Quantifying magnitudes of shear heating in metamorphic systems

Calvin A. Mako*, Mark J. Caddick

Virginia Tech, Department of Geosciences, 926 West Campus Dr., Blacksburg, VA, USA

ARTICLE INFO

Keywords:
Shear heating
Heat diffusion modeling
Heating mechanisms
Crustal rheology

ABSTRACT

Estimated magnitudes of stress and strain rate in crustal rocks suggest that shear heating should contribute significantly to the thermal budget of deforming metamorphic systems. A simple one-dimensional thermal model is used to calculate magnitudes of shear heating in ductile shear zones based primarily on quartz flow laws, with consideration of additional models for wet feldspar and biotite. We calculate shear heating for likely ranges of key parameters, so that constraints can be placed on particular natural systems based on inferred stresses, durations of deformation, shear zone widths and metamorphic temperatures. By exploring large parameter spaces, our results should be appropriate for estimating shear heating experienced by a wide range of shear zones. Heating due to ductile deformation is highly dependent on these parameters and ranges from negligible to tens of degrees for most plausible deformation scenarios, and up to 200 °C in extreme cases. Results also predict that wet feldspar rheologies should produce significantly less shear heating than quartz and biotite schist rheologies. The width of a shear heating thermal anomaly can greatly exceed the width of the shear zone itself. For example, in a 0.5 km wide shear zone that experiences a maximum temperature increase of 100 °C in its core after approximately 1 Myr of deformation, a region ~1.0 km wide heats by at least 90 °C. Thermal weakening in our quartz rheology models is very rapid, so differential stresses on the order of 100 MPa should be short lived. These results can be used to better constrain magnitudes of shear heating in natural systems and can be extrapolated to infer the contribution of shear heating to evolving orogenic systems. We apply our results to the strike-slip Davenport and Norumbega shear zones (Australia and USA, respectively), also making tentative predictions for the thrust-sense Himalayan Main Central Thrust.

1. Introduction

One of the fundamental goals of metamorphic geology is understanding and describing the sources of thermal energy that cause metamorphism in the Earth’s crust. It is well known that in orogenic settings, simple numerical models of crustal thickening and heating often do not provide sufficient heat to reproduce the highest temperatures recorded by metamorphic rocks (e.g. Jamieson et al., 1998; Penniston-Dorland et al., 2015). Shear heating is one mechanism that is commonly invoked to explain the thermal dynamics of orogenic systems as inferred from metamorphic rocks (e.g. England and Molnar, 1993; Harrison et al., 1997a, 1997b; Nabelek et al., 2001; Duprat-Oualid et al., 2013). Heat produced by mechanical work contributes to the thermal budget of deforming metamorphic systems in magnitudes that can greatly exceed that of radiogenic heat production, depending on ambient stresses and strain rates (Fig. 1; see also Nabelek et al., 2010). In this contribution, we present calculated magnitudes of shear heating based on published flow laws for quartz (Hirth et al., 2001; Platt and Behr, 2011), wet feldspar (Rybacki et al., 2006) and biotite schist (Shea and Kronenberg, 1992), accounting for thermal softening with progressive deformation. Models assume a geometrically simple strike-slip shear zone and aim to quantify the time and length scales of shear heating over a wide range of deformation conditions.

Estimates of shear heating magnitudes are often produced in large scale geodynamic models (e.g. Leloup et al., 1999; Burg and Gerya, 2005; Schmalholz and Duretz, 2015) or theoretical studies (e.g. Brun and Cobbold, 1980; England and Molnar, 1993; Nabelek et al., 2001, 2010; Duprat-Oualid et al., 2015; Platt, 2015a). These give a general sense of how much mechanical heat can be produced during crustal deformation and the significance of shear heating for crustal evolution, but are of less utility for specific scenarios in which an estimate of the shear heating contribution to a metamorphic path is needed. We present a generalized, simplified approach that is constrained in terms of parameters such as shear zone width, plate velocity, duration of deformation and shear zone temperature, which can be relatively easily quantified in natural systems. Conceptually, our model represents a strike-slip shear zone, but the resulting magnitudes of shear heating are applicable to other kinematic scenarios given appropriate...
The production of heat ($H_e$) from the conversion of mechanical energy is related to deviatoric stress and strain rate during deformation. We define shear heating as the temperature change that results solely from the balance of this mechanism of heat generation and thermal diffusion. In a simple shear geometry $H_e$ is defined by

$$H_e = 2 \sigma_{xy} \dot{\varepsilon}_{xy}$$  \hspace{1cm} (1)$$

where $\sigma_{xy}$ is the shear stress and $\dot{\varepsilon}_{xy}$ is the shear strain rate. $H_e$ in more complex, non-simple shear, strain geometries can be greater, as more components of the deviatoric stress and strain rate tensors are non-zero (see discussion and simplifications in Turcotte and Schubert, 2002; Burg and Gerya, 2005).

Both stress and strain rate can be quantified for naturally deformed rocks. Behr and Platt (2014) presented a compilation of crustal stresses calculated primarily by the grain size piezometer of Stipp and Tollis (2003) with the Holyoke and Kronenberg (2010) modification. Fig. 1 highlights that it is not exceptional for stress estimates and associated strain rates to result in mechanical heat production that exceeds typical radiogenic heat production in the crust by several orders of magnitude. Depending on the duration of deformation, the volume of deforming crust and the magnitude of thermal weakening, temperature changes due to deformation are expected to be significant.

Many numerical modeling studies demonstrate the importance of shear heating in crustal geodynamics. It is often shown that the contribution from shear heating can be on the order of 100–200 °C (e.g. Leloup et al., 1999; Burg and Gerya, 2005; Souche et al., 2013; Schmalholz and Duretz, 2015; Duprat-Oualid and Yamato, 2017), and many studies have shown that shear heating is necessary to explain the temperatures recorded by metamorphic rocks (Molnar and England, 1990; Peacock, 1992; England and Molnar, 1993; Leloup and Kienast, 1993; Harrison et al., 1997a, 1997b; Camacho et al., 2001; Nabelek et al., 2001, 2010; Souche et al., 2013). Several studies have pointed out the potential for the deforming lithospheric mantle to contribute significantly to the thermal budget of orogenic belts (Kincaid and Silver, 1996; Leloup et al., 1999). Shear heating is thus thought to be a key component of dynamic processes including shear zone formation and localization (Montési and Zuber, 2002; Kaus and Podladchikov, 2006; Duretz et al., 2015; Jacquet and Schmalholz, 2017), folding and thrusting (Burg and Schmalholz, 2008; Hobbs et al., 2008), and deep earthquakes (Kelemen and Hirth, 2007; John et al., 2009; Thielmann et al., 2015; Prieto et al., 2017; Thielmann, 2017).

High values of mechanical heat production are apparently common, and many theoretical studies demand significant heat production, so it should be expected that field-based estimates of the magnitudes of shear heating are abundant. However, very few field-based studies in metamorphic systems have conclusively quantified how much mechanical heat is produced during deformation. Field-based evidence of shear heating during brittle deformation is relatively common (e.g. Ben-Zion and Sammis, 2013; Fulton et al., 2013; Evans et al., 2014; Maino et al., 2015). However, the most well-known example is the San Andreas fault, where minimal heat flow anomalies suggest shear heating may only locally be important, if at all (Lachenbruch and Sass, 1980; Thatcher and England, 1998; Saffer et al., 2003). Conversely, Gao and Wang (2014) show that heat flow measurements across subduction zones are compatible with significant shear heating. Field-based examples of shear heating in ductile metamorphic systems that are independent of numerical modeling are rare (Scholz et al., 1979; Scholz, 1980; Leloup and Kienast, 1993; Camacho et al., 2001; Sturm, 2017; Wei et al., 2017), and many remain somewhat equivocal. The lack of obvious natural examples tends to suggest to many geoscientists that shear heating is not particularly important in natural metamorphic systems. We emphasize that although it is rare to see natural examples of shear heating in the ductile regime, many numerical modeling studies and basic rock mechanics suggest that it should be important, raising the question of why evidence for such a fundamental process is generally lacking in the rock record.

