Olivine Slurry Replenishment and the Development of Igneous Layering in a Franklin Sill, Victoria Island, Arctic Canada

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ABSTRACT

The Franklin sills and dykes on Victoria Island in the Canadian Arctic represent the sub-volcanic plumbing system to the Natkusiak flood basalts, which are associated with the late Neoproterozoic (c. 723–716 Ma) break-up of Rodinia. The Lower Pyramid Sill (LPS) is the distal end of a sill complex that may be rooted in the Uhuk Massif, a major fault-guided magmatic feeder system. The LPS is unusual for a thin (c. 21 m), shallow, tholeiitic intrusion because it displays well-developed cumulate layering similar to that seen in large layered intrusions. The LPS has an aphanitic, olivine-phryic (c. 5%) Lower Chilled Margin (LCM), a (< 1 m thick) dendritic, olivine-phryic Lower Border Zone (LBZ), a (c. 7 m thick) olivine-dominated (up to c. 55%) melanogabbro–feldspathic-peridotite zone (OZ), a thin (c. 1 m) clinopyroxene-rich cumulate gabbro (CPZ) containing sector-zoned euhedral clinopyroxene, a (c. 10 m thick) doleritic gabbro zone (DZ), a (< 1 m thick) aphyric, dendritic Upper Border Zone (UBZ) and an aphanitic, olivine-phryic (c. 5%) Upper Chilled Margin (UCM). Distinct compositional groups recognized in olivines from the OZ can be associated with specific crystal morphologies, some showing significant reverse zoning. Melt compositions were calculated through application of the olivine–melt Fe¼Mg exchange coefficient. The calculations suggest that phenocrystic and primocrystic olivine (Fo88–82) in the LCM–LBZ and lower OZ formed from melts with c. 13–10 wt % MgO. Modeling implies that reversely zoned olivine primocrysts and chadacrysts have rims in equilibrium with melts of c. 10–8 wt % MgO that were saturated only in olivine (+ minor chromite), whereas some olivine cores formed from melts as evolved as c. 6–5 wt % MgO that would have coexisted with a gabbroic assemblage. The presence of multiple olivine populations in the OZ (some reverse zoned) indicates that the LPS did not crystallize from a single pulse of melt that evolved by closed-system fractional crystallization. We propose that the reverse zoning pattern records incorporation of evolved crystals, most derived from the mushy gabbroic host, when an olivine-charged replenishment under- or intraplated the partly crystallized basaltic magma, now preserved as the DZ. The intervening CPZ may also owe its origin to the emplacement of the olivine slurry, possibly as a result of pore-scale melt mixing at this interface. The DZ shows inward differentiation trends that can be explained by in situ differentiation. The data imply that late emplacement of olivine-rich crystal slurries and in situ differentiation both played a role in the development of the layered LPS.

Key words: differentiation; Franklin; igneous layering; sill; slurry
INTRODUCTION

There is considerable petrological and geochemical evidence from active volcanoes for periodic replenishment of high-level magma chambers by the influx of new magma from depth (Murphy et al., 2000; Tepley et al., 2000; Garcia et al., 2003; Humphreys et al., 2006; Morgan et al., 2006; Ginibre & Wörner, 2007). Replenishment is also well documented in layered intrusions (Wager & Brown, 1968; Huppert & Sparks, 1980; Ballhaus & Glikson, 1989; Cawthorn, 1996; Wallace & Bergantz, 2002; Namur et al., 2010). Influxes of replenishing primitive magma can (1) prolong the life of a magma chamber (Usselman & Hodge, 1978; Annen et al., 2014), (2) modify chemical fractionation trends (O’Hara, 1977; Dungan & Davidson, 2004; Reubi & Blundy, 2008), (3) affect the cumulate stratigraphy in layered intrusions (Wager & Brown, 1968; Cawthorn, 1996; Gibb & Henderson, 2006), and (4) trigger eruption (Brown, 1956; Sparks et al., 1977; Field et al., 2013). This is widely believed that primitive replenishments play a role in the formation of cyclic–macrorhythmic layering in mafic–ultramafic intrusions (Brown, 1956; Irvine & Smith, 1967; Wager & Brown, 1968; Irvine, 1977; Huppert & Sparks, 1980; Smewing, 1981).

Huppert & Sparks (1980) developed a model to explain cyclic olivine-rich–plagioclase-rich macrorhythmic layers by ponding of dense, hot, primitive magma charges beneath a resident, more evolved basaltic magma. In the application of their model to the Rum intrusion, the contrast in composition and temperature between the two liquid layers drove vigorous convection, holding olivine in suspension in the lower layer until its eventual emplacement of largely aphyric magma that evolve by differentiation trends (Parsons, 1987; Naslund, 1989). Slow cooling of large intrusions allows significant post-cumulus textural (Hunter, 1996; Holness et al., 2007) and compositional re-equilibration (Barnes, 1986) that may obscure primary magmatic signatures. This is especially true for olivine, owing to the rapid exchange of Fe and Mg between crystal and melt at magmatic temperatures (Chakroboty, 1997), whereas ‘primary magmatic’ zoning patterns of clinopyroxene (Müller et al., 2013) and especially plagioclase (Grove et al., 1984; Morse, 1984) seem to more commonly survive the post-cumulus stage. In contrast, thin (~100 m) intrusions would cool quickly when emplaced at shallow crustal levels (Carslaw & Jaeger, 1959) and might better preserve the textural and mineral-chemical evidence of the magmatic processes by which they formed.

The sills constituting the Neoproterozoic Franklin sub-volcanic plumbing system on Victoria Island are examples of such thin, quickly cooled systems. These sills are well exposed and unmetamorphosed, and the suite has a capping, co-genetic flood basalt sequence (Natuskiak flood basalts). The excellent preservation and availability of liquid compositions allow the physical and chemical linkages between sills, feeder or transfer dykes and the capping basalts to be established. In this study, we reconstruct the differentiation history of the Lower Pyramid Sill (LPS), a thin (~21 m) sill that is prominently layered, with an olivine-enriched lower layer and a gabbroic upper layer. Our textural and mineral-chemical observations from the LPS are best explained by mixing between a resident gabbroic mush and a late invading olivine slurry, and we propose that this is how the first-order olivine-rich–plagioclase-rich macro-layering formed.

REGIONAL GEOLOGY

The Franklin Large Igneous Province

The Franklin Large Igneous Province extends for >2500 km from the west coast of Greenland as far as the Great Bear Lake and Victoria Island in northern Canada and formed during the break-up of Rodinia as Siberia separated from northern Laurentia (Heaman et al., 1992). Franklin magmatism has been dated at between c. 723 and 716 Ma using concordant U/Pb ages on baddeleyite–zircon from the Coronation sills and Franklin sills (Heaman et al., 1992; Macdonald et al., 2010). Franklin intrusions are exceptionally well exposed within the Shaler Supergroup (Rainbird et al., 1996) of the Minto Inlier on Victoria Island (Fig. 1), which is a NE–SW-trending erosional window comprising intra-continental clastic and carbonate sediments (Jones et al., 2010; van Acken et al., 2013; Thomson et al., 2014). The Shaler Supergroup is overlain by the Natuskiak flood basalts, a c. 1 km thick sequence of basaltic volcanic rocks that are related to the Franklin intrusions (Baragar, 1976; Jefferson et al., 1985; Williamson (2009) and others (Shirley, 1985; Parsons, 1987; Naslund, 1989).
Franklin sills range in thickness from a few meters to c. 150 m, are mostly conformable to bedding and, where exposures permit, are seen to be laterally continuous for tens of kilometers along strike with little change in thickness (Bedard et al., 2012). There are narrow (c. 2–10 m) contact metamorphic haloes around the Franklin sills (Nabelek et al., 2013) with wider haloes enveloping some dykes (Hryciuk et al., in preparation).

The Natkusiak flood basalts are preserved as two erosional remnants in the Holman Island syncline (Fig. 1). The lowermost extrusive unit (c. 50–100 m thick) is a primitive basalt (c. 7–11 wt % MgO) characterized by light rare earth element (LREE)–large ion lithophile element (LILE) enrichment (Bedard et al., 2013). These LREE-enriched magmas are also referred to as Type-1 magmas. This lowermost extrusive unit is locally characterized by sparse olivine phenocryst pseudomorphs. Overlying this thin unit are laterally extensive basaltic (c. 6–10 wt % MgO) sheet flows with less strongly enriched trace element signatures (Bedard et al., 2013), also referred to as Type-2 magmas. The two populations of magma compositions recognized in the volcanic sequence can be correlated with those in the underlying sills. Both magma types in the sills form gabbroic–diabasic rocks that exhibit systematic inward enrichment of FeO, TiO₂ and incompatible elements as a result of in situ fractional crystallization (Naslund et al., 2013). The LREE–LILE-enriched sills (including the LPS) mostly occur in the lower part of the Shaler Supergroup and are commonly characterized by olivine-rich bases and gabbroic upper parts. They correspond geochemically to the lower primitive Natkusiak lavas. Sills formed from the younger, less enriched magma population lack olivine-rich layers, but may be prominently plagioclase–clinopyroxene phyric.

