Emplacement of the Nain anorthosite: diapiric vs. conduit ascent

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Abstract

Estimation of settling velocities of large orthopyroxene megacrysts, found within the anorthosite intrusions, are calculated and compared to ascent rates achieved by diapirism and by conduit propagation. Calculations suggest that diapirism is far too slow to be an appropriate ascent mechanism for anorthositic crystal mush and favour conduit emplacement. The intrusions of the Nain Plutonic Suite (NPS) are located along the Abloviak shear zone (which marks the boundary between the Nain and Churchill provinces) and within the zone of juxtaposition of the Saglek and Hopedale blocks of the Nain province. These crustal weaknesses have probably controlled the emplacement and distribution of the intrusions. Contact relations between intrusions of anorthosite and their gneissic host rock provide evidence for two emplacement styles within the NPS, the first typified by strongly deformed and recrystallised rocks, the second by an outer border zone of mafic rocks. It is proposed that these differences in intrusive style are due to differences in ductility contrast between the magma and its surrounding host rocks, such that those intrusions emplaced into the thermally softened shear zone have deformed margins, whereas those intruded into the cooler Archaean crust have undeformed margins.

Introduction

The Nain Plutonic Suite (NPS) is a composite body consisting of numerous intrusions of four major types: norite-anorthosite, troctolite, ferrodiorite, and quartz-monzonite (Williams et al. 1985; Ryan 1990a) emplaced between 1.35 and 1.29 Ga (Ryan and Emslie 1994). The intrusions straddle the boundary between the Nain province, an Archaean craton, and the Churchill province, a complex terrane of largely granulite-facies gneisses that includes reworked Archaean crust, deformed Paleoproterozoic charnockitic intrusions, and a major belt of Paleoproterozoic paragneiss (Fig. 1). This boundary is tectonic, and demarcates a continental suture zone. The deformation and metamorphic imprint of the suture zone defines the Torngat orogen (Fig.1; e.g., Mengel and Rivers 1989). The Torngat orogen is the result of oblique collision between the Nain and the Rae (Churchill) provinces at ca. 1845-1822 Ma (VanKranendonk and Wardle 1997), which resulted in the formation of the Abloviak shear zone. Post-collisional isostastic uplift in the southern Rae province resulted in reactivation of the orogen at ca. 1798-1770 Ma (Van Kranendonk and Wardle 1997); this continued episodically until 1710 Ma (Scott 1998).

In stark contrast to the anorthosites of the Grenville province to the south, the NPS rocks are largely undeformed, retaining their original primary igneous mineralogy, structures, and textures (Emslie et al. 1972; Taylor 1979; Emslie and Hunt 1990; and the Nain Anorthosite Project NAP 1971-1981). Deformation features observed in some of the intrusions, particularly in some of the marginal zones, have been attributed to the emplacement of a crystal-laden mush (Wiebe 1990).

The mode of emplacement of massif-type anorthosite plutons has been a topic of study for many decades (Simmons 1964; Berg 1977; Morse 1981). Some earlier workers assumed that anorthosite intrusions were linked to metamorphic processes in high-grade rocks (upper amphibolite to granulite facies) where they are commonly found (Martignole and Schrijver 1970). However, the argument for a direct genetic link between anorthosite intrusions and high-grade rocks is no longer supported. Valley and O'Neil (1982) measured the variation in oxygen isotope composition between the margins of the Adirondack anorthosite and wollastonite-bearing skarns. They found that these margins had spectacular gradients that could only be attributed to the exchange of oxygen isotopes with circulating hot meteoric water, implying an emplacement depth of not greater than 10 km. Thermo-barometry carried out on the Laramie anorthosite (Wyoming) places its emplacement at 10-15 km depth, from estimates of crystallisation conditions of associated monzosyenites (Grant and Frost 1990). Work undertaken on the contact aureoles surrounding the NPS similarly indicates mid-crustal to upper crustal depths (6 to 14 km and ~ 645-915^{o}C) as the sites of pluton crystallisation (Berg 1977; Speer 1982). Berg (1977) suggested that ambient temperature at the time of emplacement was probably no greater than 250-300^{o}C.

A possible mechanism for ascent mechanism is diapirism. In this model, hot magma ascends by density contrast into the crust. In order to rise, the diapir must soften its wall rocks, expending thermal energy in the process. In granitoid plutons, evidence for this mechanism of ascent comes from the field observations that plutons are roughly circular, and that they appear to have marginal deformation reminiscent of that expected in diapirism (Schmeling et al. 1988; Ramsay 1989; Cruden 1990). A note of caution to this theory was first sounded by Leake (1978) to the effect that emplacement geometry is not a reliable indication of ascent mechanism. The role of dykes in transporting magma has gained in popularity in recent years as the ascent mechanism responsible for the production of large batholiths (Clemens and Mawer 1992; Petford et al. 1993). The significance of faults and shear zones in controlling not only magma emplacement but also magma ascent has been recognised by many authors (Pitcher 1979; Castro 1987; Brun et al. 1990; Schmidt et al. 1990; D'Lemos et al. 1992; Petford and Atherton 1992).

