Impact of gneissic layering and localized incipient melting upon melt flow during experimental deformation of migmatites

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Highlights :

- Purpose: examine experimental control of anisotropies on melt flow in migmatites.
- Experiments on a homogeneous gneiss and one with strong mineral layering.
- Initial anisotropy strongly controls melt flow and segregation at sample-scale.
- Initial anisotropy also controls deformation distribution and its partitioning.
- Boudinage occurs in the gneiss with mineral layering.

Title: Impact of gneissic layering and localized incipient melting upon melt flow during experimental deformation of migmatites

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Abstract:

In this study, we test experimentally the role of compositional layering as a key parameter for controlling melt flow in a natural migmatite during coaxial deformation. We performed in – situ pure-shear experiments on two natural gneisses. The first gneiss is weakly foliated with minerals homogeneously distributed. The second gneiss shows a pronounced compositional layering of alternating quartz – feldspar – rich and biotite – muscovite – rich layers. Experimental conditions were selected to obtain homogeneous melt distribution in the homogeneous gneiss and heterogeneous melt distribution in the layered gneiss. Initial melt distribution is not modified by deformation in experiments on the homogeneous gneiss, implying that melting products did not migrate from their initiation sites. In contrast, melt flowed in shear zones or in inter-boudin positions during experimental deformation of the heterogeneous gneiss. These experiments attest to the strong influence of initial gneissic layering on melting pattern, melt segregation and flow during deformation of partially molten rocks.

1. Introduction

Within the crust, partial melting is the main geological process responsible for production of leucogranites, crustal differentiation (Rudnick, 1995) and major changes in crustal strength (Rosenberg & Handy, 2005). Field evidence shows that exhumation of high-pressure metamorphic rocks or extrusion of large tectonic slices are facilitated when the crust reaches its solidus (e.g. Labrousse *et al.*, 2011; Trap *et al.*, 2011). Magneto-telluric (Hashim *et al.*, 2013) and seismic (Nelson *et al.*, 1996) surveys in the North-West Himalaya seem to confirm the presence of a partially molten crust at depth, which are responsible for large-scale lateral flow of the Eurasian crust (Caldwell *et al.*, 2009). At smaller scale, many field observations within migmatitic terrains emphasize the feedback relationship between crustal

weakening and the production, segregation and flow of silicate melt (e.g. Brown & Solar, 1998).

Melt fraction is documented as the first parameter controlling mechanical behaviour and localization of deformation of partially molten aggregates in experimental studies (e.g. Misra *et al.*, 2009). Melt fraction is inferred to have a drastic weakening effect on bulk rock strength (Arzi, 1978; Van der Molen & Paterson, 1979; Rutter *et al.*, 2006). Rosenberg & Handy (2005) showed that 10% partial melting of a dry aplite leads to a strength drop from 820 MPa down to 80 MPa in experimental conditions. This first order of magnitude in strength drop is expected to represent a two orders of magnitude drop in effective viscosity for bulk non-Newtonian behaviour with stress exponent $n=3$.

Melt viscosity also influences the rheological behaviour of partially molten rocks. It depends mostly on melt silica and water contents, as well as temperature. The addition of 1 wt% H₂O to a dry haplogranite at 800°C can induce a viscosity drop from 10^{13} Pa.s to 10^7 Pa.s (Dingwell *et al.*, 1996). Similarly, hydrated melts (1 wt % H₂O and above) maintain substantially equivalent low viscosities (between 10^6 and 10^7 Pa.s). In contrast, at similar conditions, an increase of 100°C in melt temperature, which represents the actual variation tested in our experiments, induces less than a one-order viscosity drop for hydrated haplogranite melts (Dingwell *et al.*, 1996).

Melt distribution is also a major parameter that controls rock strength, responsible for local changes on melt percentage. In particular, a strength drop occurs for a 10% melt fraction when full connectivity of melt pockets is achieved in experiments (Rosenberg & Handy, 2005). Grain-scale surface energy (Cooper & Kohlstedt, 1984; Laporte & Watson, 1995), differential stress (Daines & Kohlstedt, 1997; Zimmerman *et al.*, 1999) and finite deformation magnitude (Kohlstedt & Zimmerman, 1996) have been recognized as controlling parameters for melt distribution. In addition, Tumarkina *et al.* (2011) argued that protolith anisotropy, in

particular foliation orientation, is critical for partial melting rate. Yet, anisotropy seems to have little impact on rheological behaviour of partially molten rocks although it may influence their behaviour in sub-solidus domains (Holyoke & Rushmer, 2002). According to Marchildon & Brown (2003) the distribution of leucosomes within migmatites is controlled by protolith anisotropy rather than by regional stresses. With prograde metamorphism, incipient partial melting leads to the formation of metatexites where paleosome dominates. Metatexites are typically heterogeneous with leucosomes (crystallized melt) arranged in a planar manner and oriented parallel to the protolith planar anisotropy, i.e. compositional/gneissic layering (Marchildon & Brown, 2003; Sawyer, 2008). Thus, inherited protolith planar fabric and related melt distribution after incipient melting may be an important parameter to consider when studying strength of partially molten continental crust and related melt flow.

In this contribution, we test the impact of incipient melt distribution as a function of host rock texture (isotropic vs anisotropic) for the deformation of partially molten rocks (migmatites) and melt flow. Two sets of pure-shear experiments were performed on two natural gneisses using a Paterson apparatus. We focus on natural gneisses to avoid artificial porosity, potentially high in synthetic material (up to 15%; Misra *et al.*, 2014) that may disturb melt distribution. One natural gneiss shows a pronounced gneissic layering whereas the other is compositionally homogeneous, so that we could document changes in melt distribution as a function of the fabric of the starting gneiss for an experiment. In addition, incipient melt distribution, homogenous vs heterogeneous, was controlled using different experimental conditions, i.e. temperature and hydration, to promote water – present and muscovite – dehydration melting reactions and prevent other melting reactions, in particular, biotite – dehydration melting. Homogeneous gneiss was deformed at relatively low temperature (850°C and below), in hydrated and dry conditions, while heterogeneous gneiss

was experimented at higher temperature (850°C and above) and dry conditions. One experiment set-up (850°C, dry conditions) is common to both experiment series. The comparison of deformation patterns and melt distributions observed in our experiments allow to discuss the impact of the inherited gneissic layering and related melt distribution on melt flow and deformation of partially molten rocks.

2. Starting materials

Starting materials are two natural gneisses, one homogeneous, weakly foliated gneiss (NOP1) with no compositional layering and one heterogeneous gneiss (PX28) with a strong gneissic layering (Fig. 1). For both starting materials, we also choose experimental conditions to restrain melt production to water – present and to muscovite – dehydration melting reactions, and prevent biotite – dehydration melting. This approach required to adjust experimental conditions using different temperatures with addition of free water for some experiments. The experimental set up were designed to work with two texturally different migmatites, one with a homogenous distribution of melt (in NOP1) and one with a layered-type distribution of melt (in PX28).

