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ARTICLE

Trondhjemite leucosomes generated by partial melting of a hornblende-gabbro (Alvand plutonic complex, Hamedan, NW Iran)

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ABSTRACT

This work describes the petrogenesis of in situ and in-source trondhjemite leucosomes generated by low-degree partial melting of the Jurassic Cheshmeh-Ghasaban hornblende-gabbros in the northern Alvand batholith (Hamedan, NW Iran). These leucosomes occur in a metatexitic migmatite as patches, net-structures, veinlets, and dikes at scales ranging from a few millimetres to a few metres. They have high SiO$_2$ (mean ≈ 78 wt%) and Na$_2$O (4–5 wt%) with low Al$_2$O$_3$ (<15 wt%), K$_2$O (<1.5 wt%), Sr (33–267 ppm), Eu (0.31–0.62 ppm) and heavy REE (<4 ppm) contents similar to typical low-Al$_2$O$_3$ trondhjemites. Batch-melting models indicate these leucosomes are the result of ≈1-2% partial melting of the hornblende-gabbro, whereas thermobarometric modelling constrains their petrogenesis in the field of hornblende hornfels-facies metamorphism (pressure ≈3.0–4.5 kbar and temperature ≈700-750°C). On a wider scale, our results document natural trondhjemites generated in the upper crust by amphibole-dehydration melting of a mafic source at temperature conditions close to the solidus, filling the existing gap of data in low-pressure (P < 5kbar) and low-temperature (T < 800°C) amphibole-dominated partial melting experiments. The genesis of these trondhjemites is then discussed with respect to the Mesozoic evolution of the Alvand plutonic complex.

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Supplemental data for this article can be accessed here.

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1 Introduction

Migmatites and their felsic leucosomes are generally considered the products of high-grade metamorphism and anatexis (viz., partial melting events) with or without extraction/seggregation of generated melts affecting crustal rocks (e.g., Harris 1974; Ashworth 1979, 1985; Johannes 1988; Sawyer 2008). In particular, partial melting events characterized by melts separating from the source are considered as one of the most effective processes for the generation up to large volumes of rhyolitic/granitic melts and therefore also for the growth and geochemical differentiation of the continental crust (e.g., Clemens 1990; Vielzeuf and Ph. 1990; Stevens and Clemens 1993; Johannes and Holtz 1996; Liégeois 1998; Clemens and Watkins 2001; Kriegsman 2001; Koepe et al. 2005; Ratajeski et al. 2005; Sisson et al. 2005; Gao et al. 2016; Rossetti et al. 2020 and references therein). In general, crustal anatexis occurs through dehydration melting reactions (e.g., Clemens 1990) of minerals such as muscovite, biotite and amphibole (Gao et al. 2016), corresponding to four major groups of rocks: i) amphibole-bearing rocks, ii) amphibole-biotite-bearing rocks, iii) biotite-bearing rocks and iv) muscovite-biotite-bearing rocks. Partial melts derived from these four groups of protoliths are correlated with four groups of melts with peculiar, distinguishing chemical characteristics (Clemens 1990; Johannes and Holtz 1996; Patiño Douce 1999; Clemens and Stevens 2012; Castro 2013). With the aim to understand and directly constrain the genesis of felsic melts in closed systems through partial melting, in the last decades a great number of melting experiments were performed at different pressure (1–32 kbar) and temperature (670–1100°C) conditions (see Gao et al. 2016 and references therein). These experiments explored the partial melting of protoliths of different nature from mafic magmatic (e.g. Beard and Lofgren 1991; Castro et al. 1999; Sisson et al. 2005) and metamorphic (e.g. Laurie and Stevens 2012; Skjerlie and Patiño Douce 1995; Skjerlie and Patiño Douce, 2002) rocks to intermediate plutonic (e.g., Watkins et al. 2007) and volcanic (Conrad et al. 1988; Skjerlie and Johnston 1996) products, from greywackes (e.g., Vielzeuf and Montel 1994; Montel and Vielzeuf 1997) to metapelitic gneisses (e.g. Castro et al. 1999; García–Arias et al. 2012). A recent review of major element bulk composition of the felsic melts obtained through those experiments is proposed by Gao et al. (2016).

In particular, great attention was given, through both natural case studies and experimental evidence, to amphibole-dominated partial melting of mafic and intermediate rocks either magmatic (i.e., basalts, andesites, gabbros) or metamorphic (i.e., greenstones, amphibolites and eclogites) (e.g., Barker and Arth 1976; Barker 1979; Beard and Lofgren 1991; Sisson et al. 2005; Ratajeski et al. 2005), since the partial melting of these rocks is considered the key to understand the genesis of modern low-K tonalites and trondhjemites (e.g., Barker and Arth 1976; Beard and Lofgren 1991; Ratajeski et al. 2001, 2005; Sisson et al. 2005; García-Casco et al. 2008), dacites (e.g., Beard and Lofgren 1991), adakites (e.g., Rossetti et al. 2014), plagiogranites sensu lato (e.g. Koepe et al. 2005), and also of Archaean TTG (tonalite-trondhjemite-granodiorite) grey gneisses believed to be crucial in the development of Earth’s early continents (e.g. Barker and Arth 1976; Moyen and Stevens 2006).

This study describes the field occurrence, the petrography, and the geochemistry of a metatexite mafic migmatite suite made up of a hornblende-gabbro protolith plus in situ trondhjemitic leucosomes and melanosomes in the Cheshmeh-Ghasaban gabbros comprising the northern part of the Jurassic Alvand plutonic complex in the Hamedan region (Northern Sanandaj-Sirjan zone, NW-Iran). Petrographic investigations are integrated with mineral chemistry, whole-rock geochemistry and bulk-chemistry mass balance modelling to define the protolith source and the genesis of the in situ leucosome-melanosome suite. Classic and forward modelling thermobarometry are then applied to constrain the pressure-temperature (P-T) conditions of the migmatization and partial melting event. Results are discussed in the light of previous melting experiments developed at comparable P-T conditions for comparable source rocks. This work provides new insights into the amphibole-dehydration processes involved in trondhjemite melts generation from hornblende-bearing mafic protoliths. Results obtained in this study are also discussed in the light of the Middle Jurassic to Early Cretaceous scenario of the Alvand plutonic complex in Northern Sanandaj-Sirjan zone of Iran.

2 Geological background

The study area lies within the Sanandaj-Sirjan Zone (SaSZ) of Iran (Figure 1a), a discrete NW-SE trending geological terrane (Stöcklin 1968) ≈1000 km long and up to 100 km wide, tectonically bounded to the northeast by the Urumieh Dokhtar magmatic arc (UDMA) and to the southwest by the Zagros suture zone (Stöcklin and Nabavi 1973; Berberian and Berberian 1981; Berberian and King 1981; Berberian et al. 1982; Alavi 1994; Mohajel et al. 2016).
Golonka 2004; Ghasemi and Talbot 2006; Davoudian et al. 2008; Chiu et al. 2013; Mohajjel and Fergusson 2014). The SazSZ is interpreted to be the southwestern margin of the Iranian microcontinent (Hassanzadeh et al. 2008; Hosseini et al. 2015; Azizi et al. 2016, 2018a) and is characterized by Jurassic intrusive and volcanic rocks spatially associated with regionally metamorphosed, greenschist to amphibolite facies rocks (e.g., Berberian et al. 1981; Mohajjel et al. 2003; Ghasemi and Talbot 2006; Ahmadi Khalaji et al. 2007; Hassanzadeh and Wernicke 2016; Azizi et al. 2018a, 2018b and references therein). Comparable to the whole Iranian crust, the SazSZ basement is Cadomian (∼550 Ma) in age, as testified by the diffuse outcrops of cadomian rocks (e.g., Stöcklin and Nabavi 1973; Berberian and Berberian 1981; Berberian et al. 1982; Mohajjel et al. 2003; Golonka 2004; Ghasemi and Talbot 2006; Davoudian et al. 2008; Hassanzadeh et al. 2008; Malek-Mahmoudi et al. 2017; Shabanian et al. 2017; Moghadam et al. 2020). Since the work of Eftekharnejad (1981), the SazSZ has been divided into two parts: northern and southern, respectively.

In the northern SaSZ (NSaSZ), where the study area is located (Figure 1a), outcrops of the Jurassic metamorphic complex (Mohajjel and Fergusson 2000; Mohajjel et al. 2003; Agard et al. 2005; Hassanzadeh and Wernicke 2016) made up of Palaeozoic to Early Jurassic mafic to felsic, volcanic and volcanioclastic rocks interbedded with marbles, metasandstones, shales and slates (e.g., Mohajjel et al. 2003; Baharifar et al. 2004; Hassanzadeh and Wernicke 2016) deposited in a marine basin environment over the thinned continental Cadomian crust (e.g., Hassanzadeh and Wernicke 2016); however, medium- to high-grade Barrovian-type metamorphic rocks are locally reported (e.g., Hassanzadeh et al. 2008). During the Middle Jurassic to Early Cretaceous, the Early Jurassic metamorphic complex was affected by intense magmatic activity and plutonism (Berberian et al. 1982; Sepahi and Athari 2006; Arvin et al. 2007; Sepahi 2008; Torkian et al. 2008; Azizi and Jahangiri 2008; Mazhari et al. 2009; Shahbazi et al. 2010, 2014; Azizi et al. 2011, 2015a, 2015b, 2016, 2018a, 2018b; Mahmoudi et al. 2011; Aliani et al. 2012; Azizi and Asahara 2013; Maanijou et al. 2013; Sepahi et al. 2014, 2018, 2019; Yajam et al. 2015; Yang et al. 2018; Zhang et al. 2018a, 2018b; Sheikh Ghashlaghi et al. 2020). These intrusive bodies, such as the Alvand plutonic complex...
(APC) in the Hamedan area (e.g., Shahbazi et al. 2010; Aliani et al. 2012), produced thermo-metamorphic contact aureoles characterized by Buchan-type facies series (e.g., Baharifar et al. 2004; Agard et al. 2005; Sepahi et al. 2009; Shahbazi et al. 2010; Hassanzadeh and Wernicke 2016; Saki et al. 2020). The high-temperature and low-pressure (HT/LP) thermo-metamorphism was also responsible for local migmatization processes (e.g., Sepahi et al. 2009; Saki et al. 2020; Sheikh Gheshlaghi et al. 2020). Both metamorphic and magmatic rocks were then unconformably covered by unmetamorphosed Cretaceous limestones (e.g., Hosseiny 1999; Azizi et al. 2018a and references therein; Yang et al. 2018).

3 Local geology

The Hamedan region in the N-SaSZ (red square in Figure 1a) is characterized by regional metamorphic rocks intruded by the Middle to Upper Jurassic plutons such as the Alvand batholith (e.g. Hassanzadeh et al. 2008; Mahmoudi et al. 2011; Hassanzadeh and Wernicke 2016; Sepahi et al. 2018, 2020). Here we present an abbreviated geological summary of the Hamedan area and the APC (Figure 1b-c), based on the existing recent detailed literature (Yang et al. 2018; Azizi et al. 2018a, 2018b; Shahbazi et al. 2010, 2014; Aliani et al. 2012; Sepahi 2008, 2013; Sepahi et al. 2009, 2014, 2018, 2019, 2020; Saki 2011; Saki et al. 2012, 2020; Hassanzadeh and Wernicke 2016; Baharifar et al. 2004; Mohajjel et al. 2003; Chiu et al. 2013; Mahmoudi et al. 2011; Sheikh Gheshlaghi et al. 2020).