Detailed mapping of the margins of the NPS by Ryan (1992) found that foliated margins to the anorthosite plutons were far more widespread than previously reported (Morse 1971-1983; Wiebe 1990), and he believed that these margins were genetically related to the emplacement of the NPS. The results of this work led to the re-examination of the NPS margins by the present authors. This paper, therefore, examines the marginal relationships between the anorthosite plutons and their host rocks, and discusses the constraints placed on pluton growth by the physical properties of the magma. A model for the emplacement of the Nain anorthosites is proposed which satisfies both field observations and physical constraints.

Regional Geology

Gneisses. During fieldwork carried out as part of the first authors, Ph.D. thesis, the Archaean rocks, in the achipelago east of Nain, Labrador (Fig. 2), were examined in detail. The Archaean rocks lie on the boundary between the Saglek and Hopedale blocks (Fig.1), forming a narrow belt between the anorthosite plutons. They consist of amphibolite- to granulite-facies, migmatitic, tonalitic to trondhjemitic gneisses, basic gneisses, metasedimentary gneisses, and schlieric granitoid plutons (Royse 1997; Ryan 1991, 1993).

The gneissic foliation was found to strike north-south and dip at 50-90^{o} to the east (Fig. 2), although minor variation due to earlier deformation events complicates this pattern locally, particularly in the Dog Island area (Fig. 2). The migmatised gneisses contain a transposed foliation S0-S1 that was later regionally folded by northeast-trending folds (F2). The schlieric-granitoid gneisses, which have intruded in a lit-par-lit fashion into the migmatised gneisses, are also affected by this D2 event; hornblende and biotite define a NE- trending mineral lineation, which is parallel to the F2 fold axis. The schlieric-granitoid gneisses do not display the same complex deformational history or the pervasive migmatisation seen in the migmatised gneisses. In places, relics of potassium feldspar phenocrysts, in schlieric-granitoid gneisses, lead to a porphyroclastic (augen) texture in outcrop.

It was found that this foliation trend appeared to deflect around the margins of certain anorthosite plutons (e.g., Nukasusutok, Carey, and Sandy Islands, Fig. 2). All the dykes in the area (except the late unmetamorphosed E-W trending Nain dykes (Gower et al. 1990) crosscut the migmatitic and schlieric gneisses at a high angle but are subsequently themselves crosscut by the NPS. Suggesting that the apparent concordance of the gneissic country rocks with the anorthosites is not due to reorientation during pluton emplacement but rather to the pluton contacts paralleling the regional trend of gneissosity. In detail, the gneissic layering and local trends are truncated by the intrusive contacts. Overall map patterns indicate that the trends were probably established prior to the emplacement of the NPS (Fig. 2).

Unstrained contact-metamorphic replacement of regional garnet by cordierite and hypersthene, and sillimanite by corderite and spinel, has been recorded (Berg 1977; Speer 1982). Fieldwork in the area showed that both pseudomorphous assemblages retain the shape of the primary phase without any flattening or stretching of the original morphology, implying a low degree of strain associated with the thermal overprint in the contact aureoles (Ryan 1991,1992; Royse and Ryan 1995) and suggesting that the intrusion of the NPS did not result in large scale deformation within the country rocks.

U-Pb ages from zircon and monazite within rock units on islands east of Nain (Connelly and Ryan 1996) indicate emplacement of the schlieric granitoid gneisses at ca. 2578±3 Ma (U/Pb), emplacement of some mafic dykes at 2559±10 Ma, and mylonitisation, shearing, and granulite-facies metamorphism at ca. 2578±3 to 2549±3 Ma. This 30 Ma period of magmatism and tectonism resulted in the formation of the Nain Province (Connelly and Ryan 1996) i.e., the amalgamation of the Hopedale and Saglek blocks. A similar structural history with comparable ages of intrusion and metamorphism is found in the Okak area (Van Kranendonk 1992; Schiøtte et al. 1990) in off-shore northern Labrador (Wasteneys et al. 1992) and in the region west of Hopedale (Loveridge et al. 1987), which is indicative of a regional event corroborating the amalgamation model (Wasteneys et al. 1992). This event (D2) was probably synchronous with the emplacement of the schlieric granitoid gneiss (Royse et al. 1994; Royse and Ryan 1995) (see Table 1). The region was largely quiescent until the intrusion of the Nain Plutonic Suite (NPS) between 1350 and 1290 Ma (Ryan and Emslie 1994; Morse 1971-1983; Ryan 1990), with the majority of the anorthosite plutons being emplaced over a 17 Ma period (Hamilton et al. 1993; Hamilton 1993). The location of the NPS plutons along the boundary between the Nain and Churchill provinces and also in the zone of amalgamation between the Saglek and Hopedale blocks is significant, perhaps suggesting that crustal weaknesses may have controlled the emplacement of the Nain anorthosites.

Dykes. All the country rock gneisses described above are cross-cut by an assemblage of mafic dykes of various ages, emplaced under both ductile and brittle conditions. All the dykes have been recrystallised at amphibolite or granulite facies, and are described in detail by Ryan (1995). The occurrence of two pyroxenes and hornblende within some of the dykes is indicative of the transitional stage from amphibolite to pyroxene-hornfels facies, and is attributed to the contact-metamorphic effect of the NPS.

