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Age, petrogenesis and metamorphism of the syn-collisional Proven Igneous Complex, West Greenland

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Abstract

The Paleoproterozoic Prøven Igneous Complex (PIC) in West Greenland extends from ca. 72°15' to 73°10'N, approximately 500 km north of the subduction-related intrusive complex in the core of the +1100 km wide, asymmetric collisional Nagssugtoqidian-Rinkian Orogen. A new U-Pb SHRIMP age for the PIC of 1869±9 Ma indicates that it intruded synchronously with the main collisional phase of the orogen into the passive margin side of the collision. Sm-Nd and Lu-Hf isotopic and A-type geochemical signatures are compatible with its derivation from melted Archean lower crustal material contaminated to varying degrees by pelitic sedimentary rocks of the Karrat Group. The timing, petrogenesis and position of the PIC within the orogen support a model of collisionally induced delamination of the mantle lithosphere following initial collision. Upwelling asthenospheric mantle replacing the partially or completely detached mantle lithosphere caused widespread partial melting of lower crust that resulted in the areally extensive (~250,000 km²) Cumberland-Prøven intrusive complexes of Baffin Island and West Greenland. Emplacement of the PIC at 1.87 Ga caused a high-temperature low- to medium-pressure metamorphic aureole that contrasts the regional, overprinting higher-pressure amphibolite facies metamorphism. The consequent high-temperature garnet-orthopyroxene-biotite-bearing assemblages occurring within the margin of the intrusion in the aureole are attributed to the intrusion event. Garnet-controlled Sm-Nd and Lu-Hf ages of 1.82-1.80 Ga require efficient diffusion of these elements during orogenic reheating at this time. This age range overlaps the post-collisional, north-south shortening in the Nagssugtoqidian Orogen to the south and serves to confirm the recently proposed genetic link between these two orogens. These new data infer that garnet-controlled isochrons based on the Lu-Hf and Sm-Nd systems cannot date high-grade events in slowly cooled or significantly reheated terrains in rocks possessing other phases that close at low temperatures. © Springer-Verlag 2005.

Keywords

complex, igneous, proven, west, collisional, greenland, syn, metamorphism, petrogenesis, age

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Age, petrogenesis and metamorphism of the syn-collisional Prøven Igneous Complex, West Greenland

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Abstract

The Paleoproterozoic Prøven Igneous Complex (PIC) in West Greenland extends from ca. 72°15' to 73° 10' N approximately 500 km north of the subduction-related intrusive complex in the core of the +1100 km wide, asymmetric collisional Nagssugtoqidian-Rinkian Orogen. A new U-Pb SHRIMP age for the PIC of 1869 ± 9 Ma indicates that it intruded synchronous with the main collisional phase of the orogen into the passive margin side of the collision. Sm-Nd and Lu-Hf isotopic and A-type geochemical signatures are compatible with its derivation from melted Archean lower crustal material contaminated to varying degrees by pelitic sedimentary rocks of the Karrat Group. The timing, petrogenesis and position of the PIC within the orogen support a model of collisionally-induced delamination of the mantle lithosphere following initial collision. Upwelling asthenospheric mantle replacing the partially or completely detached mantle lithosphere caused widespread partial melting that resulted in the areally extensive ($\sim 250,000$ km²) Cumberland-Prøven intrusive complexes of Baffin Island and West Greenland.

Emplacement of the PIC at 1.87 Ga caused a high-temperature low to medium pressure metamorphic aureole that contrasts the regional, overprinting higher pressure amphibolite facies metamorphism. The consequent high temperature garnet-orthopyroxene-biotite-bearing assemblages occurring within the margin of the intrusion in the aureole is attributed to the intrusion event. Garnet-controlled Sm-Nd and Lu-Hf ages of 1.82-1.80 Ga require efficient diffusion of these elements during orogenic reheating at this time. This age range overlaps the post-collisional, north-south shortening in the Nagssugtoqidian Orogen to the south and serves to confirm the recently-proposed genetic link between these two orogens. These new data infer that garnet-controlled isochrons based on the Lu-Hf and Sm-Nd systems cannot date high grade events in slowly-cooled or significantly-reheated terrains in rocks possessing other phases that close at low temperatures.

Keywords: Paleoproterozoic; delamination; garnet diffusion; Prøven Igneous Complex; Rinkian

Introduction

The northeast Laurentian craton in the North Atlantic region comprises several anastomosing Paleoproterozoic collisional orogens that stitch together previously-stabilized Archean cratons (Hoffman 1990; Van Kranendonk et al. 1993; van Gool et al. 2002) (Fig. 1). Realistic tectonic models for formation of this region require integration of correct lithological interpretations,

complex field relationships, accurate partitioning of Archean and Paleoproterozoic deformation fabrics, geochronology, and chemical and isotopic data. In particular, the identification of Paleoproterozoic subduction-related intrusive suites is pivotal to all Paleoproterozoic tectonic models as they delineate the placement of pre-collisional subduction zones and, thus, cratonic boundaries. While there is general agreement about the nature, age, extent and significance of most Paleoproterozoic intrusive complexes in the North Atlantic region, the details of the largest of these remains enigmatic. The granitic Paleoproterozoic Prøven Igneous Complex (PIC), extending from ca. 72°15' to 73° 10' N in West Greenland (Fig. 2), and its larger counterpart, the Cumberland Batholith, Baffin Island, (northeast Canada) underlie an area of more than 250,000 km². Most existing models envision a subduction-related system that require convergence and eventual collision of two cratonic blocks to explain the occurrence of these complexes; an interpretation permitted by the present understanding of the geology along a Quebec-Labrador-Baffin Island transect (Grocott and Pulvertaft 1990; Hoffman 1990; Van Kranendonk et al. 1993; Scott 1999). However, a subduction-related origin seems at odds with its emplacement into passive margin sequences in both Canada (Piling and Lake Harbour Groups (Jackson et al. 1990)) and Greenland (Karrat Group) and is not easily reconciled with emerging tectonic models for West Greenland (van Gool et al. 2002; Connelly et al. in review). More specifically, the Nagssugtoqidian-Rinkian orogenic system appears to reflect convergence and collision of two cratonic blocks where subduction dipped southwards beneath the North Atlantic craton to produce arc-related magmatism (the Sisimiut and Arfersiorfik intrusive suites) about 500 km south of the PIC (van Gool et al. 2002).

The objectives of this study were to better understand the timing, petrogenesis and tectonic setting of the PIC and evaluate the Lu-Hf and Sm-Nd systematics of garnet in high-grade assemblages of known age relative to the intrusion. Towards these ends, we present new Sensitive High Resolution Ion Microprobe (SHRIMP), geochemical and high-precision Lu-Hf, Sm-Nd, Rb-Sr, Pb-Pb isotopic data. Our results indicate that the 1869 ± 9 Ma PIC represents partial melting of the lower continental crust variably contaminated by passive margin sedimentary rocks in the upper crust that we attribute to delamination of mantle lithosphere triggered by collisional orogenesis. Furthermore, we infer that rare earth elements (REE) and Hf diffuse at comparably rapid rates at millimeter length scales in garnet at amphibolite facies temperatures.