The peak of regional metamorphism in Hamedan area probably occurred in Early to Middle Jurassic time (Sepahi et al. 2018, 2019) after a protracted multi-phase history (e.g. Berberian et al. 1981; Izadi-Kian 2009; Hassanzadeh and Wernicke 2016; Sepahi et al. 2018). The regional metamorphism in Hamedan area predates the Middle to Upper Jurassic plutonism (e.g. Aliani et al. 2012; Chiu et al. 2013; Sepahi et al. 2018; Yang et al. 2018; Azizi et al. 2018a, 2018b). Metapelitic phyllites, slates and schists are the dominant low-grade regional metamorphic rocks (e.g. Mohajjel et al. 2003; Baharifar et al. 2004; Hassanzadeh and Wernicke 2016; Sepahi et al. 2018; Saki et al. 2020) and are generally known in literature as ‘Hamedan Phyllite’ (Mohajjel et al. 2003). They are also commonly associated with interlayered meta-psammitites, marbles, calc-silicates and metabasics (e.g., Sepahi et al. 2018). A westward increase of the metamorphic grade, up to andalusite-garnet-sillimanite schists and parageneses is documented (e.g., Sepahi et al. 2009, 2018, 2020; Mahmoudi et al. 2011). Where the Jurassic plutons were emplaced, the Hamedan phyllites underwent HT/LP thermo-metamorphism generating contact aureoles characterized by Al₂SiO₅-bearing hornfels up to migmatization (e.g., Shahbazi et al. 2010, 2014; Mahmoudi et al. 2011; Hassanzadeh and Wernicke 2016; Sepahi et al. 2018, 2019, 2020; Saki et al. 2020).

With an area of >400 km², the APC (Figure 1b-c) is one of the largest Jurassic batholiths intruding the Hamedan area in the N-SaSZ (e.g., Aliani et al. 2012). It consists of mafic to felsic intrusive rocks (Shahbazi et al. 2010; Mahmoudi et al. 2011; Aliani et al. 2012; Sepahi et al. 2018; Yang et al. 2018; Zhang et al. 2018a, 2018b) with a main body consisting of porphyritic biotite-bearing granite (e.g. Shahbazi et al. 2010; Yang et al. 2018). Leucocratic, aplitic and pegmatitic dike-swarms intruding both the Alvand complex and the surrounding contact aureole are diffusely reported (e.g., Sepahi 2008; Shahbazi et al. 2010; Aliani et al. 2012; Yang et al. 2018; Sepahi et al. 2018, 2020; Sheikh Gheshlaghi et al. 2020).

Mafic intrusions are mainly found in the northern part of the APC, at the Cheshmeh-Ghasaban (CG) locality (Figure 1b-c), however, minor outcrops are also reported in the eastern part of the APC, in the Simin-Khaku area (Eghlimi 1998). These mafic rocks are pegmatoid and microcrystalline gabbros and diorites that form stocks and dikes (e.g. Shahbazi et al. 2010; Aliani et al. 2012). Based on field observations (Eshraghi and Mahmoudi Gharai 2003) and on the primary mineral assemblages (e.g., Sepahi et al. 2009; Shahbazi et al. 2010; Yang et al. 2018; Saki et al. 2020), the CG mafic rocks were further classified as i) hornblende-bearing gabbros and diorites (hereafter Hbl-gabbro), and ii) hornblende-free olivine-bearing gabbros (hereafter Ol-gabbro) with pyroxenite lenses. Based on clinopyroxene-plagioclase thermobarometry models, Sepahi et al. (2013) calculated the emplacement conditions of these gabbroic rocks. They obtained comparable pressure conditions of 5–6 ± 2.5 kbar, with temperatures of 1200–1300°C and 900–1100°C for Ol-gabbro and Hbl-gabbro, respectively. Locally a primary contact between the Ol-gabbro and the Hbl-gabbro is observed, however, a detailed study on the field occurrence and petrography of these gabbro is still missing.

Existing U-Pb zircon ages for the APC (Shahbazi et al. 2010; Mahmoudi et al. 2011; Chiu et al. 2013; Sepahi et al. 2018, 2020; Zhang et al. 2018b) indicate a Jurassic prolonged multiple-stage magmatic history spanning at least from ~167 Ma to ~150 Ma. To note, different U-Pb zircon ages are obtained for the CG gabbros: i) Shahbazi et al. (2010) obtained 166.5 ± 1.8 Ma for the Hbl-gabbro whereas ii) two younger, discordant and different ages of 161.2 ± 1.2 Ma and 135 ± 2 Ma for the Ol-gabbro (Yang et al. 2018). No reasons are given by Yang et al. (2018) for
the two different U-Pb zircon ages obtained, and it is far beyond the scope of this work discussing these ages. However, these ages suggest that Ol-gabbro and Hbl-gabbro are not coeval.

Emplacement of the APC in the hosting Hamedan Phyllite produced a well-developed contact aureole characterized by cordierite-Al$_2$SiO$_5$ (andalusite and sillimanite)-bearing hornfels and cordierite-bearing migmatites (e.g., Sepahi et al. 2019). These metapelitic migmatites and their S-type leucosomes are mainly found in the southern (Tuyserkan) and eastern (Simin-Khaku) parts of the inner aureole (e.g., Sepahi et al. 2019, 2020; Sheikh Ghashlaghi et al. 2020; Saki et al. 2012, 2020 and references therein). U-Pb dating of zircons from the metapelitic mesosomes indicate ≈170 Ma for the migmatization and partial melting event (Sepahi et al. 2019). Recently, Saki et al. (2020) also reported the presence of mafic migmatites with in situ leucosomes occurring in Hbl-gabbro further north in the CG area. They investigated these mafic migmatites through an integration of i) field observations, ii) representative mineral and whole-rock chemistry, iii) qualitative discussion of melting reactions and iv) conventional and forward thermobarometric models. Authors qualitatively suggested the origin of the leucosome melts as the product of the partial melting of a hornblende + plagioclase + biotite assemblage in the Hbl-gabbro source. On the basis of thermobarometric modelling applied to the mafic mesosome (i.e., thermally metamorphosed gabbro), they proposed a temperature of ≈800°C and pressure of ≈3.7 kbar for the migmatization event. No age for these mafic migmatites exists, however, Saki et al. (2020) suggest their formation in the Early Cretaceous was coeval with the emplacement of the presumably younger Ol-gabbro.

4 Field observation and petrography

The CG gabbros, with an area of ≈25 Km$^2$ (Figure 1b-c), represent the most important exposure of mafic rocks in the APC. The two main magmatic types are the Hbl-gabbro and the Ol-gabbro (Eshraghi and Mahmoudi Gharai 2003; Sepahi et al. 2009; Shahbazi et al. 2010; Yang et al. 2018; Saki et al. 2020). A primary contact between these two gabbroic facies is rarely and not easily visible (Figure 2a), and often it appears as a gradually transition from the lower Ol-gabbro to the upper Hbl-gabbro. The Ol-gabbro sampled close to the contact shows coarse-grained texture and consists, in descending order of abundance, of plagioclase, olivine, clinopyroxene, orthopyroxene, biotite and opaque minerals. Olivine is commonly characterized by secondary coronas of secondary amphibole and serpentine.

The Hbl-gabbro (Figure 2b), when close to the Ol-gabbro, is locally affected by thermal metamorphism and migmatization (Figure 2a). The mafic migmatite is made up of a mesosome (i.e., a metagabbro) showing the presence of i) in-source leucosomes (sensu Sawyer 2008) in centimetre- to metre-scale veins and dykes (Figure 2c) and ii) in situ leucosomes (sensu Sawyer 2008) in millimetre- to metric-scale patches (Figure 2d), net-structure and veinlets (Figure 2e), and lenses (Figure 2f). Mafic selvedges (Figure 2c) dominated by hornblende are observed at rims of major leucosome veins and lenses. Locally, blocks of protolith and/or melanosome within the leucosomes are observed (Figure 2a,c,e,f), describing a typical schollen (or raft) migmatite fabric. With respect to the studied outcrops, the relative volume of leucosome is low (<10 vol%), thus defining this migmatite as a metatexite migmatite (Sawyer 2008). No leucosomes are found instead in the Ol-gabbro. Thermal metamorphism of gabbros was reported by Sepahi et al. (2009), while the presence of ‘agmatites’ in gabbros was already documented in the Tuyserkan geological map (Eshraghi and Mahmoudi Gharai 2003), however, only Saki et al. (2020) produced a detailed petrographic investigation of the meta-gabbros and the associated in situ leucosomes.

The Hbl-gabbro shows holocrystalline to hypidiomorphic fine- to coarse-grained ophitic-like isotropic texture (Figure 3a-b) and it is composed of euhedral to subhedral plagioclase (35–45 vol%) and hornblende (30–40 vol%), with subhedral to anhedral clinopyroxene (up to 25 vol%), biotite (∼10 vol%), quartz (∼5 vol%) and minor orthopyroxene (<5 vol%), with ilmenite,apatite, zircon and rare titanite as main accessory minerals. No olivine has been observed in the studied Hbl-gabbro samples. The thermally metamorphosed Hbl-gabbro (Figure 3c-d), a metagabbro corresponding to the mesosome of Saki et al. (2020), shows a modal mineralogy similar to the protolith; however, a gradual increase of hornblende and biotite in sheaf texture (Figure 3d) is observed when approaching leucosomes (Figure 3e-i). Furthermore, mesosomes and mesosome relics in leucosomes show i) protolith plagioclase (Pl$_3$) with subrounded morphology (Figure 3c,g,h,i), ii) biotite locally replaced by neo-formed quartz and orthopyroxene, iii) hornblende characterized by overgrowth of clinopyroxene and orthopyroxene (Figure 3f-h) and iv) protolith assemblage partially melted (Figure 3c-i). The analysed leucosome samples show a general holocrystalline, allotriomorphic to hypidiomorphic granular hololeucocratic texture (Fig. c-i), with the primary assemblage dominated by anhedral quartz (up to 40 vol%) and anhedral to euhedral plagioclase (up to 65 vol%) showing polycrystalline albite twinning, combined albite-Carlsbad
twinning and combined albite-pericline twinning. Rare granophyric textures made up of quartz and K-feldspar are observed in veins and veinlets. Rare xenomorphic clinopyroxene and orthopyroxene specimen characterized by abundant quartz intergrowth are observed within the leucosome. The melanosome (Figure 3e) is almost entirely constituted by aggregates of generally anhedral plagioclase (40–50 vol%), hornblende (10–20 vol%), clinopyroxene (30–40 vol%) and biotite (<5 vol%). All phases are characterized by xenomorphic and lobate morphology. No quartz nor orthopyroxene are observed. These melanosomes are interpreted as a melt residuum (sensu Taylor and Stevens, 2010; Rossetti et al., 2020).

