

RB-SR GEOCHRONOLOGY AND GEOCHEMISTRY OF PROTEROZOIC

BASIC INTRUSIONS IN SWEDEN AND SOUTH GREENLAND.

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Ph.D thesis

University of Edinburgh

September 1976



Declaration

I declare that this thesis is
entirely my own work, and has not
been submitted for a degree at any
other university.

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Abstract

23 dolerite dykes and sills and a syenite intrusion from central and southern Sweden, together with 7 basic dykes from south Greenland have been studied by Rb-Sr: the intrusions penetrate crust thermally unaffected since 1600 m.y., and emplacement ages lie between 1600 and 800 m.y. Dolerite ages are obtained by mineral isochrons and biotites from dolerites, and by mineral isochrons from rheomorphic granites. The isochrons depend upon a series of feldspar density fractions, but for a reliable emplacement age, all minerals should be concordant, and two or more independent ages should agree at 2σ .

In Sweden, 4 periods of dolerite intrusion are defined. Localised dykes were intruded in central Sweden at ca. 1560 and at ca. 1400 m.y. Widespread alkali-olivine dolerite sills of ca. 1250 m.y. in northern central Sweden correlate with similar activity in Finland, with the dykes of this study in south Greenland, and with widespread dolerite dykes in the Canadian Shield. Extensive dykes parallel to the 'Sveconorwegian Front' in central and southern Sweden were emplaced 1020-890 m.y. before present. The Vaggeryd syenite of southern Sweden was probably intruded ca. 1150 m.y. ago.

The ca. 1250 m.y. events correspond to a period of widespread continental fracture and basic magmatism in the North Atlantic region, clearly associated with large-scale tectonic processes. These may be due to Grenville-Sveconorwegian orogenic events, but a cause in pre- or even intra-Grenville continental separation seems more likely. Emplacement of the 1020-890 m.y. dykes is explained as due to craton fracture during post-orogenic uplift of the adjacent Sveconorwegian belt.

Initial $^{87}\text{Sr}/^{86}\text{Sr}$ studies yield variations of 0.703-0.707 for

dolerites over the whole Swedish area, of 0.703-0.712 for dolerites and more evolved rocks within single intrusions, and of 0.703-0.709 for feldspars within single dolerite samples: usually positive correlation of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ with Rb/Sr is observed. It is possible to produce the whole-rock variations largely by assimilation of rheomorphic melts of crustal rocks, while circulating hydrothermal fluids become important during crystallisation and cause contamination of minerals.

1.1 General age pattern

The oldest units of continental crust in the North Atlantic region are Archaean gneisses and supracrustal rocks in excess of 2600 m.y. old, and these are surrounded by mobile belts which developed from 1900 to 1700 m.y. From ca. 1600 m.y. onwards all these Archaean and Lower Proterozoic regions formed essentially stable cratons (fig. 1.1), into which inter alia the dolerite dykes and sills of this study were emplaced.

Mobile belts formed later than 1700 m.y. tend to be linear in contrast to the more irregular gross tectonic trends seen in the Lower Proterozoic and Archaean regions. Deformation and metamorphism in the Grenville-Sveconorwegian belt (fig. 1.1) culminated ca. 1100 m.y. ago, and the Caledonian orogeny (fig. 1.1) took place during the approximate interval 550 to 400 m.y.

Caledonian thrust sheets and/or Lower Palaeozoic cover overlie representatives of most groups of dolerite intrusions in Sweden, which are therefore pre-Caledonian and likely to be Precambrian in age. This study is in fact concerned with basic and other igneous events taking place in the 'pre-Grenville cratons' during the interval 1600 to 800 m.y.: the geological relationships of these events to tectonism in the Grenville-Sveconorwegian belt are not clear.

1.2 Pre-Grenville cratons

These regions, unshaded in fig. 1.1, are made up of crust of several different ages: an ad hoc division into 1900-1700 m.y. mobile belts and older (Archaean) regions is possible.

The oldest ages for the Archaean regions are in the range 3800 to 3400

m.y.: such ages are recorded from the north-central USA (Goldich and Hedge, 1974), coastal Labrador (Hurst et al., 1975; Barton, 1975), central-west Greenland (reviewed by Moorbath, 1975), northern Norway (Taylor, 1975) and the Kola peninsula (Tougarinov and Bibicova, 1973). A detailed chronology of events in the period 3800 to 3000 m.y. is being built up by the authors cited (recent reviews are by Moorbath, 1975 and Bridgwater et al., in press), but correlation from one area to another is still very uncertain. Large areas of the Archaean crustal regions, including almost all the Archaean in the Canadian Shield, were subject to a strong 2800- 2600 m.y. thermal and orogenic event: this is the earliest widespread tectonism clearly recognised to date in the north Atlantic region.

These pre-2600 m.y. regions formed essentially stable blocks during the 1900-1700 m.y. orogenic events, though some 'reworking' of older crust did occur: the 1900-1700 m.y. event is recorded over the whole north Atlantic region. In Canada K-Ar cooling ages fall between 1800 and 1650 m.y. (Stockwell, 1964), and in south Greenland van Breemen et al. (1974) have demonstrated major plutonism at ca. 1800 m.y., while K-Ar cooling ages are as low as ca. 1600 m.y. (Bridgwater, 1965). The events of similar age in Sweden have been reviewed by Welin (1970), who defines a depositional period up to 1900 m.y., major tectonism and plutonism 1900-1775 m.y., and a late tectonic period of uplift and granite etc. emplacement from 1775 to as late as 1540 m.y.

It is therefore clear that all the Archaean and Lower Proterozoic regions can be considered as forming stable cratons from ca. 1600 m.y. onwards, and that consideration of the tectonic regime of fracture and dolerite emplacement between 1600 and 800 m.y. need not greatly concern itself with any pre-1600 m.y. kinematic events.

1.3 The Grenville-Sveconorwegian orogenic belt

Following the 1900-1700 m.y. period, the next major tectonic event affecting the north Atlantic region was the Grenville-Sveconorwegian, occurring in Canada and Scandinavia (fig. 1.1); rocks of this age are known also from Scotland (Brook et al., 1976), arctic Canada (Sinha and Frisch, 1976), East Greenland (Hansen et al., 1974) and possibly northern France (Leutwein, 1968). In most of these areas the Grenville rocks have become involved in later deformation events.

The geology of the Grenville-Sveconorwegian belt has been reviewed by Magnusson (1965) in Scandinavia and Wynne-Edwards (1972) in Canada. Well over half the exposed belt consists of polymetamorphic gneisses, demonstrated to be at least partly of pre-Grenville age (Wynne-Edwards, 1969), while metamorphosed supracrustal rocks and granites of Grenville age form a more restricted component. Metamorphic facies are usually upper amphibolite to granulite (Touret, 1971; Wynne-Edwards, 1972), though restricted areas of low-grade metamorphism do occur (Wynne-Edwards, 1972).

Geochronologically the Grenville-Sveconorwegian belt has been rather hard to define. A comprehensive collation of 109 Rb-Sr whole-rock and U-Pb ages from within the belt is shown in fig. 1.2: the data from which the compilation is made are listed in appendix A, and do not include subsurface data from Canada or the USA, ages from Grenville gneisses within the Appalachian belt or from the Llano region of Texas. The mean 2σ uncertainty on these determinations is 52 m.y., consequently an interval of 100 m.y. is used for dividing up the population. The results above 1600 m.y. show that the age data are consistent with structural evidence in indicating a component of older rocks in the belt, both in Canada and Scandinavia. At what point results can be considered as

Grenville ages proper rather than 'relict' ages is however a matter of debate. The 'igneous' ages between 1600 and 1300 m.y. may correspond to a widespread period of intrusion of anorthosite and other more evolved rock types which took place during this time interval both within and beyond the Grenville-Sveconorwegian region. The earliest clearly metamorphic ages of the Grenville cycle appear in the 1300-1200 m.y. group, persisting down to 1000 m.y. Grenville ages that have been interpreted as igneous occur from 1300 to 900 m.y., but there is a marked concentration in the 1000-900 m.y. group; this reflects a widespread period of late tectonic granite emplacement in the Sveconorwegian portion of the belt, which has not been recorded in Canada. Many of the U-Pb zircon ages have been interpreted as being close to the actual age of metamorphic climax, but the matter of timing of metamorphic and plutonic events in detail remains an open question because of consideration of blocking temperatures of isotopic systems; in these terms many ages may be somewhat younger than the thermal maximum. Fig. 1.2 does not indicate actual thermal peaks but merely shows in what intervals the age determinations are concentrated. It is however apparent that Grenville-Sveconorwegian metamorphic and igneous events took place in the general interval 1300-900 m.y., with a slight possibility that some events may be a little older due to non-closure of isotopic systems.

Work in western Sweden by Welin and Gorbatschev (1976), Gorbatschev and Welin (1975) and Skiöld (1976) has shown that following emplacement of granites at ca. 1250 m.y., slates, sandstones and pillow lavas were laid down on an erosion surface: the slates give an Rb-Sr whole-rock isochron of ca. 1050 m.y. and were definitely deformed in at least the later stages of Grenville-Sveconorwegian orogenesis, giving K-Ar ages of 1000-750 m.y. Thus these data suggest tectonic quiescence, marine sedimentation and

basalt eruption at some time between 1250 and 1050 m.y., and this may imply that at least in western Sweden, the Grenville-Sveconorwegian orogeny was not continuous, but may have consisted of tectonic pulses separated by at least one tectonically quiet period. An alternative explanation could be that the ca. 1250 m.y. granites are essentially unrelated to orogeny, being atectonic igneous intrusions. At any rate it is possible by implication of the Swedish data that other breaks in the igneous/tectonic process, both in Scandinavia and Canada are hidden by the spread of radiometric ages from 1300 to 900 m.y.

In addition to the essential geochronological coherence of the whole belt demonstrated by Rb-Sr and U-Pb results, the K-Ar ages form a strong unifying feature. All parts of the belt in Canada and Scandinavia show a general range of K-Ar ages (all samples) from 1130 to 800 m.y. (Magnusson, 1960; Stockwell et al., 1968), testifying to cooling during this time interval (Harper, 1967). Large portions of this range can however, be observed in single areas by use of different rocks and minerals. In an area in the western part of the Grenville of Canada Macintyre et al. (1967) found a range of K-Ar ages for hornblende, biotite, muscovite and nepheline from 1070 to 840 m.y. irrespective of metamorphic grade or rock type. In southern Norway, O'Nions et al. (1969) studied K-Ar ages in metamorphic rocks and pegmatites and found ranges of 1120-970 m.y. for hornblendes, 1110-1010 m.y. for pegmatitic micas and 1040-970 m.y. for micas from metamorphic rocks. In western Sweden, Skiöld (1976) studied the slates and pillow lavas deposited during the interval 1250 to 1050 m.y., and found a range from 1000 to 750 m.y. for whole-rock slates, whole-rock pillow lavas and pillow lava hornblendes, while mica ages from one of the late-Sveconorwegian granites (Rb-Sr whole-rock age 910 ± 35 m.y.) gave 860-680 m.y. These K-Ar studies show that the

Grenville-Sveconorwegian belt was being cooled during the interval 1130 to 800 m.y., and that locally some systems did not become blocked until as late as ca. 700 m.y. Interpretation of this range of K-Ar ages into post-metamorphic uplift history is well-established in principle (Harper, 1967), but it is not clear how much of the cooling interval corresponds to actual uplift. The Rb-Sr whole-rock and U-Pb ages between 1100 and 1000 m.y. seem to make post-tectonic uplift prior to 1000 m.y. unlikely.

A tentative history of the Grenville-Sveconorwegian belt might thus be defined. Anorthosites and related granitic etc. plutons were emplaced between 1600 and 1300 m.y. ago in both America and Scandinavia. Some anorthosites may have been intruded later than this however, and this form of plutonism apparently extended through most of the Grenville-Sveconorwegian tectonic cycle (Barton and Doig, 1972; Verstevee, 1974). The main tectonic and plutonic events of the orogeny proper took place within the general time range 1300 to 1000 m.y., though there is considerable uncertainty as to detailed timing of events, and quiescent periods are probably disguised within this long interval (Skiöld, 1976). Certainly no major metamorphism was taking place after 1000 m.y., and the K-Ar ages documenting cooling from ca. 1130 m.y. onwards support this view. This cooling continued until ca. 800 m.y., even as late as 700 m.y. in some areas of the Sveconorwegian portion of the belt. At approximately 950-900 m.y. this region was affected by late-tectonic batholith emplacement, an event which has not been recorded in Canada. The data emphasise the essential coherence of the whole Grenville-Sveconorwegian belt, and extension of Grenville activity beyond this region is demonstrated by Rb-Sr and U-Pb ages from Scotland, East Greenland and arctic Canada (see appendix A), by a whole-rock K-Ar age of 987 m.y. from the Rockall Bank, West of Scotland (Miller et al., 1973), by K-Ar ages

between 1100 and 930 m.y. from basement in northern France (Leutwein, 1968), and by the numerous inliers of Grenville age in the USA (e.g. Muehlberger et al., 1966; Odom et al., 1973).

The Grenville-Sveconorwegian belt, characterised by K-Ar ages in the range 1130-800 m.y., is separated from regions giving much older K-Ar ages by a narrow tectonic zone, in Canada the Grenville Front, in Sweden the 'Schistosity Zone'. This 'Sveconorwegian Front' in Sweden is a zone 10-25 km wide consisting of a variety of rocks intensely sheared along vertical planes (Magnusson, 1965). In Canada the Grenville Front (15-20 km in width) is more complex and may be expressed as a fault, an abrupt metamorphic transition or a zone of shearing. Dalziel et al. (1969) have shown however, that the Grenville Front is fundamentally a steeply inclined zone of mylonitisation less than 2 km wide, and that the faults which intersect the front are later features unrelated to its original formation. The Grenville and Sveconorwegian Fronts thus divide regions which were cooled during the period 1130 to 800 m.y. from regions unaffected by thermal events since Archaean or Lower Proterozoic times. Actual uplift of the Grenville-Sveconorwegian belt associated with the cooling seems to have taken place after ca. 1000 m.y. This suggests that these tectonic zones were the loci of considerable relative movement between the Grenville-Sveconorwegian and pre-Grenville-Sveconorwegian 'crustal blocks' at some time during the interval ca. 1130-800 m.y., and probably post-1000 m.y. This scheme is supported by seismic work across the Grenville Front in Canada which indicates that the Front is a deep fracture penetrating into the mantle lithosphere (Mereu and Jobidon, 1971). The seismic data could be interpreted as being consistent with powerful uplift of the tectonically thickened Grenville block relative to the pre-Grenville block, though the data do not define the amount of

relative movement.

Structural evidence from Canada (Dalziel et al., 1969; Brocoum and Dalziel, 1974) indicates firstly that the mylonites of the Grenville Front zone may have been formed much earlier than the Grenville, even as early as Hudsonian (ca. 1800 m.y.) time, and secondly that they do not of themselves constitute clear evidence for shearing motions. The argument for relative uplift of the Grenville and Sveconorwegian 'blocks' is however based entirely upon the observed K-Ar ages, the association of which, in orogenic belts with uplift and cooling is widely accepted (see Harper, 1967). Thus while it is possible that, in addition to the complex history claimed for the Grenville Front, future detailed work in Sweden could also demonstrate an older original age for the 'Sveconorwegian Front', the actual origin of these tectonic boundaries does not affect the conclusion that they acted as the loci of a large portion of the relative movement of Grenville/Sveconorwegian and pre-Grenville/Sveconorwegian blocks during the general interval 1130-800 m.y.

The apparent period of uplift and cooling for the Grenville-Sveconorwegian belt from ca. 1130 to ca. 800 m.y., locally even as late as 700 m.y., is long by comparison with other orogenic belts (e.g. Dewey and Pankhurst, 1970). This would be consistent with tectonic overthickening of the crust of the Grenville belt due to continental collision and 'underplating' (Dewey and Burke, 1973), or with very slow uplift or possibly with both these features. Additionally, it is possible that some of the earlier K-Ar ages date argon blocking close to the metamorphic events and are essentially unrelated to the uplift history of the belt. It is however clear that uplift was taking place at least from 1000 m.y. onwards, as given by the falling-off in frequency of U-Pb and Rb-Sr whole-rock ages (see fig. 1.2), and the main concentration of K-Ar

determinations.

1.4 Igneous rocks of the period 1600-800 m.y. occurring in the pre-Grenville cratons

In an early review of geochronology in the midcontinental USA, Goldich et al. (1966) discussed ages of mainly felsic intrusive and extrusive rocks which fell between 1700 and 1000 m.y. and assigned them to periods of igneous activity apparently unrelated to major tectonism. Later work in the region has tended to confirm this picture: in Colorado, Peterman et al. (1968) and Vera and van Schmus (1974) recognise a period of granite emplacement at 1450-1400 m.y., while Hedge (1970) dated one batholith as late as 1040 m.y.

In eastern Canada most igneous activity in the range 1600-1300 m.y. has been referred to the so-called 'anorthosite event', already discussed in connection with the Grenville Province. Rocks associated with anorthosite in the Grenville belt have given ages between 1550 and 1100 m.y. (see appendix A), and felsic rocks associated with the Harp Lake and Michikamau anorthosites lying on the eastern coast of Canada and outside the Grenville belt have given U-Pb zircon ages of ca. 1450 m.y., while a ca. 1300 m.y. result was obtained at the nearby Nain anorthosite (Krogh and Davis, 1973).

In Sweden, the pre-Sveconorwegian crust is penetrated in two regions by granitic plutons (in one case associated with anorthosite) which give U-Pb and/or Rb-Sr ages between 1500 and 1300 m.y. (Welin and Blomqvist, 1966; Welin and Lundqvist, 1975; Larsen, pers. comm. 1975). These occurrences are treated in more detail in chapter 4.

The existence of long-lived anorthosite-granite-rhyolite activity across the north Atlantic region has given rise to speculation concerning

an abnormal heat-flow regime in the continental crust of this zone during the interval ca. 1500 to 1300 m.y. (Herz, 1969).

The pre-Grenville crust during the period 1600-800 m.y. was also affected by emplacement of alkaline complexes: these are somewhat dispersed in Scandinavia and Canada, but are concentrated in the Gardar Province of south Greenland, where alkaline magmatism took place repeatedly between ca. 1330 and ca. 1150 m.y. (Blaxland et al., in press).

Perhaps the most striking, and certainly the most widespread of the cratonic igneous rocks occurring in the pre-Grenville cratons are swarms of dolerite dykes and sills, with accompanying lavas where these have not been removed by erosion. The dolerites occur over almost all pre-Grenville regions and are of various ages. Gates and Hurley (1973) have shown that several major swarms in the Canadian Shield were emplaced at ca. 2700 m.y. and ca. 2000 m.y. Other groups, including all dolerites in the Baltic Shield must be post-1700 m.y., because they cut rocks deformed at this time; those in the range 1300-800 m.y. are shown in fig. 1.1. All evidence relating to the age of these igneous events will be discussed in chapters 3, 4 and 5 of this study. The dolerite intrusions of Sweden, together with selected dykes from south Greenland form the basis of this work, and the ages obtained are compared to similar igneous events in Canada and to ages of tectonic activity in the Grenville-Sveconorwegian orogenic belt.

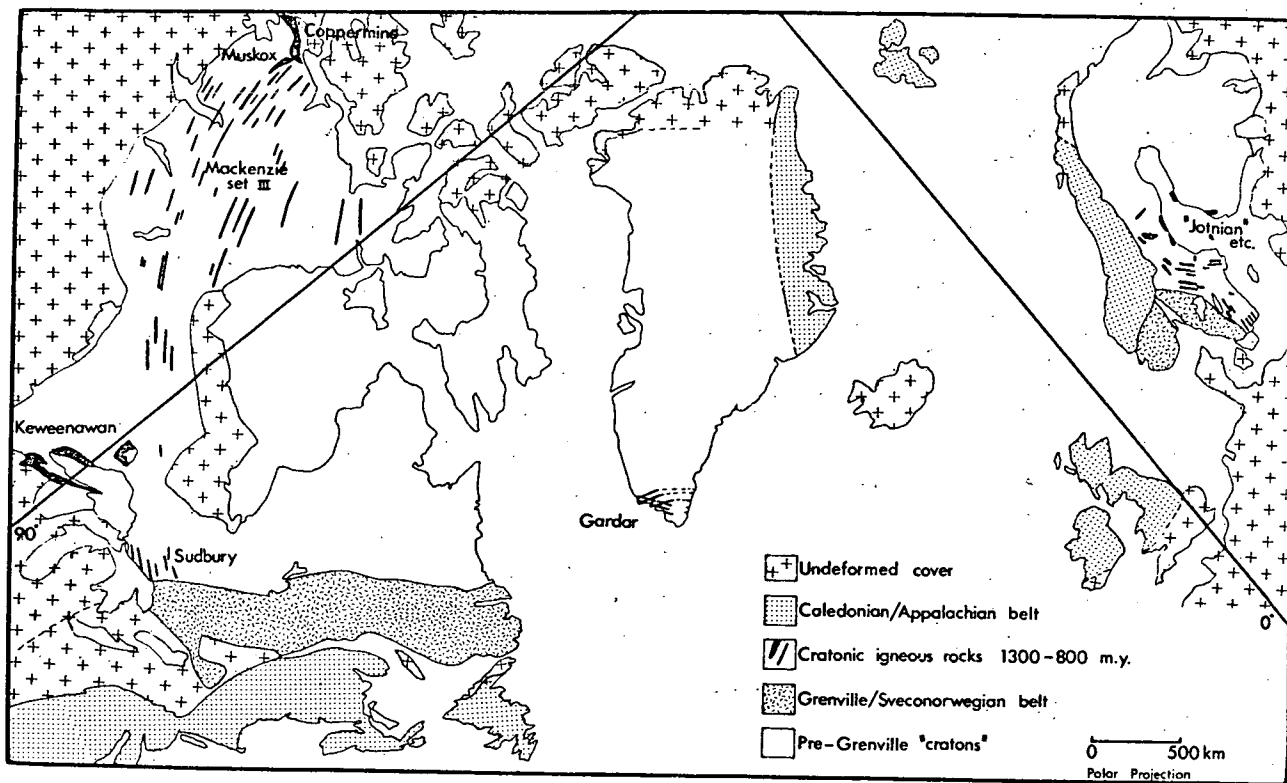


Fig. 1.1 Simplified tectonic map of the North Atlantic region showing the distribution of cratonic igneous rocks emplaced between 1300 and 900 m.y. relative to the Grenville- and Caledonian-age orogenic belts.

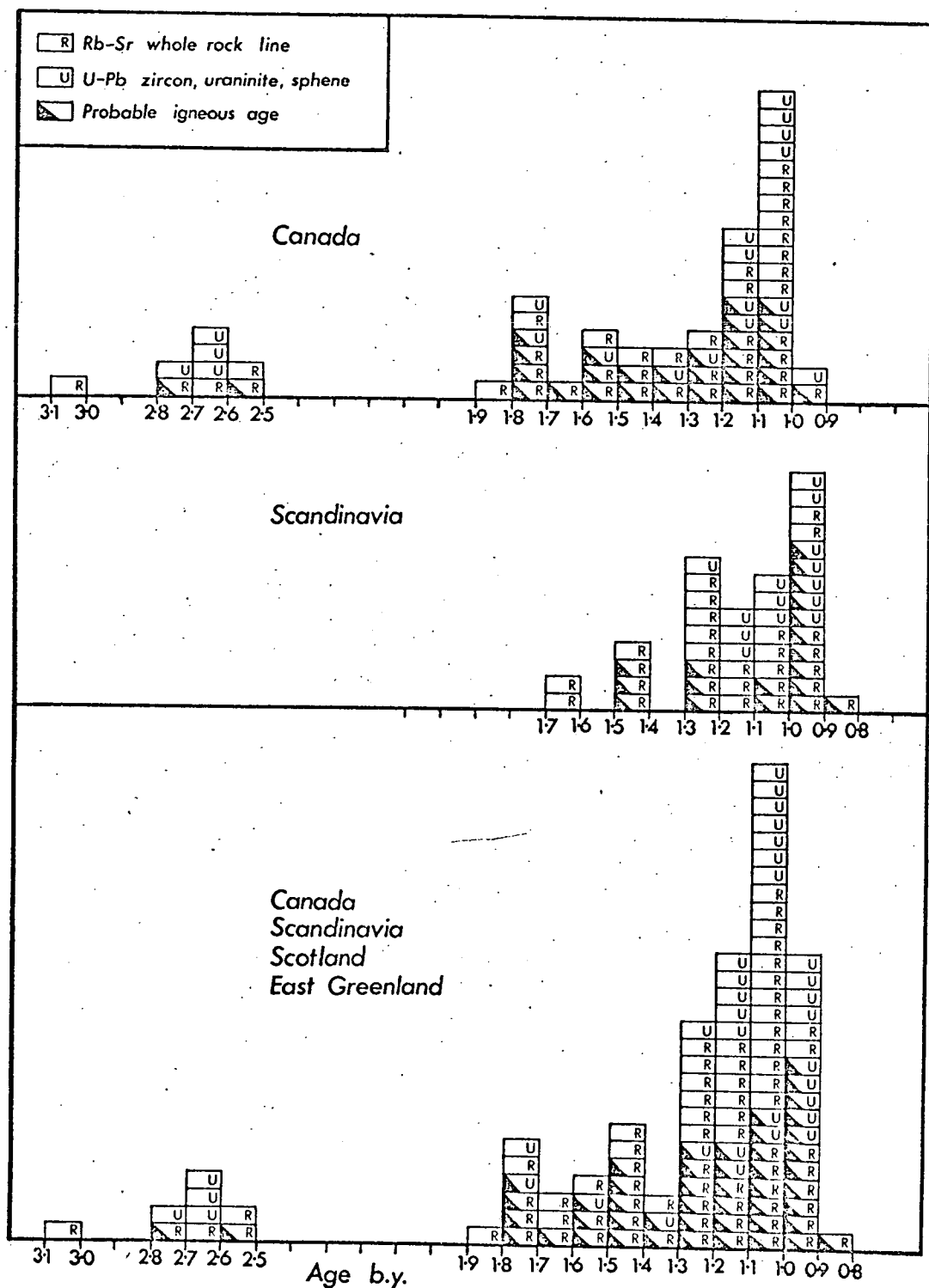


Fig. 1.2 Histograms of ages by Rb-Sr whole-rock and U-Pb methods from the Grenville-Sveconorwegian belt and related areas in the North Atlantic region. Top, from Canada only; centre, from Scandinavia only; bottom, from all North Atlantic areas.

2.1 Scope and purpose of the study

It is now widely accepted that the Caledonian orogeny in the north Atlantic region was the result of closure of an ocean by lithospheric subduction: structural, geochronological and palaeontological evidence all support this conclusion (e.g. Dewey, 1971). The period of ocean opening necessitated by this hypothesis is however of unknown age.

The preceding major orogeny, the Grenville-Sveconorwegian event is, as discussed in section 1.3, the earliest in the north Atlantic region to display a linear trend. Earlier belts have characteristically variable tectonic trends, and could have been deformed without plate movement (Sutton and Watson, 1974; Christie et al., 1975). The question therefore arises as to whether the Grenville-Sveconorwegian belt is the earliest manifestation of present-day style plate tectonics in the region.

The study of dolerites in Sweden and south Greenland was undertaken to investigate the possibility of identifying phases of major tectonism and/or ocean opening in the period between the 1900-1700 m.y. thermal events and the Caledonian (ca. 550-400 m.y.) event, as reflected by fracture and basic magmatism in hitherto stable continental areas. Additionally it was hoped to find a correlation between igneous events in the Baltic and Canadian/Greenland shield areas.

The age study presented here runs in parallel with palaeomagnetic work: Dr. J.D.A. Piper collected independently in Greenland, while in Sweden both collecting and research by Dr. G. Bylund have been closely coordinated with this study. The aims of the joint Swedish work have been twofold: first, to obtain geochronologically controlled pole positions with a view to establishing a polar wander path for the Baltic Shield in

the Upper Proterozoic, and second to investigate relationships between poles of similar age in Canada/Greenland and Scandinavia. Interpretation of the palaeomagnetism of the north Atlantic region from ca. 1300 to 800 m.y. is hampered by the problem of lack of age control on poles, particularly those from rocks of the Grenville-Sveconorwegian orogenic belt. This has led to widely diverging hypotheses for the disposition of continents, plate motions and the origin of the Grenville structural province (summarised by Irving et al., 1974). Preliminary results (Patchett and Bylund, in preparation; Bylund and Patchett, unpub. data) suggest that two main phases of dolerite emplacement in Sweden, ca. 1250 and 1000-900 m.y., give results that are useful in interpreting the disposition of north Atlantic continents at these times, and shed further light on the palaeomagnetic 'Grenville problem'.

The strontium isotope composition of basic igneous rocks forms a basis for discussion both of magma origin and of the evolution of mantle strontium through time. Peculiarities in the isotopic composition of dolerites and basalts emplaced into continental crust have been previously observed (see section 6.2), and a secondary aim of this study was to investigate by means of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios the degree of interaction between the basaltic magmas and continental crustal material, on what scales this occurs, and whether contamination plays a major role or not in the compositional evolution of the magmas.

2.2 Principal methods of research

In the past, Precambrian basic igneous rocks have been dated largely by the K-Ar method on whole-rocks, hornblendes, pyroxenes or biotites. In all cases, results are rather badly scattered (e.g. Leech, 1966), and it is likely that most samples have been affected to some degree by argon

loss.

In an early study, van Schmus (1965) dated a dolerite dyke of the Sudbury swarm in Canada (see fig. 1.1) by Rb-Sr on biotite from the dolerite and by use of a mineral isochron from a rheomorphic vein recrystallised from the country rock; the age of ca. 1250 m.y. has been substantiated by palaeomagnetic work. Gates and Hurley (1973) dated several Canadian swarms of Archaean age by use of the Rb-Sr whole-rock isochron method: despite the low Rb/Sr ratios, the whole-rock samples had developed a sufficient range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios over ca. 2500 m.y. to render a whole-rock isochron feasible.

In this study the whole-rock isochron technique is not used for dating because the samples have a low range of Rb/Sr ratios, and because it is shown that whole-rock lines may be subject to large upward errors resulting from contamination of higher Rb/Sr samples by radiogenically-enriched Sr from crustal rocks (chapter 6). Methods used comprise:-

1. Mineral isochrons from dolerite samples, usually consisting of 4-6 feldspar separates plus whole-rock, clinopyroxene and/or apatite.
2. Mineral isochrons from rheomorphic granite samples, which recrystallised at the time of dolerite intrusion; the isochrons usually consist of feldspars and whole-rock only.
3. Whole-rock--biotite pairs from dolerite samples.

The dolerite mineral isochrons depend mainly upon a series of density fractions of feldspar numbered F1 onwards such that the highest-numbered

separate is the least dense and usually has the highest Rb/Sr ratio. The whole-rock, apatite and clinopyroxene exert little influence over the isochron age.

Procedural blanks during the course of this work were 8 ng Rb and 4 ng Sr: these are significant outside estimated analytical uncertainty only in the case of clinopyroxenes with low Rb contents (less than 2 ppm), which are corrected for Rb blank contribution. Mass spectrometry, for which spiked samples were always used, was performed to the accuracy required for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, which depends upon the range of Rb/Sr (and hence $^{87}\text{Sr}/^{86}\text{Sr}$) ratios of the samples. For the anorthosites and gabbros from south Greenland a 1σ error of 0.00002-0.00003 was achieved, while most mineral and whole-rock samples from dolerites were determined to a 1σ error of 0.00002-0.00010. Biotites and samples from granites and syenites were performed under 'routine' conditions and here the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio has the estimated 1σ uncertainty of 0.04%, based on a large number of previous replicate analyses. The laboratory obtains results for the NBS 607 K-feldspar standard which agree closely with those published, and a value for the Eimer and Amend SrCO_3 standard based on 16 analyses by J. Hutchinson over the general period of this study is 0.70804 ± 0.00012 (2σ). NBS standard stoichiometric salts are used for spike calibration. Detailed analytical techniques are given in appendix B.

All results, both ages and initial ratios presented or quoted in this study are calculated using $\lambda^{87}\text{Rb} = 1.39 \times 10^{-11} \text{ yr}^{-1}$. Regression calculations for ages are performed by the method of York (1969): all errors are given at the 2σ level, and except in the rare cases where the scatter of points can be accounted for entirely in terms of experimental uncertainty, the quoted error figure consists of both an experimental and

a geological or 'scatter' component. For whole-rock--biotite pairs however, statistical regression treatment is not possible, and here the quoted error is entirely of analytical origin based upon estimated uncertainties calculated from a large number of duplicate analyses (see also section B.7). The whole-rock--biotite pairs are therefore not subject to any internal test of significance, and their interpretation must depend entirely upon whether they give geologically meaningful results.

2.3 Isotopic closure of mineral systems

Since the intrusion ages of this study are based upon various mineral systems, it is important to consider at what temperature these systems become blocked to isotopic exchange. In view of the importance of closure temperature in studies of metamorphic rocks, the topic has been considered by several authors.

A closure temperature for Rb-Sr dates on Alpine biotites is estimated at ca. 300°C from consideration of metamorphic facies and rate of uplift by E. Jäger (1973). Dodson (1973) has arrived at the same result for Alpine biotites by calculation from diffusion constants. The cooling rates for intrusive sheets given by J.C. Jaeger (1957) show that the centre of a dolerite sheet 100 m thick emplaced into cold upper crustal rocks will cool to 300°C after less than 1000 years, while for sheets 400 m and 1000 m thick the times are less than 10,000 and less than 100,000 years respectively. Since in many cases the dykes and sills of this study penetrate flat-lying continental sediments resting unconformably on granites and gneisses, then they can be regarded as being intruded into a cool upper crustal environment, and for all assumptions of thickness, biotite in a dolerite sill or dyke will cool to its closure

temperature in times which are negligible relative to the age.

Many of the ages presented in this study are based upon alkali feldspar, which forms the highest Rb/Sr point on most mineral isochrons. Hart (1964) studied resetting of mineral ages near an intrusive contact, and concluded that out of Rb-Sr and K-Ar on alkali feldspar, K-Ar on hornblende and Rb-Sr and K-Ar on biotite, the Rb-Sr alkali feldspar ages were least affected by resetting. On the other hand, Arriens et al. (1966) reported isotopic disturbance of alkali feldspar in granites where biotites were unaffected and gave the original age. This observation is in accordance with data on granophyres and granites from this study and suggests that feldspars may be open to isotopic disturbance of a kind differing from that produced by heating alone. O'Neil and Taylor (1967) have demonstrated essentially complete cation exchange between ultra-fine feldspar grains and 2-4 molal alkali chloride solutions at 1 kb fluid pressure and down to a temperature of 350°C. Reaction times were between 1 hour and a few days. Extrapolation to natural conditions where feldspars are coarser and solutions probably more dilute is uncertain, but the ease of reaction does suggest that given sufficient time, significant exchange might be expected in natural conditions at low temperatures, possibly even lower than 300°C. In this way feldspars could lose ^{87}Sr to an aqueous fluid phase at lower temperatures than those necessary to reset biotite in the same rock.

New data presented here and arguments based on published data show that biotites, and where observed, alkali feldspars from country rocks in Sweden and south Greenland have not been isotopically disturbed later than 1700-1600 m.y. On this basis the mineral ages for dolerites are considered to represent the time of emplacement rather than some isotopic resetting event. Feldspar lines for dolerites are considered to give

reliable ages only when apatite and/or clinopyroxene are concordant. In addition, ages of intrusion are considered wholly reliable only when two or more determinations agree at 2σ , and whenever possible these are performed by different Rb-Sr mineral methods (section 2.2). Given the unmetamorphosed state of the rocks, agreement of different mineral methods at 2σ rules out to a large extent any possibility of a mild isotopic disturbance, which new data show to be extremely selective in its effects (section 6.9).

3.1 Introduction

The Gardar Province of south Greenland was the subject of a classic study by Wegmann (1938), who distinguished most of the major rock types, including the dolerite dykes, or 'Brown Dykes' which form the principal subject of this isotopic study. Since that time the province has been extensively and intensively studied, with a large proportion of the interest centring on the alkaline rocks of the central intrusive complexes. Upton (1974) and Emeleus and Upton (in press) have summarised the current knowledge of the province.

Dykes to which a Gardar age has been assigned occur from Frederickshåb in the North to Kap Farvel in the South, and appear also in east Greenland (Bridgwater and Gormsen, 1969). The main Gardar area of south-west Greenland (fig. 3.1) comprises a repetitive association between central complexes of alkaline felsic rocks and dykes and dyke complexes composed dominantly of dolerites and gabbros, though dykes corresponding in composition to many of the alkaline rocks do occur.

3.2 Chronology of Gardar events

Early K-Ar work (Moorbath et al., 1960; Bridgwater, 1965) suggested that Gardar activity spanned an interval 1400 to 1000 m.y. before present. Recently however, systematic Rb-Sr isochron work has been undertaken by van Breemen and Upton (1972), Blaxland and Parsons (1975), Blaxland et al. (1975), Blaxland (1976) and Blaxland et al. (in press). These workers have demonstrated that Gardar magmatism took place within the limits 1330 to 1150 m.y. before present, and was episodic in nature.

Over 60% of the exposed Gardar rocks were emplaced at or around 1170

m.y., when intense dyking along two ENE-trending zones through Nunarssuit-Isortoq and Tugtutôq-Igaliko was accompanied by intrusion of central alkaline felsic complexes (fig 3.1). The Rb-Sr isochron ages of ca. 1170 m.y. are from the central complexes, but all dyking is either bracketed by these ages (as in the Tugtutôq-Igaliko zone), or is believed to precede the complexes by a relatively short time (as in the Nunarssuit-Isortoq zone). This Late Gardar magmatic activity was confined to the central and eastern parts of the province (fig. 3.1).

Earlier Gardar events are chronologically less well-defined. Two complexes in the area to the North and West of the region of Late Gardar activity have given ages of ca. 1250 m.y. (fig. 3.1); the Kungnât Fjeld complex at 1245 ± 17 m.y. (Blaxland and Upton, unpub. data) and the nearby Ivigtut granite at 1248 ± 25 m.y. (Blaxland, 1976). In the same area the Grønnedal-Ika complex has given an age of 1327 ± 17 m.y., while two complexes within the eastern part of the Gardar have ages of ca. 1300 m.y. (Blaxland et al., in press).

3.3 Early and Mid-Gardar basic igneous events

The isochrons for the central complexes (see preceding section) leave the following events with poorly defined ages:-

1. Intrusion of a variety of E- to NE-trending dykes in the northwestern Gardar (Ivigtut region), which post-date Grønnedal-Ika (ca. 1330 m.y.), and pre-date Kungnât and Ivigtut (ca. 1250 m.y.).
2. Intrusion of ESE-trending basic dykes in the central and eastern Gardar (Nunarssuit and Julianehåb regions), which pre-date all Late Gardar (ca. 1170 m.y.) activity (e.g. Upton, 1964).

3. Sedimentation and volcanic eruption to form the Eriksfjord Formation of the eastern Gardar, which pre-dates the oldest central complex of that region (ca. 1310 m.y.).

This study does not deal with the Eriksfjord Formation (3), but the dolerite dykes of 1 and 2 are shown in fig. 3.2, and these, together with some lamprophyre dykes from the Ivigtut region form the subject of this study. The intention was to date the dykes of the northwestern (Ivigtut) region to discover at what time within the bracket 1330-1250 m.y. they were emplaced, and to investigate the relationship (if any) between these dykes and the ESE-trending dolerite dykes in the Julianehåb region of the eastern Gardar.

The relative chronology of Gardar events in the Ivigtut region has been described by Berthelsen and Henriksen (1975), who give reference to more detailed field accounts. Table 3.1, simplified from table 8 of these authors, gives the sequence of events.

Table 3.1 General Gardar chronology for the Ivigtut region

Intrusions	Faulting
Ivigtut granite complex	
Kungnât complex	
	0° dextral
Alkaline dykes	110° sinistral
	25-50°, 160-180° dextral
Microporphyritic dolerites	
	60-80° sinistral & dextral
Olivine dolerites BD2	
BD1	
	90-120° sinistral
BDO	
Big-feldspar dykes	
Trachytic to granophyric dykes	
Basic dykes	
Lamprophyric dykes	
Grønnedal-Íka complex	

The olivine dolerite dykes predominate volumetrically over other types, and are chronologically important in the dyke sequence, as well as defining approximately the initiation of faulting activity.

The olivine dolerite dykes are divided into three generations, BDO, BD1 and BD2, which are believed to be synchronous over the whole Ivigtut region (Berthelsen and Henriksen, 1975). Earlier dykes (ED0) trend more

nearly East than later ones (BD1, BD2), which run ENE or NE. Combined with this strike change through time is one in space: all three generations, and in particular the BDO, show a progressive clockwise rotation of trend when followed from the northwestern to the southeastern part of the Ivigtut region. Thus the BDO dykes in the North-west trend $60-70^{\circ}$ (East of North), while in the South-east the dominant trend is $100-110^{\circ}$. This relationship is apparent from fig. 3.2, but see Berthelsen and Henriksen (1975) for a complete picture. These authors have explained all Gardar faulting and dyking in this region by a maximum principal stress oriented E-W, changing to NE-SW with time, and a complementary minimum principal stress changing from N-S to NW-SE; dykes developed at right angles to the minimum stress direction.

If the changing trend of the olivine dolerites from 60° in the Northwest to 110° in the Southeast is extrapolated a little further, then the next dykes to the Southeast would be expected to trend $115-120^{\circ}$. This is the observed trend of the early olivine dolerites of the Nunarssuit and Julianehåb regions of the central and eastern Gardar (fig. 3.2), and in addition to petrographic similarity, this gives grounds for suspecting that these dykes may be related to those of the Ivigtut region. In fact the dykes have been designated 'BDO' in the Nunarssuit and Julianehåb regions (e.g. Upton, 1964), though the correlation has remained doubtful.

The dykes described are commonly 10-50 m wide and are olivine dolerites composed of plagioclase, olivine (often altered), augite, opaques, biotite, apatite and occasionally interstitial quartz. The dykes are distinguished from other Gardar dolerite or gabbro dykes by their consistent ophitic texture, the rather regular development of

pegmatitic patches within the bodies, the general lack of any phenocrysts and the absence of mineral layering or lamination. Analyses of these olivine dolerites suggest a mildly alkaline basaltic magma with $\text{SiO}_2 = 46\text{-}50\%$ (Winstanley, 1975; Upton, unpub. data).

Additional dykes of this group may be present in other areas of the eastern Gardar. Olivine dolerite dykes partly of ESE trend have been reported by Allaart (1969) in the area to the East of Ilímaussaq, and many of the dykes in the region to the North of Igaliko trend ESE where they are the earliest dolerite dykes (Walton, 1965). The correlation of these dykes with those of the Nunarssuit and Julianehåb regions is however not established.

The country rocks of the Ivigtut region comprise a range of quartzofeldspathic gneisses of Archaean age (e.g. Pidgeon and Hopgood, 1975) while the Nunarssuit and Julianehåb regions are composed of Ketilidian Julianehåb granite dated at 1730 ± 20 m.y. (van Breemen et al., 1974). Mineral age studies show that the country rocks of the Gardar Province were last cooled through K-Ar and Rb-Sr blocking temperatures of minerals at ca. 1600 m.y. (Bridgwater, 1965; van Breemen et al., 1974).

3.4 Age of lamprophyre dykes from the Ivigtut region

Location fig. 3.2; sample localities section C.1; analytical data section D.1.

The two dykes studied are ultramafic rocks carrying phlogopitic biotite. This together with olivine pseudomorphs, fresh clinopyroxene and the very low feldspar content shows that they correspond to the most common type of lamprophyre in the general Ivigtut region, reported to be the earliest major dyking event (table 3.1).

In view of the olivine and feldspar alteration in these dykes and the low Rb/Sr ratios of the whole-rocks as determined by X-ray fluorescence, the unaltered biotite was separated and whole-rock--biotite pairs used for age determination. The results, 1254 ± 29 and 1276 ± 29 m.y. are subject to larger 2σ errors than usual for micas because of the low Rb/Sr ratios (ca. 1.0) of the biotites, but they suggest an age of intrusion for the lamprophyres of between 1300 and 1230 m.y. This result agrees with a K-Ar biotite age on a lamprophyre from the Ivigtut peninsula of 1275 m.y. (Larsen, 1966).

3.5 Age of olivine dolerite dykes from the Ivigtut region

Location fig. 3.2; sample localities section C.2; analytical data section D.2.

In view of the agreement between age determinations by mineral isochrons and biotites from dolerites in Sweden (chapter 4) and the Julianehåb region of the Gardar Province (section 3.6), and since the country rocks have not been not affected by any event sufficient to reset biotites later than 1600 m.y., it was thought adequate to date the olivine dolerites of the Ivigtut region by means of whole-rock--biotite pairs alone.

Biotite was separated from three samples. 161003, from the northern of two prominent BDO dykes ca. 2 km WSW of Ivigtut (fig. 3.2) gives whole-rock--biotite pairs of 1277 ± 18 and 1278 ± 18 m.y. (duplicate biotite determinations), while 161002, from the southern of these BDO dykes gives 1272 ± 18 m.y. 161001, from a ca. 100 m wide BD1 dyke ca. 3 km SW of Ivigtut (fig. 3.2) gives 1265 ± 18 and 1263 ± 18 m.y. (biotite duplicated). The results, all of which agree at 2σ , suggest an age of 1260-1280 m.y. for emplacement of BDO and BD1 dykes in the Ivigtut

region. Since the BD2 generation was intruded in some cases while BD1 was still hot (Berthelsen and Henriksen, 1975), then this age can be assigned to all three ED swarms.

3.6 Age of olivine dolerite dykes from the Julianehåb region

Location fig. 3.2; sample localities section C.3; analytical data section D.3; isochron diagrams figs. 3.3, 3.4.

The 'BD0' swarm of the Julianehåb region in the eastern part of the Gardar province consists of three prominent dykes which persist for 100 km or more. The westernmost and easternmost of these were sampled on Tugtutôq and in Julianehåb Havn respectively (fig. 3.2). The single samples were separated for biotite, apatite and feldspars and the results are shown in figs. 3.3 and 3.4.

Mineral lines of whole-rock, apatite and feldspars define ages of 1285 ± 144 m.y. (Tugtutôq) and 1244 ± 117 m.y. (Julianehåb). The high 2σ errors are caused by scatter in excess of analytical uncertainty, $[\text{SUMS}/(n-2)]^{1/2}$ values being 3.99 and 4.99 respectively. The reason for these relatively large components of 'excess' error is unclear; for Swedish samples, mineral isochrons consistently matched estimated analytical uncertainty. It seems possible that there has been some minor isotopic equilibration between zones of feldspar in the Gardar samples at some time after intrusion. Since the two mineral lines have similar slopes and identical initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios they could be combined (14 points) to yield an age of 1257 ± 83 m.y., where $[\text{SUMS}/(n-2)]^{1/2} = 4.29$; strictly however, the analyses should be retained in their single-sample groups.

Biotites from these two samples give ages of ca. 1240 m.y. (figs. 3.3, 3.4), which confirm that despite the scatter, the mineral isochrons

do in fact give the approximate age of intrusion. This agreement testifies to the minor nature of any disturbance which has affected the feldspars; the only alternative possibility, that both feldspar and biotite ages are the result of resetting at ca. 1240 m.y. is rejected because the emplacement of these dykes is the only thermal event recorded between ca. 1600 m.y. (late Ketilidian) and ca. 1170 m.y. (Late Gardar). The biotites are taken as defining the 'best' age of intrusion at ca. 1240 m.y.

3.7 Assessment of Mid-Gardar (1280-1240 m.y.) magmatic events

Table 3.2 Ages of dykes from the Gardar Province

Sample	Dyke	Number of points	Age m.y.	Initial $^{87}\text{Sr}/^{86}\text{Sr}$
81119	Lamprophyre	2	1254 $\pm$ 29	0.7051 $\pm$ 0.0006
86192	Lamprophyre	2	1276 $\pm$ 29	0.7036 $\pm$ 0.0006
161003	BDO	2	(1) 1277 $\pm$ 18	0.7061 $\pm$ 0.0006
		2	(2) 1278 $\pm$ 18	
161002	BDO	2	1272 $\pm$ 18	0.7050 $\pm$ 0.0006
161001	BD1	2	(1) 1265 $\pm$ 18	0.7026 $\pm$ 0.0006
		2	(2) 1263 $\pm$ 18	
186506	'BDO'	7	1285 $\pm$ 144	0.7049 $\pm$ 0.0005
		2	1245 $\pm$ 18	0.7046 $\pm$ 0.0006
186564	'BDO'	7	1244 $\pm$ 117	0.7050 $\pm$ 0.0005
		2	1238 $\pm$ 18	0.7046 $\pm$ 0.0006

The direct conclusions arising from the ages presented here (table 3.2) are as follows:-

1. In the Ivigtut region, the earliest dykes, lamprophyres, and the principal dykes of the area, the olivine dolerites BD0, BD1 and BD2 were probably emplaced after ca. 1280 m.y. Assuming that lamprophyres on the Ivigtut peninsula correlate with those at Arsuk, this means that at least the majority of Gardar dyking and faulting in the Ivigtut region is of post ca. 1280 m.y. age. There are thus either few or no significant events in the region between ca. 1330 m.y. (Grønnedal-Íka) and ca. 1280 m.y. (probable initiation of dyke intrusion), and this would mean that the dykes and faults are to be regarded as a series of tectonic and magmatic events shortly preceding emplacement of the Kungnât and Ivigtut complexes, rather than as following Grønnedal-Íka.

2. The ESE-trending olivine dolerites of the Nunarssuit and Julianehåb regions, known as the 'BD0' dykes, have ages of ca. 1240 m.y., which shows that they are probably part of the same magmatic events as the 1280-1240 m.y. rocks of the Ivigtut region.

Arising from 1 and 2 it is clear that the dykes of this study, together with all other dykes of the Ivigtut region and the central complexes of Kungnât and Ivigtut comprise a distinct group of magmatic events within the Gardar Province between 1280 and 1240 m.y. This Mid-Gardar 'magmatic cycle' is distinct both from those complexes giving ages greater than 1300 m.y. (Early Gardar), and from the intense

magmatic events of the Late Gardar at ca. 1170 m.y. The magmatic cycle is not, however sufficiently distinct at 2 σ from the age of one of the Igaliko complexes, dated at 1295 $\pm$ 61 m.y. (Blaxland et al., in press), and continuous activity from ca. 1330 to ca. 1240 m.y. must be considered possible.

Evidence supporting the contention that these Mid-Gardar dykes all belong to the same tectonic and magmatic regime comes from the continuously rotating trend of the olivine dolerites. Because the Labrador Sea and the Inland Ice restrict exposure to the West and East respectively, these dykes form an apparent fan-shaped array diverging from a point some 30 km West (offshore) from Arsuk (see fig. 3.2). This apparent radial pattern may have no significance tectonically, but serves to emphasise the coherence of the whole dyke swarm. The common structural and petrographic characteristics of the dykes detailed in section 3.3 are also unifying features.

Consideration of the whole range of rock types listed in table 3.1 shows that the Mid-Gardar magmatic events comprised an association between alkaline ultramafic magma (lamprophyres), copious mildly alkalic basaltic magma (BDs, Big-feldspar dykes etc.), and a variety of more differentiated magmas, both over- and undersaturated, which formed dykes and central complexes. This range of magmatic products is typical of both the Early and particularly the Late Gardar magmatic 'cycles'. The Mid-Gardar cycle at 1280-1240 m.y. is therefore to be seen as a distinct magmatic maximum within a coherent igneous province which was developed from ca. 1330 to ca. 1170 m.y.

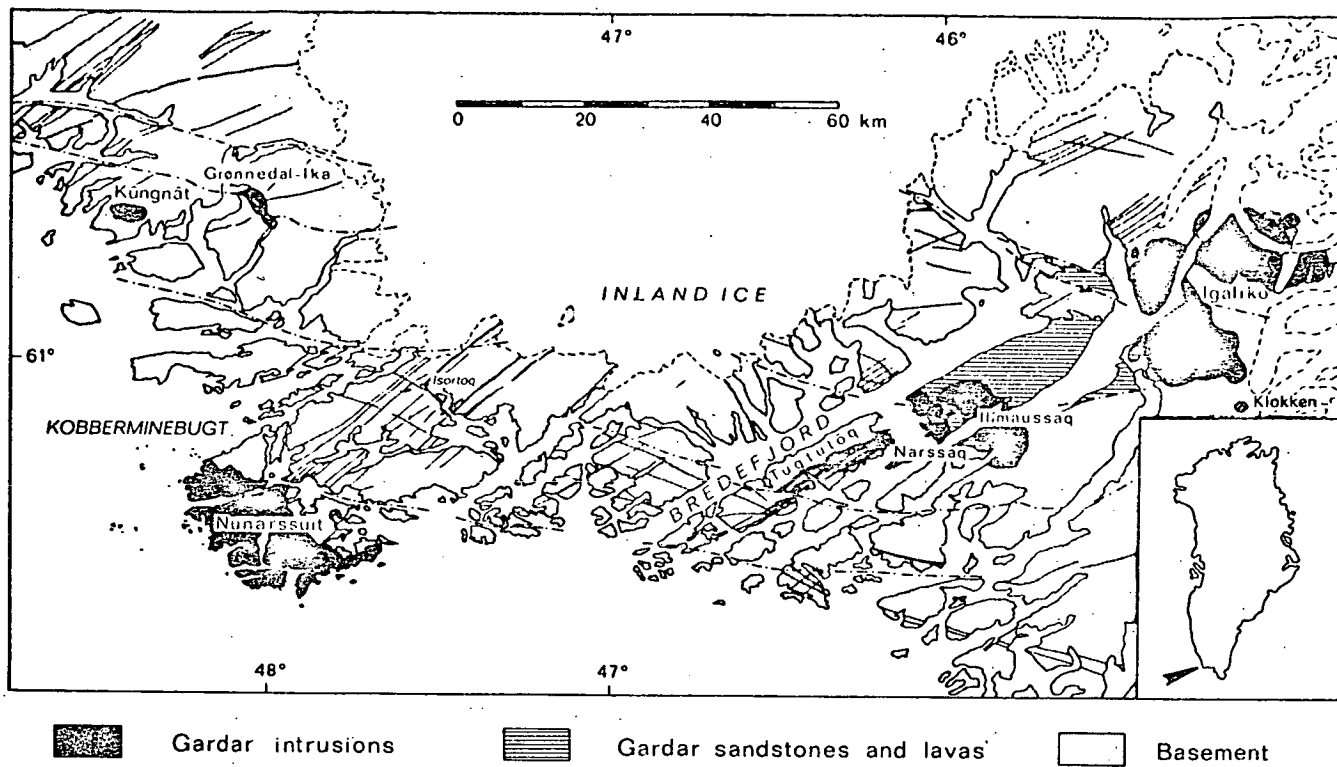


Fig. 3.1 Simplified geological map of the Gardar Province.

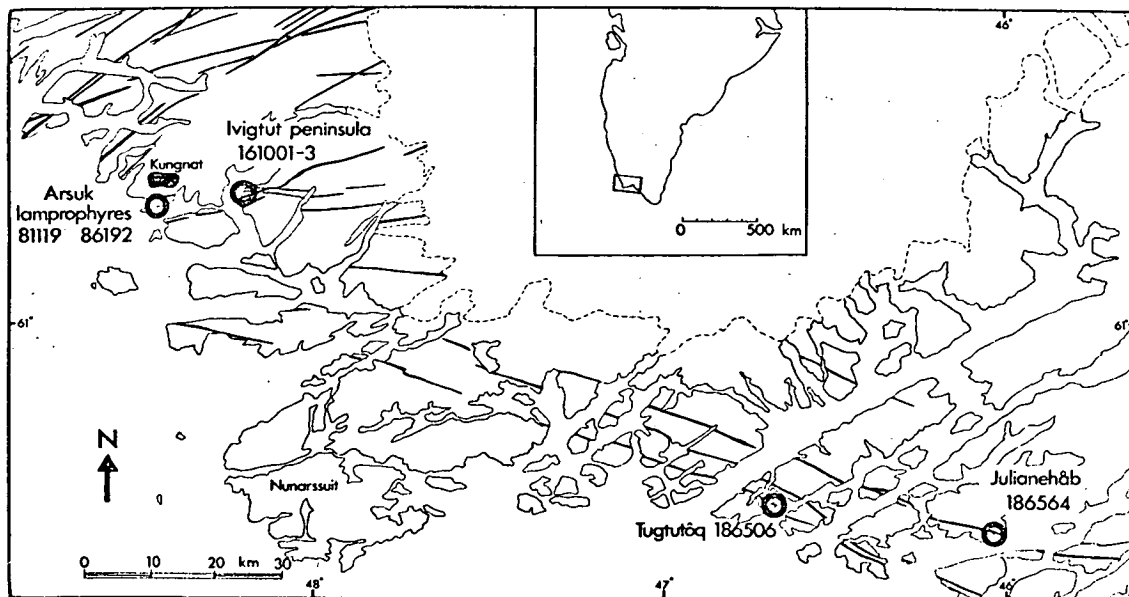


Fig. 3.2 Map of the Gardar Province showing rocks of ca. 1250 m.y. age: the olivine dolerite dykes and the Kungnát complex. The ca. 400 m diameter Ivigtut granite complex at Ivigtut is omitted for reasons of scale.

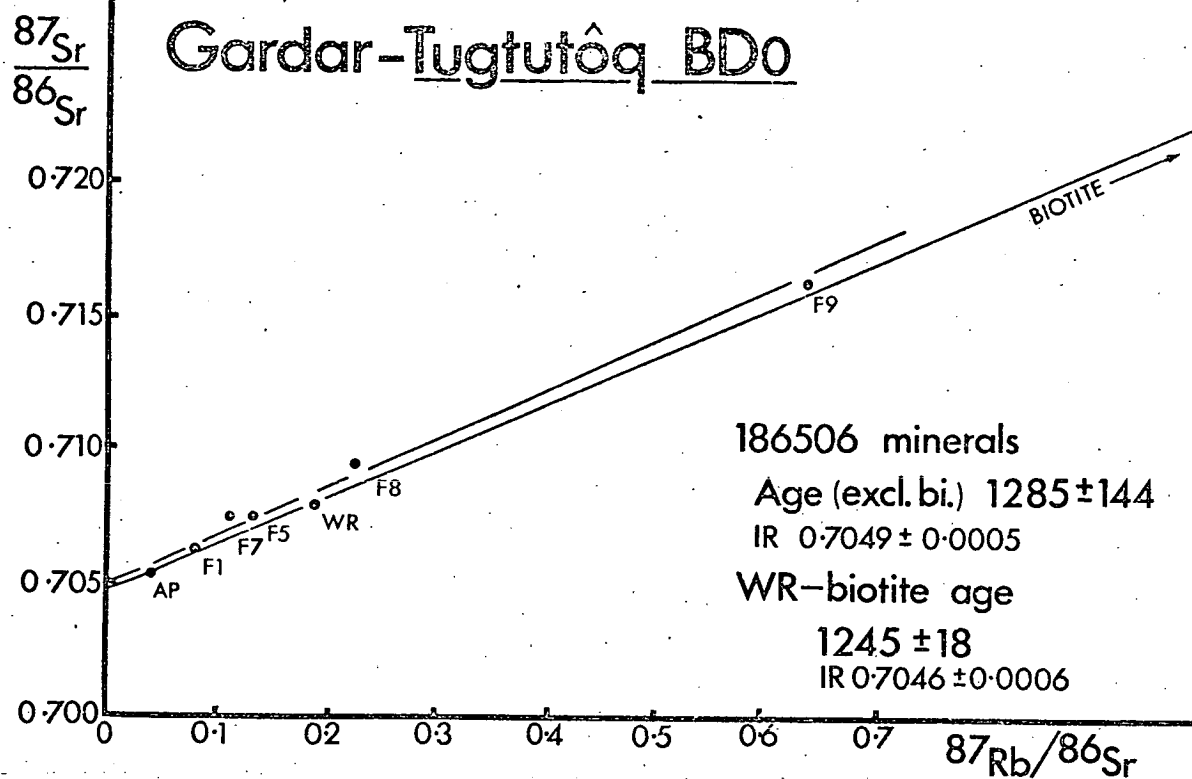


Fig. 3.3 Mineral line and whole-rock--biotite age for an ESE-trending olivine dolerite ('BDO') from Tugtutôq. For location see fig. 3.2.

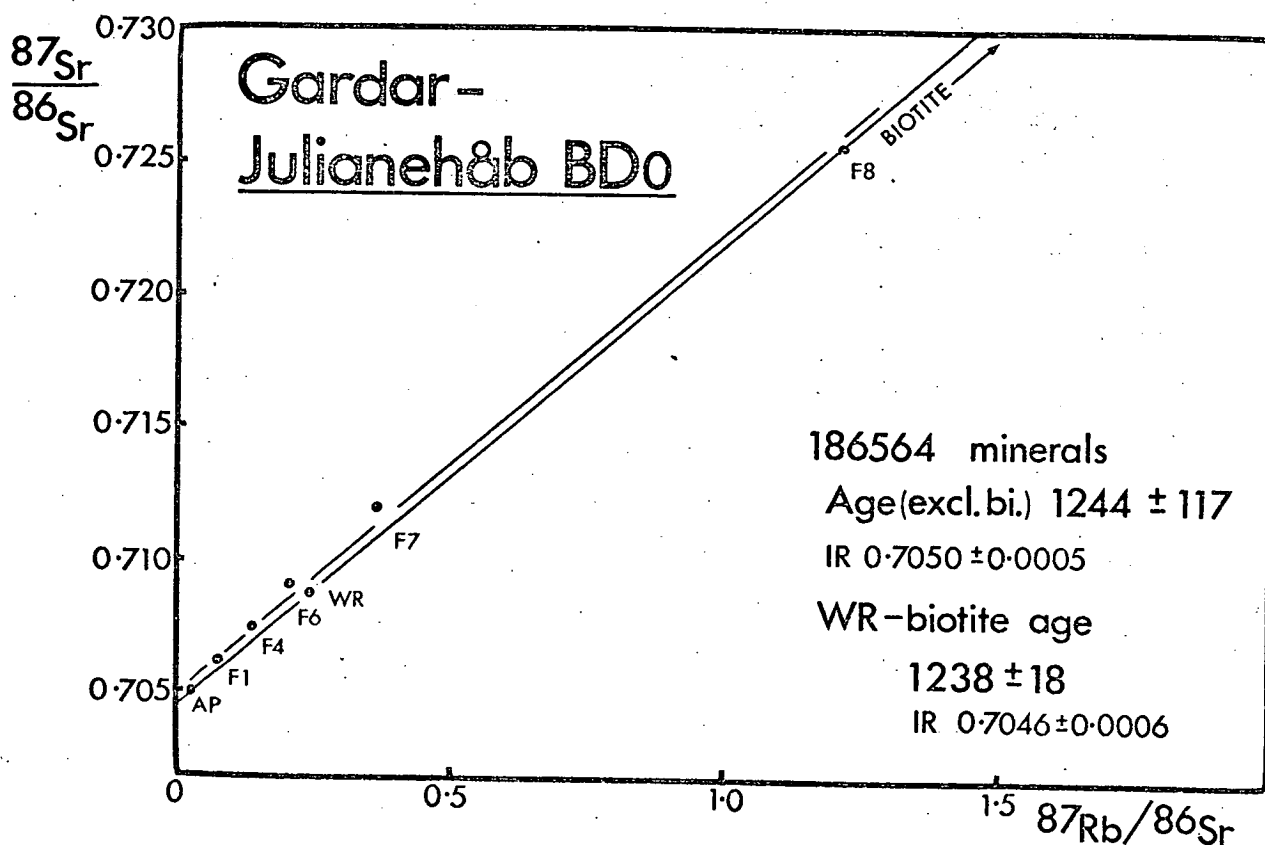


Fig. 3.4 Mineral line and whole-rock--biotite age for an ESE-trending olivine dolerite ('BDO') from Julianeh&b. For location see fig. 3.2.

4.1 Post-Svecofennian igneous and sedimentary events

U-Pb zircon studies in Finland give ages in the region of 1900 m.y., the oldest yet obtained for true Svecofennian rocks (Wetherill et al., 1962; Kouvo and Tilton, 1966). The Rapakivi granites, a major post-tectonic group of plutons, are dated by zircons at 1700 m.y. (Kouvo, 1964).

