Exhumation of a migmatitic unit through self-enhanced magmatic weakening enabled by tectonic contact metamorphism (Gruf complex, Central European Alps)

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Abstract
The Central Alpine lower crustal migmatitic Gruf complex was exhumed in contact to the greenschist-grade Chiavenna ophiolite and gneissic Tambo nappe leading to a lateral gradient of ~ 70 °C/km within the ophiolite. The 14 km long, E-W striking subvertical contact now bridges metamorphic conditions of ~ 730 °C, 6.6 kbar in the migmatitic gneisses and ~ 500 °C, 4.2 kbar in the serpentinites and Tambo schists 2–4 km north of the contact. An obvious fault, mylonite or highly sheared rock that could accommodate the ~ 8.5 km vertical displacement is not present. Instead, more than half of the movement was accommodated in a 0.2–1.2 km thick orthogneiss of the Gruf complex that was heterogeneously molten. Discrete bands with high melt fractions (45–65%) now contain variably stretched enclaves of the adjacent MOR-derived amphibolite. In turn, the adjacent amphibolites exhibit tonalitic in-situ leucosomes and dikes i.e., were partially molten. The H₂O necessary for fluid-assisted melting of the orthogneiss and amphibolites was likely derived from the tectonic contact metamorphism of the Chiavenna serpentinites, at the contact now in enstatite + olivine-grade. U–Pb dating of zircons shows that partial melting and diking occurred at 29.0–31.5 Ma, concomitant with the calc-alkaline Bergell batholith that intruded the Gruf. The major driving forces of exhumation were hence the strong regional North–South shortening in the Alpine collisional belt and the buoyancy provided by the Bergell magma. The fluids available through tectonic contact metamorphism led to self-enhanced magmatic weakening and concentration of movement in an orthogneiss, where melt-rich bands provided a low friction environment. Continuous heating of the originally greenschist Chiavenna ophiolite and Tambo gneisses + schists by the migmatitic Gruf complex during differential uplift explains the skewed temperature profile, with intensive contact heating in the ophiolite but little cooling in the portion of the now-exposed Gruf complex.

Keywords  Migmatites · Central Alps · Zircon · Fluid-present melting · Magmatic weakening

Introduction
Migmatitic units, often associated with calc-alkaline intrusions, usually characterize the hot cores of orogenic systems (Mehnert 1968; Brown 1993; Sawyer et al. 2011). Partial melting at crustal scales changes the bulk rheology of the continental crust by lowering its viscosity and hence differential stresses that may be sustained (e.g., Jamieson et al. 2011). Consequently, part of the orogenic scale net deformation partitions into such low-viscosity, melt-bearing zones (Vanderhaeghe 2009).

Melt-induced rheological weakening has long been recognized (Arzi 1978), as little as 7–10 vol. % melt may lower rock strength by an order of magnitude (Rosenberg and Handy 2005). In rocks dominated by quartzo-feldspathic mineralogies, the amount of (granitic) melt is initially determined by the availability of H₂O. The influx of aqueous fluids (1) lowers the melting point to the wet solidus and, (2) determines melt fractions, which in H₂O-saturated bulk compositions close to the granitic minimum or eutectic may vary from 0 to > 90% over a few tens of degrees. H₂O in the melt may arise either from external influx of fluids formed by metamorphic reactions, through fluid-absent muscovite melting at 720–780 °C that typically produces 7–15% melt, or by fluid-absent biotite melting, which generates larger
amounts of melt (Vielzeuf and Holloway 1988). Most orogenic systems are characterized by maximal metamorphic temperatures of 650–750 °C, suggesting that fluid-saturated melting should dominate, but also that fluid-absent muscovite melting may well be reached (Berger et al. 2008; Weinberg and Hasalova 2015).

Partial melting in the crust is thought to affect crustal dynamics by increasing buoyancy and promoting uplift. The ascent of migmatitic units is favored in compressional systems, where doming is enhanced by tectonic forces (Brown 1994; Vanderhaege 2009). Rapid uplift of “hot” migmatitic units may cause heat advection to shallower levels, lead to elevated thermal gradients and enhance prograde metamorphic reactions as e.g., described from the Gruf complex in the European Central Alps (Schmutz 1976; Galli et al. 2013), the Naos metamorphic core complex in Greece (Keay et al. 2001), the French Massif Central (Bouilhol et al. 2006), or the Karakorum metamorphic complex in Pakistan (Rolland et al. 2001). Such features are also predicted by numerical modelling (Gerya et al. 2004; Rey et al. 2011). Melt proportion in migmatitic terranes ought to depend on H$_2$O-availability, which is arguably highest at contacts and along thrusts. Uplift or overthrusting of hot migmatitic blocks may be self-enhancing when heating of adjacent rocks yields prograde fluids, which in turn increase melt fractions, lubricating initial deformation zones and amplifying focalization of deformation. The understanding of the thermo-mechanical behavior of migmatitic terranes and their contacts is hence crucial for deciphering strain distribution and thermal structures of orogens.

In this study, we investigate the contact zone between the Gruf complex, a migmatitic block of the European Central Alps and the adjacent, originally greenschist-facies units, i.e., the Chiavenna ophiolite and gneissic Tambo nappe (Fig. 1a, b). We show that this contact zone accommodated ~ 8.5 km of near-vertical relative uplift of the migmatitic Gruf complex without forming significant mylonites or bands with visible large strain or any major faults. Instead, the contact is characterized by a migmatitic orthogneiss rich in mafic enclaves derived from the ophiolitic unit, the partially molten orthogneiss acting as low-viscosity layer. The uprising migmatitic unit metamorphosed the greenschists to conditions of partial melting, promoting dehydration-reactions that provided fluids for enhanced melting. Field work, thermodynamic modelling coupled with thermobarometry, bulk rock chemistry, and Hf and U–Pb zirconology are used.

Fig. 1 a Regional map of the Central European Alps (after Galli et al. 2012). b Map of the contact area of the migmatitic Gruf complex with the Chiavenna ophiolite/Tambo nappe, which are at uppermost greenschist-facies a few km from the contact. Two sections across the contact (A-B and C-D), which accommodated ~8.5 km uplift, were studied, units plunge almost constantly 60–80° N with an ENE strike (see also Fig. 2). Isograds and Chiavenna map after Schmutz (1976). Gruf complex after Galli et al. (2013)
to characterize the contact, decipher its thermal structure, quantify the differential exhumation, and determine the timing of melting, the melt source(s), melt fractions and distribution, altogether constraining the mechanism of exhumation of a hot migmatitic terrane.

Geological setting

Migmatites in the central Alps

The investigated area is in the south-eastern corner of the Lepontine dome, the Barrovian-type metamorphic core of the European Alps (Fig. 1a). The southern Lepontine dome exposes an east–west striking, steeply dipping, 100 × 15 km wide migmatite belt (e.g., Burri et al. 2005). Migmatization occurred at 32–25 Ma (Berger et al. 2009; Galli et al. 2012) at upper amphibolite-facies conditions (T ≤ 750 °C) and most resulted from fluid-assisted melting at < 7 kbar (Nagel et al. 2002; Berger et al. 2008). Fluid-absent muscovite melting is restricted to the area around Bellinzona and to the Gruf complex (Fig. 1a; Burri et al. 2005; Galli et al. 2013), whereas there is no evidence of fluid-absent biotite melting of indisputable Alpine age.

The Gruf complex

The Gruf complex is a 15 × 10 km migmatitic unit. To the East and South it is bounded by the Oligocene, mantle-derived, calc-alkaline Bergell pluton (von Blanckenburg 1992), to the Southwest by the peraluminous, crustal-derived Novate granite (25 Ma, Liati et al. 2000), to the West by the Forcola fault, an Oligocene normal fault separating the Gruf from the Adula nappe (Ciancaleoni and Marquer 2006), and to the North by the Chiavenna ophiolite and Tambo nappe (Fig. 1a). Hitherto, the northern contact was described through the putative Gruf line, an allegedly major mylonitic zone (Schmutz 1976). However, only a few irregular, laterally limited centi- to decimetric mylonitic bands exist, which could not have accommodated the 5–10 km uplift between the Gruf and Chiavenna + Tambo units (Galli et al. 2013).