Regional geology

Previous accounts of West Greenland geology have described two Paleoproterozoic orogenic belts named the Nagssugtoqidian Orogen and Rinkian Belt, to the south and north, respectively (Fig. 1). With the boundary between them poorly defined and a general lack of geochronological data, they were principally differentiated on the basis of structural style (Escher & Pulvertaft 1976, Escher et al. 1976). The Nagssugtoqidian Orogen is a more upright, intensely-folded belt in contrast to the open structural style of the Rinkian Belt. The failure to identify a distinct boundary, combined with new structural and geochronological data that demonstrates an overall kinematic and temporal similarity has led recent investigators to infer that the Rinkian Belt represents a northward continuation of the Nagssugtoqidian Orogen (van Gool et al. 2002; Thrane et al. 2003; Connelly et al. in review). If correct, this system collectively defines a single, asymmetric, ~1100 km-wide orogenic belt of a similar scale to the present day Himalayan Orogen.

The Nagssugtoqidian-Rinkian Orogen predominantly comprises Archean orthogneisses reworked to varying extents and metamorphic grades during the Paleoproterozoic. The Nagssugtoqidian Orogen also contains Paleoproterozoic magmatic and metasedimentary rocks, the former attributed to 1.92-1.87 Ga arc magmatism related to a south-dipping subduction zone beneath a southern craton. In contrast, Archean orthogneisses of the Rinkian Belt are overlain by a very thick and widespread package of Paleoproterozoic supracrustal rocks, known as the Karrat Group. Based on lithologies, stratigraphy and extent of this group, as described mainly by Henderson and Pulvertaft (1987), we interpret this sequence to represent a Paleoproterozoic passive margin sequence.

Granulite facies rocks in the central Nagssugtoqidian Orogen were interpreted to represent the higher-grade collisional core of this asymmetric orogen. Deformation and metamorphism diminish northward in intensity and grade, respectively, such that the Rinkian Belt comprises a remarkably-consistent amphibolite-facies assemblages in which rims of plagioclase commonly occur on garnet.

The Prøven Igneous Complex (PIC)

The PIC is more than 100 km in diameter and intrudes Archean orthogneisses and the Karrat Group in the Upernavik region of the central Rinkian Belt (Fig. 2). The complex, which remains virtually unstudied, was first referred to as the Prøven granite by Escher and Pulvertaft (1968) but was

renamed by Escher and Stecher (1978) as the Prøven charnockite upon recognizing that the body contained 5-15% hypersthene. Although the dominant rock type is a medium- to coarse-grained, orthopyroxene-bearing granite, we prefer the informal term Prøven Intrusive Complex to include all phases of this suite. Where observed by us, garnet content varies from 0 to 15% and is especially abundant along the contacts where it is clear from xenoliths and transitional contacts that pelites of the host Karrat Group were variably assimilated. A weak foliation is present in the internal parts of the body whereas it is locally strongly foliated along its outer contacts. The only existing age for the complex prior to this study was an Rb-Sr whole-rock isochron age of 1860 ± 25 Ma reported by Kalsbeek (1981), indistinguishable from the U-Pb zircon date of 1869 ± 9 Ma we report below. All samples examined of the PIC during this study were margin phases and consist of feldspar + quartz + biotite + garnet + orthopyroxene + ilmenite, which appear to be in textural equilibrium (Fig. 3). The minerals lack crystal shape and the grain boundaries generally meet at 120 degrees. Monazite is locally present as small grains showing textures consistent with partial resorption. Quartz is present in the matrix and as both inclusions in and rims around garnets (Fig. 3).

Host Rocks

The PIC intrudes Archean orthogneisses and metapelites of the Paleoproterozoic Karrat Group. Typically at amphibolite facies across most of the Rinkian Belt, these rocks contain granulite facies assemblages where we observed them proximal to the PIC. Rb-Sr analyses of Archean orthogneisses collected near Upernavik yield an errorchron of ca. 1.9 Ga, in contrast to the regional amphibolite orthogneisses that yield an age of 2860 Ma (Kalsbeek, 1981). Metapelites of the Karrat Group contain orthopyroxene and garnet proximal to the PIC in contrast to the regional amphibolite facies assemblage of qtz-pl-bt-ms. Approaching the PIC from the south, the typical regional assemblage in the Karrat Group gives way to sillimanite and extensive migmatization within a zone 15 km wide along the southern contact (Henderson & Pulvertaft, 1987). Along the northern boundary, orthopyroxene-bearing fabrics in the Karrat Group are cut by dykes of the PIC.

Analytical methods

Sample preparation

Five samples of the PIC and one sample of the orthogneiss were crushed to small chips with a standard jaw crusher. A split of the crushed material was powdered in a tungsten carbide shatter box for major and trace element, as well as whole rock isotopic analyses. Mineral separates for

isotopic analyses were milled to mineral size with a disc mill, sieved, and the 200-500 micron size fraction passed through a Frantz magnetic separator. This process provided purified mineral fractions from which final mineral separates were isolated by handpicking using a binocular microscope.

Major and trace elements

Major and trace element analyses were performed on four whole rock samples of the PIC and on the sample of Archean orthogneiss. Major elements were analysed by X-ray fluorescence (XRF) on glass discs, and trace elements were analysed on an Elan 6300 quadrupole inductively coupled plasma mass spectrometer (ICP-MS) after dissolution in HF and HNO₃. All analyses were carried out at the Geological Survey of Denmark and Greenland in Copenhagen.

In situ zircon dating by Sensitive High Resolution Ion Microprobe (SHRIMP)

Sample 269457 of the PIC was dated by the SHRIMP U–Pb zircon method at the Research School of Earth Sciences, Australian National University, using the SHRIMP 1 instrument. Zircons were hand picked and placed on two-sided tape, collared, and mounted in a transparent epoxy resin. The zircon mount was ground and polished to expose any potentially inherited cores.

Cathodoluminescent images characterised the zircons and guided analyses. ²³⁸U/²⁰⁶Pb of the unknowns were referenced to the standard AS3 with an age of 1099 Ma (Paces and Miller 1993), and U was calibrated against a fragment of the standard single crystal SL13, with 238 ppm U. The ²⁰⁷Pb/²⁰⁶Pb date obtained from the five AS3 analyses interspersed with the unknowns was 1120 ± 25 Ma, within error of the ID-TIMS value. Analytical protocols, calibration of data and calculation of analytical errors are summarised by Stern (1998) and Williams (1998). Common Pb was corrected for using measured ²⁰⁴Pb and Cumming and Richards (1975) model Pb at 1870 Ma.

Sm-Nd and Lu-Hf analyses

Whole rock Lu, Hf, Sm and Nd isotope analyses were performed on four samples of the PIC. In addition, two of the PIC samples were selected for isotopic analyses of a subset of garnet, biotite, feldspar, orthopyroxene and Fe-Ti oxide phase(s). For Lu-Hf and Sm-Nd analyses, approximately 50 mg were processed for whole rocks and mineral analyses, except oxides, for which only approximately 25 mg were analysed. Two fractions of garnet were picked from each sample, one of which was leached in 7N HNO₃ for 3 hours on the hotplate with the intention to remove LREE-rich

micro-inclusions such as epidote, allanite, apatite, sphene or monazite. This procedure is not expected to eliminate all inclusions, but was intended to increase the Sm/Nd ratio of the residual leached garnet and, therefore, the precision of the calculated age (Thöni 2002). Sample 483634 contained both ilmenite and magnetite, which were combined and analysed together. The oxide fraction picked from 483641 consisted of pure magnetite. The oxide fractions were leached in cold 2N HCl for 15 minutes before digestion to remove micro-inclusions of apatite.

