Shear heating in extensional detachments: Implications for the thermal history of the Devonian basins of W Norway

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ABSTRACT

The supra-detachment basins in western Norway were formed during Lower to Middle Devonian in the hangingwall of the extensional system of detachments known as the Nordfjord-Sogn Detachment Zone (NSDZ). The basins experienced elevated peak temperatures (up to 345 °C adjacent to the main detachment fault) during the terminal stages of the extension. In this study, we show that the heat generated by the deformation of the rocks within the shear zone of the detachment, also known as shear heating, played a role in the thermal evolution of the entire region during the extension and could have induced the elevated peak temperatures of the supra-detachment basins. We numerically analyse the influence of the rheology and the deformation style within the shear zone, the rate of exhumation of the footwall, and the thickness of the sedimentary accumulation on the top of the system. The model reproduces the elevated temperatures recorded in the supra-detachment basins where locally 100 °C and 25% of the total heat budget can be attributed to shear heating. We show that this additional heat source may increase the peak temperatures of the hangingwall as far as 5 km away from the detachment.

1. Introduction

Large-scale extensional detachments, resulting from gravitational collapse of over-thickened crust, or extension driven by plate motions, have been recognized and documented in a number of places and provide a major mechanism to exhume large segments of the lower crust and mantle (Dewey, 1988; Lister et al., 1986; Manatschal, 2004; Wernicke, 1985). The main geological feature of such crustal-scale excision is the juxtaposition of lower and upper crustal rocks across relatively narrow (100 m to a few kilometre thick) damage and shear zones (e.g. Andersen and Jamtveit, 1990; Norton, 1986). Based on geological observations, the ‘simple shear model’ or modified versions of it, explains the excision of the lithosphere along low angle normal faults and shear zones. However, despite the general acceptance of this model, several important questions remain regarding the dynamics and the overall thermal evolution of crustal-scale detachments.

A detachment consists of a footwall (“lower plate”), a shear zone, and a hangingwall (“upper plate”).

1) The footwall records higher-grade synkinematic metamorphism and younger metamorphic cooling ages than the hangingwall. The footwall is often referred to as a metamorphic core complex, and in extreme cases the entire crust may be stretched off to expose mantle rocks (Manatschal, 2004; Wernicke, 1985).

2) The shear zone is usually characterized by the ductile deformation with normal-sense kinematic indicators in the mylonites, which are commonly capped by brittle fault rocks. In the case of the Nordfjord-Sogn Detachment Zone, the mylonites may be several kilometers thick and associated with displacement in the order of 100 km (Andersen and Jamtveit, 1990; Hacker et al., 2003).

3) In the upper stratigraphic level of the hangingwall, syntectonic supra-detachment basins develop along active faults (e.g. low-angle-normal faults, listric normal faults or strike-slip transfer faults) and are commonly found in tectonic contact with the shear zone (Andersen, 1998; Osmundsen et al., 1998, 2000; Seguret et al., 1989).

The thermal evolution of a crustal-scale detachment has to be considered with two distinct aspects: fast exhumation and highly-localized deformation. Pressure-temperature-time (PT-t) paths of the footwall metamorphic core complexes typically indicate fast decompression at near-isothermal conditions during exhumation from the lower to the upper crust level (Labrousse et al., 2004). The time window bracketing such isothermal exhumation is only a few million years (Hacker et al., 2010; Johnston et al., 2007) and is significantly smaller than the time required for thermally equilibrating the crust by diffusion...
(several tenths of Myr, e.g., Turcotte and Schubert, 2002). Therefore, the thermal structure of the crust after fast exhumation of a high- to ultra-high-pressure [(U)HP] terrane might be strongly perturbed by interaction between a “hot” footwall and a “cold” hangingwall.

The other characteristic feature controlling the thermal evolution of the detachment zone results from the relative movement between the foot- and the hangingwall. Large strain can accumulate inside the shear zone of the detachment, and this may have a significant effect.

Fig. 1. a) Geological map of the Nordfjord-Sogn Detachment Zone in western Norway. b) Schematic cross section (E-W) across the detachment. Heat flux introduced at the base of the Devonian basins can come from the exhumation of the footwall or from shear heating in the shear zone.
on the heat budget of the system in form of shear heating (Brun and Cobbold, 1980; Burg and Gerya, 2005; Campani et al., 2010; Hartz and Podladchikov, 2008; Leloup et al., 1999; Schmalholz et al., 2009; Scholz, 1980). Shear heating rate depends on the non-elastic strain rate and the deviatoric stress experienced by rocks during deformation. The amount of shear heating produced in a geological system is difficult to measure and modelling provides a viable tool for assessing the role of shear heating.