The Gruf complex consists mainly of steeply dipping, ENE-WSW oriented, hundreds of m to km wide bands of migmatitic orthogneisses, migmatitic alkali feldspar-sillimanite-garnet-bearing paragneisses and micaschists and associated leucogranites (Fig. 1b; for details see Galli et al. 2013). U–Pb zircon intrusion ages of the leucogranites and orthogneisses are mostly Permian (Galli et al. 2012), yet, the northernmost (enclave-rich) migmatitic orthogneiss contains abundant Cambro-Ordovician zircon cores. Migmatization occurred at 675–750 °C and 5.0–7.5 kbar (Galli et al. 2013; Oalmann et al. 2019) at 34–29 Ma (Galli et al. 2012). The alkali feldspar-sillimanite isograd, marking the occurrence of fluid-absent muscovite melting, coincides with the boundaries of the Gruf complex. Deformation is organized in a set of characteristically 0.5 m thick, syn- to late-migmatitic duc- tile shear zones that yield a top-to-the NE, dextral-normal movement of uplift of the Gruf complex with respect to the northern units (Galli et al. 2013).

The migmatitic orthogneisses are characterized by ubiquitous charnockite lenses and rare, highly residual sapphireine-opx granulites (Wenk et al. 1974; Galli et al. 2011). Metamorphic conditions of the granulites and charnockites were 900–1000 °C and 8–10 kbar (Galli et al. 2011; Oalmann et al. 2019). The age of this ultra-high-temperature (UHT) event is debated. Galli et al. (2012), based on zircon U–Pb ages in the charnockites, suggested a Permian (282–260 Ma) age, also because the Permian zircon cores include orthopyroxene and garnet. On the other hand, Schmitz et al. (2009) proposed an Alpine and Niccol et al. (2018) both a Permian and Alpine granite grade, based on 34–30 Ma monazite ages in restitic granulites.

The south-adjacent calc-alkaline Bergell pluton

The ~ 300 km² Bergell pluton is an Oligocene syn-orogenic, calc-alkaline intrusion directly north of the Periadriatic Fault System (Rosenberg 2004). It postdates collision and is thought to relate to slab break-off (von Blanckenburg and Davies 1995). In this model, collision caused underthrusting of subduction-modified mantle below the European crust, slab break-off allowed influx of asthenospheric mantle, isostatic rebound and subsequent melting of the underthrust metamorphized mantle, resulting in typical calc-alkaline plutons in a peculiar position i.e., in the subducting European plate, rather than in the overriding Adriatic plate. The isostatic rebound after slab break-off likely opened space for magma emplacement and for magma channeling and ascent along the precursor shear zone to the Periadriatic Fault (e.g., Rosenberg 2004; Galli et al. 2012).

The main intrusion phase occurred at 33–30 Ma (von Blanckenburg 1992; Oberl et al. 2004; Gianola et al. 2014), similar to the 34–29 Ma zircon rims dating the freezing of the migmatitic melt in the Gruf (Galli et al. 2012). Field studies have shown that intrusion and syn-kinematic folding of the Bergell pluton were contemporaneous with migmatization and deformation in the Gruf complex and with its uplift and juxtaposition with the Chiavenna ophiolite and Tambo nappe (Rosenberg et al. 1995; Berger et al. 1996; Galli et al. 2013).

Contact metamorphism by the Bergell pluton to the migmatitic Gruf complex is limited, the granodioritic magma (800–850 °C, Samperton et al. 2017) was not that much hotter than the Gruf complex itself (700–750 °C). A < 100 m wide contact aureole is recognizable as corundum-cordierite-bearing garnet-biotite-sillimanite-schists, an increase of
migmatization and consequent back-veining of melts from the migmatites into the pluton’s margin (Galli et al. 2013). In contrast, at its east, the Bergell pluton intruded into green-schist-facies Chiavenna ophiolite and Tambo nappe producing a 2 km wide contact aureole that reaches anthophyllite-olivine grade and a lateral thermal gradient of 60–90 °C/km (e.g. Trommsdorff and Connolly 1996).

Metamorphism at the northern contact: Chiavenna ophiolite and Tambo Nappe

The northern contact of the Gruf migmatites is mostly to the Mesozoic Chiavenna ophiolite that forms three 3–5 km long and up to 3.5 km thick bodies (Fig. 1b), which are structurally overturned, i.e. the mafics face the Gruf complex (Fig. 2). Only in its central part, the contact is to the Tambo nappe. The Chiavenna ophiolite is composed of metamorphosed basaltic and peridotitic rocks, which are epidote-amphibolites and serpentinites in the northernmost part, and of rare metacarbonates (Schmutz 1976). Approaching the contact from the North, peridotites cross the talc-in, antigorite-out, anthophyllite-in/talc-out, anthophyllite-out, enstatite-in and spinel-in isograds, while in the mafic rocks, epidote is progressively replaced by amphibole; diopside occurs, however, mostly in amphibolites adjacent to metacarbonate rocks (Schmutz 1976; Talerico 2000). Within the last few hundred meters to the contact, migmatites and a network of syn-kinematic felsic dikes occur in the amphibolites. In this study we show that part of this melt derives from the Gruf migmatites, but part also from the amphibolites, implying fluid-assisted melting of the metabasalts at 700–750 °C. Isograds are parallel to the contact and indicate an increase in temperature from ≤ 540 to ≥ 700 °C (at 4–5 kbar) over a horizontal distance of 2.0–3.5 km (Schmutz...
metric, granoblastic leucosome bands, rich in plagioclase, and garnet (+ plagioclase + quartz), alternate with centi-

cern migmatitic Gruf paragneisses and micaschists (Fig. 2a). The southern end of the cross section begins in the south-

tural discontinuity, dip steeply to the North. The descriptions

cliffs but often in forests with poor outcrops. All rock units

The Tambo nappe (Fig. 1a) consists of metapelites and Permian metagranitooids characterized by lower amphibio-
lite-facies conditions of 500–550 °C and 3–7 kbar (Baudin

and Marquer 1993). Similar to the Chiavenna ophiolite, the

The contact zone between the migmatitic Gruf complex and the Chiavenna ophiolite/Tambo nappe

This study investigates the northern contact of the Gruf (Fig. 1b) along its entire length, exposed in part in vertical

cross sections, along Val Schie-
sone in the West (A-B, Figs. 1b, 2a) and along the Denc dal

The Val Schiesone cross section

The southern end of the cross section begins in the souther-

orthogneiss constitutes the northernmost large orthogneiss

Next, the ENE-WSW striking enclave-rich migmatitic orthogneiss constitutes the northernmost large orthogneiss sheet of the Gruf complex, typically 1 ± 0.2 km thick, but here at its thinnest, ~200 m (Fig. 1b). The rock is massive to foliated and composed of quartz, plagioclase, alkali feldspar and coarse-grained biotite, with minor muscovite, chlorite, allanite/epidote, zircon, titanite and apatite. Centi- to deci-

Next, a < 100 m thick, along strike discontinuous sheet (Fig. 1b) of northern migmatitic paragneisses and micaschists occurs just south of the Chiavenna ophiolite. Paragneisses and micaschists are similar to the southern

paragneiss unit described above, except of (1) occasional cordierite, (2) generally fewer and smaller leucosome pock-

North of the contact, the southern amphibolites of the

The 700 m of amphibolite directly north of the contact display massive centimetric leucocratic patches or dikelets within small-scale ductile shear bands and intra-boudin pockets (Fig. 3c–e). These leucosomes are made of euhedral millimetric plagioclase, quartz, ± diopside, ± amphibole with minor alkali feldspar, biotite, titanite and accessory apatite, calcite and zircon and interpreted as in-situ leucosomes. Furthermore, largely undeformed centimeter to meter wide leucocratic dikes and dike networks (Fig. 3d, e) with sharp to diffuse borders occur within the first few hundred meters of the contact. The dikes can be separated into two endmem-

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massive and homogeneous, and contain epidote, titanite and rutile but no diopside. Evidence of in-situ partial melting or leucocratic dikes are absent.