Certain dykes dated at 1316±1.5 to 1327±1.5 Ma (Cadman et al. 1999) falls within the emplacement dates of the NPS and are considered to be genetically related.

Shear zones. Ductile shear zones developed under metamorphic conditions that vary from granulite- to greenschist-facies. Individual shear zones transect the gneisses and the dykes throughout the study area.

The geometric relationships between the dated basic dykes with ages between 1327 and 1316 Ma (Cadman et al. 1999) and associated shear zones indicate that these dykes are coeval with movements on the shear zones. Cadman et al. (1995) interpret a change in viscosity contrast between dykes and gneisses from the earlier to the later dykes as indicating intrusion during a period when the host-rock temperatures were declining, which suggests that the shear zones were reactivated at the time of NPS emplacement when the country rocks were hot, and continued to be active as the area cooled down.

The shear zones display evidence for at least two periods of movement. The first is probably related to the D2 Archaean event. The second, which generally post-dates peak metamorphic conditions of Mesoproterozoic age, which is found to have taken place around the time of emplacement of the NPS (Table 1). Certain smaller shear zones, e.g., on Moskie Island, record both stages of movement.

Margins of the Nain Anorthosite

The NPS exhibits two radically different types of contact with its country rocks, a feature emphasised by Ryan (1991, 1992). Some anorthosites, especially those found inland in the western and central parts of the NPS, exhibit strongly foliated margins parallel to the outline of the pluton-gneiss contact: e.g., the Mount Lister intrusion (Ryan 1991, 1993), the Bird Lake massif (Morse 1981), and the Pearly Gates intrusion (Ryan 1993). Other plutons show no signs of deformation, exhibit slightly oblique to sharply discordant intrusive contacts against their country-rock hosts, and have an abundance of locally derived host-rock inclusions. These undeformed plutons exhibit contacts that can further be divided into two types, namely (i) contacts in which easily identified igneous rocks directly abut gneisses, and (ii) contacts where the igneous-textured plutonic rocks are separated from country rocks by mafic granulites whose origin is not easily deduced, but which are intruded by the adjacent igneous rocks. The latter contact type is well developed along part of the outer zone of the Jonathon intrusion on Jonathon Island, Carey Island and Dog Island (Ryan 1991; Royse and Ryan 1995) and along the eastern side of the Paul Island intrusion on Paul Island (Berg et al. 1994). We will focus on the Mount Lister intrusion and the Jonathon intrusion as examples of the deformed and undeformed types respectively.

Internal characteristics of the Nain anorthosite have not been as well documented, as have been the coastal exposures. However, texturally the intrusions are heterogeneous at outcrop scale with reports of plagioclase crystals varying from mm's to tens of cm across (largest recorded at 70cm) and the plutons, overall appearance changes from massive to diffusely layered, to well layered, to foliated (Ryan 1997, 1998).

The Mount Lister intrusion. The eastern side of this large intrusion has been examined in shoreline outcrops along the west side of Webb Bay, and to the south of the bay (Fig. 1).

There, it is composed of coarse-grained to very coarse-grained, highly foliated, gneissic leuconorite and intensely deformed anorthosite at its margin, and less deformed to massive anorthositic and noritic rocks in the interior. The gneissose character of the margin is enhanced by primary layering between anorthositic and leuconoritic components (Fig. 3) and by the presence of deformed anorthositic xenoliths that are attenuated along strike. The characteristic rock of the deformed margin is one in which white recrystallised plagioclase surrounds augen-like relics of dark-grey primary igneous plagioclase, and coarse orthopyroxene is broken down to lozenge-shaped elongate granular aggregates retaining the original coarse-grained pyroxene in the core (Ryan 1992; Royse 1997). Gneissose olivine-clinopyroxene monzonite, dated at ca.1347 Ma (Connelly and Ryan 1993) forms a discontinuous sheath, up to 100 m wide, separating the anorthosite and norite from Archaean granulite-facies gneisses (Ryan 1993). The age of this monzonite is one of the oldest documented from the NPS, and if it represents a close approximation of the age of crystallisation of the adjacent anorthositic rocks, then this pluton and all similar foliated rocks of the NPS may be the earliest intrusions to have been emplaced. The deformed margin of the Mount Lister intrusion at Webb Bay is at least 100 m wide, and the fabric within it dips steeply to the east. Foliation planes are noteworthy in having a lineation that plunges shallowly to moderately (18^{o}-34^{o}) to the southeast, which indicates a normal oblique-slip to strike-slip movement sense. Shear sense indicators give a dextral sense of movement. This lineation geometry poses problems for any model ascribing foliation formation to simple vertical diapiric emplacement.

