Are microstructures in plutonic rocks primary or secondary?: a re-examination of the metasomatism hypothesis for the roof-sourced autoliths in the Skaergaard intrusion.

Marian B. Holness[1], Jens C. Ø. Andersen[2], Olivier Namur[3], and Troels F.D. Nielsen[4]

1. Department of Earth Sciences, University of Cambridge, Downing Street, Cambridge CB2 3EQ, UK. Email: marian@esc.cam.ac.uk

2. Camborne School of Mines, University of Exeter, Penryn Campus, Tremough, Penryn TR10 9FE, UK

3. Department of Earth and Environmental Sciences, KU Leuven, Celestijnlaan 200E, 3001 Leuven, Belgium

4. Department for Mapping and Mineral Resources, Geological Survey of Denmark and Greenland, Øster Voldgade 10, 1350 Copenhagen K, Denmark

November 2023

Abstract

The roof-derived autoliths in the floor cumulates of the Skaergaard Intrusion have been argued to have been extensively metasomatized and recrystallised, forming the foundation of the hypothesis that microstructures in plutonic rocks are essentially metamorphic. However, the augite-plagioclase-plagioclase dihedral angles and plagioclase core composition of the autoliths match with those of the roof rocks, demonstrating that they were generally solid on arrival at the floor, with no subsequent microstructural or compositional modification. Many autoliths have mafic rinds, which were used as evidence of metasomatism: these rinds fall into two groups. The rarely developed rind rock of Irvine *et al*. (1998) is most likely chilled magma infiltrating along fractures in the roof rocks, either associated directly with detachment of roof material, or occurring before final detachment. Thin mafic rims are widespread in LZc and MZ, present at the tops of the more elongate autoliths, with a corresponding felsic rim at the base of the most elongate. The close correspondence of thin rim development with autolith shape, rather than composition, is argued to be evidence that they formed as a result of differential migration of immiscible conjugate interstitial liquids: the dense Fe-rich liquid flowed downwards and ponded on the tops of impermeable autoliths, whereas its buoyant Si-rich conjugate flowed upwards and was trapped underneath. Any differences in microstructure and bulk composition of the autoliths compared to the remaining exposures of the roof sequence reflect the wider range of lithologies in the now-eroded regions of the roof.

INTRODUCTION

A number of studies have argued that the microstructures, and even compositions, of layered mafic intrusions are not primary, but have been significantly modified during cooling (e.g. McBirney & Hunter, 1995; McBirney 2009; Boudreau, 2011; Bartley *et al*. 2018; Glazner *et al*., 2020). A particular example used to support and develop this hypothesis is that of the abundant autoliths in the floor cumulates of the Skaergaard Intrusion of East Greenland, which, following Wager & Brown (1968), are universally agreed to have fallen from the inwards-solidifying intrusion roof. It has been argued that their mafic components have been extracted (by some unspecified process) to form melanocratic rinds around a relatively anorthositic centre, and that modal layering and foliation were imposed once they reached the intrusion floor (McBirney, 1979; 1987; 1995; 2009; McBirney & Sonnenthal, 1990; McBirney & Hunter, 1995), though Irvine (1987) and Irvine *et al*. (1998) argued against this interpretation.

Here we present the results of a detailed microstructural investigation of both autoliths and the remaining exposures of the intrusion roof, together with a consideration of plagioclase composition and larger-scale field observations. We also examine the melanocratic rinds, distinguishing between those described by Irvine *et al*. (1998) and attributed to the chilling of magma infiltrating along fractures in the roof sequence, and the localised development of Fe-Ti oxide-rich rinds. Our results support the contention of Irvine *et al.* (1998) that the cognate xenoliths were almost invariably entirely solid when detached from the roof and that no microstructural or compositional changes occurred after they arrived at the intrusion floor. We argue that the Fe-Ti oxide-rich rinds are bodies of an unmixed Fe-rich interstitial liquid that ponded on the tops of the impermeable autoliths. Instead of forming part of the foundation of the hypothesis that plutonic microstructures are generally extensively overprinted, both the compositions and microstructures in these autoliths are primary.

GEOLOGICAL SETTING

The Palaeogene (56.02 Ma: Wotzlaw *et al*., 2012) Skaergaard Intrusion, East Greenland (Figure 1), is associated with the opening of the North Atlantic Ocean. It comprises an 8 km x 11 km x 4 km body (Nielsen, 2004) of tholeiitic magma rapidly emplaced (Annen *et al*., 2022) in a fault-bounded opening at the unconformity between basement Archean gneisses and overlying Cretaceous-Palaeocene sediments and Paleogene flood basalts (Wager & Deer, 1939; McBirney, 1989a). It fractionated as a closed system once the chamber was filled. The intrusion is now tilted towards the south-east by ~20°, resulting in an almost complete section through the stratigraphy of the floor cumulates, accompanied by the loss of most of the intrusion roof (Wager & Brown, 1968).

The intrusion is divided into three main units: the Layered Series, Marginal Border Series (MBS) and Upper Border Series (UBS), which crystallized inwards from the floor, walls, and roof respectively and display correlated changes in liquidus assemblage. The Layered Series is subdivided into Lower, Middle and Upper Zones (LZ, MZ, and UZ, respectively), based on the absence of cumulus olivine in

the Middle Zone. The Lower Zone is further subdivided: LZa contains cumulus olivine and plagioclase; the base of LZb is marked by the arrival of cumulus augite; and the base of LZc is marked by the arrival of cumulus Fe-Ti oxides. Upper Zone is also subdivided; the base of UZb defines the arrival of cumulus apatite, while the base of UZc marks the first appearance of the mosaic form of ferro-hedenbergite inverted from β-ferrobustamite. The Layered Series is underlain by a quantity of unexposed floor cumulates termed the Hidden Zone (HZ) by Wager & Deer (1939), thought to be of the order of 150 m thick in the region of the Uttental Plateau, and with the same primocryst assemblage as LZa (Maaløe, 1976; Holness *et al*., 2007a; b; 2015).

The Upper Border Series

All remaining UBS exposures are at the extreme south of the intrusion (Figures 1, 2): the UBS is thought to have originally comprised ~1/6 of the total volume of the intrusion, with an average thickness of ~600 m (Nielsen, 2004). Its lower part has been intruded by the mafic Basistoppen sill, together with several minor granophyric sills (Figure 2). It was subdivided by Wager (1960) into Upper Border Group α, β and γ, with a hypothetical outermost portion assigned to Upper Border Group ultra-α (Figure 2), correlated with the Layered Series on the basis of plagioclase core composition (Wager & Brown, 1968; Figure 2). The first three subdivisions are equivalent to the Layered Series LZ, MZ and UZ, respectively, with ultra-α equivalent to HZ. The UBS contains much less olivine than the equivalent parts of the Layered Series, attributed by Wager & Deer (1939) to gravitationally driven loss of dense mafic grains, and by Wager & Brown (1968) to contamination by the felsic gneiss country rock. Apart from the fine-grained outermost regions, the UBS is commonly coarser-grained than the Layered Series: a similar relatively coarse grainsize of the MBS was attributed by Holness *et al*. (2022) to a predominance of *in situ* nucleation and this is likely also to apply to the UBS.

Because the UBS is richer overall in SiO_2, K_2O, P_2O_5 and incompatible trace elements compared to the Layered Series, Naslund (1984a) argued that it crystallised from a separate body of magma with a relatively early appearance of apatite primocrysts. He subdivided the UBS into an outer Upper Border Zone τ, followed by Upper Border Zones α (plagioclase, olivine and apatite primocrysts), β (plagioclase and Ca-rich pyroxene) and γ (all four minerals: Figure 2). However, re-examination by Salmonsen & Tegner (2013) confirmed the Wager (1960) subdivisions, now denoted LZ', MZ' and UZ': they differ from Wager (1960) in placing the boundary between LZ' and MZ' (Wager's α and β) at An_{50}, instead of An_{53}, with the Sandwich Horizon at An_{21} instead of An_{34} (though it should be noted that there is a variability of ~5 mol.% at any one stratigraphic level: Namur & Humphreys, 2018). They place the boundary between LZa' and LZb' at $An_{59\text{-}60}$, that between LZb' and LZc' at An_{52}, and that between LZc' and MZ' at An_{50}. The base of UZ' is placed at An_{44}, with the two boundaries between the subdivisions in UZ' at An_{40} (apatite-in) and $An_{32\pm2}$ (ferrobustamite-in). In this contribution, we use the subdivision scheme of Salmonsen & Tegner (2013) (Figure 2).

The inaccessibility of much of the terrain prevents detailed mapping of the UBS. However, the current exposure contains a clearly-defined major unconformity, with much of MZ' missing in the region of Brødretoppen (Wager & Brown, 1968; Figure 2). The complete UBS stratigraphy is only exposed on the eastern side (Salmonsen & Tegner, 2013; Figure 2).

Liquid immiscibility in the Skaergaard Intrusion

The Skaergaard mass-balanced liquid line of descent encountered the silicate liquid binode during fractionation, splitting into Fe-rich and Si-rich conjugates (McBirney & Nakamura, 1974; McBirney, 1975), at a composition equivalent to the bulk melt near the base of UZb (Nielsen *et al*., 2019). This is recorded by coexisting Si- and Fe-rich melt inclusions in primocryst apatite from UZb (Jakobsen *et al*., 2005), and the co-existence in the upper part of the Layered Series of grain-scale bodies of interstitial granophyre (formed from the solidification of the Si-rich conjugate) and interstitial intergrowths dominated by ilmenite, magnetite and clinopyroxene which represent the solidified Fe-rich conjugate (Holness *et al.,* 2011; Humphreys, 2011). On the grain scale, granophyric pockets are confined to plagioclase-rich regions whereas the ilmenite-rich pockets are surrounded by olivine and pyroxene (Holness *et al*., 2011), consistent with the different wetting properties of the immiscible conjugates (Honour *et al*., 2019).

The Si-rich liquid is buoyant and viscous whereas its Fe-rich conjugate is dense and relatively inviscid (Namur *et al*., 2015), creating the potential for differential migration. The systematic loss of an exsolved buoyant Si-rich liquid from the fractioning interstitial liquid in the mush on the chamber floor, and its addition to the overlying bulk magma, is argued to have resulted in an overall Fe-enrichment of the Layered Series from LZc upwards, with a corresponding Si-enrichment in UBS (Nielsen *et al.,* 2019). Outcrop-scale evidence of differential migration of unmixed immiscible conjugates in the mush is provided by paired, irregularly shaped, felsic and mafic bodies of late-stage segregated liquid (Figure 3; McBirney, 1975; Sonnenthal, 1992; Larsen & Brooks, 1994). In all cases, the mafic component (comprising olivine pyroxenite, with variable amounts of Fe-Ti oxides) occurs beneath the associated felsic material, either forming the lower part of a single segregation or a separate body. The paired felsic and mafic bodies are closely spatially associated with coarse-grained pegmatitic segregations that occur through much of the Layered Series (Wager & Deer, 1939; Wager & Brown, 1968; McBirney & Noyes, 1979; McBirney & Sonnenthal, 1990; Larsen & Brooks, 1994). These pegmatites are rounded and diapir-like in the lower parts of the Layered Series, but comprise layer-parallel sills and sharply defined dykes above the upper parts of MZ, attributable to a reduction in the porosity of the mush (Larsen & Brooks, 1994), and consistent with a reduction in mush thickness from ~100m in LZ to a few metres at the UZa/b boundary (Holness *et al*., 2017a).

