Petrogenesis and timing of mafic magmatism, south Taimyr, Arctic Siberia: a northerly continuation of the Siberian Traps?

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Abstract

The Siberian large igneous province (LIP) forms the world's most extensive continental exposure of basalt and has several sub-provinces surrounding it, which may be genetically related. The Taimyr peninsula of north Siberia is one of these sub-provinces and is frequently assumed to be the northerly continuation of the basalts exposed at Noril'sk, the best-studied area of the Siberian LIP. However, the correlation is uncertain.

We present new major and trace element data from 35 samples of extrusive and intrusive rocks from Taimyr, with Sr and Nd isotope data from a subset of ten. The Taimyr rocks fall into two groups with low (~7 wt.%) and elevated (~9 wt.%) MgO concentrations. The high-MgO rocks display a restricted range of initial $^{87}\text{Sr}/^{86}\text{Sr}$

(0.705 to 0.706) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.5122 to 0.5124) ratios, and share bulk silicate earth normalised rare earth element patterns strikingly similar to data observed in the ore-related Noril'sk intrusions. The remaining low-MgO group samples have a broader range with higher Sr and lower Nd isotope values and higher incompatible trace element ratios (e.g. $\text{Th}/\text{Ta} > 5.3$ and $\text{La}/\text{Sm}_n > 1.7$) similar to the crustally-contaminated Nadezhdinsky and Morongovsky suite basalts of the Noril'sk region. The major and trace element data for both groups are consistent with a process of fractional crystallisation coupled with small degrees of assimilation of incompatible-element-enriched lower crust involving different contaminants. Trace element model calculations indicate a process of magma formation at large degrees of partial melting and at pressures of less than 3 GPa, probably within the garnet-spinel transition zone or the spinel stability field of the asthenospheric mantle. We obtained an argon plateau age of ~ 252 (252.7 ± 1.5) Ma and a ~ 239 Ma total fusion age from a Taimyr lava and intrusive sample, respectively, confirming that volcanism is only partly contemporaneous with the activity of the Siberian LIP. Although this is in agreement with previous interpretations, we argue that the age difference between both events is only ~ 13 myrs and probably less (~ 5 myrs) although further investigation of the relationship is required. Our data allow correlation with distinct Noril'sk members and most importantly to the ore-bearing (Ni-Cu) intrusions implying that whole rock chemistry could have value as a prospecting tool in Taimyr.

Keywords: Taimyr, Siberian Traps, Petrogenesis, Large Igneous Province

1 Introduction

The Siberian large igneous province (LIP) can be traced from the Siberian Platform (or craton) to the Ural Mountains, and from the Kuznetsk Basin in the south, to the Arctic Circle, including the Taimyr peninsula in the north (Maysatis, 1983; Fedorenko et al., 1996; Sharma, 1997; Reichow et al., 2002, 2009). The evidence for this vast areal coverage comes primarily from the occurrence of large volumes of basaltic lavas, intrusions (mainly dykes and sills), and pyroclastic rocks (e.g. Fedorenko et al., 1996). Radiometric age data, primarily $^{40}\text{Ar}/^{39}\text{Ar}$, have been used to show that most of the activity was broadly contemporaneous at about 252 Ma, and thus represented a surge in volcanism that may have triggered the end-Permian mass extinction (e.g. Renne and Basu, 1991; Renne et al., 1995; Reichow et al., 2002, 2009; Burgess et al., 2014).

There are, however, few published compositional data for the magmatic products found in most sub-provinces of the Siberian LIP, including the Taimyr peninsula, Urals, and the Kuznetsk Basin. Such data are important to understand the petrogenesis of the Siberian LIP as a whole and for the identification of magmatic processes that control compositional diversity in each of the sub-provinces. A key question is whether the igneous bodies found in each of these sub-provinces can be correlated with the intensively studied igneous stratigraphy in the main part of the province at Noril'sk (e.g., Lightfoot et al., 1990, 1993; Fedorenko et al., 1996). Was the activity in each of these sub-provinces genetically linked, or is each sub-province characterised by distinctive processes (melting, contamination, and fractionation)? We present new $^{40}\text{Ar}/^{39}\text{Ar}$ data, whole-rock and mineralogical compositional data for suites of basaltic rocks and dolerite sills and dykes from the south Taimyr peninsula (Fig. 1). Using age constraints to determine if the lavas are contemporaneous with

the Siberian Traps volcanism, we can address whether these rocks also share any geochemical similarities with basalts of the Noril'sk region, and if correlations can be made between the intrusive rocks at Taimyr and at Noril'sk. The latter correlation is of particular interest, because the world's greatest concentrations of platinum group elements and massive Ni-Cu sulphide ores are located at Noril'sk, and similar deposits occur in Taimyr (e.g., Yakubchuk and Nikishin, 2004)

2 Geological setting

The Taimyr peninsula is located between the Kara and Laptev Seas in Arctic Russia (Fig. 1a). The peninsula is bounded to the south by the NE-SW trending ~300 km wide Yenesei-Khatanga trough, which separates it from the Siberian Platform. The Yenesei-Khatanga trough is filled with Triassic to Early Cretaceous sediments and, based on the interpretation of seismic data, the deeper part of the basin contains volcanic rocks (Vernikovsky, 1996). The Taimyr peninsula is traditionally divided into three ENE–WSW trending structural-stratigraphical domains (Zonenshain et al., 1990; Vernikovsky, 1996): North, Central and South Taimyr (Fig. 1a). North Taimyr is characterized by Precambrian basement with strongly deformed Precambrian to Early Palaeozoic flysch deposits that are intruded by Late Palaeozoic Uralian collisional granites. North Taimyr has been interpreted to be an independent crustal block (i.e., the North Kara Terrain or North Kara Massif; Metelkin et al., 2005), and detrital zircon ages indicate that it was probably part of Baltica in Early Palaeozoic time (Pease and Scott, 2009). The Central Taimyr zone, separated from North Taimyr by the Main Taimyr Thrust, contains a weakly to unmetamorphosed Late Neoproterozoic (Vendian) to Early Palaeozoic continental margin succession which unconformably overlies various Precambrian crystalline units. The Neoproterozoic to

Early Palaeozoic succession is interpreted to be the continental slope of Siberia (Inger et al., 1999; Pease and Scott, 2009) and indicates that central/south Taimyr has been a coherent part of Siberia since at least late Neoproterozoic time.

The South Taimyr domain comprises mainly folded Ordovician to Carboniferous carbonate-dominated sediments that were deposited on the passive margin of Siberia, overlain by Late Carboniferous to Triassic shallow-marine and deltaic sedimentary rocks (Fig. 1b; Inger et al., 1999). Mafic lavas, sills and dykes are present in South Taimyr and are deformed together with their sedimentary host rocks (Inger et al., 1999). Fedorenko et al. (1996) estimated that the cumulative thickness of mafic lavas and sills in south Taimyr reaches at least 2 km. The igneous rocks are usually attributed to the late Permian to early Triassic magmatic activity of the Siberian large igneous province (e.g., Gurevitch et al., 1995; Vernikovsky et al., 2003). Initial $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations of these lavas and sills indicated Triassic and Early Jurassic ages (Walderhaug et al., 2005), respectively. Dating of the same units by Reichow et al. (2009) demonstrated that eruption of the Taimyr basalts was contemporaneous with the ~252 Ma activity of the Siberian LIP.

Most tectonic models describe the Taimyr fold belt as a continuation of the Late Palaeozoic Ural orogeny (Zonenshain and Natapov 1989; Otto and Bailey 1995, Vernikovsky, 1996), which is indicated by Carboniferous to Permian collisional granites in north and central Taimyr (Vernikovsky, 1996; Pease et al., 2015). This tectonomagmatic event records the suturing of north Taimyr to central/south Taimyr. A significant transpressional deformation episode affected Taimyr in latest Triassic to earliest Jurassic times (Inger et al., 1999) and it was during this event that folding of the basalt lavas and intrusives occurred.

3 Sampling and petrography

The samples available for this study were obtained from exposures on the eastern shore of Lake Taimyr (Fig. 1b; Table S1). Limited exposure and the effects of frost shattering precluded the production of detailed stratigraphic logs such as those available for the Norilsk area (e.g., Fedorenko et al., 1996). In the sampling area the basaltic volcanic rocks overlie a Late Paleozoic deltaic succession and are preserved in the cores of two synclines, of which the northernmost offers the better exposure, particularly on the Hoffman Peninsula. The base of the igneous succession rests on Late Permian sediments and is characterised by volcanic breccias. Overlying these breccias are basaltic volcanic rocks that include rubbly debris flow deposits, and vesicular and massive lavas. The massive basaltic lavas are typically 3 to 5 m thick with well-preserved columnar jointing. Pillows and pahoehoe flow tops testify to alternation between subaqueous or subaerial conditions, compatible with the deltaic environment into which they were erupted. Mesozoic sediments are exposed further to the north but field relationships between these and the basaltic volcanic rocks are unknown and therefore no relative age relationships could be derived from bracketing sediments. Dolerite sills and infrequent dykes are also exposed in the area cutting through the Late Palaeozoic succession up to Late Permian sediments. The sills are 3 to 5 m thick and provided the bulk of samples analysed in this study. Dolerite sill and dyke samples have similar mineralogy and compositions, and are therefore referred to simply as intrusive rocks here.

Most of the volcanic rocks comprise sparsely porphyritic basalts, with an aphanitic to fine-grained, intersertal to intergranular groundmass (Fig. S1). Plagioclase is a ubiquitous phenocryst phase; most phenocrysts are subhedral but a few heavily

resorbed xenocrysts occur. In some samples, plagioclase occurs as monomineralic intergrowths of equant crystals (T98-48) or is intergrown with augite (T98-49b). These glomerocrysts are interpreted as cumulate fragments. Chemical compositions of feldspars from intrusive sample T98-41, and basalt samples T98-57 and T98-58 also included in the argon geochronology study were obtained by microprobe analyses and are listed in Table S2. Data obtained from the core and rim of individual grains are generally similar. Feldspar chemistry in dolerite sample T98-41 displays a gradual decrease from An₇₉ to An₃₆ with a small gap in composition between An₄₈ to An₄₁. Alkali feldspars are also present as small (<40 µm) interstitial grains and in intergrowth with quartz with a compositional range between Or₇₂ and Or₅₆. Feldspar composition obtained on aphyric basalt sample T98-57 and subophitic sample T98-58 display a smaller range compared with the dolerite sample and range from An₇₉ to An₄₂ but are on average ~An₇₀.

Some basalts (e.g., T98-49) contain trace amounts of iddingsite pseudomorphed after olivine, which is rarely preserved in an unaltered state (*c.f.* the dolerites; Fig. S1). Sample T98-48 contains 1-mm-long, subhedral phenocrysts of orthopyroxene, and sample T98-50 contains augite and diopside phenocrysts. The groundmass of the basalts comprises augite, titanomagnetite and randomly orientated plagioclase. Alteration of the basalts is manifested by the partial to complete replacement of olivine by iddingsite and glassy mesostasis by orange-brown clay minerals.

The majority of the samples collected from intrusions, and sample T98-49a from a thick lava, exhibit well-developed ophitic texture with large augites (1–3 mm) enclosing smaller plagioclase laths (<0.5 mm) (Table S1). Augite is anhedral to subhedral and slightly dominates over plagioclase in some of the intrusive rocks with modal abundances of 40 to 45% and 35 to 40%, respectively. Olivine occurs as

anhedral grains (≤ 0.3 mm) and may constitute up to 15 vol.% of the rock. Opaque phases (titanomagnetite) are anhedral and can reach up to 0.5 mm but are usually smaller (< 0.3 mm). The dolerites are less affected by secondary alteration than the basalts and alteration predominantly affects patchy mesostasis. The main secondary phase is iddingsite occurring along cracks and in the rims of otherwise unaltered olivine grains.