Both the whole rock powders and mineral separates were digested by flux fusion in high-purity graphite crucibles as described in Bizzarro et al. (2003). The molten sample is poured into 23 ml Savillex beakers containing approximately 10 ml of 2N HCl and appropriate amounts of isotopic tracers of ^{176}Lu , ^{180}Hf , ^{149}Sm and ^{150}Nd . The beakers are shaken until the samples are in solution after which the chemical separation is carried out in three steps: (1) cation exchange resin chemistry, which isolates heavy REE (including Lu) from middle-light REE's (including Sm, Nd) and high field strength elements (HFSE including Hf), (2) Hf is further purified from Step 1 using TEVA-spec® resin, and (3) Sm and Nd are further purified from the waste of Step 2 using Re-spec® resin. For a more detailed description of these analytical methods see Bizzarro et al. (2003). Lu-Hf ages were calculated using the Lu-176 decay constant of Scherer et al. (2001). The present day mean value of chondrites reported by Blichert-Toft and Albarède (1997) was used to calculate ϵ_{Hf} values.

Lu, Hf, Sm and Nd isotope ratios were measured on the VG AXIOM MC-ICP-MS at the Danish Lithosphere Centre. The 2 standard deviation (2sd) external reproducibility of $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ are 0.2% and 0.003% respectively, and 0.2% and 0.002% for $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$. We conservatively assign 0.5% 2sd errors on $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{147}\text{Sm}/^{144}\text{Nd}$.

Pb-Pb analyses

For Pb isotopic analyses, 20-40 mg mineral fractions were picked from sample 483641. The minerals were rinsed several times in hot HCl (except for biotite, which was rinsed in cold HCl) and dissolved in a mixture of concentrated HF and HNO_3 in Savillex capsules on a hotplate. After four days, the dissolved samples were evaporated, and a small amount of concentrated HNO_3 was added and evaporated immediately. Finally, 7N HCl was added and the samples were left on the hotplate for another 24 hours. The samples, now completely in solution, were evaporated again and taken up in 1N HBr for anion HBr chemistry. Pb was separated on columns loaded with anion

exchange resin as described in Baker et al. (2004). The natural isotopic composition of Pb was determined for all samples on the VG AXIOM MC-ICP-MS at the Danish Lithosphere Centre. For the least radiogenic samples (feldspar, biotite and orthopyroxene), a ^{207}Pb - ^{204}Pb double spike was added and the samples re-analysed to obtain higher precision isotopic data (ca. ± 100 ppm; see Baker et al. 2004). The magnetite and garnet samples did not contain enough Pb to obtain as precise analyses as for those above and were, therefore, not analysed by this method. The procedural blank was 80 pg and the SRM-981 standard was run sequentially with unknowns yielding a mean $^{208}\text{Pb}/^{206}\text{Pb}$ ratio of 2.1677.

Rb-Sr analyses

Feldspar and biotite from sample 483634 were analysed for Rb-Sr. The mineral fractions were dissolved as described for the Pb analyses. After the samples were dissolved, a mixed ^{87}Rb - ^{84}Sr spike was added and the samples were evaporated and taken into solution in 2N HCl. Rb and Sr were separated using conventional cation exchange columns and analysed on the VG Sector 54-IT thermal ionization mass spectrometer at the University of Copenhagen.

All ages and associated errors were calculated using Isoplot/Ex (Ludwig 1999).

Results

Major and trace element chemistry

Major and trace elements for the four PIC samples are presented in Table 1 and illustrated on incompatible element and REE diagrams in Figure 4. These diagrams also show ranges of averaged values for the: (a) subduction-related rocks of the Arfersiorfik intrusive suite of the central Nagssugtoqidian Orogen (Kalsbeek 2001), (b) Archean orthogneisses of the Nagssugtoqidian Orogen (Kalsbeek 2001), (c) lower crust (after Rudnick and Fountain 1995).

SiO_2 contents of 64 to 68% (Table 1) for the PIC are slightly lower than that for the sample of host Archean orthogneiss (72%) as well as the average for Archean orthogneisses in the Nagssugtoqidian Orogen (ca. 70%, Kalsbeek 2001). Elemental concentrations of K_2O and Na_2O in the PIC overlap with the analysed sample of Archean orthogneiss but are slightly higher and lower, respectively, than average values for the Archean orthogneisses in the Nagssugtoqidian Orogen (Kalsbeek 2001). Despite the wide range of compositions exhibited by the PIC samples, they are

generally enriched in incompatible elements from Ba to Hf (especially Rb and Th but not P) and less enriched in the elements Ti to Yb relative to average lower crust and Archean rocks of the Nagssugtoqidian Orogen (Fig. 4a). Although there is a greater degree of overlap with Paleoproterozoic arc rocks of the Nagssugtoqidian Orogen, contents of incompatible elements Rb to La and Zr to Y are higher in the PIC, whereas Eu and Sr are lower. In general, the PIC is strongly enriched in light REE (LREE) and middle REE (MREE), but have flatter heavy REE (HREE) patterns that overlap with average lower crust and rocks of the Nagssugtoqidian Orogen, and have significant negative Eu-anomalies (Fig. 4b).

Geochronology and Isotopes

SHRIMP

Sample 269457 of the PIC yielded simple 200-300 μm long prismatic zircons, with oscillatory zoning and without obvious inherited cores or later overgrowths. They are interpreted to represent a simple igneous population. SHRIMP analyses are summarised in Table 2, and are presented on a Tera-Wasserburg $^{238}\text{U}/^{206}\text{Pb} - ^{207}\text{Pb}/^{206}\text{Pb}$ concordia diagram in Figure 5. The amount of common Pb detected in all analyses was minor. Analysis of both centres and rims of several grains were undertaken to check for inherited components and/or variable ancient loss of radiogenic Pb. All analyses are concordant within error and yield indistinguishable $^{207}\text{Pb}/^{206}\text{Pb}$ ratios. They yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1869 ± 9 Ma (95% confidence, MSWD = 0.89), which we interpret to represent the age of crystallization of this body.

Sm-Nd

Sm-Nd isotope analyses were obtained from four whole rock samples, and mineral separates of two samples (Table 3). Replicate analyses were also performed on two of the whole rock samples. The replicate whole rock results from sample 483634 plot very close together, whereas the two results from 483641 do not overlap, most likely due to powder heterogeneity as initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of both samples agree within error. All the whole rock samples yield very negative $\epsilon_{\text{Nd}}(0)$ values that range from -34 to -24 (Fig. 6a), with $\epsilon_{\text{Nd}}(1870)$ values between -5.2 and -4.3 , corresponding to T_{DM} ages of 2660-2500 Ma. All whole rock analyses yield a statistically acceptable line corresponding to an age of 2040 ± 85 Ma (MSWD = 1.6) and an initial $^{143}\text{Nd}/^{144}\text{Nd} = 0.50987 \pm 0.00006$ (Fig 6b). As discussed below, this age has no significance but instead must reflect mixing of two components of varying Sm-Nd ratios and Nd isotopic signatures.