We base our study on the geological observations of the Nordfordj-Sogn Detachment Zone (NSDZ) of western Norway. In the top stratigraphic level of the detachment, several supra-detachment basins formed (Fig. 1) as a response to repetitive tectonic motions during the exhumation of the footwall (Osmundsen et al., 1998; Seguret et al., 1989; Seranne et al., 1989; Steel et al., 1977). These supra-detachments basins are commonly named the Devonian basins due to the presence of Devonian plant fossils in the sediments (Heegh, 1945; Kolderup, 1916, 1921, 1927). The special feature of the Devonian basins is documented in low-grade metamorphism (Seranne and Seguret, 1987; Souche et al., 2012; Svensen et al., 2001; Torsvik et al., 1988). Souche et al. (2012) showed that the temperature conditions at the burial depth of 9.1 km was as high as 345 °C which can be translated to a geothermal gradient of 38 °C/km. Additionally, the peak temperatures in the Devonian basins increase with proximity to the detachment contact and thus suggest that the lateral variation of maximum temperatures was controlled by a dynamically evolving system, most likely linked to the detachment. Was shear heating contributing to the heat budget in the system? Could it have an impact on the lateral variation of the peak temperatures within the supra-detachment basins?

In this study, we analyse the thermal evolution of the detachment zone and estimate the contribution of shear heating on the heat budget of the area. We present a compilation of available geological data from the NSDZ to build a well-founded geological model that can be studied numerically. We compare the results of our numerical studies with independent observations from the supra-detachment basins of western Norway and discuss the importance of the shear heating in this particular geological setting.

2. Geological constraints of the model

2.1. Footwall conditions and strain partitioning in the detachment

The NSDZ is one of the largest extensional detachments in the world, with a displacement of 70 to 100 km (Andersen and Jamtveit, 1990; Hacker et al., 2003). The extension was initiated during the latest stages of the Caledonian continent-continent collision when the crust reached its maximum thickness and partially crystallized to (U)HP eclogites in the Late Silurian to Early Devonian (Hacker et al., 2012; Krogh et al., 2001). The extension was initiated during the latest stages of the Caledonian metamorphism (Fig. 2) and the kinematic evolution of the NSDZ (Hacker et al., 2003; Johnston et al., 2007; Young et al., 2011). We take the maximum burial of the HP units to represent the initial depth (~60 km) where the exhumation started along the shear zone. Based on the decompression path of the HP units, we constrain our model to account for 40 km of vertical exhumation bringing the HP rocks to a depth of 20 km, at which the retrograde mineral assemblages and their cooling age have been determined by 40Ar/39Ar geochronology (Andersen, 1998; Chauvet and Dallmeyer, 1992; Cuthbert, 1991; Fossen and Dallmeyer, 1998; Young et al., 2011).

The mylonites of the NSDZ have a maximum thickness of 5.5 km north of the Hornelen basin, but the precise original thicknesses cannot be accurately determined because of excision by later faults. Marques et al. (2007) estimated the shear strain partitioning in the NSDZ and showed that the lower parts of the mylonites experienced considerable flattening across the foliation in addition to simple shear (ratio of simple to pure shear ~ 1). The upper ~2 km of the mylonites are, however, characterized mainly by simple shear, with a shear strain of approximately 20 (Andersen and Jamtveit, 1990; Hacker et al., 2003; Marques et al., 2007), used here as the bulk shear strain along the NSDZ.

2.2. The hangingwall peak temperatures

Following fluid-inclusion and mineralogical studies of the basins (Svensen et al., 2001), a recent study using Raman spectroscopy of carbonaceous material (RSMC) from Devonian plant fossils (e.g. Souche et al., 2012) documents high peak temperatures from the Solund, Hornelen, and Kvamshesten basins. The temperatures locally reached 345 °C for an estimated burial of 9.1 km in the Kvamshesten basin (Table 1). This suggests a geothermal gradient of approximately 38 °C/km, which is clearly at variance with commonly accepted geothermal estimates for a thickened orogenic crust or even for a crust in early stages (415–395 Ma) of post-orogenic extension (Fossen, 2010; Gabrielsen et al., 2005; Hacker, 2007; Spengler et al., 2009). Souche et al. (2012) concluded that the elevated basin temperatures could only be explained by an additional heat source introduced at the base of the basin.

