Peculiar characteristics of the Bjerkreim-Sokndal intrusion (Rogaland anorthosite province, S. Norway) and implications for the layering formation.

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Abstract: The Bjerkreim-Sokndal intrusion has emplaced in a deep-seated (c. 5 kbar) and high temperature environment close to anorthosite diapirs. Isomodal layers and small scale modally graded layers are particularly well developed in the intrusion. The roof zone was a mixture of melts maintained at high temperature for c. 12 m.y. by repeated intrusions of several magmas. There are ample evidence that the magma column was zoned. The cumulates are built up by a mixture of plagioclase and a mafic pole with constant modal proportions of mafic minerals. The latter property precludes sorting and magma currents. Immiscibility can also be precluded for the same reason. *In situ* crystallization at the floor of the intrusion with variation in the plagioclase nucleation rate is suggested as the dominant mechanism of cumulate formation.

Keywords: nucleation rate, *in situ* crystallization, ultra high temperature,

The main features of the intrusion

The Bjerkreim-Sokndal layered intrusion, the largest layered intrusion in Europe (c. 230 km^2), has been extensively studied thanks to exceptional outcropping conditions and accessibility, particularly in its northern part, the Bjerkreim lobe (see the review of Wilson et al., 1996; Duchesne, 2001). The intrusion took place soon after the emplacement of 3 massif-type anorthosites, at about c. 5 kbar pressure in granulite facies conditions favourable to plastic deformation of the envelope of country gneiss (see Vander Auwera et al., 2011). It contains a c. 7000 m Layered Series made up of rocks of the charnockitic or AMCG series: anorthosite, leuconorite, troctolite, norite, gabbronorite, mangerite topped by unlayered quartz mangerite and charnockite. The Layered Series is organized in megacyclic units (MCU 0 to IV) that display a progressive upward evolution from anorthosite to more mafic lithologies. There are conspicuous evidence that the magma chamber was periodically replenished with new influxes of parental magma that crystallized to form each MCU. MCU IV, with a maximum thickness of 1.8 km, is the thicker and most complete. It grades upwards from anorthosite, to leucotroctolite, leuconorite and gabbronorite. Following the terminology of Irvine (1982) with p as plagioclase, o as olivine, h as Ca-poor pyroxene (hypersthene or pigeonite), i as ilmenite, m as magnetite, c as Ca-rich pyroxene, a as apatite, and –C as cumulate, the detailed succession is p-C, poim-C, phi-C, phim-C, and phimca-C. Variation in mineral compositions and progressive evolution of Sr isotope ratio suggests that continuous assimilation took place during fractional crystallization of the MCUs (Nielsen et al., 1996). There is no upper border zones with reverse cryptic layering as observed e.g. in the Skaergaard (McBirney, 1996) or in the Fedorivka intrusion (Duchesne et al., 2006).

The thicknesses of the MCUs wedge out in a synform structure. This structure results from a syn- to post magmatic gravity-induced subsidence affecting the Layered Series and also the upper part of the massif (Bolle et al., 2000; Bolle et al., 2002). After crystallization of the first two MCUs in a funnel shape structure, the magma chamber expanded laterally over a considerable distance (Nielsen and Wilson, 1991). The leucotroctolite layer close to the base of MCU IV can be followed over more than 24 km (Michot, 1960). Given the maximum thickness of 1.8 km for MCU IV, the shape of the unfolded magma chamber in which MCU IV influx crystallized was sill-like, that is relatively thin compare to the surface extension, more resembling, relatively speaking, the Bushveld than the onion's shape of the Skaergaard chamber.

Classical characteristics of layered rocks are met in BKSK: macrocyclic units, isomodal layers (m- to tens of m-thick), small scale modally graded layers, inch-scale layering, igneous lamination (though somewhat blurred by post cumulus processes) and progressive and regressive cryptic layering. Xenoliths with impact structure are common. Erosional structures are very scarce and grain-size graded layers have not been observed.