483634. Analyses of mineral separates of garnet (leached and unleached), biotite, feldspar, orthopyroxene and the two whole rocks define an isochron corresponding to an age of 1799 ± 7 Ma (MSWD = 1.3) and an initial $^{143}\text{Nd}/^{144}\text{Nd} = 0.51003 \pm 0.00001$ (Fig. 6c). The leached garnet yields more radiogenic Nd than the unleached garnet, commensurate with LREE-rich micro-inclusions (not optically visible) in the garnet having been preferentially removed during the acid leach step.

483641. As in the previous sample, the highly-radiogenic garnet fractions control the isochron slope, but in this case the leached and unleached garnet yield two different ages. Surprisingly, the unleached garnet is more radiogenic than the leached garnet. The unleached garnet, all other mineral fractions and the two whole rock analyses define an age of 1821 ± 18 Ma (MSWD = 2.2) with an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio = 0.51002 ± 0.00002 (Fig. 6d), which are within error of the internal isochron for sample 483634. An age of 1764 ± 19 Ma (MSWD = 1.9) is obtained using the leached garnet rather than the unleached garnet. As it appears that the leaching procedure has fractionated Sm and Nd from the garnet, we defer to the age based on the unleached garnet fraction.

Lu-Hf

Lu-Hf isotope analyses were performed on the same samples and mineral separates as for Sm-Nd described above, although replicate analysis of only one whole rock was performed (Table 4). All whole rock samples have very negative $\varepsilon_{\text{Hf}}(0)$ values that range from -45 to -35 and have ε_{Hf} (1870 Ma) values between -6.7 and -5.1 (Table 4). Extrapolating to the depleted mantle evolution curve of Vervoort and Blichert-Toft (1999) derives T_{DM} ages of 2700-2600 Ma. Although all five analyses of four whole-rock samples do not form a statistically valid regression (2081 ± 310 Ma (MSWD = 8.9)), four of the analyses (omitting 483643) define a statistically acceptable line corresponding to an age of 2085 ± 68 Ma (MSWD = 1.3) with an initial ratio of 0.28140 ± 0.00001 (Fig. 7b). As with the Sm-Nd whole-rock analyses, this age is attributed to the mixing of two components with different Lu-Hf ratios and Hf isotopic compositions, as discussed below.

483634. Setting the Lu content of ilmenite to zero (it was too low to measure) and excluding the leached garnet, Lu-Hf analyses of all mineral fractions plus the two whole rock analyses yield an errorchron age of 1842 ± 15 Ma (MSWD = 12) and define an initial ratio of 0.28147 ± 0.00005 (Fig. 7c). No combination of garnet and any group of minerals with low Lu/Hf ratios define an

acceptable isochron. However, it should be noted that two point “isochrons” (garnet combined with individual low Lu/Hf phases) only vary in age from 1839 to 1843 Ma (± 9 Ma). That the leached garnet fraction falls above the Lu-Hf errorchron indicates that the leaching step fractionated Lu and Hf.

483641. As in sample 483634, the Lu content for ilmenite assumed to be zero and the leached garnet fraction is excluded. A four point isochron is defined by garnet, orthopyroxene, biotite and ilmenite (excluding the whole rock analyses) and corresponds to an age of 1820 ± 9 Ma (MSWD = 0.99) with an initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.28142 ± 0.00001 (Fig. 7d). The leached garnet plots above the line and was not included in the regression.

Pb-Pb

Pb isotope analyses were performed on five mineral fractions (biotite, feldspar, garnet, magnetite and orthopyroxene) from sample 483641 (Table 5). Feldspar, biotite and orthopyroxene have relatively-unradiogenic Pb and yield precise Pb isotope ratios with $^{206}\text{Pb}/^{204}\text{Pb} = 15.7$ to 26.1. Magnetite and garnet yielded relatively low amounts of Pb resulting in somewhat less precise Pb isotope ratios. All 5 analyses define a linear array (MSWD = 1.7) corresponding to an age of 1796 ± 4.6 Ma (Fig. 8).

Rb-Sr

Biotite and feldspar fractions were separated from sample 483634 and analysed for Rb and Sr (Table 6). The resulting two-point date is 1606 ± 5 Ma, with an initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.71119 \pm 0.00033$. As a two point regression, this is not a true isochron and the error must be higher than strictly calculated here.

Discussion

This study was undertaken to better understand the age and petrogenesis of the PIC, thereby refining models for the regional tectonic history of the North Atlantic region. Additionally, determination of garnet-controlled Lu-Hf and Sm-Nd isochron ages for rocks in a thermal aureole of known age provides the required constraints to assess diffusion in garnet.

Crystallization Age

Zircon analyses by SHRIMP yields an age of 1869 ± 9 Ma, which is interpreted to reflect the crystallization age of the PIC. Furthermore, since the growth of the garnet-orthopyroxene-biotite bearing assemblages within and around the PIC are attributed to the heat related to emplacement, this also constrains the timing of the growth of this assemblage.

Granite Petrogenesis

Geochemistry: The dominant geochemical signature of the PIC is one of general enrichment of K, Fe, incompatible elements including HFSE and LREE relative to HREE, high Fe^*/Mg ratios and significant negative Eu anomalies, all relative to average lower crust (Rudnick and Fountain 1995) and Archean orthogneisses of the Nagssugtoqidian Orogen (Kalsbeek 2001). To lesser degree, the PIC also appears distinct from the Paleoproterozoic arc rocks of the Nagssugtoqidian Orogen in the enrichment of other incompatible elements, especially Rb, Th, K, Nb, Zr, Hf, MREE, and more depleted in Eu. By contrast, the arc rocks in the Nagssugtoqidian Orogen are enriched in Sr and P. The PIC signature is also enriched in incompatible elements with respect to the well-documented Paleoproterozoic subduction-related arc rocks of the Narsajuaq Terrane of northern Quebec (Dunphy and Ludden 1998). The geochemical signature of the PIC samples is broadly similar to that of A-type granites (as outlined by Loiselle and Wones 1979; Collins et al. 1982; Whalen et al. 1987; Eby 1992). Although the petrogenesis and tectonic setting of A-type granites remain controversial and may involve different settings, the original geochemical classification highlights the unique geochemical features of this spectrum of granitoid rocks and clearly differentiates them from subduction-related granitoid rocks.

Without tectonic inference, petrogenetic models to explain the A-type signatures fall into two broad categories: 1) highly-fractionated derivatives of mantle-derived melts (Foland and Allen 1991; Turner et al. 1992; Kerr and Fryer 1993), and 2) relatively-high temperature partial melts of ($\pm$ previously depleted) continental crust (including tonalite-trondjemite-granite (TTG) suites) (Collins et al. 1982; Whalen et al. 1987; Creaser et al. 1991). Whereas geochemical data cannot easily distinguish between these two models, our whole rock Sm-Nd and Lu-Hf isotopic data discussed below favour formation of the PIC mainly via partial melting of the lower Archean crust.

