SOURCE AND EQUILIBRATION STUDIES OF XENOLITHS FROM THE CALEDONIAN GRANITES OF SCOTLAND

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SOURCE AND EQUILIBRATION STUDIES OF XENOLITHS FROM THE CALEDONIAN GRANITES OF SCOTLAND

by Peter Holden.

Thesis presented in partial fulfilment for the
degree of Doctor of Philosophy.

February 1987

This is dedicated to my father Clifford Holden who died shortly after the completion of this thesis.

CERTIFICATE

I PETER HOLDEN hereby certify that this thesis has been composed by myself, that it is a record of my own work and that it has not been accepted in partial or complete fulfilment of any other degree or professional qualification.

I was admitted to the Faculty of Science of the University of St. Andrews under Ordinance General No 12 in September 1982 and as a candidate for the degree of PhD in March 1983.

Signed ....Date..24th/2/87.........

I hereby certify that the candidate has fulfiled the conditions of the Resolution and Regulations appropriate to the Degree of PhD.

Signed..........................Date..24th February 1987.....

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ABSTRACT

The origin of the dark microdioritic inclusions within I-type granitoids has been the subject of much debate and speculation. These 'xenoliths' have been variously ascribed to restite or cognate crystallisation (autoliths), wall rocks (accidental xenoliths), exotic (mantle) magmas and immiscible liquids. The object of this study was to test between these potential sources of inclusions and processes and to investigate whether such inclusions can be used to provide information on the petrogenesis of the host pluton and the nature of the source region(s) as well as processes of compositional zoning.

The approach used was to study inclusions and their host rocks from three zoned plutons in the Caledonian of Scotland, namely Strontian, Ballachulish and Criffell. These were investigated petrographically, by whole rock major and trace element analysis and by Sr and Nd isotopes. Interpretation of the equilibration between inclusion and host benefitted from theoretical considerations of diffusion. It was found that for most 'igneous' inclusions Sr isotopes had thoroughly equilibrated, hosts and inclusions normally having identical Sr initial ratios. However the microdioritic inclusions normally retain a more primitive signature in their Nd isotopes (normally up to +2 Nd units greater than the host) precluding cognate, restite or immiscibility origins and are quite different from local country rocks. Most 'igneous' inclusions thus appear to have an exotic source of more primitive composition, presumably the upper mantle.

The appinites of the Strontian pluton were also investigated as part of a study of coexisting acid-basic melts in granitoid plutons. It is shown that these appinites have quenched liquid contacts with the granitoid hosts and some degree of mixing between appinitic magma and host granitoid has given rise to a reverse zonation pattern in composition and texture.

The isotopic data for the appinites of Strontian and for inclusions and their host granitoids in the Strontian and Criffell plutons point to a model for the initiation of granitoid magma genesis in the crust by advection of heat through transfer of basaltic magma from the upper mantle. Some of this magma is incorporated in the granitoid magma as basaltic inclusions which gradually become modified towards dioritic compositions. The characteristics of this mantle derived magma and the granitoid crustal source do not vary significantly between Strontian and Criffell in terms of Nd isotope model ages. This is a remarkable finding considering the very different tectonic settings and suggests greater similarities at deeper lithospheric levels.

The study of Nd isotopes in xenolithic inclusions in granitoids has been shown to provide at least as much additional information on the petrogenesis of granitoid plutons as the conventional study of their host rocks, but a complementary study of Sr isotopes is not particularly useful. This investigation provides some of the theoretical basis for this and presents a methodology for carrying out such investigations.

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CHAPTER 1

BACKGROUND AND OBJECTIVES

1.1 INTRODUCTION

One of the most characteristic features of granitoids compared with other igneous rock types is the common presence of inclusions variously known as XENOLITHS, ENCLAVES and AUTOLITHS. The fact that these inclusions are so typical of granitoids (they are uncommon even in their extrusive equivalents) is strongly suggestive of a close causal relationship and this idea has been espoused in many studies published over the last century. That these inclusions have petrogenetic significance is undoubted but there is no consensus on that significance. The fact that so many diverse processes have been proposed for their origins and that rigorous evaluation of these processes has not been done, led to the present study of inclusions using modern geochemical techniques.

Two current widely accepted models for the origin of these inclusions involve source region processes and magma chamber processes. Compositional zoning in plutons is also often attributed to one or other of these processes, thus studies of inclusions along with their host rocks might be expected to shed light on the nature of the source and the emplacement and crystallisation of the pluton.

The inclusions under investigation in this study, are those ellipsoidal in shape and normally darker and finer grained than their hosts, commonly 0.01 to 1.0m in length. The study specifically excludes small xenocrystal clots and schlieren-like structures. Of the available terms for these dark, fine grained masses within granitoids, inclusion and enclave are preferred as xenolith and autolith have genetic connotations. The term inclusion will be used throughout this study while genetic terms will be used when that meaning is intended.

1.2 INCLUSIONS: A SUMMARY OF PUBLISHED IDEAS

A comprehensive review of inclusion studies is not attempted here as other such reviews already exist, notably Didier (1973), Pabst (1928), Grout (1937) and Vernon (1983); this section will review general ideas presented in these publications. It is not possible to discuss these ideas without reference to the processes involved which will be discussed at greater length later and are presented in summary form only here. These ideas are presented as general themes within which many variants exist.

Accidental inclusions.

These are recognisable inclusions of local country rock accidentally included by processes such as stoping, wedging and fault-controlled permitted intrusion, as proposed by Sollas (1894) and developed by Goodspeed (1947). Despite being the simplest and most easily identified source of inclusions there has historically been considerable debate as to the amount of chemical exchange between inclusion and host magma. Assimilation of such inclusions might be expected to modify substantially the original magma compositions and as a result of assimilation the visual evidence is effectively lost.

Recognition of locally derived country rock may be straightforward when features such as lithology, bedding, foliation and even folding are present. However, this is most commonly a feature of metasedimentary inclusions and will be difficult to apply in igneous terrains. Didier (1973) pointed out that fine grained volcanic rocks as accidental inclusions might be confused with other inclusion types. Another category is the inclusion of earlier granitoids within a later granitoid magma, which should not be confused with cognate inclusions. An example of this is the inclusion of Thorr granodiorite within the Main Donegal granite (Pitcher & Berger 1972). Such features are especially useful in establishing the chronology of pluton arrays.

Cognate inclusions.

The mineralogy and igneous texture of most inclusions has often been interpreted as representing an earlier more basic phase in the evolution of a pluton. Finer grain size is possibly the result of marginal chilling. The common association of inclusion-rich tonalites and granodiorites with marginal diorites is good evidence for this process. Three types of cognate inclusion may be recognised, namely (a) more basic precursors included within their more evolved magmatic products (Didier 1973), (b) inclusions within plutons of their own chilled margins (Daly 1914, 1933), and (c) incorporation of cumulates presumably from marginal regions. Powers (1915) suggested that the rounded nature of such inclusions may be due to magmatic attrition. In further discussions of inclusion processes restite and liquid immiscibility (both types of cognate inclusions), will be considered separately and it should be assumed that the term cognate inclusion used in this study refers only to the types described in this section.

Restite inclusions.

During partial melting of the crust the refractory residuum (restite) is considered by White & Chappell (1977) to become mobilised within minimum and non-minimum melts. Fluid dynamic investigations (Van der Molen & Paterson 1979), suggest that this might occur with as little as 30% melt. Restite is expected to segregate from the melt phase during ascent and intrusion, but early marginal portions of plutons may not have undergone effective separation (unmixing). White & Chappell (1977) consider inclusion-rich margins to many granitoid plutons as being rich in restite.

Exotic magma inclusions.

The introduction of a more basic magma of exotic source (presumably upper mantle) into granitoid magma may take several forms. It might occur in the source region (Eichelberger 1975, 1978, 1980), or during ascent and emplacement (Thomas and Campbell-Smith 1932, Pitcher 1978, 1979). It has been argued that quenching of more basic magma in more acid magma results in 'pillows' (Wager & Bailey 1953). Alternatively magma may be injected as 'syn-plutonic' dykes (Pitcher and Berger 1972) with inclusions representing disrupted forms (Furman and Spera 1984). These processes represent the arrested stages of magma-mixing which in the field have rounded outlines (Blake et al. 1965), thus obviating the need for mechanical attrition.

Immiscible liquid inclusions.

Backstrom (1893) suggested that basic liquid may spontaneously segregate from a granitoid liquid the former occurring as immiscible magma droplets and appearing within the rock as small, rounded, dark and fine grained inclusion. This process was invoked by Bender et al. (1982) to explain the inclusion suite within the Cortlandt complex (New York USA).

1.3 INCLUSIONS AND PETROGENETIC PROCESSES

The apparent relationship of young granitoids and their volcanic equivalents to the subduction process is undisputed but clearly not simple (Gill 1981). In the circum-Pacific belt most granitoid plutonism appears to have a primitive source with only occasional evidence of an evolved crustal contribution (Pitcher & Cobbing 1985). The relative contributions of the subducted slab, mantle wedge and continental crust are uncertain and are difficult to model isotopically inasmuch as young crust may be indistinguishable from upper mantle.

Crustal sources for granitoid magmas have been modelled geochemically (Chappell & White 1974) and experimentally (Winkler 1976, Wyllie 1983a) and thermal conditions necessary to achieve complete melting for tonalite can be achieved in crustal collision zones (England & Thompson 1986). Isotopic characteristics of granitoids in many Paleozoic orogens where subduction is not obviously the cause strongly suggest the presence of continental crust in magma generation. An important contribution can be made to the thermal conditions of crustal melting by studying inclusions. Two end-member processes of crustal melting are considered here, namely (a) partial melting resulting in a melt with suspended restite, and (b) advection of additional heat into the crust with intrusion of mantle-derived magma. Only by studying the geochemical and isotopic characteristics of the inclusions can these processes be distinguished.

During ascent and emplacement granitoid magmas may encounter other magmas with which they may mix, at least partially (Mezger et al. 1985). During these stages they may partially assimilate wall rocks (DePaolo 1981), and Bender et al. (1982) even proposed that they may evolve an immiscible phase. The resulting inclusions in each case should be geochemically recognisable, provided they retain their compositional identity. The former two processes should show distinctive isotopic features between host granitoid and inclusions whilst immiscibility will be reflected in trace element partitioning (Bender et al. 1982).

Recent advances in the understanding of magma chamber processes in terms of rheological properties, crystallisation and convection have increased the number of mechanisms by which differentiation of granitoids are understood to occur, particularly in zoned plutons (e.g. Sparks et al. 1984). Formation of cumulates on the chamber wall may be a general process in granitoid plutons and such inclusions should be discernible by their cumulate chemistry and isotopic similarity to their host.

Recently widespread use has been made of inclusions as strain markers within plutons in attempts to understand the emplacement mechanisms, particularly diapirism. Notable studies include those of Holder (1979) on Ardara (Donegal) and Bateman (1985) on Cannibal Creek granite (Queensland, Australia). These studies make the assumptions of originally spherical geometry and similar viscosities of host granitoid and inclusion. Clearly an understanding of the inclusion's origin is necessary before making these assumptions.

One of the most common features of I-type granitoids (Chappell & White 1974) is a regular normal compositional zoning. This feature has been interpreted as being due to various processes even within the same pluton. For example, the concentric zoning of the Loch Doon pluton, Southern Scotland, was interpreted as being caused by in-situ fractional crystallisation (Tindle & Pearce 1981), by changes in the magma source (Halliday et al. 1980) or by contamination (Brown et al. 1979). It is not known whether the compositions of inclusions vary sympathetically with pluton compositions but it is normal that inclusions are less common within the evolved interiors of zoned plutons. The reasons for this are not clear but a detailed study of the inclusions may shed light on the overall processes of zonation.

The geochemical study of inclusions has an important function in granitoid petrogenesis for the reasons given above. Inclusions may be the only way to distinguish between important and fundamental petrogenetic processes including mantle and/or crustal sources, the subsequent role of mixing, wall rock contamination and immiscibility, and also in the crystal-liquid processes of magma chambers. It is remarkable that inclusions have been almost totally neglected; the number of published isotopic determinations on inclusions is a very small fraction of those on their host rocks. A major conclusion of this study is that the inclusions in any one pluton may not have a single source and that combinations of the above processes may have operated. Furthermore, inclusions do not appear to have remained in their pristine state after incorporation in the host and thus normally require careful investigation to discover their true origin. The present study is not built on a foundation of previous similar investigations, but demonstrates that some significant differences between granitoid host and inclusion remain to constrain petrogenetic interpretations. Having demonstrated this fact it is hoped that other investigations on granitoid inclusions will make further contributions such that consideration of inclusions becomes an essential part of the geochemical investigation of any granitoid pluton.

1.4 CLASSIFICATION AND NOMENCLATURE.

The most useful classification of inclusions is genetic, in which inclusions are assigned to the following categories discussed above, namely accidental, cognate, restite, exotic, or immiscible liquid. However such a genetic classification is interpretative and usually impractical without supporting geochemical and isotopic evidence. Similarly xenolith and autolith are genetic terms for those inclusions clearly of foreign derivation and those generated during the magmatic process. Goodspeed (1947) further subdivided xenoliths into epi-xenoliths, identifiable fragments of country rock which have not travelled far, and hypo-xenoliths, recognisable accidental inclusions derived from a source not observed within the pluton envelope.

In his comprehensive review of inclusions (enclaves), Didier (1973) suggested that all genetic classifications should be replaced with a simpler non-genetic system and this view is adopted in this work. All fragments of material within a granitoid, regardless of size, texture or composition, are simply referred to as inclusions (or the French enclave, see Didier 1973, p1-7). Where a genetic description can be confidently applied to an inclusion, then it may be added as a prefix e.g. xenolithic inclusion, cognate inclusion etc.

1.5 AIMS, OBJECTIVES AND METHODOLOGY.

The principal aim of this research was the systematic description and analysis of the dark inclusions (commonly termed xenoliths or autoliths) typical of those found in I-type granitoid plutons of all ages throughout the world. The inclusions were studied in three zoned plutons belonging to the Newer Granites of the Caledonian of Scotland (Read 1961).

The primary objectives of the study were the following:

1) To understand the processes by which inclusions are formed and to provide criteria for their recognition.

2) To discriminate granitoid petrogenesis models using the results of the first objective.

Dependent upon these primary objectives several secondary objectives were pursued namely:

1) To recognise the source rock or magma of inclusions.

2) To understand the mechanism of inclusion incorporation.

3) To determine the degree of subsequent modification of texture and/or composition.

4) To determine the effects of inclusions on host granitoid magmas.

5) If the inclusion can be shown to be restitic then can it be used to better constrain the nature of the source?

6) If the inclusion was derived from an exotic magma, does this represent an igneous event at this level in the crust (e.g. underplating)?

7) If inclusions can be shown to have deep source(s) then is it possible to map out regional variations in these sources using contemporaneous plutons?

8) To meet the needs of structural geologists who require to know

the history of inclusions for studies of emplacement mechanisms.

The methodology used in this study was first to map inclusion-rich and minor intrusion-rich interiors of an appropriate pluton (Strontian). Samples of inclusions and their immediate host rocks were collected from all types of inclusions in three nearly contemporaneous plutons in the Caledonian of Scotland (Strontian, Ballachulish and Criffell). These plutons were selected for their known abundance and variety of inclusions. Each pluton was located in one of the three major structural blocks of the Scottish Caledonian, namely the North-West Highlands, Grampian Highlands and the Southern Uplands.

The collected samples were studied petrographically, and major and trace element analyses were performed on several host-inclusion pairs from each pluton. Careful note of potential inclusion sources such as country rocks and minor intrusions were made with particular emphasis being placed on contemporaneous magma suites. After characterising the inclusions, Nd and Sr isotopes were determined on sub-sets of inclusion-host pairs from the Strontian and Criffell plutons.

The data were interpreted using geochemical variation diagrams, some specifically designed for the purpose. Models of Nd isotope behaviour were used to provide information on inclusion source ages. Processes of inclusion modification within granitoid magmas were studied using a combination of textural interpretation and diffusion modelling.

CHAPTER 2

GEOLOGICAL SETTING

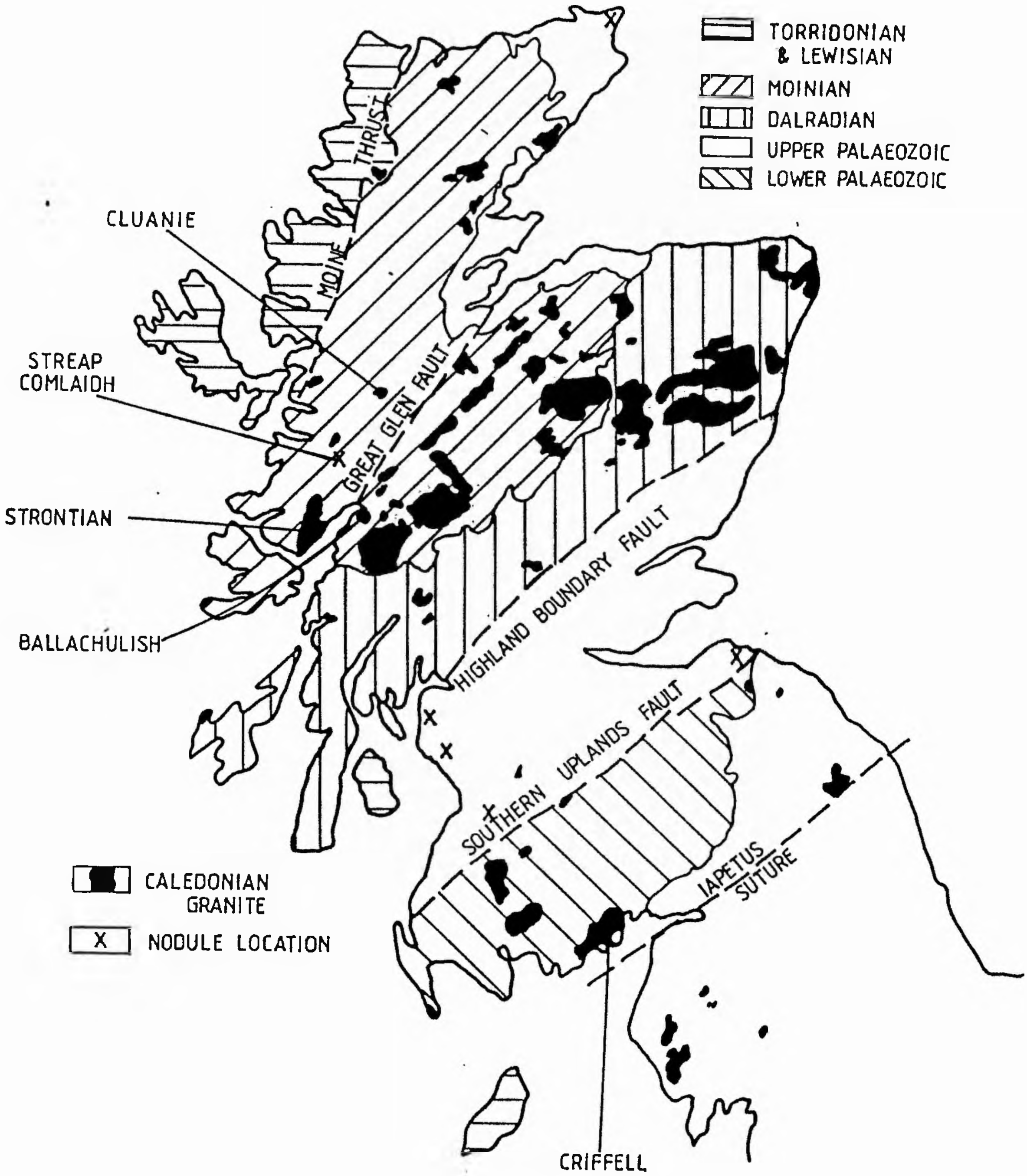
2.1 REGIONAL SETTING

The Scottish segment of the Caledonian Orogenic Belt may be divided into the orthotectonic Caledonides i.e. those rocks lying to the north of the Highland Boundary Fault Zone (HBFZ) which have undergone major tectonothermal modification during the Grampian orogeny, and the paratectonic Caledonides to the south of the HBFZ which have been much less modified (Read 1961, Watson & Dunning 1979, Watson 1984 and Fig 2.1). Deformation and metamorphism during the Grampian orogeny climaxed between 480-460 Myr (Hamilton et al. 1979 & 1980).

Granitoids of diverse composition are seen to penetrate both the metamorphic and weakly metamorphic Caledonides over 150Myrs the earliest being the pre-tectonic Carn Chuinneag granite dated at 555+/-17 Myr (Pidgeon & Johnstone 1974). In his review of Caledonian magmatism, Read (1961) on structural grounds subdivided the plutons into pre or syn-tectonic (Older Granites) and post-tectonic (Newer Granites). Whilst there is a general consensus that the Older Granites are of S-type (in the Chappell & White 1974 sense), being melts of the sialic crust, no agreement has been attained for the much more voluminous Newer Granites (Pankhurst 1979, Brown & Locke 1979, Hamilton et al. 1980, Stephens & Halliday 1984).

Attempts at comparing the Newer Granites with their time equivalents in the Lachlan fold belt of SE Australia using geochemical, isotopic and petrographic techniques have met with limited success (Pitcher 1983, Halliday & Stephens 1984). Whereas Australian granitoids fall into two distinct geochemical and

Fig 2.1 Regional map of Scottish Caledonides.



petrographic groupings (Chappell 1984), Scottish granitoids tend to fall within the I-type classification only, with no significant volume of S-type plutons (Stephens & Halliday 1984, see also Halliday et al. 1980 for S-type cores to Southern Uplands plutons). This dominance of I-type plutons has led some workers to propose that the Newer Granites were derived from a dominantly mantle source region (Brown et al. 1979, Tindle & Pearce 1983), whilst others have favoured an origin via wholesale crustal recycling (Pidgeon and Aftalion 1979, Hamilton et al. 1980, Ben Othman et al. 1984a), whilst still others favour a crustal control on an initially mantle-derived magma, thereby envisaging two separate source regions contributing to the melts (Stephens & Halliday 1980, Halliday 1983, Halliday & Stephens 1984).

Pidgeon & Aftalion (1979) in an important regional U/Pb isotopic study of the Newer Granites, discovered that zircons separated from some granitoids to the north of the HBFZ yielded upper concordia intercept ages of 800-1000Myr older than the emplacement age of the granitoid, whilst no zircon separates from south of the HBFZ yielded older age intercepts. Since an inherited zircon component can only be generated by assimilation of old crustal zircons, Pidgeon & Aftalion concluded that the HBFZ was the southerly limit of old Archaean (Lewisian) basement beneath the Caledonian cover. Subsequent Sm-Nd isotopic (Halliday 1984) and geochemical studies (Stephens & Halliday 1984) refined the earlier hypothesis, suggesting that the major geochemical-isotopic change occurs about a line running NE-SW through the Grampian Highlands (the so-named Mid Grampian Line, MGL), representing the suture of two continental basements of differing age and type, following the closure of the Iapetus Ocean in late Silurian times (Halliday et al. 1985).

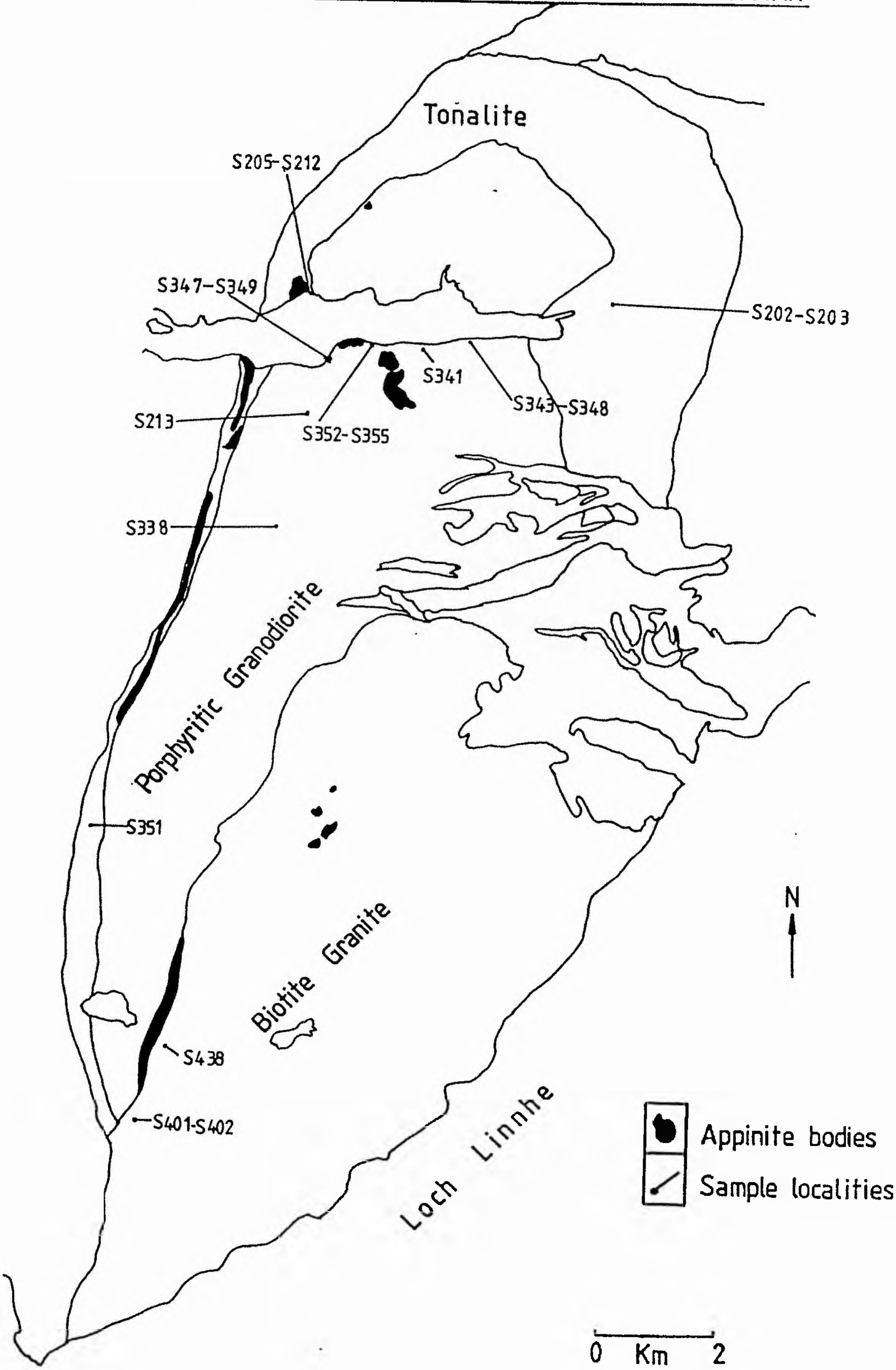
Much research effort has been expended in attempting to relate the closure of the Iapetus Ocean to geochemical trends and intrusive style of magmatism in the Caledonian (Thirlwall 1979, 1982, Brown & Locke 1979, Hamilton et al. 1980). Stephens and Halliday (1984) in assessing over one thousand major and trace element analyses have delineated 3 major geographical suites, the Cairngorm suite, the Argyll suite and the South of Scotland suite (for the Grampian Highlands, Midland Valley and Southern Uplands). Each suite consists of granitoids thought to be derived from a similar source region, with significant source differences occurring between suites.

At the outset of the project, a source-related origin for inclusions seemed probable so three plutons from different structural blocks were selected for study. These were Strontian (Northern Highlands), Ballachulish (Argyll suite) and Criffell (South of Scotland suite). No inclusion-rich pluton was found within the Cairngorm suite, a fact that may be significant. Each pluton is described in turn, with reference to their inclusions.

2.2 THE MORVERN-STRONTIAN GRANITE

The Morvern-Strontian 'granitic' mass has been the subject

Fig 2.2 OUTLINE GEOLOGICAL MAP OF STRONTIAN



of much study since the early mapping of Bailey (1917 & 1960), with more detailed mapping studies being undertaken by MacGregor & Kennedy (1932) and Sabine (1963). Small portions of the complex occur on the Geological Survey sheets 44 and 53 but the bulk of the pluton occurs on sheet 52E and in total covers an area in excess of 200Km^2.

The 'granitic' complex intrudes a metamorphosed Proterozoic supracrustal sequence, the Glenfinnan subdivision of the Moine. To the north and north-east the Moine consists of a granitic gneiss injection complex which has yielded a Rb/Sr isochron age of 1050 +/- 50Myr (Brook et al. 1976), and is thought to represent one of the few outcrops of Grenvillian age in the Scottish Caledonides. To the east the Moine country rock becomes progressively less metamorphosed until a structural break (the Sgurr Beag slide) which provides the tectonic separation of the Morar sub-division to the west, and the Glenfinnan sub-division to the east (Fig 2.2). To the south the 'granitic' complex is cut by a major transcurrent fault, the Great Glen fault (GGF). The observation of a similar dissected granitic pluton on the south side of the fault some 100Km to the north-east (the Foyers granite), lead Kennedy (1946) to propose that the GGF was a major sinistral strike-slip fault complex, and that the Strontian and Foyers granitoids were once a single pluton. Isotopic work by Pankhurst (1979), however, indicates that this was unlikely, although the central portions of each pluton do show certain isotopic similarities (Hamilton et al. 1980). Archaean rocks of the Lewisian complex are thought to provide the lower crustal basement upon which the Moine either lies uncomformably or in thrust contact (Shackleton 1979).

Structure

Where the regional strike of the local schist is normal to the strike of the pluton margin, the strike of the schists is seen to swing into parallelism with that of the pluton (MacGregor & Kennedy 1932). At the western edge of the pluton along the shores of Loch Sunart (NM 778607), the contact relations with the country rock are clearly displayed. The tonalite cuts what MacGregor and Kennedy (1932) described as an 'injection gneiss' consisting of strongly foliated silicic granulites with pods of amphibolite. The foliation in the granulite is parallel to strong foliation in the tonalite. It would appear then, that the intrusion of the pluton has deformed and re-crystallised the surrounding country rock into parallelism with the upwelling magma diapir causing distention and stretching.

In some places for example the northern shore of Loch Sunart (NM 789611), the contact between the tonalite and granodiorite is relatively sharp. At this locality the contact may be defined within 3 meters and appears to dip at a steep angle. However to the east in Glen Tarbet where the outcrop is sporadic and poor, the contact between the tonalite and granodiorite is everywhere gradational. No contact can be delineated in this area and may only be defined to within the nearest 50-100 meters. The granodiorite portion of the pluton is thus asymmetrically emplaced into the tonalite, with the western contact sharp and

Plate 2.1 Inclusion of porphyritic granodiorite in the biotite granite at NM 754447. Note the small black circle (centre picture) is a lens cap 55mm in diameter.

Plate 2.2 Typical country rock inclusion (xenolith) from Ballachulish (NN 054596).

the eastern contact diffuse (see Fig 2.2).

