Magma evolution in the upper part (Stage IV) of the Fongen-Hyllingen Layered Intrusion, central Norway

Mohamed M. ABU EL-RUS^{*,**}, J. Richard WILSON^{***} and Henning S. SØRENSEN^{****,*****}

^{*}*Physics of Geological Processes, University of Oslo, PO Box 1048 Blindern, N-0316 Oslo, Norway*

^{**}*Present address: Geology Department, Assiut University, Assiut 71615, Egypt*

^{***}*Department of Earth Sciences, Aarhus University, 8000 Århus C, Denmark*

^{****}*Present address: Danfoss A/S, DK-6430 Nordborg, Denmark*

The upper part (Stage IV) of the Fongen-Hyllingen intrusion, southeast of Trondheim, Norway, comprises a ~ 1800 m thick sequence of cumulates showing a strong normal fractionation trend from olivine gabbros at the bottom through diorites to quartz-bearing syenites at the roof. Mineral compositions show extreme ranges: olivine Fo_{74-0} ; Ca-rich pyroxene Mg_{79-0} ; plagioclase An_{60-0} . The cumulate sequence has been divided into five intervals based on cumulus assemblages. The summation method has been used to calculate successive magma compositions for these intervals based on whole-rock major element chemical analyses of 49 samples. The most primitive calculated magma composition has 51.7% SiO_2 , 5.0% MgO , 5.3% total alkalis and a Mg-number of 47.5; it lies near the base of the basaltic trachyandesite field in the Total Alkali Silica (TAS) diagram, very close to the dividing line between tholeiitic and alkaline suites. This magma evolved through the trachyandesite field into the trachyte field. The most evolved magma composition has ~ 11.4% alkalis; alkali enrichment is not due to assimilation of metapelites. There was very little iron-enrichment during fractionation and magma density decreased continuously, enabling repeated influxes of new magma to pond and to elevate the resident melt, facilitating magma stratification. The parental magma may have resided in a lower crustal chamber where it differentiated before repeatedly supplying the overlying Fongen-Hyllingen chamber.

Keywords: Layered intrusion, Summation method, Parental magma, Olivine hiatus, Density evolution

INTRODUCTION

Petrologists encounter several difficulties in reconstructing magma evolution in layered intrusions. The chemical compositions of cumulates are principally controlled by the modal proportions of the cumulus minerals and interstitial melts trapped during crystallization (e.g., Maier and Barnes, 1998; Bédard, 1994). Assimilation of material from surrounding rocks may also play a significant role in magma evolution (e.g., DePaolo, 1981, 1985; Huppert and Sparks, 1985; Marks and Markl, 2001). Injections of fresh magma into an existing chamber can modify the composition of the evolving magma (Wiebe and Snyder, 1993; Wilson and Sørensen, 1996; Wilson et al., 1996). Migration of reactive pore melts within the cumulates may result in resorption of cumulus minerals and the crystallization of new ones, and generally re-distribute mass within the intrusion (e.g., Robins, 1982; Boudreau, 1999; Brandriss and Bird, 1999). Furthermore, cumulates in lay-

ered intrusions do not preserve magma compositions but they preserve a time-integrated record of the product of fractional crystallization at, or close to, the floor or walls of the chamber.

The upper part (Stage IV) of the Fongen-Hyllingen layered intrusion is characterized by a strong normal fractionation trend extending continuously from the most primitive compositions in the intrusion to extremely evolved compositions at the roof with no significant breaks or regressions (Wilson and Larsen, 1985; Sørensen and Wilson, 1995; Wilson and Sørensen, 1996). This enables us to deduce the nature of the parental magma and its liquid line of descent, and to reconstruct the density evolution of the magma. We document the major element composition of the cumulates and the mineral chemistry in Stage IV cumulates of the Fongen-Hyllingen intrusion and discuss the evolution of the magma chamber.

THE FONGEN-HYLLINGEN INTRUSION

The Fongen-Hyllingen intrusion (FHI) is the largest layered intrusion in the Trondheim Region of the Norwegian Caledonides, located at 62°55'–63°12'N and 11°41'–11°25'E (Fig. 1). The intrusion was emplaced into folded metavolcanics and metapelites in the late Silurian (Wilson et al., 1983). A swarm of mafic dykes (commonly plagioclase-phryic) intruded these deformed country rocks before emplacement of the FHI. Mineral parageneses in a well-developed contact metamorphic aureole indicate that emplacement took place at 3.5 ± 0.5 kb (Thayssen, 1998), prior to regional amphibolite facies metamorphism and penetrative deformation (Olesen et al., 1973). The intrusion has a crystallization age of $426 \pm 8/-2$ Ma based on a zircon U-Pb age determination of the final differentiates (Wilson et al., 1983). A Rb-Sr whole-rock isochron on the same rocks gives an age of 405 ± 9 Ma (Wilson and Pedersen, 1981) which is believed to record cooling after late Caledonian regional metamorphism (Wilson et al., 1983).

The FHI occupies an area of about 160 km² and comprises a ~6 km thick sequence of cumulates spanning a lithological range from troctolites and olivine gabbros to quartz-bearing syenites. The intrusion lacks a chilled margin and coarse grained massive dioritic rocks are commonly developed at the margins (Wilson et al., 1981). A possible feeder to part of the intrusion may be present in the extreme northwest (Wilson and Olesen, 1975; Wilson et al., 1981). The rest of the intrusion comprises the layered series. Abundant large, raft-like inclusions are an important feature in FHI, particularly in the stratigraphically lower part of the layered series. Most of the inclusions are of metabasaltic hornfels, but metapelites are locally present, consistent with the lithology of the wall rocks.

The FHI has been divided into the Fongen Series in the north and the Hyllingen Series in the south. These are separated by a narrow zone consisting of strongly deformed and metamorphosed rocks (Fig. 1). Wilson and Larsen (1985) subdivided the Hyllingen Series into four evolutionary stages on the basis of mineral chemical vari-

Figure 1. (A) Location of the Fongen-Hyllingen complex in Norway. (B) Simplified geological map after Wilson and Sørensen (1996). (C) Schematic illustration of the cryptic variation along profiles normal to modal layering. The cryptic variation is shown by Fo% in olivines, Mg# in pyroxenes, An% in plagioclase and Mg# in calcic amphiboles. (D) Distribution of Stages I to IV in the Fongen-Hyllingen intrusion after Sørensen and Wilson (1995). Sample profiles F1 and H2 are used in the present study of the evolution of Stage IV.

Table 1. Representative average analyses of plagioclase cores

Sample:	Wr.F1	Wr.F14	Wr.F17	Wr.H1	Wr.H10	Wr.H13	Wr.H15	Wr.H16	Wr.H17	Wr.H18
Height:	0 m	100 m	190 m	300 m	380 m	470 m	540 m	620 m	819 m	936 m
SiO ₂	53.03	53.46	53.46	54.32	54.84	56.01	55.56	56.24	57.18	58.08
Al ₂ O ₃	29.51	28.97	28.76	28.27	28.17	27.51	27.49	27.34	26.66	26.15
FeO	0.40	0.13	0.16	0.19	0.26	0.27	0.21	0.14	0.13	0.04
CaO	11.99	11.74	11.88	10.83	10.77	10.14	10.06	9.50	8.84	8.14
Na ₂ O	4.45	4.85	4.85	5.50	5.69	5.88	5.88	6.16	6.50	6.67
K ₂ O	0.08	0.11	0.16	0.17	0.18	0.18	0.19	0.18	0.17	0.22
Total	99.46	99.26	99.27	99.28	99.91	99.99	99.42	99.56	99.48	99.30
Cations per 8 oxygens										
Si	2.415	2.434	2.434	2.462	2.467	2.518	2.513	2.533	2.575	2.619
Al	1.584	1.554	1.543	1.510	1.494	1.458	1.465	1.451	1.415	1.390
Fe	0.015	0.004	0.005	0.006	0.009	0.009	0.007	0.005	0.005	0.002
Ca	0.585	0.573	0.580	0.526	0.519	0.487	0.488	0.459	0.427	0.394
Na	0.393	0.428	0.428	0.483	0.496	0.512	0.515	0.538	0.567	0.583
K	0.005	0.006	0.009	0.010	0.010	0.011	0.011	0.011	0.010	0.013
Total	4.997	4.999	4.999	4.997	4.995	4.995	4.999	4.997	4.999	5.001
An%	59.51	56.90	57.03	51.62	50.63	48.22	48.13	45.54	42.53	39.80
Ab%	39.98	42.50	42.08	47.40	48.39	50.69	50.79	53.37	56.47	58.89
Or%	0.51	0.60	0.88	0.98	0.98	1.09	1.08	1.09	1.00	1.31

Sample:	Wr.H21	Wr.H24	Wr.H27	Wr.H31
Height:	1069 m	1200 m	1300 m	1720 m
SiO ₂	58.23	58.98	62.62	69.14
Al ₂ O ₃	26.09	25.41	24.66	19.76
FeO	0.03	0.04	b.d.l.	0.03
CaO	8.14	7.35	5.74	0.08
Na ₂ O	6.86	7.37	6.79	11.42
K ₂ O	0.24	0.06	0.06	0.12
Total	99.59	99.21	99.87	100.55
Cations per 8 oxygens				
Si	2.614	2.651	2.819	3.009
Al	1.380	1.346	1.308	1.014
Fe	0.001	0.002	0.000	0.001
Ca	0.392	0.354	0.277	0.004
Na	0.597	0.643	0.593	0.963
K	0.014	0.004	0.004	0.007
Total	4.998	5.000	5.001	4.998
An%	39.08	35.36	31.69	0.41
Ab%	59.52	64.24	67.85	98.87
Or%	1.40	0.40	0.46	0.72

Averages of usually three point analyses in the cores of each of three plagioclase grains.

b.d.l. means below detection limit.

ations. These comprise from base to top: Stage I (< 100 m thick) that consists of mostly non-layered ferrodiorites in which there is a compositional (basal) reversal; Stage II (400–1500 m) that comprises layered gabbros/diorites with fairly constant mineral compositions ending with a trend to more evolved rocks; Stage III (350–550 m) comprises a gradual regression to more primitive compositions, ending with the most primitive compositions in the intrusion; and Stage IV (800–2300 m) which exhibits a normal cryptic zoning trend with extremely evolved rocks at the top in direct contact with contact metamorphosed

country rocks forming the roof. Lateral correlation between the Fongen and Hyllingen Series has been established on the basis of the strike of modal layering, the distribution of raft-like metabasaltic inclusions and, most importantly, the continuous occurrence of Stage III (Sørensen and Wilson, 1995; Wilson and Sørensen, 1996) (Fig. 1). The lower part of Stage IV in the Ruten area of the Fongen Series contains several olivine-rich units (Fig. 1B) that were formed as a result of magma mixing during minor episodes of magma addition (Meyer and Wilson, 1999). A slightly younger intrusion of mafic to ultramafic rocks, the Treknattan Intrusion (Fig. 1B), cuts the northern part of the FHI (Sørensen and Wilson, 1996).

An important feature of the FHI is the presence of systematic lateral variations in mineral compositions. Solid solution minerals gradually become more evolved along the strike of the modal layering towards the southern margin in the Hyllingen Series. For example, olivine and plagioclase compositions vary systematically over about 7 km from Fo₂₃:Fo₇ and An₄₆:An₃₅ along the Stage II/III boundary and Fo₇₅:Fo₁₃ and An₆₃:An₄₂ along the Stage III/IV boundary (Wilson and Larsen, 1985). Wilson and Larsen (1985) suggested that these discordant relations between modal and cryptic layering developed as a result of the in situ crystallization of compositionally zoned magma along an inclined magma chamber floor.

SAMPLING AND ANALYTICAL METHODS

A total of 49 Stage IV samples were collected along two profiles (Fig. 1). Collecting appropriate samples for chemical analysis from layered intrusions may be complicated by the presence of modal layering. In the FHI, modal lay-

Table 2. Representative average analyses of olivine cores

Sample: Height:	Wr.F1 0 m	Wr.F14 100 m	Wr.F17 190 m	Wr.H15 540 m	Wr.H16 620 m	Wr.H17 819 m	Wr.H18 936 m	Wr.H21* 1069 m	Wr.H23* 1170 m	Wr.H27* 1300 m
SiO ₂	38.02	37.57	37.14	34.14	33.69	33.93	33.66	29.99	30.66	32.07
FeO	23.52	27.08	28.37	46.15	46.06	46.65	47.33	64.14	63.86	63.91
MnO	0.39	0.44	0.41	0.74	0.68	0.77	0.74	1.67	1.02	1.46
MgO	37.62	34.84	34.01	19.25	18.87	18.31	17.77	3.58	3.43	1.99
NiO	0.10	0.10	0.08	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.
Total	99.65	100.03	100.01	100.28	99.3	99.66	99.50	99.38	98.97	99.43
Cations per 4 oxygens										
Si	0.999	1.001	0.995	1.003	0.999	1.009	1.006	0.995	1.009	1.067
Fe ²⁺	0.517	0.603	0.635	1.134	1.143	1.160	1.183	1.780	1.757	1.778
Mn	0.009	0.010	0.009	0.018	0.017	0.020	0.02	0.047	0.028	0.041
Mg	1.473	1.383	1.358	0.843	0.834	0.811	0.792	0.177	0.168	0.099
Ni	0.002	0.002	0.002	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	3.000	2.999	2.999	2.998	2.993	3.000	3.001	2.999	2.962	2.985
Fo%	74.02	69.64	68.14	42.64	42.20	41.15	40.10	9.04	8.73	5.27

Averages of usually three point analyses in the cores of each of three olivine grains.