Sm-Nd and Lu-Hf whole rock isotopes: The Sm-Nd and Lu-Hf systems behave similarly in that the daughter elements (Nd and Hf) are more incompatible than the parents (Sm and Lu), with an

expected stronger coupling of the Sm-Nd system given they are both LREE. Depleted mantle model ages (T_{DM}) for samples in both systems range between 2700 and 2500 Ma, as calculated using the models of DePaolo (1981) and Vervoort and Blichert-Toft (1999) for Sm-Nd and Lu-Hf, respectively. These ages are 300-500 m.y. younger than the known ages of Archean orthogneisses in the area (Thrane et al. 2003). Furthermore, both isotopic systems yield valid isochron ages ca. 200 Myr older than their crystallization age.

The young T_{DM} age and statistically-valid but false whole rock isochrons requires mixing of two components to produce variably-contaminated batches of magma. The isotopic compositions of the two components must have been homogeneous and straddle the range of isotopic compositions of the analysed samples at 1.87 Ga and possess appropriate parent-daughter ratios to derive common back-projected isotopic compositions at their respective false isochron ages (Fig. 9).

We identify three likely sources for the PIC magmas, namely, upper mantle, Archean gneisses and metasedimentary Karrat Group. Assuming conservative isotopic and elemental compositions and melt production models for the mantle (including very undepleted signatures) and Archean gneisses, we find it impossible to explain the data solely by batch emplacement of different mixtures of these two components (Fig. 9) as it would result in older false isochron ages than obtained.

However, varying degrees of contamination of Archean crust derived partial melts by metasedimentary rocks of the Karrat Group satisfies all observations. More specifically, partial melting of a previously-depleted 2.8-3.0 Ga Archean TTG lower crust would yield magmas enriched in incompatible elements and with Nd and Hf isotopic compositions appropriate to represent the least radiogenic component at 1.87 Ga (Fig. 9). Conversely, pelitic sedimentary rocks of the Karrat Group had more radiogenic isotopic compositions at 1.87 Ga than the PIC samples and a less enriched geochemical signatures appropriate for the second component, as documented by Kalsbeek (1998).

Therefore, we conclude that homogeneous magmas produced by partial melting of depleted Archean orthogneisses of the lower crust were variably contaminated by homogeneous pelitic sedimentary rocks of the Karrat Group. Partial melting must have caused sufficient fractionation of

the parent lower crust to engender back-calculated Nd and Hf isotopic compositions of the derived lower crustal melts that intersect the unfractionated bulk Karrat group isotopic evolution at 2.04 and 2.08 Ga, respectively. This model is supported by field observations of large xenoliths of Karrat Group in the PIC intrusion, gradational boundaries between the PIC and host, and the high, but variable, garnet content of the PIC within its margins. We cannot preclude very minor amounts of mantle as a third component that homogenized with lower crustal melts before their ascent and contamination by Karrat Group (Fig. 9), but this must have been minor and in fixed proportions.

Age of the Garnet-Orthopyroxene Assemblage

Lu-Hf and/or Sm-Nd garnet geochronology is typically undertaken in regional metamorphic terrains to evaluate the timing of metamorphism. This study offers the relatively-unique condition in which garnet (+opx-bt-ilrn) formed within a high-temperature metamorphic aureole of known age, thereby permitting evaluation of garnet-controlled isochron ages relative to the age of peak temperatures. The following section discusses evidence indicating that the analyzed assemblages reflect high temperature contact metamorphism related to the emplacement of the 1.87 Ga PIC rather than either an igneous assemblage or a post-intrusive regional metamorphic assemblage.

First and foremost, mineral textures observed in all thin sections reflect a general lack of crystal shape and commonly exhibit triple junctions, both consistent with metamorphic recrystallization rather than igneous crystallization. Furthermore, the occurrence of quartz inclusions in garnet indicates that the latter overgrew a pre-existing solid-state assemblage rather than having formed from an igneous melt. While clear that the granulite assemblages in the PIC are metamorphic rather than igneous, the question remains as to whether they are related to the heat of intrusion of the PIC at 1870 Ma or an overprinting regional metamorphic event of unknown age.

Several lines of evidence indicate that the garnet-bearing granulite assemblages in the border zones of the PIC must be related to its emplacement and partial digestion of Karrat Group host. First, the host rocks around the PIC also possess granulite assemblages, in contrast to the regional upper amphibolite facies assemblages typical of most of the Rinkian Belt. This is most obvious south of the PIC where regional middle amphibolite facies rocks give way to orthopyroxene-bearing migmatites within 15 km of the southern margin of the PIC. Degree of partial melting increases northward along Uvkugsigssat Fjord approaching the southern margin of the PIC (Henderson and

Pulvertaft 1987). Second, orthopyroxene-bearing fabrics within the Karrat Group are cut by dykes of the PIC, requiring that the local granulite assemblages must be at least as old as the PIC. This point is distinct from the first as it requires that the high temperature event that caused granulites around the PIC is at least as old as the intrusion rather than being coincidentally centred around the PIC after its intrusion. Thirdly, monazite from the Karrat Group migmatites adjacent the PIC return ca.1870 Ma U-Pb ages (Connelly & Thrane unpublished data), requiring that partial melting here was coincident with the PIC intrusion.

We infer that the garnet-rich portions of the marginal PIC reflect partially-digested Karrat Group interacting with PIC magmas to produce metamorphic grt-opx-bt-ilm during high temperatures associated with the emplacement of the PIC. This assemblage is predicted for the high Fe/(Fe+Mg), low Al composition of these rocks (Table 1) at pressures of 0.5-8.0 kbar and temperatures of 650-850°C (Spear et al. 1999). These relatively-high temperatures are supported by the occurrence of partial melts as inferred by quartz moats around garnet (Fig. 3; Johnson et al. 2003)

With this high temperature assemblage spatially and temporally related to the PIC, we confidently ascribe its formation to the emplacement of the intrusion at 1869 ± 9 Ma rather than any younger, overprinting event. This requires that the ages defined by Lu-Hf and Sm-Nd isochrons cannot correspond to the formation age of this assemblage or peak temperatures in this region.

Significance of the Isochron Ages

Sample 483634 yielded only a Lu-Hf errorchron, and is not considered further. Hence, acceptable mineral isochrons defined by Sm-Nd, Lu-Hf and Pb correspond to a tight cluster of ages between ca. 1.82 and 1.80 Ga, significantly younger than the interpreted growth ages of garnet and orthopyroxene in the PIC and its aureole. Possible explanations for this discrepancy are that: (1) Nd, Hf and Pb mineral ages reflect the first closure after slow, post-crystallization cooling, or (2) the Lu-Hf, Sm-Nd and Pb systems were re-opened between 1820 and 1800 Ma during a subsequent regional metamorphic reheating and that the PIC initially cooled at a much faster rate than that following regional metamorphism.