2.1 Homogeneous gneiss: sample NOP1

The homogeneous gneiss (NOP1) is a quartzo-feldspathic two – mica gneiss, sampled in the inland of the Western Gneiss Region in Norway (N62°05'29'' E006°51'976'', UTM32). Modal composition is 45% quartz, 20% muscovite, 20% biotite, and 15% plagioclase. Accessory minerals are Fe-Ti oxides and zircons. Minerals are homogeneously distributed at sample scale with a weak foliation marked by preferred orientation of micas and elongated quartz grains and aggregates (Fig. 1a). The typical rock fabric is made of 1 mm thick and up to 5 mm long quartz aggregates (composed of grains with maximum length

about 500 μm to 1 mm) surrounded by grains of micas and plagioclase of ca. 100 μm in size and even locally smaller (Fig. 1b).

'Insert figure 1 here'

Melting tests on the homogeneous gneiss (NOP1) revealed that partial melting does not occur for $T < 825^\circ\text{C}$ probably due to paucity of free water. Melting through the biotite – dehydration reaction started for $T = 850^\circ\text{C}$. To avoid biotite melting and to promote melting through water – present reactions and homogenous distribution of melt, the starting material was hydrated prior to the deformation experiments. With the addition of free H_2O , experiments were run at temperature of 750°C to avoid biotite – related melting.

Hydration was performed at room pressure and temperature during three days in a closed glass containing distilled water. Complete and homogeneous hydration is only attained for small core sizes (diameter = 5mm). To enhance hydration, cores were plunged into distilled water and placed in a vacuum holder one hour per day. Hydrated cores were sealed within gold caps and heated at 750°C , 300 MPa for 48 hours in an internal-heated vessel. In that way, water was dissolved and trapped in the melt, which became hydrated glass when cores were quenched. This hydrated homogeneous gneiss is sample NOP1-B whereas the dry homogenous gneiss is sample NOP1-A (Table 1).

'Insert Table 1'

2.2 Heterogeneous gneiss: sample PX28

Starting material PX28 is a quartzo-feldspathic two – mica gneiss, sampled in the Sioule Valley in the French Massif Central (46.110613, 3.013235). Modal composition is 50% quartz, 20% feldspars with mainly plagioclase and little K-feldspar, 20% biotite and 10% muscovite. Accessory minerals are zircon, apatite, and monazite. PX28 is well foliated

and shows a pronounced gneissic layering defined by millimetre – thick mica – rich and quartz – rich layers (Fig. 1c). Within leucocratic layers, quartz and feldspars display a 50-100 μm grain size, with elongated biotite and few muscovite grains up to 100 μm long and 20 μm thick. In mica-rich layers, biotites and muscovites are commonly 100-500 μm long and 10-50 μm thick and reach up to 1 mm long and 1 mm thick, with some very rare grains up to 3 mm in length. Muscovite grains are preferentially located within mica – rich layers and represent less than 1% of quartzo-feldspathic layers volume. Within mica – rich layers, foliation is locally deflected by 100 μm to 1 mm thick shear zones (Fig. 1d) marked by mica deflections. Within shear zones, some small 10-50 μm -size quartz and feldspar grains occur. One experiment under hydrostatic conditions, i.e. without deformation, (PP123: 850°C, 300 MPa, 1h45 min; Table 1) was performed. Only slight modifications of PX28 fabric were observed with quartz-feldspar grain boundaries becoming more rounded and production of melt. Within shear zones, small grains became uncommon, which we attribute to melt consumption. Deformation experiments were therefore preceded by a 30 minutes heating event under hydrostatic conditions at running temperature so as to limit the influence of small – scale structures and grain – size heterogeneities on melt generation and migration.

3. Experimental methods

Pure-shear deformation experiments were performed in an internally heated gas-medium deformation apparatus (Paterson instrument, Australian Scientific Instrument) at ISTO Orléans, France. Cylinders with 5 to 15 mm in diameter and 10.3 to 28.8 mm in length (Table 1) are cored perpendicularly to the initial foliation for both natural gneisses. The smallest diameter is used for experiments on the homogeneous gneiss NOP1 with water added (starting material NOP1-B in Table 1) since complete hydration of a sample could only be completed for the small – diameter cores. Length of cores depends on the fragility of the

173 material and/or the cylinder; as for example long cores with smallest diameter were difficult
174 to drill. Although cores of both gneisses have different mineral sizes and different diameters
175 of cylinder were used, a minimum of 30 grains could always be traversed across the core so
176 that individual large grains do not affect the mechanical behaviour of the entire samples.
177 Opposite faces of cylinders were finely polished to achieve parallelism within $\pm 3 \mu\text{m}$. Cores
178 were then inserted in a column assembly made of alumina and zirconia pistons enclosed in an
179 iron jacket (Champallier *et al.*, 2008). To limit loss of fluid and isolate the sample from the
180 thermocouple, two 3 mm-thick alumina spacers were inserted between the rock sample and
181 alumina pistons. Two geometries of column are used depending on core size. For the smallest
182 cores (diameter = 5 mm), a copper cylinder was introduced between the sample and the iron
183 jacket in order to fill the gap between them (cores: diameter = 5 mm, iron jacket: diameter =
184 10 mm). The diameter of the iron jacket was reduced to 10 mm at sample edges, forming
185 “dog-bone” geometry. For bigger cores, no reduction was needed since core diameter equals
186 internal iron jacket diameter. A platinum foil (25 μm) was systematically introduced between
187 core and iron jacket to prevent any chemical reaction between melt and iron. The column was
188 placed in the Paterson apparatus to locate the sample in the isothermal zone of the heating
189 chamber (temperature controlled at $\pm 1^\circ\text{C}$ over a distance of 5 cm). Temperature was
190 measured with a K-type thermocouple situated at 3 mm from the top of the sample. Argon
191 was used as confining medium. Force was applied with a motor located below the column
192 assembly, where the top of the column assembly being fixed. Force was monitored by an
193 internal load cell with a 10 N resolution (Champallier, 2005). Experiments were performed at
194 a constant 300 MPa confining pressure, at different temperatures (750°C, 850°C and 900°C)
195 and strain rates varying from 10^{-5} s^{-1} to $1.25 \cdot 10^{-3} \text{ s}^{-1}$. Experiment names are based on (1) type
196 of material, (2) temperature ($^\circ\text{C}$) and (3) glass content (surfacic %). For the homogeneous
197 gneiss (NOP1), experiments names start with “Ho” (Table 1). For the heterogeneous gneiss

(PX28), experiments names start with “He”. For example, Ho/750/6 corresponds to the experiment performed on the homogeneous gneiss (NOP1) at 750°C with a glass content of 6%. Strain rate is kept constant over the whole duration of the Ho/750/6 experiment. Ho/850/33, He/850/19 & He/900/21-32 experiments experienced several strain rate steps. Details about experimental conditions are summarized in Table 1. Deformation began after temperature and pressure equilibrium was reached, meaning that melting reactions or glass-to-melt transformation began before deformation. Thus, deformation started in presence of melt.