Table 1

<table>
<thead>
<tr>
<th>Burial depth (km)</th>
<th>T (average) °C</th>
<th>T (local) °C</th>
<th>Distance from the detachment km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hornelen</td>
<td>9.1 ± 1.6</td>
<td>250 ± 20</td>
<td>295 ± 30</td>
</tr>
<tr>
<td>Kvamshesten</td>
<td>9.1 ± 1.6</td>
<td>250 ± 20</td>
<td>345 ± 30</td>
</tr>
<tr>
<td>Solund</td>
<td>13.4 ± 0.6</td>
<td>315 ± 15</td>
<td>284 ± 30</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Fluid-inclusion and mineralogical studies (*)</th>
<th>Raman spectroscopy of carbonaceous material (**)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Burial depth (km)</td>
<td>T (average) °C</td>
</tr>
<tr>
<td>------------------</td>
<td>---------------</td>
</tr>
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</tr>
</tbody>
</table>

*Souche et al. (2012); **Andersen et al. (2001); ***Hacker et al. (2003) The burial depth is estimated from lithostatic pressure assuming a constant rock density of 2600 kg/m³ (Svensen et al., 2001).
the hangingwall during the extension. This conclusion also is in agreement with previous description of ductile deformation of the basin-fill sediments adjacent to the detachment, which could suggest elevated temperatures in this area (Braathen et al., 2004; Norton, 1987).

2.3. Model: constraints and simplifications

The different geological aspects presented in the previous sections allow us to build a simplified geological model of the evolution of the detachment zone used in the numerical analysis (Fig. 3). The initial configuration of the model (Fig. 3a) corresponds to the thickened Caledonian crust with a Moho depth set to 70 km corresponding to approximately twice the average thickness of continental crust. We assume that the deformation was initiated by the development of a lithospheric-scale shear zone with a constant dip angle $\alpha = 30^\circ$ (Andersen and Jamtveit, 1990; Norton, 1986).

The active deformation within the shear zone and the exhumation of the footwall is considered within a time window denoted $t^*$. At that time the system still. We assume constant rates of deformation for simplicity. During deformation, the footwall translates upward along the detachment and accounts for the exhumation of the lower crustal body ($h_f = 40$ km) from 60 to 20 km depth (marked by a star in Fig. 3a and b). At the same time, the hangingwall translates downward along the detachment and creates accommodation space in the basin formed at the surface. We assume immediate erosion and sedimentation processes so that the model does not produce variations in topography. At time $t^*$, the lower crustal body is at 20 km depth and the basin is at maximum depth ($b_{max}$). Accordingly to the estimated burial depth of the Devonian basins (Table 1), we consider two values for $b_{max}$, 10 km (for the Hornelen and Kvamshesten basins) and 15 km (for the Solund basin).

An important element of the model is the style of deformation within the shear zone. Despite observations of Marques et al. (2007) on the significant component of pure shear within the lower parts of the NSDZ (Section 2.1), we assume simple shear only. This simplification reduces the number of model parameters, which are poorly constrained and might be related to the earlier stages of exhumation (Andersen et al., 1994). Thus, simple shear (Fig. 3c) accommodates the total relative displacement between the foot- and hangingwall. According to the observations and measurements along NSDZ (see Section 2.1), we set a bulk shear strain ($\gamma_{bulk}$) of 20 at time $t^*$. Given $\gamma_{bulk}$, $\alpha$, and the total displacement between the foot- and hangingwall, we estimate the thickness of the shear zone to be $D = 5$-$5.5$ km depending on $b_{max}$, which is consistent with the observations (Marques et al., 2007).

The assumption of uniform translation (and thus, without any deformation) of the foot- and hangingwall is a simplification of our model. We have, however, tested the importance of the lateral stretching of the units, and found that a moderate extension does not cause significant changes of the peak temperatures in vicinity of the detachment zone.

Normal sense reactivation of the detachment fault along the Solund and Kvamshesten segments (Fig. 1) in the Permian, and again in the Late Jurassic has been documented by palaeomagnetic and geochronological studies of fault breccias along the NSDZ (Andersen et al., 1999; Eide et al., 1997; Torsvik et al., 1992), and may have contributed to some exhumation. In this study, we assume that the Devonian basins reached their peak temperature conditions during or shortly after the major post-Caledonian extension and were not influenced by later reactivation of the detachment faults.

3. Model

3.1. Governing equations

We analyze the thermal evolution in the model by solving the transient heat balance equation:

$$\rho C_p \left( \frac{\partial T}{\partial t} + \vec{v} \cdot \nabla T \right) - \nabla \cdot (k \nabla T) = H_f + H_s,$$

where $T$ is the temperature, $t$ is the time, $k$ is the thermal conductivity, $\vec{v}$ is the rock velocity, $\rho$ is the rock density, $C_p$ is the rock specific heat capacity, and $H_f$ is the radiogenic heat production (see characteristic values in Table 2). The shear heating term $H_s$ is the product of the deviatoric stress $\tau$ and the non-elastic strain rate $\dot{\gamma}$ (Brun and Cobbold, 1980; Burg and Gerya, 2005; Turcotte and Schubert, 2002) so that:

$$H_s = \tau \dot{\gamma}$$

The hangingwall and footwall translate along the detachment without internal deformation. All the deformation is accommodated within the shear zone and thus shear heating (Eq. (2)) is only used in the shear zone.