Slow Cooling: The closure temperatures of both Sm-Nd and Lu-Hf systems in garnet remains controversial. Estimates for closure temperatures for Lu-Hf in garnets range from ca. 540 to

>700°C, whereas closure temperatures for Sm-Nd range from 480 to 900°C (Scherer et al. 2000 and references therein). Variations in closure temperature are controlled mainly by the size of the grains and, to a lesser extent, the chemical composition of the phases and cooling history. We acknowledge this uncertainty and that garnet in the PIC ranges in size from ca. 1 to 8 mm, requiring that they would have closed over a large temperature interval. However, for the sake of evaluating a crude estimate of cooling rate, we assign an average closure temperature of 700°C for Lu-Hf and Sm-Nd in garnet for this exercise. Based on Giletti (1991) and Jenkin et al. (1995), we assume a closure temperature of 400°C for the Rb-Sr system in the biotite-feldspar pair and we estimate that the PIC would have intruded at temperatures of approximately ca. 850°C. Plotting available data on a temperature-time path yields a smooth, upward-concave cooling path that spans 1870 to 1600 Ma (Fig. 10). This path would correspond to a very-slow, post-orogenic rebound of tectonically over-thickened crust after 1.87 Ga emplacement of the PIC with an average cooling rate of 1-2 °C/Myr. However, this scenario is inconsistent with four regional first-order observations: 1) the mineral assemblage of the PIC aureole requires that it intruded at shallow to middle crustal levels, implying that, without immediate crustal overthickening and/or additional heat input, this region would have cooled more quickly, 2) regional amphibolite facies metamorphism reflects higher pressures followed by rapid decompression (Grocott et al., 2004), implying that this metamorphic event is not coeval with the medium to low-pressure granulite facies metamorphism, 3) deformation associated with the regional amphibolite facies appears to be kinematically incompatible with syn-intrusive deformation in the aureole, and 4) the Nagssugtoqidian Orogen was active until at least 1820 Ma (Connelly et al. 2000), implying that the correlative Rinkian Belt was not likely experiencing slow post-orogenic rebound between 1870-1820 Ma.

Orogenic Reheating: Instead, integrated isotopic data and regional field, metamorphic and structural relationships are more compatible with the ages reflecting the time of closure during or after orogenic reheating during a younger, higher-pressure amphibolite facies metamorphic event. The uncertainty in closure temperatures of Hf and Nd in garnet precludes determining whether the garnet-controlled isochron ages represent the time of near-peak temperatures during the amphibolite facies event or points along the cooling path after this event. However, the clustering of Lu-Hf, Sm-Nd and Pb-Pb ages derived from minerals representing a range of different grain sizes and closure temperatures, regional field evidence for a change in kinematics after syn-PIC deformation (implying episodicity) and clear mineralogical and textural evidence of rapid decompression after

this second event (implying tectonically-driven post-amphibolite exhumation) collectively support an interpretation in which the isochron ages approximate the time of regional amphibolite facies metamorphism. Therefore, we infer that the regional amphibolite facies metamorphism in the Rinkian orogen occurred between ca. 1.84-1.80 Ga, an event distinct and significantly younger than the emplacement of the PIC. Regardless of whether the isochron ages represent peak amphibolite facies metamorphism or a subsequent cooling age, it is clear that the garnet-controlled Lu-Hf and Sm-Nd systems do not retain the ages of granulite facies (i.e. opx-bearing assemblages) metamorphism in either slowly-cooled or significantly reheated terrains.

A Revised Tectonic Model

Our new geochemical and isotopic data are not consistent with earlier interpretations of a subduction-related origin for the PIC (and Cumberland Batholith) (Grocott and Pulvertaft 1990; Hoffman 1990; Van Kranendonk et al. 1993; Scott 1999). Furthermore, we interpret that the PIC intruded into a passive margin side of a convergent system in which subduction dipped south beneath the North Atlantic Craton. The consequent arc-related intrusive rocks now lie in the central Nagssugtoqidian Orogen, approximately 250 km south of the proposed suture zone (Connelly et al. in review) and 500 km south of the PIC. Therefore, given the widespread presence of the Karrat Group on both sides of the PIC, the lack of geochemical and isotopic data to support a subduction-related origin for the PIC and the lack of regional geological relationships to support a second collisional zone or unrecognized continent within the Rinkian Belt, we prefer to interpret the PIC as having been generated via partial melting of the lower crust and emplaced at 1869 ± 9 Ma into the passive margin side of a colliding pair of Archean cratons.

The problem remains as to what triggered the large-scale melting to generate the PIC and the Cumberland Batholith. The timing of emplacement of the PIC is broadly coeval with the end of subduction-related intrusive rocks in the Nagssugtoqidian Orogen (from 1908 ± 4 to 1873 ± 7 Ma; Connelly et al., 2000) and the main collisional phase of the central Nagssugtoqidian Orogen (D1 estimated to be underway by 1860 Ma; Connelly et al. 2000). Within the Rinkian Belt, it is clear that the PIC intruded already deformed Karrat Group, requiring that deformation was already underway by the time the PIC was emplaced. Given that lesser-deformed PIC-related intrusions cut deformed earlier phases, it is clear that deformation continued during the emplacement of PIC such that it intruded syn-tectonically.

While details of the timing and rate of orogenic crustal thickening and eventual collapse of the Rinkian Belt remain to be determined, the spatial and temporal relationship of the PIC with collisional orogenesis is clear. We propose a model in which collision-induced crustal shortening and thickening destabilized the mantle lithosphere, initiating its partial or complete delamination from the overlying Archean continental crust. Upwelling buoyant asthenosphere replacing the vacated mantle lithosphere would have provided the heat necessary for widespread partial melting of the Archean lower continental crust. Given distances of at least 300 km between the proposed delamination beneath the Rinkian Belt and extinct subduction zone to the south and the lack of similar intrusive rocks in the intervening region, it seems unlikely to be directly related to the detachment of subducted oceanic lithosphere as proposed by Sacks and Secor (1990). Instead, it seems more likely that delamination in the Rinkian Belt would have been initiated by more local conditions. While speculative, we suggest that differential responses to shortening by stiff mantle lithosphere vs. ductile lower crust across the Moho during the pre-intrusive collisional phase may have destabilized the crust-mantle interface. This may have set the stage for detachment during either continued shortening and/or subsequent relaxation and extension. Whether the mantle lithosphere fully detaches or not may not be vital, provided pathways are provided for buoyantly upwelling asthenosphere to make contact with the lower crust. However, once initiated, the upwelling asthenosphere would induce a positive feedback that would encourage delamination and more upwelling.

Conclusions

The isotopic data and geochemical trends obtained from the 1869 ± 9 Ma PIC are compatible with its derivation from melted Archean lower crustal material contaminated to varying degrees by homogeneous pelitic sedimentary rock of the Karrat Group. No evidence exists to support a juvenile mantle source as typically envisioned for Paleoproterozoic subduction complexes throughout the North Atlantic region. The timing and position of the PIC relative to the main collisional core of the orogen (approximately 500 km to the south) support a model of collisionally-induced partial or complete delamination of the mantle lithosphere. Upwelling asthenospheric mantle represents the proposed heat source for partial melting that was widespread enough to form the 250,000 km² Cumberland-Prøven intrusive complexes on Baffin Island and West Greenland.

Sm-Nd and Lu-Hf isotopic systems in syn-intrusive (ca. 1869 Ma) garnet-orthopyroxene-biotite assemblages were reset during orogenic reheating between 1.82-1.80 Ga, an age that reflects the timing of the main, regional amphibolite facies metamorphism across the Rinkian Belt. Whether this dates peak amphibolite facies metamorphism or subsequent cooling is less certain, but it is clear that diffusion of Sm, Nd, Lu and/or Hf in garnet is fast enough that garnet-controlled isochrons based on these systems cannot date high grade events in slowly-cooled or significantly-reheated terrains that possess multiple low-T phases.