**Lower Pyramid Sill**

The LPS is located near Boot Inlet at the western end of the Minto Inlier (Fig. 1). It formed from the LREE–LILE-enriched magma population. The LPS was emplaced at a depth of c. 3 km based on its stratigraphic position.
It is one of many olivine-rich sills that were emplaced just above the Fort Collinson Formation quartz-arenite within a broad corridor along the Collingwood Hills. Collectively these olivine-enriched sills are known as the Fort Collinson Sill Complex (FCSC: Hayes et al., in preparation). The thin (c. 50–100 m) Fort Collinson marker unit is embedded in much thicker sequences of dolostone and limestone, and this mechanical discontinuity appears to have greatly facilitated intrusion of the early magma pulses in the Franklin province. The western, sill-like part (c. 40 m thick) of the Uhuk Massif (UM), located c. 50 km to the east of the LPS (Fig. 1; see Bédard et al., 2012), is another example of an olivine-enriched sill emplaced at this stratigraphic position. The LPS has a thin (c. 5–6 m) metamorphic halo surrounding it, similar to other Franklin sills (Nabelek et al., 2013).

The LPS is unusual for such a thin hypabyssal intrusion in having a well-defined basal olivine-rich cumulate layer that is overlain by a sub-ophitic doleritic gabbro (Fig. 2b). The LPS has chilled contacts against a thin septum of fissile Jago Bay Formation carbonates, only a few meters above the top of the Fort Collinson quartz-arenites (Fig. 2a). The LPS (Fig. 2b) comprises an aphanitic Lower Chilled Margin (LCM), a thin (<1 m) dendrite-textured Lower Border Zone (LBZ), a c. 7 m thick sequence of olivine cumulate melagabbro to feldspathic-peridotite with c. 40–55% modal olivine (OZ), a c. 1 m thick layer of clinopyroxene-rich gabbro (CPZ), a c. 10 m thick layer of sub-ophitic doleritic gabbro (DZ), a thin (<1 m), dendrite-textured Upper Border Zone (UBZ) and a thin (<1 m) aphanitic Upper Chilled Margin (UCM). We have taken 26 samples from a single LPS profile with sample locations constrained by global positioning system (GPS) and height above the lower contact measured in the field (data are given in the Supplementary Data electronic appendices; supplementary data are available for downloading at http://www.petrology.oxfordjournals.org). We analysed all samples for their whole-rock major element concentrations (Fig. 2b) and 22 for mineral compositions.

ANALYTICAL METHODS

Whole-rock analysis

Rock samples were reduced to powder in the rock preparation laboratory of Cardiff University. Each sample was crushed to coarse grit using a steel jaw crusher before 80 g of each sample was milled in an agate planetary ball mill to a fine powder. Afterwards, 2 g of each powdered sample was ignited for 2 h in a furnace at 900°C to burn off volatile substances and determine loss on ignition (LOI) values. The samples were then put into solution for inductively coupled plasma optical emission spectrometry (ICP-OES) and inductively coupled plasma mass spectrometry (ICP-MS) analysis using the lithium metaborate fusion method (McDonald & Viljoen, 2006). The instruments used at Cardiff University to analyse element abundances were a Jobin Horiba Ultima 2 ICP-OES system and a Thermo Elemental ZT series ICP-MS system. The full whole-rock dataset is included in the Supplementary Data electronic appendices.

Electron microprobe

In situ mineral chemical analyses were carried out using a Cameca SX-100 electron microprobe at Laval University in Québec City, Canada. The probe was fitted with five wavelength-dispersive spectrometers. All phases were analysed on polished thin sections using a focused beam (1 μm diameter), 20 nA current and a 15-kV accelerating potential. Counting times were typically 30 s on peak and background levels. Core to rim profiles were carried out on the major silicate phases. Where grain size was very fine or only a representative core composition was required, spot analyses were obtained. The dataset for each phase analysed and the analytical conditions for each mineral species are supplied in the Supplementary Data electronic appendices.

Fig. 2. (a) Photograph of the LPS showing the lower contact just above the Fort Collinson Formation quartz-rich sandstone and the upper contact against Jago Bay Formation dolostone. (b) The internal stratigraphy of the LPS showing bulk-rock variations in MgO and TiO₂ concentration (in wt %).
PETROGRAPHY

Lower Chilled Margin (LCM)
The LCM (sampled at c. 5 cm above the contact) has ≤5% euhedral olivine phenocrysts (c. 0.5–2 mm), which may contain tiny chromite microphenocrysts as inclusions. The olivine phenocrysts are surrounded by a fine-grained (c. 0.1 mm) groundmass (Fig. 3a) that grades into coarser dendrites of plagioclase (Plag) and clinopyroxene (Cpx) that cement olivine phenocrysts (Olph). Olivine (Ol) is also intergrown with the dendrites, suggesting that the melt reached the three-phase cotectic early in its crystallization history (image in crossed Nicols).

Lower Border Zone (LBZ)
In the LBZ, the proportion of olivine phenocrysts (c. 1–2 mm, also with minor Cr-spinel inclusions) increases to c. 10–15% (at c. 20 cm above the lower contact). The dendritic intergrowths of plagioclase and clinopyroxene reach c. 2 mm grain sizes with dendrites commonly nucleating on the surface of olivine phenocrysts (Fig. 3b). The rims of some olivine crystals are intergrown with the plagioclase–clinopyroxene dendrites, suggesting that the three phases co-saturated during crystallization of the LBZ (Fig. 3b). The textures, with the presence of cumulus olivine, are consistent with a weakly olivine-cumulative composition for the LBZ.

Olivine Zone (OZ)
There is a noticeable textural change c. 1 m above the lower contact of the LPS. At this height, modal olivine increases over a few centimeters to c. 40% and the groundmass texture changes from dendritic to medium-grained poikilitic (Fig. 4a). We define this modal and textural transition as the boundary between the LBZ and the OZ. The OZ is c. 7 m thick and is composed of cumulate olivine melagabbro to feldspathic-peridotite. Cr-spinel microphenocrysts are present throughout the OZ, typically as inclusions in olivine primocrysts and as grains embedded in interstitial clinopyroxene and plagioclase.

Olivine within the OZ can be broadly assigned to four textural groups: (1) euhedral olivine primocrysts (c. 0.5–2 mm), which commonly contain euhedral
Cr-spinel microphenocrysts (Fig. 5a and c); (2) chadacrystic fine-grained (c. 0.2–0.5 mm) euhedral–subhedral olivine crystals (some bearing Cr-spinel inclusions), which are enclosed by clinopyroxene oikocrysts (Figs 4a, 5a and 6f); (3) fine-grained (c. 0.2–0.5 mm) groundmass olivine crystals (some bearing Cr-spinel inclusions) surrounded by interstitial plagioclase (Figs 4b, 5a and 6f); (4) hopper olivines (c. 0.2–2 mm), which are skeletal in habit in cross-section and orthorhombic in basal sections, where they appear to enclose clinopyroxene and plagioclase (Fig. 5d–f). Hopper olivines also have anhedral margins indicating co-crystallization with plagioclase (Fig. 5d and e). We subdivide the OZ based on the spatial distribution of these olivine textures into three subzones: lower (LOZ), middle (MOZ) and upper (UOZ) olivine zone. The LOZ consists of a basal pyroxene-poikilitic olivine-melagabbro, which mainly consists of euhedral olivine primocrysts, olivine chadacrysts and groundmass olivines (Figs 4a, 5a and 6f), with only minor (c. 5%) proportions of fine-grained (<0.5 mm) hopper olivine. The MOZ is dominated by hopper olivine crystals that are typically intergrown with clinopyroxene, with subordinate primocrystic, chadacrystic, and groundmass olivine (Fig. 5b and c). The
UOZ is dominated by hopper olivine morphologies with subordinate euhedral primocrystic and groundmass olivine, and few olivine chadacrysts hosted by poikilitic clinopyroxene (Fig. 5d–f).

Clinopyroxene (c. 1–2 mm) habits range from euhedral to subhedral in the LOZ and it commonly encloses olivine chadacrysts (Figs 4a and 6). Clinopyroxene becomes more subhedral to anhedral in the MOZ, where...
intergrowths with olivine are common (Fig. 5b). In the UOZ, euhedral cumulus clinopyroxene crystals appear (c. 1–2 mm; Fig. 5d and f), substituting for the clinopyroxene olivine crystals that typify the LOZ and the anhedral clinopyroxenes present in the MOZ. Subordinate intergranular, interstitial clinopyroxene occurs throughout the OZ.

Plagioclase is anhedral throughout the OZ and is interstitial to cumulus olivine (Figs 4b and 6) and mostly interstitial to clinopyroxene, except for local co-crystallization textures (Fig. 5b, d and e). There are minor accessory amphibole and biotite, and there are traces of groundmass Fe–Ti oxides and Fe–Ni–Cu sulphides.

**Clinopyroxene Zone (CPZ)**
The CPZ is c. 1 m thick. The diffuse base of the CPZ is characterized by a drop in the mode of olivine to c. 5% and a corresponding increase in the proportion of euhedral clinopyroxene olivine to c. 30–40% (c. 0.5–1 mm). The euhedral clinopyroxene crystals are often sector-zoned (Fig. 7) and are distributed as trains of grains (Fig. 8), reminiscent of those described by Philpotts et al. (1998). Between these trains of sector-zoned clinopyroxene is a gabbroic matrix consisting of serpentinized olivine (c. 5%), euhedral–subhedral clinopyroxene and interstitial plagioclase with subordinate interstitial clinopyroxene. There is minor (<5%) magnetite and ilmenite, as well as accessory amphibole and biotite.