Completing the mathematical model of thermal evolution requires the specification of initial and boundary temperature conditions, velocities and rheological relationship. The assumption of constant translation during exhumation (Section 2.3) gives the velocities within the foot- and the hangingwall by dividing the total displacements by $t^*$. The initial and boundary temperature of the system and specification of Eq. (2) for the shear zone are discussed below.

3.2. Initial geotherm

The initial thermal situation before the extension has to be considered in the geological framework of the Caledonian orogeny (peak condition data points on Fig. 2). Thus, to obtain the initial thermal state, we perform an auxiliary modelling of the evolution of the lithosphere during the Caledonian collision. In this experiment we assume that, prior to collision, the lithosphere was in steady state and, during the orogeny, the lithosphere thickened uniformly by pure

![Fig. 3.](image_url)
shear. We used a 1D computer code LitoToastPhere (Hartz and Podladchikov, 2008) to perform the calculations.

We consider the Caledonian continent-continent collision to last for 25 My (Corfu et al., 2006; Johnston et al., 2007; Torsvik and Cocks, 2005) and to result in 100% thickening of the crust (double of the original thickness) which corresponds to an average strain rate of $8.7 \times 10^{-16} \text{s}^{-1}$. The model lithosphere consists of three layers, composed of a wet quartzite upper crust, a diabase lower crust, and a dry dunite mantle (Ranalli, 1995). The initial thickness of the upper crust is set to 16 km (Hasterok and Chapman, 2011). The lower crust extends to the Moho discontinuity with initial depths of 30, 35 and 40 km (corresponding to the model "Mxx"). The initial depth of the mantle lithosphere is set to 140 km (Artiemiya, 2006). The thermal properties used in the calculations are listed in Table 2. We performed a series of numerical tests varying the initial Moho depth and the amount of radioactive heat production and considering the influence of shear heating. We compare the model results with the data of the peak metamorphic conditions of the Caledonian nappes below the Solund basin and the Hornelen basins (Fig. 2). The best fit geotherm is obtained with the model M35-sh (Fig. 4), which accounts for shear heating during the deformation of the lithosphere. Fig. 4 also presents model results obtained with similar setups but where shear heating was ignored (M30, M35 and M40). The shear heating models are warmer by an average of 100 °C, illustrating the importance of this heat source during deformation. The results also illustrate greater dependence on the initial conditions and rock properties for the models without shear heating.

We use the preferred temperature profile M35-sh (Fig. 4) to build the initial temperature distribution. We also use this profile to constrain the bottom boundary condition during the model evolution. As the footwall moves upward during exhumation, the material incoming into the system assumed to sustain its temperature initially prescribed by M35-sh. This assumption ignores the thermal relaxation of the deep lithosphere after Caledonian orogeny, which is approximately valid for fast exhumation considered here.

### 3.3. Kinematic of the shear zone

Restoring the exact distribution of velocities and stress field within the shear zone is beyond the scope of this study. Instead, we approximate the velocities within the shear zone using two end-member approaches, in which either shear strain rate $\dot{\gamma}_{\text{cst}}$ or shear stress ($\tau_{\text{cst}}$ model) is kept constant across the shear zone. The two kinematic models correspond respectively to an upper and lower bound on the estimate of shear heating expressed by Eq. (2). Whereas the $\gamma_{\text{cst}}$ model forces uniform deformation and thus active heat production even in strong areas of the shear zone, the $\tau_{\text{cst}}$ model results in localized deformation of weaker parts of the detachment, and thus produces less heat. It is important to keep in mind that neither the $\dot{\gamma}_{\text{cst}}$ nor the $\tau_{\text{cst}}$ models reproduce exactly the actual deformation process within the Nordfjord-Sogn Detachment and the main purpose in using these two virtual kinematic models is to deliver an upper and lower bound on the shear heating effect.

#### 3.3.1. Rheology

The shear zone accumulates the entire deformation in the model (Fig. 3c) and hosts the production of shear heating. This deformation occurs in the range of 0-60 km, and thus involves the transition from brittle to ductile behaviour of rocks. The brittle behaviour is described by Byerlee’s relationship:

$$\tau_{by} = (1-p_f)\mu\gamma \int_{0}^{y} \rho dy$$

See Table 2 for definitions and values of parameters used. This constitutive relationship assumes no deformation for the stress values below critical, $\tau_{by}$, which increases with depth (Fig. 5). We approximate Eq. (3) using a power law relationship (after Nanjo et al., 2005; Turcotte and Glasscoe, 2004)

$$\tau = \tau_{by} \left(1 - \frac{\tau_{by}}{\tau_{cr}}\right)^{n_b} \text{ or } \gamma_{by} = \left(\frac{\tau}{\tau_{by}}\right)^{n_b} \gamma_{cr}$$

where $\gamma_{cr}$ is a critical strain rate and $\gamma_{by}$ is the strain rate of the brittle deformation. Here we define $\gamma_{cr}$ as the bulk strain rate imposed by the motion of the footwall and the hangingwall ($\gamma_{bulk}/t^*$. For large
power law exponent \((n_b = 20)\), the deviatoric stress is approximately equal to the critical stress \(\tau_b\) (yield strength) for a wide range of strain rate values centered about the critical strain rate \(\dot{\gamma}_c\).