Grain-scale evidence of pervasive differential migration of the immiscible liquids in the mush is provided by the presence of two fractionation trends in intercumulus plagioclase (Humphreys, 2011), attesting to a breakdown of chemical communication between the two immiscible conjugates. It is also

manifest in LZc and MZ by the development of reactive symplectites formed of anorthitic plagioclase and olivine (or pyroxene) rooted on Fe-Ti oxide grains and replacing the surrounding primocrysts of plagioclase and (less commonly) olivine and augite (the Type I symplectites of Holness *et al*., 2011). In LZc and MZ, Type I symplectites are accompanied by olivine rims separating Fe-Ti oxides from plagioclase and augite primocrysts (Figure 4a; Wager & Deer, 1939; Nwe, 1976; Nielsen *et al*., 2015). They are monocrystalline in rocks containing olivine primocrysts (primarily in LZc), with the rims extending from nearby grains in which they are in optical continuity. Where primocryst olivine is absent (i.e. MZ) the rims are polycrystalline (Holness *et al.*, 2011). The oxide-rich gabbros in LZc and MZ, containing highly irregular and extensive aggregates of Fe-Ti oxides, are notable for the recrystallisation of clinopyroxene primocrysts to form a mosaic of small grains (Figure 4): this combination of reactive symplectites, abundant anhedral oxides, olivine rims and recrystallised pyroxene, known as Paragenesis 2, is argued to result from the crystallisation of ponded Fe-rich interstitial liquid out of chemical equilibrium with the earlier-formed liquidus paragenesis (Nielsen *et al.,* 2015).

ROOF-DERIVED AUTOLITHS IN THE LAYERED SERIES

The MBS and the lowest exposed parts of the Layered Series contain xenoliths of both gneissic (Markl, 2005; Figure 5a) and basaltic country rock, together with picritic xenoliths (Figure 5b) thought to be sourced from a nearby wehrlite plug (Kays & McBirney, 1982). Some xenoliths from the overlying plateau basalts forming the roof of the intrusion (Figure 5c) are present in MZ near the south of Kraemer Ø (Irvine *et al*., 1998), but the predominant type of blocks in the Layered Series is material derived from the UBS (Figure 5d).

The roof-derived autoliths occur from the top of LZa to the middle of UZb, and are particularly abundant in LZb, LZc and the lower half of MZ (Figure 1). They concentrate in stratigraphic zones 50–100 m thick, extending laterally up to 1 km, and range in size from a few 10's of cm up to several hundreds of metres (Irvine *et al.,* 1998). They commonly appear leucocratic relative to the surrounding gabbroic cumulates (Figure 5d), with a variable shape, though most are slab-like and oriented parallel to the layering in the host gabbro, with sharply-defined margins on the grain-scale (Figure 6a). Irvine *et al.* (1998) describe an autolith in UZa with a cleanly broken upper surface, but an irregular lower margin, suggestive of incomplete solidification on arrival at the floor. The grain-size in the autoliths is generally different to that of their host, being commonly coarser-grained.

Following Wager & Brown (1968; p.136), who argued that many of the autoliths found in MZ on Kraemer Ø are derived from LZ', Irvine *et al*. (1998) divided the autoliths into three types: gabbroic troctolite; gabbroic anorthosite; and oxide gabbro. Gabbroic troctolite (scans of thin sections of typical examples are provided in the Supplementary Appendix) is a plagioclase cumulate with minor amounts of olivine (which may be poikilitic) and interstitial augite: autoliths of this lithology appear in LZa and are found at all levels up to UZb. On the basis of plagioclase composition, Irvine *et al*. (1998) assign

this lithology to the UBS equivalent of HZ, with some overlap with the LZa equivalent. Gabbroic anorthosite is relatively uncommon and is dominated by plagioclase primocrysts with sub-poikilitic augite: olivine is absent (scans provided in the Supplementary Appendix). Gabbroic anorthosite autoliths are found in LZb and MZ: Irvine *et al.* (1998) assign this lithology to the UBS equivalent of LZb. Oxide gabbro is generally coarse-grained relative both to the other autolith types and to the gabbros of the Layered Series, and contains primocrysts of plagioclase, augite, and Fe-Ti oxide (scans provided in the Supplementary Appendix). Small amounts of olivine may be present. Oxide gabbro autoliths are found only in the lower two-thirds of MZ, and are assigned by Irvine *et al.* (1998) to the UBS equivalent of LZc, although the plagioclase composition also overlaps with that of MZ* (in the MBS).

Irvine *et al.* (1998) report weak concentrations of Fe-Ti oxides, only 1–2 mm thick, along the top surfaces of many small autoliths in MZ (an example is shown in their Figure 24D). They also report that a small number of the medium-sized to large autoliths of all three different compositional categories are bordered by pale brown weathering, fine-grained, mafic to ultramafic rinds 5–15 cm thick, known as rind rock, dominated by augite, plagioclase and Fe-Ti oxides, with rare olivine (Figure 7). Rind rock has a well-defined and sharp contact against the autolith, but a gradational contact with the gabbroic host: grainsize increases from the autolith contact into the gabbroic host. Rind rock also forms small dykes in some of the larger autoliths, some of which have a coarser-grained central part: the dykes are truncated at the autolith edges. The main occurrences of this relatively rare phenomenon are in LZc and the lower half of MZ on Kraemer Ø: no rinds were recognized on stratigraphically higher autoliths. Irvine *et al*. (1998) suggest rind rock represents injections of magma into fractures that developed during disintegration of the UBS, with the large size of the autoliths on which they are found contributing to a rapid chilling and consequent fine grainsize. They further suggest there might have been some reaction between the injecting magma and the autolith.

THE HYPOTHESIS OF METASOMATIC REPLACEMENT OF THE AUTOLITHS

The first suggestion that autoliths found in the Layered Series had undergone metasomatic replacement following their arrival at the floor was made by McBirney (1979). He argued that a magnetite-rich layer surrounding a basaltic xenolith in MZ (Figure 5c) was formed by the removal of mafic components from the autolith, together with migration of mafic material from the surrounding gabbro. He suggested this metasomatism was a consequence of a difference in the oxidation states between the autolith and the liquid on the floor. No mechanism was proposed, and the only detailed work demonstrating changes in bulk composition or mineralogy was reported by Irvine (1979), who found evidence of only limited interaction with the adjoining cumulates.

The concept of metasomatic alteration was expanded by McBirney (1987) to include the roof-derived autoliths, based on the assumption that their relatively felsic nature would have prevented them sinking to the floor (Naslund, 1986), together with their more anorthitic plagioclase, and lower bulk Fe, Ti, K

and P contents compared their putative source in the UBS. It was argued that metasomatism was accompanied by wholesale recrystallisation resulting in propagation of layering present in the surrounding gabbros into the autoliths. McBirney (1989a) further developed the metasomatic hypothesis by suggesting that the mafic components removed from the autoliths form iron-rich selvages and paired bodies of anorthosite and mafic material (e.g. Figure 3), in a similar manner to his suggestion for the basaltic xenolith (McBirney, 1979). No detailed arguments were provided for this process, and it is not clear whether the iron-rich selvages are equivalent to the Irvine *et al.* (1998) rind rock.

These ideas were expanded by McBirney & Hunter (1995), who argued that microstructures in layered intrusions are essentially metamorphic, with wholesale recrystallisation caused by "an anisotropy in the environment in which the crystals grew". McBirney (2009) strengthened his earlier statements about the felsic nature of the autoliths by saying that metasomatism resulted in them becoming almost pure anorthosite, with the mechanism for this extreme version of the metasomatic hypothesis argued to be re-melting of extensively hydrothermally altered autoliths once they arrived at the floor, with their mafic components carried into the surrounding gabbro by volatile-rich liquid. A quantitative model of the metasomatic process was presented by Sonnenthal & McBirney (1998), involving diffusion in a melt phase.

Dissent was expressed by Irvine (1987), who pointed out that not only does layering *not* propagate into the autoliths from the surrounding gabbros, but there is no evidence of any intermediate stages of metasomatic replacement. Autoliths have sharply defined margins (e.g. Figure 6a), and some contain well-defined internal boundaries between units of different composition that are at a different orientation to the layering in the host gabbros. The autoliths generally have plagioclase core compositions different from that of the cumulus plagioclase of their host (Irvine *et al.,* 1998). Furthermore, differences of $\delta^{18}O$ between the autoliths and their host (Taylor & Forrester, 1979) precludes metasomatism after they left their source at the roof. The only exceptions to this are some of the very smallest autoliths, which develop a centimetric-scale diffuse feldspathic margin: Irvine *et al*. (1998) suggest this very localised replacement was perhaps facilitated by volatile-rich fluids.

CHOICE OF SAMPLES AND ANALYTICAL METHODS

Samples

We examined a suite of samples collected from the UBS by Lawrence Wager and Alex Deer during the 1935-6 expedition, with further UBS samples from the Exchange Collection collected by Alex Deer in 1966 (locations are shown in Figure 2), all housed in the Harker Collection of the Sedgwick Museum, University of Cambridge. We examined autoliths from Uttental Plateau and Kraemer Ø (locations shown in Figure 1). We also examined the autolith in UZa on the Skaergaard Peninsula interpreted by Irvine *et al.* (1998: p. 1414) as having been only partially solidified on arrival at the floor (samples

SK11-B9-9 and -10). Most autoliths were sampled in their centres, but a small sub-set were also sampled at their margin. Samples were also taken of the host gabbro within a few metres of some autoliths. Examples of rind rock were collected from the autoliths illustrated by Irvine *et al*. (1998) in their Figures 17a, 21b, 22 and 23.

The geometry of three-grain junctions

In most gabbros and dolerites, the geometry of three-grain junctions involving two grains of plagioclase and one of augite, parameterised by the median value of the clinopyroxene-plagioclase-plagioclase dihedral angle population, Θ_{cpp}, is lower than the equilibrium value of 109˚ ± 2˚, due to a cooling rate control on the relative rates of growth of the two minerals during solidification (Holness *et al.,* 2012a; Fowler & Holness, 2022). In layered intrusions, Θ_{cpp} changes in a step-wise manner due to changes in the contribution of the latent heat of crystallisation to the overall enthalpy budget on the arrival of a new liquidus phase (Morse, 2011; Holness *et al.,* 2013; Holness, 2015). In the Skaergaard intrusion, there is a step-wise increase in Θ_{cpp} corresponding to the arrival of liquidus augite (the LZa/b boundary), another at the arrival of liquidus Fe-Ti oxides (the LZb/c boundary) and a third corresponding to the arrival of liquidus apatite (the UZa/b boundary).