4 Analytical techniques

4.1 Major and trace element data

A total of 35 samples were prepared for analysis by grinding in an agate Tema mill. Major and trace element data were obtained on fusion beads and powder pellets, respectively, by X-ray fluorescence (XRF) using a Philips PW1400 wavelength-dispersive spectrometer at the University of Leicester following standard methods and calibration as described by Harvey et al. (1996). Analytical uncertainties are 1–2% and 5–10% for major and trace elements, respectively. Major and trace element analytical results for the Taimyr samples and selected reference materials are provide in Table 1 and in the supplementary data Tables S3.1 and S3.2, respectively. Rare earth elements (REE) were obtained for 29 samples by dissolving 300mg of powder in 15ml HF and 4ml HClO₄ in an open Teflon (PTFE) beaker on a hot plate until almost dry. This was followed by 4ml HClO₄ acid until complete dryness and the resultant solid dissolved in 30ml 1.7N HCl. After dissolution, REE pre-concentration was achieved using ion exchange columns; these solutions were then analysed on a JY-Ultima-2 inductively coupled optical emission spectrometer (ICP-OES) at the University of Leicester. Details of analytical procedures are given in Harvey et al.

(1996). The analyses of internal basalt standard BOB-1 indicate that precision over the period of analytical work was 8% RSD or better (Table S3.3).

4.2 ICP-MS

Trace elements for a subset of three extrusive and four intrusive samples (Table 1) were analysed using an Agilent 7500a ICP-MS instrument at the Open University, UK according to methods described by Rogers et al. (2006). Detection limits for most trace elements are generally <10 ppt, which is equivalent to <10 ppb in the rock sample. Precision is assessed from repeat analyses of selected unknown samples and an appropriate reference material and, within an individual run, is generally about 1% RSD for elements above Cs (mass 133) and 1-3% RSD for elements below this mass.

4.3 Radiogenic isotope data (Rb-Sr and Sm-Nd)

Nd and Sr isotope compositions were analysed at the NERC Isotope Geoscience Laboratory, Nottingham. Samples were dissolved using a standard HF-HNO₃ procedure. Sr and a bulk REE fraction were separated using Dowex AG-50X8 ion exchange columns. Sm and Nd were subsequently purified using LN-SPEC ion exchange columns. Sr and Nd isotope compositions were determined using a Thermo-Electron Triton mass spectrometer operating in static multi-collection mode. Samples were loaded on single Re filaments (Sr analysis) and on the Ta filament of a double Ta-Re filament assembly (Nd analysis). Eighteen analyses of the NBS987 standard across the period of analysis gave a mean value of 0.710277 ± 0.000006 (1-sigma). Data are normalised to a preferred value of 0.710250 for this standard. Thirteen analyses of an internal Johnson-Matthey Nd standard gave a mean value of

0.511109 $\pm$ 0.000005 (1-sigma), equivalent to 0.511846 for La Jolla. All Nd data are normalised to a preferred value of 0.511860 for La Jolla.

4.4. $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology

We selected a representative and unaltered extrusive (T98-58) and intrusive (T98-41) sample from two different sampling locations (Fig. 1 and Table S1) for the geochronology study. Plagioclase was extracted from the two samples by standard mineral separation techniques and cleaned with 3N HNO₃, distilled water and acetone before hand-picking. Fish Canyon Tuff sanidine (FC-1) was used as the fast neutron fluence monitor with a reference age of 28.172 $\pm$ 0.028 Ma (Riviera et al., 2011). Sample mineral separates and those of the neutron fluence monitor (including CaF₂ and K-glass) were wrapped in copper foil and loaded in quartz vials. Samples T98-41 and T98-58 were irradiated separately for 16 and 48 hrs in the unshielded McMaster Nuclear Reactor and Cd-shielded Petten HRF reactor, RODEO facilities, respectively. Argon measurements were carried out by incremental furnace heating on multi-grain plagioclase fractions at the Geochronology Laboratory, Massachusetts Institute of Technology (MIT), Boston, USA and the NERC Argon Isotope Facility at the Scottish Universities Environmental Research Centre (SUERC), East Kilbride, UK. Analytical and laboratory procedures, blanks and corrections, including data reduction used are described in Pringle (1993).

4.5 Microprobe data

Major element compositions of feldspar were determined on a JEOL JXA-8600 electron microprobe at the Department of Geology, University of Leicester, UK. The microprobe was operated using a 15-kV accelerating voltage and 30-nA beam

current with a 10- μ m beam diameter for analyses. Counting times for each element were 80 seconds; 40 seconds for characteristic X-ray peak measurement and 20 seconds each for positive and negative background. The raw data were corrected using a ZAF correction procedure and are listed in Table S2.

5. The Noril'sk volcanic succession

The most studied part of the Siberian LIP is the 3.5 km extrusive and intrusive succession at Noril'sk, some 750 km southwest of this study area. The Noril'sk extrusive succession has been divided into 11 suites and 10 sub-suites, based on petrography and geochemistry (e.g., Fedorenko, 1981; Fedorenko et al., 1996; Hawkesworth et al., 1995; Naldrett et al., 1992, 1996; Lightfoot et al., 1993; Wooden et al., 1993) and conveniently group into two main sequences: a lower, high-Ti sequence, and an upper low-Ti sequence (Fig. 2a; Lightfoot et al., 1993). The high-Ti sequence includes (upwards) the alkaline Ivakinsky (*sub-suites* Iv1-3), and the sub-alkaline Syverminsky and Gudchikhinsky (Gd1-2) suites, whereas the upper low-Ti sequence comprises the Khakanchansky tuff, and predominately tholeiitic lavas of the Tuklonsky, Nadezhdinsky (Nd1-3), Morongovsky (Mr1-2), Mokulaevsky, Kharaelakhsky, Kumginsky and Samoedsky extrusive suites (Fedorenko et al., 1996). The upper sequence is distinct from the high-Ti suite having lower Gd/Yb_n at varying La/Sm_n and Nadezhdinsky suite basalts having the highest ratios (La/Sm_n>2.0; Fig. 4) which also coincides with a marked decrease in Ni and Cu in this suite.

The intrusive bodies of the Siberian Traps are mainly found along the margins of the Siberian craton but are also well developed along or near fault lines, most notably the Noril'sk-Kharaelakh fault which hosts the Ni-Cu-platinum-group-element ore-bearing

intrusions. The intrusions are represented by a wide range of types including alkali sills, undifferentiated dolerite sills and dykes and differentiated bodies (e.g., Naldrett et al., 1992; Fedorenko et al., 1994; Hawkesworth et al. 1995; Fedorenko et al., 1996). Naldrett et al. (1992) divided the intrusions into five distinct geochemical and petrological groups; group 5 is related to the ore deposits which they further divided into the Cr-rich sub-group 5A (Fig. 3; Noril'sk-type) containing significant sulphides and the Cr-poor sub-group 5B (Lower Talnakh-type) characterised by weakly disseminated sulphides.

6 Results

6.1 Major and trace element data

The major and trace element data are presented in Table 1 and chemical characteristics listed in Table 2. Variations in TiO_2 , P_2O_5 , Fe_2O_3 , Al_2O_3 and CaO , and Cr, Ni, Cu, Nb, Zr and Y element concentrations versus MgO illustrated in Figures 2 and 3, respectively.

The Taimyr intrusive and extrusive rocks range in composition from 9.5 to 5.2 wt.% MgO, but form two distinct clusters (Fig. 2): Group A comprising only intrusive rocks, with MgO ~9 wt.%, and the more evolved Group B including all of the lavas and a few intrusives with MgO below 8 wt.%. The range in MgO and variation in Mg-numbers from 66 to 45 (Table 1) are lower than expected for primary melts from mantle peridotite, indicating that the Taimyr samples are moderately to highly evolved. TiO_2 contents (Fig 2a), with the exception of samples T98-24 and T98-49b, are low, and range from 1.13 to 0.85 wt.%; fall into the Noril'sk low-Ti upper sequence field as defined by Lightfoot et al. (1993). Sample T98-24 has a TiO_2 of 1.71 wt.% straddling the discrimination line, whereas sample T98-49b (3.22 wt.%) falls into the high-Ti

field. As MgO decreases there are expected increases in P and Fe, and incompatible elements such as Zr and Y, whereas Al_2O_3 and CaO decrease, consistent with fractionation of olivine, pyroxene and plagioclase prior to emplacement of the magmas. All Group A samples have similar major element concentrations as the Noril'sk Tukulonsky suite basalts, whereas data of Group B samples broadly overlap with data from the Noril'sk lower and upper sequence basalts.

Trace element concentrations of Cr and Ni in the Group B basalts and dolerites display a wide variation (e.g., Cr between 37–417 ppm) despite the limited range of MgO (Fig. 3); several samples are also low in Ni (<80 ppm) and Cu (<100 ppm) which is a characteristic of the Nadezhdinsky sub-suites Nd1-Nd2 (Lightfoot et al., 1993). Ni abundances in both groups display large variation, which may be related to olivine fractionation but are generally consistent with the moderate Mg numbers. In contrast, Cr in all Group A dolerites is over 350 ppm, showing high values similar to the Noril'sk Tukulonsky suite basalts and Noril'sk Type intrusives associated with ore deposits (Fig. 3a). Group A major and trace element data generally plot with the Tukulonsky suite data and mostly with data of the intrusive Group 5A although these display a wide range of values. Some scatter in the basalt data may also be related to secondary alteration (e.g., basalt sample T98-60 has the highest LOI and lowest Ni; Table 1).

Abundances of incompatible elements, such as Nb, Y, Zr, Th, Ta and rare earth elements (Table 1), are higher in the Group B basalts and dolerites, compared with Group A, consistent with the former's more evolved (MgO range between 5.26 and 7.33 wt.%) composition. Nb/Zr ratios in both groups are constant (Nb/Zr ~0.055) implying that they may be derived from a broadly similar source or processes.

However, Group A has lower Th/Ta ratios (<4.2) compared with Group B samples ($\text{Th/Ta} > 5.3$).

Rare-earth element (REE) patterns and ratios also show that there are subtle differences between the main groups of samples (Figs. 4 and 5). Group A dolerites have bulk silicate earth (BSE) normalised REE patterns with La/Sm_n between 1.26 and 1.63, and Gd/Yb_n between 1.21 and 1.43. The Group B samples in contrast have higher La/Sm_n and Gd/Yb_n and can be divided into two subgroups with differing La/Sm_n ratios of ~ 1.7 and ~ 2.5 (Fig. 5). Group B samples with $\text{La/Sm}_n \sim 2.5$ also have high La/Yb_n ratios between 4.99 and 5.92 (Table 1). The Group A samples show either no or small positive Eu anomalies (Fig. 4 and Table 1), whereas the Group B samples show mostly slightly negative Eu anomalies.

Multi-element diagrams (Fig. 4) illustrate that all Taimyr basalts and dolerites have high abundances of large-ion-lithophile elements (e.g., Ba, Rb and K) relative to the more immobile elements (e.g., Nb, Zr, Ti, and Y) and display a pronounced negative Nb anomaly. Group A dolerites show a positive Sr anomaly (cf. positive Eu), whereas in contrast the Group B samples exhibit negative Nb, Sr, P and Ti anomalies.