The contact between the biotite granite and the granodiorite or tonalite is sharp in all instances. The contacts appear sheeted in the field occurring as dykes or sheets within the granodiorite, tonalite or local country rock, generally trending east-west (Fig 2.2). Foliation parallel to the margins of the pluton, which is a feature of the tonalite and granodiorite portions, is not observed within the biotite granite. It is thus concluded that the emplacement of the biotite granite occurred some time after the forceful emplacement of the two outer portions (the fabric-forming event), by a mechanism of passive intrusion. A small inclusion of porphyritic granodiorite was found within the biotite granite at NM 754447, confirming the intrusive sequence (Plate 2.1).

Towards the outer margin of the pluton where, as has already been noted, the fabric within the granitoid increases, the inclusions within the granitoid become increasingly elongated or disc shaped. The degree of flattening may be crudely estimated by measuring the longest and shortest axes and calculating the ratio. By measuring such Z/X ratios for a number of inclusions within a small area the effect of initial shape of the inclusion may be minimised (Hutton 1982).

The shape of inclusions within the Strontian pluton depends upon their geographical location, that is, inclusions close to the margin of the pluton appear flattened whilst those remote from the outer contact are not. Furthermore, where flattened inclusions are found so also is a penetrating fabric within the

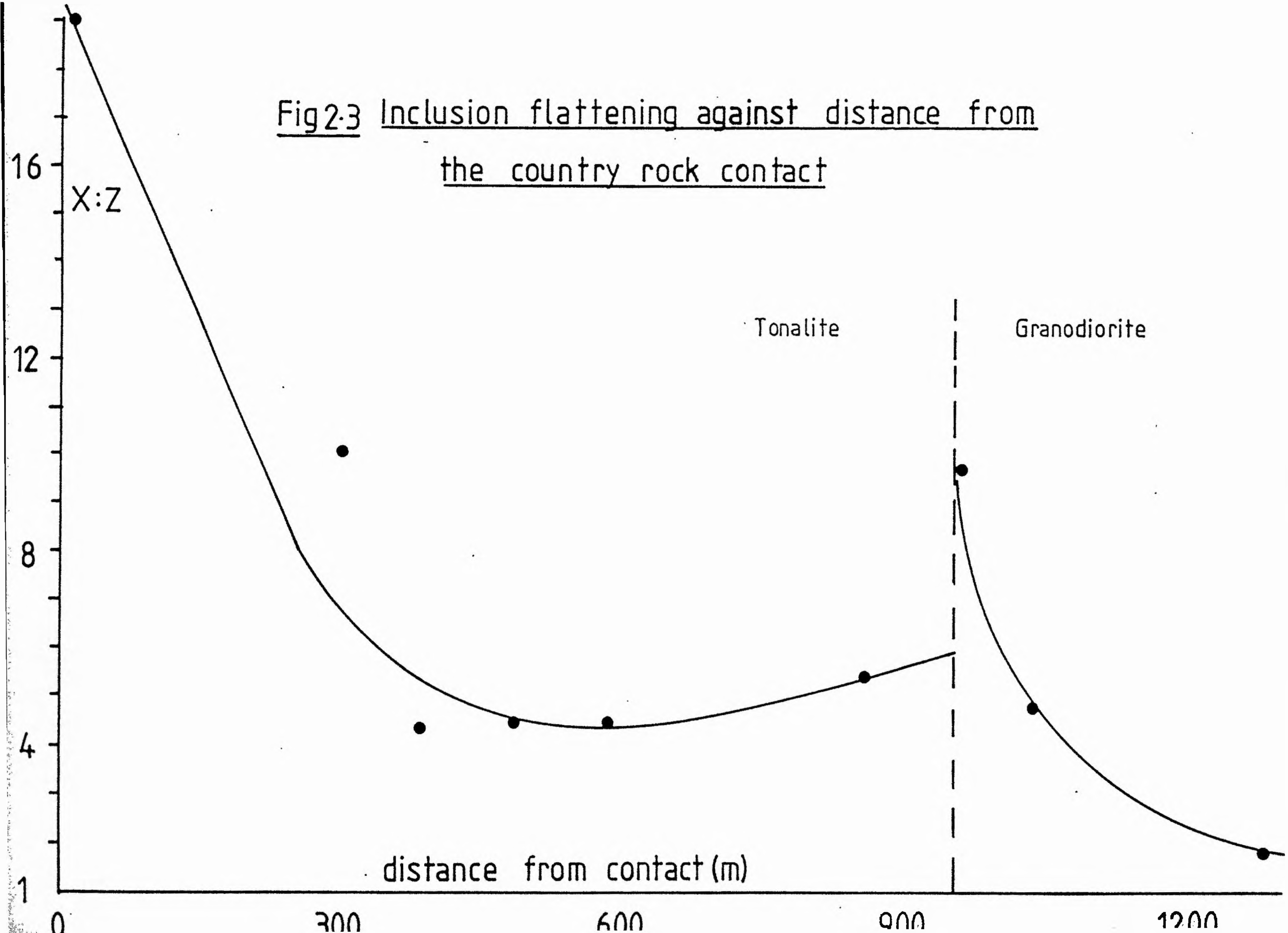
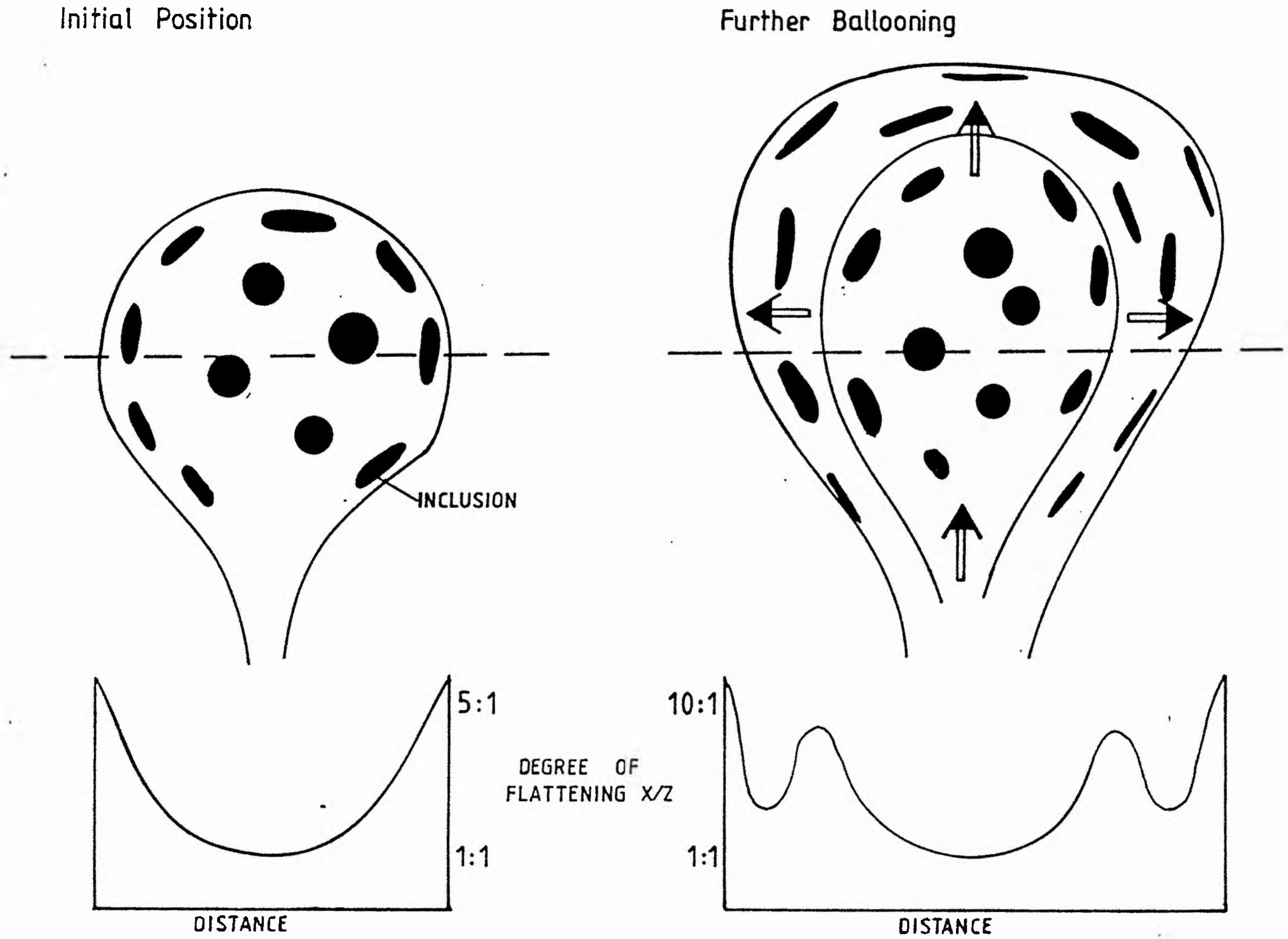


Fig 2·3 Inclusion flattening against distance from the country rock contact

Fig 2·4 Pluton 'Ballooning' and Inclusion Flattening



host, the intensity of the fabric increasing concomitantly with the degree of inclusion flattening. Also, the fabric is at all times parallel to the pluton margins.

Fig 2.3 is a plot of average inclusion elongation (Z/X) against distance from the contact, the data being collected in a traverse along the northern shore of Loch Sunart. It is clear that the degree of flattening decreases dramatically with distance from the external contact but is seen to rise abruptly again at the internal contact between the tonalite and granodiorite.

Ramsay (1975), Holder (1979) and Bateman (1985) have used similar observations on the Ardara (Donegal, Ireland), Chindamora (Zimbabwe) and Cannibal Creek (Australia) plutons respectively to suggest that these granitoids were emplaced by a ballooning mechanism. Fig 2.4 schematically represents the inflation of an idealised pluton by this 'ballooning' mechanism. (A) represents the initial magma chamber containing undeformed inclusions. Inflation of the chamber occurs due to magma influx through the relatively small conduit at the base, causing radial distention of the pluton and deformation of the inclusions and country rocks. (B) represents the inflated balloon complete with flattened inclusions and deflected country rock. Since only pure flattening is involved (Holder 1979) the volume of magma intruded by this mechanism may be estimated. Values given by Holder (1979) and Bateman (1985) suggest that around 70% of the magma in the Ardara and Cannibal Creek plutons was emplaced in this way.

Whilst a quantitative estimate of the amount of magma qualitative study has shown that pluton ballooning is consistent with the distribution of flattened inclusions within both the Strontian and Criffell plutons, but the lack of deformed inclusions within the Ballachulish pluton does not support this mechanism. Further, the central portion of the Strontian body contains no fabric associated with its intrusion, and displays all the characteristics of passive emplacement, suggesting a change in emplacement mechanism in the later stages of the pluton's formation.

Minor intrusives

Clustering around and within the Strontian pluton are numerous ultrabasic to basic minor intrusions which cut the regional Caledonian structures. These were deemed by MacGregor and Kennedy (1932) to belong to the Kentallenite-Appinite suite as defined by Bailey (1917 & 1960). From field observations on the 'appinite' inclusions within the complex, MacGregor and Kennedy concluded;

"Small xenoliths have also been stoped off the basic inclusions close to the edge of the mass. Thus the appinites are undoubtedly earlier than the granite complex, and are not intrusive."

Hence MacGregor and Kennedy (1932) concluded that the appinites occurring around and within the pluton cut the regional Caledonian structures but were earlier than the tonalite member. In MacGregor and Kennedy's scheme, the appinites represent mafic precursors to later granitoid plutonism (see Chapters 4 & 7 for

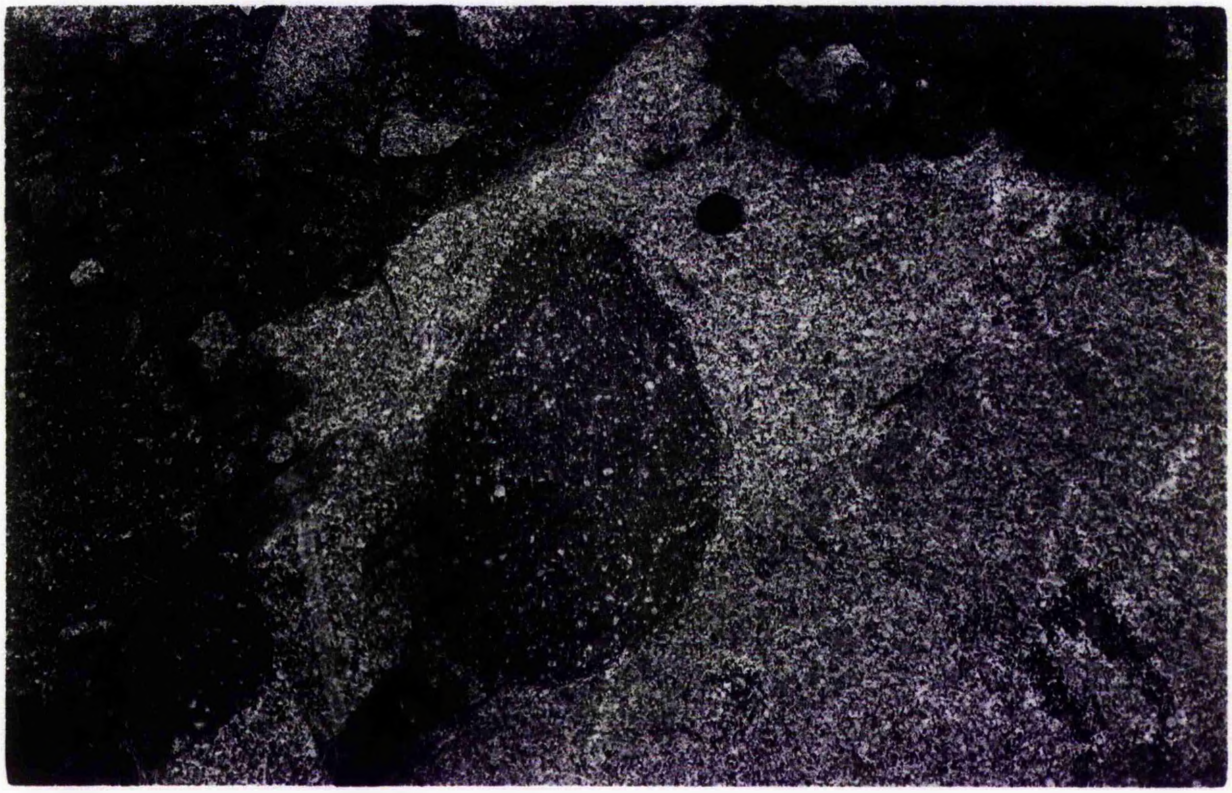


Plate 2.3 Typical ovoid micrograniticid inclusion from Strontian (NM 600792).

further treatment).

Inclusions within the Strontian pluton.

Within the intrusive calc-alkaline granitoids observed throughout the world, small discoid mafic inclusions are almost ubiquitous in the tonalite and granodiorite types (Didier 1973). Strontian is no exception with microgranitoid inclusions being one of the striking features of the outer portions of the pluton. In contrast the inner biotite granite carries few inclusions.

The small dark inclusions are almost always finer grained than their host granitoid (see Chapter 5 for exceptions). They consist mostly of plagioclase, hornblende, biotite and lesser quartz and K-feldspar, with accessory minerals dominated by apatite and sphene. Hornblende may occur as euhedral laths or in irregular aggregates associated with biotite, or as both in any inclusion (see Chapter 5). Based on field observation over the whole pluton, small inclusions may be classified into three major types:

1) Country rock with clear evidence of primary metasedimentary or regional metamorphic features (Plate 2.2)

2) Remnants of syn-plutonic intrusives which may be identified by their quench texture and irregular (cuspate) outline (see Plate 2.3).

3) Microdioritic inclusions which cannot be readily identified in the field as belonging to either Type 1 or Type 2 (Plate 2.3).

The majority of inclusions observed throughout the tonalite and granodiorite members of the pluton are those which show a range in textural features suitable for further sub-division into four sub-groups:

3a) Fine grained dark inclusions which carry few megacrysts and whose matrix is dominated by hornblende and plagioclase.

3b) Fine to medium grained inclusions which contain conspicuous porphyritic plagioclase feldspar laths similar to those observed within the host granitoid. These may or may not contain laths of hornblende.

3c) Fine grained inclusions with conspicuous ferromagnesian minerals which may occur either as aggregates or large individual euhedral crystals.

3d) ‘Felsic’ inclusions of relatively coarse grainsize but finer than the host granitoid, which are dominated by plagioclase feldspar and lesser hornblende.

With respect to their host granitoid, inclusions form dark fine grained patches of spherical to discoid shape. Contacts between host and inclusions are almost ubiquitously sharp. Where inclusions have not been subjected to significant flattening stresses the shape of the inclusion is generally spherical to ovoid, with the contacts sometimes appearing lobate in hand specimen. No discernible mineralogical contrast was observed in granitoids at or away from the inclusion contacts.

The small microgranitoid inclusions are randomly distributed throughout the tonalite and granodiorite, with no tendency for an accumulation of inclusions at the pluton margin as has been

Plate 2.4 Accumulation of inclusions: Hill Top Quarries, Strontian (NM 771575).

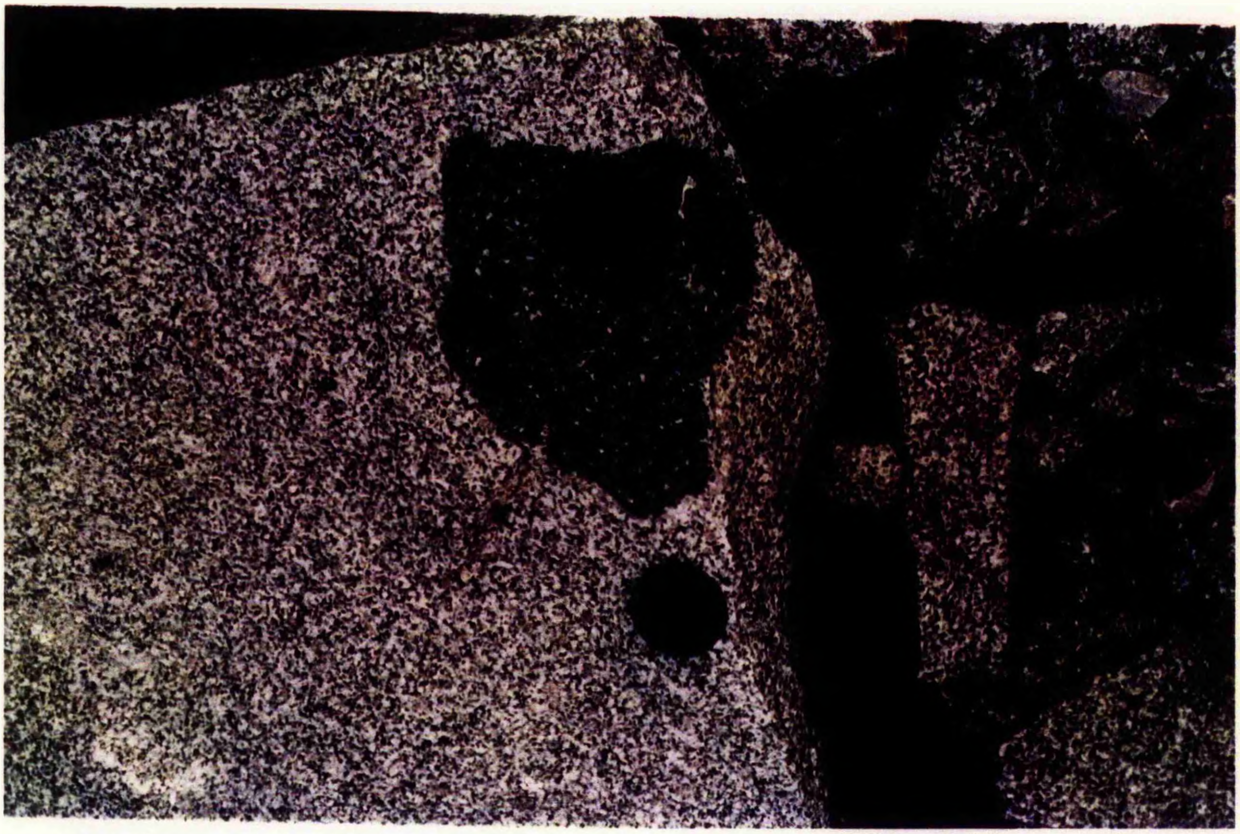


Plate 2.5 Double inclusion: Craignair Quarry, Criffell (NX 820607).

observed elsewhere (Pitcher & Berger 1972, Grout 1937, Pabst 1928, Didier 1973, Vernon 1983). Locally, an increased number of inclusions per unit area is observed. These concentrations take the form of rounded patches extremely rich in inclusions (e.g. Hilltop Quarries, NM 771575, see Plate 2.4), or as trains of inclusions strung together with the same overall dip and strike as the host granitoid fabric (e.g. the shore section of Loch Sunart, NM 792610). Texturally the inclusions within these concentrations are usually similar, though this is not always the case. Some inclusions may be coarser than the others whilst the overall colour index may be variable, but all the inclusions within these enriched zones are of igneous aspect with interlocking mineral textures and finer grainsize than their host granitoid. However frequent the inclusions become in these zones, a thin sliver of host granitoid material is always seen to separate the individuals from one another and the fact that they do not touch is said to be a feature of the non-Newtonian behaviour of the magma (Pitcher 1986, in press).

A particularly rich collection of inclusions may be observed on the north shore of Loch Sunart (NM 792610). Here large numbers of inclusions (thousands) of similar size and texture are arranged in a train trending 175-180° and dipping 50°E to the vertical. The dip and strike of the train follows precisely that of the host granitoid fabric, the long axis of the inclusions being parallel to that of the fabric. Since the direction of elongation of these inclusions matches that of the host fabric, the inclusions can be deduced to have been in-situ prior to the flattening stress which yielded the fabric.

The origin of these local accumulations is somewhat obscure and may represent the later stages of break-up of syn-plutonic bodies (Pabst 1928, Pitcher & Berger 1972, Weibe 1973, Pitcher 1979, Furman & Spera 1985). Alternatively the trains may represent the accumulation of inclusions swept from surrounding magma by convection currents (Link 1970) but no local depletion in inclusion population was apparent in the vicinity of the inclusion trains.

Further along the north shore of Loch Sunart (NM 793611 and NM 800612) irregular fine grained masses may be observed. These are very similar in structure to the so called ‘syn-plutonic’ dykes (Pitcher & Berger 1972) observed within the Thorr pluton, Donegal. Locally the fabric within the hosts is disrupted, and the granitoid becomes extremely porphyritic, with potash feldspars growing to 5cm in length. These masses do not carry a significant fabric nor are they discoid and are interpreted as a late phase of basic-intermediate intrusion after the host fabric was established, but prior to the consolidation of the pluton, allowing late stage movements to disrupt the dykes into irregular shapes in a plastic manner.

2.3 THE BALLACHULISH PLUTON.

The Ballachulish zoned pluton consists of an outer grey quartz diorite and an inner pink granite (Bailey 1917 & 1960).

The pluton intrudes a late-Proterozoic to early-Precambrian supracrustal sequence, the Appin group of the Dalradian (Fig 2.5). No up-tilting of the country rock surrounding the pluton was observed.

Structure

The marginal portion of the body, unlike those of Strontian or Criffell, contained no appreciable foliation. This observation together with the abundant country rock fragments included within the outer portion of the pluton, led Bailey (1917 & 1960) to the conclusion that the pluton had been passively emplaced, with stoping being the dominant intrusive mechanism.The external contact of the pluton is generally sharp and cross-cutting with some veining of the local country rock (e.g. at the road cut at NN 019589).

The internal contact is sharp and the pink granite cross-cuts the quartz diorite and is clearly younger (Bailey 1917 & 1960). No fabric is associated with the intrusion of the central portion which, like the outer member appears to have been emplaced passively.

Minor intrusives.

Associated with the pluton are several minor intrusives which make up the appinite-kentallenite suite of rocks. These range from explosion breccia types (Bowes & Wright 1961), through small bosses of variable augite-hornblende-biotite rich rocks, to hornblendites (Bailey 1917 & 1960). Bowes & Wright (1967) have shown that these numerous minor intrusives pre-date the invasion

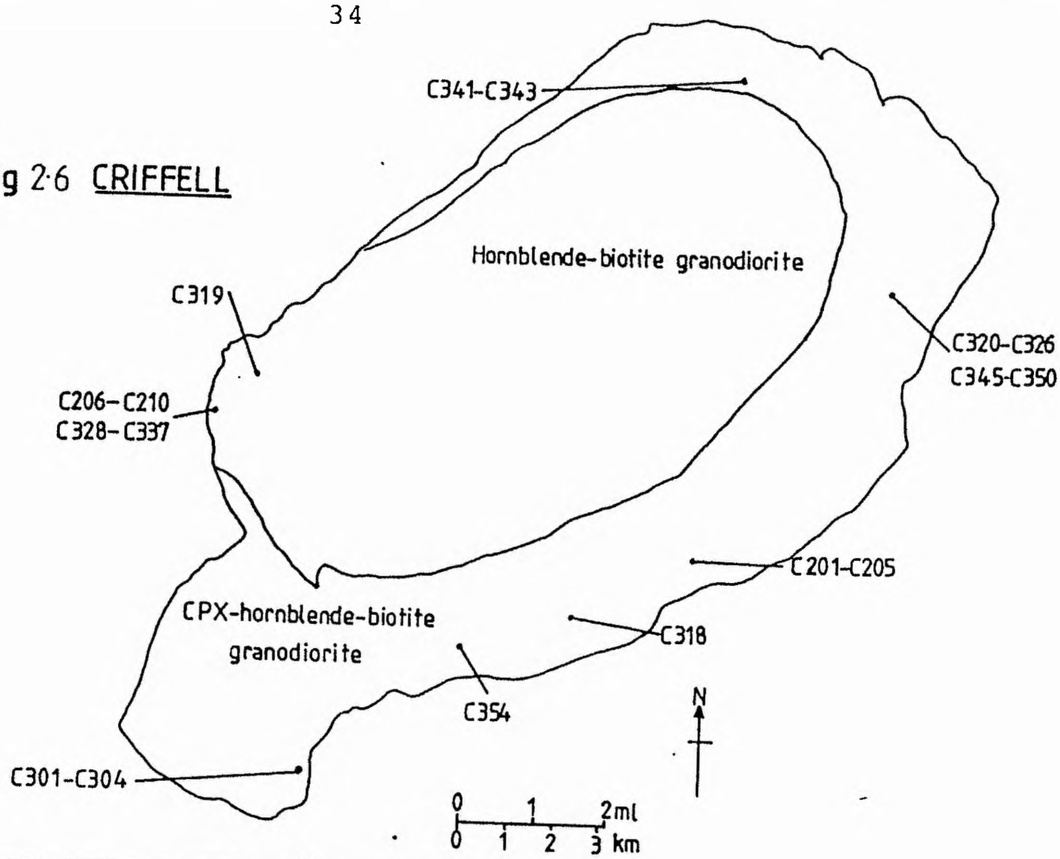


Fig 2·6 CRIFFELL

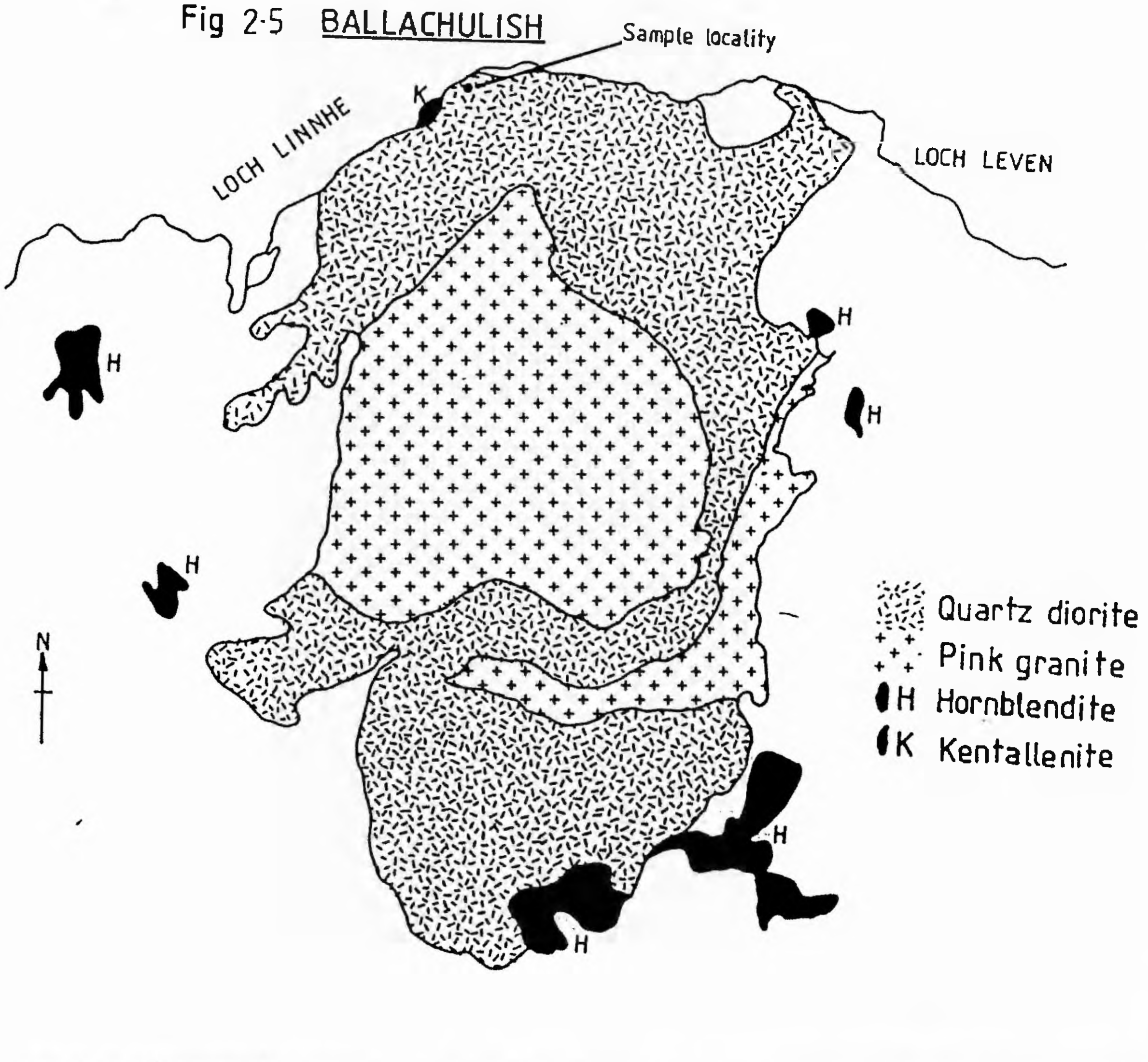


Fig 2·5 BALLACHULISH

of the granitoid. No appinites (or other minor intrusives) were found within the pluton (cf. Strontian) but several are clustered around the the country rock envelope.

Inclusions within the Ballachulish pluton.