The ductile behaviour is described by a power-law for dislocation creep:

\[
\tau = \exp(E/RTn_d) \sqrt[3]{\dot{\gamma}_b/A} \quad \text{or} \quad \dot{\gamma}_d = \tau^{n_b} A \exp(-E/RT)
\]

where \(\dot{\gamma}_d\) is the strain rate of the ductile deformation. The Arrhenius term \(\exp(E/RTn_d)\) results in a strong decrease of the stress with temperature and, for standard geotherm, with depth (Fig. 5b). The shear zone exposed today between the Devonian basins and the WGR is mainly formed of Caledonian nappes represented by meta-sedimentary units and some deformed slices of basement. The rheological weakness of the meta-sediments is approximated by flow laws for wet and dry quartzite (see Table 3).

We assume the brittle and the ductile mechanism of deformation to act simultaneously on the total shear strain rate, and it allows us to write:

\[
\dot{\gamma} = \dot{\gamma}_b + \dot{\gamma}_d
\]

For a given shear stress value across the shear zone, Eqs. (4) – (6) yield a strain rate profile. For a given strain rate value across the shear zone, a shear stress profile is obtained by solving Eq. (6) after substituting Eqs. (4) and (5) into it. Our approximation to the shear stress (Fig. 5b) follows the minimum of the two stress regimes and smoothes the transition zone.

3.3.2. Constant shear strain model \((\dot{\gamma}_{\text{cst model}})\)

The constant shear stress model results in a linear velocity profile across the shear zone (Fig. 6a) defined by the hangingwall and footwall velocities. The deformation is homogeneous and distributed across the shear zone, without localization, and with the same intensity in the “strong” or “weak” (thermally weakened) parts of the shear zone. Therefore, we expect the shear heating term (Eq. (21)) to be maximized using this kinematic model.

\[
\dot{\gamma} = \frac{\dot{\gamma}_{\text{bulk}}\tau^c}{\tau^c}
\]

The corresponding shear stress, variable across and along the shear zone, is determined at any points by substituting Eqs. (4) and (5) into Eq. (6) and solving it using Eq. (7). This approach insures a mass conservation in the model but induces large variations of shear stress across the shear zone due to the variations of rheological properties.

3.3.3. Constant shear stress model \((\tau_{\text{cst model}})\)

The motivation for the \(\tau\)-cst model is to augment the flow pattern by avoiding large stress variation across the shear zone and allowing for localization of the deformation. The localization has a direct impact on the distribution of heat produced by shear heating since large deformations mostly occur in “weak” parts of the shear zone. Thus, the \(\tau\)-cst model is used as end-member kinematic model minimizing the impact of shear heating as opposed to the \(\dot{\gamma}\)-cst model. A value of the shear stress \(\tau\), uniform across and variable along the shear zone, is obtained by integrating Eq. (6) across the shear zone (in the \(\dot{\gamma}\) direction, Fig. 6) and using the bulk strain rate \(\dot{\gamma}_{\text{bulk}}\) as a constraint:

\[
\int_0^D \dot{\gamma} dy' = \int_0^D (\dot{\gamma}_b + \dot{\gamma}_d) dy'
\]

\[
\dot{\gamma}_{\text{bulk}} = \dot{\gamma}_c + \tau^{n_b} A \exp(-E/RT)
\]

\[
\int_0^D \dot{\gamma} dy' = \int_0^D \left(\frac{\tau}{\tau_b}\right) \dot{\gamma}_c + \tau^{n_b} A \exp(-E/RT) \right) dy'
\]

**Table 3**

Material and rheological (Ranalli, 1995) parameters used in the numerical experiments.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Symbol</th>
<th>Unit</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravity</td>
<td>(g)</td>
<td>m.s(^{-2})</td>
<td>9.81</td>
</tr>
<tr>
<td>Friction coefficient</td>
<td>(\mu)</td>
<td>-</td>
<td>0.7</td>
</tr>
<tr>
<td>Pure fluid factor</td>
<td>(P_f)</td>
<td>-</td>
<td>0.4</td>
</tr>
<tr>
<td>Power law exponent</td>
<td>(n_b)</td>
<td>-</td>
<td>20</td>
</tr>
<tr>
<td>Gas constant</td>
<td>(R)</td>
<td>J.K(^{-1})</td>
<td>8.314</td>
</tr>
<tr>
<td>Pre-exponential multiplier</td>
<td>(A)</td>
<td>MPa.m(^{2}).s(^{-1})</td>
<td>(3.2 \times 10^{-12} \times 6.7 \times 10^{-8})</td>
</tr>
<tr>
<td>Activation energy</td>
<td>(E)</td>
<td>kJ.mol(^{-1})</td>
<td>(154E + 3/156 \times 10^3)</td>
</tr>
</tbody>
</table>