6.2 Sills or lavas?

Volumes of un-erupted magma associated with LIPs can be substantial and up to 10 times larger than the associated extrusive volume. The extent to which sills and dykes contribute to magma transport onto the Earth's surface in many large igneous provinces remains unclear (e.g., Muirhead et al., 2014). Intrusive rocks related to the Siberian Traps contribute an estimated 44% of the known volcanic volume and represent along the south-eastern periphery of the Siberian craton the only magmatic

expression (Zolotukhin and Al'Mukhamedov, 1988). There is no indication that crustal-scale discontinuities, such as faults or folding, control the distribution of intrusive and/or lavas in the region. However, a key feature of the Taimyr data set is the strong dichotomy between the more primitive Group A samples, which are all from intrusions, and the more evolved Group B samples which are mostly extrusive. Sills, dykes and lavas show comparable Nb/Zr ratio (~0.055) implying they may be fed from the same or similar source. Estimated magma densities calculated using bulk rock compositions do not show a significant difference between Groups A and B (Table 1). Group A (intrusive) magma densities cluster at 2620 kg/m³ whereas densities of the Group B magmas range from 2570 to 2650 kg/m³ (all calculated assuming a magma temperature of 1200°C; Table 1). However, whereas most of the lavas are aphyric or sparsely phyrlic (with plagioclase as the main phenocryst phase), most of the dolerites contain significant volumes (10–15%) of olivine 'primocrysts' with compositions ranging from Fo₄₅-Fo₇₅. These olivine crystals could increase the fractional bulk density of the magma by up to 100 kg/m³ (see discussion by Sparks and Huppert, 1984), slightly offset by co-crystallisation of less dense plagioclase. It is therefore possible, that compositional differences between intrusive and extrusive rocks may represent a form of density stratification. We must emphasise though, there is insufficient evidence to relate emplacement to either a chemically- or tectonically-controlled mechanism; considering the importance of the intrusive history in Siberia, this issue warrants further research.

6.3 Sr and Nd isotope ratios

A subset of ten samples was analysed for Sr and Nd isotope ratios. The Group A samples (higher MgO, intrusive) have a restricted range of isotope ratios, with initial

$^{87}\text{Sr}/^{86}\text{Sr}$ ranging from 0.705 to 0.706, and $^{143}\text{Nd}/^{144}\text{Nd}$ ranging from 0.5122 to 0.5124 (Fig. 6). The Group B samples (lower MgO, mostly extrusive) have higher Sr isotope ratios than samples of Group A. Three Group B samples form a tight cluster with similar Nd isotope ratios to Group A sample T98-18. Samples T98-41 and T98-50 show similar radiogenic Sr and Nd isotope ratios to the Nadezhdinsky suite basalts. There is a broad correlation between $^{87}\text{Sr}/^{86}\text{Sr}$, La/Sm and Th/Ta with each group displaying elevated (Group B) or low (Group A) ratios (Table 1). These elevated ratios may be attributed to the influence of crustal contamination or magma mixing (see discussion). The Taimyr data broadly overlap with most of the Siberian Traps Noril'sk suite data except the low-Ti Gudchikhinsky suite, which has higher radiogenic Nd ratios (Fig. 6).

6.4 Comparisons with basalts and dolerites from Noril'sk

All of the Taimyr data, except basalt sample T98-49b, plot in the low-Ti field with intrusive sample T98-24 straddling the classification boundary (Fig. 2a). The incompatible element and isotope ratios indicate strong similarities between the Taimyr samples and the Noril'sk upper sequence samples (Figs. 4 and 6). Taimyr Group A intrusive data overlap with analysed upper sequence lavas (Figs. 4, 5 and 6) but most closely resemble those of the Tuklonsky lavas and the Noril'sk Group 5A intrusives (Fig. 4) which include the Noril'sk 1 and Main Talnakh intrusions sharing similar element abundances including key features like positive Sr and negative Nb anomalies.

The Taimyr Group B samples are also compositionally distinct, and some possess the high La/Sm_(n) (Fig. 5) and radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ characteristic (Fig. 6), and low Ni and Cu abundances (Table 1) of the Nadezhdinsky basalt suite. Multi-element

patterns for Group B closely resemble those of the Nadezhdinsky and Morongovsky (sub-suite Mr1) suites (Fig. 4a). However the more radiogenic (e.g., $^{87}\text{Sr}/^{86}\text{Sr}$) Taimyr basalt sample (T98-50) is within the range of ratios reported for the two early Nadezhdinsky sub-suite Nd1 and Nd2 basalts (Wooden et al., 1993). The radiogenic compositions along with high La/Sm of these two Noril'sk suites and sub-suites are inferred to reflect varying degrees of contamination of the magmas by crustal lithologies (e.g., Lightfoot et al., 1993) and this process may also explain the observed range in the Taimyr Group B data.

The initial Nd and Sr isotopic compositions of Taimyr basalts is narrow (0.512218–0.512230 and 0.706199–0.706243) except sample T98-50 which has a significantly unradiogenic Nd but radiogenic Sr composition (0.51193 and 0.70870; Fig. 6). Sample T-98-50 also belongs to the basalt group with $\text{La}/\text{Sm}_n \sim 2.5$. The remaining three basalt samples have radiogenic isotope ratios similar to those reported (Wooden et al., 1993) for the Nadezhdinsky and Morongovsky sub-suites Nd3 and Mr1 with chemical features transitional between those of sub-suites Nd1 and Nd2, and those of all overlying basalts.

6.5 $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology

Data from the incremental heating experiments, including values for the irradiation parameter J and complete analyses, are available in the background data set (Table S4). Age spectra including K/Ca ratios for the incremental-heating experiments are presented in Figure 7 and data summarized in Table 3.

6.5.1 Taimyr basalt sample (Hoffman Peninsula)

Sample T98-58 (Group B, extrusive) was processed in two separate incremental heating experiments (Fig. 7a-b) with size fractions of 75–125 μm and 125–250 μm .

They yielded indistinguishable weighted mean plateau ages of 252.9 ± 1.4 Ma (MSWD = 0.37) and 252.5 ± 1.5 Ma (MSWD = 0.55), respectively. Slightly older ages at low temperature steps are attributed to the loss of ^{39}Ar due to irradiation-induced recoil effects. The plateaux include 19 out of 22 and 9 out of 20 steps with 87% and 77% of the released ^{39}Ar , respectively. The corresponding inverse isochron ages are concordant at 2σ with the plateau ages and show trapped compositions that are indistinguishable from air (Table 3). The error weighted mean age of both plateau ages is 252.7 ± 1.5 (2σ) Ma and considered to represent the crystallisation age of the sample.

6.5.2 Taimyr intrusive sample (Weather Station Flats)

Two separate incremental step heating experiments on size fractions of 125–250 μm and 250–500 μm were performed on plagioclase of sample T98-41 (Group B intrusive; Fig. 7c-d). Both experiments provided discordant disturbed age spectra with <20% of the total radiogenic argon released in the high temperature steps and therefore no reliable plateau age could be derived from these data. Apparent ages rise in the early stages of the gas release followed by decrease leading to a ‘pseudo-plateau’ and a subsequent further steep rise in ages at high temperatures providing a sinusoidal gas release pattern and correlating variable K/Ca (0.012 to 0.42) spectra. Younger apparent ages at low temperatures probably reflect minor alteration. The patterns of disturbance are reminiscent of argon loss and recoil, and indicate a heterogeneous distribution of Ar isotopes (e.g., Heimann et al., 1994; *Note:* Walderhaug et al. (2005) reported similar disturbed argon release pattern in their experiments; their Figure 11). The total fusion ages for these two experiments are

239.2 $\pm$ 0.6 Ma and 238.2 $\pm$ 0.7 Ma, respectively, and are interpreted to provide minimum ages of crystallisation.

7 Discussion

7.1 Argon release spectra and implications for the timing of the intrusive history in Taimyr

The total fusion ages obtained on our intrusive sample are ~10 Ma older (~18 Ma if the high temperature step ages are considered) than ages previously reported (229–227 Ma; Walderhaug et al., 2005) on Taimyr intrusives. However, high temperature steps in both experiments on intrusive sample T98-41 have low K/Ca ratios and provide indistinguishable weighted mean ages of 245.6 $\pm$ 1.4 Ma and 248.3 $\pm$ 1.6 Ma (Fig. 7b-c). Although low K/Ca at high temperature steps indicate the presence of pristine plagioclase these ages have to be considered with caution due to the low combined total amount of radiogenic argon in the high temperature steps (<20%). Argon ages are usually accepted as estimates of the true crystallisation age of a sample if the age spectrum is formed by at least three concordant, contiguous steps representing at least 50% of the ^{39}Ar released (Singer and Pringle, 1996). Although these data do not meet these criteria, useful information can be derived from the argon release pattern of these two experiments.

The sinusoidal age and correlated K/Ca spectra are striking and frequently reported in argon geochronological studies (e.g., Jessberger et al., 1974; Sebai et al., 1991; Heimann et al., 1994; De Min et al., 2003; Rosset et al., 2007) and thought to reflect discordance due to: i) post magmatic sericitic alteration of plagioclase (Sebai et al., 1991; Dalrymple et al., 1995), ii) complex ^{39}Ar and ^{37}Ar redistribution between

cryptoperthite lamellae and host (Rosset et al., 2007), iii) potassium-rich mesostasis associated with plagioclase (Jessberger et al., 1974; Heimann et al., 1994), or iv) the presence of graphic intergrowth between quartz and alkali feldspar (Foland et al., 1993; Heimann et al., 1994). Feldspars in sample T98-41 (and other dyke samples included in this study) display no sign of sericite alteration and surface scanning SEM imaging on HF etched sample surfaces (thin section) revealed no complex exsolution features. Interstitial alkali feldspars and graphic intergrowth between quartz and alkali feldspar ($<20\ \mu\text{m}$) in the vicinity of plagioclase crystals were observed. These areas were also strongly affected by HF leaching. The presence of graphic intergrowths of quartz and alkali feldspar could provide a source of discordance due to radiation-induced ^{39}Ar recoil redistribution in the sample with high and low apparent ages at low and high temperatures steps, respectively (Heimann et al., 1994; Jourdan et al., 2007). The plagioclase grain size ($>125\ \mu\text{m}$) used in our experiments is too large for recoil to be significant and there is no evidence of smaller sub-domains in the plagioclase that could act as fine grained recoil sites.

Figure 8 displays plots of basalt and dolerite apparent ages versus K/Ca and $^{37}\text{Ar}/^{39}\text{Ar}$ with ^{37}Ar primarily originating from calcium. The correlations between low and variable ages and low $^{37}\text{Ar}/^{39}\text{Ar}$ and high K/Ca in the low to medium temperature steps in dolerite sample T98-41 are interpreted to reflect degassing of K-rich intergrowths observed in this sample. Older ages are in contrast associated with high $^{37}\text{Ar}/^{39}\text{Ar}$ and low K/Ca reflecting degassing of pristine plagioclase at high temperatures. The weighted mean ages of $245.6 \pm 1.4\ \text{Ma}$ and $248.3 \pm 1.6\ \text{Ma}$ derived from these high temperature steps are tentatively interpreted to reflect more closely the crystallisation age of dolerite sample T98-41, although a definite age

cannot be inferred from these data. This interpretation is supported by the reliable age obtained on extrusive basalt sample T98-58 where tight clusters of high $^{37}\text{Ar}/^{39}\text{Ar}$ and low K/Ca indicate the presence of pristine plagioclases which are similar in composition to those observed in sample T98-41 (Fig. 8).