In Sweden all ages have been measured by Rb-Sr; the oldest is ca. 1800 m.y. (Welin et al., 1970), and many isochron ages fall in the range 1800-1700 m.y. (as reviewed by Wilson and Nicholson, 1972). There are in addition several significantly younger isochron ages: late-kinematic granites and acid volcanics in northern and central Sweden give results between 1700 and 1600 m.y. (Welin and Lundqvist, 1970; Welin et al., 1971), and an even younger group of post-kinematic granites and syenites confined to northern Sweden gives ages of ca. 1550 m.y. (Welin et al., 1971; Gulson, 1972).

The difference between the distributions of ages in Finland and in Sweden is not necessarily of geological significance, but may have a methodological cause, since two different techniques have been used in the two countries. All the rocks described however, are construed in this study as being part of a coherent Svecofennian tectonic and magmatic cycle, which terminated at or before ca. 1550 m.y.

It is with the events which followed the final stabilisation of the Svecofennian crustal segment that this study is concerned. All such events, sedimentary or igneous, are shown in fig. 4.1, and these subdivide into four types:-

1. Granite and syenite intrusions

Late cross-cutting plutons of granite or syenite cut the Svecofennian in several places. The Karlshamn group of granites in southeastern Sweden (fig. 4.1) is dated at 1455 m.y. by K-Ar and U-Pb on zircon, uraninite and monazite (Magnusson, 1960; Welin and Blomqvist, 1966). This is confirmed by K-Ar and Rb-Sr determinations from Bornholm to the South of Karlshamn (Larsen, 1971 and pers. comm., 1975). The Vaggeryd syenite of central southern Sweden (fig. 4.1) is included in this group on the basis of the age reported here. It lies wholly within the 'Schistosity Zone' marginal to the Sveconorwegian orogenic belt, and has been strongly affected by the shearing deformation of that region. In northern central Sweden intrusions of anorthosite, gabbro, rapakivi granite and pegmatite (including the Ragunda and Nordingr  complexes, fig. 4.1) are dated at 1350 m.y. (Ragunda complex) and 1415 ± 30 m.y. (Nordingr  granite), while basic members of the Nordingr  complex are apparently up to 200 m.y. older (Welin and Lundqvist, 1975).

2. 'Jotnian' continental sediments

Sediments, mainly sandstones and shales, which have been called 'Jotnian', unconformably overlie the Svecofennian crust in several places (fig. 4.1). The Dala Sandstone Formation includes a series of basalt lavas. The lower age limits of the formations are given by the local basement; in Dalarna, 1669 ± 38 m.y. (Welin and Lundqvist, 1970), in the Nordingr  region 1415 ± 30 m.y. (Welin and Lundqvist, 1975), and in the Almes kra region 1695 ± 20 m.y. ( berg, 1972). The sediments in all cases are only slightly affected by later tectonic events.

3. Dolerite intrusions

The Svecofennian crust of central and southern Sweden is penetrated by numerous swarms of dolerite sills and dykes (fig. 4.1): they are not reported from northern Sweden. Wherever intersections are observed, these are later than the granites and continental sediments of groups 1 and 2. They have previously been dated only by the K-Ar method, which yields scattered results.

4. Alkaline complexes

The locations of three post-Svecofennian alkaline complexes, Norra Kärr, Alnö and Särna are shown in fig. 4.1. No intersections with rocks of groups 1, 2 or 3 are ever observed. Many K-Ar determinations date Alnö at ca. 550 m.y., Norra Kärr has been dated by the Rb-Sr whole-rock method at 1580 ± 62 m.y. (Blaxland, in press). The Särna complex however, was of unknown age until covered by this study.

4.2 Post-Svecofennian dolerites

The results of the age work have suggested the following classification of post-Svecofennian dolerites in Sweden:-

1. An East-West trending group of dykes lying West of Stockholm in central Sweden (figs. 4.2, 4.3).
2. A locally-developed group of NNE trending dykes, the Tuna dykes, in southern Dalarna, central Sweden (figs. 4.2, 4.3).

3. A large number of sills of alkali-olivine dolerite in northern central Sweden (figs. 4.2, 4.3).
4. Groups of dykes and occasional sills characterised by their trend parallel to the margin of the Sveconorwegian orogenic belt (figs. 4.4, 4.5).

General locations for these swarms are given in figs. 4.2-4.5, together with the areas covered by the detailed geological and sample location maps of appendix C. For all analytical data see appendix D.

4.3 East-West dyke system

The East-West dyke system of central Sweden is exposed over an area 120 km (E-W) by 60 km (N-S) lying to the West of Stockholm (fig. 4.2), and intrudes Svecofennian gneisses and granites. The group consists of two giant dykes and a variety of smaller dykes.

The smaller dykes of the system (fig. C.1) are olivine-bearing and olivine-free dolerites, and include many dykes carrying plagioclase phenocrysts and occasional anorthosite xenoliths (Gorbatshev, 1961). Two of these dykes are selected for study, a 20 m olivine dolerite from Forskilen (fig. C.1) and a 2 m plagioclase-phyric dyke from Gåstorp (fig. C.1). The Hällefors giant dyke (fig. C.1) consists of a marginal olivine dolerite facies intruded by a more differentiated central facies porphyritic in pyroxene and feldspar (Krokström, 1936). Gorbatshev (1961) has drawn comparisons between the rock types present in the Hällefors dyke and in the smaller dykes. The Breven giant dyke (fig. C.2) consists of a western portion made up entirely of olivine dolerite and an eastern portion where a roughly central granophyre body occurs (Krokström,

1932); the emplacement of this felsic magma has produced a variety of hybrid rocks by interaction with the dolerite. Krokström (1932, 1936) recognised several intrusive phases in the dyke and interpreted the granophyre as a partial melt of crustal rock.

Gåstorp

Location figs. 4.2, 4.3; sample locality section C.4, fig. C.1; analytical data section D.4.

The sample SÖ-54 is from an aphyric part of a dyke carrying large plagioclase crystals, and the whole-rock--biotite pair defines an age of 1565 ± 22 m.y.

Forskilen

Location figs. 4.2, 4.3; sample locality section C.4, fig. C.1; analytical data section D.4; isochron diagram fig. 4.6.

The olivine dolerite sample SÖ-49 was separated into clinopyroxene and several feldspar fractions and the analyses define an isochron of 1594 ± 79 m.y. (fig. 4.6): the value of 2.65 for $[\text{SUMS}/(n-2)]^{1/2}$ indicates an isochron relationship. The whole-rock--biotite pair from this sample gives 1543 ± 23 m.y., which agrees at 2σ with the mineral isochron, and is interpreted as defining the intrusive age of the dyke.

Hällefors

Location figs. 4.2, 4.3; sample localities section C.5, fig. C.1; analytical data section D.5; isochron diagram fig. 4.7.

Analytical data for whole-rocks and minerals from basic rocks of the Hällefors dyke define scattered patterns on isochron diagrams and are interpreted as the result of post-intrusive isotopic disturbance (see

section 6.9). The only data that can be sensibly interpreted as having age significance come from a 6 cm rheomorphic vein (SÖ-36) derived from the country-rock gneiss at the margin of the dyke, which was separated into feldspar fractions. This is a fine grained granophyric rock hybridised to some extent with the dolerite magma, and in order to eliminate feldspars which originate from the dolerite, only the five lightest (i.e. most potassic) fractions were analysed. The five-point isochron has an age of 1551 ± 38 m.y. (fig. 4.7), and $[\text{SUMS}/(n-2)]^{1/2} = 1.32$. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, 0.717 ± 0.002 is consistent with this rheomorph having been derived from rocks considerably older than 1550 m.y., and the age of recrystallisation at 1551 ± 38 m.y. is in agreement with the inferred emplacement ages of the Gåstorp and Forskilen dykes. The Hällefors dyke is therefore also interpreted as having been intruded at this time.

Breven

Location figs. 4.2, 4.3; sample localities section C.6, fig. C.2; analytical data section D.6.

Analytical data for whole-rock and mineral samples from the eastern part of the Breven dyke, where the granophyre occurs, define divergent patterns on isochron diagrams and are interpreted as the result of similar post-intrusive isotopic disturbance to that affecting Hällefors (see section 6.9). Olivine dolerite from the eastern portion of the dyke where the granophyre is absent was separated for biotite, and whole-rock--biotite pairs from two samples, SÖ-3 and SÖ-4, define ages of 1581 ± 23 and 1575 ± 23 m.y. respectively. These agree with the results from other East-West dykes.

Summary

The reliable age results for the East-West dyke system are summarised in table 4.1, and are illustrated in fig. 4.3.

Table 4.1 Summary of age data for the East-West dyke system

Intrusion	Sample	Method	Number of points	Age m.y.	$[\text{SUMS}/(n-2)]^{1/2}$
Gåstorp	SÖ-54	WR-biotite	2	1565 $\pm$ 22	-
Forskilen	SÖ-49	Dolerite minerals	7	1594 $\pm$ 79	2.65
		WR-biotite	2	1543 $\pm$ 23	-
Hällefors	SÖ-36	Rheomorph minerals	5	1551 $\pm$ 38	1.32
Breven	SÖ-3	WR-biotite	2	1581 $\pm$ 23	-
	SÖ-4	WR-biotite	2	1575 $\pm$ 23	-

The convergence of data on ca. 1560 m.y. indicates that the dyke swarm was emplaced at this time; age differences within the 1590-1540 m.y. interval are not detectable by present techniques.

4.4 Tuna dykes

Location figs. 4.2, 4.3; sample localities section C.7, fig. C.3; analytical data section D.7; isochron diagram fig. 4.9.

In southern Dalarna, central Sweden occurs a well-defined swarm of dykes, trending NNE and mapped over an area of ca. 50 km dimensions (Hjelmqvist, 1966). These, known as the Tuna dykes include aphyric dolerite, plagioclase-phyric dolerite, porphyritic intermediate rocks and

a range of alkali feldspar-quartz porphyries. The dykes are individually less than 20 m in width, but may occur in dense swarms, composite dykes being also common.

Samples from two localities are analysed: the three with lowest Rb/Sr ratios (fig. 4.9) are from a locality where basic and intermediate rocks occur, and the three with highest Rb/Sr ratios are from a locality where alkali feldspar-quartz porphyries only are exposed. In view of the small number of samples used, three of the analyses were duplicated and their position on the isochron diagram confirmed. The line obtained (fig. 4.9) has an age of 1401 ± 50 m.y., an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.703 ± 0.002 and $[\text{SUMS}/(n-2)]^{1/2} = 4.89$. This value indicates that errors are outside analytical uncertainty, but is not so large that serious doubt of the 1401 ± 50 m.y. age should be entertained. The large error in the initial ratio disallows any conclusion as to the ultimate origin of the magmas.

4.5 Alkali-olivine dolerite sills

An association between remnants of 'Jotnian' continental sediments and intrusive olivine dolerites is well described throughout north-central Sweden. The occurrences of the sediments are shown in fig. 4.1, and the distribution of the olivine dolerites which intrude both the sediments and the underlying basement are shown in fig. 4.2. Including related occurrences in Finland, the total area affected by intrusion of the sills is ca. 500 km (E-W) by 400 km (N-S). In the Gävle area Gorbatshev (1967) has shown that some sills were emplaced prior to cementation of the sandstones, whereas others were post-cementation, but elsewhere there is no evidence as to the length of time which separates 'Jotnian' sedimentation and olivine dolerite intrusion.

In northern Dalarna the sills, which are up to hundreds of metres thick have been divided petrographically into 'Åsby' and 'Särna' types (Hjelmqvist, 1966). The Åsby type contains olivine, plagioclase, clinopyroxene, opaques, apatite, biotite and interstitial alkali feldspar: the Särna type differs from this in its low abundance of apatite and biotite and in often lacking an alkali feldspar phase. Major element analysis of a typical Åsby and a typical Särna dolerite shows that the Åsby variety has distinctly higher TiO_2 , Na_2O , K_2O , P_2O_5 and Fe/Mg ratio for the same SiO_2 content as compared to the Särna dolerite. The Särna type is confined to the westernmost parts of Dalarna, and all the sills penetrating the basement to the East of the Dala Sandstone Formation are of the Åsby type. Petrographic and chemical data for sills of the Lös-Hamra area (Lundqvist, 1968), included in the 'northern Dalarna' of this study, show that these are of the Åsby type. This applies also to the Nordingrå region (Lundqvist and Samuelsson, 1973), the Gävle region (Gorbatshev, 1967) and to the Finnish occurrences (Sederholm, 1934; Kahma, 1951). The sills and occasional dykes of Jämtland contain olivine, plagioclase, clinopyroxene, opaques, apatite, biotite and interstitial alkali feldspar, and are thus on mineralogical grounds to be compared to the Åsby dolerites of Dalarna. No layering was observed by the author in any of the intrusions sampled, but in the Nordingrå region Larson (1973) and Lundqvist and Samuelsson (1973) have described rhythmic layering in well-exposed sills. By implication this may also be present in less well-exposed intrusions inland.

Over the entire region of north-central Sweden there is a tendency for the development of 'monzonitic dolerites' and 'monzonites': these are related to dolerite by increase of the alkali feldspar content and introduction of quartz, the most felsic monzonites being granophyres of

sub-granitic composition.' In minor occurrences these rocks are often marginal to the dolerite bodies and are probably produced by hybridisation of basic magma with country rock (e.g. Hjelmqvist, 1966 p.150), but in other cases extensive areas of monzonite occur and may even constitute a central facies of the sills (e.g. Lundqvist, 1968).

The dolerite intrusions, chiefly sills, of the north-central region of Sweden and adjacent areas of Finland thus consist of a dominant alkali-olivine dolerite (Åsby and related types) associated with a subordinate olivine dolerite poorer in TiO_2 , Na_2O , K_2O and P_2O_5 (Särna type); the two varieties do not occur in the same areas and the Särna type is confined to western Dalarna, all other occurrences being identical with, or similar to the Åsby type. Both types were sampled at many localities in Dalarna and Jämtland. In Dalarna the country rock can be late-Svecofennian acid volcanics or the overlying Dala Sandstone Formation, and in Jämtland country rocks are mainly Svecofennian gneisses and granites.

Gimån

Location figs. 4.2, 4.3; sample localities section C.8, fig. C.4; analytical data section D.8; isochron diagram fig. 4.10.

In Jämtland a complex of sill-like intrusions extends over large areas in the region of the river Gimån; most of this is composed of uniform medium-grained olivine dolerite with occasional pegmatitic patches. At some contacts however, melting of granitic country rock has occurred and rheomorphic dykes and veins have intruded the dolerite; the basic magma at such localities has often become contaminated and hybridised by granitic material. The samples analysed from the Gimån area were taken from such a locality. Feldspars and whole-rock from granitic rheomorph J-33 define a

four-point isochron of 1256 ± 19 m.y. (fig. 4.10), where $[\text{SUMS}/(n-2)]^{1/2} = 0.82$ and initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.753$; such a high initial ratio is consistent with the considerable history of this granitic material from ca. 1800 to 1250 m.y. Contaminated dolerite from the same locality, J-35 defines a mineral line (feldspars and clinopyroxene) of 1204 ± 46 m.y. (fig. 4.10), with initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.714 \pm 0.001$, consistent with this rock having been heavily contaminated by older granitic material. The value $[\text{SUMS}/(n-2)]^{1/2}$ is equal to 3.58, and if clinopyroxene is omitted from the regression then the age is unchanged at 1201 ± 14 m.y., but $[\text{SUMS}/(n-2)]^{1/2}$ is reduced to 1.09. This relationship indicates that clinopyroxene is slightly discordant with the other points on the line; the age however, agrees at 2σ with the mineral isochron of J-33, and is therefore considered to be reliable.

Sundsjö

Location figs. 4.2, 4.3; sample localities section C.9, fig. C.4; analytical data section D.9; isochron diagram fig. 4.11.

To the West and Northwest of the Gimån area various petrographically similar dolerite bodies of more restricted extent occur; one of these is a dyke ca. 250 m in width at Sundsjö. The dyke is cut and slightly hybridised by a 50 cm rheomorphic dyke of granite, and two samples of this were separated into feldspar fractions for mineral analysis. J-55 defines a four-point line of 1181 ± 15 m.y., and J-56 a three-point line of 1192 ± 19 m.y. (fig. 4.11): in both cases $[\text{SUMS}/(n-2)]^{1/2}$ is less than unity, 0.35 and 0.91 respectively. This can be explained by fortuitously good alignments of small numbers of data points; therefore the true age errors are likely to be larger than indicated. The rheomorphic samples have high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (ca. 0.720), consistent with their pre-1200

m.y. isotopic history since the Svecofennian episode at 1900-1700 m.y. The whole-rock--biotite pair from olivine dolerite J-58 of the Sundsjö dyke gives 1239 ± 18 m.y., in general agreement with the ages from the rheomorph and with the results from the Gimån area. The lightest feldspar from this sample gives an apparent age of 1120 m.y.; this analysis could not be repeated due to lack of material so that its significance, if any, is unclear. In view of the agreement between the other age results, an 1120 m.y. figure must be rejected as spurious, possibly due to analytical error.

Älvho

Location figs. 4.2, 4.3; sample localities section C.10, fig. C.5; analytical data section D.10; isochron diagram fig. 4.12.

Results from the Åsby dolerite-monzonite sill at Älvho, northern Dalarna are shown in fig. 4.12. The sample DN-317 gives a mineral line (feldspars, apatite and clinopyroxene) of 1318 ± 63 m.y., with $[\text{SUMS}/(n-2)]^{1/2} = 4.76$; this value indicates a 'geological component' in the scatter of points, involving slight discordance of apatite and clinopyroxene with the feldspars. A whole-rock--biotite pair from olivine dolerite DN-325 of the same sill gives 1241 ± 18 m.y., and this agrees at 2σ with the mineral line age of DN-317.

Emådalen

Location figs. 4.2, 4.3; sample localities section C.11, fig. C.5; analytical data section D.11; isochron diagram fig. 4.13.

To the South of Älvho, the comparable sill of Emådalen also consists of dolerite and monzonite: at Emådalen itself, monzonite is a homogenous medium-grained rock developed over at least 100 m measured across the

trend of the sill, though whether it has an internal or a marginal position is not clear. Whole-rock and mineral samples from eight of these monzonites, a total of 17 points are concordant with a single isochron of 1249 ± 36 m.y. (fig. 4.13), where $[SUMS/(n-2)]^{1/2} = 1.32$.

Älvdalsåsen

Location figs. 4.2, 4.3; sample localities section C.12, fig. C.5; analytical data section D.12.

Åsby type olivine dolerite DN-51 from Älvdalsåsen, northern Dalarna, yielded a whole-rock--biotite pair of 1258 ± 18 m.y.

Mossiberg

Location figs. 4.2, 4.3; sample localities section C.13, fig. C.5; analytical data section D.13.

Åsby type olivine dolerite from Mossiberg, northern Dalarna, yielded a whole-rock--biotite pair of 1243 ± 18 m.y.

Idre

Location figs. 4.2, 4.3; sample localities section C.14, fig. C.6; analytical data section D.14.

Särna type olivine dolerite from Idre, northern Dalarna, was dated by whole-rock--biotite pairs at 1259 ± 18 m.y. (sample DN-17) and 1254 ± 18 m.y. (sample DN-21, from the same quarry).

Bornåssjön

Location figs. 4.2, 4.3; sample locality section C.15, fig. C.6; analytical data section D.15.

Särna type olivine dolerite 71-15 from Bornåssjön, northern Dalarna

defines a whole-rock--biotite age of 1274 ± 18 m.y.

Summary

The results from the alkali-olivine dolerite sills of northern central Sweden are summarised in table 4.2 and illustrated in fig. 4.3.

Table 4.2 Summary of age data for the alkali-olivine dolerite sills

Intrusion	Sample	Method	Number of points	Age m.y.	$[\text{SUMS}/(n-2)]^{1/2}$
Gimån	J- 33	Rheomorph minerals	4	1256 ± 19	0.82
	J- 35	Dolerite minerals	8	1204 ± 46	3.58
Sundsjö	J- 55	Rheomorph minerals	4	1181 ± 15	0.35
	J- 56	Rheomorph minerals	3	1192 ± 19	0.91
	J- 58	WR-biotite	2	1239 ± 18	-
Älvho	DN-317	Dolerite minerals	9	1318 ± 63	4.76
	DN-325	WR-biotite	2	1241 ± 18	-
Emådalen		WRs + minerals	17	1249 ± 36	1.32
Älvdalsåsen	DN- 51	WR-biotite	2	1258 ± 18	-
Mossiberg	DN- 74	WR-biotite	2	1243 ± 18	-
Idre	DN- 17	WR-biotite	2	1259 ± 18	-
	DN- 21	WR-biotite	2	1254 ± 18	-
Bornåssjön	71-15	WR-biotite	2	1274 ± 18	-
Mean				1244 ± 63	(2 σ)

If all ages, by whatever method are given equal weight, then the mean age is 1244 ± 68 m.y. where the error is 2σ of the 13 results. Welin and Lundqvist (1975) obtained a K-Ar 'isochron' of 1245 m.y. (two biotites independently gave this age) for dolerite sills in the Nordingrån region. Kouvo (1976) reported a U-Pb zircon discordia upper intercept at 1270 m.y. for olivine dolerite from Märket in the islands between Sweden and Finland (Suominen, in preparation). These ages are consistent with the Rb-Sr results reported here.

The close agreement of all WR-biotite pairs with a ca. 1250 m.y. figure (see table 4.2) suggests that in dolerite intrusions unaffected by later tectonic/thermal events this method of dating yields the most consistent results. In view of the agreement of ages, the isotopic data suggest either a single or a series of closely spaced magmatic events to explain the dolerite intrusions, though it must be emphasised that age differences of less than 18 m.y. would be undetectable even by biotite determinations. The evidence therefore indicates a major alkali-olivine dolerite igneous province in northern central Sweden and adjacent areas of Finland during the interval ca. 1250 ± 25 m.y. before present, this being the distribution of the whole-rock--biotite pairs, believed to provide the most accurate and consistent estimates of the emplacement ages of the sills.

4.6 Dykes parallel to the margin of the Sveconorwegian orogenic belt

The dykes (in one area sills) of this group are shown in fig. 4.4. They are grouped together primarily on the basis of the age work, but their orientation is at once a strong unifying feature. In the literature they have been treated region by region, and indeed petrographic characteristics do vary from one part of the 'swarm' to another. Since

many parts of central and southern Sweden are not covered by detailed mapping, the distribution of these dykes may actually be more continuous than fig. 4.4 suggests.

The most southerly of the groups of dolerite intrusions studied occurs in the province of Blekinge (fig. 4.4), trending NNE and mapped over a width of ca. 90 km. The dykes cut the Karlshamn granite group and consist of olivine, orthopyroxene, clinopyroxene, plagioclase, apatite, opaques and sometimes biotite and contain rare plagioclase megacrysts and sandstone xenoliths. In the Tärnö dyke, xenolithic sandstone can account for a large percentage of the volume.

In the Almesåkra region (fig. 4.4) the metasediments and granites are penetrated by long NNE-trending dykes, with a NNW trend developed to a lesser extent. The Almesåkra Formation sediments which overlie the Svecofennian rocks and consist of conglomerate, sandstone, shale and rare limestone, are intruded by large sills of dolerite sometimes over 50 m thick. The typical mineralogy of these rocks comprises orthopyroxene, clinopyroxene, plagioclase, opaques, apatite and interstitial granophyre.

The portion of the Schistosity Zone to the South of the Almesåkra region and to the West and North of the Blekinge region contains many linear dolerite bodies, known as the 'hyperites' after their orthopyroxene content (see fig. 4.4). These are noticeably parallel to the dykes of the two regions described and some of them cut granites classified by Magnusson (1958) as part of the Karlshamn group which would suggest a post-1450 m.y. age (see section 4.2). The 'hyperites' thus appear to be related to the Blekinge and Almesåkra dolerites on grounds of trend and petrography, but they are not covered by this study.

North of the Almesåkra region NNE dykes no longer appear and there is a gap in the distribution of dykes parallel to the Sveconorwegian margin,

the next known occurrence being at Årby, West of Stockholm (fig. 4.4) where a NW-trending dyke consisting of serpentinised olivine, clinopyroxene, plagioclase, opaques and apatite appears amongst the older East-West dyke system of the region. It is not clear how many of the dykes of this area which trend NW should be grouped with the Årby dyke, which is identified by its Rb-Sr age only. Lundegårdh et al. (1971) record a NNW trend for a plagioclase-phyric dolerite, a type common in the the older East-West system (Gorbatshev, 1961), and this would suggest that a NW or NNW trend is no definite evidence that a dyke in this region should be grouped with the swarms of this section.

The dolerite dykes of the Stockholm area (Ståhlös, 1969), which are indicated on fig. 4.4 ca. 100 km East of the Årby dyke, have a dominant WNW to NNW trend associated with a minor NNE trend. These dykes are not covered by this study.

In southern Dalarna a swarm of NNW trending olivine dolerite dykes occurs over a width of some 70 km (Hjelmqvist, 1966); these dykes have been referred to the group 'Åsby dolerite' and hence by implication correlated with the large sills of this type in northern Dalarna. The dykes consist of olivine (often altered), clinopyroxene, plagioclase, rare apatite, opaques, biotite and often interstitial granophyre. The NNW dykes are distinct structurally from the thick sills of northern Dalarna, which have an age of ca. 1250 m.y. (section 4.5).

In southwestern Dalarna and the neighbouring province of Värmland to the West and Southwest occur olivine-free orthopyroxene-bearing dolerites with a NNW trend (Hjelmqvist, 1966). Olivine dolerites reappear closer to the Sveconorwegian Front as straight dykes which can lie very close to, though not actually intersect the front zone (Schistosity Zone) itself (Gorbatshev, 1971). None of these dykes lying to the West of Dalarna are

studied here.

Certain NNW-trending dykes cut the Dala Sandstone Formation in northwestern Dalarna, some of which have widths greater than 100 m and are studied here (fig. 4.4). Of the two selected for study, one (Västra Nornäs) is a clean dolerite composed of serpentinised olivine, clinopyroxene, plagioclase, opaques, apatite, biotite and interstitial granophyre; the other (Bunkris) has apparently a parallel trend but is a rather strongly altered dolerite with considerable sericite, amphibole and chlorite. These dykes are among the most northerly which trend parallel to the Sveconorwegian Front, any further occurrences being covered by Caledonian thrust-sheets.

Worthy of mention also however, are the 'hyperites' which lie wholly within the Sveconorwegian block immediately West of the boundary against Svecofennian from the region of Lake Vänern northwards; they are not shown in fig. 4.4. These usually carry orthopyroxene (which may however in many cases be metamorphic in origin according to Gorbatshev, 1971), and could originally have been either dyke- or sill-like bodies. Their relationships both to the 'hyperites' of southern Sweden and to the dolerites intruding the Svecofennian are not established.

Important fracture directions in the Svecofennian crust are often oriented parallel to dolerite dykes. In the region to the West of Stockholm faults trend NW (Ståhlös, 1969; Gorbatshev, 1961), and in southern Dalarna an important fracture direction is NNW, with faults also of this orientation in northern Dalarna (Hjelmqvist, 1966).

The total extent of dykes parallel to the margin of the Sveconorwegian belt is thus some 600 km measured in a North-South direction (with an apparent gap of 150 km), by a maximum of 130 km measured across the strike.

Karlshamn

Location figs. 4.4, 4.5; sample localities section C.16, fig. C.7; analytical data section D.16; isochron diagrams figs. 4.14, 4.15.

The ca. 250 m Karlshamn dyke is the most western of three studied in the Blekinge region of southern Sweden, and intrudes Svecofennian gneisses at the sampled localities. Mineral determinations from average centre dolerite BL-3 and pegmatitic dolerite BL-18 give isochrons of 956 ± 56 m.y. (fig. 4.14) and 890 ± 25 m.y. (fig. 4.15) respectively; these ages overlap at 2σ . In both cases apatite and clinopyroxene are concordant with the isochrons and a considerable degree of confidence is therefore to be placed in the results.

Bräkne-Hoby

Location figs. 4.4, 4.5; sample localities section C.17, fig. C.7; analytical data section D.17; isochron diagrams figs. 4.16, 4.17.

The dyke 2 km East of Bräkne-Hoby intrudes Karlshamn granite and the analysed samples BL-24 and BL-27 are average dolerites from the centre of the dyke. The range of Rb/Sr ratios of the minerals is low (0-0.1) and the errors on the isochrons consequently rather high. The clinopyroxene analysis of BL-27 was repeated and the two analyses have an 'isochron relationship' to one another (due to inhomogeneity of granular sample); the mean is used for regression treatment. The ages obtained are 996 ± 58 m.y. (BL-24, fig. 4.16) and 899 ± 120 m.y. (BL-27, fig. 4.17). These are in general agreement with the Karlshamn results, though a close comparison is not possible.

Tärnö

Location figs. 4.4, 4.5; sample localities section C.18, fig. C.7; analytical data section D.18; isochron diagram fig. 4.18.

The 50 m Tärnö dyke is the central of the three studied in Blekinge, intrudes granite and consists of an intimate mixture of fine dolerite and loose arkosic sand, despite the fact that no such sediment is exposed in Blekinge at the present erosion level. The dyke also contains well-indurated xenoliths which must have been carried in the magma for longer than the loose arkosic sand: they are totally recrystallised to medium- or coarse-grained rocks consisting of quartz, feldspars, epidote, hornblende, sphene and opaques. Two of these (BL-3 and BL-ARX) were separated into mineral fractions and they generate isochrons with identical ages of 890 ± 21 and 899 ± 28 m.y. (fig. 4.18). If the xenoliths are presumed to have last recrystallised at the time of intrusion of the dolerite, then the mineral isochrons can be taken as defining an emplacement age for the dyke: this inferred age agrees with the results from Karlshamn and Bräkne-Hoby.

Forserum

Location figs. 4.4, 4.5; sample localities section C.19, fig. C.8; analytical data section D.19; isochron diagram fig. 4.20.

The ca. 50 m sill at Forserum in the Almesåkra region intrudes Svecofennian metasediments and the overlying Almesåkra Formation; the analysed samples are both from the interior of the sill. AL-6 defines a mineral isochron of 980 ± 18 m.y. (fig. 4.20), which includes clinopyroxene, apatite however lying some distance below the line; this analysis could not be repeated so that its significance is not clear. AL-12 defines a mineral isochron (including clinopyroxene) of 1071 ± 35

m.y. (fig. 4.20): according to the inferences of section 6.3, in a case where two mineral lines from dolerites of the same body do not agree at 2σ , then the younger age is generally to be preferred. In this case therefore, 980 ± 18 m.y. is taken as the 'best' age of intrusion.

Nilstorp

Location figs. 4.4, 4.5; sample localities section C.20, fig. C.3; analytical data section D.20; isochron diagram fig. 4.21.

The sill of comparable size at Nilstorp intrudes sediments at the sampled locality. The sample AL-22 is an interior medium-grained dolerite and the mineral isochron of 1005 ± 47 m.y. (fig. 4.21), which includes clinopyroxene and apatite, is in good agreement with the result from Forserum.

Årby

Location figs. 4.4, 4.5; sample locality section C.21, no detailed map; analytical data section D.21; isochron diagram fig. 4.19 (out of sequence).

The Årby dyke trends NW and is ca. 10 m wide: the country rocks are Svecofennian gneisses. The mineral isochron from SÖ-50 includes clinopyroxene and defines an age of 1016 ± 65 m.y. (fig. 4.19).

Falun

Location figs. 4.4, 4.5; sample localities section C.22, fig. C.9; analytical data section D.22; isochron diagram fig. 4.22.

The NNW olivine dolerite dykes of southern Dalarna were covered by a study of the most easterly of the group, a 50 m dyke running for over 50 km from Falun southwards. A sample from Bisberg (DS-9) defines a mineral



isochron which includes clinopyroxene of 987 ± 20 m.y. (fig. 4.22). A petrographically very similar sample DS-31 from Falun, 30 km NNW of Bispberg, gives a mineral isochron (also including clinopyroxene) of 1129 ± 77 m.y. (fig. 4.22). These ages do not agree at 2σ , and by the principle of section 6.3 the younger age should be preferred. Support for this comes from the fact that the whole-rock--biotite pair from DS-31 gives 934 ± 13 m.y., in agreement rather with the 987 ± 20 m.y. age from Bispberg than with the 1129 ± 77 m.y. result from the Falun sample DS-31. An age of ca. 950 m.y. is therefore indicated for the Falun dyke.

Västra Nornäs

Location figs. 4.4, 4.5; sample locality section C.23, fig. C.10; analytical data section D.23; isochron diagram fig. 4.23.

Dolerite D75-26 from a large dyke near Västra Nornäs, intruding the Dala Sandstone Formation defines a mineral isochron (including clinopyroxene) of 920 ± 39 m.y. (fig. 4.23), while the whole-rock--biotite pair from the same sample gives 953 ± 13 m.y. The agreement between mineral isochron and biotite ages is good evidence that the ca. 950 m.y. age is that of dyke intrusion.

Bunkris

Location figs. 4.4, 4.5; sample localities section C.24, fig. C.10; analytical data section D.24; isochron diagram fig. 4.8 (out of sequence).

A large dyke at Bunkris, lying parallel to that at Västra Nornäs, intrudes Svecofennian acid volcanics and the overlying Dala Sandstone Formation. Two altered dolerites give feldspar lines of 1549 ± 62 and 1579 ± 84 m.y. (fig. 4.8), but clinopyroxene in both cases lies considerably below the line and consequently little confidence is to be

placed in the result. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the feldspar lines are as high as 0.710 and suggest substantial addition of radiogenically-enriched Sr to the dyke: if some of this was added during feldspar crystallisation then anomalously old ages could have been produced. Clinopyroxene is the earliest phase to crystallise, and so the fact that it has a lower initial ratio than the feldspars supports the idea of contamination during crystallisation. Since the Bunkris dyke lies parallel to that at Västra Nornäs, it would under normal circumstances be assigned a similar age. The results however, despite the likelihood of contamination during crystallisation, may indicate an age as old as ca. 1550 m.y. for the Bunkris dyke, and this would mean that not all dykes mapped as lying parallel to the Sveconorwegian Front were emplaced at 1000-900 m.y. Additionally, the Dala Sandstone Formation would then have been deposited prior to ca. 1550 m.y., somewhat older than previously assumed (see section 5.6).

Summary

Age data from the dykes parallel to the Sveconorwegian Front, omitting only the data from Bunkris, are summarised in table 4.3 and illustrated in fig. 4.5.

Table 4.3 Summary of age data for dykes parallel to the margin of the Sveconorwegian orogenic belt

Intrusion	Sample	Method	Number of points	Age m.y.	$[SUMS/(n-2)]^{1/2}$	
Karlshamn	BL- 3	Dolerite minerals	8	956 $\pm$ 56	1.44	
	BL-18	Dolerite minerals	7	890 $\pm$ 25	1.93	
Tärnö	BL- 2	Xenolith minerals	5	899 $\pm$ 28	1.49	
	BL-ARX	Xenolith minerals	4	890 $\pm$ 21	1.49	
Bräkne-Hoby	BL-24	Dolerite minerals	6	996 $\pm$ 58	1.36	
	BL-27	Dolerite minerals	6	899 $\pm$ 120	2.10	
Forserum	AL- 6	Dolerite minerals	7	980 $\pm$ 18	1.70	
	AL-12	Dolerite minerals	7	1071 $\pm$ 35	2.17	
Nilstorp	AL-22	Dolerite minerals	7	1005 $\pm$ 47	2.75	
Årby	SÖ-50	Dolerite minerals	7	1016 $\pm$ 65	2.33	
Falun	DS- 9	Dolerite minerals	6	987 $\pm$ 20	1.03	
	DS-31	Dolerite minerals	6	1129 $\pm$ 77	2.71	
		WR-biotite	2	934 $\pm$ 13	-	
V. Nornäs	D75-26	Dolerite minerals	7	920 $\pm$ 39	2.50	
		WR-biotite	2	953 $\pm$ 13	-	

From the values of $[SUMS/(n-2)]^{1/2}$ it is clear that all mineral lines are to be regarded statistically as isochrons (i.e. $[SUMS/(n-2)]^{1/2}$ less than ca. 2.6). Neglecting the probably spurious age of ca. 1130 m.y. from Falun, it is clear that the results define a range of ages from 890 to ca. 1050 m.y. The limits are distinct at 2σ , and it appears that within a

general ca. 950 m.y. age for the activity there is variation in detail. If the ages were taken literally then the spread of activity would be from ca. 1050 to 890 m.y.: the effect of anomalous increase in dolerite mineral isochron age due to contamination during crystallisation cannot be discounted however, and seems to have resulted in a ca. 150 m.y. increase in the age given by the sample DS-31 from Falun. If this had occurred to a more limited extent in other samples, then part or even all of the range of age results could be explained by this means, and no variation in emplacement age would be necessary. There are, however several results between 1020 and 950 m.y., including one biotite, and these are in contrast to the ca. 900 m.y. ages from Blekinge. Therefore while the mineral contamination effect makes detailed interpretation of the age patterns difficult, it seems likely that the dykes parallel to the Sveconorwegian Front represent a spread of intrusive ages at least from ca. 1000 to 900 m.y.

4.7 Vaggeryd syenite

Location figs. 4.4, 4.5; sample localities section C.25, fig. C.11; analytical data section D.25; isochron diagram fig. 4.24.

The Vaggeryd syenite occurs within the 'Schistosity Zone' marginal to the Sveconorwegian belt, and intrudes gneisses of the Sveconorwegian block to the West and schistose Svecofennian granite to the East (Quensel, 1960). The body has dimensions of 12 km (E-W) by 60 km (N-S) and is thus elongated parallel to the shear-planes of the schistosity zone. The syenite, quartz-bearing in marginal areas and sometimes peralkaline in its interior, is itself strongly deformed along zones of shearing. Metamorphic muscovite, biotite, epidote and garnet have formed in the most deformed rocks; zones of strong deformation can however be separated by

lenses of almost undeformed syenite on a scale of hundreds of metres or kilometres.

The northern part of the body, together with sample localities, is shown in fig. C.11. Most of the collected rocks have rather low Rb/Sr ratios (less than 1.0), only the peralkaline syenites yielding higher isochron points; in view of the scanty number of analyses at high Rb/Sr, three of these were duplicated and their position confirmed.

The whole-rock age obtained, 1151 ± 67 m.y. (fig. 4.24) has a $[\text{SUMS}/(n-2)]^{1/2}$ value of 7.30, indicating that the scatter of points contains a significant 'geological component'. This is clearly to be found in redistribution of Rb and Sr during the metamorphism which has affected the syenite. The further possibility, that the 1151 ± 67 m.y. result is itself a metamorphic age (resulting from complete reequilibration of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios at 1151 m.y.) cannot be ruled out. However, if the samples V-5, V-9, V-14, V-15 and V-18 which are taken from undeformed and igneous-textured lenses are regressed, an age of 1130 ± 43 m.y. is obtained. Although this age depends strongly on the sample V-5, the correspondence of the apparent age of igneous-textured samples with that of all samples suggests that the latter does represent emplacement of the syenite and not some later event.

The error on the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.707 ± 0.002) is too large to allow any conclusions regarding the ultimate origin of the syenite magma.

Ages of whole-rock--biotite and whole-rock--muscovite pairs from schistose syenites are shown in table 4.4.

Table 4.4 Ages of metamorphic micas (in m.y.)
from the Vaggeryd syenite

Sample	WR-muscovite	WR-biotite
V- 2	948 $\pm$ 13	894 $\pm$ 13
V- 8	-	877 $\pm$ 13
V-10	934 $\pm$ 13	923 $\pm$ 13

The micas became closed systems to Rb and Sr migration at an approximate time of 915 m.y. (see fig. 4.24), and since these rocks were selected for their content of metamorphic minerals, then the results should be regarded as metamorphic ages. If the Vaggeryd syenite was deformed and metamorphosed at a relatively shallow level, then the mica ages could be close to the actual age of metamorphism, but it is more likely that they represent the time at which the rocks cooled through the relevant Rb and Sr blocking temperatures: thus the metamorphism would have taken place during some time prior to the age given by the micas. The fact that the muscovites (isotopic closure ca. 500°C; Jäger, 1973) give slightly older results than the biotites (isotopic closure ca. 300°C; Jäger, 1973; Dodson, 1973) lends support to this view.

4.8 Mineral age studies of granite country rocks

Sample localities section C.26, figs. C.7, C.8; analytical data section D.26.

Because the age results presented here for the dolerites are based on various types of mineral data, it is necessary to establish that the

ages obtained are those of intrusion and not some later thermal event. The almost completely undeformed nature of the dolerites might be considered evidence enough, but in the central and northern regions of study several K-Ar ages of ca. 1700 m.y. (Magnusson, 1960) indicate that the country rock has not been thermally affected since Svecofennian time. The uncertainty over the age of the last thermal event in southern Sweden however, prompted a mineral age study of country rocks from the Almesåkra and Blekinge regions (Växjö and Spinkamåla granites respectively).

Coarse white Växjö granite F-3 (see fig. C.8 for locality) from the Almesåkra region, lying 15 km to the North-east of Forserum, gives whole-rock--biotite ages of 1649 ± 24 and 1660 ± 24 m.y. (two analyses), and the whole-rock--orthoclase pair gives 1607 ± 97 m.y. These ages, which must be younger than the intrusion age of the granite (and thus consistent with Welin et al., 1966 and Åberg, 1972), show that this portion of crust has not been thermally affected since late-Svecofennian time.

Spinkamåla granite SPG-1 (see fig. C.7 for locality) from 13 km West of the Karlshamn dyke in Blekinge is of similar grain-size to the samples of dolerite used for dating. The whole-rock--biotite pair from this sample gives 1393 ± 23 and 1381 ± 23 m.y. (two analyses), and the whole-rock--microcline pair gives 1194 ± 68 m.y. The Spinkamåla granite is considered to belong to the Karlshamn group of granites, emplaced at ca. 1450 m.y., and a cooling age of 1390 m.y. might thus be considered reasonable for this body. On the other hand, the age of the microcline does suggest some open-system behaviour as late as 1190 m.y.; this can be related to Sveconorwegian events at this time in western Sweden. The granite sample is in any case much nearer to the Schistosity Zone than

the dolerite dykes, and since it does not show mineral ages as young as ca. 900 m.y., then the mineral isochron dates for the Blekinge dykes can be taken as defining intrusive ages.

4.9 Särna alkaline complex

Location fig. 4.1; sample localities section C.27, fig. C.6; analytical data section D.27; isochron diagram fig. 4.25.

The cancrinite-nepheline syenite of Särna (särnaite) is exposed on the hills West of Särna, northern Dalarna. The body is of circular outline, 3.5 km in diameter and is intrusive into late-Svecofennian acid volcanics. The alkaline rocks are fine to coarse grained and consist mainly of Na-orthoclase, albite, cancrinite, nepheline and aegirine. All these phases are essentially unaltered and the cancrinite is considered to be a primary mineral (Magnusson, 1923). Most rocks show mineral lamination of feldspar and/or aegirine, and schlieric structure with considerable differences in mineralogical composition from one band to another is often observed: the schlieren are strongly irregular, their direction changing on a scale of metres.

Since all whole-rock samples have Rb/Sr less than 0.15, the mineral isochron technique used for the dolerites was also applied here. The isochron (fig. 4.25) includes cancrinite, aegirine, sphene and three feldspar separates and gives an age of 287 ± 14 m.y. with $[\text{SUMS}/(n-2)]^{1/2} = 0.92$. This age is consistent with a palaeomagnetic pole for the complex suggesting an upper Carboniferous to Permian age (Bylund and Patchett, in press).

The variation in initial $^{87}\text{Sr}/^{86}\text{Sr}$ of whole-rock samples from 0.7041 to 0.7070 (see fig. 4.25) has been explained as a result of contamination of a viscous, inhomogenous alkaline magma during

emplacement (Bylund and Patchett, in press).

These authors have also related the emplacement of the complex to the alkaline magmatism of the Oslo graben in early Permian times (276 ± 2 m.y. according to Heier and Compston, 1969), since the complex lies directly on the trend of outcrop of the Oslo plutonic rocks when it is projected 150 km inland.

4.10 Summary of age results from Sweden

A summary of the igneous events dated in Sweden is given in table 4.5.

Table 4.5 Generalised ages of igneous events dated in Sweden

Number of intrusions	Range of results m.y.	'Best' age m.y.	Event	Loc. fig.
1	290	290	Särna alkaline complex	4.1
8	1130-890	1020-890	Dykes parallel to the Svecconorwegian Front	4.5
1	1150	1150	Vaggeryd syenite	4.5
8	1320-1180	1250	Alkali-olivine dolerite sills	4.3
6	1400	1400	Tuna dykes	4.3
4	1590-1540	1560	East-West dykes	4.3

The column 'best' age gives the interpreted times of emplacement of the rocks. The East-West dykes give ages that are all consistent at 2σ with an age of ca. 1560 m.y. The alkali-olivine dolerite sills give a

wide range of age results from 1320 to 1180 m.y., but biotites from all intrusions define a close group within the limits 1250 ± 25 m.y., and consequently this is chosen as the best age of intrusion for the sills. The spread of age results from the dykes parallel to the Sveconorwegian Front however, is such that many results do not overlap at 2σ . There is furthermore, no pronounced clustering of the ages as for the olivine dolerite sills, and it is probable that the results reflect at least in part a real variation in intrusive age: a reasonable period for intrusion of the swarm would be 1020-890 m.y.

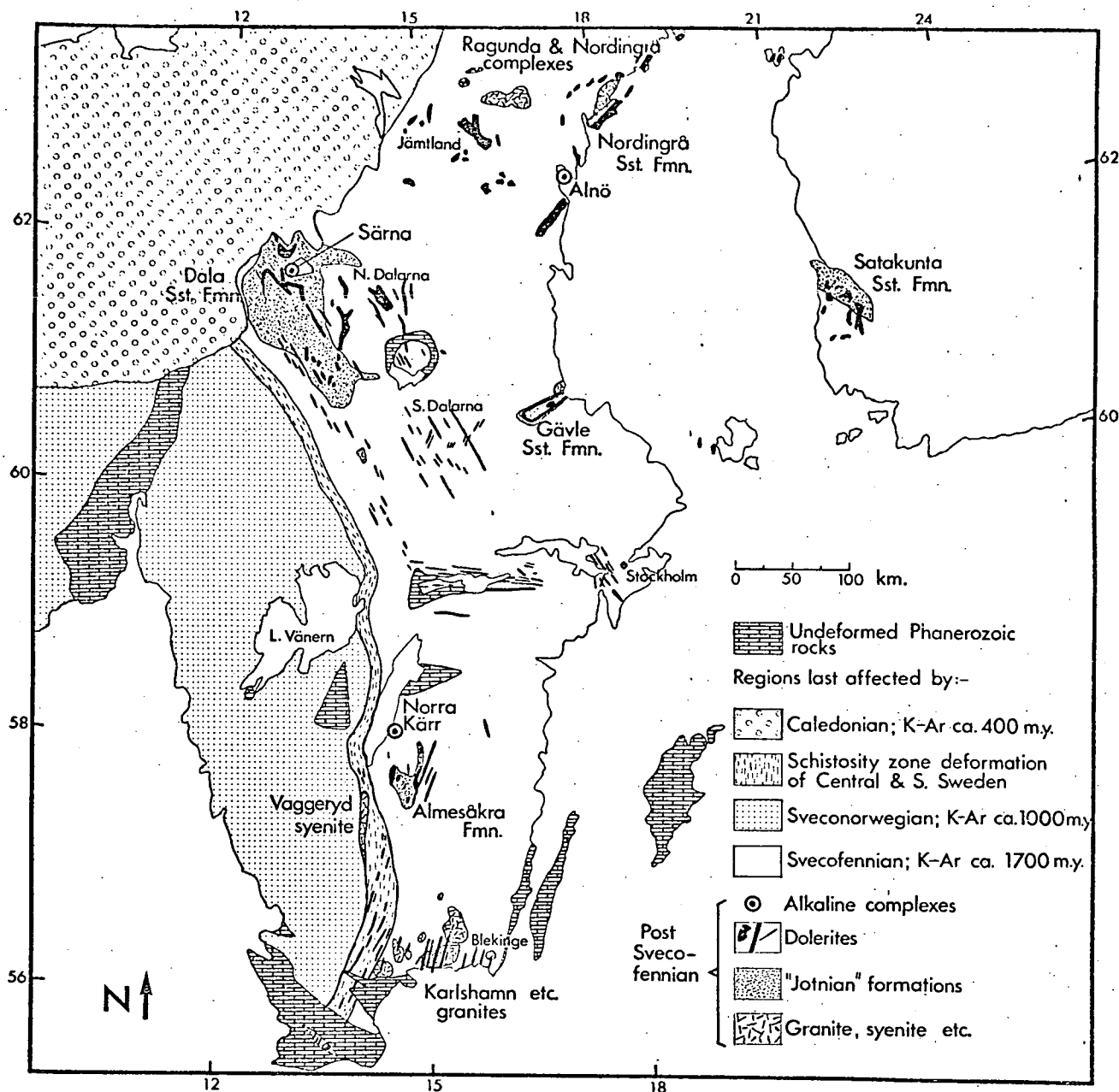


Fig. 4.1 Simplified geological map of central and southern Sweden showing post-tectonic granites and syenites, 'Jotnian' continental sediments, dolerite intrusions and alkaline complexes.

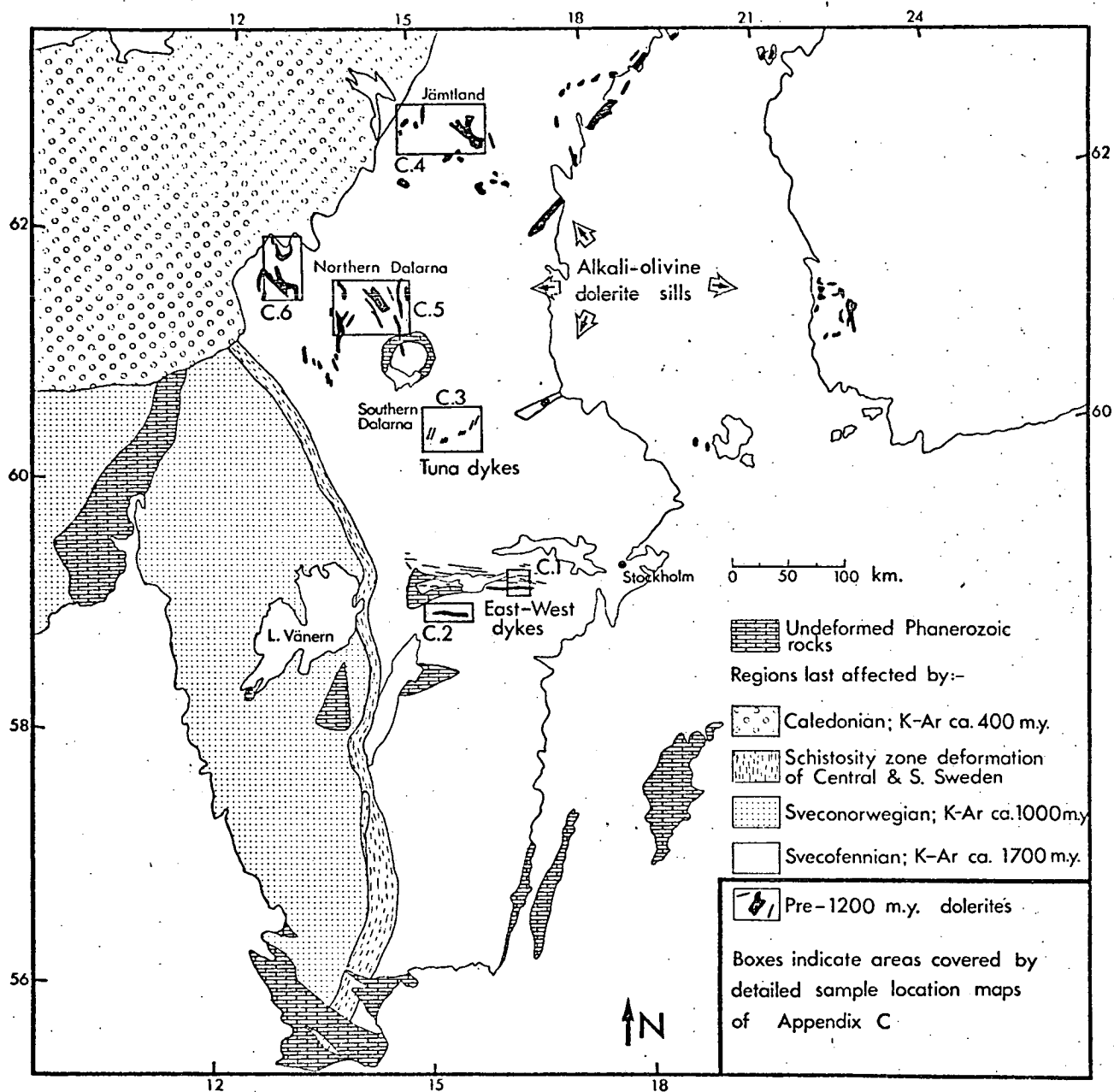


Fig. 4.2 Pre-1200 m.y. dolerites in central Sweden, showing areas covered by detailed locality maps of appendix C.

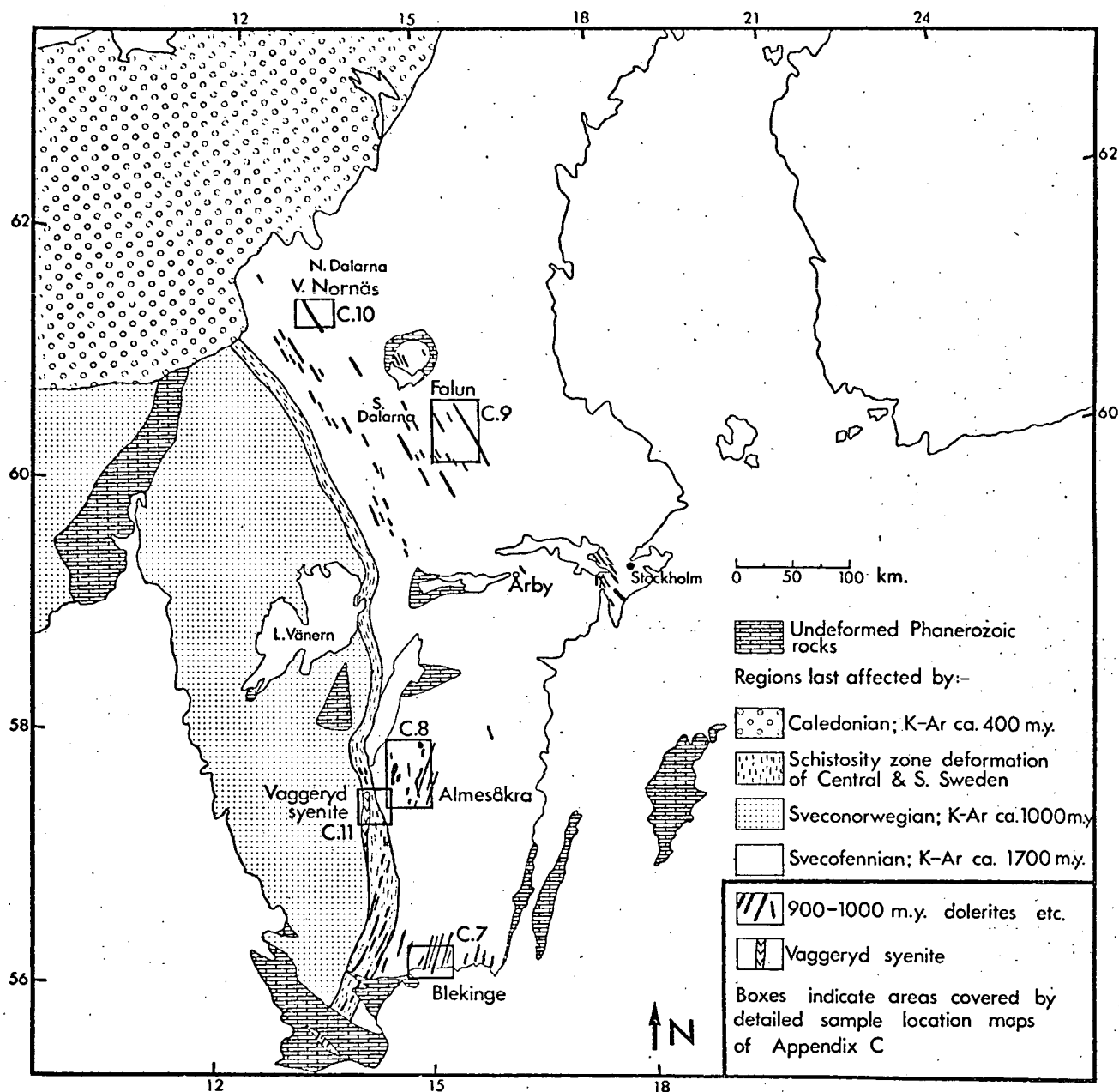


Fig. 4.4 Post-1200 m.y. dolerites of central and southern Sweden and the Vaggeryd syenite, showing areas covered by detailed locality maps of appendix C.

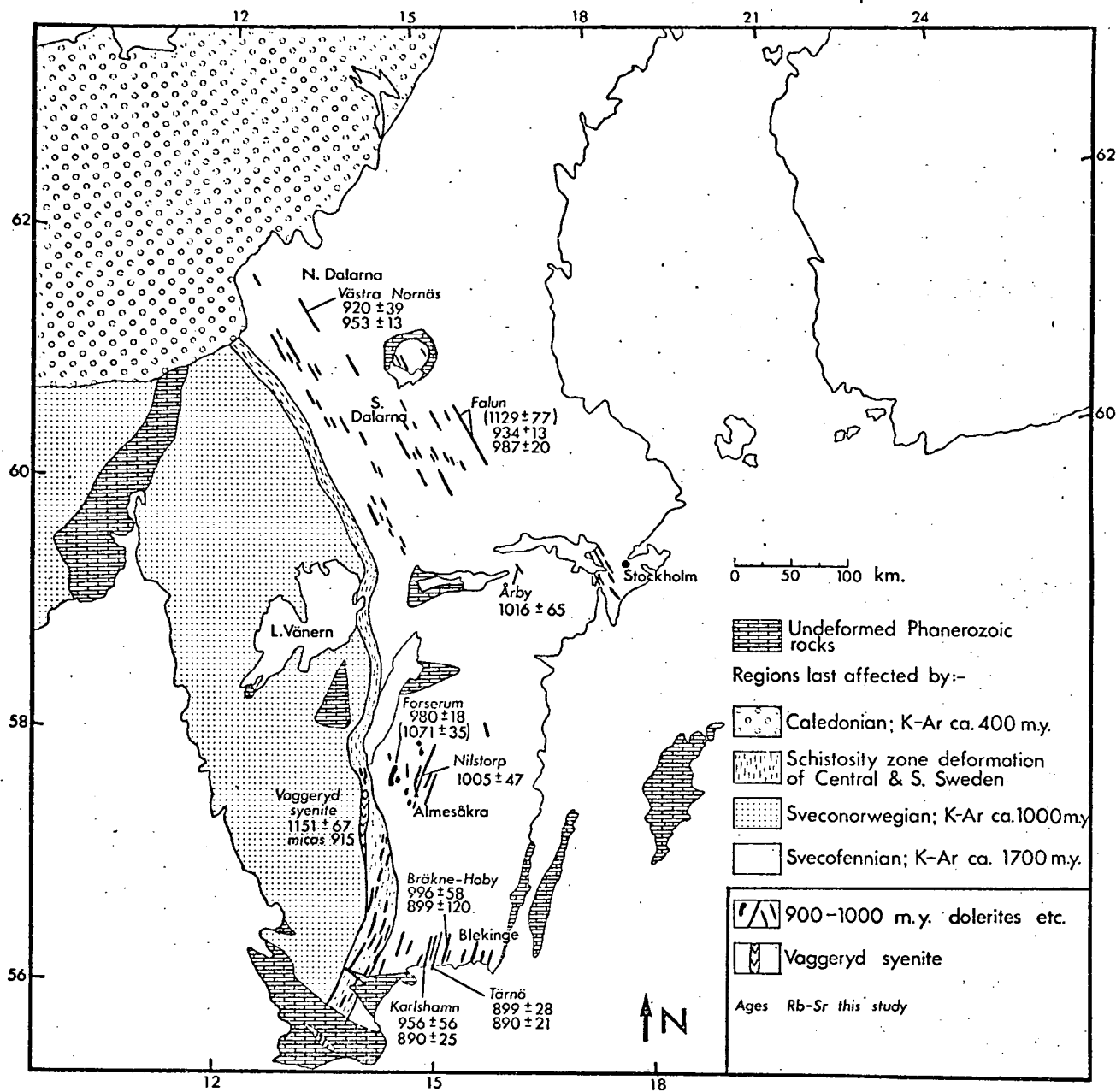


Fig. 4.5 Ages of post-1200 m.y. dolerites and the Vaggeryd syenite.

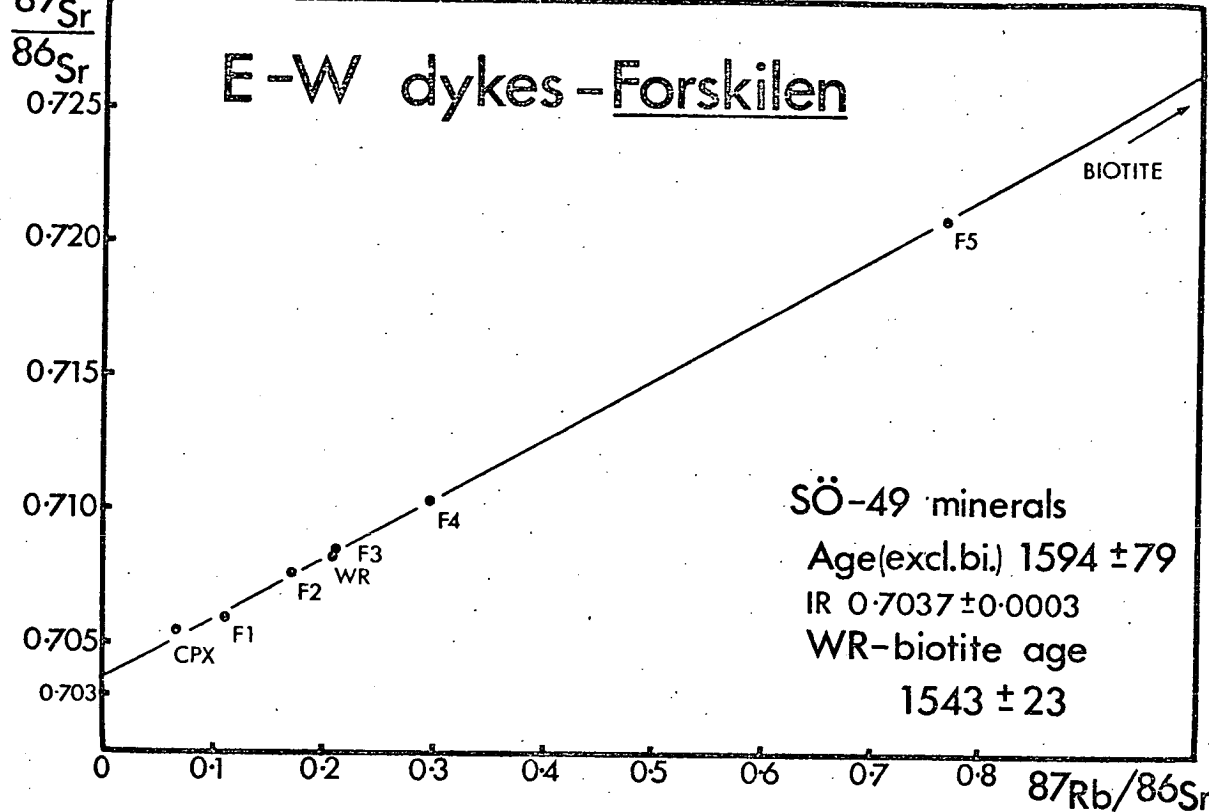


Fig. 4.6 Mineral line and whole-rock--biotite age from olivine dolerite of the Forskilen dyke, East-West system. For location see fig. 4.3.