The dearth of observations clearly revealing a natural shear heating signature is partly explained by the fact that many localized shear zones occur in terranes that have higher grade pre-existing metamorphic assemblages. Even if shear heating is significant (~100 °C) it may not...
produce an obvious metamorphic anomaly relative to the host rocks. Indeed, we might expect a heat-producing shear zone to appear as a region of lower metamorphic grade if that zone focuses deformation and fluid availability in previously metamorphosed rocks. This presents the additional challenge of estimating the ambient temperature of syn-or pre-deformational regional metamorphism that can be compared to temperatures during shear heating. To understand the necessary scale of recrystallization, consider the characteristic length scale of thermal diffusion ($L$)

$$L = \sqrt{\nu t}$$  

where $\nu$ is the thermal diffusivity ($1 \times 10^{-6}$ m$^2$/s), $t$ is time, and $L$ is the average length scale of thermal perturbation. If a shear zone is active for 5 Ma, the characteristic length scale of thermal diffusion from shear heating is 14.8 km. Syn-deformational recrystallization to lower grade metamorphic assemblages would thus have to occur ~15 km away from the actively deforming zone to infer an ambient temperature and measure the magnitude of shear heating. High grade host rocks therefore generally obscure evidence of shear heating such that shear heating is 14.8 km. Syn-deformational recrystallization to lower grade metamorphic grade if that zone focuses deformation can be inherently hard to detect solely on thermometric techniques and changes in metamorphic grade.

For example, the Carboniferous age greenschist facies mylonite zones of the strike-slip Norumbega fault system in Maine, USA (Ludman, 1997; West and Hubbard, 1997; Sullivan et al., 2013; Price et al., 2016) are hosted in amphibolite to migmatite grade rocks associated with the Devonian Acadian orogeny (West et al., 2003; Tucker et al., 2001). Even if 100 °C of Carboniferous shear heating occurred in these mylonite zones, this heating may not be readily recorded petrologically without broad-scale rehydration between the two events. Relatively pristine high grade (600–700 °C) host rocks are often even preserved close to the mylonite zones (350–400 °C), with variable and incomplete retrogression occurring in the surrounding tens of kilometers.

Thrust and normal sense shear zone geometries also present challenges for estimating the contribution of shear heating to a natural metamorphic path. A cooler hanging wall or footwall might prevent a clear fault-localized temperature increase from being observed. Because temperatures naturally vary continuously across normal or thrust sense shear zones, an initial temperature that can be compared to shear zone temperatures is difficult to establish (England et al., 1992). Furthermore, temperatures in the hanging wall, footwall and shear zone may not be petrologically recorded contemporaneously, further obscuring a thermal anomaly centered on the shear zone. However, an abundance of numerical models has shown how shear heating can influence a thrust system (Bird et al., 1975; Molnar and England, 1990; England et al., 1992; Duprat-Oualid and Yamato, 2017) and have demonstrated that field-based observations can testify to the importance of shear heating (England and Molnar, 1993; Harrison et al., 1997b; Nabelek et al., 2001, 2010).

It is often supposed that fluid flow in metamorphic systems would damp a shear heating signature such that it would not be measurable or significant. Burg and Gerya (2005) comment that “[b]ecause fluid and rock have comparable heat capacities, the mass of circulating fluids passed through rocks must be comparable to the mass of the rocks themselves to have efficient cooling effects.” Furthermore, fluid flow does not necessarily remove heat from a metamorphic system, but can transport and redistribute heat depending on the situation (Bickle and McKenzie, 1987). In a case where dehydration reactions cause upward fluid flow, fluid flow could serve to enhance an apparent shear heating signature or heat flow anomaly in overlying rocks that receive fluid and heat input. In a case where fluid flow was directed across a heat producing structure, the shear heating signature would tend to be modified and attenuated. Infiltration of cooler meteoric fluid could mitigate the effects of shear heating. Evidence for meteoric fluid in fault zones below the brittle-ductile transition has been reported in normal-sense detachment faults in the North American Cordillera, central Alps, Turkey and the Himalaya (e.g. Fricke et al., 1992; Holk and Taylor, 2007; Campani et al., 2012; Hetzel et al., 2013; Gébelin et al., 2015, 2017). Saffer et al. (2003) showed, using a coupled heat and fluid flow model, that groundwater circulation does not significantly affect temperature change due to shear heating in the upper crust (see also Lachenbruch and Sass, 1980), so it is unlikely that fluid flow in the ductile middle to lower crust, where porosity and permeability are expected to be lower, will significantly affect the magnitude of shear heating.

Given all of the above considerations, it is unlikely that constraints on magnitudes of shear heating can be obtained solely on the basis of thermometric methods. In order to overcome this challenge, we present a method for making field-based estimates of the magnitudes of shear heating using estimates of shear stress, strain rate, duration of deformation, rheology and basic shear zone geometry. This approach allows reasonable estimates of shear heating to be compared with inferences made from various thermometry techniques and allows assessment of the magnitudes of shear heating in metamorphic systems where localized metamorphic grade changes are not evident.

3. Methods

3.1. Numerical setup

We have employed a simple one-dimensional implicit finite difference model with strike-slip kinematics to calculate the magnitudes of shear heating for various shear zone parameters (Fig. 2A-C). A 1D model is sufficient for this scenario in which shear zone width is expected to be much smaller than the length of the shear zone, and the
The steepest thermal gradient is supposed to be perpendicular to the shear zone. We use a forward-time, centered-space setup to solve the heat flow equation

\[
\frac{dT}{dt} = \kappa \frac{dT}{dx^2} + \frac{H_x}{\rho C_p}
\]

(3)

where \(x\) is the distance across the model domain, \(T\) is temperature, \(H_x\) is the rate of thermal energy production by deformation (Eq. (1)), \(\rho\) is density, and \(C_p\) is heat capacity. The thermal conductivity is defined by the relationship

\[
\kappa = \frac{K}{\rho C_p}
\]

(4)

where \(K\) is the thermal conductivity (see below). The model domain consists of a fixed width shear zone at the center, over which a heat production value is assigned based on deformation parameters (Fig. 2A, B). A time step of \(1.584 \times 10^4\) years \((5 \times 10^{11} s)\) was used for all models and the model domain was discretized at a spatial resolution between 50 and 5000 m. The full width of the model domain is 200 km, which is wide enough that the shear heating anomaly does not interact significantly with the constant temperature boundary condition for all modeled shear zone widths. The spatial resolution was calculated by dividing the width of the shear zone by 10. This approach ensures that nodes are always placed at the shear zone boundaries and that the same number of nodes describes the shear zone in each model. Without this measure, the number of heat-generating nodes that define the shear zone varies incrementally rather than continuously. For example, if the spatial resolution was fixed at 100 m, shear zones of width 1010 and 1100 m would have the same number of nodes within the shear zone and therefore functionally the same width. A constant number of nodes in the shear zone ensure that heat generation scales exactly with shear zone width. Tests with substantially higher spatial and temporal resolution did not significantly improve or modify the results (see Appendix 1). A temperature increase is added to the model across the width of the shear zone at each time step according to the amount of mechanical heat generated during that time step (Eq. (1) and second term of Eq. (3)) at each node, followed by the thermal diffusion calculation. The strain rate is calculated using the relative plate velocity \((V)\) and shear zone width \((w)\) as follows:

\[
\dot{\varepsilon}_{xy} = \frac{V}{2w}
\]

(5)

See Platt (2015a). Our model is thermo-kinematic rather than thermomechanical. That is, we impose a velocity field instead of allowing the shear zone width to be controlled by the shear zone rheology. This allows us to treat shear zone width as an independent variable and study the impact of shear zone width on the magnitude of shear heating.