True, 3D, clinopyroxene–plagioclase–plagioclase dihedral angles were measured using a four-axis Leitz universal stage mounted on a James Swift monocular microscope, with a UM32 Leitz long working distance objective and a x10 eyepiece. The median value of a population of angles can be determined satisfactorily with only 25 measurements (Riegger & van Vlack, 1960), although reduction of the 95% confidence intervals about the median to less than ± 3-4˚ [calculated according to the method of Stickels & Hücke (1964)] generally requires more than 25 (Holness, 2010). We report the median values of populations of 25 - 70 individual measurements of true 3D dihedral angles in each sample.

Plagioclase compositions

The major element composition (Na, Si, Al, K, Ca and Fe) of plagioclase was determined using either a Cameca SX100 electron microprobe at the Department of Earth Sciences, University of Cambridge, or a JEOL JXA-8600 Superprobe at the Geologisk Institut, Aarhus University. Analysis at both Cambridge and Aarhus used a defocused (5μm) beam, with accelerating voltage of 15 keV and a 10nA beam current. Peak count times were 60 s for Fe and 10 s for all other elements, with the total background count times equivalent to the on-peak time for each element. Primary calibrations were undertaken using appropriate mineral standards and measurements were internally calibrated using a Smithsonian Microbeam Standard (Labradorite 115900) to ensure consistency across multiple analytical sessions. Totals <98 wt.% and >101 wt.% were discarded. Analytical uncertainties were estimated through repeat analysis of Smithsonian standards, with major (>5wt%) and minor (>0.5 wt.%) element 1σ precisions better than 0.6 wt.% and 0.1 wt.% respectively. The average composition

of plagioclase was determined in Cambridge by measuring 3 spots in the centres of 6 – 9 grains: analyses that were significantly more albitic than the rest of the data points were discarded on the assumption that an evolved outer rim had inadvertently been analysed. Plagioclase core compositions were determined in Aarhus from 2-5 grains in each thin section.

NEW OBSERVATIONS

Microstructures in the UBS

The outermost parts of the UBS (HZ' and LZa': scans of representative thin sections are provided in the Supplementary Appendix) contain abundant, randomly oriented, tabular plagioclase primocrysts with large, unzoned cores. In common with much of the rest of UBS, plagioclase is randomly oriented, with irregular grain boundaries indicative of *in situ* growth to impingement (Figure 8a): isolated strongly laminated horizons occur in LZ', defined by thin (0.3 mm) plagioclase plates up to 1 cm across (Plate 22 of Wager & Deer, 1939). Olivine commonly forms small subhedral grains but may also be ophitic (Figure 8b). Variable amounts of interstitial augite are present, with an inwards increase in mode. In HZ' and LZa', augite is commonly poikilitic (Figure 8b), and the median value of the clinopyroxene-plagioclase-plagioclase dihedral angle, Θ_{cpp}, is 84.5 – 86˚ (data are summarised in Table 1, with full data-set in the Supplementary Appendix). A sample of LZa equivalent from the topmost part of the MBS on the east margin of the intrusion (at Udløberen, Figure 2) also has the same value of Θ_{cpp} (Table 1).

The change from interstitial to primocryst clinopyroxene (the LZa'/LZb' boundary) is marked by a reduction in the number of plagioclase inclusions, a progression to a subhedral or ophitic habit (scans of representative thin sections are provided in the Supplementary Appendix), and the appearance of simple twins parallel to the long axis, indicative of a primocrystic origin (Holness *et al*., 2022). This is associated with a step-wise increase in the median value of Θ_{cpp} to 89.5 – 91˚: this is also observed in a sample of LZb equivalent from the topmost part of the MBS on the east margin of the intrusion (at Udløberen, Figure 2, Table 1, Supplementary Appendix).

Fe-Ti oxides are anhedral through much of the UBS, commonly with a skeletal or dendritic habit (scans of thin sections provided in the Supplementary Appendix). The LZb'/LZc' boundary is marked by an increase in oxide mode, with abundant small euhedral grains of both magnetite and ilmenite (Figure 8e). This is associated with a decrease in grainsize, increase of clustering, and a less compact habit of the augite, and a second step-wise increase in Θ_{cpp} to 94 – 94.5˚.

In the more evolved parts of the UBS, augite reverts to being highly anhedral (Figure 8c), with some dendritic grains (Figure 8d). Dihedral angles were not measured in UZ', as the abundance of granophyre and the highly anhedral augite habit means there are no suitable three-grain junctions. Euhedral grains of apatite are present, and locally abundant, throughout UBS (Naslund, 1984a),

particularly in pockets of granophyre, though in the more evolved parts it commonly has a hopper or skeletal habit. The boundary between UZa' and UZb' is marked by an increase in apatite mode and the appearance of apatite inclusions in primocrysts of augite and plagioclase. The lowermost horizons of the UBS contain polycrystalline aggregates of hedenbergite, interpreted as inverted β-ferrobustamite, consistent with them belonging to UZc (Salmonsen & Tegner, 2013)'.

Although the Paragenesis 2 olivine rims separating Fe-Ti oxides from plagioclase and augite primocrysts that are a prominent feature of MZ (Figure 4a) are present in the equivalent subdivision of the MBS (Holness *et al*., 2022), they are absent in UBS. Type I symplectites (Figure 4a) are rare and confined to the outermost parts of the UBS.

Interstitial quartz and granophyre are rare in HZ' and LZa', but increase through LZb', becoming abundant in LZc', MZ' and UZ'. On the grain scale, both quartz and granophyric intergrowths form interstitial bodies typically bounded by the randomly oriented tabular plagioclase grains, and commonly associated with chlorite (Figure 8f): locally, granophyre is so abundant that the euhedral plagioclase grains are entirely surrounded. The quartz component of the granophyre may form thin plates, inverted from primary tridymite due to the low lithostatic pressure (Figure 8d; Lindsley *et al*., 1969). Late-stage hydrothermal alteration is widespread.

Evidence of separation of mafic and felsic components in the Layered Series

The upper part of LZc is notable for a series of highly oxide-rich modally-graded layers, ranging in thickness from ~10 cm to ~ 30 cm (Weatherley *et al*., 2018). Locally, the sharply-defined base of these layers is underlain by a felsic layer up to 2 cm thick (Figure 9), containing plagioclase of a similar grainsize to that in the underlying gabbros, but with finer-grained augite and Fe-Ti oxides (Figure 6b). While the oxide grains in the felsic layer are associated with Type I reactive symplectites, there are no olivine rims separating oxide from adjacent pyroxene, although such rims are present both below and above the layers.

Laterally extensive, essentially planar, modally-graded layers are common in UZa and UZb, separated by up to ~20 m of homogeneous gabbro. These layers are generally 10-20 cm thick and grade from a mafic base rich in pyroxene, olivine and Fe-Ti oxides to a top rich in plagioclase. A previously unremarked feature of these modally-graded layers is that although their base is generally sharply defined, their tops are highly irregular (Figures 10a, b, c, d). The extent of the irregularity increases steadily upwards in the stratigraphy, with the felsic parts of the layers in UZb almost completely detached from the underlying layer, to form irregular wisps and blebs (Figures 10c, d).

Several reverse modally-graded layers are present in the lower part of UZa (McBirney & Noyes, 1979: Figure 10e), in the region of the intrusion in which the late-stage pegmatites form well-defined sills and dykes (illustrated in Figure 5 of Larsen & Brooks, 1994), indicative of a thin and rigid mush zone. The

reverse modally-graded layers extend laterally over ~ 100 m, and have sharply-defined mafic tops and grade downwards over 10-15 cm to more average gabbroic compositions. Locally, small and irregular anorthositic bodies lie immediately above the tops of the layers (Figure 10e). The mafic tops of these layers are dominated by Paragenesis 2 of Nielsen *et al.* (2015) (Figure 4).

The autoliths

The gabbroic troctolites (assigned to HZ' and LZa' by Irvine *et al*. (1998)) are dominated by randomly oriented plagioclase grains with irregular grain boundaries characteristic of *in situ* growth to impingement (Holness *et al*., 2019) (Figure 11a). Plagioclase core compositions are in the range An_{70} – An_{59} (Table 2). Pyroxene is abundant, comprising interstitial (commonly poikilitic, generally untwinned) augite and inverted pigeonite. Olivine has a wide range of habits and modes: where rare, it forms small rounded grains (Figure 11b, commonly enclosed by pyroxene), or rims separating augite from plagioclase (Figure 11c), but where common, it forms either well-sintered clusters of subhedral grains, or oikocrysts enclosing plagioclase chadacrysts. Representative bulk compositions are provided in Table 3. We found one autolith, hosted in LZc, in which the olivine grains were surrounded by reactive symplectites of orthopyroxene and plagioclase (Figure 11d) – the Type II symplectites of Holness *et al*. (2011), ascribed to reaction between olivine and a Si-rich interstitial liquid. Fe-Ti oxides form anhedral aggregates and oikocrysts that, in some autoliths, are associated with Type I symplectites (Figure 11e). Pockets of interstitial granophyre are rare. Values of Θ_{cpp} in the gabbroic troctolites are in the range 83.5 – 85.5˚ (Table 2, with full dataset in Supplementary Appendix).

The gabbroic anorthosites (assigned to LZb': Irvine *et al.* 1998; representative bulk compositions in Table 3) are coarser-grained than the gabbroic troctolites (see thin section scans in the Supplementary Appendix, and are characterised by euhedral randomly oriented plagioclase (with core compositions in the range An_{63} – An_{52}: Table 2), with augite forming large, commonly ophitic grains, many of which have simple (100) twins (Figure 11f). Fe-Ti oxides form anhedral interstitial aggregates, rarely surrounded by Type I symplectites. Olivine forms rare small, anhedral and sometimes interstitial, grains. Quartz is interstitial, locally abundant, and commonly in granophyric intergrowths. Granophyric pockets are always bounded by plagioclase, and variously contain green amphibole, chlorite, apatite and brown mica. Values of Θ_{cpp} in the gabbroic anorthosites are in the range 89 – 91˚ (Table 2, Supplementary Appendix). The UZa-hosted autolith suggested by Irvine *et al*. (1998) to have been only partially solidified on arrival at the floor, shown in Figure 12, is a gabbroic anorthosite. It has well-defined top and side margins: the value of Θ_{cpp} is the same within 1 cm of the top of the autolith margin (sample SK11-B9-10) as it is in the autolith centre (sample SK11-B9-9) (Table 2).