5 Analytical methods

5.1 Mineral chemistry

The major-element composition of plagioclase, hornblende, clinopyroxene, orthopyroxene and biotite from both Hbl-gabbro protolith and felsic leucosomes were analysed at the Wuhan Technology University of China (WTU), Wuhan, using a JXA-8230 electron microprobe (EMP). The accelerating voltage was 20 kV. The beam current was 30 nA and the beam diameter varied between 3 and 10 µm. For the major minerals, calculated 1 (%) precisions are (i) better than 1.5% for Si; (ii) better than 2% for Al; (iii) 1% to 5% for Ca, Mg, Fe, Mn, and Ti, applying the above-mentioned applied conditions. For Na and K, calculated 1 (%) precisions are below 5% for analyses of feldspars and clinopyroxene. Validation of mineral chemistry results was also achieved following the method presented in the Appendix of Lucci et al. (2020). Back-Scattered Electron (BSE) imaging was obtained at Institute of Geosciences, Universität Potsdam (Germany) by using a JEOL JSM-6150 scanning electron microprobe (SEM) combined with an Oxford Instrument INCAx-act detector (Energy Dispersive X-ray spectrometer, EDX), with operating conditions of 20 kV, 10 to 50 nA. Mineral structural formulae of plagioclase and biotite were calculated through the software CalcMin_32 (Brandelik 2009). Pyroxene chemistry and
structural formulae calculation was performed following the workflow from Lucci et al. (2020). Pyroxenes were classified following Morimoto (1988, 1989). Hornblende chemistry and structural formula were calculated following Miri et al. (2016) and Lucci et al. (2018), using the ACES2013 MS Excel spreadsheet (Locock 2014). Amphiboles are classified following Leake et al. (2004) and Hawthorne et al. (2012). In the following discussion, we use the mineral abbreviations recommended by Whitney and Evans (2010).

5.2 Whole-rock geochemistry

Seventeen samples from the CG Hbl-gabbro and from the in situ leucosome-melanosome suite were selected for whole-rock analysis. The samples were crushed using a steel jaw crusher and pulverized in a tungsten carbide mill. Samples were analysed for major and trace elements at Activation Laboratories (Ontario, Canada) by ICP-OES and ICP-MS (code 4Lithoresearch) with a quality control (QC) based on the repeated analyses of 14 certified reference materials. For major elements, the uncertainty (1) is estimated better than 2% for values higher than 5 wt %, and better than 5% in the range 0.1–5 wt %. For trace elements and REEs, the precision is 5% in the range 1–100 ppm and 10% in the range 0.1–1 ppm. Compositions of the samples are presented in Supplementary Table 2. The dataset is integrated with analyses of four leucosome and three mesosomes collected from the same outcrops and recently presented by Saki et al. (2020). Whole-rock compositions obtained by Saki et al. (2020) are also reported in Supplementary Table 2 for comparison.
6 Mineral chemistry

The aim of the mineral chemistry is presenting the general compositions of main phases used for major element mass balance modelling. Representative mineral compositions as obtained from electron microprobe analyses and mineral formulae are presented in Supplementary Tables 1a, 1b, 1c, 1d (for plagioclase, hornblende, pyroxenes and biotite, respectively).

6.1 Hbl-gabbro

In Hbl-gabbro, feldspar is represented by homogeneous labradoritic plagioclase (An_{57-60}Ab_{40-42}; mean An_{58}Ab_{41}) with minor (<1 mol%) orthoclase (Or) (Figure 4a). Iron content is <0.18 wt% Fe_2O_3 corresponding to Fe\(^{3+}\) < 0.006 atom per formula unit (apfu). Igneous amphibole (i.e. hornblende) presents equilibrium textures with the labradorite plagioclase and shows SiO_2 in the range 46.52–47.46 wt% (Si 6.73–6.91 apfu), Al_2O_3 ranging 7.45–8.88 wt% (Al_{tot} 1.28–1.51 apfu), FeO_{tot} 13.39–14.46 wt% (Fe_{tot} 1.64–1.75 apfu), CaO 11.54–11.81 wt% (Ca 1.78–1.85 apfu) and TiO_2 0.99–1.48 wt% (Ti 0.11–0.16 apfu), corresponding to magnesiohornblende (Leake et al. 2004; Hawthorne et al. 2012) compositions (Figure 4b). Biotite is generally associated with hornblende; however, it is the less abundant mafic phase. Biotite is characterized by TiO_2 content in the range 1.26–1.45 wt% (Ti 0.07–0.08 apfu), Fe\# (= atomic [Fe\(^{2+}\)/(Fe\(^{2+}\)+Mg)]) in the range 0.28–0.36 and Al\(^{IV}\) of 1.27–1.35 apfu and is classified as eastonite (Figure 4c; Fleet et al. 2003). According to the 10\(^{\times}\)TiO_2–(FeO+MnO)–MgO ternary system (Figure 4d) proposed by Nachit et al. (2005), the dark mica falls in the field of re-equilibrated magmatic biotites (field ‘B’ in Figure 4d; Nachit et al. 2005; Sirqueira et al. 2018). Clinopyroxene is the second most abundant mafic primary phase after hornblende and generally occurs as single crystal, however, intergrowth with hornblende are also observed. Clinopyroxene shows high Al_2O_3 (5.01–5.06 wt%; Al_{tot} 0.22 apfu) and TiO_2 (1.11–1.19 wt%; Ti 0.03 apfu) contents, with Mg\# (= atomic [Mg/(Mg+Fe\(^{2++}\)]) of 0.81–0.82, Ca of 0.84–0.85 apfu, Q + J of 1.91 and J/(J + Q) of 0.03–0.04. Analysed clinopyroxene belong to Ca-Mg-Fe pyroxenes or Quad pyroxenes (Figure 4e; Morimoto 1988, Morimoto 1989) and can be classified as a diopside-rich augite (mean Wo_{45}En_{42}Fs_{10}) (Figure 4f). Orthopyroxene is very rare, and where present it shows hypersthene-like composition (mean Wo_{41}En_{43}Fs_{35}) (Figure 4e-f) with Mg\# of 0.62, Al_2O_3 up to 0.50 wt% (Al_{tot} up to 0.02 apfu) and CaO < 1 wt% (Ca < 0.04 apfu), corresponding to Q + J 1.89–1.95 and J/(Q + J) <0.01.

6.2 Leucosome

Leucosomes consist primarily of quartz and oligoclase (An_{26-30}Ab_{58-72}Or_{02}) (Figure 4a). The only mafic minerals observed are rare completely destabilized hornblende and pyroxenes with xenomorphic habit and compositions comparable to those from Hbl-gabbro. Clinopyroxenes have high Al_2O_3 (4.56–4.72 wt%; Al_{tot} 0.20 apfu) and TiO_2 (1.11–1.18 wt%; Ti 0.03 apfu) contents, with Mg\# of 0.82, Ca of 0.86–0.87 apfu, Q + J of 1.91 and J/(J + Q) of 0.03. It can be classified as diopside-rich augite (mean Wo_{46}En_{44}Fs_{10}) (Figure 4e-f). Orthopyroxene is very rare and shows again hypersthene-like composition (mean Wo_{42}En_{41}Fs_{37}) (Figure 4e-f) with Mg\# of 0.62–0.64, Al_2O_3 up to 0.40 wt% (Al_{tot} up to 0.02 apfu) and CaO < 0.6 wt% (Ca < 0.02 apfu) corresponding to Q + J 1.94–1.96 and J/(Q + J) <0.01. Similar compositions (Figure 4) for plagioclase (An_{23-26}Ab_{71-73}Or_{02-03}) clinopyroxene (mean Wo_{46}En_{44}Fs_{10}) and orthopyroxene (mean Wo_{42}En_{43}Fs_{36}) in leucosomes were obtained by Saki et al. (2020).

6.3 Meta-gabbros: mesosome of Saki et al. (2020)

In this section, we report from the work of Saki et al. (2020) a brief summary of the mineral chemistry of mesosome layers that authors recently identified in the same studied outcrop. With the term of mesosome, authors indicate layers of meta-gabbro characterized by a higher content of hornblende and biotite in micro- to meso-crystalline sheaf texture and the presence of well-visible diffuse and irregular patches of whitish leucosome. The mesosome mineralogy (i.e. Pl + Hbl + Cpx + Bt) of the mesosome paragenesis show chemistry similar to those from the Hbl-gabbro protolith. Clinopyroxenes have high Al_2O_3 (4.56–4.72 wt%; Al_{tot} 0.20 apfu) and TiO_2 (1.11–1.18 wt%; Ti 0.03 apfu) contents, with Mg\# of 0.82, Ca of 0.86–0.87 apfu, Q + J of 1.91 and J/(J + Q) of 0.03. It can be classified as diopside-rich augite (mean Wo_{46}En_{44}Fs_{10}) (Figure 4e-f). Orthopyroxene is very rare and shows again hypersthene-like composition (mean Wo_{42}En_{41}Fs_{37}) (Figure 4e-f) with Mg\# of 0.62–0.64, Al_2O_3 up to 0.40 wt% (Al_{tot} up to 0.02 apfu) and CaO < 0.6 wt% (Ca < 0.02 apfu) corresponding to Q + J 1.94–1.96 and J/(Q + J) <0.01. Similar compositions (Figure 4) for plagioclase (An_{23-26}Ab_{71-73}Or_{02-03}) clinopyroxene (mean Wo_{46}En_{44}Fs_{10}) and orthopyroxene (mean Wo_{42}En_{43}Fs_{36}) in leucosomes were obtained by Saki et al. (2020).

7 Whole-rock geochemistry

The chemical compositions for Hbl-gabbro (n = 6), in situ leucosome (n = 5) and associated melanosome (n = 6)
were investigated for major, trace and REE element chemistry. As previously declared, the dataset is integrated with leucosomes (n = 4) and mesosomes (n = 3) from Saki et al. (2020). Data are presented in Supplementary Table 2 and graphically shown in diagrams of Figures 5 and Figures 6 where they are also compared to the existing literature on APC (Sepahi 2008; Shahbazi et al. 2010; Aliani et al. 2012; Yang et al. 2018; Sepahi et al. 2018, 2020; Sheikh Gheshlaghi et al. 2020).

7.1 Hbl-gabbro

All Hbl-gabbro samples (n = 6) show LOI < 1.2 wt%, suggesting a negligible degree of alteration. The Hbl-gabbro shows SiO₂ in the range 48.88–50.82 wt% and Al₂O₃ in the range 14.05–17.10 wt%, with variable Fe₂O₃ * (7.33–14.33 wt%) and CaO (8.4–12.85 wt%) contents. The Hbl-gabbro is characterized by high TiO₂ (1.043–4.340 wt%) and low MgO (5.16–7.71 wt%) contents with