Stress was calculated using published flow laws for quartz (Platt and Behr, 2011), feldspar (Rybacki et al., 2006) and biotite schist (Shea and Kronenberg, 1992). The quartz and feldspar flow laws have the following form.

\[
\dot{\varepsilon}_{xy} = A f_{H_2O}^{-M} \sigma_\theta^{q} \exp\left(\frac{-Q + P_n}{RT}\right)
\]

(6)

In the above, \(A\), \(r\), \(M\), \(n\), \(Q\) and \(\varepsilon\) are material constants, \(f_{H_2O}\) is water fugacity (MPa), \(d\) is grain size (meters), \(\sigma_\theta\) is differential stress (Pa), \(P\) is lithostatic pressure, \(R\) is the gas constant, and \(T\) is temperature (K). See text below for details on water fugacity and pressure. The form of the biotite schist flow law is as follows.

\[
\dot{\varepsilon}_{xy} = C \exp(\alpha \sigma_d) \exp\left(\frac{-Q}{RT}\right)
\]

(7)

\(C\), \(\alpha\) and \(Q\) are material constants. The parameters used in the above equations are shown in Table 1. Because experimental flow laws are based on axial compression tests it is not appropriate to use values of shear stress directly from flow laws (e.g. Behr and Platt, 2013). The differential stress from flow laws and piezometric relationships is related to the shear stress as follows for plane stress scenarios (Behr and Platt, 2014).

\[
\sigma_{xy} = \frac{1}{\sqrt{3}} \sigma_d
\]

(8)

Values of shear strain rate and shear stress calculated using these equations can be used to calculate rates of heat generation due to deformation (\(H_x\), Eq. (1)). At each time step the stress is recalculated with the temperature and grain size of the previous time step, which incorporates thermal softening and microstructural changes (Fig. 2). The relationship between grain size \((d)\) and differential stress \((\sigma_d)\) is as follows:

\[
d = j \sigma_d^{-b}
\]

(9)

See Table 1 for input parameters (Stipp and Tullis, 2003; Holyoke and Kronenberg, 2010; Post and Tullis, 1999). These methods are similar to those of Platt (2015a).

To calculate the water fugacity required for the flow laws, we used the “Fluids” application of the Perple_X package (Connolly, 2005; v6.7.6) with the data of Pitzer and Sterner (1994) at \(X_{CO_2} = 0\). Setting \(X_{CO_2} > 0\) would reduce water fugacity, thereby increasing shear stresses and magnitudes of shear heating. Fugacity was calculated at 50 MPa and 10°C intervals and linearly interpolated for use in the thermal model (Fig. 3). Water fugacity was updated at each time step and at every position in the thermal model using this interpolation, such that it was appropriate for the evolving thermal field. The flow laws utilized here are pressure dependent, so lithostatic pressure was calculated using the initial temperature of each model \((T)\) assuming a 20°C/km geothermal gradient and a temperature and pressure dependent density (see below). Our use of fugacity data in this way assumes water saturation at all times.

Heat capacity, density and thermal conductivity were updated at each time step and at every position in our models. Heat capacity and density were calculated using Perple_X (Connolly, 2005). Equilibrium metamorphic assemblages were calculated for the range of temperatures and pressures experienced in our thermal models and for the bulk compositions of Westerly granite (Goldich and Oslund, 1956 for quartz

Table 1
Parameters used in quartz, feldspar and biotite flow laws. Pa is Pascals, m is meters, s is seconds, J is joules, mol is moles. Pressure, water fugacity, initial temperature, shear zone width, duration and relative plate velocity are defined in each model run. References in the table are for the flow law parameters. Parameters j and b are from Holyoke and Kronenberg (2010) for quartz and Post and Tullis (1999) for feldspar. Note: we are making the implicit assumption that rocks that are described by a quartz or wet feldspar rheology have a composition (and thus Cp and density) that are approximated by the composition of Westerly granite. A similar assumption is made for biotite-rich rocks.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Quartz</th>
<th>Wet feldspar</th>
<th>Biotite</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A) ((\text{Pa}^{-m\text{m}^2\text{s}^{-1}}))</td>
<td>2.75 \times 10^{-34}</td>
<td>2.00 \times 10^{-25}</td>
<td>–</td>
</tr>
<tr>
<td>(C) ((\text{s}^{-1}))</td>
<td>–</td>
<td>–</td>
<td>1.40 \times 10^{-10}</td>
</tr>
<tr>
<td>(\alpha) ((\text{MPa}^{-1}))</td>
<td>–</td>
<td>–</td>
<td>0.15</td>
</tr>
<tr>
<td>(r)</td>
<td>1</td>
<td>1</td>
<td>–</td>
</tr>
<tr>
<td>(M)</td>
<td>1</td>
<td>3</td>
<td>–</td>
</tr>
<tr>
<td>(n)</td>
<td>3</td>
<td>1</td>
<td>–</td>
</tr>
<tr>
<td>(Q) ((\text{J mol}^{-1}))</td>
<td>105,500</td>
<td>159,000</td>
<td>89,000</td>
</tr>
<tr>
<td>(v) ((\text{m}^3\text{mol}^{-1}))</td>
<td>–</td>
<td>38 \times 10^{-6}</td>
<td>–</td>
</tr>
<tr>
<td>(j) ((\text{in Pa}^{-3}))</td>
<td>2451</td>
<td>55.5</td>
<td>–</td>
</tr>
<tr>
<td>(b)</td>
<td>1.26</td>
<td>0.66</td>
<td>–</td>
</tr>
<tr>
<td>Cp and density model</td>
<td>Westerly Granite</td>
<td>Westerly Granite</td>
<td>Average Pelite</td>
</tr>
</tbody>
</table>
and feldspar flow laws) and a biotite-rich pelite (Palin et al., 2016, their Table 3 average composition). The Holland and Powell (2004) database was used with the following solution models: feldspar (Fuhrman and Lindsley, 1988), muscovite (Coggon and Holland, 2002; Auzanneau et al., 2010), biotite (Powell and Holland, 1999), chlorite (Holland et al., 1998) and garnet (White et al., 2007). Heat capacity and density was used with the following solution models: feldspar (Fuhrman and Lindsley, 1988), muscovite (Coggon and Holland, 2002; Auzanneau et al., 2010), biotite (Powell and Holland, 1999), chlorite (Holland et al., 1998) and garnet (White et al., 2007). Heat capacity and density data (Fig. 3). Note we have not included melting in our density, heat capacity and feldspar laws (al)ow laws in Fig. 4. The wet di-water fugacity, density and heat capacity data used to calculate strain rates, stresses and thermal diffusivity as a function of temperature. Green and blue curves are the product of heat capacity and density. A) and B) are calculated for the Westerly granite composition of Goldich and Oslund (1956), C) and D) are calculated for a biotite-rich pelite composition from Palin et al. (2016). Density and heat capacity data are smoothed with a Gaussian filter to prevent thermal model instability. Curves in B) and D) are plotted for a pressure of 550 MPa. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

![Fig. 3. Water fugacity, density and heat capacity data used to calculate strain rates, stresses and thermal diffusivity as a function of temperature. Green and blue curves are the product of heat capacity and density. A) and B) are calculated for the Westerly granite composition of Goldich and Oslund (1956), C) and D) are calculated for a biotite-rich pelite composition from Palin et al. (2016). Density and heat capacity data are smoothed with a Gaussian filter to prevent thermal model instability. Curves in B) and D) are plotted for a pressure of 550 MPa.](image-url)

3.2. Assumptions associated with this methodology

Before discussing results and interpretations it is useful to clarify model assumptions and their implications. Our models represent extreme simplifications of natural shear zone processes. Most importantly in this regard, simulations use 1) a monomineralic flow law, ignoring complex rheological changes, 2) a constant shear zone width and temporally and spatially uniform strain rate, 3) simple shear strain geometry, 4) strike-slip kinematics, ignoring the potential advection of heat around dip-slip shear zones, and 5) a perfect transformation of mechanical work into heat energy without considering other microstructural energy sinks. Each of these is discussed in more detail below. We have also limited our model durations to 25 Myr because this exceeds the lifespan of most shear zones. At timescales approaching 25 Myr, 1D calculations are likely inappropriate because the temperature anomaly would be influenced by the surface boundary condition. We have endeavored to make the model as simple as possible, so that the results are generalizable to many tectonic scenarios, and thus encourage the reader to consider the assumptions and simplifications discussed below before applying our models to a particular set of deformation conditions.