The oxide gabbros (assigned to LZc' and MZ') are almost invariably coarse-grained relative to the gabbroic anorthosites (scans of thin sections in the Supplementary Appendix), with primocrysts of Fe-Ti oxides, augite (commonly with simple (100) twins: Figure 11f) and randomly oriented plagioclase with core compositions in the range An_{52} – An_{50} (Table 2). Rare grains of olivine are present as

inclusions in oxide or augite. Fe-Ti oxides form anhedral aggregates, commonly with extensive marginal interstitial growth, although small grains included in augite or plagioclase are euhedral. Interstitial pockets in plagioclase-rich regions are filled with single grains of quartz (some of which extend in optical continuity through large regions of the thin section) or granophyric intergrowths. Brown mica and amphibole are minor constituents of these quartz-rich interstitial pockets. A representative bulk composition is provided in Table 3. Type I symplectites are common in regions where there is no interstitial quartz or granophyre. The olivine rims between oxides and pyroxene that are a ubiquitous feature of LZc and MZ are only very rarely present, even within ~1 mm of the autolith margin. Values of Θ_{cpp} in the oxide gabbros are in the range 95 – 97˚ (Table 2, Supplementary Appendix).

We examined several samples from an elongate, composite autolith hosted by MZ gabbro. Autolith SK11-B8 (location shown in Figure 1) comprises three layers parallel to the long axis of the outcropping surface, the upper two of which are oxide-rich gabbro, while the base comprises a gabbroic troctolite. The value of Θ_{cpp} in the upper two layers is 95 – 95.5˚, with a corresponding plagioclase composition of An_{51}, whereas that in the lower gabbroic troctolite is 85 ± 2˚ (with plagioclase of An_{59}) (Table 2, Supplementary Appendix).

Rims, rinds and segregations associated with the autoliths

Paired segregations and their relationship with autoliths

Following the assertion by McBirney & Sonnenthal (1990) that the paired mafic and felsic segregations were formed as a consequence of metasomatism, we focussed on the spatial relationships between these features and the autoliths. Examples of paired mafic and felsic segregations in close proximity to autoliths are not common, but where present invariably display the same spatial relationships as those in the host gabbros (e.g. Figure 3). Figure 13a shows an example of a large gabbroic anorthosite autolith in LZb on Uttental Plateau, which is immediately underlain by large irregular bodies which are felsic at the top, with a sharp contact with a lower mafic portion. The mafic material forms a highly irregular network extending some metres below the autolith (c.f. Figure 3), and the felsic portion locally penetrates the autolith on irregular fractures.

Rind rock

A sample of the rind rock dykes (JC-107) cutting the gabbroic troctolite autolith shown in Figures 22 and 21B of Irvine *et al*. (1998) is fine-grained, with an almost granular microstructure (Figures 14a, b). It is dominated by augite and plagioclase, which is anhedral, randomly oriented and with low aspect ratio. The value of Θ_{cpp} is 105 ± 2˚ (Supplementary Appendix). Fe-Ti oxides form anhedral grains and

patches commonly separated from the adjacent clinopyroxene by rims of olivine associated with reactive symplectites that replace adjacent plagioclase (Figure 14c). The example of rind rock (JC-100) shown in Figure 17a of Irvine *et al.* (1998), associated with an oxide-rich gabbroic autolith found in MZ, contains not only very well-developed reactive symplectites (Figure 14d), but also small and rare interstitial pockets of quartz, surrounded by strongly normally zoned plagioclase.

The group of autoliths illustrated in Figure 23 of Irvine *et al*. (1998) is located in LZc (Figure 7), and belongs to the gabbroic anorthosite category of Irvine *et al*. (1998). A sample of rind rock (JC-59, location shown in Figure 7b) is generally fine-grained though with variable grainsize. The finest-grained parts, with an almost granular microstructure, contain abundant prismatic ilmenite with a preferred grain alignment (Figure 14e), with minor apatite forming rounded grains enclosed by plagioclase. The coarser-grained parts contain anhedral olivine with abundant rounded inclusions of plagioclase and Fe-Ti oxides (Figure 14f). Again, Θ_{cpp} is 105 ± 2˚ (Supplementary Appendix).

Mafic and felsic rims

Mafic rims, a few mm in thickness and rich in Fe-Ti oxides, occur at the tops of autoliths from LZc upwards (e.g. Figure 12a). We undertook a detailed examination of the autolith population in a ~200 m x 200 m area of LZc on Kraemer Ø. The majority of the autoliths are oxide gabbro, and range from ~10 cm to ~20 m on the longest axis. Their apparent shapes on the outcrop surface range from rounded (with an aspect ratio of ~1) to highly elongate parallel to the layering in the host gabbro (Figure 13). The rounded autoliths, regardless of their size, do not have mafic rims (Figure 13b). Mafic rims are present on the more elongate autoliths, forming an oxide-rich layer up to 1 cm thick (Figures 13c, d), found only on their top surface. The mafic rims comprise rounded pyroxene grains (generally smaller than those in the autolith and the overlying gabbro), with subordinate anhedral plagioclase, set in an interconnected aggregate of Fe-Ti oxides (Figure 6c). Well-developed polycrystalline olivine rims separate the Fe-Ti oxides from the pyroxene (similar to those shown in Figure 4), and Type I symplectites are common.

A previously undocumented feature is that the most tabular autoliths have not only an upper mafic rim, but also have a corresponding felsic rim (generally < 1 cm thick) along their base (Figures 13e, f). These felsic rims are similar to the felsic material underlying the highly oxide-rich layers in the host gabbro (Figure 9), in that they contain fine-grained pyroxene and Fe-Ti oxides, together with coarse-grained plagioclase, the Fe-Ti oxides are associated with Type I symplectites, and no olivine rims separate oxides from adjacent pyroxene (Figure 6d).

There is a systematic relationship between the (apparent) autolith shape and the presence/absence of the rims. There is a gradation from autoliths with no rims, through autoliths with an upper mafic rim, to autoliths with a paired upper mafic and lower felsic rim as the (apparent) aspect ratio of the autolith

increases (Figure 15). There is no relationship between rim thickness or composition with either size or composition of the autoliths.

DISCUSSION

Compositional evidence used to support the metasomatic hypothesis

McBirney (1989a) provides the average composition of 20 autoliths (no descriptions or localities are provided) and of 13 samples from UBS-β (MZ') "from which most are thought to have been derived" (*ibid*.): these are given in Table 4 together with a further average composition for the autoliths from Sonnenthal & McBirney (1998) (termed anorthositic blocks in their Table 2). The averaged autolith compositions from these two studies are comparable to those of the gabbroic troctolites shown in Table 3: there is no evidence that McBirney (1989a) and Sonnenthal & McBirney (1998) considered autoliths with compositions typical of the oxide-rich gabbros (e.g. Table 3). Their averaged compositions are richer in the components of anorthitic plagioclase and depleted in iron, TiO_2, K_2O and P_2O_5 compared to published averages of MZ' composition (UBS-β) (McBirney, 1989a; Naslund, 1984; Salmonsen & Tegner, 2013), and are most similar to the Salmonsen & Tegner (2013) averages of LZa' and LZb'. The CIPW norm of the average autolith compositions given by McBirney (1989a) and Sonnenthal & McBirney (1998) is richer in plagioclase than that of any of the average UBS compositions (Table 4). The average composition of plagioclase cores in the autoliths analysed by McBirney (1989a) is $An_{56.2}$. According to the position of the subdivision boundary of Salmonsen & Tegner (2013), Namur *et al*. (2014) and Namur & Humphreys (2018), such an average plagioclase composition is consistent with a source in LZb'.

The case for the autoliths having been metasomatized appears to rest entirely on the assumption that they were derived from MZ', perhaps on the basis that much of MZ' is absent from the current exposure. However, the average bulk compositions and plagioclase core compositions presented by the McBirney studies are consistent with a source in LZb'.

Where *did* the autoliths come from?

Importantly, the preserved exposure of UBS is limited. Irvine *et al.* (1998) recognised this and suggested that some of the more anorthositic autoliths may have been sourced from a now-eroded part of the roof. That many autoliths were derived from a different part of the roof than that now remaining (and hence that the large missing section of MZ' cannot be used as evidence that they were predominantly derived from MZ') is incontrovertibly demonstrated by the presence of xenoliths of the basaltic country rock in upper MZ on Kraemer Ø, requiring the complete removal of the UBS from a

now-eroded region of the roof (Figure 5c; Irvine *et al*., 1998). The remaining part of the UBS is clearly not representative of the whole.

The corollary of this evidence of localised excavation of the entire thickness of the contemporaneous UBS is that LZ' must have been the source of at least some of the autoliths in the Layered Series (in agreement with Wager & Brown (1968) and Irvine *et al.* (1998)). Both average bulk compositions and plagioclase composition in the previously published studies are consistent with those autoliths being sourced from LZb'. The plagioclase compositions reported here confirm the assignment of the three lithologies of Irvine *et al.* (1998) to HZ'/LZa, LZb' and LZc'/MZ', supported by the close correspondence of Θ_{cpp} between the autoliths and UBS.

Evidence for recrystallisation: primary vs secondary microstructures

In agreement with Irvine *et al.* (1998), we argue that the wholesale crystallisation advocated by McBirney & Hunter (1995) cannot have taken place and that the autoliths retain not only their original composition but also their original microstructures. Further evidence for retention of their original microstructures is provided by the observation that most have sharply-defined margins on the grain-scale (Irvine *et al*., 1998; Figure 6a). This sharply-defined margin is manifest also by the absence of olivine rims separating Fe-Ti oxide from pyroxene in the oxide-rich gabbroic autoliths hosted in LZc and MZ, with the change occurring over length-scales commensurate with the grainsize, together with the different values of Θ_{cpp} maintained within a few mm of the margin (observed for the top of autolith SK11-B9: Table 2). These sharply-defined margins demonstrate that the autoliths were generally entirely solid on arrival at the floor (exceptions to this include SK11-B9, Figure 12). Importantly, wholesale metasomatism necessitates the blocks to be melt-bearing, to permit significant mass transport through autoliths many 10's of metres in size (Sonnenthal & McBirney, 1998). Therefore, were the autoliths metasomatized, not only would we need to invoke re-melting, but the Θ_{cpp} of the autolith would be indistinguishable from that of its host since the cooling rate during the last stages of solidification of a metasomatized autolith would be the same as that of the host gabbro.

A further argument against the metasomatism hypothesis is that the mafic rinds, argued by McBirney to represent the mafic material removed from the autoliths, are not ubiquitous. The rind rock of Irvine *et al.* (1998) is only present on a small number of autoliths, and even where present, does not fully surround them. Furthermore, the truncation at autolith margins of dykes formed by rind rock (Irvine *et al*., 1998) is inconsistent with a metasomatic origin. The thin mafic rinds are not ubiquitous either and, where present, are confined to autolith top surfaces. Were they a result of metasomatism, they would be more prevalent, with their size and distribution correlating with autolith composition rather than autolith shape.