Inclusions within the Ballachulish pluton are concentrated in the marginal member of the quartz diorite. Inclusions within the pink granite are occasionally found but in general are rare. Both inclusions of country rock (type 1 inclusions) and microgranitoid inclusions of uncertain origin (type 3 inclusions) are common within the marginal areas, but as the internal contact with the pink granite is approached, the inclusion suite tends to be dominated by the type 3 inclusions. No syn-plutonic material was identified within the pluton despite the abundance of appinitic material surrounding the body.

A shore exposure east of the Ballachulish-Onich bridge (NN 054596) is filled with inclusions of country rock and microgranitoid type. Country rock may be observed in various states of assimilation with a tendency for the more digested varieties to superficially resemble the microgranitoid type.

At the quarries (NN 022590) only 250m from the external contact the inclusions within the quartz diorite are dominated by microgranitoid inclusions (type 3), with all four sub-types represented. Representative samples were taken from the quarry along with samples of their host granitoid.

2.4 THE CRIFFELL PLUTON

The Southern Uplands consists of a repetitive sequence of volcanic, greywacke and shale horizons. Detailed mapping by Peach & Horne (1899), Walton (1955, 1961, 1963), Weir (1974) and Legget et al. (1979), has shown that each sequence is separated by significant high angle dip-slip faults, with each sequence younging to the north west. However the older Ordovician rocks occur to the north west whilst the younger Silurian rocks occur to the south east (Peach & Horne 1899). This anomalous behaviour was interpreted by Mackerrow et al. (1976) as a fossil accretionary prism, whereby successively younger sequences of oceanic material are thrust onto the existing imbricate stack. The initially low angle faults are successively steepened as each new sequence is added.

The Criffell pluton is the southernmost granitoid in the Scottish Caledonian being close to the proposed line of the Iapetus suture (Phillips et al. 1976; Fig 2.1). Many petrographic, geochemical and isotopic data are available (Phillips 1956, Phillips et al. 1981, Stephens 1972, Halliday et al. 1980, Stephens & Halliday 1980, Halliday 1983, Stephens et al. 1985), allowing the complex to be segregated into two major components (Stephens & Halliday 1980) viz; a marginal clinopyroxene-hornblende granodiorite of relative homogeneity (pulse 1) and a second pulse (pulse 2) of biotite granodiorite to biotite-muscovite granite asymmetrically emplaced within the former (Fig 2.6). The Southern Uplands country rock surrounding the pluton has suffered considerable uptilting, indicating that the pluton forced the country rock aside during the final stages of emplacement in a manner similar to the outer portions of the Strontian complex. To the west of the pluton lies the quartz diorite complex of Bengarin Hill and associated dyke swarm (Phillips 1956).

Structure

Contacts between the outer and inner portions of the pluton were originally described as gradational (Phillips 1956, Phillips et al. 1981). Discontinuities in geochemical surfaces in the region of the internal contacts are in most cases sharp which led Stephens & Halliday (1980) to propose that the two portions were successively intruded magma pulses which had co-mingled at their mutual contact. An internal contact at NX 894684 can be observed however (Stephens pers. comm. 1983), proving that the zonation within the body is largely the product of two magma pulses with subsequent fractional crystallisation processes becoming important in the zonation pattern of pulse (2).

Like Strontian, the outer members of the pluton contain a strong foliation. Unlike Strontian, the intensity of the fabric and the concomitant flattening of microgranitoid inclusions increases from the margin inwards until the zone of contact with the second pulse is encountered, whence the intensity of the fabric decreases rapidly. This was interpreted by Phillips et al. (1981) to be the product of liquid state laminar flow in a convecting magma chamber, a view challenged by Holder (1981), who favoured pluton ballooning and associated planar flattening, similar to that envisaged for the Ardara pluton, Donegal (Holder 1979, see Fig 2.4).

Minor intrusives within the Criffell pluton.

Unlike the Strontian pluton, Criffell does not contain an abundance of minor basic to ultrabasic bodies. It does however contain numerous porphyry and porphyrite dykes (Phillips 1956) which cut the outer portions but disappear as the evolved central portion is approached. It would appear then that the intrusion of these bodies coincided with the influx of magma which constituted pulse (2). A contemporaneous quartz dioritic complex may also be found to the west of the Criffell mass (Phillips 1956) but does not intersect with the main body at any point.

Inclusions within the Criffell pluton

Inclusions are of ovoid to discoid shape similar to those observed within the Strontian pluton and are dominated by igneous textured microgranitoids of uncertain affinity (type 3 inclusions). No unambiguous country rock fragments were found except at the very margin of the intrusion where it abutted against the country rock. At Craignair Quarry (NX 820607) a double inclusion was discovered (see Didier 1973, p.228). Here one inclusion is seen to be wholly enclosed by another (Plate 2.5) and at least the outer portion of the inclusion must have been in a liquid state at some stage, in order itself to acquire an inclusion (Didier 1973).

Inclusions are abundant throughout pulse (1) and the outer portion of pulse (2) but become less frequent as the geochemically more evolved members are encountered. However, unlike the Strontian pluton, anomalously rich zones of inclusions were not observed, although this may be a feature of the exposure. At the geochemical centre of the pluton, despite extensive searching, no inclusions were observed.

Where inclusions were of discoid shape it was noticed that the prevailing fabric within the host passed straight through the inclusion without deviation. During the event which produced the fabric, the inclusion must have had very similar rheological properties to their hosts, otherwise significant deviations in the host fabric would occur in the vicinity of the inclusion (Hutton pers. comm. 1983).

Samples of inclusions and their host granitoid were collected from several localities around the pluton (see Fig 2,6). Samples of possible source rocks were also collected which included four samples of nearby Caledonian dykes and four samples from the Bengarin Hill quartz diorite.

CHAPTER 3

PETROGRAPHY OF THE HOST GRANITOIDS.

3.1 STRONTIAN

Tonalite and porphyritic granodiorite

In hand specimen the porphyritic granodiorite is easily identified with respect to the tonalite by the presence of large salmon-pink micro-perthitic potassium feldspars, which in places reach 2cm in length. Furthermore the tonalite in general has a distinctive foliation which is only poorly developed within the granodiorite. In thin section these two units are texturally similar but for the presence of K-feldspar megacrysts. Thus for convenience the two members will be described together with the important differences being highlighted as they arise.

The modal analysis shows (Fig 3.1), plagioclase as the dominant mineral, occurring as euhedral laths of 5-10mm in length (Plate 3.1). Zonation, often oscillatory, varies from An35-40 at the core to An15-20 at the rim. Sericitic alteration is often patchy, and usually at the margins. Potassium feldspar occurs mainly as microperthite, interstitial to earlier plagioclase and hornblende. The presence of mineral inclusions at the ragged periphery of these megacrystic K-feldspars and their absence from the core of the same mineral suggest an earlier phase of euhedral K-feldspar crystallisation which acted as nucleation sites for later anhedral overgrowths (Plate 3.1). In contrast, the tonalite carries only small, intersertal, anhedral, micro-perthitic K-feldspars. Quartz occurs as small intersertal pools within the tonalite, and these are slightly larger and more common in the porphyritic granodiorite.

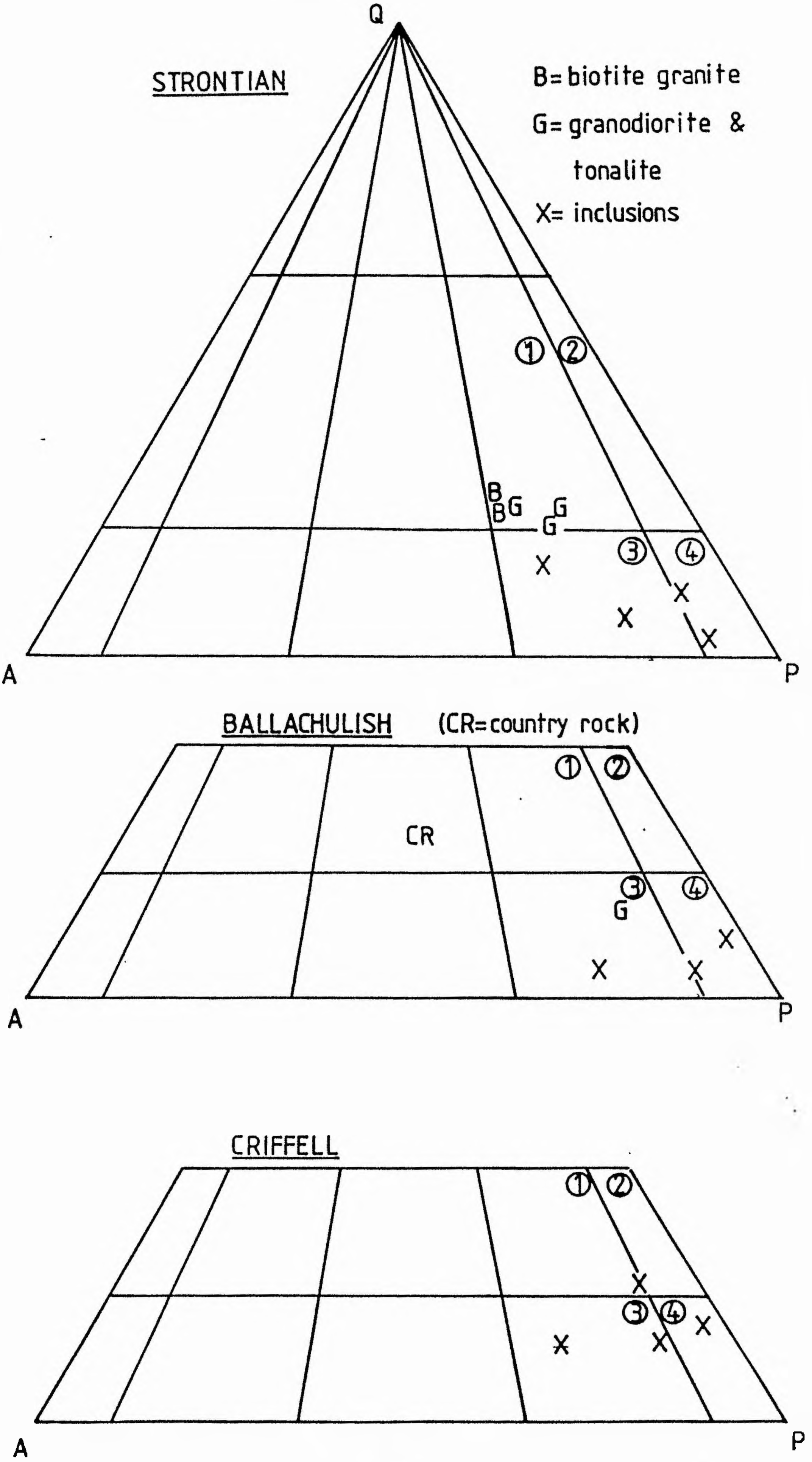


Fig 3·1 Modal analysis of hosts and inclusions. Fields of interest are
① Granodiorite ② Quartz diorite ③ Monzodiorite ④ Diorite

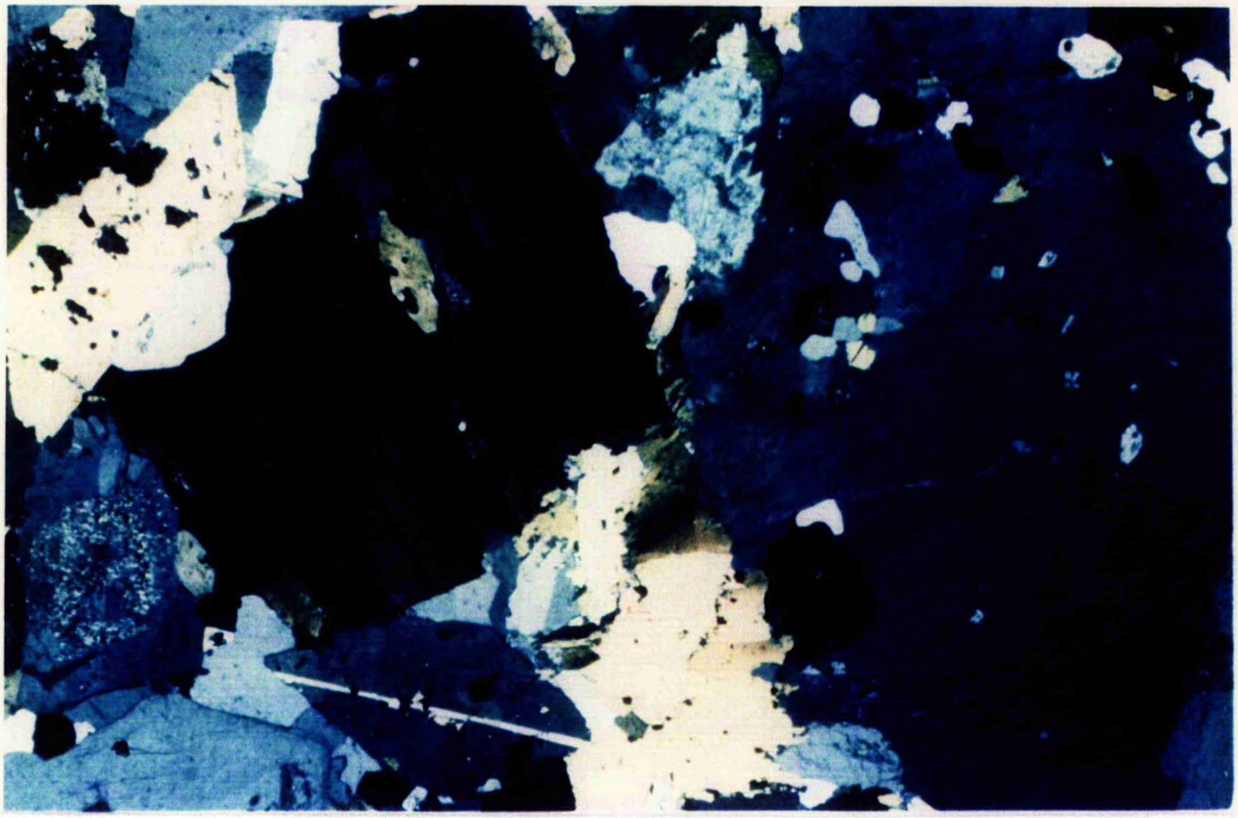


Plate 3.1 Strontian tonalite/granodiorite texture, sample S353G (X14).

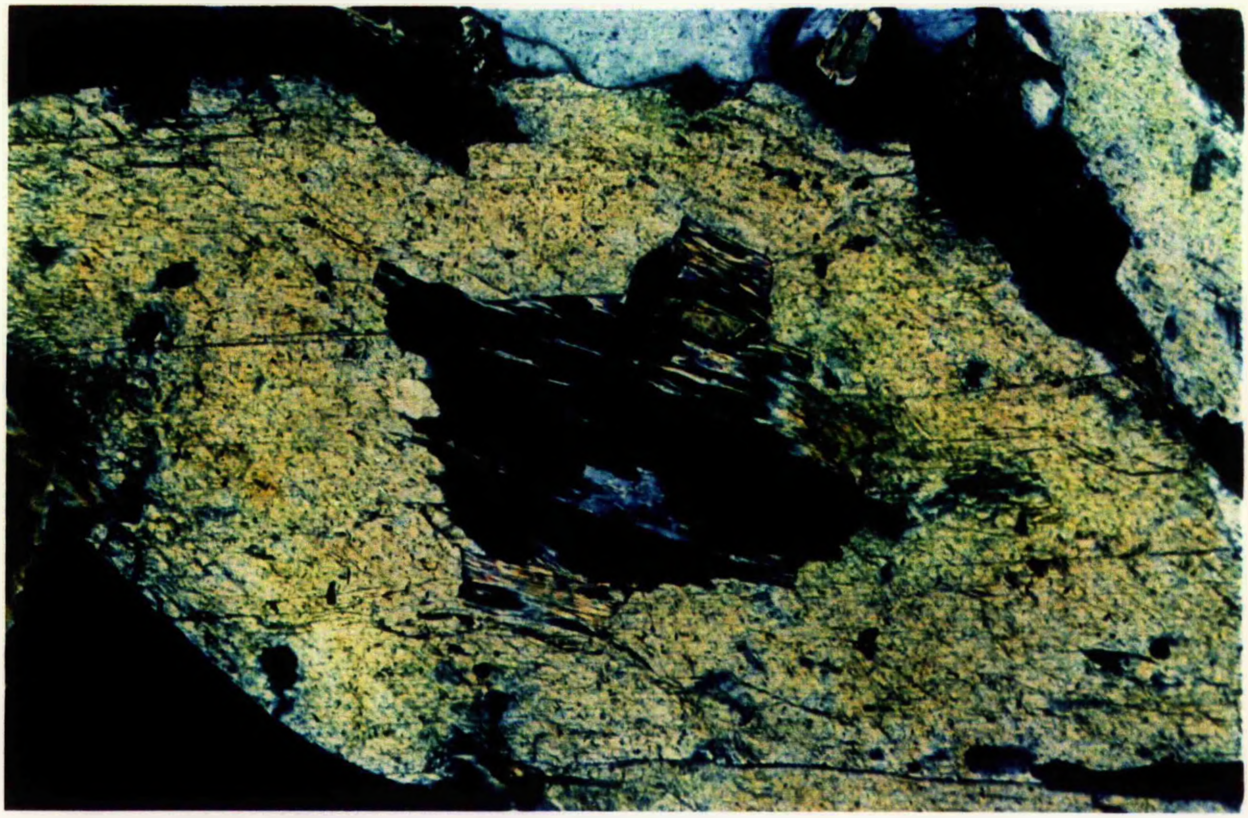


Plate 3.2 Biotite at centre of hornblende, sample S345G (X 48).

The ferromagnesian minerals are dominated by a green hornblende and a highly pleochroic biotite. Hornblende occurs as an early euhedral mineral typically 5mm in length, and as a late ragged mineral mantling partially (or mostly) resorbed clinopyroxene (augite). The later hornblende is associated with a dusting of opaques as well as sphene and biotite. These replacement textures are more common in the tonalite and decrease with increasing silica content of the rock. Biotite occurs in two forms. As an early subhedral to anhedral mica which occurs as cores to the euhedral hornblende (Plate 3.2). No reaction appears to have taken place between the two minerals, the later euhedral hornblende presumably using the biotite as a site of preferred nucleation. Alternatively the hornblendes crystallised as hollow minerals which were later infilled by biotite growth. This is an unusual but common observation throughout the pluton. It is not confined to Strontian alone; similar textures have been observed within the Moor of Rannoch pluton (Leighton, 1985), perhaps reflecting the high K and Ba contents of the magma stabilising mica in preference to amphibole (Wones 1981). The later biotite occurs as intersertal anhedral minerals between the plagioclase and early hornblende laths, and it is associated with the late hornblende as a product of clinopyroxene breakdown.

Accessory minerals include zircon, abundant sphene and apatite and less commonly allanite. Sphene occurs both as large 1-3mm euhedral prisms indicating relatively early crystallisation and in a late ragged anhedral form associated with the breakdown of augite. Colourless apatite forms equant prisms commonly enclosed by late K-feldspar but not within euhedral plagioclase laths and thus appears to have crystallised fairly late in the magmatic sequence. Zircons tend to be small and in general well formed. Allanite is associated with biotite.

Biotite Granite

The biotite granite may be classified as a granodiorite according to the nomenclature of Streckeisen (1976), however the original field term (MacGregor & Kennedy 1932) is retained to distinguish it from other types.

Plagioclase is the most abundant mineral in the rock and comprises 2-5mm euhedral to subhedral laths. Later overgrowths of albitic plagioclase often give a subhedral look to a previously euhedral mineral. Sericitic alteration is patchy and occasionally severe. Potassium feldspar (microperthite) is late and forms subhedral and anhedral crystals 3-5mm enclosing earlier euhedral plagioclase in a poikilitic fashion. At the mineral interface between the plagioclase and K-feldspar, myrmekitic intergrowths are readily observed with bulbous protuberances being convex to the K-feldspar. Quartz is extremely late in this rock filling in the spaces between the subhedral K-feldspar and the plagioclase.

The principal ferromagnesian mineral is a late biotite. This is often in close association with small (<0.2mm) opaque minerals. Later chloritisation has modified the periphery of some of the biotite grains.

Accessory minerals include abundant (mostly euhedral) sphene with lesser amounts of apatite, zircon and allanite. Occasionally a cube shaped sphene may be observed at the centre of an early euhedral and poorly zoned plagioclase. This may be a relict of restitic material brought up from the source region by the granitoid and is a texture not observed within the outer two members of the pluton.

3.2 BALLACHULISH

Outer quartz diorite

Abundant plagioclase of euhedral aspect is early and dominates the mode. Zonation of these 3-4mm crystals is weak being An35 at the core and An20 at the rim. Alteration (sericite) particularly affects the core region of the plagioclase. K-feldspar forms late poikiloblastic minerals surrounding all other minerals. Quartz likewise is late and confined to the interstices between plagioclase and hornblende laths.

Biotite is slightly more abundant than hornblende with the latter occurring principally as early euhedral minerals frequently cored by relict clinopyroxene +/- opaque minerals. Biotite on the other hand is mostly late occurring as aggregates along with later hornblende in the intersticies left by plagioclase.

Allanite is common occurring in close association with biotite and less frequently with hornblende. Sphene forms late anhedral poikilitic grains, whilst apatite occurs both as laths and as squat prisms. Zircon is generally euhedral, but tends to be less evident than other accessory minerals.

Inner pink granite

Euhedral plagioclase with white mica alteration is common and occurs as grains up to 5mm. K-feldspar forms late anhedral grains of microperthite, whilst quartz was the last mineral to crystallise, being intersertal between plagioclase and K-feldspar grains.

Biotite and replacement chlorite are the only ferromagnesian minerals. Biotite occurs late, crystallising around plagioclase and K-feldspar grains.

Allanite enclosed by biotite is common, whilst apatite and zircon form small rounded crystals. Sphene is rare and anhedral.

3.3 CRIFFELL

Petrographic descriptions of the zonation within the Criffell pluton were given by Phillips (1956). Stephens et al. (1985) reclassified and extended this earlier work and only a brief summary of the petrography of the host granitoids is presented here.

The Criffell pluton consists of a strongly foliated outer member (pulse 1) and a weakly foliated central member (pulse 2). Contacts between the two are generally gradational, but at Lotus [NX 897681] a sharp contact between the two pulses may be observed (Stephens, pers.comm. 1983).

Clinopyroxene hornblende biotite granodiorite

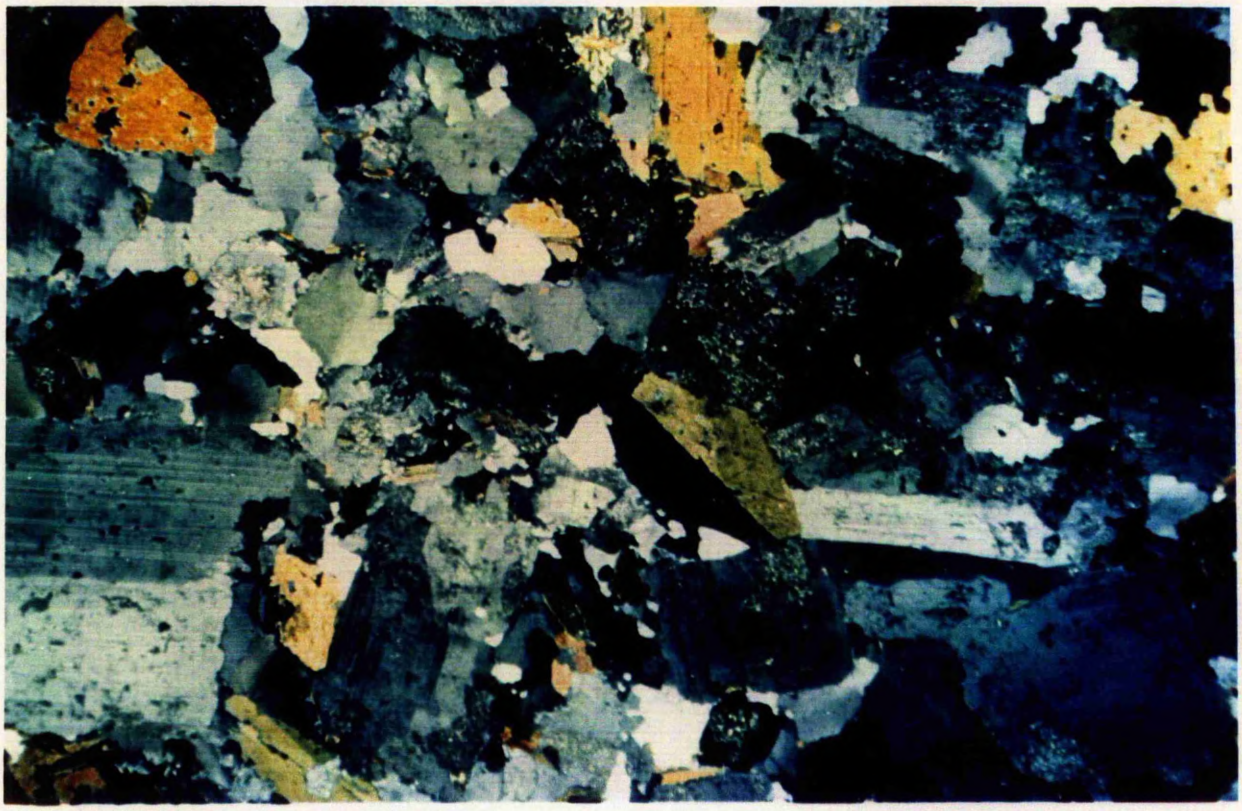


Plate 3.3 Criffell granodiorite texture, sample C350G (X 24).

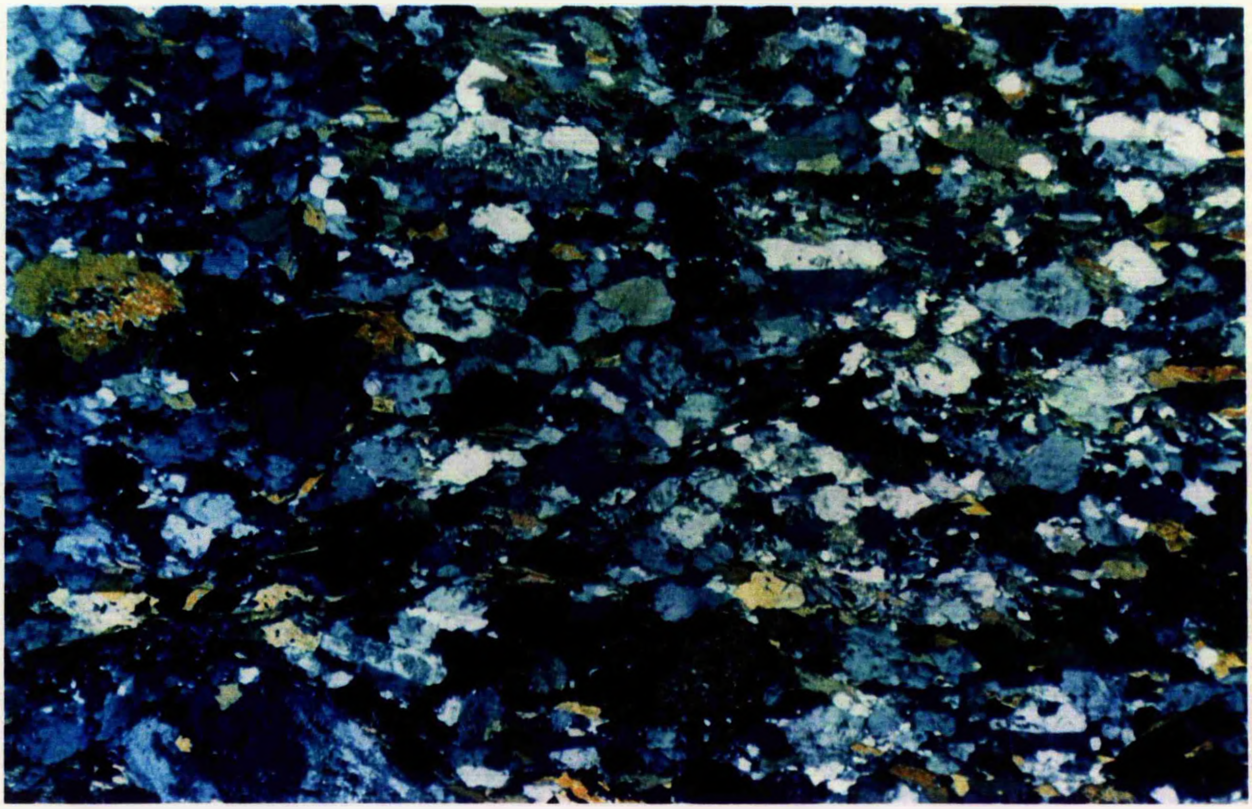


Plate 3.4 Typical mortar texture close to the inner and outer pulse contacts, sample C205G (X24).

Large euhedral plagioclase forms grains (up to 6mm) which frequently show oscillatory zoning from An35-40 at the core to An15-20 at the rim. K-feldspar occurs as late subhedral to anhedral grains of microperthite and microcline. Quartz is common and late, forming irregular shaped grains around plagioclase and K-feldspar.

Hornblende is the most common ferromagnesian mineral forming early euhedral crystals often with a core of partially or mostly resorbed clinopyroxene (Plate 3.3). Embayments around the periphery of the grains indicates that the hornblende had become unstable during the later stages of crystallisation and was being resorbed. Biotite is late, forming small (2mm) intersertal grains between plagioclase crystals.

Accessories are mostly in the form of euhedral sphene, often associated with an opaque phase. Apatite forms small squat grains enclosed within biotite and K-feldspar, whilst zircon is mostly euhedral and associated with biotite. Allanite is common.

As the contact with pulse 2 is approached, the intensity of the deformation within pulse 1 increases from the margin inwards. Close to the contact, pulse one granodiorite begins to take on the appearance of augen gneiss. A mortar texture (Plate 3.4) is developed with large plagioclase and smaller hornblende grains forming "eyes" in a matrix of comminuted quartz, biotite, plagioclase and K-feldspar. Most of the minerals show evidence of intense strain. It is within this area, at the interface between the outer and inner pulses, that inclusions attain their maximum degree of recrystallisation.

Hornblende biotite granite

A collection of samples was made from the outer portions of pulse 2 as this is the only section of this member which contains inclusions.

Early euhedral plagioclase shows oscillatory zoning from An35 at the core to An20 at the rim. K-feldspar is more common than in the outer granodiorite, and is of large (up to 15mm) euhedral grains of microperthite. Lesser microcline is also apparent as anhedral intersertal minerals. Quartz, like microcline, is late and anhedral, taking up interstitial positions.

Biotite dominates the ferromagnesian minerals as late intersertal grains. Hornblende, often being replaced by biotite, forms small (2mm) euhedral grains with embayed margins. Clinopyroxene is extremely uncommon but on occasion can be identified at the core of partially resorbed hornblende.

Accessories are similar to those of pulse 1.

