GEOCHRONOLOGY AND ISOTOPE GEOCHEMISTRY OF THE GARDAR ALKALINE COMPLEXES, SOUTH GREENLAND

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DECLARATION

I hereby declare that the material presented in this thesis is the original work of the author, except where stated to the contrary.

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ABSTRACT

Rb-Sr studies of the major alkaline igneous complexes within the Precambrian Gardar Province of south Greenland indicate that the magmatic activity is divisible into three phases, referred to as Early (∿1300 m.y.), Mid (∿1250 m.y.) and Late (∿1165 m.y.) Gardar, respectively. Initial $^{87}Sr/^{86}Sr$ ratios for the majority of the Gardar magmas were low (0.702 - 0.704) suggesting derivation from a primitive (mantle) source; some complexes have moderately elevated ratios (0.704 - 0.707) attributed to varying degrees of contamination by older granitic basement. Two complexes (Ilimaussaq and Ivigtut), both highly differentiated and having high volatile contents, had high (∿0.707) initial ratios that cannot be explained by such bulk assimilation mechanisms. For these, it is suggested that *preferential* leaching of ^{87}Sr from the surrounding crust before and during intrusion can account for the elevated values.

Rb-Sr investigations of the Red Wine (Labrador) and Norra Kärr (Sweden) agpaitic complexes show that the former (∿1350 m.y.) may well have been emplaced during the Early Gardar Period but the latter is considerably older (∿1580 m.y.). Both have been affected by the Grenville metamorphism.

Rare earth element data for the Tugtutôq younger giant dyke complex of the Gardar Province support previous suggestions that anorthosite inclusions found in the gabbro dykes represent early fractionates disrupted and carried upward by the gabbro magmas. Rare earth and other evidence suggests that the anorthosites may be related to the development of peralkaline magmas at Ilimaussaq.

1. INTRODUCTION

The Precambrian Gardar igneous province of south Greenland comprises sediments, lavas and a large variety of alkaline intrusions forming a belt about 200 km long (in the E-W direction) by 80 km, bounded by latitudes 60°30' and 61°30' and longitudes 44°50' and 49°. Popular interest in the area, which dates from the first investigations by Giesecke between 1806 and 1813 (Giesecke, 1910) has been largely centred on the two most exotic intrusions: Ivigtut, in the west of the province, is the site of a small granite pipe containing a substantial cryolite body which was, until quarrying ceased in 1962, the world's only cryolite mine; the Ilímaussaq complex, towards the east, contains a remarkable layered series of agpaitic syenites, highly enriched in rare elements such as Zr, Nb, U, Th and the rare earths, and has recently been subjected to reconnaissance mining by a Swedish company interested in the high uranium content. Besides these unusual occurrences, there are many more central-type intrusions (Emeleus and Upton, 1976, in their general review of Gardar geology, have listed some 10 major complexes) and, in addition, a wide variety of dykes and lava flows. Rock compositions range from gabbros and nepheline-rich syenites to highly siliceous granites and include both silica over- and under-saturated syenogabbros and syenites and their volcanic equivalents.

Due to the intense interest in this area and its relative accessibility, most of the central complexes and their surroundings have been thoroughly mapped, and there have been many geochronological determinations, mostly K-Ar and Rb-Sr mineral (single sample) analyses. Bridgwater (1965) summarised the geochronological data available at that time and these showed a scatter of ages suggesting that the Gardar Period represented more or less continuous magmatism during the interval 1400-1000 m.y.

before present. Recently, however, a systematic Rb-Sr isochron study on selected complexes within the Gardar Province was initiated, and the first results of this investigation (van Breemen and Upton, 1972) have thrown doubt on earlier conclusions. The desirability of a comprehensive and precise geochronological study within the Gardar was clearly evident, and the province provided an excellent locality for such a detailed Rb-Sr isochron study; good exposures with well-defined and documented field relationships and a fairly clear relative chronology based on a complex series of dyke and fault intersections made this a challenging area. In addition to the age considerations, initial $^{87}Sr/^{86}Sr$ ratio data were needed to help elucidate the nature of the origin and development of the central complexes and to help understand the genetic relationships between such chemically diverse intrusions. This thesis presents Rb-Sr isochron data on 13 of the main central complexes in the Gardar and, additionally, initial $^{87}Sr/^{86}Sr$ ratio data on two further complexes. Some of the samples for this investigation were collected by the author during a field expedition to south Greenland, led by Dr. B.G.J.Upton in the summer of 1974, but the size and nature of the terrain necessitated obtaining most of the samples from the many geologists who have, over the past several years, been engaged in mapping the area.

In south-central Sweden and in central Labrador, there are agpaitic complexes (the Norra Kärr and Red Wine complexes, respectively) which share chemical and petrographic similarities with the Ilimaussaq intrusion in the Gardar. In view of these similarities, and also of the location of all three complexes close to the presumed Grenville metamorphic "front", samples from Norra Kärr and Red Wine (collected by the author and Dr. L.Curtis of the University of Toronto, respectively) were also investigated.

Finally, rare earth element data (determined by instrumental neutron activation analysis) are reported for some Gardar rocks. These analyses are limited to samples of anorthosite xenoliths occurring in gabbroic and syenogabbroic dykes in the Tugtutôq (eastern Gardar) region, and to the host gabbros and syenogabbros. The nature and composition of the anorthosites and host-rocks has led to the inference that plagioclase fractionation was important during the development of the alkaline Gardar magmatism (Bridgwater and Harry, 1968; Upton, 1974) and the rare earth analyses (together with initial $^{87}Sr/^{86}Sr$ ratio determinations by Mr. P.J.Patchett) were made to further examine this suggestion.

Much of the work reported in this thesis has been published or is the subject of forthcoming papers. These publications, contained in the list of references, include Blaxland and van Breemen (1974), Blaxland and Parsons (1975), Blaxland et al. (1976), Blaxland (1976), Blaxland (1977), Blaxland et al. (in press), Patchett et al. (in press) and Blaxland and Upton (in preparation).

2. GEOCHRONOLOGICAL TECHNIQUES

Introduction

Geochronological data presented in this thesis consist of Rb-Sr whole-rock isochrons for 13 Gardar central complexes (Kûngnât, Ivigtut, Grønnedal-Ika, Nunarssuit (Alángorssuaq Gabbro, Helene Granite, Biotite Granite, Nunarssuit Syenite), Ilimaussaq, Igaliko (Motzfeldt, North Qôroq, South Qôroq, Late Igdlerfigssalik) and Klokken), and on the Norra Kärr and Red Wine complexes, Rb-Sr initial ratio data on whole-rock and mineral samples from other Gardar intrusions (Ivigtut, Narssaq ultramafic bodies and the Narssaq granite-syenite complex), Rb-Sr mineral isochrons for Norra Kärr and Red Wine and additional K-Ar analyses on two mineral separates from the Norra Kärr samples. All the Rb-Sr work was carried out by the author at the Scottish Universities Research and Reactor Centre in East Kilbride; the two K-Ar analyses were kindly provided by Dr. R.M.Macintyre of the same laboratory.

A note on fundamental theory of Rb-Sr geochronology

Natural Rb is composed of two isotopes, ^{85}Rb and ^{87}Rb. The latter, which amounts to approximately 27.83% of total Rb, is radioactive, producing ^{87}Sr by β- decay, the half-life for this process being in the region of 5 x 10^{10} y. The low energy of this β- decay (275 KeV) has made accurate measurement of the half-life, and hence the decay constant ($\lambda = 0.693/t_{\frac{1}{2}}$), extremely difficult, and direct attempts during the last 20 years to determine the value have produced results ranging from 4.2 x 10^{10} to 6.2 x 10^{10} y. Recently, geologists have used either of two values; one (4.7 x 10^{10} y) obtained by direct counting (Flynn and Glendenin, 1959) and the other (5.0 x

10^{10} y) obtained by geological comparison with reliable concordant U-Pb ages (Aldrich et al., 1956). At the 1974 Paris symposium on the "Status of the Decay Constants" it was suggested (Armstrong, 1974) that there should be simultaneous adoption of a new set of mutually consistent decay constants for U, Th, Rb and K, and it now seems likely that this will occur after a consensus is reached by the major geochronology laboratories, possibly at the International Geological Congress in Australia in 1976. Current opinion seems to favour the adoption of 1.42 or 1.43 x 10^{-11} y^{-1} for $\lambda_{^{87}Rb}$.

From a consideration of half-lives alone, it is clear that the Rb-Sr method does not provide an accurate age but, provided that geochronologists agree on a set of constant values (however arbitrarily determined), techniques exist to determine precise points on an internationally agreed chronological scale. For the purposes of this study, the 5.0 x 10^{10} y half-life will be adopted (corresponding to $\lambda_{^{87}Rb} = 1.39 \times 10^{-11}$ y^{-1}), as this is currently the most widely accepted value. It should be noted, however, that some values already reported in the literature (for example, by Bridgwater, 1965, on the Gardar; Wanless and Loveridge, in press, on the Seal Lake region around the Red Wine complexes) are based on the shorter half-life ($\lambda = 1.47 \times 10^{-11}$ y^{-1}), corresponding to 6% younger ages.

^{87}Sr is one of four naturally occurring stable isotopes of Sr (in order of abundance, 88, 86, 87, 84), and it is the only one produced by radioactive decay of another element. Geochemically, Rb and Sr behave differently, Rb following the behaviour of K and tending to concentrate in the upper crust while Sr follows Ca and concentrates in rocks of intermediate composition, more characteristic of the lower crust. If an igneous rock (say a granite) forms by some differentiating process from the mantle, the Sr inherited by the rock finds itself in an environment with a much higher Rb/Sr ratio than it has hitherto experienced, and so its radiogenic ^{87}Sr content (which has stayed relatively constant up to this point) increases rapidly. If the original amount of ^{87}Sr is known ($^{87}Sr/^{86}Sr$ is often in the region of 0.70 for mantle material) then the time of formation of a high-Rb rock can be determined from the equation:

$$\left(\frac{^{87}Sr}{^{86}Sr}\right) = \left(\frac{^{87}Sr}{^{86}Sr}\right)_{orig} + \left(\frac{^{87}Rb}{^{86}Sr}\right)(e^{\lambda t} - 1)$$

An original $^{87}Sr/^{86}Sr$ ratio can only be assumed if the Rb/Sr ratio of the rock (or mineral) and/or the age is sufficiently large to yield an $^{87}Sr/^{86}Sr$ increase which is large compared to the uncertainty in the initial $^{87}Sr/^{86}Sr$ value. This is generally only the case for Rb(K)-rich, Sr(Ca)-poor minerals such as the micas and K-feldspars. As Rb and Sr systems within single mineral grains are susceptible to isotopic exchange under temperature conditions far below crystallisation, minerals are not always reliable in establishing ages of igneous intrusions. Until recently, however, mineral ages constituted the only Rb-Sr data available on the Gardar complexes (Moorbath et al., 1960; Moorbath and Pauly, 1962).

The whole-rock isochron method provides a more precise approach, and, in addition, a means of recognising isotopic anomalies within an intrusion. At the time of crystallisation of a magma, Sr will, it is (probably correctly) assumed, be isotopically homogeneous, and therefore the $^{87}Sr/^{86}Sr$ ratio will be the same in all phases of the rock. However, because of the different chemical behaviour of Rb and Sr, the $^{87}Rb/^{86}Sr$ ratio will vary. With increasing time, those parts of the rock having higher $^{87}Rb/^{86}Sr$ values will also develop higher $^{87}Sr/^{86}Sr$ values according to the above equation. This is an equation for a straight line, with $(^{87}Sr/^{86}Sr)_{orig}$ as the intercept and $(e^{\lambda t} - 1)$ as the slope. A plot of $^{87}Rb/^{86}Sr$ against $^{87}Sr/^{86}Sr$ for a series of contemporaneous rock samples should, then, yield a straight line ("isochron"), the slope of which is related to age and the intercept of which provides the original (initial) $^{87}Sr/^{86}Sr$ ratio. The following sections outline the application of the method to Gardar and other related rocks.

Sample collection and preparation

The present author did not visit south Greenland until the summer of 1974, well after the current investigation had commenced, and so much of the material used in this study was supplied by other geologists who have been working in the region. Also, of course, the Gardar Province is a large area, and in places presents difficult terrain, so it was only possible to visit personally a small number of intrusions, and even then time did not always permit systematic collection. Of the samples supplied by other workers, some (for example, those from Nunarssuit and Ilimaussaq) were collected specifically for the project, whereas others (Kûngnât, Grønnedal-Ika, Igaliko, Klokken) were selected by the author, in consultation with the collector, from quite extensive systematic collections made during field mapping. The following list indicates the original collectors of the samples used in this thesis:

Kûngnât -	B.G.J.Upton (Edinburgh)
Ivigtut -	A.B.Blaxland (E.Kilbride)
Grønnedal-Ika -	C.H.Emeleus (Durham)
Nunarssuit -	J.G.Anderson (Aberdeen)
Narssaq ultramafics -	B.G.J.Upton (Edinburgh)
Narssaq complex -	A.B.Blaxland (E.Kilbride)
Ilimaussaq -	A.Steenfelt (Copenhagen)
Igaliko -	C.H.Emeleus (Durham)
Norra Kärr (Sweden) -	A.B.Blaxland (E.Kilbride)

Red Wine (Labrador) - B.G.J.Upton (Edinburgh) &
L.Curtis (Toronto)

Criteria for the selection of samples for isotopic study were as follows:

(a) samples should, generally, be representative of the volume of rock from which they were taken.

(b) xenolithic material, areas likely to be contaminated by country rock, material from post-magmatic dyke-rocks, veins etc. should be avoided.

(c) Weathered samples, or rocks showing other signs of alteration (including rocks believed to have been subjected to recrystallisation) should be avoided.

(d) Samples should be chosen so as to provide as wide variety as possible of Rb/Sr ratios, within the limitations set out in a-c above.

For whole-rock analyses, the sample preparation procedure was as follows. Selected samples were trimmed (to remove any weathered or contaminated surfaces) and ground to below approximately 150 mesh (samples were never sieved) in a steel jaw-crusher and Tema mill. Ideally, 250-500 g of crushed powder was prepared by this method (although in cases of very coarse grained rocks very much more "coarse crush" was initially processed and a split is taken for the Tema). Representative splits of about 20 g of this were taken by coning and quartering or central coring, and this is used for all future analytical work.

Mineral separation procedures

Rock samples from which individual mineral separates were required were briefly crushed in the Tema mill, washed and decanted in tap-water to remove rock flour and then sieved into appropriate size fractions using clean nylon mesh (sizes 60-200 μ). After further washing in "Analar" (British Drug Houses reagent grade) acetone, the minerals were separated by conventional heavy liquid (tetra-bromoethane and methylene iodide) and magnetic (Frantz Isodynamic magnetic separator) techniques. After separation, samples were carefully washed several times in Analar acetone.

X-ray fluorescence analysis

Under normal circumstances, splits of all whole-rock powders and mineral separates were analysed by X-ray fluorescence for Rb and Sr to enable the most suitable selection of Rb/Sr ratios for isochron study and to facilitate satisfactory sample/"spike" ratios during isotope dilution analysis (ideally, the $^{86}Sr/^{84}Sr$ and $^{87}Rb/^{85}Rb$ ratios of the sample-spike mixtures should be close to unity). In some cases, XRF data were already available for samples collected by other geologists, but for most samples XRF analyses were carried out by the author at Edinburgh University, using a Phillips PW 1212 fully automatic X-ray fluorescence spectrometer with a W or Mo tube and LiF analysing crystal and using a scintillation counter. Samples were analysed in the form of a 2-5 mm thick layer of powder on Mylar film, and machine drift was monitored using an artificial high-Rb-Sr powdered mix which was inserted as every fourth sample. Rb and Sr concentrations in each sample were calculated from the peak/background intensity ratio from a graph of peak/background ratio versus concentration for 5 rock standards covering a suitable range of Rb and Sr concentrations and for which accurate isotope dilution analyses were available. XRF data obtained by this method were generally found to be accurate to within ± 10% for concentrations above 20 ppm.

Chemical procedures for isotope dilution analysis

During the course of this investigation, the chemical procedures were modified several times in order to improve speed and lessen the risk of contamination. The following description relates to the method used for the majority of samples, and is the procedure currently in use.

Samples are processed in batches of up to 5, each batch containing a limited range of Rb/Sr ratios (so as to minimise contamination risks). Splits of 100-500 mg (depending on Rb and Sr concentration) are taken, these being weighed (to 0.01 mg) directly into a Teflon beaker* containing aliquots of ^{87}Rb and ^{84}Sr enriched ("spike") solutions. Spikes used are Oak Ridge (US) National Laboratory 99.2% ^{87}Rb and 99.783% ^{84}Sr; the spike solutions are calibrated for concentration and composition against US National Bureau of Standards stoichiometric RbCl (SRM 984) and $SrCO_3$ (SRM 987). Double-spiked Teflon beakers are prepared in advance in batches of about 80 so as to provide for an appropriate range of Rb/Sr values.

The sample is dissolved in "Suprapur" (Merck reagent grade) HF and either Suprapur $HClO_4$ or twice distilled Analar HNO_3. After evaporation to dryness, the residue is dissolved in 6N HCl and again taken to dryness. The sample is then taken up in a minimum quantity of 2.5N HCl and eluted through a cation-exchange column of Biorad AG 50W-X8 (200-400 mesh) resin. Rb and Sr cuts are collected in 50 ml Teflon or Pyrex beakers and evaporated to dryness.

All HCl used is prepared by distilling Analar HCl and diluting as required with twice distilled H_2O ("D2"). All stock solutions of 6N and

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* Static charges on Teflon and other lab-ware can be neutralised by means of a "Zerostat" anti-static pistol.

Mass spectrometry

Strontium is loaded, in the form of a phosphate, onto a single tantalum filament bead; the Sr is taken up in a small drop of 10% H_3PO_4 (containing a small amount of HCl to leach out the Sr from any resin that may have come through the column), placed on the filament and dried down by means of an electric current. Rubidium is loaded, as a chloride, onto one side filament of a triple-filament assembly.

Mass analysis is carried out on a modified AEI GEC MS12 solid source mass spectrometer, with a 12" radius and 90° sector, using a Faraday collecting cup and Cary 401 vibrating reed electrometer with a 10^{11} ohm resistor. The instrument is linked to a Data General "Nova" on-line computer and tele-printer.

For Sr analyses, 5 peaks are measured (in order, 84, 85, 86, 87, 88) using automatic peak-switching with two-second count and integration times.

Following an initial "zero" base-line measurement of 10 readings, the sequence of peaks is measured 10 times to form a "set" of data. The computer then immediately calculates, allowing for relevant decay characteristics of the Cary vibrating reed electrometer, the mean 86/84, 88/86 and 87/86 ratios, together with the appropriate standard deviation, standard error and percentage error (PE). Ratios having excessive PE values (normally greater than 0.02) are then discarded. Typically, three or more sets of data with acceptable PE values are taken from each sample load, at the end of which the computer re-tabulates the ratio data, correcting for spectrometer fractionation. (During heating of the sample in the spectrometer, there is a tendency for light isotopes to boil off more readily than heavier ones and, although this effect is generally much lower than 0.5% per mass unit, it critically affects the precise measurement of isotopic ratios, and so values are corrected so that all 88/86 ratios are equal to the first recorded value.) At this stage, the programme allows for correction of the 87/86 values for possible Rb contamination (estimated from the net 85 peak, if any). In most cases, the Rb has boiled off at a lower temperature than that used to run Sr, and so this correction is unnecessary.

As there are only two natural isotopes of Rb (85 and 87), it is not possible to correct for machine fractionation, and care is taken during analysis not to use too high a running temperature. Again, three sets of 10 scans are generally measured, but sometimes two sets with very close agreement are considered sufficient.

Evaluation and treatment of data

Raw analytical data for each sample (sample and spike weights, isotopic composition of spikes, concentration of spikes and mean values for the measured isotopic ratios of the sample-spike mix (^{85}Rb/^{87}Rb, ^{86}Sr/^{84}Sr, ^{88}Sr/^{86}Sr, ^{87}Sr/^{86}Sr) are typed into the computer which is programmed to calculate sample Rb and Sr concentrations and isotopic compositions. All Sr isotopic ratios are corrected for laboratory (mass spectrometer) fractionation effects by assuming an (internationally accepted) value of 0.1194 for the constant ^{86}Sr/^{88}Sr ratio (Faure and Hurley, 1963).

Subsequent treatment of data depends upon the purpose for which it is intended; all treatment is related to the fundamental isochron equation (p. 6). In some cases, the age is known within fairly precise limits and the sample (usually with a low Rb/Sr ratio) is analysed only to obtain the initial ^{87}Sr/^{86}Sr ratio from the isochron equation. Alternatively, an estimate of the age may be required from a single sample (usually chosen with a high Rb/Sr ratio); in this case an estimate of the initial $^{87}Sr/^{86}Sr$ ratio is used. For most purposes, however, both age and initial ratio are required from a series of whole-rock or mineral samples and the reduced analytical data are plotted in the form of an isochron diagram ($^{87}Sr/^{86}Sr$ against $^{87}Rb/^{86}Sr$) and a best-fit line is drawn through the points, this line giving the age (slope) and initial $^{87}Sr/^{86}Sr$ ratio (intercept).

Various methods of fitting a straight line to points representing a suite of cogenetic rocks have been proposed, and Brooks et al. (1972) have critically reviewed them. All major programmes designed to deal with Rb-Sr data (York, 1966; York, 1969; Brooks et al., 1968; MacIntyre et al., 1966) use a "least-squares" regression method and calculate the best estimate of the slope and intercept together with appropriate standard deviations (based on laboratory precision estimated from previous replicate analyses) and an estimate of the actual scatter about the line regressed. It is primarily in the treatment of these error estimates that the available regression treatments vary.

First, a note on laboratory error estimates is in order. This error is usually estimated from an examination of replicate analyses (not just repeated mass spectrometer runs, but repeats involving the whole chemical procedure on separate sample splits) taken over a period of time. Criteria for selection of duplicate analyses for inclusion in laboratory precision estimates must be carefully considered. Many repeat analyses are performed on samples for which the first analysis is suspect (either the point lies off the isochron or there is some doubt about part of the laboratory procedure). Such duplicate analyses are only taken into account if the second value confirms the first. Also, positive correlation between $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ may result in different splits of imperfectly homogenised samples.

Such correlation, clearly, leads to an exaggerated difference between duplicates. This problem can be avoided to some extent in estimating laboratory precision by concentrating attention on duplicate analyses with low Rb/Sr values (Natrella, 1963, has indicated that such problems are not significant at the 95% confidence level in whole-rock samples having $^{87}Sr/^{86}Sr$ ratios below 1.0). York (1969) has taken the effect of positive correlation of errors into account in his regression programme, and allows an estimate of the correlation coefficient (r) to be used in assigning errors to individual points.

All models treat the error in $^{87}Rb/^{86}Sr$ as a percentage (proportional) error, but McIntyre et al. (1966), unlike other authors, treat $^{87}Sr/^{86}Sr$ errors as a fixed value, reasoning that the range of $^{87}Sr/^{86}Sr$ commonly encountered in routine analyses straddles the machine optimum. In practice, the choice between fixed or proportional error for $^{87}Sr/^{86}Sr$ is probably unimportant, as in high $^{87}Sr/^{86}Sr$ samples, where the difference would be significant, the error in $^{87}Sr/^{86}Sr$ is very small compared with that in $^{87}Rb/^{86}Sr$.

Provided that the estimate of laboratory precision is realistic, this sets a constraint on the accuracy with which an isochron can be determined, and if a series of points being regressed do indeed represent a cogenetic suite of samples, it can be expected that the actual scatter about the line should approximate the estimate of laboratory precision. If it significantly exceeds this value, then a true isochron does not exist; Brooks et al. (1972) have proposed that lines having such excess scatter should be termed "errorchrons". The excess scatter, referred to by McIntyre et al. (1966) as "geological" scatter, existed before the commencement of laboratory procedures, and in many cases it is not possible to detect what processes were responsible for the scatter. All regression models provide an estimate of the actual scatter of points about the isochron as compared to the estimated laboratory precision. McIntyre et al. (1966) calculate a value, the Mean Square of Weighted Deviates (MSWD) which, if actual scatter exactly corresponds to estimated scatter, is equally to unity. York (1969) calculated a value equivalent to MSWD, referred to as $[SUMS/(n-2)]^{\frac{1}{2}}$. The cut-off point between isochron and errorchron must be made by each laboratory based on the degree of confidence with which the analytical precision is known. In this study, a value of MSWD or $[SUMS/(n-2)]^{\frac{1}{2}}$ is considered appropriate (see Brooks et al., 1972, for a discussion of this topic).[1]

McIntyre et al. (1966) are the only authors who have attempted to proceed with errorchron models and recalculate the best-fit line. One such model assumes that all excess scatter is distributed proportional to $^{87}Rb/^{86}Sr$ (the sort of scatter that would result from metamorphic overprinting) and another model assumes that the error is distributed independent of $^{87}Rb/^{86}Sr$ (this would apply where small differences in the initial ratio existed). However, Brooks et al. (1972) show that these geological error models may be of limited value.

For the purposes of this study, York's (1969) model has been adopted. This model is one of the most widely used for Rb-Sr isochron work, and its small size enables it to be used on the Nova computer. A programme for the McIntyre et al. (1966) regression treatment is available on the main Univac 1108 computer at the National Engineering Laboratory site, for reference and comparison purposes.

[1] See Addendum, p 212

3. RESULTS OF BLANK, STANDARD AND REPLICATE ANALYSES

All reagents used in the chemical procedure are analysed for Rb and Sr content to ensure that no significant contamination is added to the samples. Periodically, a "total blank" is measured for the whole analytical procedure by including an empty platinum dish or Teflon beaker to which a few drops of spike solutions have been added, and processing it together with sample-loaded beakers. Total blanks are typically about 5 ng for both Rb and Sr, but higher blanks were occasionally encountered. In most cases, such high blanks were traceable to imperfectly washed ion-exchange columns, and consequently this part of the procedure has been carefully reviewed. In no case was it considered necessary to apply blank corrections to sample analyses.

In order to assess intra-laboratory variations in values resulting from modification to calibrations, improvements in measuring techniques, etc., an internal standard (GGU 86035) is periodically analysed. This sample is taken from the top of a precisely defined isochron (van Breemen and Upton, 1972). Since the time of first analysis of this standard, new NBS stoichiometric salts RbCl (SRM 984) and $SrCO_3$ (SRM 987) have been adopted for more accurate spike calibrations (de Laeter et al., 1973, discuss such calibration changes) and improved estimates of the Cary 401 vibrating reed electrometer response characteristics have been incorporated into the computer programme. Table 1 lists 5 sets of duplicate analyses made on 86035 since 1970, and Fig. 1 shows that the later analyses would tend to yield lower ages. For intra-laboratory consistency, it has been agreed that all reported ages made before adoption of the NBS calibration standards (represented by the first duplicate analysis of 86035) should be reduced by 1% in order to render them equivalent to the data now being produced, and Fig. 1 illustrates the effect of this. (de Laeter et al., 1973, have reduced their ages by 1.8% for similar reasons). The only direct effect of this on the present study is to reduce the value of ages published for two Gardar intrusions, the Hviddal Dyke and Tugtutôq Central Complex (van Breemen and Upton, 1972) by 1%.

In order to assess inter-laboratory differences in reported mass spectrometer measurements, an internationally agreed standard (Eimer and Amend, MIT, standard $SrCO_3$) is analysed at regular intervals. During the course of this investigation, 38 such analyses were made (by Mr. J.Hutchinson) giving an average value of 0.70811 ± 0.00007 (1σ) for $^{87}Sr/^{86}Sr$ (Table 2). Each analysis is based on 10 sets of 10 peak-height scans. It is customary to adjust values of $^{87}Sr/^{86}Sr$ to a Eimer and Amend value of 0.7080 in order to compare values reported by different laboratories. Additionally, a standard feldspar (NBS SRM 607) is analysed periodically to monitor $^{87}Sr/^{86}Sr$ and Rb/Sr ratio measurement accuracy; values are listed in Table 3.

Table 4 lists 16 repeat analyses made routinely during the course of this present study. Repeats involving suspect runs are not included. Also listed are the percentage variations on $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ values, that is $100(x_{max} - x_{min})/x_{min}$. As can be seen from the relevant isochron diagrams, many of the repeat values show positive correlation between $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ (resulting from sample inhomogeneity) so the percentage variation shown is greater than that attributable directly to laboratory error. Also, it is necessary to note that sometimes highly exaggerated errors (for example 10.3% for 140053 $^{87}Rb/^{86}Sr$) may result from uncorrectable machine fractionation differences on duplicate analyses of substantially over-spiked, high-Sr, low-Rb samples. From approximately 50 replicate analyses available (from this and previous studies), error estimates (1σ) of 0.5% on $^{87}Rb/^{86}Sr$ and 0.05% on $^{87}Sr/^{86}Sr$ are used for routine regression treatment of isochron data in this study.

4. THE GARDAR PROVINCE OF SOUTH GREENLAND

Introduction

The Gardar Province of south Greenland has been fairly extensively studied, especially since the early 1950's, primarily as part of the mapping programme of the Grønlands Geologiske Undersøgelse (GGU), and recent comprehensive summaries of the general geology have been published by Berthelsen and Noe-Nygaard (1965), Upton (1974) and Emeleus and Upton (1976). A much simplified geological map of the Gardar Province is illustrated in Fig. 2; comprehensive 1:100000 geological map sheets of the region are published by the GGU, those of most relevance to the Gardar being 61 V.1 Syd (Ivigtut), 60 V.1 Nord (Nunarssuit), 60 V.2 Nord (Julianehåb) and 61 V.3 Syd (Narssarssuaq). More detailed maps of the individual complexes investigated here are presented in Figs. 3, 5, 7, 9, 11, 13, 15 and 17, and details of the geology relevant to this study are discussed in conjunction with the analytical data. The following is intended only as a brief description of the province.

Rocks assigned to the Gardar Period occur largely within a belt of heterogeneous Julianehåb Granite (named after the largest town in the region) basement, produced during the Ketilidian orogeny (∿1800 m.y. according to van Breemen et al., 1974). In the extreme north-west of the province, however, the basement rocks are of older, Archaean, gneisses. Berthelsen and Henriksen (1975) show various K-Ar and Rb-Sr mineral ages for these older rocks, but all reflect later periods of metamorphism; however, Kanasewich and Slawson (1964) and Ulrych (1964), on the basis of common Pb analytical data from Ivigtut galenas, concluded that the crust in this region is probably about 3000 m.y. old.

Gardar supracrustal rocks

The oldest rocks of the Gardar Period are probably the interbedded continental sandstones, basaltic lavas (mainly alkali olivine basalt and hawaiite) and pyroclastic rocks of the Eriksfjord Formation (Stewart, 1964; Poulsen, 1964; Watt, 1966; Larsen, 1973). Much of this early supracrustal sequence has been removed by erosion, and outcrops are now confined to the eastern-most region of the province (see Fig. 2), where an accumulation of over three kilometres is preserved in the down-faulted Ilimaussaq peninsula; however, there is evidence, in the form of preserved subsided blocks in Gardar intrusions, that the Eriksfjord cover was once much more extensive.

Unfortunately, the lavas all show some degree of post-crystallisation alteration, and have not been dated radiometrically, but they are bracketed in time by the cooling of the Ketilidian rocks ($\sim$1600 m.y., according to van Breemen et al., 1974) and the beginning of the Gardar intrusive period ($\sim$1300 m.y. ago, according to this present study).

Gardar dykes

Dykes of widely varying compositions were emplaced throughout Gardar time. Probably the first are two series of lamprophyric dykes occurring in the west of the province, one near Arsuk (described by Upton, 1960) and the other (described by Emeleus, 1964) in the vicinity of Grønnedal, some 20 km to the east; both trend ENE but, according to Upton (personal communication, 1975) they are petrologically quite distinct and therefore probably not directly related. These earliest dykes were followed by larger olivine dolerite dykes (named "Brown Dykes" by Wegmann, 1938, and subsequently classified by Berthelsen, 1962, into three generations: BD_0, BD_1, and BD_2) which have a generally WNW trend. These dykes, which may be up to 150 m wide, occur over much of the Gardar Province, from the Ivigtut area in the west to the Tugtutôq area in the east-central region; possibly, they may even exist in the Narssarssuaq area in the vicinity of the Igaliko complex, in the extreme east of the province (Emeleus and Harry, 1970). From this point of view, the BD dyke series provide a useful chronological reference point in the Early Gardar. Recent Rb-Sr whole-rock and mineral analyses by Patchett (1976) suggest that many of the lamprophyric dykes and the BD_0 and BD_1 dykes were emplaced ∿1250 m.y. ago; they are thus younger than the Eriksfjord Formation.

Following the BDs, numerous generations of dykes occur in restricted areas (for example, granophyres in the Ivigtut area, microporphyritic basalts and alkaline dykes in the Grønnedal area, and possibly alkali gabbros, syenogabbros and phonolites in the Igaliko area), but none of these can be correlated over a wide area and so do not in themselves provide chronologically useful reference points. The next widespread series of dykes are the so-called "Mid-Gardar" dyke swarm, particularly dense in the Tugtutôq-Narssaq region (Upton, 1962; Upton, 1964b) and the Isortoq region (Bridgwater and Coe, 1970), but also occurring, though more sparsely, in the Qaersuarssuk area (between Tugtutôq and Isortoq) and further east in the Narssarssuaq region (Emeleus and Harry, 1970). The structure and composition of these dykes has been studied most closely in the Tugtutôq area (McDonald, 1969a; MacDonald, 1969b; McDonald and Edge, 1970; McDonald and Parker, 1970; Upton, 1962; Upton, 1964a; Upton, 1964b; Upton et al., 1971). Here, the dyke-emplacement began with the intrusion of the Hviddal "giant dyke" some 500 m wide and cropping out for some 20 km down the central axis of Tugtutôq island (Upton, 1964b). This dyke shows internal variations in composition, both from margin to centre (syenogabbro to undersaturated syenite) and laterally (augite syenite to sodalite foyaite). The Hviddal dyke is succeeded by further gabbroic giant dykes (the "younger giant dyke complex") which in turn are cut by a large number of narrower dykes including dolerite, trachyte, phonolite and rhyolite. Similar dykes, including giant dykes (over 100 m wide) and sharing the common ENE trend, are found in Isortoq, although here the time relationships are less certain - the giant dykes are not demonstrably the oldest members of the swarm and there is no direct evidence to indicate that the ENE dyke emplacement in the two regions occurred at the same time.

The majority of these ENE dykes preceded a major period of salic, central-type intrusions, but a few post-date the central complexes and represent the last recorded phase of Gardar magmatic activity.

Upton (1974) has suggested that there may be a complete progression from simple dilational dykes, through composite giant dykes, to the more salic, central-type complexes. Possible evidence of this exists at Bangs Havn (about 5 km to the north-east of the Nunarssuit complex, Fig. 2) where gabbroic giant dykes appear to coalesce towards a lenticular intrusion of gabbro, syenite and quartz syenite.

Gardar central complexes

Emplacement of salic magmas in the form of central complexes took place throughout the Gardar Province, and it is clear from field relationships alone that more than one phase of magmatism is represented. The Grønnedal-Ika complex in the west, for example, pre-dates the BD_0 dykes and all but a few of the lamprophyre dykes, and thus represents one of the earliest phases of Gardar activity. Ilimaussaq (in the east), on the other hand, cuts all but the very latest ENE dykes and has widely been regarded as probably the last of the complexes to be emplaced, very late in the magmatic cycle.

The central complexes themselves vary considerably in size and composition; Emeleus and Upton (1976) have recognised 10 major complexes, ranging from the 300 m diameter pipe-like granite intrusion at Ivigtut to the large syenite complexes of Nunarssuit and Igaliko, extending over several hundred square kilometres. Smaller intrusions do occur, however, and there may well be others hidden from view - below the present erosion level, covered by the inland ice sheet or beyond the present shore-line. (Andrews et al., 1971, report extensions of the Gardar dyke system on the east coast of Greenland, north of Kap Herluf Trolle, and it seems quite reasonable to expect that Gardar intrusions exist buried beneath the intervening inland ice cap.) The composition of the central complexes varies greatly; gabbros, syenogabbros, augite syenite, nepheline syenite, quartz syenite, granites and carbonatites are all represented. Most complexes represent multiple intrusions (larger ones, for example the Igaliko complex, are composed of a number of individual centres each comprising multiple intrusions) but each complex is restricted to modally either oversaturated or undersaturated rock-types (Ilimaussaq is a notable exception - here a small granite intrusion occurs with a larger body of highly undersaturated syenites, and its place in the intrusive sequence is a matter of some controversy).

Specific details of individual intrusions are discussed together with the relevant analytical data. Descriptions of the general geology can be found in the various field geologists' accounts published by the GGU. From west to east, the major complexes are: Kûngnât (normatively under- and oversaturated syenites with gabbro and granite; Upton, 1960; Upton et al., 1971), Ivigtut (granite with associated cryolite deposit; Berthelsen, 1962), Grønnedal-Ika (nepheline syenite and carbonatite; Emeleus, 1964), Nunarssuit (oversaturated syenites with granite and gabbro; Harry and Pulvertaft, 1963), Puklen (granite and oversaturated syenite; Parsons, 1972a), Tugtutôq Central Complex (granite and oversaturated syenite; Upton, 1962), Narssaq (granite and syenite; Ussing, 1912), Ilimaussaq (slightly undersaturated syenite and highly differentiated feldspathoidal agpaites, also granite and quartz syenite; Ferguson, 1964), Igaliko (nepheline syenite, also slightly undersaturated syenite; Emeleus and Harry, 1970) and Klokken (gabbro and oversaturated syenite; Parsons, 1972b).

Evidence for magma generation and emplacement

Most attempts to explain the diversity of rock types in the Gardar Province involve differentiation from basic magmas; alkali olivine basalt magma seems to have been parental to most of the salic Gardar complexes. Although relatively high concentrations of some trace elements found to be enriched in many Gardar intrusions (for example, Ti, Zr, Nb, U) may have been attained in the mantle by the volatile flux process suggested by Bailey (1970), these concentrations were undoubtedly further raised by strong crystal fractionation of the derivative magmas (Upton, 1974).

The smaller dykes were emplaced along dilational fractures, whereas substantial stoping was important during the intrusion of the central complexes. The relative rôles of dilation and stoping in the emplacement of the larger dykes and giant dykes is debatable; Bridgwater and Coe (1970) argue that the large dykes in the Isortoq area were emplaced by stoping and do not show evidence of dilation, but xenoliths of acidic basement rocks in these basic dykes are rare, and at least some dilation must have occurred (Emeleus and Upton, 1976). Upton (1974) and Upton (in preparation) has estimated a total extension of about three kilometres across the Gardar Province associated with ENE dyke emplacement. Evidence for substantial *in situ* differentiation exists in many Gardar intrusions; marked mineral layering occurs, especially in Kûngnât (Upton, 1960), Nunarssuit (Harry and Pulvertaft, 1963), Grønnedal-Ika (Emeleus, 1964), the younger giant dykes on Tutgtutôq (Upton, 1962, 1964b), Ilimaussaq (for example, Ferguson, 1964; Ferguson, 1970a; Ferguson, 1970b; Bohse et al., 1971) and Klokken (Parsons, 1972b), together with cryptic variations; in other cases, however, (the various centres of the Igaliko complex are an example) there are sharp breaks between units, indicating multi-phase intrusion, with differentiation taking place at sub-emplacement levels. Although in most cases such units reflect a sequence progressing from less to more differentiated magmas, there are examples of the reverse sequence obtaining (for example, at Kûngnât, where a late gabbro phase follows the emplacement of a syenite complex. This has suggested the possibility that liquid fractionation may have been important during the development of some of the Gardar magmas (Bridgwater and Harry, 1968).

Abundant anorthosite inclusions occur in many of the Gardar intrusions, particularly in some of the doleritic and trachydoleritic ENE dykes (called "Big Feldspar Dykes" - BFDs, because of these inclusions). Such dykes are common over much of the province, and have been described in detail by Bridgwater (1967) and Bridgwater and Harry (1968). A sympathetic chemical relationship between dyke host-rock and anorthosite has led Bridgwater to conclude that the anorthosites were derived from a disrupted anorthosite body or bodies at depth, formed during the early stages of magma differentiation, and hence that plagioclase fractionation was important during the development of the Gardar magmas.

In addition to plagioclase fractionation, Upton (1974) explains the high Al, Fe/Mg, (K+Na)/Ca properties of the Gardar basalts in terms of variable degrees of clinopyroxene fractionation at depths between 60 and 15 km.

Gill (1973) has proposed that the compositional difference between dykes (mafic) and central complexes (salic) may be due to a density filtration process in the continental crust, a mechanism also believed to be important in other magmatic provinces (see, for example, Walker, 1975a, 1975b). According to Gill, during periods of crustal extension, dense mafic magma is admitted to the surface along tensional fissures, whereas during the intervening non-dilational periods the low density salic magma, accumulating at the top of the deep-seated magma chamber, rises to the surface while the denser material is restrained at depth. According to this model, dykes and central complexes cannot be synchronous.