* Remnants of olivines oxidized to symplectitic intergrowths of magnetite and Ca-poor pyroxene or grunerite.

b.d.l. means below detection limit.

ering is commonly developed on a scale between thicknesses of 2 to 30 cm. An attempt was made to collect "average cumulates", avoiding mafic or plagioclase-rich variants. It is, however, inevitable that some of the compositional variation reflects local modal variations. Stratigraphically the samples span from the most primitive compositions in FHI at the base of Stage IV to the most evolved compositions (quartz-bearing syenites) at the roof. There is evidence for several minor episodes of magma replenishment in the lower part of Stage IV in the northern part of the intrusion; this was responsible for the formation of several olivine-rich units as a result of magma mixing (Meyer and Wilson, 1999). Our sampling profile in the northern part of the intrusion (along F1 in Fig. 1) is located below the first olivine-rich unit, thereby avoiding problems caused by magma addition. Mineral compositions were acquired by a JEOL-733 electron microprobe at the Department of Earth Sciences, Aarhus University, using natural and synthetic glasses and minerals as standards. Operating conditions were 15 keV accelerating voltage, 20 nA beam current, 1–10 µm beam diameter, depending on the nature and size of mineral grains, and 20s counting time for each element. Results were corrected with a ZAF matrix correction program. Fe²⁺ and Fe³⁺ were distributed based on the stoichiometry. Representative mineral analyses are presented in Tables 1–5.

Major element analyses of whole rocks were determined on a Phillips PW 1220 XRF spectrometer at the Department of Earth Sciences, Aarhus University, Denmark. Analyses of SiO₂, TiO₂, Al₂O₃, Fe₂O₃ (total), CaO and K₂O were performed on glass discs, prepared by fusion with a lithium tetraborate and lithium metaborate mixture. P₂O₅ was analysed on pressed powder pellets,

whereas Na₂O, MnO and MgO were analysed by atomic absorption. FeO was determined by titration. The analytical results are presented in Table 6.

PETROGRAPHY AND MINERAL CHEMISTRY

The lateral compositional variations explained above mean that no single profile normal to modal layering covers the entire compositional range in Stage IV. The most primitive rocks at the base of Stage IV are present in the Ruten area where several olivine-rich units reflect magma mixing in the proximity of a feeder. Compositional reversals in Stage IV in this region reflect repeated events of magma addition that, combined with deformation and erosion, result in the upper part of Stage IV not being developed in the Fongen Series. The uppermost part of Stage IV is exclusively developed in the eastern part of the Hyllingen Series. In Figure 2 there is a compositional gap between the lower part of Stage IV in the Fongen Series and Stage IV in the Hyllingen Series (between the top of F1 and the base of H2 in Fig. 1D). This is a consequence of the lateral compositional variations. It is possible that this gap could be filled with additional samples from the upper part of Stage IV in the Fongen Series or north of profile H2 in the Hyllingen Series, but our present material does not cover this stratigraphic interval. We estimate that the compositional gap represents a thickness of about 110 m. This seems reasonable on the basis of the mineral compositional trends shown in Figure 3 and is consistent with the size of the gap estimated on the basis of the trace element compositional variation by Abu El-Rus (2003).

Plagioclase and Ca-rich pyroxene are cumulus

Table 3. Representative average analyses of Ca-rich pyroxene cores

Sample:	Wr.F1	Wr.F14	Wr.F17	Wr.H1	Wr.H10	Wr.H13	Wr.H15	Wr.H16	Wr.H17
Height:	0 m	100 m	190 m	300 m	380 m	470 m	540 m	620 m	819 m
SiO ₂	51.00	50.42	50.12	50.42	50.25	50.68	50.59	50.64	50.40
TiO ₂	1.05	0.85	1.10	0.85	0.95	0.80	0.70	0.65	0.73
Al ₂ O ₃	3.29	3.65	3.89	3.51	3.23	3.10	2.72	2.42	2.59
Cr ₂ O ₃	0.29	0.26	0.26	0.04	b.d.l.	0.01	0.04	b.d.l.	0.04
Fe ₂ O ₃ *	1.81	2.36	1.66	2.01	2.07	2.01	1.41	1.75	1.41
FeO	6.23	5.98	6.19	7.84	8.90	8.19	10.46	11.12	10.71
MnO	0.26	0.21	0.25	0.25	0.26	0.28	0.23	0.34	0.37
MgO	15.95	15.09	14.16	13.28	13.30	13.21	12.47	12.30	11.70
CaO	19.20	19.81	20.89	20.57	20.16	21.18	20.64	20.23	20.95
Na ₂ O	0.50	0.55	0.53	0.62	0.49	0.46	0.38	0.38	0.46
Total	99.58	99.18	99.48	99.05	99.61	99.92	99.64	99.83	99.36
Cations per 6 oxygens									
Si	1.887	1.879	1.875	1.894	1.890	1.897	1.913	1.917	1.917
Ti	0.029	0.024	0.031	0.020	0.027	0.023	0.020	0.019	0.021
Al	0.143	0.160	0.172	0.156	0.143	0.137	0.121	0.108	0.116
Cr	0.009	0.008	0.008	0.001	0.000	0.001	0.001	0.000	0.001
Fe ⁺³	0.051	0.067	0.047	0.057	0.059	0.057	0.040	0.050	0.041
Fe ⁺²	0.196	0.186	0.194	0.246	0.280	0.257	0.331	0.352	0.342
Mn	0.008	0.007	0.008	0.008	0.008	0.009	0.007	0.011	0.012
Mg	0.880	0.838	0.79	0.743	0.745	0.737	0.703	0.694	0.664
Ca	0.761	0.791	0.836	0.827	0.812	0.850	0.836	0.821	0.853
Na	0.036	0.040	0.038	0.045	0.035	0.033	0.028	0.028	0.034
Total	4.000	4.000	3.999	3.997	3.999	4.001	4.000	4.000	4.001
Wo%	40.42	42.18	44.891	44.29	42.96	44.85	43.86	42.94	44.99
En%	46.74	44.69	42.421	39.79	39.42	38.89	36.88	36.30	35.02
Fs%	12.85	13.13	12.688	15.92	17.62	16.27	19.25	20.76	19.98
Mg#	78.50	77.29	76.976	71.42	69.10	70.51	65.70	63.61	63.67

Sample:	Wr.H18	Wr.H21	Wr.H27	Wr.H31
Height:	936 m	1069 m	1300 m	1720 m
SiO ₂	50.02	50.51	49.29	47.58
TiO ₂	0.68	0.47	0.27	0.18
Al ₂ O ₃	2.82	1.53	1.19	0.59
Cr ₂ O ₃	0.06	0.03	0.04	0.04
Fe ₂ O ₃ *	1.96	0.000	0.37	1.83
FeO	12.56	18.08	23.29	28.59
MnO	0.37	0.61	0.77	1.01
MgO	11.26	9.07	4.81	b.d.l.
CaO	19.84	18.50	19.21	18.50
Na ₂ O	0.43	0.36	0.39	0.81
Total	100.00	99.16	99.63	99.13
Cations per 6 oxygens				
Si	1.904	1.974	1.974	1.978
Ti	0.020	0.014	0.008	0.006
Al	0.126	0.071	0.056	0.029
Cr	0.002	0.001	0.001	0.001
Fe ⁺³	0.056	0.000	0.011	0.057
Fe ⁺²	0.400	0.591	0.780	0.997
Mn	0.012	0.020	0.026	0.036
Mg	0.639	0.528	0.287	0.000
Ca	0.809	0.775	0.824	0.826
Na	0.032	0.028	0.030	0.065
Total	4.000	4.002	3.997	3.995
Wo%	42.62	40.92	43.35	44.07
En%	33.66	27.88	15.10	0.00
Fs%	23.72	31.20	41.55	55.93
Mg#	58.66	47.18	26.65	~ 0.00

Averages of usually three point analyses in the cores of each of three Ca-rich pyroxene grains.

Fe₂O₃ calculated in Ca-rich pyroxene assuming 4 cations and 6 oxygens stoichiometry. b.d.l. means below detection limit.

phases throughout Stage IV (Fig. 2). Olivine is present through a large part of the sequence but is absent for some 250 m at the base of H2 and in the most evolved rocks below the roof. Ca-poor pyroxene is an intercumulus phase in the lower part of F1 but is a persistent cumulus phase in the upper part of F1 and H2 to ~ 500 m below the roof. FeTi-oxides are intercumulus phases in F1 and in the lowest part of H2. At ~ 350 m they have cumulus status, continuing into the most evolved rocks at the top of H2. They are only present in small quantities in the uppermost ~ 250 m. Calcic amphibole is generally present as an intercumulus phase through the lower ~ 950 m, above which it is an early crystallising phase. Apatite joins the cumulus assemblage at about the same level as cumulus amphibole. Cumulus zircon appears at ~ 1200 m, followed successively by biotite, K-feldspar, quartz and allanite. Quartz occurs as a minor, sporadic intercumulus phase above ~ 1200 m and increases upwards, reaching some 8 vol% just below the roof.

The amount of intercumulus material is usually minor (consisting largely of Ca-amphibole and/or FeTi-oxides in the lower part) but overgrowths, particularly on

Table 4. Representative average analyses of Ca-poor pyroxene cores

Sample: Height:	Wr.F1 0 m	Wr.F14 100 m	Wr.F17 190 m	Wr.H1 300 m	Wr.H10 380 m	Wr.H13 470 m	Wr.H15 540 m	Wr.H16 620 m	Wr.H17 819 m
SiO ₂	53.65	53.44	53.27	52.35	51.58	51.84	51.54	50.89	51.22
TiO ₂	0.27	0.30	0.33	0.27	0.37	0.27	0.12	0.28	0.27
Al ₂ O ₃	0.64	1.98	1.50	1.00	1.11	1.19	0.53	1.11	1.10
Cr ₂ O ₃	0.13	0.07	0.04	b.d.l.	b.d.l.	0.06	b.d.l.	0.03	0.03
Fe ₂ O ₃ *	2.28	0.45	1.34	0.48	1.03	0.000	0.53	0.51	0.000
FeO	13.61	16.89	16.28	22.70	22.32	24.02	26.86	26.86	27.14
MnO	0.37	0.36	0.46	0.56	0.52	0.59	0.68	0.63	0.59
MgO	27.61	25.42	25.8	21.44	20.79	20.34	18.40	17.82	17.68
CaO	0.92	1.15	0.99	1.04	1.50	0.87	1.02	1.43	1.47
Na ₂ O	b.d.l.	0.03	0.03	b.d.l.	0.04	0.04	0.01	b.d.l.	0.03
Total	99.48	100.09	100.02	99.84	99.26	99.22	99.69	99.56	99.53
Cations per 6 oxygens									
Si	1.946	1.943	1.94	1.964	1.952	1.967	1.978	1.959	1.972
Ti	0.007	0.008	0.009	0.008	0.010	0.008	0.003	0.008	0.008
Al	0.028	0.085	0.064	0.044	0.050	0.053	0.024	0.051	0.050
Cr	0.004	0.002	0.001	0.000	0.000	0.002	0.000	0.001	0.001
Fe ⁺³	0.062	0.012	0.037	0.014	0.029	0.000	0.015	0.015	0.000
Fe ⁺²	0.413	0.514	0.496	0.712	0.706	0.762	0.862	0.865	0.874
Mn	0.012	0.011	0.014	0.018	0.017	0.019	0.022	0.021	0.019
Mg	1.493	1.3775	1.40	1.199	1.173	1.151	1.053	1.022	1.014
Ca	0.036	0.045	0.039	0.0416	0.061	0.035	0.042	0.059	0.061
Na	0.000	0.002	<0.001	0.000	0.003	0.003	0.001	0.000	0.002
Total	4.001	4.000	4.000	4.001	4.001	4.000	4.000	4.001	4.001
Wo%	1.80	2.31	2.02	2.12	3.10	1.80	2.13	3.01	3.13
En%	74.73	70.74	72.35	61.01	59.66	59.09	53.44	52.16	52.03
Fs%	23.47	26.95	25.63	36.87	37.24	39.12	44.43	44.83	44.84
Mg#	76.10	72.41	73.84	62.33	61.57	60.17	54.60	53.78	53.71

Sample: Height:	Wr.H18 936 m	Wr.H21 1069 m	Wr.H24 1200 m	Wr.H27 1300 m
SiO ₂	49.89	48.22	47.17	47.02
TiO ₂	0.23	0.17	0.03	0.05
Al ₂ O ₃	0.76	0.64	0.32	0.42
Cr ₂ O ₃	0.04	0.01	b.d.l.	0.01
Fe ₂ O ₃ *	0.13	0.26	1.68	b.d.l.
FeO	32.88	39.32	41.82	46.51
MnO	0.77	1.10	1.54	1.94
MgO	13.61	8.89	6.90	3.02
CaO	1.51	1.20	0.49	0.91
Na ₂ O	b.d.l.	b.d.l.	0.03	0.13
Total	99.82	99.81	99.98	100.01
Cations per 6 oxygens				
Si	1.973	1.976	1.966	2.005
Ti	0.007	0.005	0.001	0.002
Al	0.035	0.031	0.016	0.021
Cr	0.001	0.001	0.000	0.001
Fe ⁺³	0.004	0.008	0.053	0.000
Fe ⁺²	1.087	1.347	1.458	1.658
Mn	0.026	0.038	0.054	0.070
Mg	0.802	0.529	0.429	0.192
Ca	0.064	0.054	0.022	0.042
Na	0.000	0.000	0.002	0.011
Total	3.999	3.989	4.001	4.002
Wo%	3.27	2.74	1.12	2.22
En%	40.99	27.28	21.92	10.15
Fs%	55.74	69.98	76.95	87.63
Mg#	42.38	28.05	22.17	10.38

Averages of usually three point analyses in the cores of each of three Ca-poor pyroxene grains.