7.1.1 Implications for the timing of magmatism in Taimyr

The argon age obtained from a Taimyr basalt lava, together with those published by Reichow et al. (2009), confirm that lava emplacement was contemporaneous with Siberian Traps magmatism at ~252 Ma, thus extending the province to the north of Noril'sk. However, published paleomagnetic and $^{40}\text{Ar}/^{39}\text{Ar}$ evidence (Torsvik and Andersen, 2002; Walderhaug et al., 2005; Zhang et al., 2011) indicate that the intrusive activity took place between 230 and 198 Ma, thus post-dating the Siberian main activity by at least 20 myr. Torsvik and Andersen (2002) also argued that the observed re-magnetisation of Late Precambrian and Palaeozoic sediments between 230–220 Ma on the peninsula is related to thermal re-magnetisation caused by this late Taimyr magmatism. It is apparent that the emplacement of Taimyr lavas cannot be responsible for the Triassic thermal re-magnetisation if the re-magnetisation occurred at 230–220 Ma.

Our argon step heating experiments and those reported by Walderhaug et al. (2005) on Taimyr intrusive samples display highly disturbed argon release patterns. We believe that these patterns are related to the complex (e.g., subsolvus) cooling history of these intrusive rocks including alteration-related argon loss. The feldspar chemistry reported here support this interpretation (Section 3). As outlined above, alkali feldspar is present in the Taimyr intrusives, and if admixed with the plagioclase

aliquots, will ultimately result in ages that are too young. This may be common to intrusives from Taimyr, and also from other areas (e.g., Noril'sk). Based on ages obtained in the high temperature steps of Taimyr sample T98-41 we argue that these intrusives are likely closer to ages ~247 Ma narrowing the reported age gap between these and the 252 myr basalt volcanism. If this hypothesis is correct it will not only have implications on the Taimyr intrusive history but also for dated intrusives with similar disturbed release pattern from the entire Siberian Traps (e.g., Dalrymple et al., 1995) as well as other large igneous provinces. The timing of emplacement of sills and dykes is an important issue as these may have a similar if not larger volume than the Traps basalts on the Siberian craton (e.g., Zolotukhin and Al'Mukhamedov, 1988). If sill and dyke emplacement in Taimyr occurred before 247 Ma, Torsvik and Andersen's (2002) argument that magmatism induced the thermal re-magnetisation of sediments between 230 and 220 Ma needs to be reassessed. There is no other obvious cause than magmatism to induce the pervasive re-magnetisation documented in Taimyr, therefore we conclude that the re-magnetisation, like the intrusive magmatism, must also be older than envisaged. Further high-precision age constraints will help to resolve this issue.

7.2 Crustal contamination versus recycling of continental crustal

Multi-element patterns from the Taimyr igneous rocks (Fig. 4) display pronounced negative Nb and Ti anomalies (although less distinct in Group A intrusives), and high large-ion lithophile element (LILE) abundances. These element characteristics are also observed in most Siberian Traps basalts and attributed to crustal contamination (e.g., Arndt et al., 1993; Wooden et al., 1993; Reichow et al., 2005) or to partial melting of the mantle lithosphere (e.g., Lightfoot et al., 1993; Hawkesworth et

al.1995). Most of the Taimyr Group B data display strikingly similar element abundances to the Siberian Traps Nadezhdinsky and lower Morongovsky suite (Mr1 sub-suites) basalts and are clearly different from ocean island basalt (OIB) or normal (N)-MORB (Fig. 4). The samples also have lower Ti abundances than average OIB and N-MORB. Mg-numbers between 45.1–61.6 indicate that fractional crystallisation controlled at least some of the compositional variation within the Taimyr Group B igneous rocks, but it cannot easily account for the pronounced negative Nb and Ti anomalies (Fig. 4). The Group A samples also show element abundances different from OIB or N-MORB over a narrower range in Mg-numbers (63.8–66.3), thus limiting the strong involvement of fractional crystallisation in the formation of these rocks. Lightfoot et al. (1993) argued that the narrow range of Mg-numbers, and thus the chemical signatures, observed in the Nadezhdinsky to Mokulaevsky suites were inherited from the source – e.g., the subcontinental lithospheric mantle, rather than through an open system process involving crystal fractionation. Although it is difficult to distinguish whether the chemical signature of the Taimyr rocks is derived through assimilation or contamination of continental crust or directly from the source, we explore the feasibility of an assimilation process to account for the observed Taimyr data below.

The average continental crust displays high LILE (e.g., Rb, K, LREE, and Th) relative to HFSE (e.g., Nb, Ta and Ti), (e.g., Rudnick and Fountain, 1995) and consequently has characteristically high Th/Nb ratios. Whereas Th/Nb ratios will not be significantly affected by fractional crystallisation involving plagioclase, pyroxene and olivine, they are highly sensitive to assimilation of continental crust (Fig. 9). Oceanic basalts, which are unlikely to have been contaminated by continental crust, tend to have very

low Th/Nb ratios (<0.1), and low abundances of Th and Nb. This applies equally to MORB and oceanic plateau basalts such as the Ontong Java Plateau (derived by large degrees of mantle melting and with similar Nb/Zr=0.056 as the Taimyr rocks) and ocean island basalts, such as Hawaiian tholeiites (Fig. 9). All of the analysed Taimyr rocks have Th/Nb ratios above 0.2, significantly higher than expected from the closed-system crystallisation of primitive mantle-derived melts (e.g., $\text{Th/Nb}_{\text{Bulk Silicate Earth}} = 0.12$; McDonough and Sun, 1995). Although Mg-numbers in the Nadezhdinsky suite rocks vary only moderately ($\text{Mg\#} = 53\text{--}64$) the difference in Th relative to Nb is consistent with the process of contamination of magmas by continental crust and fractional crystallisation or a combination of these.

Reichow et al. (2005) argued that the observed geochemical variations in the West Siberian Basin (WSB) basalts and similar Noril'sk Nadezhdinsky suite basalts are consistent with processes involving low-pressure fractional crystallisation accompanied by wall-rock assimilation (AFC processes). In order to explain the chemical variation observed in the Taimyr data, combined wall-rock assimilation and fractional crystallisation (AFC) modelling based on the equations by DePaolo (1981) were performed (Fig 10a–c). In the AFC model calculations and in accordance with Reichow et al. (2005) three parental magmas were used. These are a primitive magma derived by 20% partial melting of a bulk silicate earth (BSE) source at 3 GPa, a primitive Ontong Java Plateau (OJP) magma (Fitton and Goddard, 2004), and averaged N-MORB (Salters and Stracke, 2004). The differences between these and the most likely contaminants are shown in Figure 10 and in Table 4. The model was tested using six different north Siberian contaminants: the Bolgokhtokhsk granodiorite (Hawkesworth et al., 1995), Taimyr flyschoid gneiss (Vernikovsky et al., 1995), Aldan granulite gneiss, south Siberian Olekma granodiorite (Jahn et al.,

1998), WSB rhyolite (Medvedev et al., 2003) and Taimyr syenite (Vernikovskiy et al., 2003). A fractionating mineral assemblage was assumed using plagioclase, clinopyroxene and olivine in the proportions of 50:30:20, respectively.

It is apparent that only contamination of a primitive magma from either a BSE source or OJP-type parental magma by Taimyr syenite or Bolgokhtokhsk granodiorite crust can account for the Group A data (Fig. 10a–b). These data are best modelled by contamination of magmas with Taimyr syenite with wall-rock assimilation between 2 and 8%.

The Taimyr Group B data lie between the AFC model lines for the WSB rhyolite and the Olekma granodiorite contaminants when using 20% melt from a BSE source and primitive OJP magma (Fig. 10a and b). This implies that the composition of the possible contaminant may lie between these. However, this is only the case when the rates of assimilation to fractional crystallisation are increased to unreasonably high values ($r = 0.7$) and after 21–45% wall-rock assimilation with an Olekma granodiorite composition. The Group B data are best described when using the WSB rhyolite as contaminant with wall-rock assimilation between 2 to 12% for the OJP magma and 4 to 12% for the primitive BSE-derived magma. Independent of the parental magma, using the Taimyr flyschoid gneiss contaminant in the model calculations requires assimilation of at least 45% ($r = 0.7$) to account for some of the Group B data. This is probably unlikely considering the low SiO_2 content (average of 49.3 wt.%), moderate initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.706202 to 0.708702) and range in Mg-numbers (45.1–61.6) of this group, which does not support high degrees of combined fractional crystallisation and assimilation. Using N-MORB as parental magma (Fig. 10c) wall-

rock assimilation of around 2% is required when contaminated by Taimyr syenite or Bolgokhtokhsk granodiorite to account for the Group B data. In contrast to the BSE and OJP parental magmas, contamination of N-MORB magma by WSB rhyolite only accounts for some of the basalt data. However, using Taimyr flyschoid gneiss requires unreasonably high rates of assimilation to fractionation ($r = 0.7$) with wallrock assimilation of at least 45%, similar to the OJP-type and 20% melt from the BSE source.

It is apparent from Figure 10, that the Taimyr basalt and dolerite data can be modelled using different parental magmas and different contaminants. The AFC model calculations were therefore re-run with the same parameters (e.g., $r = 0.2$), but for a larger element range to evaluate the most likely parental magma and contaminant pair (Fig. 11 and Table 4). Most of the Group B data can be modelled using either the primitive OJP melt contaminated by 4–12% Taimyr syenite (Fig. 11a; note a very similar result is obtained using 20% melt from the BSE source and is therefore not shown) or magma with N-MORB composition contaminated by 4–8% Taimyr syenite (Fig. 11b). However, the negative Nb and Ti anomalies are only recreated when using the primitive OJP (or BSE) parental magma. Taimyr Group A (dolerite) data are also reproduced by using an OJP-type source and 2–8% assimilation of Taimyr syenite, with 4% of assimilation fitting the data best (Fig. 11c).

Low degrees of assimilation are supported by moderate $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ ratios (Figure 6) and Mg-numbers (63.8–66.3) observed in Group A. Further evidence for smaller rates of assimilation or contamination is provided when modelling mixing between the OJP and BSE sources and a radiogenic source (e.g., WSB rhyolite with $^{87}\text{Sr}/^{86}\text{Sr}$ at

0.709072), and variable Sr/Nd ratios (Fig. 6). Between 1 and 20% mixing are required to accommodate the entire observed data range, whereas a mixture between 1 and 5% describe the less radiogenic data. The same results are obtained but using variable Sr/Nd ratios (not shown) enclosing the entire Taimyr data range using Sr/Nd between 6 and 50. Mixing between a Taimyr syenite ($^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ at 0.706440) and the OJP source reproduces only the less radiogenic Taimyr data with mixtures between 10 and 60% (Fig. 6 inset).

These results indicate that AFC processes were likely to be involved in the generation of the Taimyr basalts and dolerites. The low amounts of assimilation required to explain the data are consistent with the SiO_2 concentrations (on average 48.4 and 49.3 wt.% in Groups A and B, respectively) and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ ratios (0.705584 and 0.706911, respectively), and could be related to low temperature conditions in the upper crust (DePaolo, 1981). The AFC model presented in Figure 10 and 11 only fits the Taimyr data if at least two different contaminants are considered in the calculations, which seems reasonable when taking into account the heterogeneity of the continental crust even on a small scale (e.g., Wedepohl, 1995). While the Taimyr igneous rocks could be derived from two different sources that share the same contaminant, it is unlikely that the chemical signature of Group B samples is solely source-related as these are also low in Cu and Ni. Therefore Group B magmas must have reached sulphur saturation. An increase in magmatic sulphur can only be achieved by either extensive fractional crystallisation or by contamination of a mafic magma with continental crust. Considering the moderate Mg-numbers in these rocks, extensive fractional crystallisation could not have caused an increase in sulphur and it is therefore more likely that this was induced by contamination.

