PETROLOGICAL AND STRUCTURAL FEATURES OF THE SCHLIEREN MIGMATITES

In the El Oro Complex (EOC), the migmatitic sequence of the La Bocana unit is composed of metatexite (1-2 km thick) and garnet-bearing schlieren migmatite (7-8 km thick)(Fig. 1). These two zones correspond to two partial melting reactions related to muscovite-breakdown for the metatexite (garnet-free) and biotite-breakdown for the garnet-bearing schlieren migmatite (Riel et al., 2013). The transition zone between the metatexite and the schlieren migmatite is marked by an increase of leucosome proportion, by the progressive change of the rheological behavior of the rock, from solid-dominated to magma-dominated, and by the appearance of peritectic garnet (Riel et al., 2013).

The schlieren migmatite zone is bounded by the magmatic contact with the gabbroic unit of Piedras to the north, and by the metatexite unit to the south (Fig. 1). At the scale of the schlieren migmatite zone and excluding the metasedimentary xenoliths, the peak pressure-temperature (P-T) conditions paragenesis of the rock is reasonably homogeneous (Figs. DR2 and DR3A,B,C,D), composed of the mineral assemblage garnet + biotite + quartz + plagioclase + K-feldspar + sillimanite ± ilmenite ± rutile (see Riel et al., 2013). Partial melting is controlled by biotite-breakdown at P-T conditions ranging from 760 ± 50 °C at 5.5 ± 1 kbar at the top of the zone close to the metatexites, to 900 ± 50 °C at 8.5 ± 1 kbar at the base, close to the gabbroic intrusion (Riel et al., 2013; this study). Although the average schlieren migmatite is mesocratic (Fig. 2A), it can vary locally from leucocratic (up to 80% of SiO₂ content) to more melanocratic (down to 60% of SiO₂ content, Table DR1) (Fig. DR2E,F).
Throughout the schlieren migmatite zone, the metasedimentary xenoliths exhibit various metamorphic reactions and mechanical behavior (Figs. DR2 and DR3B, D, E, F). They vary from unaffected to partially molten, depending on the gradient of chemical potential between xenolith and surrounding schlieren migmatite. This is evidenced in Fig. DR2E where a quartzitic xenolith is partially molten on its rim while in Fig. DR2F another quartzitic xenolith is unmelted and instead is fractured and filled with leucosome produced in its surroundings. Similar behaviors can be observed for aluminous xenoliths (Figs. DR2A, DR2D, DR2G). Figure DR2C shows a folded xenolith located at the top of the schlieren migmatite zone and made of quartzo-feldspathic layers (0.5-3cm) intercalated with thin aluminous layers (<0.5 cm). Either this metatexite xenolith has been mechanically incorporated during mass-convection or was present before. The second hypothesis implies that when the temperature increased above the solidus, the metatexite was first produced and then incorporated as a xenolith in the schlieren migmatite when the temperature increased above the biotite-breakdown partial melting conditions. Even if both hypotheses are valid, the geochemical study indicates that a K₂O-rich liquid has been extracted from the schlieren migmatite prior to the biotite-breakdown melting reaction, REE. Consequently, we interpret the presence of layered quartzo-feldspathic xenoliths to indicate a two-step partial melting sequence. (1) Between 650 and 750°C, at muscovite dehydration melting reaction conditions, the metatexites formed and part of the melt was extracted to form the granitoid belt at shallow level of the crust. (2) As temperature rose between 750 and 900°C (Tₘₐₓ recorded in the schlieren migmatite), to biotite dehydration melting reaction conditions, partial melting affected most of the metatexite (except for the top 1-2km), and only resistant lithologies remained as (meta)stable xenoliths. It is worth noting that the "metatexite-like" xenoliths are only observed at the top of the schlieren migmatite zone, near the transition zone to the metatexite. Lower in the schlieren migmatite zone, only quartzitic and aluminous xenoliths are observed.
Figures DR3A to DR3E present a characteristic garnet-bearing schlieren migmatite outcrop located near the La Bocana locality. In this outcrop, curved syn-to-late-migmatitic foliations are preserved (Fig. DR3C) and could potentially represent shear zones between parts of preserved convection cells. However, a detailed structural and micro-structural study is needed to fully understand these structures.