**Dolerite Zone (DZ)**
The DZ is characterized by the absence of olivine and modal increases in plagioclase (from c. 25% in the OZ to 40% in the DZ) and clinopyroxene (from c. 30% in the OZ to c. 40% in the DZ). The DZ is also marked by a change in the morphology of plagioclase, from subhedral–anhedral interstitial textures to coarser (c. 2–3 mm) lath-shaped crystals that mostly form ophitic intergrowths with clinopyroxene. Around 5–10% of the plagioclase–clinopyroxene mode is composed of cumulus crystals that do not display intergrowth textures. Full thin section scans show domains of sub-ophitic dolerite with varying grain sizes (vari-textured), with coarser pegmatitic patches (c. 5 mm plagioclase and c. 3–4 mm clinopyroxene) and finer-grained domains (c. 2–3 mm plagioclase and c. 1–2 mm clinopyroxene) present in the DZ.

Clinopyroxene is typically euhedral–subhedral and intergrown with plagioclase, but both augite and low-Ca pigeonite also occur as separate cumulus crystals. Low-temperature alteration is more common in the DZ compared with other zones of the LPS, with plagioclase and clinopyroxene being partly altered to fine-grained albite and saussurite, or chlorite and actinolite (respectively). This is interpreted to represent deuteric alteration caused by volatile exsolution during the final stages of LPS crystallization. Olivine was not observed in the DZ, but sparse orthopyroxene grains (c. 5–10%) in the lower DZ may be pseudomorphous after olivine. Interstitial ilmenite and magnetite are present (c. 5–10%), as are interstitial sulphides (c. 5%), with associations of pyrrhotite-pyrite in the lower DZ grading into associations of pyrite-chalcopyrite in the upper DZ. Minor accessory amphibole and biotite occur throughout the DZ and may be associated with interstitial granophyre in the uppermost DZ (c. 19–20 m height). Also, in the uppermost DZ, long (c. 5–10 mm), blade-like clinopyroxene crystals are present with preferred growth directions roughly perpendicular to the upper contact of the LPS (element maps are included in the Supplementary Data electronic appendices).

**Upper Border Zone (UBZ)**
The UBZ (from c. 20 m to c. 21 m height) consists largely of fine-grained (<1 mm) plagioclase and clinopyroxene dendrites. There are also coarser subhedral–anhedral clinopyroxene crystals (c. 1–2 mm), which show both sector-zoning and complex concentric zoning, locally with sieve-textures. No olivine phenocrysts (nor their pseudomorphs) were observed in the UBZ.

**Upper Chilled Margin (UCM)**
The LPS at its upper contact with the Jago Bay Formation carbonate is similar to the LCM in that it is a very fine-grained, almost glassy, chilled contact. It contains c. 5% olivine phenocrysts (c. 1–2 mm) that have been completely pseudomorphed by secondary serpentine. The groundmass appears to consist of very fine-grained dendrites, but it is more altered in comparison with the LCM. A few amygdales (<5%; c. 2 mm) are filled with biotite and chlorite.

**MINERAL CHEMISTRY**
The main silicate phases present in each layer in the LPS were analysed by electron microprobe. Chemical zonation were mapped to constrain the crystallization history of the LPS. We use the term ‘core’ when referring to the central region of a crystal, ‘mantle’ for the intermediate part between the core and rim (often with weak compositional gradients) and ‘rim’ for the strongly zoned edges of a crystal. We use ‘primocryst’ when we refer to euhedral cumulus crystals, which are typically ≥1 mm in size, and ‘phenocryst’ for euhedral crystals in the chills and border zones.

**Olivine**
Compositional variations of olivine in the LCM, LBZ and OZ are summarized in Fig. 9. Variations in Fo content [molar Fo = 100 Mg/(Mg + Fe²⁺)] are typically correlated with olivine NiO (wt %) contents.

Olivine phenocrysts in the LCM are normally zoned with near-constant core and mantle compositions and narrow, more evolved rims. The average core and rim compositions are Fo₉₉ and Fo₅₂, respectively. The cores and mantles of olivine phenocrysts in the LCM have an average NiO content of 0.32 wt %, whereas the rims have a lower NiO content of 0.28 wt %. Some euhedral olivine phenocrysts show subtle core to mantle
increases in Fo content (c. 0.3–0.4 mol % increases) followed by decreasing Fo at the rim. A small olivine phenocryst in the LCM displays unusually strong zoning from Fo$_{86}$ (0.28 wt % NiO) in the core to Fo$_{75}$ (0.24 wt % NiO) at its rim.

In the LBZ, the average core composition of olivine phenocrysts is Fo$_{88}$ (0.29 wt % NiO). Olivine commonly shows strong normal zoning with an average rim composition of Fo$_{78}$ (0.2 wt % NiO). Locally, some olivine phenocrysts in the LBZ show weak reverse zoning from

Fig. 7. BSE (a) and element maps (b–f) of sector-zoned clinopyroxenes from the CPZ. (a) BSE image. (b) Ti, (c) Cr, (d) Ca, (e) Mg, and (f) Al element maps. The hourglass style sector-zoning is characterized by zones with 2–3 times more enrichment in both Cr and Al.
core to mantle (c. 1 mol %), but with little corresponding change in NiO content (average c. 0.32 wt %).

In the LOZ, olivine primocrysts have core compositions ranging between Fo$_{85}$ and Fo$_{82}$ (0.4–0.25 wt % NiO). Smaller olivine chadacrysts (enclosed by clinopyroxene oikocrysts) have systematically lower Fo contents and a restricted range of compositions with an average core composition of Fo$_{83}$ (0.25 wt % NiO) and an average rim composition of Fo$_{81}$ (0.2 wt % NiO). Groundmass olivine surrounded by plagioclase has still more evolved compositions with average core and rim compositions of Fo$_{81}$ (0.21 wt % NiO) and Fo$_{79}$ (0.1 wt % NiO), respectively. Most olivines in the LOZ are zoned. The larger olivine primocrysts commonly exhibit normal zoning with near-constant core and mantle compositions, and narrow Fe-rich rims as evolved as Fo$_{77}$–75. Chadacrystic olivines enclosed by clinopyroxene oikocrysts are mostly normally zoned (typically a c. 1–2 mol % change from core to rim) except for a weakly reverse-zoned olivine grain that has a core composition of Fo$_{82}$ (0.2 wt % NiO) that increases to Fo$_{83}$ (0.24 wt % NiO) at its rim. Groundmass olivines that are surrounded by plagioclase are generally normally zoned and reach very low Fo compositions at the rims (c. Fo$_{77}$–79), but may also show weak reverse zoning with core to rim increases of c. 0.5 mol % (with NiO contents <0.2 wt % for both core and rim).

In the MOZ and UOZ, the relative proportion of olivine primocrysts decreases, whereas the total olivine mode increases (up to c. 55%). Olivines enclosed by clinopyroxene oikocrysts become less common, but where present they have a restricted compositional range (Fo$_{83}$–81). In the MOZ, the average Fo composition of groundmass olivines is generally more primitive (c. Fo$_{80}$) than the average groundmass olivines from the LOZ (c. Fo$_{79}$). Subtle (c. 1 mol %) reverse zoning is observed in both groundmass and hopper olivines in the MOZ.

In the UOZ, common hopper olivines have core compositions as low as Fo$_{72}$ and commonly display reverse zoning with rims between Fo$_{85}$ and Fo$_{77}$. An olivine primocryst in the UOZ displays reverse zoning (Fig. 10), with a core to rim increase from Fo$_{77}$ (0.16 wt % NiO) to Fo$_{88}$ (0.23 wt % NiO). Groundmass olivine in the UOZ shows reverse zoning for Fo, with cores of c. Fo$_{77}$ and rim compositions ranging between Fo$_{80}$ and Fo$_{78}$ (with NiO content for core and rim typically <0.22 wt %).

**Cr-spinel**

Cr-spinel microphenocrysts enclosed by phenocrystic olivine in the LCM have compositions that depend on their location in the host olivine phenocryst. For example, a Cr-spinel in the core of an olivine phenocryst has a Cr$\#$ [= 100Cr/(Cr + Al)] of 50, whereas a Cr-spinel included in the rim of the same olivine has a higher Cr$\#$ of 58. Neither Fe$_3$# [= 100Fe$^{3+}$/(Fe$^{3+}$ + Cr + Al)] nor Fe$_2$# [= 100Fe$^{2+}$/(Mg + Fe$^{2+}$)] for these LCM Cr-spinel microphenocrysts shows any variation, however.

Cr-spinels inclusions in the LOZ have very similar compositions (Cr$\#$ 73–46; Fe$_3$# 18–10; Fe$_2$# 84–41) to
Clinopyroxene oikocrysts have near-constant core-mantle compositions (Fig. 12a), but may have narrow Fe–Ti-rich rims (Mg# c. 75; Fig. 12a; also see Fig. 6b and d). The evolved rim compositions of oikocrystic and cumulus clinopyroxene in the OZ overlap with the compositions of interstitial OZ clinopyroxene, and with those of LBZ groundmass clinopyroxene (Fig. 11).

Some LOZ clinopyroxene oikocrysts show subtle sector-zoning, visible on element maps of Cr, Al and Mg (Fig. 6c–f); Cr and Al (Fig. 6c and e) appear to show similar behaviour and are slightly enriched in the north–south sector, whereas Mg (Fig. 6f) is more enriched in the east–west sector of the clinopyroxene.