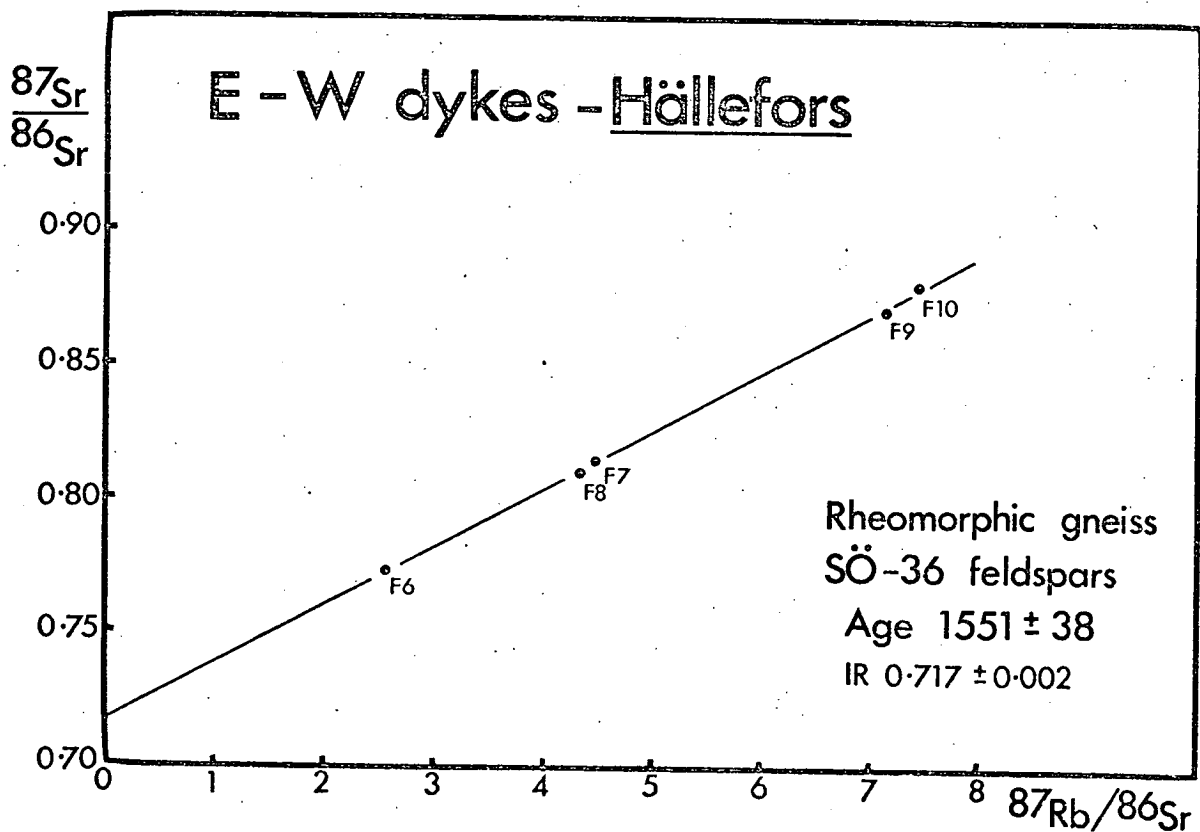


Fig. 4.7 Mineral (feldspar) line from a 6 cm rheomorphic granophyre vein at the contact of olivine dolerite with Svecofennian gneiss: Hällefors dyke, East-West system. For location see fig. 4.3.

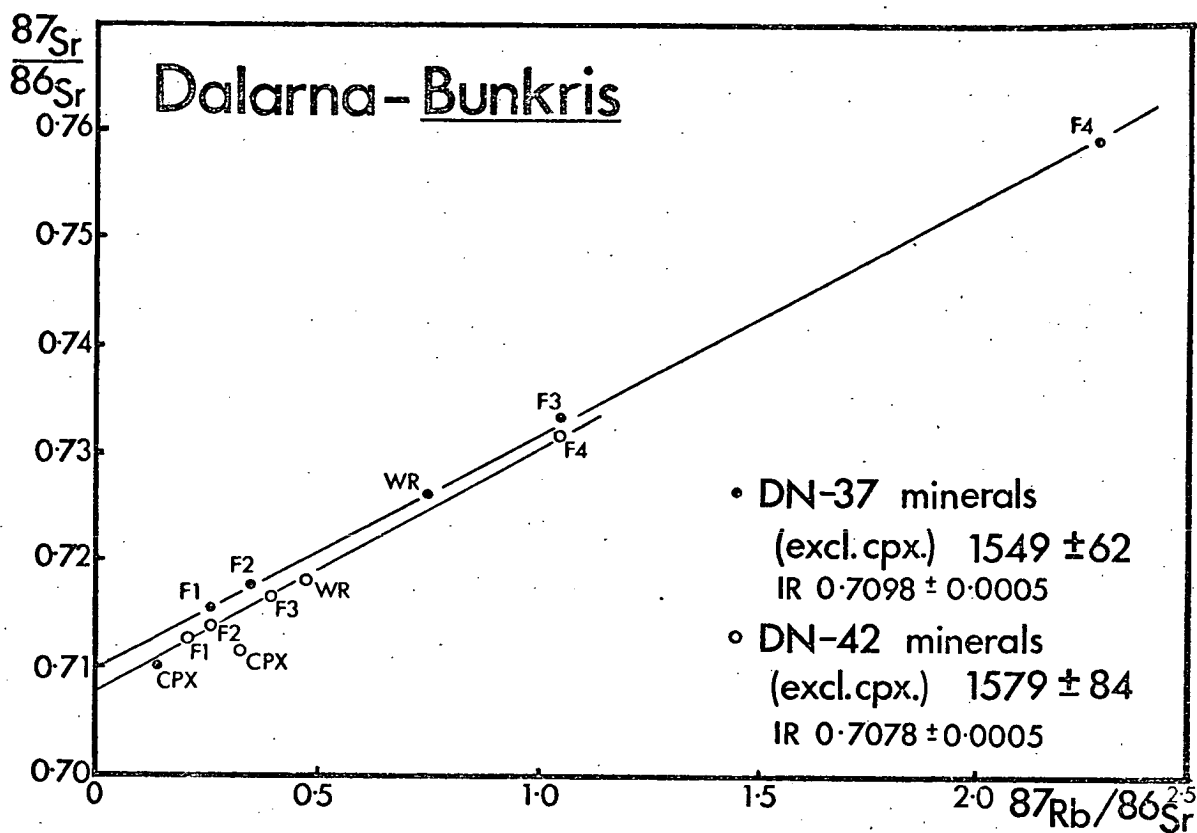


Fig. 4.8 Feldspar lines from two samples of the Bunkris dyke, northern Dalarna. Note the position of the clinopyroxenes lying below the feldspar lines. The dyke lies close to the Västra Nornäs dyke (see figs. 4.4, C.10), and trends parallel to the Sveconorwegian margin, but may be of ca. 1550 m.y. age as given by the feldspars: see text for discussion.

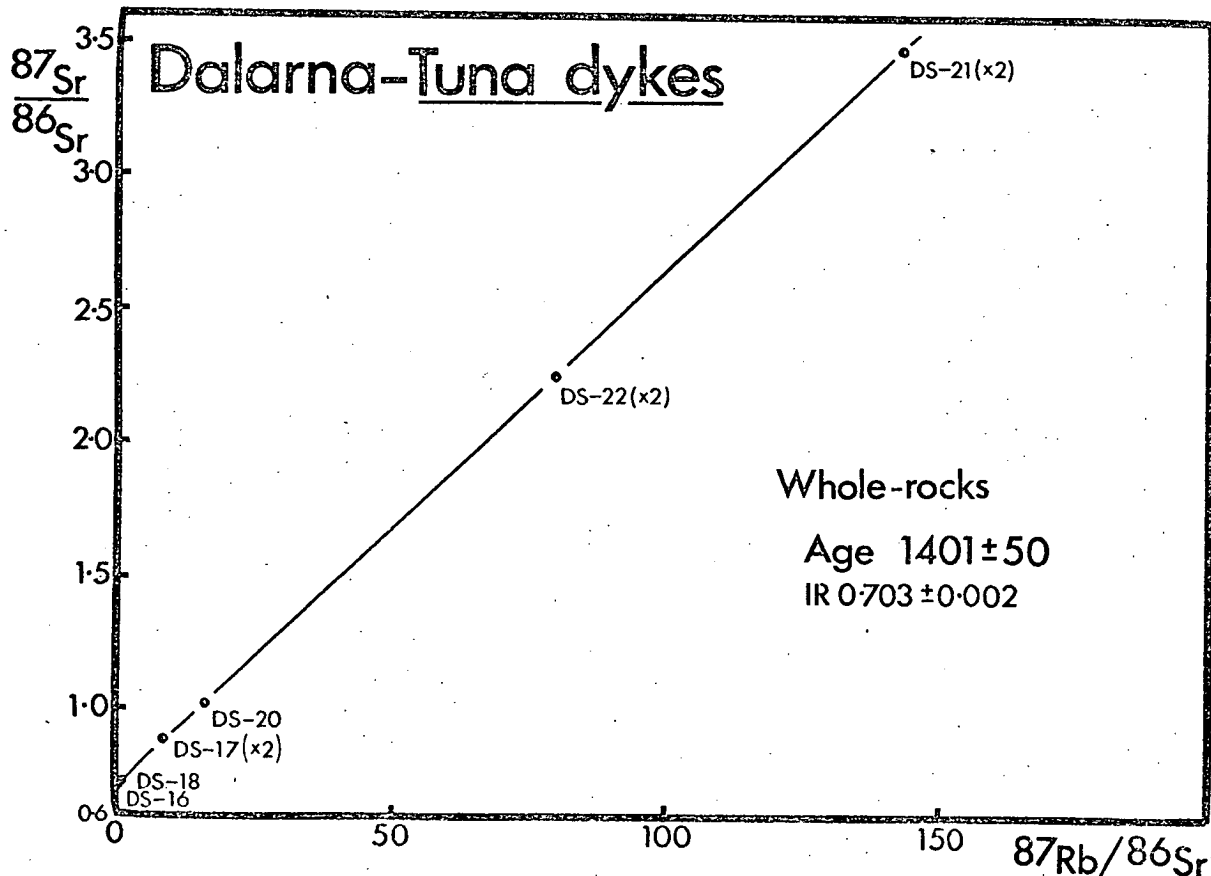


Fig. 4.9 Whole-rock line for dolerites, a trachyte and quartz porphyries from the Tuna dykes, southern Dalarna. All samples are from different dykes. For location see fig. 4.3.

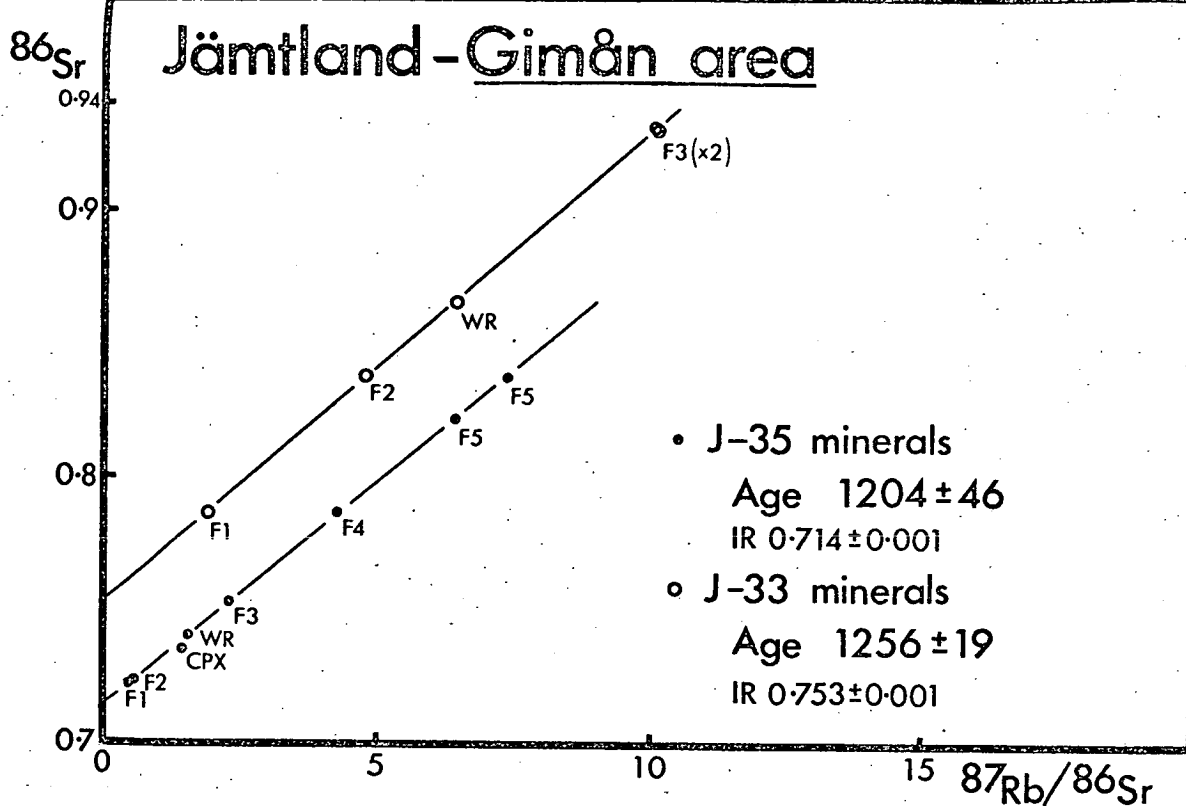


Fig. 4.10 Mineral lines for a granitic rheomorph and a related hybridised dolerite from a large alkali-olivine dolerite sill in the Gimån area, Jämtland. For location see fig. 4.3.

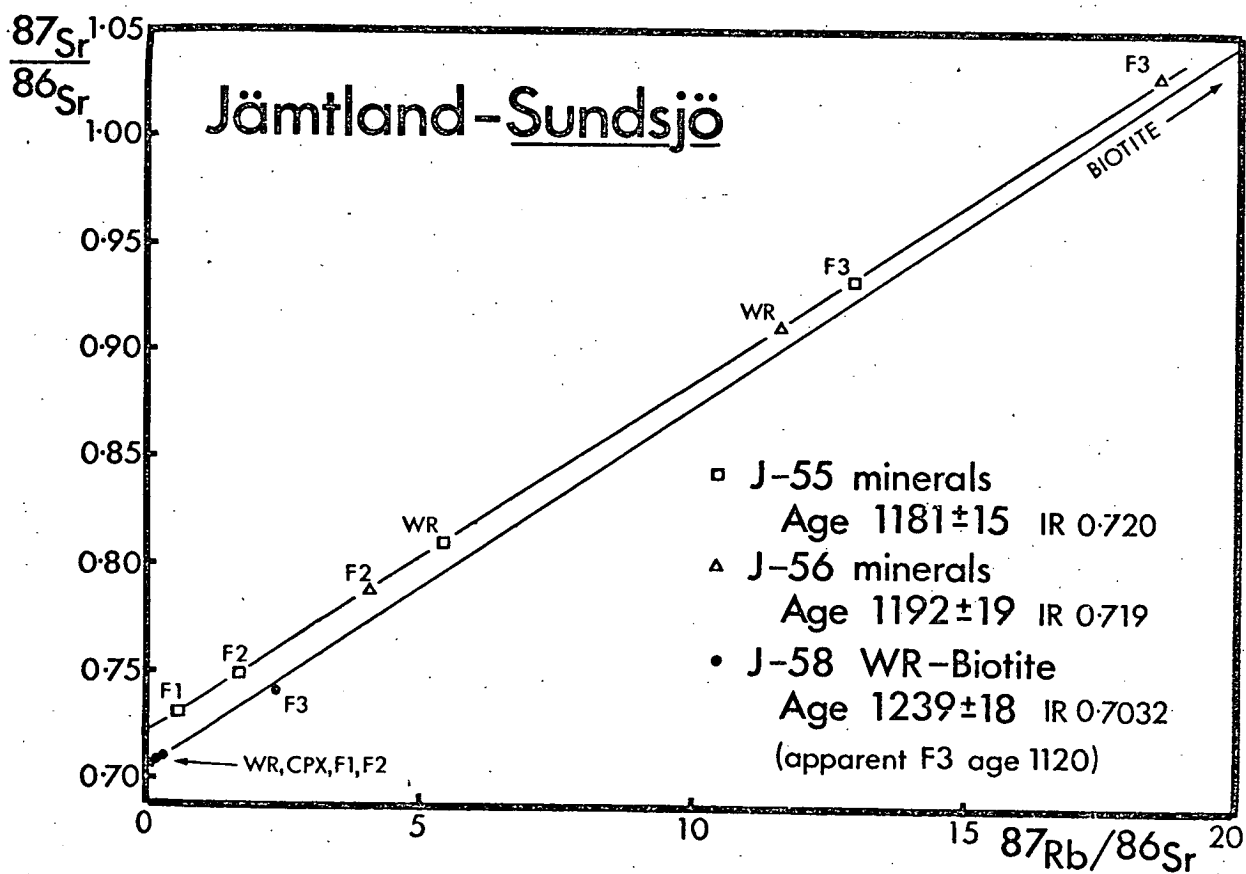


Fig. 4.11 Superimposed mineral lines for two granitic rheomorphs together with a whole-rock--biotite age from a large alkali-olivine dolerite dyke at Sundsjö, Jämtland. For location see fig. 4.3.

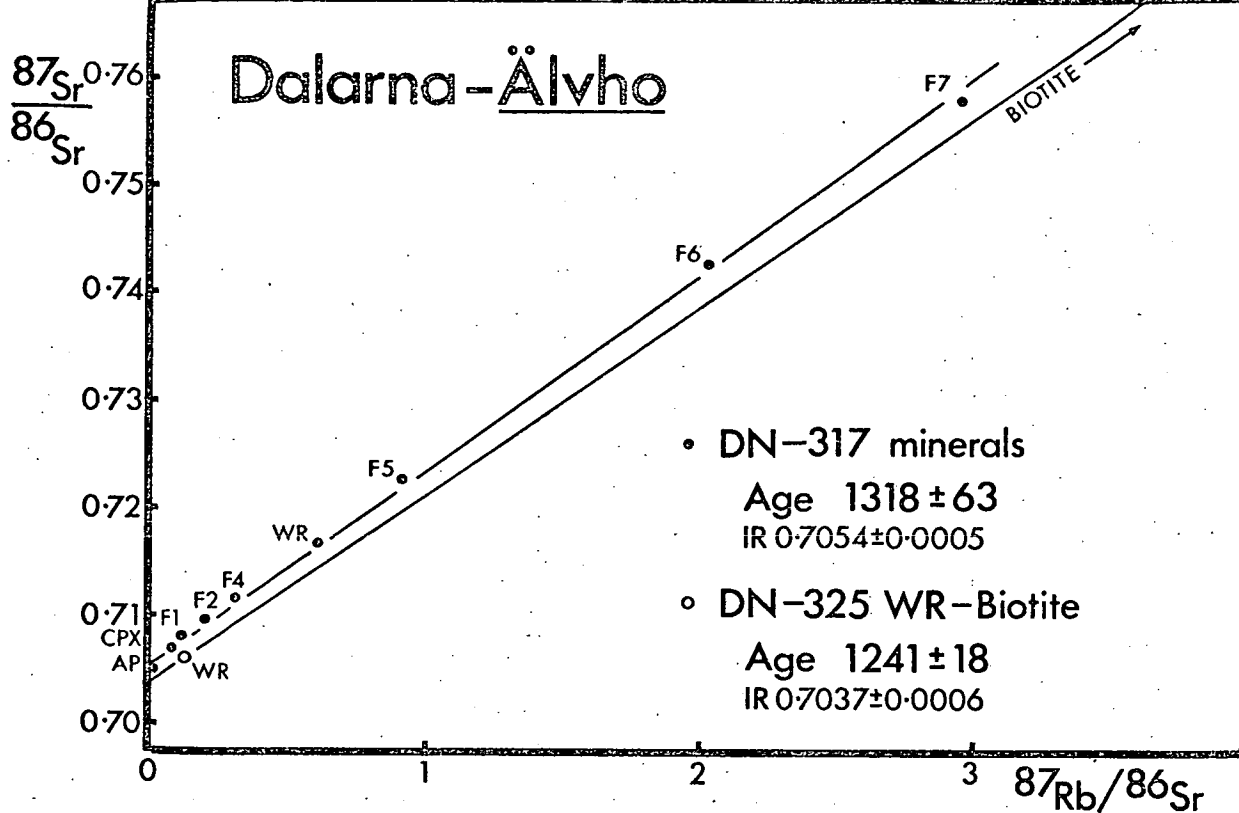


Fig. 4.12 Mineral line and whole-rock--biotite age for dolerites of the Älvho alkali-olivine dolerite-monzonite sill, northern Dalarna. For location see fig. 4.3.

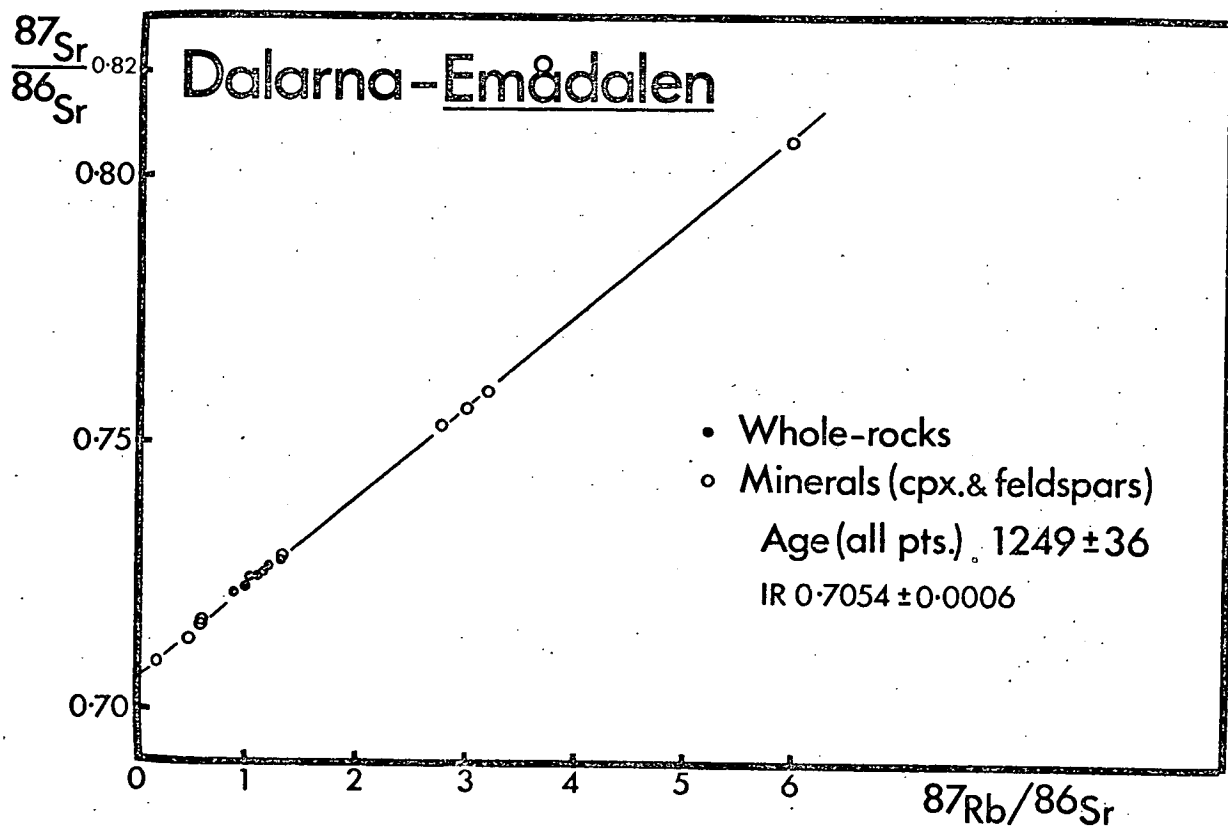


Fig. 4.13 Line for whole-rock monzonites and their minerals from the Emådalen alkali-olivine dolerite-monzonite sill, northern Dalarna. For location see fig. 4.3.

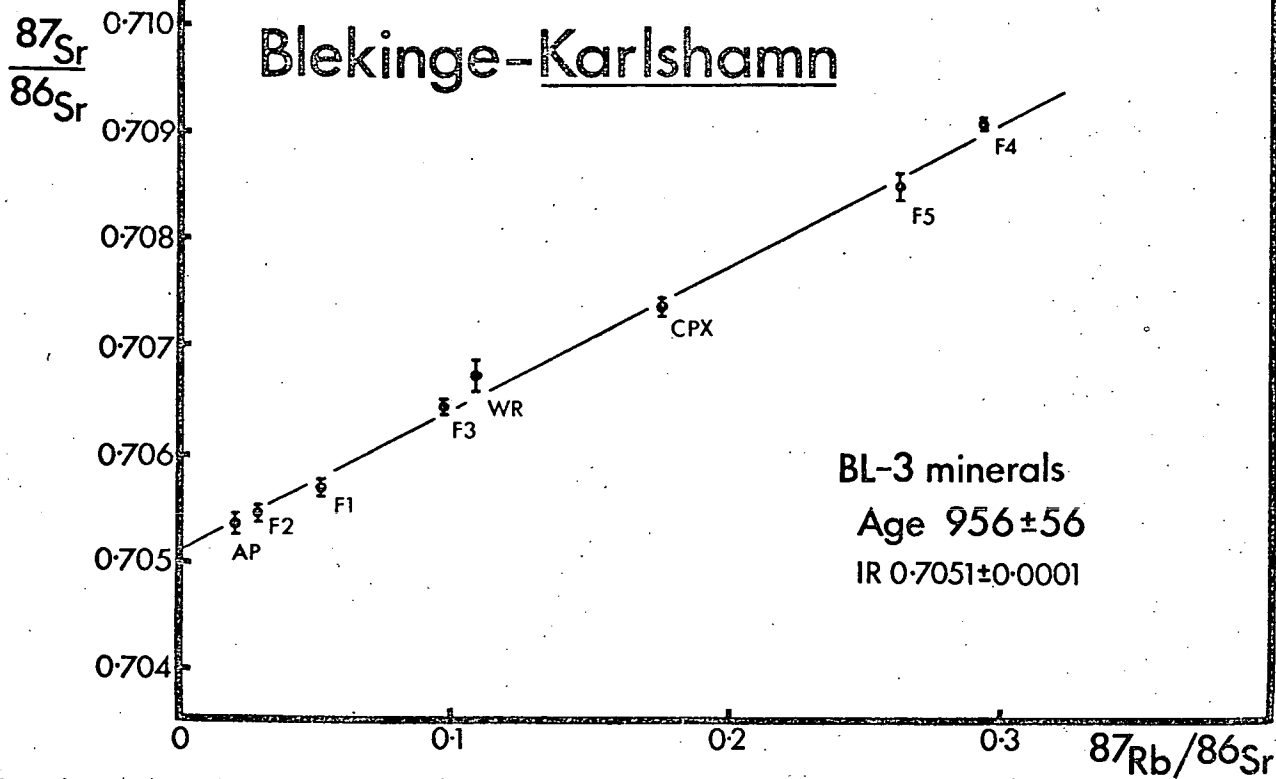


Fig. 4.14 Mineral line for olivine-hypersthene dolerite BL-3 from the Karlshamn dyke, Blekinge. For location see fig. 4.5.

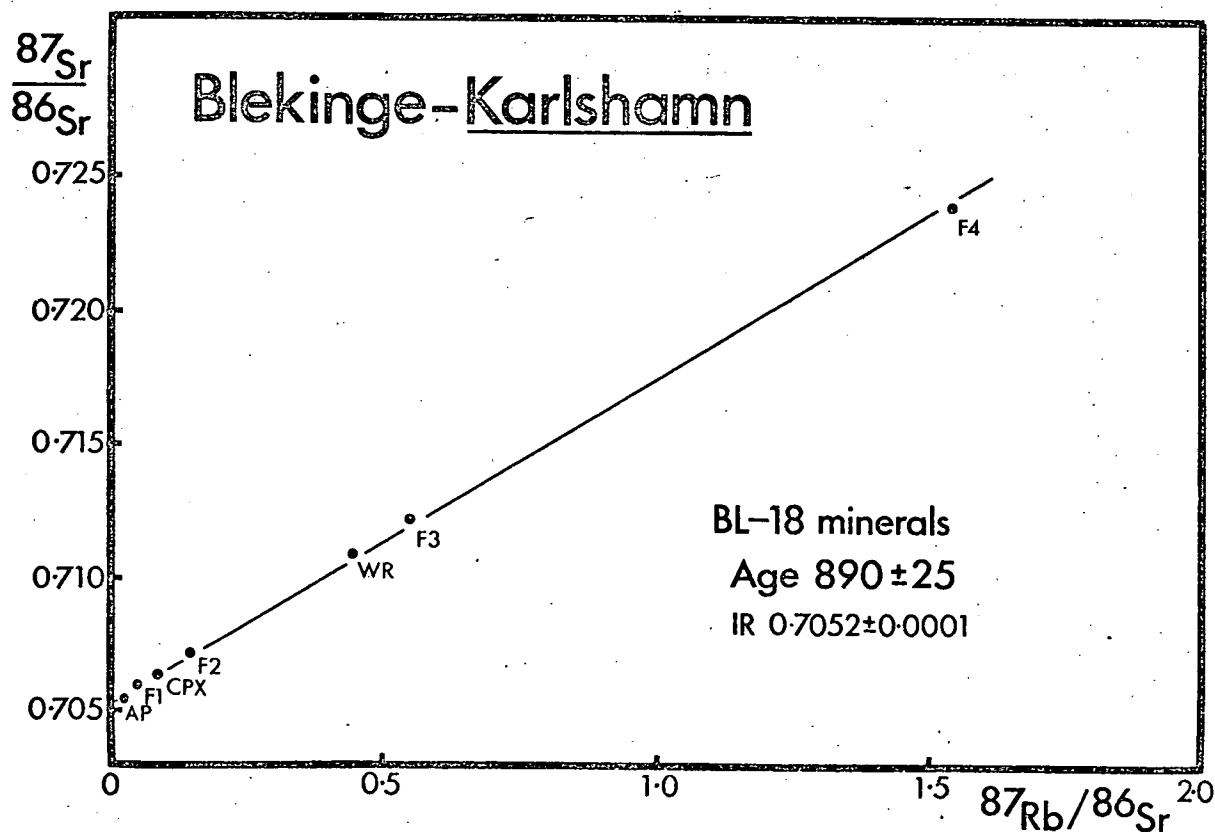


Fig. 4.15 Mineral line for pegmatitic olivine-hypersthene dolerite BL-18 from the Karlshamn dyke, Blekinge. For location see fig. 4.5.

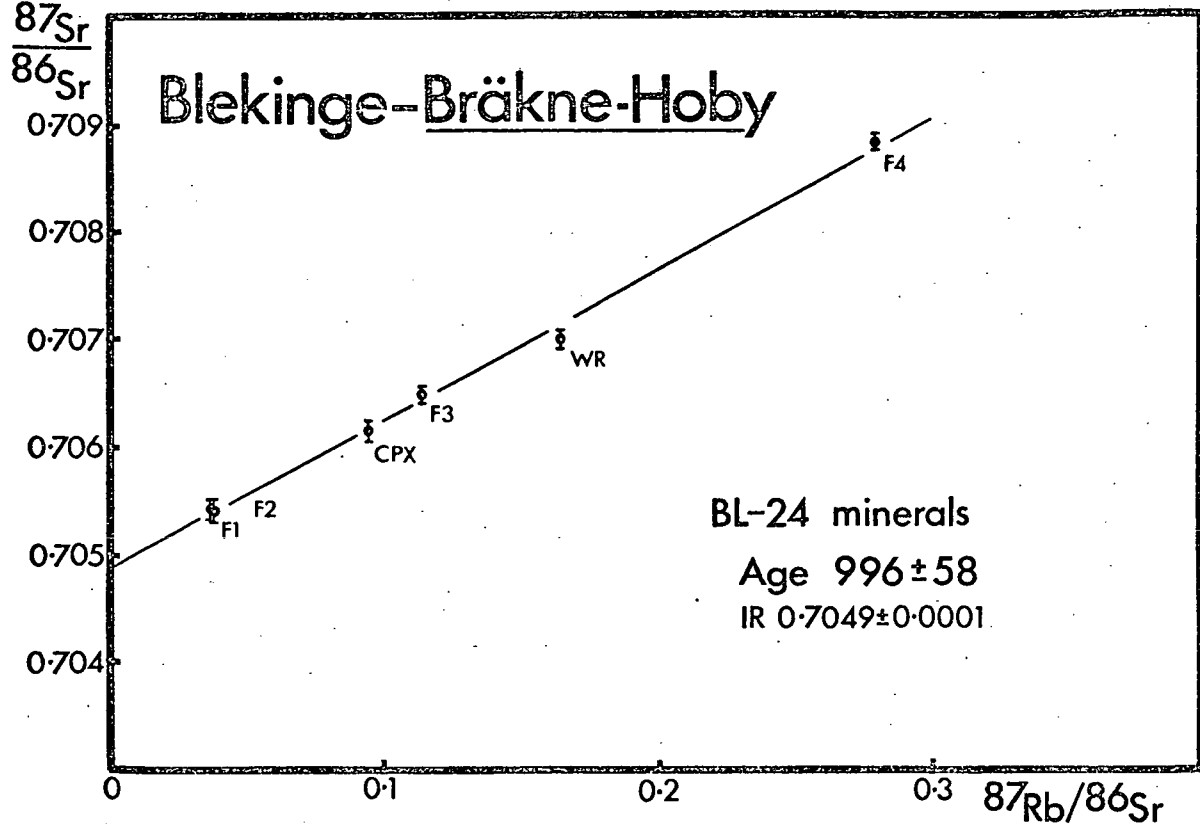


Fig. 4.16 Mineral line for olivine-hypersthene dolerite BL-24 from the Bräkne-Hoby dyke, Blekinge. For location see fig. 4.5.

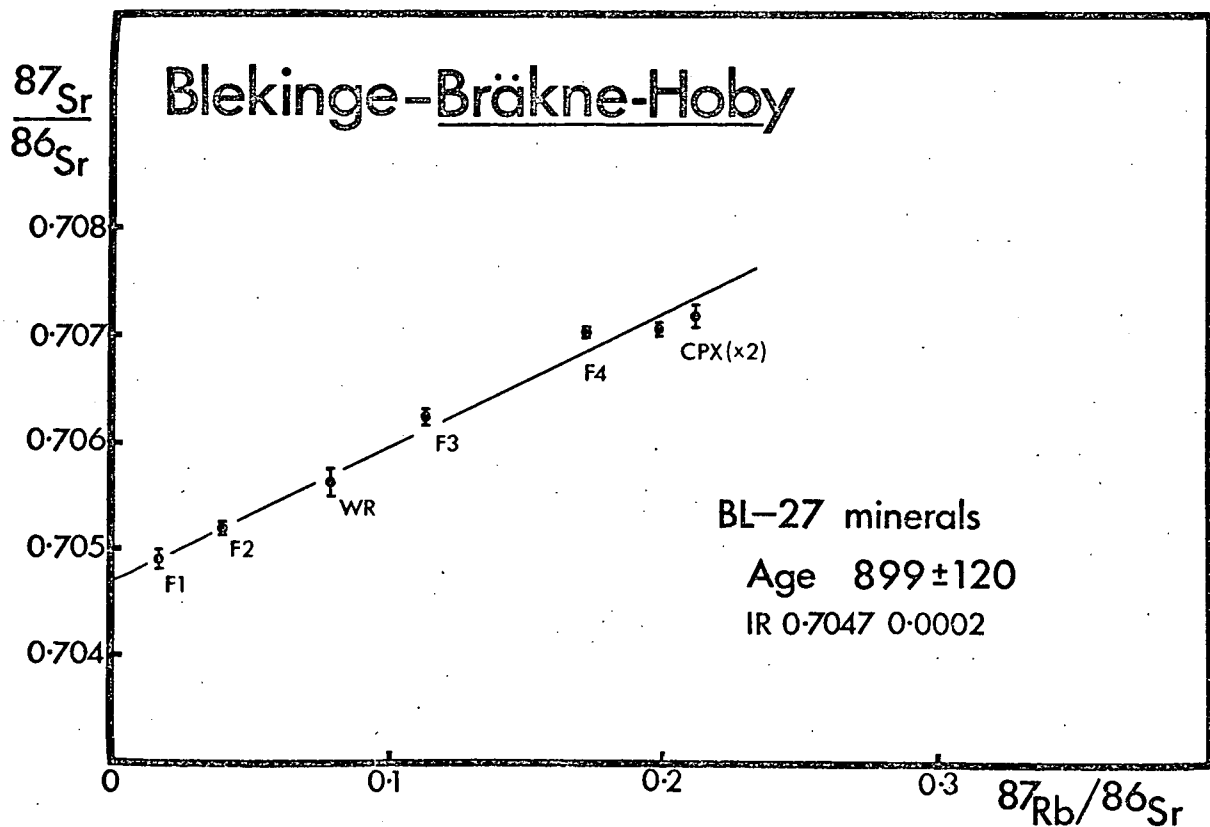


Fig. 4.17 Mineral line for olivine-hypersthene dolerite BL-27 from the Bräkne-Hoby dyke, Blekinge. For location see fig. 4.5.

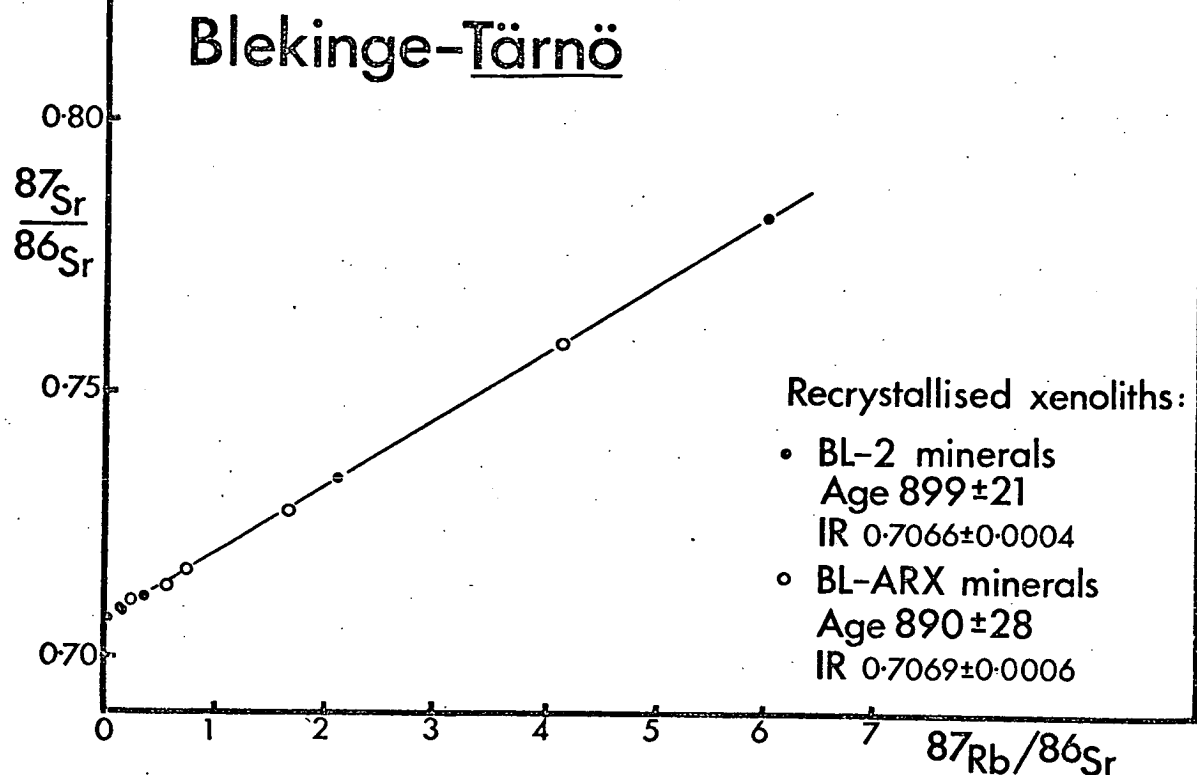


Fig. 4.18 Superimposed mineral lines for two recrystallised xenoliths from the Tärnö dyke, Blekinge. For location see fig. 4.5.

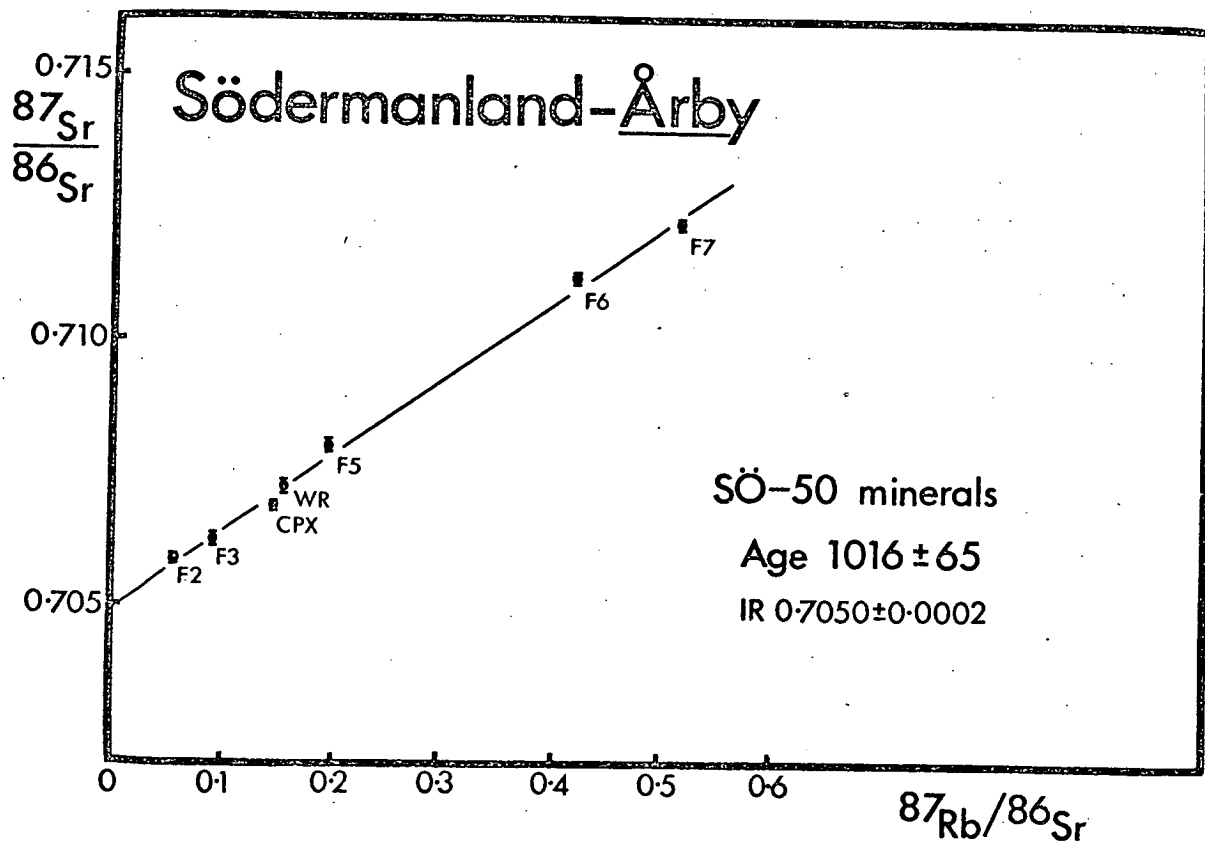


Fig. 4.19 Mineral line for olivine dolerite from the Årby dyke, West of Stockholm. For location see fig. 4.5.

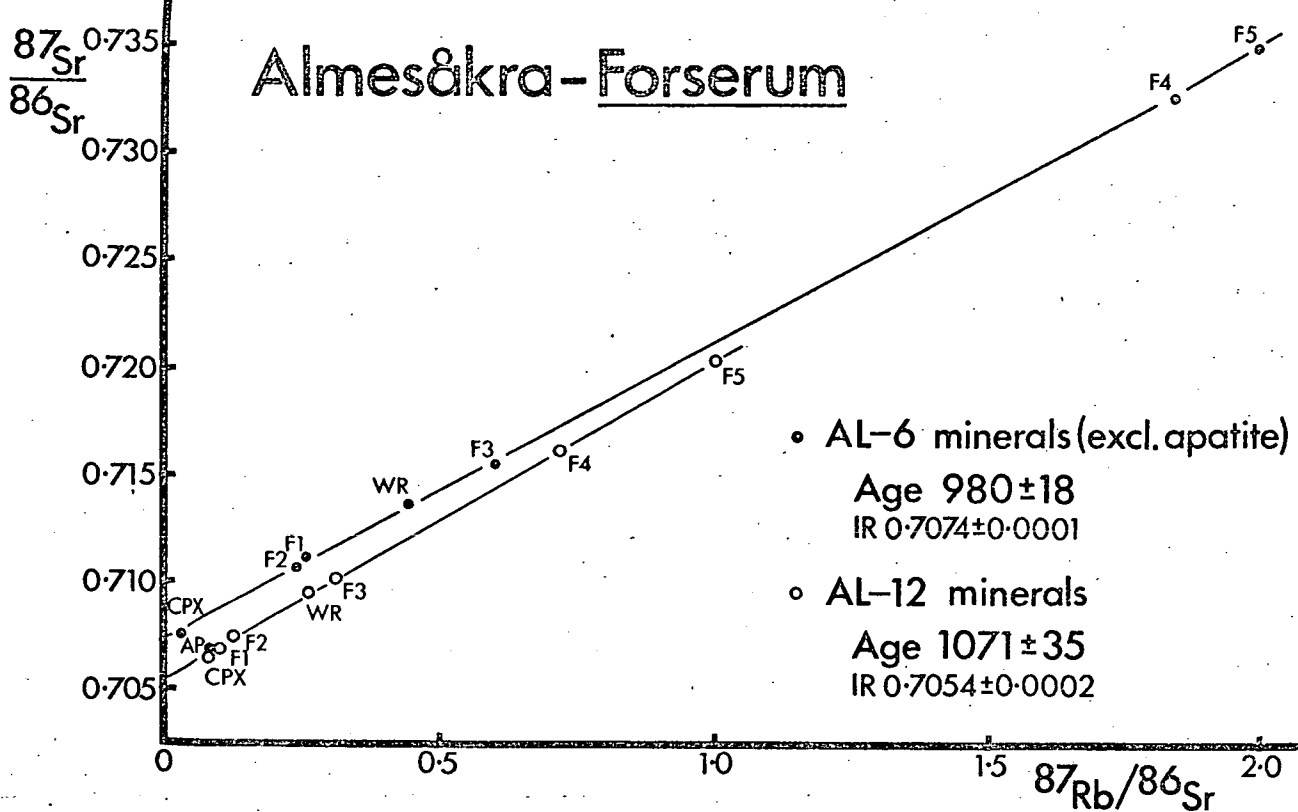


Fig. 4.20 Mineral lines for two hypersthene dolerites from the Forserum sill, Almesåkra region. For location see fig. 4.5.

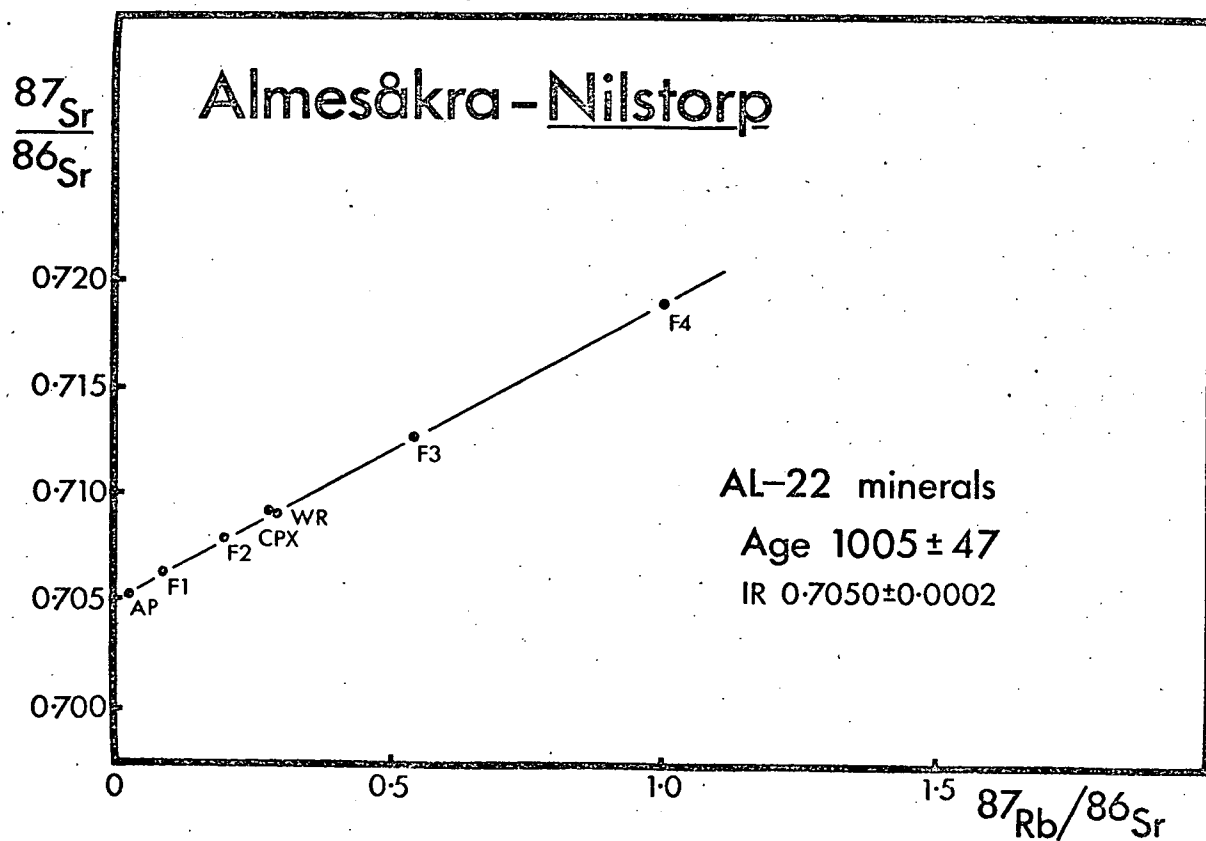


Fig. 4.21 Mineral line for hypersthene dolerite from the Nilstorp sill, Almesåkra region. For location see fig. 4.5.

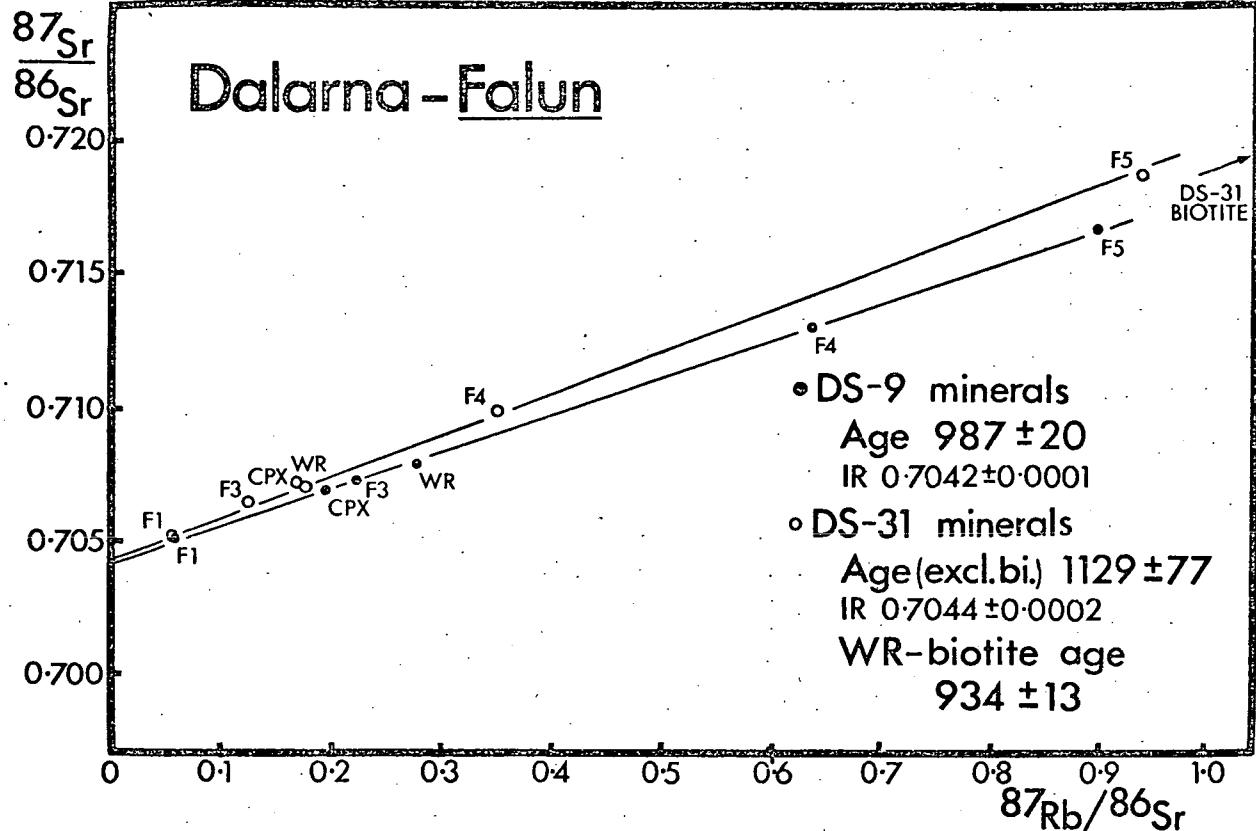


Fig. 4.22 Mineral lines and a whole-rock--biotite age for two olivine dolerites from the Falun dyke, southern Dalarna. For location see fig. 4.5.

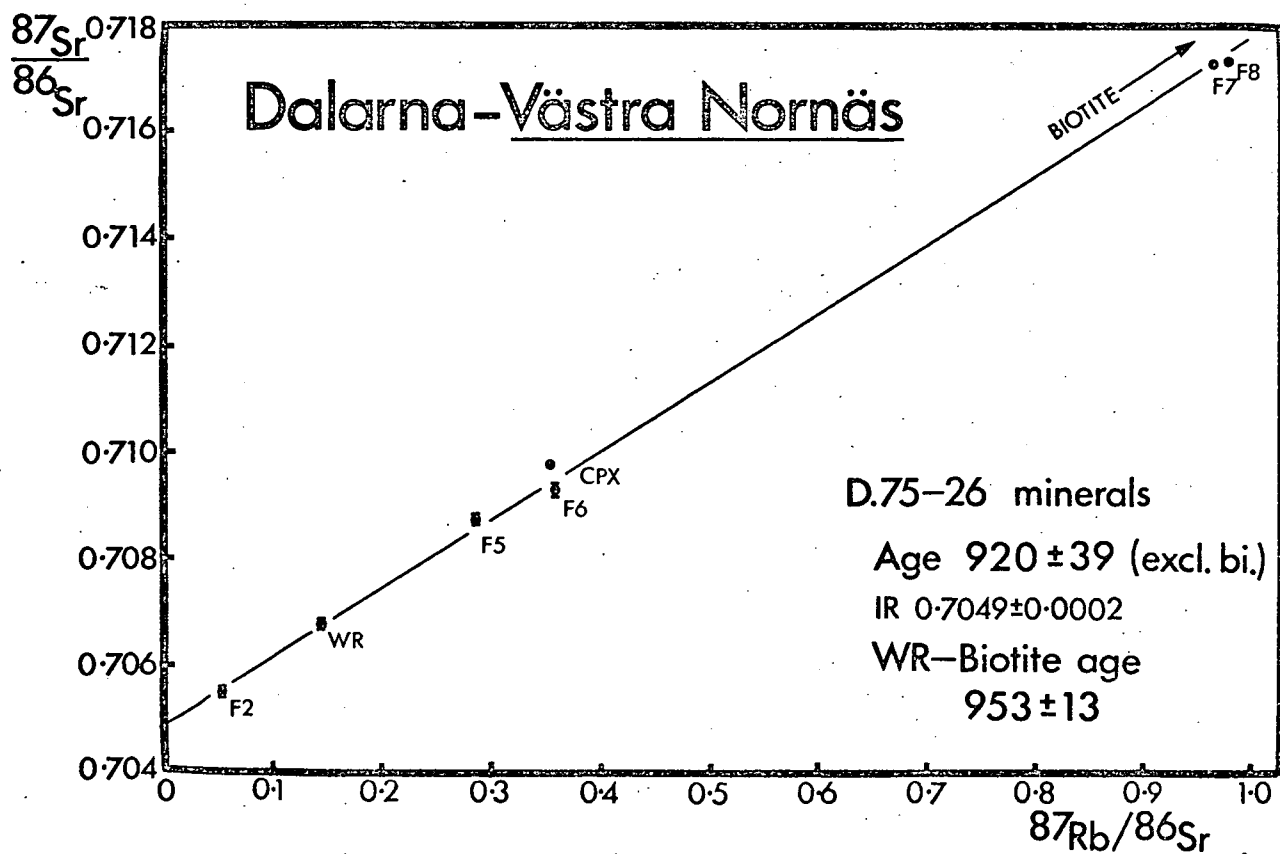


Fig. 4.23 Mineral line and whole-rock--biotite age for olivine dolerite from the Västra Nornäs dyke, northern Dalarna. For location see fig. 4.5.

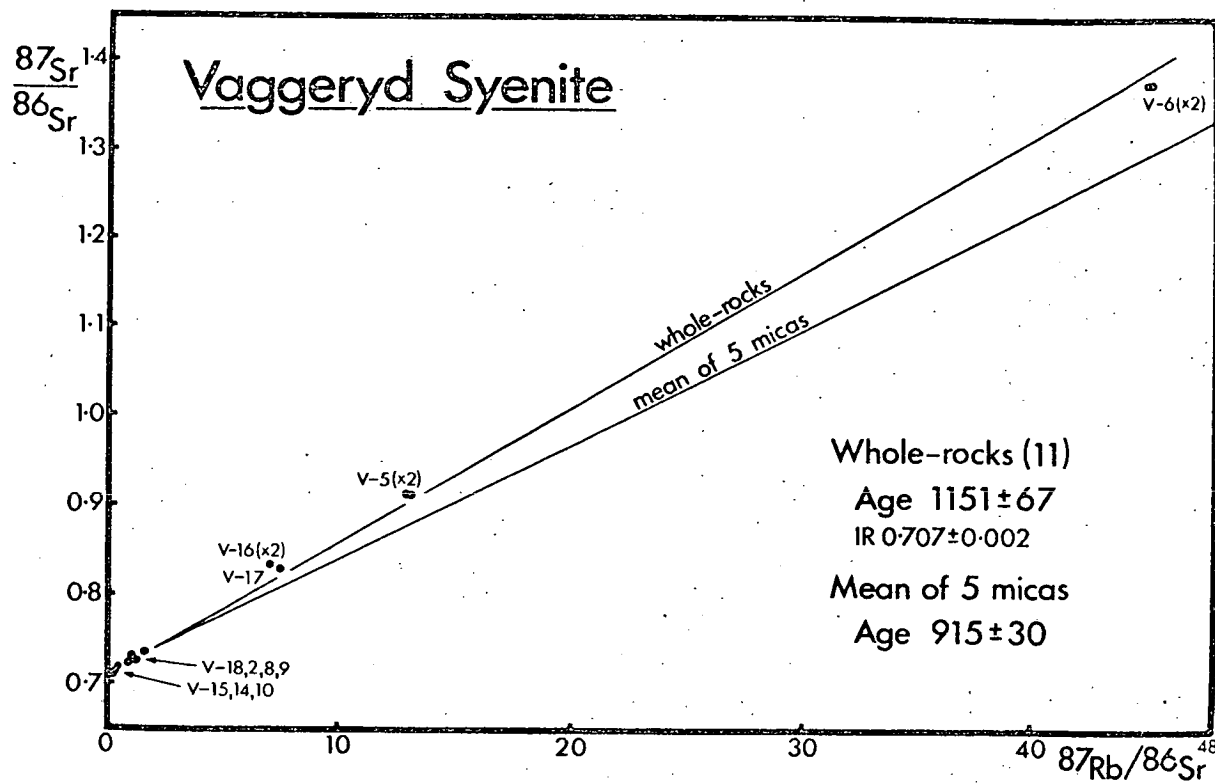


Fig. 4.24 Whole-rock line and generalised age of 5 biotites and muscovites from the Vaggeryd syenite, southern Sweden. For location see fig. 4.5.

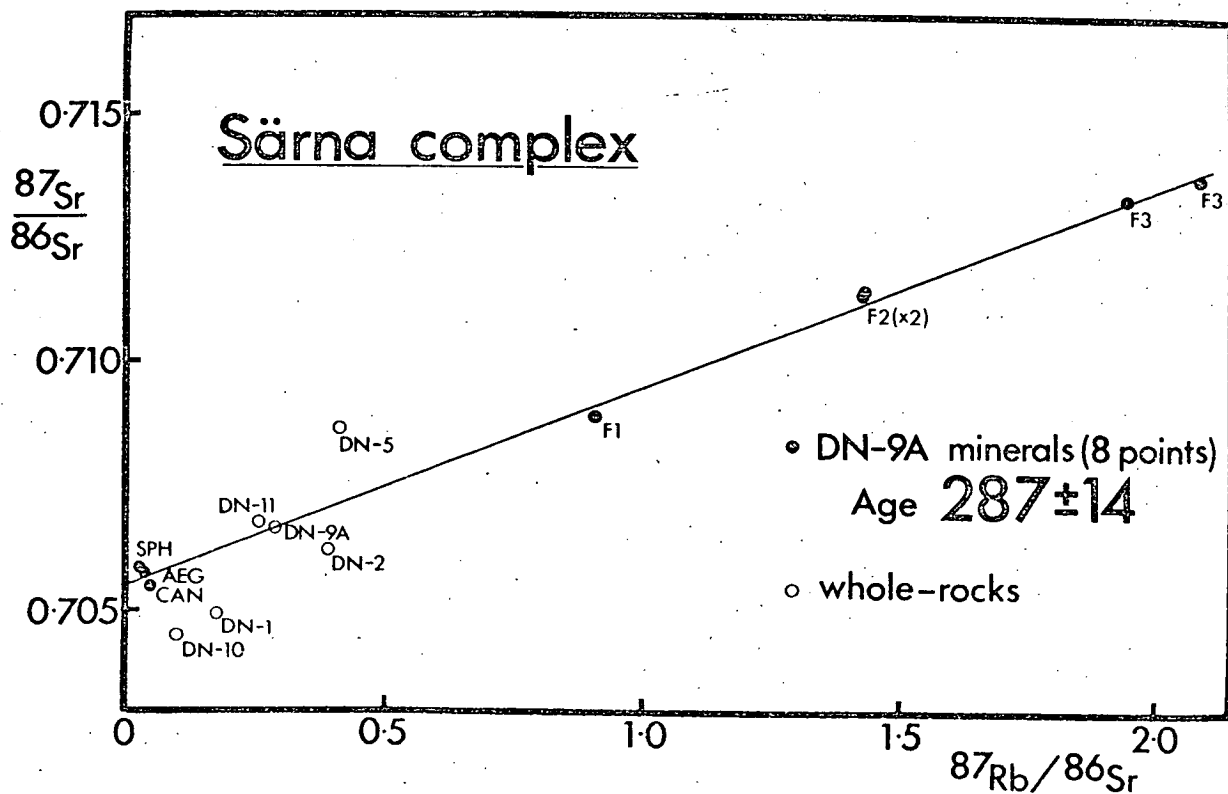


Fig. 4.25 Mineral line and scattered whole-rock analyses from the Särna alkaline complex, northern Dalarna. For location see fig. 4.1.