Rind rock: magma chilled entering fractures in the UBS

The fine grainsize of the rind rock is consistent with the high nucleation rates expected for rapid chilling and crystallisation due to magma encountering a relatively cold environment, Importantly, the rarity of autoliths associated with rind rock demonstrates the unusual conditions required to form it. The rapid chilling cannot have been due to the quenching of bulk magma around a cold, sinking, fragment broken away from the UBS, since such an origin would result in most autoliths having a complete rind. Furthermore, it would be difficult to account for the rind rock dykes by such a mechanism. We therefore agree with Irvine *et al.* (1998), that it formed by the injection of bulk magma into opening fractures in the UBS, most likely created during ongoing seismic activity, the most vigorous episodes of which were responsible for the concentration of the autoliths in certain stratigraphic horizons (and localised collapse of the MBS: Holness *et al*., 2022). The presence of rind rock on the surface of some autoliths would then be because the filled and solidified fractures represented a mechanical heterogeneity which became the site for some of the subsequent fractures resulting in complete detachment of material from the UBS.

The almost granular microstructure of the rind rock, with Θ_{cpp} close to the expected equilibrium value, is very different to that found in dolerites and is indicative of significant subsolidus textural modification. A similar microstructure is found in other examples of chilled magma in large plutons, such as the base of layers formed in the Kap Edvard Holm intrusion formed by successive batches of incoming magma that flowed along the existing chamber floor (Holness *et al*., 2012b). The chilled margin of the Skaergaard intrusion itself provides another example of a well-developed granular microstructure. The close approach to subsolidus textural equilibrium in the Skaergaard chilled margin was likely due to its early crystallisation and extended time at high temperatures, whereas the more limited subsolidus equilibration of the rind rock is most likely due to a relatively high temperature at which the UBS fractured, with slow subsolidus cooling caused by most of the heat from the intrusion passing through the roof. That high values of Θ_{cpp} are not seen in either the UBS or the autoliths is both because a fine grainsize, and consequent short diffusion distances, together with a low initial value of Θ_{cpp}, and hence large departure from equilibrium, are critical for significant subsolidus textural modification (Holness *et al*., 2012b).

Rims form by differential migration of immiscible liquids in the floor cumulates

In contrast to the rarity of rind rock, the mafic/felsic rims are widely developed around autoliths in LZc and MZ (allowing for constraints due to autolith shape: Figure 15). Given the possibility of variations in the events preceding the arrival of the autoliths at the floor, these rims must have formed after they arrived. Building on our new observations and following Holness *et al*. (2011) and Nielsen *et al*. (2015), we suggest that they are a consequence of differential migration of immiscible liquids in the crystal mush on the floor.

Outcrop-scale evidence for differential migration within the floor mush is provided by the felsic material at the base of modally-graded layers (Figure 9), and the wispy tops of modally-graded layers (Figure

10), both of which we argue are formed by the upwards-flow of the buoyant silica-rich conjugate. This liquid gets trapped under the Fe-rich base of the overlying material (Figure 9) since the permeability of the mush is not only a function of the interconnectivity of the porosity, but also of the minerals making up the mush. Capillary forces due to the non-wetting nature of Si-rich liquid against the predominantly mafic minerals at the base of the overlying modally-graded layer (Honour *et al*., 2019; Vukmanovic *et al*., 2018) would prevent further rise. In contrast, there is nothing preventing the upwards rise of felsic liquid from the modally graded layers shown in Figure 10 into the overlying, relatively plagioclase-rich, later-accumulating mush. We suggest the increasing separation of the felsic material with increasing stratigraphic height (compare Figure 10a with Figures 10c and d) is a consequence of greater mobility of this separating Si-rich conjugate as it forms an larger component of the progressively more fractionated interstitial liquid.

We suggest that the reverse modally-graded layers (Figure 10e), with their microstructure characteristic of bodies of Fe-rich liquid (Figure 4; Nielsen *et al.,* 2015), are solidified bodies of Fe-rich liquid which ponded on the top of the thin mush on the chamber floor: the low permeability of the mush and the abundance of plagioclase (against which Fe-rich liquid is non-wetting) prevented the liquid percolating downwards more than a few cm. Upwards-percolation of Si-rich liquid, exsolved from the solidifying ponded Fe-rich liquid, into subsequently deposited crystal mush accounts for the wispy and irregular felsic patches overlying the reverse modally-graded layer (Figure 10e).

We further suggest that the example of paired segregations associated with autoliths shown in Figure 13a, is evidence for ponding of migrating late-stage interstitial liquid against impermeable fully solidified material, instead of a result of metasomatism. Such paired segregations formed by the unmixing of the ponded liquid, with the dense Fe-rich conjugate sinking to the base, while the buoyant Si-rich liquid rose to the top, locally penetrating the autolith along fractures (Figure 13a).

Building on the abundant evidence of migration of unmixed immiscible conjugates in the mush, and the similarity in microstructure and composition of the mafic and felsic rims associated with the autoliths with their compositional counterparts forming the paired late-stage segregations throughout the Layered Series (e.g. Figure 3), we suggest that the thin mafic and felsic rims in LZc and MZ (Figure 13) are a consequence of ponding of these two liquids against the impermeable autoliths (c.f. Cawthorn & Street, 1994). Accordingly, the mafic rim forms by the solidification of downwards-percolating Fe-rich liquid, whereas the felsic rim forms from the upwards-percolating Si-rich conjugate. The shape of the autoliths affects the extent to which either kind of rim forms because of limited sideways flow of the interstitial liquid: the more laterally extensive the autolith, the more of a barrier it creates. That felsic bodies are found below only the most tabular autoliths is unlikely to be a consequence of a greater lateral mobility of the relatively inviscid fluid. It is more likely because of its lower abundance in the LZc mush, so it can only accumulate in visible amounts against the most tabular autoliths.

THE AUTOLITHS AS A RECORD OF ROOF MUSH EVOLUTION

The confirmation that lithology, plagioclase composition and Θ_{cpp} can be used to identify the source of the autoliths enables a greater understanding of the evolution of the UBS. Firstly, the composite autolith we examined (SK11-B8, Table 2; B8, Figure 1), comprising an upper layer of LZc' material and an underlying LZa' component, can now be understood as representing an unconformity developed within the UBS due to fragmentation. We suggest this autolith is actually upside-down, with the missing LZb' material removed during contemporaneous stoping, resulting in the juxtaposition of later-formed LZc' material on the more primitive LZa' material. A further stoping event during MZ times dislodged the entire autolith, which rotated while it sank and arrived on the floor upside-down.

We can also comment on the likely range of mush thickness in the roof zone during late MZ times, on the basis of the morphology of autolith SK11-B9 (LZb'). The irregular morphology at its base points to this autolith having been incompletely solidified at the moment it was dislodged from the roof. That the top had solidified sufficiently to lock in the primary value of Θ_{cpp}, indicates that the mush was only a few tens of metres thick, most likely due to re-melting of this outer part of the UBS immediately prior to dislodging. Conversely, the relatively anorthitic composition of the plagioclase in autoliths JC-2 and JC-12 (Table 2) is consistent with significant mush thickness during LZa and LZb times in their source region. While the value of Θ_{cpp} in these two autoliths is the same as that seen in LZb', the anorthitic plagioclase and interstitial untwinned augite are consistent with an origin in LZa'. This apparent contradiction can be resolved if these two autoliths came from close to the LZa'/LZb' boundary, in a region of thick mush with a significant distance between the step-wise change in Θ_{cpp} (denoting the horizon where solidification was essentially complete) and the arrival of augite on the liquidus of the bulk magma (the magma-mush interface) (Holness *et al.,* 2017a).

The thermal history of the UBS

In the absence of subsolidus textural equilibration, Θ_{cpp} provides an indication of super-solidus cooling rate for dolerites and gabbros. The similarity of Θ_{cpp} in the UBS (both *in situ* and autoliths) and previously reported values for the MBS (Holness *et al*., 2022) is consistent with broadly similar cooling rates. In detail, Θ_{cpp} in the HZ and LZa equivalents of MBS near the roof on the Skaergaard Peninsula is in the range 81-84˚, slightly lower than that found for LZa* at Udløberen (Table 1). Values of Θ_{cpp} in the MBS LZb equivalent on the Skaergaard Peninsula are in the range 85-87˚, again slightly lower than in LZb* on Udløberen. LZc rocks are generally absent in MBS (Holness *et al*., 2022), but Θ_{cpp} in the MZ equivalent are in the range 91-92˚ in the MBS on the Skaergaard Peninsula. The consistently slightly higher values of Θ_{cpp} in UBS, and in the topmost eastern part of the MBS, relative to those on the western wall on the Skaergaard Peninsula, are consistent with a more rapid cooling rate for the top part of the western wall. The country rock in all locations is the overlying plateau lavas, so the difference is most likely linked to the details of the geometry of the intrusion upper contacts.

Evidence of differential migration of immiscible liquids in the UBS

The autoliths provide an opportunity to constrain the extent of differential migration of immiscible liquids in the UBS as a whole. While neither the autoliths, nor the remaining exposures of the more evolved parts of LZ' or of MZ', contain the olivine rims separating Fe-Ti oxide from plagioclase and pyroxene primocrysts (Figure 4) that are ubiquitous in their Layered Series equivalents, the autoliths derived from LZb' and LZc'/MZ' contain abundant Type I symplectites, which are very rare in the UBS exposures. A re-examination of the MBS shows that although the olivine rims are present (as stated by Holness *et al*., 2022), they only occur close to pockets of granophyre. Understanding what these observations may mean in terms of differential migration of immiscible conjugate liquids requires a re-assessment of the processes resulting in the olivine rims.

The absence of olivine in MZ was attributed by Wager & Deer (1939) to raised Si activity in the bulk magma, induced by the crystallisation of magnetite (Morse *et al*., 1980). They therefore accounted for the olivine rims in MZ by arguing that they crystallised from highly evolved interstitial liquid. However, Nwe (1976) showed that the rims have an Mg/Fe ratio similar to any coexisting cumulus olivine and suggested instead that they formed by some unspecified solid-state reaction between iron-rich oxides and pyroxene. McBirney (1996) attributed the rims to reaction between oxides and pyroxene during cooling and equilibration, during which Fe and Mg from the oxides combined with Ca-poor pyroxene exsolved from augite. Similarly, they were suggested to be a result of a progressively reducing oxygen fugacity during late-stage crystallisation by Holness *et al*. (2011), which triggered the reaction magnetite + pyroxene = olivine + Ca + O_2. Nielsen *et al*. (2015) argue that they crystallised from an Fe-rich interstitial liquid rendered out of chemical equilibrium with the surrounding primocrysts due to the loss of an exsolved, buoyant, Si-rich immiscible conjugate.

Although Nwe (1976) determined that a relatively high Si activity in the interstitial liquid itself (as required for the reaction suggested by Wager & Deer) could not have been responsible for rim growth, it is possible that overgrowth of magnetite primocrysts from the interstitial liquid creates a boundary layer enriched in SiO_2, which would destabilise magnetite via the FMQ oxygen fugacity buffering reaction:

$$SiO_2\ (Liq) + Fe_3O_4\ (Mt) = 3Fe_2SiO_4\ (Fa) + ½\ O_2$$

However, this requires only the overgrowth of magnetite from the interstitial liquid, so would be expected wherever there was sufficient interstitial liquid during orthocumulus magnetite growth: it cannot account for the absence of olivine rims in the autoliths, the MBS and UBS, and in LZa and LZb.