The deformed cores were cut along the long axis, i.e. perpendicular to the protolith foliation. Thin sections were observed under polarized microscope and SEM. Glass amount after deformation was estimated from representative zones for each core (compilation of 4 to 15 SEM images). Given that glass and plagioclase have similar grey colour in SEM images, which prevented automated calculation of glass amount. Glass location was optically observed and manually drawn for each SEM image. To better resolve local glass (former melt) distribution, chemical maps are also acquired at several locations for two representative experiments (Ho/850/33 & He/900/21-32, see section 4.5). From the combination of elemental maps, mineral phases are recognized by automatic clustering using “k-means” algorithm. One pixel of each phase was manually picked as initial guess to ensure proper convergence of the algorithm. The algorithm then assigns a phase to each pixel based on his statistical similarity with those initial guess. Then pixels recognized as glass are counted to calculate glass amount. Comparisons between manual and automatic estimations of glass amount, on the same area, are in good agreement (e.g. for the zone represented in Fig. 11b: 22% obtained from manually drawn and 20% from computation).

Glass drawings together with SEM images were then analyzed by the intercept method (Intercept2003 and SPO2003 software, Launeau & Robin, 1996; Launeau *et al.*, 2010), to determine the statistical long-axis orientation of glass pockets and the amount (surfacic %) of

glass. Shear zones were optically recognized and shear zone geometry was drawn. Then shear zone orientations were measured in the same way than glass (former melt) pockets.

For each protolith, two experiments were performed at two different temperatures to achieve conditions with low and high melt fractions. Different strain rates varying from 10^{-5} s^{-1} to $1.25 \cdot 10^{-3} \text{ s}^{-1}$ were tested, depending on the mechanical response of the sample during an experiment.

4. Experimental results

4.1 Partial melting reactions in both protoliths

In non – deformed hydrated cores from the homogeneous gneiss NOP1 (after the passage in an internal-heated vessel), glass is observed around muscovite grains in contact with quartz (Fig. 2a) and contains small euhedral newly formed biotite and sillimanite. Rare spinel and corundum are sometimes observed. These microcrystalline phases are systematically oriented and do not show any typical quenching textures (disoriented acicular growth pattern for instance). We therefore consider them as melting reaction products together with the melt represented by the interstitial glass after quenching (Webb & Knoche, 1996). We interpret the presence of glass, biotite and sillimanite, between muscovite and quartz, as indicating the water – present partial melting reaction documented in metapelitic system (Clemens & Vielzeuf, 1987; Solar & Brown, 2001):

abbreviations after Kretz (1983). Spinel may be the product of the biotite – dehydration reaction:

with sillimanite as a product of reaction (1). In the KFMASH system (rock chemistry reduced to K, Fe, Mg, Al, Si and H a simplification relevant for quartzo-feldspathic rocks; Spear & Cheney, 1989), spinel and corundum may result from the reaction:

In cores of NOP1-B, glass without any associated newly formed peritectic phase is also abundant between plagioclase and quartz (same geometry than in Fig. 2b), indicating the former presence of melt and thus the following water – present partial melting reaction (Boettcher and Wyllie, 1969):

In cores from the heterogeneous gneiss PX28 and from non-modified homogeneous gneiss NOP1 (NOP1-A in Table 1), the same glass and newly formed peritectic phases association as in hydrostatic test experiments (e.g. PP123, Table 1) are observed. We therefore interpret them as produced during the initial hydrostatic stage of the deformation experiment. Glass is present around muscovite grains and along muscovite cleavages, indicating the presence of former melt in contact of muscovite and quartz (Fig. 2c). Glass is associated with newly formed peritectic grains (micrometric crystals with needle or prismatic shape) of biotite, sillimanite and K-feldspar (Figs. 2c & d), indicating the following dehydration partial melting reaction (Patiño Douce & Harris, 1998; Castro *et al.*, 2000):

Glass without newly formed peritectic phases is also observed at plagioclase-quartz grain boundaries (Fig. 2b). Considering that temperatures and pressure used in those experiments (850°C in He/850/19 and 900°C in He/900/21-32, both at 300 MPa) are too low for water – absent partial melting of Plg+Qtz to occur (reaction reached at 900°C for similar pressure, Boettcher and Wyllie, 1969), the former melt, now evidenced as glass, is interpreted as the

direct product of water – present partial melting (melt2 in reaction 4) or after interaction between melt (melt3 formed by reaction 5) and plagioclase following the reaction:

For both protoliths, biotite also shows isolated glass pockets along cleavages (Fig. 2d). This occurrence may indicate either sporadic hydrated partial melting of the biotite or incipient destabilization of biotite within a former melt that was produced by reaction (1), (4) or (5).

'Insert figure 2 here'

4.2 Mechanical data

4.2.1 The homogeneous gneiss NOP1

The two experiments on the homogeneous gneiss NOP1 were performed in different manners: experiment Ho/750/6 was performed for one single strain rate value, whereas experiment Ho/850/33 is the succession of four strain-rate steps, with an increase of strain rate at each step. Even if protocols were different, it is possible to draw first order comparison of stress evolution with strain between Ho/750/6 and the first step of Ho/850/33, both of them being performed at strain-rates in the same order of magnitude ($5 \cdot 10^{-5} \text{ s}^{-1}$ and $2 \cdot 10^{-5} \text{ s}^{-1}$) and both being deprived of any inheritance from earlier strain history. Experiment Ho/750/6 shows hardening with strain (Fig. 3a) whereas experiments Ho/850/33 first step reaches a steady state (Fig. 3b). Experiment Ho/750/6 ($>200 \text{ MPa}$ at $2 \cdot 10^{-5} \text{ s}^{-1}$, Fig. 3a) reaches higher stress than experiment Ho/850/33 ($<20 \text{ MPa}$ at $5 \cdot 10^{-5} \text{ s}^{-1}$; Fig 3b).

4.2.2 The heterogeneous gneiss PX28

At 10^{-4} s^{-1} , experiment He/850/19 reaches higher stress (circa 60 MPa Fig. 3c) than experiments He/900/21-32 (ca. 10 MPa , Fig. 3d). The same observation is true for 10^{-3} s^{-1}

(>70 MPa in He/850/19 vs ca. 30 MPa in He/900/21-32; Figs. 3c & d). At 850°C in He/850/19, the stress vs. strain curves show a stress peak at ca. 7% strain followed by the very beginning of a strength decrease (Fig. 3c). At 900°C, in experiment He/900/21-32, steady state is attained after ca. 10% strain with flow stresses about 10% of the confining pressure at 10^{-3} s^{-1} (Fig. 3d).