CHAPTER 5: SIGNIFICANCE OF POST-SVECOFENNIAN IGNEOUS ACTIVITY IN
SWEDEN, AND ITS RELATIONSHIPS TO EVENTS IN GREENLAND AND
CANADA

5.1 Introduction

In the Phanerozoic, and most clearly in the Mesozoic and Caenozoic, basic igneous activity in essentially stable continental regions has frequently been associated with the opening of oceans. In the Tertiary Brito-Arctic Province for example, widespread igneous activity took place on eastern Baffin Island, the western and eastern coasts of Greenland and the western coast of Britain; it is likely that these areas were contiguous or at least close together at the time of volcanism, but they are now separated by major ocean basins (e.g. Vogt and Avery, 1974). Other basic igneous activity in continental areas has less clear tectonic relationships: in most present-day continents (and oceans) there are several points at which magmatic activity is taking place, or has recently taken place, and these do not bear any obvious relationship to the opening of oceans. They have been called 'hot spots' implying a cause in local thermal upwelling in the mantle (e.g. Burke et al., 1973). Another possible cause of cratonic magmatism, illustrated by that taking place in northern China, may lie in lithospheric subduction at the continental margin some distance away.

If the operation of global tectonics similar to that of the present day is accepted for the Proterozoic, then continental igneous activity at this time might have had the same tectonic controls. Thus a widespread period of basic magmatism on several continents, as demonstrated in this chapter for ca. 1250 m.y., could be interpreted in terms of continental separation and ocean opening. Consideration of the tectonics of the North Atlantic

region in the interval 1600-800 m.y. however, both on a local and a continental scale, is dominated by the uncertainty as to the nature and time duration of the Grenville-Sveconorwegian orogeny. Evidence discussed in chapter 1 has shown that Rb-Sr whole-rock and U-Pb zircon ages spread from 1300 to 900 m.y. for 'Grenville events', but that in one case at least episodic rather than continuous igneous/tectonic processes can be demonstrated. An orogeny of 300-400 m.y. duration seems inherently unlikely, especially when compared to more recent examples, and it seems possible that if more sophisticated dating techniques were available, the Grenville might either break down into two or more discrete events separated by quiescent or even tensional episodes, or possibly be shown to be a single 'event' with a much shorter duration than hitherto supposed.

5.2 East-West dyke system

The East-West dykes are confined to a relatively small area of central Sweden (figs. 4.2, 4.3), where intrusion of olivine dolerite, more evolved basic rocks and granophyre took place within the interval 1590-1540 m.y. (section 4.3).

This age corresponds to other results from Sweden. Intrusion of post-tectonic granites and syenites in northern Sweden took place at ca. 1550 m.y. (Welin et al., 1971; Gulson, 1972), and in central Sweden to the immediate North and North-east of the East-West dyke region, Welin (1963) has dated a period of mineralisation by U-Pb at 1585 m.y. In southern Sweden, the alkaline complex of Norra Kärr (see fig. 4.1) has been dated at 1580 ± 62 m.y. by the Rb-Sr whole-rock method (Blaxland, in press).

The correlation with mineralisation in the same area is probably significant: the mineralisation could have been associated with the fracturing and heating of the Svecofennian basement connected with the

magmatic activity which gave rise to the East-West dyke system. Thus the emplacement of the East-West dykes can be related to anorogenic igneous and hydrothermal activity in northern, central and southern Sweden, and may be part of a widespread event at 1590-1540 m.y., though on present evidence such an event would not have encompassed very intensive activity.

5.3 Tuna dykes

The NNE-trending Tuna dykes of southern Dalarna (figs. 4.2, 4.3) show some similarities to the East-West dykes: they have a restricted locale in central Sweden and comprise a range of primitive and more evolved rocks occurring as abundant sub-parallel dykes. They were emplaced at 1401 ± 50 m.y. (section 4.4).

There are several events of similar age in the Baltic Shield. Intrusion of the Ragunda and Nordingrå gabbro-syenite-granite complexes took place between ca. 1450 and ca. 1350 m.y., and the Karlshamn group of granites of southern Sweden are dated at 1455 m.y. (see section 4.1 and fig. 4.1). In addition, two metamorphosed granite bodies in western Sweden have given ages of ca. 1400 m.y. (Skiöld, 1976), and Priem et al. (1972) have obtained a ca. 1400 m.y. result from granitic gneiss in southern Norway.

The association of the Ragunda and Nordingrå complexes is probably anorogenic, but the granites of southern and western Sweden may be associated with deformational events. Thus while the age of the Tuna dykes compares in a general way to ages of igneous activity in both the Svecofennian and the Sveconorwegian regions of the Baltic Shield, it is unclear whether these events should be considered as anorogenic and entirely pre-Grenville, or as being associated with deformational events. Certainly the period 1450-1350 m.y. does not overlap with the distribution

of 'Grenville' ages in the Grenville-Sveconorwegian belt (see section 1.3, fig. 1.2), and this may suggest that the activity is anorogenic. Whatever the cause, fracture and magmatism took place in the Svecofennian crust 1450-1350 m.y. ago, when dolerites, gabbros, anorthosites, syenites and granites were emplaced, and the dolerite to quartz-porphyry Tuna dykes are one expression of this activity.

5.4 Alkali-olivine dolerite sills

These cover a wide extent (500 x 400 km) in northern-central Sweden and adjacent areas of Finland (figs. 4.2, 4.3), and consist predominantly of large intrusions of alkali-olivine dolerite emplaced during the approximate interval 1250 ± 25 m.y. (section 4.5).

The wide area affected by this activity is at once suggestive that the sills do not represent some isolated 'hot-spot' magmatism, but must be part of more significant tectonic/magmatic events.

The only ages in the Baltic Shield comparable to those of the sills come from the Sveconorwegian orogenic belt. In western Sweden, Gorbatschev and Welin (1975) and Welin and Gorbatschev (1976) have dated granite plutons at 1240 ± 30 and 1250 ± 50 m.y. These intrusions are an intermediate stage in a complex tectonic history for the region concerned. In southern Norway, O'Nions and Baadsgaard (1971) record a metamorphic maximum at 1160-1200 m.y., while ages between 1150 and 1300 m.y. have been reported by several authors (see appendix A and fig. 1.2 for compilation). Verstevee (1974) reported a metamorphic maximum and some igneous intrusion in south-western Norway at ca. 1200 m.y. Taken cumulatively, these results indicate that metamorphism and igneous activity were taking place in the Sveconorwegian block in the range 1300 to 1150 m.y. ago.

The age of the alkali-olivine dolerite sills is thus comparable to some

metamorphic and granitic ages in the Sveconorwegian region. It has been suggested however (section 1.3), that ages from this orogenic belt should possibly not be taken too literally, and that the spread of ages from ca. 1300 to 1000 or 900 m.y. may hide quiescent or even tensional tectonic episodes. The alkali-olivine dolerite sills may therefore be associated either with orogenic events in the neighbouring Grenville-Sveconorwegian belt, or may be an expression of fracture and magmatism in the Svecofennian crust during some hitherto disguised tensional period. The question is discussed more fully in sections 5.7 and 5.8, where it is shown that the sills correspond in age to very similar activity in Greenland and Canada.

5.5 Dykes parallel to the margin of the Sveconorwegian orogenic belt

Faults and dykes running parallel to the Schistosity Zone, occurring over a width of 80-130 km, are an important tectonic feature in central and southern Sweden (figs. 4.4, 4.5). They include a considerable proportion of hypersthene-bearing dolerites, and this indicates an overall more tholeiitic character for this swarm than for most other Swedish dolerites. The ages range from ca. 1050 to 890 m.y. (section 4.6), and are believed to reflect at least in part a real variation in intrusive age; reasonable limits for the activity would be 1020 to 890 m.y. (sections 4.6, 4.10).

This age range is more obviously comparable to events in the Sveconorwegian block than for any other of the dolerite groups. Relevant ages are given in appendix A and fig. 1.2. Apart from several ages from metamorphic rocks between 1100 and 900 m.y., there is a widespread and well-documented period of granite plutonism throughout southern Norway and adjacent areas of western Sweden at 950-900 m.y. (see, among others

Killeen and Heier, 1975 and Ski8ld, 1976).

K-Ar determinations presented by Magnusson (1960) from western Sweden show a range from 1130 to 920 m.y. More systematic K-Ar work has given a range of 1000-750 m.y. for hornblendes and whole-rock spilites from western Sweden (Ski8ld, 1976). In southeastern Norway, O'Nions et al. (1969) have shown a range from 1125 to 975 m.y. for many hornblendes, muscovites and biotites. Versteve (1974) has presented Rb-Sr and K-Ar mica ages from southwestern Norway of 940-860 m.y.

These mineral ages demonstrate that the entire region of southern Norway and western Sweden (the Sveconorwegian block) at the present level of erosion underwent cooling through argon blocking temperatures during at least the interval 1130 to 800 m.y. Although literal translation of the range of K-Ar cooling ages into uplift history is probably not justified, the K-Ar ages suggest that cooling had begun slightly before 1100 m.y. There is a marked falling-off in abundance of Rb-Sr whole-rock and U-Pb deformation ages after ca. 1000 m.y., a large proportion of the ages from the Baltic Shield in the range 1000-900 m.y. being due to the late-tectonic granite batholiths. Interpretation of K-Ar ages in terms of uplift is well-established (e.g. Harper, 1967), and the data as a whole suggest uplift and cooling of the Sveconorwegian belt during at least the interval 1000-900 m.y., though possibly beginning earlier than this time, and certainly continuing later in many areas. In addition, the Rb-Sr results from micas of the Vaggeryd syenite show that the rocks of the Schistosity Zone or 'Sveconorwegian Front', marginal to the Sveconorwegian block were cooled through their mica Rb-Sr blocking temperatures at ca. 915 m.y. (section 4.7). Almost immediately to the East of the Schistosity Zone, K-Ar ages and Rb-Sr mica ages are distinctly higher (see for example the Rb-Sr mica ages in section 4.8), and this means that the Svecofennian

block remained relatively static during uplift of the Sveconorwegian. Since the Schistosity Zone separates the two provinces between which a clear break in K-Ar age occurs, then most of the relative movement must have occurred across its 10-25 km width, and the rocks of the zone must have taken up considerable near-vertical shearing motions between the ascending Sveconorwegian block to the West and the relatively cool, static Svecofennian block to the East, a hypothesis consistent with their intensely deformed nature (see section 1.3).

It is not to be expected however, that the front zone will have been able to take up all the movement involved, and stress systems are likely to have been set up in the Svecofennian region; the probable effect would be a partial uplift of the Svecofennian crust adjacent to the front zone. This would result in the Svecofennian crust, by virtue of its upwarping and consequent stretching, being subjected to tensional stresses oriented at right angles to the Sveconorwegian margin; beyond a certain distance, presumably related to the thickness and rigidity of the Svecofennian crust, this effect would die away. It seems reasonable to relate the emplacement of dyke systems parallel to the Sveconorwegian Front within the approximate interval 1020 to 890 m.y. to such a tensional stress regime associated with an upwarping and stretching of the Svecofennian crust (fig. 5.1). The consistent orientation of the dykes, their confinement to a zone 80-130 km in width adjacent to the Sveconorwegian Front, and the correspondence between their emplacement ages and the period of uplift of the Sveconorwegian block all support this scheme. Whether or not some of the dolerites (e.g. the southern 'hyperites') subsequently became involved in front zone deformation during the later stages of uplift of the Sveconorwegian does not affect the validity of the hypothesis.

The structural relationships at the Grenville Front in Canada have been discussed in section 1.3. Studies in this tectonic zone (Dalziel et al., 1969) have suggested a complex origin, possibly even an original pre-Grenville age (Brocoum and Dalziel, 1974). This implies that both the Grenville and the Sveconorwegian fronts could have had complex histories, which would be consistent with their being the boundaries of a major orogenic belt. The reasoning for relative uplift of the Sveconorwegian block however, depends only upon the distribution of K-Ar ages, and merely shows that the front zone must have been the locus of a large part of the relative movement, not that it necessarily originated at this time.

The same argument as for Sweden also applies to Canada, where the K-Ar relationships are very similar: here however, dyke swarms penetrating the pre-Grenville craton are lacking, and it must be supposed either that the stress regime differed from that in Sweden, or that the pre-Grenville crust responded to the stresses in a less brittle fashion. Some evidence may perhaps lie in the fact that at certain points at the margin of the Grenville, Lower Proterozoic formations can be traced into the orogenic belt with little or no change in structural level (Dalziel et al., 1969; Wynne-Edwards, 1972). Additionally, the Grenville Front is rather indistinct in places, being occasionally a thrust, major fault or simply a metamorphic transition (Wynne-Edwards, 1972), though Dalziel et al. (1969) reported mylonites at all observed localities. The wide shear zone, such an obvious feature of Swedish geology, is apparently absent or obscured in Canada, and this certainly argues for different tectonic conditions during uplift of the Canadian portion of the belt, possibly involving weaker uplift, a less brittle pre-Grenville crust, or both.

5.6 Age of 'Jotnian' sedimentary formations

The present geochronological work sheds some light on the age of 'Jotnian' sedimentary rocks which overlie the Svecofennian crust, though it does not clearly bracket their deposition.

These sediments consist of six major outcrops (fig. 4.1): the Nordingrå Sandstone Formation (Sobral, 1913), the Dala Sandstone Formation (Hjelmqvist, 1966), the Gävle Sandstone Formation (Gorbatshev, 1967), the Almesåkra Formation (no systematic study published), and in Finland the Satakunta Sandstone Formation (Laitakari, 1925) and the Muhos Formation lying to the North of fig. 4.1.

Direct age determinations in Sweden by K-Ar have given ca. 900 m.y. for Almesåkra Formation slates (3 measurements), and 1185 m.y. for Dala sandstone (both Magnusson, 1960). All the formations except the Muhos are cut by dolerites. In northern-central Sweden and in Finland these intrusions are of ca. 1250 m.y. age, and the Rb-Sr isochron of 1415 ± 30 m.y. for the Nordingra granite (Welin and Lundqvist, 1975) provides a lower age bracket. If the four sedimentary formations of this region of the Baltic Shield are correlatable, as assumed by a large number of geologists, then sedimentation is constrained within the approximate interval 1415 to 1250 m.y. On the other hand, the data for Bunkris (see section 4.6, fig. 4.8) suggest that this dyke may be as old as 1550 m.y.: since the intrusion is mapped as cutting the Dala Sandstone Formation, this would mean that the sediments are somewhat older than previously thought, and could not correlate with those of Nordingrå. The question remains unresolved because of the uncertainty as to the age of the Bunkris dyke: due to alteration, direct age determination by Rb-Sr on the lavas of the Dala Sandstone Formation was not possible. Palaeomagnetic data (Mulder, 1971) does however favour a pre-1250 m.y. age for both the lavas

of the Dala Sandstone Formation and the Bunkris dyke.

The age uncertainty makes an overall interpretation of the 'Jotnian' sediments difficult: they could be molasse formations deposited as a result of erosion of an early Grenville-Sveconorwegian mountain terrain, or equally they could be basin-fill sediments deposited following fracture of the pre-Sveconorwegian craton during a tensional tectonic phase in pre-Grenville-Sveconorwegian time. The latter would be similar to the examples of Burke and Dewey (1973), where basin-filling sediments on the continents are thought to represent the 'failed third arms' of triple-junction continental rifts.

Slates of the Almesåkra Formation have given K-Ar ages of ca. 900 m.y., and the sediments are cut by dolerites giving Rb-Sr mineral ages of ca. 1000 m.y.: both these results may suggest a somewhat younger age for this formation than for those to the North. The large quantities of sand mixed with dolerite in the Tärnö dyke of Blekinge, to the south of Almesåkra, is evidence for unconsolidated, probably recently deposited sediment in the area at ca. 900 m.y. The Almesåkra region too, contains cases where dolerite has become intimately mixed with sediment, in this case pebbles (the well-known 'diabase conglomerates'), and the same argument may also apply here (see Magnusson, 1965). The sediments of Almesåkra and the former sediments of Blekinge may be correlatable, in which case they represent conglomerate, sandstone, siltstone and rare limestone deposition on the continent at some time around 1000-900 m.y., an age which corresponds to that of the Stoer Group of the Torridonian Sandstone of Scotland (Moorbath, 1969). Since it has been shown in section 5.5 that post-tectonic uplift of the Sveconorwegian orogenic belt was at its maximum during the 1000-900 m.y. period, it seems possible that the sediments of southern Sweden could represent molasse formations derived

from a Sveconorwegian mountain terrain and deposited on the lower-lying Svecofennian crust during this time.

5.7 Transatlantic correlation at ca. 1250 m.y.: a widespread basic igneous episode

The widespread alkali-olivine dolerite sills of 1250 ± 25 m.y. have equivalents in the Shield areas of Greenland and Canada. Fig. 5.2 (a replica of fig. 1.1) shows the distribution of Caledonian- and Grenville-age orogenic belts in the North Atlantic region, as well as igneous rocks occurring in the pre-Grenville cratons from 1300 to 900 m.y. This interval includes igneous events of ca. 1250 m.y. and also various younger events: the Mid Keweenaw, the Late Gardar, the Vaggeryd syenite and the 1000-900 m.y. dolerites of Sweden. These younger events are discussed in sections 5.5 and 5.9.

The Mid Gardar magmatic event of ca. 1250 m.y. in south Greenland comprises an association between lamprophyres, alkaline dykes, olivine dolerites and oversaturated central complexes. The olivine dolerites resemble the ca. 1250 m.y. dolerites of the Baltic Shield, and in fact the rocks of the two regions are of strikingly similar aspect in the field.

The Sudbury dykes of the Canadian Shield (fig. 5.2) have been dated by van Schmus (1965; see 1975 for discussion) using the Rb-Sr technique, at 1245 ± 50 m.y. for dolerite biotite, 1230 ± 70 m.y. for contact-rock biotite, and 1205 ± 100 m.y. for a 3-point line for minerals from a rheomorphic vein. The Rb-Sr whole-rock work of Gates and Hurley (1973), which redefined the ages of several earlier Canadian dolerite dyke systems, did not produce any more accurate results for the Sudbury dykes than were obtained by van Schmus (1965), and the Mackenzie set III dykes (fig. 5.2) are still only inferred to be the same age as the Sudbury

dykes. Four K-Ar biotite ages for the Mackenzie set III dykes fall between 1330 and 950 m.y. (see van Schmus, 1975 for discussion). However, the very similar and consistent palaeomagnetic pole positions (Fahrig and Jones, 1969) given by the Sudbury dykes, the Mackenzie set III dykes, the Muskox intrusion and the Coppermine River Group lavas (all shown in fig. 5.2) have generally been regarded as good evidence for their having a similar age. The Coppermine River Group lavas have given an Rb-Sr age of 1284 ± 45 m.y. (Wanless and Loveridge, 1972).

It is therefore very likely that there was a widespread cratonic episode in the Canadian Shield at ca. 1250 m.y., and this should be compared to the well-dated events of Greenland and the Baltic Shield. In Scandinavia the generalised error on the magmatic events seems to be ± 25 m.y., as given by biotites, and by implication this uncertainty may also apply to south Greenland. In Canada the error seems to be ± 100 m.y., though the palaeomagnetic evidence indicates a closer grouping. The magmatic rocks of this overall episode comprise a selection of very mildly to moderately alkaline dolerites; these were associated with dykes of more differentiated compositions and central complexes only in south Greenland. The highly simplified alkali-silica relationships for all the dolerites of known or suspected ca. 1250 m.y. age are shown in fig. 5.3. The means disguise the fact that the Gardar and Åsby dolerite data define quite a large spread of results; almost all of these however, lie on the alkaline side of the Hawaiian alkali basalt-tholeiite divide. The general conclusion that the majority of the ca. 1250 m.y. dolerites are of the alkali-olivine type therefore seems reasonable. The single analysis of Särna type olivine dolerite, also clearly dated to ca. 1250 m.y., lies on the tholeiitic side of the divide: the less alkaline character of this magma type has already been noted on the basis of mineralogy.

Intrusive forms are predominantly dykes in Canada and Greenland, and almost exclusively sills in Scandinavia. The occurrence of the layered Muskox intrusion and the Coppermine River Group lavas in northwestern Canada (Baragar, 1969) is probably a reflection of the higher structural level exposed there, though there is no indication that lavas of the appropriate age ever existed in the other regions. The distinction between sills in Scandinavia and dykes elsewhere may also depend upon structural level, or perhaps some more fundamental control, but the existence of widespread dolerite intrusions of very similar or identical age is taken to imply major stressing and fracturing of the pre-Grenville cratons, regardless of the intrusive forms seen at the present level of exposure. It seems likely that such widespread magmatic events, comprising similar magma types over a distance of at least 4000 km on any continental reconstruction, must be related and have a similar tectonic cause. Additionally such widespread contemporaneous events seem, by comparison with similar basic igneous activity in the more recent past, to be related to large-scale, rather than local tectonic events.

The existence of widespread basic igneous activity on the pre-Grenville cratons at ca. 1250 m.y. is therefore probably an expression of global tectonic processes in the Upper Proterozoic. In consideration of the nature of these processes, it is necessary to take into account a somewhat broader time interval than ca. 1250 m.y.: three approaches have provided important evidence on the question of North Atlantic tectonics in the Upper Proterozoic: structural studies, geochronology and palaeomagnetism, and these are discussed in section 5.8.

5.8 Upper Proterozoic tectonics and the possible significance of

ca. 1250 m.y. basic magmatism

This section deals with evidence from tectonic studies, geochronology and palaeomagnetism as affecting the interpretation of thermal events, orogeny and possible continental movements in the North Atlantic region from ca. 1600 to 800 m.y., with a view to establishing the significance of widespread basic igneous events at ca. 1250 m.y.

Two tectonic hypotheses

There are two radically different views of the style of tectonics in the Archaean and Lower Proterozoic. The first of these is derived from a general application of uniformitarianism and holds essentially that Archaean and Lower Proterozoic orogenies took place as a result of lithospheric subduction and continental convergence. Features characteristic of present-day continental separation, such as rift structures at continental margins, the 'failed arms' of mantle-plume generated triple junctions, are recognised even as far back as pre-Hudsonian (pre-1800 m.y.) time (Burke and Dewey, 1973). It is also implied therefore that orogeny was associated with convergence of continental areas, and specifically the Grenville has been compared to recent collisional belts such as the Himalayas (Dewey and Burke, 1973).

The opposite view is derived from observation of the relationships between orogenic belts and pre-orogenic cratons. Lower Proterozoic orogenic belts, even as late as the Pan-African (700-500 m.y.), have characteristically variable tectonic trends on a continental scale, and often involve rather narrow belts of 'reworked' material lying between stable cratons (e.g. Shackleton, 1969): rocks identifiable with former oceanic crust are usually absent, and geological features can sometimes be

correlated from one craton to another (see Clifford, 1968 fig. 9). On a large scale, the cratons, which are small compared to present-day continents, often form a mosaic pattern separated by the mobile belts (Clifford, 1970): unless it is supposed that the continent became separated into several small plates which were re-united at the time of orogeny into apparently the same relative positions, it seems reasonable to suggest that the mobile belts were formed without significant plate movement. They might have been produced as a result of minor horizontal movements between cratons and related highly increased heat flow (with emplacement of granite bodies contributing to deformation) in the intervening area, the mobile belt; see Anheusser et al. (1969) and Sutton and Watson (1974). The precise mechanism and the role of the mantle in the orogenic processes are however, somewhat vaguely defined. This view of Proterozoic orogenesis receives support from African palaeomagnetism, which does not allow movement of more than $10-20^{\circ}$ between Archaean cratons since ca. 2200 m.y. ago (Piper et al., 1973), and from recent work in Canada (Christie et al., 1975) which places limitations on plate motion prior to and during the ca. 1800 m.y. Hudsonian episode.

This second hypothesis of major tectonics in the Archaean and Proterozoic implies a fundamental change in earth processes in the Upper Proterozoic or latest Precambrian, since the Caledonian and all later belts have probably been produced by the operation of plate tectonics similar in style to the present day (e.g. Dewey, 1971). The Grenville-Sveconorwegian orogeny occupies something of an intermediate position in this scheme as far as the North Atlantic region is concerned, in that it post-dates the supposed 'in situ' ca. 1800 m.y. belts but pre-dates the Caledonian. The Grenville-Sveconorwegian is, as noted in chapter 2, the earliest orogenic belt in the North Atlantic region to

exhibit a single linear trend.

Geochronology

Apart from the ages for the basic igneous events themselves, ages between 1300 and 1200 m.y. in the North Atlantic region come almost entirely from the Grenville-Sveconorwegian orogenic belt. Fig. 1.2 and appendix A show that ages generally regarded as Grenville occur within the limits 1300-900 m.y.: there is a marked concentration in the 1200-1000 m.y. period, but several results are reported in the 1300-1200 m.y. range from both Scandinavian and Canadian portions of the belt. These include both metamorphic ages and ages interpreted as giving times of plutonic crystallisation.

The possibility of a break in orogeny has already been noted, but the geochronological data as a whole imply, if interpreted literally, that Grenville-Sveconorwegian events were in progress from 1300 m.y. onwards, and were essentially continuous up to at least 1000 m.y.: this conclusion could have considerable bearing on the origin of widespread ca. 1250 m.y. basic igneous activity in the pre-Grenville cratons.

Palaeomagnetism

The two opposite views of Proterozoic, at least Lower Proterozoic tectonics are to a certain extent reflected in the interpretation of palaeomagnetic data.

The period up to 1800 m.y. is relatively obscure palaeomagnetically, but on the data available Piper (1976a and b) has suggested that Africa and North America formed a single continent at least as far back as 2160 m.y., while McGlynn et al. (1975) have pointed out the considerable differences between the polar wander paths in the 2200-1800 m.y. interval.

This controversy bears directly on the question of Lower Proterozoic tectonics, because drift of continental areas would favour the plate tectonic model, whereas coherence would support the 'in situ' model. It is fair to say however, that both Piper (1976a and b) and McGlynn et al. (1975) support the Sutton and Watson model at least as far as orogenic belts within Africa and within Canada are concerned; any disagreement is over the relative motions of these major units.

There is a reasonable consensus among palaeomagnetists that Africa, North America and Scandinavia were joined as essentially a single continental area from ca. 1700 m.y., the end of the Hudsonian-Svecofennian episode to at least 1300 m.y. (Donaldson et al., 1973; Neuvonen, 1974; Piper, 1974, 1976a; Poorter, 1976). This is derived from the general agreement of pole trends for this period, and one reconstruction for this period is the 'Proterozoic supercontinent' of Piper (1974).

There is abundant palaeomagnetic data for the ca. 1250 m.y. dolerites in Scandinavia and North America, and the Keweenawan formations yield data for North America down to 1100-1000 m.y., but after this time and up to ca. 700 m.y., almost all palaeomagnetic data for these two regions have been derived from rocks deformed in the Grenville-Sveconorwegian orogeny. Up to the present work, no poles from the pre-Grenville cratons for times between 1000 and 700 m.y. had existed. The data from the Grenville belt define a polar trend that does not correspond to any other data from North America, and a comparison of the Grenville track with Mid-Upper Keweenawan poles (ca. 1120-1000 m.y.) leads to the hypothesis that the Grenville belt became magnetised when far away from the pre-Grenville Canadian Shield, and when the two areas were converging at approximately 'normal' rates of drift. This would imply that the Grenville was a 'plate-tectonic' orogenic belt. On the other hand, Piper (1975) has shown that the

Grenville poles do correspond in a general way with the 1000-800 m.y. portion of the African polar path, and new data from Scandinavia (Patchett and Bylund, in preparation) demonstrates that the poles from rocks of the Sveconorwegian belt correspond to those of the ca. 1020-890 m.y. dykes emplaced into the pre-Sveconorwegian craton. Thus it is likely that the Grenville-Sveconorwegian track is younger than 1000 m.y., and this would mean that comparison of Grenville poles with 'cratonic' poles of 1120 to 1000 m.y. is unsound and that the orogenic belt may well have been magnetised while undergoing post-tectonic uplift in its present position relative to the pre-Grenville-Sveconorwegian cratons.

Piper (1975) interpreted the agreement between the Grenville poles and 1000-800 m.y. data from Africa as favouring his 'supercontinent' reconstruction down to 800 m.y., and the data from the Grenville and from Scandinavia certainly favour a coherent North Atlantic continent during 1000-900 m.y. The data do not however disprove the operation of continental drift, but merely show that the Grenville data no longer require this. The question must depend on palaeomagnetic poles for the period 1250 to 1000 m.y., which are not yet available from Scandinavia.

Possible significance of ca. 1250 m.y. basic igneous events

In the light of the data reviewed, three possible origins for the ca. 1250 m.y. basic igneous events in the pre-Grenville cratons can be proposed:-

1. The geochronological evidence for Grenville-Sveconorwegian events between 1300 and 1150 m.y. is neglected and it is proposed that the 'Proterozoic supercontinent' rifted and separated in the North Atlantic region at 1250 m.y.: the basic magmatism is the expression of this

phase of continental separation (Burke and Dewey, 1973). Later convergence gave the Grenville-Sveconorwegian belt from ca. 1150 m.y. onwards. It is possible to explain pre-1150, even pre-1100 m.y. 'Grenville' ages as recording deformation and intrusion associated with the 'anorthosite event', and therefore due to high heat flow etc., rather than to actual orogeny.

2. The geochronology is interpreted literally and the Grenville-Sveconorwegian orogeny had a protracted history from 1300 to 1000, or even 900 m.y. In this case the basic magmatism could have been the result of fracture of the pre-Grenville cratons caused by some climactic tectonic event: a continental collision for example, if plate tectonics were to have operated at this time, would be capable of producing the necessary 'instantaneous' stresses in the cratons.

3. The spread of ages from 1300 to 900 m.y. for the Grenville-Sveconorwegian belt is realised, but is assumed to disguise quiescent or even tensional episodes in North Atlantic tectonics. Section 1.3 discusses evidence for one of these in Scandinavia at some time between 1250 and 1050 m.y. (Skiöld, 1976), and it is implied that there may be a corresponding period in Canada, or even that there may be earlier or later quiescent episodes. Thus the continents could have broken up at ca. 1250 m.y., as given by the basic magmatism, and any necessary relative movement of Baltic and Canadian Shields would have taken place at least before the time of magnetisation of the Grenville-Sveconorwegian belt at 1000-800 m.y., but probably before the concentration of metamorphic ages beginning at 1150-1100 m.y. This hypothesis would imply that the Grenville-Sveconorwegian belt was

formed by at least two somewhat distinct tectonic episodes, ca. 1300-1250 m.y. and 1150-1000 m.y.: the coherence of the whole belt given by K-Ar ages would be a consequence of uplift and cooling following the latest events, and would not be a reflection of the age of original deformation.

At present it does not seem possible to categorically reject any of these three alternatives. However, in view of the necessity of reconciling any hypothesis with the abundant geochronological data from the orogenic belt, and because at least one phase of tectonic quiescence has been demonstrated, hypothesis 3 seems possible. On the other hand, interpretation of the 'Grenville' ages between 1300 and 1100 m.y. as due to the anorthosite, granite etc. plutonism of the 'anorthosite event' would allow a definition of the Grenville as being a post-1100 m.y. orogeny, and of the ca. 1250 m.y. fracture and magmatism as a pre-Grenville continental separation. Either of these would also be consistent with the overall alkali-olivine basalt compositions of the magmas, which may suggest an association with tensional rather than compressional tectonics.

5.9 Cratonic igneous events 1200-1100 m.y.

The remaining igneous events shown in fig. 5.2 are the Keweenawan volcanism of the Canadian Shield (1140-1120 m.y.; Silver and Green, 1972), the Late (main) Gardar activity in south Greenland (ca. 1170 m.y.; Blaxland et al., in press), and the Vaggeryd syenite in southern Sweden (1151 $\pm$ 67 m.y.; this study). The Mid Keweenawan and Late Gardar appear to be distinct, but the error on the Vaggeryd syenite is such that it could correlate with either of these events.

For these rocks, the problem of possible contemporaneity with Grenville-Sveconorwegian orogenesis is more acute than for the ca. 1250 m.y. events. The Keweenaw is indisputably a rift structure (see Burke and Dewey, 1973), the Late Gardar comprises crustal extension by dykes and a variety of alkaline rocks usually associated with rifting (see Upton, 1974), and the Vaggeryd syenite bears some resemblances to the oversaturated syenites of the Late Gardar. Three tectonic environments for the magmatism of these regions can be considered. First, they may fall in a quiescent or tensional period within the Grenville, possibly the same as that envisaged as an explanation for the ca. 1250 m.y. events. Second, they may be entirely pre-Grenville, which would be allowed by the interpretation of early 'Grenville' ages as igneous events only: this would have the advantage that all Gardar events from 1330 to 1170 m.y. (Blaxland et al., in press) could then have taken place under a similar tectonic regime, consistent with the similarities in style of activity and differentiation trends for Early, Mid and Late Gardar (Upton, 1974). The post-1200 m.y. cratonic igneous events in the North Atlantic region could then represent persisting tensional tectonics following the ca. 1250 m.y. continental separation. Third, they may have taken place at the same time as Grenville orogenic events: this alternative would be reasonable if the Grenville were a belt of the 'in situ' type, involving no plate compression.

As is the case for the ca. 1250 m.y. cratonic events, the uncertainty as to the nature, timing and duration of Grenville-Sveconorwegian events means that it is not possible to discriminate between these alternatives. Since precise contemporaneity between the Mid Keweenaw, Late Gardar and Vaggeryd events is not demonstrated however, it is not necessary in this case to envisage a single widespread period of craton fracture: the three

magmatic events could be unrelated to each other, and this would simplify to a large extent any problems of global tectonics.

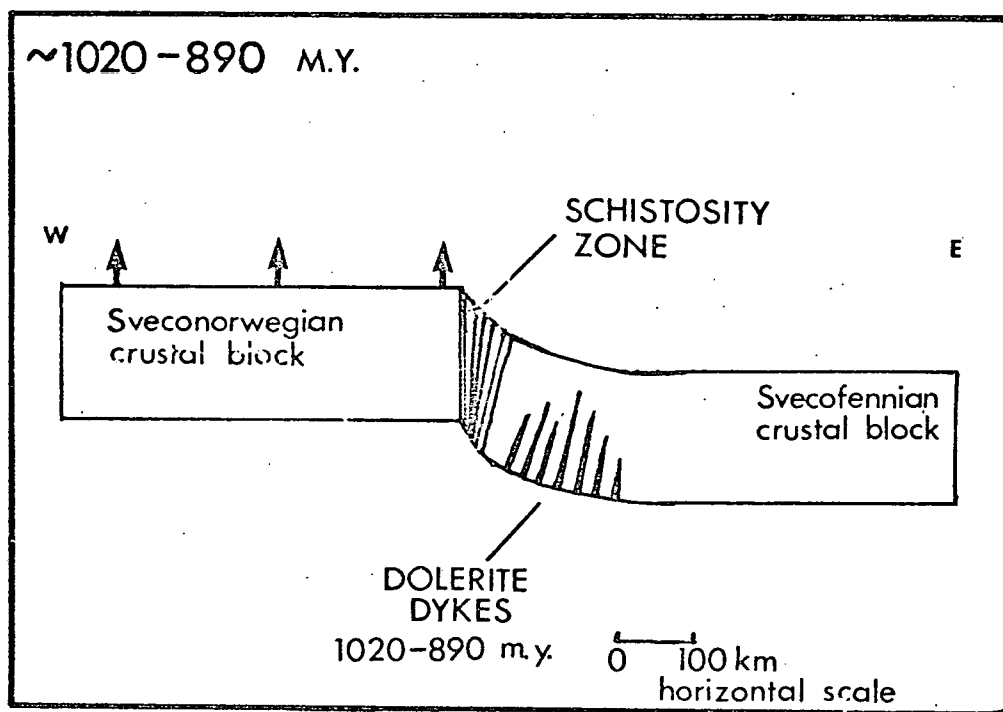


Fig. 5.1 Postulated tectonic scheme for post-orogenic uplift of the Sveconorwegian region with related tensional fracture and emplacement of dolerite dykes into the neighbouring Svecofennian region during the approximate interval 1020-890 m.y.

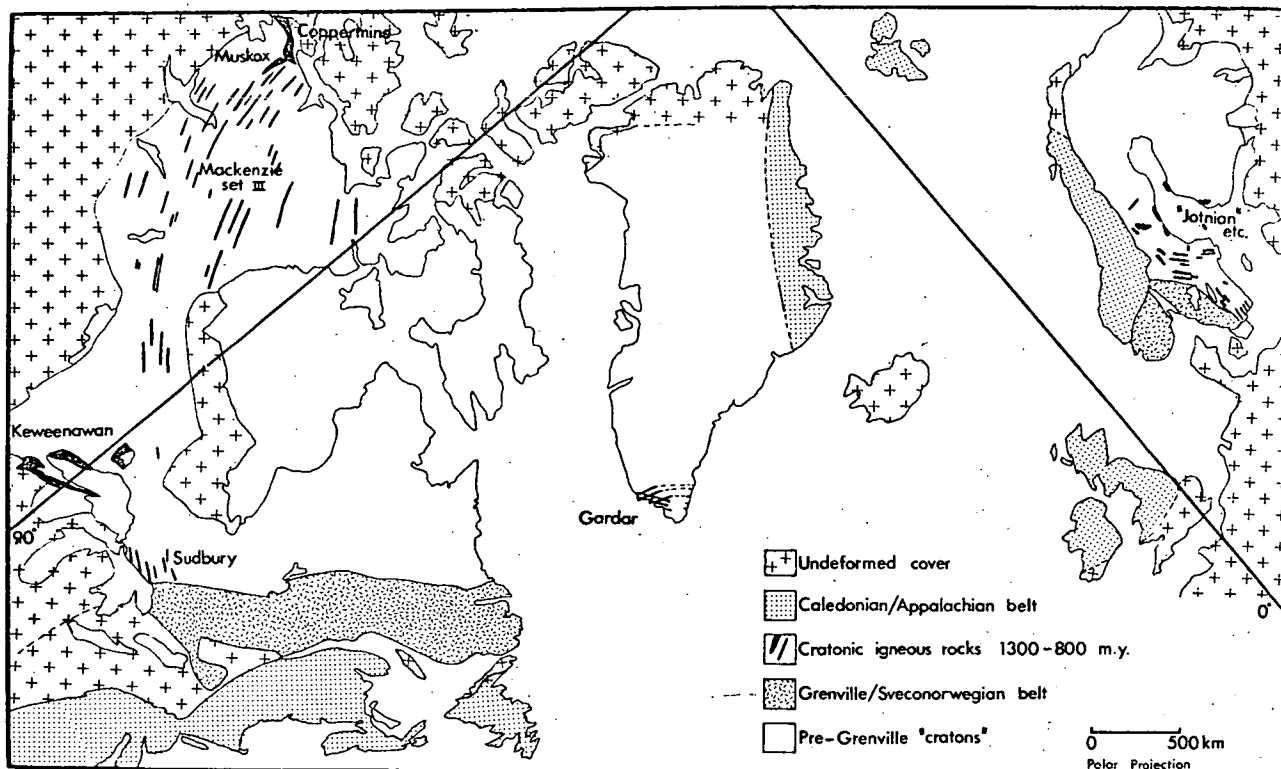


Fig. 5.2 A replica of fig. 1.1: simplified tectonic map of the North Atlantic region showing 1300-900 m.y. cratonic igneous rocks relative to the Grenville and Caledonian orogenic belts.

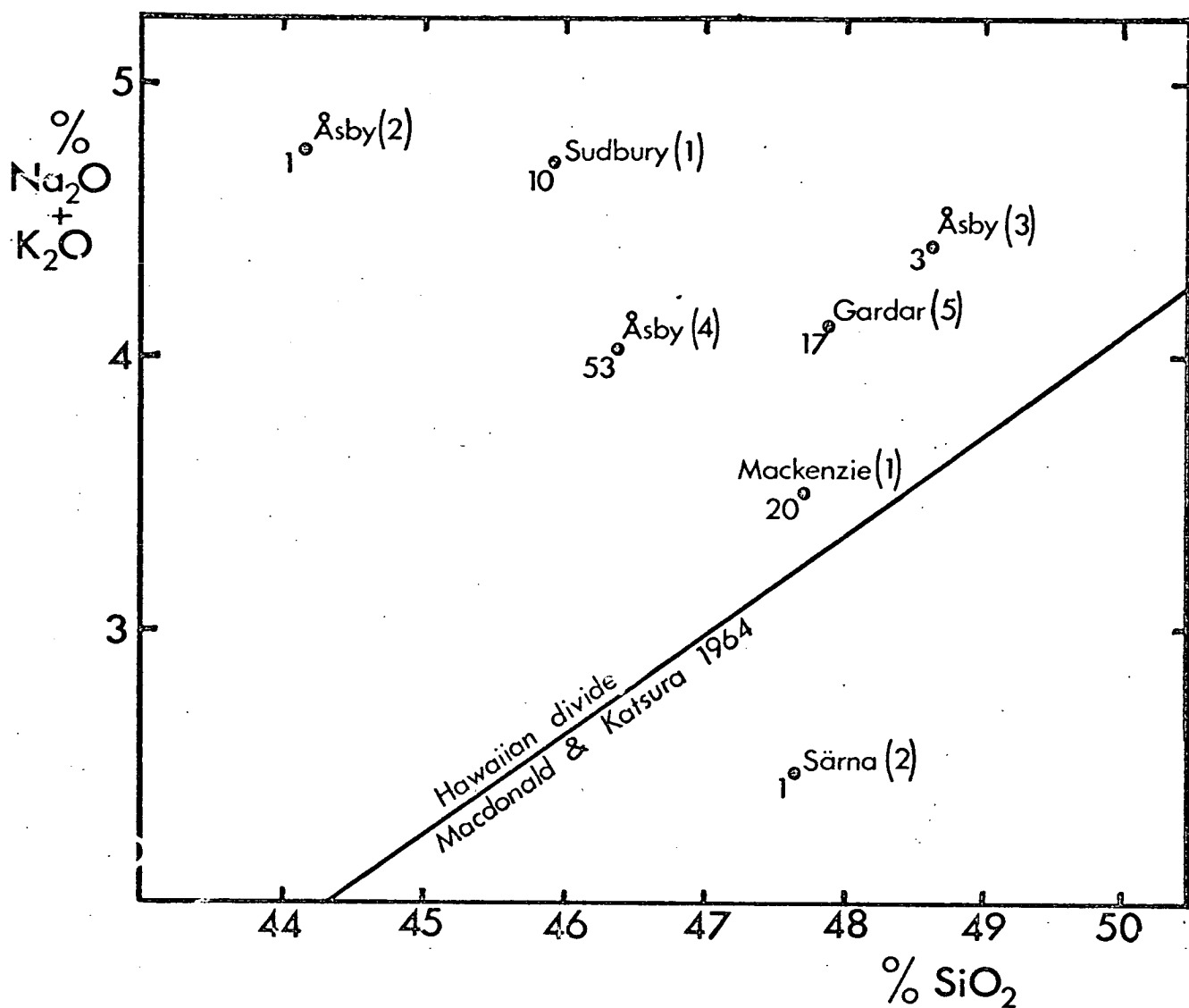


Fig. 5.3 Highly simplified alkali-silica relationships among dolerites of known or supposed ca. 1250 m.y. age: each mean represents a wide spread of results. Numbers to left of data points indicate the number of analyses in each mean, with references shown in parentheses after the locality: 1. Fahrig et al., 1965; 2. Hjelmqvist, 1966; 3. Lundqvist, 1968; 4. Lundqvist and Samuelsson, 1973; 5. Winstanley, 1975.

6.1 Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of anorthosites and gabbros from
south Greenland

Location fig. 6.1; sample localities section C.28, fig. 6.1; analytical data section D.28; isochron diagram fig. 6.2.

The Gardar Province of south Greenland has been briefly described in sections 3.1 and 3.2, and a general map of the region is given in fig. 3.1. The rocks of both Mid Gardar (ca. 1250 m.y.) and Late Gardar (ca. 1170 m.y.) include intrusions which contain xenoliths of anorthosite and megacrysts of plagioclase. These are most common in dykes of hawaiitic or mugearitic composition (often known as the 'Big-feldspar dykes' on this account), and are somewhat rarer in rocks of either less or more differentiated compositions. Bridgwater and Harry (1968) observed a sympathetic variation between the compositions of xenolith or megacryst plagioclase and host rock, and inferred a close relationship between anorthosite genesis and the differentiation of Gardar magmas. In terms of their hypothesis, magmas evolved by plagioclase fractionation, and anorthosite cumulates formed at depth. These cumulates could then be disrupted and carried upwards either by the parent magma itself or by later batches of magma.

The present work was undertaken in order to test this hypothesis in one well-documented case: that of a gabbro giant dyke occurring at the North-east end of Tugtutôq island in the neighbourhood of Narssaq (see figs. 3.1, 6.1). In this area the country rock is the Julianehåb granite dated at 1780 ± 20 m.y. (van Breemen et al., 1974), with cooling ages of ca. 1600 m.y. The rocks studied belong to the Late Gardar (ca. 1170 m.y.) magmatic phase in this area (Blaxland et al., in press).

The earliest event in the magmatic cycle was intrusion of the gabbroic to foyaitic Hviddal giant dyke of Tugtutôq, dated at 1175 ± 9 m.y. (van Breemen and Upton, 1972; age adjusted to a calibration based on NBS stoichiometric standard salts). The activity continued with emplacement of further large dykes on Tugtutôq composed of gabbro, syenogabbro and subordinate quartz-syenite (the Younger Giant Dyke Complex), and it is in the gabbroic rocks of these giant dykes that anorthositic material first appears in the area. Plagioclase megacrysts up to 15 cm occur sparsely through large parts of the dyke complex and become more common in the northern dyke at the eastern end of Tugtutôq (fig. 6.1); at Assorutit anorthosite xenoliths up to 100 m across are abundant. They are laminated with plagioclase crystals up to 15 cm and show a crude layering. The rocks, interpreted as labradorite cumulates (Upton, 1964), consist of plagioclase, olivine, ilmenite, magnetite, apatite, augite and biotite. Whereas there is evidence of slight alteration of the gabbros and anorthosites (growth of minor chlorite and epidote), there is none to suggest that the anorthosites have been penetrated or remelted by the gabbro magma or that either have been affected by fluids emanating from later intrusions. This is not the case for the contiguous gabbro on the mainland where both host rock and anorthositic material have been strongly affected by metasomatism associated with later cross-cutting alkaline complexes; for this reason these rocks were not included.

Small potassic ultramafic bodies were emplaced either within or alongside the gabbro dykes, and the whole area was then invaded by hawaiite to mugearite dykes carrying abundant plagioclase megacrysts and anorthositic material (the 'Big-feldspar dykes'). After further intrusion of trachyte, phonolite and comendite dykes activity in the Narssaq area terminated with emplacement of the Tugtutôq Central Complex, dated at 1168

± 37 m.y. (van Breemen and Upton, 1972) and the Ilímaussaq complex, dated at 1168 ± 21 m.y. (Blaxland et al., 1976).

The rocks studied are thus bracketed by the ages 1175 ± 9 and 1168 ± 21 m.y., and an age of 1170 m.y. is used for initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio calculation.

Five anorthosites and five gabbros were selected for study (fig. 6.1). They were analysed according to methods outlined in appendix B. Blanks measured at the time of analysis were 7 ng Rb and 4 ng Sr; the error in initial ratio introduced by this Rb blank is always less than 2% of the quoted overall error, so that no blank correction is made.

If the analyses (see section C.28) are regressed using errors of 0.7% on $^{87}\text{Rb}/^{86}\text{Sr}$ and 0.0035% on $^{87}\text{Sr}/^{86}\text{Sr}$ (the latter figure is the mean of in-run 1σ errors), then an age of 1150 ± 142 m.y. is obtained (fig. 6.2), with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70297 ± 0.00005 (errors at 2σ). $[\text{SUMS}/(n-2)]^{1/2} = 2.03$, indicating an isochron relationship (Brooks et al., 1972). This age agrees with the true age of the dyke at ca. 1170 m.y.

Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios can be calculated using the 1170 m.y. age, and these are shown in table 6.1.

Table 6.1 Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of anorthosites
and gabbros from south Greenland

Sample	$(^{87}\text{Sr}/^{86}\text{Sr})_0$	Sample	$(^{87}\text{Sr}/^{86}\text{Sr})_0$
Anorthosites		Gabbros	
30752	0.70300 ± 5	30755	0.70298 ± 6
30753	0.70296 ± 6	40518	0.70303 ± 6
40522	0.70288 ± 6	50224	0.70292 ± 6
50220	0.70303 ± 7	81156	0.70299 ± 5
50221	0.70298 ± 5	85977 *	0.70289 ± 6
		85977 *	0.70296 ± 6
Mean 0.70297 ± 11		Mean 0.70297 ± 8	

* Duplicated analysis: average initial ratio, 0.70293 used in mean calculation.

The 2σ error on initial $^{87}\text{Sr}/^{86}\text{Sr}$ is calculated by adding a 1.4% (2σ) $^{87}\text{Rb}/^{86}\text{Sr}$ component to the in-run 2σ standard error on $^{87}\text{Sr}/^{86}\text{Sr}$.

The anorthosites have a mean initial ratio of 0.70297, identical with that of the gabbros. One possibility to explain this agreement is that the $^{87}\text{Sr}/^{86}\text{Sr}$ of the anorthosite was adjusted to that of the gabbro during dyke emplacement. This is considered unlikely because the anorthosites have undisturbed cumulate textures and show no signs of significant alteration. Accidental derivation of the xenoliths from Archaean anorthosite, the nearest occurrence of which is ca. 100 km to the WNW, can be ruled out because these occur as deformed enclaves in

gneisses, whereas the anorthosites of this study are cumulate rocks unaffected by any deformation. Texturally, the rocks have more in common with the Proterozoic anorthosites of Labrador and south Norway; apart from having higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7031 to 0.7059 according to Heath and Fairbairn, 1968) however, the latter are generally orthopyroxene-bearing (e.g. Anderson, 1968), whereas the anorthosites of this study contain olivine, a single high-Ca pyroxene phase, apatite and biotite suggesting an alkaline affinity.

The marked concordance of the anorthosite and gabbro initial ratio data strongly suggests that the two rock-types are closely related, though not necessarily crystallised from the same magma. This conclusion is supported by rare-earth element patterns in anorthosite, gabbro and syenogabbro of the giant dyke complex (Blaxland and Upton, in preparation). Experimental studies on two uncontaminated chills of the gabbro giant dykes (Upton, 1971 and unpublished data) indicate olivine as the liquidus phase from 1-10 kb, replaced by clinopyroxene at higher pressures; plagioclase crystallisation began some 10-50°C below the liquidus temperature. Consequently, cumulates having plagioclase as the first crystallising phase, such as the anorthosites, cannot have formed from the magma represented by the gabbro dyke chills. However, Upton (1971) suggested that extended clinopyroxene fractionation at pressures higher than 10 kb from magmas similar to those of the giant dyke chills could have yielded a highly aluminous liquid which at lower pressures would have plagioclase on the liquidus.

In this way magmas generated early in the Late Gardar magmatic phase (from ca. 1200 m.y. onwards), rising slowly into the lower crust and fractionating clinopyroxene could have been responsible for the genesis of magmas from which large volumes of anorthosite crystallised at shallower

levels. These anorthosites were disrupted and carried upwards by subsequent batches of magma to appear both in the gabbro giant dykes and in various younger intrusions. The common failure of more felsic magmas to carry anorthosite can be attributed either to their insufficient density or to their generation by low-pressure fractionation above the level of anorthosite accumulation.

This hypothesis and others similar to it hold that plagioclase fractionation at depth was an important means of differentiation of Gardar magmas (Upton, 1964; Bridgwater, 1967; Bridgwater and Harry, 1968; Upton, 1971, 1974). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio data suggest a closely related origin for xenoliths of cumulate anorthosite and their host gabbros and thus lend considerable support to this argument.

6.2 Introduction to studies on dolerites from Sweden: previous work on anomalous initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in continental magmatic rocks

During the course of the work on Swedish dolerites over 100 whole-rocks were analysed, many of them specifically for the purpose of determining initial $^{87}\text{Sr}/^{86}\text{Sr}$ (or $(^{87}\text{Sr}/^{86}\text{Sr})_0$) ratios, and investigating the results of any interaction of basaltic magma with continental crust.

Basalts of oceanic regions show a restricted range of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios from 0.702 to 0.706, and these variations are in general ascribed to causes operating in the source region (e.g. O'Nions and Pankhurst, 1974). In continental regions however, there is present a 15-35 km thickness of dioritic, granodioritic, granitic, metamorphic and sedimentary rocks through which all magmas must pass en route to the upper crust or the surface. Since these rocks are almost invariably considerably older than the emplaced magmas and have higher Rb/Sr ratios,

they introduce the possibility of contamination of the magma by radiogenically-enriched strontium. This is in fact reflected in the much greater range of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios observed in continental than in oceanic magmatic rocks, though even in continental areas a high proportion of basaltic rocks still have $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values consistent with an uncontaminated mantle origin.

Fig. 6.3 shows schematic evolution paths of a basaltic magma in $^{87}\text{Sr}/^{86}\text{Sr}$ -- Rb/Sr space. If the magma changes composition (Rb/Sr increases) by differentiation alone, then no contamination by radiogenic strontium takes place, and the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ of all members of the magmatic series is the same as that of the starting magma. This is the probable evolution of a magma emplaced into oceanic crust. The opposite end member to this process is contamination by radiogenic strontium alone, so that the bulk composition of the magma (including Rb/Sr) remains essentially unchanged. This type of contamination has probably occurred for rocks of continental regions where an increase of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ is not accompanied by any concomitant change in bulk chemistry or trend of differentiation, but it is not commonly invoked. Between these two extremes are a whole range of possibilities which can be due either to varying combinations of differentiation and concurrent ^{87}Sr contamination, or more likely to contamination of the magma by material higher both in $^{87}\text{Sr}/^{86}\text{Sr}$ and Rb/Sr . Such added material, irrespective of the method of addition must be both older and more evolved chemically than the basaltic magma: most crustal materials have these characteristics.

Pushkar et al. (1972) have derived equations for contamination of magmas:-

$$\left(\frac{Sr_m}{Sr_m + xSr_c} \right) \cdot (^{87}Sr/^{86}Sr)_m + \left(\frac{xSr_c}{Sr_m + xSr_c} \right) \cdot (^{87}Sr/^{86}Sr)_c = (^{87}Sr/^{86}Sr)_{cm} \quad 6.1$$

$$\left(\frac{1}{1+x} \right) \cdot Sr_m + \left(\frac{x}{1+x} \right) \cdot Sr_c = Sr_{cm} \quad 6.2$$

where:

Sr_m	= Sr concentration of parent magma
Sr_c	= Sr concentration of contaminant
Sr_{cm}	= Sr concentration of contaminated magma
$(^{87}Sr/^{86}Sr)_m$	= isotopic ratio of parent magma
$(^{87}Sr/^{86}Sr)_c$	= isotopic ratio of contaminant
$(^{87}Sr/^{86}Sr)_{cm}$	= isotopic ratio of contaminated magma
x	= parts by weight of contaminant assimilated by 1 part of magma

Pushkar et al. (1972) applied these equations to the genesis of andesites and rhyolites, and Faure et al. (1974) modelled contamination of successive basalt lavas by granitic crustal material. They showed that all compositional differences between flows could be explained by contamination processes, and that no differentiation was necessary.

In cases where differentiation also takes place however, the Sr content of the magma cannot be modelled only in terms of mixing. The path of a magma in fig. 6.3 will be made up of a contamination component, leading to increase of both Rb/Sr and $^{87}Sr/^{86}Sr$, and a differentiation component, leading to increase of Rb/Sr alone.

Given knowledge of the start- and end-point magmas, these considerations generate two mathematical possibilities:-

1. 'Closed-system contamination': no differentiation takes place, and equations 6.1 and 6.2 both apply. Unknowns are then $(^{87}\text{Sr}/^{86}\text{Sr})_c$ and Sr_c , and the solution is a curve in these parameters with a unique value of x for each point on the curve (see Pushkar et al., fig. 3 for an example).
2. 'Open-system contamination': differentiation takes place as well as contamination, and only equation 6.1 applies. Unknowns are then $(^{87}\text{Sr}/^{86}\text{Sr})_c$, Sr_c and x , and the solution is an infinite series of curves in the $(^{87}\text{Sr}/^{86}\text{Sr})_c$ -- Sr_c plane with each curve having its own value of x (see fig. 6.14 for an example).

In the closed-system model, crustal rocks and other materials can be considered as possible contaminants if they lie on the curve and if the corresponding value of x is geologically reasonable. Additionally, if the sampled magmatic rocks show a continuous trend of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ with Sr content, then it is possible to quantify the composition of the contaminant by assuming reasonable values for $(^{87}\text{Sr}/^{86}\text{Sr})_c$ (Faure et al., 1974). In the open-system model however, almost any composition is a possible contaminant (see fig. 6.14), and the judgement of geological feasibility must be used to a greater extent. Heat considerations generally rule out higher values of x .

Early studies of anomalous $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios in magmas were made by Hamilton (1963) who found values up to 0.71 for late granophyres in the Skaergaard intrusion, and Manton (1968) who found a range from

0.704 to 0.712 for Jurassic basalts from Nuanetsi, southern Africa. In fact the widespread Jurassic basaltic rocks of the southern continents very commonly show anomalous $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios: Heier et al. (1965) reported a mean value of 0.712 for Tasmanian sills, and Compston et al. (1968) gave 0.711 for sills in Antarctica. These authors also studied K, U and Th abundances and concluded in favour of crustal contamination. To this work should be added the results of Faure et al. (1972) and Faure et al. (1974), who reported the value 0.711 for some Jurassic Antarctic basalts, though other basic rocks from Antarctica did give mantle-type initial ratios. Other studies of anomalous $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios have been made by Noble and Hedge (1969), who found vertical variations in single ash-flow sheets, explained as eruption of a magma chamber which had suffered varying degrees of contamination, and Ewart and Stipp (1968), who reported extensive work in Quaternary volcanics of New Zealand.

In all these studies the problem of the origin in detail of the contaminating material is unsolved, as is the question of the mechanism of addition. Bulk-rock assimilation of various basement rocks has generally been found unsatisfactory, largely on account of the high values of x required, and for this reason several of the authors cited have appealed to assimilation of partial melts of crustal rock. Pushkar et al. (1972) have shown that glass formed by partial melting of a granitic xenolith contained in basalt has $(^{87}\text{Sr}/^{86}\text{Sr})_0 = 0.723$, as against 0.706 for the total xenolith, and this lends support to the idea.

The more restricted problem of contamination of dolerite intrusions has been studied by Gates (1971) and Fratta and Shaw (1974), in addition to the earlier work on Jurassic sills (Heier et al., 1965; Compston et al., 1968). Gates (1971) observed mantle-type $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios in most major Precambrian dyke swarms of the Canadian Shield, but

superimposed on this was a variation in K, Rb and initial ratio dependent on the nature of the immediate country rock of the dyke and the distance of a particular sample from the contact. Fratta and Shaw (1974) observed variations in K, Rb, Li, B and Tl which were dependent upon the same controls. Both of these studies indicate that alkali contents are systematically higher where dolerite intrudes granite or syenite than where greenstone is the country rock. Additionally, the variation with distance from the contact suggests late-stage contamination, and Gates (1971) has concluded in favour of ion-exchange performed by deuteritic fluids, though Fratta and Shaw (1974) are less definite upon this question.

Variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ between minerals in a single whole-rock sample are detailed in section 6.3. Sections 6.4 and 6.5 deal with whole-rock variations within single intrusions and over the entire area of study in Sweden respectively. Sections 6.6-6.8 consider the origin of more evolved rock types, contamination mechanisms, feldspar compositional variations and the relevance of oxygen and hydrogen isotope studies. Section 6.9 deals with the effects of disturbance of mineral and whole-rock systematics long after emplacement of intrusions. General conclusions and a tentative model for the interaction of basaltic magma and continental crust are presented in section 6.10.

6.3 The variation of initial $^{87}\text{Sr}/^{86}\text{Sr}$ on a mineral scale

Many of the ages presented in this study are based upon mineral isochrons from dolerite samples. Since the ages are generally consistent with biotites and mineral isochrons from rheomorphic rocks, it is clear that mineral samples on these lines must have had identical or very similar $^{87}\text{Sr}/^{86}\text{Sr}$ ratios at the time of intrusion. This applies to

the great majority of the samples studied.

In rare cases however, discordances between mineral phases are observed. The large sill at Idre, northern Dalarna, intruded into 'Jotnian' arkosic sandstone (figs. 4.3, C.6, data section D.14) belongs to the ca. 1250 m.y. group of dolerites, and this age is given by two biotites from average rocks DN-17 and DN-21 (fig. 6.4). A sample DN-31 from a different part of the body carries significant sericite, chlorite and amphibole as well as very small amounts of interstitial granophyre and late interstitial epidote, in contrast to the average rocks. Anomalies are apparent in the analyses from DN-31. F1 and clinopyroxene both have $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values of ca. 0.704, while F3 through F5 show a trend of decreasing Rb/Sr ratio and increasing $(^{87}\text{Sr}/^{86}\text{Sr})_0$ up to 0.705. This trend of decreasing Rb/Sr is unique among the dolerites studied, and is shown to be related to sericitisation processes (section 6.8), such that most of the K (and Rb) in the rock entered sericite, while the alkali feldspar (F4, F5) is of albitic composition. Clinopyroxene containing 10% epidote has an initial ratio of ca. 0.709, indicating that the pure epidote must have a value of ca. 0.75.

Fig. 6.5 shows results from the essentially unaltered olivine dolerite of the Breven dyke of the East-West system (figs. 4.3, C.2, data section D.6). Here biotites from two rocks give ca. 1580 m.y., consistent with other ages from the swarm. Though the olivine dolerites have rather low Rb contents (less than 10 ppm), they do contain interstitial alkali feldspar, sometimes intergrown with quartz, and this feldspar occurs in zoned patches with late biotite, chlorite, serpentine and sericite in spaces between plagioclase crystals. These feldspars (F7) separated from two rocks define apparent ages of 2200-2400 m.y. Since the dyke cuts Svecofennian rocks, these ages can have no meaning, and at 1580 m.y. (the

known age of intrusion) the F7 feldspars must have had $^{87}\text{Sr}/^{86}\text{Sr} = 0.709$ as against ca. 0.704 for all other constituents of the rocks. The F7 feldspars are also remarkable in having Sr contents ca. 250 ppm higher than the F1 etc. feldspars, and this is the only case in all the dolerites studied for which an increase of Sr content with decreasing density is observed. X-ray diffraction studies on these and other feldspars show that in no case do the separates analysed from dolerites contain detectable amounts of feldspar alteration products such as sericite, zoisite, calcite etc. Therefore any anomalies in analyses of feldspar separates are due to feldspar and not to a content of secondary minerals.

These isotopic data show that late crystallising phases, including feldspar can be considerably contaminated with radiogenically-enriched strontium from outside sources. The geochemical implications of this inference, together with new compositional data on feldspars from some of the samples of interest, are discussed in section 6.8, but the principle of producing anomalously old feldspar ages by contamination during crystallisation may be applied to other less unequivocal examples where a dolerite mineral line is older at 2σ than either biotite, a rheomorphic mineral line or another dolerite mineral line. The dyke at Falun (fig. 4.5) shows a dolerite mineral isochron of 1129 ± 77 m.y., while biotite from the same sample gives 934 ± 13 m.y., in general agreement with another dolerite mineral isochron from the dyke of 987 ± 20 m.y. (see fig. 4.22). The sill at Forserum (fig. 4.5) gives two dolerite mineral isochrons of 980 ± 18 and 1071 ± 35 m.y. (see fig. 4.20). It might be suggested that the older mineral isochron in both these cases is caused by contamination of later-crystallising, higher Rb/Sr feldspar fractions with radiogenically-enriched Sr from outside the intrusion. These examples are less clearly identified than those of Breven and Idre, but whereas the

contaminated phases in the latter comprise much less than 1% of the rock, in the Falun and Forserum examples 10% or so of the feldspars are involved. The contaminated phases in the Breven and Idre examples are clearly associated with late hydrous alteration, and appear to be consistent with very late-stage contamination, but the Falun and Forserum data may suggest contamination while some 5% at least of the magma was still liquid.