CHAPTER 4

THE APPINITES OF STRONTIAN

4.1 INTRODUCTION

The term ‘appinite’ was first used by Bailey et al. (1916) to describe a diverse suite of minor intrusive, ultra-basic to basic rock-types found throughout the Moine and particularly the Dalradian rocks of the British Caledonides. Few major attempts at discerning the origin of these minor intrusives have been undertaken since the classic works of Nockolds (1941) and Deer (1938 & 1950) on the Garabal Hill and Glen Tilt complexes respectively. Field reviews were given by French (1966) and Pitcher & Berger (1972) for Donegal whilst geochemical data were presented by Hall (1967) and Wright & Bowes (1979).

Previous workers recognised the spatial and temporal relationship between appinites and major calc-alkaline plutons, and concentrated their efforts on establishing a clear genetic link between the two. In this chapter, detailed field, petrographic, geochemical and isotopic studies of four appinite bodies within the Strontian pluton were undertaken in an attempt to understand the following:

1) The source of the appinites and the nature and conditions under which the appinites were generated.

2) The genetic relationship (if any) between the appinites and surrounding granitoids.

3) The relative timing of appinite and granitoid intrusion.

4) The degree of interaction between appinites and the granitoids.

5) To test the hypothesis that stoping of the large appinites produced the small dark fine-grained inclusions.

Nomenclature

The type appinite consists of a dark coarse-grained rock comprising large euhedral hornblendes in a matrix of plagioclase with a little quartz and orthoclase. However the term 'appinite suite' was used by Bailey et al. (1960) to include a diverse range of minor intrusives from ultra-basic through basic to intermediate in character, all of which contain essential hornblende and plagioclase. Bailey et al. (1960) associated the appinites suite with other minor intrusives of a less volatile rich nature such as the pyroxene-mica-diorites of Nockolds (1941) and kentallenites of Hill & Kynaston (1900), which together with members of the appinite suite are clustered around major late Caledonian calc-alkaline granitoids.

French (1966) produced a textural classification of the appinite suite which is used here. Essentially, members of the appinite suite may be divided into two series depending upon the habit of hornblende. Those rocks containing idiomorphic hornblendes are assigned to the appinite series whilst those containing hornblendes as glomeroporphyritic aggregates are assigned to the dioritic series (Fig 4.1). Both the appinite and diorite series contain essential hornblende and plagioclase +/- quartz +/- orthoclase commonly with clinopyroxene and/or biotite, making a significant contribution to the mode. It must be emphasised that the appinite suite is extremely variable with one

THE APPINITE SUITE

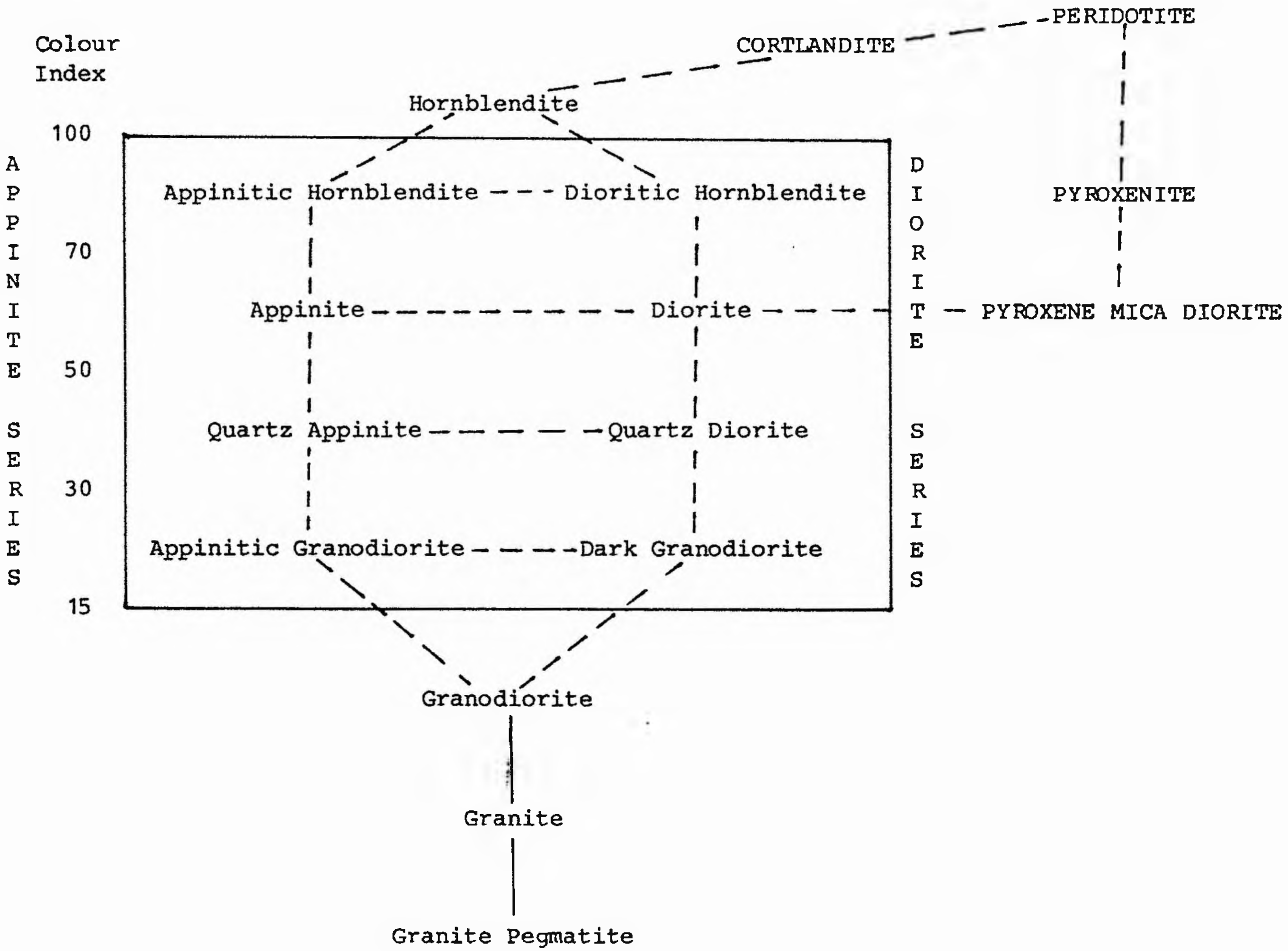


Fig 4.1 Nomenclature of appinites (after French 1966).

grouping merging into the next with in some cases, both idiomorphic and glomeroporphyritic hornblendes within the same rock.

Previous studies

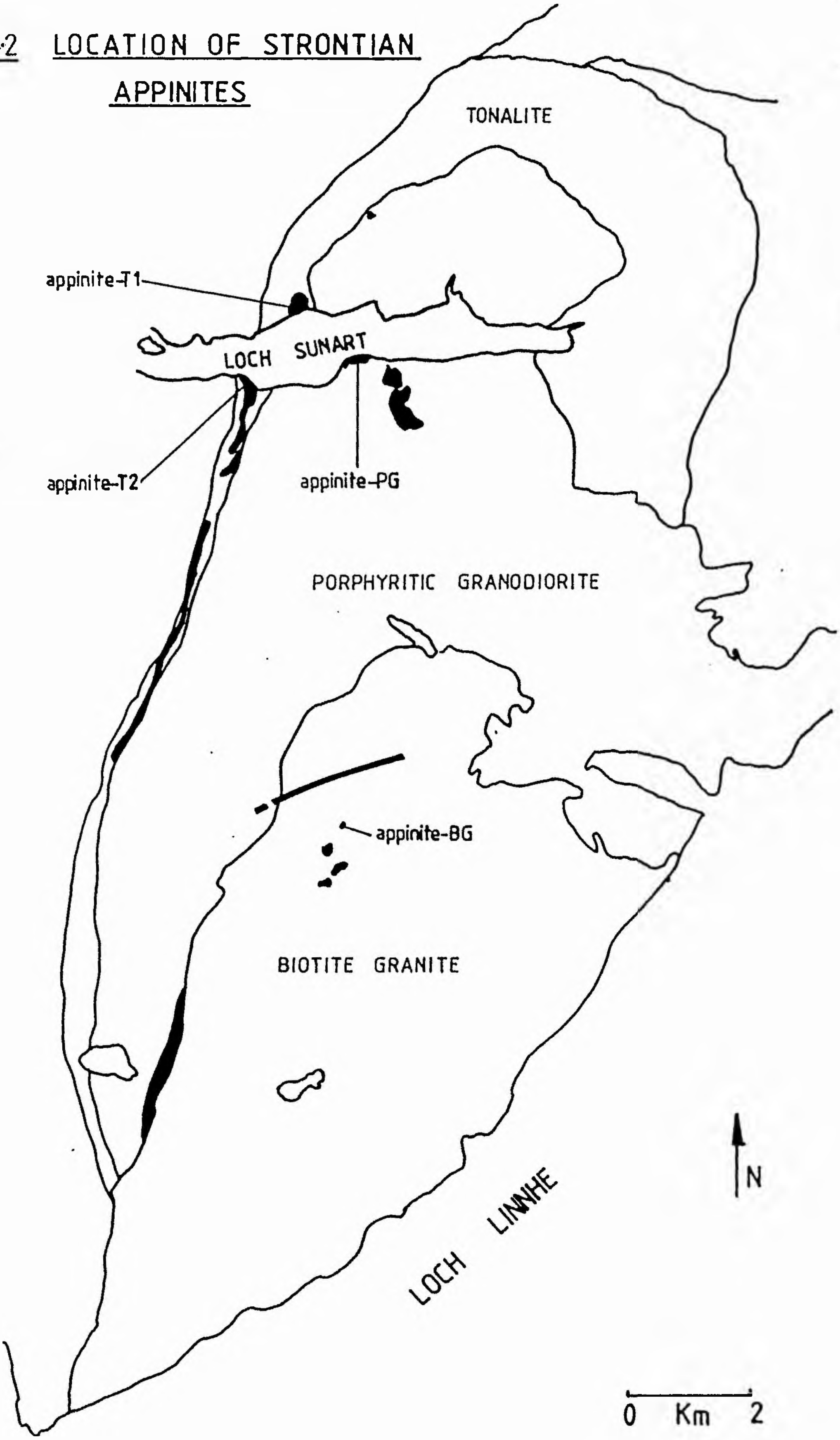
Hall (1967) found that only granitoids of a well defined composition were associated with appinites thus implying a genetic link between the two. Nockolds (1941), in his now classic study of the Garabal Hill complex, used major element criteria, notably the enriched nature of the MgO in the appinites, to suggest that these were the cumulitic products of the fractional crystallisation of a pyroxene-mica-diorite parental magma and that granitoids were the fractionation products of the same parent.

Appinites and associated ultra-basic to basic minor intrusives occur as bosses, dykes and occasionally sills. Clear association of appinites with explosive structures has been demonstrated by Bowes & Wright (1964) in the Ballachulish area, and by Pitcher & Berger (1972) in Donegal. In many respects this explosive activity, interpreted as the catastrophic release of gases from a volatile rich magma (Bowes & Wright 1964), is reminiscent of the emplacement mechanism of kimberlites (Dawson 1979 p.31-38). Here, volatile rich kimberlite magma is thought to be intruded into the crust along rapidly propagating fissures, the rate of intrusion being controlled by the rate at which magma can fill the fracture (see Spera 1984 for a discussion of intrusion mechanics). Near the surface degassing of the magma becomes rapid, building up an over-pressure above hydrostatic and any structural weakness within the region of degassing will naturally become a site of easy passage for the magma. When a structural weakness is penetrated close to the surface (1-3Km) gas-streaming may occur disrupting the enclosing country rock. Further intrusion takes place explosively as a matrix of disrupted wall-rock, magma and magma-xenolith material are emplaced together in a funnel shaped diatreme resulting in a magmatic breccia. After the initial violent phase further intrusion of magma may be ‘passive’ (Dawson 1979).

The intrusive breccias described by Bowes & Wright (1964) and Pitcher & Berger (1972) for appinites associated with the Ballachulish and Ardara plutons respectively, show many similarities to those described by Dawson (1979) for kimberlites. However unlike kimberlites, the appinites seldom carry mantle nodules (except within the more alkali-rich lamprophyres), and are associated with major calc-alkaline plutonics. Furthermore kimberlitic materials are seldom found within fold belts being usually confined to a cratonic setting, whereas appinites and lamprophyres are found within fold belts associated with the end-stages of orogenesis (Thompson et al. 1984). Thus whilst the emplacement style of the appinites may be in some ways similar to that of kimberlites the tectonic setting of these two rock-types is quite different.

The Strontian appinites are not typical of the suite however in that they seldom contain the large euhedral hornblende crystals so typical of the Appin and Donegal appinites (Bailey et

Fig 4·2 LOCATION OF STRONTIAN APPINITES



al. 1960, French 1966, Wright & Bowes 1979). In this respect the Strontian appinites are similar to the hybrids of Ach'uaine type described by Read & Phemister (1925) in Sutherland. The Ach'uaine hybrids were considered by Read & Phemister (1925) to be the product of interaction of ultra-basic and acid magmas. Such hybrids may also be found within a host granitoid at Rogart (Lockett 1986), an observation which is seldom made for true appinite. This study then is a first attempt to discover the origin of hybrids of Ach'uaine type and to relate such origins to the formation of the host pluton.

4.2 FIELD RELATIONS

Mapping by MacGregor & Kennedy (1932) and by Sabine (1963) showed that tonalite is seen to penetrate some of the appinites and that the appinite bodies outwith the pluton cut the regional structure. They concluded that the appinites were older than the tonalite but younger than the regional Moine structure and were granitoid precursors. The appinites outcropping within the granitoid were the remains of appinite bodies stoped from the country rock envelope.

Fig 4.2 brings this stoping hypothesis into question. A dyke of appinitic affinity (MacGregor & Kennedy 1932) may be seen to cut the contact between the biotite granite and the porphyritic granodiorite, and must have been emplaced after the granitoids, precluding stoping as a viable mechanism for the emplacement of this dyke.

Four appinite bodies were chosen for detailed study based

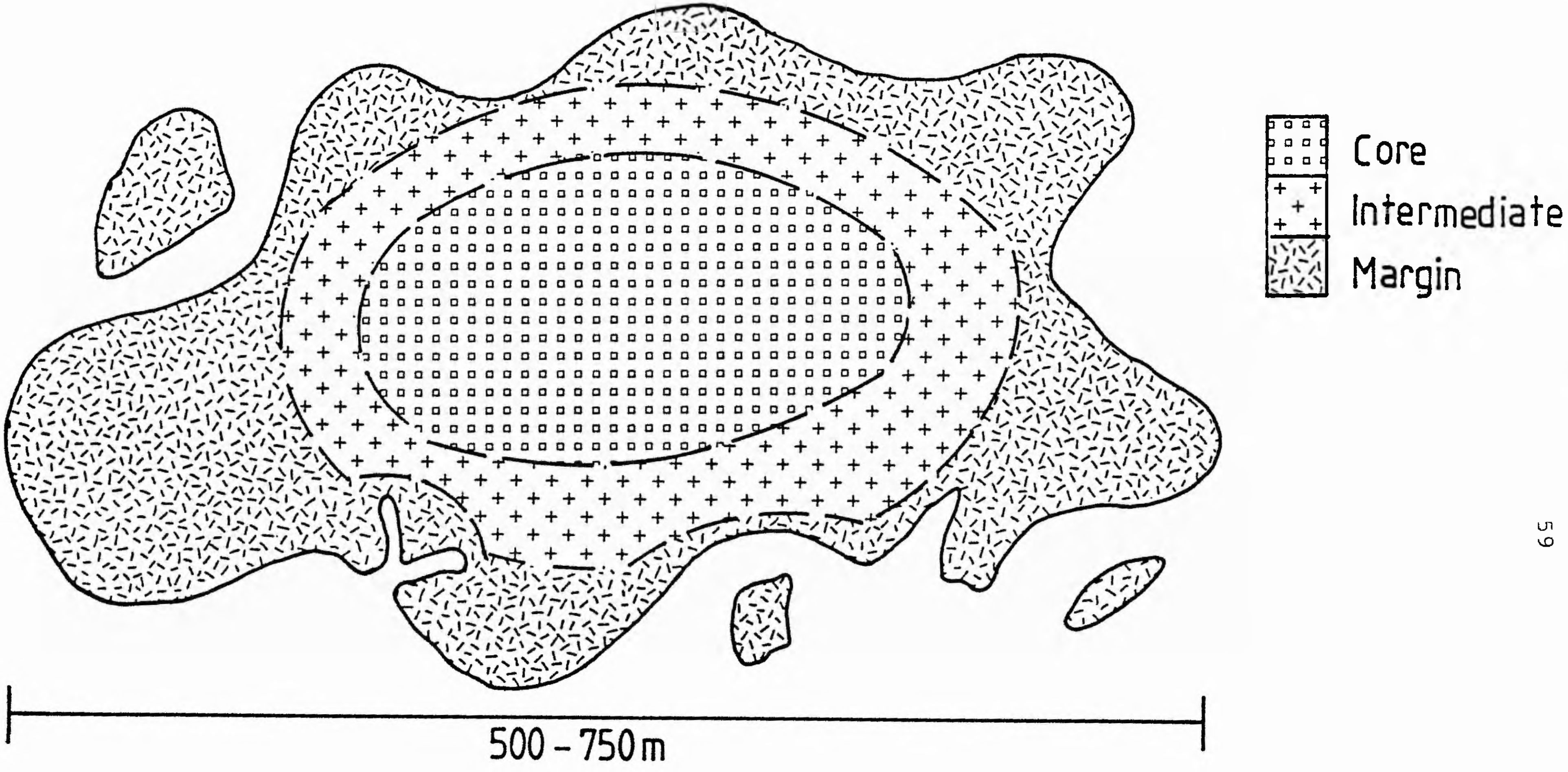


Fig 4·3 Schematic Representation of an Appinite Body upon the degree of exposure and their location (Fig 4.2). The four appinite bodies mapped two in the tonalite (appinites-T1 & appinite-T2) and one each in the granodiorite (appinite-PG) and the biotite granite (appinite-BG) show broadly similar relationships to their host granitoid. Each appinite was mapped in fine detail (with the exception due to poor exposure of appinite-BG) and the new maps are enclosed.

Field observations.

1) All four appinite bodies are between 200-750m in diameter, and are steeply-sided with irregular and often lobate contacts. In other instances near the external contacts these bodies have a lenticular pattern in outcrop (e.g. at NM761550), the constituent minerals indicating a strong foliation.

2) Within a given body there is extreme variation both in terms of colour index and texture. Overall they consist of a relatively homogeneous, coarse grained, biotite-hornblende rich appinite core (C.I.>75), surrounded by a rapid but gradational change to heterogeneous rocks of finer grainsize and more leucocratic aspect (C.I.typically between 50 to 60). Internal contacts were not observed, and the bodies are thus reversely zoned (see Fig 4.3 for schematic representation).

3) Marginal zones are extremely variable and often veined by late-stage, fine grained, aplitic material which may be observed to isolate angular fragments of appinite (Plate 4:1). Host granitoid may also be observed to vein the appinites but the

Plate 4.1 Aplite veining and isolating appinite fragments, Appinite-T2.

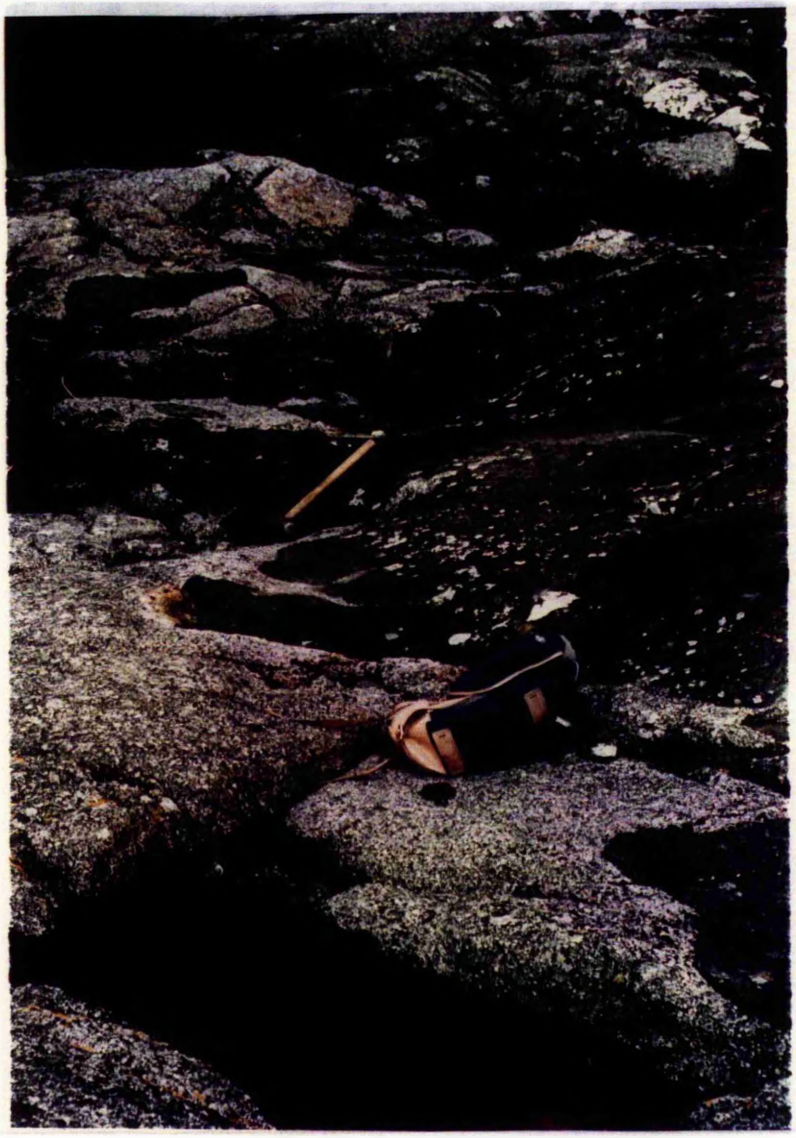


Plate 4.2 Appinite-PG contact with granodiorite (NM 749602).

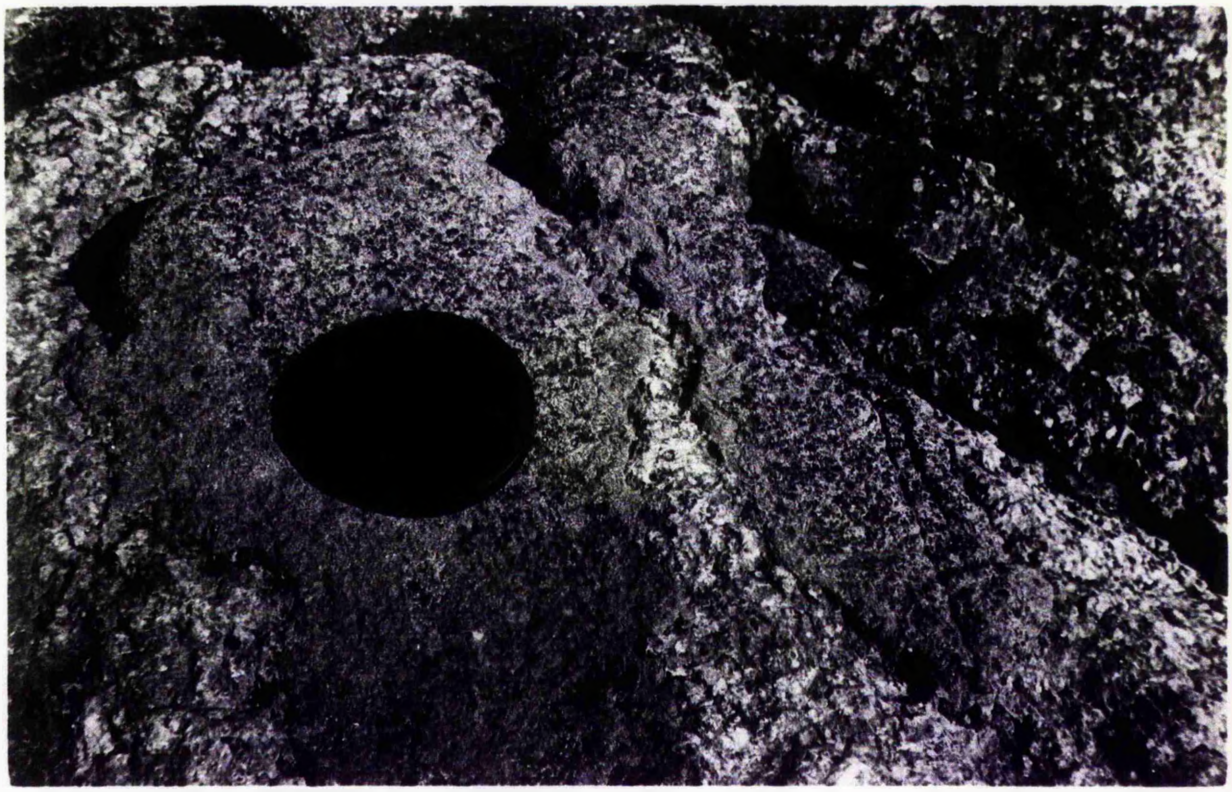


Plate 4.4 Chilled margin of 'pillow-like' structure (NM 749600),

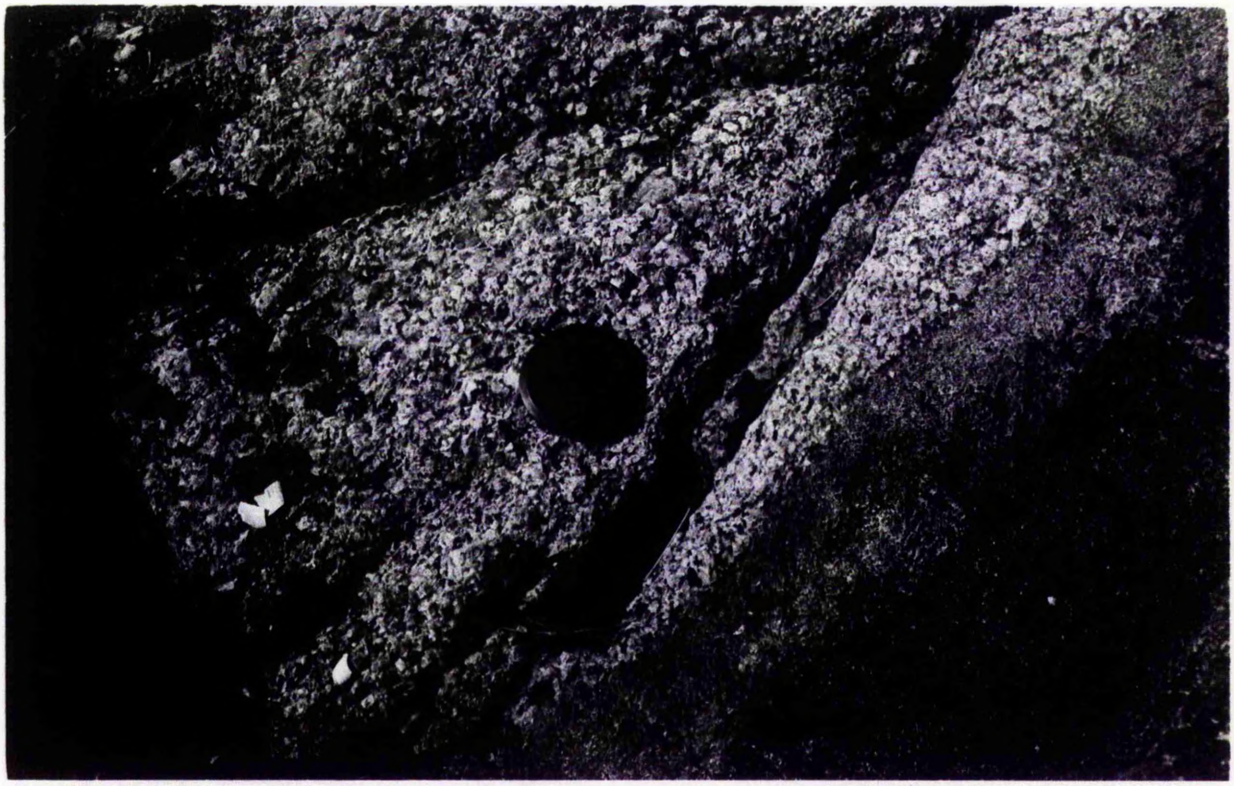


Plate 4.3 Close up of contact between appinite-PG and granodiorite. Note the finer grained 'chilled' nature of the appinite against the host.

granitoid material is nowhere seen in close association with coarse-grained central appinite.

4) Contact relations between host and appinite are sharp, irregular and often lobate as at NM749602 (appinite-PG and Plate 4.2). Here the margin of the appinite shows evidence of chilling and the rapid skeletal growth of ferromagnesian minerals (Plate 4.3). Furthermore, 10m to the west a pillow-like structure complete with cuspate margins and exhibiting rapid growth textures (Lofgren 1980) may be observed (Plate 4.4). This appears to be the product of quenching of the appinite against the host granodiorite.

5) There is no local increase in the amount of small dark inclusions observed within the surrounding host granitoid, suggesting that the appinite bodies were not necessarily the source for smaller inclusions. However, in the biotite granite the local host rock contains clots of hornblende similar to those observed within appinite-BG. Furthermore the abundance of these clots increases as appinite-BG is approached indicating interaction between the appinite and its host had taken place. The biotite granite must therefore have been in a 'liquid' condition in order to distribute the mafic clots. Conversely appinite-BG shows a distinctive chill or quench texture against the biotite granite, indicating that it also had a melt component at the time of intrusion.

6) Large (1 to 2m) rounded but irregular fragments of appinite are seen to have been 'spalled' off the major bodies.

These are bigger than average inclusion size but do not show any tendency to be associated with smaller inclusions.

7) No country rock fragments were observed in association with the appinites.

Mode of emplacement of the appinites.

Appinites distributed from around Appin and Donegal (French 1966, Pitcher & Berger 1972, Bailey et al. 1960) were intruded into country rock and are typically of small dimension (<1Km in diameter). Compared with their counterparts intruded into the country rock, appinites within the Strontian pluton are similar in size but lack associated country rock fragments. The existence of a chilled margin, the large detached fragments (1-2m), the irregular and lobate nature of the contact and the reversed zoning within each of the appinite bodies, leads to the conclusion that the appinites were injected into their host granitoid magma whilst the host was not totally consolidated. Late stage movements within the host, possibly aided by heat from the cooling appinite, promoted remelting and back-veining. On field evidence only, it would appear that the appinite intruded the granitoids under relatively quiescent conditions to form an homogeneous core region as no magmatic breccias were found. However, the force of injection may have been sufficient to promote co-mingling of crystals and magma derived from the granitoid with magma from the appinite in varying proportions. Thus the appinites within the Strontian pluton are characterised by the presence of a relatively homogeneous basic core surrounded by an heterogeneous intermediate margin consistent with magma injection into an unconsolidated granitoid coupled with varying degrees of co-mingling and chilling at the periphery.