An alternative explanation takes into account the Nielsen *et al.* (2015) interpretation of Paragenesis 2 as a consequence of reaction between plagioclase and pyroxene primocrysts and ponded Fe-rich immiscible melt which had lost appreciable amounts of its Si-rich conjugate, resulting in a decreased silica activity. The parameters likely to play an important role in the formation of the olivine rims are the volume of interstitial melt present when the two-liquid field is reached, the permeability of the mush,

689 and its orientation. Hence, their absence from LZa and LZb is a consequence of the late arrival of the
690 interstitial liquids at the binodal and therefore greater retention of both conjugate liquids within the
691 mush. Their absence from the MBS is because although differential migration of the two liquids could
692 occur, the vertical geometry means that any upwards-migrating Si-rich liquid from any particular level
693 on the wall will be replaced by liquid migrating upwards from lower levels. The situation in the UBS
694 would be complementary to that at the floor, in that any unmixed Fe-rich conjugate would flow
695 downwards, with retention of the Si-rich conjugate. Although this locally resulted in destabilization of
696 magnetite in the UBS and replacement by Fe-rich olivine and pyroxene via reaction with Si-rich liquid
697 (Nielsen *et al*., 2015), there was no ponding of Fe-rich liquid within the UBS and therefore no
698 dissolution of the earlier-formed primocrysts to create olivine rims.

699 Interestingly, Type I symplectites occur in LZc' autoliths, despite the absence of olivine rims. Clearly,
700 reactive symplectites can form in the absence of significant dissolution of the earlier-formed
701 primocrysts, suggesting either that they form by reaction with Fe-rich liquid at lower temperatures or
702 that a smaller extent of chemical disequilibrium is required. It is notable that there appear to be very
703 few reactive symplectites in the remaining exposures of UBS, with the only examples found near the
704 contact. Given the evidence we have presented earlier that the autoliths retain primary microstructures
705 and compositions, this provides further evidence that the remaining exposure of UBS are not fully
706 representative of the entire roof sequence.

707

708 **CONCLUSIONS**

709 The autoliths preserved in the floor cumulates of the Skaergaard intrusion retain primary
710 microstructures created during solidification at the roof of the intrusion, even within a few millimetres
711 of their margins. The bulk compositions and overall microstructures of the majority of the autoliths are
712 similar to those of the outer parts of the UBS, consistent with a source predominantly within the LZ'
713 subdivision.

714 Importantly, the development of mafic rims previously argued to have formed by metasomatism of the
715 autoliths is strongly associated with autolith shape, rather than composition or size and, on the more
716 tabular autoliths, are paired with an underlying felsic rim. This supports our contention that the UBS
717 autoliths were mainly completely solidified when they arrived at the floor, creating barriers to the
718 outcrop-scale migration of mobile interstitial liquid, resulting in the ponding of interstitial liquid migrating
719 through the mush. This is also consistent with the model of Nielsen *et al*. (2015) accounting for the
720 sulphide-poor PGE-Au mineralization in upper MZ, involving the ponding of dense Fe-rich immiscible
721 melts on the floor, with complementary accumulation of low density components such as plagioclase
722 and immiscible Si-rich melt under the roof. The absence of ponding of significant quantities of Fe-rich
723 liquid in the roof mush zone is supported by the absence of olivine rims separating Fe-Ti oxides from
724 primocrysts of other phases in both the autoliths and the UBS, although the presence of Type I
725 symplectites in the autoliths indicates some reaction due to differential migration in parts of the roof
726 no longer exposed.

The Skaergaard autoliths represent a major component of the arguments for microstructures in plutonic rocks being secondary, rather than primary (e.g. McBirney, 2009). While a full assessment of the arguments for the imposition of secondary microstructures on plutonic rocks by wholesale recrystallisation requires re-examination of all other examples used to support this hypothesis, we argue that the Skaergaard autoliths cannot be used in support. Far from having been entirely recrystallised, the Skaergaard autoliths are well-preserved samples of the missing majority of the UBS, offering the only opportunity to establish the true range of composition, microstructural diversity and record of stoping and autolith detachment in this otherwise unavailable part of the intrusion.

ACKNOWLEDGEMENTS

We are grateful to Sam Weatherly, Victoria Honour, Gautier Nicoli, Madeleine Humphreys and Brendan Dyck for assistance in the field. Iris Buisman is thanked for the analysis of plagioclase compositions, and we are grateful to Mike Stock for help with re-calculating the results. Gavyn Rollinson helpfully provided QEMSCAN maps to aid the interpretation of several samples. Rune Larsen and Dick Naslund are thanked for helpful and insightful reviews of an earlier version of the manuscript.

FUNDING

The collection of many of the autoliths was undertaken by JCØA during graduate work supervised by Neil Irvine, funded by the Carnegie Institution for Science, and under the supervision of Kent Brooks with financial support from the Danish Natural Science Research Council and Carlsberg Foundation. We gratefully acknowledge financial support from Trinity College, Cambridge. This work was also funded by UKRI [grant number NE/N009894/1]. For the purpose of open access, the author has applied a Creative Commons Attribution (CC BY) licence to any Author Accepted Manuscript version arising.

DATA AVAILABILITY

All data referred to in this contribution is available either as data tables or in the Supplementary Appendix.

REFERENCES

Annen, C., Latypov, R., Chistyakova, S., Cruden, A.R. & Nielsen, T.F.D. (2022) Catastrophic growth of totally molten magma chambers in months to years. *Scientific Advances*, **8**: eabq0394.

Bartley, J.M., Glazner, A.F. & Coleman, D.S. (2018) Dike intrusion and deformation during growth of the Half Dome pluton, Yosemite National Park, California. *Geosphere*, **14**: 3

Boudreau, A.E. (2011) The evolution of texture and layering in layered intrusions. *International Geology Review*, **53**: 330-353

Cawthorn, R.G. & Street, J. (1994) Vertical migration of residual magma in the Upper Zone of the Bushveld Complex. *Mineralogy and Petrology*, **51**: 345-354.

Fowler, A.C. & Holness, M.B. (2022) The formation of three-grain junctions during solidification. Part II: theory. *Contributions to Mineralogy and Petrology*, **177**: 58. doi.org/10.1007/s00410-022-01921-w

Glazner, A.F., Bartley, J.M., Coleman, D.S. & Lindgren, K. (2020) Aplite diking and infiltration: a differentiation mechanism restricted to plutonic rocks. *Contributions to Mineralogy and Petrology*, **175**: 37. doi.org/ 10.1007/s00410-020-01677-1

Holness, M.B. (2010) Decoding dihedral angles in melt-bearing and solidified rocks. *Journal of the Virtual Explorer*, **35**: paper 3, doi:10.3809/jvirtex.2010.00265.

Holness, M.B. (2015) Plagioclase growth rates control three-grain junction geometry in dolerites and gabbros. *Journal of Petrology*, **56**: 2117-2144.

Holness, M.B., Nielsen, T.F.D. & Tegner, C. (2007) Textural maturity of cumulates: a record of chamber filling, liquidus assemblage, cooling rate and large-scale convection. *Journal of Petrology*, **48**: 141-157.

Holness, M.B., Anderson, A.T., Martin, V.M., Maclennan, J., Passmore, E. & Schwindinger, K. (2007) Textures in partially solidified crystalline nodules: a window into the pore structure of slowly cooled mafic intrusions. *Journal of Petrology*, **48**: 1243-1264.

Holness, M.B., Stripp, G., Humphreys, M.C.S., Veksler, I.V., Nielsen, T.F.D. & Tegner, C. (2011) Silicate liquid immiscibility within the crystal mush: late-stage magmatic microstructures in the Skaergaard Intrusion, East Greenland. *Journal of Petrology*, **52**: 175-222.

Holness, M.B., Richardson, C. & Helz, R.T. (2012a) Disequilibrium dihedral angles in dolerite sills: a new proxy for cooling rate. *Geology,* **40**: 795–798.

Holness, M.B., Humphreys, M.C.S., Sides, R., Helz, R.T. & Tegner, C. (2012b) Toward an understanding of disequilibrium dihedral angles. *Journal of Geophysical Research*, **117**: B06207. doi: 10.1029/2011JB008902

Holness, M.B., Namur, O. & Cawthorn, R.G. (2013) Disequilibrium dihedral angles in layered intrusions: a microstructural record of fractionation. *Journal of Petrology*, **54**: 2067-2093.

Holness, M.B., Tegner, C., Namur, O. & Pilbeam, L. (2015) The earliest history of the Skaergaard magma chamber: a textural and geochemical study of the Cambridge drill core. *Journal of Petrology*, **56**: 1199-1227.

Holness, M.B., Neufeld, J.A., Gilbert, A.J. & Macdonald, R. (2017b) Orientation of tabular mafic intrusions controls convective vigour and crystallisation style. *Journal of Petrology,* **58**: 2035–2053.

Holness, M.B., Tegner, C., Nielsen, T.F.D. & Charlier, B. (2017a) The thickness of the mushy layer on the floor of the Skaergaard magma chamber at apatite saturation. *Journal of Petrology*, **58**: 909-932.

Holness, M.B., Stock, M.J. & Geist, D. (2019) Magma chambers versus mush zones: constraining the architecture of sub-volcanic plumbing systems from microstructural analysis of crystalline enclaves. *Philosophical Transactions of the Royal Society A,* **377**: 20180006. http://dx.doi.org/10.1098/rsta.2018.0006

Holness, M.B., Humphreys, M.C.S., Namur, O., Andersen, J.C. Ø., Tegner, C. & Nielsen, T.F.D. (2022) Crystal mush growth and collapse on a steep wall: the Marginal Border Series of the Skaergaard Intrusion, East Greenland. *Journal of Petrology*, **63**: 1-21.

Honour, V.C. Holness, M.B., Partidge, J.L. & Charlier, B. (2019) Microstructural evolution of silicate immiscible liquids in ferrobasalts. *Contributions to Mineralogy and Petrology*, **174**: 77.

Hoover, J.D. (1989) Petrology of the Marginal Border Series of the Skaergaard Intrusion. *Journal of Petrology*, **30**: 399–439.

Humphreys, M.C.S. (2011) Silicate liquid immiscibility within the crystal mush: evidence from Ti in plagioclase from the Skaergaard intrusion. *Journal of Petrology*, **52**: 147-174.

Irvine, T.N. (1980) Observations on layering in the Skaergaard intrusion. *Carnegie Institution of Washington Yearbook*, **79**: 257-262.

Irvine, T.N. (1987) Layering and related structures in the Duke Island and Skaergaard intrusions: similarities, differences and origins. In: Parsons, I. (ed.) *Origins of Igneous Layering*, pp. 185-245. D. Reidel Publishing Company.

Irvine, T.N., Andersen, J.C.Ø. & Brooks, C.K. (1998) Included blocks (and blocks within blocks) in the Skaergaard intrusion: geologic relations and the origins of rhythmic modally graded layers. *Geological Society of America Bulletin*, **110**: 1398-1447.