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**Figure 4.** Mineral Chemistry. a) An-Ab-Or ternary diagram showing the composition of feldspars in the Hbl-gabbro protolith (red diamonds) and in leucosomes (cyan triangles). b) Amphibole classification diagram for Ca-amphiboles (Leake et al. 2004; Hawthorne et al. 2012). c) Fe/(Fe+Mg) vs Al[IV] classification diagram for trioctahedral dark-micas (after Fleet et al. 2003). d) TiO₂-(FeO+MnO)-MgO discrimination ternary diagram for biotite (Nachit et al. 2005) indicating domain of primary magmatic biotite (field A), domain of the re-equilibrated biotite (field B) and domain of the neoformed biotite (field C). e-f) Clinopyroxene and orthopyroxene compositions plotted in Q-J and Di-Hd-En-Fs diagrams (after Morimoto 1988, 1989). In the Q-J diagram the term ‘Quad’ stands for Di-Hd-En-Fs quadrilateral and indicates the compositional field of Ca-Mg-Fe pyroxenes. EMPA data for leucosomes and mesosomes from literature (Saki et al. 2020) are also reported.
Mg# of 0.45–0.68. The low MgO content excludes the Hbl-gabbro is a primitive melt (e.g. Azizi et al. 2018a and references therein). On the total alkali versus silica (TAS) diagram (after Middlemost, E.A.K. 1994; Le Maitre et al. 2002) the studied samples fall across the ‘Gabbro’ and ‘Monzogabbro’ fields (Figure 5a) and are characterized by K$_2$O content in the range 0.93–2.57 wt% (Figure 5b). These Hbl-gabbros have 430–510 ppm Sr, 15–26 ppm Y, 87–216 ppm Zr, ≈65–130 ppm total LREE, and ≈7–12 ppm total HREE. In selected Harker diagrams for both major, trace and REE elements (Figure 6a-i), the Hbl-gabbro samples show a narrow distribution with no appreciable correlation with the SiO$_2$. The sample GH-Gb1 is highlighted with a different colour in diagrams (red diamond) because of the higher Al$_2$O$_3$, CaO, MgO, Mg# and Sr and lower Na$_2$O, K$_2$O, Fe$_2$O$_3$, Rb, Y and REE contents, with respect to other Hbl-gabbro samples (pink diamonds). The analysed Hbl-gabbros are somewhat similar to Hbl-bearing gabbro and diorites (n = 4) from the APC investigated by Shahbazi et al. (2010) for geochemistry and geochronology (black diamonds in diagrams of Figure 5a-b, Figure 6a-i), whereas no such similarity with Hbl-free Ol-gabbro analysed (n = 4) by Yang et al. (2018) is observed (grey diamonds in Figure 5a-b, Figure 6a-i). In chondrite (Ch)-normalized REE diagrams (Figure 7a; Sun and McDonough 1989), the Hbl-gabbros show fractionated patterns with slightly enriched profile characterized by La$_N$/Yb$_N$ and Dy$_N$/Yb$_N$ ranging 6.40–11.35 and 1.26–1.65, respectively. The Eu anomaly ([Eu/Eu*] = [Eu$_N$/(Sm$_N$×Gd$_N$)${}^{1/2}$]) is slightly positive (1.08) for sample GH-Gb1, whereas it is slightly negative (0.82–0.96) for the other Hbl-gabbroic rocks. In primitive mantle (PM)-normalized multi-elements diagram (Figure 7b; Sun and McDonough 1989), the GH-Gb1 sample shows a homogeneous profile characterized by slight Th, Nb, Zr, and Ti negative anomalies. Profiles of other Hbl-gabbro samples show higher variability with both negative and positive anomalies of Nb, Zr, Hf and Ti. The GH-Gb2-6 Hbl-gabbros, with respect to GH-Gb1 sample,
show: i) enriched REE- and trace-profiles (Figure 7a,b), ii) a progressive depletion of $\text{Al}_2\text{O}_3$, $\text{CaO}$, $\text{MgO}$, and $\text{Sr}$; iii) a progressive enrichment of $\text{Na}_2\text{O}+\text{K}_2\text{O}$, $\text{Fe}_2\text{O}_3$, trace and REE elements; and iv) a progressive development of a negative Eu anomaly. These chemical characteristics are compatible with a main plagioclase fractionation/extraction process developed during magma differentiation and emplacement along the same line of descent in a gabbroic cogenetic suite (e.g. Dokuz et al. 2006) starting from the GH-Gb1 parental composition.

### 7.2 Leucosome-melanosome suite

The leucosomes have high-$\text{SiO}_2$ (76.45–80.08 wt%) and very low $\text{MgO}$ (0.11–0.55 wt%), $\text{Fe}_2\text{O}_3$ (0.21–1.14 wt%), $\text{TiO}_2$ (0.03–0.14 wt%) and $\text{CaO}$ (1.27–2.89 wt%). They show $\text{Al}_2\text{O}_3$ contents of 12.38–13.42 wt% and are strongly enriched in $\text{Na}_2\text{O}$ (3.96–5.01 wt%) with respect to $\text{K}_2\text{O}$ (0.24–1.26 wt%) (Figure 5b). In the TAS diagram (Figure 5a) they fall in the ‘Granite’ field. Leucosomes are peraluminous with A/CNK (molar $\text{Al}_2\text{O}_3/\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$)
K$_2$O)) values ranging 1.01–1.22 (mean value 1.09). The leucosomes have 33–267 ppm Sr, 7–10 ppm Y, 34–76 ppm Zr, 43–53 ppm total LREE and ≈3-4 ppm total HREE. Based on the estimated modal abundances of mineral phases constituting the primary magmatic assemblage, the leucosomes in the Q-A-P diagram (Q: quartz, A: alkali-feldspar, P: plagioclase; Streckeisen 1976 in Le Maitre et al. 2002) fall in the tonalite/trondhjemite field (Figure 5c). Considering the leucocratic character (mafic phases < 10%) of these leucosomes, they can be further classified as trondhjemite (Le Maitre et al. 2002). The trondhjemite character is also confirmed in the Ab-An-Or diagram (Figure 5d; Barker 1979) built on normative components obtained through the CIPW+Hb+Bt norm of Hutchison (1975). Considering the Al$_2$O$_3$ (<15 wt%) and K$_2$O (<1.3 wt%) contents, these leucosome can be further defined as low-Al$_2$O$_3$ trondhjemites (Barker and Arth 1976). To note, in both ternary plots presented in Figure 5c and Figure 5d, the studied leucosomes fall in the field (cyan shaded area) of typical trondhjemites from west-central Trondheim Region (Nilsen et al. 2003).

Melanosomes have high MgO (11.87–16.77 wt%) and low SiO$_2$ (46.25–48.19 wt%), low Al$_2$O$_3$ (10.95–12.89 wt %), low TiO$_2$ (1.28–1.36 wt%), and variable Fe$_2$O$_3$* (9.97–11.80 wt%) and CaO (8.56–13.27 wt%) and. Na$_2$O (1.32–2.10 wt%) is always greater than K$_2$O (0.68–1.08 wt%). In the TAS diagram (Figure 5a) they fall in the ‘Gabbro’ field. The melanosomes have 231–368 ppm Sr, 14–17 ppm Y, 87–111 ppm Zr, 70–89 ppm total LREE, and ≈7-8 ppm total HREE.

The abundances of both large-ion lithophile elements (LILEs, such as Ba, Sr) and high-field strength...
elements (HFSEs, such as Zr, Y, Nb, REE) are negatively correlated with SiO$_2$ and it is generally higher in melanosome layers (Figure 6a-i). Both melanosomes and leucosomes are generally depleted with respect to the Hbl-gabbro (Figure 6a-i). The chemistry of the melanosome partly converges towards both the Hbl-gabbro (Shahbazi et al. 2010) and the Olgabbro (Yang et al. 2018) (Fig. 5, Figure 6), whereas the trondhjemite leucosomes are clearly discriminated (Figure 5a-b, Figure 6a-i) from APC monzonite, syenite, granodiorite and granite bodies (n = 60 analyses; grey circles in Figs. 5 and Figs. 6) where no high-SiO$_2$ (>76 wt%) granitoids have been reported. In contrast, aplites, leucogranite dikes, pegmatites and sapphire-bearing pegmatites in the APC (Sepahi 2008; Shahbazi et al. 2010; Aliani et al. 2012; Yang et al. 2018; Sepahi et al. 2018, 2020; Sheikh Gheslaghi et al. 2020), shows SiO$_2$ in the range ≈50-80 wt%, with a general K-rich signature (shown in Figs. 5 and Figs. 6 with the existing compositional data [n = 77] as white triangle). Only the Smk15 (Sepahi 2008) and AR15 (Sheikh Gheslaghi et al. 2020) pegmatites (Figure 5b) from the Simin-Khaku area (Figure 1b) with SiO$_2$ >76 wt%, Al$_2$O$_3$ < 15 wt%, Na$_2$O > 4 wt% and K$_2$O < 1 wt%, are comparable to the Cheshmeh-Ghasaban in situ trondhjemite leucosomes. Notably, in the Simin-Khaku area where the two pegmatites were sampled, limited and scattered occurrences of Hbl-bearing gabbro-diorite bodies are reported by Eghlimi (1998); however, only few of these outcrops were mapped suggesting therefore the presence of a higher volume of gabbroic rocks in the area (A.A. Sepahi, personal communication, November 2020).

When normalized to chondrite (Figure 7a; Sun and McDonough 1989), both leucosome and melanosome samples are enriched in LREE with La$_{N/Yb_N}$ in the range 9.19–13.53 and 7.38–9.93, respectively. Melanosome ratios (Dy$_{N/Yb_N}$ 1.40–1.55) are comparable to that of GH-Gb1 protolith (red line in Figure 7a), whereas leucosomes (cyan and blu lines in Fig. 7a) show depleted concave upward profiles with HREE patterns characterized by nearly flat profiles (Dy$_{N/Yb_N}$ 0.80–1.15). The Eu anomaly is slightly negative to positive (Eu/Eu* 0.97–1.07) in melanosome and negative to positive (Eu/Eu* 0.67–1.18) in leucosome suggesting a variable role of plagioclase in their genesis (e.g. Hu et al. 2016).

On a PM-normalized diagram (Figure 7b), melanosomes are enriched in Th, Ta, Nb, Ti and LREE with respect to that of GH-Gb1 Hbl-gabbro, whereas leucosomes show complex profiles characterized by marked positive anomalies of Th and Hf, associated with negative anomalies of Nb and Ti. This behaviour, together with trace and REE element contents, is compatible with melt segregation/evolution dominated by amphibole and plagioclase (e.g., Lucci et al. 2016; Moghadam et al. 2016; Hu et al. 2016).

7.3 Meta-gabbros: mesosome of Saki et al. (2020)

As with the ‘Mineral Chemistry’ section, we present here a short summary of the whole-rock composition of mesosome layers investigated by Saki et al. (2020). Mesosomes are characterized by SiO$_2$ in the range 49.43–50.70 wt%, high Al$_2$O$_3$ (15.69–16.25 wt%), high CaO (9.05–11.69 wt%), Fe$_2$O$_3$ (9.0–10.8 wt%), MgO (6.5–7.1 wt%, corresponding to Mg# 54–59) and TiO$_2$ (1.32–2.04 wt%). Comparable to Hbl-gabbro protolith, melanosome and leucosomes, the mesosome layers show K$_2$O concentrations (1.12–2.44 wt%) always less than Na$_2$O (2.08–2.70 wt%). In the TAS diagram (Figure 5a) mesosome fall in the ‘Gabbro’ field. Mesosome layers show high Sr (410–536 ppm), high Y (12–23 ppm), high Zr (112–224 ppm) and LREE and HREE ranging 65–134 ppm and 7–11 ppm, respectively. In K$_2$O vs. SiO$_2$ diagram (Figure 5b) and Harker diagrams (Figure 6a-i), mesosome layers overlap the Hbl-gabbros investigated in this study. In Ch-normalized REE diagram (Figure 7a; Sun and McDonough 1989), mesosomes show enriched profiles characterized by La$_{N/Yb_N}$ and Dy$_{N/Yb_N}$ ranging 6.65–9.59 and 1.13–1.47, respectively. The Eu anomaly is negative to positive (Eu/Eu* 0.83–1.27). In PM-normalized multi-elements diagram (Figure 7b; Sun and McDonough 1989), the mesosome profiles are characterized by both negative and positive anomalies of Ta, Nb, Ti and Y. In both Ch- and PM-normalized diagrams, mesosome layers profiles are not homogeneous and mainly resemble both Hbl-gabbro protolith and melanosome patterns.

8 Discussion

There are three main questions concerning CG mafic complex and associated leucosome-melanosome suite constituting the metatexite migmatite. First, do CG Hbl-gabbro protolith hosting the metatexite migmatite represents a cogenetic suite from magma differentiation via fractional crystallization processes? Second, do the leucosome-melanosome suite in the metatexite is the product of Hbl-gabbro partial melting and in case, which is the degree of the partial melting event? Finally, what were the thermobaric conditions for the trondhjemite leucosome genesis? These questions are addressed below.
8.1 Workflow for modelling the magmatic processes

In this section, we present petrogenetic models to test the fractional crystallization and the partial melting hypotheses for the cogenetic relationship between all Hbl-gabbros and for the origin of the leucosome-melanosome suite from the Hbl-gabbros.