Our most significant assumption is that we use flow laws that are appropriate for the deformation of natural shear zones. We used the dynamic recrystallization-assisted quartz dislocation (DRX) creep flow law of Platt and Behr (2011). DRX creep uses the parameters of the Hirth et al. (2001) flow law, but incorporates a grain size dependence to account for the importance of surface energy driven grain boundary migration in dynamic recrystallization. Shear heating simulations for DRX creep are compared with the Hirth et al. (2001), and Gleason and Tullis (1995) quartz flow laws in Fig. 4. The wet diffusion creep feldspar
Fig. 4. Across shear zone temperature profiles and the temporal evolution of several parameters. In the left panels (A, C, E, G) shear zone width, flow law (DRX Creep) and initial temperature are held constant and plate velocity is varied. In the right panels (B, D, F, H) plate velocity and shear zone width are constant, while flow law and initial temperature are varied. Duration is 5 Myr in all models. A) Temperature profiles for shear heating anomalies generated at 1 (blue), 3 (green) and 5 (red) cm/year from an initial temperature of 300 °C after 5 Myr of deformation. B) Temperature profiles generated using various flow laws at 5 cm/year and an initial temperature of 300 °C after 5 Myr. C–D) Temperature-time evolution at the shear zone center for 1–5 cm/year plate velocities (C) and various flow laws (D). E–F) Temporal evolution of shear stresses in the shear zone center. G–H) Evolution of heating rate due to shear heating. Note that high shear stresses and heating rates are very short lived. Dashed curves in D, F and H were calculated with initial temperatures of 400 °C at 5 cm/year and solid curves (excepting the black curve for feldspar) were calculated with initial temperatures of 300 °C at 5 cm/year. The feldspar curve (solid black) was calculated from an initial temperature of 500 °C at 5 cm/year. Feldspar typically deforms brittly at 300–400 °C. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
flow law of Rybacki et al. (2006) is used to model shear heating in feldspar controlled rheologies. We have assumed that grain size is controlled by stress during diffusion creep and our code simultaneously solves for differential stress and grain size. The biotite schist flow law of Shea and Kronenberg (1992) is used to model micaceous rocks. The bulk composition of the samples used in deformation experiments (Shea and Kronenberg, 1992) is rather unusual, so we have elected to calculate thermo-chemical data for an average biotite-rich composition from Palin et al. (2016). Our models also do not account for the rheological impact of interconnected weak phases or polymineralic rocks (e.g. Law et al., 2013; Behr and Platt, 2014; Platt, 2015a; Thielmann et al., 2015).

These factors are known to be important for the strength of natural shear zones, but using a monomineralic flow law to describe the strength and rheology of the crust or mantle is a common simplification (e.g. Law et al., 2013; Behr and Platt, 2014; Platt, 2015a; Thielmann, 2017) that allows us to map out the first-order expectations of shear heating in natural systems. To the degree that it is realistic to extrapolate from experimental to natural conditions, these flow laws approximate the rheologies that are expected for middle to upper crustal rocks.

Our model assumes that the shear zone is of fixed width, which is appropriate for some, but not all, natural shear zones. Constant width shear zones fall into the Type 3 category of Fossen and Cavalcante (2017), which are characterized by uniform strain rate across the shear zone. Both dynamic recrystallization and progressive shear zone thickening favor the maintenance of shear zone thickness or thinning (Type 2 or 3) over time. Strain hardening tends to lead to shear zone thickening (Type 1). Because we employ a constant shear zone width, our model setup is thus consistent with expectations for natural shear zones in which strain softening occurs (Fossen and Cavalcante, 2017). Heating in our model thus leads to decreased shear stresses in the shear zone center relative to its margins, in contrast to the model of Platt (2015a) in which force balance equations dictate uniform stresses and variable strain rates throughout the constant width shear zone (Platt et al., 2008).

We have employed a simple shear strain geometry in all of our models. As mentioned above, more complex strain geometries that involve, for example, a component of flattening, will have greater magnitudes of mechanical heat production (Burg and Gerya, 2005). Our results thus represent a minimum estimate for deformation scenarios with significant additional components of non-simple shear or non-plane stress.

The model is most appropriate for metamorphism during strike-slip deformation, but can also place constraints on the maximum contribution of shear heating in thrust and normal sense shear zones. Dip-slip scenarios likely feature complexly changing thermal boundary conditions and a significant advective component, unlike the present model, which assumes a single ambient temperature orthogonal to the fault at any given depth. Advection will tend to progressively heat or cool a fault zone, and in both cases we can predict how much shear heating would occur relative to our simplified model. Additional heating of a fault zone, for example associated with advection by thrusting hanging-wall blocks (Le Fort, 1975), further weakens shear zone rocks, reducing their capacity to produce heat for any given set of strain rate, strain duration and strain distribution. In contrast, cooling of the local environment, for example by footwall underthrusting, can effectively strengthen rocks and lead to an increase in the total shear-heat generation in a fault zone. In this case, although the heat generation is increased, the total temperature change of the fault is mediated by the advective cooling. Interaction with a lower temperature boundary, such as the Earth’s surface, would have the same effect. In both cases, our models therefore represent a maximum estimate of the contribution of shear heating to the temperature evolution around dip-slip faults. Explicit consideration of these factors is the subject of ongoing work.

Microstructure energy sinks are not included in our model. Dislocations in crystals store elastic strain energy, so any proportion of the work of deformation that is used to increase the number of intracrystalline or grain boundary dislocations is not dissipated as heat. However, the total microstructural stored energy is often considered insignificant compared to the total energy budget of a deforming system (Platt, 2015a; Thielmann et al., 2015).

In summary of our methods, we model a constant width, uniform strain rate shear zone in which shear stresses are calculated with a temperature dependent flow law. An implicit finite difference scheme is used to solve the heat conduction equation, with shear stress, viscous
heath generation and thermal diffusivity updated at each time step to simulate progressive shear zone weakening. Our results are most applicable to strain-softening, strike-slip shear zones in the middle to lower crust undergoing simple shear deformation, where they should provide a first order constraint on the magnitudes of shear heating in crustal shear zones.

3.3. Output representation: figure construction

The results of several individual model runs are shown in Fig. 4 to illustrate the initial output obtained from our model. The left panels (Fig. 4A, C, E, G) are results for varying convergence velocities (1–5 cm/year) at an initial temperature of 300 °C and shear zone width of 1 km with a DRX Creep rheology. The right panels (Fig. 4B, D, F, H) compare several flow laws at initial temperatures of 300 °C and 400 °C, with a fixed convergence velocity of 5 cm/year and shear zone width of 1000 m. 5 cm/year is the maximum tectonic velocity used in our models and best illustrates the difference between flow laws.