Jakobsen, J.K., Veksler, I.V., Tegner, C. & Brooks, C.K. (2005) Immiscible iron- and silica-rich melts in basalt petrogenesis documented in the Skaergaard intrusion. *Geology*, **33**: 885-888.

Kays, M.A. & McBirney, A.R. (1982) Origin of the picrite blocks in the Marginal Border Group of the Skaergaard intrusion, East Greenland. *Geochimica et Cosmochimica Acta*, **46**: 23-30.

Larsen, R.B. & Brooks, C.K. (1994) Origin and evolution of gabbroic pegmatites in the Skaergaard intrusion, East Greenland. *Journal of Petrology*, **35**: 1651-1679.

Lindsley, D.H., Brown, G.M. & Muir, I.D. (1969) Conditions of ferrowollastonite–ferrohedenbergite inversion in the Skaergaard intrusion, East Greenland. *Mineralogical Society of America Special Paper*, **2**: 193–201.

Maaløe, S. (1976) Zoned plagioclase of the Skaergaard intrusion, East Greenland. *Journal of Petrology*, **17**: 398-419.

Markl, G. (2005) Mullite-corundum-spinel-cordierite-plagioclase xenoliths in the Skaergaard Marginal Border Group: multi-stage interaction between metasediments and basaltic magma. *Contributions to Mineralogy and Petrology*, **149**: 196-215.

McBirney, A.R. (1975) Differentiation of the Skaergaard Intrusion. *Nature*, *253*: 691-694

McBirney, A.R. (1979) Effects of assimilation, *in* Yoder, H.S., Jr., ed., *The evolution of the igneous rocks, Fiftieth Anniversary Perspectives*: Princeton, New Jersey, Princeton University Press, pp. 307–338.

McBirney, A.R. (1987) Constitutional zone refining of layered intrusions. In: Parsons, I. (ed.) *Origins of Igneous Layering*, pp. 437-452. D. Reidel Publishing Company.

McBirney, A.R. (1989a) The Skaergaard layered series: I. Structure and average compositions. *Journal of Petrology*, **30**: 363–397.

McBirney, A.R. (1989b) *Geological map of the Skaergaard Intrusion, East Greenland*. In: *Eugene, Department of Geology*. University of Oregon, p.1:20 000.

McBirney, A.R. (1995) Mechanisms of differentiation in the Skaergaard Intrusion. *Journal of the Geological Society, London*, **152**: 421-435

McBirney, A.R. (1996) The Skaergaard intrusion, *in* Cawthorn, R.G. (ed.), *Layered intrusions* (Developments in Petrology 15): Amsterdam, Elsevier, pp. 147–180

McBirney A.R. (2009) Factors governing the textural development of Skaergaard gabbros: a review. *Lithos*, **111**: 1-5.

McBirney, A.R. & Hunter, R.H. (1995) The cumulate paradigm reconsidered. *The Journal of Geology*, **103**: 114-122.

McBirney, A.R. & Nakamura, Y (1974) Immiscibility in late-stage magmas of the Skaergaard intrusion. *Carnegie Institution of Washington Yearbook*, **73**: 348-352.

McBirney, A.R. & Noyes, R.M. (1979) Crystallisation and layering of the Skaergaard intrusion. *Journal of Petrology*, **20**: 487-554.

McBirney, A.R. & Sonnenthal, E.L. (1990) Metasomatic replacement in the Skaergaard Intrusion, East Greenland: Preliminary observations. *Chemical Geology*, **88**: 245–260.

Morse, S. A. (2011) The fractional latent heat of crystallising magma. *Ameri-can Mineralogist,* **96**: 682–689.

Morse, S.A., Lindsley, D.H. & Williams, R.J. (1980) Concerning intensive parameters in the Skaergaard intrusion. *American Journal of Science*, **280A**: 159-170.

Namur, O., Higgins, M.D. & Vander Auwera, J. (2015) The Sept Iles Intrusive Suite, Quebec, Canada. In: Charlier, B., Namur, O., Latypov, R. & Tegner, C. (eds.) *Layered Intrusions*. Springer. Dordrecht, Heidelberg, New York, London. p. 465-516.

Namur, O. & Humphreys, M.C.S. (2018) Trace element constraints on the differentiation and crystal mush solidification in the Skaergaard Intrusion, Greenland. *Journal of Petrology*, **59**: 387-418.

Naslund, H.R. (1984a) Petrology of the Upper Border Series of the Skaergaard intrusion, East Greenland: *Journal of Petrology*, **25**: 185–212.

Naslund, H.R. (1984b) Supersaturation and crystal growth in the roof-zone of the Skaergaard magma chamber. *Contributions to Mineralogy and Petrology*, **86**: 89-93.

Naslund, H.R. (1986) Disequilibrium partial melting and rheomorphic layer formation in the contact aureole of the Basistoppen sill, East Greenland. *Contributions to Mineralogy and Petrology*, **93**: 359–367.

Nielsen, T.F.D. (2004). The shape and volume of the Skaergaard Intrusion, East Greenland: Implications for mass balance and bulk composition. *Journal of Petrology*, **45**: 507-530.

Nielsen, T.F.D., Andersen, J.C.Ø., Holness, M.B., Keiding, J.K., Rudashevsky, N.A., Rudashevsky, V.N., Salmonsen, L.P., Tegner, C. & Veksler, I.V. (2015) The Skaerggard PGE and gold deposit: the result of *in situ* fractionation, sulphide saturation, and magma chamber-scale precious metal redistribution by immiscible Fe-rich melt. *Journal of Petrology*, **56**: 1643-1676.

Nielsen, T.F.D., Brooks, C.K. & Keiding, J.K. (2019) Bulk liquid for the Skaergaard intrusion and its PGE-Au mineralization: composition, correlation, liquid line of descent, and timing of sulphide saturation and silicate-silicate immiscibility. *Journal of Petrology*, **60**: 1853-1880.

Nwe, Y.Y. (1976) Electron-probe studies of the earlier pyroxenes and olivines from the Skaergaard Intrusion, East Greenland. *Contributions to Mineralogy and Petrology*, **55**: 105-126.

887 Pang, K.N., Li, C., Zhou, M.F. & Ripley, E.M. (2009) Mineral compositional constraints on petrogenesis
888 and oxide ore genesis of the late Permian Panzhihua layered gabbroic intrusion, SW China. *Lithos*,
889 **110**: 199-214.

890 Riegger, O.K. & van Vlack, L.H. (1960) Dihedral angle measurements. *Transactions of the*
891 *Metallurgical Society of the AIME*, **218**: 933–935.

892 Salmonsen, L.P. & Tegner, C. (2013) Crystallisation sequence of the Upper Border Series of the
893 Skaergaard Intrusion: revised subdivision and implications for chamber-scale magma homogeneity.
894 *Contributions to Mineralogy and Petrology*, **165**: 1155-1171.

895 Salmonsen, L.P., Tegner, C., Barfod, G.H. & Lesher, C.E. (2014) Compositional differences between
896 roof and floor rocks of the Skaergaard intrusion. *EGU General Assembly Conference Abstracts,*
897 2014EGUGA..1612525S

898 Sonnenthal, E.L. (1992) Geochemistry of dendritic anorthosites and associated pegmatites in the
899 Skaergaard intrusion, East Greenland: evidence for metasomatism by a chlorine-rich fluid. *Journal of*
900 *Volcanology and Geothermal Research*, **52**: 209-230.

901 Sonnenthal, E.L. & McBirney, A.R. (1998) The Skaergaard Layered Series. Part IV. Reaction-transport
902 simulations of foundered blocks. *Journal of Petrology*, **39**: 633-661

903 Stickels, C.A. & Hücke, E.E. (1964) Measurement of dihedral angles. *Transactions of the Metallurgical*
904 *Society of the AIME*, **230**: 795–801.

905 Taylor, H.P. Jr. & Forester, R.W. (1979) An oxygen isotopic study of the Skaergaard Intrusion and its
906 country rocks: a description of a 55-m.y. old fossil hydrothermal system. *Journal of Petrology*, **20**: 355-
907 419.

908 Vukmanovic, Z., Holness, M.B., Monks, K. & Andersen, J.C.Ø. (2018) The Skaergaard trough layering:
909 sedimentation in a convecting magma chamber. *Contributions to Mineralogy and Petrology*, **173**: 43.

910 Wager, L.R. (1960) The major element variation of the Layered Series of the Skaergaard intrusion and
911 a re-estimation of the average composition of the Hidden Layered Series and of the successive
912 residual magmas. *Journal of Petrology*, **1**: 364-398.

913 Wager, L.R. & Deer, W.A. (1939) Geological investigations in East Greenland, Part III. The petrology
914 of the Skaergaard Intrusion, Kangerdlugssuaq, East Greenland. *Meddeleser om Grønland*, **105**.

915 Wager, L. R. & Brown, G. M. (1968) *Igneous Rocks*. Edinburgh, Oliver and Boyd, 588 p.

916 Weatherley, S., Andersen, J.C.Ø., Holness, M.B., Dyck, B.J., Nicoli, G. & Honour, V.C. (2018)
917 Combined 'sedimentary' and in-situ origin for magmatic Fe-Ti-V deposits: new insights from the
918 Skaergaard intrusion, East Greenland. 33rd Nordic Geological Winter Meeting.
919 https://2dgf.dk/foreningen/33rd-nordic-geological-winter-meeting/ngwm-2018-abstracts/1-igneous-
920 rocks-and-processes/#5

921 Wotzlaw, J.-F., Bindeman, I.N., Schaltegger, U., Brooks, C.K. & Naslund, H.R. (2012) High-resolution
922 insights into episodes of crystallization, hydrothermal alteration and remelting in the Skaergaard
923 intrusive complex. *Earth and Planetary Science Letters*, **355–356**: 199–212.

924

925 **FIGURE CAPTIONS**

926 Figure 1 – Schematic map of the Skaergaard Intrusion (after McBirney, 1989b). The location of the
927 detailed map of the southern part of the intrusion shown in Figure 2 is outlined by the blue box. The
928 locations of the autoliths analysed as part of this study are also shown (autoliths labelled B are the
929 SK11-B samples listed in Table 2, and SK11-B9 is illustrated in Figure 12).

930 Figure 2 – Detailed map (after McBirney, 1989b) of the southern part of the Skaergaard Intrusion, with
931 the location of the samples examined for this study. The various schemes for subdividing the UBS are
932 also shown, together with the composition of plagioclase cores used to define the boundaries.

933 Figure 3 – Paired felsic and mafic segregations in the Skaergaard Layered Series. (a) Felsic wisp-like
934 segregations forming a network in LZb on Uttental Plateau. Note the corresponding network of mafic
935 wisp-like segregations immediately underlying the felsic network. Handheld GPS for scale. (b)
936 Localised zone rich in late-stage liquid segregations in upper LZc, Kraemer Ø. The felsic segregations
937 overly the associated mafic segregations, and the localised disruption of the fabric in the gabbroic host
938 indicates fluid movement in a deformable mush. (c) A pocket of segregated interstitial liquid containing
939 both felsic and mafic components. The rounded top may be related to the high viscosity of the felsic
940 liquid, in comparison to the irregularity of the lower relatively inviscid mafic conjugate. Upper LZc on
941 Kraemer Ø. (d) Close-up of the right side of the segregation in (c).