8.1.1 The origin of hbl-gabbro

To model the magmatic differentiation via fractional crystallization, based on the primary assemblages, the mineral and the whole-rock chemistry, the sample GH-Gb1 was selected to represent the possible modelled parental magma for all the studied Hbl-gabbro samples. The sample GH-Gb1, with respect to the other Hbl-gabbro samples, was chosen since the higher Al₂O₃ (17.1 wt%), CaO (12.85 wt%), MgO (7.71 wt%) and Mg# (73.8) associated with the lower Na₂O+K₂O (3.14 wt%). Furthermore, the GH-Gb1 gabbro shows higher Sr (510 ppm) and a positive Eu anomaly (Eu/Eu* 1.08). These chemical characteristics are compatible with a liquid extracted from a primary gabbroic source and not yet affected by fractionation of Ca-Al-bearing phases such as plagioclase, hornblende or augitic clinopyroxene (e.g. Dokuz et al. 2006; White et al. 2009; Lucci et al. 2016; Moghadam et al. 2020).

8.1.2 The origin of mafic migmatites

To model the partial melting process to generate the leucosome-melanosome suite, the sample GH-Gb1 was again selected to represent the possible mafic protolith. The sample GH-Gb1 is chosen since i) it was collected from layers close to the contact with Ol-gabbros, where locally the relative amount of in situ leucosomes reach the higher values (up to 10 vol%), ii) it shows REE- and trace-patterns overlapping those of melanosomes from layers close to the contact with Ol-gabbros, where iii) it presents LREE (La, Ce) and HREE (Yb, Lu) contents comparable to those of leucosomes. Considering also the high CaO and Sr and the low K2O contents together with the positive Eu anomaly, the GH-Gb1 sample could indicate the Hbl-bearing gabbro protolith source underwent partial melting to generate low-Al₂O₃ trondhjemite leucosomes (e.g., Barker and Arth 1976).

8.1.3 Major-element mass balance models

The Major-element mass balance model of Bryan et al. (1969) is used here to test and determine mineral phases involved in the Rayleigh Fractional Crystallization (FC) and Equilibrium Batch Melting (EBM) hypotheses (e.g., White et al. 2009; Lucci et al. 2016, 2020; Moghadam et al. 2016, 2020). In major-element mass balance model, it is assumed that for FC-modelling the Daughter Liquid = Parental melt – Minerals (fractionating assemblage), whereas for EBM the Leucosome = Protolith source – Minerals (residual assemblage) and the Melanosome = Protolith source – Minerals (melted assemblage). The composition of the parental melt (for FC) or of the protolith source (for EBM) is assumed as matrix b; if all elemental equations are solved for b, then i) for FC b= Daughter Liquid + Minerals, ii) for EBM b= Leucosome + Minerals and b= Melanosome + Minerals. When the compositions of Daughter Liquid, Leucosome, Melanosome and Minerals involved are known (matrix A), their proportion (matrix c) estimates can be obtained through the least square approximation. The similarity of the residuals (matrix b’ = c X A) to the parental melt or to the protolith (matrix b) is then quantified with the sum of the square of the residuals (Σᵣ²) as the following:

\[ \sum r^2 = \sum_{i=1}^{n} (b'_i - b_i)^2 \]  

FC and EBM model results are considered acceptable when Σᵣ² < 1.0. In all calculations, the proportion of daughter liquid, leucosome and melanosome is expressed with the variable F in matrix c. The FC-hypothesis was tested for all the studied Hbl-gabbro samples, whereas the EBM-hypothesis was tested for all leucosomes and melanosomes. Results are presented in Supplementary Table 3.

8.1.4 Trace and REE elements FC models

To model the trace- and REE-fractional crystallization we use the equation of Lord Rayleigh (1986):

\[ C_L = C_0 \times F^{(D-1)} \]  

where \( C_L \) is the calculated element concentration in the daughter melt for a specific degree of fractionation, \( C_0 \) is the element concentration in the parental melt, D is the calculated bulk mineral/melt partition coefficient for the chosen element and F is the daughter melt fraction, with F = 1-FC and FC is the degree of fractionation with 0< FC<1 (after Rossetti et al. 2014; Lucci et al. 2016; Moghadam et al. 2016, 2020). Considering the primary assemblage of the selected parental rock dominated by plagioclase, hornblende, clinopyroxene and biotite, following the existing literature (e.g. Rossetti et al. 2014; Lucci et al. 2016; Moghadam et al. 2016, 2020), we calculated the FC-models for Y vs Sr/Y and Yb vs Dy/Yb systems. The element bulk partition coefficients (D_{Sr}, D_{Y}, D_{Yb} and D_{Dy}) were calculated for mineral phase weight percentages as obtained from the major-element mass balance FC model, after normalization to phase density.
(Augite-Diopside Clinopyroxene 3.20 g/cm$^3$ in Deer et al. 1997a; Tschermakite Hornblende 3.44 g/cm$^3$ in, 1997b; Labradorite Plagioclase 2.68 g/cm$^3$ in, 2001; Biotite 3.1 g/cm$^3$ in, 2013). Mineral/melt partition coefficients for mafic melts were selected from the existing literature (Schnetzler and Philpotts 1970; Nagasawa and Schnetzler 1971; Ronov and Yaroshevskiy 1976; Matsui et al. 1977; Nicholls and Harris 1980; Vilmant et al. 1981; Irving and Frey 1984; Gallahan and Nielsen 1992; Sisson 1994; Hack et al. 1994; Latourrette et al. 1995; Vannucci et al. 1998; Bindeman et al. 1998; Vannucci et al. 1999; Bindeman et al. 2000). Results are presented in Supplementary Table 4a.

8.1.5 Trace and REE elements EBM models
The partial melting is the gradually and progressive melting of the protolith and it is quantified through the degree (or percentage) of melting. The partial melting processes can develop as open- or closed-system, depending on liquid extraction from or permanence in the residuum, respectively (e.g., Shaw 1970; Ersoy 2013). Field and petrographic observations (Figure 2, Figure 3) indicate that studied trondhjemites are in situ leucosomes remained in equilibrium within the residuum until end of melting. It is compatible with a batch melting process (Shaw 1970; Zou 1998; Ersoy 2013). Considering the generally absence/paucity of protolith remnants or peritectic phases within or close to the leucosomes (Figure 3) and the results obtained from the major-elements EBM models (see section 8.3), it is here suggested that the proportion of mineral phases underwent melting was that of the primary assemblage of the protolith, as expected for modal batch melting event (e.g., Shaw 1970; Ersoy 2013). To model the trace- and REE-modal equilibrium batch melting we choose the equations of Shaw (1970) for batch melts and residues, respectively:

$$C_L = \frac{C_0}{D_0 + F(1 - D_0)}$$  \hspace{1cm} (3)

$$C_S = \frac{C_0 D_0}{D_0 + F(1 - D_0)}$$  \hspace{1cm} (4)

where $C_L$ and $C_S$ are the calculated element concentrations in the produced batch melt and residue, respectively, for a specific degree of partial melting, $C_0$ is the element concentration in the protolith source, $D_0$ is the calculated bulk mineral/melt partition coefficient for the chosen element and $F$ is the fraction of melt produced during the anatexis. Comparable to the FC-model, since the same starting material (i.e., sample GH-Gb1) also the EBM-models were calculated for $Y$ vs $Sr/Y$ and $Yb$ vs $Dy/Yb$ systems. The element bulk partition coefficients ($D_{Sr}$, $D_{Y}$, $D_{Yb}$ and $D_{Dy}$) were calculated using the same procedure proposed for FC-modelling in section 8.1.3. Following the approach proposed by White (2003), different bulk partition coefficients were calculated for felsic leucosome and mafic melanosome. Mineral/melt partition coefficients were selected from the existing literature (Philpotts and Schnetzler 1970; Schnetzler and Philpotts 1970; Nagasawa and Schnetzler 1971; Ersoy et al. 1973; Ronov and Yaroshevskiy 1976; Matsui et al. 1977; Luhr and Carmichael 1980; Sisson 1991, 1994; Green et al. 1993; Ersoy and Griffin 1994; Klein et al. 1997; Bindeman et al. 1998; Vannucci et al. 1999). Results for batch melts and residues are presented in Supplementary Table 4b and 4c, respectively.

8.2 FC processes and generation of the hbl-gabbro suite
The FC hypothesis has been tested for all analysed Hbl-gabbros, assuming GH-Gb1 sample as parental melt and all the other Hbl-gabbro samples ($n = 5$) as daughter melts. Considering the primary mineralogy of the suspected parental melt, the following fractionating assemblage was investigated: An-rich plagioclase + Ca-hornblende + Ca-Al-clinopyroxene + biotite. The major-element mass balance FC model was calculated through Eq. (1) in the system $SiO_2$-TiO$_2$-Al$_2$O$_3$-$FeO^*$-MnO-MgO-CaO-Na$_2$O-K$_2$O. Results are presented in Supplementary Table 3a, where a reliable solution ($\Sigma r^2$ 0.05–0.23) for each presumed daughter is shown (Models 1 to 5).

The obtained major-elements mass balance models show that all the studied Hbl-gabbros ($SiO_2 < 51$ wt%) reflect 14–37% (mean value 21%) fractional crystallization of the Hbl-gabbroic parental melt with a crystallizing/fractionating assemblage of Pl$_{28-37}$ + Hbl$_{10.24}$ + Cpx$_{19-26}$ + Bt$_{3-5}$. A mean representative fractionating assemblage Gb ($Pl_{34} + Hbl_{18} + Cpx_{22} + Bt_8$) is calculated for the purpose of modelling.

Based on the major-element solutions, the FC modelling was applied to selected trace and REE elements (Sr, Y, Dy and Yb). Calculated partition coefficients ($K_d$) for the Gb fractionating assemblage are $D_{Sr}$ 1.10, $D_{Y}$ 0.79, $D_{Yb}$ 0.70 and $D_{Dy}$ 0.74. Results as obtained by application of Eq. (2) are presented in Supplementary Table 4a and are shown in diagrams of Figure 8. Obtained results indicate Hbl-gabbros were produced by $\approx$25% (mean value, range 10–50%) fractional crystallization, consistent with major-elements FC modelling. These results further support the working hypothesis of CG Hbl-gabbros being cogenetic and belonging to the same differentiating line of descent controlled by Rayleigh fractional crystallization.
8.3 EBM processes: protolith-leucosome-melanosome connection

The EBM hypothesis has been tested for all leucosomes \( n = 9 \) and melanosomes \( n = 6 \), again assuming the GH-Gb1 Hbl-gabbro representative of the protolith source prior to the partial melting event. As for the FC-model, considering the primary assemblage of the source rock, the Pl + Hbl + Cpx + Bt assemblage was investigated. The major-element mass balance EBM model was calculated through Eq. (1) in the system \( \text{SiO}_2-\text{TiO}_2-\text{Al}_2\text{O}_3-\text{FeO}^*-\text{MnO}-\text{MgO}-\text{CaO}-\text{Na}_2\text{O}-\text{K}_2 \).
that leucosome composition (SiO\textsubscript{2} = 0.76–0.79; Models 15 to 20) are reported. The major-elements mass balance EBM models show that leucosome composition (SiO\textsubscript{2} > 76 wt\%) reflects ≈1-2% partial melting of the Hbl-gabbro with a mineral residue of Pl\textsubscript{42–43} + Hbl\textsubscript{41–42} + Cpx\textsubscript{13–14}, whereas the melanosome chemistry (SiO\textsubscript{2} < 48 wt\%) is compatible with 3–16% (mean value 9%) residue in equilibrium with a reacting assemblage of Pl\textsubscript{40–43} + Hbl\textsubscript{33–39} + Cpx\textsubscript{11–14}.