Faulting and evidence of Gardar rifting

Major faulting occurred at intervals throughout the Gardar Period over the entire province, and indeed the positions of faults and their intersections are a major controlling factor of the ENE linear character of the present topography and of the siting and emplacement of the Gardar intrusions. A set of E-W left-lateral faults and an intersecting set of right-lateral N-S faults (NNE to NNW trending) show a substantial vertical displacement in addition to the predominantly horizontal movement. The aggregate left-lateral displacement across the Gardar is at least 15 km (Emeleus and Upton, 1976) and vertical movements of at least 2-3 km must have occurred in order to preserve the Eriksfjord supracrustal sequence on the Ilimaussaq peninsula.

The nature of the faulting, alkaline and mainly undersaturated volcanism and plutonism and the distribution of supracrustal material in the Gardar Province have led to the suggestion that the province may represent a deeply dissected ancient continental rift zone similar, perhaps, to the recent East African rift system (for example, Sørensen, 1970; Stewart, 1970; Bailey, 1974a; Upton, 1974; Upton, in preparation). Such rifting would have occurred during, or just preceding, the Grenville (Sveconorwegian) orogenic activity occurring on either side of the Atlantic, the northern front of which probably passed just to the south of Greenland.

Previous geochronological and isotopic studies

Until fairly recently, all available chronological data for the Gardar Province was based on (often conflicting) Rb-Sr mineral (single sample) and K-Ar determinations (see Table 5) and various intersections between faults, dykes and central complexes. Unfortunately, although such intersections occur in pleasingly large numbers, the province is a large one, and possible correlations, particularly between extreme west and east ends of the province, have been limited and tentative. The scatter of ages represented in Table 5 suggested magmatism throughout the period 1400-1000 m.y. or even (Bridgwater, 1965) commencing as early as 1500 m.y. before present. Although some of the ages (especially some K-Ar values) clearly contradict either each other or the field relationships, others have been assumed to be generally valid. Iliamussaq (with three relatively precise Rb-Sr mineral ages at 1086 ± 20 m.y.) was widely regarded as probably the youngest Gardar intrusion, whereas Grønnedal-Ika was thought to be the oldest, as it predated practically all other activity in the area, but the chronological position of some other complexes was largely a matter for speculation. For example, Emeleus and Harry (1970) proposed, on the basis of a single dyke intersection, that the Igaliko complex was almost certainly younger than the Grønnedal-Ika complex, and went on to suggest that the Igaliko syenites were older than the Narssaq and Ilimaussaq intrusions but younger than the giant dykes of Tugtutôq. Such tentative correlations were clearly unsatisfactory, and more recently van Breemen and Upton (1972) initiated a Rb-Sr isochron study on three of the central complexes (Table 5); these results not only give more reliable age data but provide initial $^{87}Sr/^{86}Sr$ ratios which in turn yield valuable information about the nature and origin of the magmas. Engell and Pedersen (1974) have reported a Rb-Sr isochron for the Bangs Havn complex (referred to earlier) and Patchett (1976) has very recently completed a Rb-Sr study on whole-rock and mineral samples from the "Early" Gardar lamprophyre and BD dykes, these indicating an age of ∿1250 m.y. (Table 5). However, these studies cover very few of the many Gardar intrusions and a much more extensive investigation was clearly needed.

In addition to the Rb-Sr and K-Ar work listed in Table 5, a number of papers relate to the unusual isotopic composition of the common Pb of galenas associated with the Ivigtut granite intrusion. Two of these papers (Kanasewich and Slawson, 1964; Ulrych, 1964) have already been mentioned, and these papers contain further bibliographies on the subject.

Gardar complexes investigated in this study

The following chapter considers the Rb-Sr data on the major syenite complexes in the Gardar Province. These include Kûngnât, Grønnedal-Ika (including data on some later dykes), Nunarssuit (data on four of the main units: Alángorssuaq Gabbro, Helene Granite, Biotite Granite and Nunarssuit Syenite), Narssaq Complex, Igaliko (data on four main intrusive centres: Motzfeldt, North Qôroq, South Qôroq and Late Igdlerfigssalik) and Klokken. The Ilimaussaq complex is considered separately in Chapter 6. The Sr isotopic systematics of the Ivigtut Granite and associated cryolite body are discussed in Chapter 7, and data on the small Narssaq ultramafic bodies are treated in Chapter 8. Chapters 9 and 10 summarise and discuss the initial $^{87}Sr/^{86}Sr$ and chronological relationships in the Gardar Province. Rare earth data on samples from the Tugtutôq younger giant dyke complex are presented in Chapter 13.

5. RB-SR DATA FOR THE GARDAR PROVINCE: THE MAJOR SYENITE COMPLEXES

Introduction

There have been many Sr isotopic studies on alkaline basalts and granites (see, for example, Faure and Powell, 1972) but relatively little is known about the Rb-Sr systematics of syenites. A particular feature of the Gardar Province is that syenites form the major rock type in the central intrusions; indeed, Watt (1966) estimated that 82% of the total surface area of Gardar intrusions consist of syenite or alkali granite. Moreover, within the syenite grouping, a wide variety of rock types exist, from highly undersaturated nepheline syenites and agpaites, to oversaturated syenites, associated with generally minor amounts of gabbros, syenogabbros and granites.

This chapter presents Rb-Sr data on all the major syenite complexes of the Gardar, with the exception of Ilimaussaq. This complex stands apart not only due to its highly undersaturated, peralkaline composition and the great attention this has attracted, but also due to the rather unusual Sr isotopic relationships; it is dealt with separately in the following chapter. For convenience, the syenite complexes are considered in sequence for west to east (see Fig. 2).

Kûngnât

The Kûngnât alkaline syenite complex (Fig. 3) is situated at 61°12'N, 48°15'W, at the extreme west of the Gardar Province, and has been described by Upton (1960). The complex, about 5 x 2.5 km in size, intrudes Archaean gneisses and cuts a number of early ENE lamprophyre and dolerite dykes;

similar dykes are seen to intrude the Grønnedal-Ika complex, and thus Berthelsen (1962) considered Kûngnât to have been emplaced late in the Gardar Period. The relationship of Kûngnât to the Ivigtut Granite (midway between Kûngnât and Grønnedal-Ika) is not clear from field relationships due to lack of informative dyke intersections of the granite.

The complex at Kûngnât consists of four major intrusions (Upton, 1960). A south-western marginal syenite was followed by a larger western syenite mass (Fig. 3) which, during crystallisation, became divided into two separate magma bodies by a raft-like layer of gneiss blocks which spalled off the roof, sank and became suspended in the syenite magma. Differentiation by crystal settling (producing crystal layering, rhythmic banding and feldspar lamination) occurred both above and below this gneiss raft, and syenites range from nepheline-normative to hypersthene- and hypersthene-quartz-normative and ultimately quartz-modal types (Upton et al., 1971; Upton, personal communication, 1975). Part of this western syenite unit was subsequently removed by a later, eastern syenite; this is generally more basic than its predecessor, having higher Mg and Ca contents, but showing similar *in situ* differentiation to quartz-modal syenites. Subsequent ring-faulting was accompanied by the intrusion of a gabbroic ring-dyke approximately 100 m wide and almost exactly following the external contact of the syenite complex. That this gabbro was intimately associated with the intrusion of the syenites is confirmed by the occurrence of late-stage riebeckite-bearing granite differentiates of the syenites which re-intrude both syenites and gabbro; the gabbro was emplaced before the main syenite magma had completely crystallised.

Kûngnât was one of the complexes chosen for the original Gardar Rb-Sr isochron study by van Breemen and Upton (1972), and yielded an age and initial $^{87}Sr/^{86}Sr$ ratio of 1233 ± 16 m.y. and 0.7071 ± 0.0021, respectively (Table 5). However, a repeat investigation seemed desirable for a number of reasons. First, the isochron was based on 5 samples, two from the early, western syenite and the remaining three from the late peralkaline granite sheets; the isochron age depended heavily on these last three samples, which had high Rb/Sr ratios ($\sim$50), quite atypical of the Kûngnât body as a whole. The age of around 1233 m.y. seemed somewhat anomalous in relation to the ages of other Gardar complexes (presented in the following pages) and the possibility that these granite samples had been subjected to post-crystallisation exchange of Rb and/or Sr had to be checked. Second, the precision obtained for the initial $^{87}Sr/^{86}Sr$ ratio ($2\sigma = 0.0021$) was poorer than that for many other complexes, and permitted interpretations ranging from relatively minor crustal contamination of mantle-derived magma to alternative methods involving a large crustal component - in either case, the initial ratio was above the modal value for Gardar Syenites, and there appeared to be no good reason why this should be so. Third, the possibility of significant crustal contamination of the syenites had caused van Breemen and Upton (1972) to speculate that there might be a difference between the original Sr isotopic compositions of the syenites and the gabbro ring-dyke, and analyses of the gabbros were needed to examine this possibility.

With these considerations in mind, a dozen or so samples of the Kûngnât syenites were selected from the collections of Dr. B.G.J.Upton and these were analysed for Rb and Sr (XRF). 5 samples giving a suitable range of Rb/Sr ratios were selected, one from the south-western marginal syenite, and two each from the western and eastern units (see

Fig. 3 for sample locations). Isotope dilution data for these samples are presented in Table 6, and an isochron based on them is shown in Fig. 4. The age (1245 ± 17 m.y.) essentially confirms that reported by van Breemen and Upton (1972), but the precision on the initial $^{87}Sr/^{86}Sr$ ratio (0.7041 ± 0.0008) allows a more meaningful interpretation. A value of 1.50 for $[SUMS/(n-2)]^{\frac{1}{2}}$ indicates that the line is a true isochron.

Also shown in Table 6 are analytical data for two samples of the gabbro ring-dyke (locations given in Fig. 3); initial ratios for these (calculated using an assumed age of 1245 m.y.) are 0.7039 and 0.7036, identical (within 1σ) to that of the syenites and rather higher than the majority of the other Gardar syenites. This suggests that crustal contamination may have taken place in the magma before differentiation of the different gabbro and syenite fractions, a hypothesis consistent with the general lack (apart from the gneiss zone already mentioned) of xenoliths in the syenites. It is, of course, possible that crustal Sr was incorporated preferentially into the syenite magma and that this effect has been masked by the analytical uncertainty in the initial ratio measurements. It is very difficult to evaluate how much crustal component could be accomodated in this way, but it is probably not significant. For example, assuming a 200 ppm average concentration for Sr in both syenite and gneiss, and a $^{87}Sr/^{86}Sr$ ratio of 0.712 for the gneiss at 1245 m.y. ago (a value based on measurements on the nearest studied Archaean crust by Pidgeon and Hopgood, 1975) and assuming the worst possible 2σ errors on the initial ratio measurements for the gabbro and syenite, then a maximum of about 15% crustal Sr could have been incorporated into the syenite magma. This is still probably a lesser quantity than that postulated by Upton (1960) to account for the development of a syenitic magma at the top of a differentiating mass of alkali olivine basalt. The mechanism of deep-level fractionation of the Kûngnât magma is still a matter for speculation.

In summary, the isotopic evidence presented here, together with that of van Breemen and Upton (1972) indicates that all units (three syenites, gabbro ring-dyke and late granite differentiates) are comagmatic and probably shared the same Sr isotopic composition up to the time of emplacement.

Grønnedal-Ika

The nepheline syenite and carbonatite complex of Grønnedal-Ika (Fig. 5) lies about 20 km to the east of Kûngnât, at 61°15'N, 48°W, between the naval base of Grønnedal and the Ika Fjord. Like Kûngnât, the complex intrudes foliated quartzo-feldspathic gneisses of Archaean age, and was recognised by Berthelsen (1958) as probably the earliest Gardar event in the area; the geology has been described by Emeleus (1964). The present form of the complex is elongate, measuring about 8 x 3 km, but this shape is the result of considerable post-intrusive faulting and its original outline is believed to have been oval, measuring 6 x 3½ km.

Several distinct intrusive syenites are recognised; these are classified by Emeleus into Upper and Lower Series, both containing foyaitic cumulates, and separated by a raft-like layer of gneiss, believed to have flaked off the roof of the complex during crystallisation of the syenites (*c.f.* the western syenite unit of Kûngnât). Sheets of porphyritic microfoyaite cut syenites of both series and the gneiss raft; the Upper Series syenites are also intruded by a strongly xenolithic porphyritic syenite containing fragments of the earlier syenites, rarer trachytes and gneissic country rock. Although large fragments of gneiss occur within the Lower Series syenites (Emeleus, 1964, plate 2), there is little evidence that these have been assimilated by and contaminated the syenites; where exposed, the contacts between gneiss and syenite are sharp and the degree of metamorphism very slight, whether within the smaller inclusions, the large raft or in the gneisses at the steep, outward dipping external contacts of the complex.

The carbonatites of Grønnedal-Ika, consisting principally of calcite and siderite (altered to magnetite next to later dolerite dykes) with minor apatite, sphalerite, pyrite, barite and other minerals, forcibly intruded the already solidified foyaites. Both adjoining syenites and the syenite inclusions occurring in the carbonatite show signs of extensive metasomatic alteration.

The complex is intruded by a wide variety of lamprophyric, doleritic, and oversaturated to undersaturated alkaline "trachytic" dykes. Apart from some lamprophyres (a lamprophyre xenolith was found by Emeleus in a sample from the outer syenite) all dykes post-date Grønnedal-Ika, although most pre-date the three recognised episodes of faulting which later deformed the complex. Larsen (1969) obtained K-Ar ages of 1210 ± 30 and 1170 ± 30 m.y. on two samples of nepheline syenites from the complex, but these data conflict with the more reliable age of ∿1240 m.y. for Kûngnât, which is known to cut the post-Grønnedal-Ika dykes.

For the isochron study, 11 samples of syenite and one sample of carbonatite (locations shown in Fig. 5) were selected from collections made by Dr. C.H.Emeleus. Data for these samples are listed in Table 7, and the isochron, based on all but one of the samples, is shown in Fig. 6. Syenite sample 39763A clearly falls off the isochron and is therefore not included in the regression treatment; a value of 1.26 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ for the rest of the samples indicates a true isochron. The age (1327 ± 27 m.y.) is the oldest isochron date so far reported for the Gardar Province. The low initial $^{87}Sr/^{86}Sr$ ratio (0.7032 ± 0.0004) accords well with field evidence for minimal crustal contamination.

Significantly, the initial $^{87}Sr/^{86}Sr$ ratio of the carbonatite sample (31863) is 0.7032, identical to that of the foyaites, and permits comment on the possible rôle of carbonatite magma in the genesis of alkaline rocks. On the basis of Sr isotopic data, Powell and Bell (1974) discuss the possibility of producing alkaline rocks by the mixing of two rock types, one of sialic and the other of carbonatitic composition. If mixing of the two end-members took place at Grønnedal-Ika, the resulting mix (foyaites) would have an isotopic composition intermediate between one extreme (the hypothetical "sialic" component) and the carbonatite. The identical composition of both foyaite and carbonatite does not support this hypothesis as a likely mechanism for the generation of alkaline magmas in the Gardar; it suggests, rather, a common source for both foyaites and carbonatites.

Also, four samples from the late alkaline dykes were analysed, in the hope that these would yield the age of dyke emplacement and faulting. Data for these samples are also listed in Table 7 and clearly show that the samples must have been affected by post-intrusive migration of Rb and/or Sr (ages for these samples, based on a typical "Gardar" initial ratio of 0.703, would range from 1115 to 1381 m.y.). The dykes are actually quite badly jointed, and even those few that are more than three or four metres wide show the effects of weathering along the joint surfaces; in view of this, the scatter of apparent ages is not surprising.

Nunarssuit - Introduction

The Nunarssuit complex occurs at the extreme south-west of the province (see Fig. 2) at 60°45'N, 48°W. Its visible exposure measures some 45 x 25 km, but an unknown, possibly larger, portion of the complex is obscured from view off the coast; the islands of Indre Kitsigsut and Ydre Kitsigsut indicate that it extends at least another 15 km westwards, making it possibly the largest Gardar complex. A wide variety of rock types (described by Harry and Pulvertaft, 1963) are represented, from gabbro to granite and including the voluminous Nunarssuit Syenite. There is, however, no field evidence for *in situ* differentiation of one rock type from another (although cumulate layering is very pronounced *within* units) and, if the rocks share a common origin, substantial evolution must have taken place at lower levels. Much of the exposed (north-eastern) outer contact is with the Julianehåb Granite basement, but in one or two places, the complex cuts older "greenschists" (regionally metamorphosed under amphibolite facies conditions) or the ENE dykes which occur in large numbers in the region. The relative ages of the various units which make up the complex are not completely determined by field relationships; the order of intrusion proposed by Harry and Pulvertaft (1963) is:

(a) Alángorssuaq Gabbro, (b) Kitsigsut Syenite, (c) Helene Granite,

(d) Biotite Granite, (e) Nunarssuit Syenite and (f) Late Granites.

That the Nunarssuit complex represented a late phase of magmatism within the Gardar was recognised by Harry and Pulvertaft (1963); the complex transects all but a couple of the ubiquitous ENE trending dykes in the area (*c.f.* Grønnedal-Ika which is cut by such dykes). Absolute age determinations reported by Harry and Pulvertaft (1963) and Bridgwater (1965), listed in Table 5, conflict, but the more recent isochron work of van Breemen and Upton (1972) showed that the Tugtutôq Central Complex in the east of the province (which likewise truncates ENE trending dykes) was 1168 ± 37 m.y., and they postulated that Nunarssuit would be similar in age. A Rb-Sr isochron for the small Bangs Havn intrusion (Engell and Pedersen, 1974), which is cut by Nunarssuit, gives 1211 ± 22 m.y., supporting a young age for Nunarssuit.

Samples for this study were collected by Dr. J.G.Anderson of the University of Aberdeen, and are confined to four of the major units in the northern (Alángorssuaq) part of the complex (Fig. 7); a detailed account of recent mapping in that area is given in Anderson's (1974) PhD thesis.

Nunarssuit - Alángorssuaq Gabbro

This unit (Fig. 7) is, on the basis of field relationships, demonstrably the oldest of the four units in this study, and forms a roughly arcuate outcrop about 5 km in length. Fresh exposures of rock are confined mainly to the central part of the unit and consist of olivine gabbro (plagioclase $\sim An_{50}$ and olivine with subordinate titanaugite). Much of the exposed Alángorssuaq Gabbro is, however, highly altered. Preliminary XRF analyses of some 40 samples showed that most had very low Rb/Sr ratios (below 0.03) and, in view of their rather altered condition, were considered unsuitable for an isochron study. Of those remaining, many were either altered or from areas close to the contact with either the Julianehåb Granite basement or the Helene Granite unit (see Fig. 7). Samples 131952, 131951, 131857 (all located within a few metres of the Julianehåb Granite contact) and 131967 (central part of the intrusion) are all uralitised, the olivine and pyroxene having been replaced by amphibole and biotite, and the feldspars are cloudy. Of the remainder, sample 131856 is 30 m from the basement contact whereas 131974 and 131970 are 150 and 300 m respectively from the Helene Granite contact; 131970 has also been uralitised. 131850 was taken very close to the Helene Granite contact (5 m) and is actually hybridised by the granite.

Data for all the samples are listed in Table 8, and an isochron plot is shown in Fig. 8a. Predictably, the points show excessive scatter, and a regression of all points would yield an age of 1243 ± 101 m.y. However, this value is largely weighted by the samples taken close to the contact with the Julianehåb Granite (a line through points 131856, 131952, 131967, 131951 and 131857 would give an age of about 1310 with a large statistical uncertainty), and a line through 131974, 131856, 131970 and 131850 (shown in Fig. 8a) yields an age of 1143 ± 48 m.y. The generally closer proximity of the first group of samples to the contact zones, and the high scatter observed, suggests that the Alángorssuaq Gabbro was contaminated by the Julianehåb Granite, and so these samples are considered least reliable. Of the four points in the second group, one sample (131850) has definitely been affected by the Helene Granite; nevertheless, the four points appear to lie on a concordant line (although the low value of 0.32 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ is not meaningful for a four-point line), perhaps supporting an almost contemporaneous origin for the two intrusions. The preferred age estimate of 1143 ± 48 m.y. is, however, based on atypical (contaminated) gabbros and should be regarded as a minimum figure only.

The low initial $^{87}Sr/^{86}Sr$ ratio for this unit (0.7035 ± 0.0002) is consistent with a "mantle" origin, but the relatively high concentration of Sr (several hundred ppm) and the low Rb/Sr ratio would make this rock a relatively insensitive indicator of post-genetic history.

Nunarssuit - Helene Granite, Biotite Granite and Nunarssuit Syenite

The various units of the Nunarssuit complex are thought by Harry and Pulvertaft (1963) to have been intruded by forcible injection into the surrounding country rocks, major stoping being a predominant mechanism of emplacement. Metamorphic effects in the surrounding basement rocks are largely confined to the contacts of the Nunarssuit Syenite, where a transitional zone of some 40 m exists, and there is some evidence of introduction of Na into the country rock (development of small, secondary sodic amphibole crystals).

Data for the Helene Granite, Biotite Granite and Nunarssuit Syenite are listed in Tables 9, 10 and 11 respectively. The Nunarssuit Syenite (thought to be younger than the Alångorssuaq Gabbro and Helene Granite on the basis of field relationships) is the largest unit, some 24 x 13 km in known areal extent. In general, it is a fairly homogeneous, coarse grained rock, showing spectacular mineral layering in places, consisting of perthitic feldspar (∿85%) with smaller amounts of sodic plagioclase, ferroaugite, amphibole (later phases of hornblende and sodic amphibole mantling other mafic minerals), some Fe-rich olivine and biotite. Quartz, iron ores and accessories (zircon, apatite, fluorite, carbonate) occur in places. Xenoliths of basement metasediments and metavolcanic rocks are found in places, particularly near to the transgressive contact with the older Helene Granite; these are interpreted as disrupted roof-pendants (Harry and Pulvertaft, 1963). The isochron for the Nunarssuit Syenite (shown in Fig. 8d) gives an age of 1154 ± 14 m.y. and an initial ^{87}Sr/^{86}Sr ratio of 0.7043 ± 0.0002, slightly, but significantly, higher than that of the earlier gabbro.

The age relationships between the two granite units (Helene Granite and Biotite Granite) is not entirely clear from field evidence; Anderson (1974) has found some evidence of cross-cutting of the Helene Granite by the Biotite Granite and so the Helene Granite is believed to be the older. Certainly, both granites are younger than the Alángorssuaq Gabbro and the Helene Granite is older than the Nunarssuit Syenite on the basis of clear transgressive boundaries. The Helene Granite forms an outcrop, some 15 x 12 km, to the south of the Alángorssuaq Gabbro (Fig. 7), of fairly homogeneous coarse grained rock consisting of alkali feldspar with some quartz, occasional apatite and minor mafic minerals (hornblende, sodic amphibole, clinopyroxene, Fe-rich olivine and biotite). The Biotite Granite occurs to the north of the Helene Granite and to the west of the Alángorssuaq Gabbro (Fig. 3). It is homogeneous, coarse grained, conspicuously biotitic, and consists of quartz, microcline perthite, biotite, hornblende and accessories (iron ore, fluorite, zircon, sphene, apatite). Variants of both granites occur (in volumetrically minor amounts). In general, external contacts, both with the basement and with the older Alángorssuaq Gabbro, are sharp, but the boundary between the two granite units is confused by a wedge of marginal variants (Anderson, 1974). Isochron plots for the Helene Granite and Biotite Granite (Figs. 8b and 8c respectively) yield ages (1149 ± 31 and 1162 ± 21 m.y.) which are indistinguishable from that of the Nunarssuit Syenite and the preferred age for the Alángorssuaq Gabbro. The initial $^{87}Sr/^{86}Sr$ ratio for the Biotite Granite (0.7068 ± 0.0013) is significantly higher than those for the Alángorssuaq Gabbro or Nunarssuit Syenite; the value for the Helene Granite (0.7040 ± 0.0024) is too imprecise to make any accurate comparisons (unfortunately, suitable samples with low Rb/Sr ratios for a more precise determination were not available).

Narssaq Complex

The chronological relationships of this intrusion, sometimes called "Dyrnaes-Narssaq", are well established. The complex, which measures 9 x 4 km, lies at 60°55'N, 46°5'W, just behind the town of Narssaq. It consists of several intrusions of quartz syenites and later alkali granite (Ussing, 1912; Emeleus and Upton, 1976), and truncates the Hviddal and younger giant dykes on Tugtutôq island (see Fig. 17). Indeed, the Narssaq Gabbro, which can be regarded as the first (pre-syenite) unit of the complex, is believed (Emeleus and Upton, 1976) to be a high level culmination of these dykes. Both syenite and granite units are, in the north of the complex, affected by one of the E-W trending sinistral faults, showing over three kilometres horizontal displacement. This fault, and numerous ENE dykes which also cut the Narssaq Complex, are transected by the larger Ilimaussaq intrusion which, indced, cuts the Narssaq body itself in the extreme north. Thus the age of the Narssaq Complex is bracketed by the Hviddal Dyke (1175 ± 9 m.y. according to van Breemen and Upton, 1972) and the Ilimaussaq Complex (1168 ± 21 m.y., see following chapter). It was therefore only necessary to determine the initial $^{87}Sr/^{86}Sr$ ratio for this intrusion. Samples were collected by the author from the exposed outcrops of syenite and granite above the Dyrnaes valley. Unfortunately, much of the material is to some extent weathered or affected by low grade metamorphic effects (probably as a result of proximity to the Ilimaussaq Complex), and all but a few of the samples had to be discarded for this reason. Of the 5 chosen for XRF analysis, two syenites had suitably low Rb/Sr ratios. Data for these samples (186679 and 186680) are listed in Table 12 and indicate that the initial ratios (based on an assumed age of 1170 m.y.) were 0.7038 and 0.7042, respectively.

Igaliko

The Igaliko Complex, at 61°10'N, 45°10'W (Fig. 9), comprises the largest area (∿450 km^2) of nepheline syenites exposed in the Gardar Province and intrudes the Julianehåb Granite basement and overlying Eriksfjord Formation. The complex consists of four major intrusive centres, each containing several arcuate intrusions, emplaced in a well-defined anti-clockwise direction (Emeleus and Harry, 1970). A number of satellitic nepheline syenite bodies also occur. Somewhat further from the south-eastern edge of the main complex lies the small Klokken gabbro-syenite intrusion (discussed in the following section), which is petrologically quite distict from Igaliko in that it is oversaturated.

The nepheline syenites of the four major centres have steep-sided, outward-dipping external contacts, and were emplaced by a combination of ring dyke intrusion and block subsidence (Emeleus and Harry, 1970). Although the age relationship between the earliest two centres cannot be determined by field evidence alone (the Qôroq Fjord obscures the critical boundary area), Emeleus and Harry inferred the order of intrusion to be:

(a) Motzfeldt, (b) North Qôroq, (c) South Qôroq and (d) Igdlerfigssalik.

In their original description of the geology of the complex, Emeleus and Harry (1970) numbered, chronologically as far as possible, individual intrusive units within each centre; here, this numbering system is used to identify samples in Tables 13-16 (the scale of Fig. 9 is too small to allow individual units to be distinguished, and the interested reader is referred to the 1:50000 map sheet compiled by Emeleus and Harry, 1970). Thus, for example, within the Motzfeldt centre the intrusive sequence is from SM.1 to the youngest unit, SM.5. In this study, the three earlier centres and the last four units of the Igdlerfigssalik intrusion ("Late Igdlerfigssalik") were investigated, using samples collected by Dr. C.H. Emeleus.

The majority of the Igaliko Complex intrusions are nepheline syenites consisting of perthitic alkali feldspars, nepheline, alkali pyroxenes, Fe-Mn-rich olivine, biotite, Fe-Ti oxides and apatite, with less common analcite, natrolite and alkali amphibole, and accessory eudialyte; occasional pegmatites contain a great variety of rare minerals (Ussing, 1912). In several intrusions, the amount of nepheline is slight; these units, which include SS.4, SI.1 and SI.4, are termed augite syenites. Detailed studies of the rock chemistry and mineral compositions show that quite pronounced variation occurs, as for example in units of the South Qôroq centre, where considerable ranges of olivine and alkali pyroxene compositions are found (Stephenson, 1972; Stephenson, 1974).

The nepheline syenites frequently have good feldspar lamination; less commonly the texture is more granular and occasionally mineral layering, conformable with the lamination, if present, is developed.

The Motzfeldt centre contains 5 separate units. There is relatively little mineral layering in the syenites although feldspar lamination is common. Xenoliths are not common except in unit SM.4 where rafts and smaller inclusions occur; these include syenites, microsyenites, porphyries and trachytes, but some at least are derived from the Eriksfjord Formation and older basement, and they have been recrystallised and metasomatically altered, suggesting a prolonged period of contact with active magma. The isochron for Motzfeldt, based on Rb-Sr whole-rock data listed in Table 13, is shown in Fig. 10a. A value of 4.25 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ indicates significant "geological" scatter; this may reflect patchy contamination within the intrusion or it may indicate that post-magmatic isotopic exchange (for example, by groundwater) has taken place. (Although there must have been some finite time interval between the emplacement of the individual units, such effects are considered unlikely to be in excess of analytical uncertainty, in view of the general lack of distinct chilled contacts.) The age (1310 ± 31 m.y.) is indistinguishable from that of Grønnedal-Ika, some 150 km to the west, and indicates that strongly undersaturated magmas were being produced at both ends of the Gardar Province early in its history. The low initial $^{87}Sr/^{86}Sr$ ratio (0.7024 ± 0.0015) is also indistinguishable from that of Grønnedal-Ika and other uncontaminated Gardar magmas (for example, the Hviddal Dyke and Alángorssuaq Gabbro).

A similarly old age (1295 ± 61 m.y.) for North Qôroq (Table 14, Fig. 10b) has a larger uncertainty due to the higher degree of data scatter; $[SUMS/(n-2)]^{\frac{1}{2}} = 8.51$. This could reflect a larger amount of country rock contamination in the samples, especially as the unit SN.1 (from which the majority of the isochron samples were taken) has an above average xenolith content. The initial $^{87}Sr/^{86}Sr$ ratio (0.7052 ± 0.0030) is imprecise, but there is no evidence to suggest that it should be significantly different to that of Motzfeldt.

Both the South Qôroq and Late Igdlerfigssalik centres yield isochrons (Tables 15 and 16; Figs. 10c and 10d) with insignificant scatter, having $[SUMS/(n-2)]^{\frac{1}{2}}$ values of 1.09 and 1.63 respectively. The ages (1185 ± 8 and 1167 ± 15 m.y.) are indistinguishable from each other at the 2σ level. Each unit consists of 5 arcuate sheets and pronounced mineral layering, especially in unit SS.4 of South Qôroq and units SI.4 and SI.7 of Late Igdlerfigssalik, suggest that *in situ* differentiation took place in some units before the emplacement of the next. The initial $^{87}Sr/^{86}Sr$ ratios of South Qôroq and Late Igdlerfigssalik (0.7029 ± 0.0004 and 0.7029 ± 0.0007) are statistically identical with those of the older Igaliko centres.

The difference in age between North and South Qôroq was unexpected; there is no field evidence to suggest a major break here, and petrographic and geochemical differences seem slight. Even more unexpected was the similarity in age between South Qôroq and the Late Igdlerfigssalik units. All syenite intrusions up to and including the Early Igdlerfigssalik units predated the main ENE "Mid Gardar" dyke swarm. These dykes, which can be correlated with those occurring on Tugtutôq (Upton, 1962; McDonald, 1969) and on the Ilimaussaq peninsula, are plentiful in the Igaliko area, and include "Big Feldspar Dykes" (basic dykes containing plagioclase megacrysts and anorthosite inclusions), lamprophyres, alkali trachytes and phonolites. Most of these dykes are cut by the later units of the Igdlerfiggsalik centre (units SI.4-SI.7); thus Emeleus and Harry (1970) assigned all syenites from Motzfeldt to Early Igdlerfigssalik, together with the majority of the ENE swarm, to the Mid Gardar, using the term "Late Gardar" to cover subsequent faulting, emplacement of the Late Igdlerfigssalik units and the latest few ENE trending dykes. The lack of a discernible chronological break between South Qôroq (and hence Early Igdlerfigssalik) and Late Igdlerfigssalik, shows that this time scale needs revision.

Also, unlike earlier centres, Late Igdlerfigssalik produced marked thermal metamorphism of its surroundings, including parts of the South Qôroq centre (Emeleus and Harry, 1970; Stephenson, 1976); however, samples selected from South Qôroq for this study were chosen so as to avoid such effects, and the high degree of concordance of the isochron gives no reason to suspect thermal resetting.

Klokken

The Klokken intrusion (Fig. 11) is a small (3 x 2 km) oval, layered gabbro-syenite intrusion, lying at 60°56'N, 45°05'W, reaching to within 1 km of the SE margin of the Igdlerfigssalik centre and appearing, geographically, to be the next point in the anti-clockwise Igaliko progression. However, in contrast to Igaliko, Klokken is slightly oversaturated and its most fractionated members are aplites and pegmatites of alkali granite composition; unfortunately, there is no contact between Klokken and Igaliko, neither are there any dyke intersections with Klokken, and so the relative ages cannot be determined on the basis of field evidence.

Despite its small size, the Klokken intrusion contains a great diversity of rock-types, from gabbro, through syeno-ferrodiorite to strikingly layered ferrosyenites. The geology of the body has been briefly described by Parsons (1972b) and aspects relevant to the isotopic work are reviewed by Blaxland and Parsons (1975); more detailed mineralogical work is soon to be published (Parsons, in preparation). The body has sharp contacts with the surrounding Julianehåb Granite and there exist substantial protrusions, roof pendants and rotated xenoliths, suggesting that stoping was important during intrusion. Although the transition from a marginal gabbro to syenite takes place over a short distance, there is no intrusive contact between these two units and all intermediate types may be found. The existance of a basic syenitic liquid intermediate between gabbro and the more fractionated syenites is demonstrated by a late discordant biotite-syenite which cuts (and is chilled against) the ferrosyenites, but is cut by acid aplites; fractionation of gabbro to syenite was a single, continuous event. In this respect, Klokken resembles other Gardar intrusions (*c.f.* Bangs Havn and the Tugtutôq older, Hviddal, and younger giant dyke complexes) but differs from others (*c.f.* Kûngnât and the Nunarssuit Complex at Alângorssuaq) where sharp breaks are observed between gabbros and other members.

Samples for this study were selected from the detailed collections made by Dr. I.Parsons of the University of Aberdeen, and sample locations, together with simplified geology, are given in Fig. 11. One sample (140053) was taken from the outer gabbro, and was carefully chosen to avoid, as far as possible, contamination effects by the Julianehåb Granite (some of the gabbro samples contain rounded inclusions with granophyric quartz-feldspar intergrowths) and the ubiquitous plagioclase alteration. The outer gabbro ring gives way gradually to unlaminated augite syenites (see Fig. 11); this coarse grained rock-type is relatively susceptible to weathering, poorly exposed and, for these reasons, has not been included in this study. The majority of samples come from the central part of the intrusion which comprises a remarkable layered syenite series (Parsons, 1972b; Blaxland and Parsons, 1975), containing granular and laminated syenites. The latter are of particular interest as they show "reversed" crystal settling, *i.e.* mafic minerals (hedenbergite-rich clinopyroxene, fayalite and magnetite) gradually increasing (over about 2 m) towards the top of each layer. Samples 140138, 140154, 140047, 140128, 140049, 140122, 140026 and 140093 are taken from the laminated series and, as listed, represent a series from hedenbergite-rich to alkali feldspar-rich syenites. The layered series is cut by the fine grained feeder of a late augite-biotite syenite (140081) which forms a coarser grained sheet-like body near the topographic summit of the intrusion; this represents a less fractionated phase of injection intermediate between the unlaminated syenites and gabbros.

Data for these samples are listed in Table 17, yielding a true isochron, with $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}} = 1.66$ (Fig. 12). The age (1159 ± 11 m.y.) suggests that Klokken is indeed the last point in the Igaliko progression, chronologically as well as spatially, and its age may well represent the termination of Gardar magmatism in south Greenland. The initial ratio of 0.7031 ± 0.0003 is identical (within 1σ confidence limits) to the Igaliko nepheline syenites, indicating that the magma has suffered little or no crustal contamination.

6. RB-SR DATA FOR THE GARDAR PROVINCE: THE ILIMAUSSAQ COMPLEX

The Ilimaussaq Complex is probably the best known within the Gardar, largely due to the highly differentiated agpaitic sequence and its attendant high concentrations of rare elements. In addition to several general descriptions of the geology and structure of the complex (see, for example, Ferguson, 1964; Hamilton, 1964; Gerasimovskiy, 1969; Sørensen, 1970; Ferguson, 1970a; Ferguson, 1970b; Engell, 1973), there are numerous other papers dealing with more specific aspects of the geochemistry and mineralogy; in the "Contributions to the Mineralogy of Ilimaussaq" series of publications (published in conjunction with the GGU) alone, there are, to date, 41 papers.

The present isotopic study was undertaken in conjunction with Agnete Steenfelt of the GGU, who spent some considerable time in the field at Ilimaussaq, making a collection specifically for a Rb-Sr isochron study. The author personally spent about a week at Ilimaussaq in the company of Dr. B.G.J.Upton in 1974, visiting three localities - the southern differentiated sequence at the head of Kangerdluarssuk Fjord, the north-eastern border of the intrusion just to the north of Nunasarnaq and the mineralised localities in the extreme north-west of the intrusion at Kvanefjeld. No attempt was made to duplicate Steenfelt's excellent sampling, and collection was limited to traverses across the boundary with the Julianehåb Granite and Eriksfjord Formation, for possible future investigation.

Outline of the geology of Ilimaussaq

A simplified geological map of Ilimaussaq, together with sample locations, is shown in Fig. 13. The southern part of the complex intrudes the Julianehåb Granite basement, and the northern part (to the north of Kangerdluarssuk Fjord) further intrudes what remains of the Eriksfjord sandstones and lavas, which reach their greatest observed thickness on the Ilimaussaq peninsula.

The complex is composed of three units: (1) an augite syenite, situated along the southern and western borders and in the uppermost levels of the intrusion, (2) a small oversaturated unit comprising alkali granite and alkali quartz syenite occurring as two sheet-like bodies cropping out on either side of the augite syenite in the top of the intrusion, and (3) a layered sequence of highly differentiated syenites forming the major part of the complex and consisting, in descending sequence , of pulaskite, foyaite, sodalite foyaite, *naujaite* (sodalite-nepheline syenite), *lujavrite* (arfvedsonite and aegirine-bearing nepheline syenite) and *kakortokite* (eudialyte-bearing nepheline syenite).

Different opinions exist concerning the evolution of the complex, and these have been summarised by Sørensen (1970) and Ferguson (1970b). All workers agree that the augite syenite is the earliest unit. According to Hamilton (1964), it is intruded by the alkali granite. The recent field work of Agnete Steenfelt (see Blaxland et al., 1976) has revealed xenoliths of both augite syenite and alkali granite/quartz syenite in the pulaskite and foyaite; xenoliths of the augite syenite are also found in other members of the layered sequence. The field relations thus indicate that the augite syenite had solidified before intrusion of the layered sequence.

Most workers agree that the layered sequence was formed by *in situ* crystallisation and differentiation in one magma chamber. Pulaskite, foyaite, sodalite foyaite and naujaite crystallised downwards from the roof of the magma chamber; about 600 m of naujaite formed as a flotation cumulate in which crystals of sodalite were poikilitically enclosed in much larger crystals of feldspar, eudialyte, arfvedsonite and aegirine. At least 400 m of kakortokite formed as a bottom cumulate by gravity settling, producing a remarkable black, red and white banding by preferential settling of arfvedsonite, eudialyte and feldspar/nepheline rich layers (Bohse et al., 1971). Lujavrite crystallised from the residual magma between the naujaite and kakortokite and, before its final consolidation, intruded and brecciated the lower and marginal parts of the overlying rocks.

Ferguson (1964), Ferguson (1970a), Engell (1973) and Larsen (1976) emphasise that the geochemistry of the augite syenite and the agpaites, and the chemistry of the mafic minerals contained in these rocks strongly indicates that the agpaites and the augite syenite are related, and all authors assume that the agpaites formed by differentiation from an augite syenitic magma. Ferguson (1964), Ferguson (1970a) and Engell (1973) further asserted that this differentiation took place *in situ*, although Engell's calculations show that derivation of agpaites by closed-system fractionation from an augite syenite parent would require a volume of magma greatly exceeding the size of the Ilimaussaq chamber as judged from its present exposure.

Based mainly on field evidence, Ussing (1912), Hamilton (1964) and Steenfelt (1972) argued that the augite syenite and the agpaites formed from two separate intrusive events and that the differentiation leading to the agpaitic magma took place at depth. Larsen's (1976) studies on the compositional trends of the mafic minerals support this view and further suggest that the agpaitic suite was formed by the inflow of two or possibly three pulses of magma.

The north-west corner of Ilimaussaq cuts the Narssaq granite-syenite complex (see previous chapter) which in turn post-dates the Hviddal Dyke, dated by van Breemen and Upton (1972) at 1175 ± 9 m.y. (Table 5), and so this places a maximum age limit on the age of all Ilimaussaq units. Also listed in Table 5 are three polylithionite ages, averaging 1086 ± 24 m.y., obtained by Moorbath et al. (1960); these mineral ages are clearly anomalous in view of the new data discussed in the previous chapter, and suggest that Ilimaussaq may be a very much younger body than the main group of Gardar intrusions. Also, of course, the possibility is raised that the first (augite syenite) and later (agpaitic) intrusions were separated by a considerable time interval, although this is regarded as improbable in view of the lack of any chilled marginal facies in the agpaites.