4.3 PETROGRAPHY

Petrographic descriptions of appinites from the type area and elsewhere (Bailey et al. 1960, French 1966, Pitcher & Berger 1972) indicate the variability and textural complexity of the suite. Read & Phemister (1925) in considering the similar ACH'UAINE hybrids in the Sutherland region, concluded that they were hybrid between granitic and ultra-basic end-members, and that the textural complexity was a function of magma co-mingling. Clearly in individual intrusions of small size, textural and compositional variability could be generated by a number of processes namely magma-mixing, fractionation, assimilation, variability of the intensive parameters such as volatile content and oxygen fugacity etc. It should be emphasised that although not explicitly stated, the appinite bodies intruded into Dalradian country rock studied by French (1966 and reviewed by Pitcher & Berger 1972) are in fact reversely zoned, which indicates that magma mixing is not the only mechanism for generating reversely zoned bodies.

Petrographic observations.

Appinites T1 & T2

1 Core: Since these bodies are reversely zoned, the central more basic rocks are described first, followed by descriptions of the more evolved rocks.

Large euhedral green and brown hornblendes dominate both the mode and the texture (Plate 4.5). Where free growth has been permitted the hornblendes show good idiomorphic forms and are between 5 and 8mm in size. Almost ubiquitously these hornblendes contain cores of highly pleochroic brown biotite. Clinopyroxene (augitic) cores to amphiboles are rare in the core region of appinite-T1 but are more common within appinite-T2. A later intensely pleochroic biotite may be found in the interstices between the hornblende grains, but these are not common.

Although hornblende dominates the centres of the appinites as idiomorphic crystals, it is interesting to note that these hornblendes are frequently cored by biotite; indeed those not possessing a biotite core are demonstrably later, less well formed and of a tremolitic-hornblende composition. Occasionally an augitic clinopyroxene may be observed at the core of these hornblendes, and less commonly both clinopyroxene and biotite are enclosed within the same idiomorphic hornblende. Thus biotite, which shows clear evidence of partial resorption, appears to have crystallised prior to the amphibole (see Plate 4.5a). This is not in accord with Bowen's discontinuous reaction series although it may be argued that the biotite represents the preferential replacement of augitic cores to the amphiboles. These augitic cores themselves show evidence of partial resorption by the amphibole, so replacement by biotite may give the impression that

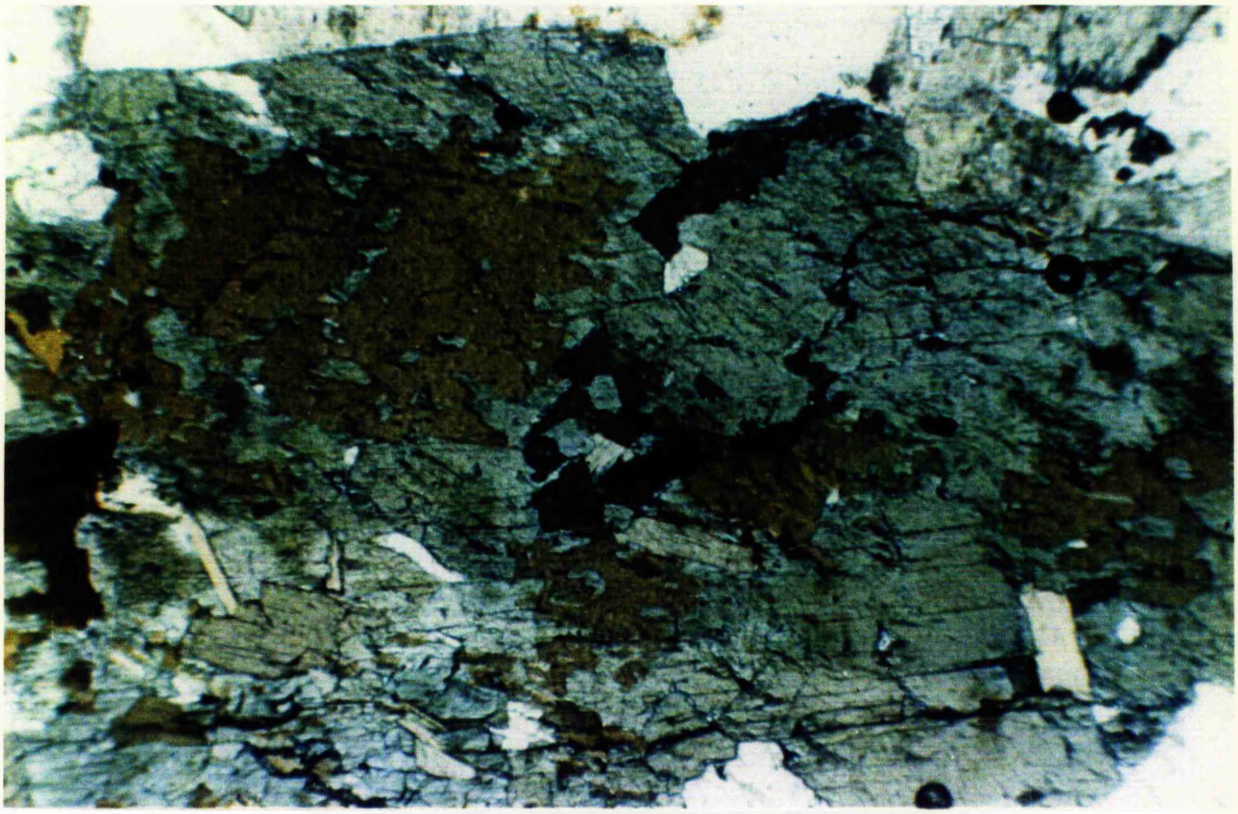


Plate 4.5a Large euhedral hornblendes with cores of biotite, sample S318A, appinite-T2 (X48).

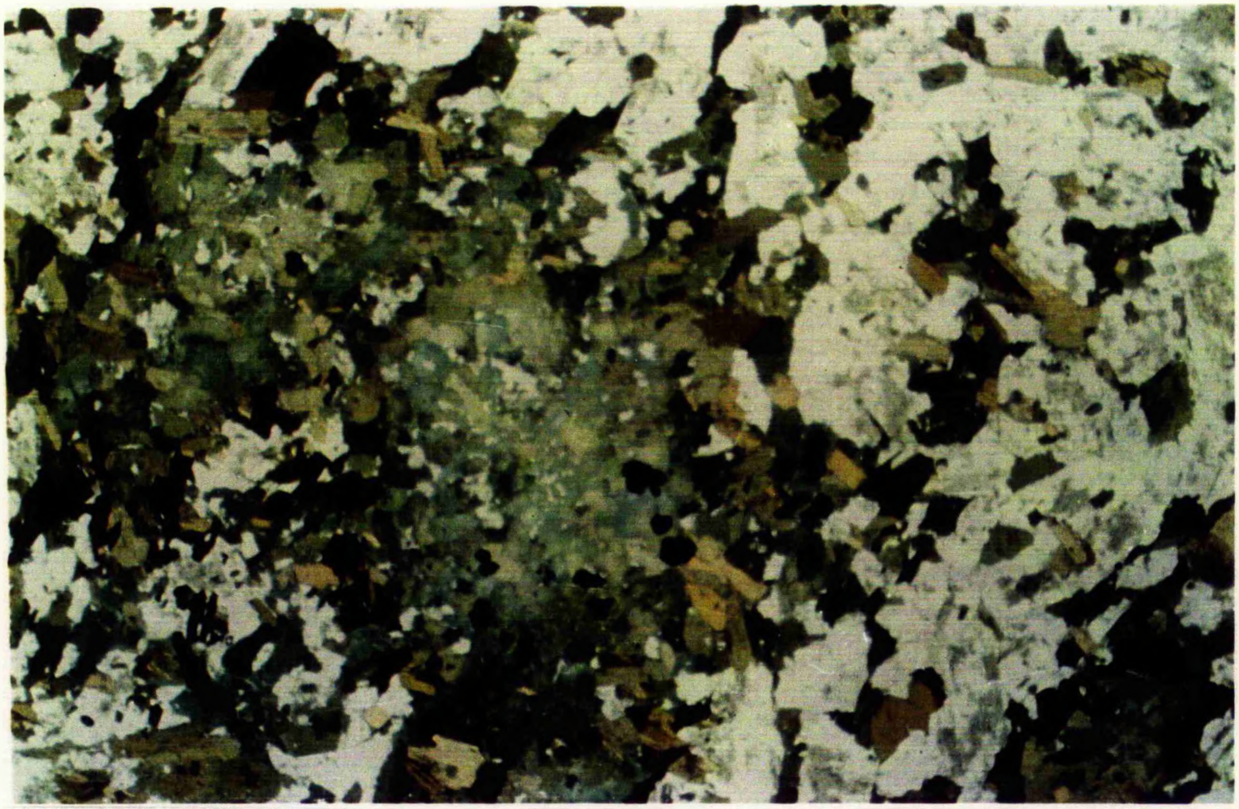


Plate 4.5b Aggregations of hornblende and biotite in appinite-PG, sample S336A (X24).

the biotite preceeded the amphibole (see for example Joplin 1935).

This apparent reversal of Bowen's discontinuous reaction series has been observed elsewhere (Read 1926, Pitcher and Berger 1972, Naney and Swanson 1980, Wones 1981, and Leighton 1985). Experimental work by Naney and Swanson (1980), observed biotite crystallisation in the order of 100 degrees above the predicted liquidus temperature in the granodiorite system. To account for this, they suggested that catalytic nucleation of the biotite upon the walls of the experimental capsule may well be important. Indeed, this catalytic effect may well be the key to early biotite nucleation as all the biotites observed within the appinites are of somewhat skeletal (rapidly grown?) aspect. Wones (1981) in considering biotite-amphibole relationships states; " in magmas containing low f(H_2O) and moderate K_2O values, biotite can follow clinopyroxene ahead of amphibole." Appinites-T1 & T2 are a classic example of this reversal of Bowen's discontinuous reaction series.

Poikilitic plagioclase (An35) is the dominant felsic mineral with much lesser quartz and micro-perthite as late interstitial minerals. Plagioclase within appinite-T1 retains a pristine aspect whilst plagioclase in appinite-T2 contains clouded sericitised cores and later unclouded albitic overgrowths. This suggests fluid interaction with the feldspars prior to consolidation as there is also a tendency for the hornblendes within appinite-T2 to be replaced by an aggregate of tremolitic hornblende or have calcite parallel to the cleavage planes.

Accessory minerals are typically euhedral (and rather large 1-2mm) apatite, with anhedral and late sphene. Zircon (euhedral) and allanite are also occasionally present.

2 Intermediate zones

As in the core region, idiomorphic hornblende dominates, retaining its characteristic core of partially resorbed biotite. These appear to be slightly less well formed than in the core region, with a tendency for the hornblende to have ragged terminations or be partially or wholly replaced by tremolite-hornblende aggregates.

Plagioclase shows the greatest difference from the core region, with many crystals showing scalloped margins and sericitic alteration perhaps indicative of an abrupt change in the physico-chemical environment. Microperthitic potash feldspar becomes more common with distance away from the core and generally occurs as a poikilitic mineral including all other minerals except quartz and late biotite. Accessory minerals are similar to those in the core region.

3 Margin

The marginal areas in contact with the host tonalite show a different texture to those a couple of meters away in the intermediate zone. In the contact area amphibole is always in the form of aggregates of tremolitic hornblende which appear to be pseudomorphs after earlier hornblende (and clinopyroxene?). Late brown biotite is generally wrapped around these aggregates (Plate

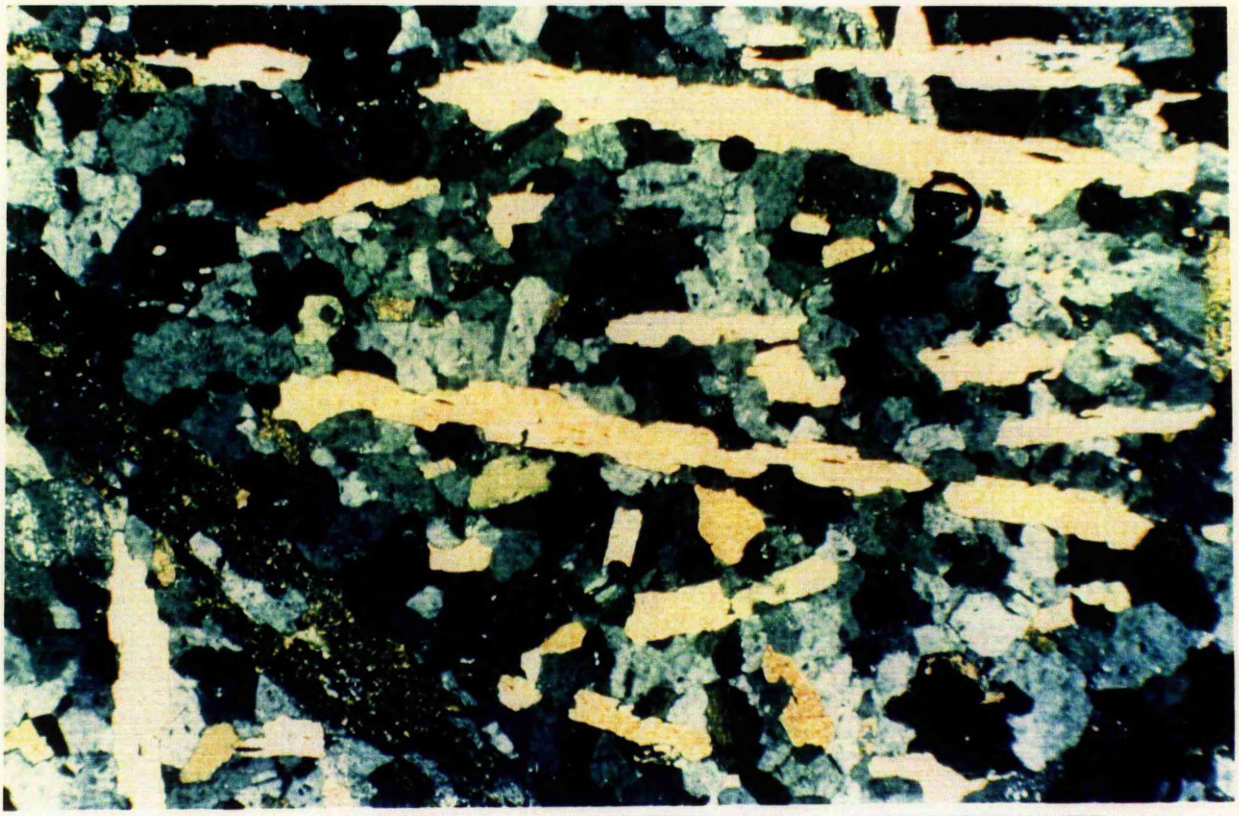


Plate 4.6 Rapid growth texture exhibited by skeletal biotite at the margin of appinite-PG, sample S427A (X48).

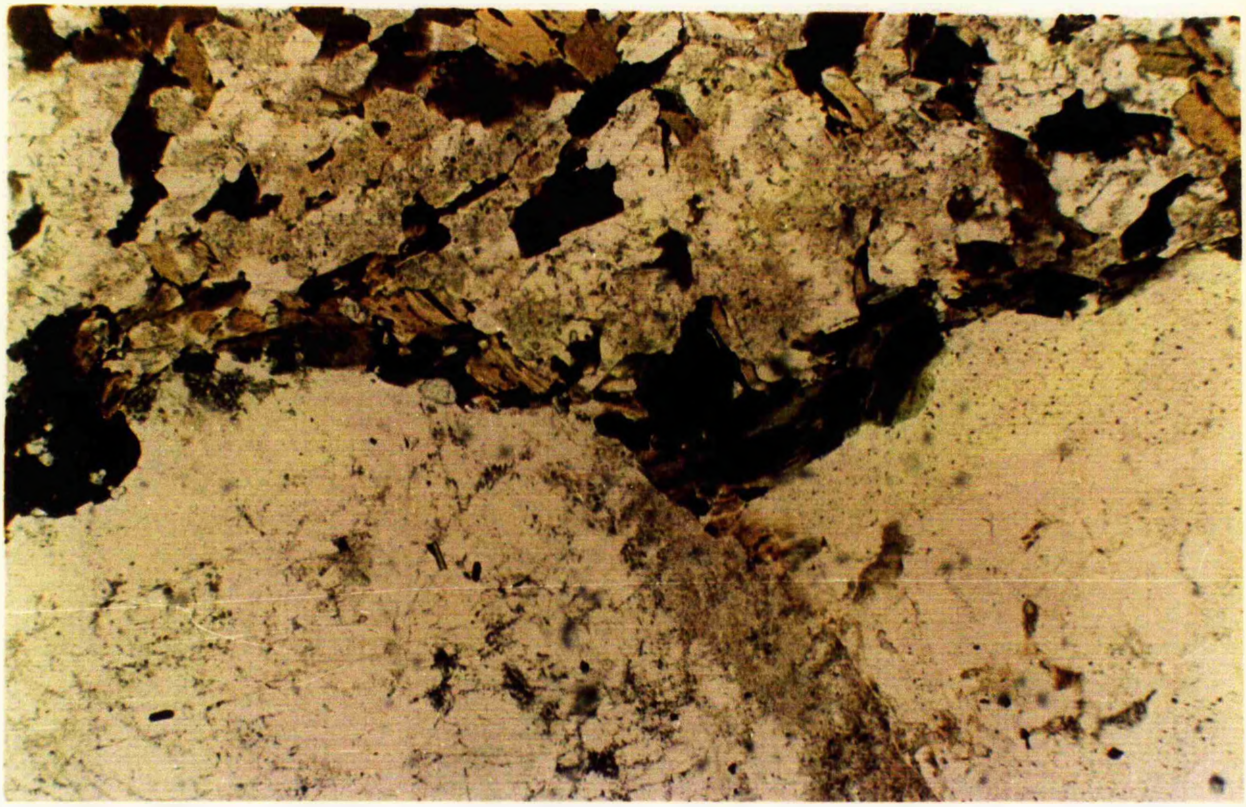


Plate 4.7 Appinite-PG contact with granodiorite. Note that biotite is nucleated on the host granodiorite grains (X48).

4.6). Microperthite is common and like plagioclase and quartz forms the granular matrix between the knots of amphibole.

Plagioclase megacrysts, free from inclusions, may be observed close to the contact with the host granitoid. These are optically identical to those within the host tonalite and contain similar oscillatory zoning. Within appinite-T2 a fabric pervades the body but is particularly intense at the margins and the fabric is seen to wrap around the plagioclase megacrysts indicating the existence of the megacrysts prior to the fabric forming event. Apatite which in the core and intermediate regions is euhedral, is acicular within the contact region and included within all other matrix minerals but not within plagioclase megacrysts or the centres of amphibole aggregates. In this respect the texture at the margin of the appinites is similar to those described for small dark inclusions (Chapter 5).

The dioritic series.

Those appinites outcropping within the inner two members of the pluton do not contain abundant idiomorphic hornblendes or megacrysts of plagioclase in their marginal zones. Occasionally apparently relict idiomorphic hornblende with a core of biotite may be seen, which suggests that the aggregates are the product of replacement of primary amphibole by aggregates of tremolitic hornblende (see Plate 4.5b).

Core.

The texture is dominated by late biotite which overgrows any pre-existing textures. Amphibole is either a felted mass of tremolite commonly associated with green biotite, or aggregates of tremolite-hornblende pseudomorphs after hornblende and possibly clinopyroxene, giving the rock a glomeroporphyritic texture.

Poikilitic plagioclase with a little orthoclase fills the interstices between the amphibole aggregates and often these plagioclases are turbid and sericitised. Accessory minerals are mainly late anhedral sphene and a euhedral apatite up to 1mm across.

Intermediate zone.

Feldspar, particularly potash feldspar, is modally much greater than within the core. Occasionally partly resorbed but still euhedral hornblende crystals are preserved within poikilitic plagioclase grains. Late microperthite commonly contains myrmekitic intergrowths where it abuts earlier plagioclase.

For the most part the amphiboles are again aggregated and associated with calcite, magnetite and sphene as reaction products of earlier ferromagnesian mineral breakdown. Accessory apatite is much smaller than within the core.

Margins.

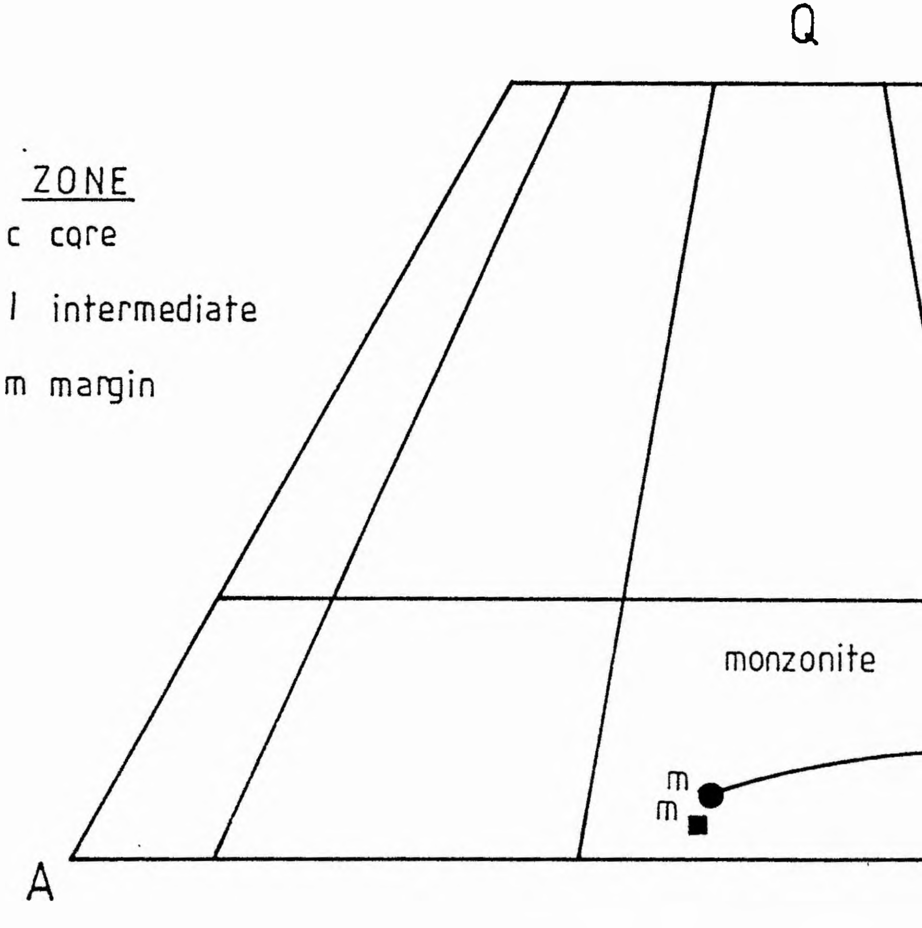
At the zone of contact between appinites-PG & -BG with their host granitoids, a very distinctive texture is developed (Plate 4.4), where skeletal biotites up to 1cm long are randomly distributed throughout the rock. At the very margin of appinite (3), biotite is observed as having nucleated on mineral grains in the granodiorite (Plate 4.7), and it may be concluded that the host was largely consolidated at the time of appinite emplacement. Further confirmation of this time sequence comes from the broken nature of quartz crystals within the host granitoid close to the appinite.

The appinite matrix is of plagioclase, potassium feldspar and quartz with a granular texture, ferromagnesian minerals being either knots of amphibole or skeletal biotite. Small acicular apatites are abundant being included within all the matrix minerals except the amphibole aggregates. Within appinite-BG the biotite is a green variety whereas the biotite within appinite-PG is an intensely pleochroic brown. No megacrysts of plagioclase were found within either appinites-PG or -BG.

Discussion.

At Strontian the overall gross similarities of the appinites far outweigh the differences in their amphibole morphology. Indeed, the dioritic series rocks contain relict idiomorphic hornblendes essentially identical to those within the rocks of the appinite series, which together with their similar outcrop pattern, suggests that the only major difference between the two rock-types is the conversion of idiomorphic hornblende to aggregates of tremolitic hornblende. Late stage fluids (possibly H_2O and CO_2 rich) derived from the granitoid may have been the medium of alteration as suggested by the turbid nature of the

Fig 4·4 Appinite modal



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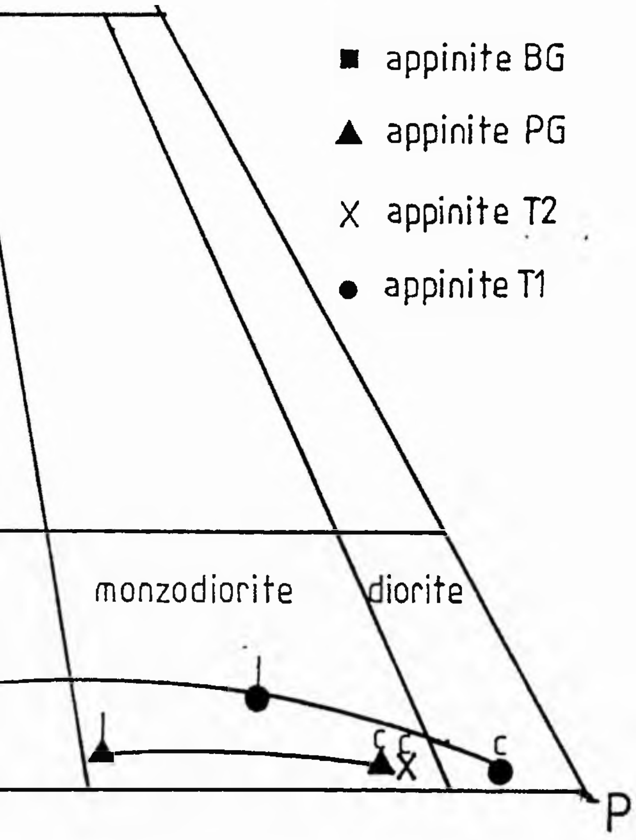

■ appinite BG
▲ appinite PG
X appinite T2
● appinite T1
monzodiorite
diorite
c
c c
c
P

plagioclase grains within rocks of the dioritic series.

All four appinites show similar chill textures against their host, often with the contacts being typical of the mutual coexistence of two viscous liquids (Wager & Bailey 1953, Furman & Spera 1985).

Fig 4.4 shows that appinite marginal zones have the highest modal percentage of K-feldspar whilst the core region has the lowest modal K-feldspar and a more refractory mineralogy. This reverse zoning cannot simply be explained by a mechanism of in-situ fractional crystallisation from the margins inward as the lowest temperature minerals are found closest to the cooling surface. Other possible processes must be sought to explain the zonation and it is worthy of note that the core regions are generally homogeneous in both texture and composition within all four bodies.

In summary, appinites occur within all three members of the Strontian pluton as rounded 'pipes', with the presence of a marginal chill at the periphery of each appinite yielding strong evidence for the existence of an appinite liquid, which may or may not have interacted with its host. The lobate nature of the contacts (Plate 4.2) and the chilled nature of the appinite abutting its host, are powerful evidence that the appinite intruded into a still unconsolidated granitoid magma and was either younger than, or contemporaneous with, its host. Appinites within the Strontian pluton are not the stoped remains of granitoid precursors intruded into the Moine country rock as suggested by MacGregor & Kennedy (1932), but are basic magmas coexisting in space and time with their granitoid hosts.

Interaction of the appinites with their host rocks in terms of geochemical exchange, and the significance of the resulting rocks to the generation of microgranitoid inclusions, are dealt with in Chapter 7.

CHAPTER 5

FIELD RELATIONS AND PETROGRAPHY OF THE INCLUSIONS.

5.1 NATURE OF INCLUSIONS

Inclusion distribution

A common observation within zoned granitoid plutons is that inclusions are concentrated in the less evolved intermediate portions of the pluton, generally the periphery of the pluton (Pabst 1928, MacGregor & Kennedy 1932, Thomas & Campbell-Smith 1932, Bailey 1917 & 1960, Pitcher 1978 & 1979). Furthermore the observation within some granitoids, that country rock fragments are contained within a narrow marginal zone whilst dark microgranular inclusions are more evenly distributed, lead to the conclusion that the latter are derived from the former by processes of assimilation (Michel-Levy 1893, Daly 1914 & 1933, Nockolds 1933, Sherlock & Hamilton 1958, Bateman et al. 1963).

Rounded microgranitoid inclusions are, in general, evenly distributed throughout a given member of a pluton, but local accumulations have been described. Pabst (1928) described inclusion 'swarms' in the Sierra Nevada batholith which cover from many hundreds down to a few square meters, whilst Link (1970) described similar 'trains' within the Half Dome quartz monzonite (Yosemite, National Park USA). Link proposed that these trains were granitic flotsam caught in the dead zone of convecting magma in the same way as flotsam collects in the dead zones of flowing rivers. Other workers have proposed that these swarms are the product of the break-up of dyke material (Adams & Barlow 1910, Clapp 1921, Dixely 1925, Wegmann 1963, Ganser & Gyr 1964, Pitcher & Berger 1972, Weibe 1973, Taylor 1976, Reid et al.

1983, Furman & Spera 1984). If the latter suggestion is correct the relative chronology of dyke emplacement to the melting episode within the granitoid source region is not easily determined. Wegmann (1963) showed that pre-existing dyke swarms within the source region (and unrelated to the cause of melting) may be attenuated and broken to form inclusions. Eichelberger (1975) suggested that sheets of mantle-derived material may provide the heat necessary for crustal melting, that is dyke emplacement and granitoid magma production were synchronous. Pitcher & Berger (1972) demonstrated that dyke intrusion occurred late in the intrusion history of the Thorr granitoid (Donegal, Ireland), but prior to the consolidation of the pluton. They suggested the term ‘syn-plutonic’ to describe such dyke occurrences. It would appear therefore, that the intrusion of mantle-derived material may take place throughout the production and emplacement of a granitoid magma, yielding ample opportunity for the formation of inclusions both randomly distributed and in locally concentrated patches.

Inclusion shape

Fragments of country rock within a pluton seldom have a rounded outline but may occasionally have diffuse boundaries due to processes of assimilation (Bowen 1922). However igneous inclusions and inclusions of uncertain origin are generally rounded (Didier 1973). The rounded nature has been interpreted as the product of mechanical attrition and chemical assimilation

(Powers 1915), as a primary feature due to mechanical break-up (MacGregor & Kennedy 1932) or as the product of magma chilling processes similar to those observed in the formation of pillow basalts (Wager & Bailey 1953, Windley 1965). Pabst (1928) noted that in some plutons the sphericity of a given inclusion depended upon the angle at which it was viewed and its distance from the pluton margin. The increasingly lenticular nature of inclusions has allowed the mechanism of intrusion of the host pluton to be investigated (Ramsay 1975, Holder 1979, Hutton 1981, Bateman 1985) using deviations from sphericity as strain markers.