Fig. 5. a) Brittle failure approximated by power law relationship between shear stress and shear strain rate. The power law expression (Eq. (3)) approaches Byerlee’s brittle failure (where \(\tau_b(\dot{\gamma})\) is function of depth) for large power law exponent (\(n_b = 20\) in this study). b) Depth distribution of the shear stress within the shear zone \(\dot{\gamma}_{\text{cst model}}\) and \(\tau = 3\) MPa. A single stress profile is drawn for dry and wet quartzite at the initial conditions. The brittle-ductile transition is smoothed (illustrated by yellow) because of the use of approximation presented on Fig. 5a and Eq. (4). Lateral variation of the temperatures at the end of the exhumation (\(t^*\)) scatters the shear stresses for a given depth as illustrated by the gray shade for the wet quartzite simulation.
The strain rate profile is then calculated from Eqs. (4) and (5) and it results in localized deformation in weak parts. Depending on the dominant deformation mechanism, rock weakens at the upper part of the shear zone in the brittle regime (where $\tau_{\text{by}}$ is smaller, Eq. (3)) and at lower part of the shear zone in the ductile regime (where temperature is higher, Eq. (5)). Therefore, highly localized deformations occur along the boundary with the hangingwall in the upper part of the shear zone, and along the boundary with the footwall in the lower part of the shear zone (Fig. 6b).

The velocity profile resulting from the constant shear stress assumption (Couette flow, Fig. 6b) respects the no-slip conditions at the boundaries of the shear zone. However, significant rheological variations along the shear zone result in velocity fields characterized by a non-vanishing divergence and lead to a violation of mass and momentum conservation condition. To limit these effects, we augment the kinematic profile by adding a Poiseuille flow contribution (Fig. 6c) determined by keeping the integrated mass balance constant along the shear zone. This results in a non-zero pressure gradient in the shear zone. We note that due to adding the Poiseuille flow component the original assumption of the constant shear stress across the shear zone is not strictly satisfied. Yet, we choose to refer to this model as $\tau$-cst throughout the manuscript, as the strain

![Diagram](image-url)
rates related to the Poiseuille flow are significantly smaller than the total strain rates (Fig. 6c, d). The resulting velocity profile of the $\tau$-cst model is the sum of Couette and Poiseuille flows as illustrated Fig. 6d.

3.4. FEM strategy

The model domain is 80 km deep and 120 km wide (Fig. 3). We use a triangular mesh generated with Triangle (Shewchuk, 1996,
to discretise the model domain and accurately represent the boundaries separating materials of different thermal properties. The mesh is updated at each time step accordingly to the subsidence of the basin on the top of the model.

Eq. (1) is implicitly solved with a finite element code using the Streamline Petrov-Galerkin scheme. A similar discretisation has been documented and used for geodynamical problems and we refer to the paper of Braun (2003) for more technical details. Temperature and velocity fields are linearly interpolated using 3-nodes elements. The temperature-dependent thermal conductivity is calculated using temperatures from the previous time step. We consider zero flux on the lateral boundaries and Dirichlet boundary conditions on the top and bottom (Section 3.2).

Shear zone related calculations are performed on an auxiliary regular grid onto which the modelled temperature is projected. The regularity of this grid allows accurate integration of Eqs. (4) – (6) and/or (8) across the shear zone to define the stress field, strain and strain rate. The resulting velocities and shear heating are then interpolated back to the main grid each time step.

4. Results

We performed a systematic study by varying the time of the exhumation \( t^* \) (3, 5, 10, and 15 My), the shear zone rheology (wet and dry quartzite), the kinematic model of the shear zone \( \gamma \text{-cst} \) and \( \tau \text{-cst} \), and the maximum basin depth \( b_{\text{max}} \) (10 and 15 km). All the simulations are repeated twice, with and without accounting for shear heating.

4.1. Temperature evolution and peak conditions

For any \( t^* \), the exhumation of the footwall and the subsidence of the hangingwall leads to an unsteady temperature field. Thermal relaxation is reached by 20 to 30 My after \( t^* \). The different thermal stages of the model are illustrated in Fig. 7a-c. Note that for the similar initial and final geometry and for the same distribution of the thermal properties of rocks, the initial and final geotherms (Fig. 7a and c) will be equivalent. The peak temperature profile (Fig. 7e), however, depends greatly on the parameters tested in our numerical experiments.

To facilitate the comparison of our model results with geological observations (Section 2), we extract the peak temperature of each point during the evolution of the model up to thermal relaxation (Fig. 7e). The peak temperatures are reached at different time for different localities, often well after \( t = t^* \) (Fig. 7d).