7.3 Partial melting and the nature of parental melt(s)

A key feature of the Taimyr igneous rock is their variable and high abundance of LILE and presence of negative Nb and Ti anomalies (Fig. 4). This contrasts with the signature of ocean island basalt data derived from a primitive mantle source. The nature of the mantle source of the Taimyr volcanic rocks is illustrated in a plot of Nb/Y versus Zr/Y (Fig. 12); as these ratios are insensitive to low pressure fractional crystallisation and crustal contamination (Fitton et al., 1997). Nb is markedly depleted in normal-mid ocean ridge basalt (N-MORB) relative to ocean island basalt (OIB) because of previous melt extraction events. A low-Nb, depleted plume component is compositionally distinct from the N-MORB source (Fitton et al., 1997). Most of the Taimyr Group A data plot within or just below the lower bound of the Iceland array close to bulk silicate earth (BSE). In contrast, Group B data plot along or below the lower bound of the Iceland array and, except for samples T98-24 and T98-25, at higher Zr/Y and Nb/Y ratios. A similar data range is also observed in other continental large igneous provinces. Crustal contamination or recycling of continental crust may account for the shift towards higher Zr/Y as observed in Taimyr and other large igneous provinces basalts (e.g., Parana) but it is unlikely to reproduce the Taimyr Nb data if the parental magma was N-MORB.

High and low Nb/Y are derived from a mantle source by small and large degrees of partial melting in the presence of residual garnet, respectively (Fitton et al., 1997). The degree of partial melting can be estimated by using REE data, in particular the light REE which are sensitive to the effects of partial melting. Reichow et al. (2005) used a plot of La/Yb versus Gd/Yb to illustrate the difference in degree of melting of

different mantle source compositions at pressures of 3 and 4 GPa to explain variations observed in the WSB basalt data. The WSB basalts have low Ni and Cu abundances together with high La/Sm, and the pronounced negative Nb and Ti anomalies distinctive of the Nadezhdinsky suite basalts that are also observed in the Taimyr Group B data. The mantle source compositions illustrated in Figure 13 are identical to those used in Reichow et al. (2005) (OJP, primitive mantle (BSE) and depleted mantle (N-MORB)) and melting trend lines are calculated assuming non-modal batch melting with the same parameters as described there. The majority of the Taimyr data plot close to the modelled trend of partial melting for mantle peridotite with OJP mantle or the BSE source composition, at a pressure of 3 GPa. However, the degree of partial melting differs for the two sources and lies between 0.1–10% for the OJP source and 1–30% using a BSE source composition. Taimyr basalt and dolerite data which occur between the melting curves of spinel and garnet peridotite (Fig. 13 inset) suggests derivation within the garnet–spinel transition zone for at least some of the samples. These data are best described with melting of mantle peridotite containing spinel and garnet in proportions near 50:50, and 15 and 25% partial melting. Calculating primary magma composition using PRIMELT2.XLS software by Herzberg and Asimow (2008) demonstrate that melting occurs at 4 to 2 GPa under high ambient mantle temperatures and melt fractions ~30% (Table S5). These calculations support our modelling, but also revealed the influence of pyroxene fractionation potentially leading to elevated FeO contents and mantle source temperatures that are too high (Herzberg et al., 2007). Nevertheless, these conditions are suggested by both model calculations and indicate internal consistency, even with pyroxene fractionation affecting the results for the Taimyr samples.

The degree of crustal contamination recorded by the Taimyr rocks is considered to be low (<12%) on the basis of our element and radiogenic isotope model calculations (Figs. 6, 10 and 11), therefore the Gd/Yb ratio should remain unaffected, even at high degrees of contamination. The partial melting model can be used to identify the most likely source and pressure regime responsible for the generation of the Taimyr igneous rocks. It is apparent from Figure 13 that the Taimyr dolerites and basalts may be derived from the same source but with different degrees of melting. Although the absolute degree of partial melting may be debated it is apparent that the Taimyr basalts and dolerites are not derived from a depleted source (N-MORB). The mantle source that best reproduces the Taimyr data is most similar in composition to that of BSE or the OJP-type sources.

7.4 Petrogenesis of the Taimyr basalts and dolerites

Both Groups A and B were derived by large degrees of partial melting of their mantle source(s), as indicated by their low Gd/Yb ratios (Fig. 5). However, the mineral assemblage (e.g., plagioclase and pyroxene) and chemical data from these samples imply that their petrogenesis was also controlled by processes within deeper (Group A) and low-pressure shallow-level (Group B) magma chambers, respectively. The presence of olivine in Group A samples supports this interpretation. Nevertheless, both groups share similar Nb/Zr ratios but differ in Th/Ta and magma densities calculated from the whole rock data. The former may imply that these two groups share a common source. It is evident that these data do not represent those of primary magmas and evolved in magma chambers isolated from their source regions. Melt evolution included fractional crystallisation and AFC processes.

Group B is most likely derived from a primitive (BSE) or an OJP-like mantle source, rather than a depleted (N-MORB) source. This becomes clear when trying to recreate the Ti and Nb signature of the basalts by AFC processes from an N-MORB source (Fig. 11). The lithosphere to the north of the Taimyr peninsula and in the West Siberian Basin region is substantially thinner than beneath the Siberian craton (e.g., Nikishin et al., 2002). This is consistent with the large degrees of melting required of the primitive mantle (OJP-like or BSE) source. These primitive melts underwent fractional crystallisation combined with wall-rock assimilation at shallow crustal levels. The Group B data are best described by modelling combined wall-rock assimilation and fractional crystallisation (AFC processes) with 8–12% assimilation of a contaminant similar in composition to Taimyr syenite.

In a similar way to Group B basalts, Taimyr Group A (intrusives) was derived by large degrees of melting and subsequent fractional crystallisation at shallow crustal levels. This is supported by low La/Yb ratios and REE abundances (Fig. 4, Table 1) which are lower than those of the Taimyr Group B. The source for the dolerites is unlikely to be depleted (N-MORB) mantle and rather a primitive mantle source (e.g., OJP-type), similar to that proposed for the Taimyr Group B, is favoured. The Taimyr dolerite data are best modelled by combined wall rock assimilation of Taimyr syenite (2–4%) and fractional crystallisation. These degrees of assimilation are smaller than those suggested for the basalts. Based on olivine data, Sobolev et al. (2007) argued that ~20% of recycled crust was involved in the formation of the Siberian Traps lavas. Our binary mixing calculations using radiogenic isotopes agree with the involvement of recycled crustal material. The Taimyr Group A data are best described using the initial radiogenic isotope composition of primitive OJP basalts implying that this could

also be a likely source for this group. Fitton and Goddard (2004) argued that the primary OJP magma was derived from a peridotitic mantle source from which about 1% (by mass) of continental crust had been previously extracted. Based on our radiogenic isotope mixing calculations we infer that a similar process produced the Taimyr intrusive (Group A) rocks.

7.5 Correlating Taimyr and Siberian Traps volcanic units

A distinctive feature of LIPs, including the Siberian Traps, is the dispersal of magmas over long distances with individual lavas extending over 1000 km (Self et al., 2008; Ernst, 2014). Whether lavas were fed from a single source travelling subaerial or through magma transport at depth and fed through multiple vertical dyke systems remains open (Elliot and Fleming, 2004). The volcanic activity on the Taimyr peninsula is often related to that of the Siberian Traps (e.g., Gurevitch et al., 1995). As demonstrated above, Taimyr basalts (Group B) are contemporaneous with and share chemical similarities with the Siberian Traps, in particular the Noril'sk Nadezhdinsky suite basalts (Fig. 4), and the dolerites of Group A closely resemble the Tuklonsky basalts and associated intrusives. This supports the idea of similar conditions for their formation and a possible relationship. From data presented here we argue that Taimyr basalt (all Group B) with $\text{La}/\text{Sm}_n < 2$ (including the dated sample) can be correlated with the Nadezhdinsky and Morongovsky sub-suites Nd3 and Mr1 whereas basalts with $\text{La}/\text{Sm}_n > 2$ can be correlated with Nadezhdinsky sub-suites Nd1 and Nd2. The Nadezhdinsky and Morongovsky suites preceded the main stage volcanism at Noril'sk commencing with high La/Sm basalts and by implication geochemically similar Taimyr samples are therefore potentially older; however, this hypothesis requires confirmation. Trace element data and model calculations

presented here indicate a derivation of the Taimyr basalts by large degrees of partial melting. These melts were subsequently modified through fractional crystallisation and contamination by continental crust in shallow crustal levels. The short duration and inferred large degrees of partial melting may imply the involvement of a mantle thermal anomaly (plume) in the formation of the Taimyr basalts. Characteristics suggestive of a mantle plume origin for a flood basalt province, like the Siberian Traps, include emplacement of large amounts of magma in a short time interval, associated topographic doming and linkage to a hotspot track (Campbell and Griffiths, 1990). Czamanske et al. (1998) argued that evidence for uplift is lacking in the accessible sedimentary cover immediately beneath the Siberian Traps, and thus arguing that there was no lithospheric uplift associated with a buoyant mantle plume. However, Saunders et al. (2005) suggest that doming due to an ascending mantle plume ('Siberian plume') occurred in the area of the West Siberian Basin rather than beneath the thick and stable Siberian craton. Alternatively, the plume may have been intrinsically dense due to an abundant pyroxenite (eclogite) component, and this, associated with erosion of the base of the lithosphere by the plume, may have minimised surface uplift (Sobolev et al., 2011).

Post-magmatic sedimentary sequences in Taimyr show that the lavas were being eroded during the Triassic (Zhang et al., 2013, 2015). Any mechanism addressing the apparent post-magmatic uplift in Taimyr and subsidence in the West Siberian Basin has to account for the apparent short duration for the bulk of the Siberian Traps volcanic activity confirmed by numerous studies (e.g., Renne and Basu, 1991; Campbell et al., 1992; Reichow et al., 2002, 2009; Kamo et al., 2003). Morgan (1981) proposed that the Siberian flood basalt eruptions might be linked to the 'Jan Mayen'

hot spot, with the 'Siberian plume' moving north relative to the continental lithosphere and forming a trail of progressively younger volcanic activity. Evidence to support this idea is lacking, as age determinations on basalts to the north of the Siberian Traps are rare. Our age determinations for Taimyr and those obtained on similar rocks located on the New Siberian Islands further east (Kuzmichev and Pease, 2007), indicating contemporaneous activity with the main Siberian Traps. However, younger volcanic activity is found in the Franz Josef Islands and Svalbard part of the Cretaceous High Arctic LIP (Maher, 2001). Smirnov and Tarduno (2010) reconstructed the position of the Siberian Traps at the time of their eruption and constrain it to the same mantle region which supplied the ~60 Ma old North Atlantic Igneous Province. Whether or not such protracted volcanism linked to the Siberian Traps activity extends north into the Arctic and towards Iceland remains unproven.

The Taimyr dolerite (Group A) data share some similarities with data from the Siberian Traps basalts at Noril'sk, such as the REE, Zr, Ti, Y abundance of the Tuklonsky suite. However, none of the Siberian Traps suites match all the intrusive chemical characteristics of the Taimyr dolerites. These are only matched by Group 5A intrusives related to the ore bearing formations including the Noril'sk 1 intrusion which was dated at 251 myr (Kamo et al., 1996). It has been argued above, that the dolerites underwent different conditions of formation from the Taimyr or Siberian Traps basalts. Although the dolerite chemical characteristics imply large degrees of melting, which might be in accordance with a derivation from a mantle plume, the inferred Early Triassic ages by Walderhaug et al. (2005) would question any genetic linkage to the 252 Ma volcanic activity of the Siberian Traps and plume. If the Taimyr basalts represent the Permo-Triassic activity of the Siberian plume, the gap of at

least 20 myr between the Taimyr basalt and the dolerite emplacement reported by Walderhaug et al. (2005) requires explanation. However, as argued above these younger ages may not represent the true crystallisation ages of these rocks and further geochronology studies to resolve this issue are required.