The Jonathon Intrusion. The western margin of the Jonathon Island intrusion (Berg and Briegel 1983), 35 km northeast of Nain, represents the second type of pluton-gneiss contact. This olivine-bearing leuconoritic to noritic intrusion is bordered by a discontinuous zone of layered, granular, mesocratic, and melanocratic basic rocks, dominated by olivine gabbro and olivine gabbro-norite, that are well displayed in ice-scoured outcrops along the shorelines of Jonathon, Carey, and Dog islands (Fig. 1). These basic zones are up to 500 m wide and separate the anorthositic pluton from the Archaean quartzofeldspathic gneiss. These rocks are a greenish brown to black weathering and characterised by the presence of elongate “stringers” and podiform pyroxene aggregates. Layering composition can vary greatly across the border zone from pale green pyroxenite, dark brown lherzolite to leucogabbro/anorthosite. Unlike the regional basic gneisses, the layered rocks abutting the Jonathon Island intrusion lack significant hydrous minerals and contain poikilitic olivine (Fig. 4). Brecciated fragments can be found within the marginal zone, particularly on Carey Island (Fig. 4). The layering in the marginal zone is locally chaotic and displays slump folding, and slides, and is crosscut by olivine-bearing mafic dykes (Ryan 1991; Royse and Ryan 1995; Royse 1997). Large orthopyroxene pods and pegmatitic xenoliths of leuconorite (which on the northern shore of Dog Island are up to a metre in width) crosscut the layering within the marginal series. These are interpreted as the products of bottom or wall accumulation within the magma chamber during its early evolutionary stages.

The marginal layered rocks of the Jonathon intrusion lack the intense and pervasive tectonic foliation that characterises the margin of the Mount Lister intrusion and appear to have many features in common with the outer border zone of the Kiglapait intrusion (e.g., they are mafic, granular, layered and contain mafic xenoliths). The margin of the Michikamau intrusion in western Labrador is also reported to have similar layered rocks along its margin (Ryan, 1991). The marginal rocks of both these intrusions have been subject to considerable debate (Morse 1969; Wheeler 1942, 1960; Berg 1977).

Rheological constraints

Ascent rates from megacrysts. Large orthopyroxene megacrysts, varying in size from a few centimetres up to a metre in width, are found within anorthosite massifs (Emslie 1975; Morse 1975; Berg et al. 1994). It should be noted that not all giant orthopyroxene crystals found within the NPS are megacrysts. Some are an integral part of the pegmatitic leuconoritic rocks (Ryan 1991; Royse and Ryan 1995). Geobarometry carried out by Emslie (1975) suggested that the megacrysts crystallised at ~15-20 kbar and 1120 ^{o}C, approximating to 40-60 km depth. Wiebe (1986) carried out further experimental work and suggested that these megacrysts crystallised at a maximum pressure of 11 kbar. Since Berg (1977) inferred that the pluton crystallised at mid crustal depths at around 6 km, the anorthositic crystal mush must have ascended some 34 km.

Dymek and Gromet (1984) and Owens and Dymek (1995) have suggested that similar pyroxene megacrysts found within the Labrieville massif and Saint Urban Anorthosite, Québec, are products of in-situ crystallisation. However, in the Nain complex, the occurrence of unexsolved high-Al megacrysts in alkali-basalt dykes (Weibe 1986) with identical chemical features to the plagioclase-bearing megacrysts (Emslie 1975), and because pyroxene megacrysts are the most primitive material in several massifs (including the NPS), having the most primitive initial isotopic ratios of Sr and Nd (Ashwal 1993); It is considered that the orthopyroxene megacrysts found in the NPS have characteristics that support a deep-seated origin.

These megacrysts are used here to estimate the minimum ascent rates of the anorthositic crystal mush by balancing frictional and buoyancy forces acting on the orthopyroxene megacrysts while settling at its terminal velocity (Spera 1980).

The rheological condition of an anorthositic crystal mush must be considered if we are to obtain an estimate of its ascent rate. The rheology is dependent on a complicated mix of several factors: strain rate, composition, temperature, pressure, and crystal, bubble and water content (Webb and Dingwell 1990; Weinberg and Podladchikov 1995). Knowledge of the crystal state is also critical in ascertaining its rheological behaviour (Fernandez and Gasquet 1994; Kerr and Lister 1991). Most magmas ascend through the crust at sub-liquidus temperatures (Sparks et al. 1977) and therefore will contain crystals. The rheological behaviour of magma changes significantly during progressive crystallisation (Kerr and Lister 1991; Fernandez and Gasquet 1994; Petford and Koenders 1998). As such, a model relying on Newtonian behaviour (i.e., where a fluid's viscosity is independent of the rate of shear or velocity gradient) would give spurious results (Miller and Paterson 1999).

Assuming non-Newtonian behaviour of the magma, the terminal velocity [V_n] of orthopyroxene megacrysts can be calculated using the equation below (Spera 1980):

$$V_n = 0.344\ (\Delta\rho\ g / \rho_l)^{5/7}\ (\rho_n / \mu)^{3/7}\ (R_n - (3\ k\ \sigma_0 / 4\ \Delta\rho\ g))^{8/7} \qquad [\ 1\ ]$$

where ρ_n = density of the megacryst, k = a dimensionless constant equal to about 5, R_n = the radius of the megacryst, ρ_l =the density of the magma or liquid that surrounds the megacryst, g = acceleration due to gravity, $\Delta\rho$ = the change in density, and μ = viscosity of the magma. σ_0 = the yield strength which can be approximated to the volume fraction of crystals within the surrounding liquid by the equation:

$$\sigma_o = K_1\Phi^3 \qquad [\ 2\ ]$$

where K_1 = an empirically derived constant having the dimensions of stress = 3×10^4 Pa and Φ = volume fraction of crystals. σ_o is also dependent on the size of the solid fraction (Jeffrey and Acrivos 1976; Fernandez and Gasquet 1994).