Inclusion size

In the roof zones of many plutons, inclusions of country rock may be of large dimension (usually less than 1Km^3), and may or may not be disturbed from their original structural position (Pitcher & Berger 1972). Microgranular inclusions are of a much more restricted size range seldom exceeding 1m in diameter (Grout 1937, Ferstater & Borodina 1977).

The restricted size spectrum of microgranitoid inclusions has been used to investigate the rate of ascent of granitoid magmas on the premise that the host will have to travel upwards at least as fast as the settling velocity of the inclusions if they are not to settle out (Oxburgh & MacCrae 1984). This approach requires that the host granitoid be of Newtonian rheology and assumes that the maximum average inclusion size is proportional to ascent velocity of the host. Sparks et al. (1977a) challenge this assumption claiming that granitoids are non-Newtonian fluids (see also Pitcher 1979) containing a yield strength, the restricted inclusion size range being a function of this yield strength and not ascent velocity of the magma.

Texture

Textural observations can be useful evidence of the origin of inclusions. Those inclusions which contain perfect granoblastic textures and regular banding may with confidence be assigned to a sedimentary or metasedimentary origin. Inclusions which are only partially recrystallised are most likely to have been derived from an igneous source (Didier 1973). Ferromagnesian minerals will recrystallise with greater ease than will feldspars, thus an igneous rock containing pyroxene and feldspars may well suffer uralitisation but retain an overall igneous texture (Harker 1939).

Grainsize

Inclusions of coarser grainsize than the enclosing granitoid are unusual (Didier 1973). The generally fine to medium grainsize of inclusions has led to the proposition that processes of recrystallisation are important in decreasing the grainsize (Joplin 1935). This would result from the break-up of porphyroblasts, mostly of hornblende, as the country rock fragments continue to recrystallise. Alternatively Wager & Bailey (1953) claim that the fine grainsize is the product of chilling of a basic magma intruded into the liquid granitoid a view supported by Eichelberger (1975, 1978 & 1980), Vernon (1983 &

1984) and Cantagrel et al. (1983).

Much recent work (eg Dowty 1980, Lofgren 1980 and Donaldson 1976) on the morphology of minerals crystallised from magmas, has emphasised the effect of growth kinetics upon crystal shape. That is, those crystals grown rapidly (i.e. a large degree of undercooling) produce spherulitic, dendritic, skeletal or acicular morphologies depending upon a complex interplay of internal structure, nucleation rates and degree of undercooling. Idiomorphic minerals are the product of low degrees of undercooling with low nucleation rates (Lofgren 1980).

Many workers (Harker 1939, White & Chappell 1977, Vernon 1984) have noted the acicular nature of apatite within microgranular inclusions. Vernon (1983) in reviewing the textures of inclusion suites within the Lachlan fold belt (SE Australia), attributed the acicular apatites and zircons to high level quenching (rapid growth) processes indicating that the inclusions represented fragments of disrupted chilled magma.

Macroscopic igneous textures are notable for the interlocking of the constituent minerals due to their sequential growth. Metamorphic minerals on the other hand, are seldom well formed or skeletal and characteristically form porphyroblasts with sieve textures. The observation of rapid growth textures, or an interlocking groundmass provides some evidence of a magmatic precursor to the inclusion

Presence of feldspar megacrysts

Oscillatory zoned plagioclase feldspars may be considered to have grown in the presence of a melt phase, the zonation patterns resulting from a complex interplay of growth kinetics and melt diffusion rates to the growth surface (Vance 1962, see also Hibbard 1981 for a magma mixing origin). The presence of such megacrysts within inclusions suggests therefore a magmatic precursor but the notion of such megacrysts growing in equilibrium with the host granitoid melt (whilst still within the inclusion) cannot be discounted.

The presence of K-feldspar megacrysts is much more equivocal with the observation of orthoclase growing across inclusion-host garnitoid boundaries indicative of a late metasomatic origin (Didier 1973, p.76).

Thus whilst the presence of oscillatory zoned plagioclase megacrysts within inclusions suggests a magmatic precursor, K-feldspar megacrysts may or may not yield corroborative evidence.

Indicator minerals.

Grout (1937) considered the following minerals as indicative of a sedimentary or metasedimentary origin:

Cordierite, Kyanite, Sillimanite,
Andalusite, Graphite and Staurolite.

However, Didier (1973) expressed doubts about the uniquely metasedimentary origin of such minerals, pointing out that minerals such as cordierite may well occur as primary phases within some granitoids.

5.2 INCLUSION PETROGRAPHY

Inclusions within the Strontian, Ballachulish and Criffell plutons may be placed within one of three major petrographic types viz:

Type 1. Inclusions which may be readily identified as sedimentary or metasedimentary in origin by the presence of relict structures and metamorphosed texture.

Type 2. Inclusions which, through the presence of a chilled margin or rapid growth textures, may be identified as the product of syn-plutonic injection of a basic magma into the host granitoid.

Type 3. Dark, mostly fine-grained inclusions (generally referred to as autoliths), whose origin cannot be readily ascertained in the field.

Type 1 and Type 2 inclusions are rare within the forcefully emplaced Strontian and Criffell plutons and Type 1 inclusions are common only within the outer portions of the passively emplaced Ballachulish pluton. The inclusion suites of the Strontian, Criffell and Ballachulish plutons are dominated by the dark fine-grained bodies (of exotic origin?) classified as Type 3 which may be further divided into four sub-types thus;

Type 3(a). Fine grained quartz dioritic inclusions containing no megacrysts (e.g. Plate 5.1).

Type 3(b). Fine grained quartz dioritic inclusions containing quartz or feldspar megacrysts (e.g. Plate 5.3).

Type 3(c). Fine grained quartz dioritic inclusions containing megacrysts of ferromagnesian minerals +/- feldspar

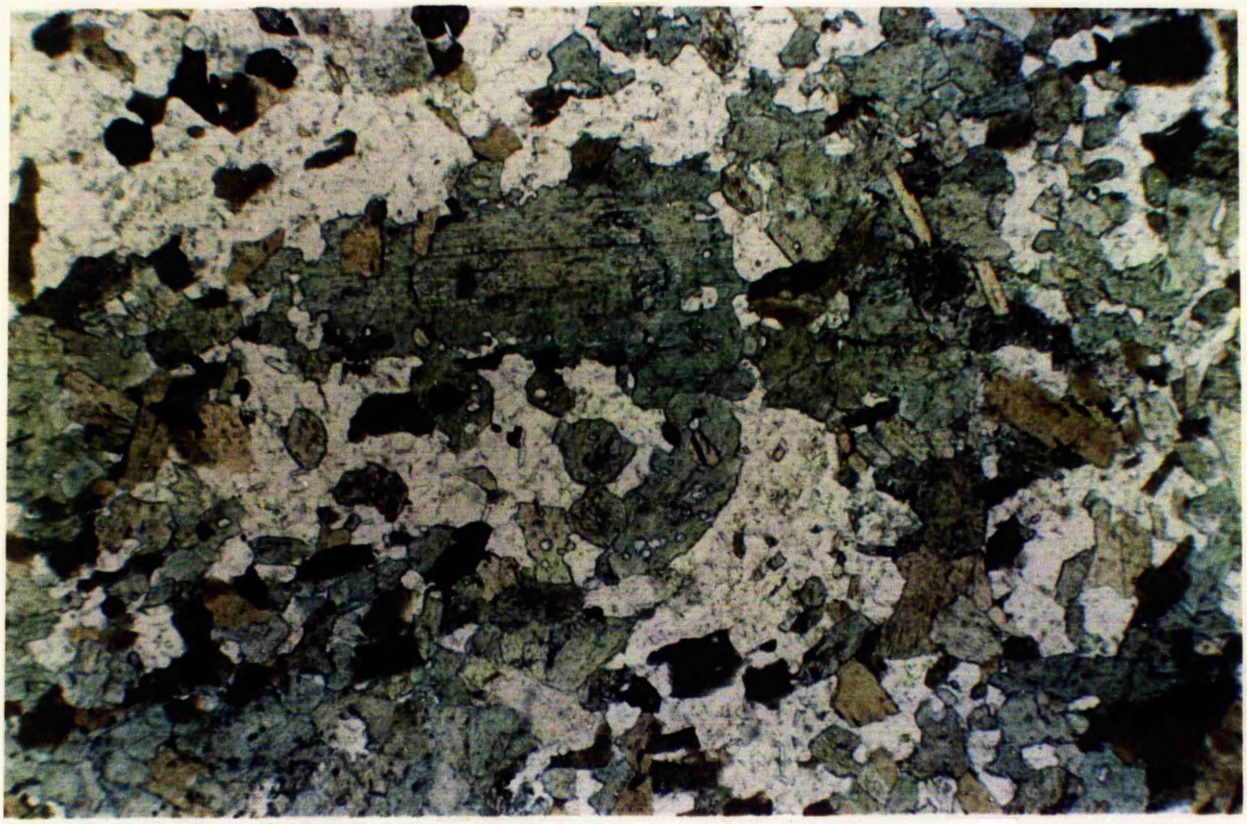


Plate 5.1 Texture of Strontian S(3a) inclusions, sample S210X (X48).

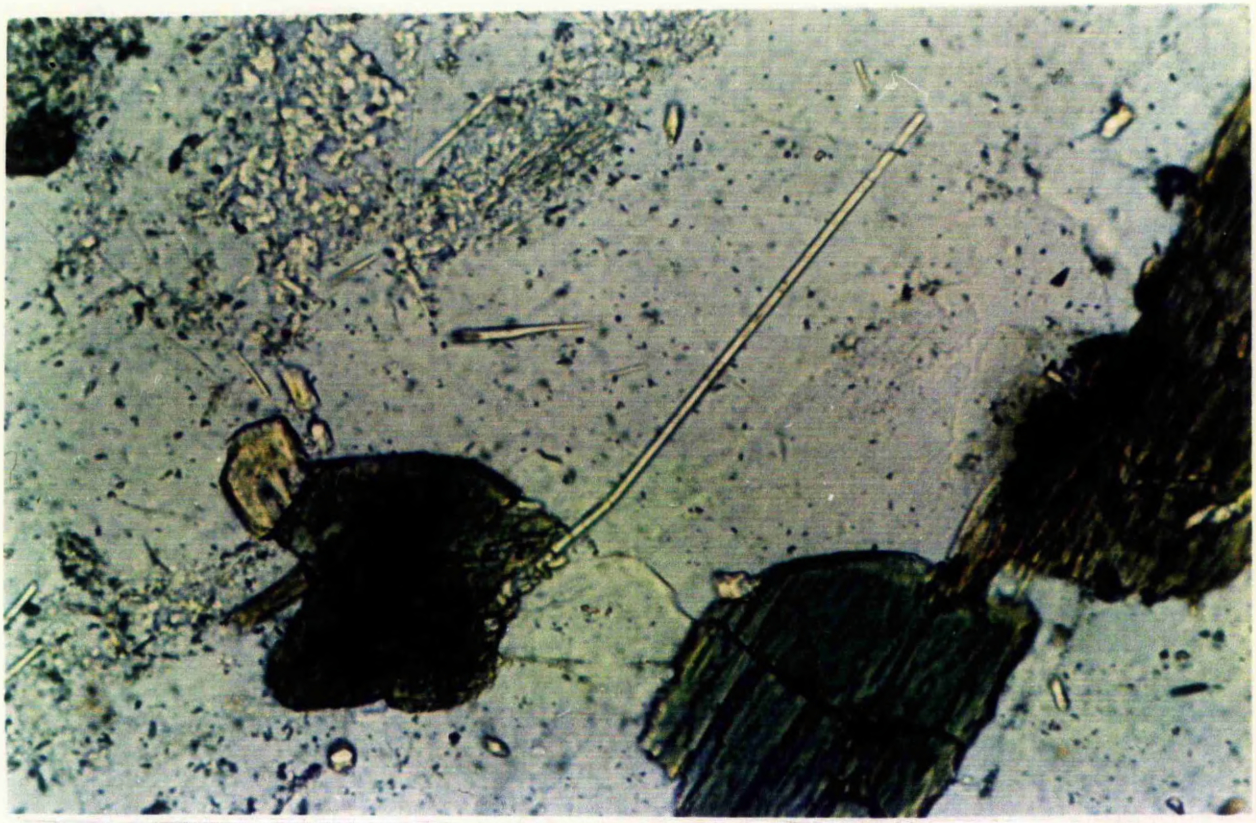


Plate 5.2 Habit of apatite within microgranitoid inclusions, sample S209X (X120).

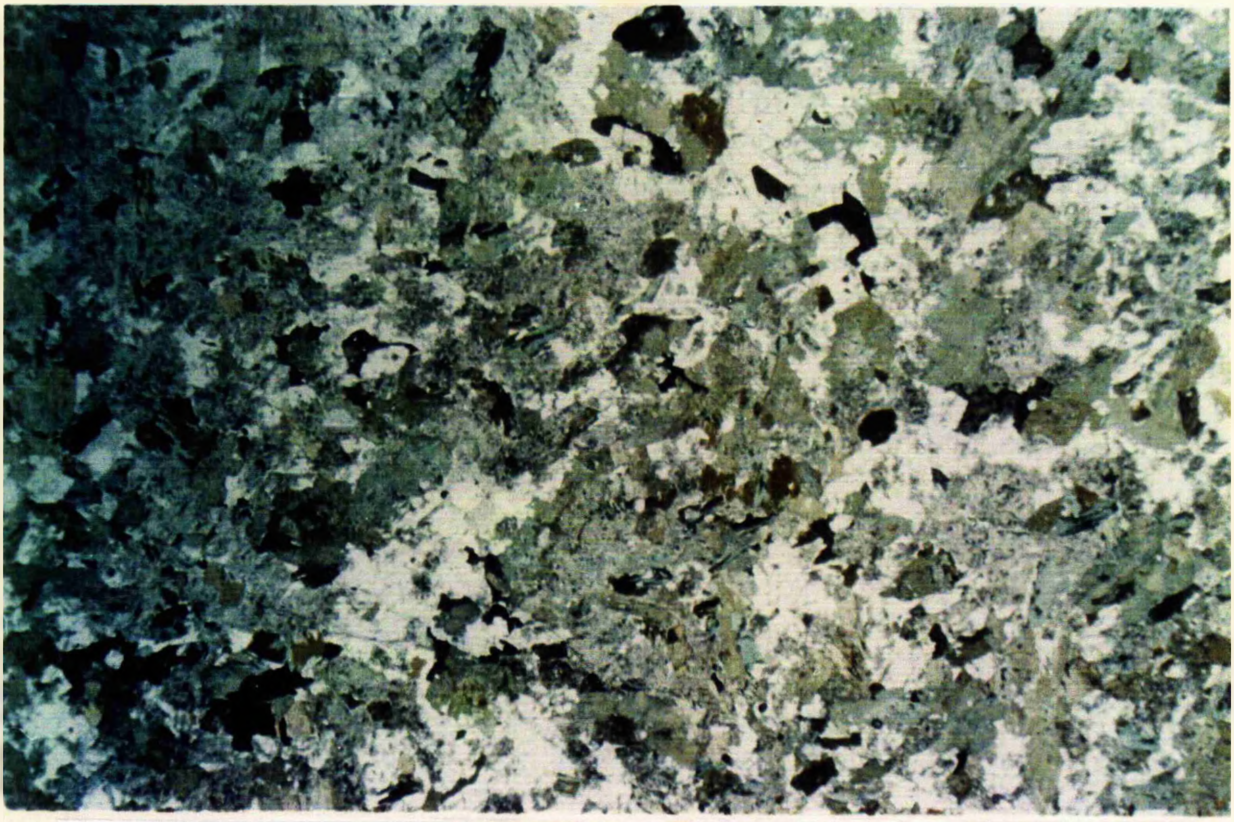


Plate 5.3 Texture of B(3a:1) inclusions, sample B204X (X48).

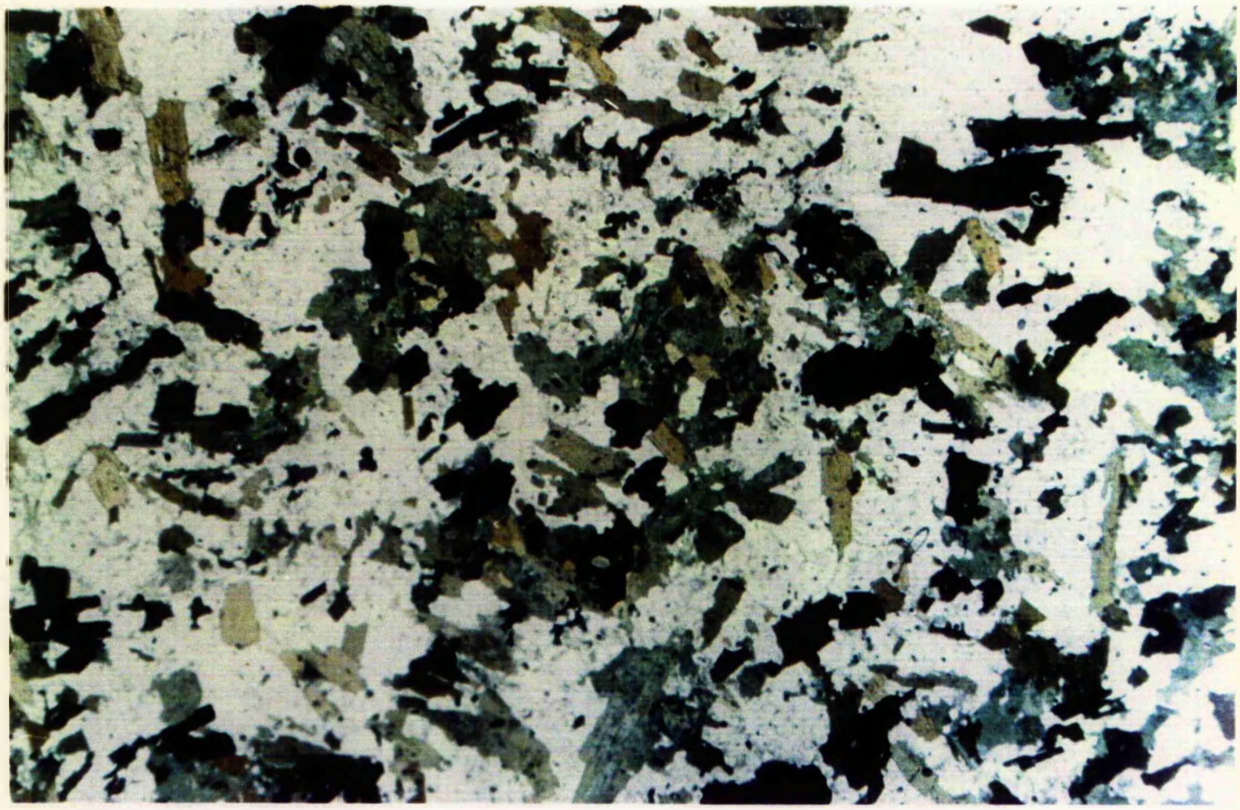


Plate 5.4 Texture of B(3a:2) inclusions, sample B202X (X48).

megacrysts (e.g. Plate 5.4).

Type 3(d). Medium to coarse grained inclusions which may or may not be quartz dioritic.

Each of the primary types and sub-types may be further divided on the basis of mineral assemblage or texture. Petrographic data for inclusion suites from the Strontian, Ballachulish and Criffell plutons are presented in tabulated form (Tables 5.1 to 5.4), for the purposes of comparison. Photomicrographs represent typical features of each of the major groupings.

(1) Inclusions which fall within Types 3(a) to 3(d) for all three plutons ubiquitously contain plagioclase crystals which are compositionally zoned. Typically the smaller plagioclase grains within the matrix show simple normal zoning whereas the megacrysts tend to have a complex core region with multiple embayments and oscillatory zoning superimposed on a normal zonation trend. There is often a sharp compositional break between the core region and the outer non-oscillatory zoned region.

(2) Types 3(a) to 3(d) inclusions frequently contain relict or pristine myrmekite intergrowths. These are more frequently observed within inclusions which have undergone low degrees of

CLASS 3(a) FINE-GRAINED NON-PORPHYRITIC INCLUSIONS.

Felsic minerals.

	Quartz	Plagioclase	K-feldspar.
S(3a)	Anhedral	Subhedral, zoned An40-20, alteration confined to core.	Poikiloblastic microperthite +/- lesser microcline.
B(3a:1)	as S(3a)	as S(3a) with lesser alteration.	"
B(3a:2)	"	"	"
C(3a:1)	"	as S(3a)	"
C(3a:2)	"	"	"

Ferromagnesian minerals.

	Clinopyroxene	Hornblende	Biotite.
S(3a)	Rare, cores to hornblende.	Subhedral hornblende often as aggregates.	Late, less common than hornblende.
B(3a:1)	Common augite cores to hornblende.	Similar to S(3a).	"
B(3a:2)	CPX rare.	Lesser hornblende than B(3a:1).	More common than hornblende.
C(3a:1)	As B(3a:1).	As S(3a).	As S(3a).
C(3a:2)	As S(3a)	As B(3a:2)	As B(3a:2).

Table 5.1 Petrographic comparisons between fine-grained non-porphyritic inclusions from Strontian (S), Ballachulish (B) and Criffell (C).

Accessory minerals.

	Apatite	Sphene	Zircon	Allanite
S(3a)	Acicular, within all other minerals.	Common, anhedral, & late.	Anhedral.	Rare, anhedral, within biotite.
B(3a:1)	As S(3a).	As S(3a), associated with CPX.	As S(3a).	As S(3a).
B(3a:2)	As S(3a).	As S(3a).	As S(3a).	As S(3a).
C(3a:1)	As S(3a).	As B(3a:1).	As S(3a).	Common within biotite.
C(3a:2)	As S(3a).	Anhedral, associated with hornblende.	As S(3a).	As C(3a:1).

Textural features.

	Grainsize	Foliation	Grain boundary terminations	Photomicrograph plate numbers
S(3a)	Fine to medium.	Strong to moderate.	Often 120°.	Plate 5.1
B(3a:1)	"	Weak.	Few 120°, interlocking.	Plate 5.3
B(3a:2)	"	Weak.	As B(3a:1).	Plate 5.4
C(3a:1)	"	V. strong to strong.	As S(3a).	
C(3a:2)	"	Moderate to weak.	As B(3a:1).	

Table 5.1 Cont.

CLASS 3(b) FELDSPAR MEGACRYST BEARING INCLUSIONS.

Felsic minerals.

	Quartz	Plagioclase	K-feldspar.
S(3b)	Small, anhedral.	Megacrysts complexly zoned (An40-20); matrix normal zonation; sericitic alteration patchy.	Small intersertal microperthite.
B(3b)	"	"	"
C(3b:1)	"	"	"
C(3b:2)	"	"	More common than C(3b:1).

Ferromagnesian minerals.

	Clinopyroxene	Hornblende	Biotite.
S(3b)	Very rare.	Common anhedral, often as aggregates.	Late, intersertal.
B(3b)	Common as cores to hornblende.	"	"
C(3b:1)	As B(3b).	"	"
C(3b:2)	None observed.	Uncommon, small, euhedral.	Common as late tablets.

Table 5.2 Petrographic characteristics of feldspar megacryst bearing inclusions.

Accessory minerals.

	Apatite	Sphene	Zircon	Allanite
S(3b)	Acicular, within all minerals except plagioclase megacrysts.	Late, anhedral.	Rare, anhedral.	Rare.
B(3b)	"	"	"	"
C(3b:1)	"	"	"	"
C(3b:2)	"	"	"	"

Textural features.

	Grainsize	Foliation	Grain boundary terminations
S(3b)	Fine to medium.	Moderate, passing around megacrysts.	Often 120° but also interlocking.
B(3b)	As S(3a).	Weak.	Interlocking.
C(3b:1)	As S(3a).	V. strong to strong, passing around megacrysts.	Mostly 120°, often with mortar texture.
C(3b:2)	As S(3a)	Weak.	Interlocking.

Table 5.2 Cont.

CLASS 3(c) FERROMAGNESIAN MEGACRYST BEARING INCLUSIONS.

Felsic minerals.

	Quartz	Plagioclase	K-feldspar.
S(3c)	Anhedral and rare.	Subhedral to anhedral, common matrix mineral, simple zonation from An40-An20.	Small anhedral.

Ferromagnesian minerals.

	Clinopyroxene	Hornblende	Biotite.
S(3c)	Rare, found at centre of hornblende aggregates.	Subhedral, aggregates form dominant petrographic feature.	Late, often wrapped around hornblende aggregates.

Accessory minerals.

	Apatite	Sphene	Zircon	Allanite
S(3c)	Acicular included within all other minerals.	Common, euhedral.	Anhedral.	Rare, anhedral, within biotite.

Textural features.

	Grainsize	Foliation	Grain boundary terminations	Photomicrograph plate numbers
S(3c)	Fine to medium.	Strong to. moderate.	Interlocking.	Plate 5.7.

Table 5.3 Petrographic characteristics of ferromagnesian megacryst bearing inclusions.

CLASS 3(d) COARSE-GRAINED FELSIC INCLUSIONS.

Felsic minerals.

	Quartz	Plagioclase	K-feldspar.
S(3d)	Anhedral and late.	Euhedral to subhedral, often as complexly zoned megacrysts.	Poikiloblastic, microperthite.

Ferromagnesian minerals.

	Clinopyroxene	Hornblende	Biotite.
S(3d)	None.	Euhedral minerals dominate with occasional core of biotite.	Late, less than hornblende.

Accessory minerals.

	Apatite	Sphene	Zircon	Allanite
S(3d)	Equant, euhedral but not common.	Common, euhedral.	Anhedral.	Rare, anhedral, within biotite.

Textural features.

	Grainsize	Foliation	Grain boundary terminations
S(3d)	Medium to coarse.	Weak.	Interlocking.

Table 5.4 Petrographic characteristics of the corase-grained felsic inclusions. Note, no similar inclusions were found within the Ballachulish and Criffell plutons.

matrix recrystallisation than those thoroughly recrystallised, indicating that the myrmekite was a pre-hornfelsing texture which was progressively removed by recrystallisation processes.

(3) Ferromagnesian minerals in Types 3(a) to 3(d) are dominated by green hornblende either as euhedral minerals or as replacement products of clinopyroxene. Magnetite +/- sphene appears to result from clinopyroxene breakdown. Biotite occurs late in all inclusions and may be replacing hornblende in some instances. At Strontian, biotite may be frequently observed at the centre of euhedral hornblende suggesting either a prior precipitation from a melt (Wones 1981) or the later infilling of earlier hollow hornblende.

(4) Apatite is common throughout all inclusion classes and host granitoid. Within Types 3(a) to 3(c), apatite occurs principally with an acicular habit, frequently being hollow with a length to breadth ratio often in excess of 20:1, a feature commonly observed within volcanic and hypabyssal rocks (Wyllie et al. 1962). In Types 1, 3(d) and within host granitoids themselves, apatite appears as equant grains which are seldom hollow and with a length to breadth ratio which rarely exceeds 3:1. In Types 3(b) and 3(c) where plagioclase or ferromagnesian megacrysts may be observed, acicular apatites are included within all the matrix mineral phases including plagioclase, amphibole and biotite, but are not observed as inclusions within the megacryst phase. This suggests that the megacrysts precipitated prior to apatite and

later matrix formation.

(5) The matrix minerals may be recrystallised to differing degrees. The degree of recrystallisation of the inclusions is strongly dependent on the intensity of foliation within the host granitoid, that is, where a strong fabric is present within the host then the inclusions are strongly recrystallised.

(6) Modal analysis shows that the inclusions in general are not mafic (as they are commonly referred to in the literature). Few indeed have a colour index greater than 50. The inclusion Types 3(a) to 3(d) are of intermediate mineralogy, slightly more basic than their host granitoids.

5.4 DISCUSSION

Significance of apatite habit

Acicular apatite is commonly observed within volcanic and hypabyssal magmatic rocks, but is uncommon in the plutonic equivalents (Wyllie et al. 1962). The acicular and particularly hollow texture within minerals has been attributed to rapid growth of the crystal along a preferred crystallographic axis (Lofgren 1980) and is an indication of a quench environment.

In the types of inclusion containing megacrysts it is noteworthy that acicular apatites are found included within all the matrix minerals but are not included within the inner portions of the megacrysts. Thus the megacrysts must have precipitated a significant proportion of the mineral prior to a quenching of the residual material which formed the matrix. Furthermore where acicular apatites are observed to be included within one or more of the matrix minerals, differential movement during grain boundary readjustment has bent or broken these apatites but not recrystallised them to more equant forms (see Plate 5.2).

Thus if acicular apatites are the products of high degrees of undercooling and rapid growth, then Type 3 inclusion precursors were magmatic and these magmas were at some stage chilled prior to a deformation/recrystallisation event which produced matrices of granoblastic texture.

Significance of oscillatory zoned plagioclase

Normal and oscillatory zoned plagioclase crystals (including megacrysts) occur throughout inclusion Types 3(a) to 3(d). Vance (1962) in an appraisal of oscillatory zoning within plagioclase feldspars stated:

"The abrupt transition from oscillatory core to normally zoned rim can, it is believed, be explained only in terms of magmatic processes."

Later work by Sibley et al. (1976) developed further the ideas of Vance (1962) which in turn were based on earlier pioneering work of Harlof (1927). Sibley et al. contend that the oscillatory zones were the product of a complex interplay between

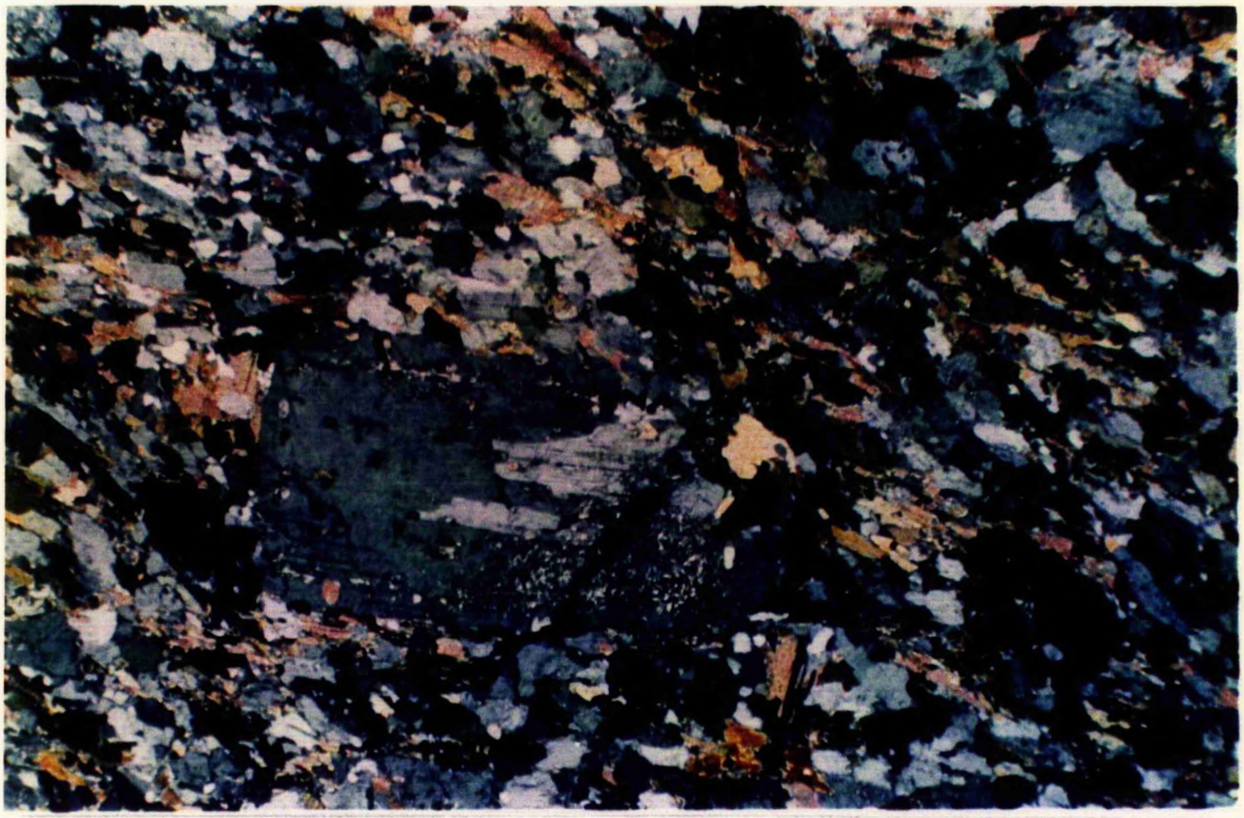


Plate 5.5 C(3b:1) inclusions, sample C204X (X48). Note the fabric passes around the plagioclase megacrysts.