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Figure captions

Fig. 1. Trans-Hudson orogens (ca. 1.87 – 1.78 Ga) and Archean cratons of the North Atlantic region before the Cretaceous drift.

Fig. 2. Geological sketch map of the central Rinkian orogen (from Henderson and Pulvertaft 1987).

Fig. 3. Plane- and cross-polarised light photomicrographs of sample 483641 showing the high temperature assemblages.

Fig. 4. Chondrite normalized (Sun and McDonough 1989) trace element variation diagram (a) and REE diagram (b) for the PIC samples compared to arc rocks and basement gneisses in the central Nagssugtoqidian Orogen (Kalsbeek 2001). The basement gneiss collected close to the PIC is also plotted. The composition of the lower continental crust is taken from Rudnick and Fountain (1995).

Fig. 5. Tera-Wasserburg concordia diagram for zircons from the Prøven intrusive complex (sample 269457). Error crosses are 1σ .

Fig. 6a. Whole rock Nd isotope evolution diagram
b. Sm-Nd whole-rock isochron
c. Sm-Nd mineral isochron of sample 483634
d. Sm-Nd mineral isochron of sample 483641. The leached garnet () is not included in the calculation of the age.

Fig. 7.a. Whole rock Hf isotope evolution diagram
b. Lu-Hf whole rock isochron
c. Lu-Hf mineral isochron of sample 483634. The leached garnet () is not included in the calculation of the age.
d. Lu-Hf mineral isochron of sample 483641. The leached garnet () is not included in the calculation of the age.

Fig. 8. Pb-Pb mineral isochron for sample 483641

Fig. 9. Epsilon vs. time diagram showing the evolution of the PIC samples compared to Archean Lower Crust, Karrat Group and a partial melt of depleted mantle (DM). The thin dashed line in the $\epsilon(\text{Hf})$ diagram represents sample 483643, which was not included in the calculation of the isochron age. It is clear that the false isochron ages and the discrepancy in the epsilon values at the time of crystallization is a consequence of a lower crustal melt being contaminated by Karrat Group metasedimentary rocks.

Fig. 10. T-t history of the Prøven igneous complex. The solid line represent the reheating path, the dashed line show the slow cooling path.

Fig. 1

Fig. 2

Fig. 3

Fig. 4

Fig. 7

Fig. 8

Fig. 10

Table 1.
Major and trace element data for the PIC

Sample	483634	483639	483641	483643	483638
	Prøven	Prøven	Prøven	Prøven	Orthogneiss
Major elements					
SiO ₂ , wt%	67.6	66.3	64.7	68.2	72.2
TiO ₂	0.54	1.19	0.76	0.64	0.09
Al ₂ O ₃	15.3	13.3	15.2	14.3	14.2
Fe ₂ O ₃	0.96	1.03	0.80	0.96	0.55
FeO	3.30	5.30	4.19	3.56	0.81
MnO	0.05	0.02	0.05	0.04	0.01
MgO	1.25	2.28	1.54	1.41	0.30
CaO	3.14	3.12	2.53	2.65	1.26
Na ₂ O	3.26	2.53	2.40	2.90	3.29
K ₂ O	3.08	2.92	6.04	3.52	5.89
P ₂ O ₅	0.15	0.36	0.34	0.23	0.07
Volatiles	0.55	0.66	0.80	0.88	0.27
Trace elements					
Sc, ppm	15.7	16.0	13.8	15.3	1.64
V	41.6	90.3	74.2	58.5	17.6
Cr	33.0	67.9	41.8	38.5	4.16
Co	27.3	29.8	26.1	27.6	26.6
Ni	12.6	30.9	16.9	18.2	2.7
Cu	13.6	23.4	15.8	16.9	1.8
Zn	64.9	106	69.5	63.3	18.1
Ga	20.4	24.4	19.0	18.7	15.9
Rb	131	71	171	99	161
Sr	155	376	197	143	213
Y	25.5	26.3	14.6	29.7	2.2
Zr	227	488	199	195	100
Nb	13.7	25.2	16.8	14.3	0.394
Cs	0.426	0.119	0.133	0.164	0.0846
Ba	696	1,089	1,084	574	965
La	41.4	211	57.1	42.4	18.7
Ce	81.8	419	119	86.6	29.5
Pr	9.12	44.1	13.2	9.37	2.85
Nd	32.5	147	49.2	33.5	9.47
Sm	5.81	19.2	8.95	6.28	1.37
Eu	1.10	2.07	1.40	1.11	0.986
Gd	6.64	19.1	8.69	6.94	1.50
Tb	0.876	1.67	0.889	1.09	0.131
Dy	5.01	7.23	4.00	6.10	0.611
Ho	0.945	0.960	0.578	1.06	0.0826
Er	2.84	2.65	1.55	2.84	0.233
Tm	0.386	0.234	0.156	0.360	0.0282
Yb	2.47	1.29	0.854	2.13	0.178
Lu	0.336	0.172	0.124	0.280	0.0272
Hf	5.87	11.5	5.16	5.26	2.84
Ta	3.10	2.56	1.19	2.06	1.07
Pb	26.9	45.5	30.6	30.2	31.6
Th	10.5	69.4	12.9	10.9	0.416
U	1.17	2.02	1.55	2.29	0.297
ID data					
Sm	6.812	19.51	10.02	6.340	
Nd	38.92	150.2	58.61	33.50	
Lu	0.366	0.162	0.134	0.305	
Hf	5.304	11.00	7.220	7.239	

Major elements analysed by XRF on fused glass discs (Na by AAS). Volatiles: loss on ignition corrected for uptake of O during oxidation of Fe. Trace elements analysed on an Elan 6300 quadrupole ICP-MS. ID analyses from VG AXIOM MC-ICP-MS.

Table 2

SHRIMP U/Pb zircon analyses for 269457, 72°47'W 55°37'N

spot	site	U	Th	Th/U	common	²³⁸ U / ²⁰⁶ Pb	²⁰⁷ Pb / ²⁰⁶ Pb	²⁰⁷ Pb / ²⁰⁶ Pb	%disc
	type	ppm	ppm		²⁰⁶ Pb%	ratio	ratio	age	
1.2	m,osc,p,fr	363	224	0.62	0.07	2.990 ± 0.071	0.1118 ± 0.0020	1829 ± 33	2
1.3	m,osc,p,fr	416	295	0.71	0.15	3.044 ± 0.057	0.1155 ± 0.0007	1888 ± 11	-3
2.2	m,osc,p,fr	241	107	0.44	0.04	2.926 ± 0.064	0.1155 ± 0.0014	1888 ± 22	0
3.2	m,osc,p,fr	522	382	0.73	0.02	2.938 ± 0.064	0.1144 ± 0.0010	1870 ± 15	1
4.2	m,osc,p	208	91	0.44	0.23	2.968 ± 0.059	0.1133 ± 0.0014	1853 ± 22	1
4.3	e,osc,p	254	134	0.53	0.04	3.134 ± 0.073	0.1126 ± 0.0011	1842 ± 18	-3
5.2	m,osc,p	520	362	0.70	0.03	2.951 ± 0.056	0.1132 ± 0.0011	1851 ± 17	2
5.3	e,osc,p	414	271	0.65	0.02	3.054 ± 0.074	0.1140 ± 0.0008	1864 ± 12	-2
6.2	e,osc,p,fr	326	160	0.49	0.06	2.997 ± 0.070	0.1151 ± 0.0010	1881 ± 16	-1
7.1	e,osc,p	357	168	0.47	0.16	3.033 ± 0.067	0.1138 ± 0.0011	1861 ± 17	-1
7.2	m,osc,p	407	264	0.65	0.12	2.995 ± 0.058	0.1144 ± 0.0008	1870 ± 12	-1
8.1	e,osc,p	281	152	0.54	0.11	3.045 ± 0.077	0.1156 ± 0.0016	1889 ± 24	-3
9.1	e,osc,p	412	250	0.61	0.09	3.026 ± 0.063	0.1142 ± 0.0008	1867 ± 12	-1

p=prism, e=end, m=middle, osc=oscillatory finescale zoning, fr=crystal fragment, spot #.#=grain.analysis

corrected with 1860 Ma model Pb of Cumming and Richards, 1975

A first analytical session (analyses 1.1 to 6.1) was abandoned due to instrumentation problems; thus the data presented here is from a second session after the problem had been rectified starts with analysis 1.2.