Biotite-bearing assemblages (Pl+Hbl+Cpx+Bt) were used as testbed to validate and identify limitations of the presented EBM modelling (Models 21 and 22, Supplementary Table 3d). Batch melt models produce acceptable results (Σr = 0.38) but with a negligible role of biotite (<4%) in leucosome genesis, whereas residuum models completely fail in reproducing the melanosome chemistry.

These results emphasize the main role of the Ca-Al-rich (i.e. Tschermak-bearing) mineral phases, with respect to phyllosilicates, in controlling the partial melting of the gabbroic rocks, and are in general agreement with amphibole-dehydration melting reactions proposed by Beard and Lofgren (1991). Furthermore, calculated mineral assemblages at equilibrium with batch melts (i.e. leucosomes) and residues (i.e. melanosomes) both overlap the primary mineral modal abundances of the Hbl-gabbroic protolith, confirming the modal-type equilibrium batch melting process (e.g. Shaw 1970). For the purpose of modelling, mean representative assemblages Lc (1% melt, Pl\textsubscript{43} + Hbl\textsubscript{42} + Cpx\textsubscript{14}) and Mln (9% residuum, Pl\textsubscript{42} + Hbl\textsubscript{37} + Cpx\textsubscript{12}) are calculated for leucosomes and melanosomes, respectively.

Considering the two obtained mean equality relations:

\[ \text{Hbl-gabbro source} = \text{Melt}_1 + \text{Pl}_{43} + \text{Hbl}_{42} + \text{Cpx}_{14} \quad \text{Eq. (Lc)} \]

\[ \text{Hbl-gabbro source} = \text{Residuum}_9 + \text{Pl}_{42} + \text{Hbl}_{37} + \text{Cpx}_{12} \quad \text{Eq. (Mln)} \]

The following observations are possible: i) the relationship between leucosome and melanosome is ≈1:9, and ii) through a preliminary application of the row reduction method, the sum of leucosome and melanosome compositions corresponds to ≈5 wt% with respect to the original gabbroic source.

Based on the major-element modelling results, the modal EBM modelling was again applied to Sr, Y, Dy, and Yb elements. Calculated partition coefficients (K_{Dj}) are i) for the Lc reacting assemblage D\textsubscript{Sr} = 1.15–1.55, D\textsubscript{Y} = 1.09–1.16, D\textsubscript{Dy} = 1.17–1.18 and D\textsubscript{Yb} = 1.06–1.14, and ii) for the Mln reacting assemblage D\textsubscript{Sr} = 0.84–0.94, D\textsubscript{Y} = 1.00–1.01, D\textsubscript{Dy} = 1.03 and D\textsubscript{Yb} = 1.01. Results obtained for each model by application of Eq. (3) for leucosome and Eq. (4) for melanosome are presented in Supplementary Table 4b and 4c, respectively and in Figure 9. EBM-curves (maximum and minimum) calculated for both batch melts and residues, confirm all leucosomes are the product of modal batch melting with 1–10% degree of partial melting of the Hbl-gabbro protolith. Highly variable results are instead obtained for melanosomes. However, considering the melanosome chemistry comparable to the Hbl-gabbro source, their narrow distribution (Figure 9) around the protolith and the calculated KD values close to 1, impose an overcautious evaluation of results obtained for residues through the Eq. (4). In any case, trajectories of calculated EBM-curves for residues match the melanosome distribution in both Y vs Sr/Y and Yb vs Dy/Yb systems (Figure 9), confirming in our opinion the residuum character of the studied melanosomes.

### 8.4 The partial melting event: thermobaric estimates

A preliminary thermobarometric estimate of the migmatization process of the CG Hbl-gabbros was proposed by Saki et al. (2020). In their works, Saki et al. i) qualitatively investigated biotite vs. hornblende plus plagioclase dehydration melting reactions (see Saki et al. 2020 and references therein), and ii) through the application to mesosomes of hornblende-plagioclase thermobarometry (Holland and Blundy, 1994) and the pseudosection method, they propose the partial melting event occurred at ≈800°C and 3.7 kbar. However, no estimates of the volume of melts generated and the degree of partial melting of the mafic protolith were calculated. In the light of new results from this work such as i) fabric, microfabric and mineral textures investigations, ii) the clear trondhjemite signature of the investigated in situ and in-source leucosomes, iii) the negligible involvement of biotite in a partial melting event mainly controlled by Ca-Al-bearing phases and iv) the very low (≈1-2%) degree of partial melting of the Hbl-gabbro source, it is our opinion that a refining of the pressure-temperature (P-T) conditions of the migmatization event and metatexite forming is needed.

The P-T conditions attained during the partial melting are here constrained through i) microfabric and mineral textures characterization, ii) mineral chemistry thermobarometry (e.g. Larocque and Canil 2010) as derived from hornblende in mesosomes ii) zircon saturation thermometry (Watson and Harrison 1983) and monazite-
Figure 9. A) Y vs Sr/Y and b) Yb vs Dy/Yb diagrams illustrating the genesis of the leucosome-melanosome suite equilibrium batch melting (EBM) of primary Pl+Hbl+Cpx assemblage, starting from gabbroic protolith sample GH-Gb1. Bulk rock/melt partition coefficients used for EBM-models are reported in diagrams. The calculated EBM-curves represent, for every model, the Shaw (1970) modal batch melting solutions calculated for minimum $K_{D}$ (Min) and for maximum $K_{D}$ (Max) values. Batch melts stands for leucosomes, whereas residues are melanosomes. The percentages indicate the amount of liquid fraction. Symbols as in Figure 5.
pointing out the necessity of the revision of the Pl-Hbl of plagioclase-hornblende equilibrium pairs, therefore orthopyroxene overgrowths (Figure 3c,f,g,h). These textures showing also clinopyroxene- and irregular shapes (Figure 3c,f,g,h). These textures can be attributed to hornblende ± plagioclase dominated melting reactions (e.g. Rushmer 1991; Pattison 1991; Thompson and Ellis 1994). However, these textures do not permit a correctly identification of plagioclase-hornblende equilibration pairs, therefore pointing out the necessity of the revision of the Pl-Hbl thermometry estimates produced by Saki et al. (2020).

Following the workflow proposed by Rossetti et al. (2017), and considering the mafic nature of the protolith and the excess of Ti content in the system as indicated by the presence of ilmenite (Figure 3a,c,g,h) and titanite (Figure 3a), we approached classic thermobarometry using the Ti-in-hornblende (Otten 1984) geothermometric and the Al\textsuperscript{VI}-in-hornblende (Larocque and Canil 2010) barometric models as derived from composition of hornblende from mesosomes. The thermobarometric estimates are presented in Supplementary Table 5.

The compositions of the hornblende from mesosomes, with Ti 0.11–0.18 apfu and Al\textsuperscript{VI} 0.17–0.26 apfu, provide P = 3.0–4.4 ±1.0 kbar (weighted mean 3.6 ±0.6 kbar, ±1σ standard deviation of the weighted mean, MSWD = 0.30, n = 10; Al\textsuperscript{VI}-in-Hbl barometry) and T = 682–760 ±25°C (weighted mean 718 ±22°C, ±1σ standard deviation of the weighted mean, MSWD = 1.5, n = 10; Ti-in-Hbl thermometry) (Figure 10). The relative variation of Al\textsuperscript{VI} is also compatible with crystallization during a decomposition process (e.g., Zhang et al. 2014; Hu et al. 2016). These results converge with the shallow crust conditions proposed by Saki et al. (2020) but lower temperatures for the partial melting event.

To better understand the anatexitic event, we integrated zircon saturation thermometry (Watson and Harrison 1983) and monazite- (REE-) solubility model (Montel 1993) applied to leucosome bulk chemistry and iv) forward modelling thermo-barometry (i.e., pseudosection method) applied to the protolith composition using the software program Perple_X (Connolly 2005).

The contact between mesosome and leucosome is characterized by i) plagioclase crystals with rounded and irregular shapes (Figure 3c,f,g,h,i) and ii) destabilized hornblende grains showing also clinopyroxene-orthopyroxene overgrowths (Figure 3c,f,g,h). These textures can be attributed to hornblende ± plagioclase dominated melting reactions (e.g. Rushmer 1991; Pattison 1991; Thompson and Ellis 1994). However, these textures do not permit a correctly identification of plagioclase-hornblende equilibration pairs, therefore pointing out the necessity of the revision of the Pl-Hbl thermometry estimates produced by Saki et al. (2020).

Following the workflow proposed by Rossetti et al. (2017), and considering the mafic nature of the protolith and the excess of Ti content in the system as indicated by the presence of ilmenite (Figure 3a,c,g,h) and titanite (Figure 3a), we approached classic thermobarometry using the Ti-in-hornblende (Otten 1984) geothermometric and the Al\textsuperscript{VI}-in-hornblende (Larocque and Canil 2010) barometric models as derived from composition of hornblende from mesosomes. The thermobarometric estimates are presented in Supplementary Table 5.

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To better understand the anatexitic event, we integrated zircon saturation thermometry (Watson and Harrison 1983) and monazite- (REE-) solubility model (Montel 1993) applied to leucosome bulk chemistry. Results are presented in Supplementary Table 2. Leucosomes show Zr ranging between 34 and 76 ppm (mean 62 ppm), corresponding to T\textsubscript{Zr} of 684–735°C (mean value 718°C) (Figure 10). Leucosomes are characterized by LREE (La+Ce+Pr+Nd +Sm+Gd) in the range 44.2–55.0 ppm (mean 51.1 ppm). The LREE-solubility temperatures (T\textsubscript{REE}) were calculated for variable H\textsubscript{2}O contents. The two models show a general convergence (T\textsubscript{REE} = T\textsubscript{Zr} ± 15°C) at T = 710–737°C for a water content in the range 0.1 wt%, here assumed as near-anhydrous conditions, or fluid-absent condition (e.g. Gao et al. 2016 and references therein). Only GH-L1 and GH-L3 leucosomes show convergence (T\textsubscript{REE} = T\textsubscript{Zr} ± 20°C) at T = 684–698°C for hydrous conditions (estimated H\textsubscript{2}O up to 5 wt%). In agreement with Watson and Harrison (1983) and Montel (1993) and the existing literature on felsic melts (e.g. Miller et al. 2003; Rossetti et al. 2013; Lucci et al. 2018), these models indicate the temperature of leucosomes at the onset of crystallization, and furthermore they should resemble the temperature conditions of the anatectic source (e.g. Rossetti et al. 2013). The convergence at T 718°C between for mesosomes (Ti-in-Hbl) and leucosomes (T\textsubscript{Zr} and T\textsubscript{REE}), in our opinion clearly identify the temperature conditions of the partial melting.