'Insert figure 3 here'

4.3 Microstructures at sample scale (1mm-1cm)

4.3.1 Microstructures in the homogeneous gneiss NOP1

To a first order, the deformation distribution in both experiments (Ho/750/6 & Ho/850/33) is mostly homogeneous given the near barrel shaped end-cores with smooth edges and the lack of penetrative shear zones cross-cutting either sample (Fig. 4). In the low – glass fraction experiment Ho/750/6, little strain localization is observed, as shear zones are rare and small (only 8 observed) and connect to sample edges (Fig. 4a). Geometrically these shear zones occur as planes with angles greater than 25° to the experimental compression direction (black arrows in Fig. 4). However, the large range of angles and the small number of shear zones prevent any statistical analysis. In high glass fraction experiment Ho/850/33, only five shear zones are observed and they are all connected to sample boundaries without penetrating more than a few mm into the sample (shear-zones reported in Fig. 4b were first observed on SEM images).

'Insert figure 4 here'

4.3.2 Microstructures in the heterogeneous gneiss PX28

In the layered gneiss (PX28), deformation at core – scale is more heterogeneous as primarily shown by the jagged edges of core samples (Fig. 5). In the experiment with low glass content, run at 850°C (He/850/19), deformation is mainly accommodated by two core-scale shear zones (in red in Fig. 5a). Deflection of the pre-existing foliation along the two shear zones defines “shear zone boudins” as classified by Goscombe *et al.* (2004). Shear zones orientations were systematically measured (angle to the experimental compression direction) from image analysis and the minerals they affect systematically noted. We distinguish mica-mica, mica-quartz, and quartz-quartz shear zones. Their offset being the same order of magnitude or smaller than aggregate sizes, the association of minerals on each side of shear zones is interpreted to be the shear zone initiation site. In He/850/19 experiment, shear zones concentrate in or along mica – rich layers (Fig. 5a), as only two are observed along quartz – quartz contacts (Fig. 6a). In experiment with the higher melt fraction, conducted at 900°C (He/900/21-32), shear zones are distributed within or between micaceous and quartzo-feldspathic layers (Figs. 5b & 6b).

‘Insert figure 5 here’

The average orientation of shear zones is close to 45° within both experiments (Fig. 6). In details, in He/850/19, shear zones lack an angle to experimental compression direction less than 20°, and most angles are between 20° and 60° with a mean for the population of about 45° (Fig. 6a). Shear zones with larger angles to the experimental compression direction develop mainly in micas-rich layers. At 900°C and higher melt fraction (He/900/21-32), shear zones have lower incidence values with a more pronounced peak in the 40-45° (Fig. 6b).

‘Insert figure 6 here’

In He/900/21-32, some quartz-feldspars layers are truncated into few millimetres-long symmetric boudins while the mica – and glass – rich layers tend to occur within boudin-necks

(Figs. 5b & 7a). Here, boudins are mostly symmetrical with little slip along a single shear band on the inter-boudin surface (no slip boudinage from Goscombe *et al.*, 2004).

'Insert Figure 7 here'

4.4 Microstructures at grain-scale (μm -mm)

4.4.1 Microstructures in the homogeneous gneiss NOP1

In Ho/750/6, mica grains were deformed by syn-experimental kink bands (Fig. 8a, red lines in Fig. 4), whereas Ho/850/33 contains only a few kink bands. Some biotites with sigmoid-like shapes are observed in both samples (Figs. 8c & d).

In experiment Ho/750/6, quartz grains are broken (“crystal breakage” in Fig. 4) and the related inter-grain voids are generally empty, lacking glass (empty cracks in Fig. 8b). At both temperatures (750 & 850°C), quartz grains may be completely broken into pieces of different sizes, with each fragment being surrounded by glass, forming an interconnected glass – filled fracture network (Figs. 8c, d and e). Such networks inside quartz grains have larger expansion in Ho/850/33 (Fig. 8e) than in Ho/750/6 (Fig. 8c). In these networks, glass is generally free of newly formed peritectic phases (Fig. 8e).

'Insert figure 8 here'

4.4.2 Microstructures in the heterogeneous gneiss PX28

A major difference between both experiments (He/850/19 & He/900/21-32) is that muscovite is not observed anywhere in He/900/21-32, which means that muscovite completely reacted during the experiment. As the only known reaction involving muscovite is dehydration melting, we propose that muscovite entirely disappeared by dehydration melting.

In both experiments, biotites present sigmoid-like shapes (Fig. 7a). No kink-bands are

observed. Some biotites are broken in small pieces ($<20\text{ }\mu\text{m}$) and aligned along a plane that we interpret as a shear zone (Fig. 7a). Biotites oriented perpendicularly to their initial orientation and surrounded by glass, also containing newly formed peritectic phases, are observed in He/900/21-32 (Fig. 7b), implying flow of biotite during deformation.

Newly formed peritectic phases are mainly observed in glass films along mica grains and oriented parallel to the foliation. However in both experiments, changes in newly formed peritectic phase orientation are observed along glass veins inside quartz grains (e.g. in He/900/21-32 in Fig. 7c). This change in orientation of newly formed peritectic phases is interpreted as marking flow of the peritectic phases-bearing melt.

Quartz grains show syn-deformation lobate boundaries and grain-size reduction (Figs. 9a & b), which are assumed to be due to partial consumption during melting reaction. Outside of mica – rich layers, quartz and feldspar grains show offsets and relative displacements of grains with respect to each other without evidence of cataclastic deformation of grains, nor intracrystalline plasticity (Fig. 9c). The observation that glass fills all pores and boundaries between the matrix grains supports a melt-assisted granular flow. Quartz and feldspar breakage is observed in few locations (e.g. Fig. 7a), within shear zones. In He/900/21-32 experiment, very few glass-filled cracks within quartz and feldspar grains are observed (Fig. 7c).

‘Insert figure 9 here’

4.5 Glass distribution

4.5.1 Glass distribution in the homogeneous gneiss NOP1

In experiments on the homogeneous gneiss (NOP1), glass content at 850°C is 5.5 times greater than in experiment at 750°C (33% vs 6% respectively, Fig. 10 & Table 1). Glass is

located (1) at grain boundaries as 20-100 μ m thick films (Fig. 11b), (2) along micas cleavages forming few μ m-size pockets (black arrows in Fig 11b) or (3) in quartz grains as injections (blue triangles in Fig. 4; Fig. 11c). Glass injections, depleted in newly formed peritectic minerals, are generally thicker at the boundary of their host grain or grain fragment (Figs. 8e & Fig. 11c) and narrower within quartz grains, which allows determination of glass movement direction, i.e. from mica grains toward quartz grain aggregates. Glass-filled cracks within quartz-supported network (e.g. Figs. 8d & 11d) is systematically poor in newly formed peritectic phases (in blue in Fig. 11d). Even if the zone illustrated in Fig. 11c is located between two shear zones (thick dashed lines), its glass content is less than the zone in Fig. 11d, located away of any shear zone. The fact that shear zones do not correspond with high glass fraction areas is observed over the whole sample (e.g. Figs. 8d & 10b) and in Ho/750/6. No clear relation between glass distribution and shear zones localization can be identified from sample examination.