In a Bingham fluid (i.e., a fluid which exhibits a yield strength which must be exceeded before flow can start) the megacrysts of orthopyroxene will have to reach a minimum radius before they can sink through the magma. Timoshenko and Goodier (1970) and later Sparks et al. (1977) suggested that the minimum radius can be calculated by relating the yield strength of the liquid (as in equation 2) to the density ratio between the liquid and the crystal. Thus:

$$r^* = 3\ K\ \sigma_o / 4\ \Delta\rho\ g \qquad [\ 3\ ]$$

where r* = the critical radius below which the crystal will not sink, K= a dimensionless constant equal to about 5, g = acceleration due to gravity, and $\Delta\rho$ = the change in density.

Orthopyroxene megacrysts within the NPS have been observed over a metre in length (Emslie 1975; Morse 1975; Ashwal 1993), but on average they range between 10 , and 50 cm in radius. Orthopyroxene megacrysts can be assumed to have a density of approximately 3.4g/cm^3 (Deer et al. 1985). The composition of the parental magma to massif-type anorthosites has long been a topic of debate (Morse 1982; Emslie et al. 1994) It has been suggested that the presence of Al-Fe rich gabbros found in close proximity to anorthosite plutons, e.g., the NPS (Royse et al. 1999), and the Laramie, Harp Lake, Adirondacks , and Kiglapait intrusions (Olson and Morse 1990), are important in the quest to find a parental magma to massif-type anorthosites. These liquids could not produce the required amounts of plagioclase directly (Fram and Longhi 1992; Emslie et al. 1994) , and therefore they must be mechanically enriched. One method of doing this is by a polybaric crystallisation history (Longhi and Ashwal 1985; Emslie et al. 1994). An Al-Fe rich magma at high pressures, which was already fractionating olivine, orthopyroxene, and spinel at the base of the crust, could end up with a magma containing a high proportion of plagioclase due to the appearance of a peritectic point at 15kbar. If this magma is emplaced into mid-crustal regions, it would crystallise plagioclase alone on the liquidus.

Therefore, for simplification, the magma will be assumed to be predominantly plagioclase-rich, and hence to have a density of 2.5 -2.7g/cm^3, dependent on its crystallinity (Bottinga and Weill 1970; Sparks and Huppert 1984; Marsh 1995). A value of 2.6 g/cm^3 was taken as the most appropriate estimate for the anorthositic crystal mush , and 3.4 g/cm^3 for the orthopyroxene megacrysts, giving a density contrast of 0.8g/cm^3.

The viscosity of the magma is taken from the experimental results of Cranmer and Uhlmann (1981) and Hummel and Arndt (1985). Since the average An value is An_{55} (Morse 1978), and the ambient temperature was ~1120^{o}C (Emslie 1975), a viscosity of 1×10^6 Pa.s is appropriate. However, this does not take into account the crystal content (i.e., the proportion of solid material in the magma) of the magma, nor changes of viscosity with depth (Cserepes 1993). It is possible to account for the solid-fraction by calculating the solid-fraction-dependent effective viscosity (Marsh 1981; Vigneresse et al. 1996); if the solid fraction is taken as between 5-30%, then viscosity can be calculated as between 1×10^6 and 5×10^6 Pa.s using the equation:

$$\mu_e = \mu\,(1-1.67\,F)^{-2.5} \qquad [4]$$

where μ_e = the effective viscosity, μ = viscosity of the liquid, and F = the solid fraction. Table 2 lists several terminal velocities (settling rates) of orthopyroxene megacrysts at differing crystallinities. Crystallinities greater than 50% were not considered, as at that point the magma changes from a mushy magma to a body which is mostly solid (Marsh 1988). In granite magmas, when crystallinities reach 30 to 35%, enclaves stop sinking, and other gravity processes become less efficient (Lejeune and Richet 1995).

From equations 1-4, terminal velocities within the crystallinity range 0-50% were found to vary between 8.2cm/s and 5.0 cm/s. These velocities give minimum ascent rates for the magma, since, if the megacrysts are not to remain neutrally buoyant, the magma must travel faster than these terminal velocities. Whichever mechanism is invoked for the emplacement of anorthosite, it must allow for an ascent rate faster than these terminal velocities.

Diapiric ascent. If an anorthositic crystal mush ascended through the crust diapirically then the rate of ascent can be approximated by Stokes' equation for the velocity of a solid sphere moving through a Newtonian fluid. The equation can be extended to Newtonian, and inviscid spheres by accounting for the effects of the viscosity and, therefore the drag of the magmatic body itself (Batchelor 1967, eq. 4.9.30):

$$V=1/3\ (\Delta\rho\ g\ d^2/\ \mu_c)(\ \mu_c + \mu_m\ /\ \mu_c + 3/2\ \mu_m) \qquad [5]$$

where V = ascent rate, d = thickness of the drag zone (approximately equal to the radius of the body), g = acceleration due to gravity, $\Delta\rho$ = the change in density, μ_m = viscosity of the plutonic body, and μ_c = viscosity of the country rock.