**Fig. 3** Partial melting across the contact a Southern migmatitic Gruf paragneiss with a melt filled shear band, Val Schiesone. b Enclave-rich band within the migmatitic orthogneiss, Denc dal Luf. c Chiavenna amphibolite 500 m north of the contact with tonalitic hornblende + diopside – bearing leucosomes, Val Schiesone. d Close-up of crossing and mixing dikes in the Chiavenna amphibolite, the leucocratic dikes are granitic while the darker dikes are intermediate granodiorite dikes, Val Schiesone. e Dike network within the Chiavenna amphibolite about 300 m from the contact, Val Schiesone

**The Denc dal Luf cross section**

The enclave-rich migmatitic orthogneiss south of the Denc dal Luf peak forms a steeply dipping, ~1.0 km thick sheet (Figs. 1b, 2b), which fabric changes from massive to
strongly foliated towards the northern contact. The orthogneiss has the same coarse-grained, biotite-rich, quartzo-feldspathic mineral assemblage as in Val Schiesone. Evidence for migmatization is hard to identify in the field because of the often homogeneous, massive character, a lack of stromatic textures and a limited color contrast between leucosomes and palaeosomes. Where the rock is more foliated, partial melting can be recognized by cross-cutting relationships. This orthogneiss hosts centi- to decimetric mostly amphibolitic enclaves (Fig. 3b) overwhelmingly concentrated in 1–30 m thick bands parallel to the main foliation, their frequency increasing towards the northern contact. Within the enclave-rich bands, the orthogneiss is more massive and near the enclaves enriched in biotite and plagioclase and some amphibole.

The enclaves themselves are angular, rounded, or strongly elongated, often fractured and intruded by granitic leucosomes from the host orthogneiss (Fig. 3b). Enclaves are mostly mafic but occasionally calc-silicate, ultramafic or pelitic. Mafic enclaves are composed of amphibole, plagioclase, biotite with minor chlorite. Chemical interaction with the granitic matrix of enclave-rich bands results in crystallization of abundant biotite, often concentrated at the enclave rim, and of poikiloblastic amphibole, plagioclase, quartz, epidote and titanite at the expense of the polygonal amphibolite assemblage.

The northern migmatitic paragneisses and micaschists north of the enclave-rich orthogneiss are strongly foliated and composed of granoblastic quartzo-feldspathic leucosome bands alternated with biotite, garnet and sillimanite-rich darker bands that locally bear muscovite and cordierite. These metasediments contain 1–10 m long lenses of foliated diopsidite-bearing amphibolite.

Next, a garnet-bearing two-mica leucogranite has the topographic shape of a “wolf’s tooth” giving the name “Denc dal Luf”. This leucogranite is fine to coarse-grained, massive to slightly foliated and composed of quartz, plagioclase, alkali feldspar, muscovite, garnet ± biotite, chlorite, apatite, ilmenite and zircon.

Peridotitic rocks of the Chiavenna ophiolite occur with about 200 m thickness immediately north of the two-mica leucogranite. The first 20 m show massive quartzo-feldspathic leucosomes that intruded and brecciated the strongly serpentinitized, now chlorite-rich ultramafic rocks. Locally, these leucosomes are sheared into < 5 cm thick mylonites. Otherwise, the olivine-enstatite-chlorite-spinel metaperidotites are slightly foliated and homogeneous. They contain < 1 m thick meta-rodingite dikes.

At the northern end of the cross section, two-mica paragneisses and micaschists of pelitic to psammitic composition constitute the southern end of the Tambo nappe. These metasediments are strongly foliated, fine- to middle-grained and composed of quartz, biotite, muscovite, sillimanite and plagioclase, with minor chlorite, apatite and zircon.

**Orientation of the main structures and kinematics**

Planar and linear structures in the Gruf complex, Chiavenna ophiolite and Tambo nappe display similar orientations. The main foliation in the contact-near Gruf dips 60–80° to the NNW (Fig. 4) and 50–70° to the N-NNW in the Chiavenna ophiolite and Tambo gneisses + schists (Schmutz 1976; Galli et al. 2013). Deviations are locally observed in the ophiolite, where foliations deflect and wrap around highly competent metaperidotite masses.
In the Gruf complex and Tambo nappe, a stretching lineation is defined by elongated biotite in the enclave-rich migmatitic orthogneiss and by elongated biotite and sillimanite in the migmatitic paragneisses and micaschists. In the Chiavenna ophiolite, elongated hornblende or tremolite define the stretching lineation in the amphibolites or peridotites. These lineations plunge predominantly 20–50° to the E-ENE, similar to the rest of the Gruf complex and also Bergell pluton (Galli et al. 2013).

Along the contact, deformation is expressed as 1–20 cm thick, isolated or conjugated, steeply dipping ductile shear zones. Three sets of shear zones are identified: (1) dipping to the NNW with stretching lineations dominantly plunging 20–40° to the NE to ENE and top to the SW sense of shear (Fig. 4a), (2) dipping N with stretching lineations plunging 20–50° E to ENE and top to the NE sense of shear (Fig. 4b), and (3) dipping NE with stretching lineation plunging 10–60° E to ENE and top to the NE sense of shear (Fig. 4c). Together, they describe a syn-migmatitic, top the NE dextral-normal movement. Similar shear zones and kinematics are reported from the rest of the Gruf complex and the Bergell pluton (Davidson et al. 1996; Galli et al. 2013), the Novate granite (Ciancaleoni and Marquer 2006), the southeastern Adula nappe (Meyre et al. 1998) and the Southern Steep Belt (Schmid et al. 1989). In the enclave-rich migmatitic orthogneiss and in the Chiavenna amphibolites, these shear zones show in-situ leucosomes accumulated at junctions and inter-boudin structures (Fig. 3c) and are commonly intruded by syn-kinematic leucocratic dikes (Fig. 3e). All these observations testify for deformation contemporaneous to migmatization.

**Shape analysis of enclaves in the enclave-rich migmatitic orthogneiss (Gruf complex)**

Aspect ratios of enclaves (Fig. 5a) in the migmatitic orthogneiss were measured within the main foliation and parallel to the mineral stretching lineation in a N-S profile at Denc dal Luf. Furthest from the contact, aspect ratios are 2–4, then increase to 6 and finally to 12 in the last tens of meters. This increase correlates with a stronger sub-solidus fabric in the orthogneiss host (Fig. 5b, c), more dynamically recrystallized quartz and more deformed micas towards the contact. In the northernmost 30 m, quartz and biotite sizes are strongly reduced and feldspars generally fractured with in-fills of quartz and chlorite/epidote indicating deformation under lower amphibolite-facies conditions.

These observations show strain concentration in the enclave-rich bands and a strong near- or sub-solidus strain gradient across the enclave-rich migmatitic orthogneiss. Remarkably, in the more southern enclave trails, enclaves are almost undeformed, which is attributed to their incorporation into the orthogneiss in presence of ample melt.

The northwards increasing enclave deformation and strain localization down to middle and lower amphibolite-facies conditions demonstrate that deformation continued during cooling, which resulted from juxtaposition of the “hot” Gruf with the originally greenschist-facies northern units.

**Geochemistry**

A total of 59 samples from the Gruf complex, the Chiavenna ophiolite and the Tambo nappe were sampled and analyzed for major and trace elements (analytical methods and data in supplementary material).

**Chiavenna amphibolites**

The Chiavenna amphibolites, which represent oceanic crust of the Valais ocean, are akin to N-MORB (Talerico 2000) with 46–50 wt% SiO₂, 1–2 wt% TiO₂, 5–9 wt% MgO, 10–14 wt% CaO and < 0.7 wt% K₂O (Fig. 6), traces are factor 2–3 more enriched than N-MORB (Fig. 6h).

**Mafic enclaves within the migmatitic orthogneiss (Gruf complex)**

Major element compositions of the mafic enclaves are mostly in the range of the Chiavenna amphibolites, but several have slightly higher SiO₂ and lower CaO-contents (Fig. 6). In addition, some enclaves are distinctly FeO and TiO₂-rich (Fig. S5 in supplementary material) suggesting a more fractionated MORB-magma not yet sampled in the Chiavenna amphibolites. Trace-element patterns of the enclaves are similar to the Chiavenna amphibolites except for a tenfold increased LILE (Fig. 6h). This enrichment in LILE and to a lesser extend LREE correlates with K₂O, enclaves with the lowest K₂O being most similar to the Chiavenna amphibolites. In thin section, high K₂O and LILE correlate with the abundance of neo-formed biotite flakes in the enclaves. This systematics suggests that the enclaves are pieces of Chiavenna amphibolite mechanically eroded and incorporated into the host biotite-orthogneiss. The enrichment in SiO₂, K₂O and LILE results from chemical interaction between enclaves and orthogneiss, in particular from granitic melts derived from the migmatitic orthogneiss.