In the El Oro Complex, the garnet-bearing migmatites were previously described as diatexite on the basis of their magmatic appearance by Riel et al. (2013). However, the modelled melt% of the garnet-bearing migmatites from this study and using whole-rock composition from Riel et al. (2013) is relatively low (5-15%, Table DR5), which is incompatible with a standard diatexite for which coherence is lost due to high melt% (generally > 50%). This apparent discrepancy may be first explained by an underestimate of the melt% which is strongly dependent on the water-content used to model the system at $P-T_{\text{max}}$ conditions. Additional water from the migmatite itself or from an external source would enhance partial melting (water-fluxed melting) and result in a higher melt-content (Weinberg and Hasalová, 2015). In the El Oro Complex, we did not find any evidence of aqueous fluids entering the migmatites throughout the garnet-bearing schlieren migmatitic unit: there is no evidence of enriched network of leucosome with diffuse contact or with lack of peritectic minerals. Moreover, the first step of muscovite-breakdown melting reactions occurred at increasing temperature. As shown in the T-XH$_2$O diagrams in Figure DR10, this strongly dehydrated the rock prior biotite-breakdown melting reactions thus precluding influx melting directly from the garnet-bearing migmatites or from the gabbro. Therefore, another mechanism has to be considered to explain the formation of a migmatite with magmatic appearance at relatively low melt-content. In migmatitic orthogneiss from the Gföhl Unit (Bohemian Massif), Hasalová et al. (2008a) proposed that melt infiltration from an external source passing pervasively at grain boundaries can change the microscopic and macroscopic appearance of the rock by progressive disintegration of former layering to produce an apparent
magmatic microstructure. The main limiting factor for their model was attributed to the decrease of $P-T$ conditions during melt infiltration. In the El Oro Complex such a phenomenon could have occurred at increasing temperature conditions, during muscovite-breakdown melting and upward transfer of the related melts to form the schlieren migmatites prior biotite-breakdown melting. This model is supported by the migmatitic study of Riel et al. (2013) showing that the drop of solid framework continuity occurs at the bottom of the muscovite-breakdown melting zone before the appearance of peritetic garnet related to biotite-breakdown melting reactions.

Another feature that has been observed in aluminous xenoliths in the garnet-bearing schlieren migmatite is presented in Figs. DR3E and DR3F. The xenoliths are partially molten and exhibit segregated leucosome and adjacent residuum. On the rim of the xenolith we lose track of the leucosome which becomes mixed in the surrounding schlieren migmatite (Fig. DR3F). As no post-migmatitic deformation overprints the xenolith and the surrounding schlieren migmatite, this shows that the rheology of the xenolith and of the surrounding schlieren migmatite were different during migmatization. In the xenolith, segregation of the melt into a leucosome allowed melt circulation and extraction. In contrast, in the surrounding schlieren migmatite the leucosome is mainly pervasively distributed within the residuum (Fig. DR2E) or in low pressure zones such as fractured quartzite (Fig. DR2F).

Structural evidence of upward/downward motion has also been observed in the schlieren migmatitic unit. Figure DR4 shows a large unmelted, folded metaquartzite layer within the schlieren migmatite forming an eroded sheath fold (Fig. DR4A-B). The hinge of the fold trends towards the south (originally upward) and indicates vertical motion within the schlieren migmatite during partial melting (Fig. DR4C) consistent with vertical mass-flow during convection.
GEOCHEMICAL METHODOLOGY

Whole-rock major element analyses and REE concentrations analyses of unmelted metasedimentary rocks, garnet-bearing migmatites, leucosomes and granitoids were performed at the ISTerre laboratory. Concentrations of major elements were determined by ICP-AES using a Perkin Elmer 3000 DV spectrometer using the method of Cotten et al. (1995) and REE using an Agilent 7500ce quadrupole ICP-MS following the method of Chauvel et al. (2011). The results are presented in Tables DR1 and DR2.