Fig. 5 demonstrates how multiple calculations such as those shown in Fig. 4 are combined to illustrate the magnitudes of shear heating developed over parameter spaces that would be appropriate for evaluating individual shear zones that might be encountered in the field. Figs. 6, 7 and 8 consist of 40 × 40 arrays contoured for temperature and stress experienced at the shear zone center for a given shear zone width, tectonic velocity, duration of deformation and flow law. These figures consist of a series of 25 Myr model runs (as in Fig. 4) with shear zones of varying fixed widths. Over the course of each model run the maximum model temperature \( T_{\text{max}} \), shear stress at the center of the model and width of the resultant thermal anomaly (defined below) were recorded. This 40 × 40 resolution adequately captures the variability of shear heating across the parameter spaces of interest. Fig. 9 consists of 20 × 20 arrays in which each ‘pixel’ represents an individual finite difference run of a certain duration with a fixed shear zone width, but with different initial temperatures and strain rates.

It is also valuable to consider the width of the thermal anomaly produced by shear heating. The widths of the thermal anomalies corresponding to 90% and to 50% of the maximum temperature change experienced in the core of a shear zone \( (\Delta T_{\text{max}}) \) were recorded at the same time intervals as \( T_{\text{max}} \) and shear stress (Fig. 10). For example, if the maximum temperature in the model domain at a given time step is 400 °C after heating from an initial temperature of 300 °C \( (\Delta T_{\text{max}} = 100 °C) \), the widths of the model domain that exceed 390 °C (90% of \( \Delta T_{\text{max}} \)) and exceed 350 °C (50% of \( \Delta T_{\text{max}} \)) were recorded. To account for the discreteness of the model nodes, an interpolation of the temperature profile across the shear zone was made using a first order spline. These results can be used to assess how much temperature variation can be expected within a shear zone that has significant shear heating and how far away from the shear zone significant temperature changes are expected.

4. Results

Progressive thermal softening and temperature dependent thermal properties are key aspects of our model and have a significant impact on the resulting shear heating calculation (Fig. 6). To illustrate the importance of these factors, models were run with constant thermal diffusivity (Fig. 6A and C, calculated with commonly used values of thermal diffusivity \( [1 \times 10^{-6} \text{ m}^2/\text{s}] \), density \( [2700 \text{ kg/m}^3] \) and heat capacity \( [1000 \text{ J K}^{-1} \text{ kg}^{-1}] \)) and with constant shear stresses (Fig. 6A and B). For the constant stress models, shear stress is fixed at a value calculated for an initial temperature of 300 °C. A model that includes no thermal softening and temperature-independent thermal diffusivity produces 50–250 °C of deformation-induced heating for a 3 cm/year convergence velocity after 10 Myr (Fig. 6A). If a temperature dependent thermal diffusivity (calculated from temperature dependent \( k, \rho \) and \( C_p \), as in Fig. 3) is used without thermal softening, deformation produces as much as 400 °C of heating under otherwise similar conditions in narrow (0.5–1 km wide) shear zones (Fig. 6B). These results highlight the potential inappropriateness of shear heating models that employ a constant shear stress. In this case, the fact that thermal diffusivity decreases
Duration of Deformation (Myr)

Width of Shear Zone (km)

Temperature (°C)

Shear stress (MPa)

- Temperature (°C)
- Shear stress (MPa)
with increasing temperature leads to a thermal runaway and extremely high predicted temperatures.

Inclusion of thermal softening in a constant thermal diffusivity model (Fig. 6C, shear stresses are recalculated at each time step) substantially decreases the effects of shear heating to 40–110 °C after 10 Ma. In this case, use of a constant thermal diffusivity results in 10–30 °C less shear heating than models that include a temperature dependent thermal diffusivity (compare Figure 6C and 7B). Note also that the range of thermal diffusivities we use is 25–50% smaller than the most commonly used value (Fig. 3B, D). Our values are probably more appropriate than the ‘typical’ fixed thermal diffusivity used elsewhere because our heat capacities and densities are calculated directly for the predicted mineral assemblages at temperatures and pressures of interest. We thus concur with previous studies that the choice of thermal properties and their temperature dependencies has a significant effect on the results of thermal modeling studies (Whittington et al., 2009; Nabelek et al., 2010; Duprat-Oualid et al., 2013).

4.1. Quartz dislocation creep

Magnitudes of shear heating can vary greatly depending on the conditions of deformation. The most dramatic heating occurs in the first ~1 Myr in all models, with heating rates that can initially exceed 100 °C/Myr (Fig. 4G, H), but then drop very rapidly. For all initial temperatures, velocities and flows laws modeled, heating rates are reduced to less than 50 °C/Myr after 1 Myr of deformation and long-term heating rates are 1–10 °C/Myr. Significant shear heating does not require extremely prolonged deformation and the effects of shear heating are continually diminished during deformation (Fig. 4).

Total temperature increases due to viscous dissipation in a quartz dislocation creep rheology range from a few degrees to > 200 °C. 1–10 km wide shear zones that are deformed at initial temperatures of 300–400 °C and fast convergence rates of 3–5 cm/year can experience 15–90 °C of shear heating in 1 Myr and 50–175 °C of shear heating in 10 Myr (Figs. 7B, C, E, F; 9C, F and I). Slow convergence of 1 cm/year,
Fig. 9. Predicted final temperatures for shear zones of 500, 2000 or 5000 m width after 1, 5 and 10 Myr, calculated with a quartz flow law. These figures can be used to compare inferred plate velocities, temperatures and stress and constrain an initial temperature of deformation. For example, (E) shows possible magnitudes of shear heating for a 2000 m wide shear zone that has deformed for 5 Myr. If $T_{\text{max}}$ of 500 °C (black lines) and shear stresses of 10 MPa (grey lines) can be inferred, the intersection of the corresponding lines indicates an initial temperature of ~455 °C (45 °C of shear heating) and strain rate of $2 \times 10^{-13}$ s$^{-1}$ (~3.5 cm/year). Inferred temperatures corresponding to approximate deformation conditions in the Main Central Thrust (F) and Norumbega fault zone (J, K) are shown in red shading (see text for details). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
which is more typical of continental shear zones, predictably results in far less shear heating, even with low initial temperatures. For example, 0.5 km wide shear zones with initial temperatures of 300–400 °C are predicted to experience 10–25 °C heating in 1 Myr and 30–60 °C in 10 Myr (Figs. 7A, D, 9A, B, and C). Initial hotter rocks (starting temperature of 500–600 °C) experiencing fast convergence rates (3–5 cm/year) in 1–10 km wide shear zones produce 4–30 °C heating in 1 Myr and 15–80 °C in 10 Myr (Fig. 7H, I, K, L). Initial temperatures of 500–600 °C with slower convergence rates (1 cm/year) produce temperature changes of only 5–15 °C after 10 Myr. This could be considered negligible in terms of contribution to metamorphic paths or relative to the resolution of most petrologic techniques. However, it is important to note that a wide range of deformation conditions can result in the same final temperatures. Shear zones that start at 400 °C could reach temperatures of 500 °C in less than 10 Myr with fast convergence rates (3–5 cm/year). To make meaningful estimates of the contribution of shear heating, metamorphic temperature constraints must thus be coupled with estimates of the duration of deformation, convergence velocity and shear stress.

The greatest weakening, or shear stress reduction, occurs in the models with the lowest initial temperatures and highest plate velocities, as expected. Indeed, weakening is most rapid in models with high initial stresses (Fig. 4E–F). The initial shear stress in a 0.5 km wide shear zone at 300 °C and 5 cm/year convergence velocity is 86 MPa, which is reduced to 35 MPa by 1 Myr (Fig. 7C). In contrast, the same shear zone experiencing a velocity of 1 cm/year experiences a reduction in shear stress from 59 MPa to 46 MPa in 1 Myr. In shear zones with initial temperatures > 400 °C and velocities from 1 to 5 cm/year, initial shear stresses are less than 32 MPa and shear stress reductions are less than ~10 MPa in most cases. Above initial temperatures of 500 °C shear zones experience less than 1 MPa weakening. Shear zones with lower convergence velocities and higher initial temperatures are predicted to better maintain their strength over time, but in general have low shear stresses.