In the CPZ, euhedral cumulus clinopyroxene, which are often sector-zoned, have core compositions similar to those of OZ clinopyroxene (Mg# c. 83; Fig. 11), whereas rims extend to compositions that are more Fe-rich (Mg# 70–30) than clinopyroxene rims in the underlying OZ. The sector-zoning of clinopyroxene is apparent on element maps (Fig. 7), with different zones showing weak contrasts in Cr and Al abundance (Fig. 7c and f), whereas Ca and Mg show no clear inter-zone differences (Fig. 7d and e). Cr2O3 typically decreases from c. 0.6 wt % in the cores to c. 0.2 wt % in the rims, whereas Al2O3 increases from c. 1.8 wt % in the cores to c. 2 wt % at the rims. TiO2 contents also increase slightly from the cores (c. 0.4 wt %) to the rims (c. 0.5 wt %), which contrasts with the much greater increase in TiO2 (to c. 1 wt %) seen in OZ clinopyroxene rims. We suspect that the deficit in TiO2 of CPZ clinopyroxene rims may be due to sequestration of TiO2 into interstitial ilmenite in the case of the CPZ. The CPZ grades upward over c. 0.8 m into slightly more evolved clinopyroxene core compositions (Mg# 80), which have reversely zoned mantles (up to Mg# 83), suggesting interaction with a more MgO-rich melt. The presence of wide, strongly evolved rims on all forms of clinopyroxene in the CPZ suggests that all have interacted with abundant, late, evolved pore melts during the final stages of crystallization, possibly suggesting higher trapped melt fractions in these rocks.

In contrast to the muted variations of clinopyroxene compositions in the OZ, clinopyroxene in the DZ shows systematic up-section cryptic variations (Fig. 11). At the base of the DZ, clinopyroxene compositions overlap with those of clinopyroxene from the OZ and CPZ (Mg# c. 84). A cumulus-textured clinopyroxene at the base of the DZ (Fig. 14a) shows complex zoning, with a core composition of Mg# 84 that increases outward gradually to Mg# 85 and then shows a sudden decrease out to Mg# 63 in the rim. This pattern of Mg-enrichment in the mantle of cumulus clinopyroxene occurs sporadically in the DZ (Fig. 11). More generally, clinopyroxene displays strong normal core to rim zoning throughout the DZ with changes in Mg# ranging between c. 10 and 20 mol %. Rims of euhedral clinopyroxene overlap with the compositions of evolved interstitial clinopyroxene in the DZ. Interstitial clinopyroxene from the DZ is more evolved than interstitial clinopyroxene from the OZ.

 Clinopyroxene

The variation of Mg# [molar Mg# = 100 Mg/ (Mg + Fe2+ + Mn)] of clinopyroxene versus stratigraphic height is summarized in Fig. 11. The Mg# of groundmass clinopyroxene in the LBZ is relatively low (Mg# 77–70) compared with clinopyroxenes of the overlying OZ. Most LBZ clinopyroxene crystals are normally zoned (decreasing Mg# towards the rims), but some are reversely zoned, with Mg# increasing by up to c. 3 mol % towards the rim (Fig. 12b).

In the OZ, clinopyroxene core compositions show little variation with stratigraphic position (Fig. 11) with Mg# 86–83 and average Cr2O3 of c. 0.8 wt % and average TiO2 of c. 0.4 wt % (Figs 12a and 13).
In the upper DZ, clinopyroxenes (including bladed clino-
pyroxene crystals) have core compositions of Mg# 75–70 and strong normal zoning at the rims (to Mg# 30–20). The peak in clinopyroxene Fe-enrichment, just below the UBZ, may represent a type of ‘sandwich’ horizon.

The UBZ contains a rare, euhedral (c. 1 mm), sector-
zoned clinopyroxene with compositions of Mg# 81 in the crystal core (Fig. 14b), with a narrow Fe–Ti-enriched rim (element maps of this crystal are included in the Supplementary Data electronic appendices). This crystal is sieve-textured with mantle depletion in Mg# (Mg# 70–67; Fig. 14b) relative to the rims, which may indicate a thermal spike, or a reaction with more primitive melt. Interstitial clinopyroxene is intergrown with plagioclase and has evolved compositions (Mg# c. 67).

Fig. 11. Stratigraphic column showing clinopyroxene compositional variations in molar magnesium number \( [\text{Mg#} = 100 \frac{\text{Mg}}{\text{Mg} + \text{Fe}^{2+} + \text{Mn}}] \) through the LPS. The grey shaded rectangle in the OZ indicates the range of poikilitic and cumulus clinopyroxene core compositions in the OZ (between Mg# 85 and 82), illustrating the restricted range in composition.

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Plagioclase
Plagioclase anorthite \([\text{An}; \text{molar An} = 100 \frac{\text{Ca}}{\text{Ca} + \text{Na} + \text{K}}]\) content versus stratigraphic height is shown in Fig. 15. Plagioclase dendrites within the LCM groundmass have relatively low An values, averaging \( \text{An}_{64} \). In the LBZ, the coarser plagioclase dendrites become slightly more primitive with compositions of \( \text{c. An}_{71} \).
Plagioclase has core compositions of An\textsubscript{73-68} throughout the OZ. Most grains show normal zoning with sporadic occurrences of reverse zoning. Plagioclase rim compositions extend to An contents as low as An\textsubscript{58}, with the lowest An values adjacent to olivine crystals. Core compositions of interstitial plagioclase in the CPZ (Fig. 16a) are between An\textsubscript{72} and An\textsubscript{70}, whereas rims are strongly evolved (c. An\textsubscript{40}). CPZ plagioclase core compositions overlap the composition of OZ plagioclase. In the CPZ, minor euhedral cumulus plagioclase is more primitive with core and mantle compositions up to An\textsubscript{76}.

Plagioclase core and mantles are typically c. An\textsubscript{70-60} in the DZ, with strongly evolved rims as low as An\textsubscript{45}. Rare, cumulus-textured plagioclase in the lower DZ has some of the most primitive An compositions (An\textsubscript{77-76}; Fig. 16b) in the LPS. In some plagioclase crystals in the middle DZ, the cores are mantled by higher An plagioclase, suggesting interaction with a more primitive melt prior to the initiation of in situ fractional crystallization.

The core and rim compositions with the lowest An values occur in the upper DZ (Fig. 15) where clinopyroxene has the lowest Mg\# (Fig. 11), corresponding to the inferred position of the sandwich horizon.
MgO, closer to the LPS chill composition. The most evolved olivine phenocrysts rims yield model melts of c. 6–5 wt % MgO. In the LBZ, phenocrystic olivines have cores equivalent to model melts of c. 11 wt % MgO, and have narrow, Fe-rich rims equivalent to model melts of c. 5 wt % MgO.

Olivine primocrysts cores from the LOZ yield model melts between c. 13.5 and 10 wt % MgO, whereas their rims yield a melt range of c. 8–6 wt % MgO. Olivine chadacrysts enclosed by clinopyroxene oikocrysts yield magmatic model melts between c. 10 and 8.5 wt % MgO. A reversely zoned olivine chadacryst has a core in equilibrium with a model melt of 9.8 wt % and a rim in equilibrium with a model melt of 10.2 wt % MgO. Groundmass olivines in the LOZ are in equilibrium with model melts ranging between c. 7 and 5 wt % MgO. In the UOZ, primocrystic olivine and hopper olivine cores yield evolved model melts ranging between c. 8 and 5 wt % MgO. A reversely zoned olivine primocryst (Fig. 10) in the UOZ has a core in equilibrium with a model melt of c. 6.3 wt % MgO, whereas its rim yields a model melt of c. 8.5 wt % MgO. Groundmass olivines in the MOZ and UOZ yield model melts of c. 8 wt % MgO.

Climopyroxene

To calculate the MgO and FeO of the melt in equilibrium with clinopyroxene we used the clinopyroxene Fe/Mg ratio and a clinopyroxene–melt $K_a$ calculated using equation (14) of Bédard (2010). Interstitial LBZ clinopyroxene yields evolved model melts of c. 5–3.5 wt % MgO. In the LOZ and MOZ, clinopyroxene oikocryst cores yield primitive model melts between c. 9 and 7 wt % MgO, as do euhedral cumulus clinopyroxene cores in the UOZ. All clinopyroxene rims in the OZ yield Fe-rich compositions with model melts <7 wt % MgO. The cores of cumulus clinopyroxenes from the CPZ yield compositions (c. 9–6 wt % MgO), similar to the range of model melts recorded by OZ clinopyroxene, whereas their evolved rims and interstitial clinopyroxene in the CPZ yield very evolved model melts with <5 wt % MgO. Model melts calculated from the DZ clinopyroxene cores show cryptic up-section variation with melt MgO broadly ranging between c. 8 and 1 wt %. In the UBZ, dendritic clinopyroxenes correspond to model melts ranging between c. 7 and 1 wt % MgO. A coarser, sector-zoned clinopyroxene in the UBZ yields model melts of c. 6–5 wt % MgO, with values of c. 2 wt % MgO in the sieve-textured mantle and more primitive melts of c. 5 wt % at its rims.

DISCUSSION

Constraints provided by the mineral-chemical data and inverse melt compositions

Olivine and clinopyroxene in the LPS yield a broad range of model melt MgO compositions. When coupled with the textural observations, these allow us to track the chemical evolution of melts within the LPS. The cores of euhedral olivine phenocrysts and primocrysts
within the LCM, LBZ and LOZ yield the most primitive model melt MgO compositions (c. 13-5–10 wt % MgO), equal to or higher than the MgO content of their host chills (10–3 wt %), but similar to the most primitive chilled margins from other Type-1 Franklin sills (Bédard et al., 2013). The majority of olivine phenocryst rims from the LCM, as well as olivine phenocryst cores from the LBZ, are, however, in approximate equilibrium with the LCM composition. The model melt compositions in equilibrium with olivine chadacrysts included in clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %, but similar to the most primitive chilled margins from other Type-1 Franklin sills (Bédard et al., 2013). The majority of olivine phenocryst rims from the LCM, as well as olivine phenocryst cores from the LBZ, are, however, in approximate equilibrium with the LCM composition. The model melt compositions in equilibrium with olivine chadacrysts included in clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions. Because these clinopyroxene oikocrysts have MgO values between c. 10 and 8 wt %. The melt values calculated to be in equilibrium with the most primitive poikilitic clinopyroxene (c. 9–8 wt % MgO) fall within this range of melt compositions.