Morse (1983) noted that there are many different sizes of pluton within the NPS and that the exact dimensions of many of the individual plutons are unknown (Ryan et al. 1997, 1998). However, Emslie (1980) suggested a maximum diameter of 10 km for the Harp Lake complex, and the Susie Brook slab is known to be ~10×20 km in area. It would seem reasonable, therefore, to assume that the anorthositic plutons within the NPS have a maximum diameter of 10 km. Country rock viscosity is assumed to be similar to that of typical lower crust, around 10^{22} Pa.s (Cathles 1975; Clemens 1998), and the viscosity of the magma is as equation 1. The density of the country rock is taken as $2.74 g/cm^3$ (Royse 1997), and that of the anorthositic crystal mush as $2.6 g/cm^3$.

Using the above data, the ascent rate works out at approximately 3×10^{-9} cm/s. This is extremely slow, and it would take approximately 31 Ma for the magma to rise some 30 km. Succeeding bodies ascend faster than the initial diapir because drag is reduced. For example, if the diapir were half the initial size it would ascend through the crust 25 times faster (Marsh 1982; Weinberg 1996), reducing the ascent rate to 7.5×10^{-8} cm/s. However, this is still slower by 8 orders of magnitude than the terminal velocities of megacrysts calculated using equation 4.

If thermal constraints are taken into consideration (Spera 1980; Whitehead and Helfrich 1991; Bejan 1993; Marsh 1995; Yoshinobu et al. 1998), the argument against diapiric ascent of these bodies is strengthened. As Marsh (1982) suggested, if a diapir is to retain a reasonable speed of ascent, its wall rocks must be heated to at least its solidus, which for granitic lower crust is in the region of ~650°C (Yoshinobu et al. 1998). This will effectively reduce the wall rock viscosity , and consequently the drag exerted on the body. This requires the pluton to convect vigorously in order to maintain heat at its margins (Marsh 1982). From the above considerations, it would appear that, for a rise of 30 km (estimated from orthopyroxene megacrysts within the anorthosites), likely ascent rates are such that it would take far too long a time to reach current levels in the crust. Gravitational ascent is therefore only viable if the pluton rises short distances, and has a significant density contrast.

The above calculation assumes that the surrounding country rock behaves according to Newtonian principles. However, Kirby (1983), and Wilks and Carter (1990) showed, in laboratory experiments using natural strain rates, pressures, and temperatures, that crustal rocks can behave as power-law fluids. This modification would not make a significant difference to the calculated ascent rate, but it would cause significant characteristic deformational effects in the country rocks (Rubins 1993; Paterson et al. 1996; Miller and Paterson 1999). The surrounding crust will have strain aureole widths significantly smaller than those produced if the crustal rock behaved in a Newtonian manner. The strain aureole width is controlled by several factors, e.g., viscous flow, stoping , and assimilation of wall rocks, as well as doming of the roof. Diapirs rising within Bingham or non-Newtonian fluids will have aureole widths that decrease with increasing viscosity, such that the flow concentrates around the diapir (Weinberg 1994), and the margins would thus be expected to deform at a faster rate than the core (Weinberg and Podladchikov 1995), resulting in non-Newtonian diapirs being surrounded by a strongly sheared margin.

Conduit ascent. An alternative mechanism to diapirism is for ascent to take place via fractures or conduits to mid-crustal depths. If the anorthositic crystal mush rose through the crust by fracture propagation, it is possible to approximate this mathematically by using simple pipe flow equations to calculate the flow velocity [V] (Delaney and Pollard 1981):

$$V = g\ \Delta\Delta\ w^2 / 12\ \mu_m \qquad [6]$$

where w = width of the dyke or fracture , and μ_m = viscosity of the magma. The buoyancy contrast (g ΔΔ) between the magma, and the country rock is the driving force in both equations 5 and 6. In equation 6, the buoyancy contrast is related inversely to the viscosity of the magma (μ_m), but in equation 5 it is linked to the viscosity of the country rock (μ_c). Therefore, the more viscous the magma, the slower the rate of ascent by conduit propagation. To estimate the ascent rate, it is first necessary to compute the critical dyke width. Several authors have calculated this for granitic magmas: Petford et al. (1993) suggest that for granite magmas with viscosities between 10^4 Pa.s and 10^8 Pa.s, rising 30-20 km through the crust, critical dyke widths are between 2 and 20m; Corry (1988) estimates felsic dyke widths of 1-4m; and Clemens and Mawer (1992) calculate a critical dyke width of 2.7m for a magma with a viscosity of 10^5 Pa.s through a 20 km long fracture. Noritic dyke widths in the study area vary from <0.1m to 18m (Cadman and Ryan 1994). Estimates for melt densities, and viscosities were used as above.