The inference from the standpoint of dating is that younger dolerite mineral isochrons should generally be preferred in a case where two determinations differ at 2σ . There is no reason in principle why all dolerite mineral lines should not have been affected by contamination, though the agreement of results from one intrusion to another (see tables 4.1-4.3) is good evidence for the general reliability of ages.

6.4 The variation of whole-rock initial $^{87}\text{Sr}/^{86}\text{Sr}$ within a single intrusion

Almost all intrusions studied exhibit variations in $(^{87}\text{Sr}/^{86}\text{Sr})_0$: often these are unsystematic in detail, but in some cases patterns of initial ratio related to position of sample in the intrusion or to Rb/Sr ratio are observed.

The Karlshamn dyke of the Blekinge region (figs. 4.5, C.7, data section D.16) is ca. 250 m in width and consists of medium- to coarse-grained olivine hypersthene dolerite. Samples from the interior of the dyke, two average rocks (BL-3 and BL-15) and a pegmatitic sample (BL-18) have $(^{87}\text{Sr}/^{86}\text{Sr})_0 = \text{ca. } 0.705$ (fig. 6.6). The evolution of Rb/Sr ratio from the average rocks to the pegmatite without change in initial ratio illustrates the operation of differentiation. The remaining samples in fig. 6.6 were taken at progressively decreasing distances from the contact

of the dyke with granite, where the country rock has been remelted and visible mixing has taken place ca. 10 cm into the dolerite. These samples have suffered an increase in both Rb/Sr and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ as a result of assimilation of the older granite, and an apparent whole-rock age for the samples of ca. 1540 m.y. is obtained. The contact samples alone define an age of ca. 1600 m.y. These ages bear no relation to the intrusive age of the dyke, given by minerals at ca. 900 m.y.: the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios of the whole-rocks at this time range from 0.705 to 0.717.

A more subtle example of initial ratio variation is given by the sill at Forserum in the Almesåkra region (figs. 4.5, C.8, data section D.19), made up of hypersthene-bearing dolerite intruding Svecofennian metasediments and the overlying Almesåkra Formation. All the samples analysed are from a quarry ca. 25 m deep and 300 m across, which is small in relation to the volume of the sill. The sample AL-2 (fig. 6.7) is a chill against quartzite, and the remaining samples are essentially from random locations in the interior of the body. As with the Karlshamn dyke a whole-rock apparent age (ca. 1310 m.y.) is obtained which bears no relation to the real age of intrusion: the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values at 1000 m.y. range from 0.7054 to 0.7076.

The sill at Älvho in northern Dalarna (figs. 4.3, C.5, data section D.10) belongs to the ca. 1250 m.y. group of intrusions and consists of a fairly continuous variation from olivine dolerite to central monzonitic rocks. The analyses show the same pattern as at Forserum; samples with higher Rb/Sr have higher $(^{87}\text{Sr}/^{86}\text{Sr})_0$, up to 0.713 (fig. 6.8), and the apparent whole-rock age is ca. 1460 m.y. Fig. 6.9 demonstrates the nature of this variation: initial ratio increases with Rb concentration as well as with Rb/Sr ratio, and decreases with Sr concentration. Change of

initial ratio is therefore associated with change in both Rb and Sr: precisely similar diagrams could be given for the Forserum and Karlshamn examples.

These within-intrusion variations in $(^{87}\text{Sr}/^{86}\text{Sr})_0$ are very unlikely to have been produced at the source of the magmas because the intrusions are relatively small and consist of a single magmatic phase; since these small batches of magma must have ascended through at least 25 km (assuming mantle origin), it is likely that mechanical mixing would have effectively homogenised any source-related isotopic variation.

It is therefore necessary to invoke crustal contamination by the methods outlined in section 6.2. The Karlshamn example, where there is visible mixing of basic magma and crustal rock at a contact, illustrates the type of variation produced by straightforward hybridisation. The Forserum and Älvho examples are similar in the large variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ involved and in the small scale (less than 1 km) over which they are developed. This leads to the inference that the contamination occurred relatively late in the flow and emplacement of the intrusions, because it did not become homogenised throughout the magma. At Forserum the chill AL-2 has the lowest initial ratio and Rb/Sr ratio: although this sample has an Sr content ca. 1.5 times higher than samples from the interior of the sill and would therefore be less sensitive to contamination, this fact may nevertheless suggest that the contamination which has affected the interior parts of the sill took place after solidification of the chill, though whether this occurred during magma flow or when the sill was 'in place' is not clear. Gates (1971) and Fratta and Shaw (1974) have favoured 'in place' contamination, possibly by circulating fluids, but if the analysed chill at Forserum is representative of the other margins of the body, then contamination

slightly prior to emplacement of the interior of the sill seems likely. This would not imply a multiple-phase intrusion, but merely that the chill was solidified while magma, contaminated to varying degrees during flow, was still passing into the sill. Ålvho shows a similar picture: no chills were analysed here, but DN-316 and DN-325 are medium and coarse olivine dolerites respectively from near the margins of the sill. Internal samples are both more evolved (up to 64% SiO_2 , Lundqvist, 1968) and have higher $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios. Again, relatively late-stage contamination by older crustal material must be postulated in order to prevent homogenisation through the magma.

These variations therefore demonstrate variable contamination of magmas at a relatively late stage during flow and emplacement. They show that the whole-rock isochron method for dolerite bodies, even when sampling is confined to a small area, is likely to yield upwardly-biased ages and is not reliable as a general dating technique. The covariation of Rb/Sr and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ suggests that contamination processes may be a major factor in the evolution of basaltic magmas in a continental crustal environment, and this question will be discussed in section 6.6. Sections 6.7 and 6.8 deal with mechanisms of contamination.

6.5 The variation of whole-rock initial $^{87}\text{Sr}/^{86}\text{Sr}$ over the entire area of study

Fig. 6.10 is a compilation of results from 84 whole-rocks of the three important age groups ca. 1560, ca. 1250 and 1000-900 m.y. The initial ratios are listed in appendix E: the only samples omitted are certain rocks from the ca. 1560 m.y. group, believed to have suffered isotopic disturbance long after intrusion (see section 6.9): if shown, they would tend to plot at high initial ratios. Apart from the high

$(^{87}\text{Sr}/^{86}\text{Sr})_0$ values shown by the contact rocks and some monzonites, the majority of 'dolerites' define a spread of initial ratios which lies partly within and partly above the limits for the mantle source region. This gives rise to two generalisations: first, variations between dolerites proper over the whole Swedish area are much lower than those observed locally for associated more evolved rocks, and second, the initial ratios of dolerites are on average slightly higher than those allowable in uncontaminated mantle-derived rocks. This is true both for linear and non-linear mantle strontium evolution. The second of these relationships suggests widespread contamination processes, while the first implies either that only small amounts of contamination took place, or that any contamination was well homogenised through the magma; this may suggest deeper processes than for the within-intrusion variations of section 6.4.

All the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values can be standardised onto one age by correcting for mantle strontium evolution. Fig. 6.11 shows $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios corrected to an arbitrary value of 900 m.y., plotted against $^{87}\text{Rb}/^{86}\text{Sr}$ (ca. $3 \times \text{Rb}/\text{Sr}$). The samples divide into three groups: contact rocks visibly contaminated at margins of intrusions, monzonites of Emådalen and Älvho, and dolerites with $^{87}\text{Rb}/^{86}\text{Sr}$ less than 0.65. The contact rocks illustrate the effects of country rock assimilation on Rb/Sr and $(^{87}\text{Sr}/^{86}\text{Sr})_0$, and thus correspond to the within-intrusion patterns of section 6.4.

Figs. 6.12 and 6.13 show plots of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ vs. Rb and Sr concentrations respectively. The samples subdivide in much the same way as for Rb/Sr, with contact rocks and monzonites occupying similar fields in all three diagrams. The 66 dolerites show a positive correlation of $^{87}\text{Rb}/^{86}\text{Sr}$ and Rb concentration with $(^{87}\text{Sr}/^{86}\text{Sr})_0$ where the

correlation coefficient $r = \text{ca. } 0.7$ in both cases. When initial ratio is plotted against Sr concentration however, $r = -0.18$ and indicates no correlation (at $n = 66$, there is significant correlation when r is ± 0.23). Therefore the correlation of initial ratio with Rb/Sr can be explained solely in terms of a correlation with Rb, while Sr varies independently.

Such a pattern could be produced in two basic ways: by partial melting or by crustal contamination. Partial melting processes in the mantle can yield liquids with a sympathetic variation between Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$, in simple terms because the $^{87}\text{Sr}/^{86}\text{Sr}$ of the source rock, or any part of the source rock depends upon its Rb/Sr ratio. Variations such as figs. 6.11 and 6.12 could thus be produced as a result of pooling of analyses from different regions and age groups with systematically different Rb/Sr and initial ratios. Apart from the fact that within-intrusion variations such as that at Forserum (section 6.4) are partly responsible for the trend of figs. 6.11 and 6.12, inspection will show that the two principal age groups, ca. 1250 and 1000-900 m.y. both show a similar positively correlated distribution to that of the entire composite sample. On this basis, standardisation of all results as a means of increasing n seems justified, and in view of the fact that the trend is defined both by regional and highly localised variations, its production in the mantle source region seems unlikely.

Crustal contamination thus seems the most likely explanation for the trend; the basaltic magmas assimilated crustal material by some means and moved along oblique paths in fig. 6.3. The contact rocks at the present level of exposure probably illustrate the type of change involved. The large number of samples might seem to offer an ideal opportunity for statistically-based modelling of mixing processes after the principles outlined in section 6.2. Such calculations however, require a knowledge

of initial and final $^{87}\text{Sr}/^{86}\text{Sr}$ and Sr concentration: since the Sr concentration of the dolerites varies from 200 to 700 ppm, and is uncorrelated with $(^{87}\text{Sr}/^{86}\text{Sr})_0$, no modelling is possible.

The wide variation of Sr concentration from 200 to 700 ppm could be due either to widespread plagioclase fractionation or to variations in the extent and type of mantle partial melting. There is field evidence of quite extensive plagioclase fractionation in the ca. 1560 m.y. dolerites, and certain of the 1000-900 m.y. group contain rare plagioclase megacrysts. The limited chemical data available for the ca. 1250 m.y. dolerites (Hjelmqvist, 1966; Lundqvist, 1968; Lundqvist and Samuelsson, 1973) suggest that as liquids they could be in equilibrium with olivines of the compositional range Fo_{65-80} , calculated according to the partition coefficient of Roeder and Emslie (1970). Since estimates of the composition of upper mantle olivine fall between Fo_{88} and Fo_{94} (e.g. Fo_{89} , Ringwood 1975 p.488), it seems likely that the ca. 1250 m.y. dolerites have suffered some degree of fractionation before being emplaced at the present level of the crust.

There is thus evidence that the dolerites of this study represent a range of compositions affected to varying degrees by crystal fractionation. The correlation of Rb with $(^{87}\text{Sr}/^{86}\text{Sr})_0$ on the other hand, has been shown to indicate widespread crustal contamination of the basaltic magmas. Bearing in mind the limitations of composite treatment of samples from a wide region and different age groups, it nevertheless seems a reasonable generalisation that the basaltic magmas have followed oblique paths in fig. 6.3, but that in addition fractionation has operated to increase Rb/Sr independently of $(^{87}\text{Sr}/^{86}\text{Sr})_0$. The trends for the 66 dolerites in figs. 6.11-6.13 therefore do not define the actual paths of the contamination processes, but are the generalised 'resultants' of

contamination and differentiation effects.

6.6 Contamination versus differentiation for the origin of more evolved rock types

More evolved rock types associated with dolerite dykes and sills fall basically into two classes:-

1. Granitic rheomorphs and monzonitic dolerites, granophyres etc. at the contacts of dolerite with country rock. These include the 'contact rocks' of figs. 6.11-6.13 and are clearly produced as a result of melting and hybridisation with the country rock.
2. Larger areas of monzonite and granophyre, often the central parts of intrusions.

Not included in these groups are the central series porphyritic basic rocks of the Hällefors dyke; Krokström (1936) has interpreted these as derived from olivine basalt magma by fractional crystallisation. Since the central series forms a coherent magmatic unit, and similar rocks occur in other dykes of the East-West system, an origin by fractional crystallisation seems entirely reasonable.

The rocks with which this section is concerned are the larger occurrences of monzonite and granophyre. Three of these have been studied: the granophyre of the Breven dyke is shown in section 6.9 to have suffered an isotopic disturbance long after intrusion and is accordingly not discussed here; the other two occurrences are the monzonites of Emådalen and Älvho (figs. 4.3, C.5). Hjelmqvist (1966) has given examples of local hybridisation and monzonite development, and has implied that

this mechanism may explain the larger occurrences as well.

The monzonites of Emådal are developed over at least 100 m measured across the trend of the dolerite sill (Hjelmqvist, 1966), and may have a marginal or a central position. They have Rb = ca. 100 ppm, Sr = ca. 250 ppm and contain up to 10% quartz. The 8 samples analysed have $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios from 0.7047 ± 0.0010 to 0.7064 ± 0.0008 , and form the cluster at lower initial ratio in the 'monzonite field' of figs. 6.11-6.13. These values are similar both to dolerites in general and to dolerite from the same traverse across the sill at Emådal (0.7046 ± 0.0006). The slight difference between some monzonites and the dolerite does not allow any significant contamination by 1700 m.y. old crustal rocks (all local units are of this age), and it is concluded that the monzonites must be derived largely by differentiation from basaltic magma.

The monzonites at Älvho form a local central facies of the dolerite sill: they are granophyric rocks with SiO_2 up to 64% (Lundqvist, 1968). The $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios have been discussed in section 6.4: they range from 0.7082 to 0.7124 and are the points scattered to higher initial ratios in the 'monzonite field' of figs. 6.11-6.13. These values are higher than both dolerites in general and the dolerites at Älvho. The 9 samples from the body, monzonites and dolerites, show a positive correlation of Rb/Sr with $(^{87}\text{Sr}/^{86}\text{Sr})_0$ (fig. 6.9), interpreted as being due to crustal contamination (section 6.4).

The contamination processes can be modelled for this case of variation within a single body, according to the equation 6.1 (section 6.2, from Pushkar et al., 1972). The starting and finishing materials are DN-325, olivine dolerite, Sr = 414 ppm, $(^{87}\text{Sr}/^{86}\text{Sr})_0 = 0.7037$, and DN-328, monzonite, Sr = 206 ppm, $(^{87}\text{Sr}/^{86}\text{Sr})_0 = 0.7124$. These samples are those with the lowest and highest initial ratios, and have a margin-centre

relationship to one another (fig. C.5). Using these values, equation 6.1 becomes

$$(^{87}\text{Sr}/^{86}\text{Sr})_c = \frac{3.6018}{x\text{Sr}_c} + 0.7124 \quad 6.3$$

and is represented graphically in fig. 6.14. Six of the country rock acid volcanics from the immediate area (1669 $\pm$ 38 m.y.; Welin and Lundqvist, 1970) are plotted and correspond to values of x between 0.6 and 2.0. These values of x from equation 6.3 are shown in table 6.2, together with the concentrations of Rb, Sr and SiO₂ that would be produced by addition of x parts country rock to 1 part dolerite in closed-system assimilation, according to equation 6.2 (section 6.2).

Table 6.2 Possible contaminants of Älvho
dolerite

Sample	x	Rb ppm	Sr ppm	SiO ₂ % *
67056	0.81	108	277	60.0
67066	1.87	121	250	63.7
67067	1.00	127	231	60.2
67068	0.94	133	241	59.8
67069	1.17	131	251	61.1
67070	0.67	132	263	57.9
Actual DN-328		161	206	63.9 **

* Calculated using an average of 13 country rocks (Lundqvist, 1968).

** SiO₂ content of a similar sample from the same locality (Lundqvist, 1968).

The figures indicate that in no case is the contamination allowable in a closed-system (no differentiation) situation according to initial ratios capable of raising the Rb content or reducing the Sr content of the dolerite to that of the monzonite. The same applies essentially to SiO₂, and other major-element oxides such as TiO₂ and MgO give the same conclusion. It will be shown in section 6.8 that in general partial melts of country rocks are more likely material for assimilation than whole-rocks, but incorporation of any material having Svecofennian or older Rb-Sr systematics (i.e. older than 1700 m.y.) into the Älvho

dolerite leads essentially to the same result. Furthermore, since the acid volcanics are the youngest basement rocks in the area, the difficulty cannot be avoided by postulating assimilation of younger rocks with lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Partial melts of Svecofennian rocks will however, retain Svecofennian isotope systematics only if no isotopic equilibration with the residual material of the source rock takes place. Since the residual materials tend to be the low Rb/Sr, and hence low $^{87}\text{Sr}/^{86}\text{Sr}$ constituents, the effect of such equilibration would be to reduce the $^{87}\text{Sr}/^{86}\text{Sr}$ of the melt. In these circumstances it would still be possible to achieve the whole-rock variation observed at Älvho by contamination alone.

Thus while field evidence indicates local hybridism of dolerite and crustal rock to produce intermediates, the monzonites at Emådalen can be shown to be largely the result of differentiation processes. Similar rocks at Älvho could have been produced entirely by contamination of dolerite magma with partial melts of crustal rock that had equilibrated isotopically with their source residuum, and differentiation processes are essential only if this equilibration had failed to occur. A partial origin for the monzonites by differentiation is however favoured by comparison with Emådalen, where the rocks are similar in petrography and mode of occurrence. This twofold origin for the evolved rock types corresponds to that inferred for the more limited variation of the dolerites in section 6.5.

6.7 Contamination mechanisms

Sections 6.3 to 6.6 have demonstrated that contamination of basic magma by material of crustal origin has taken place throughout the region of study and probably in all three major age groups. This

conclusion has made no implications as to the mechanism of addition of the crustal material.

Three distinct methods for contamination of the basic magmas can be proposed:-

1. Bulk assimilation of crustal rocks, probably involving assimilation of xenoliths contained in the magma.
2. Assimilation of partial melts (rheomorphs) of crustal rocks.
3. Addition of any selected materials, e.g. alkalis plus radiogenically-enriched strontium, by aqueous fluids circulating through both crustal rock and dolerite.

All the processes are essentially compatible with the Sr isotope data; however, section 6.8 presents evidence on their relative importance.

6.8 Evaluation of contamination mechanisms: field evidence, theoretical considerations, compositions of feldspars and oxygen and hydrogen isotope studies

Contamination of whole-rocks

Bulk rock assimilation provides a means of adding to the basic magma material richer in SiO_2 , Rb, ^{87}Sr etc., and could therefore produce all the variations ascribed to contamination on a whole-rock scale. If this process had occurred to any extent, then xenoliths in various stages of assimilation might be expected to occur. One swarm of dykes of the 1000-900 m.y. group does contain quartzite and arkose xenoliths,

which are very abundant in one particular dyke, but no xenoliths of any kind have been observed by the author in any other dolerite bodies, nor are they reported in the literature. This amounts to a powerful argument against assimilation of whole-rocks.

On the other hand, where dolerite has intruded granites or gneisses, partial melting of the country rock has very often occurred, and such partial melts usually penetrate the dolerite for up to 10 m from the contact; rarely, rheomorphic veins are observed 20 m or more inside the intrusions. Those seen by the author vary in width from millimetres to 50 cm and are usually composed largely of granophyric intergrowths with little or no mafic constituents: they probably approximate ternary minimum melts in the system quartz-albite-orthoclase and thus have an essentially rhyolitic composition (see Barker, 1970).

The computations of Jaeger (1957) show that a stationary sheet of basaltic magma 100 m thick, intruded well below the surface and initially at its liquidus temperature of ca. 1100°C will raise the temperature in the adjacent 5 m of country rock (initially at 0°C) rather suddenly to between 670 and 600°C and hold it there for longer than 100 years. The case of a stationary sheet could apply to high crustal levels, but at greater depths it is likely that flow of magma through a conduit or fissure would raise the country rock temperature still further, and temperatures in excess of 800°C do not seem inconceivable. The data of Tuttle and Bowen (1958) for the system quartz-albite-orthoclase show that ternary minimum melts appear at 780°C for $P_{\text{H}_2\text{O}} = 0.5 \text{ kb}$ (ca. 1.5 km depth), falling to 660°C for $P_{\text{H}_2\text{O}} = 3.0 \text{ kb}$ (ca. 9 km depth). Although the assumption of $P_{\text{H}_2\text{O}} = P_{\text{total}}$ may not be applicable to certain lower crustal rocks, the resulting increased solidus temperature will be offset by the higher

initial temperature of the wall rocks at this depth. It therefore seems likely on theoretical grounds that partial melting of country rocks by heat from basic magma will be a common phenomenon over a large range of crustal depths.

A ternary minimum melt of crustal rock will contain SiO_2 , Rb, ^{87}Sr etc. in greater concentration than the basic magma. Additionally, the work of Pushkar et al. (1972) on a partially melted xenolith (see section 6.2) has shown that the melt phase can be enriched in ^{87}Sr above the level of the crustal source rock, and this gives a greater contaminating potential to such liquids.

Mixing of dolerite and rheomorphic material is observed in some instances in this work, and the 'contact rocks' (i.e. hybridised dolerites) of figs. 6.11-6.13 are examples of this effect. Mixing of rheomorph with dolerite could take place with relative ease in a flowing magma, and assimilation at depth might be quickly homogenised by mechanical means, so that no trace of the contaminating material would remain to be seen at the present level of exposure. There is no geological evidence contradicting the suggestion that partial melting and assimilation is likely to have taken place at most crustal levels.

Considerations of temperature at contacts, ease of assimilation and field evidence as exposed today thus show that rheomorph assimilation is more likely than whole-rock assimilation as a method for contamination of basaltic magmas on a whole-rock scale. Furthermore, since rheomorphs are always of a very evolved (rhyolitic) composition, lower values of x are required to achieve any given effect on the basaltic magma than would generally be necessary for whole-rock assimilation. This scheme, with its implication that given sufficient time, partial melting of the wall rock is inevitable, carries the corollary that the more

contaminated magmas may have moved slowly through the crust, while essentially uncontaminated magmas were translated rapidly from mantle to upper crustal levels. Additionally, magmas passing through feeder conduits or fissures would be expected to suffer more contamination than those emplaced rapidly as single phase ('flash') intrusions.

Contamination of minerals

Assimilation processes, whether of whole-rocks or partial melts are adequate to explain all variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ on a whole-rock scale. It is however possible that some ^{87}Sr and alkali contamination is to be explained by addition from fluids circulating through the dolerite and adjacent crustal rocks. Additionally, the isotopic data of section 6.3 show that late-crystallising phases in dolerites can have much higher $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios than early-crystallised phases, and argue for contamination occurring at a late stage in the solidification of the rock. It seems inherently unlikely that a silicate liquid, trapped in a 99% solid medium, should be able to transmit contamination from sample to sample, and more mobile fluids seem to be necessary. In this regard the association between late feldspars, with or without quartz, and secondary hydrous phases may be fundamental to the understanding of these late-stage effects. Late aqueous fluids, causing alteration of olivine and pyroxene to serpentine/chlorite/amphibole and plagioclase to sericite could transmit contamination through an intrusion with comparative ease.

Most dolerites contain some hydrous alteration products, and in the Breven olivine dolerites, the late feldspar and quartz are closely associated with hydrous minerals. The Idre sample DN-31 offers useful evidence on late-stage processes. Here the late feldspar intergrown

with quartz consists entirely of albite (fig. 6.15d, analyses 38,39) in contrast to the Na + K feldspar of all other samples studied: most of the potassium in the rock has been concentrated in sericitisation of the labradorite (see analyses 35,36,37, appendix F). This apparent partition between the alkali elements could have two explanations: either the sericite formed at the same time as the albite, or the feldspar initially crystallised as an Na-K composition, and the sericite was produced at a later time when removal of potassium from the feldspar also took place. If the albite and the sericite were crystallised at the same time, then the P-T conditions must have been such that muscovite was stable. The process of sericitisation is chemically complex, involving transfer of components such as K and Ca, but an upper stability limit for sericite lying somewhere in the general region of the muscovite curve (fig. 6.16; Evans, 1965) is probably a reasonable assumption. Fig. 6.16 also gives the upper stability limits for muscovite + quartz (Evans, 1965) and epidote at the fO_2 of the haematite-magnetite and nickel-nickel oxide buffers (Licu, 1973), which are probably the two fO_2 levels most likely to correspond to the geological situation (see Eugster and Wones, 1962). The minimum melting curve for granite (Tuttle and Bowen, 1958) is also shown. This melting curve applies to granophyres in general, which are of true granitic composition (Barker, 1970), and therefore also to the interstitial granophyres of this study, though it would not apply specifically to the quartz-albite intergrowths of DN-31. Coexistence of sericite and interstitial granophyre is observed in many dolerites besides those of Idre and Breven, and if these two phases crystallised contemporaneously, then a fluid pressure higher than 2 kb is implied (fig. 6.16). This would not necessarily mean that the dolerites solidified at 6 km depth

or more, but merely that when most of the intrusion was solid, high fluid pressures were attained in the remaining spaces. The most evolved residue of the magma crystallised in these spaces as granophyre or feldspar alone, and the high fluid pressure caused hydrous alteration of previously-crystallised phases. If the hydrothermal fluids had equilibrated isotopically with crustal rocks, and high but erratic fluid pressure in and around the dolerite intrusions would assist this, then ^{87}Sr contamination of both the hydrous minerals and the late feldspars would be expected. Almost all observed country rocks, including the arkosic sandstone at Idre, would be able to supply the necessary radiogenically-enriched Sr.

Microprobe examination of feldspars in rocks showing more normal isotopic compositions in feldspars and other minerals, such as EL-3 from Karlshamn and AL-12 from Forserum (fig. 6.15b and c) show a complete range of plagioclase compositions from labradorite to oligoclase, at which point there is an apparent gap over the anorthoclase range before the more K-rich alkali feldspars. Exsolution of alkali feldspar grains is however, rather common (e.g. analyses 24 and 25, fig. 6.16b, are exsolved albite and orthoclase from the same grain) and the apparent gap in compositions is probably of little significance in terms of crystallisation history. The feldspars from the contaminated rocks SÖ-3 and SÖ-4 show a similar range of alkali feldspars, probably partly due to exsolution (e.g. the albites 55, 56 and 59, fig. 6.15e); the plagioclase compositions however, are in contrast to those of 'normal' rocks in that there is no continuous zoning towards the alkali feldspars, and inter-grain boundaries between plagioclase and alkali feldspar tend to be extremely abrupt. This compositional gap exists despite deliberate searching for intermediate compositions, and

furthermore cannot be due to exsolution phenomena since andesine and oligoclase compositions do not exsolve. The relationship suggests that the fluids from which the alkali feldspars of SÖ-3 and SÖ-4 crystallised were not in equilibrium with the remainder of the rock, and this is consistent with the ^{87}Sr contamination which they display. The alkali feldspars of these two samples are unique in this study because they have higher Sr contents than the main plagioclases of the samples, in opposition to normal differentiation trends (see section 6.3): this could also be interpreted as evidence that their composition was controlled from outside the dyke via the medium of hydrothermal fluids.

A pilot study was undertaken to investigate whether such contamination processes involving hydrous fluids could be recognised on the basis of δD and $\delta^{18}\text{O}$ values. This should in principle be possible whenever the fluids involved are not of primary magmatic origin.

Hornblende DN-304 from monzonite at Emådalen (fig. C.5), a dolerite-monzonite sill emplaced into rhyolitic lavas gives a δD value of -102.3 ± 0.8 (2 analyses), slightly more negative than most igneous hornblendes which fall between -50 and -85 (Taylor, 1974). This may suggest an influence of water foreign to the magma, but the difference is not large enough to be unequivocal, particularly as hornblende D/H fractionations vary with Fe/Mg ratio, and the composition of this hornblende is not known.

$\delta^{18}\text{O}$ data for two feldspars from the sample SÖ-4 of the Breven dyke are given in table 6.3.

Table 6.3 $\delta^{18}\text{O}$ data for standards and feldspars

Sample	Anal.1	Anal.2	Mean $\delta^{18}\text{O}$
NBS-28	+ 9.6	-	+ 9.6 (+ 9.6)*
VQS-2	+13.0	+13.4	+13.2 $\pm$ 0.2 (+13.4)*
SÖ-4 F1	+ 6.9	+ 6.5	+ 6.7 $\pm$ 0.2
SÖ-4 F7	+ 7.2	+ 7.2	+ 7.2 $\pm$ 0.0

* Accepted standard values in parentheses

The feldspar F7 of SÖ-4 is one of those defining an anomalous age (see section 6.3, fig. 6.5). F1 is a calcic labradorite with a normal $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio, while F7 is a late interstitial Na to K feldspar on X-ray diffraction and microprobe studies. Whether F7 is more or less potassic does not affect the $^{18}\text{O}/^{16}\text{O}$ fractionation, which depends only on Si/Al ratio (O'Neil and Taylor, 1967). The waters in equilibrium F1 and F7, calculated by the curves of O'Neil and Taylor (1967), are indistinguishable within the experimental errors at any one temperature. This means that assuming equilibrium, both feldspars could have crystallised from, or equilibrated with, the same source if their temperatures of formation differed by less than ca. 100°C. However, the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio data indicate that the fluids from which F7 crystallised had become contaminated by ^{87}Sr from the country rock granites: this contamination occurred after crystallisation of F1, and the fluids did not equilibrate with F1. The oxygen isotope data do not contradict this inference because the waters in equilibrium with granite at subsolidus temperatures could very easily be indistinct from those in

equilibrium with F1 and F7. On this hypothesis, the agreement in $\delta^{18}\text{O}$ between F1 and F7 would be fortuitous, since F1 would be of magmatic origin, while the isotopic composition of F7 would be controlled by that of the country rock. Additionally, palaeomagnetic data (Bylund and Patchett, unpub. data) indicate a rather low latitude (less than 20°) for the Baltic Shield at the time of emplacement of this dyke, and this would tend to reduce the contrast between values of magmas and ground waters (see Taylor, 1974).

Thus although late-stage aqueous fluids carrying, among other constituents radiogenically-enriched Sr, are inferred on grounds of petrography and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios, they cannot be clearly recognised by means of hydrogen and oxygen isotopes in the samples studied.

6.9 Effects of a post-intrusive isotopic disturbance:

the East-West dyke system

Strong isotopic disturbance of mineral or whole-rock systems usually accompanies metamorphism. It is an implicit assumption in the age study on Swedish dolerites that since the minerals of the country rocks became closed isotopic systems at ca. 1700 m.y., and the dolerites show no signs of metamorphism, then the mineral ages must represent dolerite intrusion. This section illustrates the effects of a later mild isotopic disturbance on basic and granophyric rocks of the Breven and Hallefors dykes of the East-West system: this event is not reflected in any observable metamorphism.

The East-West dyke system has been shown in section 4.3 to be of ca. 1560 m.y. age (see fig. 4.3 for summary). At Breven, biotites from unaltered olivine dolerite give 1575 ± 23 and 1581 ± 23 m.y., and a

rheomorphic vein from Hällefors gives a 5-point feldspar isochron of 1551 ± 38 m.y.

Analysis of whole-rock samples from the porphyritic central series of the Hällefors dyke (fig. C.1) define a scatter on an isochron diagram (fig. 6.17), while minerals from a central series sample (SÖ-26) and a marginal series sample (SÖ-21) are also discordant (fig. 6.18). The lightest feldspar fraction of SÖ-21 defines an apparent age of 1670 m.y., but clinopyroxene lies considerably below the line and the age may therefore be biased upward and is to be regarded with scepticism (see section 6.3). Mineral separation of SÖ-26, the central series sample, was complicated by the fact that all feldspars with Rb/Sr greater than the whole-rock apparently had magnetic inclusions and were consequently discarded during magnetic cleaning of the feldspar fractions on a magnetic separator. Nevertheless a confused pattern is apparent in the isotopic analyses (fig. 6.18): the feldspar fractions and whole-rock are very poorly aligned along a ca. 900 m.y. line with an initial ratio of ca. 0.715, the clinopyroxene lying well below this feldspar trend.

The scattering of whole-rock analyses can readily be related to the extensive propylitic alteration which these rocks have suffered; this may have taken place at any time after dyke intrusion. The alignment of central series feldspars along ca. 900 m.y. suggests that there may have been a thermal or hydrothermal event long after dyke emplacement, and this could also explain the scattered whole-rock data. The whole-rock SÖ-31 (fig. 6.17) has an apparent initial ratio of less than 0.7 at 1550 m.y., and since the lowest possible terrestrial value at this time is ca. 0.701 (see Faure and Powell, 1972 p.26), then the present isotopic composition of SÖ-31 must have been acquired at a time considerably later than 1550 m.y. In fact, in order to have a 'possible'

$(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio SÖ-31 cannot have had its present isotopic composition for longer than ca. 1300 m.y., and so there must have been an isotopic disturbance after 1300 m.y. before present.

Analysis of whole-rock and mineral samples from the Breven dyke (fig. C.2) also define a complicated picture. Regression analysis of whole-rock data for 6 samples of the central granophyre gives an 'age' of 1531 ± 96 m.y., while inclusion of of basic and hybrid whole-rocks (14 points in total) leads to a 1524 ± 59 m.y. result (fig. 6.19). $[\text{SUMS}/(n-2)]^{1/2}$ values are 5.19 and 6.05 respectively and indicate significant discordance outside analytical error.

Feldspar-quartz fractions from four granophyres define six-point apparent isochrons which vary in age between 1422 and 1517 m.y., (fig. 6.20): the apparent age bears no systematic relationship either to geographical position or to grain-size of the granophyre sample. The extremes of these mineral 'ages' do not overlap at 2σ , and if all results are taken at face value, the Breven dyke was formed by at least three entirely separate intrusive events, one of dolerite (1580 m.y.) and two of granophyre (ca. 1510 and 1420 m.y.), the whole spanning some 150 m.y. It is believed however, that the Breven granophyre represents a coherent magmatic unit emplaced in a single intrusive event, and therefore alternative explanations for the differing mineral 'isochrons' must be sought. If the granophyre is assumed to be 1580 m.y. in age, as given by the dolerite biotites, and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios calculated accordingly, then the apparent initial ratios for the four samples are arranged in the same order as their apparent mineral isochron age (table 6.4); that is, the lower the apparent age, the lower the initial ratio at 1580 m.y.

Table 6.4 Initial ratios at 1580 m.y. and apparent
mineral isochron ages for Breven granophyres

Sample	Mineral	$[SUMS/(n-2)]^{1/2}$	$(^{87}Sr/^{86}Sr)_0$
	age m.y.		
SÖ-13	1422 $\pm$ 23	1.10	0.691 $\pm$ 0.005
SÖ- 8	1425 $\pm$ 40	2.69	0.696 $\pm$ 0.002
SÖ-18	1500 $\pm$ 19	0.93	0.701 $\pm$ 0.004
SÖ-12	1517 $\pm$ 26	2.14	0.707 $\pm$ 0.003

This relationship suggests that a mild thermal or hydrothermal event at a time considerably later than dyke emplacement could have caused loss of radiogenic ^{87}Sr from the granophyre whole-rocks (leading to lowered apparent initial ratios), and have effected a total or partial reequilibration of the mineral separates to younger ages.

Three granophyres and one hybrid rock have impossibly low $(^{87}Sr/^{86}Sr)_0$ ratios of less than 0.7 for 1580 m.y., and as is the case for the Hällefors dyke, these samples must have acquired their present isotopic composition at some time after inferred dyke emplacement at ca. 1580 m.y. In fact, for the three granophyres a reduction of the dyke emplacement age to 1530 m.y. would restore the initial ratios to values greater than 0.701, and so if the dyke was intruded as late as this there would be no anomaly in granophyre initial ratios. This is not the case, however, for the sample SÖ-10, a fresh hybrid whole-rock sample analysed in duplicate. This has an $(^{87}Sr/^{86}Sr)_0$ value of 0.694 at 1580 m.y., and its present isotopic

composition would have to have been acquired at least as late as 1350 m.y. before present in order for the initial ratio to have a 'possible' value of greater than 0.701. There is therefore evidence that the Breven dyke, as well as the Hällefors dyke has suffered an isotopic disturbance at least 200 m.y. after its intrusion.

The variation in mineral isochron ages of granophyres can be explained by consideration of the mineral separation process for these rocks. The separates F1 to F5 are a series of density fractions (see section B.3 for method) of grains 45-84 μ in diameter ground from granophyres consisting mainly of albite, quartz and granophyric intergrowths on a scale mostly finer than 45 μ . This size disparity means that there will be many composite grains of quartz and alkali feldspar in the 45-84 μ sample, and the density of these will vary over a range which includes that of albite. Consequently the separates F1 to F5 are related to each other by an increasing proportion of alkali feldspar to other constituents, and any trend on an isochron diagram will be governed by the alkali feldspar. If in addition, the alkali feldspar has suffered an incomplete isotopic disturbance at some time after crystallisation, then the above technique would be expected to yield a variety of mineral isochron ages which would vary with the degree of disturbance of the alkali feldspar and have no age significance whatsoever. This scheme is consistent with the data in as much as those rocks which have the most disturbed alkali feldspar have also lost the most ^{87}Sr . A δD value on hornblende from granophyre SÖ-12 of -97.1 ± 0.4 (2 analyses) is slightly outside the range of normal igneous hornblendes (-50 to -85, Taylor, 1974) and this supports an influence of waters from outside the dyke, possibly meteoric in origin. This hornblende has been analysed for Rb-Sr and falls well below the feldspar

line of SÖ-12, so that it has been more strongly affected by the isotopic disturbance than the alkali feldspar. Although the Fe/Mg ratio of the hornblende is not known, the δD result supports the idea of hydrothermal activity at the time of isotopic disturbance of the granophyres.

Since granophyre samples have probably lost radiogenic Sr, the whole-rock ages of ca. 1530 m.y. are to be regarded only as minimum estimates of the time of dyke emplacement, and this is consistent with an intrusion age of ca. 1580 m.y., as given by biotites. Additionally, the loss of ^{87}Sr from the granophyres obscures any relationships that might have shed light on the origin of the felsic magma.

Only the biotites of Breven and the rheomorph feldspar isochron of Hällefors agree with other determinations of the age of the East-West dyke system. All other data from whole-rocks and minerals are divergent and give no coherent results; this has been ascribed to a relatively mild thermal or hydrothermal disturbance later than dyke emplacement. On isotopic grounds this disturbance must have taken place after 1300 m.y., and the scattered mineral data from Hällefors may suggest a Sveconorwegian (ca. 900 m.y.) age. This age would be reasonable on geological grounds since the dykes were emplaced before any Sveconorwegian events, and lie only 60-100 km from the margin of the Sveconorwegian belt (see fig. 4.3). Evidence from further south however, suggests that Sveconorwegian reheating did not extend for much more than ca. 20 km into the Svecofennian crust: this apparent contradiction can be avoided to some extent by suggesting that the isotopic disturbance was effected by a hydrothermal event rather than a simple 'reheating'. The existence of dykes and fractures of the 'Sveconorwegian system' at 1000-900 m.y. in the area of the East-West

dykes is evidence for some activity at this time, and the Breven dyke is cut by a NW-trending fault probably of this age (see fig. C.2). The fact that biotites and certain other mineral systems have not been affected suggests that the resetting was very selective and localised in nature; a mild hydrothermal event seems the best way to explain the effects observed. The δD value of -97 for a hornblende affected by the isotopic disturbance lends some support to this idea.

The scattered patterns of whole-rocks and minerals produced in the Breven and Hållefors dykes by later isotopic disturbance can be taken as evidence that no disturbance has occurred in other groups where isotopic data are concordant.

6.10 Conclusions

The conclusions of studies on isotope geochemistry of basic igneous bodies in continental crust can be summarised:-

1. Anorthosite xenoliths and host gabbros from Tugtutôq, south Greenland have identical $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios to five figures, implying a related origin. This conclusion supports hypotheses of plagioclase fractionation as a major means of differentiation of magmas in the Gardar Province.
2. While many samples of dolerite dykes and sills in Sweden show concordant whole-rock and mineral $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values, there are important systematic variation patterns. These are inferred to be caused by contamination of basic magmas with material of crustal origin. Variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ are observed on a regional scale, within a single intrusion and between minerals of a single

sample: generally, the smaller the scale of observation, the larger the variation and the later in the emplacement and cooling history the contamination must have occurred.

3. Three possible contamination mechanisms are considered: bulk assimilation, partial melt assimilation and contamination by means of aqueous fluids. Bulk assimilation is isotopically possible but geologically unlikely because of the lack of any observed xenoliths in dolerite. Assimilation of partial melts of crustal rock is isotopically possible, requires smaller volumes of contaminate and is consistent with the frequent rheomorphism seen at the present level of exposure. Contamination from aqueous fluids is essential to explain variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ between minerals where the contaminated phase is less than 1% of the rock, and petrographic and geochemical studies show that late contaminated phases are frequently associated with hydrous alteration products, implying an influence of hydrothermal fluids. Such fluids cannot be unequivocally identified by means of oxygen and hydrogen isotope data.

4. These considerations lead to a model for interaction of basaltic magma and continental crust whereby ascending magmas assimilate local partial melts from the wall rocks. The amount of melting and assimilation depends upon the speed of movement of the magmas and the volume of magma passing through a conduit: magmas translated rapidly to the upper crust could escape contamination, whereas slow movement or prolonged flow through a fissure is likely to lead to a more or less contaminated magma. The relatively large variations of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ over short distances suggest either that the

magmas were not efficient at mixing themselves or more likely that assimilation was most significant at upper crustal levels: almost all bodies contain some samples with mantle-type initial ratios. Contamination of magmas by circulating fluids cannot be ruled out, and during solidification there is probably increased opportunity for this effect (possibly consequent upon rising fluid pressure with solidification), which can lead to anomalous $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios in certain late-stage minerals.

5. Post-intrusive disturbance of whole-rock and mineral systematics is only identified in one case, the East-West dykes, and the strongly divergent patterns produced suggest that such disturbance is absent from other intrusions giving concordant age results.

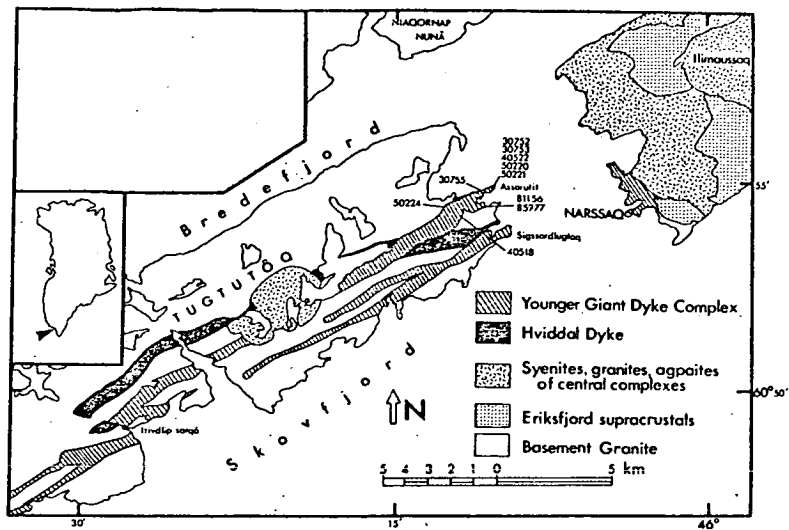


Fig. 6.1 Simplified geological map of the Tugtutôq-Narssaq area, south Greenland, showing localities for anorthosite and gabbro samples.

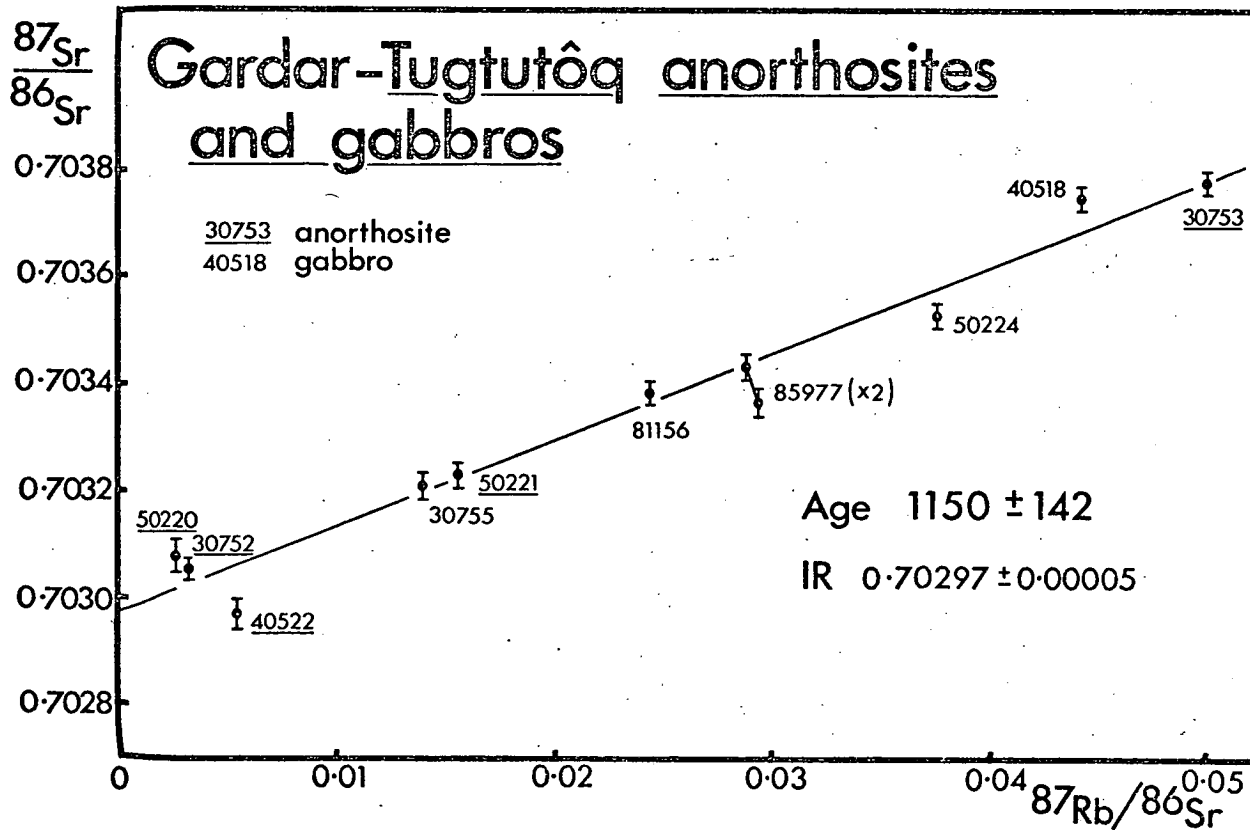


Fig. 6.2 Isochron diagram for results from the Tugtutôq anorthosites and gabbros, showing essential agreement of the low-range isochron age with the 'true' age of ca. 1170 m.y.

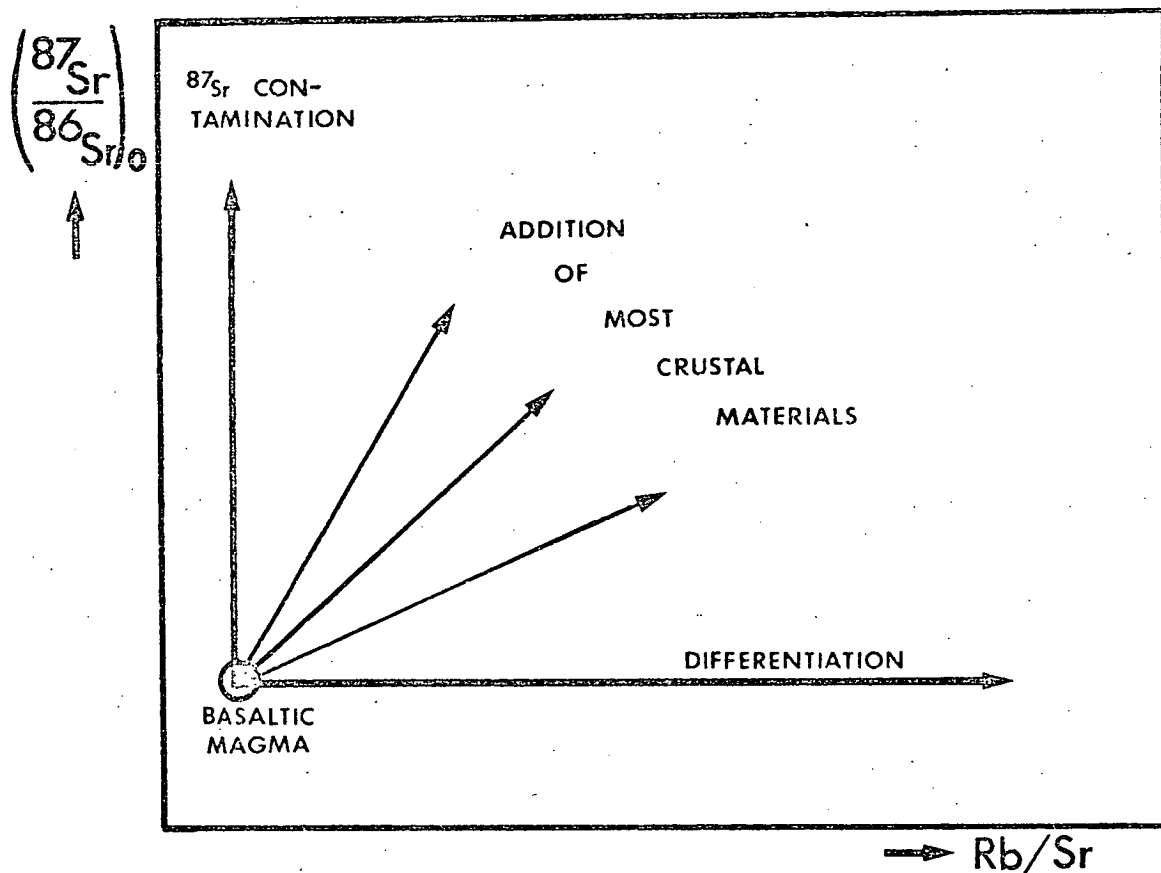


Fig. 6.3 Schematic evolution paths for basaltic magmas in a continental crustal environment. End-member processes are differentiation alone and ^{87}Sr contamination alone; all intermediates are possible, involving either of (1) a combination of differentiation and ^{87}Sr contamination, or more likely (2) addition of higher Rb/Sr, higher $^{87}\text{Sr}/^{86}\text{Sr}$ material.

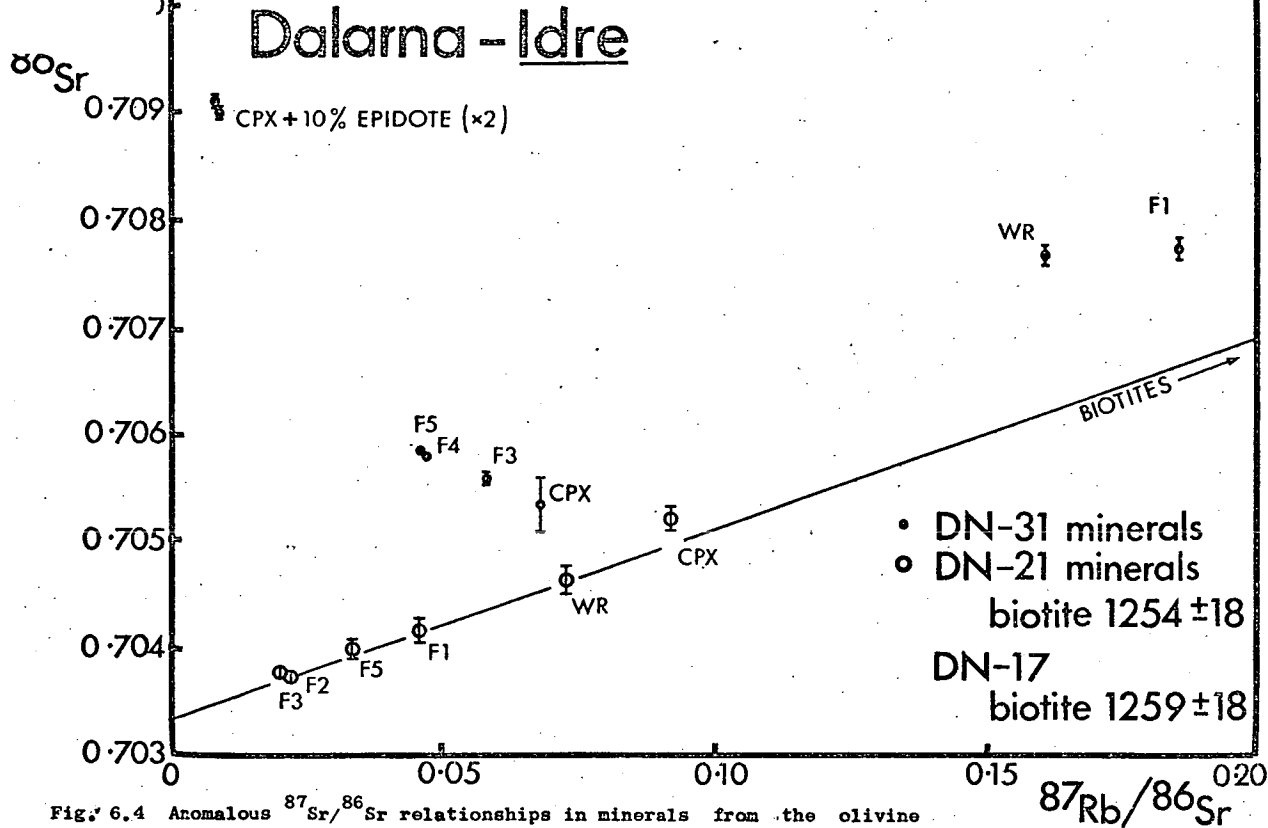


Fig. 6.4 Anomalous $^{87}\text{Sr}/^{86}\text{Sr}$ relationships in minerals from the olivine dolerite sill at Idre, northern Dalarna. For location see fig. 4.3. The intrusive age of ca. 1250 m.y. is defined by two biotites. Late feldspars F4 and F5 of DN-31 show a small increase in $(^{87}\text{Sr}/^{86}\text{Sr})_0$ over the plagioclase and clinopyroxene, and the analysis of CPX+10%EPIDOTE demonstrates an initial ratio of at least 0.75 for the late interstitial epidote.

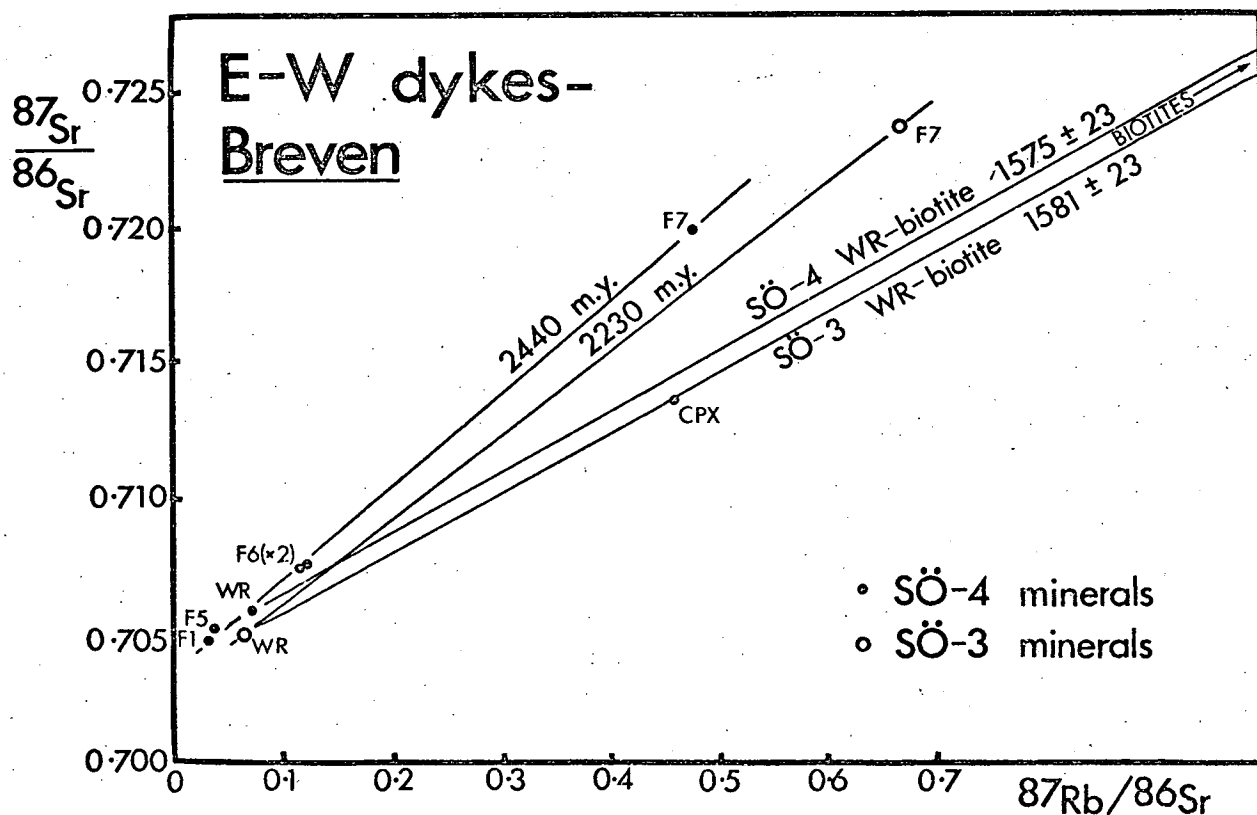


Fig. 6.5 Anomalous $^{87}\text{Sr}/^{86}\text{Sr}$ relationships in minerals from olivine dolerite of the Breven dyke, East-West system. For location see fig. 4.3. The ca. 1580 m.y. intrusion age is given by two biotites. Late alkali feldspars (F7) have $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios of up to 0.709, defining apparent ages of over 2000 m.y.

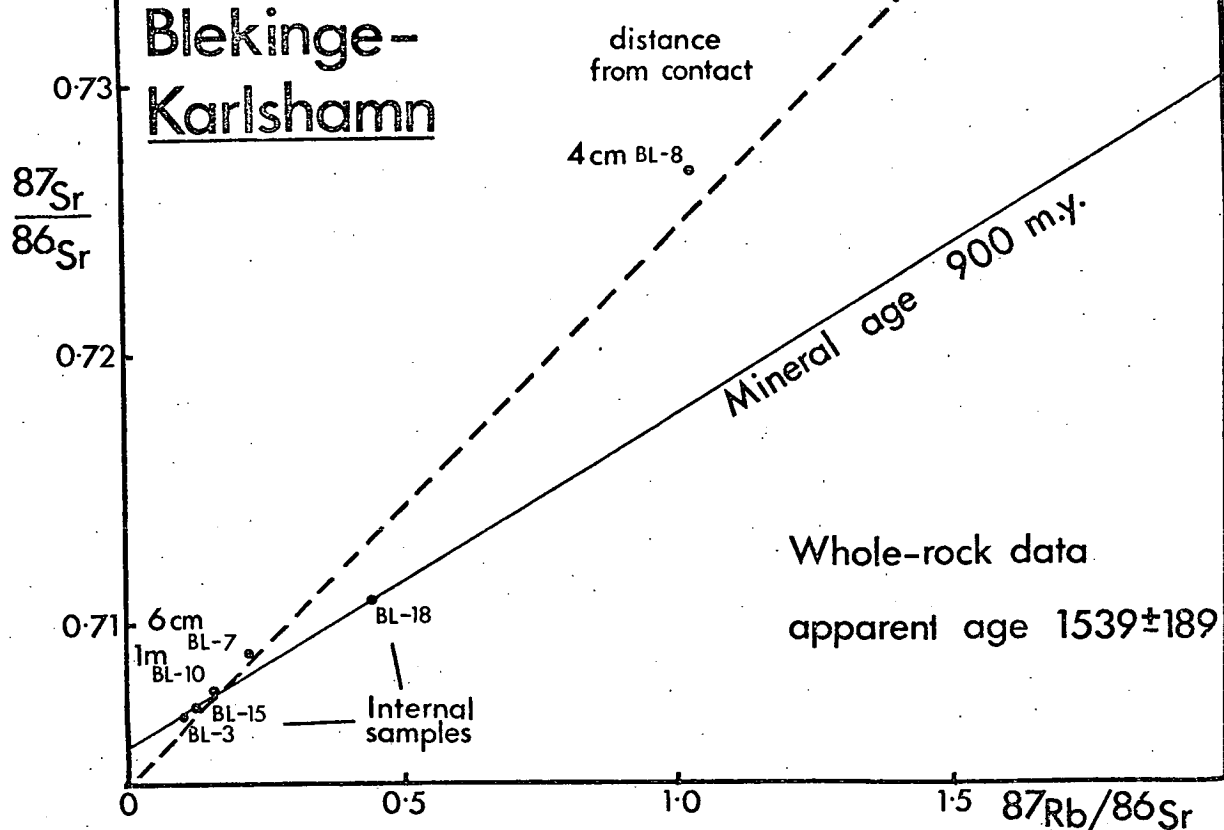


Fig. 6.6 Anomalous $^{87}\text{Sr}/^{86}\text{Sr}$ relationships in whole-rocks of the Karlshamn dyke, Blekinge. For location see fig. 4.5. Internal samples BL-3, BL-15 and BL-18 lie on the ca. 900 m.y. mineral line and have initial ratios of ca. 0.705, while samples taken at progressively decreasing distances from a contact with granite have $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios up to 0.716, defining an apparent age of 1500-1600 m.y.

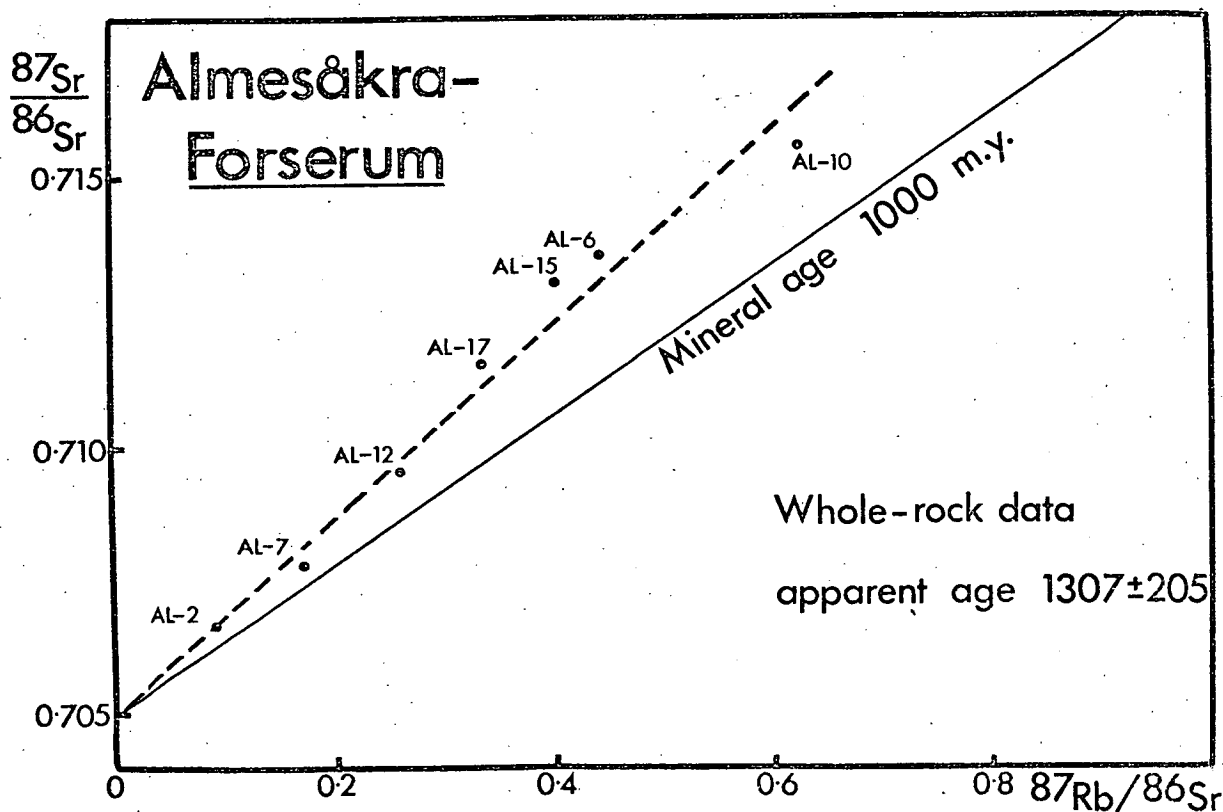


Fig. 6.7 Anomalous $^{87}\text{Sr}/^{86}\text{Sr}$ relationships in whole-rocks of the Forserum sill, Almesåkra region. For location see fig. 4.5. The age of ca. 1000 m.y. is given by mineral lines. The whole-rock sample AL-2 is a chill against quartzite, and the other samples are from random locations in the interior of the sill: the apparent whole-rock age is ca. 1300 m.y.