To refine the thermobarometric evolution up to migmatization and partial melting of the Hbl-gabbro, we applied forward modelling thermobarometry to the protolith Hbl-gabbro using the Perple_X 6.9.0 software (Connolly 2005; downloaded June 21 2020, http://www.perplex.ethz.ch). The calculation was carried out in the model system NCKFMASHT (Na\textsubscript{2}O-CaO-K\textsubscript{2}O-FeO*-MgO-Al\textsubscript{2}O\textsubscript{3}-SiO\textsubscript{2}-H\textsubscript{2}O-TiO\textsubscript{2}) assuming i) total iron as FeO to meet the presence of ilmenite, and ii) analysed loss on ignition (LOI) as pure H\textsubscript{2}O fluid. The following solid-solution models were used: Augite (G) for clinopyroxene (Green et al. 2016), Opx (W) for orthopyroxene (White et al. 2014), Amph (DPW) for Ca-hornblende (Dale et al. 2005), Bio (TCC) for biotite (Tajcmanova et al. 2009), Feldspar for feldspar (Fuhrman and Lindsley 1988) and melt (G) for melt (Green et al. 2016). Ilmenite (ilm) end-member is also considered in the calculation. Further details can be found in the file solution.dat included in the PERPLEX software. The reference database is the file hp62ver.dat, the updated version of the thermodynamic dataset of Holland and Powell (2011). The pseudosections were calculated for T = 650–900°C and P = 1.5–5.0 kbar. The mineral phases and melt were expressed in weight % (wt%) to allow a direct comparison with results obtained from the EBM modelling.

A representative simplified pseudosection calculated for the bulk chemistry of sample GH-Gb1 is shown in Figure 10. The isopleths of the melt (wt%) as obtained from the forward modelling are also shown (Figure 10) and are used as a proxy of the thermobaric evolution during the migmatization and partial melting event. The P-T conditions of the anatectic of the CG Hbl-gabbro can be evaluated considering the mineral assemblage, the isopleths for melt and the thermobaric estimates from classical thermobarometry models. The presence of PI +Hbl+Cpx+Opx+Bt in mesosome where in contact with
leucosome, the amount of produced melt <5 wt%, the composition of hornblende in mesosome corresponding to 3.7 kbar and ≈720°C (mean values), the bulk Zr and LREE content of leucosomes corresponding to T 680–740°C for a highly variable H2O (0.1–5 wt%) content, collectively constrain the partial melting event at ≈3.0–4.5 kbar and 700–750°C, at the upper limit of the hornblende hornfels facies conditions in agreement with the works of Slagstad et al. (2005) and Hansen et al. (2015). Assuming a crust density of ≈2700 kg/m3, the obtained pressure corresponds to depths 11–17 km for the gabbro migmatization and genesis of in situ trondhjemite leucosomes.

Following the approach of Nouri et al. (2018), the obtained results are then tested projecting the CG protolith-metagabbro genetic relationship in appropriate petrogenetic systems (e.g., Thompson 1988; Wyllie et al. 1989; Johannes and Holtz 1996; Castro 2013). Based on the CaO-(FeO+MgO)-(SiO2/4) ternary diagram for mafic-felsic magmatic systems (Figure 11a) modified from Castro (2013) from the original CMAS tetrahedron (CaO, MgO, Al2O3, SiO2) proposed by Wyllie et al. (1989), the protolith-metagabbro genetic relationship is clearly confirmed by i) the analysed Hbl-gabbro samples plotting in the compositional space of ‘sources’, ii) the leucosomes falling within the field of silica-rich (SiO2 >70 wt%) melts along the narrow granitic melt array and iii) the melanosomes matching the composition of a solid residuum dominated by Hbl+Cpx phases. Furthermore, when the trondhjemite leucosomes are projected in the normative Qz-(Ab+Or)-An water-saturated ternary system (Figure 11b) of Johannes and Holtz (1996), they all plot i) between the 2kbar and 5kbar Qz-Pl cotectic lines, ii) at a temperature <800°C and iii) close to the eutectic point for the Qz-(Ab+Or) binary system, confirming independently the thermobaric estimates obtained in this study.

In summary, the integration of classic and forward thermobarometry modellings (Figure 10) with appropriate phase-equilibrium projections (Figure 11), suggests that the partial melting of the Hbl-gabbro occurred at HT/LP conditions at temperatures of

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**Figure 10.** Results from classic thermobarometry compared with those obtained from forward modelling thermobarometry. Representative P-T pseudosection calculated for the GH-Gb1 Hbl-gabbro protolith in the system NCKFMASHT (Na2O-CaO-K2O-FeO*-MgO-Al2O3-SiO2-H2O-TiO2) using the Perple_X software. Dashed yellow lines are isopleths indicating the calculated amount of generated melt (wt%). Yellow star is the mean thermobaric estimate obtained through Ti-in-Hbl thermometry and Al[VI]-in-Hbl barometry models (see text for references) applied to hornblende in mesosomes. Zircon saturation thermometry (Tzr) mean value (blue line) and range (cyan field) calculated for leucosome compositions is also presented. The P-T range of Cheshmeh-Ghasaban trondhjemite leucosomes is indicated by the white dashed ellipse. Mineral abbreviations follow Whitney and Evans (2010).
≈700-750°C in the middle-upper Hamedan crust. The low volume (≈1-2 wt%) of melts together with their temperature close to the solidus, their very localized distribution and the limited interconnection between melt pockets and veins, however, suggests that the partial melting event did not evolve up to reach the critical melt percentage of ≈ 8 wt% necessary to trigger the liquid percolation or melt connectivity transition and to develop an efficient melt-drainage and extraction (Vigneresse et al. 1996; Rabinowicz and Vigneresse 2004; Rosenberg and Handy 2005; Brown 2007, 2013; Gao et al. 2016).

8.5 Cheshmeh-Ghasaban trondhjemites vs. experimental investigations

The primary result of this work is the demonstration that in situ trondhjemite leucosomes at CG locality formed near the solidus, at relatively low temperatures (≈700-750°C) and mid-to-upper crustal pressures (3.0–4.5 kbar) by amphibole-dominated low-degree (≈1-2 wt%) partial melting of a Hbl-Bt-bearing gabbroic source rock. Why could these mafic migmatites represent a valuable step forward in the comprehension of partial melting processes?

The anatexis in crust materials is usually developed through dehydration melting (e.g. Clemens 1990) of three main hydrous minerals such as amphibole, biotite and muscovite (Gao et al. 2016 and references therein). Based on typical mineral assemblages, Gao et al. (2016) classified protoliths in four major groups: i) amphibole-bearing mafic rocks, ii) amphibole-biotite-bearing mafic to intermediate rocks, iii) biotite-bearing non-mafic magmatic and metamorphic rocks and iv) biotite-muscovite-bearing metapelitic gneiss, metapelites and metagraywackes. Since the absence of muscovite in the CG gabbroic suite, experimental melts derived by muscovite- and biotite-muscovite-bearing protoliths will be excluded in the hereafter discussion. Furthermore, in agreement with the obtained thermobaric estimates and coherently with the garnet-free character (e.g., Beard and Lofgren 1991; Vielzeuf and Montel 1994; Gao et al. 2016) of the studied migmatites, high-pressure experiments will be excluded too. Only experimental melts produced at low-pressure (LP) (P < 5kbar) conditions by partial melting of amphibole- (Amph), biotite- (Bt) and amphibole-biotite (Amph-Bt-) bearing source rocks (Beard and Lofgren 1991; Holtz and Johannes 1993; Sisson and Grove 1993; Vielzeuf and Montel 1994; Patiño Douce and Beard 1995; Singh and Johannes 1996a, 1996b; Springer and Seck 1997; Patiño Douce 1997; Montel and Vielzeuf 1997; Castro et al. 1999; López and Castro 2001; Koester et al. 2002; Grant 2004; Spicer et al. 2004; Watkins et al. 2007) will be here considered. The melting experiments at 7 kbar of a Hbl-Bt-gabbro presented in the work of Sisson et al. (2005) and partly revised by Ratajeski et al. (2005) are also considered because of the protolith resembling the CG Hbl-gabbro.

The CG trondhjemite leucosomes, in the CaO-Na₂O-K₂O (Figure 12a) and normative Or-An-Ab (Figure
diagrams, plot in the field ‘Amph’ representative of melts from LP (P < 5kbar) Amph-dehydration melting (Beard and Lofgren 1991; Sisson and Grove 1993; Patiño Douce and Beard 1995; Springer and Seck 1997; López and Castro 2001), confirming the EBM modelling that pointed out a Hbl+Pl+Cpx reacting assemblage with non-participation of biotite in the partial melting process. A comparison between CG leucosome compositions and the chemistry of melts from LP Amph-dehydration experiments is also presented in Figure 13 and hereafter discussed.

The selected experimental runs cover temperature in the range 850–1100°C for P = 1–5 kbar. Compositional variations of these experimental melts are shown in representative SiO$_2$ vs. (MgO+FeO$^*$) and SiO$_2$ vs. CaO diagrams (Figure 13). Through the application of the ‘Color Fill Contour’ graphic tool in OriginPro 8.5.0 software, experimental melting temperatures of 900°C and 1000°C are reported in diagrams with isolines (dashed black lines). Clemens (2006) demonstrated how the volume and the chemistry of melts produced by partial melting reactions are mainly controlled by temperature. In particular, Gao et al. (2016) in their review, highlighted how melts derived by partial melting of amphibole-bearing sources, for a fixed pressure value, show significant changes in their compositions coherently with temperature variations. At high temperatures (1000–1115°C) the generated melts show metaluminous signature and low SiO$_2$ (mean 60.47 wt %, n = 20) content together with high MgO (mean 2.99 wt%, n = 20), FeO$^*$ (mean 6.96 wt%, n = 20) and CaO (mean 6.56 wt%, n = 20) values. At low temperatures (850–875°C), experimental melts show metaluminous to peraluminous character and high SiO$_2$ (mean 72.97 wt%, n = 10) content opposite to low MgO (mean 0.43 wt%, n = 10), FeO$^*$ (mean 1.86 wt%, n = 10) and CaO (mean 3.45 wt%, n = 10) values. Experimental runs cooler than 800°C, close to the solidus, are usually not attempted because of the small size of generated melt volumes precluding reliable bulk chemistry analyses (Sisson et al. 2005). However, it is expected that such low-temperature (T < 800°C) melts would show an even higher SiO$_2$ content together with an even lower MgO, FeO$^*$ and CaO contents. However, these expected compositional characteristics fully overlaps with the CG studied trondhjemite leucosomes, generated at T = 700–750°C and P = 3.0–4.5 kbar and showing a peraluminous signature together with a very high SiO$_2$ (mean 78.06 wt%, n = 9) and very low MgO (0.24 wt%, n = 9), FeO$^*$ (0.48 wt%, n = 9) and CaO (mean 2.12 wt%, n = 9) (Figure 13). A further use of the ‘color fill contour’ tool applied to the experimental

![Figure 12](image_url)
melts plus CG leucosomes permitted to also draw the 800°C melting temperature isoline (dashed red line in Figure 13). In this light with respect to LP amphibole-dehydration melting processes, the CG leucosomes from the CG mafic migmatites i) confirm the existing experimental observations, ii) validate the expected compositional trend for melts generated at T < 800°C and iii) contribute to improve our knowledge on amphibole-dominated partial melting event close to the solidus.

Lastly, the CG leucosomes show i) a chemistry with high-SiO₂ (>70 wt%), low Al₂O₃ (<15 wt%), low K₂O (<1.5 wt%), low Sr (33–267 ppm), low Eu (0.31–0.62 ppm) and diluted HREE contents, ii) absence of hornblende in their assemblages, iii) associated residues (i.e. melanosomes) with increasingly enrichment in plagioclase and clinopyroxene over hornblende and iv) a Hbl-gabbro protolith source. It is worth noting these characteristics match the scenario proposed by Barker and Arth (1976) for low-Al₂O₃ trondhjemite melts generated by partial melting of an amphibole-bearing source such as a hornblende-gabbro.