In sample cores from both experiments, the main population of glass films and pockets orientation is almost perpendicular to the experimental compression direction: 40% of the glass pockets and films are between 80 and 120° in Ho/750/6, 26% in Ho/850/33 (Dark grey population in rose diagrams in Fig. 10). A second population of glass orientations is present in Ho/750/6 and is oblique to the experimental compression direction (Fig. 10a). In both experiments, glass injections in quartz (glass location 3, see above) are sub-parallel to experimental compression direction (light grey population in rose diagrams in Fig. 10).

'Insert figures 10 and 11 here'

4.5.2 Glass distribution in the heterogeneous gneiss PX28

In order to discuss glass distribution in heterogeneous gneiss PX28, several zones have been carefully mapped as illustrated in Fig. 12. In He/850/19, one zone encompassing one

quartz-rich layer and one mica-rich layer have been delineated. Size of the zone (5mm x 5 mm) was determined so that enough grains were considered for statistical analysis of the rose diagrams (more than 800 values for each). For He/900/21-32, two distinct zones were chosen. One (Zone 1) is representative of the upper part of the core with alternations of 2 to 3 mm thick, continuous quartz-rich layers with thinner continuous mica-rich layers. Zone 2 was chosen to represent the lower part of the sample with disrupted mica-rich horizons within a quartz-rich matrix. Once again, the size of each zone was determined so that enough grain boundaries were mapped.

In experiments on the heterogeneous gneiss (PX28), mean glass content is $19\pm6\%$ at 850°C (He/850/19) and ranges from 21 ± 7 to $32\pm6\%$ at 900°C (He/900/21-32; Fig. 12 and Table 1). Glass content has been measured across quartz – and mica – rich layers alternations (Fig. 12). Glass content in mica – rich layers (22% in He/850/19, Fig. 12a, and 26% in He/900/21-32, Fig. 12b) is systematically higher than in quartz-rich layers (13% in He/850/19, Figure 12a, and 14% in He/900/21-32, Fig. 12b). At 850°C (in He/850/19), glass is present around muscovite grains, which are still present. At 900°C (i.e. He/900/21-32), muscovite is not observed anymore. Glass is present along and within biotite grains (Figs. 13b & c). In quartz – rich layers, glass is mostly present at grain boundaries (Fig. 13c & d), but in smaller amount than in mica – rich layers (13% in He/850/19 and 14% in He/900/21-32; Fig. 12). In He/900/21-32, glass is present in inter-boudin position within quartz – rich layers (inter-boudin illustrated in Figs. 7a, 13b and located in Fig. 5), where local high glass-fraction is reached (up to 41% in Fig. 13b). In both experiments, melt (observed as glass) is collected within shear zones as illustrated by the example given in Fig. 7a. Glass is also present along cracks in quartz or plagioclase (Figs. 7c & white arrows in 13c) and at quartz-plagioclase grain boundaries (Fig. 13d). Newly formed peritectic phases are present in glass pockets or glass films near micas (Fig. 13). Orientation of remaining micas flakes and solid reaction

products evidence that they were sucked toward dilatant sites like boudin-neck and shear zones together with the moving melt (Figs. 7b, c & d). Heterogeneous distribution of glass is documented in He/900/21-32. Fig. 12b shows two different zones that have different glass fractions of $21\pm7\%$ (zone 1) and $32\pm6\%$ (zone 2). Zone 1 contains a 1 - 3 mm – thick quartz – rich layer and is relatively poor in glass. Zone 2 contains more micaceous layers and shows a greater glass content.

In both experiments considering the mica – rich and quartz – rich layers together, glass areas have a strong preferred orientation normal to compression direction (rose diagrams in Fig. 12). Nevertheless glass orientation varies between mica – rich and quartz – rich layers. In He/850/19 (Fig. 12a), the dominant population of glass veins runs parallel to experimental compression direction in quartz-rich layers (low angle bins in rose diagram), while dominant population in mica-rich layers is perpendicular to experimental compression direction and actually parallel to mica elongation axis. In He/900/21-32, orientation patterns are more complex. Within the upper quartz layer of zone 1, a significant population of veins with low angles to the experimental compression direction also appears (16% of all measured values), while they are absent in the most mica-rich sub-layer despite representing 26% of all measured values for the entire mica rich layer. In zone 2, glass is mainly oriented perpendicularly to experimental compression direction (Fig. 12b).

‘Insert figures 12 and 13 here’

5. Discussion

5.1 Limits of this study

Our aim is to investigate the effect of protolith initial texture and related incipient melt distribution on melt flow during deformation. For this purpose, two sets of experiments on two natural protoliths with particular experimental conditions were performed. Experimental

temperatures were chosen to get water – present and water – absent muscovite melting reactions. A drawback of this approach to experimental temperature control is that these changing conditions had an impact on melt viscosity that could not be measured in our experiments. Nevertheless, published data about haplogranitic melts (Dingwell et al., 1996) show that above 1 wt % H₂O and 700°C, impacts of temperature and hydration are limited and effective viscosities are between 10⁷ and 10⁴ Pa.s, which is more than 6 orders of magnitude below the effective viscosity of surrounding rock (10¹³ Pa.s for a dry Westerly granite (Carter & Tsenn, 1987) strained at 1. 10⁵ s⁻¹ and 750°C).

Strain rate is another parameter that changes between experiments. The experiments with different strain rate steps (Ho/850/33, He/850/19 & He/900/21-32) show that stress increases with strain rate (Fig. 3). For instance, experiment Ho/850/33 exhibits an increase from 9.2 to 31 MPa when the strain rates increases from 1*10⁻⁵ to 1.25*10⁻³ s⁻¹. Stress is multiplied by 3.3 whereas strain rate is multiplied by 125, thus stress increase with strain rate remains limited. Strain rates therefore seems to be a second-order.

5.2 The meaning of glass locations and glass amounts

Until now, we have been describing glass locations and amounts in our experiments. However, our aim is to look at melt distribution and melt flow associated with deformation of partially molten rocks. Thus, it is important to understand what “glass” represents. When performing deformation, the temperature was higher than 750°C, which is far above the glass to melt transition (Webb and Knoche, 1996). This state means that during deformation, melt was present, and that the two gneisses were indeed partially molten. The transition from melt to glass occurs at the end of the deformation event, when the oven is turned off leading to rapid temperature decrease. Glass would therefore represent the melt, as usually considered (e.g. Arzi, 1978; Misra *et al.*, 2009). Following this logic, glass locations and glass amount

represent melt locations and melt amounts, respectively. Thus, we will now speak of melt instead of glass.