Fig. 14. (a) Plane-polarized light photomicrograph of a euhedral clinopyroxene crystal from the DZ, with the corresponding microprobe traverse (Nos 22–26). The clinopyroxene shows outward increase of Mg# and Cr$_2$O$_3$ in its mantle before developing normal zoning at its rim, suggesting that it may have equilibrated with a more primitive melt during crystallization. (b) BSE image of a clinopyroxene crystal from the UBZ that has a sieve-textured core–mantle. The microprobe traverse (Nos 10–20) shows complex zoning with a high-Mg# core followed by low-Mg# mantles and outwardly decreasing Mg# at the rims, a pattern that is mimicked by Al$_2$O$_3$. The primitive melt values calculated from the cores of olivine and clinopyroxene are in stark contrast to the values for melts calculated from olivine cores in the UOZ. Melt values in equilibrium with UOZ olivine cores are as low as c. 5 wt % MgO, whereas the rims extend to melt values of c. 8 wt % MgO. These reversely zoned trends are seen in hopper olivines, groundmass olivines and rare primocrystic olivines in the UOZ. Reversely zoned primocrystic olivines contain Cr-spinels that mirror the Fo trend (Fig. 10). Cr-spinel hosted within the low-Fo olivine core has lower Cr# and higher Fe2# and Fe3# than the Cr-spinel hosted by the more primitive olivine rim. This pattern suggests that the Cr-spinel in the primocryst rim formed from a less evolved melt than the one in the core, as also implied by the host olivine primocryst zoning.

In addition to the common occurrence of reversely zoned olivines in the UOZ, groundmass olivines become more primitive upward in the OZ. In the LOZ, groundmass olivines typically have compositions of Fo$_{75}$ (c. 5 wt % MgO model melt). This increases to Fo$_{90}$ (c. 8 wt % MgO melt model melt) in the MOZ and UOZ. MOZ and UOZ groundmass olivines are therefore in equilibrium with the rims of reversely zoned UOZ olivines, suggesting that they may have crystallized from the same melt. The shift to more primitive groundmass olivine compositions upward through the OZ may indicate a change in the crystallization conditions or differing proportions of trapped melt in this region of the OZ.
Recorded clinopyroxene model melt variations at the base of the DZ are in equilibrium with OZ clinopyroxenes (c. 9–7 wt % MgO), before becoming more evolved up-section to reach minimum MgO contents (c. <4 wt % MgO) in the upper DZ (sandwich horizon). Rims to euhedral clinopyroxenes in the DZ are in equilibrium with DZ interstitial clinopyroxenes (c. 4–1 wt %). Throughout the LPS, clinopyroxene rims locally record extreme Fe–Ti-enrichment, presumably as a result of extensive in situ differentiation of pore melts (see Humphreys, 2009). If the interstitial clinopyroxenes throughout the LPS crystallized from late, evolved pore melt then this implies that the larger clinopyroxenes may also have crystallized from or equilibrated with this melt. The
upward evolution in clinopyroxene rim compositions (eventually to values of Mg\# c. 20; Fig. 11) is matched by shifts in the plagioclase rim–interstitial plagioclase compositions (c. An40; Fig. 15) suggesting the presence of a sandwich horizon, possibly formed by migration and pooling of evolved pore melt (Boudreau & Philpotts, 2002) in the last part of the sill to solidify (i.e. the mid–upper DZ). A coarse UBZ clinopyroxene crystal yields a broader range of model melts (c. 7–4 wt % MgO) compared with dendrites in the LBZ (c. 4–3 wt % MgO) and may therefore be a ‘xenocrystic’ crystal that was carried in with the first pulse of LPS magma.

PELE modeling
To help constrain the crystallization history of the sill, we ran the PELE program (Boudreau, 1999) using the LPS chill (c. 10.3 wt % MgO) as a starting composition. A fractional crystallization model was run at a pressure of c. 0.8 kbar, corresponding to the estimated height of the overlying stratigraphic column. Melt CO\_2 was set at 0.2 wt %, S at 0.12 wt % and H\_2O at 0.2 wt %, and chromite was excluded from the assemblage (justifications are provided in the Supplementary Data electronic appendices). PELE predicts a liquidus at 1245°C where Fo\(_{98}\) olivine crystallizes, values similar to the core compositions of olivine phenocrysts in the LCM. After c. 8% olivine was extracted, PELE predicts that the melt co-saturates in clinopyroxene (Mg\# 87) and plagioclase (An\(_{80}\)) at c. 7.5 wt % MgO (1175°C), at which juncture the composition of olivine is Fo\(_{84}\). PELE predicts that olivine reaches compositions of Fo\(_{75}\) (similar to primocryst rims in the LOZ and the most evolved primocryst cores from the UOZ) at c. 1115°C, at which point the sill would be 40% solidified, comprising 11% olivine + 16% plagioclase + 13% clinopyroxene + 60% pore melt with an MgO content of 6 wt %. In conclusion, the crystallization pathway predicted by PELE is broadly compatible with the observed compositions and textures of olivine in the LPS OZ. However, the most primitive model melts calculated to be in equilibrium with clinopyroxene (c. 9 wt %) exceed the melt MgO composition at which clinopyroxene becomes co-saturated according to the PELE model (c. 7.5 wt %). Conversely, the model melts calculated to be in equilibrium with most DZ clinopyroxene are in accord with the PELE results. We now discuss the role that fractional crystallization may have played in the differentiation of the LPS.
fractional crystallization by sequential crystallization against the floor (Figs 9, 11 and 15). Given the extreme compatibility of nickel in olivine (Arndt, 1977; Hart & Davis, 1978), nickel in accumulated olivine should show a marked and rapid up-section decrease, a pattern that is not observed in the LPS OZ. Alternatively, the uniform compositions of the various olivine morphologies in the LPZ OZ could be explained if these olivines remained in suspension and re-equilibrated with melt (equilibrium crystallization path) prior to deposition, as in the model proposed by Huppert & Sparks (1980) and Tait (1985). This scenario also fails to explain the observations, because in such a model the deposited olivines should be essentially unzoned, or at least show a consistent normal zoning pattern, whereas olivines in the LPS OZ show marked diversity of composition and zoning pattern. The common preservation of hopper olivine habits in the LPS OZ also seems inconsistent with this scenario, because hopper olivines would have recrystallized during settling to become more euhedral (Welsch et al., 2012).

Nor are the modal abundances of olivine in the LPS compatible with the amounts of olivine predicted to crystallize from the magma by the PELE modeling. The olivine present in the OZ (when redistributed throughout the LPS) represents c. 15–20% of the volume of the LPS (assuming constant thickness). This is far in excess of the c. 8% olivine-only crystallization predicted by PELE. This suggests that the amount of olivine in the LPS OZ was not extracted by in situ crystallization of a magma similar to the LPS chill. PELE models with a starting composition corresponding to some of the most primitive Franklin chills (c. 13 wt % MgO) crystallize slightly more olivine (c. 10–15%) prior to co-saturation in plagioclase and clinopyroxene and provide better fits to the observed modal abundances of olivine in the LPS OZ. It might be possible to explain the observed olivine modal abundances if the LPS was filled with a more primitive magma (in comparison with the observed chilled margin composition), or if the initial melt pulse carried a significant cargo of olivine crystals (c. 20%). However, neither of these alternative models is compatible with the variety of olivine zoning patterns and their organized distribution within the OZ.

In conclusion, we contend that the common occurrence of reversely zoned olivines in the UOZ and the presence of evolved Cr-spinel inclusions in the low-Fo olivine cores of such grains (Fig. 10) cannot be explained by in situ fractional crystallization of a single melt pulse and require the operation of an additional mechanism during sill formation and differentiation.

The significance of reversely zoned olivine
The survival of strongly zoned olivine in the LPS, particularly the reversely zoned olivines, and the heterogeneous olivine compositions from single thin sections imply preservation as a result of rapid cooling. The presence of heterogeneous olivine compositions within single thin sections is reminiscent of the diversity of
olivine Fo compositions observed in Hawaiian picrites (Garcia et al., 2003). Preliminary olivine Fe:Mg diffusion calculations following the method of Costa & Dungan (2005) suggest that the residence time of olivine at magmatic temperatures (with a closure temperature of c. 1150°C) was between c. 2 and 5 years. Following the method of Carslaw & Jaeger (1959) for ‘simple’ conductive heat loss of a sill to wallrock at 75°C, between 3 and 6 years are needed to cool magma to its solids, similar to the time calculated above. Using a method similar to that described by Cawthorn & Walraven (1998) yields a very similar cooling rate. These calculations indicate that the LPS cooled rapidly and imply that opportunities for re-equilibration were limited. Consequently, we conclude that the mineral-chemical signatures recorded in the LPS are primary igneous features.