Radiometric age and initial $^{87}Sr/^{86}Sr$ ratios

Preliminary XRF analyses of 42 Ilimaussaq samples for Rb and Sr indicated that Rb/Sr ratios of the major rock units increase in the order: augite syenite (0.1-0.4), kakortokite (0.5-12), pulaskite (5.5-12.5), lujavrite (10-42), alkali granite (15-35), naujaite (20-65); this is consistent with the bulk chemistry of the rock-types. Of these samples, 14 were selected for isotope dilution analysis; they comprise three augite syenites, 7 kakortokites, one pulaskite, two lujavrites and one alkali granite. These were chosen to represent all major rock-types (except the naujaites, which have extreme Rb/Sr ratios and might unduly weight subsequent regression treatment) and also to include sufficient samples of two rock types (augite syenite and kakortokite) to enable the possibility of separate origins of these facies to be investigated.

Sample locations are indicated in Fig. 13, Table 18 lists the analytical data and Fig. 14 displays these data on an isochron diagram. As can be seen from Fig. 14, the alkali granite sample (86339) clearly falls below the isochron line and so is not included in any of the regression analyses discussed below. Regression of the remaining samples (Table 18) yields an age of 1188 ± 30 m.y. and an initial ratio of 0.7060 ± 0.0021, but a value of 5.63 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ indicates that this would not be a true isochron. The second regression treatment listed in Table 19 excludes the marginal augite syenite samples. Here, the $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ value is significantly reduced, although still indicating some "geological" scatter. The difference in age (now 1168 ± 21 m.y.) and, more especially, the initial ratio (0.7096 ± 0.0022) is pronounced. Clearly, in the first analysis, the augite syenites, having a lower initial $^{87}Sr/^{86}Sr$ ratio, had weighted the lower end of the isochron, giving a falsely old age.

In fact, the initial $^{87}Sr/^{86}Sr$ ratios of the three augite syenite samples (calculated by extrapolating a 1168 m.y. isochron back through the points) are 0.7064, 0.7037 and 0.7034 for samples 150781, 150782 and 150783 respectively. Bearing in mind the contamination that has affected this marginal facies, these values are very similar to the low ratios found in most other Gardar complexes.

The final regression treatment listed in Table 19 indicates the results obtained by considering only kakortokites, a single unit interpreted as a series of cumulates formed in a well mixed (and hence isotopically homogeneous) magma chamber. Thus, if the observed scatter of points in the second regression analysis is due to differences in the age and/or initial ratio of the different rock units represented, the third regression treatment would be expected to yield a significantly lower value for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$. It does not, and so the second regression (Table 19) is considered most reliable.

The point for the alkali granite sample (86339)(lying below the isochron) yields an apparently younger age and suggests that small-scale migration of radiogenic ^{87}Sr occurred after initial crystallisation of the highly peralklaine rocks (*c.f.* the 1086 m.y. ages for polylithionites reported by Moorbath at al., 1960).

The 1168 ± 21 m.y. age for the Ilimaussaq agpaites marks the end of major magmatism in the central part of the province (Tugtutôq and Narssaq area) and, together with the 1175 ± 9 m.y. age for the Hviddal Dyke, shows that the augite syenite and agpaites must have intruded in close succession. The initial ^{87}Sr/^{86}Sr ratio of the agpaites (0.7096 ± 0.0022) is, however, significantly greater than that of the augite syenites ($\sim$0.703).

Origin of the agpaites

The observed difference between the initial ratio of the augite syenite and the agpaites can only be consistent with a model of *in situ* differentiation of both these units from one magma if some mechanism existed whereby either (a) the agpaitic magma generated a high ^{87}Sr/^{86}Sr ratio after the initial cooling of the early augite syenite or (b) contamination by radiogenic strontium occurred after this initial crystallisation. Taking into account the average Rb/Sr ratio of the agpaites, it would require at least 50 m.y. for the agpaitic magma, in a fully differentiated state, to generate sufficient radiogenic strontium to account for the observed difference between the two units. The existence of such a gap is clearly ruled out by the new age data. The possibility of *in situ* contamination of a residual agpaitic magma by sources of radiogenic strontium can be rejected on the basis of field relationships (Fig. 13) as (a) the agpaites appear to have been largely separated from the older basement by a sheath of the earlier augite syenite, and (b) the roof zone of the intrusion almost certainly lay wholly within the Eriksfjord Formation, consisting of basalts (with low $^{87}Sr/^{86}Sr$ ratios) and quartzites (with low Sr concentrations). The possibility of exchange with groundwater is being investigated; the author has supplied arfvedsonite separates from the agpaites and whole-rock powders of the augite syenites to Dr. S.M.F.Sheppard for oxygen and hydrogen isotopic analysis. Preliminary results show that, although the agpaites show a record deuterium enrichment, the $\delta^{18}O$ values are "normal" and do not suggest interaction with heated groundwater.

On the basis of these arguments, the possibility of *in situ* differentiation of the agpaites (Ferguson, 1964; Ferguson, 1970a; Engell, 1973) is rejected. However, the proximity of the agpaites both in space and time strongly supports previous suggestions that the two are related and probably derived from the same parent magma. In such a hypothesis, the Sr isotopic composition of the augite syenite is similar to that of most other Gardar syenites, and it remains only to explain the high initial ratio of the agpaites.

For the observed initial ratio to have been generated wholly within the differentiating magma at depth, again a period of at least 50 m.y., and probably 100-150 m.y. would be required. Although such a lengthy history is considered unlikely for the Ilimaussaq magmas, it should be noted that some Gardar complexes are around 1300 m.y. old (see previous chapter), and if a magma produced at this time differentiated gradually in a deep-seated magma chamber, the $^{87}Sr/^{86}Sr$ ratio would be of the order of 0.710 at the time of intrusion of Ilimaussaq.

A simpler model envisages the introduction of radiogenic Sr into the agpaitic magma, either in the deep-seated chamber or during intrusion through the crust. Traditionally, introduction of radiogenic strontium can be explained by (a) complete or partial melting of crustal material, (b) assimilation of crust before or during cooling of the magma, or (c) contamination by reaction with wall-rock.

The undersaturation and extreme peralkalinity of the Ilimaussaq magma makes it difficult to postulate a derivation from silicic, aluminous, crustal material, and hence (a) is discounted. In view of what is known about the isotopic composition of the surrounding Julianehåb Granite basement (van Breemen et al., 1974) and the tendency for $^{87}Sr/^{86}Sr$ and Rb-Sr ratios to decrease at depth (for example, Heier, 1964; Heier, 1973) direct assimilation can be ruled out because of the enormous amounts of crust that would be required.

Heier (1964), citing data on the Skaergaard intrusion published by Hamilton (1963), suggested that *preferential* leaching of radiogenic ^{87}Sr may occur during the heating of silicate rocks. During heating, potassium-rich minerals (K-feldspars and micas) are often the first to break down and as these minerals are also the major Rb-bearing phases they will have the highest $^{87}Sr/^{86}Sr$ ratios. The extremely reactive agpaitic magmas of Ilimaussaq may well have reacted with the walls and roof of the magma chamber, and with descending blocks of crust during intrusion, selectively removing radiogenic ^{87}Sr from unstable positions in Rb lattice sites. As the agpaitic magma must have differentiated from a large volume of parent magma, the leaching process presumably occurred in the roof zone of the deep magma chamber. Evidence for continuous enrichment of radiogenic ^{87}Sr near the top of a magma chamber has been presented by Noble and Hedge (1969), and van Breemen et al., (1975) have described similar effects in the roof zones of peralkaline granite and syenite intrusions.

7. RB-SR DATA FOR THE GARDAR PROVINCE: THE IVIGTUT GRANITE AND ASSOCIATED CRYOLITE BODY

The Ivigtut Granite is a pipe-like body some 300 m in diameter, occurring at 61°12'N, 48°11'W on the south-eastern coast of Arsuk Fjord, and is one of the smallest complexes in the Gardar Province. This body has attracted attention since the first mineralogical surveys were made in south Greenland by Giesecke in 1806-1813 (Giesecke, 1910) due to the rich mineralisation associated with the intrusion. The ore material, largely cryolite and siderite, occupies a sizeable proportion of the present outcrop of the granite (Fig. 15) and, in fact, has been the world's major source of cryolite since 1859.

Various accounts of the geology of the Ivigtut Granite and the surrounding rocks have been published, notably those by Wegmann (1938), Callisen (1943) and Berthelsen (1962). More detailed accounts of the mineralogy and petrology of the granite and the associated cryolite are provided by Pauly (1960) and J.Bailey (in preparation). A memoir dealing with the Ivigtut region of south Greenland has been compiled by the GGU (Berthelsen and Henriksen, 1975). Previous ideas about the age of the granite have been based on rather tenuous cross-cutting relationships between the granite, a series of alkaline dykes and the much larger Grønnedal-Ika complex to the east; the granite has been assumed by various authors to have been emplaced during the middle or late phases of Gardar activity (for example, Emeleus, 1964). Rb-Sr (single sample) analyses (Table 5) by Moorbath and Pauly (1962) gave an age estimate of 1330 m.y. (recalculated to a ^{87}Rb decay constant of $\lambda = 1.39 \times 10^{-11} y^{-1}$) for this granite.

Geology and setting of the Ivigtut Granite

Like the nearby complexes at Kûngnât and Grønnedal-Ika, the Ivigtut Granite intrudes Archaean (>2500 m.y.) acidic gneisses and gabbro-anorthosite-bearing gneisses. The body is almost entirely surrounded by an intrusion breccia up to 70 m in width, consisting of blocks and fragments of surrounding gneiss in a matrix of fine grained granite. About 200 m to the east of the main granite body there is a small, oval outcrop of another breccia, the Bunkebreccia, consisting entirely of cataclastic gneissic material and interpreted by Berthelsen (1962) as being a satellite body of the main granite and produced by a gas-drilling process. Berthelsen concludes from the relationships with several generations of Gardar dykes that the intrusion breccia and Bunkebreccia were contemporaneous, and he points to evidence of strong local structural control over the contacts of the intrusion and of the presence of a cupola-shaped joint system within the exposed granite suggesting that the present erosion surface is very near to the top of the original intrusion.

The granite itself forms an almost circular outcrop with vertical contacts with the surrounding intrusion breccia and country rocks. Since the active quarrying operation ceased in 1962, the mine area, which included almost the entire oval outcrop of the granite, has been flooded and accounts of the geology of the granite are based on information obtained when the mine was active. Samples used for this study were selected from a collection made by the author during July 1974, and were taken from discarded blocks excavated during quarrying. The samples include two distinctly different types of granite.

The first group ("grey granites", samples 186699, 186700, 186701, 186703, 186735) consist of pale grey to pink granite composed of large (may be >1 cm) subhedral to anhedral microperthite and anhedral quartz in a finer groundmass of quartz, microperthite and ferromagnesian minerals (mostly biotite with some amphibole and opaques) occurring in clusters of small plates. (Callisen, 1943, reports alkali amphibole and pyroxene, but these have not been identified in the present samples.) Accessory minerals include zircon, fluorite and opaques. Sericitisation of feldspars in this group is quite pronounced in places and all samples are fractured to some extent, the fracture traces being outlined by thin (generally <1 mm) veins, mainly of fine grained quartz, feldspar and biotite.

The second group ("white granites", samples 186729, 186730, 186732, 186733) are white, albitised, medium grained granites consisting mainly of quartz, K-feldspar and albite. Some of the feldspar has been sericitised and in places ivigtite mica is an alteration product. Accessory minerals are the same as in the first group, but with the conspicuous addition of cryolite and some siderite. The cryolite occurs as interstitial patches, often quite large, and in some samples it may account for almost 20% of the rock, although 10-15% is typical. Textures similar to those observed in these "white granites" are seen in the younger Nigerian granites (Jacobsen et al., 1958), where cryolite is also an accessory phase, and have been interpreted as signifying extensive or complete recrystallisation resulting from pervasions of fluorine-rich, late-stage hydrothermal fluids (Bowden and van Breemen, 1970; van Breemen et al., 1975; Bowden, personal communication, 1975).

The cryolite body occupied the central portion of the granite outcrop and extended downwards for almost 150 m. From observations on a series of tinguaitic dykes that cut the Ivigtut Granite, Berthelsen (1962) concluded that the cryolite formed after the granite was already solid.

The tinguaitic dykes are extremely altered and mineralised as they approach the area of the mine. The mineralogy of the cryolite body was briefly described by Pauly (1960) who suggested that the temperature of emplacement of the cryolite ore was not over 500°C, based on ZnS-FeS equilibrium temperatures. Many papers relate to the unusual Pb isotopic composition of the Ivigtut galenas; the $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ ratios fall significantly below the single-stage growth curve and Kanasewich and Slawson (1964) have proposed two- and three-stage models suggesting that the galenas were derived from ordinary leads in excess of 3000 m.y. old. There is thus some evidence for a basement component in the mineralising fluids.

Radiometric age and initial $^{87}Sr/^{86}Sr$ ratios

On the basis of XRF determinations for Rb and Sr, 9 samples with suitable Rb/Sr ratios were chosen for isotope dilution analysis. Also, two samples of cryolite ore, one of fluorite ore and two separates (different size fractions) of interstitial cryolite from one of the samples (186730) were analysed in order to compare initial $^{87}Sr/^{86}Sr$ ratios.

$^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ values for the granite whole-rock samples (Table 20) yield an isochron (shown in Fig. 16) with an age of 1248 ± 25 m.y. and an initial ratio of 0.7125 ± 0.0048. A value of 2.76 for $[SUMS/(n-2)]^{\frac{1}{2}}$ indicates that the observed scatter about the line approximates (perhaps is slightly in excess of) that to be expected from analytical uncertainty. Table 20 also shows the analytical results for the ore samples (cryolite and fluorite) and the separates of interstitial cryolite from sample 186730. The initial $^{87}Sr/^{86}Sr$ ratio values listed for these samples are calculated using an age of 1248 m.y.

The 1248 ± 25 m.y. age for Ivigtut is statistically identical with that of Kûngnât, making these the only two central complexes which give a "Mid Gardar" age. (Engell and Pedersen, 1974, report an age of 1211 ± 22 m.y. for Bangs Havn but, as discussed later, this may need confirmation before being totally accepted.) Patchett (1976) has now shown that at least some of the many lamprophyre dykes in this western region and also the BD series of olivine dolerite dykes are probably $\sim$1250 m.y. old (Table 5) and the smaller dykes in the immediate vicinity of Ivigtut and Grønnedal may also be related to a Mid Gardar series of fault movements that deformed the Grønnedal-Ika complex (Emeleus, 1964); there is thus evidence for substantial magmatic activity in the middle of the Gardar Period.

Origin of the Ivigtut body

The initial $^{87}Sr/^{86}Sr$ ratio of 0.7125 ± 0.0048 for the Ivigtut Granite is the highest recorded for any of the Gardar complexes. This value is largely determined by isochron points for the white granite samples (Fig. 16) and Table 20 shows that this group is greatly enriched in Sr relative to the grey granites. (A separate regression analysis using only grey samples yields a similar age but very imprecise initial ratio.) The high ratio cannot be directly linked to secondary enrichment by the main ore body as the three samples of analysed ore have rather uniform lower values (Table 20) than the granite (actually, the differences are not significant at the 2σ level). (The relatively low initial ratios for the ore material accord with similar values, listed in Table 5, obtained for samples of baryto-celestite by Moorbath and Pauly, 1962.) However, it is immediately apparent from the data in Table 20 that the interstitial cryolite from sample 186730 (the lowest point on the isochron) had a trace element and isotopic composition differing greatly from that of the main ore. Both Rb and Sr are highly enriched (up to several hundred times) in the interstitial cryolite and, most important, the initial $^{87}Sr/^{86}Sr$ ratio of this phase (∿0.7196) is substantially higher than either the granite whole-rock (∿0.7125) or the ore body itself (∿0.7095).

The modal proportion of interstitial cryolite in the white samples is somewhat variable, but if it amounted to 10% (a minimum value) then the initial ratio of the remaining 90% of the whole-rock would be about 0.710; if the cryolite was as high as 20% (a maximum figure) then the remaining rock would have had a value of 0.705. It is clear that the whole-rock granite samples had, prior to the addition of the interstitial cryolite phase, an initial $^{87}Sr/^{86}Sr$ ratio no higher that 0.705-0.710, and possibly considerably lower as some isotopic reequilibration may have taken place during the introduction of the interstitial phase. (The preferential ^{87}Sr enrichment by interstitial cryolite of the white granites - low points on the isochron - would tend to give an artificially young age. However, this effect is small, and well within the stated statistical uncertainty.) Extensive recrystallisation involving the cryolite phase can be ruled out as this would be expected to produce an isotopically homogeneous rock.

This places certain constraints on the method of introduction of the interstitial cryolite phase, and two mechanisms seem possible. First, the cryolite may represent a late-stage residual fluid circulating freely through the interstices of the largely solid granite. Such a fluid would come into contact both with the surrounding older country rock and with xenolithic blocks within the granite, enabling radiogenically enriched Sr to be leached out. Second, the cryolite phase may have resulted from the interaction of the residual liquid in the cooling granite with invading fluorine-rich hydrothermal fluids from below (Sørensen, 1950), such fluids having had the opportunity to preferentially leach ^{87}Sr in a similar manner to that described for Ilimaussaq in Chapter 6. This second hypothesis is appealing in that it allows for a highly reactive fluid to come into contact with a relatively large amount of older crust, whereas the first model is largely limited to reaction within the bounds of the cooling granite mass. This model would also be consistent with the very high Rb and Sr contents and higher ^{87}Sr/^{86}Sr ratio of the interstitial cryolite phase relative to the main ore as the interstitial phase could have been earlier and leached out a proportionately greater amount of basement strontium.

When the substantial contribution of radiogenic ^{87}Sr introduced during the final stage of crystallisation of the granite is taken into account, the initial ^{87}Sr/^{86}Sr ratio of the granite magma probably lies within the limits observed for other Gardar magmas, although the uncertainties associated with any quantitative calculations are large. As shown in Chapter 5, two other granites in the Gardar Province, the Biotite Granite and Helene Granite of the Nunarssuit Complex to the south of Ivigtut, yield initial ratios of 0.7068 ± 0.0013 and 0.7040 ± 0.0024 respectively, and it seems likely that the Ivigtut magma had a similar ratio before the start of the mineralisation process.

8. Rb-Sr DATA FOR THE GARDAR PROVINCE: THE NARSSAQ POTASSIC ULTRAMAFIC BODIES

In addition to the predominantly salic large central complexes discussed in the preceding three chapters, and the generally more basic dyke swarms, there exist smaller intrusive bodies of Mg-rich, Si-Al-poor composition; among these are a group of mica-peridotites and mica-pyroxenites in the vicinity of Narssaq and also in the north-east of Tugtutôq (Fig. 17). Those in the proximity of Narssaq, described by Upton and Thomas (1973), probably merge at depth to form a single sill-like structure, and crop out adjacent to the Narssaq Gabbro (believed to represent the termination of the younger giant dyke complex of Tugtutôq). The contact relationship between ultramafic and gabbroic bodies is equivocal, but Upton and Thomas believe the gabbro to be the older. Both bodies are cut by dykes of the ENE swarm, which are in turn truncated by the Ilimaussaq complex. The age of the ultramafic rocks, thus bracketed by the ages of the Hviddal Dyke and Ilimaussaq (Table 5) must be ∿1170 m.y. The rocks are composed mainly of magnesian olivine, diopside, phlogopite mica and Fe-Ti oxides, and chemically, the composition combines relatively high concentrations of typically early-crystallising elements (Mg, Ni, Co, Cr) with high concentrations of late-crystallising and "incompatible" elements (K, Ba, Zr, Rb, V, Ce, Nb) (Upton and Thomas, 1973).

Clearly, these represent a separate category of Gardar magmas that cannot have originated from low pressure fractionation of the relatively Mg-poor magmas that were intruded, almost simultaneously, in much larger volumes in the same place. Initial $^{87}Sr/^{86}Sr$ ratios were desired from these rocks to test if they had had a different isotopic history. Four samples (from two of the bodies) of pyroxenite were selected (locations given in Fig. 17); in three cases the whole-rocks were analysed but, in view of the relatively high Rb/Sr ratios (Table 21), mineral separates (pyroxene and apatite) were taken so as to improve precision on the initial $^{87}Sr/^{86}Sr$ ratio estimate. Table 21 lists the Rb-Sr data for these samples, and also shows a calculated initial $^{87}Sr/^{86}Sr$ ratio, based on an assumed age of 1170 m.y. 2σ errors on these values, based on errors of 0.5% and 0.05% on $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ respectively, vary between 0.0007 and 0.0009. The initial ratios are, at the 2σ level, identical at 0.703, thus placing them among the lowest of the values found for the larger central complexes.

Possible origins for the ultramafic bodies have been considered by Upton and Thomas (1973). An origin involving the assimilation of crustal rocks would not account for the unusual chemistry unless, they suggest, a magma of carbonatitic composition was involved. They point out that this would still not account for, amongst other things, the high Ni and Cr concentrations. The Sr isotopic data from Grønnedal-Ika (Chapter 5, Table 7) give no support to such mixing schemes and, furthermore, the Grønnedal-Ika carbonatite has an initial ratio identical with that of the ultramafites, suggesting that the two magmas originated in isotopically very similar source regions. Other possible origins, considered by Upton and Thomas (1973) to be consistent with the geochemistry, were (1) small-scale partial melting of an upper mantle source region containing phlogopite, (2) a form of "zone-refining" process whereby the ascending ultramafic magma scavenges incompatible elements from the surrounding mantle material and (3) fractional crystallisation of garnet and clinopyroxene from a picritic mantle partial melt. On field and petrological grounds, they favour the last of these mechanisms, noting that as the intrusion of the ultramafics follows immediately (both in space and time) the ascent of large volumes of basaltic magmas in the Tugtutôq-Narssaq region, incipient melting of mantle, as envisaged in (1), seems unlikely.

The Sr isotopic data do not permit a distinction between these three models, but do give some information on the provenance of the magmas. The low initial $^{87}Sr/^{86}Sr$ values compare with the lowest values reported by Barrett and Berg (1975) for mantle-derived South African kimberlites, and fall low in the known range of Gardar ratios. On these grounds, it seems unlikely that any significant amount of crustal strontium could have been incorporated into the potassic ultramafic magmas before or during intrusion.

9. INITIAL ^{87}SR/^{86}SR RATIOS FROM THE GARDAR PROVINCE AND PETROGENESIS

Although estimates of the initial isotopic composition of magmas cannot be used directly to determine the mode of genesis, such data are useful in identifying possible source areas - different regions of the earth's mantle and crust, having varying Rb/Sr ratios, are characterised by different ^{87}Sr/^{86}Sr ratios. This, indirectly, often enables a more critical assessment of possible modes of petrogenesis, by eliminating those relying on sources with unsuitable Sr isotopic composition.

Recent oceanic basalts are generally regarded as providing the best estimate of the strontium isotopic composition of the mantle. For example, Hedge et al. (1970) reported a weighted average value of ^{87}Sr/^{86}Sr for 90 oceanic basalts as 0.7035 and Peterman and Hedge (1971) show that most oceanic basalts have values between 0.702 and 0.706. Fig. 18 depicts the probable evolution of ^{87}Sr/^{86}Sr in the basalt source regions, assumed to be within the upper mantle, and is based on an estimated initial ^{87}Sr/^{86}Sr ratio of 0.69899 for achondrite meteorites (Papanastassiou and Wasserburg, 1969), and the above-mentioned values for recent oceanic basalts. The shape of the shaded area (basalt source regions) assumes that the mantle has had an essentially constant (though heterogeneous) Rb/Sr ratio; if the Rb/Sr ratio decreased with time, then this area would be convex upwards (Hart, 1969; Hart and Brooks, 1970). The initial ^{87}Sr/^{86}Sr ratios for the Gardar complexes are plotted on this diagram, the relevant portion being expanded in the upper left inset. To maintain clarity in the diagram, individual complex names and error bars are omitted; this information is depicted in Figs. 19 and 20. Also shown in Fig. 18 is an evolution line for "typical" continental crust 2700 m.y.

old (Faure and Powell, 1972) and a minimum evolution line for the Julianehåb Granite basement. This line is based on an age of 1800 m.y. and minimum Rb/Sr value of 0.2 suggested by the recent studies of van Breemen et al. (1974); this represents the best estimate available of the likely Rb/Sr composition of the basement surrounding the southern and eastern Gardar complexes, but it should be noted that the basement is very heterogeneous and van Breemen et al. (1974) have only examined the area to the south-east of the Gardar outcrops.

On the basis of Fig. 18, the Gardar complexes can be divided into three groups with low (0.702 - 0.704), intermediate (0.704 - 0.707) and high (over 0.707) initial ratios. These groupings may be related to observed compositional trends; on an alkali-silica diagram ($Na_2O + K_2O$ versus SiO_2) the majority of Gardar magmas follow a common evolutionary path from basalts (gabbros) to rocks of trachytic and syenitic composition, then diverge towards either oversaturated or undersaturated compositions (Upton, 1974).

Low $^{87}Sr/^{86}Sr$ ratios (0.702 - 0.704)

Most of the Gardar centres fall into this group. Fig. 19 shows only values determined by Rb-Sr isochron methods at East Kilbride, and indicates uncertainties at 2σ; however, other available data show that additional Gardar bodies fall into this group, including the Narssaq ultramafic bodies (∿0.703, Rb-Sr single samples, Chapter 8), Bangs Havn (0.7034 ± 0.0002, Rb-Sr isochron; Engell and Pedersen, 1974) and olivine gabbros and included anorthosite xenoliths from the younger giant dyke complex of Tugtutôq (∿0.7030, Rb-Sr single samples; Patchett et al., in preparation).

Field, petrographic and geochemical evidence suggests that most Gardar magmas evolved by differentiation, either *in situ* or at lower crustal levels, from basaltic parent magmas. Such a process is consistent with these low, mantle-type initial ratios found for the majority of gabbros, syenogabbros and syenites, representing the chemical trend up to the point of divergence. The only complexes which belong to this petrological grouping and for which the $^{87}Sr/^{86}Sr$ ratios may be over 0.704 are Kûngnât and North Qôroq. However, the statistical uncertainties for these syenite centres (Table 22; Fig. 19) are large in both cases, straddling the 0.704 cut-off, and so these values are not considered anomalous. Furthermore, determinations on the Kûngnât gabbro ring-dyke (Chapter 5; Table 6) lie within the isochron error-bar and suggest that the Kûngnât magma did indeed have a low "mantle-type" initial ratio.

Within the stated 2σ limits, all these low $^{87}Sr/^{86}Sr$ ratios (with the exception of the Nunarssuit Syenite) could be identical at about 0.7033; it is more likely, however, that a range of values really exixts, and that this range represents varying degrees of contamination by older crust relatively enriched in radiogenic ^{87}Sr. Neither the precision of the data, nor the control over possible contamination effects during the magma's ascent through the crust allows sensible speculation as to whether the Gardar magmas evolved from an isotopically uniform mantle source. However, especially in view of the fact that stoping is widely considered to have been a predominant method of emplacement, it seems reasonable to expect that at least some degree of radiogenically enriched Sr from the basement would have mixed with the pristine Gardar magmas, and that such variable amounts of crustal contamination could account for the entire observed variation. Low values reported here for the gabbros and syenites may also apply to many of the Late Gardar dykes (see Fig. 21); Upton (1962) and Emeleus and Harry (1970) believe that these dykes may well represent preliminary tappings of the bodies of magma which later gave rise to the central complexes.

In summary, it appears that the majority of the Gardar syenites and related magmas have low initial $^{87}Sr/^{86}Sr$ ratios consistent with a derivation from a primitive (upper mantle) source.

Intermediate $^{87}Sr/^{86}Sr$ ratios (0.704 - 0.707)

The silica oversaturated trend in the Gardar leads to granitic rocks. Of the Gardar central complexes containing major granites (Ivigtut, Nunarssuit Helene Granite, Nunarssuit Biotite Granite, Tugtutôq Central Complex, Narssaq Complex, Ilimaussaq), isochron initial $^{87}Sr/^{86}Sr$ ratios have only been determined for those of Ivigtut and Nunarssuit. The unusually high initial ratio of the Ivigtut Granite has already been treated in Chapter 7. Of the two Nunarssuit granites studied here, the Helene Granite has yielded an imprecise initial ratio (Table 22; Fig. 19) but the Biotite Granite is significantly enriched in ^{87}Sr when compared to the adjacent Alángorssuaq Gabbro and Nunarssuit Syenite. Also, although the Nunarssuit Syenite (oversaturated) has an initial ratio (0.7043 ± 0.0002) not inconsistent with a mantle origin, it is nevertheless significantly higher than those of the other, undersaturated, syenite intrusions, and the Alángorssuaq Gabbro.

It seems reasonable to suppose from the present data that the various units of the Nunarssuit complex were intruded in fairly quick succession in the order: Alángorssuaq Gabbro → Helene Granite → Biotite Granite and Nunarssuit Syenite (see Fig. 21). The Alángorssuaq Gabbro is the only one with an initial ^{87}Sr/^{86}Sr ratio low enough to have crystallised rapidly from uncontaminated (although undoubtedly fractionated) mantle source material. For the others, two hypotheses can be considered:

(a) *Closed system enhancement of ^{87}Sr.* The syenite and granite were derived by differentiation from a basaltic parent magma having a low ^{87}Sr/^{86}Sr ratio, over a long period of time, thus enabling the accumulation of radiogenic ^{87}Sr by *in situ* decay of ^{87}Rb. The Late Gardar ages would represent only the final isotopic reequilibration in these units.

(b) *^{87}Sr derived from crust.* All Nunarssuit units include at least some older crustal material (either a direct origin by partial melting in the lower crust or assimilation before or during intrusion) and the elevation of the ^{87}Sr/^{86}Sr ratio reflects the degree of crustal content. Selective radiogenic ^{87}Sr enrichment (similar to that described for Ilimaussaq, Chapter 6) may be important in such a process.

The first hypothesis would involve the separation of all the major units during Early Gardar times, followed by separate development before final emplacement during the Late Gardar. This differentiation could be a continuous process or, more plausibly, a relatively rapid event interrupted by an intermediate episode (or episodes) of solidification. If the Nunarssuit Syenite has an average Rb/Sr ratio of 0.3 (a rough, but perhaps reasonable, estimate) and the Biotite Granite, 1.0, and if it is assumed that the differentiation of the syenite and granite took place from a gabbroic parental magma as a linear function of time during the 150 m.y. between Early and Late Gardar phases, then the excess radiogenic ^{87}Sr produced during this period would correspond to the higher initial ratios observed.

The Alángorssuaq Gabbro, having a low Rb/Sr ratio, would have produced relatively little radiogenic ^{87}Sr during this period. The objections to this hypothesis lie in its complexity; in order for the high initial ratios to be generated within the magma, a lengthy 150 m.y. cooling period of separate pockets of magma or a multi-stage crystallisation-remelting model must be envisaged. Furthermore, although such a mechanism is possible for Nunarssuit, such a process could not have occurred at Igaliko where all centres, both Early and Late Gardar, have low initial ratios (the only possible exception, North Qôroq, is Early Gardar and thus did not have time available to generate a large excess of ^{87}Sr).

Model (b) is considered more probable. The possibility of a lower crustal origin for the magmas is difficult to evaluate without knowledge of the isotopic composition at great depths. The Julianehåb Granite (Ketilidian) basement probably had an average ^{87}Sr/^{86}Sr value of about 0.710 at 1200 m.y. before present (data from van Breemen et al., 1974) and it is considered unlikely that deep crustal values were higher than this. The uncertainty on the initial ratio for the Helene Granite is too large to allow useful discussion, but it is possible that the Biotite Granite (0.7068 ± 0.0013) was derived from the lower crust; the Nunarssuit Syenite, however, is unlikely to have been generated in this region. The possibility of crustal contamination of mantle-derived magma should therefore be considered.

If contamination due to stoping and assimilation of older crust occurred, then the Biotite Granite, with the highest initial ratio, was the unit most affected. This is by far the smallest unit of the complex (the relatively small arcuate outcrops of Alángorssuaq Gabbro and Helene Granite are believed to have once been much more extensive before obliteration by the Helene Granite and Nunarssuit Syenite respectively) and is interpreted as a late-crystallising H_2O-rich differentiate of the Helene Granite magma (Anderson, 1974). As the Biotite Granite was clearly one of the latest Nunarssuit magmas to be emplaced, it can be assumed that the surrounding country rock through which the magma ascended had already been pre-heated by the earlier units, and this heat, together with the extra water content of the sub-solvus granite magma, would have enabled a relatively large amount of material to be assimilated.

Such a hypothesis is further supported by chemical evidence (Anderson, 1974). Most chemical and mineralogical characteristics of the Biotite Granite are consistent with an origin as a late-stage derivative from the Nunarssuit Syenite-Helene Granite parent magma. However, the magnesium content of the Biotite Granite (0.2 - 0.6%) is strikingly higher than that of the Helene Granite (<0.15%). Such MgO-enrichment is regarded as strong evidence for bulk digestion of the Alángorssuaq Gabbro and pre-Gardar greenschists; ample field evidence for such assimilation is documented by Anderson (1974).

If such evidence of contamination by greenschists and gabbros is accepted, it seems highly likely, indeed inevitable, that contamination by the Ketilidian granite basement must have taken place at slightly deeper levels; this would significantly raise the $^{87}Sr/^{86}Sr$ ratio but the chemical identity of the assimilated basement would be largely masked by the Biotite Granite magma. In this regard, it has been noted (Blaxland et al., in press) that in the field the Biotite Granite has characteristics more akin to the pre-Gardar allochthonous granites to the north of Nunarssuit (Berthelsen and Henriksen, 1975) than to other Gardar granites.

Contamination in the earlier, larger and H_2O-undersaturated magmas was, apparently, less pronounced. In addition to the parameters already mentioned, the relative Sr concentration in the respective magmas would affect the degree to which the contamination was reflected in the $^{87}Sr/^{86}Sr$ value; the Alángorssuaq Gabbro, for example, with a high Sr content, would have been insensitive to the introduction of radiogenically-enriched crustal Sr. When these various factors are taken into account, the hypothesis of bulk assimilation seems most plausible for those Gardar complexes having $^{87}Sr/^{86}Sr$ ratios in the range 0.704 - 0.707.

High $^{87}Sr/^{86}Sr$ ratios (above 0.707)

The high alkali, low silica trend reaches its extreme in the Ilimaussaq complex, composed predominantly of agpaitic rocks. The high initial ratio (0.7096 ± 0.0022) determined for these agpaites (Chapter 6) cannot be explained in terms of bulk assimilation or of melting in the lower crust, and strongly suggests that highly reactive magmas of this type are able to *preferentially* leach radiogenic ^{87}Sr from older crust. Similarly, in the Ivigtut Granite, which has the highest initial ratio of the province (0.7125 ± 0.0048), it is proposed (Chapter 7) that fluorine-rich liquids could have extracted radiogenically enriched Sr from the surrounding basement. In this case, an interstitial cryolite phase with a high concentration of radiogenically enriched Sr (∿0.719) could adequately account for the elevated $^{87}Sr/^{86}Sr$ value observed.

Both Ilimaussaq and Ivigtut represent instances where fractionation has produced extremely reactive fluids, capable of leaching incompatible elements from the crust. It is geologically, and perhaps economically, interesting that these are the only Gardar complexes containing substantial mineralised zones (both having been the subject of serious mining investigations). The possibility that $^{87}Sr/^{86}Sr$ ratios can identify magmas that may have scavenged ore elements from surrounding country rocks could have important implications in commercial prospecting.

Comparison with previous studies and conclusions

In their recent book "Strontium Isotope Geology", Faure and Powell (1970) comment that:

> "The petrogenesis of alkaline rocks remains an important but unsolved problem"

For a long time, alkaline rocks have been recognised as a special group of igneous rocks which, although consituting a very small proportion of igneous rocks as a whole (Daly, 1933; Barker, 1974), are a recurring feature both geographically and throughout time. Several mechanisms for alkaline magma genesis have been proposed by various authors (Sørensen, 1974, p. 534-539) though none of them can be successfully applied to all alkaline rocks. Essentially, the magma may be generated by (a) some form of mixing process (including processes involving carbonatitic magmas), (b) partial melting in the crust or upper mantle, or (c) fractional crystallisation, possibly from an alkali basalt parent. There may be, of course, a more complex mechanism involving more than one of these processes (for example, Vollmer, 1976). The results of this study support a close genetic relationship, probably largely through differentiation at depth from a common source material, of most of the major rock-types in the Gardar, including carbonatites, gabbros, syenogabbros, syenites and nepheline syenites; however, a significant crustal component in the granitic rocks is indicated.

Although the close relationship between carbonatites and alkaline rocks is widely recognised, their consanguinity within an intrusive complex is difficult to prove on the basis of Sr isotopic evidence. Some earlier data (Bell and Powell, 1970) show rather lower initial $^{87}Sr/^{86}Sr$ ratios in carbonatites than in corresponding alkaline rocks, and, where similar values are reported (for example, Powell et al., 1966) the level of precision is too low to permit definite conclusions. In the Grønnedal-Ika complex, an initial $^{87}Sr/^{86}Sr$ ratio of 0.70318 for the carbonatite sample (31863) is identical to the 0.7032 ± 0.0004 value for the associated nepheline syenites and is as good Sr isotopic evidence for the consanguinity of these units as might reasonably be expected.

The possibility of producing alkaline rocks by mixing carbonatitic and sialic rock-types, such as proposed by Holmes (1950), has been discussed by Powell and Bell (1974); these data from Grønnedal-Ika contraindicate such a process in the Gardar. Mixing (massive assimilation) processes of other kinds have undoubtedly led to the production of alkaline rocks in other areas. For example, Barker and Long (1969) at Brookville, New Jersey; Hurley et al. (1966) and Vollmer (1976) in western Italy. However, these, and other such studies, concern rocks with higher $^{87}Sr/^{86}Sr$ ratios than reported here for the Gardar, and the geological and petrological evidence favours such interpretations. In the Gardar Province, on the other hand, in the only complexes with sufficient ^{87}Sr to warrant such a hypothesis (Ilimaussaq and Ivigtut), field and geochemical evidence militates against mixing.

The production of alkaline rocks by partial melting in the upper mantle and lower crust has been propounded by Bailey (1974a, 1974b) and some Sr isotopic evidence suggests that the process may operate in certain instances, especially in the African Rift System (Powell and Bell, 1974).

Bailey's hypothesis is appealing in areas where there is little evidence of basaltic magmas that could be parental to alkaline rocks (the African Rift); however, such a model should not be too hastily applied to other regions, as emphasised by Barker (1969, 1970), who believes that most alkaline rocks are probably generated at sub-crustal levels, a view in accord with the data presented here.

Indeed, the Gardar Province contains abundant evidence for the widespread genesis and ascent of large volumes of basaltic magma. In contrast, the Nigerian alkaline province (mostly peralkaline granites with some syenites; Jacobsen et al., 1958) and the White Mountains (New Hampshire) complexes (mildly alkaline granites and syenites; Chapman, 1968) contain no such overt evidence of a basaltic parent. A recent isotopic study on the Nigerian Province (van Breemen et al., 1975) showed that unmodified syenites had low (mantle) $^{87}Sr/^{86}Sr$ values, in sharp contrast to the high and variable ratios in peralkaline rocks (attributed to radiogenic ^{87}Sr enrichment in the roof zone during the late stages of magmatic differentiation) and consistently intermediate ($\sim$0.707) values in biotite granites (significant crustal contribution). In both the Gardar and Nigerian provinces, high initial $^{87}Sr/^{86}Sr$ ratios were associated with mineralisation.

10. AGE PATTERNS AND REVISED CHRONOLOGY OF THE GARDAR PROVINCE

The Rb-Sr isochron ages now available for the Gardar central complexes (summarised in Table 22 and depicted as a cumulative frequency diagram in Fig. 20) suggest a bimodal distribution of ages (with peaks of magmatism at 1300 and 1165 m.y., the latter comprising the majority of Gardar complexes), with only the relatively small Kûngnât and Ivigtut bodies having intermediate ages, indicating a smaller magmatic episode at about 1250 m.y. before present. However, the number of undated igneous events (dykes and lavas) remains large, and the widespread and recurring dyke-injection and faulting in the region suggests that magmatic activity may have been intermittant over a period in excess of 150 m.y.

The data allow a (somewhat arbitrary) division into Early, Mid and Late Gardar phases, and a generalised chronological scheme based on these divisions is presented in Fig. 21. This shows the relative chronological position, based on field and radiometric evidence, of the major Gardar igneous rocks; the arrows do not imply genetic connexions, only age relationships, question-marks being used where evidence is speculative. For convenience, the Gardar Province has been divided into four geographic regions: Kûngnât-Ivigtut-Grønnedal, Nunarssuit-Isortoq, Tugtutôq-Narssaq and Igaliko.