PRESSURE-TEMPERATURE ESTIMATES

In order to complete a previously compiled dataset (Riel et al., 2013), we selected two garnet-bearing samples MA-10-24 [79.929W-3.663S] and TO-10-10 [79.973W-3.643S] (Fig. DR1) located in the northernmost part of the migmatitic unit, 1km and 0.1 km from the contact with the gabbro, respectively. These two samples are garnet-bearing mesocratic schlieren migmatites, in which the leucosome is homogeneously distributed within the rock, resulting in a granitic texture. The selected area for microprobe mapping of sample MA-10-24 is a melanocratic schlieren composed of the characteristic mineral assemblage of biotite + sillimanite + garnet + quartz + plagioclase + K-feldspar + ilmenite + rutile + graphite (Fig. DR5A).

In sample TO-10-10 (Fig. DR6A), garnet shows elongated and embayed shapes. Plagioclase is strongly recrystallized in a micrometric quartz + sericite assemblage. For microprobe mapping, we selected an area where unaltered plagioclase is preserved. The selected zone is composed of the characteristic mineral assemblage of garnet + quartz + plagioclase + K-feldspar + quartz + ilmenite + graphite (Fig. DR6A). Note that this sample is the only one in which sillimanite has not been observed.

Pressure-temperature ($P-T_{\text{max}}$) conditions were estimated using forward modeling and based on the results of standardized microprobe maps on MA-10-24 and TO-10-10 samples. The X-ray
compositional maps were acquired at ISTerre using a JEOL JXA-8230 electron microprobe. Analytical conditions for spot analyses were 15 keV accelerating voltage, 10 nA specimen current and ~3 seconds counting time. Standards used were Fe$_2$O$_3$ (Fe), MnTiO$_3$ (Mn, Ti), diopside (Mg, Si), orthoclase (Al, K), anorthite (Ca), and albite (Na). The compositional mapping was carried out at 200 ms per pixel (see De Andrade et al., 2006, for a detailed statistical evaluation). Higher specimen current of 100 nA was employed for mapping. The X-ray maps were standardized using XMapTools program (Lanari et al., 2014). This software allows identification of mineral phases and conversion of X-ray images (intensities of recorded counts) into oxide wt% concentrations, using the spot analyses as standards. Structural formulae are then estimated using the functions available in the software (Lanari et al., 2014). Subsequently, pressure and temperature were estimated using thermodynamic forward modeling based on the whole-rock composition of MA-10-24 and TO-10-10 samples (Table DR1) and the program Perple_X’07 (Connolly, 2005) with the internally consistent thermodynamic dataset of Holland and Powell (1998) update TC55. Details of the solid-solutions used here are presented in Table DR3. The water-content used for the calculations corresponds to the minimum amount of water necessary to saturate the subsolidus assemblage just below the solidus with no free water (see Table DR5 and Fig. DR7). Moreover, T-XH$_2$O diagrams show that the saturating water-content at subsolidus conditions is not pressure-dependent (Fig. DR7). Fe$^{3+}$ is not considered in our calculation as because in graphitic metapelite subjected to migmatization, Fe$^{3+}$ contained in biotite at sub-solidus conditions is subsequently released during partial melting, resulting in the formation of CO$_2$ and FeO-bearing peritectic minerals (Cesare et al., 2005), i.e. in this case garnet in the schlieren migmatites (Riel et al., 2013).