The parental magma was a jotunite, that is a hypersthene ferrodiorite (the Tjörn chill). Its composition is reported in Table 1. Liquidus temperatures, and densities (and viscosities) have been calculated by the MELTS algorithm (Ghiorso and Sack, 1995) assuming 5 kbar, fO_2 at QFM+1, 0-0.5 $\%H_2O$ in the Tjörn composition (Vander Auwera and Longhi, 1994). The liquidus temperature in hydrated composition is lower than in dry condition and in the range of experimental data of Vander Auwera and Longhi (1994). Noteworthy are the density values (ρ = 2.66-2.63) similar to the plagioclase density (An_{45-50}, ρ = 2.62-2.63). Plagioclase has thus little tendency to sink or float in the magma.

A slowly cooled intrusion

The roof of the intrusion is not preserved but the highest rocks in the stratigraphy of the massif are quartz mangerite and charnockite, suggesting a melt zone formed by a mixture of residual melts fractionated from the layered series, various melted products from the country rock gneiss and new intrusions of evolved magma (Duchesne and Wilmart, 1997; Wilson and Overgaard, 2005; Bolle and Duchesne, 2007) as well as of jotunitic magma (Wiebe, 1984). U-Pb zircon dating on quartz mangerites from the upper part of the intrusion has shown (Vander Auwera et al., 2011) that zircon cores crystallized at 931 ±7 Ma, most probably from residual liquids (olivine quartz mangerite) fractionated from the Layered Series, are mantled with 919 ±8 Ma zircon possibly coming from a different 2-pyroxene quartz mangeritic magma. At field scale there is no evidence of intrusive contacts between the two rock types (Duchesne and Wilmart, 1997). So it cannot be precluded that the new magma completely mingled with the resident magma when the later was still partly liquid. Moreover, jotunitic pillows in the quartz mangerites (Wiebe, 1984; Duchesne and Wilmart, 1997) testify the intrusion of at least a high temperature magma in this part of the magma chamber. All these points indicate a very long time interval (up to 12 m.y.) for the complete crystallization of the whole intrusion. This characteristic can be related to the deep-seated emplacement of the intrusion and the thermal evolution of the region. Sapphirine-bearing granulites in the country gneiss envelope indicate ultrahigh temperature conditions at c. 1000°C and c. 7.5 kbar at c. 1010 Ma followed by near-isothermal high temperature decompression to P <5.5 kbar at the time of emplacement of the anorthosites and BKSK intrusion (Drüppel et al., 2013). Thermal modelling of the whole anorthosite province (Westphal et al., 2003) showed that at 5 km from the contact between BKSK and the envelope a peak temperature of 860°C was attained 6 m.y. after the intrusion of the adjacent EGOG anorthosite and that the temperature was still above 750°C after 11 m.y., all the more so since this thermal modelling does not take into account the repeated intrusions of magmas in the upper part of the massif. This supports the hypothesis that the upper part of BKSK could still be above solidus at c. 920 Ma., that is 12 m.y. after the emplacement of the anorthosite massifs at 932 Ma.

A zoned magma chamber

The magma chamber was saucer-shaped with very low values of the slopes (a few degrees) and there are consistent evidence that the magma column was zoned with progressively more evolved liquids towards the top (Nielsen and Wilson, 1991). This zoning was formed essentially by repeated influxes on primitive magmas at the beginning of each MCU. These influxes mix only with a small quantity of the resident magma to form hybrids, most of the resident magma being elevated on top of the new influx. This is particularly well demonstrated in MCU III and IV in which the mineral association regresses to poim-C with the most primitive olivines. Another evidence of a stratified magma column is the occurrence of a sulphide-rich orthopyroxenite layer at the base of MCU III (Jensen et al., 2003). Finally the development of an acidic melt zone at the roof of the intrusion on top of the melts giving rise to the layered series contributed to the zoning. The density evolution in the melt column can be evaluated by means of the cumulate succession: after a relatively short sequence of p-C, pi-C and poim-C, the dominant cumulate contain Fe-Ti oxide minerals (see below), the crystallization of which together with assimilation of country rocks decrease the density of the melts. This situation favours the development of a double-diffusive stratification of the magma chamber (see the review of Campbell, 1996) with convection limited to the melt layers.