8.6 Implications for the heat source

Based on our documentation of i) textural and microtextural evidence, ii) mineral and whole-rock chemistry, iii) FC and EBM modelling, iv) thermobarometric constraints and v) existing literature, we propose a two-step conceptual model (Figure 14) to describe the emplacement and evolution of the Hbl-gabbro suite and the subsequent migmatization event leading to metatexite mafic migmatite forming and trondhjemite leucosome production, with respect to the local geology.

During the first phase (166.5 ± 1.8 Ma; U-Pb zircon age, Shahbazi et al. 2010) the Cheshmeh-Ghasaban Hbl-gabbro suite (pink balloon in Figure 14a) intruded into the low-grade metamorphic rocks of the Hamedan Phyllite metamorphic complex (Mohajjel et al. 2003; Hassanzadeh and Wernicke 2016), which overlies a Cadomian crust (e.g., Stöcklin and Nabavi 1973; Berberian and Berberian 1981; Berberian et al. 1982; Mohajjel et al. 2003; Golonka 2004; Ghasemi and Talbot 2006; Davoudian et al. 2008; Hassanzadeh et al. 2008; Malek-Mahmoudi et al. 2017; Shabanian et al. 2017; Moghadam et al. 2020). The Hbl-gabbro suite is comprised of Hbl-gabbros and Hbl-monzogabbros, which belong to the same line of descent controlled by Rayleigh fractional crystallization (Figure 14b). In particular, the FC models for major, trace and REE elements show that the Hbl-gabbros and monzogabbros could be the result of 20–25% (mean value) fractional crystallization of Hbl-gabbro (sample GH-Gb1) with a fractionating assemblage of plagioclase, hornblende, biotite, and biotite (Pl₂₈₋₃₇ + Hbl₁₀₋₂₄ + Cpx₁₉₋₂₆ + Bt₃₋₅).

In the second phase, the cooled and solidified Cheshmeh-Ghasaban Hbl-gabbro (grey balloon in Figure 14c) underwent re-heating and local thermal metamorphism, which initiated migmatization and partial melting (Figure 14c) with production of in situ trondhjemite leucosomes and associated melanosomes (Figure 13d). Pressure-temperature estimates obtained through classic and forward modelling of the thermo-barometric conditions applied to the protolith-leucosome suite indicate that migmatization and
Figure 14. Conceptual model for the genesis of trondhjemite leucosomes by partial melting of the Cheshmeh-Ghasaban Hbl-gabbros in Hamedan area, northern Sanandaj-Sirjan zone of Iran. a) Emplacement of the Hbl-gabbro suite at $\approx 167$ Ma (Shahbazi et al. 2010) in the Hamedan Phyllite. b) Schematic representation of the cogenetic origin of Hbl-gabbro suite; the colour gradient and the red arrow indicate the differentiation trend from Hbl-gabbro to monzogabbro via Rayleigh fractional crystallization (FC). c) Emplacement of the coeval to younger Alvand granite batholith $\approx 165-161$ Ma (Shahbazi et al. 2010; Mahmoudi et al. 2011; Chiu et al. 2013; Zhang et al. 2018b) and of the younger (161 vs. 135 Ma; see text for explanation) Cheshmeh-Ghasaban Ol-gabbro (Yang et al. 2018). Contact aureoles indicate the thermal effect on hosting rocks to generate hornfels and to possibly trigger migmatization and partial melting of the previously emplaced Hbl-gabbro. Pressure-temperature conditions as derived by thermobarometry modelling for trondhjemite melts generation, are reported. d) Sketch showing the metamixte mafic migmatite at the outcrops scale; Hbl-gabbro protolith, leucosome and melanosome are indicated with grey, white and black colour, respectively. e) Schematic sketch showing the equilibrium batch melting (EBM) process through a modal melting of the primary Hbl+Pl+Cpx assemblage. Mineral abbreviations (Cpx: clinopyroxene, Opx: orthopyroxene, Bt: biotite, Pl: plagioclase, Hbl: hornblende) follow Whitney and Evans (2010).
trondhjemite melt production occurred at a relatively low-temperatures (700–750°C) at pressures of 3.0–4.5 kbar (mean 3.6 kbar), corresponding to a depth of 11–17 km (mean 14 km). These results are consistent with the existing pressure estimates (range 2.4–5.4 kbar, mean 3.7 kbar; Saki et al. 2020) from thermobaric modelling of mesosomes, but suggest lower temperatures for in situ melt production through the integration of zircon saturation thermometry (Watson and Harrison 1983) and monazite- (REE-) solubility model (Montel 1993) applied to the leucosome compositions.

Regarding the relative low volume (<10 vol%) of leucosomes observed in field (Figure 2a-f), it has been independently confirmed by both i) EBM modelling where the high SiO2 (>76 wt%) leucosome chemistry reflects ≈1-2% partial melting of the Hbl-gabbro protolith (sample GH-Gb1) with a mineral residue of plagioclase, hornblende and clinopyroxene (Pl42-43 + Hbl41-42 + Cpx3-14) (Figure 14e), and ii) P-T pseudosection calculated for the GH-Gb1 Hbl-gabbro protolith in the system NCKFMASH (Figure 10), where the forward modelling and the classic thermobarometry models converge to a melt production <5 wt%, close to the solidus conditions (i.e. the melt-in curve in Figure 10). These results, together with the leucosome field occurrence characterized by i) a very localized distribution and ii) a general limited interconnection between melt micro-patches, pockets and veins, are compatible with an embryonic partial melting event that did not reach the production of the critical melt percentage (≈ 8 wt%) to develop efficient and diffuse melt percolation and extraction (Vigneresse et al. 1996; Rabinowicz and Vigneresse 2004; Rosenberg and Handy 2005; Brown 2007, 2013; Gao et al. 2016).

Which was the heat source that triggered the migmatization of the Hbl-gabbro?

It is worth noting the agreement of our thermobaric results, and in particular our pressure estimates (P 3–4.5 kbar, corresponding to a depth of 11–17 km), for the genesis of the trondhjemite melts with i) the depth of 13–17 km reached by the Hamedan Phyllite surrounding the Alvand plutonic complex (Agard et al. 2005), ii) the pressure conditions (P 5–6 ±2.5 kbar) for the emplacement of Cheshmeh-Ghasab Hbl-gabbro (Sepahi et al. 2013), iii) the P-T peak conditions (T ≈ 750°C and P ≈ 4 kbar, corresponding to depth of ≈15 km) for hornfels and metapelitic migmatites belonging to the Alvand contact aureole and therefore representing the P-T conditions of the emplacement of the voluminous (≈400 km2) Alvand granite (e.g., Sepahi 2008, 2013; Sepahi et al. 2009, 2013, 2020; Saki et al. 2012, 2020; Baharifar et al. 2004, 2019; Shahbazi et al. 2014; Sheikh Ghashlaghi et al. 2020), and iv) the P-T conditions (T 1200–1300°C and P 5–6 ±2.5 kbar) for the Ol-gabbro emplacement (Sepahi et al. 2013).

The Alvand granite and the Ol-gabbro are both likely candidates for the heat source since they: i) are younger than the Hbl-gabbro (Shahbazi et al. 2010; Mahmoudi et al. 2011; Chiu et al. 2013; Yang et al. 2018), ii) intruded the Hamedan Phyllite at the same depth of the Cheshmeh-Ghasab Hbl-gabbro, and iii) were capable to produce a thermal disturbance of at least 750°C necessary for the genesis of the trondhjemites batch melts. However, it is not possible to clearly distinguish the heat source triggering the partial melting. The uncertainty mainly derives by the lack of geochronological data for the trondhjemite leucosomes, and by the only ambiguous existing age(s) (161 Ma vs 135 Ma, U-Pb zircon age) reported by Yang et al. (2018) for the Ol-gabbro. The authors, in fact, report for the same gabbroic sample two distinct magmatic zircon populations presenting both concordant ages of 161.2 ±1.2 Ma and 135 ± 2 Ma, respectively. This peculiar result however is not discussed or explored by authors, therefore producing difficulties in the correct definition of the magmatic age of the Ol-gabbro.

Considering the ages of the Hbl-gabbro (≈167 Ma, U-Pb zircon age; Shahbazi et al. 2010), the granites of the Alvand batholith (≈165-161 Ma, U-Pb zircon ages; Shahbazi et al. 2010; Mahmoudi et al. 2011; Chiu et al. 2013), and the age(s) of the Ol-gabbro (Yang et al. 2018), in our opinion there are only two main possible scenarios: i) a Middle Jurassic partial melting event induced by the magmatic activity of the Alvand complex with a near coeval intrusion at 165–161 Ma of Alvand granites and Ol-gabbros, or ii) an Early Cretaceous partial melting event triggered by the intrusion at 135 Ma of Ol-gabbros. Further geochronological investigations are required to constraint the age of the migmatization of Hbl-gabbro and therefore to correctly present the Mesozoic evolution of the Hamedan area and northern Sanandj-Sirjan zone.

9 Concluding remarks

The CG Hbl-gabbro suite represents a valuable natural laboratory to decipher migmatization and amphibole-dominated partial melting of a mafic source in upper crust conditions. Significant outcomes of this work are:

i) The Middle Jurassic CG Hbl-gabbro (~167 Ma) intruded in the Hamedan Phyllite, locally shows the evidence of in situ leucosome-melanosome suite. The geochemical characteristic of leucosomes indicates they are low-Al2O3 high-SiO2 trondhjemite melts. Associated melanosomes present a mafic character dominated by plagioclase and clinopyroxene over hornblende.

ii) The Rayleigh FC modelling for major and trace + REE elements, outlines that the CG Hbl-gabbroic rocks
are cogenetic from a unique line of descent controlled by fractional crystallization (≈10–40%) of a Pl_{28-37} + Hbl_{10-24} + Cpx_{19-26} + Bt_{3-5} primary assemblage.

iii) The EBM modelling for major and trace + REE elements, indicate that in situ trondhjemite leucosomes and associated melanosomes in studied metatexite migmatite are the product of ≈ 1–2% degree of partial melting of the Hbl-gabbro protolith. Proportions of primary Hbl+Pl+Cpx underwent melting were those of the primary gabbroic assemblage. Biotite was not involved in the partial melting reactions, thus clearly indicating a leucosome genesis through amphibole-dehydration melting process.

iv) Application of classic and forward modelling thermobarometry, together with independent phase-equilibria projections, constrained the migmatization event and the genesis of CG trondhjemite leucosomes at the mid-to-upper crust (P 3.0–4.5 kbar, 11–17 km at depth) and relatively low temperature (≈700–750°C) conditions.

v) The CG in situ and in-source leucosomes with their Qtz+Pl assemblage and their chemistry confirm the temperature-controlled compositional trends of melts from low-pressure amphibole-dehydration melting experiments, and finally show the potential to decode amphibole-dominated partial melting at T < 800°C and close to the solidus, filling the existing experimental gap.

vi) The voluminous Alvand granititic batholith and the Cheshmeh-Ghasaban Ol-gabbro younger intrusions are both possible candidates for the heat source triggering the partial melting event.

In conclusion, the results obtained in this study and the comparison with experimental melts from amphibole-dehydration melting provide compelling evidence for the petrogenesis of the studied in situ trondhjemite leucosomes through partial melting of a Hbl-gabbro protolith. Further geochronological investigations are however required to better constrain the age of this partial melting event and enable us to more fully understand its implications for the Mesozoic geodynamic scenario of the northern Sanandaj-Sirjan zone of Iran.

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