Oxide maps based on a microprobe study are shown in Figs. DR5 and DR6. For MA-10-24, the X$_{Alm}$ content of garnet (Fig. DR5F) exhibits a well-defined zoning from 0.63 in the core, 0.64-0.65 in the mantle and up to 0.74 in the outer rim. In contrast, the X$_{Grt}$ content of garnet does not show
evidence of zoning and has a value of ~0.04 (Fig. DR5E). Within the matrix, $X_{Mg}$ content of biotite varies from 0.62 to 0.58 (Fig. DR5C) and the Ti-content from 0.2 to 0.34 a.p.f.u (Fig. DR5B). The highest values of $X_{Mg,bt}$ in biotite (0.67) are observed in small rounded inclusions of biotite within garnet (Fig. DR5C). This high $X_{Mg}$ content >0.66 is above the maximum $X_{Mg}$ modelled using forward modeling (0.62, Fig. DR8) and likely represents biotite that formed during peritectic garnet crystallisation at the expense of pre-existing biotite, which is supported by the fact that the Ti-content of these inclusions is relatively low (0.28). Note that the part of the garnet with the highest $X_{Alm}$ content (0.74) is in contact with the biotite crystal with the lowest $X_{Mg}$ and Ti-content (0.54 and 0.22, Fig. DR5 B,C, white circle). This area shows that the low $X_{Mg}$ in biotite crystallized from a K-feldspar crystal likely destabilized by late fluid and/or felsic melt interactions during retrogression (Fig. DR5C). The $X_{Ab}$ content of plagioclase varies from 0.53 to 0.58, with the lowest values (0.53) found in crystals included in garnet (Fig. DR5D).

In the map for sample TO-10-10, the $X_{Alm}$ content of garnet (Fig. DR6F) shows little zonation from core to rim and the average $X_{Alm}$ content is 0.65 to 0.68. The $X_{Mg}$ content of biotite shows two ranges of values: from 0.57 to 0.60 in matrix crystals and from 0.64 to 0.67 in crystals included in garnet (Fig. DR6C). As for MA-10-24, the $X_{Mg}$ of biotite inclusions in garnet is higher than the maximum $X_{Mg}$ modelled (0.60, Fig. DR8) and have a lower Ti-content (0.27) than the one recorded in the matrix (0.30) and thus is likely to indicating metastability. Plagioclase is highly saussuritized (Fig. DR6D). The $X_{Ab}$ content was only measured in the core of the bottom plagioclase crystal (Fig. DR6D) and shows values around 0.62.

At the first order, the $P-T_{max}$ conditions of both samples (MA-10-24 and TO-10-10) are constrained by the P-T field of the characteristic mineral assemblage between 840 and 900 °C and 6.0 to 10 kbar (Figs. DR8 and DR9). Representative compositions of plagioclase, biotite and garnet extracted from the quantified X-ray maps and used to restrict the $P-T_{max}$ conditions are
shown in Table DR4. The isopleths of solid solutions (garnet, biotite, plagioclase) are shown in Figs. DR8 and DR9 and provide P-T estimates during partial melting.

For sample MA-10-24, the $P-T_{\text{max}}$ compositional equilibrium is obtained using the representative values of mineral phases in the matrix and in the garnet core (Table DR4) at 880 ± 50 °C and 8 to 9 kbar with: $X_{\text{Alm}}$ content = 0.64 (garnet core), $X_{\text{Grs}}$ content = 0.04 (garnet core), $X_{\text{Mg}}$ content = 0.59 and Ti-content = 0.30 of biotite in matrix and $X_{\text{Ab}}$ content = 0.57. For sample TO-10-10, garnet shows little zoning (Fig. DR6E,F). The isopleths of garnet, biotite and plagioclase and their representative composition (Table DR4) predict the observed garnet to be stable at $P-T_{\text{max}}$ of 880 ± 50 °C between 7 and 9 kbar, with: $X_{\text{Alm}}$ content = 0.66 (garnet core), $X_{\text{Grs}}$ content = 0.045 (garnet core), $X_{\text{Mg}}$ content = 0.59 and Ti-content = 0.30 of biotite in matrix (Fig. 5f). However, the $X_{\text{Ab}}$ content of 0.61 is higher than the one modelled at $P-T_{\text{max}}$ conditions of 0.55 and is characteristic of lower temperature conditions (< 720°C, Fig. DR9). This likely shows that: either the plagioclase was retrogressed during cooling, crystallized at lower temperature or altered during saussuritization.