4.2. Heat produced by shear heating

The distribution of total heat produced by shear heating within the detachment shear zone is presented in Fig. 8 (example presented for \( t^* = 3 \) My). The results for \( \gamma \text{-cst} \) model show a maximum heat produced in the upper 10-20 km, which corresponds to the largest stress field at the brittle-ductile transition. The results for \( \tau \text{-cst} \) model are characterized by two areas with high heat production (corresponding to areas of localized deformations), one in the upper brittle domain, and one in the lower ductile domain. Irrespective of the different spatial distribution of the heat production in the two kinematic models, the amplitude remains within the same range.

Shear heating may be locally two orders of magnitude higher than the heat produced by radioactive decay in the upper crust. This effect can be seen as a punctual heat source during the active deformation in the detachment zone.

4.3. Temperature anomaly produced by shear heating

We illustrate (Fig. 9) the temperature anomaly produced by shear heating by plotting the difference of modelled peak temperatures obtained with and without shear heating. For dry quartzite models and \( t^* = 3 \) Ma, shear heating can introduce more than 100 °C at the base of the basin. The impact depends on the rheology and on \( t^* \), but remains significant (within 30-80 °C) for slower exhumation and weaker rheology in the shear zone.

In the model configurations presented here, a value of the pore fluid factor of \( p_1 = 0.4 \) in Eq. (3) is assumed along the shear zone. The effect of pore-fluid pressure on the effective stress is still a matter of debate (e.g. Nur and Byerlee, 1971; Sibson, 1994). However, the impact of varying the pore fluid pressure within acceptable range is not critical with respect to thermal effects related to shear heating. The two kinematic models, \( \gamma \text{-cst} \) and \( \tau \text{-cst} \), were introduced in Section 3.3 as an upper and lower bound to estimate shear heating.

Fig. 9 shows that for the same parameters, \( \gamma \text{-cst} \) model is hotter than \( \tau \text{-cst} \) model. However, despite the large difference in the model velocity profiles (Fig. 6), the estimated temperature anomaly produced by shear heating is on the same order. A difference of approximately 30 °C separates at most the temperature anomaly obtained from both models, which shows that the style of deformation within the shear zone does not influence much the amplitude of the induced temperature anomaly in the system.

5. Comparison with geological observations

5.1. Shear strain partitioning

Fig. 10 shows the strains along the cross sections A, AA, B, and BB (Fig. 8) at the end of the simulations. Independent of the rheology and the cross section position, the strain across the shear zone, described by the \( \tau \text{-cst} \) model, localizes mostly along the upper part, beneath the hangingwall with maximum strain higher than 200 and along the basal part of the shear zone next to the footwall with maximum strain around 35-60. Strains higher than 100 are indeed speculated in the upper part of the NSDZ (Hacker et al., 2003). However, there is no convincing field evidence for high-strain environment at the base of the shear zone such as produced by the \( \tau \text{-cst} \) model. This ductile shear localization can be seen as an artefact of our simplified shear zone geometry (e.g., the strict rigidity of the footwall). The \( \tau \text{-cst} \) model also produces a strain gap in the middle of the shear zone, reducing the effective thickness of the shear zone. This is in contrast with the deformation pattern observed along the NSDZ, which exhibits continuous and large strain (20) over several kilometres (Andersen and Jamtveit, 1990).

In contrast to \( \tau \text{-cst} \), the \( \gamma \text{-cst} \) model satisfies the observations on continuous strain within the NSDZ, but lacks the high shear localization in the upper part of the detachment. Although the two models...
differ in the style of deformation and in correlation to observations, the resulting heat produced within the shear zone based on those models does not differ significantly. Thus, we can conclude that exact reproduction of the flow within the shear zone has limited importance while considering the thermal budget of the NSDZ.

5.2. Implications for the supra-detachment basins thermal history

In Fig. 11, we compare the modelled peak temperatures with the geological data (Section 2.2). Svensen et al. (2001) documented average peak temperatures for the entire Kvamshesten and Hornelen basins (horizontal lines on Fig. 11a-c) and 315 ± 15 °C for the Solund basin (horizontal lines on Fig. 11d-f). Souche et al. (2012) reported the local estimates of peak temperature of the sediments as a function of the distance to the detachment (data points on Fig. 11).

The maximum reported burial depth of the Devonian sediments is 9.1 km for the Kvamshesten and the Hornelen basins and 13.4 km for the Solund basin (Svensen et al., 2001). Thus, we compare the geological data with the modelled peak temperatures taken at 9.1 km (Fig. 11a-c) and 13.4 km (Fig. 11d-f). Fig. 11 presents the modelled temperature curves for three different times (Fig. 11a-c) and two kinematics within the shear zone (γ-cst and τ-cst) in each sub-figure.