Kûngnât-Ivigtut-Grønnedal

Detailed discussions of the chronology of this area, based mainly on field relationships, have been presented by Berthelsen (1958), Upton (1960), Emeleus (1964) and Berthelsen and Henriksen (1975). The earliest major magmatic event in this area was the intrusion of the Grønnedal-Ika complex; evidence of earlier dyke intrusion (for example, inclusion of a lamprophyre xenolith in an early Grønnedal-Ika syenite) is sparse. A group of lamprophyric dykes near Arsuk is believed by Upton (1960) to be the earliest Gardar intrusions in the Kûngnât area, and are dissimilar to the lamprophyres in the Grønnedal area. However, mineral ages (whole-rock-phlogopite dates) of 1254 ± 29 and 1276 ± 29 m.y., recently obtained by Patchett (1976) suggest that these were not older than Grønnedal-Ika and are probably related to the Mid Gardar phase of activity that produced the complexes at Kûngnât and Ivigtut. Patchett has also obtained whole-rock-biotite ages of 1272 ± 18 and 1278 ± 18 m.y. for two BD_0 (olivine dolerite) dykes and 1264 ± 18 for a BD_1 dyke, all from the Ivigtut peninsula, thus supporting the existence of a distinct Mid Gardar tectonic-magmatic episode. The earliest major faulting in the area post-dates the BD_0 dykes (Berthelsen and Henriksen, 1975). Following this faulting, in the Grønnedal area, were a number of basaltic dykes and four phases of alkaline (trachytic and phonolitic) dykes interspersed with two further faulting episodes (Emeleus, 1964). Jacobsen (1966a and 1966b) reports that a series of thin granophyric dykes which probably immediately preceded the intrusion of the Ivigtut Granite cut representatives of all the Gardar dykes in the Ivigtut-Grønnedal area. If this is the case, then the age of the Ivigtut Granite (including post-granitic tinguaite dykes and the cryolite deposit) effectively marks the end of known magmatic activity in this northwestern region of the province. The final phase of faulting (Group 3 of Emeleus, 1964), although undated, is unlikely to be substantially younger than the rest of the tectonic activity in the area.

Nunarssuit-Isortoq

The relative age relationships of the various units within the Nunarssuit Complex are indicated in Fig. 21. The relative ages of the Biotite Granite and Nunarssuit Syenite cannot be determined from field and radiometric evidence but Anderson (1974) believes that the Biotite Granite may be the younger, representing a final, residual, H_2O-saturated magmatic phase.

The relative positions of the various ENE dykes and giant dykes in the Nunarssuit-Isortoq region are difficult to evaluate fully. The Nunarssuit Complex cuts representatives of all major ENE dykes in the area and also cuts the Bangs Havn Complex. A Rb-Sr isochron of 1211 ± 22 m.y. has been obtained for the latter centre by Engell and Pedersen (1974); this is consistent with the field relationships but also suggests that at least some of the ENE dykes are significantly older than very similar igneous activity (known to be about 1170 m.y. old) in the Tugtutôq-Narssaq region to the east. (There is some evidence that at least some of the Isortoq dykes, including the giant dykes, are younger than Bangs Havn - D.Bridgwater, personal communication to J.Patchett, 1974 - but a substantial number of them are inferred to be older or contemporaneous.) The possibility that the emplacement of similar basic dykes in the two areas of Nunarssuit-Isortoq and Tugtutôq-Narssaq took place essentially simultaneously seems too likely to be dismissed on the basis of one Rb-Sr isochron determination, reported from another laboratory using procedures quite different from those employed at East Kilbride. For this reason, the Bangs Havn value has been ignored while constructing Fig. 21; however, in this figure the ENE dykes of Nunarssuit-Isortoq are depicted as being chronologically offset from those of Tugtutôq-Narssaq to emphasise that there is no direct evidence that they are contemporaneous.

Tugtutôq-Narssaq

Here, representatives of each of the three Gardar magmatic phases occur. The Eriksfjord Formation, probably the oldest magmatic event of the Gardar Province (although its chronological relationship to Grønnedal-Ika in the west in unknown) occurs in the immediate vicinity of Narssaq and on the Ilimaussaq peninsula. The BD dykes, although cropping out on Tugtutôq, are not seen in the vicinity of the Eriksfjord Formation but are shown by Rb-Sr data to be younger. Patchett (1976) has obtained Rb-Sr mineral "errorchrons" of 1285 ± 144 and 1244 ± 117 m.y. on samples from two BD_0 dykes from Tugtutôq and Julianehåb respectively, with (probably reliable) whole-rock-biotite ages of 1245 ± 18 and 1238 ± 18 m.y. for these samples (errors are 2σ of the mass spectrometer data means). Most magmatic activity in the area occurred at about 1170 m.y. ago in the Late Gardar; indeed, the earliest (Hviddal Dyke) and latest (Tugtutôq Central Complex and Ilimaussaq) dated events of this phase are chronologically identical at the 1σ level. The termination of magmatism here is not positively known (one or two small dykes transect the Ilimaussaq Complex) but it is thought unlikely that it was substantially later than about 1160 m.y. before present.

Igaliko

Here again, the Eriksfjord Formation is the earliest recorded Gardar event. Following this, but preceding the Motzfeldt centre of Igaliko, are the small satellite syenite centres of East Motzfeldt and North Motzfeldt

(Emeleus and Harry, 1970). Also truncated by the 1310 ± 31 m.y. Motzfeldt centre is a E-W dyke thought be Emeleus and Harry to be possibly an eastern extension of a BD_0. If this is so, then the BD dyke series must have had a much longer history of emplacement than the relatively short period in the Mid Gardar suggested by the whole-rock-biotite ages of Patchett (1976). Alternatively, the dolerite dyke may belong to a pre-BD dyke series emplaced contemporaneously with the earliest lamprophyres in the Grønnedal area. Other intrusions which are possibly of Early Gardar age include the satellite syenite stocks of Narssarssuaq, Tunugdliarfik and Østfjordsdal, all cut by members of the South Qôroq centre (Fig. 9). The first of these cuts the Eriksfjord Formation, but the other two do not crop out against the supracrustal rocks. The Østfjordsdal Syenite truncates a group of alkaline trachyte dykes, petrographically similar to the "Fox Bay" group of dykes reported by Ussing (1912) and found on the coast of Igaliko Fjord.

The local relationships of individual intrusions in and around the Igaliko complex are described in the relevant sections of Chapter 5. The termination of magmatism in the region is represented by Klokken (1159 ± 11 m.y.) and a few ENE dykes which cut the Late Igdlerfigssalik centre of the Igaliko Complex (probably contemporaneous with similar dykes cutting Ilimaussaq).

General conclusions

The age data presented here indicate that alkaline magmatism, producing a highly varied series of rock-types, occurred in the Gardar Province intermittently throughout a period in excess of 150 m.y. There does not appear to be any simple relationship between age and geographic location, mode of emplacement or the recognised petrological division into saturated and undersaturated types; indeed, very similar rock-types were emplaced in the same place over a great time span (Igaliko), similar rocks were emplaced in widely separated areas at roughly the same time (Grønnedal-Ika and Igaliko) and dissimilar rocks were emplaced in close proximity almost contemporaneously (for example, Ilimaussaq and Narssaq, Igaliko and Klokken). The Late Gardar complexes are associated with abundant ENE dykes (especially in the Nunarssuit-Isortoq and Tugtutôq-Narssaq regions); although there is no such widespread evidence of dykes in the Early Gardar (during the emplacement of Grønnedal-Ika, Motzfeldt and North Qôroq), evidence of tension exists in the form of faulting and the generally NW trending lamprophyre and olivine dolerite dykes in the west of the province.

Clearly, the recurrence of similar rock-types throughout the Gardar Period must indicate broadly similar tectonic conditions. The relationship between tensional stress and the alternating intrusion of (mafic) dykes and (salic) central complexes has already been mentioned by Gill (1973), and hypotheses that the Gardar Province may represent an ancient rift zone, thus far expressed in rather tentative terms (Sørensen, 1970; Stewart, 1970; Bailey, 1974a; Upton, 1974) can be restated along more definite lines in the light of the new age data (Upton, in preparation). It seems probable that three separate tensional events are represented by the Early, Mid and Late Gardar magmatic episodes, respectively.

Although the Eriksfjord Formation has not been dated, the Early Gardar age that is always tacitly assigned to it is supported by a recently obtained Rb-Sr isochron age of 1355 ± 92 m.y. (Wanless and Loveridge, in press) for the very similar Seal Lake group of supracrustal rocks ..

in central Labrador (see following chapter). Thus, a thick sequence of Early Gardar lavas and sandstones, rapidly accumulating in fault-bounded graben, may well represent the initiation of a ∿1300 m.y. episode of rifting, terminating with the emplacement of the Early Gardar nepheline syenites.

A Mid Gardar (∿1250) period of tension is amply demonstrated (in the west of the province) by widespread lamprophyre and olivine dolerite dykes and later transcurrent faulting. Again, the tensional period may have terminated with the emplacement of the salic (and, in contrast to the Early episode, *oversaturated*) magmas at Kûngnât and Ivigtut, although it is not possible to demonstrate that these complexes post-date the faulting.

Finally, the Late Gardar period produced, in the south of the province (Nunarssuit to Igaliko), widespread ENE faulting (giving rise to the marked topographic lineation observed today) and dilation accompanied by the intense ENE dyke emplacement, especially in Tugtutôq and Isortoq where combined dilation has been estimated by Upton to be about three kilometres. Again, as in the case of the Early and Mid Gardar phases, activity ceased with the widespread emplacement of large volumes of salic magmas.

If this generalised hypothesis of rifting in the Gardar Province is correct, it could well have implications beyond the bounds of south Greenland. On a pre-drift map of the North Atlantic, the western extension of the predominantly ENE trend of the Gardar lineation (which runs essentially parallel to the assumed position of the Grenville "Front") reaches to central Labrador. Here, in the Seal Lake region, accumulations of sediments and lavas similar to the Eriksfjord Formation and agpaitic central-type complexes similar to Ilimaussaq are to be found. This suggests that the pattern of rifting proposed for the Gardar Province may extend into this area of Canada.

Similarly, the eastern extension of the Grenville Front extends into Scandinavia where an isolated agpaitic complex, Norra Kärr, occurs in southern Sweden. Although this is considerably further from the pre-drift position of the Gardar than is the Seal Lake area, the possibility again existed that there might be a chronological relationship. Rb-Sr isotopic studies of both Labrador and Swedish agpaitic complexes are discussed in the following two chapters.

11. RB-SR STUDY OF THE RED WINE COMPLEXES, CENTRAL LABRADOR

The Red Wine alkaline province (Fig. 22) comprises a number of small (less than 5 km diameter) agpaitic bodies, and an area of oversaturated peralkaline gneisses, occurring between latitudes 53°50'N and 54°15'N and longitudes 60°30'W and 62°50'W in central Labrador and situated within the Grenville Front Tectonic Zone (Wynne-Edwards, 1972). Mapping in the area has been carried out by the Brinex Corporation of Montreal, by S.K.Singh of the University of Ottawa (Singh, 1972) and, more recently, by geologists of the University of Toronto and the Canadian Geological Survey (Currie, 1973; Curtis et al., 1974; Currie et al, 1975; Curtis, 1975).

These complexes are interesting in the present context because of their similarities to the Ilimaussaq intrusion. First, on a pre-drift map, the Red Wine and Ilimaussaq complexes are relatively close (about 500 km apart) and lie close to the inferred Grenville Front. Second, the petrology and geochemistry of the peralkaline syenites of both complexes are similar (Currie et al., 1975) - both are agpaitic in the sense advocated by Edgar (1974). Third, in both cases the agpaitic rocks are closely associated with thick sequences of supracrustal basalts and sandstones; in the case of Ilimaussaq, the Eriksfjord Formation is known to have been at least three kilometres thick, whereas the Seal Lake Group in the Red Wine area may be over 10 km (Brummer and Mann, 1961). Fourth, preliminary ages reported for both the supracrustal rocks and the agpaites suggest that they may well have been contemporaneous with the Gardar sequence. For the Seal Lake Group, K-Ar ages of 960 ± 60 m.y. (Leech et al., 1963), 975 m.y. (Wanless et al., 1965), 865 m.y. (Wanless et al., 1966) and 843

± 125 m.y. (Wanless et al., 1967) probably reflect post-formational metamorphic effects, but a more recent Rb-Sr study reports an age of 1355 ± 92 m.y. (adjusted to $\lambda = 1.39 \times 10^{-11}\ y^{-1}$) based on a 9-point isochron (Wanless and Loveridge, in press). Although the absolute age of the Eriksfjord Formation has yet to be determined, it is widely inferred to be not greatly older than the ∿1310 m.y. lower limit placed on it by the age of the Motzfeldt intrusion. K-Ar determinations on the North Red Wine complex (sometimes referred to as the Joan Lake complex) range between 1000 and 1400 m.y. (Singh, 1972) and a more recent Rb-Sr (4-point isochron) determination by Teledyne Isotopes Incorporated gives an estimated age of 1264 m.y. (error not listed; Currie et al., 1975).

In view of the similarities to the Gardar Province, it seemed desirable to conduct a more thorough Rb-Sr isotopic study than had hitherto been applied to the Red Wine agpaites, in order to determine the primary intrusion age of the complexes and the effects of subsequent metamorphism on the Rb-Sr systematics.

Outline of geology

In contrast to the Gardar Province, where the generally good exposures and relative lack of severe structural deformation have enabled the field relationships to be determined in some detail, the interpretation of the geology of the Red Wine Province is hampered by complex structural relationships, the effects of regional metamorphism and the general paucity of good exposure. No comprehensive account of the geology of the area has yet been published, and the following brief summary is based on the summaries given by Currie et al. (1975), Curtis (1975) and personal communications from Dr. J.Gittins and Dr. L.Curtis (1975).

The Red Wine Province is situated close to the boundary between the Churchill and Grenville Provinces, directly to the south of the Naskaupi Fold Belt (Fig. 22). This belt contains the Seal Lake Group, tholeiitic basalt flows and sills intercalated with continental sediments (conglomerates, quartzites, shales, etc.) estimated at $\sim$7 km (Baragar, 1969) and over 10 km (Brummer and Mann, 1961). Although these rocks are complexly folded, metamorphic grade is low, ranging from prehnite-pumpellyite facies (Baragar, 1974) to Greenschist facies (Brummer and Mann, 1961). The Seal Lake Group overlies, probably unconformably, the Letitia Lake Group of quartz-feldspar porphyries, tuffs and sediments. Most outcrops of this Letitia Lake Group lie within a major zone of cataclasis 10-15 km wide, to the south-east of which occur the polymetamorphic "Red Wine Peralkaline Gneisses" of much higher grade. The Red Wine agpaitic intrusions occur as a series of tectonically deformed pods with sinuous extensions forming a NE trending belt crossing this zone of cataclasis (Fig. 22).

The age and structural relationships are equivocal. The oldest rocks in the area are inferred to be the "Wapustan Gneiss", occurring to the south-east of the Red Wine Gneisses and inferred by Curtis (1975) to be older by virtue of their polymetamorphic history (they reached upper amphibolite grade at some stage in their history). The Letitia Lake Group has yielded K-Ar ages of 1350 and 1430 m.y. (Emslie, 1970) and a Rb-Sr age of 1562 m.y. (adjusted to $\lambda = 1.39 \times 10^{-11}\ y^{-1}$; Wanless and Loveridge, 1972). Curtis (1975) believes that the Red Wine Peralkaline Gneisses may be related to the agpaites; they are bracketed in age by the agpaites (which intrude the gneisses) and the Letitia Lake Group (which occurs as inclusions in the gneisses).

The present configuration of the agpaitic bodies is probably largely controlled by post-intrusion folding and elongation. There are two main complexes (Fig. 22) referred to as North Red Wine and South Red Wine respectively (Singh, 1972. has called the North Red Wine complex "Joan Lake" complex). Other smaller, elongated bodies occurring to the north-east of these major centres are believed to be closely related metamorphosed dykes or hybrid rocks (metamorphosed fenites). The North and South Red Wine complexes are shown in Fig. 23, which indicates the rock types represented (after Currie et al., 1975). Petrographically, these can be divided into two groups: (a) nepheline syenites and malignites (melanocratic nepheline syenites) retaining relict igneous textures (commonly trachytoid or blastoporphyritic) and composing some 20% of the complexes, and (b) amphibolite facies leucocratic to melanocratic agpaitic gneisses which envelop and swirl around pods of the syenite. This last group is divided (Fig. 23) into blue melanocratic gneiss, green melanocratic gneiss and leucocratic gneiss, based mainly on the respective contents of arfvedsonite and alkali pyroxene. All the agpaites consist predominantly of albite, alkali feldspar, nepheline, alkali amphibole (arfvedsonite), alkali pyroxene, aenigmatite, and lesser quantities of eudialyte, apatite, pectolite and some other accessory minerals. Eudialyte and aenigmatite are major phases in some samples. All types show metamorphic texture, though the only metamorphic minerals observed are the series of aluminous (aegirine-jadeite) pyroxenes. Contacts between rock-types are quite abrupt, but with no unequivocally intrusive relationships. Metasomatic effects, including hybridisation, occur around the alkaline rocks and distinct alkaline mafic minerals can be found up to 10 km from the complexes.

No detailed hypothesis concerning the original evolution and intrusion of the Red Wine agpaitic magmas has been propounded but Curtis (1975) believes, on the basis of observed mineral assemblages, that metamorphism took place between 4 and 7 kb and 450-675°C.

Isotopic investigation

A pilot investigation on a reconnaissance suite of 7 samples (collected by Dr. B.G.J.Upton and Dr. S.Gandhi; locations shown in Fig. 22, data listed in Table 23) suggested an age of 1285 ± 276 m.y. However, repeat analyses on separate splits of inhomogeneous whole-rock powders suggested a younger metamorphic age (replicate analyses on samples LD-35 and LD-42 give "isochrons" of $\sim$1000 and $\sim$800 m.y. respectively). A more detailed study was clearly desirable, and Dr. J.Gittins who, with colleagues from the University of Toronto and the Canadian Geological Survey, has been working on the petrology of these complexes, was approached regarding the possibility of obtaining samples from his, much more extensive collections.

As a result, 13 samples of the "igneous" nepheline syenites and malignites (5 from North Red Wine, 6 from South Red Wine and two from an isolated outcrop some 6 km to the east of South Red Wine) were selected from collections made by Dr. J.Gittins and Dr L. Curtis of the University of Toronto. Locations are indicated in Figs. 22 and 23, and whole-rock analytical data are listed in Table 24. The isochron diagram for these samples (shown in the inset of Fig. 24) shows that the values do indeed define a straight line, with the exception of P8-175 and P7-029. The first of these samples is an amphibole-pyroxene rock (melteigite) with extremely low Rb and Sr contents and may have been formed as a "sweat" during the metamorphism of the complex (Curtis, 1975); the second point would yield an unrealistic age with any possible choice of initial

$^{87}Sr/^{86}Sr$ ratio and so both of these values are omitted from the regression treatment. Regression of the remaining 11 points yields an age of 1345 ± 75 m.y. and an initial ratio of 0.7021 ± 0.0103. A high value (13.42) for $[SUMS/(n-2)]^{\frac{1}{2}}$ is not surprising in view of the metamorphic history of the complexes and the relatively small hand-samples used for the whole-rock isochron.

To further investigate the metamorphic effects on the Rb-Sr systematics, mineral separates were prepared for two of the samples; one (P8-312) was from mid-way along the isochron, the other (P8-311) close to the intercept. In each case, four minerals or groups of minerals were separated, and the results of the isotope dilution analyses are listed in Table 24. The isotopic ratios of the minerals and the respective whole-rock points are plotted in Fig. 24 and the parameters obtained on regression are listed in Table 25. The ages obtained (969 ± 64 and 1024 ± 74 m.y.) are indistinguishable from each other (even at the 1σ level) and in each case the mineral isochron has a much smaller slope and yields a higher initial ratio than the whole-rock (1345 m.y.) line (see Fig. 24).

Discussion

The two parallel (within 1σ) mineral isochrons are interpreted as metamorphic "resetting" ages imposed upon igneous rocks formed at 1345 ± 75 m.y. ago. The excess scatter of points about the least-squares lines (Table 25) is accountable in terms of this later metamorphic effect. Within the whole-rock systems, this scatter is likely to be essentially random, and thus the 1345 m.y. age and 75 m.y. 2σ uncertainty value are considered to be reliable estimates of the primary intrusion. The ∿1000 m.y. metamorphic ages, typical of K-Ar and Rb-Sr ages for the Grenville metamorphism, represent the time at which the agpaites became a closed system with respect to Rb and Sr, probably after the peak of Grenville metamorphism. Unfortunately, the whole-rock initial $^{87}Sr/^{86}Sr$ ratio (0.702 ± 0.010) is not sufficiently precise to describe the source of the agpaitic magmas.

The 1168 ± 21 m.y. age for the Ilimaussaq complex in the Gardar Province (Chapter 6) is apparently very much younger than the 1345 ± 75 m.y. age for the Red Wine complexes. However, it is now known that the Gardar Period covered a considerable interval of time and the Red Wine age is indistinguishable from those of some other Gardar intrusions (Grønnedal-Ika, Motzfeldt, North Qôroq). Moreover, there is evidence in the Gardar that very similar magmas were emplaced at the same location at widely separated time intervals (for example, the Igaliko Complex), and it may simply be fortuitous that the only known agpaitic complex in south Greenland is Late Gardar in age whereas the observed occurrences in the Red Wine region were intruded in "Early Gardar" time.

The age obtained here for the Red Wine agpaites is also indistinguishable from the 1355 ± 92 m.y. Rb-Sr age for the Seal Lake Formation (Wanless and Loveridge, in press), confirming a chronological link between the two rock groups similar to that existing between the Early Gardar Eriksfjord Formation and older alkaline complexes.

The present geochronological data thus support a correlation across the Davis Strait between the Red Wine/Seal Lake area and the Gardar Province. The Gardar region is interpreted (Upton, in preparation) as a product of three distinguishable rift systems marked by the intrusion of dykes, transcurrent faulting and the emplacement of alkaline central complexes. The suggested extension of the Gardar Province into the Labrador area would imply that a similar tectonic environment must have obtained during the intrusion of the Red Wine plutons which would, in this scheme, belong to the earliest of the Gardar rift regimes. Currie et al. (1975) have noted a number of differences between the Red Wine and Gardar provinces, principally the apparent absence of dykes and the relative paucity of alkaline rocks in the Red Wine area, as compared to the great abundance of these intrusions in the Gardar, and they suggest that this may be due to the present level of exposure or to subsequent obliteration by later metamorphic effects. It is worth noting, however, that in the Early phase of magmatism in the Gardar Province, dykes are not at all abundant (most occur in the Mid and Late phases) and only three intrusive centres of this age have been recorded. The relatively complex structural relationships of the Red Wine area and the lack of detailed mapping make further correlation uncertain, but it appears that this region of Labrador may well have been involved in an extension of the Early Gardar rift system, whereas the later Gardar events were more restricted in occurrence. Subsequent compressional tectonics during the Grenville metamorphism have substantially altered the Red Wine rocks, but did not reach the Gardar Province.

12. RB-SR STUDY OF THE NORRA KÄRR COMPLEX, SOUTHERN SWEDEN

A small, elliptical area of peralkaline rock occurs about 15 km north of the small town of Gränna on the west shore of Lake Vättern at 58°06'N, 14°40'E in southern Sweden. This rock body, named Norra Kärr after the farm situated in the middle of the outcrop, measures about 1200 m (N-S) by 400 m (E-W) and occurs in a region of heterogeneous granitic-gneissic Precambrian (Svecofennian) rocks collectively known as the Växjö Granite.

The area was first described by Törnebohm (1906) and has since been the subject of a number of studies, notably by Gavelin (1912), Backlund (1932), von Eckermann (1942) and Adamson (1944). More recently, the genesis of the body has been the subject of strong controversy when Koark (1960, 1969) fundamentally challenged the age relationships and igneous origin postulated by the previous authors and reaffirmed by von Eckermann (1968).

The present interest in the complex derives from its various similarities to the Ilimaussaq and Red Wine complexes: agpaitic composition, proximity to the Grenville (Sveconorwegian) Front, and possibly similar age to Ilimaussaq (von Eckermann, 1968).

Rock types occurring at Norra Kärr

Fig. 25 (a sketch map of the complex after Adamson, 1944) provides a general indication of the rock types present and sample locations. The area is poorly exposed and the more recent studies of Koark and von Eckermann show that the position of the outer contacts may be subject to revision; the inferred enveloping zone of fenite and the direct contact with the Växjö Granite indicated in Fig. 1 are major points of contention.

The main rock type of the body is *grennaite* (named after the nearby town of Gränna), a fine grained, often schistose, greyish rock composed of alkali feldspar, nepheline, aegirine, eudialyte and catapleite. Other rock types in the complex occur as large pods within the main mass of grennaite and are, by contrast, coarse grained. *Lakarpite* (afvedsonite-albite-nepheline syenite) occurs at two localities (Fig. 25) and exists in both massive "igneous-looking" and in cataclastically deformed states. *Pulaskite* generally has a massive character whereas *kaxtorpite* (pectolite-eckermannite-aegirine-nepheline syenite) has a marked schistose foliation.

The grennaite and coarse grained rocks all share an agpaitic composition but the relationship of the lakarpites, pulaskites and kaxtorpites to the main mass of grennaite is not clear. Törnebohm (1906) believed them to be older inclusions caught up in the grennaite magma and Adamson (1944) tended to agree, though he drew attention to some field relationships which could suggest an intrusive relationship.

The country rocks around the peralkaline body show a zone of fenitisation, estimated by Adamson (1944) to be 25-100 m wide, but thought by von Eckermann (1968) to be narrower.

Controversial structural relationships

On the basis of field, petrographic and mineralogical studies, Adamson (1944) proposed that the peralkaline rocks of Norra Kärr are part of a volcanic neck or plug, and suggested that the coarse grained components may be early crystallising fractions that have been disrupted and included in the grennaitic magma. This hypothesis of intrusive origin is in broad agreement with most other authors. Although exposed contacts between grennaite and country rock (Växjö Granite, according to these authors) are rare, von Eckermann (1968) examined several temporary sections. In general, the contacts are reported to be sharp; fenitisation of the country rock is intense near the contact and fenitised xenoliths of country rock are found in the grennaite. Foliations within the peralkaline rocks are interpreted as protoclastic (Törnebohm, 1906; Adamson, 1944) and primary crystallisation features (Gavelin, 1912). von Eckermann (1968) described some of the foliation as "unmistakable flow structure". According to Adamson (1944) the observed foliations in the grennaite conform perfectly with the outer contact of the body.

A sharply contrasting opinion of the origin of the Norra Kärr body has been expressed by Koark (1960, 1969), who disagreed with much of the evidence put forward by the other authors and ascribed the complex to a metamorphic origin. He claimed that the immediate country rock is *not* Växjö Granite but a series of granitic, quartz dioritic and schistose gneisses which in places are intruded by the Växjö Granite; thus the presumed age relationship between agpaites and Växjö Granite was possibly invalid. Koark also disagreed with previous interpretations of the foliation in the grennaite. Contradicting Adamson (1944), he said (Koark, 1969) that the foliation in the grennaite maintained a N-S strike right up to the outer contacts at the north and south ends of the complex, and thus threw doubt on a direct igneous origin. Koark's detailed petrofabric studies caused him to conclude that the observed foliations are best explained in terms of metamorphic schistosity.

Isotopic investigation

In view of these conflicts, a fresh approach to the study of Norra Kärr would be helpful, and the present isotopic study was devised to shed further light on two questions: (a) is it probable that the Norra Kärr alkaline rocks had a primary igneous origin, and if so, at what time were they emplaced? (b) to what extent have the Rb-Sr systematics of the rocks been affected by regional and/or thermal metamorphism?

Previously, von Eckermann (1968) reported a K-Ar age for Norra Kärr of 1020 m.y. (compare new K-Ar values listed in Table 28) but no details of the sample were provided and the method of analysis and interpretation was not explained; similarly, he reports a K-Ar age of 1500 m.y. for the Växjö Granite, much younger than the 1740 m.y. Rb-Sr isochron age reported by Welin et al. (1966), the 1695 ± 40 m.y. Rb-Sr isochron age for the probably related) Småland Porphyries (Åberg, 1972) or the more recent minimum (cooling) age estimates of 1650 m.y. (Patchett, 1976). As Koark (1969) rightly pointed out, von Eckermann's values must be treated with some scepticism.

At the beginning of this present investigation, a pilot study, based on samples collected by Dr. W.W.Boyd (and lodged in the collections of the University of Edinburgh) yielded whole-rock Rb-Sr data listed in Table 26. However, it soon became apparent that this collection had been biassed towards mineralogically interesting samples at the expense of representative material and was not suitable for a precise Rb-Sr isotopic study; a further collection was clearly needed. With this in mind, the present author made a brief visit to Norra Kärr in the summer of 1974 as a member of an excursion organised by Professor Henning Sørensen and collected numerous samples of the various alkaline rock types for isotopic analysis. No attempt was made to further investigate the field and structural relationships treated at length by previous authors. Largely on the basis on Rb/Sr ratios (XRF values), 10 samples (∿1 kg) were selected for isotope dilution analysis. Mineral separates from two of the samples (lakarpites) were also analysed by Rb-Sr and K-Ar techniques. (It is assumed that the coarse grained units of the complex are likely to be at least as old as the grennaites and would thus be sensitive to any metamorphic episode affecting the main complex. It is, of course, possible that they intrude the grennaite, but if this is the case, the age difference is likely to be small.)

Table 26 lists whole-rock Rb-Sr data for the 7 samples from the original University of Edinburgh collections. A graphical plot of these is hardly justified; the age (omitting sample NK-10 which falls far below any reasonable line) is 1539 ± 240 m.y. with a value for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ of 17.03. These data are considered to be superceded by those listed in Table 27, for samples from the author's own collection. Fig. 26 shows an isochron plot for the whole-rock samples (main diagram) and mineral separates (inset diagram). It is immediately apparent that the whole-rock point for sample SW-19 falls below the best-fit line through the rest of the whole-rock points and so it is omitted from the regression treatment. All other points yield an age of 1580 ± 62 m.y. and an initial $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ ratio of 0.7072 ± 0.0035. The degree of scatter is indicated by a value of 7.12 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$. No regression data are given for the mineral analyses but reference lines of 1580 m.y. and 1250 m.y. are shown to provide some indication of the relationship of the mineral points to the whole-rock isochron age.

Interpretations and genesis of Norra Kärr

The value of 7.12 for $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ for the regression of 9 whole-rock samples indicates that the scatter about the best-fit line exceeds that to be expected from analytical uncertainty; either (a) Norra Kärr was not an isotopically homogeneous system at the time of the event represented by the 1580 m.y. age, or (b) the body has been subsequently isotopically disturbed. Of course, both (a) and (b) may be true.

The plot of points for mineral separates from samples SW-7 and SW-8 shows that there is considerable scatter and no line providing a meaningful age can be drawn. It is, however, apparent that a line connecting any one of the mineral points with the appropriate whole-rock point would give an age lower than the 1580 m.y. whole-rock "age" and, in some cases, lower than the 1250 m.y. reference isochron (the lowest point would give an age of about 1040 m.y.). It may therefore be assumed that at some time after 1580 m.y. ago, these samples were indeed isotopically disturbed, but that this did not result in isotopically homogeneous mineral systems. Furthermore, such disturbances might be expected to have introduced inhomogeneities on the scale of the hand samples used for the whole-rock isochron, and it is quite possible that the observed scatter on this isochron is entirely due to such effects. (Some scatter, of course, might well be attributed to an initial variation in the $^{87}Sr/^{86}Sr$ ratios due to differential contamination by older crustal material; however, such initial variation could hardly account totally for the degree of scatter now observed, as this would imply that the samples had a range of initial $^{87}Sr/^{86}Sr$ ratios of 0.05 or more).

Alternatively, it might be argued that the observed whole-rock scatter is greater than can be accounted for by the subsequent metamorphism, and that the Norra Kärr body was therefore not isotopically homogeneous at 1580 m.y. before present. If this was so, the 1580 m.y. age would not be so meaningful, simply representing a somewhat fortuitous alignment of points formed by the incomplete reequilibration of the Rb and Sr isotopes during a metamorphic event. It is difficult to estimate the degree of scatter attributable to the later metamorphic effects, but the $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ value for the whole-rock isochron is not extreme, and indeed isochrons with higher values have readily been interpreted as recording igneous events.

If a metamorphic genesis is claimed for Norra Kärr, two questions arise: (a) what was the nature of the rock before metamorphism, and (b) what was the nature of the metamorphic event itself? In answer to the first question, it is difficult to postulate a sedimentary body having the necessary agpaitic composition, and it seems much more likely that the body was igneous, or possibly the result of substantial agpaitic metasomatism (presumably connected with igneous activity). If the $\sim$1580 m.y. age is a metamorphic age, then it must be interpreted as a substantial event, strong enough to produce an almost concordant whole-rock isochron and much more pronounced than the subsequent metamorphic effects recorded by the mineral data. There is no evidence for such an event at this time. The surrounding rocks (Växjö Granite), although heterogeneous and perhaps representing a rather complex history, have yielded a tentative whole-rock isochron age of 1740 m.y. (Welin et al., 1966) and a value of 1695 ± 40 m.y. has been reported for the associated Småland Porphyries (Åberg, 1972) conforming well to ages recorded for the regional Svecofennian regeneration (as described by, for example, Magnusson, 1965); the Norra Kärr event is clearly post-Svecofennian.

The metamorphic event indicated by the mineral data is most likely the Sveconorwegian (Grenville) tectonism, which has yielded several K-Ar ages in the region of 950 m.y. (Magnusson, 1965); indeed, the 1020 m.y. K-Ar age reported by von Eckermann (1968) may reflect this episode. In an attempt to verify von Eckermann's value, two arfvedsonite separates from the lakarpite samples were analysed by the K-Ar method (by Dr. R.M. Macintyre). These gave ages (Table 28) of 1359 and 1456 m.y. (probable error is about ± 2%), consistent with incomplete argon loss during the Grenville metamorphism. The inset map in Fig. 25 indicates the boundary between the Grenville (west) and Svecofennian (east) regions, and shows that it passes just to the west of Norra Kärr, perhaps explaining why the Rb-Sr and K-Ar systems of the mineral phases were not completely reset by the Grenville event. Recent Rb-Sr whole-rock-mineral ages (two-point isochrons) of 1649 ± 24, 1660 ± 24 and 1607 ± 97 m.y. have been obtained on a sample of Växjö Granite collected from the vicinity of Järsnäs, about 40 km due south of Norra Kärr and only slightly further from the supposed Grenville boundary (Patchett, 1976). These data, indicating a late Svecofennian age, show that the mineral systems in this sample could not have been substantially affected by the Grenville metamorphism - further evidence of the border-line position of the Norra Kärr body. Between the Svecofennian and Grenville metamorphic episodes, there is nothing to suggest the existence of a further tectonic episode of the strength needed to produce the 1580 m.y. Rb-Sr isochron of Norra Kärr.

The preferred interpretation of the isotopic data is that the age of 1580 ± 62 m.y. represents an original intrusion age and that subsequent metamorphism has caused slight deviation of whole-rock points about the isochron. This interpretation is consistent with most of the reported field evidence indicating an intrusive nature but dominant metamorphic textural relationships within the complex. The remaining apparent contradictions between textural and mineralogical evidence may possibly be explained by crystallisation from a melt under various pressure conditions (outlined by Sørensen, 1974).

Unfortunately, the initial $^{87}Sr/^{86}Sr$ ratio (0.7072 ± 0.0035) is not sufficiently precise to give information about the possible origin of the agpaitic magma without further information on the surrounding crustal material. It is hoped that further isotopic studies in the immediate country rocks around Norra Kärr and in the Växjö Granite will further elucidate the magmatic and metamorphic history of the area.

Comparison with Rb-Sr data from other agpaitic complexes

It is clear that the 1580 ± 62 m.y. age for Norra Kärr rules out the direct chronological link with Ilimaussaq and the Red Wine complexes that had been postulated before this study (see, for example, Doig, 1970; Vartiainen and Woolley, 1974). Other alkali complexes in Sweden that have been considered as possible contemporaries of Norra Kärr (and Ilimaussaq!) are Särna in west-central Sweden (Magnusson, 1923) and Almunge, to the east of Uppsala (Gorbatschev, 1960), the latter having many petrological similarities to Norra Kärr. Patchett (1976) reports a Rb-Sr mineral isochron age of 247 ± 14 m.y. for Särna, and two K-Ar biotite ages of 1580 and 1780 m.y. (Doig, 1970 and Garbatschev, 1970, respectively) have been obtained for Almunge. However, this last complex may have been metamorphosed (A.Woolley, personal communication, 1976) and so these K-Ar values are not considered reliable. A Rb-Sr isochron study is needed before possible correlations with Norra Kärr can be seriously discussed.

13. RARE EARTH ELEMENT DISTRIBUTION PATTERNS IN THE TUGTUTÔQ YOUNGER GIANT DYKE COMPLEX, GARDAR PROVINCE

Introduction - Significance of anorthosite in the Gardar Province

This chapter discusses a short pilot investigation of the rare earth element (REE) distribution in samples of the gabbro-syenogabbro younger giant dyke complex of Tugtutôq, and of anorthosite inclusions within these dykes.

The majority of Gardar magmas are interpreted as products of differentiation of primitive basaltic parent magmas. The exact nature of these, or of the fractionation mechanisms, is unknown but it has been suggested (Upton, 1974) that the peralkaline character of the observed Gardar rocks could have been achieved by a combination of clinopyroxene and feldspar fractionation. The link between early plagioclase (anorthosite) fractionation and the production of alkaline magmas (Yoder, 1970) is particularly attractive when considering the Gardar situation, as there is widespread evidence of a large mass (or masses) of anorthositic material underlying the exposed intrusions.

Bridgwater (1967) and Bridgwater and Harry (1968) have summarised the available evidence on the nature and occurrence of plagioclase megacrysts and larger anorthosite inclusions, found ubiquitously throughout the Gardar Province. Although they occur most commonly in rocks of hawaiitic and mugearitic composition, they are found also in more salic and more basic rocks. Spatially, they are most concentrated in the Kobberminebugt-Isortoq and Tugtutôq-Narssaq areas, and chronologically they occur throughout Gardar time, being found in Early Gardar olivine dolerite (BD) dykes, and in the latest intrusions (for example, the Ilimaussaq agpaites), but the majority are concentrated within the Late Gardar (Fig. 21) ENE-trending dykes.

Several lines of evidence have suggested a cognate relationship between anorthosite xenoliths and host-rocks (Bridgwater, 1967; Bridgwater and Harry, 1968): (a) the anorthite content of the included plagioclases (An_{80-35}) shows a sympathetic compositional variation with that of the host-rock (the inclusions are characteristically more calcic than the host feldspars); (b) there appears to be a relationship between the size and form of inclusions and the nature of the host-rock, there being a gradational sequence from massive granular anorthosites (found in olivine gabbros) to single, less calcic, plagioclase megacrysts (found typically in the "Big Feldspar Dyke" trachydolerites); (c) two main types of massive anorthosite are found: the first is granular and frequently altered, possibly by the action of volatiles; these have been interpreted as having formed as a flotation cumulate at the top of a cooling magma chamber (Bridgwater and Harry (1968); the second is laminated and shows no such alteration, being interpreted as a bottom cumulate in a relatively well-differentiated Gardar magma (Upton, 1961; Upton, 1964).

The possible correlation with massive Precambrian anorthosite bodies in the Nain and Michikamau areas of Labrador (Emslie, 1969; Morse, 1969) has been discussed by Bridgwater (1967) and Bridgwater and Harry (1968). However, unlike the Gardar inclusions, the anorthosites in Labrador (and in most other world locations) are associated with sub-alkaline rather than alkaline and peralkaline rocks (Emslie, 1973). Although the genetic relationship postulated for the Gardar anorthosites and alkaline rocks seems highly plausible, specific geochemical and isotopic evidence to support this hypothesis has been lacking. Two lines of study seemed worth pursuing:

(a) $^{87}Sr/^{86}Sr$ *isotopic study.* In previous chapters, evidence has been presented that within a number of Gardar intrusions, rocks representing different stages of the differentiation/fractionation process possess, to within the limits of analytical uncertainty, identical initial ^{87}Sr/^{86}Sr ratios, suggesting a minimum of contamination by radiogenically enriched crust during the magma's development. If the anorthosites represent an early crystallising phase within the developing alkaline magma chamber(s) underlying the Gardar Province, then they could be expected to possess the same initial Sr isotopic composition as the host magma (provided, of course, that the time interval between crystallisation of the very low Rb/Sr anorthosites and the relatively Rb-enriched host-rock was not sufficiently large to allow substantial accumulation of ^{87}Sr).

(b) *Comparative REE distribution study.* If the anorthosites and host-gabbros are fractionation products derived ultimately from a single magma, it might be expected that this magma, during differentiation, would show relatively gradational chemical trends. In particular, the REEs tend to concentrate in the residual liquid phase (solid-liquid distribution coefficients for the major rock-forming minerals in the anorthosites, gabbros and syenogabbros are generally lower - often considerably so - than unity). Thus the REE distribution patterns observed in a series of related differentiates should be broadly consistent with (i) crystallisation from a magma with gradually changing REE concentration and (ii) the appropriate solid-liquid distribution coefficients for the crystallising minerals. Any major inconsistencies would tend to contradict the supposed cognate relationship. Furthermore, as feldspars tend to preferentially incorporate Eu^{2+} rather than trivalent REEs, the anorthosites and plagioclase-rich early cumulates would be expected to show a positive Eu anomaly, thus reducing the relative Eu concentration of the residual magma and leaving a substantial negative Eu anomaly for the most differentiated Gardar magmas.