The sensitivity of the modelled characteristic mineral assemblage field to the bulk-composition is an important feature as this allow to extrapolate the $P-T_{\text{max}}$ conditions throughout the whole schlieren migmatitic unit. To study the effect of the bulk-composition in the schlieren migmatites, we reported in Fig. DR10 the boundaries of the stability field of the modelled characteristic mineral assemblage Gt + Bt + Qtz + Pl + Sil + Kfs ± Ilm ± Ru of all studied samples from Riel et al. (2013) and this study (Table DR5). The figure shows that the position of the characteristic mineral assemblage fields is relatively insensitive to studied bulk-rock compositions. Moreover, Riel et al. (2013) showed that schlieren migmatite melt removal/addition (from -10% up to +50 of melt %) has little or no effect on the $P-T_{\text{max}}$ estimates. Therefore, we are confident that the $T_{\text{max}}$ conditions in the whole schlieren migmatitic unit were between 760 and 900°C.
In contrast, the absolute pressure conditions are less constrained and can vary from 5 to 10 kbar (within error bars). Because the inherent error estimates on pressure using thermodynamic modelling is large (± 2 kbar), a key aspect of this study was to retrieve a more precise anatectic metamorphic gradient. As schlieren migmatites are not or only weakly affected by post-migmatitic deformation and the characteristic mineral assemblage is consistent across the schlieren migmatitic layer (Riel et al., 2013), we reconstructed the pressure in the schlieren migmatite using the relative structural distances between studied samples from the metatexite for which the pressure is better constrained (e.g. $P\text{-}T_{\text{max}}$ estimates in Fig. DR10).

**DETAILED NUMERICAL BASIS**

We developed a 1D numerical model, which implicitly solves the conductive heat equation and takes into account heat production by radioactive elements, melting enthalpy of muscovite and biotite, temperature-dependent thermal conductivity and capacity, variable melting temperature and melt extraction (1).

\[
\rho C_p(T) \frac{dT}{dt} = \kappa(T) \frac{\partial^2 T}{\partial x^2} + H + \Delta EMs + \Delta EBg + \Delta EMigration
\]

where $\rho$ is the metasediments density (kg.m$^{-3}$), $C_p(T)$ is the heat capacity (J.kg$^{-1}$.K$^{-1}$), $T$ the temperature (K), $t$ the time (s), $\kappa(T)$ the thermal conductivity (J.s$^{-1}$.m$^{-1}$.K$^{-1}$), $H$ the radiogenic heat production (J.s$^{-1}$.m$^{-3}$), $\Delta EMs$ and $\Delta EBg$ the heat captured by the melting enthalpy of muscovite and biotite, respectively (J.s$^{-1}$.m$^{-3}$), and $\Delta EMigration$ the heat transport due to melt extraction (J.s$^{-1}$.m$^{-3}$). $\rho$ and $C_p$ are considered constant across the model while $\kappa(T)$ and $C_p(T)$ are temperature dependent (Whittington et al., 2009). The values of these parameters are summarized in Table DR6. We use an implicit conservative scheme to solve this equation. The resulting matrix system is inverted using LAPACK (Anderson et al., 1999) and BLAS libraries (Blackford et al., 2002). We used a Picard iteration scheme to determine the melting temperature.
of mica at every time step. The modelled section is 29km thick and discretized over 290 nodes; time step is 100 years. We used a constant temperature boundary condition at the surface and a constant heat flux or temperature condition at the (top of gabbro) bottom boundary of the model. The initial thermal gradient is computed at equilibrium with the boundary conditions. We simulate the gabbroic intrusion by imposing an initial temperature of 1200 °C between 25 and 29km depth. Then, we maintain this temperature for a variable amount of time at the bottom boundary of the model (but not at the top of the gabbro). The average composition of the starting unmelted crustal material contains 23% of muscovite and 23% of biotite which represents the characteristic content of these minerals in the unmelted metasedimentary rocks of the EOC. In our model we assume linear melting of muscovite between 680 and 720°C and between 780 and 900°C for biotite.