8 Conclusions

- The Taimyr lavas are contemporaneous with the bulk activity of the Siberian LIP at ~252 Ma, thus confirming Siberian Traps volcanism in the north. Published paleomagnetic and $^{40}\text{Ar}/^{39}\text{Ar}$ evidence suggested that the intrusive activity post-dates the lavas by at least 20 myr. Our new argon data indicate that these younger ages are related to a complex cooling history and that the intrusive activity was likely closer to ~247 Ma, narrowing the reported age gap. Further age determinations are required to confirm this.
- The Taimyr lavas share striking similarities to the Noril'sk Nadezhdinsky and Morongovsky suites which preceded the main stage of Siberian Traps volcanism. This implies that lavas were either transported over large distances, fed through extensive sub-horizontal sills or similar melts were formed concurrently in magma chambers located in Taimyr and the Siberian craton. In contrast to this, the intrusives have more similarities with intrusives related to ore-bearing formations such as the Noril'sk 1 intrusion. This is of particular importance as the Noril'sk district hosts one of the world's largest Ni-Cu-PGE deposits and similar deposits could potentially be located in Taimyr.
- The Taimyr lava and intrusive data imply derivation by large degrees of melting from a primitive mantle source that probably experienced previous melt extraction. These melts underwent fractional crystallisation at shallow

depths, though the presence of olivine in intrusive rocks suggests that their parental magma chamber was hosted slightly deeper. These melts underwent crustal contamination by different contaminants but which did not exceed 12% in both cases forming two distinct groups with low and high MgO content.

9 Acknowledgements

We would like to thank V. Pease and G. Fitton for their very thorough and constructive reviews. The authors are grateful for discussions and suggestions by E. Eide, C. Herzberg, I. Parsons, R. Walker and H. Walderhaug. Reviews of an earlier version by P. Lightfoot and an anonymous reviewer are greatly acknowledged. The $^{40}\text{Ar}/^{39}\text{Ar}$ analysis at the Massachusetts Institute of Technology (MIT), Cambridge, USA and the Scottish Universities Environmental Research Centre, East Kilbride, UK, and M.K. Reichow were supported under the Natural Environment Research Council (NERC) grant NE/C003276/1. Radiogenic isotope analyses were supported by NERC Isotope Geosciences Facilities Steering Committee grants IP/933/1106 and IP/1003/1107. The samples used for this study were collected during fieldwork organised and largely funded by the Swedish Polar Research Secretariat, for which we are very grateful. This paper is dedicated to the late R.A. Scott who is greatly missed.

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Figure and table captions

Figure 1: a) Generalized geological map of the Taimyr peninsula and surrounding areas modified after Zonenshain (1990) and Vernikovsky (1996). Abbreviations are as follows: N, C, S are North, Central, and South Taimyr domains, respectively, and MTT = Main Taimyr Thrust. Sampling area and geological map in b) is shown as rectangle. b) Simplified geological map of central Taimyr partially redrawn after Bezzubtsev et al. (1983), Inger et al. (1999) and CASP field observations. Basalt sample from 'Knobby Island', and dolerite sill and dyke sampling locations are

indicated as open circles. Remaining basalt sampling locations are indicated as rectangle (see Table S1 for details).

Figure 2: Selected major element data (wt.%) versus MgO (wt.%) for basalt and dolerite samples from the Taimyr peninsula. Data for the Siberian Traps lower and upper sequences and the Noril'sk Group 5A intrusives with similar MgO values are shown for comparison. Siberian Traps data source: Lightfoot et al. 1990, 1993; Wooden et al., 1993; Noril'sk Group 5A: Hawkesworth et al., 1995. Note: samples T98-24 and T98-25 are taken from the same sill.

Figure 3: Selected trace element data (ppm) versus MgO (wt.%) for basalt and dolerite samples from the Taimyr peninsula. Data for the Siberian Traps lower and upper sequences and the Noril'sk Group 5A intrusives with similar MgO values are shown for comparison. Data sources as in Figure 2 and Noril'sk Group 5A are from Hawkesworth et al., 1995.

Figure 4: Bulk Silicate Earth normalised (BSE, after McDonough and Sun, 1995) multi-element diagram for a) Taimyr, Nadezhdinsky (Nd) suite and Morongovsky sub-suite Mr1 basalts; b) Taimyr dolerites in comparison with data from the Noril'sk Nd-suite basalts and Group 5a intrusives; c) averaged data of the Siberian Traps suites and Group 5A intrusives (Lightfoot et al., 1990, 1993; Wooden et al., 1993; Hawkesworth et al., 1995); d) Ocean island basalt (Kilauea: Chen et al., 1996; Norman and Garcia, 1999); N-MORB (Saunders and Tarney, 1984; Salters and Stracke, 2004); average Ontong Java Plateau (Kroenke-type) basalt (Fitton and Godard, 2004).

Figure 5: Variations in $\text{La}/\text{Sm}_{(n)}$ versus $\text{Gd}/\text{Yb}_{(n)}$ (n denotes BSE) observed in the Taimyr and the Siberian Traps Noril'sk igneous rocks, and Group 5A intrusives. Ratios from various oceanic settings are also shown to indicate relative depth and degree of melting. Inset: highlighting the observed similarity between Taimyr and Noril'sk Nadezhdinsky (Nd1-3) and Morongovsky (Mr1-2) subgroup data. Data sources as in Figure 4.

Figure 6: Initial (at 252 Ma) radiogenic Sr and Nd variations in the Taimyr lavas showing displacement of the data to the right of the Bulk Silicate Earth (BSE; McDonough and Sun, 1995) mantle at 252 Ma. Binary model calculations using primitive Ontong Java Plateau (OJP) basalt (inset) and BSE sources mixed with radiogenic West Siberian Basin rhyolite and Taimyr syenite (T. syenite; only shown for OJP) at high and low Sr/Nd ratios. Taimyr data source Vernikovsky et al. (2003).

Figure 7: Mineral (plagioclase) age spectra showing $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages and related K/Ca ratios for basalt (a, b) and dolerite (c, d) samples as a function of cumulative percentage of ^{39}Ar released. All ages are relative to Fish Canyon sanidine feldspar standard at 28.172 ± 0.028 Ma (Rivera et al., 2011) and errors quoted at 2σ including uncertainty on the age of the monitor. TF denotes: total fusion age.

Figure 8: Apparent ages versus $^{37}\text{Ar}_{(\text{Ca})}/^{39}\text{Ar}_{(\text{K})}$ and K/Ca for plagioclase separated from Taimyr basalt (a) and dolerite (b) samples. All ^{37}Ar originates from calcium and the correlation between low and variable ages and low $^{37}\text{Ar}/^{39}\text{Ar}$ and high K/Ca are therefore interpreted to reflect degassing of K-rich intergrowths (mesostasis) observed in these samples. Older ages are in contrast associated with high $^{37}\text{Ar}/^{39}\text{Ar}$

and low K/Ca reflecting degassing of pristine plagioclase as seen for the Taimyr basalt. Note the difference in scale for K/Ca.

Figure 9: Th/Nb and Th variations in Taimyr and Siberian Traps igneous rocks.

Trends of crystal fractionation and combined with contamination are indicated. Data for Hawaii OIB, Ontong Java Plateau basalt (OJP) and bulk silicate earth (BSE) are also shown for comparison. Data sources as in Figure 4.

Figure 10: AFC model calculations (DePaolo, 1981) for La/Ba and La/ Nb for the

Taimyr basalts and dolerites. The fractionation assemblage is plagioclase, clinopyroxene and olivine in modal abundances of 50:30:20. Parental magmas are: (a) 20% melt of BSE source, (b) primitive OJP melt and (c) average N-MORB.

Contaminants are: Aldan Granulite Gneiss and Olekma granodiorite: Jahn et al. (1998); WSB Rhyolite: Medvedev et al. (2003); Taimyr flyschoid gneiss and syenite: Vernikovsky et al. (1995) and Vernikovsky et al. (2003), respectively, and Bolgokhtokhsk granodiorite: Hawkesworth et al. (1995). Assimilation to fractionation rates ($M_a/M_c=r$) are shown for $r=0.2$ (solid line) and 0.7 (dashed line, shown only for Taimyr flyschoid gneiss and Olekma granodiorite). Increments on AFC trend lines for $r=0.2$ correspond to 2%, 4%, 8% and 12%, whereas for $r=0.7$ only 21% and 45% are shown. Values and model parameters used are given in Table 4.

Figure 11: Multi element diagram showing the Taimyr basalt and dolerite data range and pattern of magma derived by 20% melting of BSE, primitive Ontong Java Plateau melt as suggested by Fitton and Godard (2004) and averaged N-MORB (Salters and Stracke, 2004). Dashed lines represent the three magmas after combined

assimilation and fractionation (AFC process, $r = 0.2$). Percentages on trend lines indicate the amount of assimilated wallrock. Contaminants used are Taimyr syenite (Vernikovsky et al., 2003) and WSB rhyolite (Medvedev et al., 2003). Data and model parameters are given in Table 4.

Figure 12: Nb/Y versus Zr/Y after Fitton et al. (1997) for the Taimyr igneous rocks, compared with other flood basalt provinces, OIB and MORB. Data sources as in Figure 6; Columbia River: Hooper and Hawkesworth (1993); Paraná: Peate and Hawkesworth (1996); Emeishan: Xu et al. (2001).

Figure 13: La/Yb versus Gd/Yb (normalised to bulk silicate earth: McDonough and Sun, 1995) displaying the range of REE in the Taimyr basalts and dolerites, and possible correlation with partial melting. Data from Kilauea, N-MORB (for clarity only indicated by dashed line), Ontong Java Plateau and the Siberian Traps are also shown for comparison (data sources as in Fig. 4). Trend lines are calculated for an OJP mantle, initial BSE and depleted (N-MORB) mantle source assuming non-modal batch melting. Bulk distribution coefficients are calculated using mineral proportions for garnet peridotite at 3 and 4 GPa after Walter (1998) and spinel peridotite given by McKenzie and O’Nions (1991). Melting of the OJP mantle source for garnet peridotite and spinel peridotite at 3 GPa is indicated by trend line (a) and (b), respectively. Partial melting within the garnet-spinel transition zone at 3 GPa is modelled for the initial BSE source (inset). Note the difference in degree of melting between the two sources. Distribution coefficients used for the REE are from Hanson (1980) and Hart and Dunn (1993).

Figure S1: Photomicrographs of representative Taimyr Group A and B samples.

Table 1: Major (in wt.%) and trace (in ppm) element, and radiogenic Sr and Nd isotope data for the Taimyr igneous rocks.

Table 2: Selected chemical characteristics of Taimyr Group A and B samples, and the Noril'sk Tuklonsky suite basalts and Group 5A intrusives.

Table 3: $^{40}\text{Ar}/^{39}\text{Ar}$ ages for Taimyr igneous rocks. All ages are relative to sanidine feldspar standard FCs at 28.172 Ma (Rivera et al., 2011).

Table 4: Results and parameters for the extended AFC model calculations.

Table S1: Sampling locations of the Taimyr basalts and dolerites, and sample characteristics.

Table S2: Feldspar major element data for samples T98-41, T98-57 and T98-58.

Table S3.1: XRF major element results and currently recommended values for international reference material. Batch IDs indicate individual runs.

Table S3.2: XRF trace element results and currently recommended values for international reference material. Batch IDs indicate individual runs.

Table S3.3: Rare earth element analyses of the internal reference material BOB-1 by ICP-OES.

Table S4: Argon isotope data.