Fe₂O₃ calculated in Ca-poor pyroxene assuming 4 cations and 6 oxygens stoichiometry. b.d.l. means below detection limit.

plagioclases and locally on pyroxenes, are common. Petrographic observations indicate that the total amount of postcumulus material is generally 2-10% and the rocks are adcumulates or mesocumulates according to the nomenclature of Irvine (1982).

The broadly gabbroic rocks in the lowermost 200 m vary from troctolites, olivine gabbros and olivine gabbronorites to local anorthositic layers (with some dunitic layers in the olivine-rich units). When plagioclase has a composition An_{<50} the rocks become dioritic (e.g., ferrodiorites, olivine ferrodiorites). When K-feldspar and quartz become intercumulus phases the rocks are quartz-bearing monzodiorites, grading to quartz-bearing monzonites when the proportion of K-feldspar increases. The uppermost rocks are quartz-bearing syenites and locally quartz syenites. Most of the rocks have cumulate textures, but when K-feldspar becomes an early crystallising phase, modal layering is no longer developed and the rocks can be texturally classified as hypidiomorphic granular. The thickness of the uppermost non-layered rocks has been estimated on the basis of a consistent orientation of layering from its uppermost occurrence to the roof.

Table 5. Representative average analyses of calcic amphiboles

Sample:	Wr.F1	Wr.F14	Wr.F17	Wr.H1	Wr.H10	Wr.H13	Wr.H15	Wr.H16	Wr.H17
Height:	0 m	100 m	190 m	300 m	380 m	470 m	540 m	620 m	819 m
SiO ₂	42.10	42.23	41.74	42.02	41.61	41.89	40.50	40.82	40.86
TiO ₂	1.93	1.47	1.97	3.57	1.27	3.47	1.43	3.00	1.70
Al ₂ O ₃	13.83	13.40	13.53	11.56	14.34	11.51	14.81	12.73	13.28
Cr ₂ O ₃	0.29	0.13	0.06	0.10	b.d.l.	0.04	0.01	0.04	0.03
Fe ₂ O ₃	0.000	1.09	0.000	0.000	0.000	0.000	0.000	0.000	0.000
FeO	8.83	10.30	10.24	11.57	14.47	15.08	15.99	16.08	16.65
MnO	0.13	0.14	0.12	0.10	0.14	0.14	0.10	0.21	0.22
MgO	14.36	13.56	14.06	12.65	10.86	11.08	9.37	10.33	9.72
CaO	12.05	11.71	11.63	11.46	11.42	11.54	11.19	10.40	10.97
Na ₂ O	2.47	2.09	2.51	2.37	2.04	2.21	2.01	2.57	2.66
K ₂ O	0.82	0.73	0.70	0.72	0.75	0.66	0.87	0.55	0.59
Total	96.81	96.85	96.56	96.12	96.90	97.62	96.28	96.73	96.68
Cations per 23 oxygens									
Si	6.198	6.235	6.180	6.347	6.233	6.301	6.169	6.230	6.244
Ti	0.214	0.163	0.219	0.406	0.143	0.393	0.164	0.345	0.196
Al ^{iv}	1.802	1.765	1.820	1.653	1.767	1.699	1.831	1.770	1.756
Al ^{vi}	0.598	0.566	0.541	0.405	0.765	0.341	0.829	0.521	0.636
Cr	0.034	0.015	0.007	0.012	0.000	0.004	0.002	0.005	0.004
Fe ³⁺	0.000	0.121	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ²⁺	1.087	1.272	1.268	1.461	1.813	1.897	2.037	2.053	2.127
Mn	0.016	0.018	0.015	0.013	0.018	0.018	0.013	0.027	0.028
Mg	3.151	2.985	3.103	2.848	2.425	2.483	2.127	2.350	2.213
Ca	1.900	1.853	1.845	1.855	1.832	1.860	1.827	1.700	1.796
Na	0.704	0.598	0.720	0.827	0.591	0.645	0.593	0.762	0.787
K	0.154	0.138	0.132	0.139	0.143	0.123	0.169	0.108	0.115
Total	15.858	15.729	15.849	15.966	15.729	15.763	15.762	15.870	15.902
Mg#	0.744	0.701	0.710	0.661	0.572	0.567	0.511	0.534	0.510

Sample:	Wr.H18	Wr.H24	Wr.H27	Wr.H31
Height:	936 m	1200 m	1300 m	1720 m
SiO ₂	42.19	40.35	40.45	41.35
TiO ₂	2.87	1.90	2.19	1.77
Al ₂ O ₃	10.51	10.34	9.58	6.97
Cr ₂ O ₃	0.10	b.d.l.	b.d.l.	0.03
Fe ₂ O ₃	0.000	0.000	0.000	0.000
FeO	21.29	27.69	29.18	33.58
MnO	0.28	0.53	0.57	1.01
MgO	7.86	3.20	2.75	b.d.l.
CaO	9.89	10.68	9.92	9.98
Na ₂ O	1.73	1.60	1.63	1.51
K ₂ O	0.67	0.95	1.01	1.12
Total	97.39	97.24	97.28	97.32
Cations per 23 oxygens				
Si	6.513	6.450	6.525	6.849
Ti	0.333	0.229	0.265	0.220
Al ^{iv}	1.487	1.550	1.475	1.151
Al ^{vi}	0.424	0.399	0.346	0.210
Cr	0.013	0.000	0.000	0.004
Fe ³⁺	0.000	0.000	0.000	0.000
Fe ²⁺	2.749	3.704	3.936	4.651
Mn	0.037	0.072	0.078	0.141
Mg	1.808	0.760	0.662	0.000
Ca	1.636	1.830	1.714	1.770
Na	0.561	0.561	0.510	0.485
K	0.133	0.194	0.208	0.237
Total	15.694	15.749	15.718	15.718
Mg#	0.397	0.170	0.144	~ 0.000

Plagioclase

Plagioclase cores (average of 3 points in each of 3 grains) exhibit a fairly systematic compositional variation with stratigraphic height (Fig. 3) from An_{59.5} at the bottom of Stage IV to nearly pure albite (An_{0.4}) in the quartz-bearing syenites (Fig. 3). This extremely sodic plagioclase occurs as discrete subhedral grains and is not a product of exsolution from alkali feldspar. The plagioclase composition changes from ~ An₆₀ to An₃₂ through the lower ca. 1300 m, and from An₃₂ to An_{0.4} through the upper ~ 500 m. The orthoclase content of plagioclase increases gradually upwards from Or_{0.5} to Or_{1.4} at 1069 m and then decreases abruptly to Or_{0.4} when alkali-feldspar becomes a cumulus phase (Table 1). Total iron in plagioclase (0.40-0% FeO^{TOT}) shows an overall decrease with falling An-content (Table 1). In thin section, some plagioclases exhibit normal zonation at the margins of grains with composition differences up to 7 An%. Weak oscillatory zoning in plagioclase is locally developed, particularly in the most evolved rocks. The sub-parallel trends for plagioclase, pyroxene and olivine compositions throughout the sequence in Figure 3 indicate that these phases are essen-

Averages of usually three point analyses in the cores of each of three calcic amphibole grains.

Fe₂O₃ calculated in calcic amphibole assuming 16 cations and 23 oxygens stoichiometry. b.d.l. means below detection limit.

Table 6. Whole-rock analyses (in wt%) of 49 samples from

Sample no.	Height (m)	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O
Wr.F1	0	44.49	0.20	10.48	2.19	12.38	0.23	20.59	7.17	1.62
Wr.F2	1	43.78	0.37	4.67	2.21	13.93	0.29	24.21	8.28	0.69
Wr.F3	4	43.30	0.20	9.09	0.93	14.09	0.24	21.27	6.39	1.34
Wr.F4	11	49.73	0.62	8.23	1.05	7.99	0.23	13.94	14.66	0.90
Wr.F5	14	52.18	0.21	19.99	0.19	5.55	0.13	7.28	9.14	3.51
Wr.F6	17	50.84	0.36	16.78	0.66	5.65	0.15	9.14	10.57	2.64
Wr.F7	23	48.31	0.24	17.22	0.77	7.92	0.15	12.42	10.28	2.60
Wr.F8	29	49.24	0.29	17.03	0.94	6.67	0.14	10.54	11.23	2.68
Wr.F9	32	49.19	0.43	12.95	1.25	8.09	0.19	13.30	11.59	2.04
Wr.F10	40	49.20	0.40	5.49	1.66	13.19	0.32	20.68	8.11	0.78
Wr.F11	42	49.26	0.32	18.33	0.67	5.81	0.12	10.05	12.06	2.63
Wr.F12	60	46.28	0.19	11.84	0.97	12.07	0.23	17.74	7.14	1.98
Wr.F13	85	50.42	0.31	15.34	0.21	8.09	0.17	12.71	9.76	2.50
Wr.F14	100	50.67	0.38	17.52	0.48	5.60	0.13	8.87	13.27	2.67
Wr.F15	140	51.52	0.50	14.34	0.34	6.96	0.18	11.03	12.38	2.26
Wr.F16	175	48.03	0.29	14.30	1.37	9.46	0.19	14.00	9.17	2.28
Wr.F17	190	48.90	0.32	17.41	0.82	5.12	0.12	10.63	12.76	2.22
Wr.H1	300	51.05	1.11	16.89	1.35	7.60	0.18	7.64	10.81	3.08
Wr.H2	311	51.63	0.78	17.74	1.72	6.90	0.18	7.45	9.87	3.29
Wr.H3	320	52.06	0.96	21.03	1.50	3.77	0.11	4.50	11.67	3.95
Wr.H4	325	51.35	1.38	18.36	1.74	6.35	0.16	7.13	10.40	3.39
Wr.H5	332	50.62	1.55	19.39	2.36	6.83	0.15	5.72	9.25	3.48
Wr.H6	342	51.47	1.38	20.08	1.45	6.09	0.13	5.95	8.21	3.80
Wr.H7	358	47.92	3.76	16.76	2.71	9.52	0.19	7.17	7.78	3.25
Wr.H8	365	45.58	5.48	17.08	3.66	9.89	0.20	5.45	8.69	3.15
Wr.H9	372	46.32	4.87	19.13	3.81	8.72	0.16	3.93	8.78	3.57
Wr.H10	380	46.21	4.01	17.41	5.56	9.30	0.18	5.11	8.93	3.22
Wr.H11	425	46.21	5.47	15.05	2.80	10.44	0.22	6.73	9.42	2.80
Wr.H12	450	45.69	4.87	14.38	4.26	10.62	0.22	6.90	10.01	2.71
Wr.H13	470	49.23	2.87	19.43	2.93	7.16	0.14	4.01	9.12	3.91
Wr.H14	515	46.41	2.94	17.77	3.88	9.27	0.19	4.44	9.79	3.60
Wr.H15	540	47.14	3.00	13.37	3.08	12.14	0.26	7.15	10.93	2.80
Wr.H16	620	44.29	3.75	16.89	2.29	14.65	0.23	5.71	6.50	3.42
Wr.H17	819	43.22	2.55	13.40	3.13	13.42	0.16	2.41	11.55	5.04
Wr.H18	936	50.63	1.75	17.82	1.20	10.53	0.22	3.91	7.76	4.09
Wr.H19	980	49.01	1.75	15.73	1.35	11.87	0.32	4.58	8.16	3.92
Wr.H20	1025	45.99	2.11	14.79	1.67	16.45	0.35	4.77	7.37	3.43
Wr.H21	1069	51.62	0.62	19.30	2.26	9.83	0.23	1.99	7.35	4.78
Wr.H22	1120	45.38	1.66	14.79	2.42	18.11	0.38	3.57	7.03	3.26
Wr.H23	1170	51.71	1.20	16.59	1.79	12.90	0.34	1.87	6.41	4.57
Wr.H24	1200	52.77	1.33	16.93	1.78	11.32	0.29	1.65	6.41	4.74
Wr.H25	1225	55.63	0.8	18.56	1.42	7.74	0.22	0.71	6.64	5.54
Wr.H26	1270	49.53	1.34	15.48	2.86	14.74	0.32	1.48	5.63	4.68
Wr.H27	1300	52.42	1.00	16.79	2.49	12.46	0.37	1.17	5.25	5.86
Wr.H28	1411	53.15	0.91	17.54	1.81	8.77	0.27	1.25	6.48	5.67
Wr.H29	1477	59.89	0.44	16.85	1.55	5.23	0.19	0.55	2.54	6.08
Wr.H30	1577	61.60	0.37	16.97	1.58	4.65	0.17	0.53	1.90	5.67
Wr.H31	1720	62.40	0.35	16.90	1.35	3.35	0.12	0.07	1.17	5.55
Wr.H32	1790	62.03	0.37	16.84	1.12	4.58	0.16	0.06	1.35	4.96

tially in mutual equilibrium.