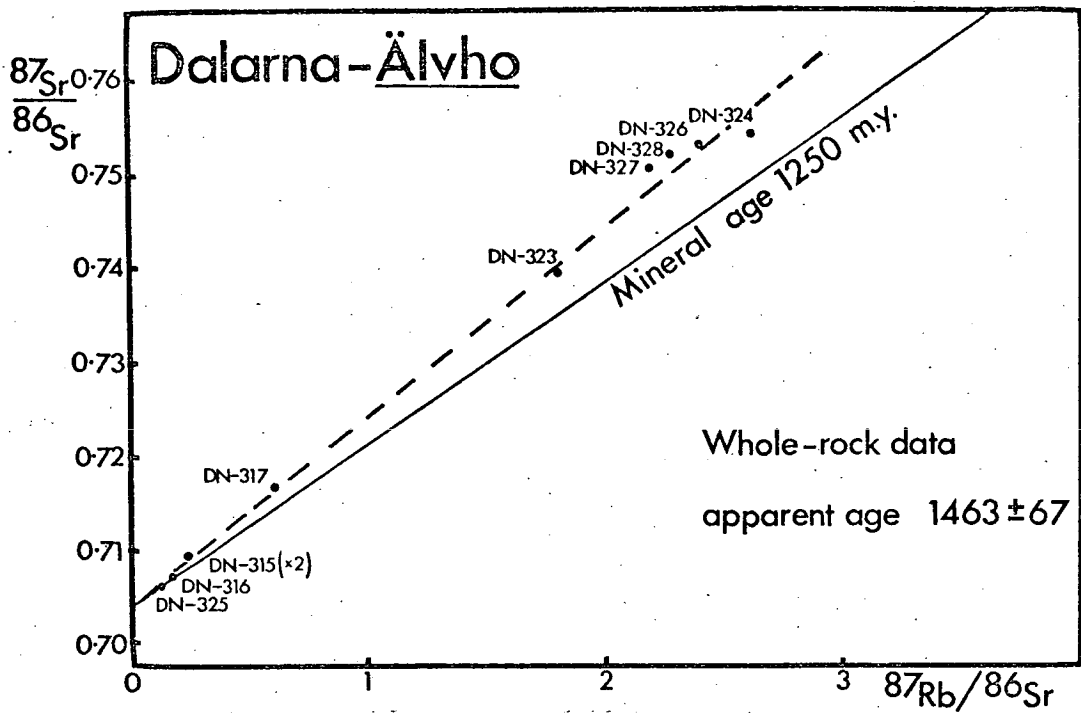


Fig. 6.8 Anomalous $\frac{87\text{Sr}}{86\text{Sr}}$ relationships in dolerite and monzonite whole-rocks of the Älvho sill, northern Dalarna. For location see fig. 4.3. The age of intrusion is given by minerals at ca. 1250 m.y., and the apparent whole-rock age is ca. 1460 m.y.

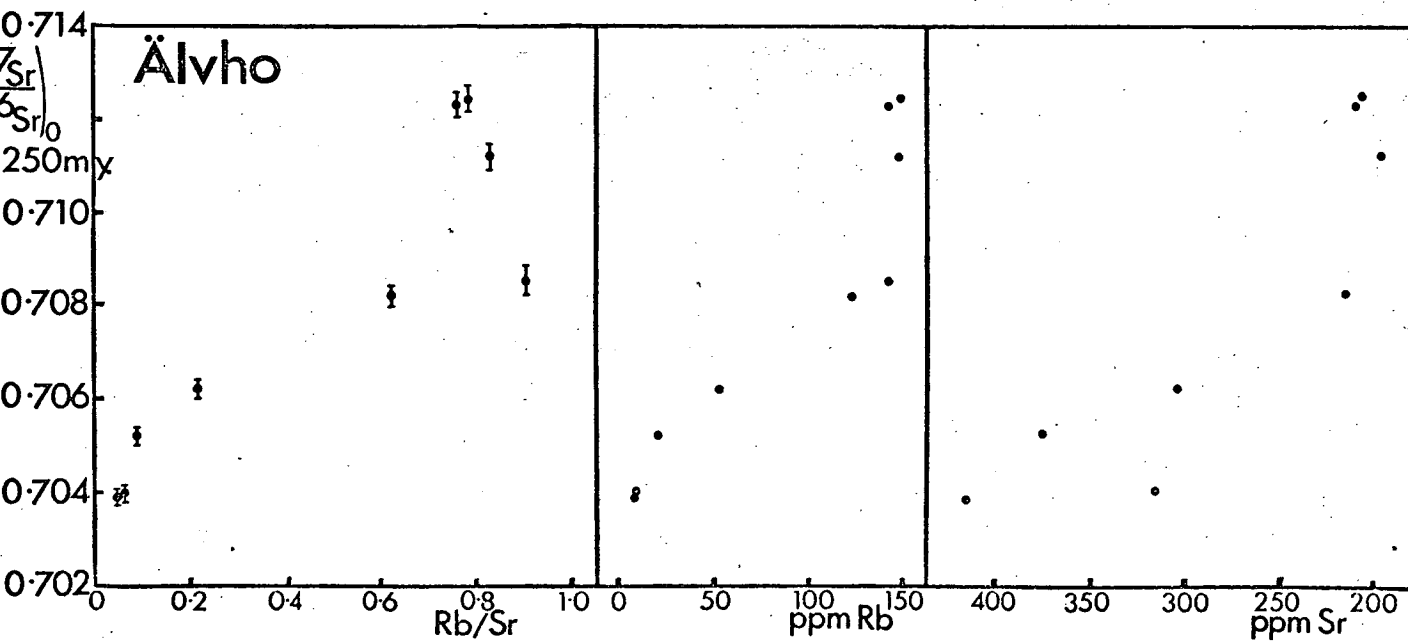


Fig. 6.9 Relationship of $\left(\frac{87\text{Sr}}{86\text{Sr}}\right)_0$ ratios (see appendix E) in whole-rocks of the Älvho sill to Rb/Sr, Rb concentration and Sr concentration. A correlation is observed with all three parameters.

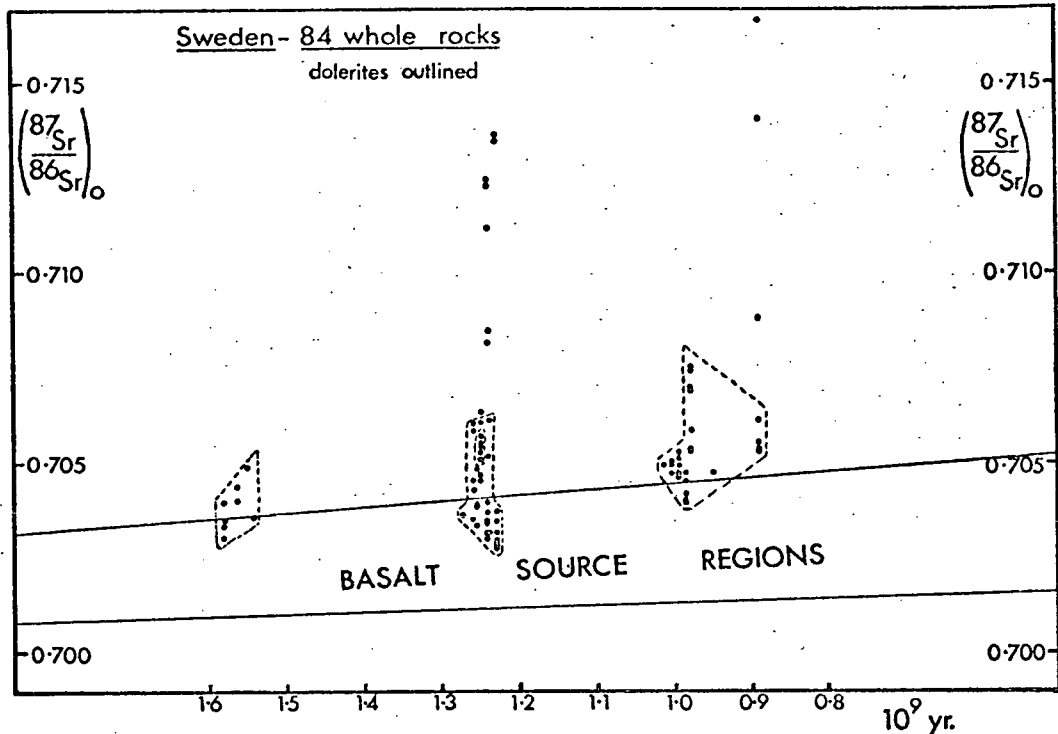


Fig. 6.10 $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios (see appendix E) for 84 dolerites and associated rocks from Sweden at the time of intrusion plotted relative to the range of possible $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the mantle source regions. The diagram assumes linear strontium evolution in the mantle (i.e. Rb/Sr does not change through time), but the relationships are almost unchanged if non-linear evolution is the case.

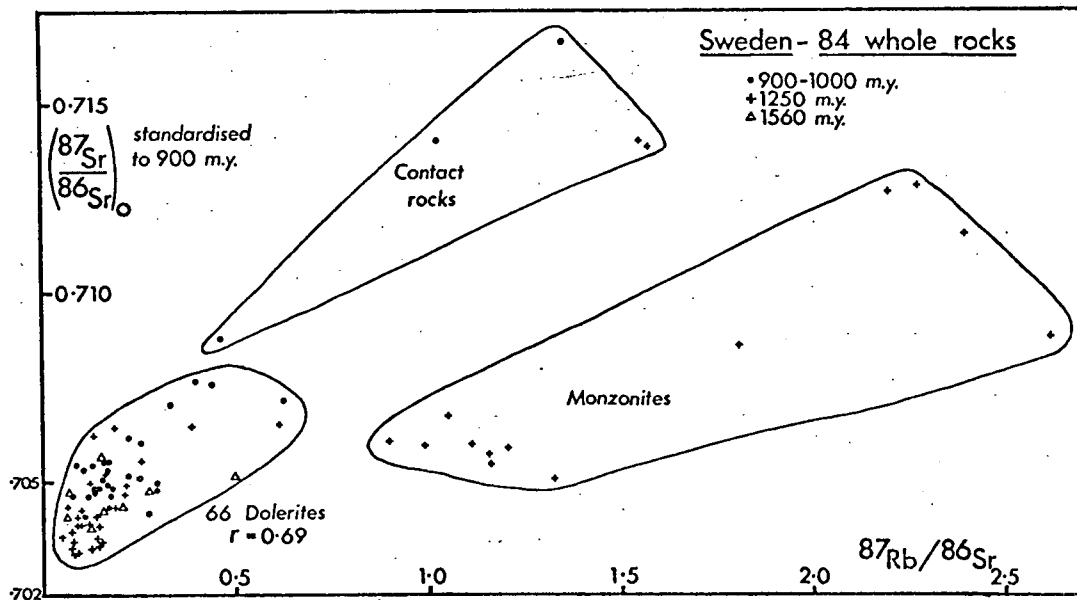


Fig. 6.11 $(^{87}\text{Sr}/^{86}\text{Sr})_0$ values (see appendix E) for 84 whole-rocks from Sweden corrected for linear mantle growth of ^{87}Sr to an arbitrary age of 900 m.y. plotted against $^{87}\text{Rb}/^{86}\text{Sr}$ ratio (= ca. $3 \times \text{Rb}/\text{Sr}$). Monzonites of ca. 1250 m.y. and contact hybrid rocks define distinct fields, and the 66 dolerites (arbitrarily, $^{87}\text{Rb}/^{86}\text{Sr}$ less than 0.65) show a positive correlation of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ with $^{87}\text{Rb}/^{86}\text{Sr}$ where the correlation coefficient $r = 0.69$.

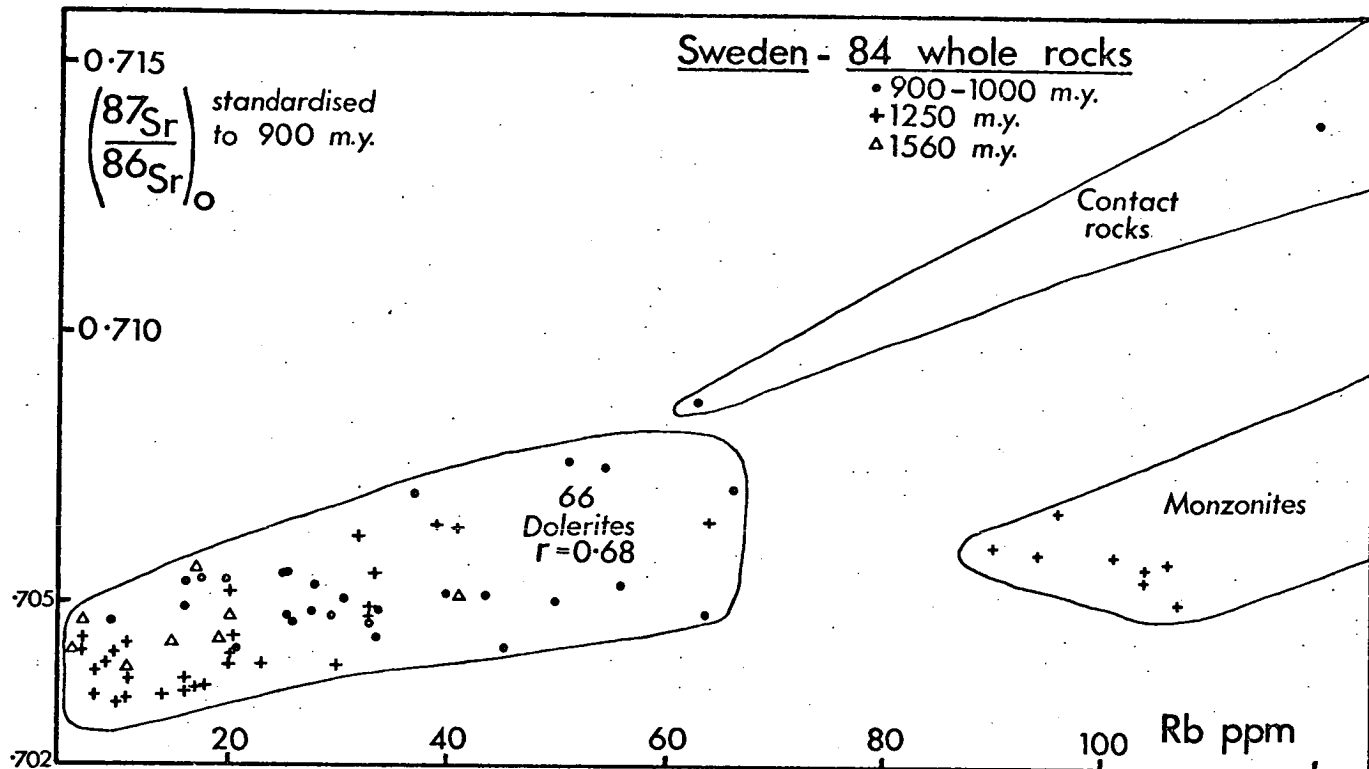


Fig. 6.12 $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ values (see appendix E) for 84 whole-rocks from Sweden corrected for linear mantle growth to 900 m.y. plotted relative to Rb concentration. Fields as in fig. 6.11. The 66 dolerites show positive correlation of $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ with Rb where the correlation coefficient $r = 0.68$.

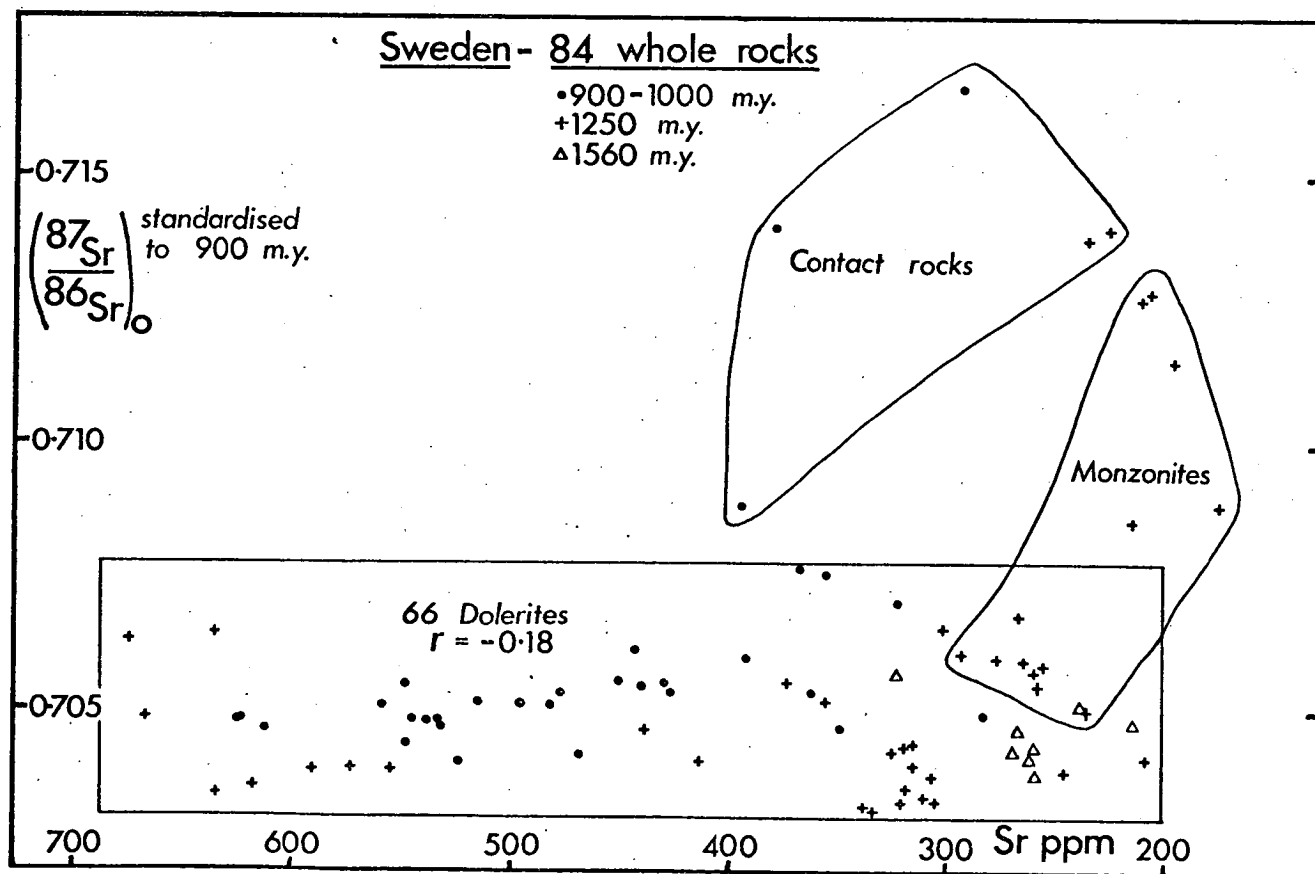


Fig. 6.13 $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ values (see appendix E) for 84 whole-rocks from Sweden corrected for linear mantle growth to 900 m.y. plotted relative to Sr concentration. Fields as in fig. 6.11. The 66 dolerites show no correlation of $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ with Sr where the correlation coefficient $r = -0.18$.

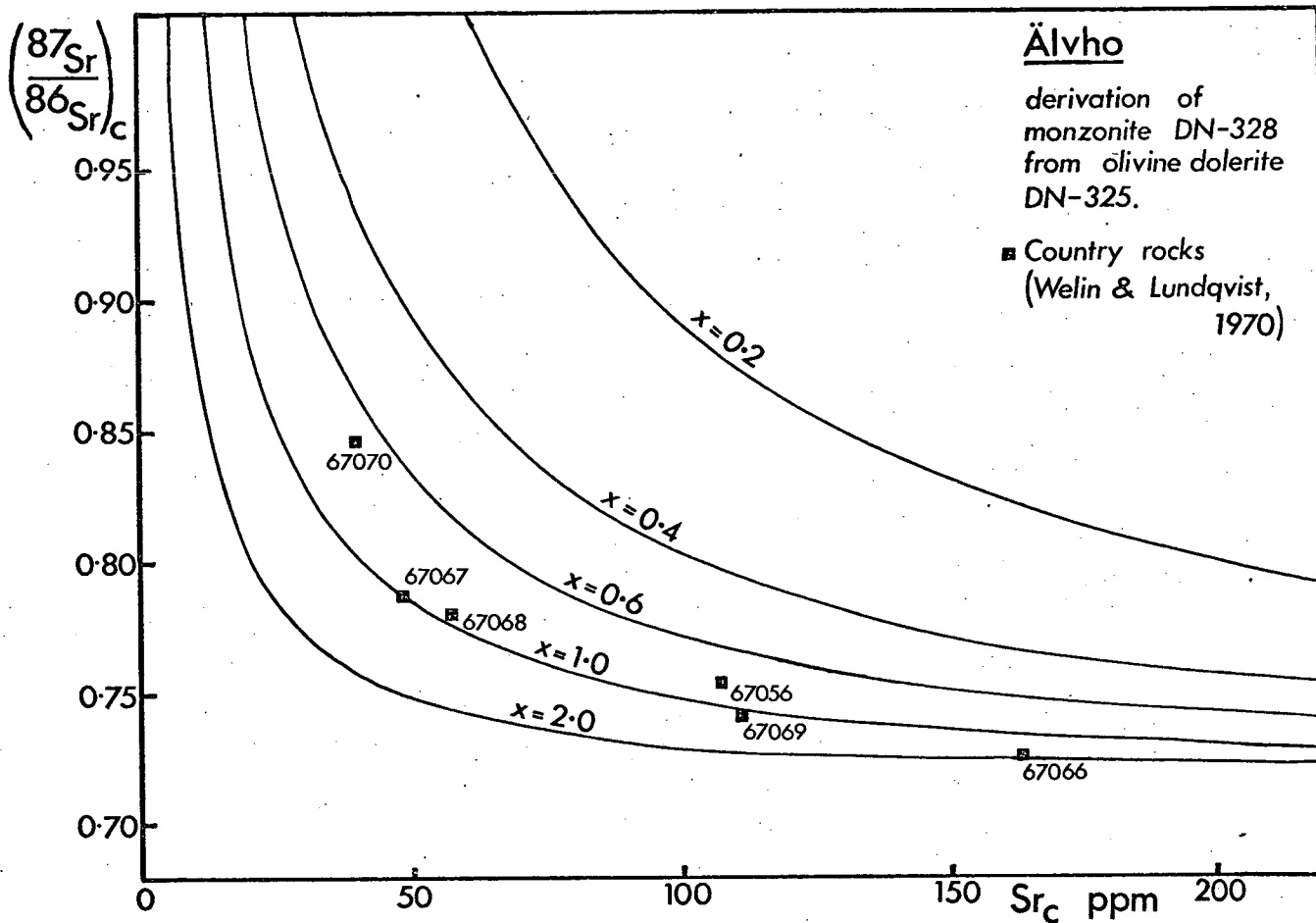
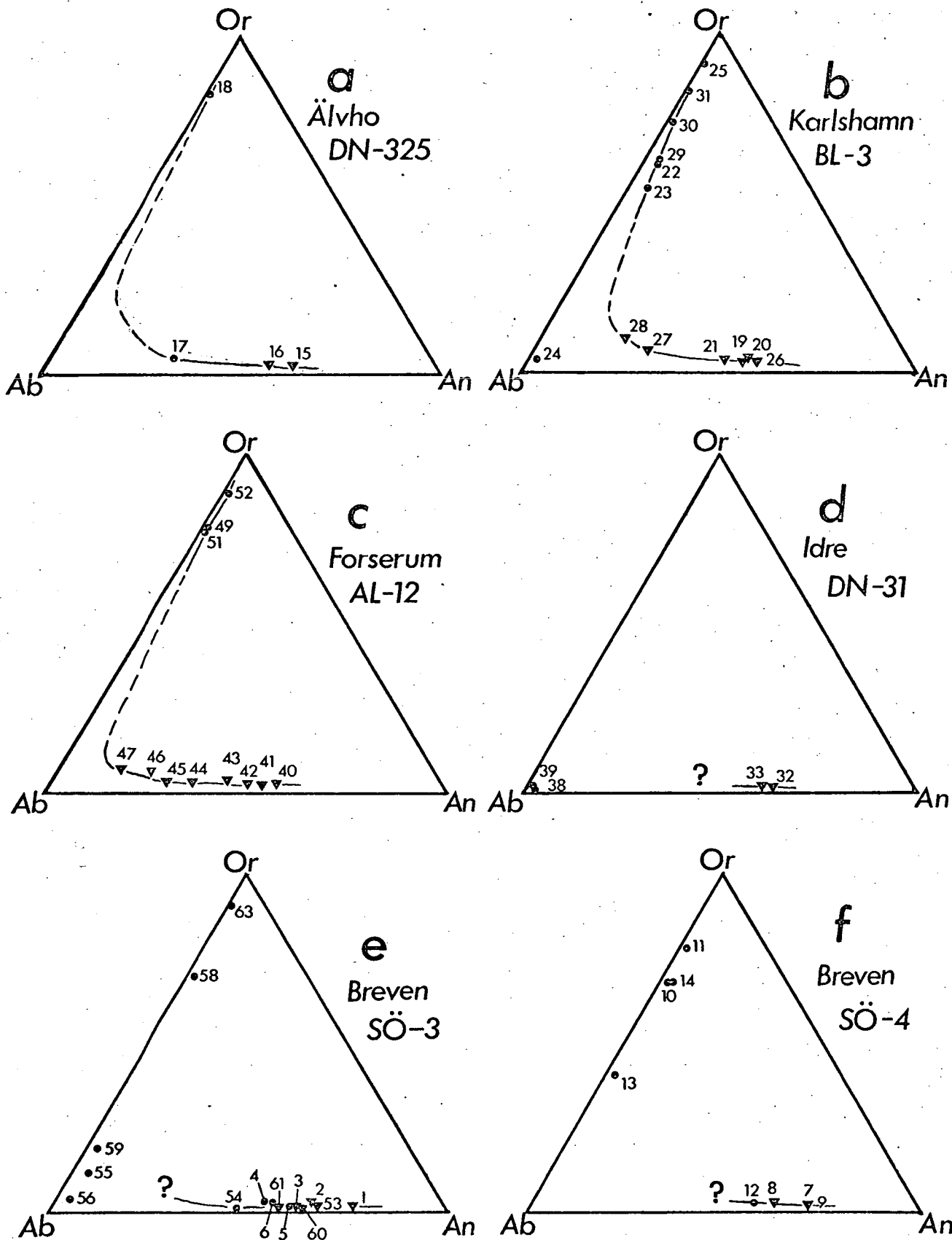


Fig. 6.14 Curves calculated from equation 6.3 for derivation of Älvho monzonite DN-328 from olivine dolerite DN-325, showing the x values (parts of contaminant to 1 part magma) required for various assumptions of $^{87}\text{Sr}/^{86}\text{Sr}$ and Sr concentration in the contaminant. The x values required by the 6 country rocks are shown in table 6.2.



• interstitial feldspar, no multiple twinning
 ▼ multiple-twinned (plagioclase) feldspar

Fig. 6.15 Compositions of feldspars in 6 dolerite samples. BL-3 and AL-12 give essentially normal feldspar lines, while DN-31, SÖ-3 and SÖ-4 show isotopic anomalies in feldspars and other minerals. For analytical data see appendix F.

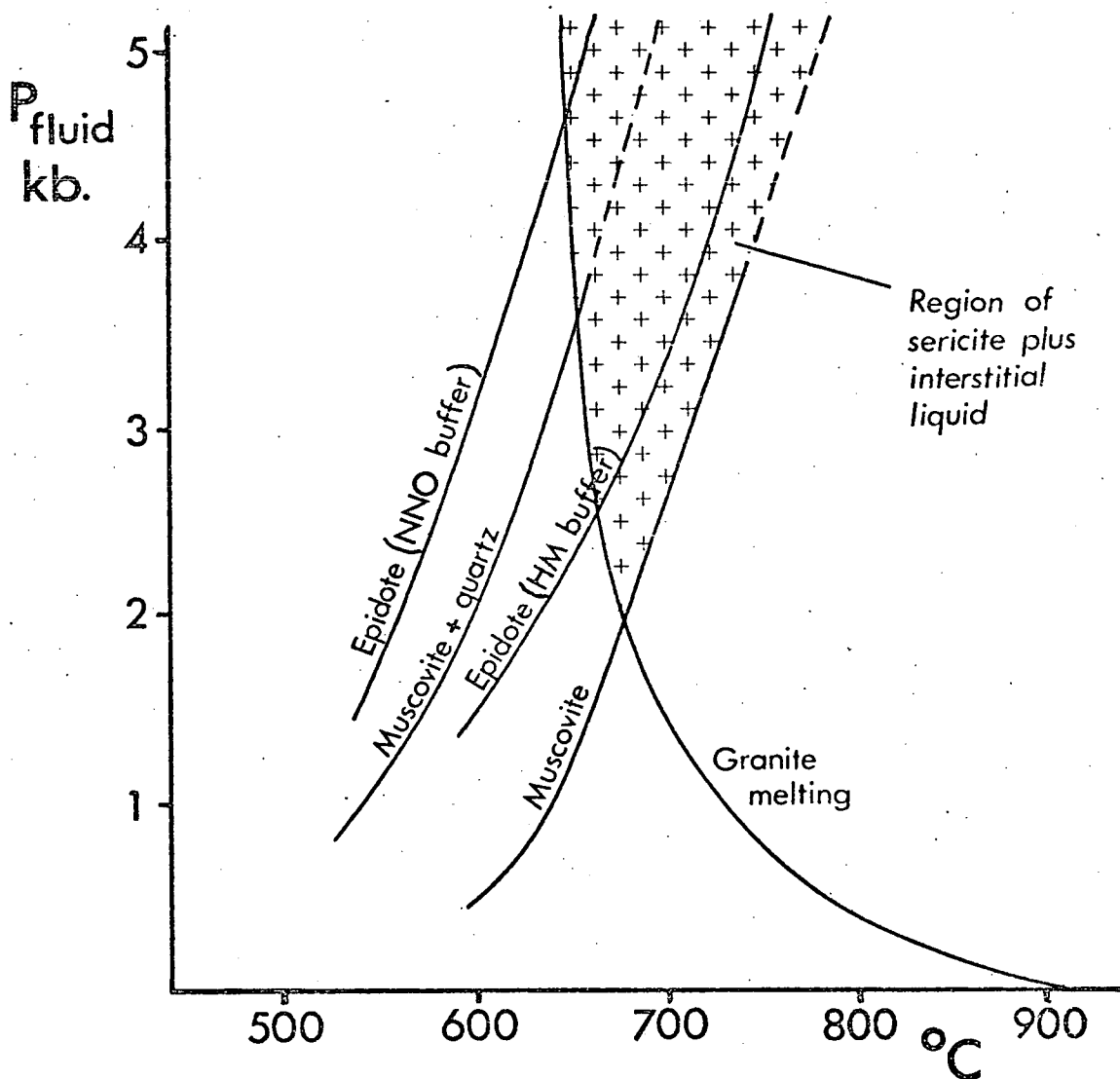


Fig. 6.16 Curves for the breakdown of muscovite and muscovite + quartz (Evans, 1965), of epidote at the nickel-nickel oxide and haematite-magnetite buffers (Liou, 1973), and for the melting of granite (Tuttle and Bowen, 1958). The granite melting curve is assumed to apply to the late interstitial liquids in many dolerites: see text for discussion.

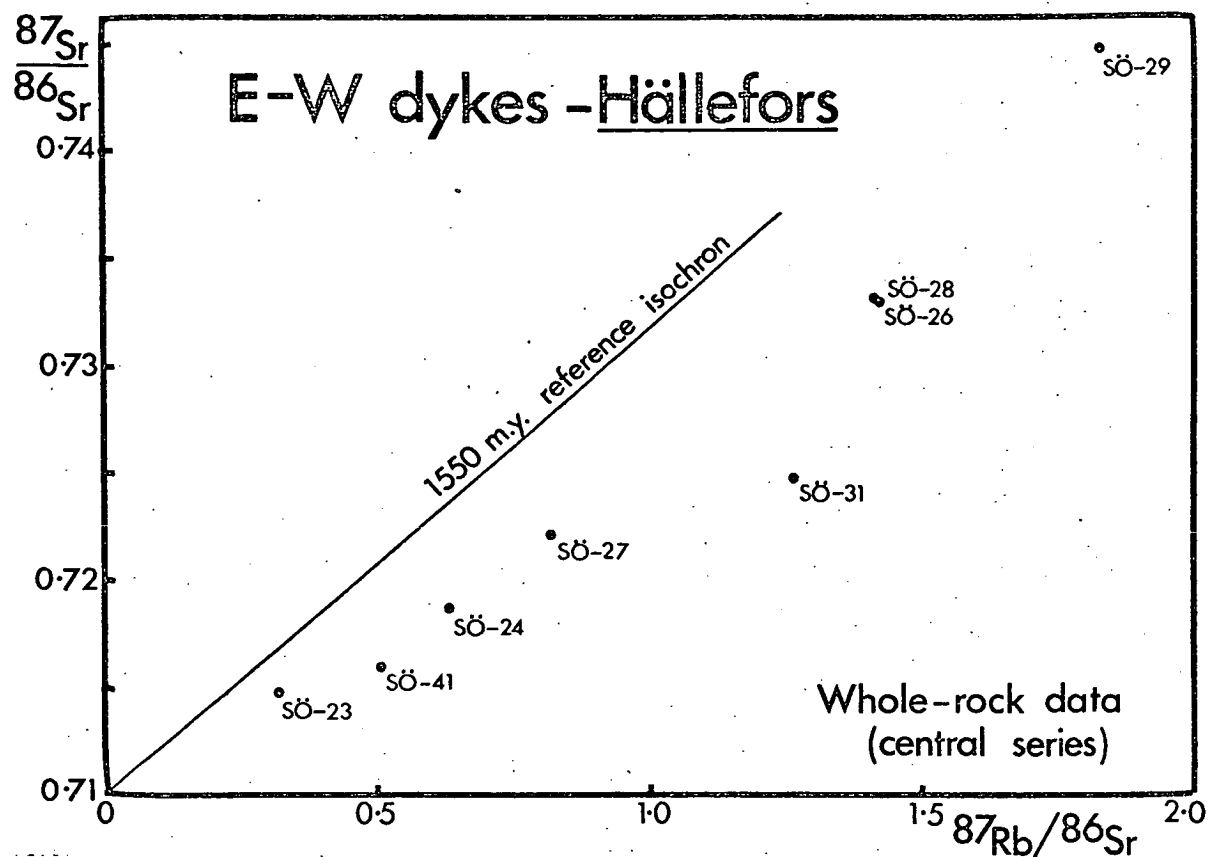


Fig. 6.17 Scattered pattern of whole-rock analyses from basic rocks of the Hällefors dyke, East-West system. See text for discussion.

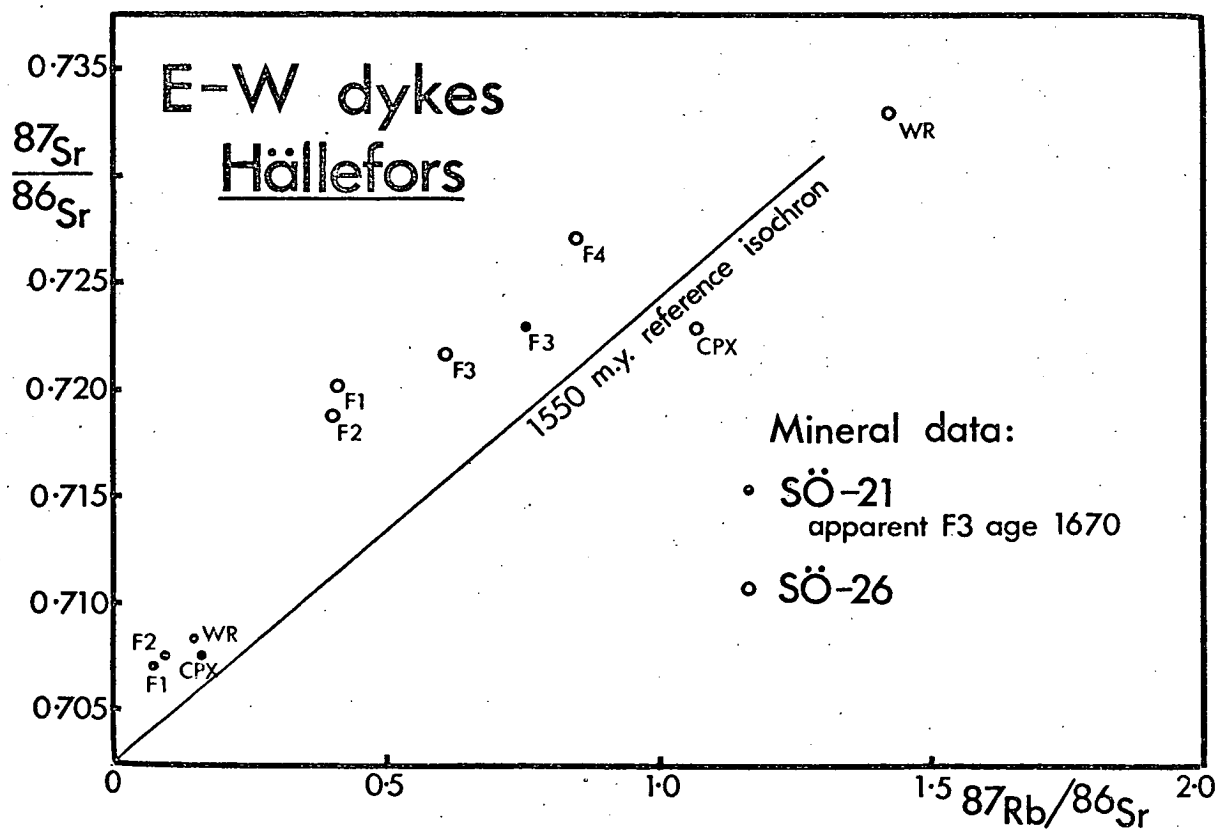


Fig. 6.18 Scattered pattern of mineral analyses from basic rocks of the Hällefors dyke, East-West system. See text for discussion.

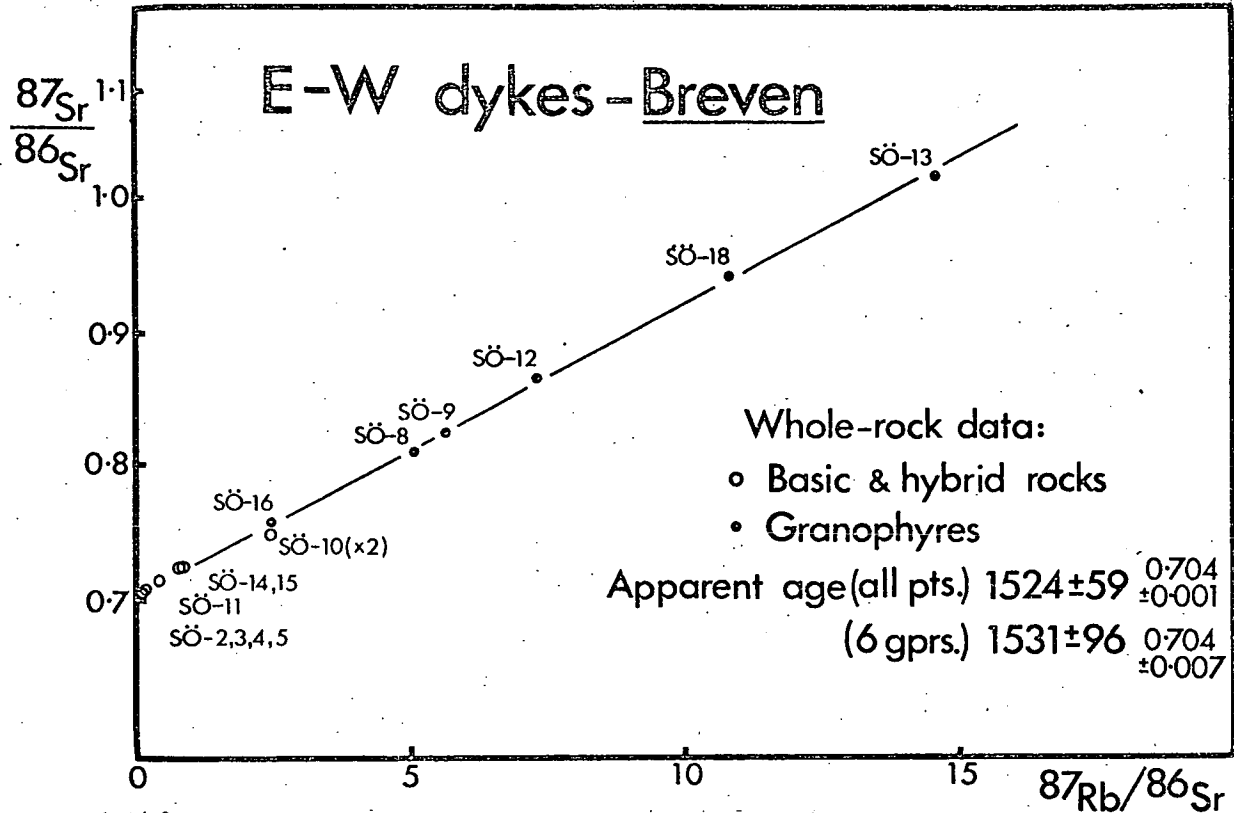


Fig. 6.19 Apparent age defined by whole-rock dolerites and granophyres of the Breven dyke, East-West system. See text for discussion.

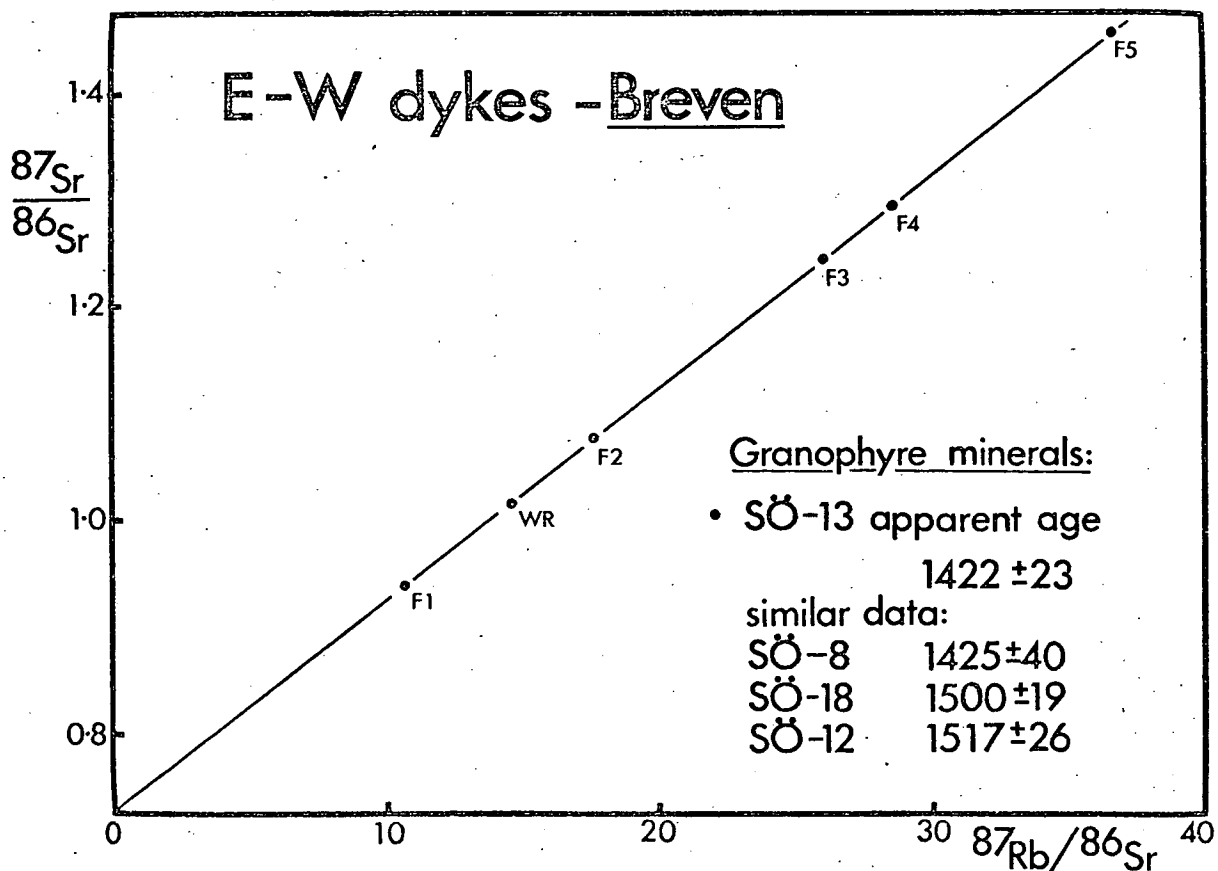


Fig. 6.20 Anomalous mineral lines for quartz-alkali feldspar density fractions from Breven granophyres, East-West dyke system. See text for discussion.

CHAPTER 7: SUMMARY OF CONCLUSIONS

7.1 Methodology: Rb-Sr dating of dolerite intrusions

1. For many dolerites the range of Rb/Sr (and hence also $^{87}\text{Sr}/^{86}\text{Sr}$) of whole-rocks is such that the $^{87}\text{Sr}/^{86}\text{Sr}$ needs to be extremely accurately determined in order to obtain an isochron. On the other hand, higher Rb/Sr ratios in whole-rocks are often associated with, or even a result of, assimilation of more evolved crustal material. Since this usually involves correlated contamination of the Rb/Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, whole-rock ages can be biased upward by large amounts. For these two reasons the whole-rock isochron method is not used for dating purposes.

2. The methods used for dating comprise:-

- a. Mineral isochrons from dolerite samples.
- b. Mineral isochrons from remelted country rocks.
- c. Whole-rock--biotite pairs from dolerite samples.

Both mineral isochron techniques depend mainly upon a series of feldspar density fractions separated from a fine grain-size; for dolerites, apatite and/or clinopyroxene are also analysed. The mass spectrometry for dolerite mineral isochrons is performed as a compromise between accuracy and speed: generally 1σ errors on $^{87}\text{Sr}/^{86}\text{Sr}$ are between 0.00002 and 0.00010, and are adequate for the range of ratios on the isochrons. It is shown that feldspar lines for dolerites can be subject to upward errors due to contamination by radiogenically-enriched strontium at a late stage in crystallisation. For this reason, feldspar lines for dolerites are considered reliable

only when apatite and/or clinopyroxene are concordant, and when the age is confirmed either by one of the other methods, or by an additional dolerite mineral isochron.

3. It is shown, both by published and new data that the country rocks were everywhere last thermally affected at 1700-1600 m.y. The mineral dates for dolerites are therefore to be regarded as defining ages of intrusion rather than any regional thermal event.
4. Where several biotites have been determined, they define the age of igneous activity to within 25 m.y., though determinations by all three methods generally agree at 2σ . The ages obtained are generally older than published K-Ar ages on the same groups of intrusions, particularly when these were performed by the whole-rock method.

7.2 Isotope geochemistry of basic igneous bodies in a continental crustal environment

1. Sometimes $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios measured on whole-rock samples show no variation within an intrusion. Examples of this are the anorthosites and gabbros from south Greenland (section 6.1), and several dolerite bodies where late-crystallising pegmatites demonstrate increase of Rb/Sr ratio by differentiation without any change in $(^{87}\text{Sr}/^{86}\text{Sr})_0$.
2. On the other hand many dolerites show variation of $(^{87}\text{Sr}/^{86}\text{Sr})_0$: these are considered on two scales; within a single intrusion and over the entire area of study. Random variation of initial ratio does

occur, but usually a correlation with Rb/Sr is observed, both on a local and on a regional scale. Visible assimilation of crustal rock at intrusive contacts involves correlated increase of Rb/Sr and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ in the dolerite.

3. In some cases variations of initial ratio between minerals in a single dolerite sample are observed, mainly involving large increases in the late-crystallising phases, such as alkali feldspar and epidote. This type of contamination is frequently associated texturally with late hydrous alteration (sericitisation, amphibolisation, serpentinisation, etc.) in the dolerites.
4. Correlated increase of Rb/Sr and $(^{87}\text{Sr}/^{86}\text{Sr})_0$ is best achieved through contamination by more evolved crustal material. Field evidence, consideration of contact temperatures, temperatures of melting of crustal rocks and the volumes of contaminate required show that incorporation of partial melts of crustal rock is a more likely mechanism for this effect on a whole-rock scale than bulk assimilation. It is possible that contamination from aqueous fluid plays some role in producing the initial ratio variations on a whole-rock scale, and this process is required to explain the contamination of late-stage minerals.
5. Although most dolerites are non-porphyritic, arguments based on observed phenocrysts, megacrysts and xenoliths of cumulate plagioclase, as well as Fe/Mg ratios show that most exposed rocks have been affected by some fractionation of mafic phases and/or plagioclase. Fractionation leads to increase of Rb/Sr, SiO_2 etc. without change in

$(^{87}\text{Sr}/^{86}\text{Sr})_0$, while contamination by crustal material increases both Rb/Sr and initial ratio. In the light of the variations observed, the compositional evolution of the dolerites in Sweden is inferred to be a result of a combination of differentiation and contamination processes. Such a combined origin can also be demonstrated for the more evolved rock types of certain intrusions.

6. Almost all bodies carry some samples with mantle-type $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratios (less than 0.705), no matter how wide the overall whole-rock variation. This may suggest that contamination was quantitatively most important at upper crustal levels, and the large within-intrusion variations of initial ratio support this view. The ability of a magma to mix itself during flow is difficult to assess however, and it is possible that contamination at deep levels could have been homogenised through the magma without appreciable effect on the overall $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio as measured at the present level of exposure.
7. It is shown that granitic interstitial liquids, equivalent to the interstitial granophyre of many dolerites, could coexist with hydrous alteration products such as sericite and epidote at P_{fluid} higher than 2 kb. Accordingly, aqueous fluids under high but erratic fluid pressure, moving through both dolerite and country rock, could introduce radiogenically-enriched Sr into late liquids (which crystallise as interstitial alkali feldspar or granophyre), and cause the observed late hydrous alteration of solid phases. This scheme is shown to be consistent with compositional data on interstitial feldspars.

8. Mild isotopic disturbance of dolerite and associated rocks long after intrusion produces strongly divergent whole-rock and mineral results. This suggests that no disturbance has occurred wherever concordant results are obtained by the various dating methods.

9. The $(^{87}\text{Sr}/^{86}\text{Sr})_0$ data for dolerites in Sweden demonstrate a profound influence on the basaltic magma of the continental crustal environment through which they pass, and within which they cool. Addition of material from crustal rocks makes a significant contribution to the compositional evolution of the magmas, and this contamination occurs right up to the final solidification of the dyke or sill.

7.3 Ages of igneous events and their tectonic significance

1. East-West trending dykes in central Sweden were emplaced at ca. 1560 m.y. before present (fig. 4.3).
2. North-North-East trending (Tuna) dykes in central Sweden were emplaced at 1401 ± 50 m.y. before present (fig. 4.3).
3. Widespread alkali-olivine dolerite sills in northern-central Sweden and western Finland have an age of ca. 1250 ± 25 m.y. (fig. 4.3). Petrographically similar dykes in the Gardar Province of south Greenland also give this age, and in south Greenland the ca. 1250 m.y. igneous event includes a variety of other rock types.

4. The Vaggeryd syenite of southern Sweden was emplaced at 1151 ± 67 m.y. (fig. 4.5), and was last cooled through Rb-Sr blocking temperatures of micas at ca. 900 m.y. This is a minimum age for the deformation and metamorphism which has affected the syenite.
5. Olivine dolerite and hypersthene dolerite dykes oriented parallel to the margin of the Sveconorwegian orogenic belt were emplaced into the Svecofennian crust during the approximate interval 1020-890 m.y. (fig. 4.5).
6. The Särna alkaline complex in northern-central Sweden was emplaced at 287 ± 14 m.y., and is related in space and time to the Oslo Graben tectonic and magmatic activity.
7. The igneous events in Sweden at ca. 1560 and ca. 1400 m.y. occupy restricted areas, but do correlate with syenitic, granitic or hydrothermal events in other parts of Sweden. The relationship of these events to major tectonics is however obscure.
8. The widespread ca. 1250 m.y. basic igneous events of Scandinavia and south Greenland are shown to correspond in age to the widespread Sudbury-Mackenzie set III dykes and Coppermine River Group lavas of the Canadian Shield (fig. 5.2). All these events comprise mainly mildly alkaline olivine basalt magmas, and it is suggested that they are tectonically related and correspond to a widespread fracturing of the 'pre-Grenville cratons'. Three causes of this fracturing and magmatism are envisaged: first, continental separation at 1250 m.y. entirely preceding the Grenville orogenic events; second, fracturing of the

pre-Grenville cratons due to Grenville orogenic events; and third, tensional tectonics during an essentially undetected interlude in Grenville orogenesis. The uncertainty over the nature, timing and duration of Grenville events prohibits a definite conclusion.

9. The ca. 1150 m.y. age for the Vaggeryd syenite could correspond to either of intensive alkaline magmatism in the Gardar Province of south Greenland at 1170 m.y., or to Mid Keweenawan igneous activity in America at 1140-1120 m.y. All three events are not however, contemporaneous and a single widespread period of fracture and magmatism is not envisaged in this case. These events could also have taken place (1) in pre-Grenville time, (2) during the Grenville events or (3) during a break in Grenville events. It is not possible to discriminate between these alternatives at the present time. A pre-Grenville age would necessitate a rejection or reinterpretation of a large amount of geochronological data from the orogenic belt.
10. The 1020-890 m.y. dykes of Sweden are apparently unrepresented in Greenland or Canada. They are explained as the result of tensional fracture of the pre-Sveconorwegian craton due to the proceeding uplift of the neighbouring Sveconorwegian 'block', as given by K-Ar ages (see fig. 5.1 for mechanism).
11. The basic igneous events dated therefore comprise:-
 - a. Ca. 1560 and 1400 m.y. events, correlatable with other igneous activity, but whose tectonic association is not yet clear.
 - b. Ca. 1250 m.y. events, corresponding to widespread fracturing of the pre-Grenville cratons in Scandinavia, Greenland and Canada.

This fracturing may be related to a climactic tectonic event in the Grenville-Sveconorwegian belt, but could equally be part of a phase of continental separation either of pre-Grenville or intra-Grenville age.

- c. Ca. 1020-890 m.y. events, suggested as caused by fracturing of the pre-Sveconorwegian craton during post-deformational uplift of the Sveconorwegian orogenic belt.

Although the ca. 1250 m.y. episode may be related to ocean opening, the interpretation of its significance is made difficult by the problem of the nature and duration of Grenville-Sveconorwegian orogenic events. The present work, and the palaeomagnetic studies which run in parallel to it, contribute to the knowledge of the global tectonics of the period 1600-800 m.y., but do not make possible a clear identification of a phase of pre-Grenville or pre-Caledonian ocean opening.

ACKNOWLEDGEMENTS

The author is grateful to Dr. O. van Breemen and Dr. B.G.J. Upton, his joint supervisors, and to A. Blaxland for much useful advice and discussion at all stages of this work. Assistance during field work and later discussions with Dr. G. Bylund of the University of Lund in Sweden are gratefully acknowledged. Staff of the University of Lund kindly made available samples BL-ARX, D75-26, F-3 and SPG-1.

During analytical work, much advice and material help were kindly provided by J. Jocelyn for sample preparation and mineral separation, G.R. Angell for X-ray fluorescence analysis, J. Hutchinson for mass spectrometry and processing of Rb-Sr data, J. Borthwick for oxygen and hydrogen isotope studies and Dr. P.G. Hill for electron microprobe work.

The advice and assistance of C. Chaplin and the technical staff of the Grant Institute of Geology, Edinburgh is gratefully acknowledged.

The author is grateful to G. Yarwood, whose help and advice greatly facilitated the final production of this thesis.

In particular, the author acknowledges the continued support of his wife, without whose forbearance a large proportion of the work would not have been possible.

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APPENDIX A: RB-SR WHOLE-ROCK AND U-PB AGES FROM THE GRENVILLE-

SVECONORWEGIAN CROGENIC BELT OF THE NORTH ATLANTIC REGION

In the following list all ages are quoted using the decay constant $\lambda^{87}\text{Rb} = 1.39 \times 10^{-11} \text{yr}^{-1}$. Age errors are given at 2σ , and the letter G after an age indicates that the authors interpreted the result as defining, or being very close to the igneous crystallisation of the rocks studied. R = Rb-Sr whole-rock line; U = U-Pb zircon, sphene or uraninite (usually zircon) age. O = Ontario; Q = Quebec; N = Norway; S = Sweden; Sc = Scotland; EG = East Greenland.

A.1. Canada and northern USA

No. of pts.	Age and method	Rock type and general location	Reference
8	1092 $\pm$ 20	R gneiss, Adirondack, USA	Hills & Gast, 1964
-	1310 $\pm$ 15	G U metarhyolite, Madoc, O	Silver & Lumbers, 1965
-	1250 $\pm$ 25	G U diorite, Madoc, O	'
-	1125 $\pm$ 25	G U granite, Madoc, O	'
-	1050 $\pm$ 20	G U pegmatite, Madoc, O	'
6	1330 $\pm$ 70	G R granite, North Bay, O	Hart et al., 1966
7	1725 $\pm$ 16	G R granite, French R., O	Krogh & Davis, 1967
10	1497 $\pm$ 10	G R granite, L. Muskoka, O	'
2	1100	U granite, French R., O	Tilton & Grünenfelder, 1968
4	2600	R granite, North Bay, O	Krogh & Davis, 1968
6	1130 $\pm$ 10	G U gneiss, Adirondack, USA	Silver, 1968
1	1100 $\pm$ 15	U syenite, Adirondack, USA	'

2	1090	±	10	U anorthosite, Adirondack, USA	
2	1080	±	10	U norite, Adirondack, USA	
1	1070	±	20	U anorthosite, Adirondack, USA	
2	1020	±	10	U ore bodies, Adirondack, USA	
7	1055	±	31	R gneiss, Adirondack, USA	Heath & Fairbairn, 1968
6	1285	±	41	G R nepheline syenite, Madoc, O	Krogh & Hurley, 1968
4	1230	±	72	R gneiss, Madoc, O	
6	1103	±	39	R gneiss, Madoc, O	
5	1096	±	50	G R syenite, Madoc, O	
4	1056	±	33	R metasediments, Madoc, O	
7	1035	±	60	R gneiss, Madoc, O	
7	1016	±	39	G R granite, Madoc, O	
2	2700			U pegmatite, Grenville F., O	Krogh & Davis, 1969
8	1700			G R granite, Grenville F., O	
3	1700			G U pegmatite, Grenville F., O	
7	1590			G R granite, Grenville F., O	
2	1550			G U granite, Grenville F., O	
4	2500	±	100	R granite, Grenville F., Q	Krogh et al., 1969
10	1790	±	95	R gneiss, Georgian Bay, O	Krogh & Davis, 1970
2	1730			U gneiss, Grenville F., O	Krogh et al., 1970
14	2759	±	170	G R granite, Grenville F., Q	Wanless et al., 1970
3	1025			G U pegmatite, French R., O	Krogh & Davis, 1972
6	1662	±	47	G R granite, Georgian Bay, O	Wanless & Loveridge, 1972
6	1120	±	46	G R granite, Madoc, O	
6	1143	±	48	G R gneiss, Morin, Q	Barton & Doig, 1972
13	1124	±	54	G R gneiss, Morin, Q	
7	961	±	40	G R granite, Morin, Q	
11	1576	±	38	R metasediments, Morin, Q	Barton & Doig, 1973

5	1460	±	116	G R alaskite, Morin, Q	
11	1091	±	102	R gneiss, Morin, Q	
9	1513	±	82	G R gneiss, Lac St. Jean, Q	Frith & Doig, 1973
10	1482	±	144	R gneiss, Lac St. Jean, Q	
5	1379	±	132	R gneiss, Lac St. Jean, Q	
5	1080	±	54	G R granite, Lac St. Jean, Q	
5	1066	±	24	G R granite, Lac St. Jean, Q	
4	1043	±	20	R monzonite, Lac St. Jean, Q	
6	1845	±	11	R gneiss, French R., O	Krogh & Davis, 1973
5	1035			R gneiss, French R., O	
2	2600			U gneiss, French R., O	Krogh & Davis, 1974
3	2600			U gneiss, Chibougamau, Q	
1	2600			U gneiss, Gagnon, Q	
2	994			U gneiss, Gagnon, Q	
6	1258	±	48	G R gneiss, Shawinigan, Q	Barton & Doig, 1974
8	1163	±	102	G R adamellite, Shawinigan, Q	
10	3000			R gneiss, Chibougamau, Q	Frith & Doig, 1975
6	1745	±	46	G R granite, Chibougamau, Q	
4	1112	±	66	R gneiss, Chibougamau, Q	
10	2548	±	94	G R syenite, Cabonga, Q	Doig, 1975
9	1083	±	18	R gneiss, N. Ellesmere I.	Sinha & Frisch, 1976

A.2. Scandinavia

No. of pts.	Age and method	Rock type and general location	Reference
3	920 $\pm$ 20	G U pegmatite, Rogaland, N	Kulp & Eckelmann, 1957
9	930	G U pegmatites, Bohus, S	Welin & Blomqvist, 1964
7	1057 $\pm$ 150	R gneiss, Tafjord, N	Brueckner et al., 1968
2	1030	U gneiss, Rogaland, N	Michot & Pasteels, 1969
3	950 $\pm$ 20	G U mangerite, Rogaland, N	
5	1616 $\pm$ 76	R gneiss, Bamble, N	O'Nions & Baadsgaard, 1971
6	1167 $\pm$ 100	R gneiss, Bamble, N	
6	1160	U gneiss, Bamble, N	
11	1100	U pegmatite, Bamble, N	
5	1265 $\pm$ 40	R metasediments, Kongsberg, N	O'Nions & Heier, 1972
5	1160 $\pm$ 48	R gneiss, Stavanger, N	Heier et al., 1972
7	1405 $\pm$ 38	G R gneiss, Telemark, N	Priem et al., 1972
6	1253 $\pm$ 160	R gneiss, Holsavatnet, N	Brueckner, 1972
5	1004 $\pm$ 90	G R granite, Hamlagrøvatnet, N	
5	961 $\pm$ 60	G R granite, Bamble, N	
5	1290 $\pm$ 50	R gneiss, Rogaland, N	Michot & Pasteels, 1972
4	1100 $\pm$ 30	U gneiss, Rogaland, N	
2	1090 $\pm$ 30	U gneiss, Rogaland, N	
2	1040 $\pm$ 20	U gneiss, Rogaland, N	
4	975 $\pm$ 20	U gneiss, Rogaland, N	
5	975 $\pm$ 20	U adamellite, Rogaland, N	
2	990 $\pm$ 20	G U farsundite, Rogaland, N	
2	957 $\pm$ 43	G U farsundite, Rogaland, N	

5	1031	+	35	G R	granite, Jotunheimen, N	Priem et al., 1973a
11	1128	+	23	R	gneiss, Telemark, N	Priem et al., 1973b
7	907	+	46	G R	granite, Telemark, N	
7	1643	+	88	R	gneiss, Røldal, N	Andresen et al., 1974
5	1289	+	80	R	metasediments, Røldal, N	
6	943	+	100	R	metasediments, Røldal, N	
6	1484	+	60	R	gneiss, Rogaland, N	Verstevee, 1974
9	1220	+	100	U	gneiss, Rogaland, N	
14	1206	+	70	G R	syenite, Rogaland, N	
10	1004	+	28	R	gneiss, Rogaland, N	
22	947	+	26	R	gneiss, Rogaland, N	
25	860	+	30	G R	monzonite, Rogaland, N	
6	1243	+	160	R	granite, Stavanger, N	Andresen & Heier, 1975
5	913	+	28	G R	granite, Flå, N	Killeen & Heier, 1975
5	900	+	35	G R	granite, Iddefjord, N	
7	1250	+	50	G R	granite, Åmål, S	Gorbatshev & Welin, 1975
7	1240	+	30	G R	granite, Åmål, S	Welin & Gorbatshev, 1976
6	1405	+	40	G R	granite, Dalsland, S	Skiöld, 1976
9	1400	+	80	G R	granite, Dalsland, S	
8	1050	+	40	R	slate, Dalsland, S	
8	910	+	35	G R	granite, Bohus, S	

A.3. Scotland and East Greenland

No. of pts.	Age and method	Rock type and general location	Reference
7	1162 $\pm$ 85	R metaseds., Scoresby S., EG	Hansen et al., 1974
4	1060 $\pm$ 40	R granite, Scoresby Sund, EG	
8	950	U gneiss, Scoresby Sund, EG	
11	1050 $\pm$ 46	R gneiss, Ardgour, Sc	Brook et al., 1976

APPENDIX B: ANALYTICAL TECHNIQUES

B.1. The Rb-Sr method

Natural rubidium is composed of two isotopes, ^{85}Rb and ^{87}Rb : the latter, comprising 27.83% of total Rb (Catanzaro et al., 1969), decays by emission of an electron (β -decay) to ^{87}Sr with a decay constant (λ) in the region of $1.4 \times 10^{-11} \text{ yr}^{-1}$. Direct determination of the decay constant ($\lambda = 0.693/\text{half-life}$) has been hampered by the low energy of the decay process, and three different values are now in use among isotope geologists. The value $1.47 \times 10^{-11} \text{ yr}^{-1}$ was determined by direct counting (Flynn and Glendenin, 1959), while $1.39 \times 10^{-11} \text{ yr}^{-1}$ was obtained by geological comparison with reliable U-Pb ages (Aldrich et al., 1956); additionally the value $1.42 \times 10^{-11} \text{ yr}^{-1}$ has come into use, based on revised U decay constants (Jaffey et al., 1971). For this study the decay constant $1.39 \times 10^{-11} \text{ yr}^{-1}$ (corresponding to $t^{1/2} = 5.0 \times 10^{10} \text{ yr}$) will be adopted, as being the one most generally used.