Table S5: PRIMELT2 primary magma solutions (after Herzberg & Asimow, 2008)

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Fig. S1a
Photomicrograph of sample T98-50 (Hoffman Peninsula, group B). Aphyric basalt containing one large, euhedral clinopyroxene (augite). Plagioclase, augite and magnetite are rock forming minerals. Plain polarised light (ppl). Bulk rock MgO = 6.57 wt.%.

Fig. S1b
Photomicrograph of sample T98-56 (Hoffman Peninsula, group B). Basalt showing aphyric texture with randomly distributed, subhedral plagioclase and anhedral augite crystals occurring in equal proportions. Magnetite makes about 5% of the total abundance. Filled vesicle is also present. Chlorite indicates some alteration. Left side: crossed polars (cpl); right side: ppl. Bulk rock MgO = 7.28 wt.%.

Fig. S1c
Photomicrograph of sample T98-49(b) (Hoffman Peninsula, group B). Basalt showing glomerophyric texture with cluster of plagioclase and augite. Groundmass minerals are plagioclase, augite and magnetite (cpl). Bulk rock MgO = 5.26 wt.%.

Fig. S1d
Photomicrograph of sample T98-24 (Thick Sill, group B). Dolerite showing subophitic texture with plagioclase, Ti-augite, olivine and large skeletal magnetite (ppl). Bulk rock MgO = 5.81 wt.%.

Fig. S1e
Photomicrograph of sample T98-12 (Site 6, group A). Dolerite showing subophitic texture with small plagioclase enclosed by large augite crystals. Olivine are anhedral and usually fresh in appearance. Magnetite makes 5–7% of the total mineral abundance. Left side: ppl; right side: cpl. Bulk MgO = 8.89 wt.%.

Table 1 Major (in wt.%) and trace element (in ppm) data for the Taimyr igneous rocks.

Location	Hoffman Peninsula										Knobby Island	Weather Stn Flats	Craggy Island Sill				
Sample no.	T98-60 Group B (Basalt)	'Site 48' T98-58 Group B (Basalt)	'Site 16' T98-57 Group B (Basalt)	T98-56 Group B (Basalt)	T98-55 Group B (Basalt)	T98-53 Group B (Basalt)	'Site 17' T98-51 Group B (Basalt)	T98-50 Group B (Basalt)	T98-49(b) Group B (Basalt)	T98-49(a) Group B (Dolerite)	T98-42 Group B (Basalt)	T98-43 Group B (Basalt)	T98-41 Group B (Dolerite)	T98-35 Group A (Dolerite)	T98-34 Group A (Dolerite)	T98-32 Group A (Dolerite)	T98-31 Group A (Dolerite)
SiO ₂ (wt.%)	49.47	48.30	47.68	48.96	48.72	50.78	50.54	50.98	47.86	50.48	48.85	48.15	50.39	49.03	48.37	48.65	48.69
TiO ₂	1.00	1.06	1.13	1.07	1.09	0.91	1.08	1.01	3.22	1.04	0.98	1.03	0.94	0.94	0.89	0.94	0.90
Al ₂ O ₃	15.45	15.93	16.37	15.59	15.70	15.33	15.75	15.32	13.62	15.10	16.03	15.53	15.77	16.21	15.87	15.94	16.07
Fe ₂ O ₃	10.21	11.46	11.80	11.71	11.64	10.03	10.52	10.65	14.40	10.76	10.05	11.49	10.20	10.80	10.86	10.88	10.89
MgO	7.00	7.16	6.55	7.28	7.29	7.06	6.43	6.57	5.26	6.62	6.74	7.42	7.28	8.40	8.62	8.64	8.85
CaO	8.95	11.55	11.85	11.21	11.32	10.78	10.03	10.28	10.68	10.26	11.42	11.35	10.56	11.57	11.38	11.50	11.21
Na ₂ O	3.07	2.16	2.21	2.33	2.21	2.10	2.37	2.45	2.64	2.37	2.18	2.09	2.59	2.38	2.31	2.31	2.31
K ₂ O	1.14	0.21	0.19	0.29	0.27	0.72	1.03	1.00	0.66	1.02	0.44	0.24	0.99	0.38	0.39	0.39	0.44
P ₂ O ₅	0.13	0.14	0.14	0.14	0.14	0.12	0.14	0.14	0.34	0.14	0.13	0.14	0.14	0.10	0.10	0.10	0.10
MnO	0.16	0.16	0.16	0.17	0.17	0.13	0.15	0.16	0.20	0.15	0.14	0.16	0.15	0.16	0.16	0.16	0.16
LOI	2.89	1.32	0.99	0.46	0.92	1.72	1.24	0.55	0.51	0.82	1.90	1.46	0.12	0.16	0.03	-0.06	0.18
Total	99.5	99.4	99.1	99.2	99.5	99.7	99.3	99.1	99.4	98.8	98.9	99.1	99.1	100.1	99.0	99.4	99.8
Mg# ^a	60.6	58.4	55.5	58.3	58.5	61.3	57.0	58.1	45.1	58.7	60.1	59.2	61.6	63.6	64.1	64.1	64.6
Magma Density (kg/m ³)	2572	2624	2629	2620	2622	2586	2582	2582	2647	2586	2600	2625	2584	2615	2619	2619	2618
Eu/Eu* ^a	-	0.86	0.88	0.85	0.84	0.85	0.88	0.84	0.89	0.99	0.85	0.84	0.94	-	1.11	1.15	1.08
Ba (ppm; XRF)	1120	141	120	168	167	272	329	332	218	328	278	166	348	146	175	154	184
Cr (XRF)	85	136	74	68	64	146	92	30	160	35	151	75	370	405	367	403	443
Cu (XRF)	19	116	113	105	106	59	69	69	230	82	81	105	84	108	111	109	108
Nb (XRF)	7.9	5.7	6.7	5.6	5.5	8.5	8.3	8.4	19.5	9.0	7.7	6.6	6.4	2.7	4.1	3.7	2.8
Ni (XRF)	11.0	120	117	106	107	49.2	40.7	39.8	106	40.5	88.1	109	71.2	93.4	98.8	101	109
Rb (XRF)	35	4	2	16	9	10	18	20	17	22	13	8	23	8	8	8	9
Sc (XRF)	33	32	36	34	33	32	35	33	30	35	31	35	31	34	34	36	33
Sr (XRF)	438	217	228	224	224	236	257	263	377	265	273	229	321	265	271	257	267
V (XRF)	212	229	258	251	234	201	229	228	468	218	200	246	204	239	237	253	231
Y (XRF)	25	23	22	26	27	25	29	28	42	28	25	26	23	19	19	21	21
Zn (XRF)	70	79	89	86	85	77	89	87	117	89	83	87	76	67	68	70	68
Zr (XRF)	120	106	111	109	108	119	140	136	231	137	120	105	110	62	59	62	59
Th (ICP-MS)	n.a.	1.71	n.a.	n.a.	n.a.	n.a.	n.a.	3.17	n.a.	n.a.	n.a.	1.77	1.90	n.a.	n.a.	0.652	n.a.
Ta (ICP-MS)	n.a.	0.323	n.a.	n.a.	n.a.	n.a.	n.a.	0.491	n.a.	n.a.	n.a.	0.325	0.355	n.a.	n.a.	0.201	n.a.
U (ICP-MS)	n.a.	0.581	n.a.	n.a.	n.a.	n.a.	n.a.	0.877	n.a.	n.a.	n.a.	0.612	0.463	n.a.	n.a.	0.152	n.a.
La (ICP-OES)	n.a.	10.13	10.56	10.14	9.88	15.80	16.67	16.32	16.62	21.83	16.51	11.24	15.09	n.a.	5.07	5.46	4.72
Ce (ICP-OES)	n.a.	20.99	25.83	21.33	23.40	33.62	35.97	36.64	32.67	48.45	35.13	25.67	32.39	n.a.	10.50	12.60	10.40
Pr (ICP-OES)	n.a.	2.85	3.26	2.90	2.91	3.76	4.31	4.31	3.52	5.70	3.98	3.04	3.54	n.a.	1.34	1.72	1.42
Nd (ICP-OES)	n.a.	12.43	13.34	12.30	12.33	15.26	17.39	17.00	15.86	26.53	14.89	12.07	14.57	n.a.	7.31	7.92	6.72
Sm (ICP-OES)	n.a.	3.60	3.71	3.54	3.59	3.66	4.31	4.33	3.82	7.39	3.99	3.61	3.62	n.a.	2.40	2.36	2.34
Eu (ICP-OES)	n.a.	1.09	1.14	1.06	1.06	1.02	1.20	1.16	1.05	2.29	1.09	1.05	1.07	n.a.	0.85	0.98	0.78
Gd (ICP-OES)	n.a.	4.15	4.26	4.14	4.13	3.64	4.03	4.10	3.42	6.79	3.70	3.99	3.33	n.a.	2.26	2.88	2.08
Dy (ICP-OES)	n.a.	4.03	4.25	3.97	3.96	3.58	4.13	4.09	3.62	6.11	3.82	3.85	3.25	n.a.	2.48	3.10	2.31
Er (ICP-OES)	n.a.	2.46	2.66	2.49	2.36	2.13	2.62	2.62	2.18	3.34	2.44	2.60	1.92	n.a.	1.51	1.88	1.36
Yb (ICP-OES)	n.a.	2.25	2.37	2.21	2.20	1.92	2.27	2.23	2.01	2.74	2.07	2.20	1.73	n.a.	1.49	1.72	1.36
Lu (ICP-OES)	n.a.	0.27	0.32	0.26	0.28	0.27	0.30	0.28	0.26	0.32	0.29	0.29	0.23	n.a.	0.21	-	0.19
Nb/Zr	0.066	0.054	0.061	0.051	0.051	0.071	0.060	0.062	0.084	0.066	0.064	0.063	0.059	0.044	0.069	0.060	0.048
Zr/Y	4.76	4.64	5.03	4.20	4.01	4.84	4.87	4.90	5.46	4.84	4.86	4.04	4.85	3.16	3.05	3.00	2.84
Th/Ta	-	5.29	-	-	-	-	-	6.46	-	-	-	5.45	5.35	-	-	3.24	-
(La/Yb) _{C1} ^a	-	3.07	3.03	3.12	3.06	5.59	4.99	4.99	5.62	5.42	5.44	3.48	5.92	-	2.32	2.16	2.35
(La/Sm) _{BSE} ^a	-	1.76	1.78	1.80	1.73	2.70	2.42	2.36	2.73	1.85	2.59	1.95	2.61	-	1.32	1.45	1.26
(Gd/Yb) _{BSE} ^a	-	1.50	1.45	1.52	1.52	1.53	1.44	1.49	1.38	2.01	1.45	1.47	1.56	-	1.23	1.36	1.24
⁸⁶ Sr/ ⁸⁷ Sr _m ^a	n.a.	0.706368	0.706346	n.a.	n.a.	n.a.	n.a.	0.709114	n.a.	n.a.	n.a.	0.706552	0.707938	n.a.	n.a.	0.705726	n.a.
⁸⁶ Sr/ ⁸⁷ Sr _i	n.a.	0.706199	0.706243	n.a.	n.a.	n.a.	n.a.	0.708702	n.a.	n.a.	n.a.	0.706202	0.707207	n.a.	n.a.	0.705402	n.a.
¹⁴³ Nd/ ¹⁴⁶ Nd _m	n.a.	0.512482	0.512471	n.a.	n.a.	n.a.	n.a.	0.512155	n.a.	n.a.	n.a.	0.512485	0.512049	n.a.	n.a.	0.512753	n.a.
¹⁴³ Nd/ ¹⁴⁶ Nd _i	n.a.	0.512229	0.512218	n.a.	n.a.	n.a.	n.a.	0.511928	n.a.	n.a.	n.a.	0.512230	0.511828	n.a.	n.a.	0.512409	n.a.