Olivine

Olivine is restricted to two stratigraphic intervals in Figure 3. It is a major constituent in the lowermost part of Stage IV in F1, varying in composition from Fo₇₄ at the base to Fo₆₈ at 190 m. It is absent at the base of H2 and more iron-rich olivine (Fo_{42.6}) reappears at 540 m. A similar hiatus in olivine composition has been described in several other layered intrusions; e.g., Fo₅₂–Fo₃₆ in Skærgaard (Wager and Brown, 1968; McBirney, 1989), Fo₅₅–Fo₃₁ in the Potato River intrusion (Klewin, 1990) and

Fo₆₂–Fo₅₀ in the Bjerkreim–Sokndal intrusion (Wilson et al., 1996). Olivine reaches almost pure fayalite at ~ 1600 m and is absent in the uppermost ~ 200 m. The most Fe-rich olivines in FHI are commonly oxidized to symplectitic intergrowths of magnetite and Ca-poor pyroxene or grunerite.

The most magnesian olivines contain up to 0.15% NiO, decreasing to below detection limits at Fo_{42.6}. MnO contents increase with the iron content to reach 1.7% MnO in the most fayalitic olivines. Some representative analyses are given in Table 2.

Stage IV in the Fongen-Hyllingen layered intrusion

K ₂ O	P ₂ O ₅	H ₂ O ⁺	Total	Mg#
0.09	0.02	0.87	100.33	74.78
0.07	0.02	1.48	100.00	75.60
0.09	0.01	2.09	99.04	72.91
0.11	0.01	1.65	99.12	75.67
0.19	0.01	1.80	100.18	70.04
0.16	0.01	2.04	99.00	74.25
0.12	0.01	0.78	100.82	73.65
0.13	0.02	0.93	99.84	73.80
0.12	0.02	1.24	100.41	74.56
0.07	0.01	1.00	100.91	73.65
0.09	0.01	0.83	100.18	75.51
0.11	0.01	1.71	100.27	72.38
0.14	0.01	0.93	100.59	73.69
0.13	0.01	0.78	100.51	73.85
0.12	0.01	0.80	100.44	73.86
0.13	0.01	1.07	100.30	72.51
0.13	0.01	1.97	100.41	78.73
0.16	0.02	0.87	100.76	64.18
0.20	0.04	0.79	100.59	65.81
0.19	0.02	0.50	100.26	68.03
0.25	0.03	0.75	101.29	66.68
0.20	0.02	0.82	100.39	59.89
0.18	0.04	1.55	100.33	63.53
0.22	0.04	1.08	100.40	57.31
0.22	0.03	1.01	100.44	49.55
0.25	0.05	0.98	100.57	44.55
0.21	0.02	0.71	100.87	49.48
0.22	0.03	0.76	100.15	53.47
0.17	0.02	0.96	100.81	53.66
0.30	0.07	0.69	99.86	49.96
0.25	0.03	0.68	99.25	46.06
0.20	0.05	1.01	101.13	51.22
0.24	0.06	1.94	99.97	41.00
0.23	2.28	0.89	98.28	24.25
0.31	0.96	0.77	99.95	39.83
0.29	1.40	1.45	99.83	40.75
0.36	1.15	2.04	100.48	34.08
0.48	0.84	0.99	100.29	26.52
0.47	1.27	2.03	100.37	26.00
0.76	0.52	1.44	100.1	20.53
0.75	0.47	1.30	99.74	20.62
0.83	0.33	0.84	99.26	14.05
0.67	0.54	2.00	99.27	15.18
0.88	0.19	1.61	100.49	14.34
0.83	0.33	0.74	97.746	20.26
4.52	0.08	0.95	98.87	15.79
5.72	0.01	1.52	100.69	16.89
6.07	0.03	1.31	98.67	3.59
5.99	0.04	0.85	98.35	2.28

Ca-rich pyroxene

Ca-rich pyroxene is present as a cumulus phase throughout Stage IV. In the pyroxene quadrilateral (Fig. 4), Ca-rich pyroxenes extend from Mg-rich diopside to the hedenbergite end-member with fairly constant Ca-contents, covering a Mg# range of 78-0 (Fig. 3). The most magnesian Ca-rich pyroxene occurs as a cumulus phase in olivine-rich units at the base of Stage IV, whereas the most Fe-rich compositions are found as remnants mantled by Ca-amphibole (ferro-edénite) in the quartz-bearing syenites at the top of Stage IV. Thin exsolution lamellae of Ca-poor pyroxenes are common, as are small blebs

Figure 2. Cumulate stratigraphy of Stage IV in the Fongen-Hyllingen intrusion along profiles F1 and H2 (modified after Wilson and Sørensen, 1996). Dotted area is a compositional gap between profiles F1 and H2. F1, H2a, H2b, H2c and H2d are stratigraphic intervals used in the calculation of the composition of the parental magma at the base of Stage IV, as explained in Figure 8.

Figure 3. Stratigraphic variation in the compositions of cumulus plagioclase, olivine, Ca-rich pyroxene, Ca-poor pyroxene and calcic amphiboles [both cumulus (closed circles) and intercumulus (open circles)] in profiles F1 and H2 through Stage IV in the FHI. Black triangle: data from Wilson and Larsen (1985). Most points represent the average of three analyses in the cores of each of three grains in the sample. Bars give maximum standard deviation (2σ) which is contained within the symbols where not indicated. "Unsampled interval" is the unsampled gap between profiles F1 and H2.

Figure 4. Compositional variation of the mafic silicates (Ca-rich pyroxenes, calcic amphiboles, Ca-poor pyroxene and olivines) in Stage IV. Selected tie lines join coexisting phases. Numbers on tie lines denote stratigraphic height in Figure 3. Note the extremely evolved mineral compositions at the top of Stage IV. Open circles, intercumulus. Dots, cumulus.

of optically continuous brown Ca-amphibole. The analyses in Table 3 represent subsolidus compositions since they were performed with a focused beam between Ca-poor pyroxene exsolution lamellae.

The Al-content of the Ca-rich pyroxenes decreases from 3.29% Al_2O_3 in the most magnesian diopsides to 0.6% Al_2O_3 in the most Fe-rich hedenbergites (Table 3). Titanium (1.1–0.18% TiO_2) shows an erratic declining trend with stratigraphic height. Chromium decreases upwards from 0.3–0.1% Cr_2O_3 in F1 whereas it is < 0.1% in H2. Sodium does not show a systematic trend with stratigraphic height although the highest Na_2O contents (0.8%) are present in the most Fe-rich compositions (Table 3).

Ca-poor pyroxene

Cumulus Ca-poor pyroxene occurs in the interval from 100 to ~ 1300 m. Its composition covers a wide Mg# range (76–13.5). The most magnesian compositions are near the base of Stage IV (Fig. 3). The most iron-rich compositions are commonly replaced partially or totally by grunerite. The Wo-content ranges from 3.3 to 1.1% and does not vary systematically with stratigraphic height. Al_2O_3 (2.42–0.32%) and TiO_2 (0.58–0.02%) do not vary systematically with stratigraphic height although the most Fe-rich compositions are generally low in these components. MnO shows an overall upward increase from 0.37 to 1.98% (Table 4).

Thin Ca-rich pyroxene lamellae are developed parallel to (100) in the Ca-poor pyroxene host. Inverted pigeonite, with characteristically abundant Ca-rich pyroxene lamellae and blebs, occurs sporadically in Fe-rich samples above ~ 1100 m. Some representative analyses of Ca-poor pyroxenes (subsolidus compositions) are given in Table 4.

Figure 5. Compositional variation of primary igneous calcic amphiboles from Stage IV in the nomenclature scheme of Leake (1978). Arrowed line shows the trend of fractionation. Fe_2O_3 is calculated assuming 16 cations and 23 oxygens stoichiometry. Dotted area is the composition variation of the calcium amphiboles in the entire Fongen-Hyllingen intrusion from Wilson et al. (1981).

Amphibole

Calcic amphibole is almost always present as an intercumulus phase through the lower part of Stage IV where it commonly forms oikocrysts, and it becomes an early crystallizing phase at ~ 950 m. Mg# of Ca-amphibole decreases systematically from 74 at the base of Stage IV to ~ 0 at 1730 m (Fig. 3). The trend of iron-enrichment with stratigraphic height is continuous and essentially parallel with that of the other mafic phases, regardless of the cumulus status of the Ca-amphibole. In terms of their compositional classification, Ca-amphiboles extend from pargasite through ferroan pargasite, ferroan pargasitic hornblende, ferro-pargasitic hornblende and ferro-edenitic hornblende to ferro-edenite in the nomenclature scheme of Leake (1978) (Fig. 5). TiO_2 (1.4–3.9%), Al_2O_3 (6.9–14%) and Na_2O (1.2–2.9%) vary unsystematically with stratigraphic height until ~ 540 m, after which they all decrease systematically upwards and reach their minimum values near the roof (Table 5). MnO increases steadily upwards from < 0.1% to 1.2% throughout Stage IV. The composition of Ca-amphibole usually varies considerably within individual samples, particularly Mg# (Fig. 3), but also TiO_2 and Al_2O_3 .

Alkali feldspar

Alkali feldspar first occurs as a minor interstitial phase above 1300 m and gradually increases in proportion until it joins the cumulus assemblages at 1477 m. The alkali feldspar is principally orthoclase microperthite and the Or:Ab ratio in reintegrated compositions (using a defo-

Figure 6. Whole rock compositional variation in Stage IV using the total alkalis versus silica (TAS) diagram of Middlemost (1994). The “alk”/“sub-alk” curved line separates the alkaline and sub-alkaline fields (after Irvine and Baragar, 1971). The filled circles indicate the bulk composition of cumulates. The fields enclosed by dotted and dashed lines indicate the compositional ranges along profiles F1 and H2 respectively. Open circles represent the magma composition estimated at successive levels in Stage IV (Table 7). The line joining these indicates the liquid line of descent in Stage IV.

cused beam with a diameter between 10 and 50 μm) ranges from 10:1 to 25:1. The An-content is usually < 0.1%.

FeTi-oxides

FeTi-oxides occur as minor intercumulus phases in the lowest 190 m of Stage IV and in the lower ~ 50 m of H2. Above this their abundance increases steady upwards, reaching a maximum (8.5 vol%) between 380 and 819 m, and then declining rapidly, being almost absent in the uppermost part of Stage IV. The FeTi-oxide enrichment in the middle part of the crystallization sequence is consistent with the fractionation of magma with tholeiitic affinity (Hunter and Sparks, 1987, 1990; Toplis and Carroll, 1996). Titanomagnetite and ilmenite are both present until 1440 m, above which titanomagnetite is only present in minor quantities whereas ilmenite continues into the uppermost part of Stage IV. Ilmenite shows no extensive oxy-exsolution and its solidus composition is most probably reintegrated in defocused beam analysis (10–20 μm). It has a low hematite content (usually < 1 mol%) throughout Stage IV. The Mn-content is roughly constant ($1 \pm 0.20\%$ MnO) until 1170 m, where it abruptly increases to 3.14 MnO%. The increase in Mn-content of ilmenite is related to increase in Mn/Fe ratio of the silicate melt, to decrease in temperature (Neumann, 1974) and/or to increase in oxygen fugacity (Czamanske and Mihálik, 1972). MgO (1.8–0%) in ilmenite decreases continuously upwards. Titanomagnetite, on the other hand, is

Figure 7. Stratigraphic variation in the major elements (wt% oxide) along profiles F1 and H2 in Stage IV of the FHI.

extensively oxy-exsolved and its original composition cannot readily be integrated by defocused beam analyses. However, Thy (1981) found a slight systematic fall in MgO in titanomagnetites with fractionation when recalculated to their probable original compositions.