Cumulate bulk composition

Another very intriguing characteristic of the BKSK intrusion is the bulk composition of its cumulate rocks. In the most abundant rock types, pih-C, pihm-C and pihmca-C, analyses of 44, 14 and 41 samples respectively (Duchesne and Charlier, 2005) showed that each composition is a mixture between a plagioclase pole and a mafic pole. In pih-C, the mafic pole is made up of 45% hemo-ilmenite + 55% Ca-poor pyroxene; in pihm-C, the mafic pole is 37% ilmenite + 49% Ca-poor pyroxene + 14% magnetite; and in pihmca-C, the modal composition of the mafic pole is 40% Ca-poor pyroxene + 9% Ca-rich pyroxene + 36% Fe-Ti oxides + 14% apatite. This interpretation was questioned by Robins and Chiodoni (2007) who suggest that these relationships resulted from spurious correlations. Merging their analyses with the previous analyses of Duchesne and Charlier (2005), the 2-pole relationship was however confirmed on a total of 124 pih-C samples (Fig. 1)(Duchesne and Charlier, 2007). It was thus concluded “the major factor responsible for the modal layering in BKSK is the variation in modal ratio of plagioclase to the sum of the mafic phases” (Duchesne and Charlier, 2005). The proportion of trapped liquid in pih-C was determined by means of trace element compositions (Charlier et al., 2005). It can attain 25% very locally (close to the margin) but it is usually less than 10% (the rocks are adcumulates). The trapped liquid fraction has thus little influence on the major elements, except on P_2O_5 and K_2O contents (see fig. 9 in Duchesne and Charlier, 2005).

Implications for the layering formation

The concept of mafic pole implies that there is no mechanical sorting of the mafic minerals although their densities vary from 3.2 for apatite to 5.2 for magnetite. The Stokes’s law would predict a deposition speed approximately 5 times larger for magnetite than for apatite. Gravity controlled processes can thus be precluded. The vigorous downward plunging currents coming from the vertical walls of the intrusion with lateral transport and deposition first suggested by Wager and Brown (1968) to explain modally graded layers have not operated here. All the more improbable is this mechanism since the vertical walls of the sill-like magma chamber were more than 24 km apart when MCU IV crystallized.

It is classically considered (see e.g. Morse, 2008)) that nucleation begins at the roof in a supercooled sheet of magma which is gravitationaly instable and descends through the magma chamber to finally wash the floor. In the BKSK intrusion, this view can be questioned. Firstly is a melt roof zone episodically reheated by injection of magmas a favourable site to initiate nucleation? Secondly, can the currents implied by this mechanism maintain a double diffusive stratification of the chamber? And finally are magma movements in descending currents or plumes compatible with the fact that the mafic minerals are not sorted when they reach the floor? All these objections point to nucleation at the (inclined) floor of the magma chamber where heat is lost more efficiently than through the roof of the intrusion.

Lack of sorting of mafics and density similarities of plagioclase and magma have led Duchesne and Charlier (2005) to propose *in situ* growth by heterogeneous nucleation at the topmost layer of cumulus grains in direct contact with the magma (Sparks et al., 1985). Complementary evidence of *in situ* growth is provided by the "stick-like" inclusion of Fig. 2.

In this model, variation in the plagioclase mode is due to variation in nucleation rate and degree of supersaturation. What causes these variations remains enigmatic. In isomodal layers a constant degree of supercooling could result from a balance between conductive heat loss and heat production (latent heat of crystallization) and could depend on the thickness of the double-diffusive layer. In modally graded layers, continuous increase in supersaturation due to faster heat loss than heat production resulting from thinner layers could explain the upwards increase in plagioclase content.

Inference on immiscibility

During the leuconoritic (pih-C and pihm-C) and gabbronoritic (pihmca-C) stages of BKSK, the two pole concept precludes immiscibility of Si-rich and Fe-rich melts, though this phenomenon is observed in ferrobasaltic intrusions (e.g. Sept-Iles intrusion, Charlier et al., 2011). Indeed, immiscibility would give cumulates with two different modal proportions of minerals, that is two different mafic poles instead of the one observed at each leuconoritic or gabbronoritic stages. Immiscibility of a Fe-Ti -(P) oxide melt, currently invoked for the formation of Fe-Ti deposits (see the review of Force, 1991) and nelsonite (Duchesne, 1999), is also irreconcilable with the concept: apatite and Fe-Ti oxide minerals remain linked to the pyroxenes in the mafic pole and have thus no tendency to segregate to form oxide (+ apatite) layers, pods or dykes.