The minimum rates of ascent for an anorthositic crystal mush, into the crust calculated on the above basis, are for dyke widths of 2m, 0.09 cm/s, and for dyke widths of 18m, 7.56cm/s. This is comparable to the ascent rates of 0.1cm/s calculated for fracture propagation of granites by Clemens and Mawer (1992), and for the Cordillera Blanca batholith by Petford et al. (1993), and is of the same order of magnitude as the ascent rate estimated from the settling velocity of orthopyroxene megacrysts. As the ascent rate estimated here is the minimum rate possible, it is conceivable that the width of the dyke or fracture may have been larger than 20m.

Petford and Koenders (1998) have shown that for a crystal mush moving through a dyke, simple Newtonian flow will not apply. It is possible to account for the effects of suspended crystals by using the work of Mctigue and Jenkins (1992), and Nott and Brady (1994). However, it should be noted that deviation from simple Newtonian flow is unlikely to be significant until the magma has stopped ascending, and emplacement (pluton filling) begins (Petford and Koenders 1998).

Discussion

The above arguments indicate that diapirism would be too slow to be a viable mechanism of ascent for an anorthositic crystal mush. If the orthopyroxene megacrysts observed in the NPS (Emslie 1975; Morse 1975; Wiebe 1986) are to reach their present level in the crust after crystallising at depths of 40-60 km (Emslie 1975; Wiebe 1986), the magma must ascend at a faster rate than the settling velocity of the megacrysts. Moreover, if diapiric ascent had occurred, evidence for vertical displacement should be recorded both in the country rocks, and within the deformation aureole itself (Marsh 1982; Mahon et al. 1988; Bateman 1984). Such effects have not been observed anywhere in the NPS.

The dyke propagation mechanism, on the other hand, is consistent with models that require rapid emplacement of anorthosite into the crust (Morse 1982; Fran and Longhi 1992; Berg 1980). For instance, Morse (1982) suggested that a basaltic liquid hypersaturated in plagioclase, if rapidly emplaced, could produce massif-type anorthosites. The anorthosite plutons were intruded over a period of 17 Ma (Hamilton et al. 1993; Hamilton 1993). The rate at which magma was emplaced can thus be estimated to be in the order of 0.01 km^3 per year (areal extent, and depth can be approximated from Bouger anomaly maps). This rate is slower, but comparable to, those recorded for the build-up of Phanerozoic rift zones (3-5 km^3 per year) and of continental crust (0.4 km^3 per year) (White and McKenzie 1989). Supporting the hypothesis that NPS anorthosites were emplaced rapidly, dyke propagation is the most likely mechanism for the emplacement of the NPS.

There are many examples in the literature where the emplacement style, and the location of plutons are considered to be controlled by major faults, and shear zones (Castro 1987; Brun 1990; Schmidt et al. 1990; Hutton and Reavy 1992; Petford and Atherton 1992). The great tonalite sill of southeastern Alaska and British Columbia (Ingram and Hutton 1994) is considered to have exploited an active shear zone, as is the Strontian granite, which is thought to have been emplaced dextrally within a releasing bend of the Great Glen fault (Hutton 1988).

It is proposed that the fundamental control on the emplacement, migration, and ascent of an anortositic crystal mush within the Nain area is the Abloviak shear zone, which represents a plane of weakness within the Precambrian crust that would be mechanically favourable to dyke propagation, since rocks within the shear zone's medial plane would have been deformed repeatedly, and therefore be weak (Hutton 1992). It is likely that the magma intruding into an established shear zone will locally increase displacement rates in that area. Such instability, once created, would draw more magma into the zone until the pluton had crystallised, and the viscosity contrast within the zone had disappeared.

Another zone of structural weakness, which could easily be exploited by incoming magma, is the zone of amalgamation between the northern Saglek, and southern Hopedale blocks, as pointed out by Connelly and Ryan (1996).

The margins of the Mount Lister intrusion display a dextral sense of movement. Many major N-S shear zones, particularly in the Nuksasusutok Island area, were reactivated in a dextral sense at the time of NPS emplacement, which is confirmatory evidence of a genetic link between shear-zone activity and anorthosite emplacement. The plutons, thus fed by the dykes, would initially assume a sill-like form, but continued supply of magma would cause the plutons to expand, and assume their present form.

The strike-slip movement observed in both the Lister massif, and the Pearly Gates intrusion indicates that there is no change in the crustal level at which the plutons were emplaced. This is confirmed by the fact that there is no difference in the contact aureoles surrounding those plutons with foliated margins, and those without. This suggests that the change in the appearance of the plutons from east to west across the NPS is likely to be related solely to changes in the ductility of the country rock into which they are intruded.

Similar conclusions were reached by Davies (1982) to account for differences in the emplacement features observed in certain Pan African granites. Those granitic plutons that were emplaced into the Ajjaj shear zone are elongate, and concordant with strike. Conversely, those granites emplaced outside the shear zone into brittle wall rocks are rectilinear bell-jar shaped cauldrons. A related observation was made by Corriveau and

Leblanc (1995), who suggested that magma emplacement styles within the Grenville province of Québec were controlled by host rock rheology. The form, and shape of the Nain anorthosites are considered, therefore, to reflect the mechanism of ascent, the tectonic conditions during emplacement, and the physical properties of the crust into which they were emplaced, rather than any change in crustal level.