Nier (1938) reported the following isotopic composition for 'normal' strontium:-

^{88}Sr	82.56 atom %
^{87}Sr	7.02 atom %
^{86}Sr	9.86 atom %
^{84}Sr	0.56 atom %

^{87}Sr is the only isotope produced by radioactive decay of another nuclide, and ^{84}Sr , ^{86}Sr and ^{88}Sr remain constant through time. The ^{87}Sr in a rock or mineral is composed of that originally present at the time of its formation plus the radiogenic ^{87}Sr produced since that time by the decay of ^{87}Rb , as expressed by the relation

$$^{87}\text{Sr}_{\text{present}} = ^{87}\text{Sr}_{\text{initial}} + ^{87}\text{Rb}_{\text{present}}(e^{\lambda t} - 1) \quad \text{B.1}$$

where t is the time since closure of the isotopic system and λ is the decay constant. It is easier in practice to deal with ratios than with concentrations, so this equation is divided throughout by ^{86}Sr

$$(^{87}\text{Sr}/^{86}\text{Sr})_p = (^{87}\text{Sr}/^{86}\text{Sr})_0 + (^{87}\text{Rb}/^{86}\text{Sr})_p(e^{\lambda t} - 1) \quad \text{B.2}$$

This is the equation for a straight line or isochron (Nicolaysen, 1961), where the intercept is the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and the slope is given by

$$\alpha = e^{\lambda t} - 1 \quad \text{B.3}$$

From this the age t is given by

$$t = 1/\lambda \ln(1 + \alpha) \quad \text{B.4}$$

For samples with very high present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios it is possible to calculate an age by assuming a value (e.g. 0.71) for $(^{87}\text{Sr}/^{86}\text{Sr})_0$. For lower present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios however, this assumption, if incorrect, could lead to large errors in age, and four or more points, defining an isochron must be determined. By definition all samples on an isochron are assumed to have equilibrated isotopically with each other at some definite time, when they all acquired the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. This value will have increased in direct proportion to $^{87}\text{Rb}/^{86}\text{Sr}$ according to equation B.2, so that at any later time

(e.g. the present) the samples will define a straight line in $^{87}\text{Rb}/^{86}\text{Sr}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ space, yielding an age and an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

The isochron method is generally applied either to whole-rock systems which equilibrated isotopically at the time of intrusion, metamorphism or diagenesis, or to mineral systems from a single whole-rock sample which generally equilibrated at the time of last crystallisation.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is measured by direct determination on a mass spectrometer, but the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio depends upon the ^{87}Rb and ^{86}Sr contents of the sample and cannot be measured directly. These concentrations are obtained by the method of isotope dilution: this involves adding a known amount of 'spike' (highly enriched in ^{87}Rb or ^{84}Sr) to a known amount of sample and measuring the resulting ratio of spike to common isotope ^{85}Rb or ^{86}Sr . The quantities of sample in grams, and spike in micromoles are known, as well as the isotopic composition of the spike; from the observed isotope ratios it is therefore possible to calculate the micromoles/gm of ^{87}Rb and ^{86}Sr in the sample, which directly yield the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio.

B.2 Sample collection

Collection of samples from 30 intrusive bodies in Sweden, mainly dolerite sills or dykes, was undertaken in May-June 1974, and the Greenland material was collected from many different dykes in July-August of the same year. In the event, material was analysed from 24 separate intrusions in Sweden, and 7 in Greenland: locations of samples and some observational data are given in appendix C. Samples from the lamprophyres of the Ivigtut region and the Younger Giant Dyke Complex (both in Greenland) were collected by Dr. B.G.J. Upton. Samples F-3, SPG-1, EL-ARX,

71-15 and D75-26 were kindly provided by staff of the University of Lund in Sweden.

In Greenland natural glaciated exposures were used in all cases, but in Sweden natural exposures are both rare and weathered, and roadcuts were used whenever possible. Weathered surfaces, joint planes and alteration features were either avoided or removed from the samples in the field.

Generally 15-25 whole-rock samples were collected from each intrusion to be studied. Traverses across intrusions were taken in the few cases where this was possible, and country rocks remelted by heat from the basic magma were collected whenever a contact was exposed. For the Vaggeryd syenite sample size was 2-3 kg, but since all other intrusions are undeformed, sample size for Rb-Sr purposes is not an important criterion; generally, samples were 0.5 kg, increasing with grain-size to 2 kg.

B.3 Sample preparation and mineral separation

For mineral age study the samples showing the least alteration of primary phases were selected on the basis of thin-section studies. These, together with samples intended only for whole-rock work were trimmed to remove any remaining weathered surfaces, cleaned under the tap and rinsed with acetone. After drying they were crushed to chips less than 10 mm in a steel jaw-crusher; this machine was vacuum-cleaned, brushed and when necessary washed between samples to prevent cross-contamination. The resulting gravel was coned-and-quartered, and 200-400 gm removed for finer crushing; the remainder was stored for future mineral separation. A 250 c.c. 'Tema' mill made of sintered Al_2O_3 was used for pulverising the whole-rock samples down to less than 30 microns; a 20 gm split of the resulting powder was taken by coning-and-quartering, and used for all analytical work.

For mineral separation, the jaw-crushed whole-rock was pulverised in the Tema mill for the time necessary to produce a maximum yield of the grain-size required.

Separations of apatite, clinopyroxene and feldspars from dolerite samples were performed on a grain-size of 45-84 μ , sieved with nylon cloths. A fine size is necessary both to avoid composite grains and to ensure separation of the various compositional zones of feldspar, on which the dolerite mineral isochron technique used in this study depends. Clinopyroxene and apatite were separated in the usual way and are at least 98% pure. The feldspars were sunk as fractions of progressively decreasing density in tetrabromoethane diluted with acetone; each stage in the separation was left until no further grains sank (generally 1 hour), the sink run off at the base of the funnel, and more acetone added and mixed to allow the next lightest fraction to settle. The separates were then cleaned at high setting in a Frantz magnetic separator (to remove composite grains), and the result was a series of feldspar fractions numbered F1 onwards in order of decreasing density and usually increasing Rb/Sr ratio. Between 2 and 10, usually 5 feldspar fractions were separated in this way, and any alkali feldspar or interstitial granophyre is concentrated in the lighter fractions.

Biotites were separated from dolerites, of which they always comprise less than 1% if present, by using the whole grain-size range from the Tema mill in order to maximise yield. The material (0.5-2 kg) was carefully washed in a large porcelain bowl and when clean, further water was added, the sand agitated and then allowed to settle slowly. The dynamic properties of the biotite grains cause them to concentrate in the surface layers, from which they can be removed with a spatula: 50-100 such settling operations yield sufficient biotite for one analysis. The

biotite-rich material is then dried, sieved into grain-size fractions and repeatedly sifted on laboratory filter papers, where the biotite is separated by adhering to the paper. This process results in a concentrate which is 30-90% biotite and heavy liquids can then be used for a fine density separation. Due to the risk of contamination from other micas, no dolerite biotites were put through the magnetic separator. The sample obtained is 0.005-0.5 gm of 90-100% biotite and is adequate for Rb-Sr analysis.

Mineral separations for other rocks, such as granite country rocks and Vaggeryd syenite were performed by a combination of standard techniques using coarser grain-sizes (up to 142 microns) than for dolerites.

B.4 Chemical procedures

Before use involving samples, standards or spike solutions, all laboratory ware was heated overnight in reagent-grade 6N HCl, followed by 2 x distilled H₂O. All apparatus not in current use was wrapped in cling-film and stored in clean drawers.

Spike solutions at East Kilbride are prepared in batches from Oak Ridge or Harwell ⁸⁴SrCl₂ (Harwell, 99.79% ⁸⁴Sr) and ⁸⁷RbCl (Harwell, 99.21% ⁸⁷Rb). The separate spike solutions are stored in 500 c.c. polythene bottles insulated against evaporation. This supply is decanted into 50 c.c. polythene drip-bottles which are used for weighing into beakers for analyses. Combined Rb and Sr spikes are made in PTFE beakers in batches of 60-100, evaporated and stored to be used as required. The concentrations of the spike solutions are calibrated against standard solutions of NBS 984 RbCl and NBS 987 SrCO₃ salts each time that spiked beakers are prepared. The isotopic composition of the spike is measured by direct determination on the mass spectrometer.

Some whole-rock samples early in the study were analysed for approximate Rb and Sr concentrations on the automated Philips X-ray fluorescence spectrometer at Edinburgh. For later samples however, the Rb and Sr concentrations were estimated from those of previously analysed similar samples. This was facilitated by the standardised mineral separation procedures for dolerites, enabling Rb/Sr ratios to be predicted accurately enough for the purposes of spiking.

Samples (0.005-0.3 gm) were weighed into the pre-spiked PTFE beakers and soaked in 5-15 ml cold HF for between 3 and 12 hours (specifications of reagents are given in section B.5). Most acid attack of the samples takes place in this cold stage and the covered beakers are then heated overnight at 150-180°C to complete the dissolution and ensure isotopic equilibration of the sample with the spikes. The next stage in the procedure requires an oxidation of the sample, since fluorides are insoluble in HCl: early samples were oxidised with HClO_4 (1-4 ml), but HNO_3 (2-6 ml) was used for some 50% of the analyses reported here. In both cases the beakers were heated at 180°C for 2-3 hours to ensure complete reaction, and the acid then fumed off leaving an HCl-soluble sample.

The sample was then dissolved in 5-10 ml 2.5N HCl, evaporated down to 1-2 ml, diluted 50% with H_2O and placed onto the ion-exchange column, capable of separating by elution the various cations in a mixture. The columns were made of glass, were unjointed, and the resin (13 gm of Bio-Rad 50W-X8 200-400 mesh) was supported by a porous glass sinter disc built into the column. The columns were cleaned between samples with the following procedure: 50 ml 6N HCl, 50 ml 6N HCl, rinse out resin and resettle in H_2O , 50 ml 6N HCl, and 100 ml 2.5N HCl to restore the column to its operating height. The sample was eluted through the column, and Rb

and Sr fractions were collected and evaporated to dryness. A single elution was found to be sufficient to separate the strontium from calcium.

B.5 Analytical blanks

Table B.1 gives details of reagent grade and measured natural Rb and Sr blanks. Distillation of H_2O was at boiling temperatures in glass, and of acids at sub-boiling temperatures in quartz.

Table B.1 Reagent grade and analytical blanks

Date	Type of blank	Reagent grade	N_{Rb} ng 10 ml ⁻¹	N_{Sr} ng 10 ml ⁻¹
8. 5.75	HF	Merck Suprapur	0.07	0.10
20.11.74	HClO ₄	Merck Suprapur	0.21	1.00
8. 5.75	6N HCl	Analar 1xdist.	0.04	0.58
20.11.74	2.5N HCl	Analar 1xdist.	0.03	0.04
20.11.74	H ₂ O	2xdist.	0.01	0.10
Date	Type of blank	Average of	N_{Rb} ng (total)	N_{Sr} ng (total)
30. 4.75	columns	2	2.30	0.20
17.10.74	overall	1	11.60	4.60
29. 5.75	overall	3	8.40	3.70
18. 8.75	overall	3	7.20	3.50

The Analar 2 x distilled HNO_3 was not measured directly in this study, however the overall blank of 18.8.75 (average of three determinations in the preceding month) included HNO_3 in place of HClO_4 and shows a clearly comparable result to 29.5.75 (also average of three), which was measured using HClO_4 ; this indicates that the 2 x distilled HNO_3 is at least of comparable grade to the Merck HClO_4 .

A summation of all the sources of blank in table B.1 leads to an overall blank of 2.7 ng ^{87}Rb and 1.9 ng ^{87}Sr . Excesses of reagent used in some cases could raise these figures to ca. 4.0 and 3.5 respectively, but the observed overall blanks are nevertheless somewhat higher than those predicted on this basis. The origin of this additional blank is difficult to identify: sources could be erratic column blanks or the loading procedure.

An overall blank of 8 ng Rb and 3.5 ng Sr, under which condition all the work on dolerites was done, produces significant contamination only in the case of clinopyroxenes; for all other samples blank errors are well within analytical uncertainties. A correction to the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio for clinopyroxenes of up to 2.0% is made where this would exceed the 1 σ estimated uncertainty of 0.7% on $^{87}\text{Rb}/^{86}\text{Sr}$, as in the case of the dolerites from Blekinge, southern Sweden. All other pyroxenes have higher Rb contents, and are hence less subject to blank errors; for this reason no correction is made.

B.6 Mass spectrometry

The mass spectrometer used at East Kilbride is a modified AEI-GEC MS-12 solid-source machine with a 12 inch radius and 90° sector. The collector system is a Faraday cup combined with a Cary 401 vibrating-reed electrometer using a 10¹¹ ohm resistor. This resistor is calibrated and

peaks are continuously corrected for signal decay at the total peak-switching time in use (4.5 secs.). The system for input of data to the on-line computer was built by G.Ross, and the programmes for continuous treatment of data were written by J.Hutchinson.

The mean value of 16 analyses of the Eimer and Amend SrCO_3 standard made by J.Hutchinson during the course of this study was 0.70804 ± 0.00012 (2σ).

The sample Rb and Sr fractions were loaded onto tantalum filaments mounted on nickel posts set in glass beads. These beads were cleaned between samples in weak HCl and 2 x distilled H_2O . The filaments were outgassed for 15 minutes in high vacuum at a heating current of 4.2 amps, at least 1 amp higher than currents used during analyses.

The Rb fraction was loaded with 2 x distilled H_2O onto both side filaments of a triple-filament assembly using a cleaned Pyrex capillary tube. The side-filament current used to vapourise the alkali was 0.4-1.0 amps, and the ionising centre-filament current was 2.2-2.5 amps. The two-stage process minimises fractionation of the two Rb isotopes during the run. Between 2 and 5 sets of 10 scans over the peaks ^{85}Rb and ^{87}Rb were taken for each analysis.

The Sr fraction was loaded onto a central filament with a weak solution of Merck Suprapur H_3PO_4 (1-3%) in 1 x distilled 6N HCl, using a cleaned Pyrex capillary tube. Filament current used during runs was 2.6-3.0 mps, giving a 50-90% intensity on the Cary 3V scale. A separate source block (in which the bead is mounted for running) was used for all samples with $^{87}\text{Sr}/^{86}\text{Sr}$ less than 0.74, in order to avoid contamination from highly radiogenic samples. All source blocks were cleaned on average after 20 samples. ^{85}Rb was continuously monitored during the runs, and the ^{87}Sr peak corrected for any ^{87}Rb present, which was however,

generally negligible.

The number of sets of 10 scans taken over the peaks 84, 85, 86, 87 and 88 depends upon the accuracy required for the analysis. For the anorthosites and gabbros from South Greenland, 20-40 such sets were taken, yielding a consistent 1σ error on $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.00003. For all samples from dolerites, 10 sets or so were taken and the 1σ error on this ratio varies between 0.00002 and 0.00015. For samples with higher isotopic ratios 3-5 sets were taken, and these analyses are subject to the standard laboratory error of 0.04%.

B.7 Analytical errors

General estimates of laboratory error for samples lying in the normal isochron range are calculated from replicate analyses; tables B.2 and B.3 give details of duplicate analyses performed in this study.

Table B.2 Duplicate analyses for $^{87}\text{Rb}/^{86}\text{Sr}$
in routine analysis

Sample	Anal.1	Anal.2	% difference
SÖ- 10 WR	2.4385	2.4145	0.99
SÖ- 18 WR	10.8098	10.7625	0.44
SÖ- 54 WR	0.2740	0.2811	2.50
DS- 17 WR	8.5049	8.4709	0.40
DS- 21 WR	142.6810	143.1550	0.33
J- 33 F3	10.0399	10.0715	0.32
DN-306 WR	1.3167	1.3236	0.52
V- 5 WR	12.9124	13.1728	2.00
V- 6 WR	45.2645	45.0589	0.46
DN- 9A F2	1.4260	1.4222	0.27
Mean			0.82

Table B.3 Duplicate analyses for $^{87}\text{Sr}/^{86}\text{Sr}$
in routine analysis

Sample	Anal.1	Anal.2	% difference
SÖ- 10 WR	0.74808	0.74777	0.04
SÖ- 18 WR	0.93985	0.94005	0.02
SÖ- 54 WR	0.71015	0.71008	0.01
DS- 17 WR	0.87978	0.88071	0.01
DS- 21 WR	3.47919	3.47085	0.24
J- 33 F3	0.93093	0.92995	0.10
DN-306 WR	0.72779	0.72783	0.01
V- 5 WR	0.91321	0.91283	0.04
V- 6 WR	1.37728	1.37762	0.02
DN- 9A F2	0.71145	0.71135	0.01
Mean			0.05

Omitted from tables B.2 and B.3 are duplicate analyses showing positive correlation of $^{87}\text{Rb}/^{86}\text{Sr}$ with $^{87}\text{Sr}/^{86}\text{Sr}$ ('isochron shifts') in excess of 1.0% and 0.1% respectively, which are due to sample inhomogeneity: duplicates on granular samples (minerals) tend to fall into this category. Samples with $^{87}\text{Rb}/^{86}\text{Sr}$ less than 0.1 are also omitted as these can be subject to excessive Rb blank errors.

Laboratory errors based on a much larger number of replicate analyses by other workers at East Kilbride suggest values of 0.7% for $^{87}\text{Rb}/^{86}\text{Sr}$ and 0.04% for $^{87}\text{Sr}/^{86}\text{Sr}$, and the mean error estimates presented here

(tables B.2, B.3) agree with these figures. Since 0.7% and 0.04% are based on a much larger sample, they are used in this study whenever an estimated precision is required.

The 0.7% (1σ) error is assigned to all $^{87}\text{Rb}/^{86}\text{Sr}$ results. The 0.04% (1σ) error is assigned to $^{87}\text{Sr}/^{86}\text{Sr}$ values whenever no internal precision is calculated from the Sr run, in practice for $^{87}\text{Sr}/^{86}\text{Sr}$ greater than 0.75 (granitic rocks etc.). For dolerite samples the in-run 1σ $^{87}\text{Sr}/^{86}\text{Sr}$ errors for all the points on the isochron are averaged and used as a percentage error: this is always less than 0.02%.

The 0.04% uncertainty is assigned to whole-rock--biotite pairs yielding an age error of between 13 and 23 m.y.; the reliability of analyses of less than 0.05 gm of biotite is confirmed by two duplicated analyses from the Gardar Province. 161001 biotite gave 1265 and 1263 m.y., and 161003 gave 1277 and 1278 m.y. The very close agreement of the ages is considered to be fortuitous, but they do show that results of this type are reproducible. The age and error on whole-rock--biotite pairs contains no internal statistical test of significance as does an isochron, and the merits of any particular result must be assessed entirely on geological grounds.

B.8 Processing of data

The ratios $^{85}\text{Rb}/^{87}\text{Rb}$, $^{86}\text{Sr}/^{84}\text{Sr}$, $^{87}\text{Sr}/^{86}\text{Sr}$ (corrected for ^{87}Rb) and $^{88}\text{Sr}/^{86}\text{Sr}$ are calculated by the on-line computer during the runs and all Sr isotope ratios are standardised onto one value of $^{88}\text{Sr}/^{86}\text{Sr}$.

The programme for the 'isotope dilution calculation' removes from the ratios the effect of the small amount of ^{85}Rb (0.79%) present in the ^{87}Rb spike, and ^{86}Sr (0.104%), ^{87}Sr (0.017%) and ^{88}Sr (0.086%)

present in the ^{84}Sr spike. The programme standardises all Sr isotope ratios to $^{88}\text{Sr}/^{86}\text{Sr} = 8.37521$ (Nier, 1938), and calculates $^{87}\text{Rb}/^{86}\text{Sr}$, ppm Rb and ppm Sr by the principle outlined in section B.1.

The programme for two-error regression of potential isochron data is slightly modified after York (1966, 1969). Analytical errors entered are all at the 1σ level and are explained in section B.7: all quoted age and initial ratio errors are at the 2σ level. The value $[\text{SUMS}/(n-2)]^{1/2}$ is reported for each set of data regressed: SUMS itself is a quantity derived as the sum of the squares of all deviations of points from the best-fit line. If $[\text{SUMS}/(n-2)]^{1/2} = 1$, the deviations correspond exactly to estimated analytical uncertainty; for higher values of $[\text{SUMS}/(n-2)]^{1/2}$, the scatter about the line must contain an additional non-analytical component, usually of geological origin. Brooks et al. (1972) have suggested a $[\text{SUMS}/(n-2)]^{1/2}$ value in the region of 2.6 as defining the approximate upper limit for which data can be regarded statistically as defining an isochron; this limit is used in a general way in this study.

APPENDIX C: LOCATION AND DESCRIPTION OF SAMPLES

Map references refer to the Swedish national grid. All samples from Greenland refer to Geological Survey of Greenland sample files. The letter S against a sample number denotes that a thin-section is available. For details of collecting methods see appendix B; analytical data are listed in appendix D under the same numbers as used in this section (i.e. data for rocks of section C.5 is listed in section D.5).

C.1 Arsuk lamprophyre dykes (fig. 3.2)

- 81119 Centre of a 3 m mica-rich ENE-trending lamprophyre. Wave-cut platform exposure, Kajartalik, 2 km W of Arsuk.
- 86192 Centre of a 1.5 m banded ENE-trending lamprophyre. Coastal exposure, Fortuna Havn, 1 km E of Arsuk.

C.2 Dolerite dykes of the Ivigtut region (fig. 3.2)

- 161001 S Medium-coarse olivine dolerite, 9 m from contact of 100 m 'BD1' or 'Grønnedal Master Dyke'. Coastal cliff exposure 6 km SW of Ivigtut.
- 161002 S Medium-coarse olivine dolerite, centre of of 50 m 'EDO' (southern of two). Coastal exposure 3 km WSW of Ivigtut.
- 161003 S Pegmatitic olivine dolerite 2 m from contact of 30 m 'BDO' (northern of two). Coastal cliff exposure 3 km W of Ivigtut.

C.3 Dolerite dykes of the Julianehåb region (fig. 3.2)

- 186506 S Medium-coarse olivine dolerite, near centre of 50 m 'BDO' (western of three prominent dykes). Coastal exposure on Itivdlingujuk, western Tugtutôq.

186564 S Medium olivine dolerite, near centre of 90 m 'BDO' (continuous with the central of the three on Tugtutôq). Blasted exposure, W side of Julianehåb Harbour.

C.4 Gåstorp, Bålgviken and Forskilen dykes (fig. C.1)

- SÖ-49 S Medium olivine dolerite at centre of 20 m dyke. Natural exposure 50 m W of Forskilen 10G SO 873672.
- SÖ-53 S Plagioclase-phyrlic dolerite of 3 m dyke. Railway cutting, Bålgviken 10G SO 833690.
- SÖ-54 S Fine dolerite, associated with dolerite carrying plagioclase megacrysts and anorthosite xenoliths in a 3 m dyke. Natural exposure 200 m SW of Gåstorp 10G SO 761700.

C.5 Hällefors dyke (fig. C.1)

- SÖ-21 S Medium-coarse olivine dolerite of southern marginal series. Roadcut SW of Hälleforsnäs 10G SO 841570.
- SÖ-23 S Medium-coarse laminated 'porphyrite'; plagioclase and pyroxene phenocrysts in a dark groundmass, intrudes marginal series. SW of Hälleforsnäs 10G SO 842570.
- SÖ-24 S Coarse porphyrite, unlaminated. As SÖ-23.
- SÖ-26 S Fine-medium porphyrite with reddened feldspars. Roadcut SW of Hälleforsnäs 10G SO 843571.
- SÖ-27 S As SÖ-26, 50 m further NE.
- SÖ-28 S As SÖ-27, 100 m further NE.
- SÖ-29 S Medium porphyrite with reddened feldspars. Roadcut 200 m NE of SÖ-28.
- SÖ-31 S Fine porphyrite with reddened feldspars. Roadcut SE of Hälleforsnäs 10G SO 873569.

SÖ-36 Fine granophyre of 6 cm rheomorphic vein at contact between fine dolerite of southern marginal series and country rock gneiss. Sample contains small pieces of dolerite chill. Roadcut SE of Hälleforsnäs 10G SO 882560.

SÖ-41 S Coarse laminated porphyrite. Quarry S of Svalboviken 10G SO 767591.

C.6 Breven dyke (fig. C.2)

SÖ- 2 Coarse olivine dolerite, near western end of dyke. Roadcut 9F NV 148379.

SÖ- 3 S As SÖ- 2. Roadcut 9F NV 148380.

SÖ- 4 As SÖ- 3.

SÖ- 5 Coarse olivine dolerite, serpentinitised and chloritised. Roadcut, Bottorp 9E NV 230380.

SÖ- 8 Coarse granophyre. Natural exposure W of Högsätter 9F NO 246377.

SÖ- 9 Coarse granophyre, with incipient weathering. Natural exposure W of Högsätter 9F NO 245377.

SÖ-10 S Coarse dolerite hybridised by granophyre. Natural exposure W of Högsätter 9F NO 244377.

SÖ-11 As SÖ-10. Natural exposure 9F NO 244378.

SÖ-12 S Coarse granophyre. Natural exposure W of Johannisberg 9F NO 319350.

SÖ-13 S Fine granophyre, vein cutting SÖ-12.

SÖ-14 S Coarse dolerite-granophyre hybrid. Roadcut W of Johannisberg 9F NO 320352.

SÖ-15 Coarse dolerite, rusty. Natural exposure at eastern end of dyke near Eriksberg 9F NO 386340.

- SÖ-16 Medium granophyre. Natural exposure 15 m W of SÖ-15.
- SÖ-18 Fine granophyre, vein cutting medium granophyre 2 m from SÖ-16.

C.7 Tuna dykes (fig. C.3)

- DS-16 Porphyritic dolerite: phenocrysts of plagioclase in an ophitic groundmass. Centre of a 7 m dyke at Halvarsgårdar 60.4°N, 15.5°E.
- DS-17 S Porphyritic intermediate rock: phenocrysts of alkali feldspar and hornblende in a trachytic groundmass. Centre of a 5 m dyke 10 m from DS-16.
- DS-18 Aphyric dolerite. Centre of a 5 m dyke 100 m E of DS-17.
- DS-20 S Quartz-K-feldspar porphyry, somewhat reddened. Centre of a 3 m dyke, natural exposure 1 km N of Österby 60.4°N, 15.6°E.
- DS-21 S Quartz-K-feldspar porphyry, somewhat reddened. Centre of a 7 m dyke 25 m E of DS-20.
- DS-22 Quartz-K-feldspar porphyry, reddened. Centre of a 6 m dyke 15 m E of DS-21.

C.8 Gimån area sills (fig. C.4)

- J- 4 Fine olivine dolerite 1.5 m from contact with granite. Roadcut 18G SV 531583.
- J-12 Medium olivine dolerite. Roadcut 18G SV 567550.
- J-18 S Medium olivine dolerite. Roadcut 18G SV 588548.
- J-19 Medium olivine dolerite. Roadcut 18G SV 472606.
- J-25 Medium olivine dolerite. Roadcut 18G SV 455655.
- J-29 S Pegmatitic dolerite. Roadcut 18G SV 457641.
- J-33 Medium granophyric rock near contact of olivine dolerite with granite, occurrence suggests rheomorphic origin. Roadcut 18G

SV 430641.

- J-34 S Medium dolerite with red feldspars, hybrid between dolerite and rheomorph. As J-33.
- J-35 S As J-34.
- J-36 Fine olivine dolerite 2 m from contact. 18G SV 428639.
- J-41 Pegmatitic dolerite. 18G SV 427731.

C.9 Sundsjö dyke (fig. C.4)

- J-55 S Medium-coarse rheomorphic granite of 50 cm dyke cutting olivine dolerite. Roadcut on highway E75 2 km S of Sundsjö 18F NV 086798.
- J-56 S Fine-medium rheomorphic granite. As J-55.
- J-58 S Coarse olivine dolerite. Centre of dyke at same locality.

C.10 Älvho sill (fig. C.5)

- DN-315 S Medium-coarse olivine dolerite. Riverbed exposure at Älvho 61.5°N, 14.7°E.
- DN-316 Medium olivine dolerite. 4 m W of DN-315.
- DN-317 S Medium-coarse dolerite with slightly reddened feldspars. 4 m NW of DN-315.
- DN-323 Medium-coarse monzonite. Slightly weathered roadcut 500 m W of hill Pilkalampinoppi 61.6°N, 14.7°E.
- DN-324 Medium-coarse monzonite. Slightly weathered roadcut 1 km S of hill Siikamäki 61.5°N, 14.7°E.
- DN-325 S Medium-coarse olivine dolerite, near eastern margin of sill. Roadcut 1.2 km S of hill Siikamäki.
- DN-326 Medium-coarse monzonite. Natural exposure on western flank of hill Siikamäki.

DN-327 S As DN-326.

DN-328 As DN-326.

C.11 Emådalen sill (fig. C.5)

DN-301 Medium-coarse monzonites, from a large homogenous body either to DN-312 within or marginal to olivine dolerite. Samples are spread at regular intervals along the river Aman and along a roadcut at Emådalen 61.3°N, 14.7°E.

DN-329 S Medium-coarse olivine dolerite. Roadcut 400 m S of river at Emådalen.

DN-330 S Fine-medium olivine dolerite with interstitial granophyre, 1 m from contact with acid volcanics. Roadcut 500 m S of river at Emådalen.

C.12 Älvdalsåsen sill (fig. C.5)

DN-51 S Medium olivine dolerite. 1500 m S of river Österdalälven, in stream Ugsiån; 2 km S of Älvdalsåsen 61.4°N, 13.8°E.

DN-56 Medium olivine dolerite. 200 m S of Österdalälven, in stream Ugsiån.

DN-60 Fine biotite-rich olivine dolerite cutting coarse. 150 m S of Österdalälven.

DN-64 S Coarse olivine dolerite. 140 m S of Österdalälven.

DN-67 Extremely coarse olivine dolerite developed for ca. 100 m S from Österdalälven.

C.13 Mossiberg sill (fig. C.5)

DN-69 S Medium olivine dolerite. Roadcut on northern lakeshore at Mossiberg 61.5°N, 13.9°E.

DN-71 S Fine olivine dolerite rich in biotite, associated with rheomorphic granite veins. Northern lakeshore, natural exposure close to contact.

DN-74 S Coarse olivine dolerite. Roadcut at lakeshore 50 m W of DN-69.

DN-78 Coarse olivine dolerite. Natural exposure at NW corner of lake, probably near contact.

C.14 Idre sill (fig. C.6)

DN-17 Olivine dolerite. Quarry at foot of hill Vålåberget on north side of lake Idresjön, 5 km SE of Idre 61.8°N , 12.8°E .

DN-21 As DN-17.

DN-24 Olivine dolerite, serpentinised. Natural cliff exposure on S side of lake Idresjön 5 km SE of Idre.

DN-30 S Olivine dolerite, serpentinised. 50 m W of DN-24.

DN-31 S Olivine dolerite, serpentinised. 20 m W of DN-30.

C.15 Bornåssjön sill (fig. C.6)

71-15 S Olivine dolerite. Roadcut 500 m S of lake Bornåssjön 61.5°N , 13.0°E .

C.16 Karlshamn dyke (fig. C.7)

BL- 3 S Medium-coarse olivine-hypersthene dolerite. Centre of dyke in quarry, 1.5 km S of Karlshamn 3E NO 909229.

BL- 7 Medium dolerite, 6 cm from eastern contact with granite. Roadcut across entire 250 m width of dyke 1 km S of Karlshamn 3E NO 911236.

BL- 8 Medium dolerite with reddened feldspars, indicating contamination. 4 cm from contact, 6 cm along strike from BL- 7.

- BL- 9 Fine-medium dolerite 0-2 cm from contact hybridised by melted granite. Adjacent to BL- 8.
- BL-10 Medium dolerite 1 m from eastern contact. Same traverse as BL- 8,9.
- BL-15 Medium-coarse olivine-hypersthene dolerite. Interior sample, 50 m from western contact, roadcut 3E NO 910236.
- BL-18 Pegmatitic patch, feldspar-rich with prominent needles of clinopyroxene. Interior of dyke 20 m from western contact, roadcut 3E NO 910236.

C.17 Bräkne-Hoby dyke (fig. C.7)

- BL-24 S Medium-coarse olivine-hypersthene dolerite. Centre of dyke in quarry 5 km NE of Bräkne-Hoby 3F SV/NV 109351.
- BL-27 S Medium-coarse olivine-hypersthene dolerite. Centre of dyke 15 m S of BL-24.
- BL-31 Two-phase fine dolerite. Loose block in quarry 3F SV/NV 109350.
- BL-32 Pegmatitic dolerite with turbid plagioclase. Loose block in same quarry as BL-31.
- BL-35 Fine dolerite. 1 m dyke cutting granite, part of a swarm equivalent to the 50 m dyke of the quarry 6 km to the N. Dyke carries xenoliths/xenocrysts of granite and K-feldspar. Roadcut S of Bräkne-Hoby 3F SV/NV 083258.
- BL-36 Fine dolerite. 20 cm from contact of 6 m dyke in the same traverse as BL-35, 3F SV/NV 085256.
- BL-40 S Fine dolerite. Centre of the 6 m dyke of BL-36.

C.18 Tärnö dyke (fig. C.7)

- BL- 1 Fine dolerite contaminated with ca. 50% fine, loose arkosic sand. Centre of 50 m dyke on highest point of the island Tärnö, 7 km SE of Karlshamn, no map ref. available.
- BL- 2 S Totally recrystallised xenolith consisting of quartz, alkali feldspar and epidote. Centre of dyke on northern coast of Tärnö, 1.5 km N of BL- 1.
- BL-ARX S As BL-2, from Tärnö. Sample on loan, no detailed locality available.

C.19 Forserum sill (fig. C.8)

- AL- 2 Fine dolerite, chill against quartzite of Almesåkra Formation. Quarry 2 km NE of Forserum 6E NV 702971.
- AL- 6 S Medium hypersthene dolerite, interior of sill. Quarry, as AL- 2.
- AL- 7 S As AL- 6.
- AL-10 S Coarse hypersthene dolerite of interior of sill, with interstitial red granophyre. Quarry, as AL- 2.
- AL-12 S Medium hypersthene dolerite, interior of sill. Quarry, as AL- 2.
- AL-15 S Pegmatitic dolerite with sericitised plagioclase and interstitial red granophyre. Quarry, as AL- 2.
- AL-17 S As AL-15.

C.20 Nilstorp sill (fig. C.8)

- AL-21 Medium hypersthene dolerite. Small quarry on highway 33 Nässjö-Eksjö 6E NO 907888.
- AL-22 S Medium hypersthene dolerite. 4 m from AL-21.

AL-29 Fine dolerite 20 cm from contact with siltstone of the Almesåkra Formation. Roadcut 800 m W of the small quarry, 500 m E of farm Nilstorp 6E NO 900888.

C.21 Årby dyke (no detailed location map, see figs. 4.4, 4.5)

SO-50 S Medium olivine dolerite, serpentinitised. Centre of 5 m NW-trending dyke in railway cutting 900 m S of Årby Hållplats 10G SO 832714.

C.22 Falun dyke (fig. C.9)

DS- 1 S Fine dolerite of 1 m satellitic dyke 15 m NE of the main dyke. Roadcut, Bispberg 60.3°N, 15.8°E.

DS- 2 Medium olivine dolerite of main dyke at Bispberg, nearest exposed to western contact.

DS- 9 S Medium olivine dolerite, centre of main dyke at Bispberg.

DS-29 Medium olivine dolerite, approximately central. Natural exposure on eastern shore of lake St. Vällån, 2 km W of Falun 60.5°N, 15.6°E.

DS-31 S Medium olivine dolerite, approximately central. 10 m across strike from DS-29.

C.23 Västra Nornäs dyke (fig. C.10)

D75-26 S Olivine dolerite, serpentinitised. Centre of dyke running 2 km W of Västra Nornäs 61.4°N, 13.1°E.

C.24 Bunkris dyke (fig. C.10)

DN-37 S Dolerite with blades of clinopyroxene, possibly phenocrysts, and reddened and sericitised feldspars. Centre of ca. 500 m

dyke, roadcut at Bunkris 61.4°N, 13.4°E.

DN-42 S Dolerite, feldspars only slightly reddened. 40 m S of DN-37.

C.25 Vaggeryd syenite (fig. C.11)

- V- 2 S Strongly schistose coarse quartz-rich syenite with considerable muscovite, biotite and epidote. Roadcut 16 km N of Vaggeryd at Röhshult 6D NO 451883.
- V- 5 Igneous-textured coarse syenite, mainly K-feldspar, black amphibole and quartz. Roadcut 8 km N of Vaggeryd, 1 km NW of Krängsberg 6E NV 495820.
- V- 6 S Fine-medium schistose syenite, mainly K-feldspar, black amphibole and quartz. Roadcut 7 km N of Vaggeryd, 500 m SW of Krängsberg 6E NV 498811.
- V- 8 S Igneous-textured medium-coarse syenite with a weak directional fabric. Roadcut 2 km SSE of Vaggeryd 6E SV 492716.
- V- 9 More feldspathic facies of V- 8 locality.
- V-10 Strongly schistose syenite, very similar to V- 2. Roadcut 6 km E of Vaggeryd 6E SV 539722.
- V-14 S Mafic syenite within feldspathic syenite, mainly hornblende and plagioclase. Roadcut 3 km E of Vaggeryd 6E NV 517726.
- V-15 As V-14.
- V-16 Strongly schistose syenite, similar to V-2. Roadcut 2 km E of Vaggeryd 6E NV 511728.
- V-17 As V-16.
- V-18 S Igneous-textured syenite, massive, mainly perthite. Roadcut 1 km E of Vaggeryd 6E NV 498730.

C.26 Country rocks from southern Sweden (figs. C.8, C.7)

- F- 3 S Coarse grey Våxjö granite. Roadcut near Järnsås, 11 km NE of Forserum (no map ref. available).
SPG-1 S Medium Spinkamåla granite. Quarry W of Ugglesjön, 14 km NW of Karlshamn 3E NO 795324.

C.27 Särna alkaline complex (fig. C.6)

- DN- 1 S Fine-medium schlieric syenite. Natural exposure on crest of SSE ridge 900 m from summit of hill Siksjöberget 61.7°N, 12.9°E.
DN- 2 S Fine-medium schlieric syenite. 30 m from DN- 1.
DN- 5 S Medium-coarse schlieric syenite. 5 m from DN- 2.
DN-9A Feldspathic portion of DN-9, coarse (pegmatitic) syenite. 4 m from DN- 5.
DN-10 Medium laminated non-schlieric syenite. Roadcut 2 km S from summit of Siksjöberget.
DN-11 Medium laminated syenite. Natural exposure 500 m E of summit of hill Ekorråsen, 3 km W of Siksjöberget.

C.28 Anorthosites and gabbros from south Greenland (fig. 6.1)

- 30752 Coarse olivine-bearing anorthosite with crude lamination, from northern marginal gabbro of the most northerly of the Younger Giant Dykes, Assorutit, eastern Tugtutôq.
30753 As 30752.
30755 Olivine gabbro from northern marginal zone. 500 m W of Assorutit.
40518 Leucocratic olivine gabbro, southern gabbro zone. E end of lake Krydssø, 2 km WSW of coast.

- 40522 Relatively olivine-rich anorthosite, northern marginal zone.
Assorutit.
- 50220 Anorthosite, northern marginal zone. Assorutit.
- 50221 As 50220.
- 50224 Olivine gabbro, northern marginal zone. 2 km WSW (inland) of
Assorutit.
- 81156 Olivine gabbro, southern marginal zone. Shore exposure 1 km S
of Assorutit.
- 85977 As 81156.

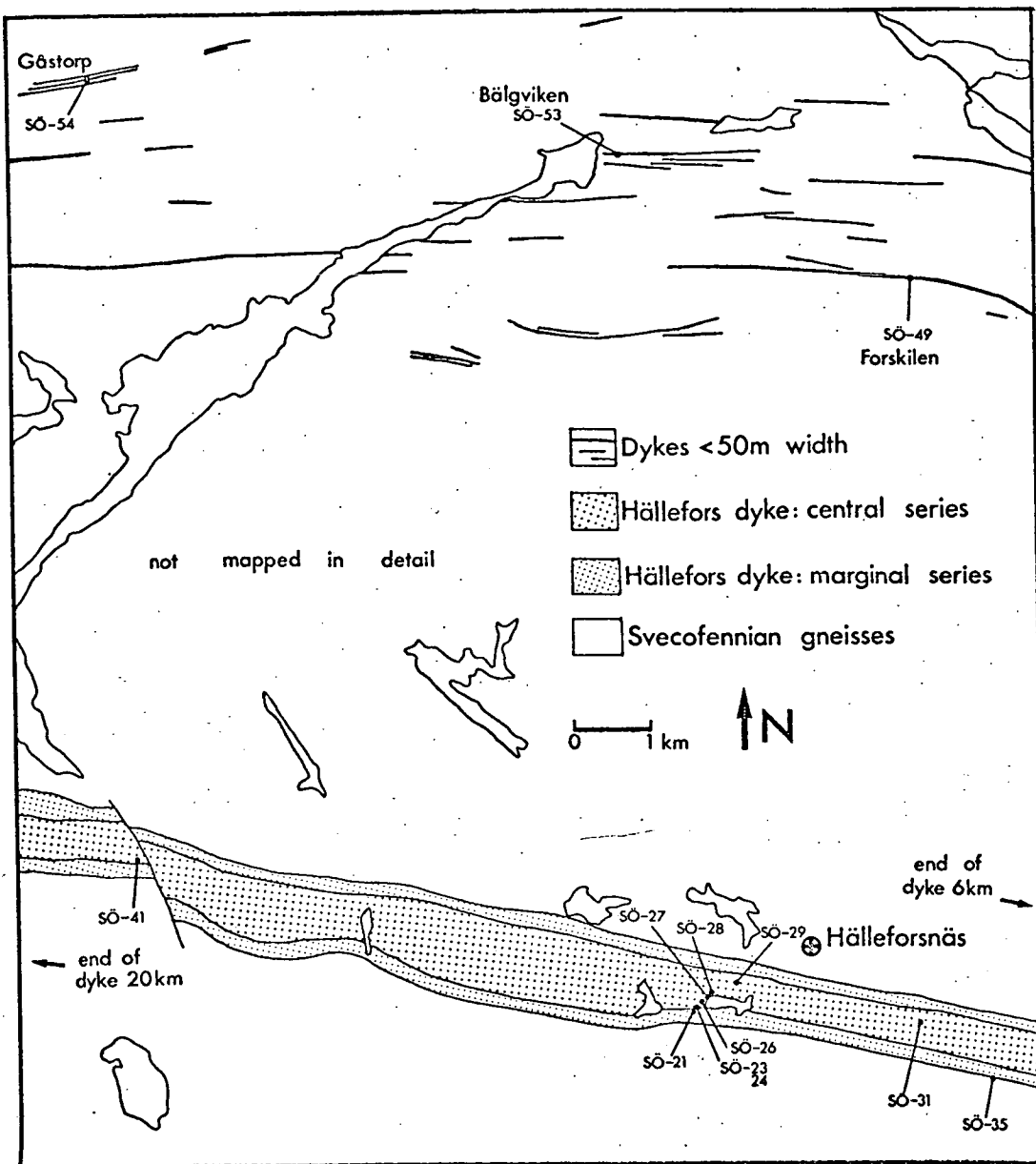


Fig. C.1 Sample localities for the Gästorp, Forskilen and Hällefors dykes, East-West system. For map location see fig. 4.2.

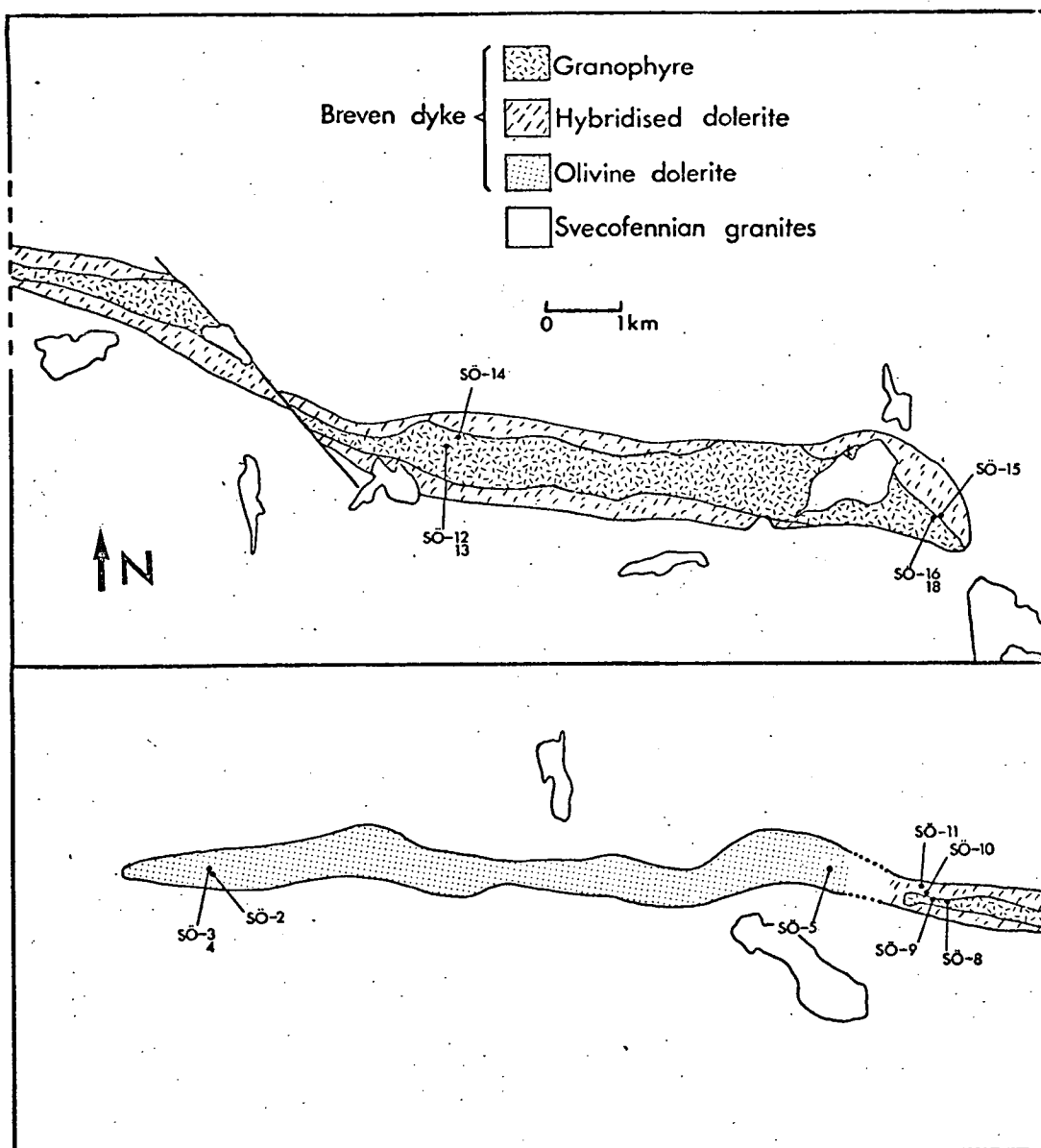


Fig. C.2 Sample localities for the Breven dyke, East-West system. For map location see fig. 4.2.

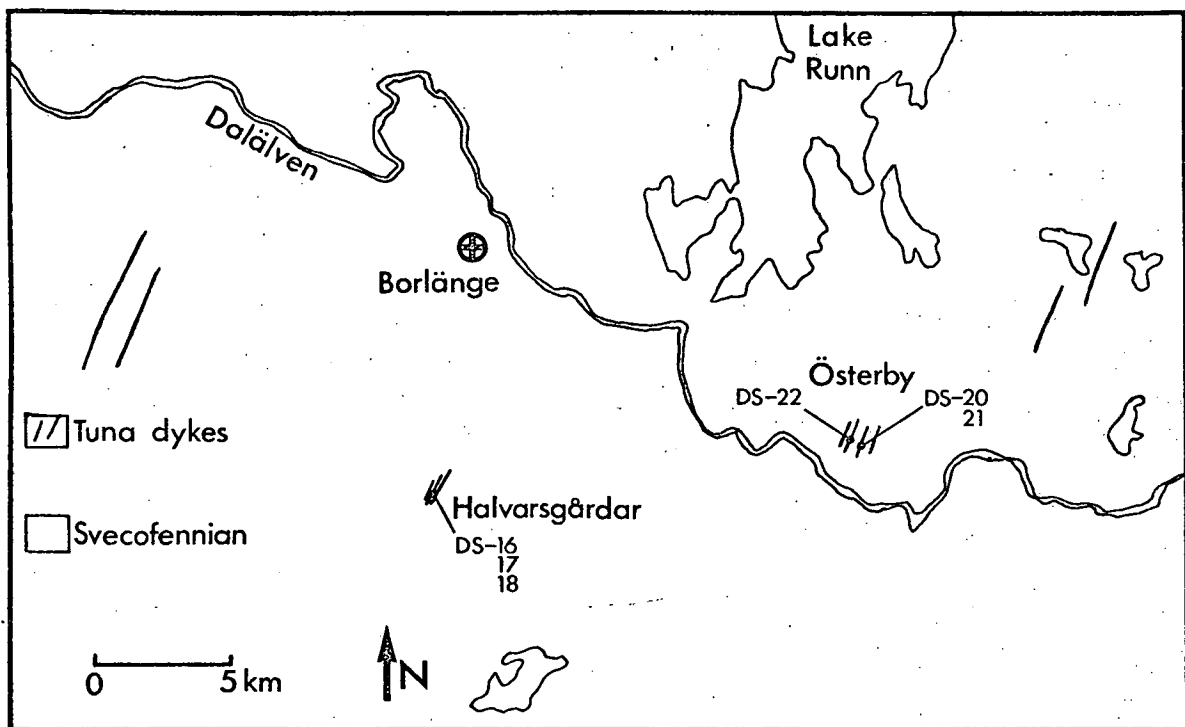


Fig. C.3 Sample localities for the Tuna dykes, southern Dalarna. For map location see fig. 4.2.

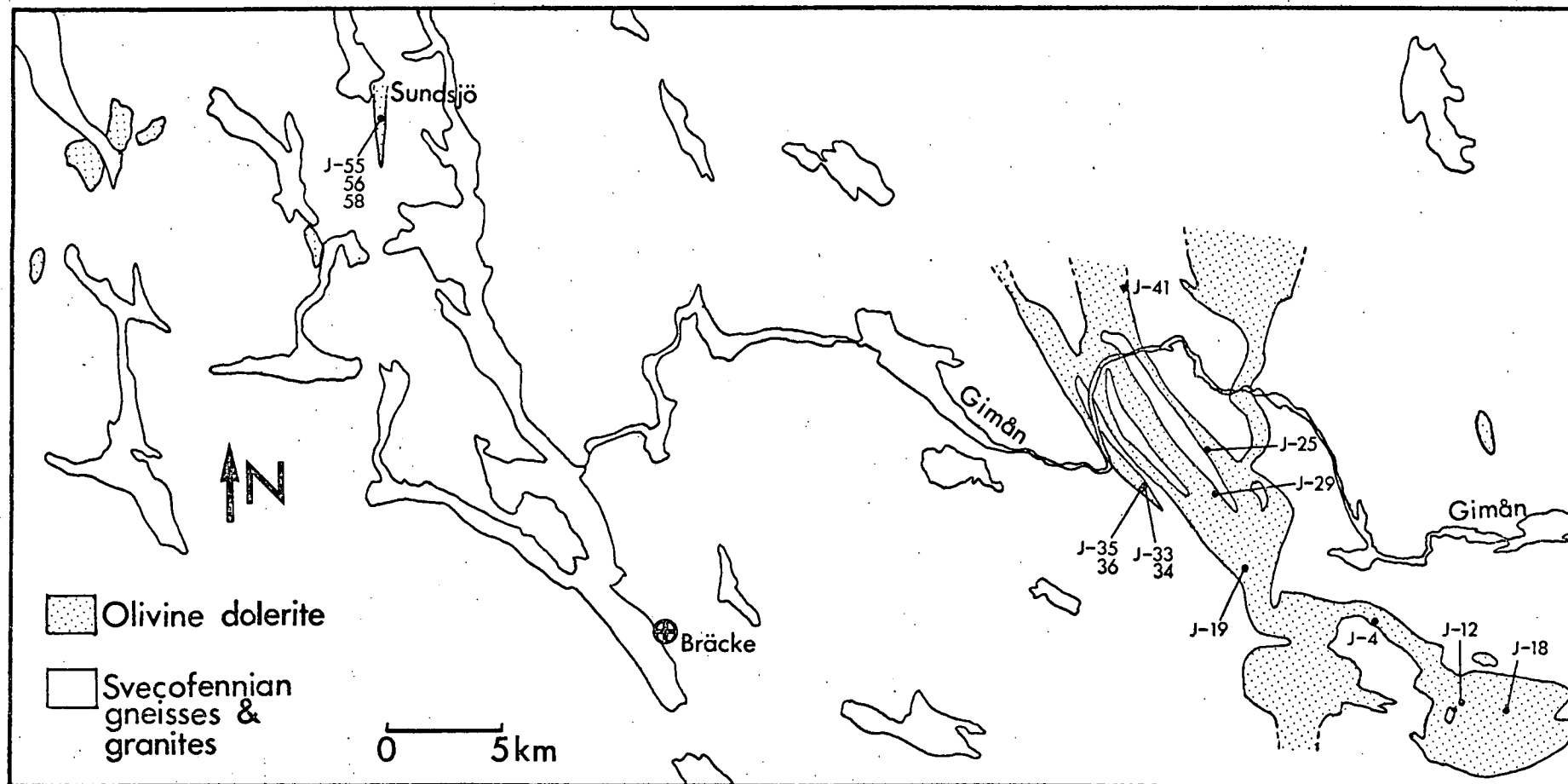


Fig. C.4 Sample localities for the Gimån and Sundsjö olivine dolerites, Jämtland. For map location see fig. 4.2.

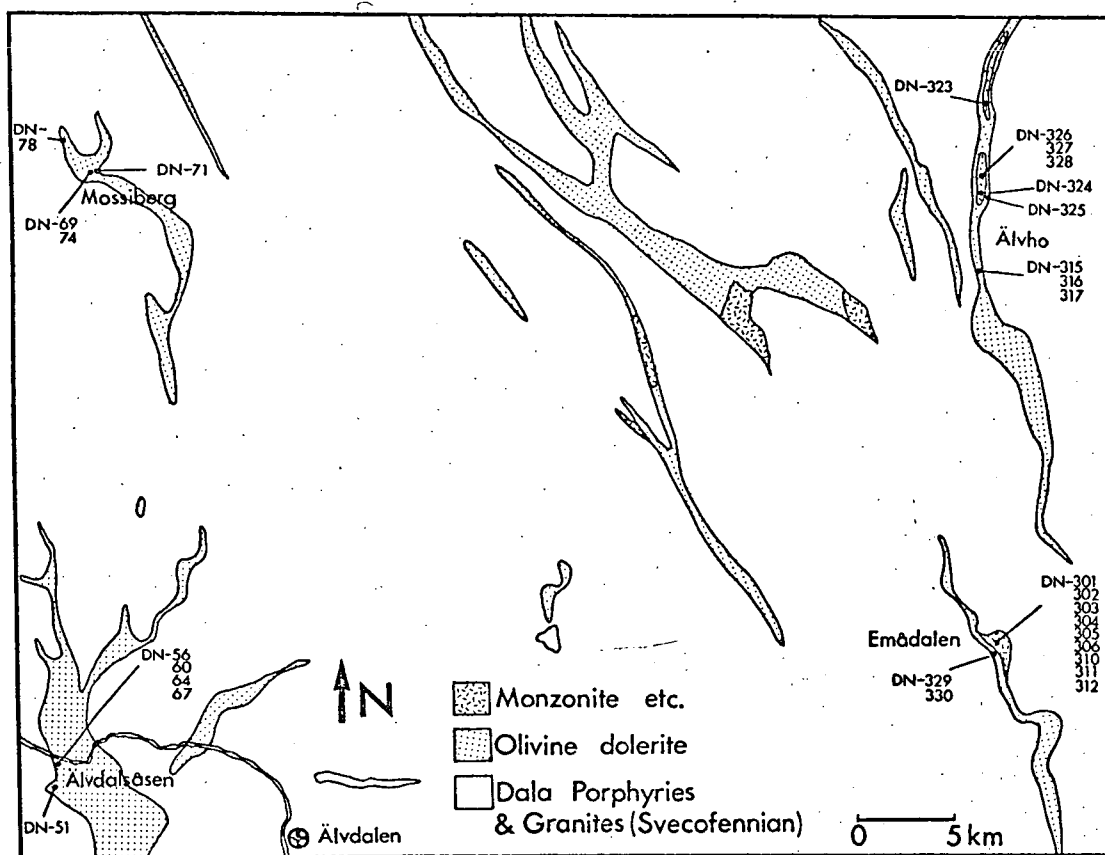


Fig. C.5 Sample localities for the Älvho, Emådalen, Älvdalsåsen and Mossberg olivine dolerite sills, northern Dalarna. For map location see fig. 4.2.

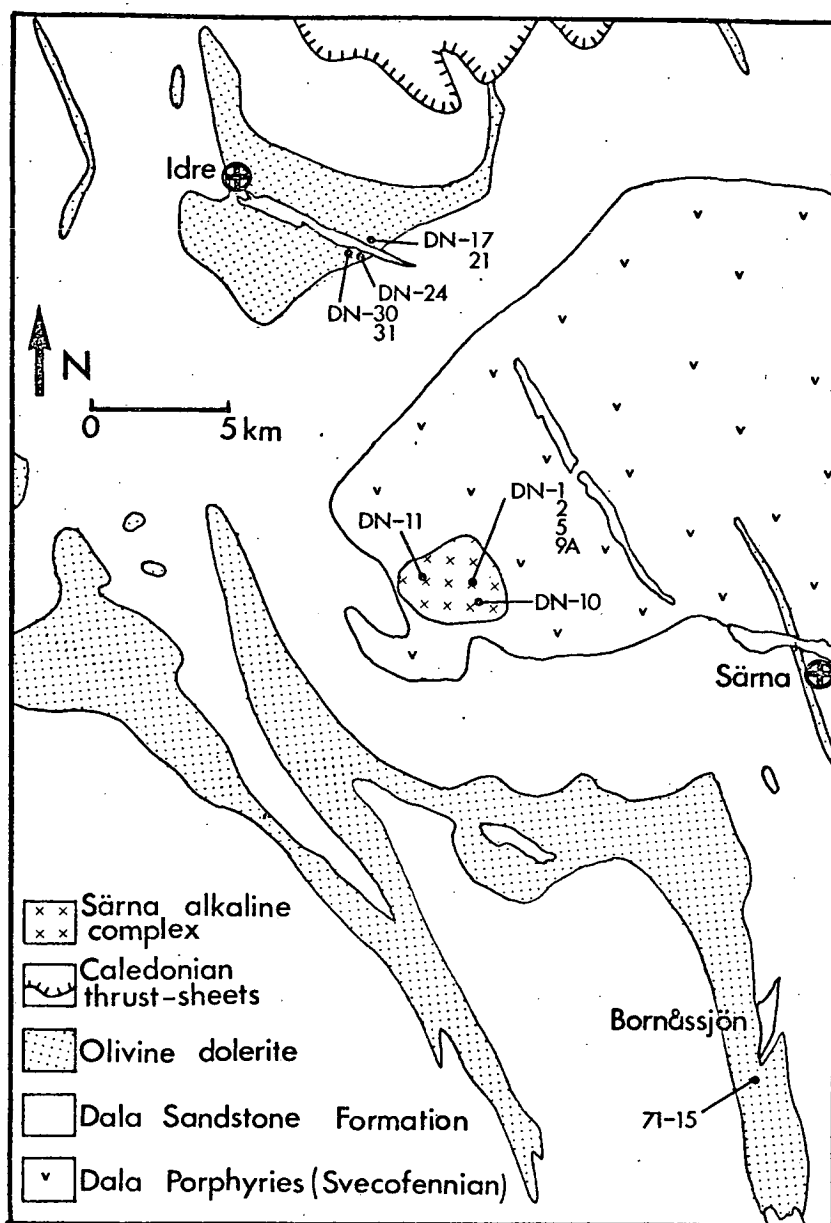


Fig. C.6 Sample localities for the Idre and Bornåssjön olivine dolerite sills, northern Dalarna. Sample localities for the Särna alkaline complex are also shown. For map location see fig. 4.2.