Conclusions

The most original character of the BKSK massif is the depth of intrusion and the (U)HT of its environment, the later resulting from the proximity with anorthosite diapirs. The roof was melted and periodically intruded by jotunitic and acidic magmas. The consequence was a very low cooling rate. The form of the magma chamber where layering was the most developed (MCU III and IV) had a high aspect ratio with walls quite apart and a gently-sloping floor. The magma chamber was zoned with upward silica enrichment of the melt and decreasing densities. This situation permitted the development of a stable double-diffusive layered structure not disturbed by currents of large amplitude, and *in situ* crystallization on the floor. All these conditions were required to develop the very peculiar type of layering, the 2-poles cumulates.

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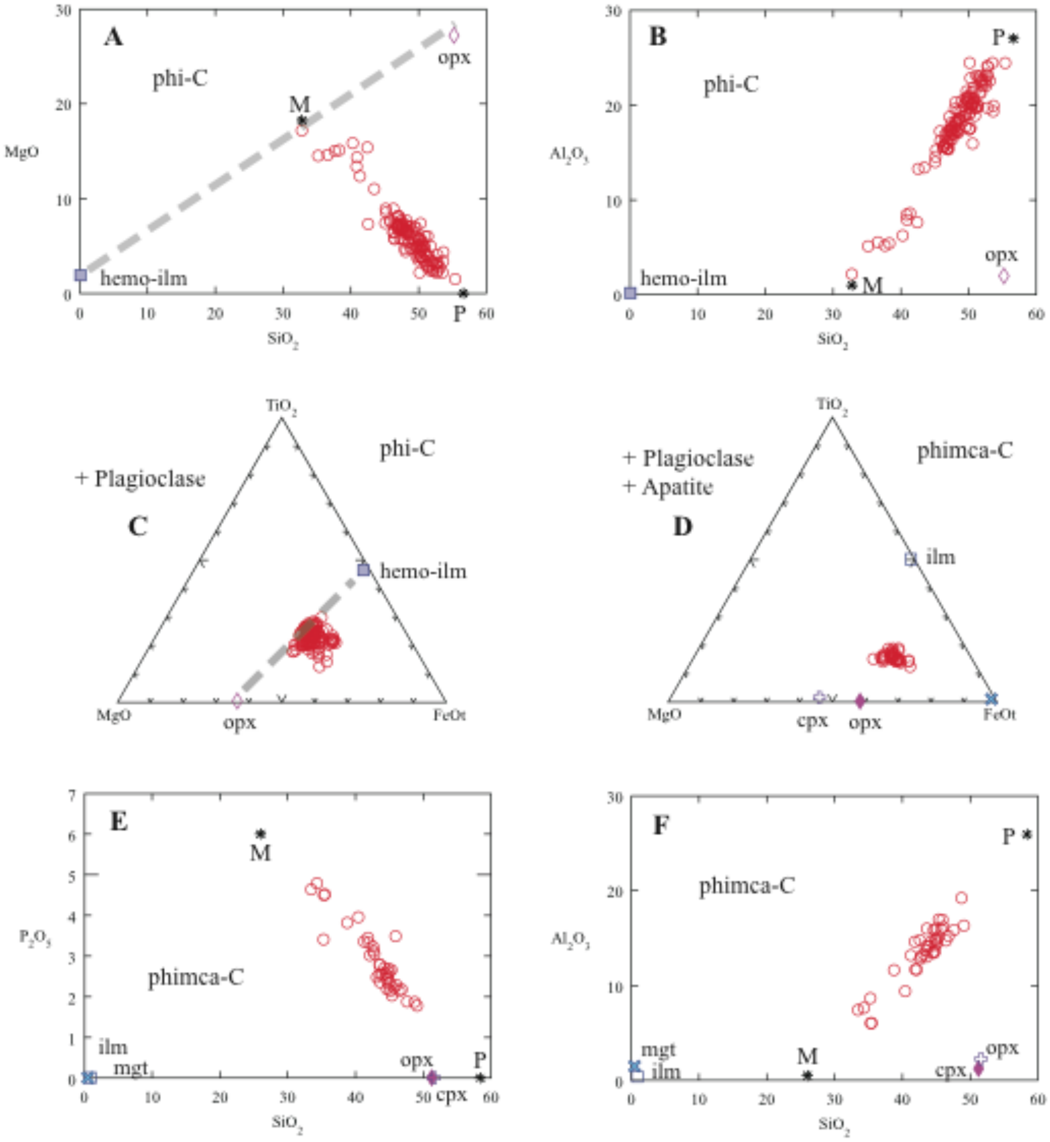