Reversely zoned olivines are present throughout the OZ but they are particularly common in the UOZ. In the LOZ, olivine chadacrysts enclosed by clinopyroxene preserve sublateral reverse zoning, indicating reaction with primitive melts containing c. 10 wt % MgO. According to the PELE modeling such melts would be saturated only in olivine + chromite. One of the most extreme examples of a reversely zoned olivine in the UOZ has an evolved core (Fo87) that formed from a low-MgO (c. 6 wt %) melt that would have been in equilibrium with an olivine + clinopyroxene + plagioclase assemblage. The core of this olivine also contains an evolved (low-Cr#, high-Fe3#, high Fe2#) Cr-spinel that must have formed from a melt that had a low Mg# and high concentrations of ferric iron. In contrast, the Fo80 rim of this reversely zoned olivine indicates formation from a more primitive melt with c. 8 wt % MgO, and it contains Cr-spinel inclusions with compositions (high Cr#, low Fe3#, low Fe2#) that also imply an outward shift to more primitive melt compositions. Reverse zoning (rimward Mg# and An increases) is also seen in both clinopyroxene and plagioclase (Figs 11 and 15) throughout the LPS. We hypothesize that these reverse mineral-chemical zoning trends in clinopyroxene and plagioclase may possibly be related to the same mechanism that produced the reversely zoned olivines in the OZ.

The different textural types of olivine and their distinct compositions, when considered together with the presence of reverse zoning (in both olivine + Cr-spinel inclusions) in the OZ, strongly suggest that the OZ rocks are hybrids, representing a mixture of multiple generations of crystallization products and liquids. Below, we attempt to constrain the different components involved in the differentiation of the LPS using the mineral and melt chemistry in combination with the PELE modeling results.

The origin of the OZ–DZ duality

The source of the high-MgO melts—an olivine slurry?

The presence of reversely zoned olivines in the OZ excludes an origin by simple fractional or equilibrium crystallization of a single pulse of melt. We propose that this reverse zonation is the result of an intra-sill melt replenishment event and that the primitive rims record the arrival of a replenishing magma into the LPS. The range of primitive melt compositions calculated from the reversely zoned rims of olivine in the OZ is between c. 10 and 8 wt % MgO. These melts resemble the LPS chilled margin composition and would have had about the same composition as the melts from which most of the primitive olivine primocrysts in the LOZ formed. Some of the olivine primocrysts are too primitive to have crystallized from a melt similar to the LPS chilled margin, however, and may be exotic to the LPS. The highest model melt MgO contents (up to c. 13.5 wt %) recorded in the OZ may represent the replenishing melt, but it is also possible that the high-Fo primocrysts that yielded these high melt MgO contents are unre-equilibrated antecrysts derived from a less-evolved magma that was actively fractionating olivine somewhere upstream as it flowed towards the LPS.

The melt–crystal systematics leads us to infer that a magma containing olivine primocrysts was emplaced into the LPS as a replenishing olivine slurry. In this scenario, the primitive carrier melt is responsible for the reverse zoning of OZ olivines, with the low-Fo cores of these grains being relicts of the crystal mush that was present in the sill before the slurry arrived. The late arrival of a crystal–liquid slurry carrying a cargo of primitive olivine primocrysts would account for the excess modal olivine in the LPS as a whole compared with the PELE fractional crystallization model. On the other hand, the diverse zoning patterns observed in the OZ, particularly the reverse zoning, appear inconsistent with models invoking simple redistribution of entrained olivine carried in by an initial magma. In this context, the low-Fo hopper olivines in the MOZ and UOZ might also reflect contamination of the primitive slurry by host DZ rocks. Radiogenic and δ34S isotope data (Hayes et al., in preparation) show that the OZ is not in isotopic equilibrium and represent separate intrusive events, strongly supporting a multiple intrusion model.

We propose that a slurry of olivine primocrysts was emplaced immediately above the dendritic LBZ (at c. 1 m above the lower contact) and beneath the DZ, to form the LPS OZ. We speculate that the boundary between the (solidified) dendritic LBZ and the relatively buoyant feldspar-rich resident mush above was an important mechanical discontinuity, facilitating emplacement of a slightly denser olivine slurry. The primitive melt in the replenishing magma would have partly mixed with the resident, evolved, gabbroic mush. Some of the evolved olivines in the OZ may be true relicts, whereas others may record more extensive, earlier hybridization steps. Because the primitive olivine primocrysts with normal zoning patterns typify the LOZ, we suggest that this sub-zone is the least-hybridized facies of the OZ and mostly contains olivines carried in by the replenishing slurry. Reversely zoned mantles to some olivine phenocrysts from the LBZ (Fig. 9), which we infer are derived from the first magma pulse emplaced...
into the LPS, may also have developed as a result of olivine slurry emplacement, through heating and impregnation of a porous floor. However, reversely zoned mantles are also observed in LCM olivine phenocrysts, so we cannot rule out the possibility that the reversed zoning signatures were created further upstream in the plumbing system prior to emplacement in the LPS. It is also possible that the melt directly in contact with host dolostone may have increased its Mg/Fe by assimilating some of the wallrock dolostone (see Gaeta et al., 2009; Di Rocco et al., 2012; Hayes et al., submitted).

Reversely zoned olivines and chromite microphenocrysts are observed in many of the olivine-cumulate sills in the Minto Inlier, suggesting that late olivine slurry emplacement may be common in Type 1 Franklin sills (Hayes et al., in preparation).

The DZ—a relict of the resident mush?

Thin hypabyssal sills and dikes (such as the LPS) are typically considered to have been emplaced near instantaneously and to have cooled rapidly. As a result they should show limited internal differentiation. They commonly have doleritic textures, which would have hindered crystal–liquid separation, so giving them only a weak cumulate component (Gunn, 1966; Philpotts et al., 1998). The compositions of model melts in equilibrium with DZ clinopyroxene cores are typical of cotectic melt compositions (c. 7.5 wt % MgO). The compositions, textures, scarcity of cumulus minerals (<10%) and abundance of Fe–Ti oxides of the LPS DZ are therefore typical of relatively quickly crystallized hypabyssal sills.

If the evolved cores of reversely zoned olivines in the UOZ are relics of the earlier gabbroic material filling the LPS, as argued above, then we can use their composition to constrain the DZ composition prior to replenishment. The most evolved olivine core (c. Fo75) in the UOZ that has a reversely zoned rim is in equilibrium with a model melt composition of c. 6 wt % MgO. The PELE and Fe–Mg modeling indicate that plagioclase and clinopyroxene joined olivine on the DZ liquidus at c. 7.5 wt % MgO. This implies that the low-Fo (Fo75) cores in the UOZ that should have been part of a three-phase cotectic gabbroic assemblage. If the DZ crystallized from a melt similar to the LPS LCM, then the PELE modeling suggests that at the moment when the replenishing event occurred, the DZ would have comprised 11% olivine, 16% plagioclase, and 13% clinopyroxene with 60% pore melt. Henceforth we refer to this assemblage as the resident mush.

One disparity between this model and the observations is that no olivine has been observed in the DZ, whereas PELE predicts c. 10% to have been present in the pre-replenishment LPS. There is commonly a thin olivine–gabbro subzone at the base of the DZ in other Franklin sills of this type (e.g. Bédard et al., 2012), so we infer that the sparse (c. 5–10%) orthopyroxene from the basal LPS DZ may be pseudomorphous after olivine. This implies that the interstitial melt in the LPS DZ became sufficiently evolved to have reached the olivine–orthopyroxene peritectic, which is consistent with the highly evolved compositions of crystal rims in the DZ (e.g. clinopyroxene Mg# 20; Fig. 11).

The primitive plagioclase (An75) and clinopyroxene (Mg# 85) crystals in the DZ that are now embedded in a mass of later in situ crystallization products may be relics of early DZ crystallization, antecrystic phases that were carried into the LPS by the first pulse of magma, or grains entrained from a syn-replenishment hybridization zone into the DZ above. Reversely zoned mantles in cumulus-textured clinopyroxene in the CPZ and DZ, and localized reverse zoning in cumulus plagioclase in the DZ, may record the arrival of the OZ replenishing magma, but this remains conjectural.

Following olivine slurry emplacement, we suggest that the DZ continued to evolve, reaching its most extreme mineral compositions (Figs 11 and 15) and highest proportions of accessory phases (Fe–Ti oxides, amphibole, biotite) at a high-level sandwich horizon, a pattern suggestive of progressive in situ fractional crystallization (see Boudreau & Philpotts, 2002). Although we cannot unequivocally determine whether the DZ sandwich horizon developed prior to or after olivine slurry emplacement, we will argue below that it post-dates it.

Mixing and hybridization between the resident DZ and replenishing magma

We propose that a partly crystalline DZ (c. 40% solid) was resident in the LPS when a replenishing olivine-charged magma was emplaced. The replenishing magma must have contained a high melt fraction (c. 50%) as otherwise flow would have been impeded (Paterson, 2009). We further propose that mixing between the slurry and resident magma led to the formation of a hybrid melt with a composition of c. 10–8 wt % MgO, and we infer that the most primitive reversely zoned olivine rims and overgrowing clinopyroxene oikocrysts crystallized from this hybrid melt.

Clinopyroxene oikocryst cores have similar Mg# throughout the OZ, indicating that they all formed from melts with c. 9–8 wt % MgO. Conversely, the rims of reversely zoned olivines become less forsteritic up-section, from Fo83 in the LOZ to Fo89 in the UOZ, where olivine rims are calculated to have been in equilibrium with c. 8 wt % MgO melts. The lack of differentiation in OZ clinopyroxene core compositions suggests that clinopyroxene co-saturated when the local pore melt reached the same point in its compositional evolution. The model melts calculated to be in equilibrium with OZ clinopyroxene exceed their cotectic melt MgO content (according to PELE), suggesting that these clinopyroxenes did not form through fractional crystallization from the LPS LCM melt composition. Instead, OZ clinopyroxene saturation appears to have been triggered by magma hybridization at c. 9 wt % MgO.