942 Figure 4 – photomicrographs of Paragenesis 2, taken under crossed polars. (a) Sample 591633 from
943 LZc, showing the olivine rims separating Fe-Ti oxides (ox) from augite (cpx). Arrows point to examples
944 of Type I symplectites. Scale bar is 500 μm long. (b) The top of the reverse modally-graded layer
945 shown in Figure 12e, displaying olivine rims separating Fe-Ti oxides from recrystallised clinopyroxene.
946 The rims are monocrystalline and in optical continuity with nearby olivine primocrysts (examples are
947 arrowed), but polycrystalline where there are no nearby olivine primocrysts. Scale bar is 1 mm long.

948 Figure 5 – Xenoliths and autoliths in the Skaergaard Layered Series. (a) xenolith of Precambrian
949 gneissic country rock near the contact on Uttental Plateau. (b) Accumulation of xenoliths of ultramafic
950 material near the contact on Uttental Plateau. Rifle for scale. (c) Group of metabasic xenoliths (rusty
951 orange colour) derived from the plateau lavas overlying the intrusion, in MZ in the southern part of
952 Kraemer Ø. The arrows show thin mafic rims. Hammer for scale. (d) Autolith derived from the UBS in
953 LZb on Kraemer Ø. Note the minor deflection of modal layering underneath the autolith indicative of
954 limited and localised compaction. Hammer for scale.

955 Figure 6 – Scans of thin sections, showing contacts between autolith and the host gabbro. The
956 sections are all oriented so the images are the right way up. All scale bars 5 mm long. (a) the lower,
957 sharply defined, contact between autolith SK11-B1 and host gabbro (LZc).The autolith comprises
958 primocrysts of plagioclase and (rare) olivine (some of which extends to form thin rims separating
959 plagioclase and augite: the location of the rim shown in Figure 11c is arrowed), with interstitial augite,
960 whereas the LZc host rock comprises primocrysts of plagioclase, olivine, augite and Fe-Ti oxides.
961 Dihedral angles form clearly distinct populations in autolith and host (Table 2). (b) The felsic layer
962 immediately underlying the oxide-rich mafic base of a modally graded layer in LZc (collected from
963 close to those shown in Figure 9). The approximate position of the base and top of the felsic layer are
964 shown by the dashed red lines. (c) Example of a mafic rind developed at the top of a autolith. The
965 autolith is an oxide-rich gabbro and the host gabbro is LZc. (d) Example of a felsic rind developed at
966 the base of an oxide-rich gabbro autolith hosted by LZc gabbro (typical of those shown in Figures 13e,
967 f). The approximate positions of the base of the autolith and the lower contact of the felsic rind are
968 shown by the red dashed lines. Note the finer grainsize of the augite in the felsic rind compared to the
969 host gabbro.

Figure 7 – (a) Rind rock surrounding some of a group of blocks in LZc (stated to be MZ by Irvine *et al*. (1998) before the LZc/MZ boundary was re-mapped), viewed looking towards SE. This location is that of Figure 12F of Irvine *et al.* (1998). Hammer for scale. (b) The setting of the blocks shown in (a), re-drawn and simplified from Figure 23 of Irvine *et al.* (1998). Pale grey denotes autoliths, with rind rock shown as dark grey. The lines denote Fe-rich modal layering. The location of rind rock sample JC-59 is shown by the circle, and that of Figure (a) is shown by the red box.

Figure 8 – photomicrographs of the UBS. Locations of the samples are shown in Figure 2. (a) Randomly oriented plagioclase with irregular grain margins indicative of *in situ* growth to impingement. Olivine (ol) forms clusters of anhedral grains. Sample EG 1724. Scale bar is 1 mm long. (b) Randomly oriented plagioclase with anhedral primocrysts of olivine, together with poikilitic clinopyroxene (cpx) and interstitial Fe-Ti oxides (ox). Sample EG 2274. Scale bar is 1 mm long. (c) UZ' gabbro, with abundant interstitial granophyre (gr), euhedral Fe-Ti oxides, and clusters of anhedral clinopyroxene. Sample EC 49. Scale bar is 1 mm long. (d) The clinopyroxene in other regions of the same sample as (c) is dendritic. The quartz in the granophyre forms plates (elongate when viewed edgewise), indicative of an origin as tridymite. Scale bar is 1 mm long. (e) LZc' gabbro with abundant euhedral grains of Fe-Ti oxides. The interstitial augite forms aggregates of anhedral grains. Sample EC 37. Scale bar is 1 mm long. (f) Highly granophyric material from the inner parts of the UBS, with abundant chlorite (chl) in the centres of the granophyric patches, which are sufficiently large to isolate individual euhedral plagioclase grains (pl). Sample EG 1876. Scale bar is 1 mm long.

Figure 9 – Modal layering in the upper part of LZc on Kraemer Ø. (a) The well-defined oxide-rich bases of several of the modally-graded layers are immediately underlain by a strongly felsic layer (examples are arrowed). This outcrop also contains a number of autoliths, aligned with their long axes (in 2D cross section) parallel to the layering. Those autoliths marked with an asterisk have no mafic top rims, while that marked with a white circle has a well-defined mafic rim along its top surface. (b) Closer view of the modally-graded layering showing the well-defined felsic layers immediately underlying their oxide-rich base (arrowed).

Figure 10 – Modally-graded layers in UZ of the Skaergaard intrusion. (a) In UZa, the isolated modally-graded layers have irregular upper surfaces, with the felsic material forming elongate fingers into the overlying host homogeneous gabbro (an example is arrowed). The pole on the right side of the image is 1 m long. (b, c, d) In UZb, the irregularity of the tops of the modally-graded layers is greater than in UZa, with increasing extent of detachment of the felsic material from the underlying layer with stratigraphic height. This detachment is complete in the upper parts of UZb, with the felsic component forming irregular separate wisps and lenses (d). (e) A reverse modally-graded layer in UZa, near the layer-parallel pegmatite sills shown in Figure 5 of Larsen & Brooks (1994). Note the felsic regions (arrowed) immediately overlying the layer. Hammer handle for scale.

Figure 11 – photomicrographs of autoliths (locations shown in Figure 1). (a) JC31, a gabbroic troctolite containing highly anhedral olivine (ol). The plagioclase grains are randomly oriented, with non-planar grain boundaries with no orientation relationship with likely growth faces, indicative of substantial *in situ* growth to impingement. Photographed under crossed polars. Scale bar is 1 mm long. (b) JC24A, a gabbroic troctolite containing subhedral olivine as isolated grains or clusters, together with interstitial augite. Photographed in plane polarised light. Scale bar is 1 mm long. (c) SK11-B1, a gabbroic troctolite containing interstitial augite that is locally rimmed with olivine (arrowed). Photographed under crossed polars. Scale bar is 500 μm long . (d) JC13, a gabbroic troctolite, in which the olivine primocrysts are locally rimmed with Type I symplectites. Photographed in plane polarised light. Scale bar is 250 μm long. (e) JC105, an oxide-rich gabbro, with Type I symplectites (examples are arrowed). Olivine rims separate the oxide from augite, with a rare example of an anhedral olivine grain enclosed by the augite. Photographed under crossed polars. Scale bar is 1 mm long. (f) JC23, an oxide-rich

gabbro, showing the (100) simple twins in augite primocrysts. Photographed under crossed polars. Scale bar is 1 mm long.

Figure 12 – Autolith SK11-B9 (location, B9, shown in Figure 1). (a) viewed looking SE, with the top of the autolith at the left, and the northern side running along the base of the image. Note the sharply defined and straight side of the autolith. The location of sample SK11-B9-10 is shown by the asterisk. Note the thin Fe-rich margin along the top of the autolith (arrowed). (b) the northern side of the autolith, viewed looking east, up the stratigraphy. The base is highly irregular and lobate, with patchy felsic margins.

Figure 13 – (a) Large relatively felsic autolith in LZb (the width in this field of view is ~10m) underlain by a group of irregular bodies of segregated late-stage liquid, separated into an upper, volumetrically dominant, felsic portion (asterisks) and a lower mafic portion (white arrows pointing up). The gabbro underlying these contains an irregular network of mafic segregations (white arrows pointing down). The felsic part of the segregation fills an irregular fracture in the autolith (black arrows). The remainder of the autoliths shown here are from LZc on Kraemer Ø. (b) A pair of rounded autoliths (the smaller of the pair is marked with an asterisk) with no compositional variations at their margins. (c) An elongate autolith (lower part of image) has a well-defined mafic rim on its top surface, whereas smaller, more rounded, autoliths (two are marked by arrows and the third by an asterisk) have no compositional changes at their margins. (d) An elongate autolith with a well-defined mafic rim on the top contact, and a poorly-defined felsic rim locally developed (immediately above the scale bar). The asterisk shows a rounded autolith with no rim. (e) A group of autoliths comprising a trio of highly elongate autoliths with both mafic and felsic rims (the felsic rims are arrowed). The autolith marked with an asterisk has no rims. (f) A highly elongate autolith with well-developed paired mafic and felsic rims. The felsic rim at the base is arrowed. This image shows only a portion of the autolith, which extends a further metre to the right.

Figure 14 – Photomicrographs of rind rock. (a), (b) and (c) Rind rock forming dykes cutting the autolith illustrated in Figures 22 and 21B of Irvine *et al.* (1998) under crossed polars (a) and in plane polarised light (b). Note the close approach to a granular microstructure and absence of a grain preferred orientation for the plagioclase. Oxides are either ilmenite (elongate) or magnetite (anhedral). (c) shows olivine rims separating magnetite from clinopyroxene, and some small and very fine-grained reactive symplectites (arrowed). Scale bar is 1 mm long in (a) and (b), and 250 μm long in (c). (d) Well-developed reactive symplectites and olivine rims associated with anhedral aggregates of Fe-Ti oxides in rind rock associated with an LZc' autolith in MZ, illustrated in Figure 17a of Irvine *et al*. (1998). Plane polarised light. Scale bar is 500 μm long. (e) Rind rock associated with an autolith in LZc and illustrated in Figure 23 of Irvine *et al*. (1998), displaying a preferred orientation of elongate ilmenite grains. Plane polarised light, Scale bar is 500 μm long. (f) Same sample as (e), showing olivine oikocrysts with inclusions of plagioclase and Fe-Ti oxides in the more coarse-grained part. Plane polarised light. Scale bar is 1 mm long.

Figure 15 – The apparent aspect ratio of 88 autoliths in LZc on Kraemer Ø, measured on outcrop surfaces approximately perpendicular to the layering, divided according to whether the autoliths show no rims, mafic rims only (on their top surfaces), or paired mafic and felsic rims.