Fig. 1

Figure 1: Characteristic features of 2-pole cumulates. A and B: Harker diagrams in which 124 samples of phi-C compositions plot on a straight line joining a plagioclase (P) and a mafic pole (M) made up of Ca-poor pyroxene (opx) and hemo-ilmenite (hemo-ilm). C: the same compositions projected in MgO-TiO_2-FeO_t diagram. D: 41 phimca-C compositions in MgO-TiO_2-FeO_t diagram . E: 41 phimca-C compositions plotted in P_2O_5-SiO_2 Harker diagram showing the role of apatite in the mafic pole (M). F: the same compositions in Al_2O_3-SiO_2 diagram showing the mixing relationship between the plagioclase pole (P) and the mafic pole (M). Abbrev.: cpx: Ca-rich pyroxene, ilm: ilmenite, mgt: magnetite. For more details see Duchesne and Charlier, 2005)

Fig. 2

Figure 2: A thin inclusion ("stick-like") of fine grained noritic gneiss in layered gabbronorite along the western shore of the Teksevatn (MCU IV, BKSK intrusion). The inclusion is resting oblique to the layer surface and has not deformed the layering plane. Sedimentation has not played any role. It is coated by a thin rim of mafic minerals resembling the thin mafic layer that is interrupted. The coating of the inclusion can have formed when it was falling in the melt or after it was deposited. Note the small scale modally graded layer at the bottom of the picture. Top of the picture towards the top of the intrusion.

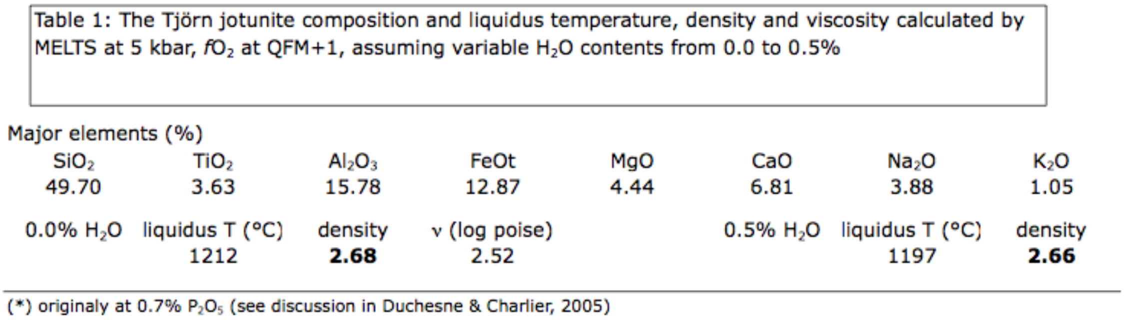

Table 1: The Tjörn jotunite composition and liquidus temperature, density and viscosity calculated by MELTS at 5 kbar, fO_2 at QFM+1, assuming variable H_2O contents from 0.0 to 0.5%

Major elements (%)

SiO_2	TiO_2	Al_2O_3	FeOt	MgO	CaO	Na_2O	K_2O
49.70	3.63	15.78	12.87	4.44	6.81	3.88	1.05
0.0% H_2O	liquidus T (°C)	density	ν (log poise)		0.5% H_2O	liquidus T (°C)	density
	1212	**2.68**	2.52			1197	**2.66**

(*) originaly at 0.7% P_2O_5 (see discussion in Duchesne & Charlier, 2005)