^a Mg# = [100xMgO]/(MgO+FeO) assuming Fe₂O₃/FeO of 0.15; Magma density in kg/m³ calculated at 1200°C; Eu/Eu* calculated with Eu* = 10^[0.5xlog(Sm/Gd)]; BSE and C1 denote Bulk Silicate Earth and carbonaceous chondrite after McDonough and Sun (1995); n.a. = not analyzed; m = measured; i = initial age 252 Ma

Table 1 Major (in wt.%) and trace element (in ppm) data for the Taimyr igneous rocks.

Location	Craggy Island Sill			Thick Sill						"Site 6"				Dolerite Beach (N)				Lr. Relaxed River
Sample no.	T98-30 <i>Group A</i> <i>(Dolerite)</i>	T98-28 <i>Group A</i> <i>(Dolerite)</i>	T98-27 <i>Group A</i> <i>(Dolerite)</i>	T98-25 <i>Group B</i> <i>(Dolerite)</i>	T98-24 <i>Group B</i> <i>(Dolerite)</i>	T98-20 <i>Group A</i> <i>(Dolerite)</i>	T98-19 <i>Group A</i> <i>(Dolerite)</i>	T98-18 <i>Group A</i> <i>(Dolerite)</i>	T98-17 <i>Group A</i> <i>(Dolerite)</i>	T98-16 <i>Group A</i> <i>(Dolerite)</i>	T98-14 <i>Group A</i> <i>(Dolerite)</i>	T98-13 <i>Group A</i> <i>(Dolerite)</i>	T98-12 <i>Group A</i> <i>(Dolerite)</i>	T98-9 <i>Group A</i> <i>(Dolerite)</i>	T98-8 <i>Group A</i> <i>(Dolerite)</i>	T98-5 <i>Group A</i> <i>(Dolerite)</i>	T98-4 <i>Group A</i> <i>(Dolerite)</i>	T98-1 <i>Group A</i> <i>(Dolerite)</i>
SiO ₂ (wt.%)	48.36	48.53	48.36	48.98	49.83	48.09	48.18	48.26	48.39	48.11	48.03	48.26	47.99	48.86	49.25	48.88	48.07	47.83
TiO ₂	0.86	0.89	0.91	0.99	1.71	0.97	1.03	1.00	1.02	0.88	0.85	0.86	0.91	0.89	0.93	0.93	0.92	0.88
Al ₂ O ₃	16.15	15.90	15.84	16.48	14.03	15.25	15.31	15.30	15.46	15.91	15.94	15.94	15.78	16.12	16.21	16.18	15.80	15.98
Fe ₂ O ₃	10.68	10.85	10.86	10.30	12.50	11.13	11.07	11.05	11.07	10.75	10.82	10.80	10.89	10.86	10.92	10.79	10.75	10.62
MgO	8.99	8.70	8.80	7.33	5.81	9.36	8.85	8.85	9.22	9.03	9.48	9.28	8.89	8.84	8.65	8.45	8.57	8.73
CaO	11.43	10.98	11.08	11.44	9.87	11.01	10.97	11.12	11.23	11.31	11.35	11.21	11.62	11.43	11.59	11.13	11.29	11.47
Na ₂ O	2.25	2.27	2.29	2.47	3.04	2.27	2.27	2.27	2.23	2.34	2.13	2.27	2.32	2.21	2.28	2.31	2.23	2.27
K ₂ O	0.35	0.52	0.50	0.69	0.91	0.43	0.51	0.51	0.43	0.36	0.36	0.37	0.35	0.43	0.38	0.57	0.51	0.40
P ₂ O ₅	0.09	0.09	0.09	0.11	0.18	0.11	0.11	0.11	0.11	0.09	0.09	0.09	0.10	0.10	0.10	0.10	0.10	0.09
MnO	0.16	0.16	0.16	0.16	0.21	0.16	0.16	0.16	0.17	0.16	0.16	0.17	0.17	0.16	0.17	0.16	0.16	0.16
LOI	-0.07	0.39	0.42	0.75	0.81	0.35	0.65	0.43	0.43	0.03	0.02	-0.02	-0.08	0.15	0.07	0.22	1.04	0.38
Total	99.2	99.3	99.3	99.7	98.9	99.1	99.1	99.1	99.8	99.0	99.2	99.2	98.9	100.0	100.5	99.7	99.4	98.8
Mg# ^a	65.4	64.3	64.6	61.5	51.1	65.4	64.3	64.3	65.2	65.4	66.3	65.9	64.7	64.7	64.0	63.8	64.2	64.9
Magma Density (kg/m ³)	2620	2615	2617	2601	2601	2625	2620	2620	2624	2620	2625	2621	2624	2619	2618	2612	2618	2620
Eu/Eu* ^a	-	1.12	1.10	1.11	0.96	1.04	1.02	1.00	1.04	1.05	1.14	1.09	1.12	1.11	1.10	-	-	1.14
Ba (ppm; XRF)	148	552	193	314	307	161	227	157	180	148	145	154	149	221	160	273	264	282
Cr (XRF)	389	383	396	417	37	481	452	457	448	400	384	387	395	380	372	385	357	358
Cu (XRF)	99	108	108	101	163	100	101	102	94	102	96	103	106	104	106	104	100	96
Nb (XRF)	2.6	2.7	2.7	3.6	7.1	4.0	4.7	3.5	3.3	3.1	2.2	3.0	2.7	3.4	3.1	3.2	3.2	3.6
Ni (XRF)	111	103	111	93.9	51.5	156	144	139	141	109	114	114	106	102	103	100	102	108
Rb (XRF)	7	14	12	21	28	12	15	17	12	6.2	7.9	8.0	6.2	11	7.4	14	17	7.6
Sc (XRF)	33	34	32	32	34	34	35	36	32	31	32	32	34	32	34	33	34	32
Sr (XRF)	247	293	257	334	306	253	256	252	259	270	269	287	266	272	264	277	266	262
V (XRF)	217	234	234	235	327	240	235	241	220	220	207	216	230	224	238	231	225	215
Y (XRF)	19	19	19	21	34	20	21	21	21	18	18	19	19	19	20	21	20	19
Zn (XRF)	67	67	69	68	119	75	75	71	73	67	65	66	68	68	68	67	67	70
Zr (XRF)	56	58	60	65	123	68	72	72	70	59	56	58	60	59	62	61	61	64
Th (ICP-MS)	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	1.05	n.a.	n.a.	n.a.	n.a.	n.a.	0.614	n.a.	n.a.	n.a.	n.a.
Ta (ICP-MS)	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	0.253	n.a.	n.a.	n.a.	n.a.	n.a.	0.212	n.a.	n.a.	n.a.	n.a.
U (ICP-MS)	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	0.269	n.a.	n.a.	n.a.	n.a.	n.a.	0.144	n.a.	n.a.	n.a.	n.a.
La (ICP-OES)	n.a.	4.74	5.05	5.56	10.55	5.58	5.96	5.63	5.83	4.08	3.99	4.32	4.31	4.95	4.75	n.a.	n.a.	5.02
Ce (ICP-OES)	n.a.	10.53	11.76	11.12	21.42	11.71	12.67	12.34	12.85	9.42	9.57	10.35	9.64	11.40	10.62	n.a.	n.a.	10.06
Pr (ICP-OES)	n.a.	1.37	1.47	1.48	2.76	1.57	1.64	1.65	1.68	1.19	1.23	1.33	1.30	1.53	1.37	n.a.	n.a.	1.34
Nd (ICP-OES)	n.a.	6.52	6.77	7.16	12.83	7.60	7.53	7.89	7.88	6.13	6.18	6.75	6.58	7.03	6.87	n.a.	n.a.	6.49
Sm (ICP-OES)	n.a.	2.14	2.11	2.39	4.25	2.28	2.40	2.34	2.24	1.92	1.85	2.00	2.00	2.36	2.20	n.a.	n.a.	2.09
Eu (ICP-OES)	n.a.	0.82	0.83	0.90	1.37	0.84	0.88	0.83	0.82	0.74	0.75	0.80	0.81	0.83	0.78	n.a.	n.a.	0.83
Gd (ICP-OES)	n.a.	2.36	2.51	2.58	4.52	2.68	2.89	2.73	2.58	2.41	2.17	2.52	2.43	2.19	2.15	n.a.	n.a.	2.36
Dy (ICP-OES)	n.a.	2.73	2.62	2.83	4.83	2.88	3.01	2.88	2.63	2.41	2.40	2.54	2.73	2.49	2.49	n.a.	n.a.	2.60
Er (ICP-OES)	n.a.	1.60	1.62	1.66	2.87	1.62	1.70	1.64	1.68	1.49	1.44	1.53	1.63	1.50	1.41	n.a.	n.a.	1.54
Yb (ICP-OES)	n.a.	1.49	1.43	1.55	2.73	1.57	1.64	1.55	1.54	1.39	1.39	1.47	1.51	1.47	1.38	n.a.	n.a.	1.46
Lu (ICP-OES)	n.a.	0.20	0.19	0.23	0.41	0.24	0.24	0.23	0.22	0.21	0.21	0.22	0.22	0.21	0.20	n.a.	n.a.	0.22
Nb/Zr	0.047	0.046	0.045	0.056	0.057	0.058	0.064	0.049	0.048	0.053	0.040	0.051	0.045	0.058	0.049	0.053	0.052	0.056
Zr/Y	3.02	3.00	3.13	3.14	3.61	3.45	3.52	3.38	3.37	3.28	3.11	3.05	3.20	3.05	3.06	2.91	3.08	3.44
Th/Ta	-	-	-	-	-	-	-	4.15	-	-	-	-	-	2.90	-	-	-	-
(La/Yb) _{C1} ^a	-	2.17	2.41	2.44	2.63	2.42	2.47	2.47	2.57	2.00	1.96	2.01	1.94	2.29	2.34	-	-	2.35
(La/Sm) _{BSE} ^a	-	1.39	1.50	1.46	1.55	1.53	1.55	1.51	1.63	1.33	1.35	1.35	1.35	1.31	1.35	-	-	1.51
(Gd/Yb) _{BSE} ^a	-	1.29	1.42	1.35	1.35	1.39	1.42	1.43	1.36	1.41	1.26	1.39	1.30	1.21	1.26	-	-	1.31
⁸⁶ Sr/ ⁸⁷ Sr _m ^a	n.a	n.a	n.a	n.a	n.a	0.706170	n.a	0.706545	n.a	n.a	n.a	n.a	n.a	0.706028	0.705657	n.a	n.a	n.a
⁸⁶ Sr/ ⁸⁷ Sr _i	n.a	n.a	n.a	n.a	n.a	0.705665	n.a	0.705870	n.a	n.a	n.a	n.a	n.a	0.705613	0.705368	n.a	n.a	n.a
¹⁴³ Nd/ ¹⁴⁶ Nd _m	n.a	n.a	n.a	n.a	n.a	0.512629	n.a	0.512509	n.a	n.a	n.a	n.a	n.a	0.512739	0.512645	n.a	n.a	n.a
¹⁴³ Nd/ ¹⁴⁶ Nd _i	n.a	n.a	n.a	n.a	n.a	0.512332	n.a	0.512239	n.a	n.a	n.a	n.a	n.a	0.512407	0.512366	n.a	n.a	n.a

^a Mg# = [100xMgO/(MgO+FeO)] assuming Fe₂O₃/FeO of 0.15; Magma density in kg/m³ calculated at 1200°C; Eu/Eu* calculated with Eu* = 10^[0.5xlog (SmxGd)]; BSE and C1 denote Bulk Silicate Earth and carbonaceous chondrite after McDonough and Sun (1995);

n.a. = not analyzed; m = measured; i = initial age 252 Ma