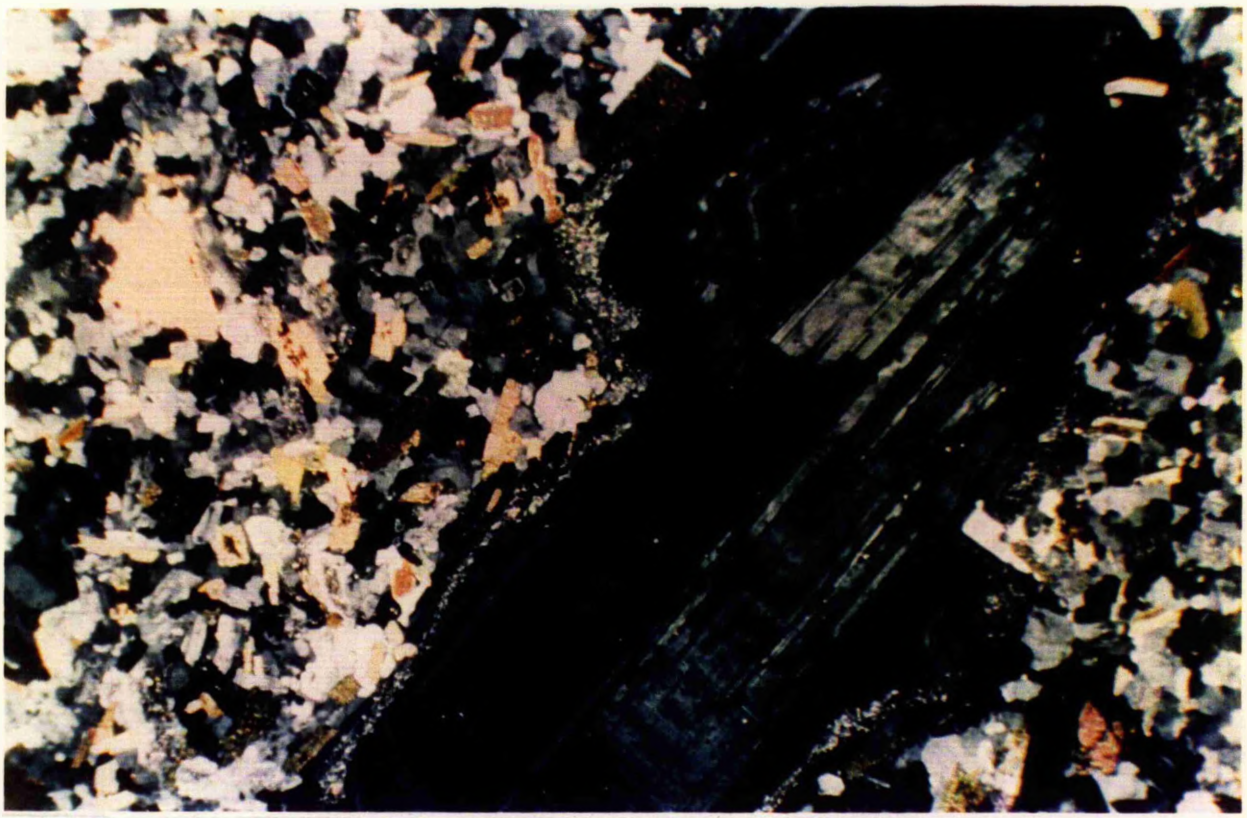


Plate 5.6 C(3b:2) inclusions, sample C328X (X48). Large complexly zoned plagioclase megacrysts in a weakly foliated matrix.

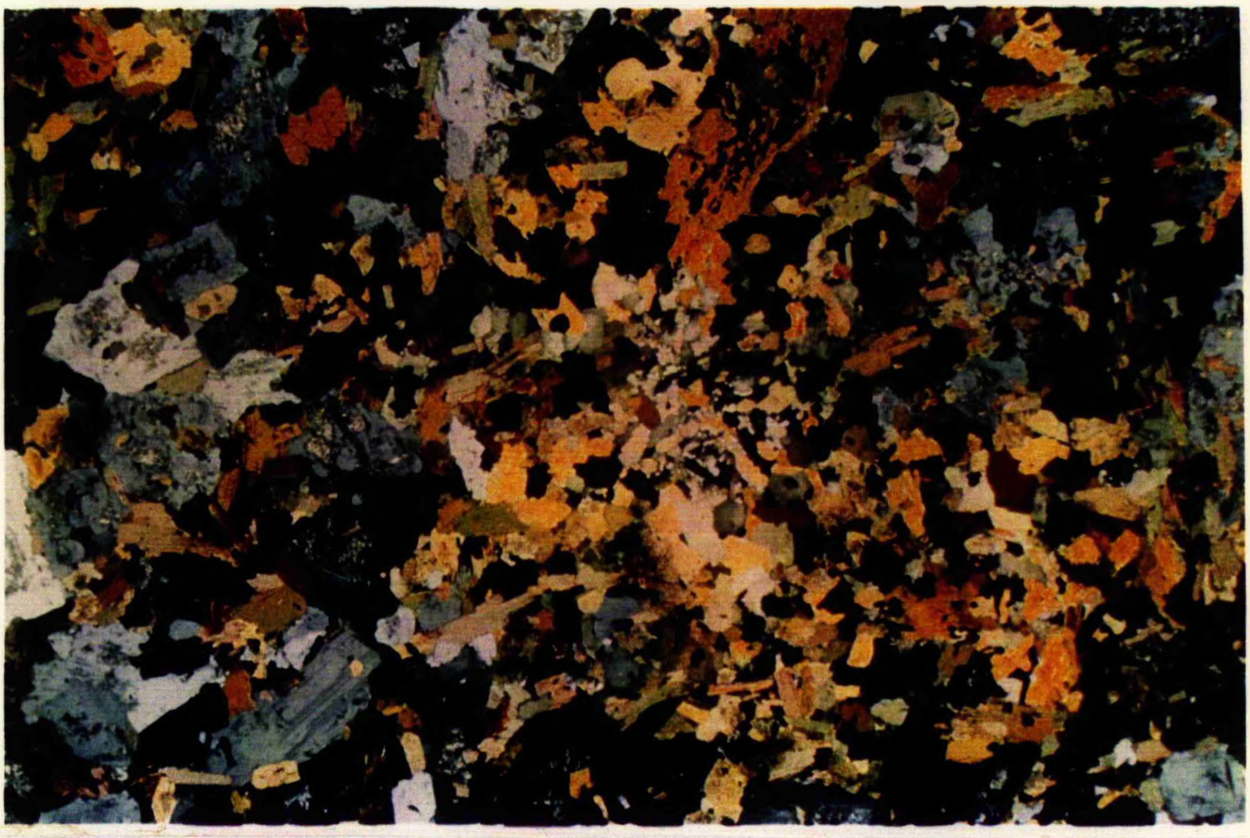


Plate 5.7 Texture of S(3c) inclusions, sample S355X (X48). Aggregated hornblendes for the dominant feature of the rock.

the rate of crystal growth and the rate to which the locally depleted magma surrounding the feldspar could be replenished by diffusion processes. Where the rate of growth exceeds that of diffusion, then continued growth will gradually reduce the surrounding magma in the anorthite component favouring more albitic precipitation. Diffusion processes replenish the anorthite component close to the feldspar growth surface permitting crystallisation of the more anorthitic end-member once again to proceed. Repetition of this cycle produces the oscillatory zones observed within the feldspar. The recognition of oscillatory zoned feldspars therefore, is good evidence of feldspar crystallisation from a melt.

White & Chappell (1977) in considering linear variations on Harker diagrams proposed that the granites of south-eastern Australia were products of partial melting of sediment-dominated or igneous-dominated crustal source regions, with inclusions within these granitoids representing refractory fragments of the source (restite). Furthermore plagioclase grains within the granitoids may contain a restitic component. That is, grains which exhibit a central core of relatively homogeneous and high anorthite composition, mantled by less calcium-rich and complexly zoned plagioclase, could have been derived in two stages, namely an unzoned (or weakly zoned) metamorphic core brought up from the source region (restite), followed by growth of an oscillatory zoned magmatic portion crystallised from the granitoid magma. By this argument, all the plagioclase grains within the inclusions should be of the poorly zoned type not having been through the magmatic stage, they remaining solid throughout the melting of the source region and the collection and emplacement of the host granitoid.

The common observation of oscillatory zoned plagioclase megacrysts (with no simply zoned core, e.g. Plate 5.6) within inclusions Types 3(b)-3(d) collected from all three plutons, suggests that these inclusions were not fragments of unmelted source region and as such cannot be described as restites.

Significance of myrmekite

At the interface between plagioclase and K-feldspar intergrowths of oligoclase and quartz (myrmekite) protrude into microperthite. These myrmekitic intergrowths were interpreted as an exsolution phenomenon similar to the development of perthitic textures where albite exsolves from potassium feldspar under subsolidus conditions (Shelley 1964, 1973). Hibbard (1979) reinterpreted these intergrowths as primary magmatic features, albeit late in the crystallisation sequence. When a magma crystallises minerals undersaturated in volatile components such as plagioclase, the dissolved volatile content of the magma will increase as crystallisation proceeds to a point where a vapour phase will exsolve from the magma due to saturation. The exsolution of a vapour phase will necessitate an increase in volume and/or pressure within the crystallising body. Hibbard (1979) believes that local zones of over-pressure within the crystalline mush will be relieved by "microtectonic" events causing local oversaturation of the crystallising species (oligoclase) which results in rapid growth. The vermicular quartz characteristic of myrmekite is the result of "forced" crystallisation due to growth rates of oligoclase being in excess of the rate at which silica could diffuse away from the growth site. Later infilling of the remaining pores by K-feldspar grows around these myrmekitic protuberances making it appear that the intergrowths were the product of subsolidus exsolution and not primary crystal growth. Hibbard's (1979) model for the development of myrmekite is entirely consistent with a magmatic origin for inclusions, with perhaps volatile infiltration from the host granitoid aiding primary myrmekitic growth which is subsequently destroyed during deformation and grain boundary readjustment.

Concluding observations

It was noted early in the chapter that individual textural criteria will not unequivocally permit the recognition of the source of any inclusion (except perhaps for Type 1 metasediments), but that consistent results for numerous observations on differing petrographic criteria might permit such conclusions (Grout 1937). For Type 3 inclusions of uncertain affinity, the acicular and hollow nature of the apatites, the oscillatory zoned plagioclase feldspar (both megacrysts and matrix minerals) and the presence of primary myrmekitic intergrowths, strongly suggest a magmatic precursor to the inclusions. Furthermore the presence of acicular apatites within matrix (oscillatory zoned) plagioclases, does not support the restite hypothesis as these minerals are demonstrably later than the magmatic apatites, which would not be expected to survive regional metamorphic conditions within the lower crust.

No evidence from any of the Type 3 inclusions observed within the Strontian, Criffell or Ballachulish plutons supports an origin by transformation of country rock. Grain boundary readjustment varies from the barely perceptible to the nearly granoblastic, with large variations occurring within a given type. That is, the grain boundary readjustment is solely a function of the strength of fabric within the host and as such, the psuedo-granoblastic texture of an inclusion is not a function of its starting material but a product of the intrusion mechanism. Inclusions not so deformed have interlocking textures characteristic of igneous rocks.

In conclusion, therefore, Type 3 inclusions are the product of magmatic crystallisation which at some stage in their history has been rapid. The relationships between the host magma and the inclusion precursors are discussed in Chapter 6.

CHAPTER 6

GEOCHEMISTRY OF INCLUSIONS AND HOSTS

Introduction

This study is concerned with the geochemistry of inclusion-host pairs from the Strontian (19 pairs), Ballachulish (17 pairs) and Criffell (26 pairs) plutons. The investigation included major and trace element analysis mostly of inclusion-host pairs but also some crustal rocks (appinite data will be discussed separately in Chapter 7). Rb-Sr isotopes were determined on a subset of 25 samples and Sm-Nd isotopes on a further subset of the Rb-Sr subset. All the data are presented in Appendix 2.

6.1 ELEMENT AND ISOTOPIC ABUNDANCES

Host granitoids: a) Major oxides.

The combined AFM plot for the Strontian, Ballachulish and Criffell plutons is given in Fig 6.1. All three adhere closely to a typical calc-alkaline trend with Ballachulish being a little Fe_2O_3 enriched. The K_2O vs. SiO_2 plot, (Fig 6.2) shows all three plutons to fall in the high-K calc-alkaline field of Peccerillo & Taylor (1976), typical of the Caledonian granitoids of Scotland (Stephens & Halliday 1984). A measure of the peraluminous index $Al_2O_3/Na_2O + K_2O + CaO$ (A/CNK) reveals that Strontian and Ballachulish have A/CNK values entirely <1.0 whilst values for the Criffell pluton are between -0.8 & +1.2 (Fig 6.3), a feature interpreted by Halliday et al. (1980) and Stephens & Halliday (1980) with the benefit of isotopes as indicating an increasing

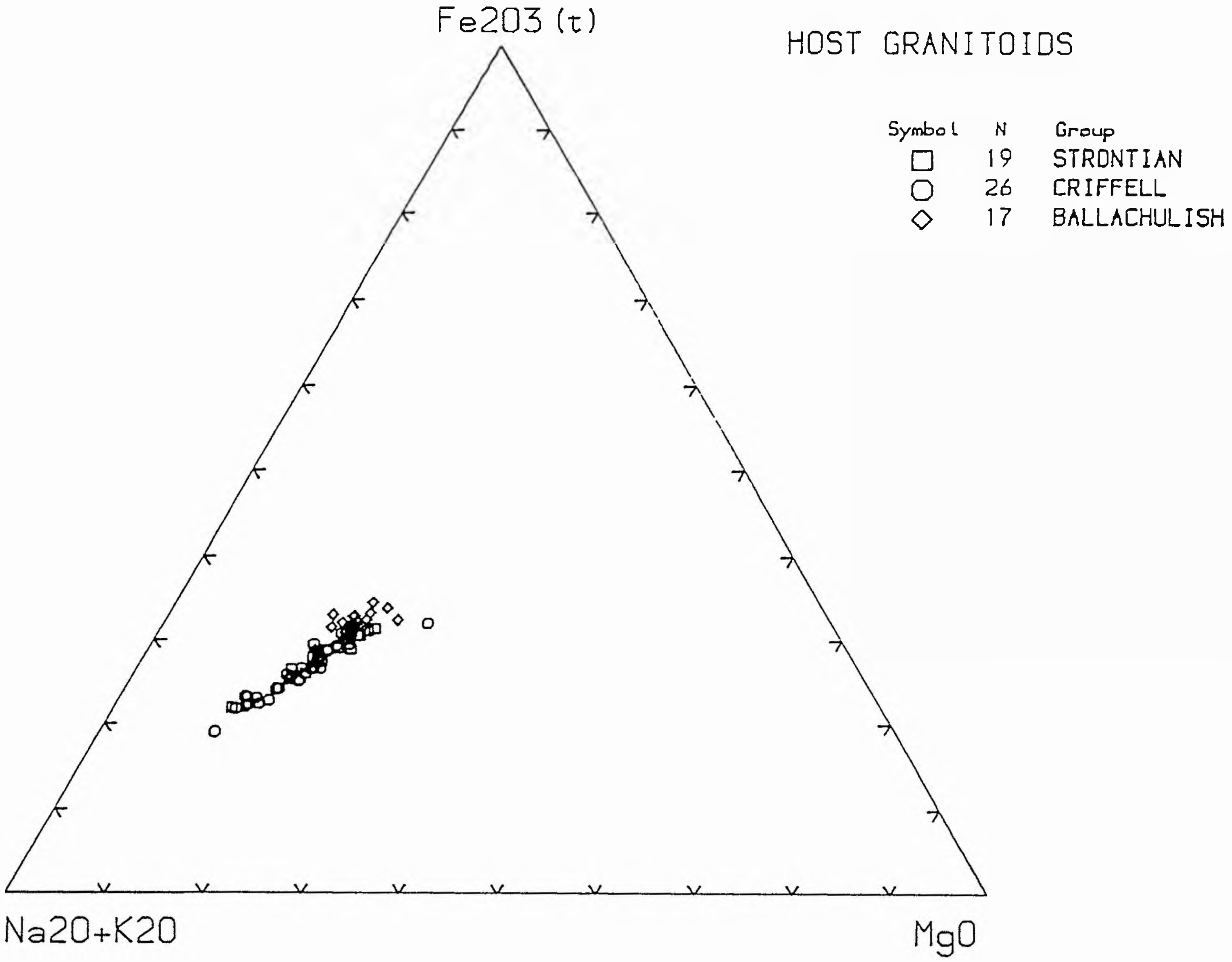


Fig 6.1 AFM plots for Strontian, Ballachulish and Criffell hosts.

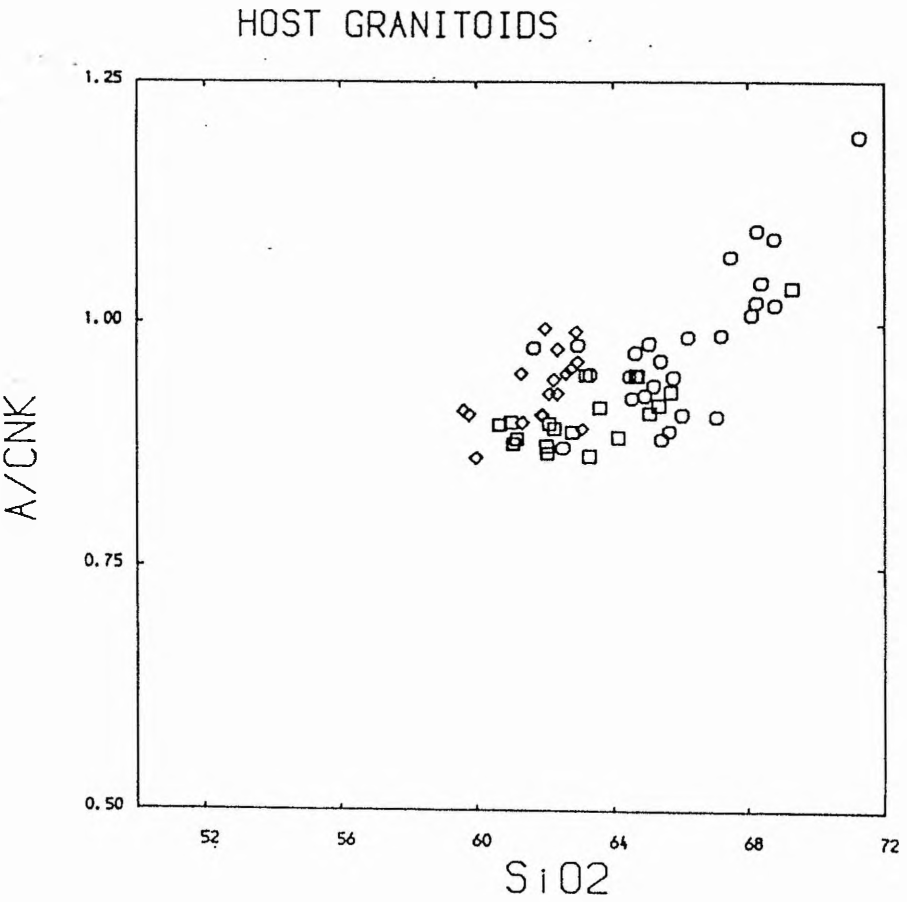


Fig 6.3 A/CNK (see text for definition) against SiO_2 for host plutons (Symbols as per Fig 6.1).

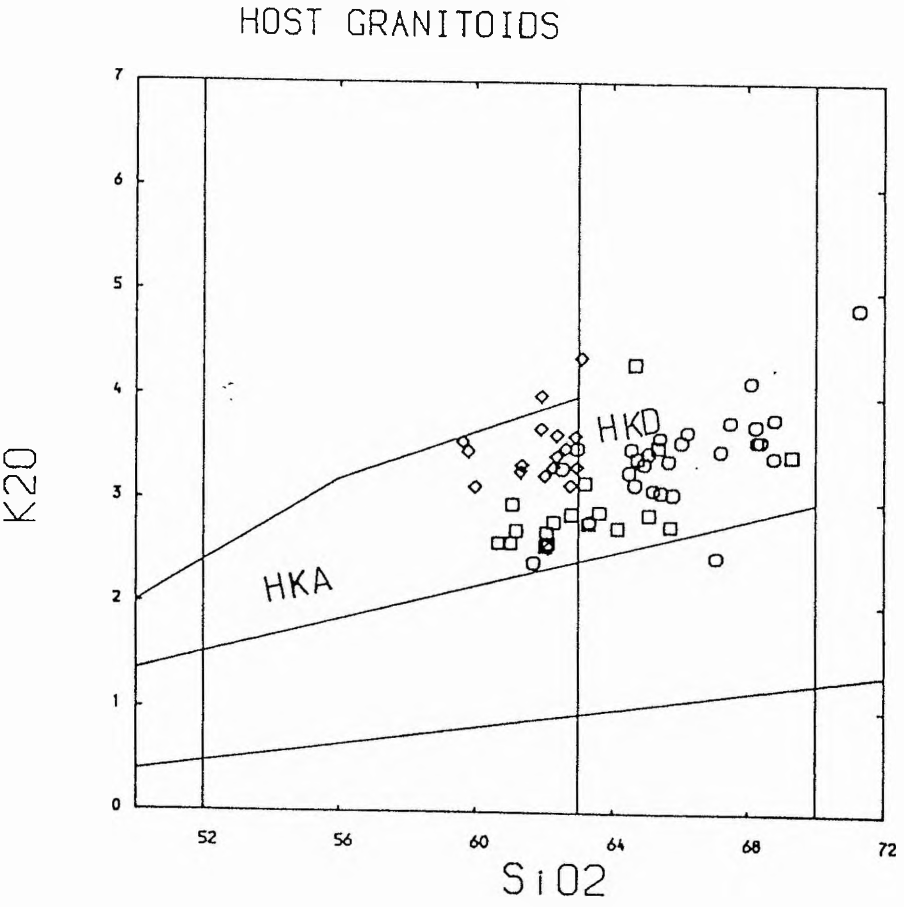


Fig 6.2 K_2O against SiO_2 for all pluton hosts (symbols as per Fig 6.1). HKA= High-K andesite, HKD= High-K dacite.

crustal input to a mixed mantle-crust melt in the more evolved parts of the pluton.

b) Trace elements

The most striking geochemical feature of the Scottish granitoids is their high Sr and Ba concentrations with respect to their contemporaneous Caledonian eqivalents in Australia and N. America (Stephens & Halliday 1984). Sr and Ba concentrations within the Strontian and Ballachulish plutons are typically in the range 600-1300 ppm whilst Criffell exhibits somewhat lower concentrations at the 300-1200 ppm level.

High field strength elements (HFSE) are represented by Zr and Nb. Like the other incompatible elements concentrations for the geographically close Strontian and Ballachulish plutons are similar (Zr=120-260 ppm, Nb=6-16 ppm) but are slightly lower within the Criffell pluton (Zr=100-200 ppm, Nb=3-15 ppm). Similarly the rare earth elements (REE) as typified by Ce and to some extent Y have comparable concentrations within the Strontian and Ballachulish plutons (Ce=50-150 ppm, Y=10-27 ppm) whilst Criffell has slightly higher LREE and slightly lower HREE (Ce=50-170 ppm, Y=8-15 ppm) respectively.

c) Isotope ratios.

Isotopic determinations (for Rb-Sr or Sm-Nd) on the Strontian and Criffell plutons have been given by Pankhurst (1979), Hamilton et al. (1980), Halliday et al. (1980) and Halliday (1984), whilst no such isotopic data are available for the Ballachulish pluton. These data are combined with with new data in this study.

Initial Sr isotope ratios from within the Strontian pluton are within the range 0.70523 to 0.70596 for the outer tonalite-granodiorite and 0.70649 to 0.70731 for the central biotite granite. At Criffell initial Sr isotopes range from 0.70522 to 0.70620 for the early pulse and the outer portion of the later pulse to 0.70684 for the central biotite granite. Ballachulish initial Sr isotope ratios are in the narrow range 0.70424 to 0.70439 for the quartz monzonite.

ϵNd_t values are in the range from +0.6 to -2.1 for the Strontian tonalites and granodiorites and from -3.2 to -6.3 for the central biotite granite. For Criffell ϵNd_t range from +0.4 to -2.0 for the outer portion and from -3.1 to -6.5 for the central biotite granite. At Ballachulish two quartz monzonite samples yield identical ϵNd_t values of -2.7 (taking an assumed age of 400Myr).

Inclusions

a) Major oxides.

Compared with their host granitoids, inclusions from all three plutons show much greater scatter on an AFM plot (Fig 6.4) suggestive that they are true xenoliths whereas autoliths would be expected to back project along the calc-alkaline trend through the hosts in a 'tight' array. Again in contrast with the host rocks the Harker plot for K_2O (Fig 6.5) shows the inclusions to have great scatter within each pluton with some inclusions of shoshonitic composition (Peccerillo & Taylor 1976) for all three

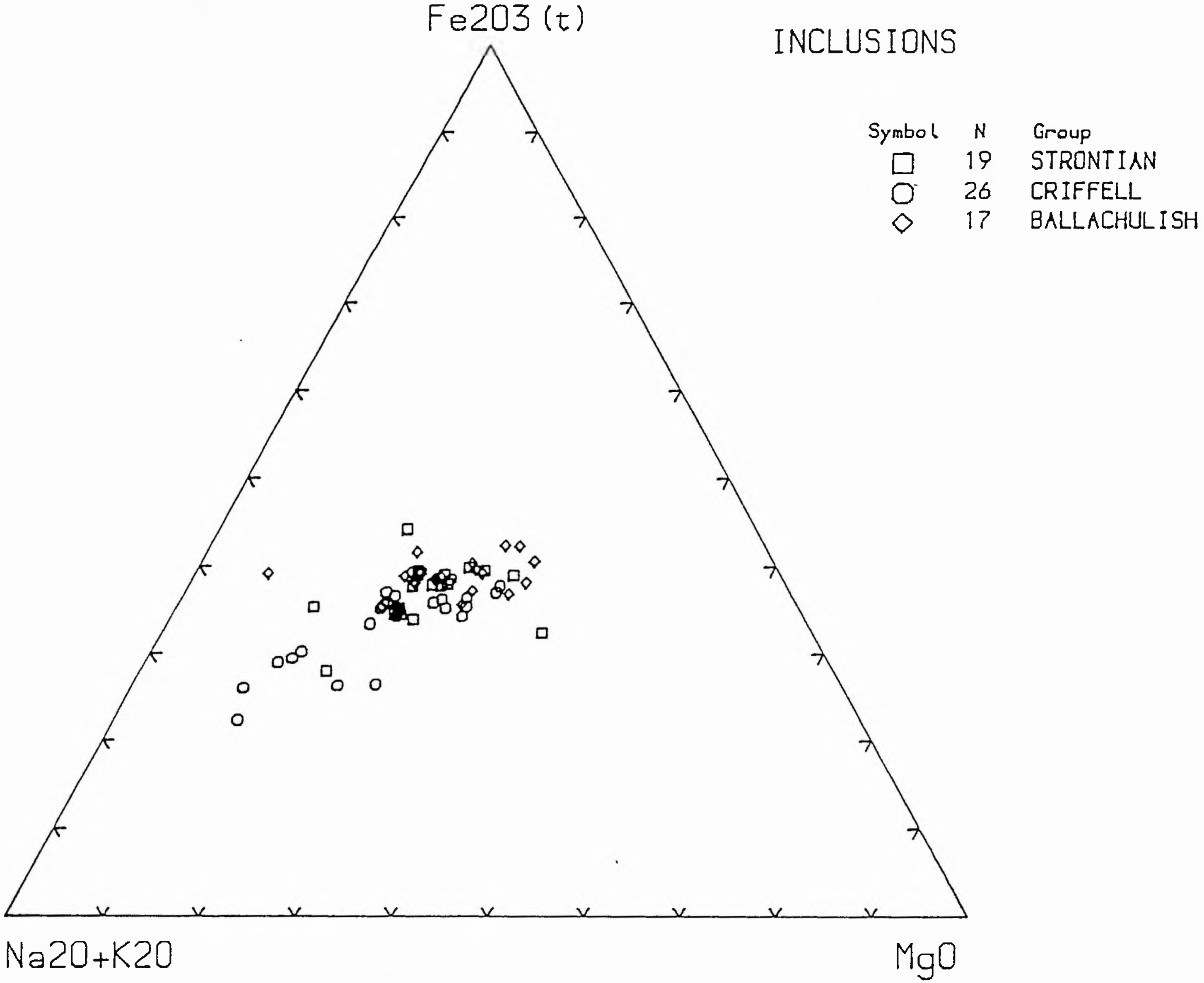


Fig 6.4 AFM plots for all inclusions.