Neither of these two approaches can yield unequivocal answers to the question of a cognate relationship for the Gardar anorthosites. Unrelated rock-types may possess identical or similar $^{87}Sr/^{86}Sr$ ratios, and many varying igneous rock sequences may share similar REE distribution patterns. However, positive evidence from two such lines of enquiry could add substantial weight to the field and petrographic evidence outlined above.

The results of an initial $^{87}Sr/^{86}Sr$ ratio study (undertaken by J. Patchett) are reported by Patchett et al. (in press), and the following pages describe the REE investigation.

Geology of the Tugtutôq younger giant dyke complex

In the Tugtutôq-Narssaq area, anorthosite inclusions are common in the olivine gabbros of the "younger giant dyke complex" (see Fig. 17), occurring particularly in the Assorutit region and in the southern area of the mainland outcrop (Narssaq Gabbro). This dyke complex includes two, presumed related, giant olivine gabbro dykes; a southern dyke some 800 m broad bifurcates some four kilometres WSW of Sigssardlugtoq, and a northern dyke (in general wider than its southern neighbour) runs the whole length of the island, cutting across the Hviddal Dyke ("older giant dyke complex") and in turn being cut by the Tugtutôq Central Complex. Cumulative textures and mineral layering (dipping towards the centre of the dyke) indicate that these rocks are at least to some extent products of *in situ* differentiation. According to Upton (1964) the area has been tilted to the ENE, such that progressively younger cumulates are encountered as one traverses the strike from Itivdlip sarqa to Assorutit. The Narssaq Gabbro outcrop on the mainland is thus interpreted as being a high level sill-like termination of the Tugtutôq dykes, its form probably influenced by the unconformity between the Ketilidian basement and the Gardar supracrustal rocks which crop out in that area.

The predominant rock type is olivine gabbro, consisting (Upton, 1964) of plagiocalse (commonly 50-70%, An_{70-60}) and olivine (15-30%) with subordinate (interstitial) titanaugite, biotite, apatite, alkali feldspar and sulphides. Picritic layers within the banded sequences may have up to 70% (or occasionally even more) olivine, and in such layered sequences the modal proportions of these minerals are, of course, variable. The gabbros are normally chilled against the Julianehåb Granite basement, although in places marginal rheomorphs can be seen and some hybridisation occurs between the dyke selvedge and the granite rheomorphs (such hybridisation rarely extending for more than a metre into the dyke).

Towards the northeast end of the northern dyke, a lenticular central facies of syenogabbro (grading upward into syenite) occurs, believed to have formed by *in situ* differentiation from the gabbroic dyke magma (Upton, personal communication, 1975). Modal compositions are very variable in this strongly layered facies, the predominant minerals being plagioclase (An_{35-20}), augite, biotite, ilmenomagnetite, apatite and olivine.

The anorthosite inclusions from the Assorutit region are laminated (Upton, 1961, 1962, 1964) of the variety believed to have formed by bottom accumulation, and consist of plagioclase (An_{61-55}, 80-100%) with <20%

of ferromagnesian minerals (olivine, clinopyroxene, biotite and chlorite).

The Tugtutôq giant dykes are cut by the "Big Feldspar Dykes" (mainly hawaiitic and mugearitic dykes containing abundant plagioclase inclusions) which in turn are followed by smaller, more salic dykes and finally the Tugtutôq Central Complex and the Narssaq Complex. This latter intrusion, together with the Ilimaussaq Complex to the east, may have been responsible for the observed alteration of the mainland (Narssaq Gabbro) extension of the dyke complex, and for this reason samples used in this study are generally restricted to the island of Tugtutôq.

Samples include three anorthosite inclusions (85964, 50221, 40523), four gabbros from the chilled margins of the northern dyke (86106, 86107, 40488, 85954), a sample of the mainland Narssaq Gabbro (81152), four gabbros from the northern dyke (30650, 30642, 85977, 40470), four from the southern dyke (30658, 30638, 30636, 86055) and two samples of the syenogabbro (85976, 50551). Sample locations are shown in Fig. 17.

REE determinations: (a) A note on REE geochemistry

The rare earth elements comprise metals with atomic numbers 57-71, that is lanthanum, cerium, praesodymium, neodymium, promethium, samarium, europium, gadolinium, terbium, dysprosium, holmium, erbium, thulium, ytterbium and lutetium respectively. These elements form a transition series with successive elements filling the 4f electron shell. All but promethium occur in trace amounts in nature, and all have a very similar chemical behaviour.

The absolute abundance of the REEs in terrestrial matter generally decreases with increasing atomic number and, moreover, the abundance of even atomic numbered elements is greater than that of adjacent odd elements, such that a plot of abundance against atomic number shows a "zig-zag" pattern for most geological samples. This clearly tends to obscure any superimposed geological fractionation effects and so, for convenience, absolute abundances are usually displayed normalised with respect to the appropriate elemental abundances in chondritic meteorites (Coryell et al., 1963), a plot of the logarithm of chondrite normalised ratios against atomic number being the common form of display. Chondrite values normally used are those of Frey et al. (1968) for an average of 9 chondrites, Masuda et al. (1973) for the Leedey chondrite or Nakamura (1974) for an average of 10 "representative" chondrites. In this study, the values of Frey et al. (1968) are used.

All the rare earth elements are commonly trivalent, though some exist in more than one oxidation state (Eu^{2+} being an important example). Relative REE distribution in geological meterials is largely governed by the respective ionic radii ($La^{3+} = 1.22Å$, $Lu^{3+} = 0.99Å$). In the mantle resevoir, the heavier REEs are likely to substitute in the available cation positions ($Fe^{2+} = 0.83Å$ and $Mg^{2+} = 0.78Å$) most easily and so these elements will be relatively depleted in most surface terrestrial samples. Measurement of REE solid-liquid distribution coefficients for common rock-forming minerals (for example, Schnetzler and Philpotts, 1970) indicate that these values are highly variable (being a function of temperature, pressure, bulk composition and oxygen fugacity) and, moreover, they display various degrees of preference for light or heavy REEs. For most major minerals, however, REE distribution coefficients are much less than unity; garnet is one notable exception (tending to concentrate the heavy REEs and thus leave the residual liquid relatively enriched in light REEs) and REEs may also substitute for Ca^{2+} in the apatite lattice. Of the major minerals relevant to this study, only clinopyroxenes are likely to have distribution coefficients approaching unity, and the REE content of the cumulate rocks may be a function, not only of the incoming and outgoing cumulate mineral phases, but also of the composition of the "trapped" liquid in the intercumulus phase.

Europium may exist as Eu^{2+} (Goldschmidt, 1958) and as such is more readily incorporated into the feldspar lattice, thus giving rise to Eu anomalies in rock sequences involving feldspar fractionation. The size of the Eu anomaly is a function, not only of the crystal chemistry, but also of the oxidation state of the crystallising magma (for example, Drake, 1972; Morris et al., 1974), Eu^{2+}/Eu^{3+} increasing with decreasing oxygen fugacity. In this regard, it is noteworthy that the Gardar magmas have moderately low to very low oxygen fugacities (Upton, 1974; Larsen, 1976).

REE determinations: (b) Sample preparation and irradiation

Representative splits of powdered samples (collected by Dr. B.G.J. Upton and prepared in the laboratories of the University of Edinburgh) were made available for this study. Determination of 7 REEs (La, Ce, Sm, Eu, Tb, Yb and Lu) were carried out by thermal neutron activation and gamma-ray spectrometry (Gordon et al., 1968), utilising the photopeaks listed in Table 29.

Approximately 100 mg of each pulverised rock sample and rock standard to be irradiated was accurately weighed into a 1 ml polythene ampoule (pre-cleaned in warm 2N HCl and twice-distilled H_2O) and this was then heat-sealed. Standard REE solutions (containing $\sim$100 μg ml^{-1} REE) had previously been prepared from Johnson Matthey "Specpure" oxides (La_2O_3, CeO_2, Sm_2O_3, Eu_2O_3, Tb_4O_7, Yb_2O_3, Lu_2O_3)by fusing in lithium metaborate and dissolving in dilute HNO_3 (Whitley, 1974; 1976). For each element to be analysed, a suitable quantity of the appropriate standard solution was transferred, by means of an Oxford micro-pipette (capable of delivering 10 μl with a precision of ± 1%) to an ampoule containing approximately 100 mg of Specpure SiO_2 and this was then dried slowly in an oven. This procedure ensured a similar distribution of samples and standards in the ampoules - an important point when considering detector-ampoule geometry during counting. Each ampoule was then identified by means of a number engraved with a hot scriber. The ampoules were individually wrapped in aluminium foil and then further wrapped in packages of about 10 ampoules suitable for inclusion in heavy-duty polythene irradiation bottles. During the course of this study, three irradiations were made; the 18 samples were arbitrarily divided into two groups in order to allow sufficient time for a suitable counting schedule (Group 1 consisted of samples 30642, 40470, 40488, 40523, 50221, 81152, 85964, 85976, 85977, 86055, two separate splits of BCR-1; Group 2 consisted of samples 30636, 30638, 30650, 30658, 35954, 86106, 86107, 85976 (repeat) 40551, two splits of BCR-1) and, in addition, a third irradiation (of Group 1 samples) was necessary to determine La and Sm as counting problems invalidated the results in the first irradiation. Duplicate samples of BCR-1 (USGS basalt standard) were included in both batches as was sample 85976, in order to test accuracy and reproducibility.

Thermal neutron activation was carried out in the reactor at East Kilbride. The polythene irradiation bottle was lowered into the central vertical stringer (CVS) in the core of the reactor and irradiated for three or six hours at full power (300 kW, corresponding to a thermal neutron flux of about 3.6×10^{12} n cm^{-2} sec^{-1}). After six hours irradiation, the polythene ampoules softened, and there was some danger that they might split;

a three hour irradiation time was therefore preferred. Whitley (1974) reported that the variation in flux density across a sample container with 33 samples, each with an attached iron foil monitor, was ± 3.5%. This is regarded as being within the limitations of accuracy imposed by the counting statistics and therefore flux monitors were not used during this investigation.

REE determinations: (c) Counting procedures

After irradiation, the samples remained in the reactor overnight to allow the short-lived activity to decay; subsequently, the samples were stored in lead containers in a radioactive isotope store until first required for counting (usually a period of three days) during which time the major background irradiation (of 15h ^{24}Na) had subsided to a safe handling level.

Just before the first counting period was due to begin, the samples were unwrapped and the aluminium foil carefully removed. Each ampoule was washed in dilute HNO_3 followed by H_2O, this process being repeated until there was no detectable activity on swabs taken from the ampoule. Between counting periods, the samples were stored in a safe located a suitable distance from the laboratory where the detectors are housed.

All counting was carried out on either (a) a 25 cm^3 coaxial Ge(Li) detector or (b) a 40 cm^3 Ge(Li) detector, both having resolution of approximately 2.3 KeV at 1.332 MeV. Both detectors were used either with a Data General Nova 1210 mini-computer programmed for use as a 4000 channel analyser or a PDP-8E mini-computer programmed to operate as two 2000 channel analysers. Data were taken as teletype printout and (on the PDP-8E) as a plot of counts versus channel number. On the Nova 1210, a fast tape punch was used to store data for later plotting on the PDP-8E.

Table 30 lists the counting schedule for the elements analysed. For each group of elements, the counting period was chosen so as to optimise the peak/background ratios. For the short-lived isotopes (La and Sm), each sample and standard was counted twice and the final results averaged; for the other isotopes, only one count was taken. In the case of Ce and Tb, the resolution of the detectors is insufficient to resolve the photopeaks of these elements from those of other elements having similar energies (^{59}Fe and ^{152}Eu, respectively). In such cases, both peaks were measured and another, interference-free, peak of the interfering element was also measured. The ratio of the two peaks of the interfering element were measured on an irradiated sample of a pure standard, thus enabling the required fraction to be "stripped" from the offending combined peak.

Before each day's counting, the detector was calibrated using a ^{226}Ra source and locating the 186, 242, 295, 352, 609, 1126, 1238 and 1764 KeV photopeaks. At the beginning of each set of counts, the sample-to-detector geometry was arranged so as to optimise the count rate while maintaining a detector "dead-time" below 20%. The sample was placed in a perspex sample-holder which can then be slotted into one of several positions in an aluminium cylinder which fits snugly over the detector. Constant counting geometry can then be maintained for all samples and standards during the run. When counting for ^{141}Ce, it is also necessary to count the 143 and 1099 KeV peaks for ^{59}Fe in an irradiated sample of Fe-wire in the same geometrical position as the samples (for peak-stripping purposes).

REE determinations: (d) Calculation of results

After identification of photopeaks for each element being determined (from a plot of total counts versus channel number), the total counts (net peak area above background) was determined from the teletype output. For this purpose, the background was averaged, typically over four or five channels, a small programme being used to calculate the count rate (counts per minute) and standard deviation of each photopeak. The concentration of each element (in ppm) was then calculated (again using a computer for the repetitive calculations) from the following expression:

$$\text{concentration (ppm)} = \frac{(\text{cpm})_{sa} \times e^{\lambda t_{sa}} \times S}{W \times (\text{cpm})_{st} \times e^{\lambda t_{st}}}$$

where, $(\text{cpm})_{sa}$ = counts per minute for sample
$(\text{cpm})_{st}$ = counts per minute for standard
S = weight (in μg) of element in standard
W = weight (in g) of sample
t_{sa} = time after irradiation of sample count
t_{st} = time after irradiation of standard count
λ = decay constant for active nuclide

The programme calculates concentration (ppm) with standard deviation and chondrite-normalised values (Frey et al., 1968) with standard deviation.

For the elements Ce and Tb, it is first necessary to "strip" interfering photopeaks (of Fe and Eu respectively) as outlined above. This procedure is carried out on the count rate (counts per minute) measurements.

REE determinations: (e) Presentation of data

Table 31 lists the REE concentrations determined for the four separate samples of USGS standard basalt BCR-1, together with the standard deviations and the "recommended" values quoted by the USGS (Flanagan, 1973). Standard deviations listed for the four analyses are based solely on counting statistics and do not include other errors that may affect the final accuracy. Such errors could include (a) incorrect standard concentration (considered to be significant in the case of Ce and Tb), (b) inaccuracies of sample weighing (considered negligible), (c) inaccuracies in pipetting standard solutions (likely to be small), (d) flux inhomogeneities during irradiation (less than 3.5%, see above) and (e) uneven distribution of samples in ampoules or variations in counting geometry (attempts to avoid these may not be wholly successful).

The results show that the average of BCR-1 values (Table 31) is, in all but two cases, within about 5% of the figure quoted by Flanagan (1973). For Ce and Tb the values are about 25% high; as this deviation is consistent in two separate irradiations using two independant sets of standards, it is considered likely that this error is attributable to a concentration error in the primary CeO_2 and Tb_4O_7 solutions used. These errors should be borne in mind when considering the REE patterns (Fig. 27).

REE concentrations of the Tugtutôq samples (and of the chondrite values of Frey et al., 1968) are listed in Table 32 and shown graphically (as chondrite normalised values) in Fig. 27. Table 32 does not list individual standard deviations; the general precision is indicated by the replicate data on BCR-1 and the duplicate analyses of sample 85976. Individual samples of the younger giant dyke complex shown in Fig. 27 can be identified by means of the letter at the left hand margin. Also shown in Fig. 27 are chondrite normalised REE data from the Ilimaussaq Complex, calculated from raw data published by Gerasimovskiy and Balashov (1968). The values shown for Ilimaussaq are not single sample data but averages (listed in Table 33) for each particular rock type.

Figs. 28a and 28b show graphically the nature of the europium anomaly in the samples from the younger giant dyke complex. Fig. 28a indicates the ppm excess (or depletion) of Eu, whereas Fig. 28b shows the *degree* of enrichment (Eu/*Eu, that is actual Eu divided by "expected" Eu estimated by interpolation); both figures use La (a "normal" REE for which relatively precise data are available) as the abscissa.

Discussion: (a) REE distribution in the younger giant dyke complex

Any quantitative discussion of REE distribution patterns must involve a detailed study of the petrology and geochemistry of the rock samples, including estimates of the distribution coefficients between liquid and the various crystallising phases at each stage of development. Such a discussion is beyond the scope or intention of the present pilot study, but the data do allow a qualitative appraisal of the likely origin of the anorthosite inclusions, and throw some light on the possible evolution of magma in the Tugtutôq-Narssaq region.

The main features observed in the chondrite normalised REE patterns for the Tugtutôq samples (represented by circular symbols in Fig. 27) are as follows:

(a) All samples are enriched in the light REEs, the chondrite normalised values for the lightest REEs being consistently higher (by an order of magnitude) than the heaviest.

(b) The samples show a fairly evenly distributed range of absolute REE concentrations, increasing by an order of magnitude in the order: anorthosites → gabbro cumulates → gabbro chills → syenogabbros.

(c) There is a very striking positive Eu anomaly in most of the samples, this anomaly decreasing with increasing abundance of REEs. The anomaly is most marked in the anorthosites, shows a gradation from strongly positive to almost zero in the cumulate gabbro samples, is very small or absent in the chill samples, and may even be negative in the syenogabbro cumulates.

Some spurious features of the plot are believed to be largely due to analytical error and uncertainty. These include an apparent positive Ce anomaly (attributable to the consistent error in standard concentration referred to earlier) and an apparent divergence of sample trends between Yb and Lu; this is considered to be due to the often poor counting statistics for Yb.

The gabbro chill samples represent the best available evidence for the composition of the magma which gave rise to the gabbro and syenogabbro cumulates. They are enriched in light REEs (a common feature of alkali basalt magmas) and show a substantial range of REE abundance values, increasing by a factor of ∿2 on the order 86106 → 86107 → 40488 → 85945. Moreover, this REE abundance increase is towards the ENE, accompanying a general chemical trend towards rocks lower in Mg and enriched in all elements incompatible in the olivine and plagioclase lattices (Upton, personal communication, 1976). This is in accord with Upton's (1964) suggestion that the dyke complex has been tilted towards the ENE. The chill samples also show a very slight (but possibly significant) positive Eu anomaly. This probably derives from occasional inclusions of small plagioclase phenocrysts found in all chill samples. If these samples had been totally aphyric, then there may not have been any Eu anomaly, or even a negative one.

The gabbro and syenogabbro cumulates show a wide variation of absolute

REE abundance, exceeding both extremes of the chill sample range; unlike the chills however, there is no simple geographic relationship for these cumulate rocks. The *relative* light REE enrichment also tends to increase slightly with increasing absolute concentration (average La/Yb ratios for the 5 lowest gabbros, the four highest gabbros and the two syenogabbros are 10.6, 15.1 and 16.3 respectively). These observations are entirely consistent with progressive crystallisation from a differentiating magma. Early crystallisation, from a magma similar to that represented by the chill samples and involving mainly olivine and plagioclase, would tend to concentrate relatively little REEs, thus producing a solid fraction with a REE content lower than the chill and enriching the residual liquid in absolute REE concentrations. As crystallisation proceeded, clinopyroxene and apatite became cumulate phases (though the latter rarely exceeds 3%). These minerals, having higher solid-liquid distribution coefficients than the previous phases and crystallising from a magma already enriched in REEs, incorporated higher relative amounts of REEs; such changes may be gradual or abrupt.

Figs. 28a and 28b show the changing relative size of the Eu anomaly with increasing REE concentration. Fig. 28a shows the actual ppm excess of Eu (calculated by interpolating a value for the "expected" Eu concentration based on adjacent REE abundances). Such a calculation is only an approximation, but the plot clearly defines a negative slope, with the largest excess Eu residing in those samples with lowest absolute REE content. Two samples (a gabbro and a syenogabbro) show an apparent negative anomaly (that is, fall below zero on the plot). Though this effect may well be real, it might just derive from the larger errors in calculation involved for interpolated Eu values for the high-REE samples. The absolute excess of Eu in the samples does not provide a totally realistic appraisal of the Eu anomaly as any given amount of Eu would constitute a much larger Eu anomaly in a rock with low REE abundance (anorthosite) than one with a high REE content (syenogabbro). Fig. 28b shows the excess Eu as a ratio to the expected Eu content and well illustrates the progressive decrease of the Eu anomaly with increasing absolute REE content.

Discussion: (b) Relationship between anorthosites and gabbros

Studies by Philpotts et al. (1966) and Green et al. (1972) on anorthosites from Tchitogama Lake, Quebec and Lofoten-Vesteraalen, Norway, respectively, indicate that both these anorthosites, although having different REE distribution patterns, are depleted in REEs relative to the Gardar samples and have a greater relative light REE enrichment (greater by an order of magnitude in the Quebec samples). Such data support the assumption that alien anorthositic xenoliths derived from an unrelated body at depth might be expected to have dissimilar REE distribution patterns to the Gardar host-rocks. However, this is not the case in the Tugtutôq samples, where quite similar REE patterns are observed in anorthosite and gabbro samples. Moreover, Patchett et al. (in preparation) show that, based on analyses of 5 anorthosites and 5 host gabbros, the initial $^{87}Sr/^{86}Sr$ ratios are identical (0.70297 ± 0.00011 and 0.70297 ± 0.00008, respectively). It thus seems entirely reasonable to examine the REE case for a close genetic relationship.

If the magmas represented by the gabbro chill samples had experienced extensive plagioclase fractionation of the type envisaged by Bridgwater (1967), then they would have suffered a substantial Eu depletion, and the REE distribution trace should reflect a negative Eu anomaly. It does not;

within the appropriate analytical uncertainty limits, there is no significant Eu anomaly (possibly a slight positive one) and, although this could be in part attributable to sporadic plagioclase phenocrysts within the chill zone, it is doubtful that a substantial negative anomaly could have been obliterated in all four analysed samples. This suggests that the anorthosites did not originate directly from the magma represented by the host gabbros. Experimental studies reported by Upton (1971) and Patchett et al. (in preparation) have already shown that such a direct relationship is improbable. At pressures of 1-10 kb, olivine is the liquidus phase for magma compositions represented by the gabbro chill; at higher pressures, clinopyroxene replaces olivine, but plagioclase only begins to crystallise 10-15° below the liquidus. However, it has been suggested that at higher pressures (>10 kb), clinopyroxene fractionation of the chill composition could yield a magma which could, at subsequently higher levels, have plagioclase on the liquidus.

Thus field, experimental, Sr isotope and REE data would be consistent with a multi-stage model. An early volume of magma, having evolved by extended clinopyroxene fractionation at depth, subsequently rose to a secondary sub-emplacement level where plagioclase fractionation gave rise to anorthosite accumulation. Later batches of magma then intruded from the same deep source, disrupting the anorthosites and carrying them to the surface.

Discussion: (c) Comparison with REE patterns from Ilimaussaq

Gerasimovskiy and Balshov (1968) determined the abundance of all REEs (except Tm and Lu) by X-ray methods in over 30 representative samples of all the major rock-types of the nearby Ilimaussaq Complex. They concluded that the total REE content (ppm) increased in the sequence: augite syenites (315), naujaite (820), foyaite and sodalite foyaite (1350), kakortokite (1550), green lujavrite (4760), black lujavrite (7750). Total REE contents of the pulaskite and quartz syenite averaged 710 ppm and of the alkali granite 2430 ppm. Moreover, their data indicated an accompanying increase in the relative light REE enrichment in the order: augite syenite → alkali granite → agpaites. However, there are numerous inflexions in this trend if the differentiation sequence is examined in detail. Gerasimovskiy and Balshov (1968) quote values for $\Sigma Ce/\Sigma Y$, defining ΣCe as La+Ce+Nd+Sm+Eu and ΣY as Gd+Tb+Dy+Ho+Er+Yb+Y; unfortunately, these values cannot be used for direct comparison with the Tugtutôq samples as all REEs were not determined for the latter. Consequently, in Table 33, La/Yb ratios are calculated for the mean REE values reported by Gerasimovskiy and Balashov (1968), and these show that the light REE enrichment increases by a factor of ∿3 from the augite syenites to the lujavrites. There is evidence of a similar, but much less pronounced, trend in the Tugtutôq samples (Table 32).

In Chapter 6, the agpaitic rocks of the Ilimaussaq Complex were regarded as having formed by differentiation at depth from a magma similar to the augite syenite envelope into which they were intruded; the possibility of *in situ* differentiation from the augite syenite was rejected on geochemical and isotopic grounds. The REE data of Gerasimovskiy and Balashov (1968) are consistent with this hypothesis, as they indicate a substantial enrichment in REE concentration in the agpaites - more than could be accounted for by crystallisation of the relatively small quantity of augite syenite inferred from present exposures. Gerasimovskiy and Balashov's averaged data has been chondrite normalised and plotted in Fig. 27

for comparison with the Tugtutôq samples. The augite syenite has a REE distribution similar to the later fractionates (gabbros and syenogabbros) of the Tugtutôq dyke complex (perhaps somewhat obscured in the diagram due to the inclusion of extra data points for the Ilimaussaq plot), suggesting that the two rock types may well be derived from similar source materials. Moreover, the augite syenite has a distinct positive Eu anomaly; the agpaites, in contrast, have a moderate to strong negative anomaly, indicating that very substantial Eu depletion preceded differentiation of the agpaitic magma. Again, this depletion could not have resulted from *in situ* crystallisation of the small quantity of augite syenite (the REE content of which is modest by comparison to the agpaites) and fractionation at depth is indicated.

The Ilimaussaq agpaites crop out a mere 10 km to the east of the Assorutit area of Tugtutôq and certainly they were intruded very soon after the younger giant dyke complex (Ilimaussaq is dated at 1168 ± 21 m.y. and the younger dykes are bracketed by this age and the older Hviddal Dyke at 1175 ± 9 m.y.). It seems highly likely, therefore, that the xenolithic feldspar fractionates that have been displayed so prominently on Tugtutôq are related to the intense feldspar fractionation which is inferred by the REE patterns to have occurred at Ilimaussaq. The combination of REE and Sr isotopic studies (Chapter 6) enable speculation on the development and intrusion of magmas in this area. Following the intrusion of the younger giant dyke complex (and the Narssaq Complex), the first, augite syenitic, intrusive pulse of Ilimaussaq magma may have just cut through the easterly end of the differentiating (plagioclase fractionating) magma chamber that underlay Tugtutôq and Narssaq. This would enable a predominantly augite syenitic magma to mix with large amounts of Eu-depleted liquid under Ilimaussaq. In the roof zone of this enlarged eastern part of the magma chamber, liquids differentiating towards agpaitic compositions could react with the pre-heated basement rocks and preferentially leach ^{87}Sr (Chapter 6). According to this hypothesis, the differentiation and radiogenic Sr enrichment of the agpaites took place *after* intrusion of the augite syenite.

To summarise, the REE data support previous suggestions (for example, Upton, 1974) that the peralkalinity of the Gardar magmas is intimately related to intense plagioclase fractionation and indicates that, as suggested by the close temporal and spatial relationship, the Late Gardar intrusions of the Tugtutôq-Narssaq-Ilimaussaq area must be regarded as products of a large interrelated system of magma generation and evolution.

14. SUMMARY OF CONCLUSIONS

Rb-Sr study of Gardar central complexes

Precise Rb-Sr whole-rock isochron studies (data summarised in Table 21) of 13 separate alkaline intrusive centres in the Precambrian Gardar igneous province of south Greenland reveal that Gardar magmatic activity in this region extended over the period 1330-1150 m.y. before present. This activity is divisible into three fairly well-defined phases, referred to respectively as Early (∿1300 m.y.), Mid (∿1250 m.y.) and Late (∿1165 m.y.) Gardar. The Early Gardar is represented by the Eriksfjord Formation of supracrstal basalts and sandstones (now observed only in the easternmost parts of the province), and by the nepheline syenite intrusions at Grønnedal-Ika (west) and the two oldest centres at Igaliko (east). In addition, at least some of the early basic dykes belong to this phase of activity. The Mid Gardar period is more widely represented by lamprophyre and olivine dolerite dykes (Patchett, 1976), but only two central-type complexes (the Ivigtut Granite and Kûngnât syenite intrusions), both in the west of the region, were emplaced at this time. The Late Gardar phase was responsible for the majority of the central complexes; this includes all the major units at Nunarssuit, the central complex on Tugtutôq together with the giant dyke system, the Narssaq Complex, Ilimaussaq, the last two centres at Igaliko and the Klokken intrusion. In addition to this central-type intrusive activity, most of the widespread ENE dykes were emplaced at this time.

The initial $^{87}Sr/^{86}Sr$ ratios of the central complexes can be divided into three groups. Low ratios (0.702 - 0.704) characterise the majority of the gabbros, syenogabbros and syenites and suggest that the parental Gardar magmas were derived from a primitive (mantle) source and suffered little or no crustal contamination and did not experience a prolonged pre-cooling history with elevated Rb/Sr ratios. Low ratios are also reported for the Narssaq ultramafic bodies and the anorthositic inclusions found in the Tugtutôq giant dykes (Patchett et al., in press).

Intermediate (0.704 - 0.707) ratios are found in some complexes, particularly the granites of the Nunarssuit Complex. Here it is suggested that contamination of a low $^{87}Sr/^{86}Sr$ magma by older crust explains the elevated ^{87}Sr content; this hypothesis is supported by field and chemical evidence. The possibility that such enrichment in radiogenic Sr was produced *in situ* during a lengthy cooling history is rejected as implausible.

High initial ratios (above 0.707) occur in two complexes, Ilimaussaq and Ivigtut. At Ilimaussaq (0.710), the high initial ^{87}Sr content is restricted to the later agpaitic rocks, the earlier augite syenite "shell" having typically low Gardar values. On the basis of field and chemical evidence, bulk assimilation of crustal material can be rejected, and the high radiogenic Sr content of the agpaites is believed to result from the preferential leaching by the highly reactive agpaitic magma, of ^{87}Sr from unstable Rb lattice sites in the granitic basement. A similar explanation is proposed for the high ($\sim$0.712) initial ratio for the Ivigtut Granite. Here, an interstitial cryolite phase has a significantly higher ($\sim$0.720) ratio than the whole-rock granite, believed to have resulted from circulation through and reaction with the surrounding Archaean basement during cooling of the granite. The famous Ivigtut cryolite body itself has an initial ratio ($\sim$0.709) slightly lower than the granite and was thus not directly responsible for the ^{87}Sr enrichment.

The Rb-Sr data are consistent with the hypothesis that the Gardar central complexes evolved by fractional crystallisation of an alkali basalt parent during an extended period of tectono-magmatic activity. The respective association of basic dyke emplacement, faulting and central (salic) intrusion are believed to reflect ancient phases of continental rifting.

Rb-Sr studies of similar agpaitic complexes

A whole-rock isochron for the Red Wine agpaitic complexes of central Labrador, which in pre-drift times were situated some 500 km to the west of the Gardar Province and on the opposite side of the Grenville "Front", yields an age of ∿1350 m.y., indicating that they may well have belonged to the Early Gardar phase of magmatism. However, two mineral isochrons (∿1000 m.y.) indicate that the complexes have been affected by the Grenville metamorphism.

A similar study of the contrversial Norra Kärr body in southern Sweden (also near the Grenville or Sveconorwegian Front, but some 2000 km distant from the pre-drift position of the Gardar region), suggests an intrusion age of ∿1580 m.y. and again the mineral systems reflect the Grenville metamorphism. The magmatic origin of the complex, recently challenged, is strongly supported by the Rb-Sr data.

Rare earth element study of Tugtutôq younger giant dyke complex

A study of REE distribution in anorthosite inclusions, gabbros and syenogabbros of the younger giant dyke complex support a cognate relationship between anorthosites and host-rocks. All samples are light REE enriched and show a progressively higher absolute concentration of REEs in the sequence: anorthosites → gabbros → syenogabbros. A slight positive Eu anomaly in the chilled gabbro samples indicates that it is unlikely that the anorthosites (large positive Eu anomaly) crystallised directly from this magma but rather that they crystallised from a related body of magma which had earlier risen into the crust, undergone plagioclase fractionation, and was then disrupted by the intruding gabbro magmas; this hypothesis is consistent with field, experimental and Sr isotopic evidence. The REE patterns of the nearby Ilimaussaq agpaites (Gerasimovskiy and Balashov, 1968) show a similar enrichment in the light REEs, and a strong negative Eu anomaly confirms that they could not have fractionated directly from the marginal augite syenite magma, but must be derived from a magma that has undergone extensive feldspar fractionation. The REE data support other lines of evidence suggesting a large interrelated system of magma generation and injection during Late Gardar time.

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Table 1

Internal Laboratory Standard (Hviddal Dyke, GGU 86035)

Duplicate analyses

Date	Analyst	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
01-04-70	1	219.0	19.40	34.6100	1.27610
30-08-70	1	226.0	20.30	34.0700	1.27330
22-11-72	1	226.83	19.765	34.8012	1.27457
		226.40	19.695	34.8904	1.27703
02-07-73	1	226.13	19.850	34.7947	1.27705
		225.94	19.921	34.6377	1.27524
19-02-74	2	227.14	19.774	35.0980	1.28076
		226.85	19.687	35.2086	1.28114
15-03-75	2	227.01	19.661	35.2760	1.27997
		228.75	19.735	35.4137	1.28003

Analysts: 1 O.van Breemen and J.Hutchinson
2 A.B.Blaxland

Table 2

Eimer and Amend (MIT) Standard $SrCO_3$

Measurements made during period of study

Date	$^{87}Sr/^{86}Sr$	Std. Error	Date	$^{87}Sr/^{86}Sr$	Std. Error
25-09-73	0.70818	0.00003	01-07-74	0.70816	0.00003
27-09-73	0.70821	0.00004	16-07-74	0.70819	0.00001
09-10-73	0.70818	0.00005	17-07-74	0.70814	0.00001
08-11-73	0.70817	0.00003	21-08-74	0.70815	0.00003
08-11-73	0.70817	0.00005	21-08-74	0.70809	0.00002
06-12-73	0.70816	0.00003	07-09-74	0.70800	0.00004
07-01-74	0.70812	0.00004	07-09-74	0.70806	0.00005
30-01-74	0.70821	0.00005	07-09-74	0.70798	0.00003
01-03-74	0.70810	0.00005	07-09-74	0.70796	0.00003
29-03-74	0.70808	0.00003	07-09-74	0.70807	0.00003
31-03-74	0.70809	0.00003	07-09-74	0.70798	0.00004
31-03-74	0.70817	0.00003	07-09-74	0.70793	0.00004
31-03-74	0.70809	0.00001	09-12-74	0.70804	0.00004
01-04-74	0.70808	0.00002	13-03-75	0.70810	0.00003
25-04-74	0.70811	0.00009	14-05-75	0.70804	0.00009
10-05-74	0.70816	0.00005	04-07-75	0.70814	0.00003
25-06-74	0.70809	0.00005	04-07-75	0.70812	0.00003
27-06-74	0.70813	0.00002	04-07-75	0.70818	0.00004
28-06-74	0.70818	0.00003	04-07-75	0.70806	0.00004

Mean of 38 analyses = 0.70811 ± 0.00007 (1σ)

Analyst: J.Hutchinson

Table 3

NBS Standard Feldspar (SRM 607)

Measurements made during period of study

Date	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
21-05-74	513.0	64.81	7.915	24.00	1.2000
	518.9	64.82	8.006	24.28	1.2003
28-05-75*	524.7	65.41	8.022	24.33	1.2013
	523.3	65.38	8.004	24.28	1.2007
	521.5	65.35	7.979	24.20	1.2011

* Heated to 105°C for 3 h and cooled in dessicator before weighing

NBS values: Rb = 523.90 ± 1.01 ppm
Sr = 65.49 ± 0.32 ppm
$^{87}Sr/^{86}Sr = 1.20039 \pm 0.00020$
(errors at 2σ)

Analysts: O.van Breemen and J.Hutchinson

Table 4

Routine repeat analyses made during period of study

Sample No.	Location	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Percent variation $^{87}Rb/^{86}Sr$	Percent variation $^{87}Sr/^{86}Sr$
54186	Motzfeldt	1.5351 1.5572	0.73104 0.73089	1.4	0.02
63720	Motzfeldt	17.7464 17.7793	1.01849 1.00416	0.2	1.40
58028	Motzfeldt	25.9430 26.1495 26.2342	1.17338 1.17344 1.18435	1.1	0.90
59757	North Qôroq	36.4700 35.9349	1.35287 1.34630	1.5	0.50
58245	South Qôroq	13.3297 13.3171	0.94399 0.94369	0.1	0.03
41974	Igdlerfigssalik	21.3291 21.5606	1.05563 1.05671	1.1	0.10
154267	Nunarssuit Syenite	1.942 1.953	0.7358 0.7359	0.6	0.01
154434	Nunarssuit B.Granite	2.6875 2.6791	0.75031 0.75012	0.3	0.03
LD-41	Red Wine (Labrador)	7.0211 6.9828	0.84485 0.84387	0.5	0.10
86035	Hviddal Dyke	35.0980 35.2086	1.28074 1.28114	0.3	0.03
86035	Hviddal Dyke	35.2760 35.4137	1.27997 1.28003	0.4	0.00
140053	Klokken	0.0263 0.0290	0.70339 0.70353	10.3	0.02
140081	Klokken	0.2194 0.2160	0.70668 0.70652	1.5	0.02
150782	Ilimaussaq	0.7830 0.7783	0.71661 0.71637	0.6	0.03

continued

Table 4 (continued)

Sample No.	Location	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Percent variation $^{87}Rb/^{86}Sr$	 $^{87}Sr/^{86}Sr$
SW-19	Norra Kärr	101.2610 100.6090	2.68831 2.66819	0.6	0.75
P8-841	Red Wine (Labrador)	60.5470 59.6072	1.81488 1.80897	1.5	0.33

Table 5

Previous geochronological and isotopic data from Gardar rocks

Locality	Isotopic technique	Age *	$(^{87}\mathrm{Sr}/^{86}\mathrm{Sr})_o$	Ref
Kûngnât	Rb-Sr biotite	1240 ± 150		1
"	Rb-Sr wr isochron	1233 ± 16	0.7071 ± 0.0021	2
Ivigtut (granite)	Rb-Sr wr (mean of 2)	1330 ± 20		3
" "	Rb-Sr biotite (m.o.3)	1261 ± 20		3
" "	K-Ar biotite (m.o.2)	1187		3
" "	Rb-Sr baryto-celestite		0.7077 ± 0.0015	3
" "	" " (replicate)		0.7094 ± 0.0015	3
" "	" " (replicate)		0.7071 ± 0.0010	3
Grønnedal-Ika	K-Ar biotite	1210 ± 30		4
"	K-Ar biotite	1170 ± 30		4
Nunarssuit Syenite	K-Ar biotite	1128 ± 30		5
" "	K-Ar augite	1355 ± 150		5
Nunarssuit b.granite	Rb-Sr biotite	1211 ± 30		6
Bangs Havn	Rb-Sr wr isochron	1211 ± 11	0.7034 ± 0.0001	7
Gardar Dykes:				
Qaersuarssuk^1	K-Ar augite	1435 ± 80		5
Ivigtut^2	K-Ar biotite	1275 ± 25		8
Mato^1	K-Ar wr	1066 ± 32		9
Narssaq^3	K-Ar biotite	1025 ± 70		5
"	K-Ar augite	1075 ± 50		5
Arsuk^2	Rb-Sr wr-phlogopite	1254 ± 29	0.7051 ± 0.0006	10

continued

Table 5 (continued)

Locality	Isotopic technique	Age *	$(^{87}Sr/^{86}Sr)_o$	Ref
Arsuk[2]	Rb-Sr wr-phlogopite	1276 ± 29	0.7036 ± 0.0006	10
Ivigtut (BD_1)[1]	Rb-Sr wr-biotite	1265 ± 18	0.7027 ± 0.0006	10
" "	" (replicate)	1263 ± 18	0.7027 ± 0.0006	10
Ivigtut (BD_0)[1]	Rb-Sr wr-biotite	1272 ± 18	0.7050 ± 0.0006	10
Ivigtut (BD_0)[1]	Rb-Sr wr-biotite	1277 ± 18	0.7061 ± 0.0006	10
" "	" (replicate)	1278 ± 18	0.7061 ± 0.0006	10
Tugtutôq (BD_0)[1]	Rb-Sr wr-biotite	1245 ± 18	0.7046 ± 0.0006	10
Julianehåb (BD_0)[1]	Rb-Sr wr-biotite	1238 ± 18	0.7046 ± 0.0006	10
Hviddal Dyke	Rb-Sr wr isochron	1175 ± 9	0.7024 ± 0.0010	2
Tugtutôq Cent. Comp.	Rb-Sr wr isochron	1168 ± 37	0.6983 ± 0.0139	2
Ilimaussaq	Rb-Sr polylithionite	1095 ± 24		1
"	" "	1077 ± 24		1
"	" "	1086 ± 24		1

* All Rb-Sr ages are based on a decay constant $\lambda = 1.39 \times 10^{-11}$ y^{-1} (thus some ages listed here are 6% higher than originally quoted in the source reference.

[1] Dolerite

[2] Lamprophyre

[3] Anorthosite inclusion

References: 1. Moorbath et al. (1960)

2. van Breemen and Upton (1972) (values of ages from this paper have been reduced by 1% for reasons explained in Chapter 3).