Fig. 10 shows the size of the region that heats by at least 50% (Fig. 10A) or 90% (Fig. 10B) of ΔT_{max} experienced at the center of a shear zone. The width of the thermal anomaly is relatively independent of initial temperature and plate velocity; examination of the full suite of initial temperatures and plate velocities explored by our models reveals that for the width > 90% of ΔT_{max} calculation, 95% of models return a value within 0.7 km of the mean value for any given shear zone width and duration of deformation. 95% of models return a width within 1.6 km of the mean for the width > 50% of ΔT_{max} calculation. The thermal anomaly widths for all model runs in Fig. 7 were therefore averaged to produce Fig. 10, which approximates the thermal anomaly width for any initial temperature or convergence velocity, as a function of shear zone width and deformation duration. In all models, a zone wider than the shear zone experiences ΔT > 50% of ΔT_{max} ΔT > 90% of ΔT_{max} is only reached over a zone wider than the shear zone in less than 10 Myr if the shear zone is less than 5 km wide. In 1–2 km wide shear zones, a zone wider than the shear zone achieves ΔT > 90% of ΔT_{max} in 0.5–2 Myr. To illustrate with an example, a 0.5 km wide shear zone that experiences a maximum temperature increase of 100 °C in its core after 1 Myr of deformation would experience heating of at least 90 °C over a region ~1.0 km wide (Fig. 10B). In contrast, a 10 km wide shear zone experiencing 100 °C of heating in its core after an identical duration of shearing would experience only 50 °C of heating over a region ~10 km (Fig. 10A). Thus, for all but the widest shear zones explored, highest temperatures associated with shear heating are generally expected to be localized to the shear zone itself, but significant heating is expected to occur across a much wider area. For narrow shear zones, there may be no significant temperature variation within the shear zone itself (Figs. 4A, B, 10) and the thermal anomaly may extend far into the ‘wall rock’, even in cases that experience a significant amount of shear heating. The extreme end-member of this behavior would be a brittle shear zone of zero effective width. An ambient temperature may thus only be estimated on the order of tens of kilometers away, especially at long timescales of deformation (Fig. 10).

4.2. Wet feldspar diffusion creep

Diffusion creep is expected to be the dominant feldspar deformation mechanism for the temperatures and strain rates (1.6 × 10^{-12}–1.6 × 10^{-14} s^{-1}) that we have modeled (Fig. 8A–C), assuming water saturation. Brittle deformation of feldspars is expected to occur at lower temperatures (300–400 °C; Passchier and Trouw, 2005, p. 58), so our models cannot be used for feldspar deformation at these temperatures. The magnitudes of shear heating for a feldspar diffusion creep rheology (Rybicki et al., 2006) are relatively modest (Fig. 8A–C). Even at relatively high convergence rates of 5 cm/year, temperature increases of 30–60 °C are expected after 10 Myr in 0.5–10 km wide shear zones. At 1–3 cm/year 5–40 °C heating is predicted, which is well within the uncertainty of many petrologic thermometers. These magnitudes would be much greater for feldspar dislocation creep or dry diffusion creep. For parts of the crust that are dominated by wet diffusion creep of feldspar (which is probably the most common mechanism at the conditions of interest here) at convergence rates of 1–3 cm/year, shear heating might be considered negligible.

4.3. Biotite flow law

The biotite flow law of Shea and Kronenberg (1992) predicts that biotite schists are stronger than quartz-rich rocks above temperatures of 300–400 °C, depending on strain rate. Additionally, biotite schists are predicted to experience less softening with increasing temperature because their rheology is apparently less temperature dependent. As a result, the expected magnitudes of shear heating for biotite-rich rocks are on the order of 100–200 °C for a wide range of deformation conditions (Fig. 8D–F). Higher initial temperature models do not show significantly less temperature change than initial temperatures of...
400 °C. If the Shea and Kronenberg (1992) flow law can be extrapolated to micaceous rocks in general, shear heating might be extremely significant in phyllonites at much higher temperatures than are predicted for quartz-rich rocks.

5. Discussion

The qualitative features of our quartz dislocation creep simulations are effectively independent of the chosen flow law (Fig. 4, right panels), although the flow laws of Platt and Behr (2011), Hirth et al. (2001) and Gleason and Tullis (1995) produce different absolute amounts of shear heating. Models using the Hirth et al. (2001) flow law produce the most shear heating by as much as 50 °C, but the long-term stresses and heating rates for all three flow laws are not substantially different, despite significant differences immediately following the initiation of deformation (Fig. 4). Our models reproduce the behavior one expects from progressive thermal weakening: shear heating causes an exponential decrease in shear stress and the magnitude of heat generation. Narrow shear zones produce greater temperature increases and shear stress reductions than wide shear zones that accommodate an identical convergence velocity (e.g. Platt, 2015a). Lower initial temperature shear zones produce more shear heating.

Our results are broadly consistent with previous numerical studies. Simulated shear heating magnitudes of ~100 °C are commonly reported in the literature. Burg and Gerya (2005) estimated temperature increases of 100–200 °C in the crust after ~5 Myr of deformation. Nabelek et al. (2010) suggested that shear heating is necessary for mid-crustal migmatization, and our results reproduce temperatures that could lead to migmatization in high initial temperature (500–600 °C) shear zones with moderate to high convergence rates (3–5 cm/year) (Figs. 7 and 9). Jaquet et al. (2017) estimated 100–150 °C of shear heating in shear zones that are active for 1.5–3 Ma, similar to our results. We closely reproduce the results of Platt (2015a, 2017) when using constant thermal properties, but temperature-dependent thermal properties result in final temperatures that are ~30 °C greater than the results of Platt (2015a) for similar deformation conditions (Platt, 2017, Table 2, Column 1). The ability to compare our results with a variety of deformation conditions and shear zone sizes is a major advantage for field-based metamorphic investigations.

Our models predict shear heating magnitudes as great as 200 °C (Fig. 7C), but in most realistic circumstances temperature increases due to deformation are likely to be much smaller. Shear heating > 100 °C is only possible at high plate velocities of > 5 cm/year or low initial temperatures (300–400 °C) for the wet quartz rheology we have used (Figs. 7 and 9). Many natural shear zones are also much narrower (0.001–0.1 km) than the shear zones we have modeled here (0.5–25 km), as in demonstrated cases of strain localization (e.g. Gueydan et al., 2005; Agard et al., 2011). We stress that although narrow shear zones are capable of inducing substantial shear heating if they accommodate sufficient displacement, many ‘small’ shear zones accommodate a far smaller percentage of total plate velocity and may thus be incapable of producing substantial heating. For example, a 100 m wide shear zone deformed at a relatively fast strain rate of $1 \times 10^{-12} \text{s}^{-1}$ would have a convergence velocity of only 0.63 cm/year (Eq. (5)), which is unlikely to produce substantial shear heating in most circumstances.

Shear heating can produce significant weakening of the crust (e.g. Hartz and Podladtchikov, 2008). Our results for quartzoil dislocation creep rheologies indicate that weakening is rapid, such that shear stresses on the order of 100 MPa generally cannot be sustained for > 1 Myr. Stresses of this magnitude may not represent the long-term strength of a shear zone and deformation at high stresses is expected to be very transient. Indeed, the higher the initial stress, the more rapidly weakening occurs, independent of flow law used for shear heating simulations. However, high shear stresses (100+ MPa) inferred from grain size piezometry on naturally deformed rocks are not uncommon (Fig. 1, e.g. Behr and Platt, 2014). Assuming the validity and applicability of piezometric relationships, high stresses may represent very short-lived deformation events that accumulate to form high strain zones. Other explanations for apparently high stresses could be a lack of microstructural steady state, more significant microstructural energy sinks than are currently recognized, or high exhumation rates that cause continual cooling of deforming systems. In any case, shear heating, which is a fundamental process of rock mechanics, is expected to induce significant softening independent of reaction softening, rheological transitions or the development of a crystallographic preferred orientation.