**Tonalitic patches, veinlets and granitic dikes in the southern Chiavenna amphibolites**

Felsic patches in the amphibolites that are texturally in-situ leucosomes, veinlet networks and dikes are quartzo-feldspathic and vary in their mafic minerals. The endmember approaching in-situ leucosome has amphibole ± cpx, almost no biotite and tonalitic compositions while the endmember...
granitic dikes have abundant biotite and no amphibole. Yet, all compositions in between also exist (Fig. 6). Concomitant melting in two principally different source rocks is also confirmed by zircon ages and chemistry (in particular $\varepsilon$Hf, see below). Chemically, the most siliceous biotite-rich dikes are similar to the enclave-rich orthogneiss matrix and have higher $\text{SiO}_2$, $\text{K}_2\text{O}$ and LILE but lower $\text{FeO}^\text{T}$, $\text{MgO}$, $\text{MnO}$, $\text{TiO}_2$ and also HREE than most of the tonalitic dikes (Fig. 6). The granitic dikes tend to be slightly peraluminous and crystallize minor muscovite while the hornblende-bearing dikes are meta-aluminous.

The amphibolite-derived leucosomes and most tonalitic dikes are comparatively low in $\text{K}_2\text{O}$ and LILE, and have HREE similar to N-MORB (Fig. 6), also V, Cr and Ni correspond to N-MORB values. The more leucocratic dikes and leucosomes have 55–60 wt% $\text{SiO}_2$ (anhydrous), too little compared to experimental near-solidus melts of MORB (65–73 wt%, France et al. 2010) or to oceanic plagiogranites thought to derive from fluid-assisted melting of oceanic crust (on average $70.7 \pm 5.7$ wt%, Spulber and Rutherford 1983). We posit that the undeformed in-situ leucosome patches and syn-kinematic dike networks (Fig. 3c–e) provide clear textural evidence for melts. It appears most likely that the observed compositions are cumulative (in plagioclase and amphibole) and have suffered extraction of granitic melts during crystallization. At the extreme, a few leucocratic plagioclase-rich dikes have $\text{MgO}$ and $\text{FeO} < 2$ wt% and up to 24 wt% $\text{Al}_2\text{O}_3$ (Fig. 6), i.e., are strongly plagioclase cumulative, consistent with a $\text{Na}_2\text{O}$-enrichment that cannot be derived through mixing. Nevertheless, most veinlets and dikes are granodioritic, due to mixing orthogneiss- and MORB-derived partial melts, trace element patterns are coherent with such mixing.

In summary, the tonalitic melts present in the southernmost Chiavenna amphibolites are derived from N-MORB sources (confirmed by $\varepsilon$Hf, see below) while most of the
granitic dikes and veinlets are from a meta-granitic, i.e. continental crust source rock, mixed to various degrees with the partial melts of the amphibolite.

**Pressure–temperature conditions across the contact**

Metamorphic P–T conditions in the migmatitic paragneisses and micaschists of the Gruf complex, as well as in the micaschists of the Tambo nappe (sample locations in Fig. S1 in supplementary material) were determined from pseudosections calculated with Perple_X (Connolly 2005), the thermodynamic data of Holland and Powell (2011), and mineral solution models as given in the supplementary material. H₂O-contents for the Gruf gneisses and schists were estimated from loss of ignition, which represents a lower bound, increasing H₂O-contents by a few wt% does not substantially change phase relations. For the un-migmatitic Tambo samples, H₂O-saturation was used. Complementary garnet-Al₂SiO₅-quartz-plagioclase barometry (GASP, Koziol and Newton 1988) was applied to the sillimanite-saturated rocks, using an error of ±0.6 kbar. Chiavenna amphibolite in-situ leucosome and tonalitic dike crystallization

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**Fig. 6** a–f Chemical variations in the Chiavenna amphibolites, in the enclaves of the enclave-rich migmatitic orthogneiss and in the various leucosomes and dikes found in the amphibolites. For the correlation of dike type with mineralogy and source rock see text. g Spider diagrams of the gneissic + granitic and h basaltic and tonalitic lithologies of the Gruf-Chiavenna contact. 23 of 31 amphibolite analyses are from (Talerico 2000, dark green dots)
conditions were calculated from Al-in-hornblende barometry (Schmidt 1992) and amphibole-plagioclase thermometry (Holland and Blundy 1994). Mineral compositions were measured by electron microprobe, when zoned, mineral rim compositions were used in the calculations (data and further pseudosections in supplementary material).

**PT conditions in the southern migmatitic Gruf paragneisses and micaschists**

Two migmatitic paragneisses from the Val Schiesone cross section (samples SC02 and SC03) were investigated. Sample SC02 is composed of garnet (1–2 vol%), biotite, plagioclase, quartz, sillimanite and ilmenite. Garnets display an increase of Mn and decrease of Mg towards the rim, but no Ca zoning. Plagioclase is homogenously $X_{An} = 0.24$. The pseudosection (Fig. 7a) yields 705–735 °C and 4.4–7.5 kbar, the GASP barometer 5.9–7.3 kbar.

Sample SC03 displays a similar mineralogy but bears also alkali feldspar. Garnet (1–2 vol%) is almandine rich and homogeneous, plagioclase is Ca-poor ($X_{An} = 0.25$) and unzoned. In pseudosection, the observed assemblage is stable at 720–740 °C and ~4–8 kbar. Pressures calculated using the GASP barometer for this temperature range yield 5.9–7.2 kbar.

Calculated conditions of 720–735 °C and 5.9–7.2 kbar are in excellent agreement with previous estimates of 675–750 °C and 5.0–7.5 kbar for the Gruf complex (Galli et al. 2013; Oalmann et al. 2019).

**PT conditions in the northern migmatitic paragneisses and micaschists**

One migmatitic paragneiss of the Denc dal Luf section (sample DC30) and one of the Val Schiesone (sample SC17) were investigated. Sample DC30 is composed of rare garnet (< 0.5 vol%), biotite, muscovite, plagioclase, quartz, sillimanite and ilmenite. Garnet displays a slight increase in Mn and decrease in Mg towards the rim, Ca is constant, plagioclase is homogenously $X_{An} = 0.30$. The calculated pseudosection (Fig. 7b) yields this paragenesis at 680–700 °C and 5.0–6.6 kbar, the GASP barometer 5.3–5.7 kbar.

Sample SC17 displays a similar mineralogy but lacks also alkali feldspar. Garnet (1–2 vol%) is almandine rich and homogeneous, plagioclase is Ca-poor ($X_{An} = 0.25$) and unzoned. In pseudosection, the observed assemblage is stable at 720–740 °C and ~4–8 kbar. Pressures calculated using the GASP barometer for this temperature range yield 5.9–7.2 kbar.

Calculated conditions of 720–735 °C and 5.9–7.2 kbar are in excellent agreement with previous estimates of 675–750 °C and 5.0–7.5 kbar for the Gruf complex (Galli et al. 2013; Oalmann et al. 2019).

**PT conditions of melting in the southern Chiavenna amphibolites**

Two in-situ leucosomes (samples MP02 and SC12) and two tonalitic dikes from the amphibolites (SC31 and SC13) were investigated. Both leucosomes contain euhedral Mg-hornblendes with $X_{Mg} = 0.67$ and 0.52, respectively, and 6.6 ± 0.1 Si per formula unit (pfu). Hornblende rims have $1.77 ± 0.11$ and $1.69 ± 0.11$ Al pfu yielding pressures of $5.4 ± 0.5$ and $5.1 ± 0.5$ kbar (MP02 and SC12, Fig. 7d). Plagioclase has $X_{An} = 0.37 ± 0.03$ and $0.31 ± 0.04$, hornblende-plagioclase thermometry yields 665 ± 0.3 °C in MP02 and 730 ± 0.4 °C in SC12.

In the tonalitic dikes, euhedral amphiboles are Mg-hornblende with an $X_{Mg} = 0.70$ and 6.7–6.8 Si per formula unit. Al-contents of $1.53 ± 0.08$ and $1.60 ± 0.09$ pfu in amphibole rims yield pressures of $4.6 ± 0.4$ in sample SC31 and $4.3 ± 0.4$ kbar in sample SC13. Temperatures calculated from hornblende rims and the fairly homogeneous plagioclase ($X_{An} = 0.31–0.32$ for both samples) are $710 ± 40$ °C in SC31 and $695 ± 25$ °C in SC13. Leucosomes and tonalite dikes commonly display texturally late greenish (actinolitic) amphibole overgrowths richer in Si and Mg but poorer in Altotal than the euhedral grains, interpreted as retrograde and fluid-mediated.