The compositions of groundmass olivines in the MOZ are more primitive than LOZ groundmass olivines
and are in equilibrium with reversely zoned rims in the UOZ. The transition to more primitive groundmass olivine in the MOZ–UOZ suggests that these groundmass olivines crystallized from hybrid magma percolating up through this mass as the OZ cumulates became compacted. As such, the UOZ rim compositions were buffered by throughflow from hybrid pore melt expelled from below. However, compaction of the OZ may have been limited by the relatively rapid cooling rate of the LPS (<6 years). We suggest that the UOZ is a strongly hybridized facies of the OZ that retains undissolved traces of the resident mush, now represented by plagioclase and clinopyroxene inclusions in hopper olivines (Fig. 5d and f), evolved olivine cores and evolved Cr-spinel inclusions (Fig. 10). The less common MOZ hopper olivines (Fig. 5b) may also have crystallized from hybrid magma, possibly suggesting a weaker contamination signature in the MOZ.

**Evidence for rapid crystallization in the center of the LPS**

**Hopper olivine morphologies in the UOZ**

The UOZ is characterized by abundant hopper olivine morphologies (Fig. 5d–f), some of which are reversely zoned. The hopper olivines of the UOZ that enclose clinopyroxene and plagioclase (Fig. 5d and f) are also inferred to record ingestion of the DZ mush by the olivine slurry. Experiments carried out by Donaldson (1976) show that olivine crystal morphology is related to the cooling rate and melt MgO composition, and that hopper olivine morphologies form at high growth rates. Huppert & Sparks (1980) suggested that hopper and elongate olivines could be produced by rapid cooling when a hot, primitive basaltic magma quenches beneath a cooler basaltic melt. A very similar mechanism was proposed more recently for the origin of cyclic harrstic layering in the Rum intrusion (O’Driscoll et al., 2007). The hopper olivines in the UOZ of the LPS seem to indicate rapid cooling, even though the UOZ is near the middle of the LPS (at c. 8 m height) where cooling rates should be the slowest. We suggest that the hopper olivine in the LPS OZ formed when the hotter, primitive olivine slurry intruded a cooler resident gabbroic mush. In this scenario, the ‘supercooled’ hybrid melt crystallized olivine (hopper-type) around the xenocrystic cores sourced from the resident gabbroic mush.

**Sector-zoned clinopyroxene in the CPZ**

Above the OZ there is a striking modal and textural change as olivine cumulates grade rapidly into a thin layer dominated by euhedral, sector-zoned clinopyroxene (Fig. 7). The passage from olivine-dominated cumulates to a clinopyroxene-rich cumulate assemblage (CPZ) could be interpreted as being due to fractional crystallization with co-saturation in clinopyroxene briefly preceding co-saturation with plagioclase. However, as discussed above, our preferred interpretation is that the OZ formed as a replenishment into a resident gabbroic mush (DZ). This suggests that the CPZ may be related to this event, rather than representing an intermediate fractional crystallization step.

The development of sector-zoning in clinopyroxene has been attributed to rapid growth brought about by rapid cooling, or quenching by magma decompression and volatile exsolution (Nakamura, 1973; Brophy et al., 1999). However, as emphasized in the discussion about the origin of hopper olivines, the CPZ is roughly in the middle of the sill (at c. 8–9 m height), where cooling rates should be slowest. Also, clinopyroxene was not a liquidus phase upon emplacement of the initial LPS melt (assumed to have had a composition corresponding to the LCM), so decompression and volatile exsolution are not plausible causes. In the context of the OZ emplacement model we suggest that the rapid growth textures and abundant clinopyroxene modes of the CPZ record compositional super-saturation induced by pore-scale melt mixing. If the DZ was underplated by an olivine slurry, then the OZ–DZ interface may have been overprinted by the migration of pore melts from one system into the other. We speculate that the buoyant, hybrid melts from the OZ percolated upwards, possibly driven by compaction of the OZ below, and infiltrated the base of the DZ, where pore melts would have had cotectic compositions. The phase topology (Fig. 18) suggests that mixtures of two such melts could result in a clinopyroxene-only saturated melt (Onuma & Tohara, 1983; Bédard, 1993). Clinopyroxenes that form from such compositionally super-saturated melts may have crystallized rapidly, much as we suggested for the hopper olivine of the UOZ. This interpretation for the origin
of the CPZ is consistent with the distribution of sector-zoned clinopyroxene into 'channel-like' trains (Fig. 8). It is also supported by the compositions of the sector-zoned clinopyroxene, which have core compositions (Mg# 85–83) that overlap with core compositions of clinopyroxene oikocrysts in the OZ below (Fig. 11). Because the latter are interpreted to have crystallized from the hybrid melt of the OZ, this compositional similarity is consistent with derivation of the sector-zoned clinopyroxene from pore melt expelled from the OZ.

Model—the development of igneous layering in the LPS

Based on the textural and mineral-chemical evidence presented above, we propose that the LPS records three main magmatic stages (Fig. 19): (1) initial magma (a dilette crystal slurry) emplacement followed by fractional crystallization; (2) replenishment and partial hybridization with an olivine slurry; (3) late-stage pore melt migration, and limited re-equilibration with cooling, trapped pore melt.

Stage 1—initial magma emplacement and fractional crystallization

The presence of c. 5% modal olivine phenocrysts (c. 1–2 mm) in the LPS chills suggests that the initial magma pulse probably carried small amounts of olivine, although we cannot exclude the possibility that higher proportions of olivine were concentrated in the center of the flow (Bhattacharji, 1967; Simkin, 1967). It is also possible that the initial magma carried a cargo of clinopyroxene and plagioclase, as evidenced by the presence of rare cumulus crystals in the DZ. Model melts in equilibrium with the most primitive olivine phenocrysts in the LCM have higher MgO contents (at c. 13 wt % MgO) than the LPS chill (10-3 wt % MgO). As such, some of the olivine carried in with the first pulse may be antecrystic relics (or products of an earlier crystallization step) carried downstream as the magma fractionated. After emplacement, this basaltic magma crystallized from both margins, producing the LBZ and UBZ, where high cooling rates formed plagioclase and clinopyroxene with dendritic habits that overgrew and cemented olivine phenocrysts. The olivine phenocrysts of the LBZ have core compositions that are in equilibrium with the LCM melt composition, suggesting that they are derived from this magma. The lack of any olivine crystals in the UBZ suggests that any phenocrysts that were present settled fast enough to evade entrapment by the downward solidification front.

The intergrowth of olivine with clinopyroxene and plagioclase dendrites in the LBZ suggests that the initial magma rapidly reached the three-phase cotectic after emplacement. If the LCM accurately reflects the melt composition, then only c. 8% olivine extraction is needed to reach three-phase saturation according to the PELE model. However, if the olivine phenocrysts in the chill are 'cumulate' phases, then the LCM composition is slightly biased towards higher MgO contents and <8% crystallization may be needed to reach three-phase saturation. If we deduct the average olivine core composition of LBZ olivines from the LCM (considering a 10% olivine mode for the LBZ), then the residual melt would contain c. 6 wt % MgO, similar to our estimate of the resident mush composition using the Fe = Mg modeling. The olivine that crystallized from this initial magma pulse into the LPS may have settled to form a first-stage olivine cumulate, but this cannot be determined with certainty. If such a basal olivine-rich layer developed at this stage then there may have been a weak S-shaped profile with the most evolved material concentrated in the upper third of the sill, as seen in other intrusions (Shirley, 1985; Meurer & Boudreau, 1998) and in thick lava flows (Bédard, 1987; Boudreau & Philpotts, 2002). The sandwich horizon defined by the clinopyroxene and plagioclase data from the LPS (Figs 11 and 15) may have developed early, or might have developed after emplacement of the olivine slurry (see below).

Stage 2—the emplacement of an olivine slurry and hybridization with the basal DZ

The mineral zonation patterns and compositions from the LPS OZ suggest that a crystal-charged olivine slurry was injected near the base of the LPS, beneath most of the buoyant DZ and above the more consolidated, dendritic-textured LBZ. This dynamic slurry could have scour ed out and mixed with weakly consolidated (olivine-enriched?) host material resting on the sill floor, with only the better-consolidated, largely solidified LBZ adjacent to the cooling surface surviving. The relative coarseness of the LBZ dendrites (Fig. 3b) suggests relatively slow cooling rates, possibly owing to the emplacement of the hot olivine slurry above. If this is correct, it would imply that the olivine slurry arrived not long after the initial magma was emplaced into the LPS.

The olivine slurry would have been denser than a partly solidified, feldspar-rich dolerite, and so it would probably have underplated the host DZ, much as was proposed previously for the replenishment of magma chambers (Huppert & Sparks, 1980; Tegner et al., 1993). On the basis of major element geochemical systematics, Hayes et al. (in preparation) calculated that the olivine slurry contained c. 20–25% olivine, making it slightly denser than a resident mush that was c. 40% solidified. Also, the 40% solidified resident mush may have been mechanically strong owing to formation of feldspar chains (Philpotts et al., 1998), and so the weakest part of the LPS may have been the interface between the dendritic LBZ and the overlying partly crystalline resident mush. We infer that the emplacement of the olivine slurry at the base of the LPS created the S-type basal MgO bulge (Fig. 2b). During emplacement, we suggest that the incoming olivine slurry mixed to some extent with the resident mush (corresponding to the missing basal DZ) and created a hybrid melt with c. 10–8 wt % MgO. The evolved olivine cores (and evolved Cr-spinel inclusions) and possibly the gabbroic inclusions in hoppier olivines
are interpreted to be partially digested relicts of this basal DZ. The high-An cumulus plagioclase and high-Mg# cumulus clinopyroxene in the DZ may be remnants of the primitive base of the DZ, much of which may have mixed with and dissolved into the feldspar- and pyroxene-undersaturated olivine slurry. However, the DZ may have a dacitic complex history prior to OZ emplacement, so we cannot exclude the possibility that the uncommon primitive plagioclase and clinopyroxene from the DZ are unrelated to OZ formation. Olivines in the MOZ and especially the UOZ commonly have hopper-type morphologies (Fig. 5b–f). These may record the chilling effect of extensive hybridization, coupled with heat loss to the cooler DZ host, similar to models for the harristic layering of the Rum intrusion (Huppert & Sparks, 1980; O’Driscoll et al., 2007).