5.3 Potential link between stress vs strain curves and microstructures

Experiments on the homogeneous gneiss NOP1 show different stress vs. strain curves: the low temperature and low melt-content experiment (Ho/750/6) shows hardening whereas the experiment performed at higher temperature and melt content (Ho/850/33) shows steady state at high finite strain (Fig. 3). Hardening behaviour was previously observed in experiments for highly crystallized magma (crystal fractions larger than 0.5 and up to 0.8; Arbaret *et al.*, 2007; Forien *et al.*, 2011). From these experiments performed in torsion, the authors concluded that strain was mainly accommodated by the development of synthetic normal Riedel – type shear zones that, associated with Trust – type shear bands, formed an anastomosed network of shear zones delimitating lenses of less deformed melt-bearing zones. In the shear zones, intragranular fracturing (i.e. crystal breakage) was identified (Arbaret *et al.*, 2007). This crystal breakage is inferred to be related to intense stress at contact between grains in localized zones submitted to large shear strains (Forien *et al.*, 2013). In addition, they documented shear hardening in experiments where the crystals suffered intense fracturing, whereas steady – state deformation occurred in experiments without any generalized crystal breakage (Forien *et al.*, 2011). In our series of experiments on the homogeneous gneiss NOP1, quartz – grain breakage (green dashed line in Fig. 4a and cracks in Fig. 8b) is observed in Ho/750/6 while it is absent for the deformation experiment at higher temperature with a larger melt fraction (Ho/850/33, Fig. 4b). Even if shear zones and crystal breakage are less developed in Ho/750/6 than in the study of Forien *et al.* (2011), we conclude that strain hardening behaviour of Ho/750/6 probably results from localized shear zones

together with regions of grain breakage that mainly accommodated the bulk sample deformation.

Experiments on the heterogeneous gneiss PX28 also exhibit different behaviours (Fig. 3). He/900/21-32 experiment is characterized by stable low shear stress at large strains, whereas He/850/19 seems to show a peak of high stress followed by the very beginning of shear softening. A similar behaviour with shear softening following the high – stress peak was experimentally reproduced in crystal-melt mixtures by Picard *et al.* (2011). They showed that this behaviour was associated with the intense development of discrete Riedel shear zones that localized the deformation. Numerous small shear zones are observed in He/900/21-32. Based on this previous study, we conclude that the rheological behaviour exhibited by the experiment He/850/19 is controlled by the early initiation of shear zones that subsequently developed over the entire volume of the deformed core. In He/900/21-32, shear zones are less developed and do not transect over the entire volume of the sample, so the strain-softening behaviour is absent.

The difference between He/850/19 showing strain softening and He/900/21-32 showing a steady state could also be explained by their difference in shear zone orientations.

Indeed, shear zones development and geometry are indicative of mechanical properties of the studied aggregates (Veermer and De Borst, 1984; Veermer, 1990). Development of compactant shear zones is expected to induce weakening of the overall material, while dilatant shear zones are associated with bulk hardening (Veermer and De Borst, 1984). Compactant (respectively dilatant) shear zones (with negative, respectively positive, compaction angles) are expected to develop at angles higher (respectively lower) to σ_1 (equivalent to the experimental compression direction in our study) in the Roscoe model (Veermer, 1990) considered as relevant for high confining pressures, as in our experiments (Le Pourhiet, 2013). The observed distributions could reflect this contrast with shear zones

developing at higher angle in He/850/19 than in He/900/21-32. Their distribution in quartz – or mica – rich layer could also explain these differences. In He/900/21-32, most low-angle shear zones were measured in quartz layers, while high-angle shear zones were mostly measured in mica – rich layers in both He/900/21-32 and He/850/19 experiments. Within the layered gneiss, melt produced by mica partial melting is present in greater amounts in mica-rich layers, indicating that partial melting occurred preferentially in mica – rich layers. In addition, microstructural observations show that melt primarily located in the sheared micaceous layers flowed toward the quartz-feldspathic layers and filled the low-angle shear zones. Thus, one could consider that shear zones developing in mica – rich layers release melt, and therefore are compactant, while shear zones developing in quartz – rich layers trap melt, and therefore dilate. In He/900/21-32, development of dilatant shear zones in quartz and compactant shear zones in mica promote drainage of melts from mica – rich layers to quartz – rich layers without significant weakening, while in He/850/19, shear zones in mica – rich layers prevail and anastomose in a sample-scale compacting shear zone inducing overall weakening.

5.4 Impacts of gneissic layering on melt flow

Melt distribution within partially molten rocks first reflects the initial distribution of melt production sites. In our experiments, we focused on water – present and muscovite – dehydration melting reactions to test the change in melt distribution during deformation of two kinds of partially molten rocks: one rock with incipient melt homogeneously distributed (NOP1) and one with a heterogeneous melt distribution that mimics gneissic layering of the protolith (PX28).

In the case of the homogeneous gneiss NOP1, sites of melt production are directly related to muscovite locations. Muscovite is homogeneously distributed at sample scale,

564 which means that in both experiments (Ho/750/6 & Ho/850/33) melt is produced all over the
565 sample. This interpretation is corroborated by the fact that melt contents calculated in
566 different locations are equal at the sample scale. Results of the experiments performed on the
567 homogeneous migmatite show that melt movement is limited to melt – filled injections in
568 quartz grains at μm -scale and melt does not flow from initiation sites into possible pathways
569 such as shear zones. Shear zones are observed in the deformed homogeneous gneiss (Fig. 4).
570 However, they do not correspond with high melt areas, implying that they do not concentrate
571 melt, which means that melt did not move toward shear zones. The lack of melt movement
572 may be explained by (1) limited melt – pressure gradients within the sample, (2) limited
573 connectivity of the melt or (3) a high melt viscosity. A high viscosity may be rejected because
574 the melt was hydrated, which means that its viscosity is low (Shaw, 1965). Melt connectivity
575 could not be estimated in this study. However, previous studies on hydrated partially molten
576 granite or aplite, i.e. two homogeneous rocks, show that melt is connected as soon as 7% melt
577 is present (see in Rosenberg and Handy, 2005). Mean melt content in Ho/750/6 is really close
578 to 7% (Table 1), even if it may be less locally. Thus, melt connectivity is likely to have
579 occurred, at least in some portions of the sample volume. However, melt connectivity is not
580 sufficient to promote melt flow. A driving force is needed. Therefore, limited melt transport
581 more probably relates to limited melt – pressure gradient within the sample. As melt is
582 produced homogeneously at sample scale, there is no melt concentration gradient and thus no
583 melt pressure gradient is created. The lack of melt – pressure gradient probably prevents melt
584 movement. Thus, the melt distribution in the homogeneous gneiss NOP1 is mainly controlled
585 by the location of melting reactants, which is controlled by the initial protolith fabric and the
586 initial spatial distributions of the grains for the different mineral components (known for static
587 conditions, Mehnert *et al.*, 1973).

In contrast, melt is heterogeneously distributed in experiments on the heterogeneous gneiss PX28 (He/850/19 & He/900/21-32). Evidences of strong melt flow as exemplified by strong alignment of peritectic minerals and the presence of melt+peritectic minerals in shear zones and interboudin positions are observed for the deformed layered migmatite (PX28; Fig. 7). In these experimentally deformed rock samples, gneissic layering strongly enhanced the production and coalescence of melt along mica – rich layers. This enhancement causes a strong pressure gradient between the two compositionally different layers, i.e. between mica – rich and quartz – rich layers. This heterogeneous distribution of melt increases the strength difference and enhances deformation partitioning between mica – rich and quartz – rich layers. This contributes to localizing deformation along shear zones and boudin-necks acting as low melt – pressure sites into which melt migrates efficiently.