It is useful to consider how pulsed or transient deformation might affect magnitudes of shear heating. Increased strain rate is accompanied by an increase in stress, so the total heat production (Eq. (1)) increases exponentially. Although the same amount of finite strain is produced for episodic and continuous deformation, there is an exponential increase in the total work of deformation, and thus shear heating, in the episodic case. Therefore, modeling continuous deformation provides a minimum estimate of the magnitudes of shear heating in episodic deformation. Preliminary examination of this process with our model confirms this expectation (see Fig. A2), but this is beyond the scope of the present study. In cases where a finite strain is accomplished by short-lived episodes of fast strain rate, the total work of deformation is greater for the episodic case than for continuous strain rate, for a power law rheology.

It should be noted that nothing in our analysis constitutes direct and absolute evidence for shear heating in natural shear zones. However, given that shear heating is a fundamental physical process, its presence should be assumed: (1) in the absence of unambiguous evidence for isothermal conditions when the conditions of deformation would suggest significant viscous dissipation, or (2) unless it can be demonstrated that microstructural processes are significant sinks for mechanical energy. For example, Ar/Ar thermochronology might be used to demonstrate an absence of shear heating by showing that there is no difference in cooling age across and outside a shear zone in which significant heating is expected. Detailed geochronology, coupled with petrologic constraints, could also be used to demonstrate isothermal conditions during deformation. However, these tests need to be coupled with estimates of how much shear heating would be expected for a given set of deformation conditions: our modeling shows that in some cases the amount of heating could be negligible while in others it should be significant.

Stresses inferred from grain size piezometry can play an important role in understanding the potential magnitudes of shear heating in a given deformation zone. In the absence of detailed information about the duration of deformation and convergence velocity, stress estimates can be used to calculate strain rates and magnitudes of shear heating. Conversely, tectonic velocity and the duration of deformation are sufficient to infer the magnitude of shear heating provided the final metamorphic temperature is known.

We explore observations from three shear zones to illustrate how our modeling results can be applied to natural systems. Two examples are strike-slip and one is a thrust-sense shear zone.

5.1. Example 1: Davenport shear zone, Australia

Camacho et al. (2001) presented evidence for shear heating in the Davenport Shear zone, Musgrave Block, Australia. Shear zone rocks
recorded temperatures of 700 °C, which is 200–300 °C greater than host rocks ~1 km away from the shear zone, where Ar/Ar ages of biotite are not reset to the ages of the shear zone. According to their modeling, these conditions can be produced in 0.003–0.3 Myr at strain rates of $10^{10}$–$10^{11}$ s$^{-1}$ and shear stresses of 100 MPa. Although the Davenport shear zone is composed of eclogites, which are likely to be very strong, our modeling suggests that stresses of ~100 MPa are unlikely to be maintained for 0.3 Myr. For many published flow laws, a 200 °C increase in temperature equates to an order of magnitude drop in strength (see Nabelek et al., 2010, their Fig. 4). Our modeling confirms the short timescale necessary to produce the observed contrast between shear zone and host rock temperatures. For example, after 0.2 Myr of deformation the width that is > 50% of $\Delta T_{\text{max}}$ is approximately 2 km (Fig. 10); thus, temperatures 1 km away from the shear zone core are predicted to have been 550–600 °C if $\Delta T_{\text{max}}$ is 200–300 °C. Maintenance of biotite 1 km from the shear zone at $T$ below a nominal closure temperature for Ar diffusion thus implies less than 0.2 Myr of deformation. Indeed, short timescales are compatible with diffusion speedometry on garnets in the shear zone (Camacho et al., 2009).

5.2. Example 2: the Norumbega fault zone, Central Maine

A sample from a 500 m wide mylonite zone in the strike-slip Norumbega fault zone (44.343048°N, 69.413339°W) has been examined to determine possible magnitudes of shear heating. Regional thermochronologic data suggests that this part of south-central Maine was at temperatures of 250–300 °C while the mylonite zone was active at 290–300 Ma (West and Lux, 1993; West and Hussey, 2016). Monomineralic quartz veins within the mylonite zone have a recrystallized grain size of 12.3 ± 0.5 μm, indicating shear stresses of 37–40 MPa (Eqs. (8) and (9), Table 1). Quartz exhibits both grain boundary bulging and subgrain rotation recrystallization microstructures that indicate 350–400 °C deformation temperatures (Hubbard and Wang, 1999; Stipp et al., 2002; Price et al., 2016). These shear stresses and temperatures could be produced in 0.5–2 Myr with a 3–5 km/year convergence velocity in a host rock initially at 300 °C (Fig. 7B–C). Alternatively, 1 Myr of deformation would produce the observed temperatures and stresses for 1,5–5 km/year convergence over a wider range of initial temperatures (Fig. 9J). 5 Myr of deformation would produce the observed stresses and temperatures at lower convergence rates (1–3 cm/year, Fig. 9K).

Swanson (1994) estimated finite shear strains of $\gamma = 30$ for this part of the Norumbega fault zone, which would correspond to 3–5 Myr of deformation at 3–5 km/year convergence velocities. All of these observations indicate that 30–100 °C of shear heating could have occurred in this mylonite zone, despite the fact that localized metamorphic grade increases are not recognized here. However, as discussed above, petrologic evidence for shear heating can be enigmatic and equivocal. In the Norumbega case, the host rocks record ~375 Ma amphibolite to migmatite grade metamorphism (West et al., 2003), which is likely to mask a recognizable shear heating signature at ~290 Ma. Hence, our results provide the first quantification of likely shear heating in this setting.

In this example we have used constraints on the shear stress, deformation temperature, shear zone width and finite strain. This is more information than is strictly necessary to predict magnitudes of shear heating, but in practice such an over-constrained system is valuable in that multiple independent observations are used to verify the conclusion. The well constrained regional cooling history for south-central Maine is particularly useful in this regard. By comparing the regional temperature with the shear zone deformation temperatures, shear heating of 50–150 °C could be inferred, but on its own this inference is somewhat equivocal. There is sufficient uncertainty in thermochronologic data and in associating quartz recrystallization regimes with temperature that this would not constitute strong evidence for shear heating. Combined with estimates of shear stress, finite strain and our modeling results, the Norumbega Fault zone may be a convincing example of substantial shear heating.

5.3. Example 3: the Main Central Thrust, Himalaya

Our model is designed primarily to describe shear heating due to strike-slip deformation but can also place useful limits on heating in compressional settings, subject to several additional assumptions. The metamorphic characteristics of the Himalayan crystalline core are intimately tied to the evolution of faults which now juxtapose a series of northward-dipping sheets that exhibit contrasting provenance, metamorphic and deformational histories (e.g. Ahmad et al., 2000; Kohn et al., 2003; Vannay et al., 2004; Searle et al., 2008). The Main Central Thrust (MCT), which was active throughout the early-Miocene and may have experienced additional deformation since then (e.g. Hodges et al., 1996; Tobgay et al., 2012), has previously been implicated in the genesis of a spatially associated inverted metamorphic gradient through mechanisms including (1) advection of heat in a hanging wall block (e.g. Le Fort, 1975) that likely reached kyanite- to migmatite-grade before exhumation (e.g. Harris et al., 2004), (2) shear heating (e.g. England and Molnar, 1993; Harrison et al., 1997a, 1997b), (3) reorganization of the orogenic core after peak metamorphism, effectively deforming pre-existing isograd structures (e.g. Brunel and Kienast, 1986; Searle et al., 1988), or (4) a combination of these factors (e.g. England et al., 1992). Here we aim to reassess the likely magnitude of shear heating, given the dynamic, temperature-dependent rock-strength approach described above. We assume a constant ~2 km wide shear zone (e.g. Searle et al., 2008; Law et al., 2013), using geologic estimates of slip rates of 2–3 km/year (Kohn et al., 2004) for on the order of 10 Myr (Searle et al., 2008). Major uncertainties associated with these simplifications are that shear zone width and deformation rate are fixed over time, that we ignore estimates placing a pure shear component at up to 10–30% (e.g. Law et al., 2004; Jessup et al., 2006), and that we ignore the complex lithologies present within the MCT. Quartz grain size piezometry constrains stresses of 6–11 MPa within ~1.5 km above the MCT, with locally higher stresses inferred (Law et al., 2013; note correction to plane stress, Eq. (7)). Quartz c-axis fabrics indicate deformation temperatures of 540–580 °C (Law et al., 2013). Our results show that these conditions of deformation could result in shear heating of 10–50 °C in the core of the shear zone, depending primarily on strain rate and initial temperature (Fig. 9F). This is associated with a final shear stress at 10 Myr of ~6 MPa (grey lines, Fig. 9F).