In summary, partial melting within the southern Chiavenna amphibolites yields temperatures around 700 °C, in agreement with a wet amphibolite solidus of 670–680 °C (at 5 kbar, Poli 1993). Pressures range from 5.4 to 4.3 kbar, are statistically moderately distinct and likely represent the depth range of 16–20 km over which these melts crystallized.
Zircon

Ca. 1–10 kg of fresh rock were crushed and sieved to < 500 µm, only the Chiavenna amphibolite in-situ leucosome SC12 had less material. Zircons were separated using a Wilfley table and a magnetic separator. 30–150 zircons per sample were handpicked, mounted in epoxy, polished to expose longitudinal center sections and imaged by cathodoluminescence. In leucosome MP02, a drill core from a polished riverbed surface, zircons were directly measured in thin section. In zircons U–Pb and Lu–Hf isotopes and trace elements were analyzed by LA-ICP-MS at ETH Zurich. Results are summarized in (Table 1), sample locations and details on analytical methods, data reduction and analyses are given in the supplementary material.

Enclave-rich migmatitic orthogneiss

Five samples far from the enclave-rich bands and four enclave-rich bands have been dated. In all samples,
zircons are elongated and prismatic, 70–380 µm long and 20–150 µm wide (Fig. 8a, b). In the enclave-free zones these display oscillatory-zoned cores truncated by bright, homogenous, 5–25 µm wide rims. In the enclave-rich bands, rims are often oscillatory and 10–100 µm wide (Fig. 8b) suggesting larger degrees of Alpine melting. Three of nine samples have 10–40% oscillatory zoned cores with Siluro-Ordovician ages of 480–400 Ma. The other cores are mostly

Table 1 Summary of zircon ages, εHf, and melt fraction derived from the zircon melt-o-meter

| Sample | Inherited and detrital zircon cores | Alpine zircon and overgrowth | Melt-o-meter (T = 730 °C) |
|--------|------------------------------------|-----------------------------|--------------------------|
|        | 238U/206Pb dates εHf | Carboniferous-Triassic | Ages εHf | [Zr]bulk | M | FPZ | F_melt |
|        | [Ma] | Carboniferous-Triassic | [Ma] | MSWD | [ppm] | |
| Gruf orthogneisses far from enclave-rich bands | | | | | |
| DC11  | 316 to 210 | − 7.2 to − 4.3 | 30.6 (.7) | 3.0 | − 7.5 to − 3.2 | 191 | 1.3 | 0.22 (.05) | 0.32 (.07) |
| DC23  | 460 to 404, 315 to 215 | − 5.4 to − 4.3 | No age | − | − 11.9 to − 8.6 | 248 | 1.3 | 0.06 (.02) | 0.11 (.04) |
| DC24  | 296 to 231 | Not analyzed | 29.5 (1.3) | 3.3 | Not analyzed | 260 | 1.3 | 0.11 (.02) | 0.22 (.04) |
| DC25  | 479 to 402, 305 to 206 | − 5.4 to − 1.6 | 29.9 (7) | 2.9 | − 8.2 to − 0.6 | 138 | 1.3 | 0.16 (.02) | 0.17 (.02) |
| DC29  | 457 to 412, 289 to 153 | − 5.2 to − 1.2 | 31.1 (5) | 5.9 | − 6.9 to − 1.7 | 134 | 1.3 | 0.61 (.07) | 0.63 (.07) |
| Gruf granitic matrix of enclave-rich bands | | | | | |
| DC52  | 316 to 203 | − 5.7 to − 2.5 | 30.6 (.5) | 4.7 | − 9.9 to − 6.4 | 238 | 1.4 | 0.41 (.06) | 0.66 (.10) |
| DC14  | 318 to 177 | Not analyzed | 30.0 (.5) | 1.3 | Not analyzed | 105 | 1.4 | 0.61 (.06) | 0.44 (.05) |
| DC63  | 650 to 404, 316 to 215 | − 6.3 to − 2.9 | 30.7 (.5) | 2.8 | − 10.2 to − 6.7 | 197 | 1.4 | 0.38 (.06) | 0.51 (.08) |
| DC75  | 316 to 215 | − 4.7 to − 1.9 | 30.8 (.5) | 2.0 | − 9.0 to − 5.5 | 132 | 1.4 | 0.58 (.05) | 0.52 (.05) |
| Gruf southern migmatitic paragneisses | | | | | |
| CP04  | 603 to 411, 297 to 219 | − 6.8 to + 3.1 | 30.9 (.8) | 0.2 | − 17.0 to − 1.4 | − | − | − | − |
| VC5   | 674 to 424, 265 to 220 | − 7.1 to − 5.2 | 29.1 (.6) | 1.4 | − 22.9 to − 7.3 | − | − | − | − |
| Migmatitic paragneisses and micaschists north of the enclave-rich orthogneiss | | | | | |
| SC17  | 1133 to 514, 333 to 211 | − 21.5 to − 4.6 | No age | − | − 18.2 to + 9 | − | − | − | − |
| DC30  | 854 to 510, 275 to 245 | − 13.2 to + 2.3 | 28.8 (.9) | 0.3 | − 15.7 to − 0.8 | − | − | − | − |
| Granitic dikes in the southern Chiavenna amphibolites | | | | | |
| SC14  | 523 to 426, 295 to 219 | − 4.1 to − 2.0 | 30.4 (.6) | 2.7 | − 9.3 to − 4.6 | − | − | − | − |
| SC38  | 662 to 540, 271 to 206 | − 3.4 to + 0.7 | 30.8 (.6) | 2.5 | − 15.9 to − 4.2 | − | − | − | − |
| Tonalitic dikes in the southern Chiavenna amphibolites | | | | | |
| SC13  | − | − | 30.6 (.5) | 2.2 | − 1.0 to + 3.6 | − | − | − | − |
| SC31  | − | − | 30.2 (.5) | 1.2 | − 5.1 to − 1.6 | − | − | − | − |
| In-situ leucosomes in the southern Chiavenna amphibolites | | | | | |
| SC12  | − | − | 30.4 (.6) | 1.0 | + 9.8 to + 13.4 | − | − | − | − |
| MP02 (in -situ dating) | − | − | 31.5 (.8) | 0.7 | Not analyzed | − | − | − | − |
| Northernmost amphibolite (Chiavenna ophiolite) | | | | | |
| S01   | − | − | 38.5 (.7) | 1.3 | + 10.1 to + 13.0 | − | − | − | − |
| Gruf two-mica leucogranite (Denc dal Luf) | Permian age | | | | |
| DC33  | 690 to 449, 286 to 246 | Not analyzed | 266 (5) | 1.7 | Not analyzed | − | − | − | − |

Bold numbers correspond to calculated ages.
Fig. 8 Cathodoluminescence images of representative zircons from the Grufl-Chiavenna contact lithologies. a Strongly deformed migmatitic orthogneiss, Denc dal Luf. b Strongly molten (~65%) matrix of enclave-rich band in the migmatitic orthogneiss, Denc dal Luf. c Southern migmatitic paragneiss, Cappella di Pizzo. d Northern migmatitic metapelite, Chiavenna amphibolite, Val Schiesone. e granite dike, Val Schiesone. f Tonalitic dike, Val Schiesone. g In-situ leucosome in Chiavenna amphibolite, Val Schiesone. h Epidote-bearing amphibolite, Piuro. Further samples in the supplementary material.
Permian, averaging 280 Ma, similar to two orthogneiss samples of Galli et al. (2012) that yielded core age peaks at 450 and 290 Ma. The nine rims are Oligocene in age yielding 31.1–29.5 Ma (typically ± 0.7 Ma, Table 1, Fig. 9 a, b and supplementary material), interpreted as the age of migmatization. Moreover, samples DC75 and DC25 yield for <10% of the concordant rim spots a second age cluster at around 25–26 Ma.

Eighty percent of the Siluro-Ordovician cores have an εHf of −3.3 to 1.0, the Permian cores an εHf of −7.2 to
− 1.2. Alpine zircon rims scatter widely between − 11.9 and − 0.6 (Table 1). Using the 176Hf/177Hf ratio of the Permian cores, we calculated the 176Hf/177Hf evolution based on a 176Lu/177Hf ratio of 0.015 (Griffin et al. 2002). This leads to an εHf of − 9.7 to − 4.5 at 30 Ma, which corresponds to − 90% of the Oligocene rims (Fig. 10a, b). This result is consistent with dissolution of the Permian zircon cores during Alpine partial melting, the Oligocene 176Hf/177Hf scattering simply reflecting the zircon core 176Hf/177Hf heterogeneity.