Conclusions

Calculation of settling velocities of orthopyroxene megacrysts found within many of the plutons of the NPS place a minimum rate of ascent for an anorthositic crystal mush, which is many orders of magnitude faster than that calculated for diapiric emplacement. The monomineralic nature of anorthosite, and its high solidus temperature would mean that, as with basalts, anorthosites would have to be emplaced relatively rapidly into the crust in order to reach mid-crustal levels in a geologically realistic period of time. It is deduced that only dyke propagation would provide a sufficiently rapid ascent rate.

It is suggested that the Abloviak shear zone, and the zone of juxtaposition between the Saglek, and Hopedale blocks of the Nain province acted as structurally weak pathways for the passage of anorthosite, which was dilationally emplaced along steep pre-existing shear zones, and solidified completely at mid-crustal levels. Continued emplacement of the anorthositic crystal-mush gave rise to high-strain zones in the earlier formed marginal material. The inter-relationship between cavities tectonically created by movement along the shear zone and by internal magma buoyancy forces resulted in the complex geometries observed in the plutons of the NPS. Differences in emplacement style across the NPS from west to east can be accounted for by changes in the ductility contrast between the anorthosite, and the country rocks, those intrusions which were emplaced into thermally softened rocks of the Abloviak shear zone having foliated margins and those emplaced in the east into cold Archaean crust having undeformed margins.

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Table 1: Geochronology of the Nain area gneisses; data compiled from Connelly and Ryan (1996), Hamilton (1993), Karlson et al. (1993) and Cadman et al. (1999)

U-Pb Age (Ma)	Event	Deformation and Metamorphism.
1276±23	Nain LP and HP dykes	
1350-1290 (1312-1309 Age of plutons in study area)	Emplacement of the Nain Plutonic Suite	Thermal resetting of U-Pb ages 1311±7 -1305 ±5Ma associated with pyroxene hornfels facies contact metamorphism; remobilisation of shear zones
1327±2.5-1316±1.5	Second phase of dyke emplacement	
1667±75	Emplacement of the Bridges intrusion	
2060±14 – 2020±17	Pink granite emplacement	
2578±3-2550±3		Mylonitisation (D2) of migmatitic gneisses and granulite facies metamorphic event
2559±10	First phase of dyke intrusion	
2578±3	Lit par lit injection of granitoid gneisses	Enclaves of older material reset thermally; granitoid gneiss emplaced at amphibolite facies
3600 ±21 – 2670±3	Intrusion of early migmatitic gneiss	Early foliation preserved in boundinaged mafic dykes; migmatisation followed by D1 event and granulite facies metamorphism

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Table 2: Settling rates V_n of orthopyroxene megacrysts within an anorthositic crystal mush (from equation 1), where R_n = radius of the megacryst r*= the critical radius (see equation 3), σ_o = yield strength values derived from equation 2 and Φ = the volume fraction of crystals.

R_n (cm)	r*(cm)	σ_o(dyne cm^{-2})	Φ	V_n(cm/s)
50	0	0	0	8.19
50	0.14	30	0.1	8.17
50	2.19	468.7	0.25	7.79
50	17.57	3750	0.5	4.99

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Figure Captions

Fig. 1: Map of field study area. Letters refer to specific units or islands: H= Hettash Intrusion, PM= Portmanvers Run anorthosite, T= Tigalak intrusion, NILI= Newark Island intrusion, L= Mount Lister Massif, J = Jonathan intrusion, BLM= Bird Lake Massif, A= Akpaume intrusion, B=Barth Island intrusion, PG= Pearly Gates intrusion, R=Reid Brook intrusion, TI= Tunungayaluk intrusion, C= Carey Island, K= Kikkertavak intrusion, S = Sandy Island and N= Nukasusutok Island. The inset outlines the major tectonic divisions in Labrador (see text for details).

Fig. 2: Diagram depicting foliation trends within the study area; see text for detail. Stereonets show: A contoured equal area plot of the migmatised gneiss, B the distribution of F2 axial planes (0) and fold axes (+), line represents the position of the regional fold axis, C, poles to layering in the granitoid gneiss, showing position of the regional F2 axis, and D, stereographic projection of the mineral lineation in the granitiod gneiss and minor fold axis (+) of F2 folds in the Dog island area (northern map section). Dyke orientations are also depicted on rose diagrams. The change in orientation between the northern and southern map is due to the presence of large-scale shear zones in the south.

Fig. 3: Photograph from the margin of the Lister Massif in the Webb Bay area, displaying strongly foliated margin containing podiformed and recrystallised plagioclase surrounding original primary plagioclase crystals. Also note the attenuation of orthopyroxene which accentuates the gneissic character of the rocks.

Fig. 4: Photograph of the western margin of the Jonathon Island intrusion taken on Carey Island. Margin contains layered norite to gabbro-norite rocks. Brecciated noritic

fragments are surrounded by leuconorite. In the centre of the photograph note the xenoliths of mafic gneiss, which contain podiformed hornblende.