5.5 Comparison with natural migmatites

In nature, incipient partial melting leads to the formation of metatexite, where paleosome dominates, and where crystallized melt appears as leucosomes arranged as elongated pockets that parallel the protolith planar fabric (Fig. 14). A similar pattern of melt distribution parallel to protolith foliation is observed in our experiments for both homogenous and heterogeneous starting materials but with differences in melt flow during experiment. We propose that the particular orientation of melt perpendicular to the compression direction reflects a strong control of the initial orientation of reactants. This hypothesis is compatible with the process of microstructure development proposed by Holyoke and Rushmer (2002), (Figs. 13 and 14 in Holyoke and Rushmer, 2002)

In the case of the homogeneous gneiss NOP1, melt distribution is not disturbed by deformation and melt does not migrate. Melt remains located in films perpendicular to shortening direction (Figs. 10 & 12), as already observed by Holyoke & Rushmer (2002),

which is opposite to results from numerous previous studies where melt concentrates within cracks and grain boundaries that are parallel to the shortening direction (e.g. van der Molen & Paterson, 1979; Rutter & Neumann, 1995).

For the heterogeneous migmatite (PX28), the early melt distribution is modified during deformation and subsequent melt migration. Shearing along melt-bearing micaceous layers, shear bands development and boudinage provide efficient pathways for melt migration. The co-location of melt products with such positions is observed within natural migmatites where due to competency contrast between various layers, leucosome is located in dilatant structural sites, such as the spaces between boudins, in pressure shadows, or in fractures in the more competent layers of migmatites (Figs. 14b, c). Occurrences of leucosomes, which derived from crystallized melt, at the neck of boudins are described in numerous terrains (e.g. Port Navalo French Armorican Massif: Marchildon & Brown, 2003; British Columbia: Sawyer, 2008; Western Gneiss Region Norway: Ganzhorn *et al.*, 2014) and may be compared to melt flow observed in experiments on the heterogeneous gneiss PX28, even if the scales are different. Shear zones and more particularly the ones that transect mica-rich and quartz-rich layers, build connection between high-pressure and low-pressure zones, and thus, enhance melt flow as observed in nature with crosscutting leucosomes feeding granite dyke networks.

6. Conclusion

Comparison of pure-shear experiments on a homogeneous gneiss and a heterogeneous layered gneiss shows that initial compositional layering of the protolith have a great influence on melt distribution and melt flow. In the homogeneous gneiss, the main result is the lack of melt migration toward melt collection zones during deformation experiments. In contrast, experiments on the strongly layered gneiss show deformation partitioning between mica –

rich and quartz – rich layers, creation of interconnected pathways that drain and collect melt efficiently. The presence of gneissic layering with the alternation of fertile and non-fertile layers enhances melt – pressure gradients, deformation partitioning with boudinage and shear zone development, driving melt flow through the partially-molten rock. Consequently melt flow within a metatexite is greatly enhanced if the protolith of the partially molten rock shows an inherited gneissic layering.

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Figure Captions

Figure 1: Initial structure of the two starting materials: the homogeneous gneiss NOP1 (a, b) and the heterogeneous gneiss PX28 (c, d). a) Cross-polarized image of NOP1 thin section showing the homogeneous distribution of phases. b) Grain size differences between quartz aggregates and plagioclase+mica aggregates in NOP1 (cross-polarized image). c) Cross-polarized image of thin section showing mineral layering in PX28. d) Mica-rich layer showing shear-zones in PX28. Pl: plagioclase, Qtz: quartz, Bt: biotite, Ms: Muscovite.

Figure 2: SEM images of textures related to partial melting reactions. a) Glass between muscovite and quartz in hydrated undeformed homogeneous gneiss NOP1 (NOP1-B in Table 1), indicating water – present partial melting. b) Glass between quartz and plagioclase indicating the following water – present partial melting reaction: $\text{plg} + \text{qtz} + \text{H}_2\text{O} = \text{Melt}$. The same geometry is observed in cores from NOP1. c) Glass between muscovite and quartz in the experiment at 900°C on the heterogeneous gneiss, indicating the muscovite – dehydration partial melting. A similar reaction occurs in NOP1 (NOP1-A). d) Newly-formed peritectic phase flow around biotite. Pl: Plagioclase, Ms: muscovite, Bt: biotite, Qtz: quartz, Gl: glass.

Figure 3: Stress vs. strain curves for both sets of experiments. Numbers labelled on curves are strain rates.

Figure 4: Sample-scale deformation in the homogeneous gneiss NOP1 at 750°C (a) and 850°C (b). Black arrows: the experimental compression direction. Dashed line rectangles: Locations for SEM images from Fig. 8.

803

804 Figure 5: Sample - scale deformation in the heterogeneous gneiss PX28 at 850°C (a) and
805 900°C (b). For each experiment: left: SEM image, middle: photo of the thin-section, right:
806 drawing indicating the main deformation structures. a) He/850/19 experiment showing two
807 sample-scale shear zones. b) He/900/21-32 experiment showing local shear zones and
808 boudinage of quartz – rich layer. Red lines associated to arrows: local shear zones. Black
809 arrows: the experimental compression direction. Dashed line rectangles: Locations for SEM
810 images from Fig. 7.

811

812 Figure 6: Shear zone orientations in experiments on the heterogeneous gneiss PX28. N:
813 number of shear zone used in the diagram. Mean: average orientation of shear zones. Colour
814 corresponds to the location of shear zone in regards to adjacent grains (e.g. quartz/quartz
815 means that the shear zone is between quartz grains). σ_1 : experimental compression direction.

816

817 Figure 7: Grain-scale deformation in experiments on the heterogeneous gneiss PX28. a)
818 Biotite showing a sigmoid-like shape and broken biotites aligned along a plan representing a
819 shear zone. Location in Fig. 5a. b) Biotite and newly formed peritectic phase flow with glass
820 in quartz – rich layer at the location of boudinage (see Fig. 5b). d) Change in newly formed
821 peritectic phases orientation indicating flow of glass and peritectic phases. Fi: initial
822 orientation (corresponding to the foliation) of newly formed peritectic phases. c) Glass with
823 newly formed peritectic phase along cracks in plagioclase grains. All newly formed peritectic
824 phases are oriented parallel to crack walls. The light grey phase in contact between glass and
825 plagioclase is a K-rich phase, which may be due to reaction between glass and plagioclase.

826 Bt: biotite, Qtz: quartz, Gl: glass, Pl: plagioclase, Kfsp: K-feldspar.

Figure 8: Grain – scale deformations in experiments on the homogeneous gneiss NOP1. a) Kink bands in micas. Dashed lines: axial plan of kink bands. b) Empty cracks in quartz grain. The curved crack along a glass filled crack is due to decompression. c) & d) Glass filled fracture network in quartz grains. c) White dashed lines: contours of glass filled fracture network. Red dashed line: shear zone. d) Arrows: glass filled fracture network in quartz. e) Glass filled injections. Dash lines highlight injections shape. Locations of b), c), d) and e) are given in Fig. 4. Ms: muscovite. Bt: biotite, Gl: glass, Qtz: quartz, Pl: plagioclase.

