of subducting crust are studied through the unidimensional channel flow model. Both density and large-scale rheology are integrated into a single parameter, the exhumation number α , whose profile along subduction plane completely determines the flow pattern. The maximum possible net flow through a given section is conversely proportional to $\alpha^{1/2}$. For a material whose eclogitization results in a large viscosity decrease, similar to the rocks studied in the Bergen Arcs ('burial resistant' crust), the α -profile within the partially eclogitized zone goes through a maximum. Such a high- α crust precludes the burial of large amount of crust, the partially eclogitized zone act, therefore, as a bottleneck to the flow. For high incoming flow of crust, the excess material, which cannot flow through the bottleneck, is forced backwards and starts exhuming. The exhuming zone that forms expands upwards at a constant rate controlled by the difference between upstream incoming flow and maximum flow in the α -max zone. The

eclogite proportion for which α is maximum is also the maximum eclogite fraction of the crust that is exhumed. Additionally, we show that really high exhumation rates, of the order of plate velocities, are incompatible with stable subduction regime. The presence of a weak layer on top of the channel favours burial for small channel widths, but enhances exhumation for large channel widths.

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