WHOLE ROCK CHEMISTRY

In the TAS diagram of Middlemost (1994) (Fig. 6), Stage IV cumulates cover a wide range of compositions from the peridot gabbro field with 43.8% SiO_2 , through the fields of gabbros, gabbroic diorites, monzogabbros and monzodiorites to syenites with up to 62.4% SiO_2 and almost 12% $\text{Na}_2\text{O} + \text{K}_2\text{O}$. A single sample plots in the foid gabbro field. The overall trend extends from gabbroic rocks in the subalkaline field to syenites, well within the alkaline field.

The variation in major oxide abundances with stratigraphic height is shown in Figure 7. Since the rocks in Stage IV are dominantly accumulates, the change in distribution of oxides is essentially attributed to changes in the modal assemblages rather than to variations in the amount of trapped liquid. In the lowermost 60 m of Figure 7 most major oxides cover a wide range and vary unsystematically with stratigraphic height. This is be-

lied to reflect local variations in modal composition where samples do not represent "average cumulates". Above 60 m all major oxides show fairly systematic trends with stratigraphic height. SiO_2 increases slightly from 48% at 175 m to 52% at 320 m, followed by a fall to 43% at 819 m and then an erratic increase to ~ 53% at ~ 1400 m. Above this there is an abrupt increase to 60–62% in the uppermost samples where K-feldspar is a liquidus phase (1450–1800 m; Fig. 7). Al_2O_3 (12–20%) is fairly constant, whereas titanium increases dramatically from 0.2% TiO_2 at the base up to a maximum of 5.47% TiO_2 in the interval 358–450 m, after which it erratically falls, reaching about 0.35% at the roof. The Fe_2O_3 variation is similar to that of TiO_2 (Table 6 and Fig. 7) and both are largely related to the abundance of FeTi-oxides in the cumulates. FeO shows no systematic variation with stratigraphic height, increasing from 5.6% at 100 m to 14.7% at 620 m before falling to about 4% at the top of Stage IV. MgO falls gradually from 18% at 60 m to 0.06% at the roof. CaO decreases from 12% at 60 m to 1.4% at the roof, whereas Na_2O increases from 2% to 6%, decreasing slightly to 5% at the roof. K_2O increases slightly, but systematically from 0.16% at the base of Stage IV to 0.83% at ~ 1477 m, after which it increases abruptly to 4.5% when alkali feldspar becomes a liquidus phase. K_2O reaches 6% at the top of Stage IV. P_2O_5 is fairly constant and low (< 0.07%) up to 800 m, after which it increases dramatically to 2.3% when apatite becomes a cumulus phase. The subsequent systematic decrease in P_2O_5 reflects the gradual depletion of phosphorus in the residual melt as a result of the crystallization of apatite.

DISCUSSION

Parental magma composition

Estimation of the parental magma composition for the Fongen-Hyllingen intrusion is hampered by the lack of chilled margins and contemporaneous mafic dykes. As is typical for layered intrusions, whole rock analyses do not provide magma compositions but represent the solid products of fractional crystallization of the evolving magma. The method of summation of rock compositions throughout the intrusion can, however, be used to calculate the composition of the parental magma, as has been done, for example, for the Kiglapait intrusion (Morse, 1981), the Basistoppen Sill (Naslund, 1989) and the Potato River intrusion (Klewin, 1990).

A prerequisite for application of the summation method is that crystallization took place in a closed system, i.e. without tapping from or repeated influx into the magma chamber. In principle, the magma should not be

contaminated during crystallization since this involves the transfer of material into the system. If the magma chamber remained closed, then summation of all of the rock compositions in the intrusion should give the bulk composition of the parental magma(s) (e.g., Blundy, 1997).

The FHI crystallized in a magma chamber that was repeatedly replenished during a large part of its evolution and the summation method therefore cannot be applied to the intrusion as a whole. It is, however, possible to estimate magma compositions in Stage IV because most of this sequence, as inferred from the strong normal fractionation trend in cumulus minerals (Fig. 3), crystallized when magma addition ceased or became insignificant (Wilson and Larsen, 1985). The presence of extremely evolved rocks with low temperature mineral assemblages at the top of Stage IV implies that tapping from the magma chamber was not significant.

A basic premise for the summation method involves mass proportions; the relationship between stratigraphic thickness and magma volume should be estimated. As in any one-dimensional section through a three-dimensional body, this is very difficult to assess. However, Stage IV has a considerable lateral extent (35–40 km) and a maximum thickness of ca. 1800 m. We have no convincing arguments to justify weighting any part of the profile more than any other, and consequently, as a first approximation, we assume that stratigraphic thickness in Figure 7 is directly proportional to volume. This seems justifiable in view of the large aspect ratio (i.e., length/thickness) of Stage IV (in the vicinity of 20).

The lateral compositional variations in the Hyllingen Series, where mineral compositions become systematically more evolved along the strike of modal layering approaching the southern margin (Wilson and Larsen, 1985), mean that sample profiles normal to the layering cover more or less complete stratigraphic sections, depending on the location. This, however, only becomes significant in the southernmost ca. 5 km in the Hyllingen Series (Wilson and Larsen, 1985) and, in view of the large aspect ratio, we consider that these lateral variations can be disregarded in our estimation of the liquid line of descent. Our profiles through Stage IV (F1 and H2) together cover the entire compositional range, from the most primitive cumulates in the intrusion at the base to the most evolved rocks at the roof.

In a profile normal to modal layering, the products of progressive fractional crystallization from initially homogeneous magma will usually be indistinguishable from those produced by fractional crystallisation of compositionally zoned magma. However, Stage IV of the FHI does not represent the products of crystallisation of a single influx of homogeneous magma. It has been convinc-

Figure 8. Modelling of successive parental magma compositions in Stage IV based on profiles F1 and H2. The boundaries between the stratigraphic intervals are drawn where there are marked changes in the trend for one or more oxide caused by changes in the cumulus assemblage. The construction and use of the regression lines is explained in the text. Some overlapping data points are omitted for the sake of clarity in the figure. Cumulate stratigraphy after Wilson and Sørensen (1996). Minerals are indicated as follows: Plag, plagioclase; Cpx, Ca-rich pyroxene; Ol, olivine; Opx, Ca-poor pyroxene; Mt/il, FeTi-oxides; Amp, amphibole; Ap, apatite; Zr, zircon; Bio, biotite; Qz, quartz; Kf, alkali feldspar; Alla, allanite. Solid lines, cumulus status; dotted lines, intercumulus status.

ingly argued that the magma in the FHI chamber was compositionally zoned, at least by the time that Stage IV developed (Wilson and Larsen, 1985). The calculated magma compositions therefore represent a series of hypothetical bulk compositions. Large volumes of magma with the calculated compositions were probably not present in the chamber, but they do reflect the liquid line of descent of the magma column that was in the chamber during crystallisation of Stage IV.

We have adopted an approach similar to that of Klewin (1990) for the Potato River layered intrusion. Klewin only considered the upper part of the intrusion, above the level where major magma addition ceased.

We have divided Stage IV into five intervals (F1, H2a, H2b, H2c, H2d) in Figure 8, each of which represents a continuous compositional trend for as many major element oxides as possible. This corresponds to important changes in the cumulus assemblage. The base of interval

Table 7. Compositions and densities of the melts derived by reverse stratigraphic summation in Stage IV of the Fongen-Hyllingen layered intrusion

Crystallization interval	F1 0-300m	H2a 300-380m	H2b 380-800m	H2c 800-1477m	H2d 1477-1790m
Stratigraphic height (m)					
SiO ₂	51.70	52.22	52.35	54.52	62.74
TiO ₂	1.68	1.95	1.93	1.21	0.39
Al ₂ O ₃	16.34	16.87	16.77	16.91	17.24
Fe ₂ O ₃	2.13	2.26	2.36	2.25	1.43
FeO	9.82	10.12	10.17	9.62	4.54
MnO	0.23	0.23	0.24	0.26	0.16
MgO	4.99	3.14	2.97	1.79	0.31
CaO	7.44	6.81	6.64	5.53	1.78
Na ₂ O	4.02	4.43	4.49	5.02	5.68
K ₂ O	1.29	1.55	1.62	2.21	5.69
P ₂ O ₅	0.35	0.42	0.46	0.71	0.04
Total	99.99	100.00	100.00	100.03	100.00
CIPW norm					
Q	0.00	0.00	0.00	0.00	1.09
Or	7.27	8.75	9.10	12.83	31.98
Ab	34.61	37.90	38.41	43.24	49.24
An	22.55	21.01	20.24	16.06	5.03
Di	10.65	8.83	8.67	5.41	3.85
Hy	7.34	6.78	6.61	10.88	5.38
Ol	10.18	8.19	8.50	4.43	0.00
Mt	3.18	3.58	3.55	3.20	2.22
Il	3.29	3.82	3.68	2.26	0.97
Ap	0.97	1.18	1.26	1.71	0.24
ρ g/cm ³ *	2.70	2.63	2.6	2.54	2.40
Mg#	47.5	35.6	34.2	24.9	10.8

* The density has been calculated assuming that the parent melt contained 1 wt% H₂O, increasing gradually with fractionation to 3 wt% at the base of interval H2c (where cumulus amphibole joined the cumulus assemblage) and finally to 4% in the uppermost part of the magma column (interval H2d).

H2a marks the cessation of olivine as a cumulus phase; the base of H2b coincides with the entry of cumulus FeTi-oxides; the base of H2c coincides with the appearance of cumulus apatite as well as the change in role of calcic amphibole from an intercumulus to a cumulus phase. The start of interval H2d coincides with the arrival of alkali feldspar as an early crystallizing phase. The average compositions of the intervals are obtained by reading off the values from the stratigraphic mid-points of “best-fit” regression lines and the values are weighted in accordance with the thickness of the stratigraphic interval (Fig. 8). In most cases this gives values that are extremely close to simple arithmetic mean values of the intervals.

A series of magma compositions have then been calculated for Stage IV by summing the weighted data in a sequence opposite to that of their crystallization (i.e., from the roof down) and subsequent normalization to 100%. The results (Table 7) represent the bulk composi-

tion of magma that remained in the magma chamber before each of the intervals began to crystallize, and therefore approximate to the liquid line of descent. The most primitive composition (F1 in Table 7) is as close as we can get to that of the Fongen-Hyllingen parental magma and lies near the base of the basaltic trachyandesite field in the TAS diagram with 51.2% SiO₂ and 5.3% alkalis (Fig. 6), very close to the line separating alkaline from sub-alkaline magma compositions. The liquid line of descent in Stage IV extends through the basaltic trachyandesite and trachyandesite fields into the trachyte field with 62.3 wt% SiO₂ and 11.2 wt% alkalis.

Using a value of 0.3 for $Kd_{FeO/MgO}^{OL}$ (Roeder and Emsile, 1970), the Mg-number of melt in equilibrium with the most magnesian olivine in Stage IV (Fo₇₄) is 0.455, slightly lower than 0.475 of the calculated parental melt at the beginning of interval F1. A possible explanation for this slight discrepancy is that the solidus composition of cumulus olivine usually changes to less magnesian compositions as a result of re-equilibration with intercumulus melt, the so-called “trapped liquid shift” (Barnes, 1985, 1986; Chalokwu and Grant, 1987; Paktunc, 1987; Chai and Naldrett, 1992). Assuming an initial porosity in the essentially adcumulate rocks of 10%, the solidus composition of cumulus olivine was unlikely to have been more magnesian than Fo₇₅ using the curves of Barnes (1986). Olivine with this composition should crystallize in equilibrium with a melt with Mg-number 0.474, very similar to that of the initial melt (F1) calculated by summation here. This Mg-number is not, however, a candidate for a primary basaltic melt that usually has an Mg-number around 0.70 (Green et al., 1974), implying that the FHI initial melt was relatively evolved. It therefore seems likely that the FH parental magma evolved from a mantle-derived composition in an underlying magma chamber before entering the FH magma chamber. We will return to this topic later.

The FHI magma was, however, subjected to contamination as a result of the assimilation of metapelitic country rocks (Sørensen and Wilson, 1995; Tegner et al., 2005). There is a fairly good correlation between Sr-isotopic ratios and mineral compositions in Stage IV, with contaminated signatures in the most evolved rocks. This indicates that assimilation of country rocks dominantly took place at the roof of the magma chamber, whereas crystallisation took place from the floor upwards (Wilson and Sørensen, 1996). Sørensen and Wilson (1995) estimated that the maximum amount of assimilation was about 8% in the most evolved rocks, decreasing to negligible amounts in the most primitive assemblages. The most evolved sample at the top of Stage IV (Wr.H32; Table 6) contains 62.03% SiO₂, 4.96% Na₂O and 5.99%

K₂O. The precise composition of the contaminant is hard to estimate, but a granitic partial melt of metapelites is likely to have more SiO₂ and less alkalis than this (S-type granite from Table 10-1 in McBirney (1993) contains 73.48% SiO₂, 2.63% Na₂O and 4.63% K₂O). Subtraction of 8% of a composition of this type will decrease the SiO₂ and increase the alkalis in the recalculated composition of Wr.H32, and consequently in the recalculated magma compositions at the bases of the other intervals. It is therefore clear that the alkali-rich nature of the final differentiates in FHI cannot be explained by the assimilation of pelitic material. The FHI parental magma must consequently have had a relatively evolved composition, with a low Mg-number and elevated contents of alkalis.