Our estimate of shear heating is dramatically smaller than previous estimates such as the > 500 °C presented in Fig. 4 of England and Molnar (1993), which was calculated for a fixed 250 MPa shear stress on a fault of zero thickness. We note that our result overlooks complicating factors such as heat advection in the hanging wall, which effectively acts to heat the fault and weakens rocks accordingly. Conversely, consideration of pure shear in the Himalayan crystalline core acts to increase the effects of shear heating (compare for example Eq. (1)a–d in Burg and Gerya, 2005). Given that the proportion of pure shear can be fixed in the range 10–30% (though this may have evolved with time) and that the importance of advection scales with deformation rate, we may thus either over-estimate the shear heating component by ignoring advection if thrusting was rapid or may underestimate it in cases in which thrusting was slow. The balance between these two
additional parameters will be the focus of future work concentrating on normal and thrust fault systems.

5.4. Subduction zones

Shear heating has been invoked to explain thermal-metamorphic observations (Peacock, 1992; Penniston-Dorland et al., 2015) as well as surface heat flow (Gao and Wang, 2014) in subduction zones. Our modeling results are not directly applicable to subduction zones for two reasons. First, water fugacity is significantly higher at higher pressures than in the examples shown here. This would act to lower shear stresses so that, all else being equal, our models likely over-estimate the magnitudes of shear heating in subduction zones. Second, it is unlikely that a quartz rheology is appropriate for subduction shear zones, which are expected to be composed of serpentinites, peridotites, eclogites and metasedimentary melange. Stronger peridotites and eclogites have greater potential for shear heating than quartz dominated rocks, whereas mélanges and serpentinites might be expected to be weaker. More focused modeling could improve constraints on the amount of shear heating possible in subduction zones (e.g. Peacock et al., 1994; Peacock, 2003).

6. Conclusions

Magnitudes of shear heating can range from negligible to > 100 °C, depending on the conditions of deformation for a quartz-dominated rheology. Shear heating > 100 °C is generally only achievable for relatively low initial temperature (300–400 °C) with high convergence velocities (3–5 cm/year), while shear zones with slow convergence velocities (1 cm/year) only produce heating > 50 °C in relatively long-lived low initial temperature (300 °C) cases.

We have provided a method for predicting expected magnitudes of shear heating based primarily on field observations (Figs. 7 and 9), for a wide range of possible shear zone conditions. This approach is an alternative to mineral thermometry methods of quantifying shear heating. We argue that a distinct localized metamorphic grade increase indicative of shear heating is almost always difficult to observe for several reasons. High grade shear zone host rocks, diachronous metamorphic assemblages, lack of fluid flow and absence of deformation far from a shear zone at ambient temperature can prevent localized metamorphic grade changes from being well recorded mineralogically. Without syn-deformational metamorphic recrystallization outside the thermal anomaly (10–20 km away), there is no baseline for comparing shear zone temperatures to regional ambient temperatures at the time of shearing. These factors may explain apparent ‘anti-shear heating’ signatures sometimes observed in localized shear zones, where a low grade high strain zone overprints higher grade rocks (e.g. Ersklev and Sutter, 1990; West and Hubbard, 1997; Ballèvre et al., 2000). Additionally, we have shown that the length scale of the thermal anomaly associated with shear heating is expected to be much wider than the shear zone itself in many cases. Temperatures are not expected to vary significantly within smaller shear zones even if shear heating is significant (Fig. 10). This means that in terranes containing multiple closely spaced active shear zones, temperature anomalies from shear heating can be superimposed. The thermal structure of highly deformed terranes may thus be strongly influenced by shear heating without producing good evidence of localized metamorphic grade changes. Shear zones that are short lived and have higher than expected strain rates are an exception to this statement. Therefore, a lack of obvious thermal anomaly around a fault zone does not necessarily constitute evidence that shear heating is an unimportant process.

Progressive thermal weakening is an important process that is often neglected in numerical models of shear heating. Shear strength decreases exponentially with increasing temperature for most published flow laws. Thus, shear heating is a self-limiting process. Our models that use a quartz dislocation creep rheology show that weakening due to shear heating is rapid, with the most significant weakening occurring in the first 0.1–1 Myr of model runs. High shear stresses on the order of 100 MPa are not expected to be maintained for significant durations in the presence of shear heating (c.f. Platt, 2015a) and therefore may not represent the long-term strength of the deforming crust, especially in crustal-scale strike-slip shear zones. High observed shear stresses (Fig. 1) may represent transient deformation, highly localized deformation, continual cooling by an extrinsic process (large fluid fluxes or advection) or the presence of unaccounted energy sinks in deforming metamorphic systems. In any case, we have provided first order predictions of magnitudes of shear heating that can be used to help better understand other processes that also control the thermal budget of the deforming crust.

Acknowledgements

This work was supported in part by a 2015 Sigma Xi Grant in Aid of Research and a scholarship from the Virginia Tech Graduate Student Assembly to CAM and National Science Foundation award EAR-1250470 to MJC. Dr. Sylvia Duprat-Oualid and an anonymous reviewer provided very constructive comments on this manuscript. We are grateful to Dr. Philippe Agard for his careful and constructive editorial handling. We thank Dr. Ryan Pollyea at Virginia Tech for providing computational resources and offering constructive feedback throughout the process. Dr. Rick Law at Virginia Tech is thanked for his insightful comments on this manuscript and the ideas therein. Dr. John Platt at the University of Southern California was helpful in discussing our model setup and results. The Metamorphic Processes Group at Virginia Tech (including Kirkland Broadwell, Dr. Besim Dragovic, Dr. Jen Gorce, Allie Nagurney, Sarah Ulrich and Lisa Whalen) is thanked for much useful feedback and criticism.

Appendix 1

We have tested the spatial and temporal resolution of our model to optimize precision and run time, finding that increased resolution did not improve or significantly change the model results (Fig. A1). In all cases, substantially increased resolution changes the resulting temperatures by much less than 0.1 °C. Shear stress is calculated directly from temperatures, so it also is not significantly changed by increased spatial resolution or temporal resolution. We also examine the effects of pulsed, rather than continuous deformation and an example of the difference between the two is shown in Fig. A2.
Fig. A2. All figures within this paper are calculated with constant strain rate over time. Periods of pulsed strain separated by deformation quiescence result in greater total work of deformation, even if the integrated strain is identical in pulsed or steady state scenarios. As an example, a shear zone of width 0.5 km deforming continuously at 3 cm/year for 5 Myr heats to the red curve in panel A. Recalculating with the same finite strain and 5 Myr duration, but with all deformation occurring only every tenth time step increases final temperature as shown by the blue curve in panel A. The evolution of $\Delta T_{\text{max}}$ and stress are shown in panels B and C, respectively.

References


West, D.P., Lux, D.R., 1993. Dating Mylonitic Deformation by the 40Ar·39Ar Method: An Example From the Norumbega Fault Zone, Maine. 120. pp. 221–237.