3. Moorbath and Pauly (1962)

Table 5 (continued)

References (continued):

4. Larsen (1969)
5. Bridgwater (1965)
6. Harry and Pulvertaft (1963)
7. Engell and Pedersen (1974)
8. Larsen and Møller (1968)
9. Bridgwater (1970)
10. Patchett (1976)

Table 6

Rb-Sr whole-rock data for Kûngnât

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$(^{87}Sr/^{86}Sr)_o$
81145	77.1	278.4	0.277	0.8021	0.71748	-
81127	74.3	134.0	0.555	1.6091	0.73230	-
86193	97.2	109.3	0.888	2.5812	0.74971	-
81103	107.3	52.4	2.046	5.9774	0.80902	-
86176	238.3	18.8	12.706	39.1939	1.38416	-
* 86182	10.1	570.7	0.018	0.0512	0.70478	0.70388
* 86186	10.4	555.7	0.019	0.0541	0.70456	0.70362

* Samples of gabbro ring-dyke not included on isochron. Initial ratio calculations based on an assumed age of 1245 m.y.

Table 7

Rb-Sr whole-rock data for Grønnedal-Ika

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr	Age (m.y.)
Carbonatite						
31863	22.1	14793.1	0.005	0.0043	0.70326	-
Syenites						
27074	102.3	1100.8	0.093	0.2690	0.70789	-
27207	98.9	684.0	0.145	0.4183	0.71139	-
27193	70.6	427.3	0.165	0.4781	0.71214	-
27179	107.3	400.0	0.268	0.7769	0.71758	-
27205	162.8	489.6	0.333	0.9637	0.72150	
* 39763A	161.4	236.2	0.683	1.9630	0.73491	-
58315	190.7	212.7	0.897	2.6067	0.75131	-
27168	224.6	232.3	0.967	2.8113	0.75522	-
126706	233.2	211.8	1.101	3.2024	0.76171	-
27088	277.1	244.7	1.132	3.2947	0.76537	-
27170	276.9	194.4	1.424	4.1508	0.78122	-
** *Late Alkaline Dykes*						
39754	191.1	35.2	5.435	16.1058	0.95467	1115
27151	189.4	29.2	6.476	19.4078	1.07383	1362
39794	280.9	37.0	7.591	22.9073	1.14706	1381
27007	313.8	30.3	10.364	31.6437	1.27337	1285

* Point lies off isochron and is not included in regression treatment.

** Late alkaline dyke samples are not included in isochron data. Age calculations based on an assumed initial ratio of 0.703.

Table 8

Rb-Sr whole-rock data for Alángorssuaq Gabbro

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
131974	1.3	211.7	0.006	0.0175	0.70390
131856	8.8	596.6	0.015	0.0425	0.70403
131856 (r)	9.4	554.2	0.017	0.0489	0.70423
* 131952	24.5	556.0	0.041	0.1274	0.70609
131970	24.0	272.6	0.088	0.2544	0.70758
* 131967	16.7	157.3	0.106	0.3075	0.70901
131850	110.6	746.2	0.148	0.4287	0.71039
* 131951	40.4	241.1	0.167	0.4844	0.71211
* 131857	101.9	562.1	0.181	0.5241	0.71297

r = repeat analysis.

* = point not included in regression treatment.

Table 9

Rb-Sr whole-rock data for Helene Granite

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
131926	85.1	195.5	0.455	1.2612	0.72407
154530	136.1	82.5	1.649	4.8056	0.78264
* 154512	210.4	67.2	3.129	9.1712	0.83948
154566	138.4	40.1	3.449	10.1322	0.86410
* 154347	140.1	25.6	5.470	16.2025	0.95034
154590	184.6	24.2	7.627	22.8582	1.07481

* = point lies off isochron and is not included in regression treatment.

Table 10

Rb-Sr whole-rock data for Nunarssuit Biotite Granite

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
154434	138.7	149.7	0.926	2.6875	0.75031
154434 (r)	138.2	149.6	0.925	2.6791	0.75012
154404	142.2	126.6	1.123	3.2603	0.75950
154411	143.4	122.2	1.174	3.4155	0.76316
* 154401	167.2	142.7	1.172	3.4086	0.76527
154415	181.7	97.8	1.859	5.4230	0.79487
131993	201.2	97.3	2.069	6.0427	0.80577
154418	116.9	52.5	2.226	6.4913	0.81154
131997	171.6	65.8	2.609	7.6257	0.83123

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Table 11

Rb-Sr whole-rock data for Nunarssuit Syenite

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
154498	63.3	295.0	0.215	0.6206	0.71443
154232	81.6	269.4	0.303	0.8767	0.71845
154487	79.9	244.4	0.327	0.9462	0.71948
154325	85.2	190.7	0.447	1.2934	0.72528
154267	95.7	142.0	0.674	1.9526	0.73587
154267 (r)	95.1	142.0	0.670	1.9418	0.73576

r = repeat analysis.

Table 12

Rb-Sr whole-rock data for Narssaq Complex

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$(^{87}Sr/^{86}Sr)_o$
186679	59.4	611.7	0.097	0.2811	0.70838	0.7038
186680	63.5	535.9	0.119	0.3429	0.70979	0.7042

Initial ratio calculations based on an assumed age of 1170 m.y.

Table 13

Rb-Sr whole-rock data for Motzfeldt

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
58017A (SM.5)	101.2	1454.8	0.070	0.2013	0.70651
58017 (SM.5)	142.5	583.4	0.244	0.7073	0.71454
58039 (SM.4)	212.9	435.0	0.489	1.4188	0.72855
58025 (SM.5)	144.1	278.3	0.518	1.5015	0.73011
54146 (SM.1)	176.6	333.6	0.529	1.5351	0.73104
54146 (SM.1) (r)	179.2	333.7	0.537	1.5572	0.73089
58024 (SM.5)	187.0	341.6	0.547	1.5872	0.73072
54134 (SM.1)	239.3	150.4	1.591	4.6384	0.78694
58069 (SM.4)	226.4	107.6	2.104	6.1481	0.81266
58400 (SM.3)	185.7	72.6	2.558	7.4984	0.84831
63721 (SM.3)	69.5	24.0	2.896	8.5080	0.86189
63720 (SM.3)	200.5	33.7	5.953	17.7464	1.01849
58028 (SM.4)	235.3	27.4	8.576	25.9430	1.17338
58028 (SM.4) (r)	235.6	27.3	8.644	26.1495	1.17344
58028 (SM.4) (r)	236.9	27.3	8.678	26.2342	1.18435

r = repeat analysis.

Intrusive units (SM.1 - SM.5) are identified according to the scheme of Emeleus and Harry (1970).

Table 14

Rb-Sr whole-rock data for North Qôroq

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
59764 (SN.2)	88.7	1027.0	0.086	0.2498	0.70759
155097 (SN.2)	94.4	683.0	0.138	0.4000	0.71016
59762 (SN.4)	133.6	411.8	0.324	0.9298	0.72062
52226 (SN.1)	104.2	323.2	0.322	0.9344	0.72366
54185 (SN.4)	207.4	325.5	0.637	1.8488	0.74679
52246 (SN.1)	216.1	108.3	1.996	5.8302	0.80672
155036 (SN.1)	250.5	77.8	3.221	9.4705	0.87476
52221 (SN.1)	231.0	53.6	4.309	12.7537	0.94200
155005 (SN.1)	236.6	34.6	6.821	20.4760	1.07720
59757 (SN.1)	403.4	34.0	11.857	36.4700	1.35287
59757 (SN.1) (r)	403.3	34.5	11.690	35.9349	1.34630

r = repeat analysis.

Intrusive units (SN.1 - SN.4) are identified according to the scheme of Emeleus and Harry (1970).

Table 15

Rb-Sr whole-rock data for South Qôroq

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr
58220 (SS.4B)	50.3	1412.9	0.036	0.1030	0.70490
59673 (SS.4B)	100.9	1267.2	0.080	0.2304	0.70690
127025 (SS.4A)	122.8	361.0	0.340	0.9851	0.71942
58150 (SS.4A)	148.2	247.1	0.600	1.7393	0.73127
46269 (SS.3)	187.9	178.0	1.056	3.0673	0.75371
58269 (SS.3)	193.4	138.1	1.400	4.0779	0.76994
127075 (SS.2)	195.3	97.1	2.011	5.8706	0.80147
46255 (SS.2)	206.5	97.8	2.111	6.1671	0.80507
46250 (SS.2)	194.5	75.4	2.580	7.5489	0.82802
58267 (SS.2)	200.6	63.1	3.181	9.3381	0.85710
* 58245 (SS.2)	372.9	82.8	4.504	13.3297	0.94399
* 58245 (SS.2) (r)	305.0	67.8	4.499	13.3171	0.94369
52235 (SS.2)	198.6	40.6	4.892	14.4832	0.94519

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Intrusive units (SS.1 - SS.4) are identified according to the scheme of Emeleus and Harry (1970).

Table 16

Rb-Sr whole-rock data for Late Igdlerfigssalik

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
87111 (SI.4)	61.7	903.1	0.068	0.1977	0.70650
43961 (SI.7)	67.6	651.0	0.104	0.3005	0.70817
43803 (SI.5)	73.9	76.9	0.962	2.7928	0.74714
43839 (SI.5)	224.6	210.7	1.066	3.0992	0.75328
43957 (SI.5)	87.7	63.0	1.392	4.0523	0.76940
43838 (SI.7)	148.2	84.0	1.763	5.1406	0.78730
43857 (SI.5)	91.0	43.7	2.084	6.0841	0.80214
* 41921 (SI.5)	186.4	83.2	2.240	6.5388	0.79288
41974 (SI.6)	253.4	35.5	7.138	21.3291	1.05563
41974 (SI.6) (r)	255.7	35.5	7.203	21.5606	1.05671

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Intrusive units (SI.4 - SI.7) are identified according to the scheme of Emeleus and Harry (1970).

Table 17

Rb-Sr whole-rock data for Klokken

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
140053	12.2	1340.7	0.009	0.0263	0.70339
140053 (r)	13.6	1352.7	0.010	0.0290	0.70353
140081	57.3	755.1	0.076	0.2194	0.70668
140081 (r)	52.3	700.4	0.075	0.2160	0.70652
140183	69.6	373.1	0.186	0.5397	0.71196
140018	70.7	180.5	0.393	1.1352	0.72213
* 140194	20.5	26.8	0.765	2.2203	0.73720
140049	50.2	61.7	0.813	2.3599	0.74128
140138	38.3	37.5	1.022	2.9697	0.75068
140122	86.2	78.2	1.102	3.2039	0.75477
* 140026	103.4	99.4	1.040	3.0249	0.75676
140047	52.5	25.2	2.086	6.0910	0.80270
140093	118.1	55.8	2.114	6.1748	0.80374
140128	57.5	23.9	2.409	7.0438	0.81771

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Table 18

Rb-Sr whole-rock data for Ilimaussaq

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr
Augite syenites					
150781	132.7	889.2	0.149	0.4321	0.71345
150782	94.0	347.5	0.270	0.7830	0.71661
150782 (r)	93.6	348.4	0.269	0.7783	0.71637
150783	98.4	337.8	0.291	0.8433	0.71718
Kakortokites					
150792	239.7	521.5	0.460	1.3331	0.73172
150728	222.1	139.7	1.590	4.6354	0.78849
150731	122.4	74.7	1.638	4.7746	0.78589
150734	194.9	119.8	1.626	4.7402	0.78573
150725	274.6	54.7	5.017	14.8608	0.94940
150732	532.7	89.3	5.968	17.7607	0.99917
150791	519.2	69.4	7.477	22.3976	1.06832
Pulaskite					
150685	207.9	20.3	10.257	31.1526	1.21645
Lujavrites					
150795	593.3	60.8	9.755	29.5947	1.20450
150794	644.0	56.9	11.315	34.5918	1.28653
Alkali granite					
* 86339	390.3	21.1	18.529	58.4952	1.63932

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Table 19

Regression parameters for data sets from Ilimaussaq

Samples used *	No. of samples	Age (m.y.)	Initial ratio	$[SUMS/(n-2)]^{\frac{1}{2}}$
All samples	13	1188 ± 30	0.7060 ± 0.0021	5.63
All samples except augite syenites	10	1168 ± 21	0.7096 ± 0.0022	3.65
Kakortokites only	7	1150 ± 28	0.7105 ± 0.0023	3.32

* Sample 86339 (alkali granite) is excluded from all three regression treatments.

Table 20

Rb-Sr whole-rock and mineral data for the Ivigtut Granite

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$(^{87}Sr/^{86}Sr)_0$
Granite whole-rock samples						
186730	443.4	178.7	2.482	7.2734	0.84038	-
186733	635.7	215.7	2.947	8.6558	0.86343	-
186732	663.4	169.2	3.922	11.5744	0.91253	-
186729	647.1	128.0	5.055	15.0080	0.97590	-
186735	477.8	79.4	6.017	17.9637	1.03435	-
186699	373.9	54.4	6.879	20.6030	1.06727	-
186700	298.2	30.0	9.925	30.2415	1.25052	-
186703	387.2	31.6	12.247	37.7234	1.36868	-
186701	493.8	40.1	12.308	37.9121	1.36807	-
Ore samples						
Cryolite A	0.1	5.9	0.016	0.0472	0.71062	0.7098
Cryolite B	0.1	38.0	0.003	0.0095	0.70894	0.7088
Fluorite	69.7	539.6	0.129	0.3740	0.71611	0.7096
Cryolite separates						
186730(a)	10.7	425.7	0.025	0.0725	0.72083	0.7196
186730(b)	37.0	473.1	0.078	0.2267	0.72360	0.7197

Initial ratio calculations based on an assumed age of 1248 m.y.

Sample 186730(a) was separated from a 61-106 μ size fraction; 186730(b) was separated from a 106-145 μ size fraction.

Table 21

Rb-Sr whole-rock and mineral data for Narssaq ultramafic bodies

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$(^{87}Sr/^{86}Sr)_o$
85913 (wr)	108.46	600.4	0.181	0.52286	0.71152	0.70294
85913 (px)	0.79	695.5	0.001	0.00330	0.70289	0.70284
30770 (wr)	136.78	376.9	0.363	1.05115	0.71942	0.70218
30770 (px)	1.92	450.2	0.004	0.01233	0.70271	0.70251
40407 (wr)	12.79	460.7	0.028	0.08031	0.70479	0.70347
40407 (px)	0.50	327.2	0.002	0.00438	0.70317	0.70310
40407 (ap)	0.08	1568.7	0.000	0.00014	0.70319	0.70318
85941 (px)	1.39	424.0	0.003	0.00945	0.70345	0.70330
85941 (ap)	1.69	1771.9	0.000	0.00276	0.70323	0.70319

wr = whole-rock

px = clinopyroxene

ap = apatite (with some olivine)

Initial ratio calculations based on an assumed age of 1170 m.y.

Table 22

Summary of Rb-Sr whole-rock isochron data for Gardar central complexes

Intrusive centre	No. of samples	$[SUMS/(n-2)]^{\frac{1}{2}}$	Age (m.y.)	$(^{87}Sr/^{86}Sr)_o$
KÛNGNÂT	5	1.50	1245 ± 17	0.7041 ± 0.0008
IVIGTUT GRANITE	9	2.76	1248 ± 25	0.7125 ± 0.0048
GRØNNEDAL-IKA	11	1.26	1327 ± 17	0.7032 ± 0.0002
NUNARSSUIT				
Alángorssuaq Gabbro	4	0.32	1143 ± 48	0.7035 ± 0.0002
Helene Granite	4	3.04	1149 ± 31	0.7040 ± 0.0024
Nunarssuit B. Granite	7	1.16	1162 ± 21	0.7068 ± 0.0013
Nunarssuit Syenite	5	0.27	1154 ± 14	0.7043 ± 0.0002
HVIDDAL DYKE *	8	3.88	1175 ± 9	0.7024 ± 0.0010
TUGTUTÔQ CENTRAL COMPLEX *	6	10.41	1168 ± 37	0.6983 ± 0.0139
ILIMAUSSAQ	10	3.65	1168 ± 21	0.7096 ± 0.0022
IGALIKO				
Motzfeldt	12	4.25	1310 ± 31	0.7024 ± 0.0015
North Qôroq	10	8.51	1295 ± 61	0.7052 ± 0.0030
South Qôroq	11	1.09	1185 ± 8	0.7029 ± 0.0004
Igdlerfigssalik	8	1.63	1167 ± 15	0.7029 ± 0.0007
KLOKKEN	10	1.66	1159 ± 11	0.7031 ± 0.0003

* Values from van Breemen and Upton (1972); ages reduced by 1% for reasons explained in Chapter 3. Values in third column are for MSWD (defined by McIntyre et al., 1966).

$\lambda_{Rb} = 1.39 \times 10^{-11}\ y^{-1}$. All uncertainties are quoted at 2σ.

Table 23

Rb-Sr whole-rock data for Red Wine Complexes (Reconnaissance samples)

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr
LD-37	16.3	172.2	0.095	0.2742	0.71282
LD-38	31.5	119.2	0.264	0.7665	0.72057
LD-40	31.6	62.6	0.503	1.4661	0.75569
LD-36	262.4	177.4	1.479	4.3156	0.78936
LD-35	280.1	176.0	1.591	4.6425	0.79358
LD-35 (r)	285.2	155.8	1.831	5.3486	0.80451
LD-35 (r)	294.4	152.8	1.927	5.6278	0.80761
LD-42	155.6	88.1	1.765	5.1246	0.79673
LD-42 (r)	162.7	85.9	1.894	5.5323	0.80179
LD-42 (r)	165.1	84.6	1.952	5.7011	0.80271
LD-41	206.8	86.8	2.382	6.9828	0.84387
LD-41 (r)	203.6	85.0	2.395	7.0211	0.84484

r = repeat analysis.

Table 24

Rb-Sr whole-rock and mineral data for Red Wine Complexes

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Whole-rock samples					
P8-311	174.3	529.1	0.329	0.9543	0.71851
* P8-175	14.3	12.2	1.170	3.4357	0.85907
* P7-029	232.9	93.4	2.493	7.2633	0.78046
P6.992	170.9	44.3	3.859	11.4054	0.92698
P6-994	175.6	44.8	3.924	11.5940	0.92488
P6-993	153.5	34.4	4.461	13.2536	0.98152
P8-313	216.4	29.7	7.288	21.8990	1.10237
P8-176	159.9	21.5	7.428	22.3417	1.11203
P8-312	223.4	30.0	7.441	22.4686	1.15406
P8-289	170.1	22.7	7.510	22.5951	1.11571
P7-064	176.2	23.0	7.667	23.1049	1.13310
P7-026	322.7	37.0	8.723	26.2751	1.12770
P8-841	178.5	9.5	18.882	60.5470	1.81488
P8-841 (r)	176.7	9.5	18.598	59.6072	1.80897
Mineral separates (P8-312)					
Eudialyte	43.6	193.2	0.225	0.6614	0.85356
Arfvedsonite	35.3	25.5	1.387	4.0823	0.88753
Albite	118.1	1.3	88.494	394.1870	6.22346
Microcline	641.1	5.1	124.582	698.3000	10.28940

continued

Table 24 (continued)

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Mineral separates (P8-311)					
Mafics	49.7	368.5	0.135	0.3905	0.71068
Pyroxene	316.2	151.0	2.095	6.1085	0.78894
Nepheline/ Feldspar (a)	401.0	73.4	5.461	16.1786	0.95297
Nepheline/ Feldspar (b)	440.8	41.8	10.550	31.8092	1.13840

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

Table 25

Regression parameters for data sets from Red Wine

Samples used	No. of samples	Age (m.y.)	Initial ratio	$[SUMS/(n-2)]^{\frac{1}{2}}$
Whole-rocks	11	1345 ± 75	0.7021 ± 0.0103	13.42
P8-312 minerals	5	969 ± 64	0.8400 ± 0.0077	11.16
P8-311 minerals	5	1024 ± 74	0.7047 ± 0.0051	9.88

Table 26

Rb-Sr whole-rock data for Norra Kärr (Reconnaissance samples)

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
NK-5	356.4	458.1	0.778	2.2610	0.75384
NK-11	229.9	162.7	1.412	4.1273	0.81009
NK-6	235.2	97.9	2.402	7.0464	0.85202
NK-12	330.3	135.1	2.444	7.1903	0.87972
NK-4	253.4	88.4	2.867	8.4303	0.87443
NK-2	244.1	27.2	8.985	27.4873	1.29393
NK-10	8054.9	609.8	13.2085	38.2479	0.71625

Table 27

Rb-Sr whole-rock and mineral data for Norra Kärr

Sample No.	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Pulaskite					
SW-12	270.7	557.2	0.486	1.4100	0.73963
Kaxtorpite					
SW-17	200.8	357.0	0.563	1.6338	0.74683
Lakarpite					
SW-8	188.1	482.5	0.390	1.1305	0.72893
SW-7	161.0	194.4	0.828	2.4087	0.76051
SW-13	271.5	150.0	1.810	5.2953	0.82123
Grennaite					
SW-5	87.5	21.2	4.136	12.2928	0.98639
SW-3	122.5	19.4	6.328	19.1050	1.15165
SW-6	262.0	27.2	9.642	29.6180	1.33798
SW-15	299.6	23.9	12.518	39.2451	1.56253
* SW-19	256.8	8.8	29.319	101.2610	2.68831
* SW-19 (r)	255.7	8.8	29.178	100.6090	2.66819
Mineral separates (SW-7)					
Fluorite	133.1	1264.0	0.105	0.3049	0.71587
Arfvedsonite	33.5	77.9	0.430	1.2478	0.73810
Albite	161.4	57.5	2.808	8.2344	0.84506
Microcline	536.2	72.8	7.362	22.1800	1.12970

continued

Table 27 (continued)

Sample No	Rb ppm	Sr ppm	Rb/Sr (wt)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Mineral separates (SW-8)					
Fluorite	133.7	4842.4	0.028	0.0799	0.70960
Arfvedsonite	25.2	92.3	0.273	0.7902	0.72810
Albite	253.9	194.4	1.306	3.8021	0.77156
Microcline	525.7	71.1	7.390	22.2016	1.10019

r = repeat analysis.

* = point lies off isochron and is not included in regression treatment.

N.B. Purity of mineral separates is estimated at >98% for fluorite and arfvedsonite and >80% for albite and microcline (the main contaminant being nepheline).

Table 28

K-Ar data for Norra Kärr

Sample No.	K (wt %)	$^{40}Ar^*$ (10^{-4}std.cm^{3}g^{-1})	$^{40}Ar^*/^{40}Ar_T$	Age (m.y.)
SW-7 (arfvedsonite)	1.97	1.567	0.991	1359
SW-8 (arfvedsonite)	1.89	1.661	0.991	1456

$\lambda_e = 0.584 \times 10^{-10}\ y^{-1}$

$\lambda_\beta = 4.72 \times 10^{-10}\ y^{-1}$

$^{40}K = (1.22 \times 10^{-4})K_T$

$^{40}Ar^*$ = radiogenic Ar

Potassium was determined by flame photometry and argon isotope ratios were measured on a MS-10 mass spectrometer at S.U.R.R.C., East Kilbride.

Estimated analytical error on the ages ("probable error") = ± 2%

Analyst: R.M.Macintyre

Table 29

Nuclides utilised for activation analysis of rare earth elements

Element	Target isotope	Isotopic abundance	Product nuclide	$t_{\frac{1}{2}}$	Cross-section (barns)	Measured photo-peak (KeV)
La	^{139}La	99.9	^{140}La	40.22 h	8.9	1596
Ce	^{140}Ce	88.5	^{141}Ce	33 d	0.6	145
Sm	^{152}Sm	26.6	^{153}Sm	47 h	210	103
Eu	^{151}Eu	47.8	^{152}Eu	12 y	5900	1408
Tb	^{159}Tb	100	^{160}Tb	72.1 d	46	963 + 966
Yb	^{174}Yb	31.8	^{175}Yb	4.21 d	55	396
Lu	^{176}Lu	2.6	^{177}Lu	6.7 d	2100	208

Table 30

Counting schedule for rare earth element determinations

Time after irradiation	Count * time	Nuclide	$t_{\frac{1}{2}}$	Measured peak (KeV)	Interfering peak	2nd peak of int. element
4 d	30 or 60 m	^{140}La	40.22 h	1596		
		^{153}Sm	47 h	103		
10 d	100 m	^{175}Yb	4.21 d	396		
		^{177}Lu	6.7 d	208		
30 d	200 m	^{141}Ce	33 d	145	143 (^{59}Fe)	1099
60 d	360 m	^{160}Tb	72.1 d	963 + 966	963 (^{152}Eu)	1408
		^{152}Eu	12 y	1408		

* Count times given are for samples and rock standards. Shorter counting times are used for REE standards.

Table 31

REE determinations on USGS standard basalt BCR-1 (concentrations in ppm)

	La	Ce	Sm	Eu	Tb	Yb	Lu
*1	26.16 ± 0.75	69.09 ± 2.60	7.12 ± 0.07	2.09 ± 0.06	1.39 ± 0.14	3.22 ± 0.41	0.45 ± 0.03
*2	25.71 ± 0.69	67.58 ± 2.09	6.50 ± 0.07	2.07 ± 0.06	1.31 ± 0.14	3.71 ± 0.45	0.54 ± 0.05
*3	26.56 ± 1.11	64.52 ± 0.70	5.86 ± 0.06	1.97 ± 0.06	1.36 ± 0.13	3.57 ± 0.56	0.60 ± 0.05
*4	26.99 ± 1.09	64.94 ± 0.65	5.72 ± 0.05	2.02 ± 0.06	1.20 ± 0.12	3.30 ± 0.55	0.52 ± 0.05
Mean	26.36 ± 0.45	66.53 ± 2.17	6.30 ± 0.64	2.04 ± 0.05	1.32 ± 0.09	3.45 ± 0.23	0.53 ± 0.06
USGS value	26	53.9	6.6	1.94	1.0	3.36	0.55
percent deviation	1.4	23	4.5	5.2	32	2.7	3.6

* Errors are indicated at the 1σ level, and represent counting statistical errors only.

USGS values are "recommended" values from Flanagan (1973).

Table 32

REE concentrations (ppm)

Sample No.	La	Ce	Sm	Eu	Tb	Yb	Lu	La/Yb	Figure ref.
Anorthosites									
85964	5.67	15.30	0.89	1.08	0.33	0.32	0.09	17.7	A
50221	9.09	19.63	1.77	1.19	0.40	0.26	0.06	35.0	B
40523	10.69	21.58	1.76	1.27	0.57	0.41	0.07	26.1	C
Gabbro chill									
86106 *	26.58	67.71	3.54	1.29	0.43	1.67	0.29	15.9	K
86107	25.99	73.82	5.78	2.55	1.41	2.36	0.35	11.0	J
40488	37.51	94.10	9.70	3.06	1.24	3.24	0.44	11.6	M
85954	44.14	111.55	8.59	3.21	1.59	3.37	0.72	13.1	O
Gabbro									
86055	13.33	42.03	3.03	1.76	0.48	1.33	0.13	10.0	D
81152	14.78	37.24	3.17	1.89	0.77	1.14	0.15	13.0	E
30650	16.22	45.46	3.65	1.90	1.06	2.48	0.37	6.5	F
30642	20.64	54.32	4.87	2.05	0.58	1.63	0.21	12.7	H
85977	20.09	66.02	6.34	2.61	0.91	1.82	0.25	11.0	G
30636	23.59	61.32	4.60	2.31	0.87	1.74	0.28	13.6	I
30658	34.24	94.69	8.03	3.27	1.29	2.01	0.40	17.0	L
40470	38.51	95.46	10.6	3.09	1.40	3.15	0.48	12.2	N
30638	45.40	128.17	11.13	3.97	1.66	2.56	0.69	17.7	P
Syenogabbro									
85876	60.97	176.19	12.58	4.95	1.88	3.36	0.59	18.1	R

continued

Table 32 (continued)

Sample No.	La	Ce	Sm	Eu	Tb	Yb	Lu	La/Yb	Figure ref.
85976 (r)	71.72	180.88	16.94	5.59	1.99	3.63	0.42	19.8	R
40551	63.01	167.00	12.49	4.43	3.19	4.62	0.68	13.6	Q

r = repeat analysis.

* = sample hybridised with basement granite.

Chondrite values of Frey et al. (1968) are as follows:

La	0.330 ± 0.013
Ce	0.88 ± 0.01
Sm	0.181 ± 0.006
Eu	0.069 ± 0.001
Tb	0.047 ± 0.001
Yb	0.200 ± 0.007
Lu	0.034 ± 0.002

Table 33

Average REE concentrations (ppm) for Ilimaussaq units

Rock unit	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Ho	Er	Yb	La/Yb
Augite syenite	49	110	14	51	11	5.2	10	2.5	-	5.3	4.6	9.2
Kakortokite	330	550	65	250	48	6.7	43	8.0	9.0	24.0	19.5	16.9
Naujaite	166	305	33	140	28	6.3	21	-	-	9.3	7.9	21.0
Lujavrite	1500	2260	257	800	140	18.5	120	23	-	58	48	31.3
Sodalite Foyaite	320	603	76	233	52	5.5	47	7.5	-	17.5	13	24.6
Alkali Granite	485	917	111	378	73	7.5	54	-	-	33	24	20.2

All values averaged from data reported by Gerasimovskiy and Balashov (1968)

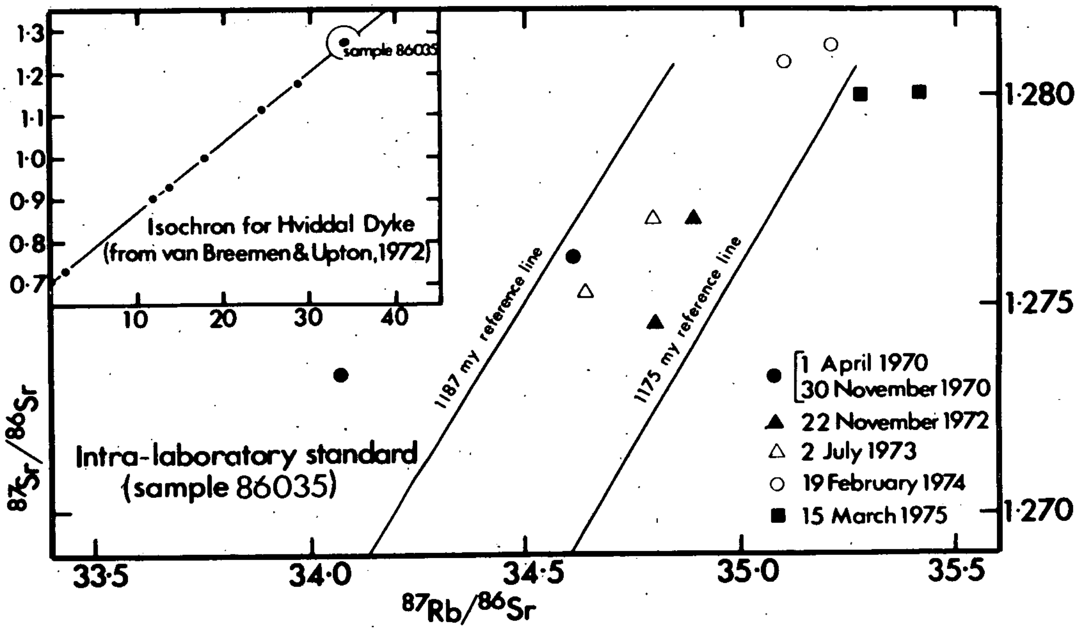

Fig. 1. Isochron plot of successive analyses of internal standard 86035 (selected from isochron, shown in inset, of van Breemen and Upton, 1972). Reference lines of 1175 and 1187 m.y. illustrate the 1% age reduction discussed in Chapter 3.

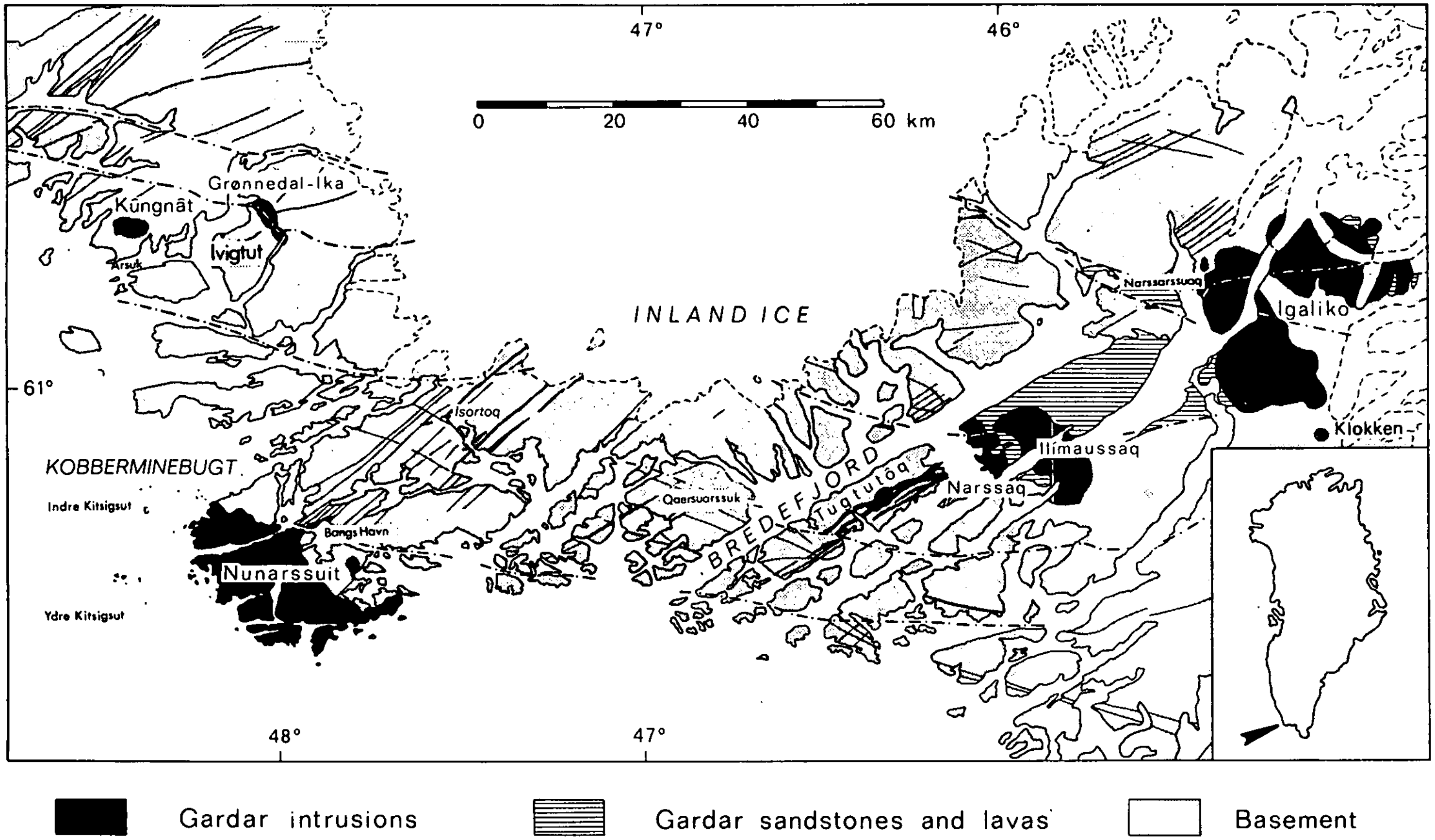

Fig. 2. Simplified geological map of south Greenland, showing the distribution of igneous rocks in the Gardar Province

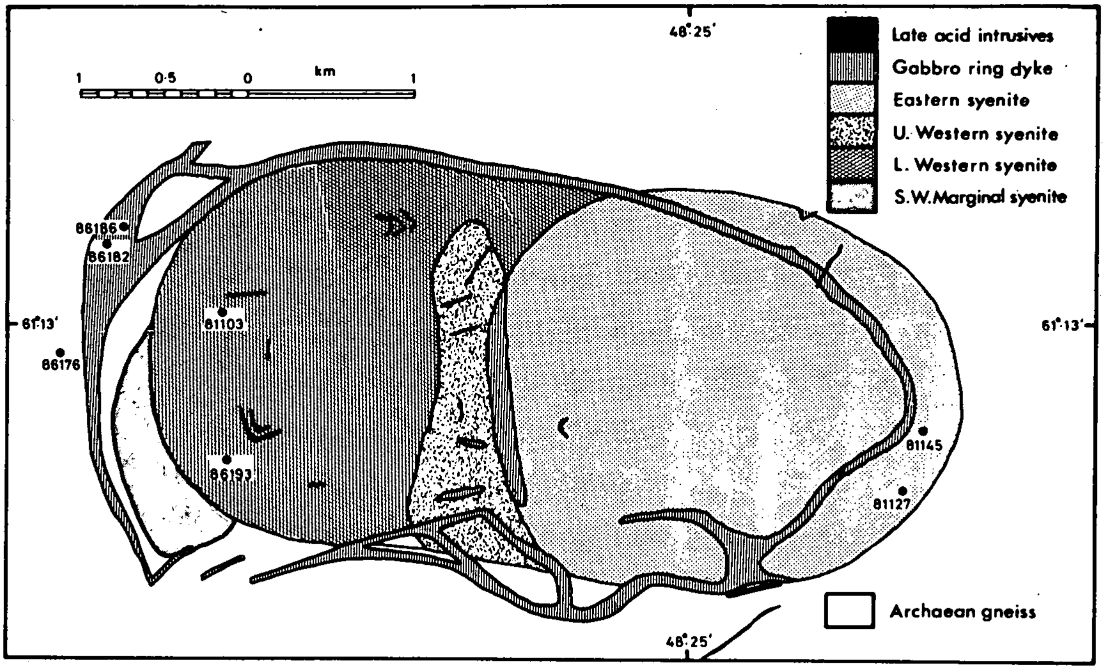

Fig. 3. Simplified geological map of the Kûngnât Complex, showing location of samples used in this study.

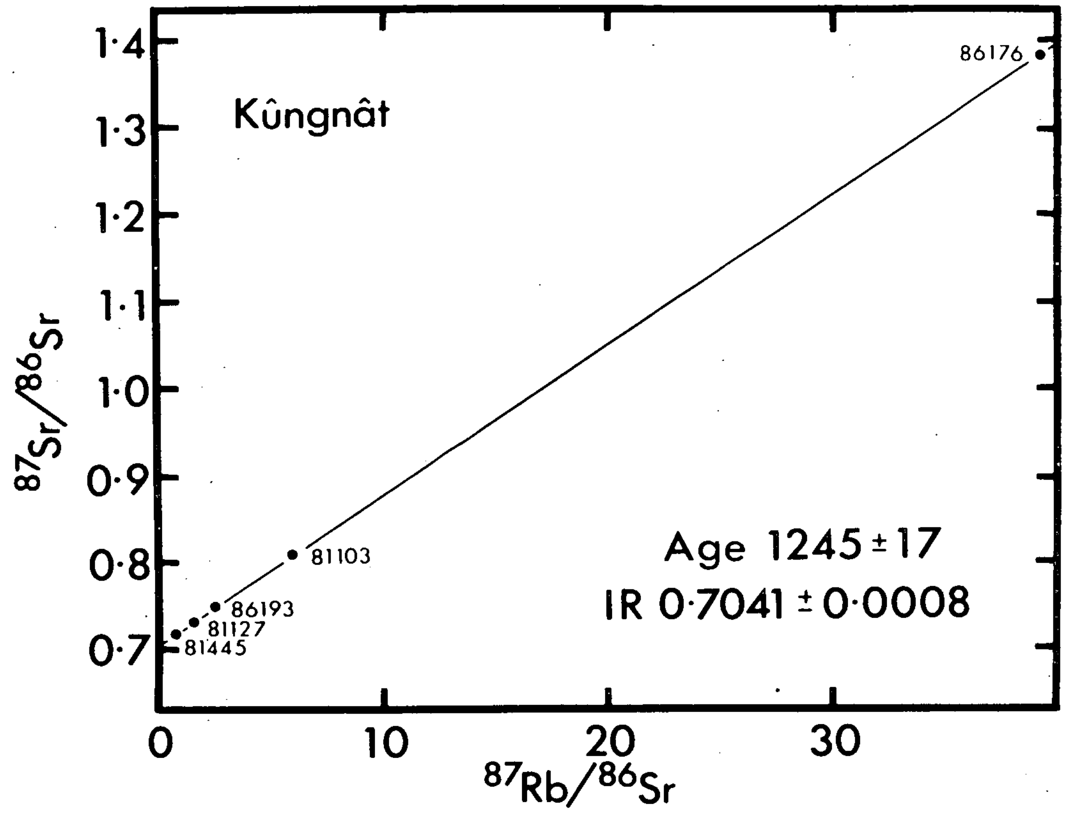

Fig. 4. Rb-Sr whole-rock isochron diagram for the Kûngnât Complex. See Table 6 for details of samples used in the regression calculation.