Differentiation trend

The estimated magma composition at the base of Stage IV lies very close to the boundary separating the alkaline and subalkaline fields in a TAS diagram (Fig. 6 and Table 7). Nevertheless, some features of the FHI fractionation trend, such as the co-precipitation of Ca-poor and Ca-rich pyroxenes, the presence of an olivine gap and the development of quartz-bearing late differentiates, are not features consistent with differentiation of an alkali basalt parent (Best, 1982).

On the basis of mineralogical and chemical observations the FHI differentiation trend could be considered as intermediate between typical tholeiitic and calc-alkaline trends. Tholeiitic affinity is expressed by the coexistence of Ca-rich and Ca-poor pyroxenes, the reaction relationship between olivine and Ca-poor pyroxene, the presence of an olivine hiatus, and the presence of a considerable thickness of quartz-bearing late differentiates (Best, 1982). The relatively early appearance of intermediate An-plagioclase as a dominant cumulus phase is also consistent with tholeiitic fractionation (Grove and Baker, 1984). On the other hand, calc-alkaline affinity is marked by the rarity of ultramafic cumulates and the early crystallization of FeTi-oxides with subsequent iron-depletion and silica-enrichment in the melt. Calc-alkaline affinity is also supported by the hydrous nature of the FHI parental melt, as is implied by the presence of calcic amphibole as an intercumulus and then cumulus phase, and the presence of biotite in the late differentiates (Kennedy, 1955; Osborn, 1959; Sisson and Grove, 1993a). The dissociation of dissolved water could lead to high f_{O_2} and early oxide precipitation (Osborn, 1959).

Iron enrichment

In an AFM diagram (Fig. 9a), the calculated FHI magma

Figure 9. Calculated melt compositions in Stage IV are shown in (a) AFM diagram and (b) SiO₂-FeO_{total} plot. F1, H2a, H2b, H2c and H2d are the estimated compositions of melts at the base of stratigraphic intervals 0 m, 300 m, 800 m and 1411 m respectively from Table 7. The termination of the FHI trend is marked by the most evolved analysed sample (Wr.H32). The base of interval H2b marks the arrival of cumulus magnetite. Evolutionary trends shown for comparison are: (1) Kiglapait intrusion (Morse, 1981); Skærgaard intrusion according to (2) Wager and Brown (1968) and (3) Hunter and Sparks (1987); (4) Potato River intrusion (Klewin, 1990); (5) Guadalupe complex (Best, 1963).

compositions show very minor initial Fe-enrichment in the interval from F1 to H2b, after which they move towards the alkali corner. In this regard the fractionation trend shows calc-alkaline rather than tholeiitic affinities. In a total iron vs. silica diagram (Fig. 9b) there is again only slight Fe-enrichment in the early stages, after which Fe-depletion is accompanied by silica-enrichment. Silica enrichment reaches 62% (63.1% on an anhydrous basis) in the most evolved rock (Wr.H32) which is taken as representing the final magma composition.

Compared to the differentiation trends of the Skærgaard (Wager and Brown, 1968) and Kiglapait intrusions (Morse, 1981), there is hardly any iron-enrichment in the magma during fractionation of FHI (Fig. 9a, b). The FHI melt is markedly lower in FeO_{total} for a given SiO₂ than

Figure 10. T - f_{O_2} path for the FHI calculated from coexisting titanomagnetite-ilmenite pairs after Thy (1981). Solid circles represent the T - f_{O_2} conditions indicated from the analysed compositions of titanomagnetite-ilmenite pairs in FHI. SKI and KI are calculated parameters for Skærgaard (McBirney and Naslund, 1990) and Kiglapait (Morse, 1980) intrusions respectively. NNO and FMQ buffers are calculated after O'Neill (1987a, 1987b) at 1 bar pressure. Note that the T - f_{O_2} path for the FHI is sub-parallel to the FMQ buffer whereas the T - f_{O_2} paths for Skærgaard and Kiglapait are not parallel to the FMQ buffer. This is consistent with crystallization in a system open to oxygen for FHI and in systems closed to oxygen for SKI and KI (Toplis and Carroll, 1996).

calculated melts in both the Skærgaard and Kiglapait intrusions (Fig. 9b) and is clearly more comparable with the Potato River and Guadalupe trends. It is noteworthy that the initiation of cumulus magnetite crystallization marks the end of absolute (albeit slight) iron-enrichment in the evolving magma (e.g., Hunter and Sparks, 1987, 1990) and iron depletion accompanies silica enrichment after this.

In a temperature-oxygen fugacity plot (Fig. 10), the Skærgaard and Kiglapait intrusions define T - f_{O_2} paths that are not parallel to FMQ buffer (Toplis and Carroll, 1996) whereas the FHI follows a T - f_{O_2} path parallel to and above it (Thy, 1981). This behaviour is consistent with crystallization in a system open to oxygen for FHI and closed to oxygen for the Skærgaard and Kiglapait intrusions (Toplis and Carroll, 1996). Magma evolution in FHI was interrupted by repeated influxes, assimilation of wall rocks and subsequent mixing between uncontaminated replenishing magma and contaminated resident magma (e.g., Wilson et al., 1981; Wilson and Sørensen, 1996), resulting in the system not being closed to oxygen. Oxygen exchange with the surrounding host rocks and the numerous country rock inclusions kept the f_{O_2} near the NNO buffer curve, but f_{O_2} was not high enough to, for example, produce a significant aegirine component in Ca-rich pyroxene (Thy, 1981; Wilson et al., 1981). The Skærgaard and Kiglapait intrusions, on the other hand, represent the crystallization products of single magma injections under conditions more closed to oxygen (Toplis and Carroll, 1996).

Water content

The hydrous nature of the FHI parental magma is evident from the significant thickness of cumulates containing substantial amounts of calcic amphibole and in the crystallization of biotite in the uppermost part of Stage IV. The presence of abundant cumulus FeTi-oxides at a relatively early stage (at ~ 350 m) and crystallization of cumulus olivine over large intervals in Stage IV (0 to 190 m and 540–1070 m) are also evidence for the hydrous nature of the FHI melt. The dissolved H_2O in the magma would produce a high f_{O_2} , leading to relatively early and abundant oxide precipitation (Osborn, 1959) and expanding the primary liquidus volume of olivine at the expense of other minerals (Kushiro, 1969; Sisson and Grove, 1993a). The early appearance of plagioclase of intermediate composition (An_{60}) as a major cumulus phase, and the coexistence of cumulus Ca-rich and Ca-poor pyroxenes in relatively primitive cumulates are, however, inconsistent with high water content in the FHI parental melt. High water contents in basaltic magma suppress the liquidus volume of plagioclase relative to those of olivine and iron oxides (e.g., Wasserburg, 1957; Kushiro, 1969, 1975; Spulber and Rutherford, 1983; Sisson and Grove, 1993a, 1993b) and lower the activity of silica in the melt, leading to destabilisation of Ca-poor pyroxene relative to olivine (Sisson and Grove, 1993a). The absence of cumulus amphibole at the base of Stage IV implies that the amount of water in the parental melt was less than 3 wt% which is the minimum water content required to crystallize amphibole from basaltic magma (Holloway and Burnham, 1972). The P_{H_2O} in the melt would gradually have built up due to the crystallization of anhydrous phases in the early stages of fractionation, leading to crystallization of cumulus calcic amphibole at ~ 950 m. The disappearance of Ca-poor pyroxene (at ~ 1300 m), the crystallization biotite, and the coexistence of orthoclase and albite in the roof syenites are also evidence that Stage IV crystallized under elevated P_{H_2O} conditions. The presence of subsolvus (two feldspar) syenitic cumulates in the uppermost part of Stage IV indicates $P_{H_2O} \geq 5$ kb (Yoder et al., 1957; Morse, 1970). This is higher than the maximum lithostatic pressure during emplacement of 4 kb indicated by the presence of a broad andalusite zone in metapelites in the contact metamorphic aureole (Bucher and Frey, 1994; Thyssen, 1998). This in turn implies that the magma must have been close to H_2O -saturation during the final stages of differentiation. Wilson et al. (1981) documented the presence of cummingtonite-grunerite amphibole mantling and/or replacing Ca-poor pyroxene towards the top of Stage IV. Formation of cummingtonite-grunerite

Figure 11. (A) Schematic representation of the forsterite-fayalite-silica phase diagram (based on Figures 4-14 in McBirney, 1993). (B) Expanded section of the Fo-Fa-SiO₂ phase diagram (distorted and not to scale) illustrating formation of the olivine hiatus in Stage IV (see text).

amphibole probably indicates P_{H_2O} greater than about $0.8P_{H_2O}$ (Wood and Carmichael, 1973).

Phosphorus enrichment

A distinct feature of FHI is the abundance of cumulus apatite in the central part of Stage IV. The amount of apatite in the first apatite-bearing cumulates at ~ 800 m is quite large (2.28% P₂O₅, Table 6), after which it gradually declines to reach very small amounts at 1411 m (0.33% P₂O₅, Table 6 and Fig. 7). Below ~ 800 m the presence of apatite is largely restricted to minute inclusions in cumulus silicates, particularly plagioclase. This variation in abundance of apatite most probably reflects the change in solubility of phosphorus during evolution of melt. The low P₂O₅-content in the calculated parental melt (0.41%, Table 7) was not adequate to precipitate cumulus apatite in the lower part of Stage IV. The minute apatite inclusions below ~ 800 m most probably reflect local P₂O₅ saturation in the melt in contact with growing crystals, since phosphorus has sluggish diffusion relative to other elements in silicate melts (Green and Watson, 1982). Due to the incompatible behaviour of phosphorus in the early silicate minerals (e.g., Henderson, 1968) and the decreasing solubility of apatite in the melt with falling temperature and increasing SiO₂ content (e.g., Watson, 1979; Harrison and Watson, 1984), the amount of phosphorus in the FHI melt might have reached the weight percent level before the melt experienced saturation in apatite. The maximum amount of P₂O₅ in the calculated liquid line of descent for Stage IV is 0.71% P₂O₅ in interval H2c (Table 7), less than values in the calculated melts of the Skærgaard (1.3% P₂O₅; McBirney and Naslund, 1990) and Kiglapait (1.31% P₂O₅; Morse, 1981) intrusions. The accumulation of apatite would result in the gradual depletion of P₂O₅ in the

residual liquid and eventually to amounts below the level for apatite saturation in the melt coexisting with the uppermost cumulate (0.04% P₂O₅, Table 7); apatite saturation occurs in felsic melts at level as low as 0.14 % P₂O₅ (Watson and Capobianco, 1981).

Olivine hiatus

A feature of Stage IV is that olivine ceases to crystallize over a considerable stratigraphic interval whereas Ca-poor pyroxene is present over a very wide range. At the base of Stage IV in Figure 2 the most magnesian olivine is Fo₇₄ and the most iron-rich composition before the olivine gap is Fo₆₈. However, elsewhere in the intrusion olivine compositions reach Fo₆₀ before this mineral temporarily ceases to crystallise (Wilson and Sørensen, 1996). When olivine reappears it has a composition of Fo₄₃, after which it evolves continuously to reach pure fayalite. Cumulus Ca-poor pyroxene crystallises over a wide compositional range, from Mg#72 to Mg#10. In contrast, there is an olivine gap in the compositional interval Fo₆₀ to Fo₄₃. In the forsterite-fayalite-SiO₂ phase diagram the behaviour of olivine depends essentially on the position of the fractionating melt with respect to the olivine-orthopyroxene field boundary. A possible course of magma evolution in FHI is shown in Figure 11. Parental magma (A) composition that was in equilibrium with Fo₇₄ evolved following a curve across the olivine field as progressively less-magnesian olivine is removed (A to B in Fig.11). The crystallisation of both olivine (Fo₇₀ to Fo₆₀) and Ca-poor pyroxene (Mg#opx₇₂ to Mg#opx₆₂) before the olivine gap is not in strict accordance with the phase diagram. It would appear that the melt moved along the olivine - Ca-poor pyroxene boundary curve (from B to C in Fig. 11) which acted as a cotectic rather than as a reac-

tion curve, although most olivines in this interval are rounded or embayed and enclosed by Ca-poor pyroxene or, in some cases, by calcic amphibole or Ca-rich pyroxene. When olivine reached a composition of Fo₆₀ (at C in Fig. 11) it ceased to crystallise and the melt entered the Ca-poor pyroxene field in Figure 11. The melt now embarked on a curved path across the orthopyroxene field and crystallized gradually more Fe-rich Ca-poor pyroxene and not olivine, resulting in the olivine hiatus. The absence of samples with coexisting of Ca-poor pyroxene and quartz without olivine implies that SiO₂-enrichment in the melt was not enough for it to reach the orthopyroxene-silica boundary. Instead, the melt remained in the pyroxene field and followed a curved path until it encountered the olivine-orthopyroxene boundary again at point E. Since the olivine-orthopyroxene boundary changes from a reaction boundary to a cotectic boundary across the En-Fs join (D in Fig. 11) (Morse, 1980; McBirney, 1993), olivine (Fo₄₃) and Ca-poor pyroxene (En₅₅) crystallized together, ending the olivine hiatus. The melt then moved along the olivine-orthopyroxene cotectic boundary until it reached point F where orthopyroxene ceased to crystallize and quartz joined the assemblage together with progressively Fe-richer olivine. Pure fayalite (now largely a symplectic intergrowth of magnetite and Mg-free grunerite) and quartz coexist at ~ 1600 m, indicating that the melt reached the fayalite - quartz boundary at G in Figure 11.