Table 3

Sm-Nd concentration and isotopic data

sample	Sm(ppm)	Nd(ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	2 σ	$^{143}\text{Nd}/^{144}\text{Nd}$	2 σ	T _{DM} (Ga)	T _{CHUR} (Ga)	$\epsilon_{\text{Nd}}(0)$	$\epsilon_{\text{Nd}}(1.87)$	±
483634	6.812	38.92	0.1062	0.0005	0.511291	0.000007	2.62	2.26	-26.3	-4.6	0.3
483634-2	6.288	36.18	0.1054	0.0005	0.511281	0.000008	2.62	2.26	-26.5	-4.6	0.3
483639	19.51	150.2	0.0788	0.0004	0.510924	0.000007	2.50	2.21	-33.4	-5.2	0.2
483641	10.025	58.61	0.1037	0.0005	0.511253	0.000006	2.62	2.26	-27.0	-4.7	0.3
483641-2	10.51	63.44	0.1005	0.0005	0.511234	0.000007	2.57	2.22	-27.4	-4.3	0.3
483643	6.338	33.54	0.1146	0.0006	0.511408	0.000009	2.66	2.27	-24.0	-4.3	0.3
483634:											
biotite	1.111	6.533	0.1031	0.0005	0.511271	0.000017	2.58	2.22	-26.7	-4.2	0.5
feldspar	1.452	11.50	0.0766	0.0004	0.510936	0.000008	2.45	2.15	-33.2	-4.4	0.3
garnet	8.868	7.671	0.7023	0.0035	0.518332	0.000006	1.61	1.71	111.1	-10.3	1.0
garnet (leached)	8.011	3.672	1.328	0.007	0.525761	0.000006	1.72	1.76	256.0	-15.6	1.7
orthopyroxene	5.644	33.01	0.1037	0.0005	0.511240	0.000008	2.63	2.28	-27.3	-5.0	0.3
483641:											
biotite	0.9939	7.145	0.0844	0.0004	0.511047	0.000008	2.47	2.15	-31.0	-4.1	0.3
feldspar	1.206	9.065	0.0807	0.0004	0.510984	0.000009	2.47	2.16	-32.3	-4.4	0.3
garnet	9.669	10.56	0.5558	0.0028	0.516682	0.000006	1.57	1.71	78.9	-7.3	0.8
garnet (leached)	8.955	10.06	0.5407	0.0027	0.516330	0.000009	1.48	1.63	72.0	-10.6	0.8
orthopyroxene	3.537	23.04	0.0931	0.0005	0.511129	0.000006	2.54	2.21	-29.4	-4.6	0.2

Sample numbers refer to the files of the Geological Survey of Denmark and Greenland. Errors on $^{147}\text{Sm}/^{144}\text{Nd}$ are ca. 0.5%. The depleted mantle model age (TDM) assumes the model of DePaolo (1981).

Table 4

Lu-Hf concentration and isotopic data

sample	Lu (ppm)	Hf (ppm)	$^{176}\text{Lu}/^{177}\text{Hf}$	2σ	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	T_{CHUR} (Ga)	$\epsilon_{\text{Hf}}(0)$	$\epsilon_{\text{Hf}}(1.87)$	$\pm$
483634	0.3656	5.304	0.009787	0.00005	0.281791	0.000005	2.2	-34.7	-5.3	0.2
483634-2	0.3550	5.195	0.009703	0.00005	0.281780	0.000006	2.2	-35.1	-5.6	0.3
483639	0.1621	11.00	0.002093	0.00001	0.281487	0.000005	2.2	-45.4	-6.4	0.2
483641	0.1340	7.220	0.002635	0.00001	0.281499	0.000006	1.9	-45.0	-6.7	0.2
483643	0.3049	7.239	0.005980	0.00003	0.281662	0.000005	2.1	-39.3	-5.1	0.2
483634:										
garnet	6.517	0.7493	1.242	0.006	0.324894	0.000011	1.9	1477.2	25.9	8.2
garnet (leached)	6.471	1.128	0.8144	0.0041	0.310881	0.000016	1.9	994.1	13.6	5.7
ilmenite	0	6.692	0	0.00005	0.281506	0.000016	2.0	-44.8	-3.2	0.6
orthopyroxene	0.1322	4.305	0.004360	0.00002	0.281644	0.000010	2.1	-39.9	-3.7	0.4
483641:										
biotite	0.02081	1.010	0.002925	0.00001	0.281593	0.000028	2.0	-41.7	-3.7	1.0
garnet	7.232	1.373	0.8623	0.0043	0.311245	0.000012	1.8	1006.9	-33.8	5.9
garnet (leached)	6.656	1.539	0.6138	0.0031	0.303280	0.000049	1.9	725.2	-3.5	5.6
ilmenite	0	4.430	0	0.00001	0.281466	0.000011	2.1	-46.2	-4.5	0.4
orthopyroxene	0.1687	4.139	0.005785	0.00003	0.281682	0.000006	2.1	-38.5	-4.2	0.3

Sample numbers refer to the files of the Geological Survey of Denmark and Greenland. Errors on $^{176}\text{Lu}/^{177}\text{Hf}$ are ca. 0.5%, for the ilmenite the absolut error is set to be equal to the error of the whole rock.

Table 5. Pb isotope data for sample 483641

	$^{206}\text{Pb}/^{204}\text{Pb}$	2s%	$^{207}\text{Pb}/^{204}\text{Pb}$	$\pm 2\text{s}\%$	$^{208}\text{Pb}/^{204}\text{Pb}$	2s%
Feldspar	15.723	0.015	15.304	0.015	35.393	0.015
Biotite	16.421	0.020	15.383	0.020	35.704	0.020
Orthopyroxene	26.092	0.020	16.444	0.020	45.970	0.020
Magnetite	35.036	0.200	17.402	0.200	71.217	0.200
Garnet	63.448	0.300	20.550	0.300	49.360	0.300

Table 6.

Rb-Sr isotope data for sample 483634

	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	2σ (%)	$^{87}\text{Sr}/^{86}\text{Sr}$	2σ (%)
Biotite	1918	25.66	422.6	0.342	10.46	0.004
Plagioclase	50.92	88.66	1.668	0.791	0.7497	0.007