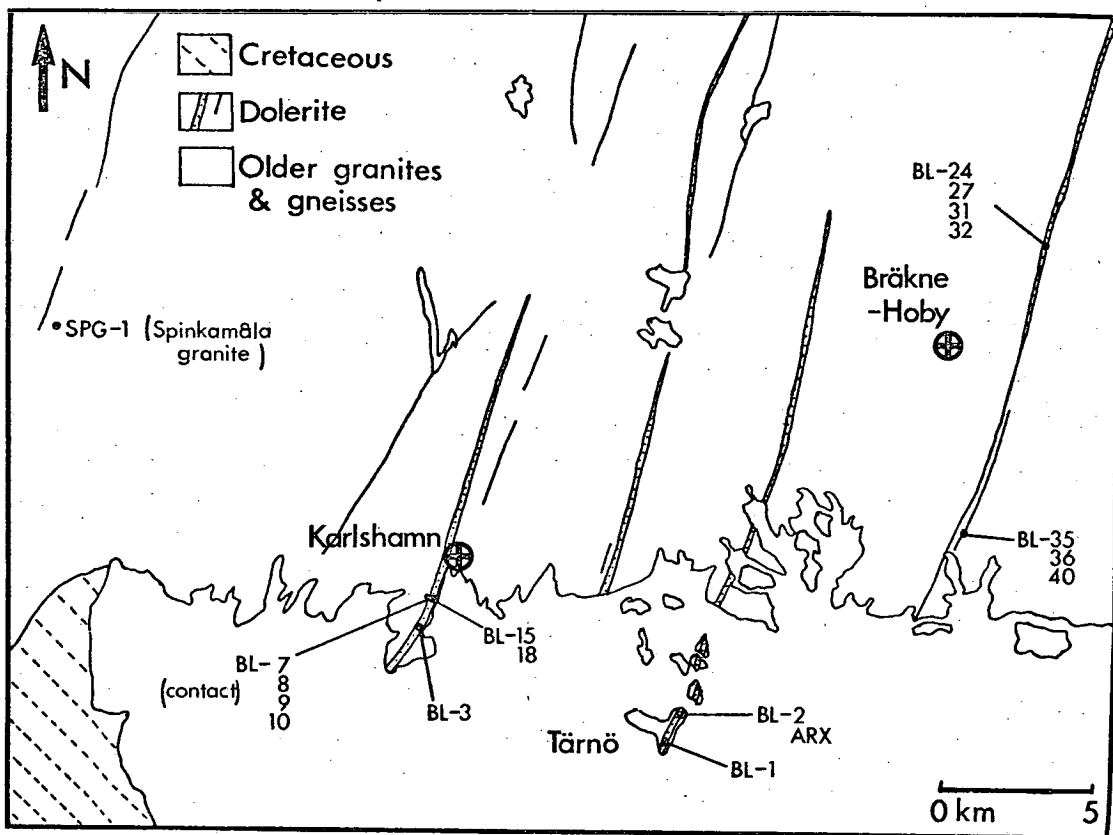


Fig. C.7 Sample localities for the Karlshamn, Tärnö and Bräkne-Hoby dykes of Blekinge, lying parallel to the Sveconorwegian Front. For map location see fig. 4.4.

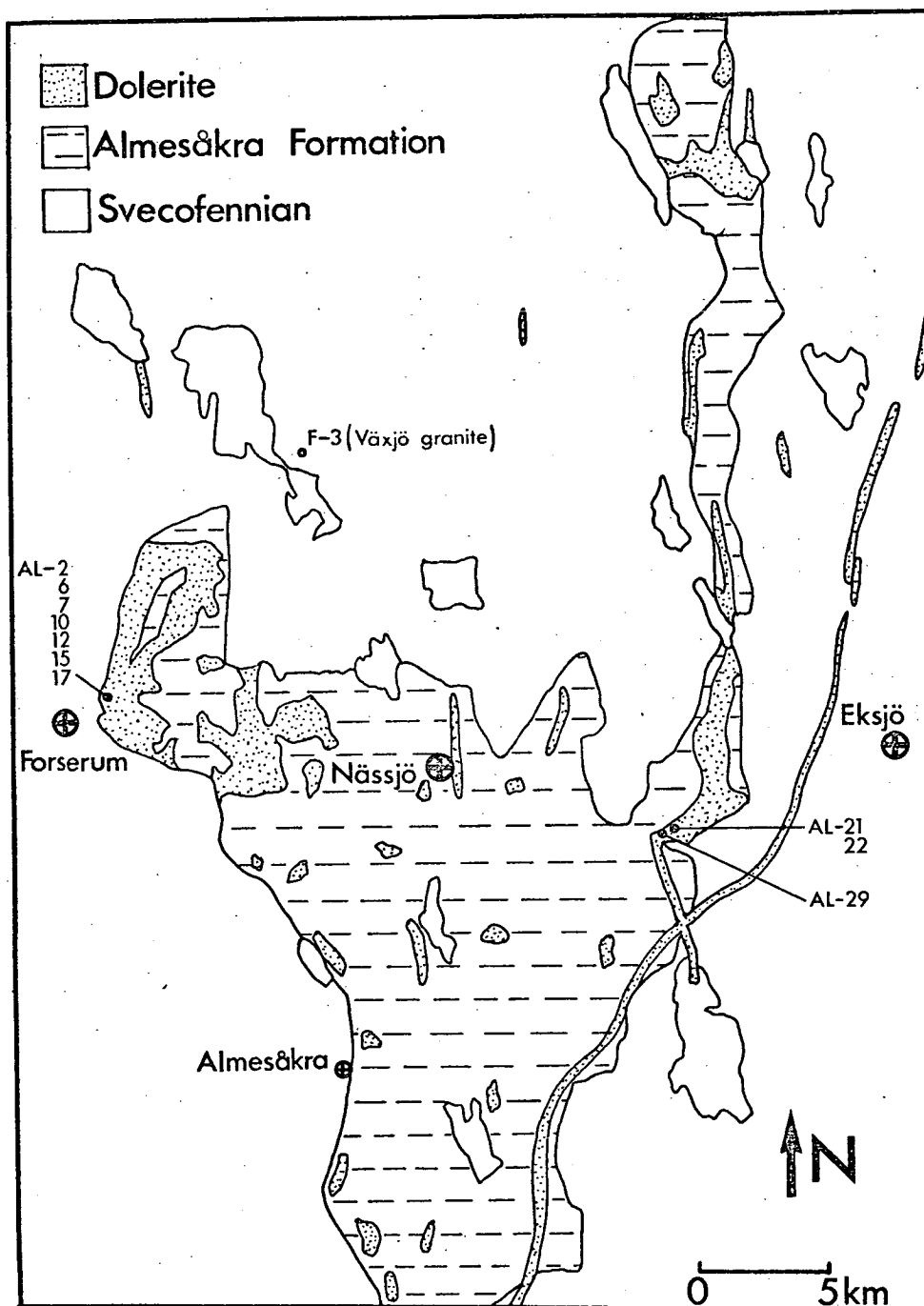


Fig. C.8 Sample localities for the Forserum and Nilstorp sills of the Almesåkra region, showing the association with dykes parallel to the Sveconorwegian Front. For map location see fig. 4.4.

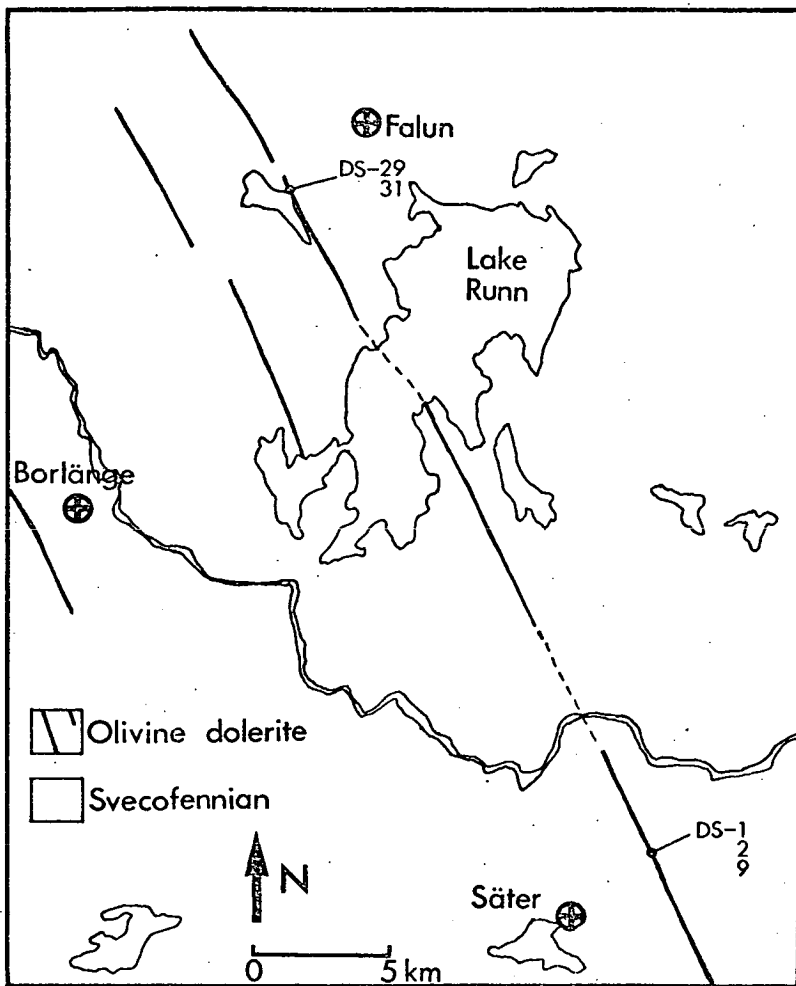


Fig. C.9 Sample localities for the Falun dyke of northern Dalarna, lying parallel to the Sveconorwegian Front. For map location see fig. 4.4.

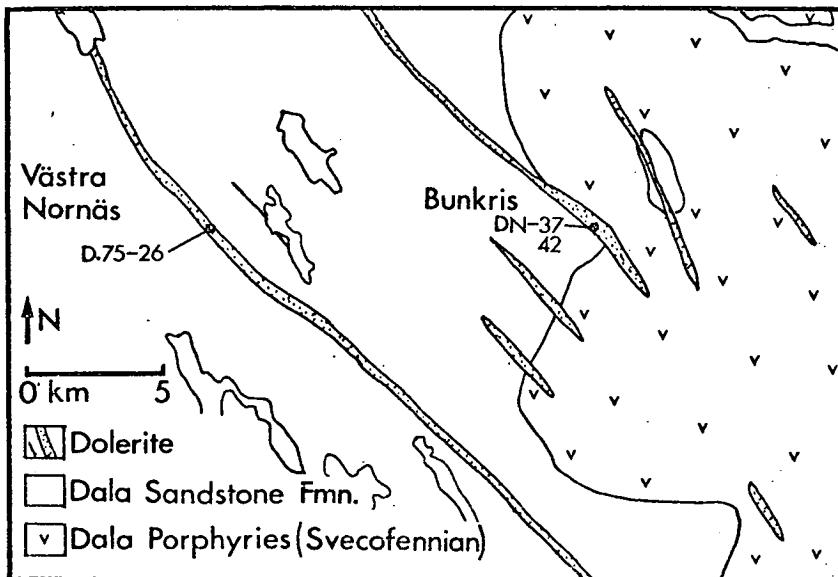


Fig. C.10 Sample localities for the Västra Nornäs and Bunkris dykes of northern Dalarna, mapped as lying parallel to the Sveconorwegian Front. For map location see fig. 4.4.

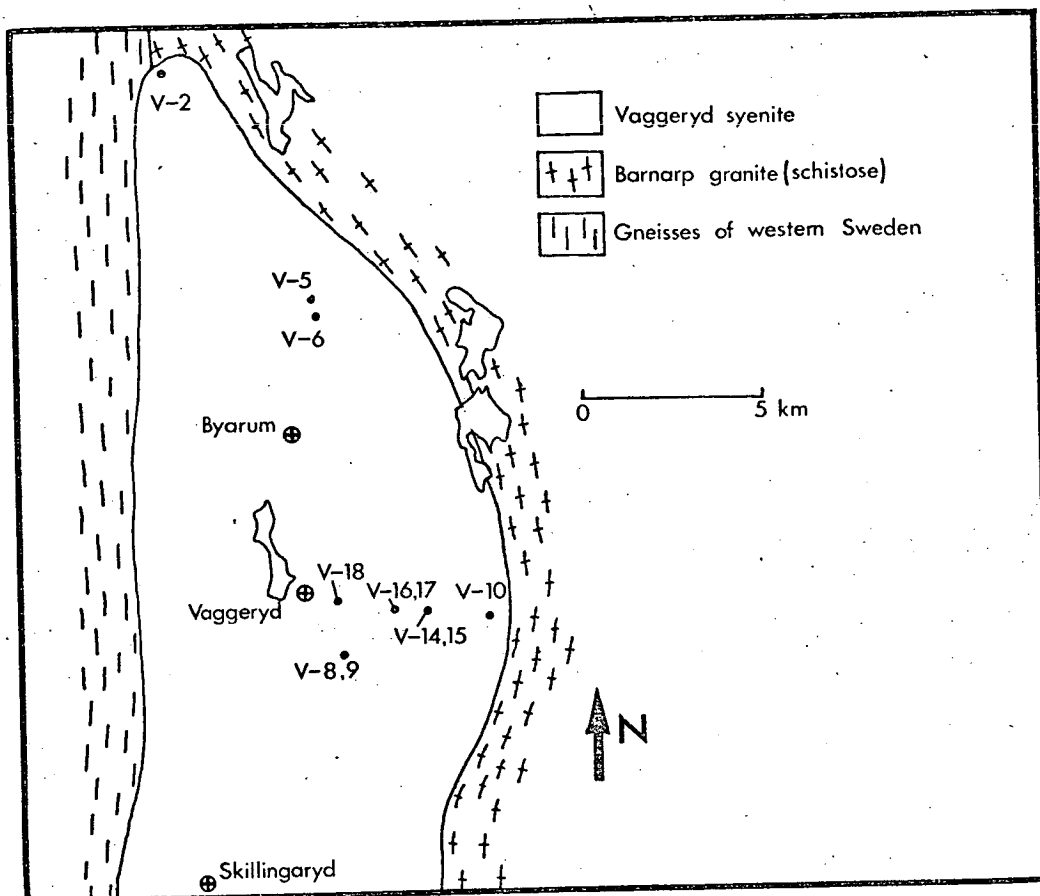


Fig. C.11 Sample localities for the Vaggeryd syenite: the northern half of the body is shown. For map location see fig. 4.4.

APPENDIX D: RB-SR ANALYTICAL DATA

Rb-Sr analytical data are presented here for each intrusion in the order in which they first appear in the text. Section numbers correspond to those of appendix C.

Analytical errors indicated for $^{87}\text{Sr}/^{86}\text{Sr}$ are at the 1σ level; where no error is shown, an assigned 1σ error of 0.04% is used.

Repeated analyses are indicated by * ; where appropriate, means of duplicates are used for interpretation purposes (see section B.7 for discussion).

Sample code:-

AB	= albite	AEG	= aegirine
AP	= apatite	BI	= biotite
CAN	= cancrinite	CPX	= clinopyroxene
EP	= epidote	F	= feldspar
HB	= hornblende	MCL	= microcline
MUSC	= muscovite	CR	= orthoclase
SPH	= sphene	WR	= whole-rock

D.1 Arsuk lamprophyre dykes (fig. 3.2)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
81119	lamprophyre	WR	114.6	1123.9	0.2952	0.71027
		BI	371.8	357.4	3.0250	0.75826
86192	lamprophyre	WR	109.0	1305.4	0.2415	0.70793

BI 344.3 332.5 3.0102 0.75747

D.2 Dolerite dykes of the Ivigtut region (fig. 3.2)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
161001	Dolerite	WR	7.5	449.7	0.0484	0.70351
'	'	BI*	243.8	18.4	41.1491	1.43243
'	'	BI*	242.2	19.4	38.5807	1.38587
161002	Dolerite	WR	24.9	287.2	0.2511	0.70947
'	'	BI	661.2	18.4	126.8910	2.96909
161003	Dolerite	WR	29.1	259.8	0.3247	0.71195
'	'	BI*	428.5	12.3	122.3730	2.89775
'	'	BI*	434.9	12.2	125.7240	2.95939

D.3 Dolerite dykes of the Julianehåb region (fig. 3.2)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
186506	Dolerite	WR	24.2	372.7	0.1882	0.70792 $\pm$ 9
'	'	F1	18.1	643.9	0.0814	0.70622 $\pm$ 14
'	'	F5	24.4	527.4	0.1338	0.70743 $\pm$ 8
'	'	F7	19.2	493.8	0.1123	0.70748 $\pm$ 3
'	'	F8	34.1	437.5	0.2254	0.70941 $\pm$ 11

	F9	76.1	345.1	0.6388	0.71628 ± 8
	AP	6.2	444.6	0.0403	0.70529 ± 14
	BI	292.2	45.4	19.2310	1.04021
186564 Dolerite	WR	24.4	287.3	0.2462	0.70884 ± 10
	F1	15.0	553.9	0.0781	0.70626 ± 7
	F4	24.4	499.6	0.1410	0.70753 ± 7
	F6	33.3	467.6	0.2064	0.70919 ± 11
	F7	52.5	411.0	0.3696	0.71206 ± 8
	F8	105.8	253.1	1.2121	0.72560 ± 9
	AP	4.4	400.2	0.0319	0.70517 ± 13
	BI	469.9	32.1	45.5641	1.49545

D.4 Gåstorp, Bålgviken and Forskilen dykes (fig. C.1)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
SÖ-49	Dolerite	WR	18.7	260.2	0.2081	0.70822 ± 9
		F1	18.0	464.4	0.1124	0.70591 ± 7
		F2	25.7	431.7	0.1720	0.70767 ± 6
		F3	30.7	414.8	0.2138	0.70858 ± 6
		F4	39.2	382.5	0.2966	0.71037 ± 7
		F5	75.8	286.8	0.7662	0.72090 ± 7
		CPX	0.6	23.9	0.0668	0.70543 ± 7
		BI	230.1	58.5	11.6454	0.95619
SÖ-53	Dolerite	WR	40.8	237.6	0.4976	0.71546 ± 12
SÖ-54	Dolerite	WR*	20.3	214.7	0.2740	0.71015 ± 6

WR*	20.9	215.2	0.2811	0.71008 ± 5
BI	409.1	12.8	115.8990	3.25270

D.5 Hällefors dyke (fig. C.1)

Sample	Rock type	Material	Rb ppm	Sr ppm	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr
SÖ-21	Dolerite	WR	16.9	323.4	0.1511	0.70828 ± 7
		F1	12.5	472.3	0.0765	0.70706 ± 4
		F2	16.0	475.7	0.0975	0.70751 ± 4
		F3	106.2	406.0	0.7580	0.72283 ± 8
		CPX	1.4	24.9	0.1627	0.70751 ± 6
SÖ-23	Porph.dolerite	WR	36.0	324.6	0.3212	0.71484
SÖ-24	Porph.dolerite	WR	39.9	182.7	0.6326	0.71870 ± 4
SÖ-26	Porph.dolerite	WR	73.5	150.7	1.4155	0.73302 ± 10
		F1	64.5	454.4	0.4113	0.72008 ± 3
		F2	58.8	424.0	0.4018	0.71870 ± 3
		F3	66.4	315.1	0.6104	0.72153 ± 5
		F4	85.7	304.0	0.8458	0.72703 ± 4
		CPX	11.5	31.3	1.0677	0.72280 ± 6
SÖ-27	Porph.dolerite	WR	59.5	211.1	0.8171	0.72210 ± 8
SÖ-28	Porph.dolerite	WR	89.7	184.6	1.4099	0.73311 ± 3
SÖ-29	Porph.dolerite	WR	78.3	124.5	1.8271	0.74477 ± 7
SÖ-31	Porph.dolerite	WR	79.9	183.5	1.2622	0.72470 ± 4
SÖ-36	Melted granite	F6	72.3	82.3	2.5597	0.77282
		F7	137.7	90.2	4.4645	0.81396

		F8	132.3	89.5	4.3175	0.80986
		F9	235.7	97.4	7.1127	0.87075
		F10	252.9	100.4	7.4139	0.88015
SÖ-41	Porph.dolerite	WR	35.7	203.3	0.5088	0.71574 $\pm$ 3

D.6 Breven dyke (fig. C.2)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
SÖ- 2	Dolerite	WR	10.9	259.3	0.1212	0.70576
SÖ- 3	Dolerite	WR	5.9	260.5	0.0656	0.70480
		F7	150.1	651.1	0.6682	0.72379 $\pm$ 1
		BI	374.6	35.6	32.5527	1.42646
SÖ- 4	Dolerite	WR	6.8	267.1	0.0737	0.70561 $\pm$ 3
		F1	5.3	443.7	0.0346	0.70459 $\pm$ 9
		F5	6.1	454.2	0.0388	0.70501 $\pm$ 9
		F6*	18.8	469.2	0.1159	0.70723 $\pm$ 10
		F6*	20.3	472.0	0.1243	0.70738 $\pm$ 3
		F7	114.4	696.1	0.4762	0.71988 $\pm$ 5
		CPX	1.7	10.6	0.4625	0.71353 $\pm$ 18
		BI	371.4	35.0	32.8865	1.43197
SÖ- 5	Dolerite	WR	14.9	268.7	0.1600	0.70706
SÖ- 8	Granophyre	WR	111.0	63.8	5.0859	0.80860
		F1	29.7	43.8	1.9677	0.75005
		F2	43.9	38.0	3.3602	0.77879
		F3	107.0	38.6	8.1586	0.87579

		F4	164.0	36.5	13.3401	0.97869
		F5	209.9	32.1	19.6484	1.10392
SÖ- 9	Granophyre	WR	116.8	61.0	5.6013	0.82227
SÖ-10	Hybrid dol.	WR*	89.1	106.1	2.4385	0.74808
		WR*	90.1	108.2	2.4145	0.74777
SÖ-11	Hybrid dol.	WR	30.0	200.0	0.4337	0.71470
SÖ-12	Granophyre	WR	195.0	78.8	7.2712	0.86864
		F1	85.9	48.0	5.2424	0.83018
		F2	157.5	38.1	12.2724	0.97934
		F3	336.8	42.0	24.3766	1.23178
		EP	19.8	1550.9	0.0369	0.71697
		HB	230.5	26.1	26.9277	1.23964
SÖ-13	Granophyre	WR	161.1	32.9	14.5774	1.01501
		F1	88.8	24.7	10.6511	0.93812
		F2	105.4	18.0	17.5280	1.07586
		F3	150.2	17.6	25.9805	1.24485
		F4	160.2	17.1	28.5824	1.29414
		F5	215.5	18.3	36.6435	1.45574
SÖ-14	Hybrid dol.	WR	43.2	140.3	0.8930	0.72347
SÖ-15	Dolerite	WR	52.0	181.6	0.8297	0.72273
SÖ-16	Granophyre	WR	112.2	130.5	2.4985	0.75878
SÖ-18	Granophyre	WR*	174.5	47.8	10.8098	0.93985
		WR*	174.2	47.9	10.7625	0.94005
		F1	35.2	26.3	3.9045	0.79663
		F2	58.0	24.3	7.0148	0.86319
		F3	89.8	31.7	8.3469	0.89109
		F4	244.6	50.0	14.6046	1.02335
		F5	326.9	54.7	17.9350	1.09236

D.7 Tuna dykes (fig. C.3)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
DS-16	Dolerite	WR	27.5	566.7	0.1405	0.70615
DS-17	Trachyte	WR*	184.3	63.8	8.5049	0.87978
		WR*	184.5	64.1	8.4709	0.88071
DS-18	Porphy. dolerite	WR	43.3	455.6	0.2749	0.70841
DS-20	Qz-porphyry	WR	257.6	47.3	16.2237	1.01198
DS-21	Qz-porphyry	WR*	295.4	7.6	142.6810	3.47919
		WR*	297.0	7.6	143.1550	3.47085
DS-22	Qz-porphyry	WR*	290.0	12.2	78.9671	2.23765
		WR*	290.0	12.1	79.7054	2.25405

D.8 Gimán area sills (fig. C.4)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
J- 4	Dolerite	WR	16.5	315.4	0.1511	0.70569 $\pm$ 11
J-12	Dolerite	WR	14.3	319.0	0.1293	0.70518 $\pm$ 4
J-18	Dolerite	WR	10.4	335.8	0.0892	0.70432 $\pm$ 6
J-19	Dolerite	WR	8.3	303.7	0.0788	0.70428 $\pm$ 4
J-25	Dolerite	WR	10.8	337.8	0.0924	0.70441 $\pm$ 5

J-29	Dolerite pegm.	WR	6.9	207.1	0.0958	0.70542 $\pm$ 8
J-33	melted granite	WR	254.1	116.6	6.4017	0.86535
"	"	F1	68.8	106.7	1.8818	0.78638
"	"	F2	179.4	110.4	4.7623	0.83773
"	"	F3*	438.1	129.0	10.0399	0.93093
"	"	F3*	439.9	129.1	10.0715	0.92995
J-34	Hybrid dol.	WR	126.5	233.9	1.5702	0.74100 $\pm$ 7
J-35	Hybrid dol.	WR	119.7	225.3	1.5416	0.74064 $\pm$ 2
"	"	F1	67.1	419.9	0.4627	0.72256 $\pm$ 8
"	"	F2	68.5	351.2	0.5653	0.72379 $\pm$ 7
"	"	F3	181.6	232.1	2.2740	0.75357 $\pm$ 9
"	"	F4	280.8	192.3	4.2583	0.78676 $\pm$ 20
"	"	F5*	359.3	143.5	7.3335	0.83730 $\pm$ 8
"	"	F5*	338.7	154.8	6.3994	0.82188
"	"	CPX	7.8	15.7	1.4387	0.73575
J-36	Dolerite	WR	16.0	310.2	0.1493	0.70557 $\pm$ 7
J-41	Dolerite pegm.	WR	8.6	244.9	0.1016	0.70530 $\pm$ 4

D.9 Sundsjö dyke (fig. C.4)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
J-55	Melted granite	WR	105.7	57.2	5.4033	0.80987
"	"	F1	15.1	69.6	0.6285	0.73057
"	"	F2	38.8	64.7	1.7419	0.74397
"	"	F3	226.9	52.0	12.8961	0.93316

J-56	Melted granite	WR	113.5	28.9	11.5724	0.91163
'	'	F2	38.4	27.4	4.0840	0.78762
'	'	F3	210.3	33.9	18.5096	1.02971
J-58	Dolerite	WR	15.5	319.1	0.1405	0.70564 $\pm$ 20
'	'	F1	5.6	518.7	0.0312	0.70371 $\pm$ 4
'	'	F2	43.3	474.0	0.2642	0.70774 $\pm$ 10
'	'	F3	222.1	267.2	2.4123	0.74118
'	'	CPX	1.3	26.5	0.1407	0.70563 $\pm$ 5
'	'	BI	489.6	22.6	70.0857	1.92065

D.10 Älvho sill (fig. C.5)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
DN-315	Dolerite	WR*	32.0	371.7	0.2488	0.70914
'	'	WR*	34.5	375.5	0.2659	0.71024
DN-316	Dolerite	WR	20.4	316.4	0.1863	0.70723
DN-317	Dolerite	WR	63.8	302.2	0.6117	0.71688
'	'	F1	35.7	839.8	0.1231	0.70808 $\pm$ 5
'	'	F2	42.3	588.8	0.2079	0.70963
'	'	F4	56.8	516.5	0.3185	0.71168 $\pm$ 5
'	'	F5	121.1	384.4	0.9132	0.72274 $\pm$ 4
'	'	F6	178.2	255.0	2.0286	0.74280
'	'	F7	219.7	215.9	2.9582	0.75804 $\pm$ 5
'	'	AP	1.3	359.3	0.0104	0.70484 $\pm$ 6
'	'	CPX	1.1	34.1	0.0910	0.70690

DN-323 Monzonite	WR	133.4	214.8	1.8030	0.73978
DN-324 Monzonite	WR	156.7	173.9	2.6137	0.75438
DN-325 Dolerite	WR	20.1	414.0	0.1405	0.70617 $\pm$ 6
	BI	492.0	23.8	66.6938	1.86438
DN-326 Monzonite	WR	161.2	195.6	2.3941	0.75311
DN-327 Monzonite	WR*	157.3	210.6	2.1698	0.74985
	WR*	156.3	205.3	2.2116	0.75149
DN-328 Monzonite	WR	160.8	206.0	2.2692	0.75220

D.11 Emådalén sill (fig. C.5)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
DN-301 Monzonite		WR	103.7	259.4	1.1580	0.72539
DN-302 Monzonite		WR	104.0	261.0	1.1546	0.72556
DN-303 Monzonite		WR	96.1	266.7	1.0445	0.72467
DN-304 Monzonite		WR	94.3	277.1	0.9859	0.72285
		F1	90.1	438.2	0.5955	0.71548
		F2	215.1	209.0	2.9919	0.75657
		CPX	2.0	32.7	0.1802	0.70854 $\pm$ 5
DN-305 Monzonite		WR	90.3	292.5	0.8949	0.72140
DN-306 Monzonite		WR*	107.1	235.7	1.3167	0.72779
		WR*	107.7	235.9	1.3236	0.72783
		F1	77.6	393.7	0.5705	0.71544
		F2	134.7	296.7	1.3163	0.72898
		F3	231.4	113.5	5.9555	0.80610

	CPX	4.5	27.4	0.4727	0.71276
DN-310 Monzonite	WR	106.4	256.3	1.2027	0.72654
	F2	195.8	206.0	2.7622	0.75300
DN-311 Monzonite	F2	205.8	187.2	3.1968	0.75963
DN-312 Monzonite	WR	101.2	264.2	1.1098	0.72501
DN-329 Dolerite	WR	32.5	431.4	0.2178	0.70836
DN-330 Dolerite	WR	39.0	292.3	0.3862	0.71287

D.12 Älvdalsåsen sill (fig. C.5)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	
DN-51	Dolerite	WR	41.3	635.3	0.1880	0.70938 $\pm$	8
		BI	405.8	25.6	49.8434	1.58542	
DN-56	Dolerite	WR	23.0	555.5	0.1196	0.70562 $\pm$	5
DN-60	Dolerite	WR	32.5	438.7	0.2142	0.70807 $\pm$	4
DN-64	Dolerite	WR	28.9	666.8	0.1252	0.70678 $\pm$	6
DN-67	Dolerite pegm.	WR	31.9	675.6	0.1368	0.70827 $\pm$	4

D.13 Mossiberg sill (fig. C.5)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
DN-69	Dolerite	WR	11.1	618.2	0.0521	0.70409 ± 5
DN-71	Dolerite	WR	19.5	572.1	0.0984	0.70524 ± 5
DN-74	Dolerite	WR	17.9	633.9	0.0814	0.70448 ± 6
		BI	307.1	56.7	16.0849	0.98333
DN-78	Dolerite	WR	30.1	590.2	0.1474	0.70608 ± 7

D.14 Idre sill (fig. C.6)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
DN-17	Dolerite	BI	398.3	45.9	26.2324	1.16654
DN-21	Dolerite	WR	7.7	306.8	0.0725	0.70462 ± 10
		F1	8.4	527.4	0.0461	0.70416 ± 10
		F2	4.1	521.6	0.0227	0.70373 ± 4
		F3	3.6	504.0	0.0204	0.70376 ± 5
		F5	5.5	475.5	0.0337	0.70399 ± 8
		CPX	1.1	35.2	0.0918	0.70520 ± 10
		BI	488.0	41.0	36.5710	1.34661
DN-24	Dolerite	WR	11.2	322.6	0.1004	0.70563 ± 9
DN-30	Dolerite	WR	7.3	318.0	0.0663	0.70514 ± 6
DN-31	Dolerite	WR	19.8	356.4	0.1604	0.70766 ± 8

		F1	36.8	572.9	0.1857	0.70772 $\pm$ 7
		F3	7.7	384.9	0.0580	0.70557 $\pm$ 5
		F4	5.1	311.3	0.0472	0.70578 $\pm$ 3
		F5	5.1	317.3	0.0463	0.70583 $\pm$ 3
		CPX	0.8	32.0	0.0681	0.70535 $\pm$ 28
		CPX+10%EP*	0.8	227.9	0.0096	0.70898 $\pm$ 7
		CPX+10%EP*	0.7	237.9	0.0088	0.70920 $\pm$ 6

D.15 Bornåssjön sill (fig. C.6)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
71-15	Dolerite	WR	10.1	314.5	0.0926	0.70527 $\pm$ 10
		BI	166.8	41.4	11.8833	0.91600

D.16 Karlshamn dyke (fig. C.7)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
BL- 3	Dolerite	WR	16.0	427.8	0.1081	0.70673 $\pm$ 13
		F1	15.3	869.2	0.0511	0.70570 $\pm$ 6
		F2	8.9	892.9	0.0287	0.70545 $\pm$ 7
		F3	28.6	857.5	0.0965	0.70645 $\pm$ 4
		F4	79.9	787.0	0.2938	0.70908 $\pm$ 5

		F5	72.8	803.0	0.2624	0.70850 $\pm$	11
		AP	4.0	542.1	0.0215	0.70536 $\pm$	8
		CPX	3.4	55.6	0.1749	0.70737 $\pm$	7
BL- 7	Dolerite	WR	34.3	443.9	0.2237	0.70893 $\pm$	6
BL- 8	Hybrid dol.	WR	133.8	379.3	1.0220	0.72682 $\pm$	6
BL- 9	Hybrid dol.	WR	135.9	292.2	1.3489	0.73354 $\pm$	4
BL-10	Dolerite	WR	25.1	451.9	0.1604	0.70754 $\pm$	6
BL-15	Dolerite	WR	19.6	439.0	0.1292	0.70703 $\pm$	4
BL-18	Dolerite pegm.	WR	55.8	361.6	0.4464	0.71091 $\pm$	6
		F1	15.3	752.4	0.0588	0.70601 $\pm$	4
		F2	35.4	673.1	0.1523	0.70721 $\pm$	3
		F3	92.3	487.1	0.5483	0.71222 $\pm$	8
		F4	170.0	320.8	1.5356	0.72403 $\pm$	4
		AP	3.5	338.2	0.0302	0.70547 $\pm$	15
		CPX	1.1	32.7	0.0970	0.70635 $\pm$	4

D.17 Bräkne-Hoby dyke (fig. C.7)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	
	type		ppm	ppm			
BL-24	Dolerite	WR	16.0	282.5	0.1641	0.70701 $\pm$	5
		F1	7.9	585.9	0.0392	0.70541 $\pm$	7
		F2	8.1	593.8	0.0393	0.70544 $\pm$	4
		F3	20.9	530.8	0.1140	0.70648 $\pm$	5
		F4	41.6	431.0	0.2792	0.70882 $\pm$	5
		CPX	0.8	25.7	0.0957	0.70616 $\pm$	7

BL-27	Dolerite	WR	9.5	347.5	0.0792	0.70562 $\pm$ 10
'	'	F1	4.0	654.6	0.0178	0.70491 $\pm$ 7
'	'	F2	8.9	634.0	0.0406	0.70520 $\pm$ 3
'	'	F3	22.9	585.6	0.1132	0.70623 $\pm$ 6
'	'	F4	33.0	557.5	0.1713	0.70703 $\pm$ 4
'	'	CPX*	1.3	18.0	0.2116	0.70719 $\pm$ 9
'	'	CPX*	1.2	18.1	0.1978	0.70706 $\pm$ 5
BL-31	Dolerite	WR	63.5	622.2	0.2953	0.70849 $\pm$ 3
BL-32	Dolerite pegm.	WR	27.9	478.3	0.1691	0.70743 $\pm$ 9
BL-35	Dolerite	WR	43.4	495.5	0.2537	0.70830 $\pm$ 6
BL-36	Dolerite	WR	25.8	612.4	0.1220	0.70616 $\pm$ 7
BL-40	Dolerite	WR	29.4	624.7	0.1364	0.70650 $\pm$ 5

D.18 Tärnö dyke (fig. C.7)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
BL- 1	Hybrid dol.	WR	62.7	395.5	0.4592	0.71454 $\pm$ 3
BL- 2	Xenolith	WR	116.3	1722.3	0.1953	0.70848
'	'	F1	39.0	288.8	0.3912	0.71119
'	'	F2	91.2	124.0	2.1332	0.73348
'	'	F3	259.1	125.3	6.0274	0.78236
'	'	EP	3.8	11870.3	0.0009	0.70705
'	'	SPH	5.8	103.5	0.1630	0.70891
BL-ARX	Xenolith	WR	78.2	382.6	0.5918	0.71384
'	'	F1	41.2	157.0	0.7595	0.71621

F2	108.3	185.9	1.6894	0.72788
F3	253.4	177.6	4.1473	0.75875
SPH	7.4	80.1	0.2689	0.71082

D.19 Forserum sill (fig. C.8)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
AL- 2	Dolerite	WR	17.4	548.1	0.0916	0.70663 $\pm$ 10
AL- 6	Dolerite	WR	54.3	355.9	0.4415	0.71355 $\pm$ 5
'	'	F1	71.1	798.8	0.2578	0.71109 $\pm$ 7
'	'	F2	61.4	731.6	0.2427	0.71070 $\pm$ 9
'	'	F3	90.4	438.2	0.5970	0.71547 $\pm$ 3
'	'	F4	156.8	247.2	1.8391	0.73254 $\pm$ 3
'	'	F5	165.2	240.6	1.9924	0.73476 $\pm$ 6
'	'	AP	14.9	503.8	0.0857	0.70682 $\pm$ 3
'	'	CPX	1.5	137.4	0.0310	0.70776 $\pm$ 3
AL- 7	Dolerite	WR	25.3	430.0	0.1704	0.70779 $\pm$ 3
AL-10	Dolerite	WR	66.1	307.0	0.6238	0.71559 $\pm$ 8
AL-12	Dolerite	WR	35.4	392.7	0.2609	0.70951 $\pm$ 8
'	'	F1	28.2	790.1	0.1034	0.70685 $\pm$ 2
'	'	F2	32.6	738.8	0.1276	0.70745 $\pm$ 8
'	'	F3	59.3	554.7	0.3095	0.71009 $\pm$ 8
'	'	F4	100.1	403.6	0.7185	0.71615 $\pm$ 4
'	'	F5	120.9	351.3	0.9973	0.72074 $\pm$ 5
'	'	CPX	1.0	33.4	0.0842	0.70654 $\pm$ 6

AL-15	Dolerite	WR	51.0	368.2	0.4009	0.71309 $\pm$	8
AL-17	Dolerite	WR	37.1	322.8	0.3326	0.71150 $\pm$	8

D.20 Nilstorp sill (fig. C.8)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	
AL-21	Dolerite	WR	39.8	514.6	0.2239	0.70816 $\pm$	9
AL-22	Dolerite	WR	49.9	482.1	0.2992	0.70909 $\pm$	3
'	'	F1	31.7	991.6	0.0926	0.70627 $\pm$	4
'	'	F2	58.4	831.1	0.2033	0.70793 $\pm$	4
'	'	F3	114.4	607.4	0.5454	0.71268 $\pm$	5
'	'	F4	161.1	467.2	0.9989	0.71901 $\pm$	3
'	'	AP	6.1	692.0	0.0255	0.70523 $\pm$	3
'	'	CPX	4.8	49.1	0.2818	0.70921 $\pm$	9
AL-29	Dolerite	WR	25.3	537.9	0.1362	0.70659 $\pm$	4

D.21 Årby dyke (no detailed locality map, see figs. 4.4, 4.5)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	
SO-50	Dolerite	WR	30.6	558.9	0.1584	0.70721 $\pm$	10
'	'	F2	20.4	1029.4	0.0573	0.70585 $\pm$	7
'	'	F3	29.0	906.8	0.0926	0.70622 $\pm$	8

F5	47.0	685.9	0.1984	0.70798 $\pm$ 10
F6	69.7	477.6	0.4221	0.71114 $\pm$ 5
F7	79.8	445.8	0.5179	0.71214 $\pm$ 6
CPX	4.0	76.8	0.1491	0.70686 $\pm$ 7

D.22 Falun dyke (fig. C.9)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
DS- 1	Dolerite	WR	20.6	523.8	0.1136	0.70557 $\pm$ 7
DS- 2	Dolerite	WR	33.5	548.2	0.1769	0.70667 $\pm$ 4
DS- 9	Dolerite	WR	45.1	469.5	0.2781	0.70793 $\pm$ 9
'	'	F1	19.6	923.7	0.0614	0.70513 $\pm$ 10
'	'	F3	49.8	642.4	0.2245	0.70731 $\pm$ 6
'	'	F4	76.0	343.3	0.6396	0.71302 $\pm$ 2
'	'	F5	88.5	284.6	0.9006	0.71671 $\pm$ 13
'	'	CPX	4.9	70.2	0.1999	0.70693 $\pm$ 4
DS-29	Dolerite	WR	33.7	534.9	0.1825	0.70727 $\pm$ 3
DS-31	Dolerite	WR	32.9	532.4	0.1788	0.70702 $\pm$ 6
'	'	F1	18.1	933.6	0.0560	0.70514 $\pm$ 4
'	'	F3	31.2	708.4	0.1273	0.70645 $\pm$ 4
'	'	F4	60.6	498.7	0.3515	0.70994 $\pm$ 5
'	'	F5	96.1	295.3	0.9425	0.71872 $\pm$ 9
'	'	CPX	3.2	55.0	0.1703	0.70729 $\pm$ 10
'	'	BI	638.6	12.3	185.7540	3.13210

D.23 Västra Nornäs dyke (fig. C.10)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
D75-26	Dolerite	WR	27.7	546.7	0.1467	0.70672 ± 5
'	'	F2	18.4	948.7	0.0562	0.70549 ± 8
'	'	F5	51.6	520.9	0.2865	0.70872 ± 8
'	'	F6	59.1	478.1	0.3577	0.70927 ± 10
'	'	F7	99.1	298.1	0.9630	0.71728 ± 5
'	'	F8	98.6	292.0	0.9778	0.71733 ± 6
'	'	CPX	2.2	18.0	0.3534	0.70973 ± 5
'	'	BI	526.6	16.3	106.1220	2.11941

D.24 Bunkris dyke (fig. C.10)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
DN-37	Dolerite	WR	69.7	268.0	0.7543	0.72584 ± 13
'	'	F1	57.1	630.5	0.2621	0.71527 ± 4
'	'	F2	66.9	556.2	0.3481	0.71748 ± 8
'	'	F3	130.5	362.3	1.0444	0.73305 ± 8
'	'	F4	201.1	255.8	2.2857	0.75832 ± 5
'	'	CPX	1.0	21.0	0.1354	0.70984 ± 14
DN-42	Dolerite	WR	46.4	281.2	0.4777	0.71795 ± 6

F1	44.5	618.8	0.2082	0.71254 ± 8
F2	51.6	582.2	0.2568	0.71346 ± 5
F3	69.6	516.2	0.3906	0.71644 ± 6
F4	138.1	378.2	1.0589	0.73161 ± 6
CPX	2.5	21.9	0.3299	0.71138 ± 13

D.25 Vaggeryd syenite (fig. C.11)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
V- 2	Syenite	WR	91.8	439.5	0.5433	0.71633
'	'	MUSC	245.6	54.6	13.2340	0.88466
'	'	BI	431.2	23.2	57.5954	1.42941
V- 5	Syenite	WR*	114.0	26.1	12.9124	0.91321
'	'	WR*	115.9	26.0	13.1728	0.91283
V- 6	Syenite	WR*	257.9	17.6	45.2645	1.37728
'	'	WR*	258.5	17.7	45.0589	1.37762
V- 8	Syenite	WR	46.2	100.2	1.3363	0.72778
'	'	BI	192.6	9.1	65.7502	1.51796
V- 9	Syenite	WR	61.3	105.3	1.6897	0.73482
V-10	Syenite	WR	146.2	338.5	1.2527	0.73116
'	'	MUSC	454.1	35.4	38.9782	1.22457
'	'	BI	824.1	28.6	93.2640	1.91886
V-14	Syenite	WR	34.6	325.1	0.3082	0.71102
V-15	Syenite	WR	34.0	342.2	0.2876	0.71196
V-16	Syenite	WR*	44.6	18.4	7.1137	0.83419

		WR*	44.4	18.4	7.0682	0.83026
V-17	Syenite	WR	41.4	16.2	7.4742	0.82865
V-18	Syenite	WR	50.5	130.2	1.1242	0.72357

D.26 Country rocks from southern Sweden (figs. C.8, C.7)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
SPG-1	Granite	WR	252.8	245.6	2.9946	0.76586
"	"	AB	106.5	198.4	1.5585	0.74146
"	"	MCL	523.5	231.6	6.6156	0.82518
"	"	BI*	699.9	81.4	26.1083	1.21762
"	"	BI*	688.2	85.7	24.2931	1.17857
F-3	Granite	WR	145.4	533.3	0.7900	0.72308
"	"	OR	275.9	524.6	1.5265	0.73972
"	"	BI*	697.4	32.3	72.8096	2.39260
"	"	BI*	682.7	29.3	79.6339	2.56348

D.27 Särna alkaline complex (fig. C.6)

Sample	Rock type	Material	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
DN- 1	Syenite	WR	46.4	738.6	0.1816	0.70492
DN- 2	Syenite	WR	47.0	343.3	0.3961	0.70622 $\pm$ 4

DN- 5	Syenite	WR	59.1	409.5	0.4173	0.70865 $\pm$ 3
DN-9A	Syenite	WR	88.1	866.7	0.2941	0.70663
		F1	59.6	189.8	0.9089	0.70890
		F2*	114.8	232.9	1.4260	0.71145
		F2*	115.2	234.5	1.4222	0.71135
		F3*	167.7	249.7	1.9438	0.71331
		F3*	167.2	231.9	2.0876	0.71374
		SPH	22.8	2260.6	0.0292	0.70585
		CAN	34.9	1898.9	0.0531	0.70546
		AEG	2.2	157.7	0.0407	0.70572
DN-10	Syenite	WR	31.4	863.4	0.1054	0.70449 $\pm$ 3
DN-11	Syenite	WR	54.7	600.8	0.2636	0.70676 $\pm$ 3

D.28 Anorthosites and gabbros from south Greenland (fig. 6.1)

Sample	Rock	Material	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
	type		ppm	ppm		
30752	Anorthosite	WR	2.15	1807.1	0.003447	0.703053 $\pm$ 23
30753	Anorthosite	WR	31.12	1800.8	0.049981	0.703776 $\pm$ 23
40522	Anorthosite	WR	3.27	1603.1	0.005588	0.702969 $\pm$ 31
50220	Anorthosite	WR	1.78	1840.1	0.002796	0.703077 $\pm$ 33
50221	Anorthosite	WR	9.96	1849.2	0.015578	0.703232 $\pm$ 25
30755	Gabbro	WR	6.42	1328.7	0.013965	0.703210 $\pm$ 26
40518	Gabbro	WR	16.69	1095.4	0.044073	0.703747 $\pm$ 24
50224	Gabbro	WR	14.81	1141.8	0.037514	0.703531 $\pm$ 26
81156	Gabbro	WR	10.00	1187.6	0.024361	0.703386 $\pm$ 21

85977 Gabbro

WR* 11.43 1127.5 0.029314 0.703369 $\pm$ 26

WR* 11.17 1124.9 0.028708 0.703432 $\pm$ 29

APPENDIX E: INITIAL $^{87}\text{Sr}/^{86}\text{Sr}$ RATIOS OF DOLERITES FROM SWEDEN

There follows a list of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for whole-rock dolerites and associated rocks from Sweden. Values are given at the time of intrusion (plotted in figs. 6.9 and 6.10) and corrected for linear mantle growth to 900 m.y. (plotted in figs. 6.11-6.13). Errors are calculated using (1) a 2σ estimated uncertainty of 1.4% on $^{87}\text{Rb}/^{86}\text{Sr}$ plus (2) either of the in-run 2σ error or an estimated 2σ uncertainty of 0.04% on $^{87}\text{Sr}/^{86}\text{Sr}$. Age groups:- 1 = ca. 1560 m.y.; 2 = ca. 1250 m.y.; 3 = 1000-900 m.y.

Age	dyke/sill	Sample	$(^{87}\text{Sr}/^{86}\text{Sr})_0$	$(^{87}\text{Sr}/^{86}\text{Sr})_0$ 900 m.y.
1	Breven	SÖ- 2	0.7031 $\pm$ 6	0.7038
1	'	SÖ- 3	0.7034 $\pm$ 6	0.7041
1	'	SÖ- 4	0.7040 $\pm$ 6	0.7047
1	'	SÖ- 5	0.7035 $\pm$ 6	0.7042
1	Hällefors	SÖ-21	0.7049 $\pm$ 2	0.7056
1	Forskilen	SÖ-49	0.7036 $\pm$ 2	0.7043
1	Bälgviken	SÖ-53	0.7044 $\pm$ 4	0.7051
1	Gåstorp	SÖ-54	0.7041 $\pm$ 2	0.7048
2	Idre	DN-21	0.7034 $\pm$ 2	0.7037
2	'	DN-24	0.7039 $\pm$ 2	0.7042
2	'	DN-30	0.7040 $\pm$ 1	0.7044
2	'	DN-31	0.7049 $\pm$ 2	0.7052
2	Bornåssjön	71-15	0.7036 $\pm$ 2	0.7040
2	Älvdalsåsen	DN-51	0.7061 $\pm$ 2	0.7065

2		DN-56	0.7035 $\pm$ 1	0.7039
2		DN-60	0.7043 $\pm$ 1	0.7047
2		DN-64	0.7046 $\pm$ 1	0.7050
2		DN-67	0.7059 $\pm$ 1	0.7062
2	Mossiberg	DN-69	0.7032 $\pm$ 1	0.7035
2		DN-71	0.7035 $\pm$ 1	0.7039
2		DN-74	0.7031 $\pm$ 1	0.7034
2		DN-78	0.7035 $\pm$ 1	0.7039
2	Emådalén	DN-301	0.7051 $\pm$ 9	0.7055
2		DN-302	0.7053 $\pm$ 9	0.7057
2		DN-303	0.7064 $\pm$ 8	0.7067
2		DN-304	0.7056 $\pm$ 8	0.7060
2		DN-305	0.7057 $\pm$ 8	0.7061
2		DN-306	0.7047 $\pm$ 9	0.7050
2		DN-310	0.7055 $\pm$ 9	0.7058
2		DN-312	0.7056 $\pm$ 9	0.7059
2		DN-329	0.7045 $\pm$ 6	0.7049
2		DN-330	0.7061 $\pm$ 7	0.7065
2	Älvho	DN-315	0.7052 $\pm$ 6	0.7056
2		DN-316	0.7040 $\pm$ 6	0.7043
2		DN-317	0.7062 $\pm$ 7	0.7065
2		DN-323	0.7082 $\pm$ 10	0.7086
2		DN-324	0.7085 $\pm$ 13	0.7089
2		DN-325	0.7037 $\pm$ 3	0.7041
2		DN-326	0.7112 $\pm$ 12	0.7115
2		DN-327	0.7123 $\pm$ 11	0.7127
2		DN-328	0.7124 $\pm$ 12	0.7128
2	Gimån	J- 4	0.7031 $\pm$ 3	0.7034

2	J-12	0.7029 $\pm$ 1	0.7033
2	J-18	0.7028 $\pm$ 1	0.7031
2	J-19	0.7029 $\pm$ 1	0.7033
2	J-25	0.7028 $\pm$ 1	0.7032
2	J-29	0.7037 $\pm$ 2	0.7041
2	J-34	0.7135 $\pm$ 5	0.7139
2	J-35	0.7136 $\pm$ 4	0.7140
2	J-36	0.7030 $\pm$ 2	0.7033
2	J-41	0.7035 $\pm$ 1	0.7039
2 Sundsjö	J-58	0.7032 $\pm$ 4	0.7035
3 Tärnö	BL- 1	0.7088 $\pm$ 1	0.7088
3 Karlshamn	BL- 3	0.7054 $\pm$ 3	0.7054
3	BL- 7	0.7061 $\pm$ 2	0.7061
3	BL- 8	0.7141 $\pm$ 3	0.7141
3	BL- 9	0.7167 $\pm$ 3	0.7167
3	BL-10	0.7055 $\pm$ 1	0.7055
3	BL-15	0.7054 $\pm$ 1	0.7054
3	BL-18	0.7053 $\pm$ 2	0.7053
3 Bräkne-Hoby	BL-24	0.7050 $\pm$ 1	0.7050
3	BL-27	0.7046 $\pm$ 2	0.7046
3	BL-31	0.7048 $\pm$ 1	0.7048
3	BL-32	0.7053 $\pm$ 2	0.7053
3	BL-35	0.7051 $\pm$ 2	0.7051
3	BL-36	0.7046 $\pm$ 2	0.7046
3	BL-40	0.7048 $\pm$ 1	0.7048
3 Forserum	AL- 2	0.7054 $\pm$ 2	0.7055
3	AL- 6	0.7075 $\pm$ 2	0.7076
3	AL- 7	0.7055 $\pm$ 1	0.7055

3		AL-10	0.7070 $\pm$ 3	0.7071
3		AL-12	0.7059 $\pm$ 2	0.7060
3		AL-15	0.7076 $\pm$ 2	0.7077
3		AL-17	0.7070 $\pm$ 2	0.7070
3	Nilstorp	AL-21	0.7051 $\pm$ 2	0.7052
3		AL-22	0.7050 $\pm$ 1	0.7051
3		AL-29	0.7047 $\pm$ 1	0.7048
3	Årby	SÖ-50	0.7050 $\pm$ 2	0.7051
3	Falun	DS- 1	0.7040 $\pm$ 2	0.7041
3		DS- 2	0.7042 $\pm$ 1	0.7043
3		DS- 9	0.7041 $\pm$ 2	0.7042
3		DS-29	0.7048 $\pm$ 1	0.7049
3		DS-31	0.7046 $\pm$ 1	0.7046
3	V.Nornäs	D75-26	0.7048 $\pm$ 1	0.7048

APPENDIX F: RESULTS FROM GEOCHEMICAL STUDIES OF FELDSPARS

F.1 X-ray diffraction studies

There follows a list of feldspar separates which were run with quartz standard on a Philips PW 1051 X-ray diffractometer to check for the presence of secondary minerals: sericite, zoisite, epidote and calcite. No detectable amounts of these are present, even in DN-37 F2, which is from a sample heavily sericitised in thin-section; it appears that the magnetic cleaning of the separates at 0.75-1.0 amps has removed any sericite present. Most separates yield complicated traces, often suggesting the presence of two or more feldspars, or of compositions intermediate between those for which data is given in the powder data file. Those suffixed L give simple traces for labradorite; no interpretation is attempted for the more alkaline feldspars.

SO- 4 F6	DS- 9 F5
SO- 4 F7	DS-31 F2 L
DN-31 F1 L	DS-31 F4
DN-31 F4	D75-26 F8
DN-31 F5	BL- 3 F2 L
DN-37 F2 L	BL- 3 F4
	BL-18 F4

F.2 Microprobe studies of feldspars and sericites

The following analyses of feldspars and sericites were performed on the Cambridge instruments 'Microscan' electron microprobe at Edinburgh. The sericites give low total analyses due to volatile content. Several of the feldspars show totals below an acceptable 98%: this can be accounted for

either by the fact that Ba (and Sr) was omitted from several analyses (in K-feldspars BaO is up to 1.3%), or by analysis of interstitial patches in the section which may not have ground down to a perfect surface. No systematic deficiency in any element was found in any of the low-total analyses. Numbers missing from the sequence (5 in all) were rejected due to unacceptably high $\text{TiO}_2/\text{MgO}/\text{FeO}$, indicating that the analyses were not of pure feldspar. P = multiple-twinned plagioclase; I = interstitial feldspar without multiple twinning.

Analysis	1	2	3	4	5
Sample	SÖ-3	SÖ-3	SÖ-3	SÖ-3	SÖ-3
	P feldspar	P feldspar	P feldspar	I feldspar	I feldspar
SiO_2	49.44	51.69	53.39	54.75	53.26
Al_2O_3	31.23	29.47	28.77	27.58	28.56
TiO_2	0.03	0.07	0.10	0.11	0.15
MgO	0.08	0.11	0.06	0.06	0.05
FeO	0.64	0.66	0.46	0.48	0.42
CaO	15.36	13.19	12.37	10.86	12.33
Na_2O	2.70	3.83	4.43	5.17	4.51
K_2O	0.21	0.38	0.31	0.56	0.33
BaO	0.00	0.00	0.00	0.09	0.06
SrO	0.03	0.03	0.03	0.06	0.05
Total	99.72	99.42	99.91	99.71	99.73

Analysis	6	7	8	9	10
Sample	SÖ-3	SÖ-4	SÖ-4	SÖ-4	SÖ-4
	I feldspar	P feldspar	P feldspar	P feldspar	I feldspar
SiO ₂	54.45	49.75	52.44	50.16	63.37
Al ₂ O ₃	28.05	30.80	28.76	30.27	19.08
TiO ₂	0.05	0.11	0.13	0.06	0.07
MgO	0.05	0.05	0.07	0.09	0.02
FeO	0.47	0.53	0.50	0.58	0.20
CaO	11.38	14.60	12.32	14.33	0.59
Na ₂ O	4.93	3.32	4.22	3.25	3.29
K ₂ O	0.40	0.22	0.48	0.30	11.11
BaO	0.44	0.00	0.00	0.00	0.69
SrO	0.07	0.06	0.06	0.06	0.06
Total	99.88	99.43	98.98	99.11	98.48

Analysis	11	12	13	14	15
Sample	SÖ-4	SÖ-4	SÖ-4	SÖ-4	DN-325
	I feldspar	I feldspar	I feldspar	I feldspar	P feldspar
SiO ₂	63.84	53.06	65.43	63.95	52.05
Al ₂ O ₃	18.64	28.27	18.91	18.92	29.48
TiO ₂	0.07	0.10	0.09	0.08	0.13
MgO	0.02	0.04	0.03	0.02	0.04
FeO	0.27	0.61	0.41	0.26	0.35
CaO	0.39	11.70	0.44	0.58	12.67
Na ₂ O	2.22	4.61	6.51	3.28	4.10
K ₂ O	12.86	0.48	6.90	11.21	0.34
BaO	0.31	0.01	0.29	0.41	0.03
SrO	0.04	0.03	0.07	0.07	0.11
Total	98.65	98.92	99.08	98.79	99.30

Analysis	16	17	18	19	20
Sample	DN-325	DN-325	DN-325	BL-3	BL-3
	P feldspar	P feldspar	I feldspar	P feldspar	P feldspar
SiO ₂	53.77	59.77	63.46	52.68	53.33
Al ₂ O ₃	28.31	24.38	18.43	28.01	27.87
TiO ₂	0.10	0.11	0.56	0.16	0.13
MgO	0.05	0.04	0.01	0.06	0.07
FeO	0.33	0.34	0.29	0.41	0.45
CaO	11.60	6.47	0.12	11.20	11.39
Na ₂ O	4.76	7.30	1.76	4.82	4.76
K ₂ O	0.46	0.78	13.78	0.58	0.53
BaO	0.06	0.12	0.58	-	-
SrO	0.10	0.12	0.04	-	-
Total	99.54	99.42	99.03	97.91	98.52

Analysis	21	22	23	24	25
Sample	BL-3	BL-3	BL-3	BL-3	BL-3
	P feldspar	I feldspar	I feldspar	I feldspar	I feldspar
SiO ₂	53.65	64.33	64.01	66.80	63.67
Al ₂ O ₃	27.72	18.81	19.12	18.99	17.96
TiO ₂	0.10	0.02	0.09	0.02	0.03
MgO	0.03	0.04	0.02	0.03	0.01
FeO	0.31	0.27	0.25	0.13	0.06
CaO	10.08	0.79	1.06	0.36	0.05
Na ₂ O	5.33	4.01	4.67	11.23	0.99
K ₂ O	0.53	10.26	9.18	0.60	15.21
Total	97.74	98.53	98.40	98.17	97.97

Analysis	26	27	28	29	30
Sample	BL-3	BL-3	BL-3	BL-3	BL-3
	P feldspar	P feldspar	P feldspar	I feldspar	I feldspar
SiO ₂	51.50	58.51	61.12	63.55	63.97
Al ₂ O ₃	28.43	23.67	22.77	18.99	18.15
TiO ₂	0.12	0.07	0.08	0.11	0.02
MgO	0.05	0.06	0.05	0.02	0.02
FeO	0.45	0.30	0.32	0.21	0.35
CaO	11.78	5.94	4.62	0.98	0.25
Na ₂ O	4.52	7.52	7.98	3.83	2.76
K ₂ O	0.47	1.02	1.68	10.10	12.43
Total	97.32	97.09	98.61	97.78	97.94

Analysis	31	32	33	35	36
Sample	EL-3	DN-31	DN-31	DN-31	DN-31
	I feldspar	P feldspar	P feldspar	sericite	sericite
SiO ₂	63.77	51.27	51.95	48.30	45.66
Al ₂ O ₃	17.94	28.75	28.51	29.69	30.63
TiO ₂	0.02	0.05	0.05	0.03	0.04
MgO	0.01	0.03	0.08	0.24	0.56
FeO	0.34	0.51	0.50	1.28	4.07
CaO	0.08	12.53	12.04	0.57	0.63
Na ₂ O	1.82	4.15	4.43	1.79	0.38
K ₂ O	13.88	0.26	0.31	8.98	10.32
Total	97.85	97.55	97.87	90.87	92.29

Analysis	37	38	39	40	41
Sample	DN-31	DN-31	DN-31	AL-12	AL-12
	sericite	I feldspar	I feldspar	P feldspar	P feldspar
SiO ₂	46.77	66.15	67.47	52.53	53.57
Al ₂ O ₃	30.97	19.19	17.35	26.97	26.91
TiO ₂	0.08	0.02	0.04	0.09	0.06
MgO	0.66	0.02	0.01	0.37	0.10
FeO	3.17	0.15	0.17	1.25	0.72
CaO	0.43	0.44	0.44	11.40	10.86
Na ₂ O	0.80	11.66	11.35	4.70	5.14
K ₂ O	10.05	0.07	0.06	0.43	0.47
Total	92.93	97.69	96.88	97.73	97.81

Analysis	42	43	44	45	46
Sample	AL-12	AL-12	AL-12	AL-12	AL-12
	P feldspar	P feldspar	P feldspar	P feldspar	P feldspar
SiO ₂	54.11	55.58	56.56	59.08	59.98
Al ₂ O ₃	26.00	25.66	24.36	23.49	21.83
TiO ₂	0.09	0.11	0.09	0.09	0.09
MgO	0.37	0.08	0.07	0.11	0.03
FeO	1.25	0.69	0.54	0.55	0.35
CaO	9.94	8.99	7.22	5.81	4.57
Na ₂ O	5.46	6.09	7.25	7.80	7.95
K ₂ O	0.45	0.66	0.58	0.63	1.11
Total	97.67	97.84	96.67	97.55	95.91

Analysis	47	49	51	52	53
Sample	AL-12	AL-12	AL-12	AL-12	SO-3
	P feldspar	I feldspar	I feldspar	I feldspar	P feldspar
SiO ₂	62.37	62.02	63.03	61.82	51.62
Al ₂ O ₃	21.13	18.37	18.00	17.88	29.51
TiO ₂	0.04	0.02	0.00	0.03	0.08
MgO	0.08	0.01	0.02	0.03	0.29
FeO	0.47	0.06	0.11	0.15	0.79
CaO	3.04	0.23	0.19	0.01	13.69
Na ₂ O	9.13	2.34	2.39	0.62	3.75
K ₂ O	1.07	13.38	13.45	15.34	0.25
BaO	-	-	-	-	0.00
SrO	-	-	-	-	0.03
Total	97.33	96.41	97.17	95.83	100.02

Analysis	54	55	56	58	59
Sample	SÖ-3	SÖ-3	SÖ-3	SÖ-3	SÖ-3
	P feldspar	I feldspar	I feldspar	I feldspar	I feldspar
SiO ₂	56.89	68.59	65.73	64.92	67.46
Al ₂ O ₃	26.79	19.67	19.99	18.58	19.25
TiO ₂	0.05	0.06	0.07	0.05	0.08
MgO	0.17	0.08	0.18	0.49	0.05
FeO	0.61	0.67	0.52	0.70	0.34
CaO	9.73	0.68	0.59	0.34	0.41
Na ₂ O	6.14	9.52	9.86	2.98	9.20
K ₂ O	0.10	2.02	0.39	11.16	3.39
BaO	0.03	0.27	0.29	0.43	0.24
SrO	0.11	0.04	0.10	0.10	0.04
Total	100.61	101.60	97.70	99.73	100.45

Analysis	60	61	63
Sample	SÖ-3	SÖ-3	SÖ-3
	P feldspar	P feldspar	I feldspar
SiO ₂	52.59	53.83	63.51
Al ₂ O ₃	29.05	28.10	18.41
TiO ₂	0.10	0.06	0.01
MgO	0.05	0.06	0.02
FeO	0.45	0.45	0.11
CaO	12.67	11.64	0.00
Na ₂ O	4.33	4.82	0.51
K ₂ O	0.17	0.21	15.80
BaO	0.02	0.00	1.33
SrO	0.02	0.07	0.06
Total	99.45	99.23	99.75