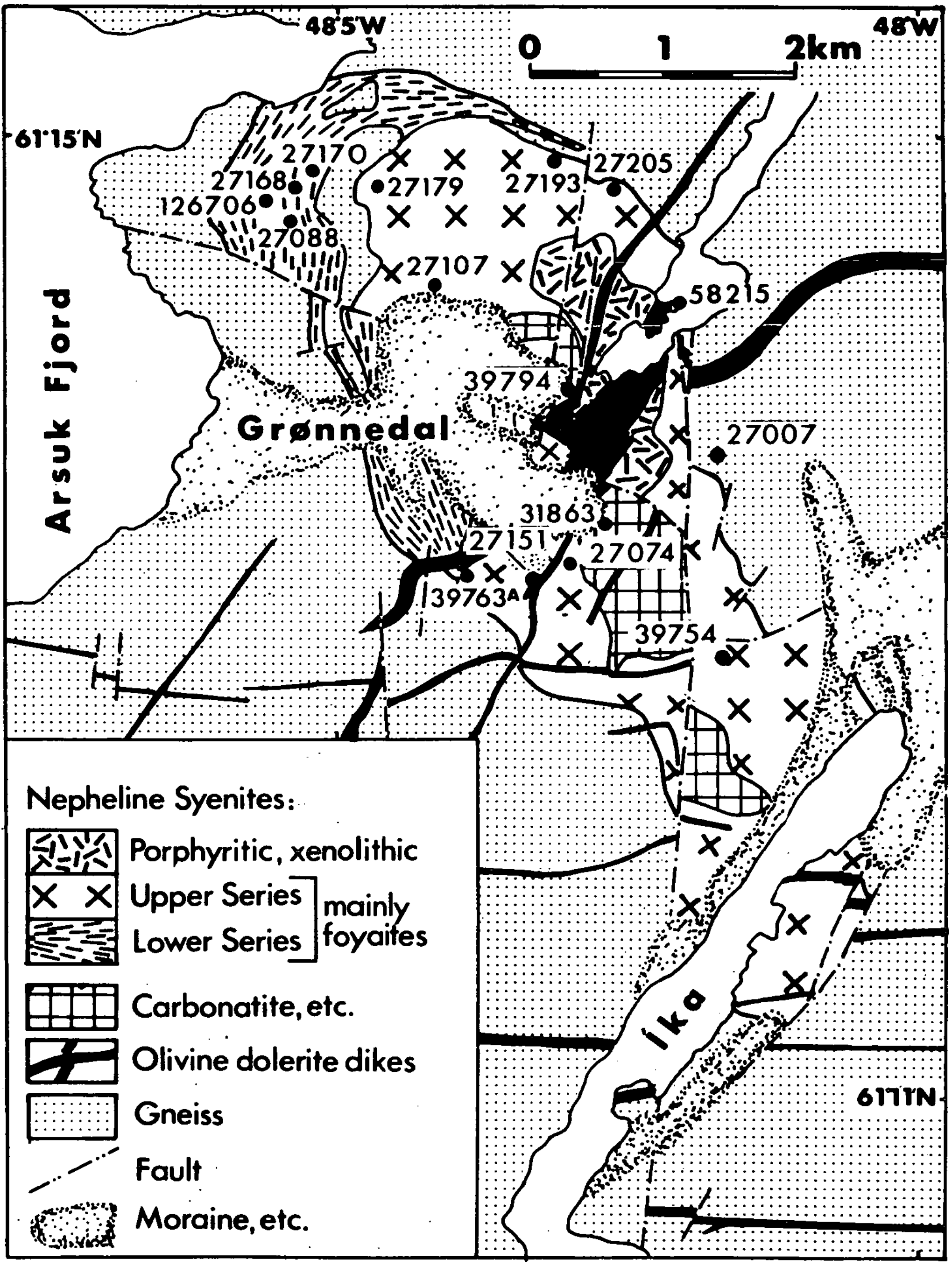

Fig. 5. Simplified geological map of the Grønnedal-Ika complex, showing location of samples used in this study. (From Blaxland et al., in press)

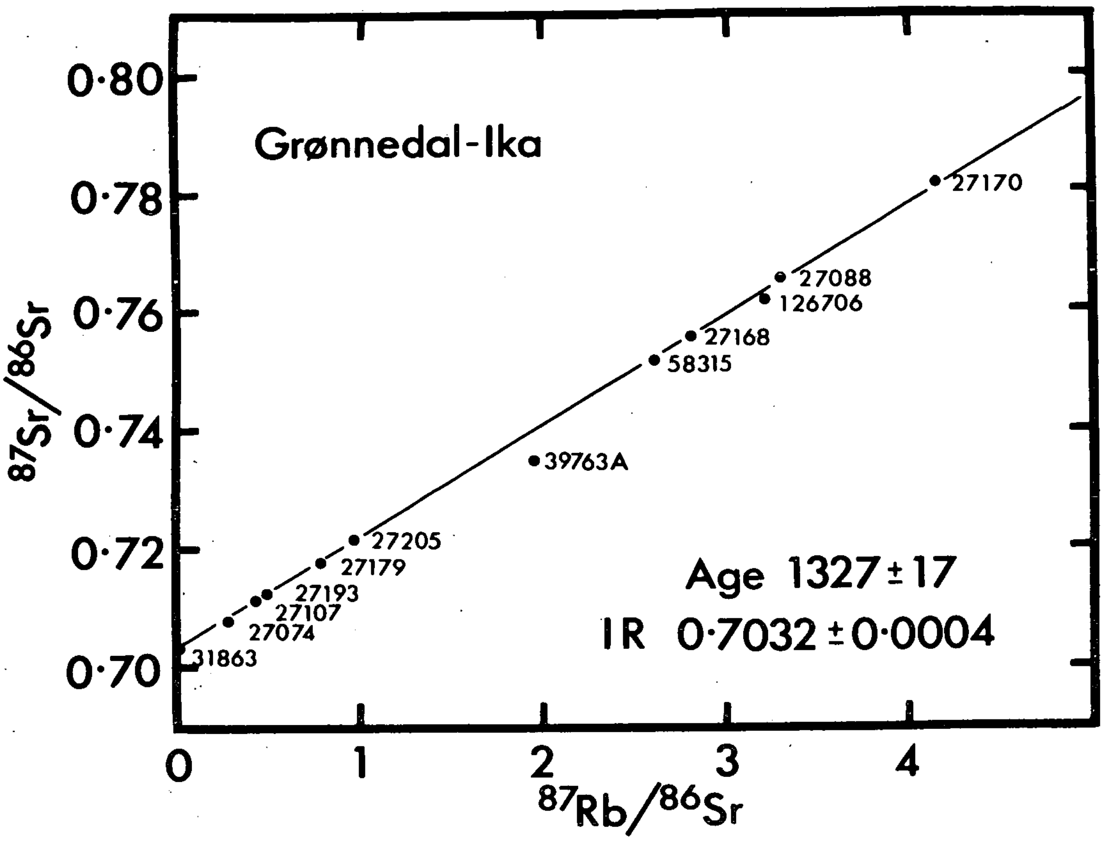

Fig. 6. Rb-Sr whole-rock isochron diagram for the Grønnedal-Ika Complex. See Table 7 for details of samples used in the regression treatment.

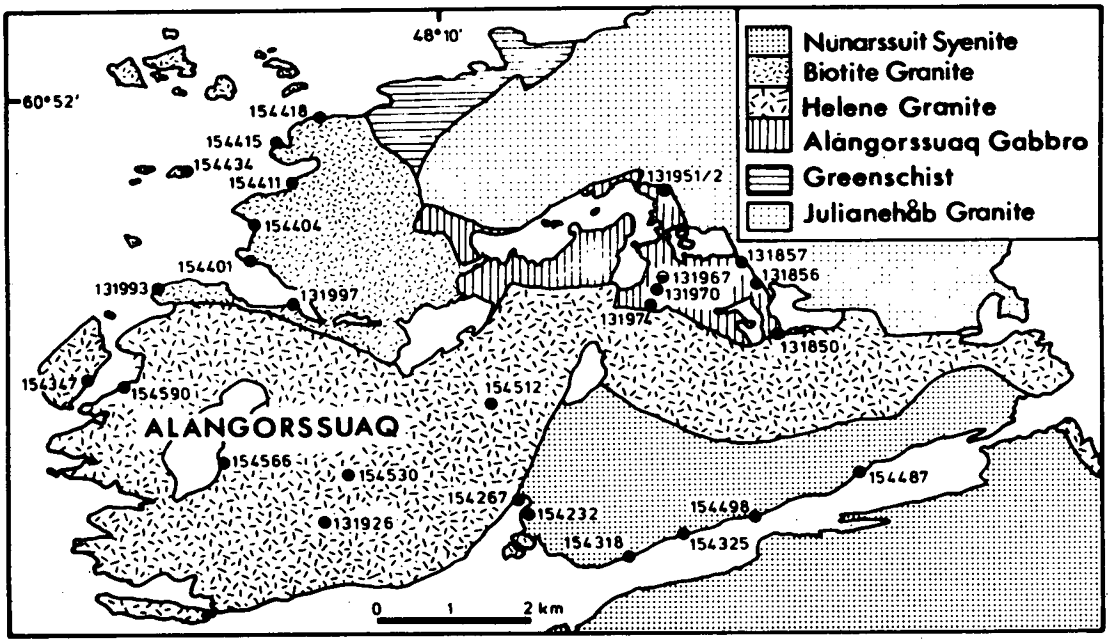

Fig. 7. Simplified geological map of the northern part of the Nunarssuit Complex, showing location of samples used in this study. (From Blaxland et al., in press)

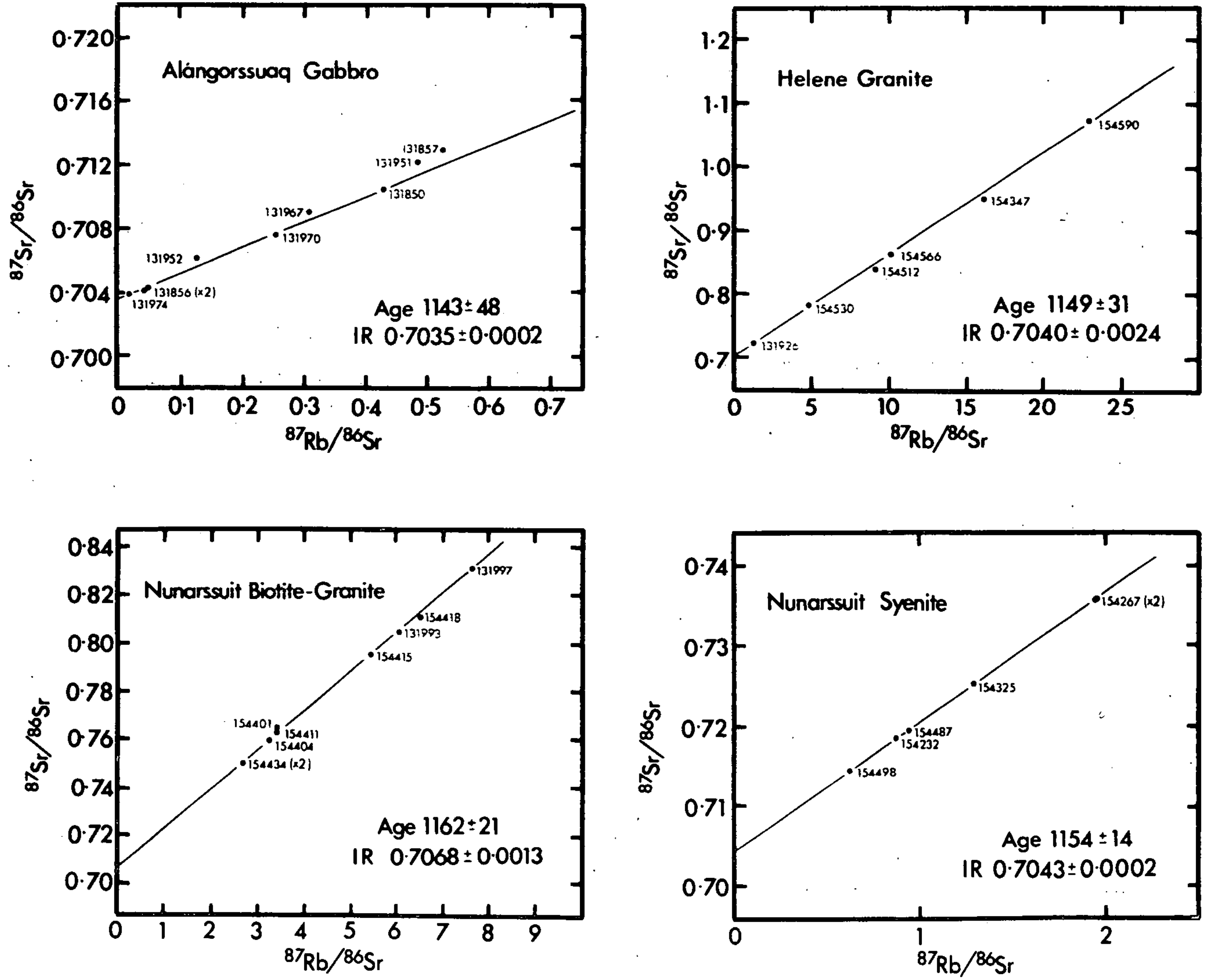

Fig. 8. Rb-Sr whole-rock isochron diagrams for the Nunarssuit Complex: (a) Alángorssuaq Gabbro, (b) Helene Granite, (c) Biotite Granite and (d) Nunarssuit Syenite. See Tables 8-11 for details of samples used in the regression treatment. The isochron for Alángorssuaq Gabbro is subject to special interpretation (p. 39-40).

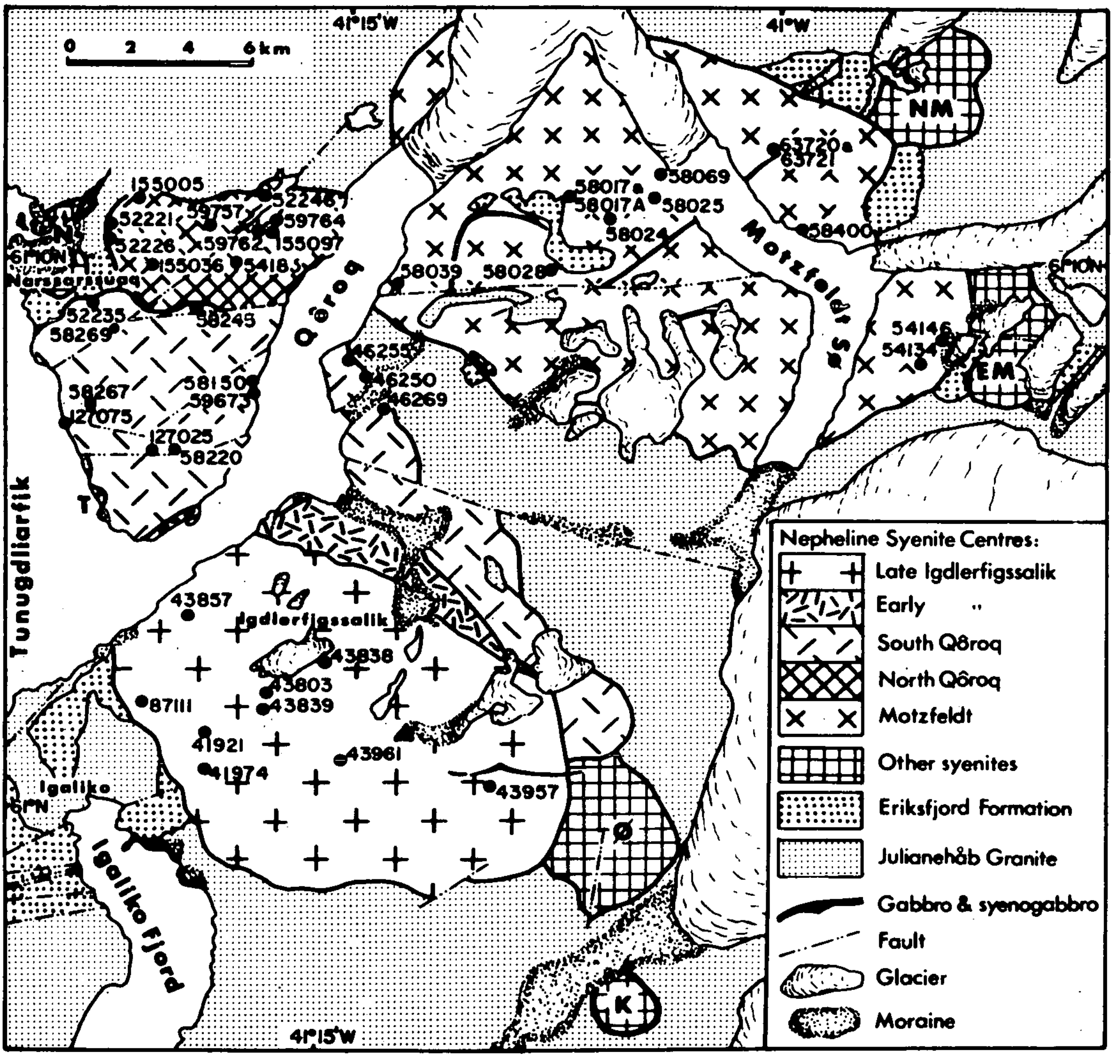

Fig. 9. Simplified geological map of the Igaliko Complex, showing location of samples used in this study. Intrusions referred to as "other syenites" in key are: NM = North Motzfeldt, EM = East Motzfeldt, N = Narssarssuaq, T = Tunugdliarfik, Ø = Ostfjordsdal, K = Klokken (see Emeleus and Harry, 1970 and Parsons, 1972b). (From Blaxland et al., in press)

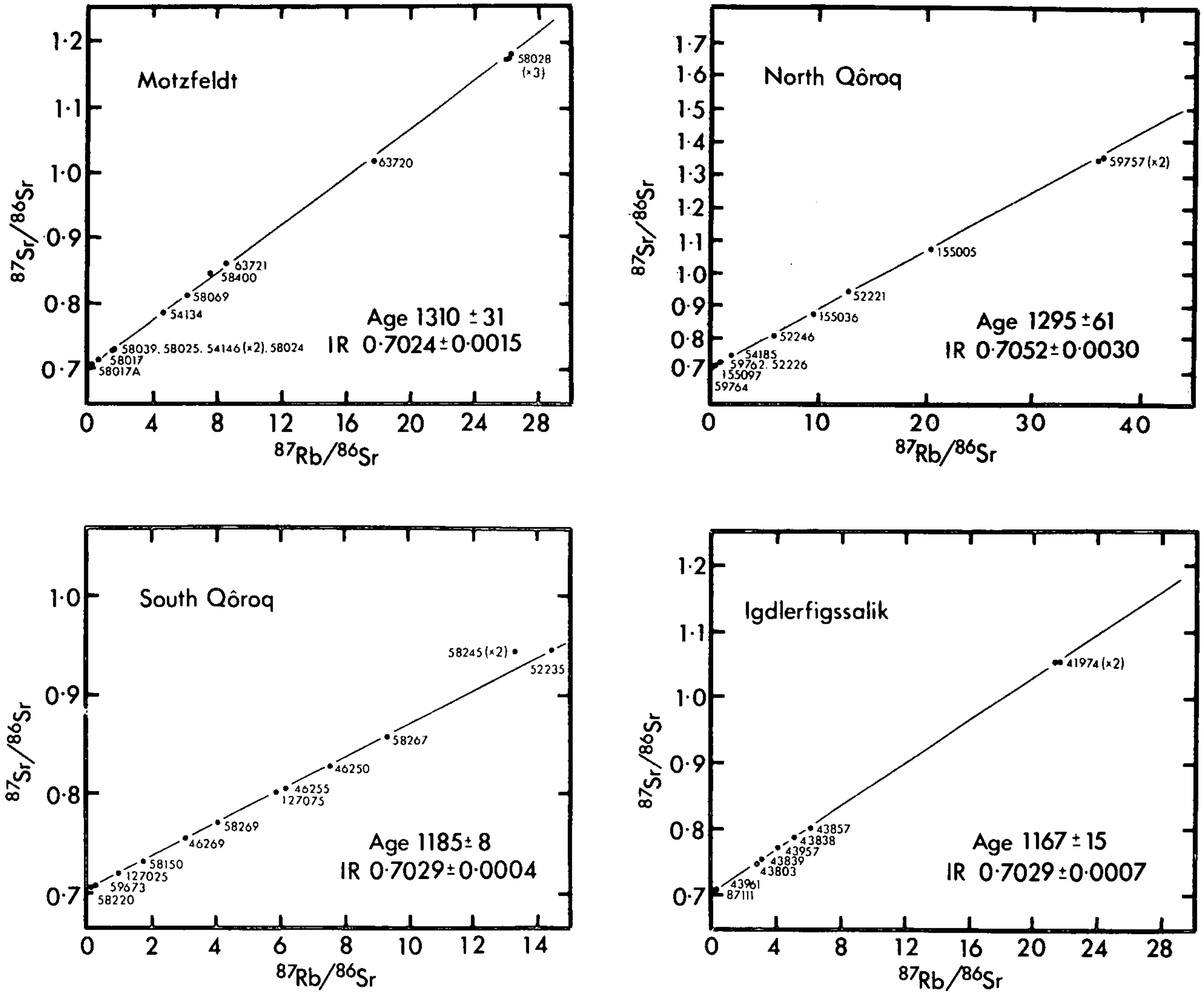

Fig. 10. Rb-Sr whole-rock isochron diagrams for the Igaliko Complex: (a) Motzfeldt, (b) North Qôroq, (c) South Qôroq and (d) Late Igdlerfigssalik. See Tables 13-16 for details of samples used in the regression treatment.

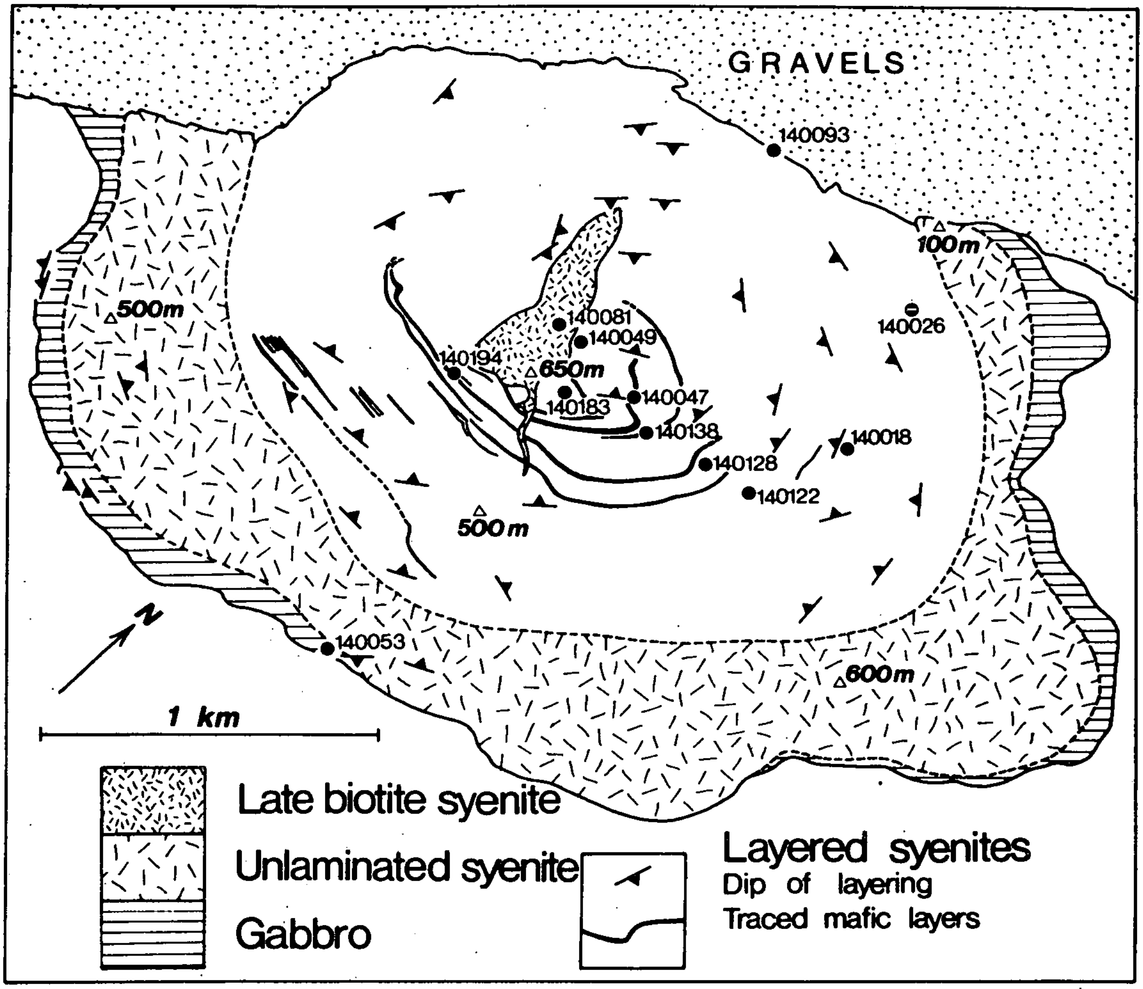

Fig. 11. Geological map of the Klokken Complex, showing location of samples used in this study. (From Blaxland and Parsons, 1975)

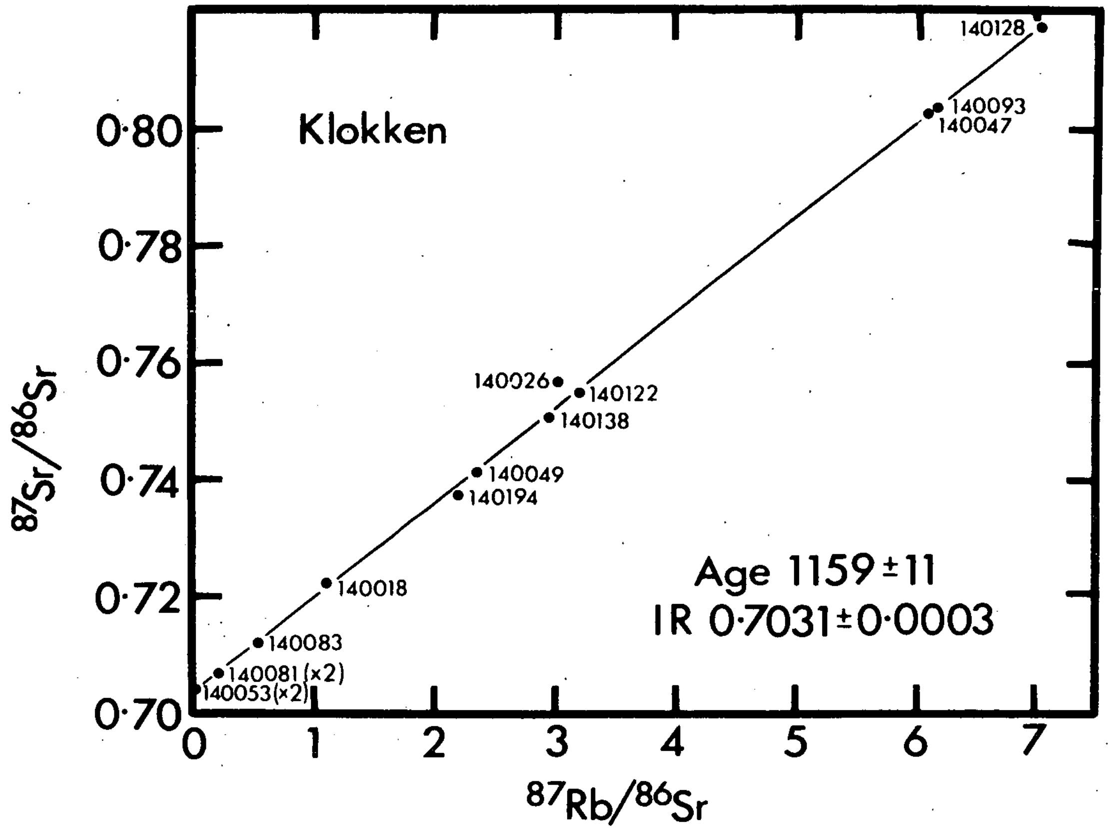

Fig. 12. Rb-Sr whole-rock isochron diagram for the Klokken Complex. See Table 17 for details of samples used in the regression treatment.

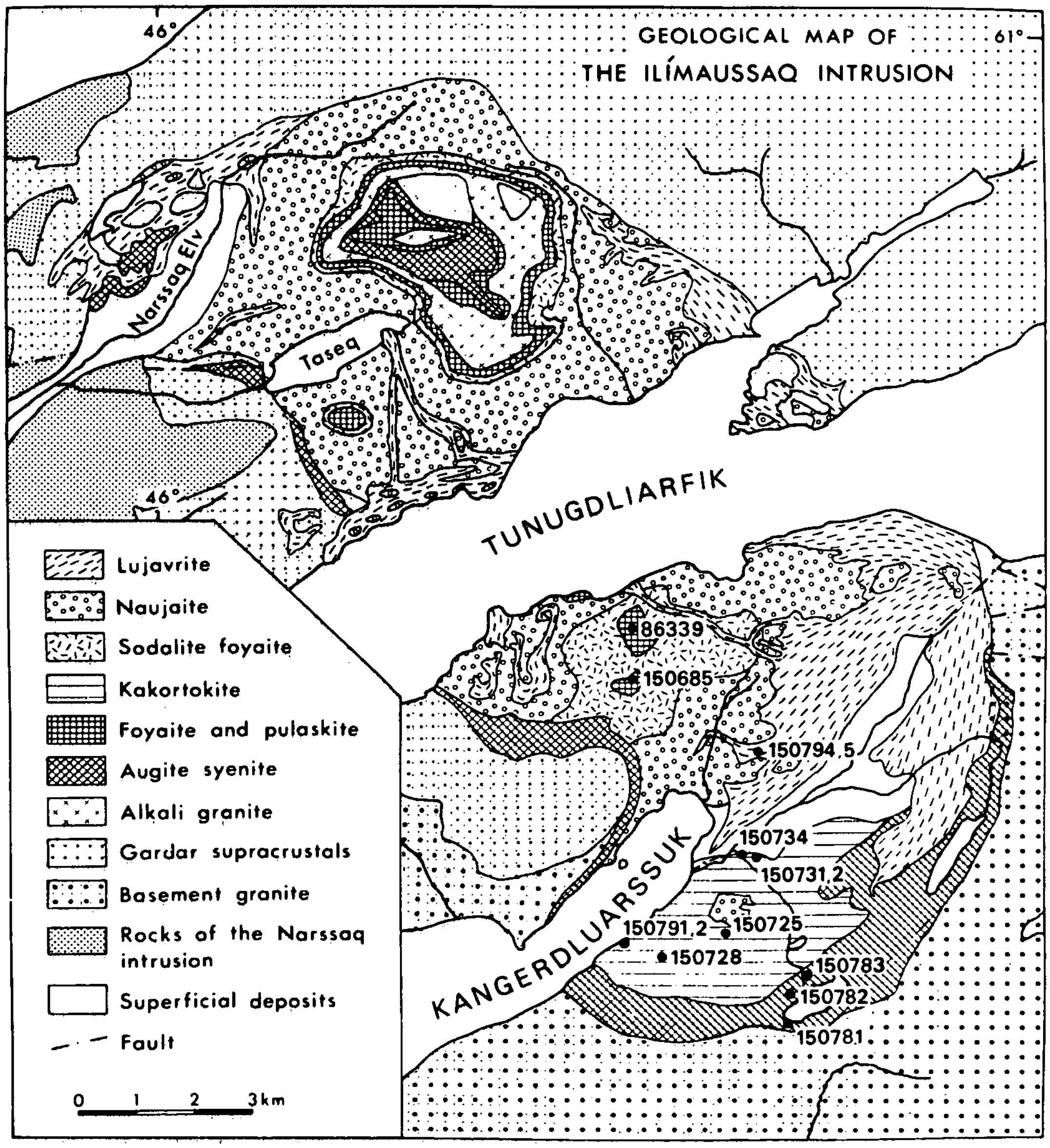

Fig. 13. Geological map of the Ilimaussaq Complex, showing location of samples used in this study. (From Blaxland et al., 1976)

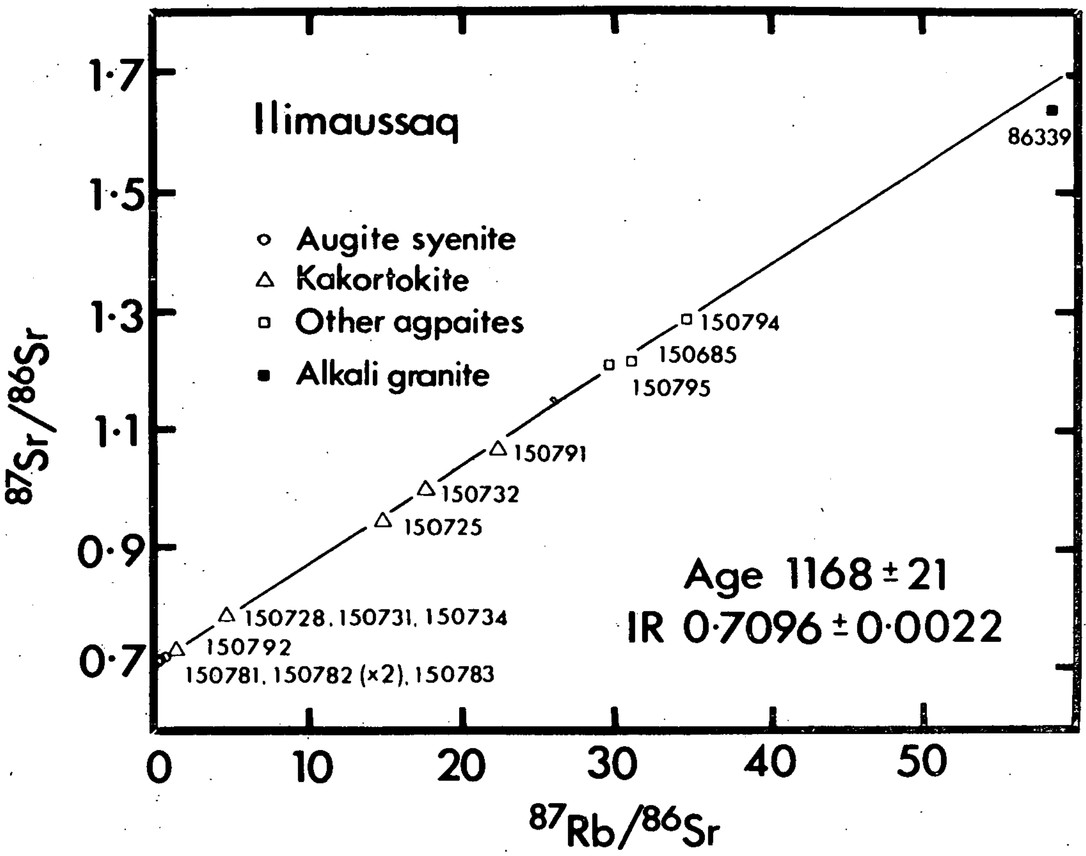

Fig. 14. Rb-Sr whole-rock isochron diagram for the Ilimaussaq Complex. See Tables 18 and 19 for data pertaining to the regression treatments.

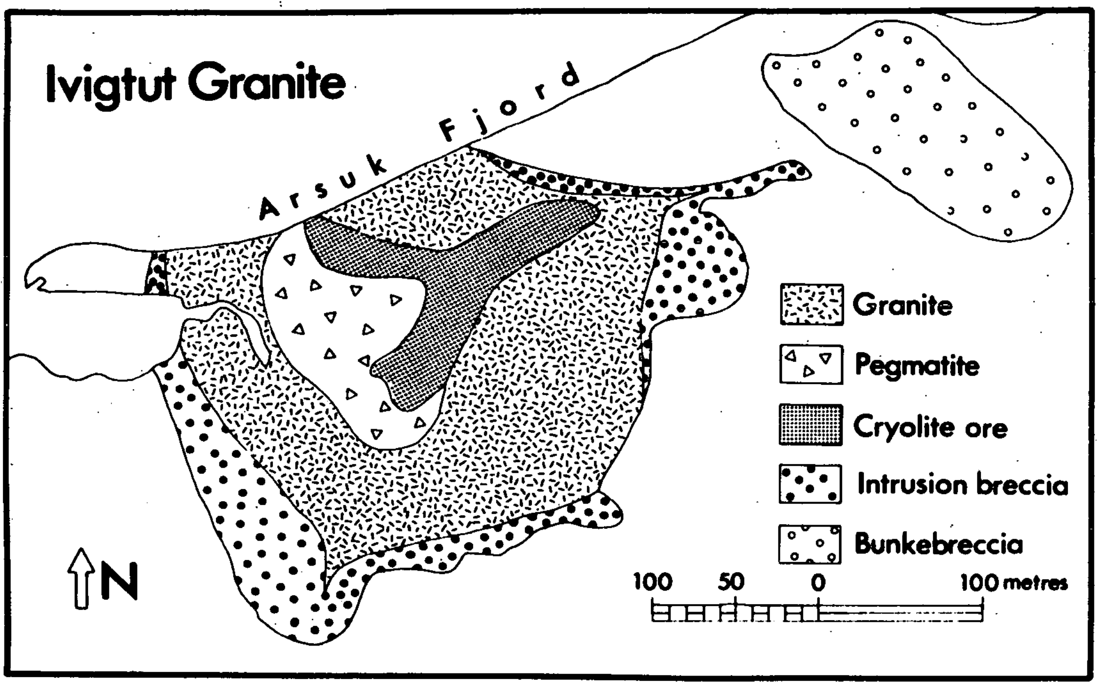

Fig. 15. Simplified geological map of the Ivigtut Granite Complex.

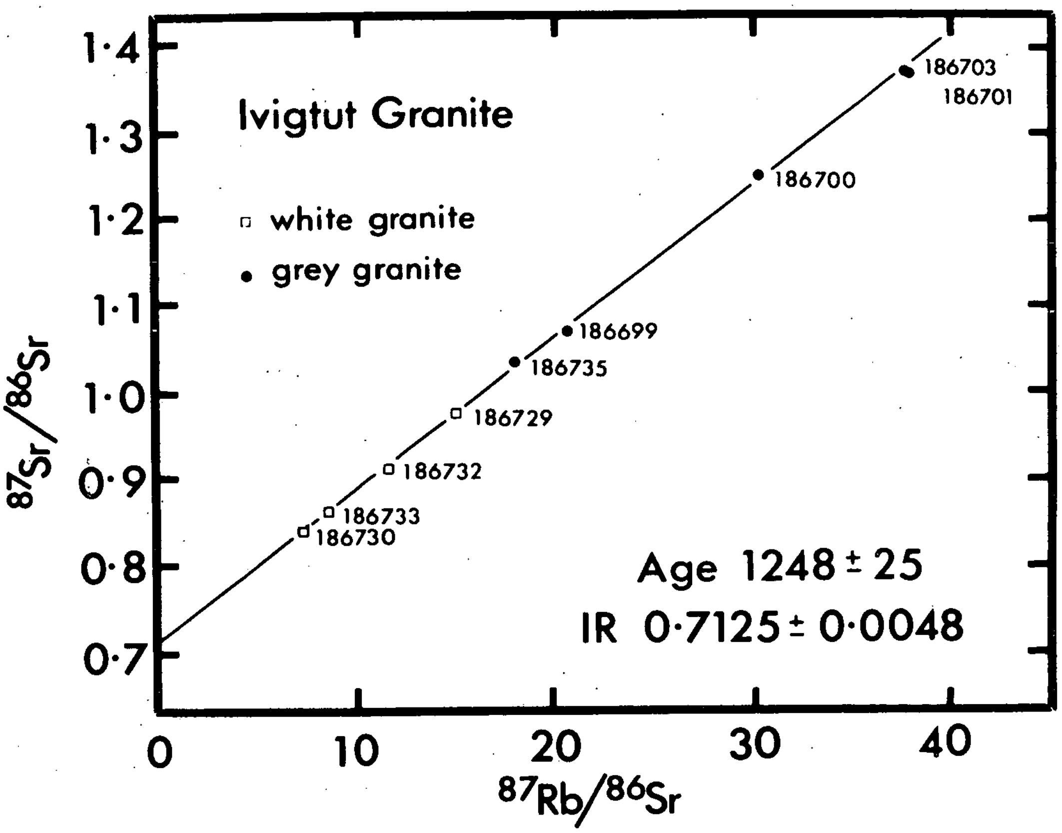

Fig. 16. Rb-Sr whole-rock isochron diagram for the Ivigtut Granite Complex. See Table 20 for details of samples used in the regression treatment.