Density evolution in Stage IV

Densities of the melt at successive levels in Stage IV have been calculated according to the method of Bottinga and Weill (1970) using the partial molar volumes of Lange and Carmichael (1987) and the compositions of melts in Table 7. It emerges that the densities of calculated melts progressively decrease during fractionation (Table 7 and Fig. 12a). Assuming an initial water content of 1 vol% H₂O, which would be a conservative value considering the abundance of the calcic-amphibole in the intrusion, the calculated melt at the base of Stage IV had a density of ~ 2.70 g/cm³, decreasing to ~ 2.51 g/cm³, if the water content remained constant, and to ~ 2.40 g/cm³ if it increased to 4 vol% H₂O in the most evolved melt at the roof. This implies that the fractionation density (the total density of chemical components in the melt phase that are incorporated into the fractionating crystalline phases; Sparks and Huppert, 1984) was generally higher than the density of the melt. Such a continuous decrease in melt density is not consistent with the density change during fractionation of anhydrous basaltic magma, which involves an initial decrease in magma density to a minimum

Figure 12. (a) Densities for calculated magma compositions in Stage IV vs. water content. F1, H2a-H2d are as in Figure 9. Wr.H32 represents the most evolved composition at the roof of Stage IV. The heavy dashed arrow indicates the density evolution of the FHI magma, assuming that H₂O is concentrated in the melt with differentiation. (b) Schematic density-differentiation path for dry basaltic melt (after Sparks and Huppert, 1984). A density minimum characterizes the point where plagioclase begins to crystallize (+pl) together with olivine and a density maximum is reached where FeTi-oxides join the crystallising assemblage (+mt-il). (c) The proposed density-differentiation path of the FHI melt in Stage IV. Note that the magma density decreases throughout differentiation. The densities in the present study have been calculated by the Bottinga and Weill (1970) method using the partial molar volumes and thermal expansion data of Lange and Carmichael (1987) and compressibility data from Kress and Carmichael (1991). The calculations were performed assuming that the partial molar volume of a component in the melt is independent of bulk composition (Bottinga and Weill, 1970; Nelson and Carmichael, 1979; Lange and Carmichael, 1986).

while olivine ± pyroxene fractionate, followed by an increase to a maximum after plagioclase joins the fractionating assemblage, and then a decrease when FeTi-oxides crystallise (Sparks et al., 1980; Stolper and Walker, 1980; Sparks and Huppert, 1984) (Fig. 12b).

In FHI, the evolved nature of the parental magma meant that olivine, plagioclase and Ca-rich pyroxene crystallised simultaneously during the early stages of fractionation. The relatively low density of the parental magma (~ 2.70 g/cm³ compared with ~ 2.80 g/cm³ for anhydrous basalts) meant that the fractionation density was higher than the magma density from the beginning of

crystallisation (Fig. 12c). Initiation of FeTi-oxide crystallisation at H2b contributed to the fractionation density. The relatively high, and increasing, water content contributed to the low density of the evolving melt (Fig. 12a).

This density evolution is important since the discordant relations between modal and cryptic layering in FHI, explained by Wilson and Larsen (1985) as resulting from the crystallisation of compositionally zoned magma along an inclined floor, requires that the magma became compositionally zoned from an early stage of development in the magma chamber. Two important processes which could have contributed to this are (a) the repeated influx of dense magma along the floor of the chamber, elevating the resident magma column and (b) the release of buoyant, relatively evolved liquid from the crystallising assemblage which migrated towards the top of the magma chamber. The density calculations presented here imply that both these processes appear to have been possible during the early evolution of the Fongen-Hyllingen magma chamber.

Origin of the Fongen-Hyllingen parental magma

It is clear that the composition of the FHI parental magma, as indicated by the compositions of the most primitive olivine (Fo₇₄) and plagioclase (An₆₀), is far too evolved to represent a primary mantle melt. However, the isotopic compositions of the most primitive rocks from FHI (initial Sr-isotope ratio = 0.70308; ϵ_{Nd} in the range +4.94 to +4.27; Sørensen and Wilson, 1995) are close to mantle signatures, implying very restricted contamination (Raymond, 1995). Abu El-Rus (2003) found that the trace element signature of calculated melts in equilibrium with the most primitive FHI cumulates are similar to those of low-Ti continental flood basalts presented by Fodor et al. (1990). The low Zr/Y ratios (1.78–3.16), together with the slight LREE-enrichment (Abu El-Rus, 2003) for these melt compositions also imply that garnet was not present in the mantle source, implying derivation from within the spinel-bearing peridotite zone of the upper mantle (Green et al., 1989).

Partial melts of peridotitic mantle, derived above the garnet zone, are believed to have fed a lower crustal magma chamber where fractional crystallisation took place with little or no assimilation. Extensive olivine fractionation may have been accompanied by some pyroxene and/or calcic plagioclase removal to reduce the CaO and Al₂O₃ values in the magma. The elevated alkali contents in F1 (4.02% Na₂O and 1.29% K₂O) are largely responsible for the unusual fractionation trend towards trachytic magma compositions in a TAS-diagram (Fig. 6). The elevated Na₂O implies that there was not much plagioclase

fractionation at depth, whereas the high K₂O is believed largely to be a primary signature, reflecting a K-rich mantle source.

The metavolcanic and metapelitic country rocks into which the FHI was emplaced belong to the early Ordovician Fundsjø Group and have immature oceanic arc and marginal basin affinity (Grenne et al., 1999). The FHI was emplaced in late Silurian times during the protracted collision phase of the Scandinavian Caledonides. Intrusion took place after one deformation episode but before the main (Scandian) regional deformation and metamorphic event. However, a tholeiitic basaltic dyke swarm in the vicinity of the FHI indicates local dilation of up to 25% (Thjømøe, 1995) shortly prior to emplacement of the layered intrusion. This may reflect development of an extensional segment during the predominantly compressional regime due to the oblique collision between the converging margins of Baltica and Laurentia (T. Grenne, personal communication).

CONCLUSIONS

- (1) Mineral compositions in Stage IV cumulates of the Fongen-Hyllingen layered intrusion cover extreme ranges. Olivine spans compositions from Fo₇₄ to pure fayalite with a hiatus in the interval Fo₆₀ to Fo₄₃. Ca-rich pyroxene and plagioclase are both present throughout the cumulate stratigraphy and span compositions from Mg#cpx79-0 and An₆₀₋₀ respectively. Ca-poor pyroxene is present over a wide interval, and the assemblage is joined successively by FeTi-oxide, Ca-amphibole, zircon, biotite, quartz, K-feldspar and allanite. The most evolved rocks are quartz-bearing syenites that occur in contact with country rock amphibolites at the roof.
- (2) Calculation of the parental magma composition for Stage IV of the Fongen-Hyllingen layered intrusion by the summation method gives values of 51.7% SiO₂, 5.0% MgO, 5.3% total alkalis and a Mg-number of 47.5. This composition lies near the base of the basaltic trachyandesite field in a TAS-diagram very close to the tholeiitic-alkaline dividing line.
- (3) This magma evolved through the trachyandesite field into the trachyte field. The final calculated magma composition has ~ 11.4% alkalis. The alkali enrichment was not due to assimilation of metapelite but rather fractional crystallization of a slightly alkali-rich parental magma.
- (4) The evolved parental magma composition led to the early crystallisation of FeTi-oxides so that there was very restricted Fe-enrichment in the magma during fractionation. Magma density decreased continually

during crystallisation. While Stage IV crystallised essentially as a closed system, repeated magma addition during the earlier stages of magma chamber evolution led to extensive compositional zoning of the magma, partly because new magma was repeatedly able to elevate the resident magma column.

- (5) The parental magma was derived from an underlying chamber where extensive fractionation of mantle-derived magma took place before repeatedly feeding the Fongen-Hyllingen chamber.

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REFERENCES

- Abu El-Rus, M.A. (2003) Trace element modelling of magma evolution in the Fongen-Hyllingen Intrusion, Trondheim region, Norway. *Journal of Mineralogical and Petrological Sciences*, 98, 47-76.
- Barnes, S.J. (1985) The petrography and geochemistry of komatiite flows from the Abitibi greenstone belt and a model for their formation. *Lithos*, 18, 241-270.
- Barnes, S.J. (1986) The effect of trapped liquid crystallization on cumulus mineral compositions in layered intrusion. *Contributions to Mineralogy and Petrology*, 93, 524-531.
- Bédard, J.H. (1994) A procedure for calculating the equilibrium distribution of trace elements among the minerals of cumulate rocks, and the concentration of the trace elements in the coexisting liquids. *Chemical Geology*, 118, 143-153.
- Best, M.G. (1963) Petrology of the Guadalupe complex, southwestern Sierra Nevada foothills, California. *Journal of Petrology*, 4, 223-259.
- Best, M.G. (1982) *Igneous and metamorphic petrology*. pp. 630. W.H. Freeman, San Francisco, USA.
- Blundy, J. (1997) Experimental study of a Kiglapait marginal rock and implications for trace element partitioning in layered intrusions. *Chemical Geology*, 141, 73-92.
- Bottinga, Y. and Weill, D.F. (1970) Densities of liquid silicate systems calculated from partial molar volumes of oxide components. *American Journal of Science*, 269, 169-182.
- Boudreau, A.E. (1999) Fluids fluxing of cumulates: The J-M reef and associated rocks of the Stillwater Complex. *Montana. Journal of Petrology*, 40, 755-772.
- Brandriss, M.E. and Bird, D.K. (1999) Effects of H₂O on phase relations during crystallization of Gabbros in Kap Edvard Holm Complex, east Greenland. *Journal of Petrology*, 10, 1037-1064.
- Bucher, K. and Frey, M. (1994) *Petrogenesis of metamorphic rocks*. pp. 318, Springer-Verlag, Berlin, Heidelberg, Germany.
- Chai, G. and Naldrett, A.J. (1992) The Jinchuan ultramafic intrusion: Cumulate of a high-Mg basaltic magma. *Journal of Petrology*, 33, 277-303.
- Chalokwu, C.I. and Grant, N.K. (1987) Re-equilibration of olivine with trapped liquid in the Duluth complex, Minnesota. *Geology*, 15, 71-74.
- Czamanske, G.K. and Mihálik P. (1972) Oxidation during magmatic differentiation, Finnmarka complex, Oslo area, Norway: Part 1, The opaque oxides. *Journal of Petrology*, 13, 493-509.
- DePaolo, D.J. (1981) Trace element and isotopic effects of combined wallrock assimilation and fractional crystallization. *Earth and Planetary Science Letters*, 53, 189-202.
- DePaolo, D.J. (1985) Isotopic studies of processes in mafic magma chambers: I. The Kiglapait Intrusion, Labrador. *Journal of Petrology*, 26, 925-951.
- Fodor R.V., Sial A.N., Mukasa S.B. and McKee E.H. (1990) Petrology, isotope characteristics, and K-Ar ages of the Maranhao, northern Brazil, Mesozoic basalt province. *Contributions to Mineralogy and Petrology*, 104, 555-567.
- Green D.H., Edgar A.D., Beasley P., Kiss, E. and Ware, N.G. (1974) Upper mantle source for some hawaiites, mugearites and benmoreites. *Contributions to Mineralogy and Petrology*, 48, 33-43.
- Green, T.H. and Watson, E.B. (1982) Crystallization of apatite in natural magmas under high pressure, hydrous conditions, with particular reference to 'orogenic' rock series. *Contributions to Mineralogy and Petrology*, 79, 96-105.
- Green, T.H., Sie, S.H., Rayan, C.G. and Cousens, D. R. (1989) Proton-microprobe determined partitioning of Nb, Ta, Zr, Sr, and Y between garnet, clinopyroxene and basaltic melt at high pressure and temperature. *Chemical Geology*, 74, 201-216.
- Grenne, T., Ihlen, P.M. and Vokes, F.M. (1999) Scandinavian Caledonide Metallogony in a plate tectonic perspective. *Mineralium Deposita*, 34, 422-471.
- Grove, T.L. and Baker, M.B. (1984) Phase equilibrium controls on the tholeiitic versus calc-alkaline differentiation trends. *Journal of Geophysical Research*, 89, 3253-3274.
- Harrison, T.M. and Watson, E.B. (1984) The behavior of apatite during crustal anatexis: equilibrium and kinetic considerations. *Geochimica et Cosmochimica Acta*, 48, 1467-1477.
- Henderson, P. (1968) The distribution of phosphorus in the early and middle stages of fractionation of some basic layered intrusions. *Geochimica et Cosmochimica Acta*, 32, 897-911.
- Holloway, J.R. and Burnham, C.W. (1972) Melting relations of basalt with equilibrium water pressure less than total pressure. *Journal of Petrology*, 13, 1-29.
- Huppert, H.E. and Sparks, R.S.J. (1985) Cooling and contamination of mafic and ultramafic magmas during ascent through the continental crust. *Earth and Planetary Science Letters*, 74, 371-386.
- Hunter, R.H. and Sparks, R.S.J. (1987) The differentiation of the Skaergaard intrusion. *Contributions to Mineralogy and Petrology*, 95, 451-461.
- Hunter, R.H. and Sparks, R.S.J. (1990) The differentiation of the Skaergaard intrusion. Reply to McBirney and Naslund. *Contributions to Mineralogy and Petrology*, 106, 248-254.
- Irvine, T.N. (1982) Terminology for layered intrusions. *Journal of*