All the results obtained without shear heating (Fig. 11c and f) show low peak temperatures and fail to reproduce the geological data. In addition, these modelled temperatures show a relatively small lateral variation through the basins. The lateral temperature variation does not exceed 40 °C in a range of 10 km away from the detachment contact. That low thermal gradient illustrates the role played by the hot rising footwall on the heating of the basin. This heat source is not sufficient to induce significant variations of the peak temperatures in the supra-detachment basins.

Fig. 11. Comparison of model temperatures and geological data for the Kvamshesten and the Hornelen basins (a-c, max. burial depth 9.1 km) and for the Solund basin (d-f, max. burial depth 13.5 km). Data from Svensen et al. (2001) present average peak temperatures for the entire basins, Souche et al. (2012) presents local peak temperatures. γ-cst and τ-cst corresponds to the two kinematic models used for the shear zone. t* is the time window for the development of the detachment zone. (c) and (f) present the model results obtained without shear heating; (b) and (e) present the model results obtained with wet quartzite rheology for the shear zone, and (a) and (d) with dry quartzite. The maximum depth of the basin (hmax) was set to 10 km in (a-c) and 15 km in (d-f).
The models including shear heating (Fig. 11a, b, d and e) show significantly higher peak temperatures and a stronger dependence on the proximity to the shear zone. Wet quartzite models, however, do not show good qualitative agreement with observations (Fig. 11b and e). Dry quartzite models (Fig. 11a and d) result in higher temperatures in the shear zone. The model results were tested against peak temperatures of Svensen et al. (2001) and to the local estimates of Souche et al. (2012).

The results obtained with dry quartzite (Fig. 11a and d) show about 100 °C or more of lateral variation of peak temperatures within 10 km distance from the detachment. This clearly indicates the significance of shear heating by locally increasing the temperature in the sediments at the vicinity of the detachment fault. These models reproduce the geological data of 295 °C in the Hornelen basin. In comparison to a maximum of 220 °C calculated without shear heating, shear heating alone accounts for as much as 25% of the thermal budget of the sediments.

Figs. 9 and 11 show that the influence of shear heating depends greatly on the distance to the shear zone, and is in accordance with the geological observations from Souche et al. (2012). The numerical results of the present study show that rocks located more than 5-10 km away from the detachment are not affected by shear heating.

5.3. PT-t path and retrograde overprint of rocks within the shear zone

The studies of PT-t paths representative for the exhumation of HP units within the NSDZ show isothermal or temperature-increasing decompression (Hacker et al., 2003; Johnston et al., 2007; Labrousse et al., 2004) illustrated in Fig. 12. Can shear heating lead to temperature increase or support isothermal condition during exhumation of the HP units? Below we demonstrate that shear heating is unlikely an explanation for such an effect and other mechanisms should be found to explain the phenomenon.

We compare these observations with our model results. We select several markers with initial depth between 60 and 25 km and reconstruct their PT-t paths. In our estimations, we use lithostatic condition to calculate pressure, which makes our pressure estimates as upper bounds (as the background extensional stress may induce underpressure) and conclusions of this section more robust. Markers initially shallower than 45 km exhibit strong thermal overprint related to shear heating. The temperature increase during the exhumation of these markers, up to 100 °C, compares with some of the geological data but is observed at much shallower depths in the model. In contrast, markers initially deeper than 45 km, comparable to the initial decompression depth of the HP units, monotonically cool during decompression.

The model results show the limited effect of shear heating in producing temperature increase at the early stage of the decompression of the HP units found in the NSDZ. Either these units were originally at shallower depth and were subjected to overpressure (Petrini and Podladchikov, 2000), or were possibly affected by heat generation during retrograde metamorphism reactions (Connolly and Thompson, 1989; Haack and Zimmermann, 1996).

6. Conclusions

We have developed a simple numerical model to understand the thermal budget and the role of shear heating in crustal-scale extensional detachments. Using geological observations of the NSDZ we have developed a conceptual model of evolution of such a system. The model was augmented by constraining the non-steady-state geotherm associated with the end of Caledonian orogeny. We have built two simplified kinematic models for the deformation within the shear zone. The model results were tested against peak temperature conditions documented within three Devonian basins of western Norway. Our numerical study has shown:

1. Shear heating within the highly-deformed NSDZ can generate a temperature increase up to 100 °C in the detachment shear zone and is a necessary mechanism to explain the elevated peak temperatures within the Devonian basins of western Norway.

2. Shear heating may account for 25% of the thermal budget of the supra-detachment basins and may affect the peak temperatures 5 to 10 km away from the detachment shear zone.
3. The amount of shear heating produced in the system is predominately dependent on the exhumation rate and the rheological parameters of the rocks forming the shear zone.

4. Shear heating has a limited impact on the documented heating during decompression of the HP units in the NSDZ.

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