For the models that include the thermal effects of convection, we use a parametrization of convection based on the Nusselt-Rayleigh number relationship of Turcotte and Schubert (2002):

\[ Nu = 0.120 Ra^{1/3} \]

Where Nu and Ra are the Nusselt and Rayleigh number, respectively. We estimate the Rayleigh number using the physical parameters of the garnet-bearing schlieren migmatitic unit:

\[ Ra = \frac{\rho_0 g \alpha \Delta T h^3}{\eta \kappa} \]

where \( \rho_0 \) is the density (2900 kg.m\(^{-3}\)), \( g \) is the gravity acceleration (9.81 m.s\(^{-2}\)), \( \alpha \) is the thermal expansivity (3\( \times \)10\(^{-5}\) K\(^{-1}\)), \( \Delta T \) (100 °C), is the temperature variation across the convective layer, \( h \) is the thickness of the convective layer (8000 m), \( \eta \) is the viscosity (10\(^{13}\) to 10\(^{16}\) Pa.s) and \( \kappa \) is the thermal diffusivity (10\(^{-6}\) m\(^2\).s\(^{-1}\)). Knowing that the Nusselt number is the ratio between convective and conductive heat flux:
\[ \text{Nu} = \frac{Q}{k \frac{\Delta T}{h}} \tag{4} \]

where \( Q \) is the convective heat flux and \( k \frac{\Delta T}{h} \) is the conductive heat flux, we can estimate the convective heat flux by using equation (2) and (4). Finally, we sum up the convective and conductive heat fluxes and calculate the corresponding hyperdiffusivity (\( K \)):

\[ K = \frac{hQ}{\Delta T} + k \]

Using viscosity range between \( 10^{16} \) and \( 10^{13} \) Pa.s, we determine a hyperdiffusivity between 8 and 84 \( \text{J.s}^{-1}.\text{m}^{-1}.\text{K}^{-1} \).

As our P-T estimates record the \( P-T_{\text{max}} \) conditions during partial melting, we save the maximum temperature reached at any given depth during the simulation. Therefore in Fig. 5 each curve represents the maximum temperature recorded over the duration of the model. Model results show that the time scale of change of the temperature gradient exceeds 1Ma (Fig. DR10), which implies that the kinetics of metamorphic reactions can be neglected and the \( T_{\text{max}} \) recorded by the model can directly be compared to the thermo-barometric data.