**Southern migmatitic Gruf paragneisses and micaschists**

Two samples (CP04 and VP5) have been dated. Zircons are elongated and prismatic, sub-euhedral, 60–210 μm long and 30–100 μm wide. They show oscillatory zoned or homogeneous cores surrounded by oscillatory zoned to homogeneous, up to 40 μm thick bright rims (Fig. 8c). Texturally, more than three growth generations are recognizable in many zircons.

Ca. 70% of the zircon cores crystallized between 674 and 411 Ma with an εHf of − 13.3 to + 0.5. 15% have Permo-Triassic ages of 297–219 Ma and εHf = − 7.1 to + 3.1. Zircon rims yield ages of 30.9 ± 0.8 (Fig. 9c) and 29.1 ± 0.6 Ma, interpreted as the age of migmatization. The εHf values of the Alpine rims spread between − 22.9 and − 1.4 (Table 1), reflecting the high 176Hf/177Hf variability of the detrital zircon cores in the metasediments.

**Northern migmatitic paragneisses and micaschists**

Two samples (DC30 and SC17) have been dated. Zircons are elongated and prismatic, mostly rounded, 75–230 μm long and 40–130 μm wide (Fig. 8d). Cores show age populations from 1133 to 510 Ma with a clear peak at 680–500 Ma and with an εHf of − 16.9 to + 10.2. Again, 10% of the core and rim analyses display concordant Permo-Triassic ages of 275–211 Ma, with an εHf of − 21.5 to + 2.3. Somewhat surprising, most zircon rims have the same Permo-Triassic age range, suggesting re-melting during Permian times and only very limited or no melting during Alpine times. Four Alpine discordant spots in sample DC30 yield 28.8 ± 0.9 Ma (Fig. 9d and Table 1).

**Granitic dikes in the southern Chiavenna amphibolites**

Two granitic dikes (samples SC14 and SC38) have elongated to prismatic zircons, 80–300 μm long and 20–80 μm wide. Zircons exhibit oscillatory zoned cores wrapped by dark, 10–60 μm wide oscillatory zoned rims (Fig. 8e). Some cores are 662–590 Ma (SC38) and 523–426 Ma (SC14), while 40% of the cores are Permo-Triassic (295–206 Ma) with an εHf of − 4.1 to + 0.7. Zircon rims and rim-less zircons yield concordant ages of 30.4 ± 0.6 and 30.8 ± 0.6 Ma, taken as the age of the dike crystallization (Fig. 9e and supplement), and have εHf of − 15.9 to − 4.2.

**Tonalitic dikes and in-situ leucosomes in the southern Chiavenna amphibolites**

Two tonalitic dikes (samples SC13 and SC31) have zircons that are mostly elongated and prismatic, 20–250 μm long.
10–60 µm wide with bright oscillatory-zoned cores and 10 µm wide, dark rims (Fig. 8f). Cores and rims are indistinguishable in age, concordant spot analyses yield 30.6 ± 0.5 and 30.2 ± 0.5 Ma, respectively, interpreted as the dike intrusion age (Fig. 9f–n and supplementary material). Zircons in sample SC13 and SC31 yield an εHf of −1.0 to +3.6 and −5.1 to +1.6, respectively (Fig. 10). Clearly, εHf and bulk rock chemistry suggest that tonalitic versus granitic dikes had different sources. The tonalitic dikes indicate a mixture of (MOR-) amphibolite- and gneiss-derived melts while the granitic dikes indicate gneiss-derived melts from continental crust sources i.e., the adjacent Gruf complex.

Two in-situ leucosome zircons (SC12 and MP02) are rounded to slightly elongated, 20–60 µm long, 10–40 µm wide and have homogenous bright (SC12) or dark (MP02) cores yielding an age indistinguishable from the occasional dark rims (Fig. 8g). SC12 yields a concordant age of 30.4 ± 0.6 Ma (Fig. 9g–o), while concordant and discordant spot analyses in MP02 yield 31.5 ± 0.8 Ma. This age represents the age of in-situ melting in the amphibolites, the εHf of +8.3 to +13.4 corresponds to values expected for a MOR source.

### Northern Chiavenna amphibolite

Zircons are dark, homogeneous, 10–50 µm long, 10–30 µm wide (Fig. 8h) and yield an Eocene concordant age of 38.5 ± 0.7 Ma (Fig. 9h–p). The εHf of +10.1 to +13.0 is consistent with a MOR-signature (Fig. 10). The calculated age suggests that the Chiavenna epidote-amphibolites are unaffected by the Oligocene tectonic contact metamorphism, this age reflecting the timing of the regional nappe formation (Schmid et al. 1996).

### Two mica leucogranite at Denc dal Luf

Zircons are elongated, prismatic, eu- to subhedral, 150–300 µm long and 30–80 µm wide. Cores are moderately luminescent and display oscillatory, sector or planar zoning. Rims are dark and < 10 µm wide. Permo-Triassic cores and rims represent 75% of the concordant spot analyses and yield an age of 266 ± 3 Ma (Table 1), whereas no concordant analyses of Alpine age occur.

### Zircon trace-element chemistry

Zircons from most lithologies have Th/U of 0.02–1.0, only the metasedimentary zircons cluster at Th/U = 0.002–0.01 (Fig. 11a). Hf-concentrations are low in the zircons of the Chiavenna N-MORB amphibolites and their partial melts, but still overlap with the lowermost end of the orthogneiss zircons (Fig. 11a). Chondrite normalized Rare Earth Element abundances show congruent ranges for all lithologies and LaN/LaN scattering to > 90% between 10−2 and 10−5. The metasedimentary zircons have no Ce-anomaly (Ce/Ce*, Ce* = CeN/(LaN0.6PrN)0.5) while most other rocks have positive ones, scattering from 2 to > 100 in each rock type (for REE plots see supplementary material). Eu-anomalies (Eu/Eu*, with Eu* = EuN/(SmN0.5GdN)0.5, Fig. 11b) are weak in the Chiavenna amphibolites, in their in-situ melts and in their tonalitic dikes, but well expressed in the orthogneisses and the granitic dikes derived thereof. Overall, source and partial melt zircons are chemically similar for the orthogneisses—granitic dike pair and for the amphibolites, in-situ leucosomes and tonalitic dikes.

### Fraction and distribution of anatectic melt in the enclave-rich orthogneiss

The fraction of newly precipitated zircon (FPZ) during the youngest melting event in a given rock can be used to quantify the fraction of granitic melt (Fmelt) present at the time when the granitic host rock became a closed system. This method is based on Zr-solubility in the granitic melt (Boehnke et al. 2013), requires the relative zircon rim volume derived from image analysis, melt composition, and bulk [Zr] (Mintrone et al. 2020). We determined Alpine melt fractions for four granitic matrix samples of enclave-rich bands in the orthogneiss, and five from enclave-free localities (Denc dal Luf cross section, Fig. 5). For each sample, we used the same averaged melting temperature of 730 °C across the orthogneiss, the compositional parameter M was taken from the melt compositions obtained from thermodynamic modelling.

Zircons in the granitic matrix of the enclave-rich bands have 31.1–29.5 Ma old, up to 100 µm thick, oscillatory-zoned rims (Fig. 8b). Rim-fractions determined as rotational volumes are 0.41 ± 0.06, 0.61 ± 0.06, 0.38 ± 0.06 and 0.58 ± 0.05 corresponding to melt fractions Fmelt = 0.66 ± 0.10, 0.44 ± 0.05, 0.51 ± 0.08 and 0.52 ± 0.05, respectively (Table 1). Zircons in the enclave-free domains have Alpine rims up to 25 µm thick (Fig. 8a). Rim-fraction in four samples yield narrow Gaussian distributions with a full half width (FWHM) of 0.06–0.22. These rims correspond to melt fractions of 0.32 ± 0.07 (2 S.E.), 0.11 ± 0.04, 0.22 ± 0.04 and 0.17 ± 0.02. A further sample (DC29) collected near the contact yields heterogeneous rim volume fractions mostly between 0.40 and 0.80, but also contains 15% homogeneously Alpine zircons. The average rim fraction amounts to 0.61 ± 0.07, corresponding to a melt fraction of 0.63 ± 0.07 (Table 1).