Figure 9: a) Photomicrograph showing lobate grain boundaries along quartz grains in contact with glass (Gl). b) SEM image showing lobate boundaries along quartz grains in contact with glass. c) Photomicrograph and related schematic drawing (d) showing melt distribution along quartz grain boundaries and intergranular displacements of quartz grains with respect to each other. This might argue for melt assisted granular flow.

Figure 10: Glass distribution in experiments on the homogeneous gneiss NOP1. Glass is represented in black in the drawings. The rose-diagrams represent glass film or pocket orientations. Dark grey population: glass films and pockets perpendicularly oriented to the experimental compression direction (black arrows). Light grey: orientation population corresponding to glass injection. The second population in a) corresponds to melt films and pockets with an oblique orientation to the experimental compression direction. Dashed red lines in b): shear zones (corresponding to the ones in Fig. 8d) showing no increase in melt content at their vicinity.

Figure 11: Local glass distribution in Ho/850/33. a) Location of b, c, and d in Ho/850/33 core. Location of Fig. 10b is also reported. Thick black dashed lines: shear zones defined from Fig.

852 4b. b) Glass associated to micas. White arrows: glass films along muscovite – quartz
853 boundaries often associated with newly formed peritectic phases (blue). Glass with newly
854 formed peritectic phases is only found close to micas. Black arrows: glass always associated
855 to newly formed peritectic phases along mica cleavage. c) Glass filled crack (white arrow)
856 parallel to the experimental compression direction (black arrows) in quartz grain. Thin dashed
857 black lines mark the shape of the crack. Thick oblique dashed lines: shear zones defined from
858 Fig. 4b. Glass in not particularly accumulates at the vicinity of shear zones. d) Glass filled
859 cracks free of newly formed phase in quartz supported network.

860 Glass estimates correspond to blue and red colours. ϕ : newly formed peritectic phases. Thick
861 vertical black arrows: experimental compression direction. Bt: biotite, Ms: muscovite, Gl:
862 glass, Kfs: K-feldspar, Pl: plagioclase, Qtz: quartz. Other: accessory minerals and K-rich
863 phase in contact between plagioclase and melt.

864

865 Figure 12: Glass distribution in experiments on the heterogeneous gneiss PX28. Glass is
866 represented in black in the drawings. The rose-diagrams represent glass film or pocket
867 orientations. a) Greater glass amount in mica – rich layer than in quartz – rich layer. Glass
868 films and pockets oriented perpendicularly to the experimental compression direction (black
869 arrows) in mica – rich layer and parallel to the experimental compression direction in quartz –
870 rich layer. b) High variability of glass amount at sample scale and between mica – rich and
871 quartz – rich layers. Black arrows=experimental compression direction.

872

873 Figure 13: Local glass distribution in He/900/21-32. a) Location of the different areas
874 illustrated in b), c) and d). b) Great glass amount associated with newly formed peritectic
875 phases and small biotites in an inter-boudin position (defined in Fig. 5b). Green dashed lines
876 in a) and b) correspond to the boundary of inter-boudin area as defined in Fig. 5b. c) Glass

veins, free of newly formed phase, in quartz grain (white arrow). Glass is also present within plagioclase grain. d) Glass films, free of newly formed peritectic phase, at quartz – plagioclase boundaries.

Glass estimates correspond to blue and red colours. ϕ : newly formed peritectic phases. Other: accessory minerals and K-rich phase in contact between plagioclase and melt. Bt: biotite, Ms: muscovite, Gl: glass, Kfs: K-feldspar, Pl: plagioclase, Qtz: quartz.

Figure 14: a) Outcrop photograph showing a metatexite where the onset of partial melting is marked by ca. 20 cm long lense-shaped leucosomes that lie parallel to the planar fabrics of the protolith, Fuping complex (North China Craton). b) Stromatic metatexite of paragneiss where boudinage of quartz-rich competent layers led to melt migration toward boudin-neck (white arrows), southern Velay dome (French Massif Central, France).c) Stromatic migmatite with a set of leucosomes that parallels the gneissic layering and a second set of leucosomes located in shear zones. Melt migrates toward shear zones, Archean Snowbird tectonic zone (Canada, courtesy from Philippe Goncalves).

Table caption

Table 1: Experimental conditions used in deformation experiments and parameters derived from the experiments. PP123 is a hydrostatic experiment. Experiment names are built on the following schema: “type of material/temperature (°C)/melt content (surfacic percentage)”. “Ho” means homogeneous gneiss and “He” means heterogeneous gneiss. In the particular case of He/900/21-32, it is not possible to define a meaningful mean main content as melt content varies a lot through the sample (see Fig. 12). Thus we use both estimates made in two different zones (see part 4.5).

*Figure1
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*Figure2

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a) He/850/19: $\dot{\gamma}=10^{-4}$ - 10^{-3} s^{-1} , Strain=20%, Glass=19.2 $\pm$ 5.7%

b) He/900/21-32: $\dot{\gamma}=10^{-4}$ - 10^{-3} s^{-1} , Strain=27%, Glass=21 $\pm$ 7 - 32 $\pm$ 6%

*Figure6
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***Figure8**
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*Figure9

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a) Ho/750/6: $\gamma=5 \cdot 10^{-5} \text{ s}^{-1}$; Strain=20% ;
Glass=6 $\pm$ 3%

b) Ho/850/33: $\gamma=10^{-5}-1.25 \cdot 10^{-3} \text{ s}^{-1}$; Strain=11% ;
Glass=33 $\pm$ 4%

*Figure11

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*Figure12
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***Figure13**
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a He/900/21-32

b Melt=41%

c Melt=20%

d Melt=8%

***Figure14**
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Table1

experiment name	experiment number	starting material	jacket	size diameter (mm length (mm)		T (°C)	Strain rate (s ⁻¹)	duration (hour)	Strain (%)	melt content (%)
Protolith: NOP1 (Homogeneous)										
Ho/850/33	PP218	NOP1-A	Fe+Pt	15	15.19	850	10 ⁻⁵	0.6	2.1	
							5*10 ⁻⁵	0.3	3	
							2.5*10 ⁻⁴	0.06	3	
							1.25*10 ⁻³	0.02	3	33
Ho/750/6	PP315	NOP1-B	Fe+Cu	5	10.29	750	2*10 ⁻⁵	2.87	20	6
Protolith: PX28 (Heterogeneous)										
-	PP123	-	Fe+Pt	14.91	19.42	850	0.0	1.75	0	unknown
He/900/21-32	PP111	-	Fe+Pt	15	28.8	900	10 ⁻⁴	0.28	10	
							10 ⁻³	0.05	17	21-32
He/850/19	PP113	-	Fe+Pt	15	21.74	850	10 ⁻⁴	0.23	8.4	
							10 ⁻³	0.02	8.8	19