- Petrology, 23, 127–162.
- Irvine, T.N. and Baragar, W.R.A. (1971) A guide to the chemical classification of the common volcanic rocks. *Canadian Journal of Earth Sciences*, 8, 523–548.
- Kennedy, G.C. (1955) Some aspects of the role of the water in rock melts. In the symposium “Crust of the Earth”. Geological Society of America, Special Papers, 62, 489–503.
- Klewin, K.W. (1990) Petrology of the Proterozoic Potato River Layered Intrusion, Northern Wisconsin, USA., *Journal of Petrology*, 31, 1115–1139.
- Kress, V.C. and Carmichael I.S.E. (1991) The compressibility of silicate liquids containing Fe_2O_3 and the effect of composition, temperature, oxygen fugacity and pressure on their redox states. *Contributions to Mineralogy and Petrology*, 108, 82–92.
- Kushiro, I. (1969) The system forsterite–diopside–silica with and without water at high pressures *American Journal of Science*, 267A, 269–294.
- Kushiro, I. (1975) On the nature of silicate melt and its significance in magma genesis: regularities in the shift of the liquidus boundaries involving olivine, pyroxene and silica minerals. *American Journal of Science*, 275, 411–432.
- Lange, R. and Carmichael, I.S.E. (1986) Densities of K_2O – Na_2O – CaO – MgO – Al_2O_3 – TiO_2 – SiO_2 liquids. *EOS Transactions, American Geophysical Union*, 67, 1273.
- Lange, R. and Carmichael, I.S.E. (1987) Densities of Na_2O – K_2O – CaO – MgO – FeO – Fe_2O_3 – TiO_2 – SiO_2 liquids: new measurement and derived partial molar properties. *Geochimica et Cosmochimica Acta*, 51, 2931–2946.
- Leake B.E. (1978) Nomenclature of amphiboles. *American Mineralogist*, 63, 1023–52.
- Maier, W.D. and Barnes, S.J. (1998) Concentrations of rare earth elements in silicate rocks of the Lower, Critical and Main Zones of the Bushveld Complex. *Chemical Geology*, 150, 85–103.
- Marks, M. and Markl, G. (2001) Fractionation and assimilation processes in the alkaline augite syenite unit of the Ilimaussaq intrusion, south Greenland, as deduced from phase equilibria. *Journal of Petrology*, 42, 1947–1969.
- McBirney, A.R. (1989) The Skærgaard Layered Series: I. Structure and average compositions. *Journal of Petrology*, 30, 363–398.
- McBirney, A.R. (1993) *Igneous petrology*. Second edition pp. 508, Jones and Bartlett Publishers, Boston, USA.
- McBirney, A.R. and Naslund, H.R. (1990) The differentiation of the Skaergaard intrusion. A discussion of Hunter and Sparks (Contrib Mineral Petrol 95:451–461). *Contributions to Mineralogy and Petrology*, 104, 235–247.
- Meyer, G.B. and Wilson J.R. (1999) Olivine-rich units in the Fongen-Hyllingen Intrusion, Norway: implications for magma chamber processes. *Lithos*, 47, 157–179.
- Middlemost, E.A.K. (1994) Naming materials in the magma/igneous rock system. *Annual Review of Earth and Planetary Sciences*, 37, 215–224.
- Morse, S.A. (1970) Alkali feldspars with water at 5 kb pressure. *Journal of Petrology*, 11, 221–253.
- Morse, S.A. (1980) Kiglapait Mineralogy II: Fe–Ti oxide minerals and the activities of oxygen and silica. *Journal of Petrology*, 21, 685–719.
- Morse, S.A. (1981) Kiglapait geochemistry IV: The major elements. *Geochimica et Cosmochimica Acta*, 45, 461–479.
- Naslund, H.R. (1989) Petrology of the Basistoppen Sill, east Greenland: a calculated magma differentiation trend. *Journal of Petrology*, 30, 229–219.
- Nelson, F.M. and Carmichael I.S.E. (1979) Partial molar volumes of oxide components in silicate liquids. *Contributions to Mineralogy and Petrology*, 71, 117–124.
- Neumann, E.-R. (1974) The distribution of Mn^{2+} and Fe^{2+} between ilmenites and magnetites in igneous rocks. *American Journal of Science*, 274, 1074–1088.
- Olesen, N.Ø., Hansen, E.S., Kirstensen, L.H. and Thyrsted, T. (1973) A preliminary account on the Selbu-Tydal area, the Trondheim region, Central Norwegian Caledonides. *Leidske Geologische Medelingen*, 49, 259–276.
- O'Neill, H.St.C. (1987a) The quartz–fayalite–iron and quartz–fayalite–magnetite equilibria and the free energies of formation of fayalite (Fe_2SiO_4) and magnetite (Fe_3O_4). *American Mineralogist*, 72, 67–75.
- O'Neill, H.St.C. (1987b) The free energies of formation of NiO , CaO , Ni_2SiO_4 . *American Mineralogist*, 72, 280–291.
- Osborn, E.F. (1959) Role of oxygen pressure in the crystallization and differentiation of basaltic magma. *American Journal of Science*, 257, 609–647.
- Paktunc, A.D. (1987) Differentiation of the Cuthbert Lake ultramafic dikes and related mafic dikes. *Contributions to Mineralogy and Petrology*, 97, 405–416.
- Raymond, L.A. (1995) The study of igneous, sedimentary and metamorphic rocks. pp. 742, Wm. C. Brown Communications, Inc., USA.
- Robins, B. (1982) Finger structures in the Lille Kufjord layered intrusion, Finnmark, northern Norway. *Contributions to Mineralogy and Petrology*, 81, 290–295.
- Roeder, P.L. and Emsile, R.F. (1970) Olivine–liquid equilibrium. *Contributions to Mineralogy and Petrology*, 29, 275–289.
- Sisson, T.W. and Grove, T.L. (1993a) Experimental investigations of the role of H_2O in calc-alkaline differentiation and subduction zone magmatism. *Contributions to Mineralogy and Petrology*, 113, 143–166.
- Sisson, T.W. and Grove, T.L. (1993b) Temperatures and H_2O contents of low-MgO high-alumina basalts. *Contributions to Mineralogy and Petrology*, 113, 167–184.
- Sørensen, H.S. and Wilson, J.R. (1995) A strontium and neodymium isotopic investigation of the Fongen-Hyllingen layered intrusion, Norway. *Journal of Petrology*, 36, 161–187.
- Sparks, R.S.J., Meyer, P. and Sigurdson, H. (1980) Density variation amongst mid-ocean ridge basalts: implications for magma mixing and the scarcity of primitive lavas. *Earth and Planetary Science Letters*, 46, 419–430.
- Sparks, R.S.J. and Huppert, H.E. (1984) Density changes during the fractional crystallization of basaltic magmas: fluid dynamic implications. *Contributions to Mineralogy and Petrology*, 85, 300–309.
- Spulber, S.D. and Rutherford, M.J. (1983) The origin of rhyolite and plagiogranite in oceanic crust: An experimental study. *Journal of Petrology*, 24, 1–25.
- Stopler, E. and Walker, D. (1980) Melt density and average composition of basalt. *Contributions to Mineralogy and Petrology*, 74, 7–12.
- Tegner, C., Wilson, J.R. and Robins, B. (2005) Crustal contamination in basalt and jotunite: constraints from layered intrusions. *Lithos*, 83, 299–316.
- Thayssen, F. (1998) En petrologisk og mineralkemisk undersøgelse af metamorfe bjergarter i den sydlige del af Fongen-Hyllingen Intrusionen, Norge. pp. 110, Masters Thesis,

- Department of Earth Sciences, University of Aarhus, Denmark.
- Thy, P. (1981) Titanomagnetite and ilmenite in the Fongen-Hyllingen basic complex, Norway. *Lithos*, 15, 1-16.
- Thjømøe, P.A. (1995) En geologisk undersøgelse af mafiske gange, ved Ramfjellet-Tofjellet området, Trondheim Regionen, Norge. pp. 108, Masters thesis, Department of Earth Sciences, University of Aarhus, Denmark.
- Toplis, M.J. and Carroll, M.R. (1996) Differentiation of ferro-basaltic magmas under conditions open and closed to oxygen: implications for the Skaergaard intrusion and other natural systems. *Journal of Petrology*, 37, 837-858.
- Wager, L.R. and Brown, G.M. (1968) Layered Igneous rocks. pp. 588, Oliver and Boyd, Edinburgh, UK.
- Wasserburg, G.J. (1957) The effects of H₂O in silicate system. *Journal of Geology*, 65, 15-23.
- Watson, E.B. (1979) Apatite saturation in basic to intermediate magmas. *Geophysical Research Letters*, 6, 937-940.
- Watson, E.B. and Capobianco, C.J. (1981) Phosphorus and the rare earth elements in felsic magmas: an assessment of the role of apatite. *Geochimica et Cosmochimica Acta*, 45, 2349-2358.
- Wiebe, R.A. and Snyder, D. (1993) Slow, dense replenishments of a basic magma chamber: the layered series of the Newark Island layered intrusion, Nain, Labrador. *Contributions to Mineralogy and Petrology*, 113, 59-72.
- Wilson, J.R. and Olesen, N.Ø. (1975) The form of the Fongen-Hyllingen gabbro complex, Trondheim region, Norway. *Norsk Geologisk Tidsskrift*, 55, 423-439.
- Wilson, J.R., Esbensen, K.H. and Thy, P. (1981) Igneous petrology of the synorogenic Fongen-Hyllingen layered basic complex, south-central Scandinavian Caledonides. *Journal of Petrology*, 22, 584-627.
- Wilson, J.R., and Pedersen, S. (1981) The age of the synorogenic Fongen-Hyllingen complex, Trondheim region, Norway. *Geologiska Foreningens i Stockholm Förhandlingar*, 103, 429-435.
- Wilson, J.R., Hansen, B. and Pedersen, S. (1983) Zircon U-Pb evidence for the age of the Fongen-Hyllingen complex, Trondheim region, Norway. *Geologiska Foreningens i Stockholm Förhandlingar*, 105, 68-70.
- Wilson, J.R. and Larsen, S.B. (1985) Two-dimensional study of a layered intrusion - the Hyllingen Series, Norway. *Geological Magazine*, 122, 97-124.
- Wilson, J.R., Robins, B., Nielsen, F.M., Duchesene, J.C. and Vander Auwera J. (1996) The Bjerkreim-Sokndal layered intrusion, southwest Norway. In *Layered intrusions* (Cawthorn, R.G. Ed.). pp. 531, Elsevier Science B.V., Netherlands, 231-255.
- Wilson, J.R. and Sørensen, H.S. (1996) The Fongen-Hyllingen layered intrusive complex, Norway. In *Layered intrusions* (Cawthorn, R.G. Ed.). pp. 531, Elsevier Science B.V., Netherlands, 303-329.
- Wood, B.J. and Carmichael, I.S.E. (1973) P_{total}, P_{H₂O} and the occurrence of cummingtonite in volcanic rocks. *Contributions to Mineralogy and Petrology*, 56, 297-303.
- Yoder, H.S., Stewart, D.B. and Smith, J.V. (1957) Ternary feldspars. *Carnegie Institution of Washington Yearbook*, 56, 206-214.

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