In summary, the overall degree of Alpine melting in this orthogneiss was mostly within 11–32%, but the enclave-rich bands and the contact had 44–66% melt, requiring the addition of 4–6 wt% of H2O. We surmise that the melt-rich bands enabled incorporation of the enclaves and, more importantly,
allowed ample movement and uplift that was not recorded in any rock texture. We postulate that most of the uplift was accommodated in the melt-rich state, yet, deformation continued to sub-solidus conditions, leading to enclave deformation at a relatively late stage, and to the observed deformation gradient that corresponds to cooling from the North.

**Discussion: pressure–temperature–time evolution**

**Pre-uplift conditions**

For the northernmost Gruf complex, we obtain metamorphic conditions of 720–735 °C and 5.9–7.2 kbar (Fig. 7), conditions that fit well the 675–750 °C and 5–7.5 kbar reported for the rest of the Gruf complex (Galli et al. 2013; Oalmann et al. 2019), including the cordierite-bearing symplectites in the residual sapphire-granulites (720–740 °C, 7–7.5 kbar, Galli et al. 2011, 2013; Oalmann et al. 2019). These pressures also fit well with those of the adjacent Bergell pluton (Davidson et al. 1996). There is hence no metamorphic gradient across the Gruf complex within the precision of thermobarometry, other than the contact metamorphism leading to 775 °C in the last 100 m towards the Bergell intrusion (Galli et al. 2013).

Nevertheless, the Gruf migmatites host widespread charnockites and rare ultrahigh-temperature residual granulites formed at > 920 °C and 8.5–9.5 kbar, but in Permian times (260–282 Ma, Galli et al. 2012). Alpine conditions that led to zircon rims of 30–33 Ma, were only of upper amphibolite-facies, precisely at the muscovite-out (720–740 °C, Berger et al. 2008; Galli et al. 2013). We acknowledge that Schmitz et al. (2009) propose a 34–30 Ma age for the restitic granulites based on Oligocene monazite inclusions in sapphire, whereas Niccol et al. (2018) propose both a Permian and an Alpine granulite-grade, arguing that monazite resetting to 34–30 Ma would not be possible at 720–740 °C. We counter these arguments by (1) the experiments of Williams et al. (2011) who show that monazite may lose any prior age information through fluid-aided alteration that is essentially an in-situ recrystallization at temperatures as low as 450 °C, far below diffusion closure temperatures, and (2) the
prograde contact metamorphism towards the concomitant Bergell intrusion which requires the Gruf to be at ≤ 750 °C at 33–30 Ma, (3) a complete absence of any indication of major re-melting of the granulites that would have to have accompanied an Alpine > 900 °C event of regional significance. Moreover, we note that all but one temperature calculated by Ti-in-zircon thermometry for the Oligocene zircon rims in the residual granulites gave 650–750 °C (Oalmann 2017), indicative of Alpine upper amphibolite-facies conditions. Furthermore, neither Schmitz et al. (2009) nor Nicollet et al. (2018) propose any tectonic scenario allowing for 900 °C at Alpine times. We also wonder how the calc-alkaline Bergell granodioritic melts should stagnate and solidify at country-rock temperatures of 900 °C.

The pre-uplift conditions of the Chiavenna ophiolite and Tambo nappe to the North of the Gruf complex are of uppermost greenschist-facies. In their North, peridotites and metabasalts are serpentinites and epidote-chlorite-amphibolites formed at ≤ 540 °C and 4.0–4.5 kbar. The paragneisses and micaschists of the Tambo nappe at 3–4 km from the contact with the Gruf complex contain biotite + muscovite + garnet + staurolite + chlorite (+ quartz + plagioclase), indicating ≤ 550 °C (Schmutz 1976). Andalusite + staurolite schists 1.5 km north of the contact yielded a pressure of 4.2 kbar ± 0.3 kbar (Schmutz 1976; Dymoke and Sandiford 1992).

**Temperature–profile across the Gruf–Chiavenna/Tambo contact**

At the northern margin of the Gruf complex, temperatures decrease slightly from 720 to 735 °C to 700 °C from the southern to the northern migmatitic paragneisses (Fig. 12a). Within the first 500 m north of the contact, the partly molten
southern Chiavenna amphibolites in Val Schiesone reached 665–730 °C. In the Denc dal Luf profile the non-migmatitic micaschists of the Tambo nappe reached 610–680 °C (Fig. 12a) and peridotites at olivine + enstatite-grade require 680–700 °C. Nevertheless, a proper enstatite-in isograd cannot be drawn, as enstatite occurs only in eight locations along the 14 km contact, some of these associated with metacarbonates (Schmutz 1976). Northwards, the Chiavenna amphibolites cross the epidote-in isograd at 1.8 km from the contact, while peridotites cross the enstatite-out/anthophyllite-in, talc/in-anthophyllite-out and serpentine-in at 1.0, 2.2 and 2.5 km from the contact (in prolongation of the Val Schiesone profile, Schmutz 1976, Fig. 1b). Similarly, paragneisses and micaschists of the Tambo nappe progressively cross the staurolite-in and sillimanite-out isograds at 1.2 and 1.5 km from the contact (near the Denc dal Luf profile, Schmutz 1976, Fig. 1b).

In summary, within 2.0–3.5 km north of the enclave-rich migmatitic Gruf orthogneiss temperatures steadily decrease from ~730 °C to ≤ 540 °C, which corresponds to a thermal gradient of 70 ± 10 °C/km. Further north, the gradient is 15–20 °C/km, mainly defined by isograds in the peridotite (Fig. 12a). The elevated thermal gradient calculated for the tectonic Chiavenna-Tambo contact zone is comparable to the 70 °C/km characterizing contact aureoles from calc-alkaline tectonic Chiavenna-Tambo contact zone is comparable to the mean temperature and a symmetric temperature profile (Fig. 12a). This yields a pressure decrease of about 1.4 kbar across the 0.2–1.2 km thick orthogneiss and ~ 1 kbar across the next 0.5 to 1 km. In total, the pressure decrease is 2.4 ± 0.8 kbar over < 2 km, yielding an apparent overall pressure gradient of 1–2 kbar/km (Fig. 12b) and a total uplift of 8.5 ± 3.0 km. The mostly 1 km thick enclave-rich migmatitic orthogneiss accommodated ca. 60% of this differential uplift between the Gruf and the northern units, i.e., ca. 5 km, in agreement with the steeply dipping foliation and the syn-migmatitic, dominantly top to the NE, dextral-normal movement along the contact. The other 3.5 km uplift occurred in part in the northern metapelites, and in part in the Chiavenna amphibolites and Tambo gneisses + schists. Within the orthogneiss, this uplift (of 5 km) occurred along melt-rich shear planes.

**Timing of migmatization and diking**

Zircons rims in the contact-near migmatitic lithologies of the Gruf complex crystallized at 29–31 Ma, identical to the partly molten paragneisses, Permian orthogneisses and leucogranites from the rest of the Gruf complex (Galli et al. 2012). Zircon rim ages of 31–34 Ma are also reported for the UHT charnockites (Galli et al. 2012), and of 30–34 Ma for the residual sapphire-bearing granulites (Oalmmann et al. 2019). The latter also have monazite ages of 30–34 Ma (Schmitz et al. 2009; Nicollet et al. 2018). Some of these monazites are located in post-peak symplectites (Oalmmann 2017; Nicollet et al. 2018) formed at amphibolite-facies conditions (Galli et al. 2011). Hence, the 29–31 Ma age range is the time of Alpine partial melting under upper amphibolite-facies conditions of 720–735 °C.

In the Chiavenna amphibolite leucosomes, zircons mirror the strongly positive εHf values of the MORB host and yield ages of 30.4–31.5 Ma, the age of partial amphibolite melting (Fig. 9). Ar–Ar ages of amphiboles from the amphibolites yielded similar ages of 35–30 Ma (Talerico 2000). Zircons from the syn-kinematic dikes intruded into these amphibolites also give intrusion ages of 30–31 Ma (Fig. 9).