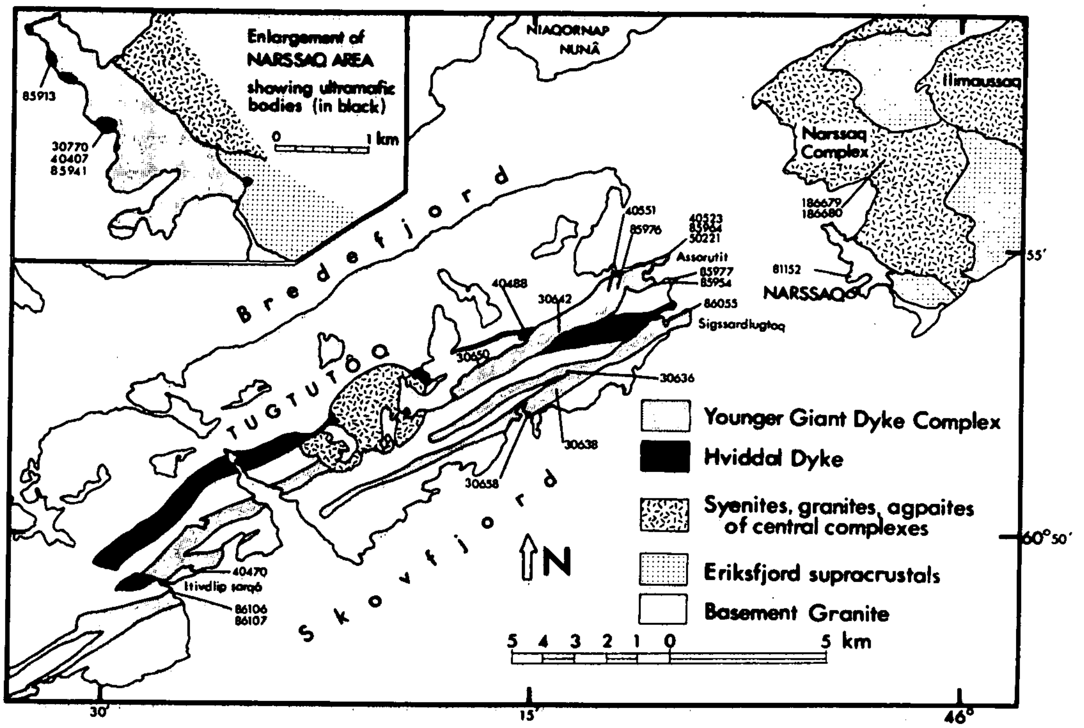

Fig. 17. Simplified geological map of the Tugtutôq-Narssaq area, showing location of samples used in this study from the Narssaq Complex (Chapter 5), the Narssaq ultramafic bodies (Chapter 8) and the younger giant dyke complex (Chapter 13).

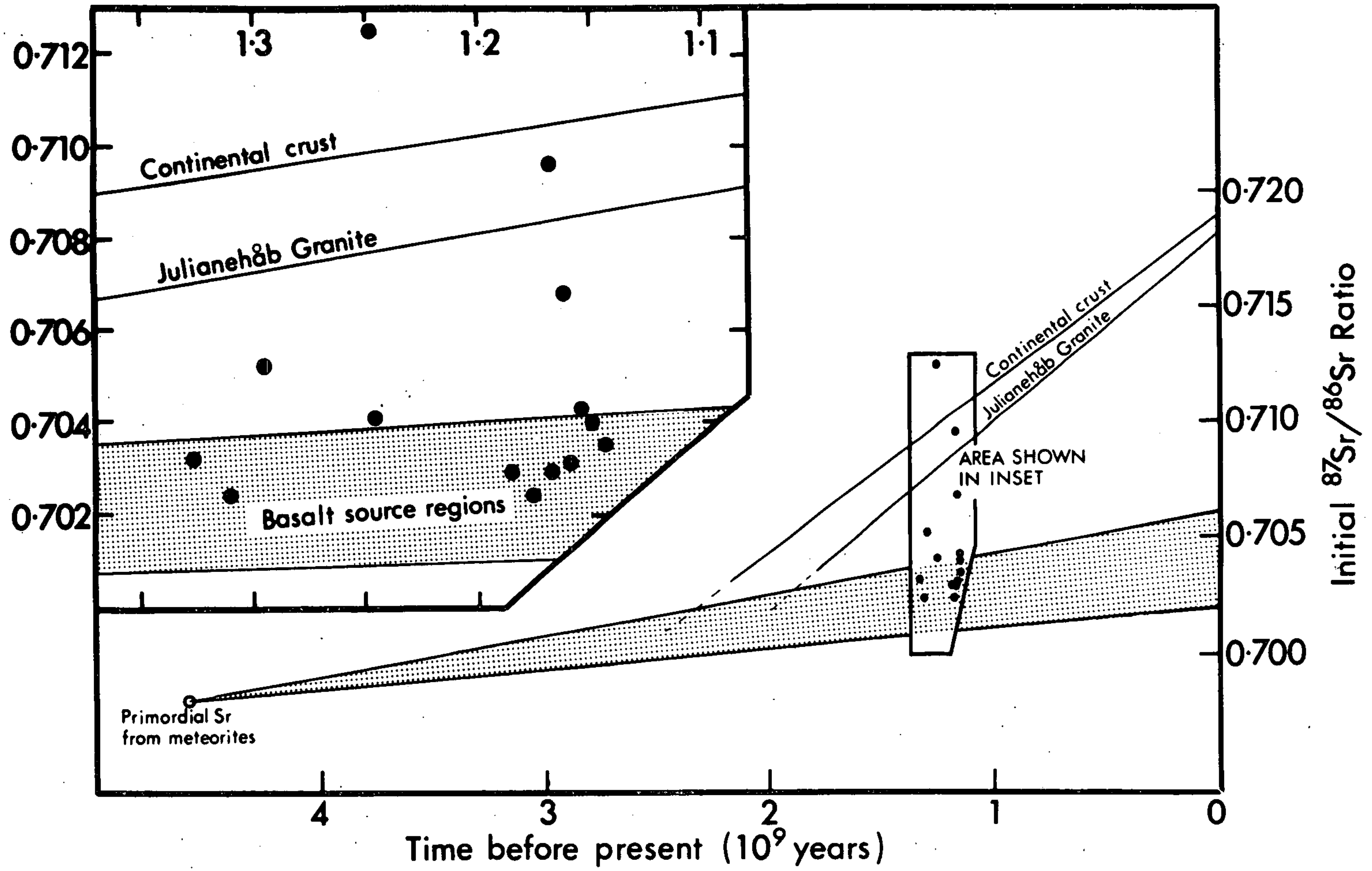

Fig. 18. Relationship between age and initial $^{87}Sr/^{86}Sr$ ratio for Gardar complexes. Individual intrusions can be identified by reference to Figs. 19 and 20.

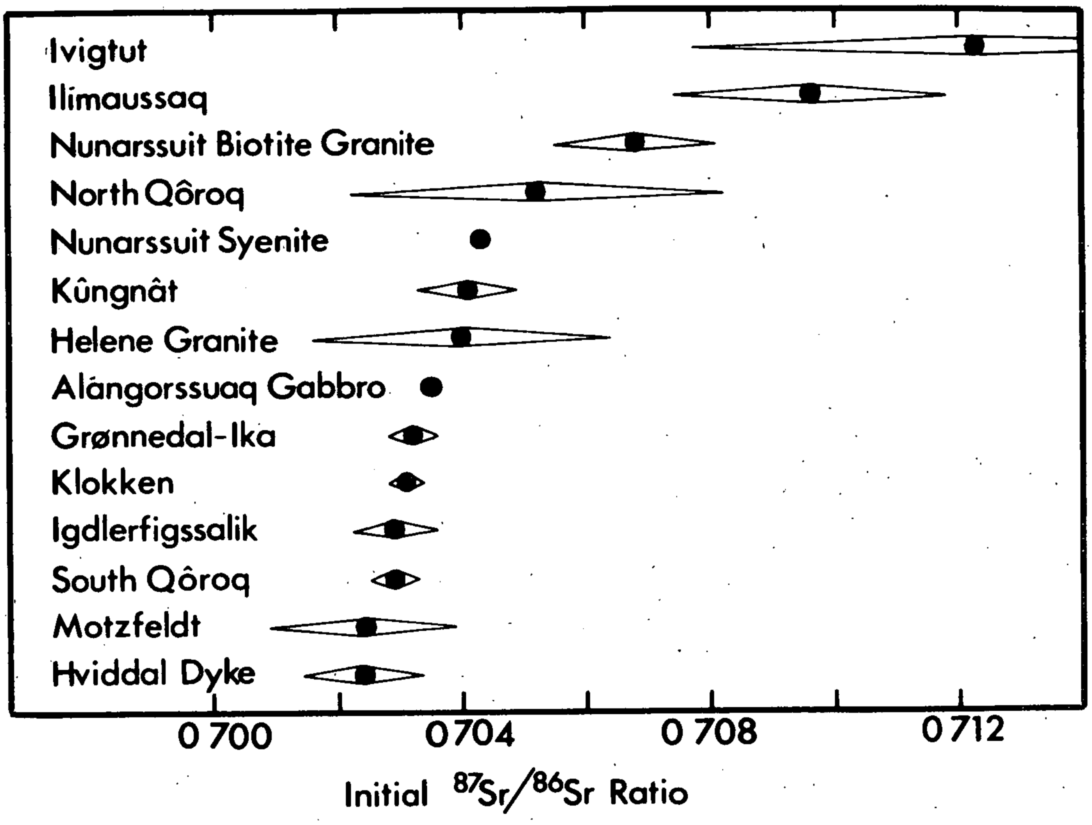

Fig. 19. Cumulative frequency plot of Gardar initial $^{87}Sr/^{86}Sr$ ratios listed in Table 22. Errors are indicated at the 2σ level.

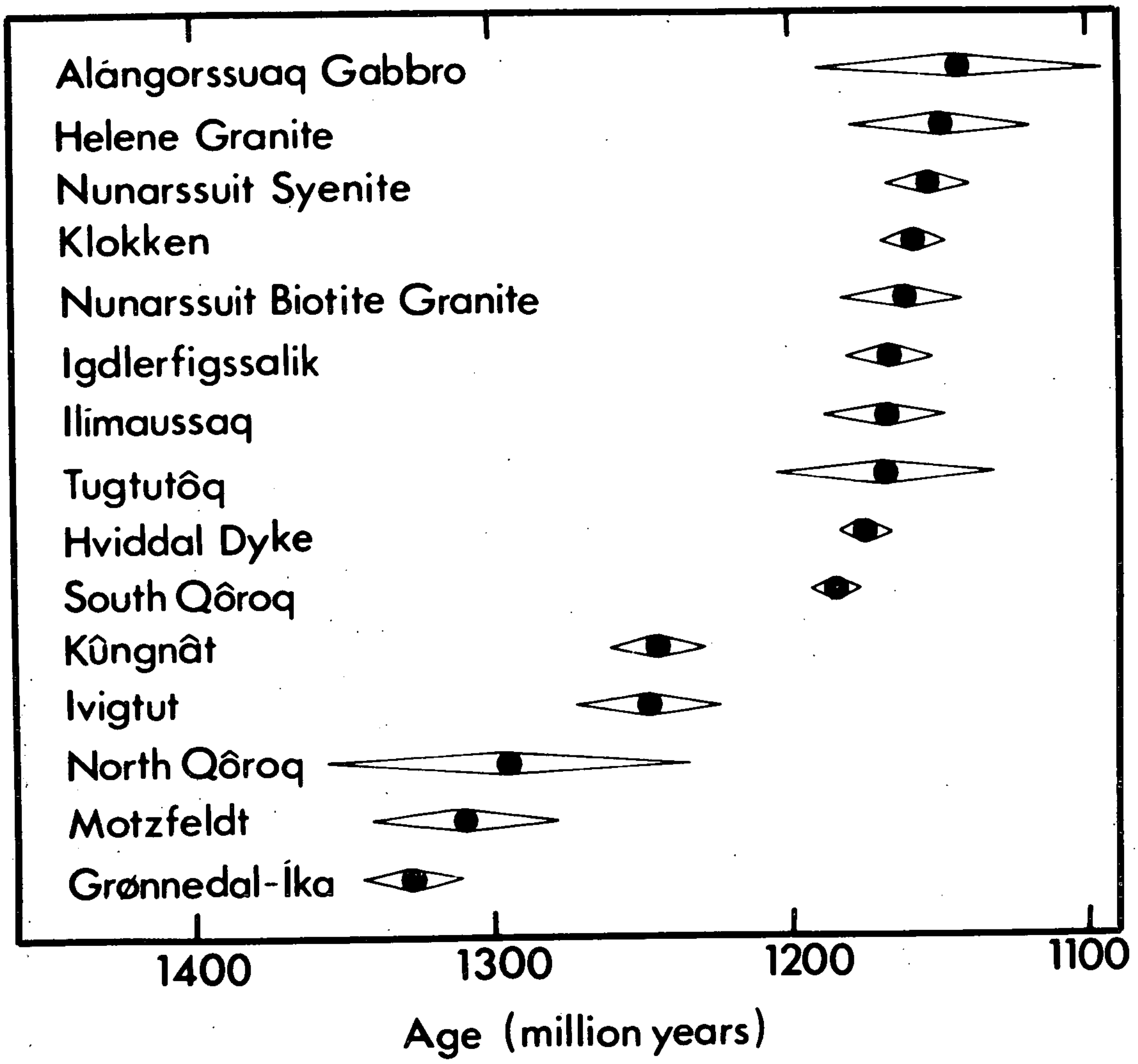

Fig. 20. Cumulative frequency plot of Gardar ages listed in Table 22. Errors are indicated at the 2σ level.

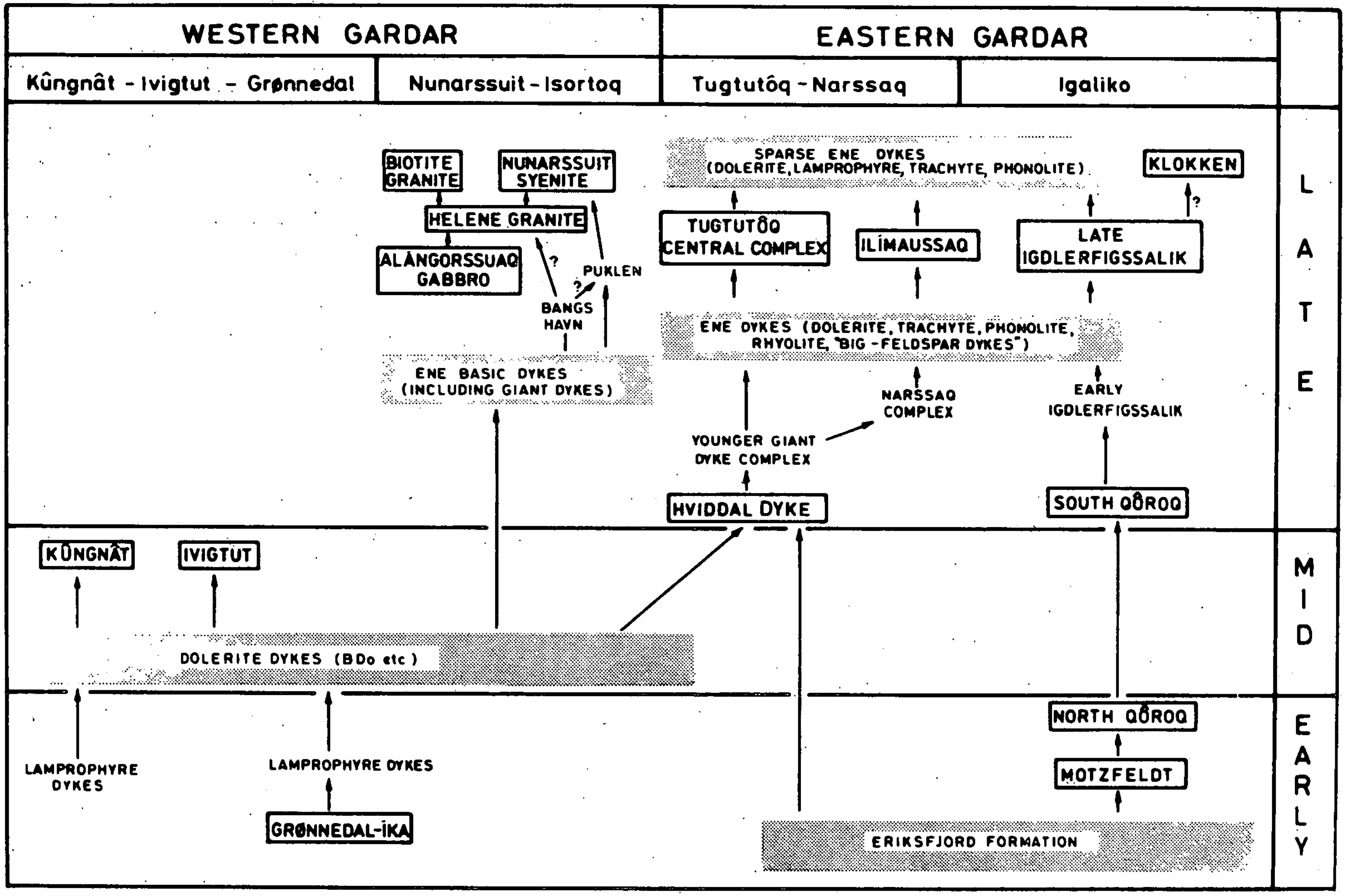

Fig. 21. Revised chronological scheme for the Gardar Province. The division into Early, Mid and Late phases is somewhat arbitrary, based on Fig. 20. Only relative chronology is shown; an arrow indicates that the top event is known to be younger than the bottom event on field or isotopic evidence. A box indicates a Rb-Sr isochron age (Table 22).

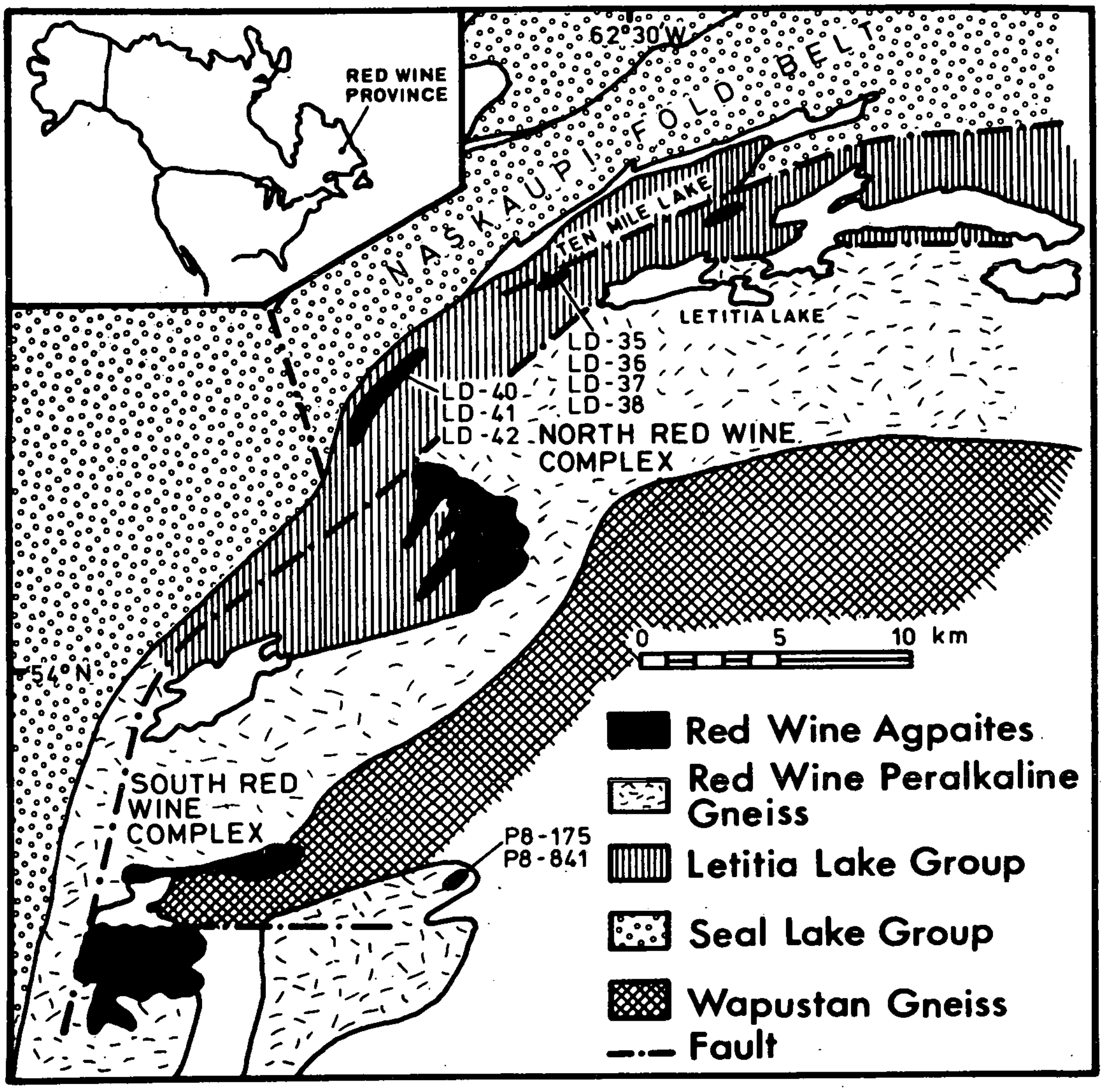

Fig. 22. Generalised geological map of the Red Wine alkaline province, central Labrador. (Based on information from Currie et al., 1975, Curtis, 1975 and personal communications from Dr. L.Curtis and Dr. J.Gittins, 1975)

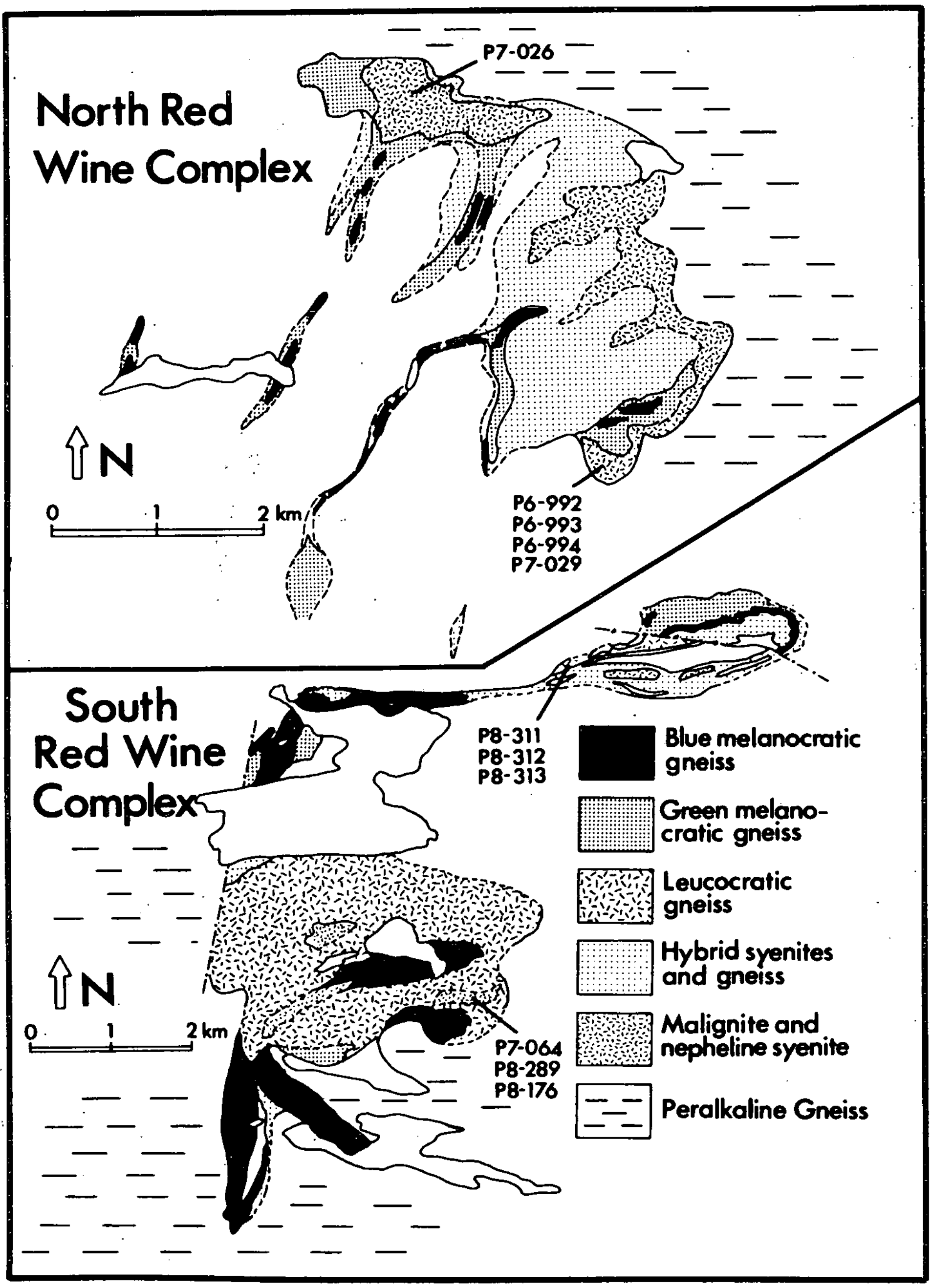

Fig. 23. Geological maps of the North and South Red Wine Complexes, showing location of most samples used in this study (samples P8-175 and P8-841, and the reconnaissance samples lie off this map and are indicated on Fig. 22). (Based on sketch maps of Currie et al., 1975)

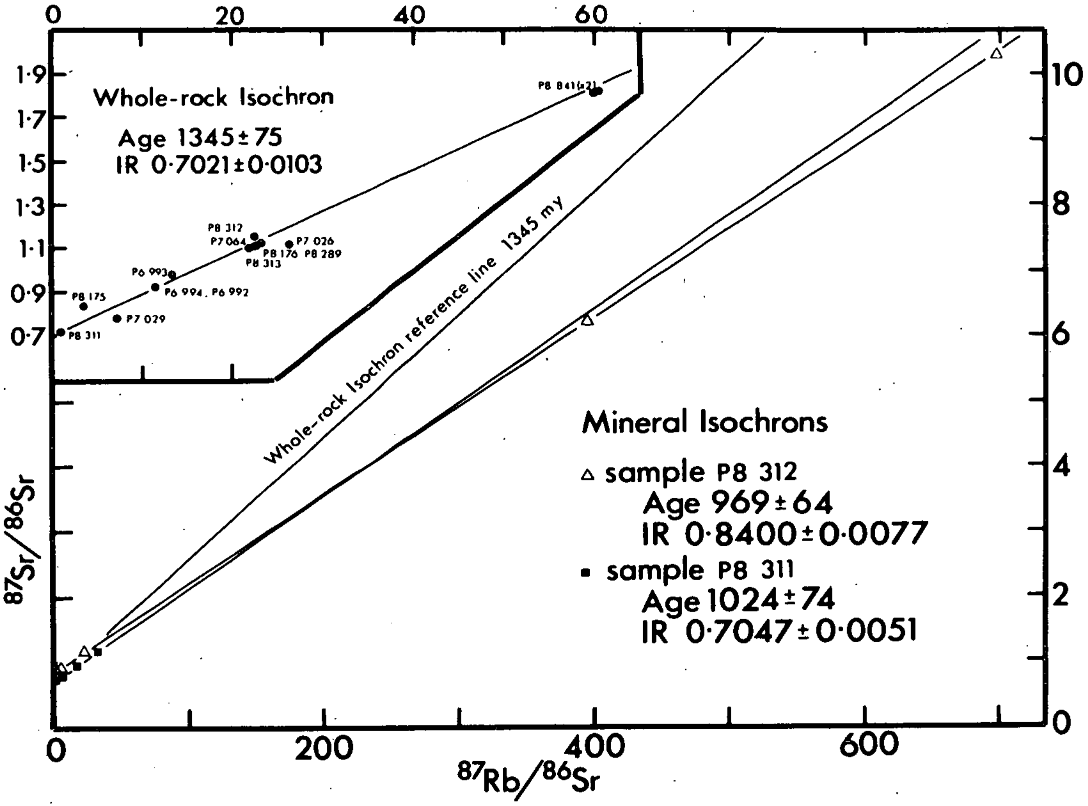

Fig. 24. Rb-Sr whole-rock (inset diagram) and mineral (main diagram) isochron plots for the Red Wine Complexes. See Tables 24 and 25 for data pertaining to the regression treatments.

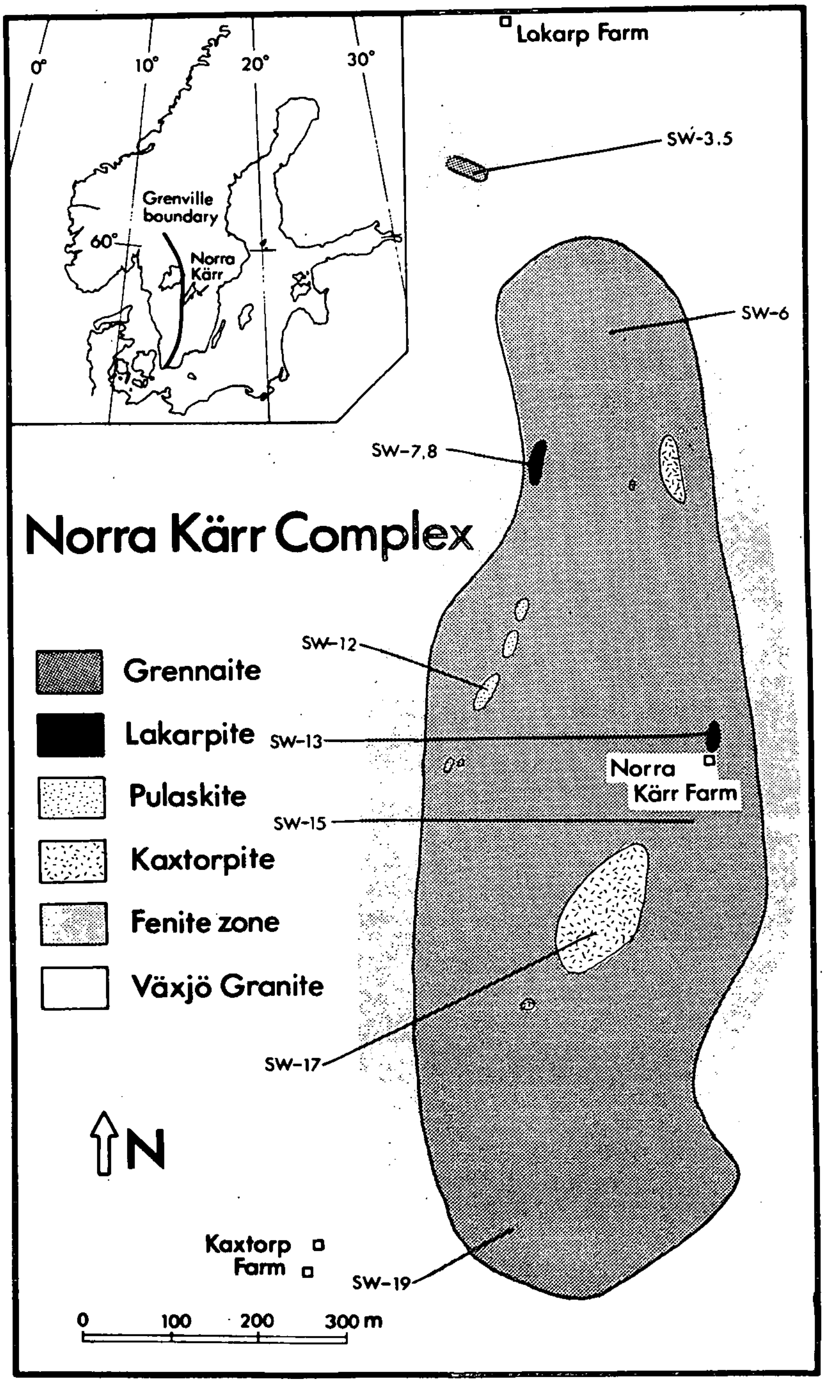

Fig. 25. Simplified geological sketch-map of the Norra Kärr Complex, showing location of samples used in this study. (Based on map of Adamson, 1944)

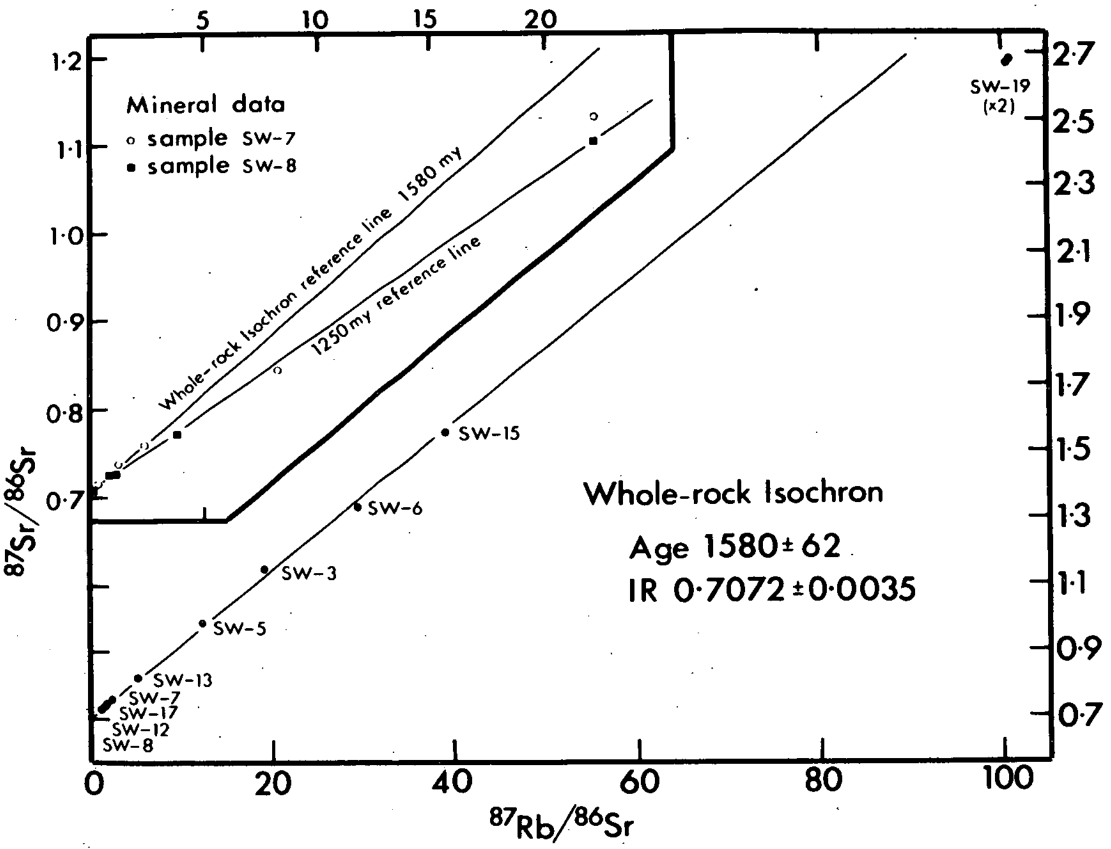

Fig. 26. Rb-Sr whole-rock isochron diagram and (inset) mineral isochron plot (for samples SW-7 and SW-8) for the Norra Kärr Complex. See Table 27 for details of samples used in the regression treatment.

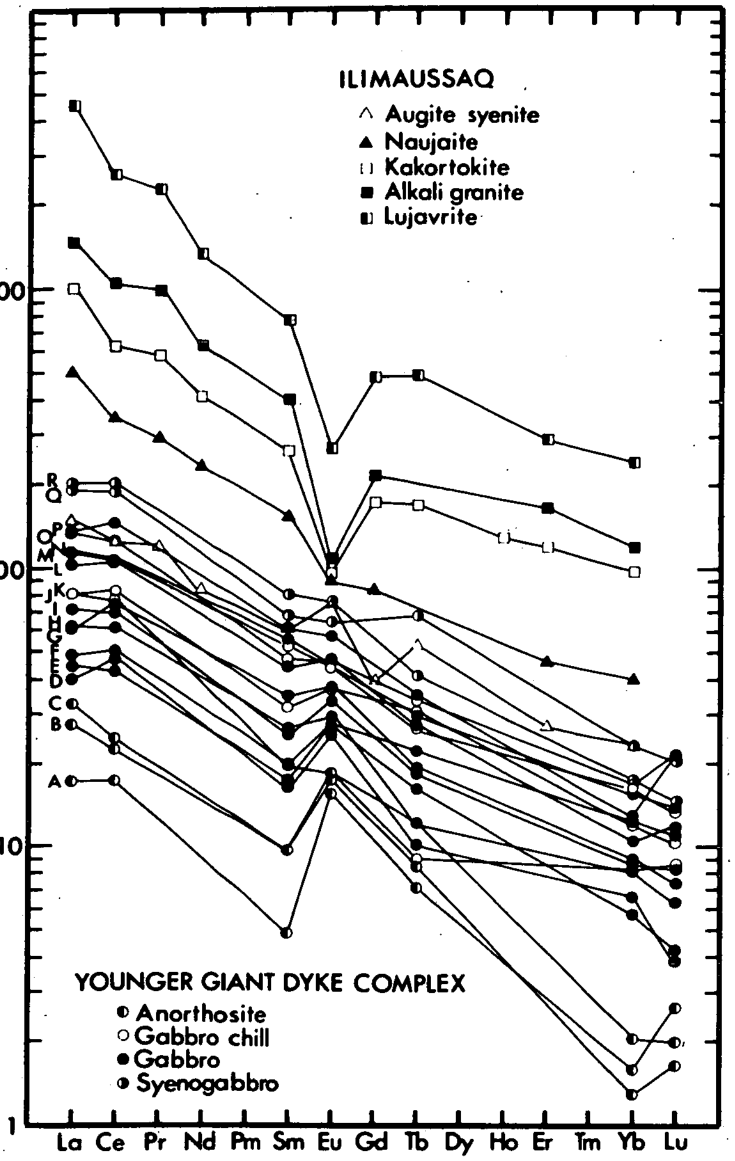

Fig. 27. Chondrite normalised REE distribution in samples from the Tugtutôq younger giant dyke complex (Table 32) and in the Ilimaussaq rock units (averaged values from Gerasimovskiy and Balashov, 1968; see Table 33).

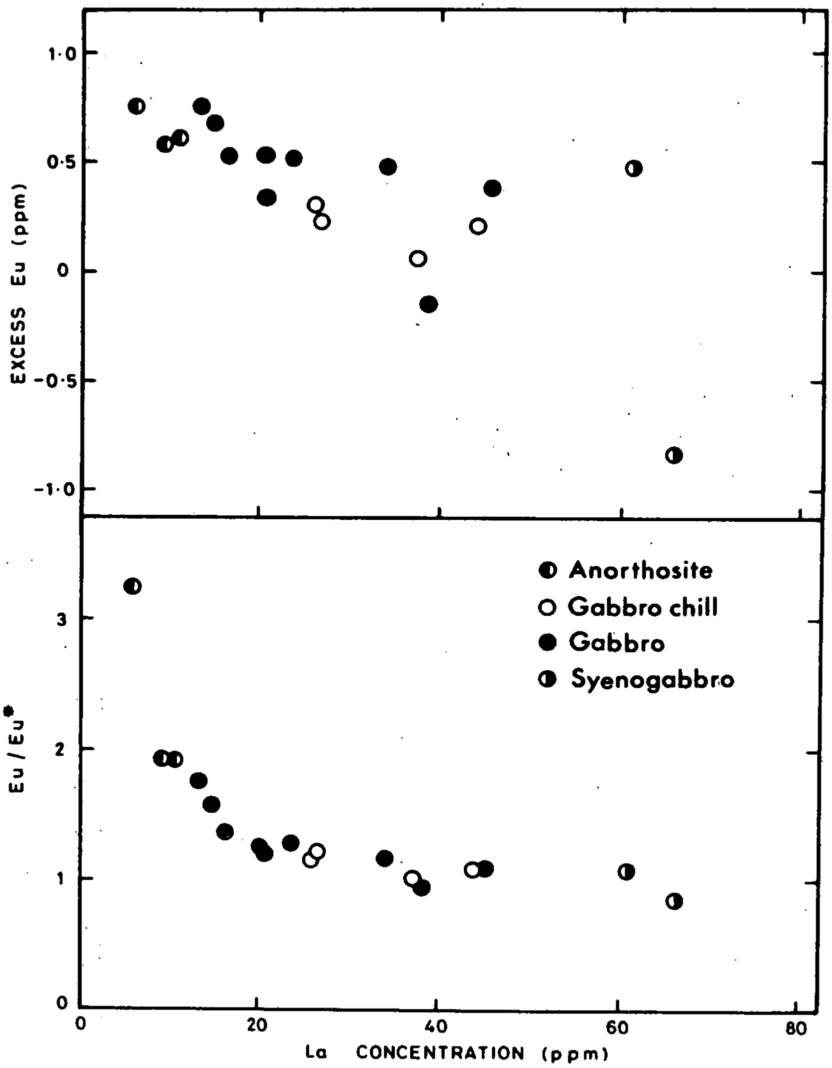

Fig. 28. Diagrammatic representation of Eu anomaly in samples from the Tugtutôq younger giant dyke complex. (a) shows actual excess Eu (ppm above interpolated Eu* value; (b) shows the ratio of actual to expected (Eu*) abundance. Both diagrams use La concentration as an approximate indicator of total REE content.

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ADDENDUM

In the text, "MSWD" (McIntyre et al., 1966) is equated to $[\mathrm{SUMS}/(n-2)]^{\frac{1}{2}}$ as stated by Brooks et al. (1972). According to R.J. Pankhurst (personal communication, 1976), this is incorrect and MSWD in fact is equivalent to SUMS/(n-2). Consequently, the values listed in the third column of Table 22 should be squared before use in a Chi-square or F statistical test. In particular it should be noted that estimates of scatter for the Motzfeldt and North Qoroq regression lines are thus greatly increased.

The York (1969) regressions take into account actual scatter of data as outlined in York's 1966 paper.