Stage 3—late-stage processes
After emplacement of the olivine slurry and hybridization with the basal DZ host the hybrid melt in the resulting OZ had a composition of c. 10–8 wt % MgO. This range may represent heterogeneities in the amount of admixed DZ material. Alternatively, the variability of the MgO content may reflect post-hybridization fractional crystallization of the hybrid melt as the OZ solidified. During this consolidation and cooling phase, interstitial and poikilitic clinopyroxene and plagioclase, began to form. The cores of the most primitive clinopyroxenes that envelop the reverse-zoned olivines are in equilibrium with a maximum melt MgO content of c. 9 wt %, recording clinopyroxene co-saturation from the hybrid melt. We do not yet have a clear explanation for why clinopyroxene appears earlier on the OZ crystallization path in comparison with the PELE model results.

It is plausible to infer that pore melt from the OZ was expelled by compaction to mix with the overlying DZ, and we suggest that this is the origin of the sector-zoned clinopyroxene in the CPZ. At this juncture, the melt being expelled from the OZ would have had
compositions roughly corresponding to the compositions of the reversely zoned olivine rims and clinopyroxene in the UOZ (c. 8 wt % MgO). When it penetrated the overlying DZ, this OZ-derived pore melt would have mixed with more evolved pore melts, possibly triggering rapid growth of sector-zoned clinopyroxene (Fig. 7).

We suspect that mixing occurred in dissolution channels in what is now the CPZ and that clinopyroxene
formed by this reaction overgrew and replaced the olivine-bearing gabbroic matrix of the CPZ (Fig. 8).

When permeability dropped to the point at which interstitial melts could no longer move, the trapped pore melts began to evolve via in situ fractional crystallization (Humphreys, 2009; Holness et al., 2011). Permeability drops may be linked with inwardly migrating solidification fronts (Marsh, 1996). In OZ rocks this generated narrow Fe-rich rims on olivine primocrysts, evolved groundmass olivines, Fe–Ti enriched rims on clinopyroxene, sodic rims on plagioclase, and a cortège of minor phases (Fe–Ti oxides, mica, amphibole, and sulphides). The effects of in situ differentiation are most prominent just beneath the UBZ in the DZ, where clinopyroxene (Fig. 11) and plagioclase (Fig. 15) cores reach their most evolved compositions at a sandwich horizon. We tentatively suggest that this sandwich horizon developed late in the crystallization history of the LPS and post-dates the emplacement of the olivine slurry. This is based on the constraints provided by our data, which suggest that the LPS was only c. 40% solidified at the time of olivine slurry emplacement, rather than the >70% solidification required to produce the strongly evolved or fractionated mineral compositions seen in the sandwich horizon.

Implications for the differentiation and formation of cyclic–macrorhythmic layering in large layered intrusions

The LPS preserves first-order igneous layering that is similar to cyclic–macrorhythmic cumulate layering observed in large layered intrusions (Eales & Cawthorn, 1996; Wilson, 2012). The rapid cooling rate of the thin LPS preserves fine details of mineral textures and zoning that provide insights into the differentiation and layer-forming mechanisms. Similar mechanisms may also have operated in larger, more slowly cooled, magma chambers. We have presented evidence that suggests that the LPS OZ was emplaced as an olivine slurry near the base of a pre-existing gabbroic mush. The prominent layering seen in the LPS is in marked contrast to most dolerite-textured gabbroic sills. For example, the 150 m thick Beacon sill of the Ferrar suite in
Table 4. Representative plagioclase compositions

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Antarctica (Zieg & Marsh, 2012) shows only limited internal differentiation, and does not develop significant modal layering, even though it is seven times thicker than the c. 21 m LPS. Our data imply that the first-order cumulate layering structure (OZ–DZ duality) of the LPS was not formed by in situ closed-system differentiation of a single magma pulse, and that the LPS is a composite intrusion that records multiple intrusive pulses, a conclusion that is supported by our unpublished isotopic data (Beard et al., in preparation; Hayes et al., in preparation). Similar textural and phase relationships have been observed in the Lower, Critical and Main Zones of the Bushveld Complex, which have been linked to replenishment (Eales et al., 1991; Mitchell et al., 1998). Other intrusions and volcanic products also bear imprints of variable crystal cagnos, as shown by in situ isotopic modeling (Davidson et al., 2007; Font et al., 2008; Martin et al., 2010).

Our results bear on a debate that has created very polarized views in the literature about how magma chambers form and differentiate. The DZ of the LPS (and most other sills of the Franklin suite) shows a D-shaped profile, with the development of an Fe-Ti-enriched zone (sandwich horizon) as a result of inward crystallization, much as argued by Shirley (1985) and others (Latypov, 2009). However, the OZ–DZ duality and S-shaped profile of the LPS as a whole appear to have formed through multiple injections, as advocated by Marsh (2004, 2013). We emphasize that the demonstration that some types of layering (in this instance, a basal olivine-rich layer) formed by emplacement of a crystal-charged slurry does not imply that magma cannot differentiate by in situ fractional crystallization.

The late emplacement of an olivine slurry in the LPS has other implications for magma chamber evolution. It seems axiomatic that such an event could occur only if the incoming slurry was driven by some type of magmatic overpressure. As the olivine-charged slurry is injected into a previously existing sill, space needs to be created to accommodate it. Either the sill inflates by floor subsidence and/or roof uplift, or an equivalent volume of pre-existing resident mush needs to be expelled. This remobilized gabbroic mush could be forced onwards to more distal zones as the overpressure allows the sill to expand laterally, it could be expelled into parasitic dykes, or it could be injected up-section to erupt at the surface if favourable structures are present to facilitate this. A multiple emplacement process similar to what we have documented in the LPS could also have operated in other basaltic provinces.
and possibly other settings, explaining common glomerocrystic and disequilibrium crystal–melt assemblages (Larrea et al., 2012; Passmore et al., 2012; Leuthold et al., 2014).

We also suggest the formation of a clinopyroxene-enriched layer at the interface between the OZ and DZ systems as a result of post-emplacement migration of interstitial melt. The possible development of the CPZ by mixing between OZ and DZ pore melts is similar to the mechanism of formation proposed for gabbros in the Rum intrusion (Bédard et al., 1988) and pyroxenites in the Bay of Islands complex (Bédard, 1991), and is similar to that proposed for the formation of high-Mg# clinopyroxene in actively spreading mid-ocean ridge crust (Lissenberg & Dick, 2008), emphasizing the need to better understand the effects of igneous metasomatic processes during the solidification of igneous bodies.

CONCLUSIONS

The Lower Pyramid Sill (LPS) forms part of the sill-dominated Franklin magmatic plumbing system that is well exposed in the Minto Inlier of Victoria Island, Arctic Canada. The thin (c. 21 m) LPS is remarkably well layered, with a 7 m thick layer of olivine-cumulate melagabbro–feldspathic peridotite (OZ) that is capped by a thin (c. 1 m) layer of sector-zoned clinopyroxene-rich cumulate gabbro (CPZ) and a 10 m thick layer of sub-ophitic doleritic gabbro (DZ). The LPS OZ is subdivided based on textures into a pyroxene-poikilitic lower OZ (LOZ), a middle OZ (MOZ) and an upper OZ (UOZ), with MOZ and UOZ rocks characterized by abundant hopper olivine morphologies. The absence of systematic cryptic mineral compositional variations in the LPS OZ, the high proportions of modal olivine, the presence of reversely zoned olivines, and the variety of olivine morphologies within the OZ appear to preclude formation by in situ fractional crystallization of a single pulse of magma. The LPS OZ is best explained as the result of an intra-sill mixing event between an invading olivine slurry and a resident 40% solidified gabbric mush. The dense olivine slurry was emplaced just above the largely solidified LBZ, and beneath the DZ. During its emplacement, the primitive, MgO-rich melts (c. 13–10 wt %) mixed with the more evolved resident mush (c. 6 wt % MgO pore melt) that was already present in the LPS. Mixing produced a hybrid magma, with the highest proportions of DZ contaminant in the UOZ. As the OZ solidified, interstitial melt percolated up through the compacting olivine cumulate, forming hopper olivine (and local reversed zoning) and clinopyroxene. The CPZ that separates the OZ and DZ is dominated by enigmatic sector-zoned clinopyroxene that may have formed when hybrid melt expelled from the OZ reacted with more Fe-rich pore melt in the DZ. The DZ shows inward fractional crystallization trends culminating in a sandwich horizon just beneath the UBZ. Our differentiation model for the LPS implies that crystal-slurry replenishment and in situ differentiation were both involved in the development of the LPS OZ–DZ duality. Such a style of magma emplacement and differentiation may also be important in the construction of cyclic–macrorhythmic layering of larger layered intrusions.

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SUPPLEMENTARY DATA

Supplementary data for this paper are available at Journal of Petrology online.

REFERENCES