The northernmost Chiavenna epidote-amphibolites formed at ≤ 550 °C contain tiny zircons with strongly positive εHf values that crystallized at 38.5 ± 0.8 Ma. This
Eocene age is consistent with the metamorphic age of 37.1 ± 0.9 Ma from a similar amphibolite sampled 2.5 km north of the contact to the Gruf complex (Liati et al. 2003). 37–38 Ma is hence the age of the regional uppermost green-schist-facies metamorphism of the Chiavenna ophiolite.

Altogether, the zircon data show that fluid assisted partial melting in the Gruf complex and southern Chiavenna amphibolites as well as the intrusion of the syn-kinematic dikes therein are 29.0–31.5 Ma in age. This suggests that differential uplift between the migmatitic and ophiolitic units of ca 8.5 km was concomitant with the emplacement of the calc-alkaline Bergell pluton (30–33 Ma, von Blanckenburg 1992; Oberli et al. 2004; Gianola et al. 2014).

Concordant zircon spot analyses from the Gruf complex obtained from newly formed zircon grains show a Gaussian distribution with a peak at 30.6 Ma and a FWHM of ~3 Ma (i.e., ±1.5 Ma), representing ~70% of the precipitated zircons. Nevertheless, some of the enclave-rich orthogneisses exhibit a few younger concordant zircon rim analyses (~10%) of 29–25 Ma (Fig. 9b) that overlap with the 25–29 Ma ages of pegmatite and aplite dikes that cut across structures in the Gruf complex (Liati et al. 2003; Oalmann 2017). Prolonged fluid flow is likely also responsible for a few younger zircon spot analyses in the granitic dikes and leucosomes of the Chiavenna amphibolites, which show a FWHM of 2–3 Ma (Fig. 9).

**Discussion: mechanism of exhumation**

**Melt enhanced uplift**

Previously, the uplift of the Gruf complex was thought to be accommodated along the Gruf line, a putative large-scale mylonite zone (Schmutz 1976). Our detailed mapping and profiles show that mylonites are scarce, at most tens of meters long and typically a few to ten cm thick. Our pressures derived across the contact show that 5 of the 8.5 km uplift were accommodated within the enclave-rich migmatic orthogneiss (Fig. 12b), which is massive to slightly foliated and hosts mostly undeformed enclaves. The amount of anatetic melt at 29.0–31.5 Ma calculated from the zircon melt-o-meter (Mintrone et al. 2020) is typically 11–32% in most of the orthogneisses, but increases to 44–66% within enclave-rich bands. Such elevated proportions of melt yield low viscosity zones that easily accommodate large displacements but at the same time would not register any syn-mafic (or syn-migmatitic) deformation (Fig. 3b).

The enclave-rich orthogneisses may form such elevated quantities of melt at T < 750 °C only through fluid-saturated melting. Similarly, in-situ partial melting in the adjacent amphibolites also requires fluid-saturation. The source of this H2O is readily identified as the serpentinites of the Chiavenna ophiolite, their prograde metamorphism to enstatite + olivine grade liberating their ~8 wt% H2O. The mineral isograds in the meta-peridotites are parallel to the Gruf contact zone, suggesting that heat advected from the Gruf complex caused the dehydration reactions, a positive feedback mechanism where heat from the uplifted block generates H2O-rich fluids that enhance melting, which in turn facilitates uplift (Fig. 13). Advecting metamorphic fluid deeply into hot terranes may be difficult (Catwright and Buick 1998; Weinberg and Hasalova 2015), but as noted by Sawyer (2010), fluids are drawn into large-scale deformation zones, triggering fluid-fluxed melting and enhancing melt production (Grenier et al. 2008; Weinberg et al. 2009). Therefore, we propose that the enclave-rich migmatic orthogneisses, where uplift and hence deformation concentrated, acted as a sink for the prograde fluids of the tectonic contact metamorphism.

**Imbrication along the uplift zone**

In the migmatic orthogneiss, the mafic enclaves suggest that the Chiavenna amphibolites were mechanically eroded and fragments incorporated into the zones of maximal movement during deformation. Also, the granitic dikes derived from the Gruf gneisses and emplaced in the Chiavenna amphibolites imply melt injection into the amphibolites near the contact zone. Likely, both dehydration induced fluids and large melt fractions in the enclave-rich orthogneiss did facilitate hydraulic fracturing in the Chiavenna amphibolites (Fig. 3c–e).

On a 100–1000 m scale, the northern paragneisses + schists form a laterally discontinuous sheet (Fig. 1b) and are also imbricated with Chiavenna amphibolites. Furthermore, a slice of deformed Permian two-mica leucogranite in the Denc del Luf profile (Fig. 2b) likely stems from the Gruf complex that contains several Permian leucogranite sheets (Galli et al. 2013). The stratigraphic provenance of the northern paragneisses + schists instead is ambiguous. These may constitute (1) a pre-Permian metasediments slice of the Gruf complex, (2) Mesozoic metasediments of the Valais ocean analogous to those in the northwestward prolongation of the Chiavenna unit, the Misox zone, (3) an originally autochthonous sedimentary cover of the Mesozoic Chiavenna ophiolite, similar to the metacarbonates now contained within the amphibolites, or (4) could derive from the Tambo nappe, which would require a movement opposite to that of the uplift tectonics, hence deemed unlikely. The zircons in these metasediments have cores of 1133–510 Ma and of 275–211 Ma, rims are also mostly Permo-Triassic. In general, the zircons are irregularly rounded, and only the very few and thin Alpine rims are idiomorphic (four discordant spots of 28.8 ± 0.9 Ma, Fig. 9d). It is hence possible that the in the Permian magmatically
recrystallized zircons were sedimented in the Mesozoic and our preferred and probably least complex stratigraphic position would be that of a Mesozoic oceanic sediment.

**Large scale tectonics and driving forces**

Melt-enhanced buoyancy is commonly invoked to explain exhumation of lower to mid-crustal blocks in high-grade terranes (Vanderhaege 2009). Buoyancy-driven upward flow of lower continental crust is thought to lead to migmatitic domes with a core of anatectic granites (Teyssier and Whitney 2002; Rey et al. 2011). In such a model, the buoyancy force relates to the amount of melt present, which in turn depends on temperature and the availability of fluids, which may be considerably enhanced during convergent tectonics (Vanderhaege 2009).

The syn-kinematic migmatization and exhumation of the Gruf complex occurred during Alpine collision at 29–32 Ma and relate to N-S shortening and E-W stretching (Fig. 13a, b), as indicated by the regional orientation of stretching and mineral lineations. In the migmatites, the amount of Alpine melt estimated from zircons is 10–30%. Higher melt fractions (44–66%) are restricted to the enclave-rich bands of the orthogneiss. This suggests that migmatite buoyancy, even if tectonically amplified, is unlikely the major driving mechanism of exhumation.

In many orogenic systems worldwide, migmatitic units are instead associated with large plutons (Faure et al. 1999; Corsini and Rolland 2009), pointing to a close relation between uprising batholite forming melts and exhumation of migmatite terranes. In the Gruf complex, migmatization, deformation and uplift are coeval to the emplacement of the adjacent Bergell pluton. We propose that the rise of
the Bergell magmas played a key role during the exhumation of the Gruf complex, which was squeezed between two magma lubricated sheets, the enclave-rich migmatitic orthogneiss to north and the Bergell pluton to the east and south (the West being buried in a valley that likely hides the continuation of the Forcola fault).

Conclusions

The Gruf complex is a lower crustal polymetamorphic migmatitic unit that has been differentially exhumed by about 8.5 km. There is no recognizable Alpine metamorphic gradient across the Gruf complex suggesting a rather homogeneous “en block” uplift and exhumation. The northern uplift zone is exceptionally well preserved and exposed, which allows to demonstrate that (1) uplift concentrated within 1–2 km and (2) was facilitated by locally large melt fractions that (3) in turn were fed by fluids derived from tectonic contact metamorphism of adjacent serpentinites. Consequently, we propose a self-enhancing mechanism in which initial uplift caused heating and dehydration of the serpentinites, which then lead to a steadily increasing melt fraction across the uplift zone, efficiently facilitating and concentrating movement therein. The thermal gradient of 70 ± 10 °C across this tectonic contact metamorphism zone is similar to that of larger scale batholiths in the middle to upper crust. The major driving forces of uplift were the contemporaneous Bergell pluton and the N-S shortening of the Alpine orogeny, leading to a wedge-like, high temperature tectonic exhumation of the migmatitic block and a consequent thermal perturbation of adjacent crust at shallower levels.

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