Based on our modeling results (\( K_{\text{max}} = 18 \) and heating = 3.5 m.y.) we find that the total heat transferred to the continental crust is of \( 1.5 \times 10^{13} \) J. This is slightly higher than the total energy stored in a 4 km thick batch of gabbroic magma (\( 1.1 \times 10^{13} \) J). The supply of the needed extra heat can be attributed to an uplifted hot mantle along the South American margin during Permian-Triassic times (Riel et al., 2013).
Figure DR1. Sample localities and $P$-$T_{\text{max}}$ estimates in the El Oro Complex. (1) This study, (2) data after Riel et al. (2013). Grt: garnet, Ilm: ilmenite, Ru: rutile.
Figure DR2. Photographs of the garnet-bearing migmatitic unit. See position of photographs on lithological column on right-hand side. A, Aluminous xenolith at the top of the garnet-bearing schlieren migmatitic unit. B, Quartzitic xenolith in typical homogeneous garnet-bearing migmatites. C, Folded quartzo-feldspathic xenolith. D, Partially molten aluminous xenolith. E, Partially molten quartzitic xenolith surrounded by a leucocratic rim with a diffuse contact with the garnet-bearing migmatites. F, Fractured quartzitic layer filled by a quartz- and feldspar-rich leucosome. G, Homogeneous garnet-bearing migmatite at the bottom of the schlieren migmatitic unit. H, Biotite-poor, garnet-bearing migmatite near the contact with the gabbroic unit.
Figure DR3. Representative photographs of the garnet-bearing migmatitic unit. A, 1000 m² outcrop of typical garnet-bearing migmatites. B, large quartzitic xenolith surrounded by a more leucocratic rim of garnet-bearing migmatites. C, curved late-migmatitic foliation. D, Quartzitic xenolith shape. E, Partially molten aluminous xenolith in a homogeneous garnet-bearing migmatite. F, Detail of E. While within the xenolith the leucosome is segregated, in the boundary the leucosome is diffused into the garnet-bearing mesocratic migmatite.
Figure DR4. Domal structure in the schlieren migmatites defined by a metaquartzite block. A, B, photographs of folded metaquartzite layer embedded in schlieren migmatite. The white arrow shows the upward direction. C, 3D sketch of the formation of the dome structure.
Figure DR5. X-ray map of MA-10-24 garnet-bearing migmatite. A, Mineralogy of the studied area. B, Ti-content of biotite (a.p.f.u.). C, $X_{\text{Mg}}$ content of biotite ($\text{Mg}/(\text{Fe} + \text{Mg})$). D, $X_{\text{Ab}}$ content of plagioclase ($\text{Na}/(\text{Na} + \text{Ca})$). E, $X_{\text{Grs}}$ content of garnet ($\text{Ca}/(\text{Ca} + \text{Fe} + \text{Mg})$). F, $X_{\text{Alm}}$ of garnet ($\text{Fe}/(\text{Ca} + \text{Fe} + \text{Mg})$). E and F show how the garnets have a narrow rim enriched in Fe. Mineral abbreviations used in figures and tables are from Kretz (1983).
Figure DR6. X-ray map of TO-10-10 garnet bearing migmatite. A, Mineralogy of the studied area. B, Ti-content of biotite (a.p.f.u.). C, $X_{\text{Mg}}$ content of biotite ($\text{Mg}/(\text{Fe} + \text{Mg})$). D, $X_{\text{Ab}}$ content of plagioclase ($\text{Na}/(\text{Na}+\text{Ca})$). E, $X_{\text{Grs}}$ content of garnet ($\text{Ca}/(\text{Ca} + \text{Fe} + \text{Mg})$). F, $X_{\text{Alm}}$ of garnet ($\text{Fe}/(\text{Ca} + \text{Fe} + \text{Mg})$). Mineral abbreviations used in figures and tables are from Kretz (1983).
Figure DR7. T-XH₂O diagrams of MA1024 and TO1010 samples at 5 and 9 kbar. This pseudosections are calculated using whole-rock composition reported in Table DR5, at fixed pressure and variable water-content from 0.0 to 4.0 wt%. The white arrows show the saturated water-content of the rock at subsolidus conditions which is used for the calculation of MA1024 and TO1010 P-T pseudosections (Figs DR8 and DR9). Note that in the case of the El Oro Complex the saturated water-content at subsolidus is not depending on pressure (for considered range of interest 5 to 9kbar).
Figure DR8. Pseudosection of MA-10-24 sample. Preferred $P-T_{\text{max}}$ estimate is indicated by a black box and is based on the intersection of the characteristic composition of the minerals presented in Table DR4. Modelled mineral proportions at $P-T_{\text{max}}$ are available in Table DR5. Mineral abbreviations used in figures and tables are from Kretz (1983).
Figure DR9. Pseudosection of sample TO-10-10. Preferred $P-T_{\text{max}}$ estimate is indicated by a black box. Modelled mineral proportions are available in Table DR5. Note that similar temperature range to those in Fig. DR10, but higher P, are expected from the relative structural positions between the two samples, this one being deeper in the schlieren migmatite.
Figure DR10. Compilation of modelled characteristic mineral assemblage fields of studied garnet-bearing schlieren migmatites. The modelled fields are based on bulk-rock composition reported in Table DR5. $P-T_{\text{max}}$ estimates of each sample are indicated by grey crosses. Mineral abbreviations used in figures and tables are from Kretz (1983).
Figure DR11. Modelled metamorphic gradient for various periods of $T_{\text{max}}$ integration time. The model is pseudo-convective with a $K_{\text{max}}$ of 25 J.s$^{-1}$.m$^{-1}$.K$^{-1}$ and a heating period of 2 Ma. The results suggest that $T_{\text{max}}$ metamorphic gradient is weakly affected by the change of $T_{\text{max}}$ integration time until 1 m.y. Note that the initial perturbation in the gabbro is rapidly diffused. $P$-$T_{\text{max}}$ estimates of each sample are indicated by grey crosses.
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