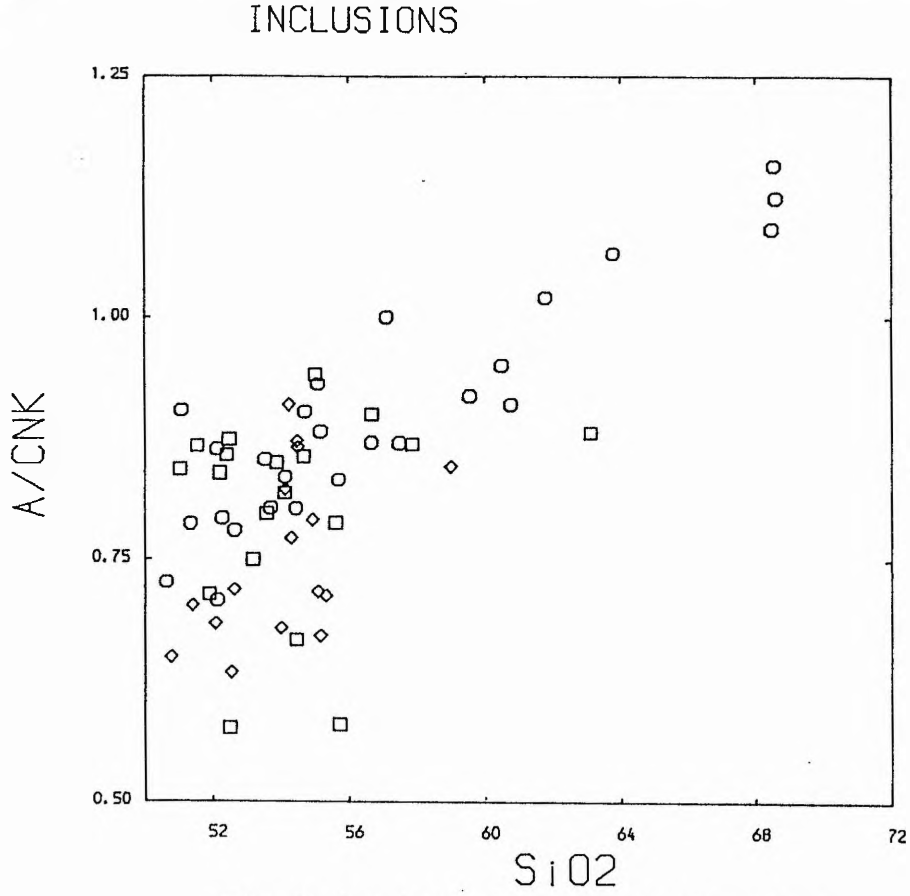


Fig 6.6 A/CNK for all inclusions.

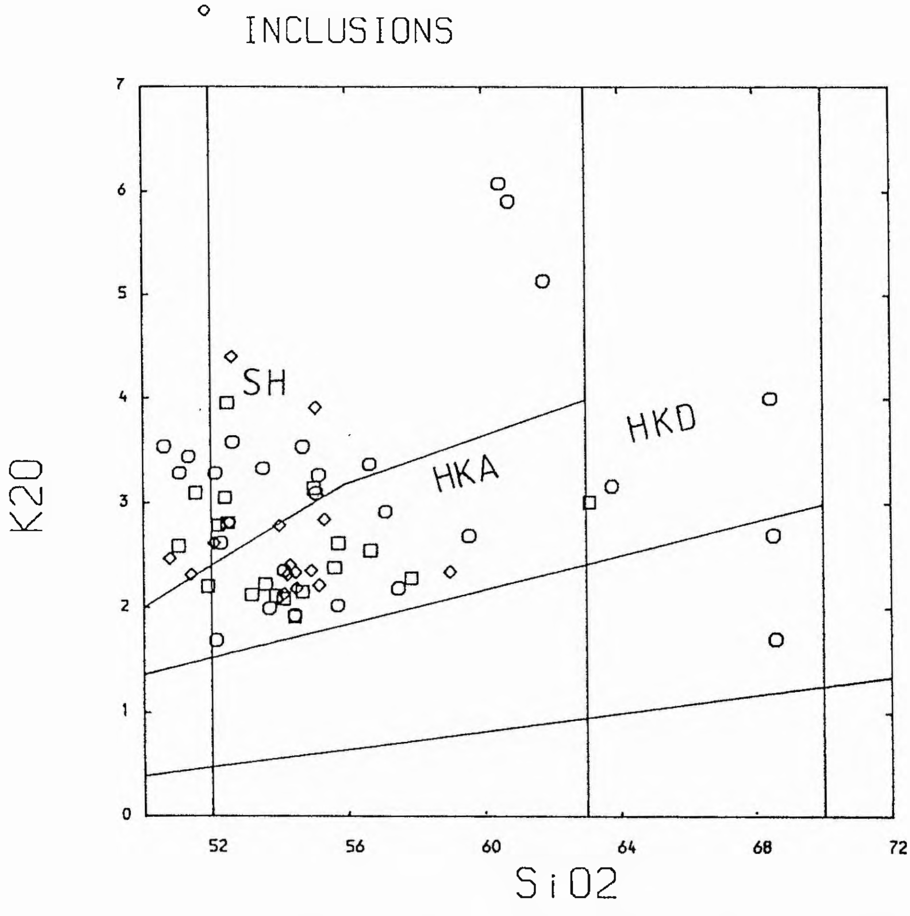


Fig 6.5 K_2O against SiO_2 for all inclusions (symbols as per Fig 6.4). HKA=High-K andesite, HKD= High-K dacite, SH=Shoshonite.

plutons. Similarly the plots of peraluminous index (A/CNK) against SiO_2 (Fig 6.6) show great scatter for inclusions, with a tendency for inclusions to cluster into intra-pluton groups. However the fact that the overall majority of inclusions plot to low A/CNK values (typically < 0.9) suggests an original igneous parentage (Chappell & White 1974, White & Chappell 1977).

b) Trace elements.

The most striking feature of the trace element geochemistry of the inclusions is the great range of concentrations exhibited for each element compared with host granitoid. Elements such as Sr and Ba tend to show a similar range in concentration as their host rocks (typically 600-1300 ppm for Strontian and Ballachulish and 300-1200 ppm for Criffell) whereas other elements contrast strongly with the host granitoid.

Inclusion HFSE and REE abundances show much greater scatter than their host granitoids, for example Zr and Nb concentrations for Strontian hosts are 150-250 ppm and 6-15 ppm respectively whilst for inclusions the values are 150-430 ppm for Zr and 7-26 ppm for Nb. Ce and Y concentrations at Criffell are 50-170 ppm and 8-15 ppm for hosts and 100-550 ppm and 5-38 ppm for inclusions. Furthermore it may be seen that inclusion concentrations for incompatible elements, HFSE and REE all tend to much higher concentrations for inclusions over hosts. The high and variable nature of inclusion trace element concentrations at a roughly constant SiO_2 level suggests that their abundances are not solely controlled by processes of fractional crystallisation.

It would appear then that inclusion trace element contents are controlled by a combination of parental magma composition, possibly fractionated prior to incorporation into the granitoid magma, and subsequent host-inclusion equilibration.

c) Isotope ratios

With respect to their host granitoids, inclusions exhibit a similar range of initial $^{87}Sr/^{86}Sr$ with host-inclusions pairs often having identical (within experimental error) isotopic compositions. However, in all three plutons the range of Sr isotopic compositions is greater for inclusions (0.70444 to 0.70605 for Strontian inclusions against 0.70523 to 0.70596 for Strontian hosts, 0.70518 to 0.70638 for Criffell inclusions against 0.70522 0.70620 for Criffell hosts and 0.70424 to 0.72699 for Ballachulish inclusions against 0.70424 to 0.70439 for Ballachulish hosts).

Sm-Nd isotopes were determined on a further subset of the Sr isotopes samples from the Strontian and Criffell plutons (microgranitoid inclusions) and from the Ballachulish pluton (metasedimentary inclusions). ϵNd_t values for microgranitoid inclusions, whilst overlapping the range exhibited by their host granitoids, tend to have <u>more</u> radiogenic Nd isotopes (ϵNd_t of -0.8 to +3.0 for Strontian inclusions and +0.6 to -2.1 for Strontian hosts, -0.4 to +1.9 for Criffell inclusions and -2.0 to +0.4 for Criffell hosts).

This observation has profound implications for the origin of microgranitoid inclusions as shifts in initial Nd isotopic signature cannot be brought about by processes of fractional crystallisation alone. Furthermore Nd isotopic compositions determined upon 3 metasedimentary inclusions from the Ballachulish pluton yield less radiogenic Nd isotopes than the host (ϵNd_t=-7.6 to -13.1 for metasedimentary fragments and -2.7 for Ballachulish hosts) indicating that microgranitoid inclusions cannot be the product of the partial digestion of such country rock fragments.

Summary

The Strontian, Ballachulish and Criffell plutons belong to the high-K calc-alkaline suite of orogenic granitoids typical of the Caledonian of Scotland. Inclusions contained within these plutons exhibit greater scatter in both elemental concentrations and isotopic composition. This large compositional scatter and the observation of intra-pluton inclusion clustering (e.g. Fig 6.6, A/CNK vs. SiO_2) suggests that inclusions may be subdivided into geochemical subgroups, and that inclusions cannot be described in terms of a single source, a finding consistent with petrographic observation (Chapter 5).

Sr and Ba concentrations and Sr isotopic compositions show similar ranges in both host granitoid and microgranitoid inclusions for all three plutons. This contrasts the HFSE and REE which show a greater range of concentration for inclusions over hosts and suggests that processes of equilibration have influenced both the Sr elemental and isotopic budgets to a greater degree than the HFSE, REE and Nd isotopes. Host-inclusion equilibration processes are now investigated.

6.2 MECHANISMS OF ELEMENTAL EXCHANGE

In natural systems where a chemical potential exists, processes of equilibration will occur to neutralise that potential (Bowen 1922). Whatever the ultimate source of inclusions, be they accidental, restitic or exotic, the likelihood is that they may be regarded (to a first approximation) as solid, sub-spherical bodies surrounded by granitoid magma. Being generally of small dimension it may be assumed that once incorporated these inclusions would rapidly attain thermal equilibrium with their host and as such, exchange during the chemical equilibration of the inclusions would occur largely in the solid state (Bowen 1922).

The rate at which the inclusions can be converted to a stable mineralogical assemblage depends upon the rate at which the necessary elemental constituents can be supplied to the site of metamorphism (Vernon 1976). The following factors may be important to the rate at which mineralogical equilibrium is attained between the host granitoid and its inclusions:

1)Temperature: in most natural processes the rate of reaction can be expressed in terms of an exponential dependence with temperature and as such temperature has a first order effect on reaction rate.

2)Grainsize: the smaller the grainsize the faster a reaction will proceed due to the increased surface area.

3)Diffusion: the migration of elements down an energy potential is primarily controlled by temperature, the magnitude of the chemical potential and the medium through which diffusion occurs.

4)Metasomatism: infiltration of elements through the pore fluid of a solid medium, effectively increases the rate of supply of elements required for metamorphic change. Infiltration by a metasomatising fluid is primarily controlled by the temperature of the host minerals, the porosity of the aggregate and the rate of fluid passage (which in turn is a function of the pressure drop across the aggregate).

5)Chromatographic effects: the passage of metasomatic fluids through a mineral aggregate containing exchangable ionic species will induce a concentration profile within that fluid (Hofmann 1972; see later). Hence over a long path length, metasomatic fluids may become depleted in exchangable ions slowing the effective equilibration rate in that region, and in effect exerting a rate controlling influence on the metasomatism.

Diffusion within silicates.

Diffusion is a random process whereby an atom (or molecule) may change its position with respect to its matrix. In solids, diffusion occurs when an atom attains sufficient vibrational energy to jump to the next available lattice site or suitable vacancy. In liquids and gases such motion is less constrained by the rigidity of the matrix and as a consequence the rates of diffusion in solids are several orders of magnitude slower than in their less ordered counterparts.

Bowen (1928) was the first to recognise the potential importance of diffusion as a petrological process. He envisaged minerals growing in a liquid with the constituent elements diffusing through the liquid to the site of crystallisation. More recent work has shown that if the rate of crystallisation exceeds that of diffusion then zoned minerals may result (Hibbard 1981). The rate at which new minerals may grow at the expense of old ones can also be diffusion controlled (Vernon 1976, p.78), often limiting the extent to which old assemblages may equilibrate with new conditions (Atherton 1976, Cliff et al. 1985).

Particularly useful modern reviews of diffusion in silicates have been given by Hofmann (1980), Henderson (1983) and Maaloe (1985), but for a more mathematical treatment of the subject, the reader is referred to Crank (1975).

Types of diffusion.

(a) Volume diffusion.

Volume diffusion, that is the transfer of material through the body of a solid object, may occur in response to one of three different driving forces; resultant elemental migration may also be defined by one of three different mechanisms:

1) Chemical diffusion: is the case in which atoms diffuse down a concentration profile, for example the concentration of Ca within a plagioclase feldspar is often variable with the tendency for the core region to be richer in Ca than the margin. Diffusion down the chemical gradient will spontaneously occur once the gradient is established

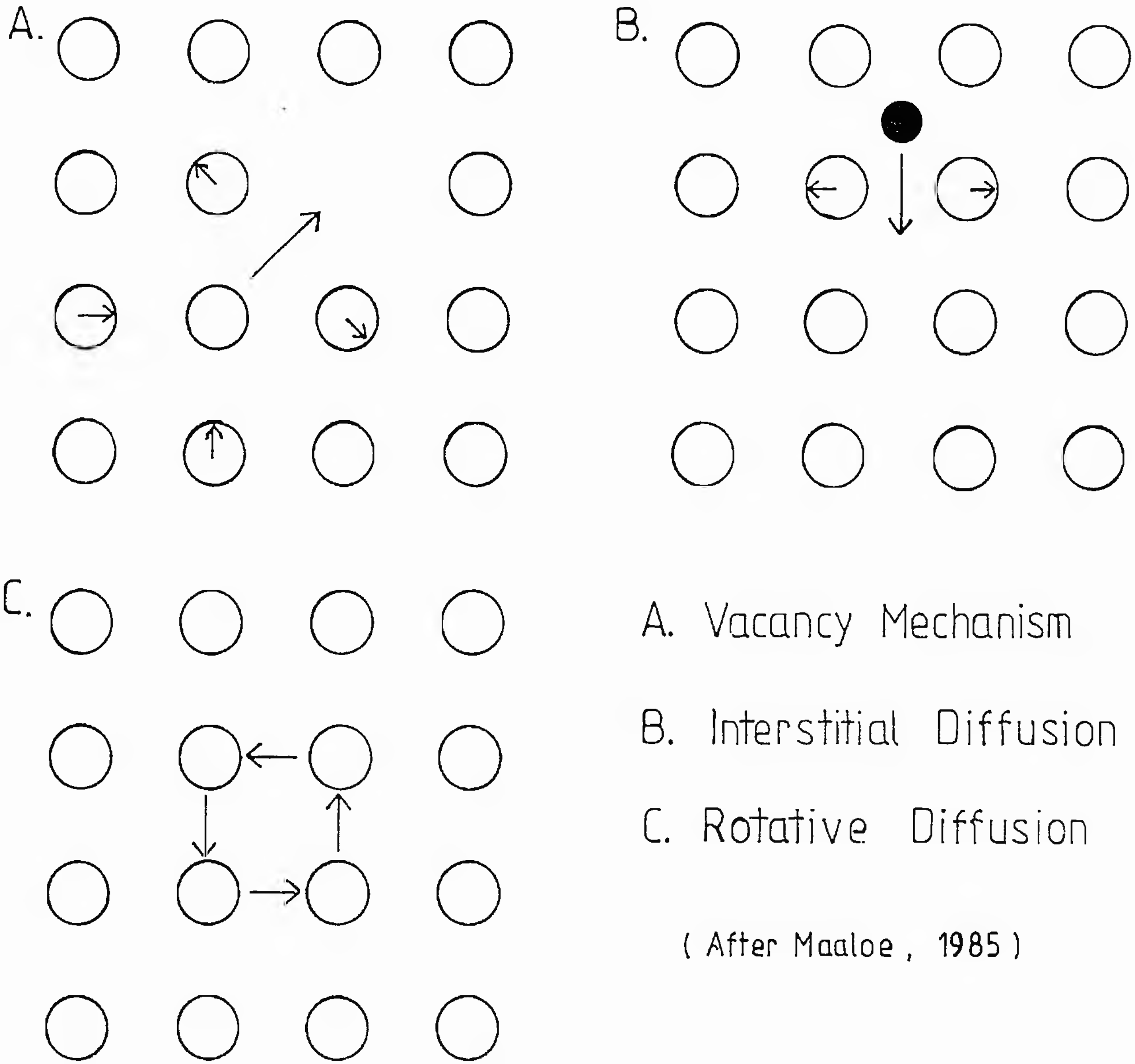
until such time as the gradient is nullified. The observation, however, of zoned plagioclase grains within igneous rocks indicates that such mass transport mechanisms are slow, not going to completion before cooling inhibits diffusion to vanishingly small rates. If, however, the element in question is of low concentration in the matrix, the chemical diffusion coefficient will approximate the self-diffusion coefficient.

2) Self-diffusion: may be described in terms of one of two sub-types;

a) Tracer diffusion: in which the only concentration gradient is an isotopic one. For example, plagioclase feldspars within an ultrabasic body may yield O isotopic compositions typical of crystallisation from a mantle-derived melt. Infiltration of meteoric fluids which have an O isotopic signature similar to that of the continental crust, may exchange oxygen isotopes between the fluid and the feldspar. This could profoundly alter the O isotopic composition of the feldspar without affecting its optical characteristics. In effect no mass transfer has occurred, only an attempt to neutralise the O isotopic gradient between the plagioclase and the pore fluids.

b) Intrinsic diffusion where there are no concentration gradients (isotopic or chemical) and diffusion describes the rate of change (or exchange) of

Fig6·7 MECHANISMS OF DIFFUSIVE TRANSPORT THROUGH SOLIDS



A. Vacancy Mechanism

B. Interstitial Diffusion

C. Rotative Diffusion

(After Maaloe , 1985)

positions for similar species within an homogeneous matrix.

3) <u>Interdiffusion</u> where the diffusion of one species depends directly upon the diffusion (in the opposite direction) of another, eg. Fe diffusion into an olivine is dependent on Mg diffusing out of the olivine.

As has already been indicated, the mode of diffusion strongly affects the diffusion coefficient. Diffusion through a solid crystal lattice may occur in one of three ways (these types being summarised in Fig 6.7)

a) <u>Vacancy mechanism.</u> No crystal is perfect, having holes or vacancies within the lattice structure. The number of 'free' sites varies but is generally in the order of 0.001% (Kittel 1956, Maaloe 1985). The diffusing species may move into the next available vacancy if it has sufficient kinetic energy, leaving behind a new vacancy with the net effect that the vacancy migrates through the crystal.

b) <u>Interstitial diffusion.</u> Atoms occupying interstitial positions between network forming atoms may jump to an unoccupied adjacent interstitial site without major structural disturbance. This movement requires an activation energy to overcome the forces of repulsion generated by the network-forming atoms. Clearly for minerals such as biotite mica, diffusion of K through loosely held interstitial sites will be much more rapid than for Fe, Mg or Si which make up the structure of each layer.

c) <u>Rotative diffusion:</u> this occurs only in open structured minerals of fairly simple type (eg Halite). Simultaneous jumping of four atoms may be envisaged if the kinetic energy (for activation of the move) is high enough.

(b) Grain boundary diffusion

Typically the outer layers of a mineral are less well ordered than in the interior. Vacancies and structural defects are likely to be substantially more common, allowing the diffusion rate to increase. Lateral diffusion may also be aided by an imperfect interface with the adjacent mineral where local disorder may be high. It follows therefore that for an aggregation of minerals (such as an inclusion), the rate of diffusion through the body will increase as the grainsize decreases due to the corresponding increase in surface area (and the surface area to volume ratio).

Disorder at the mineral-air/liquid interface is likely to be even more common than in the grain boundary case where one mineral grain abuts against another without the presence of a volatile medium. As a consequence activation energies required to reach the next available site are reduced and diffusion rate will increase. Such a case in which elements migrate from a mineral grain into a thin veneer of liquid (or gas) is often refered to as surface diffusion and a special case of the grain boundary mechanism.

Element charge-size effects

The rate at which an element can diffuse through a solid medium is also a function of its ionic characteristics and the nature of the medium through which migration is occurring. For instance an element (e.g. silicon) bound tightly in a framework silicate is not expected to diffuse rapidly due to the high energy required for the Si to break its existing bonds and jump to the next available lattice site. Thus the following parameters are generally important in determining diffusivities for individual elements.

1) Charge. Elements with high charge (eg Nb+5) diffuse slowly through solid silicate matrices due to the deforming effect the ion has on the framework. Hence movement of such a highly charged ions requires the deformation of the whole network structure, an event which happens only rarely. Ions with low charge do not deform the matrix as much and may jump to the next available lattice site at a lower energy threshold than the highly charged species.

2) Size. For elements of similar charge, the smaller the ion the less deformation a structure will have to undergo to allow passage. Therefore small ions will diffuse faster than large ones of the same charge.

3) Structure. For a given element the nature of the medium through which diffusion occurs is also important. Vacancy sites may have a specific maximum size which varies from mineral to mineral and even within minerals themselves. For instance diffusion of K through mica may occur rapidly along the plane of

the silicate sheets but may be difficult in a direction normal to the sheets.

In summary, the chemical exchange of elements between an inclusion and its surrounding granitoid magma is dependent on the temperature of reaction, the diffusivity of the species of interest, the duration of such temperatures, the presence of a pore fluid, the grainsize of the inclusion matrix and the structure of the constituent minerals. For the purposes of modelling such behaviour, idealised calculations are now performed using the assumptions that the inclusion was enclosed within a magma at 1000° celsius for 100,000 years after which time cooling was rapid (see Bowen 1922), and that inclusions are spherical and typically 20cm in diameter.

6.3 SIMPLIFIED MATHEMATICAL TREATMENT

Diffusion.

Considering linear diffusion, i.e. diffusion in one direction only, Fick defines his first law as:

$$J = -D\frac{\partial C}{\partial X} \qquad -(1)$$

(Crank 1975) where D is the diffusion coefficient (cm^2 /sec), C is the concentration ($moles/cm^3$) and X the co-ordinate (cm). J is the flux of the diffusing species past a plane at distance X from the origin (in $moles/cm^2/sec$). D's are in the order of 10^{-7} cm^2/sec for silicate liquids and 10^{-12} cm^2/sec for silicate minerals compared with 10^{-1} cm^2/sec for gases (Maaloe 1985).

Fick's first law is applicable to a steady-state system with an homogeneous matrix. Introducing a time function leads to the derivation of Fick's second law:

$$\frac{\partial C}{\partial T} = D\frac{\partial^2 C}{\partial X^2} \qquad -(2)$$

for the general case, where D may be dependent on concentration C.

Useful petrological inferences may accrue from calculating linear diffusion profiles with time, i.e. what proportion of a given species will have diffused a given distance in a given time. Hofmann (1980) gives the simple relationship:

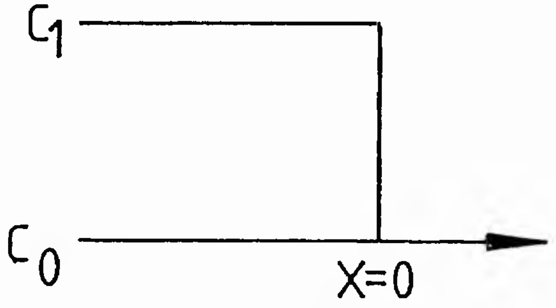
$$X = \sqrt{(Dt)} \qquad -(3)$$

where X is the distance travelled (cm), D the diffusion coefficient (cm^2/sec) and t is the time elapsed (sec). However, since the inclusions have been enclosed in a magma at relatively high temperature for a finite time, it is useful to consider what percentage of material exchange was permissible prior to the closure temperature being reached (see Dodson 1973 for discussion of closure temperatures).

Fig 6.8 sets the boundary conditions for linear diffusion between an inclusion and host. Crank (1975) has shown that the solution to such a problem is given by:

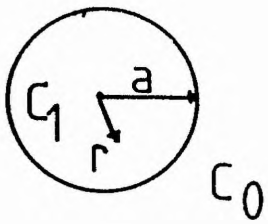
$$C_{(x,t)} = \frac{C_1 - C_0}{2}\left[1 - erf\left(\frac{X}{2\sqrt{(Dt)}}\right)\right] \qquad -(4)$$

Fig 6·8



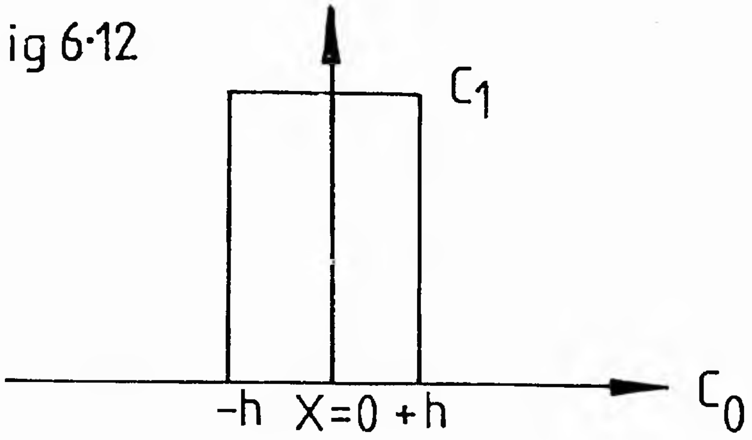
Boundary conditions for equation (4)

Fig 6·10



Boundary conditions for equation (7)

Fig 6·12



Boundary conditions for equation (9)

Fig 6·9 Equilibration In One Dimension (distances are in cms)

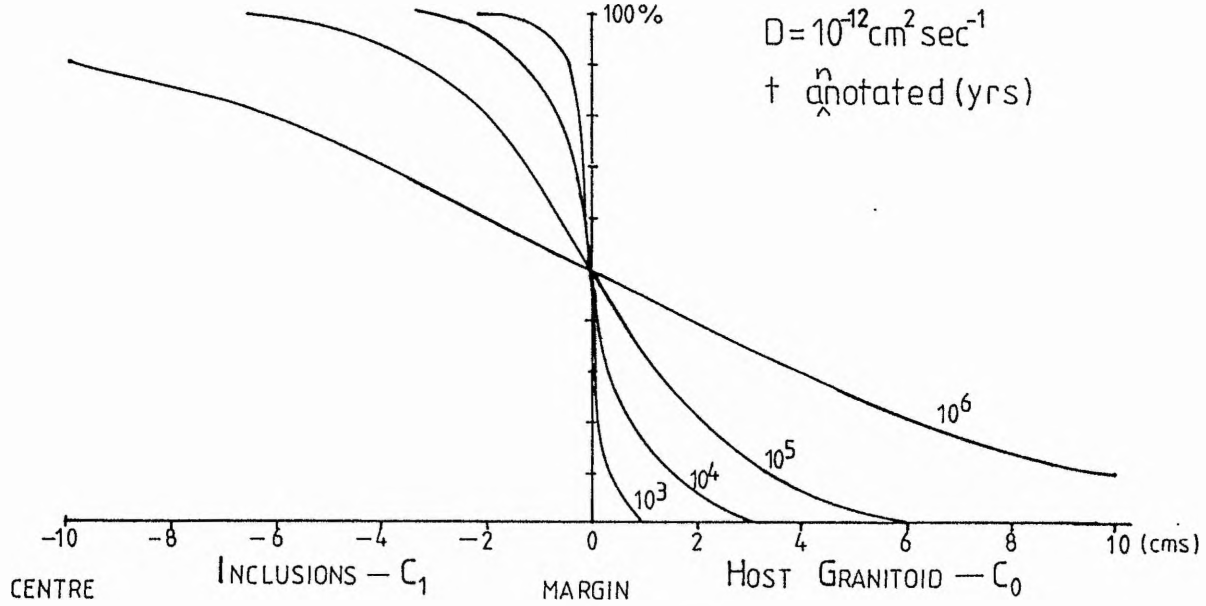
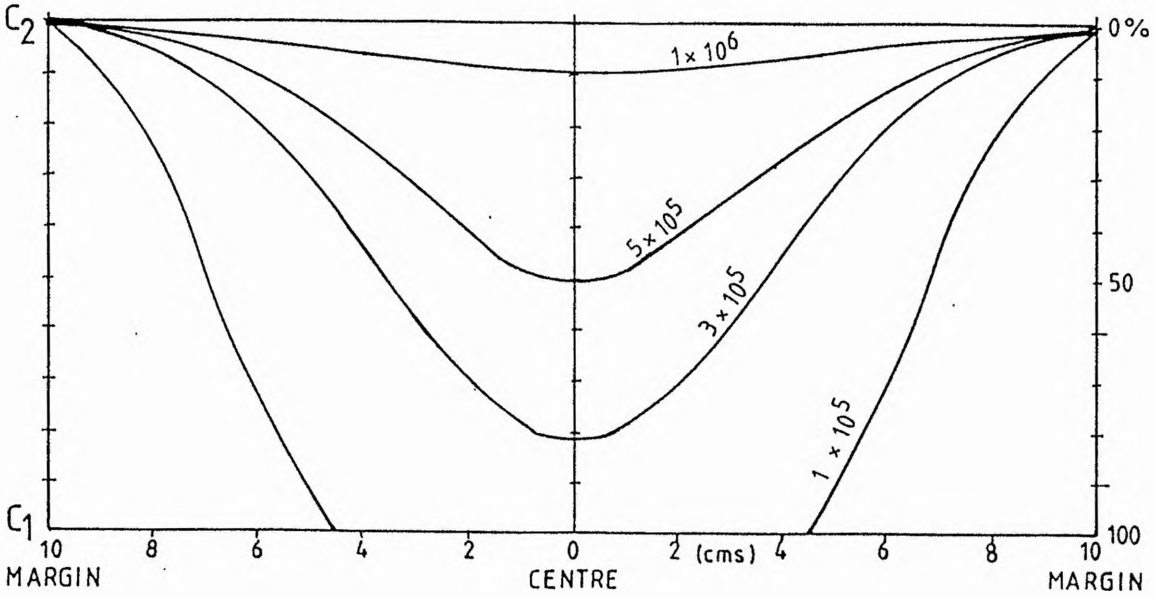


Fig 6·11 Equilibration of a Spherical Inclusion - where
$D = 10^{-12} cm^2 sec^{-1}$ t annotated (yrs)

Significant equilibration occurs in 3×10^5 yrs



where C1 is the concentration within the inclusion, Co is the concentration within the host (the rest as for 3), and erf is the error function which is the solution to the integral:

$$erf=\frac{2}{\sqrt{\pi}}\int_{0}^{x} e^{t^2}\,\partial t \qquad -(5)$$

There is no analytical solution to this integral but the value may be derived by numerical methods such as those of Pierce & Foster (1963) and falls between the values +/- 1 (see Maaloe 1985, p.243-252)

For an inclusion of 20cm diameter and using the boundary conditions in Fig 6.8 inserted into equation (4) we obtain Fig 6.9 . This indicates that equilibration would occur in the order of 10^5 years (perfect equilibration would take an infinite time span), for a species whose D value was $1*10^{-12}$ cm^2/sec. Whilst this calculation is useful, it is more applicable to sills and dykes with angular aspect ratios whose diffusion profile may be realistically estimated in one dimension. Clearly this is not an accurate representation of the case in question where we postulated the inclusion to be a spherical solid in a 'liquid' matrix. A more realistic model would be to evaluate radial diffusion in a sphere.

Fig 6.10 sets out the boundary conditions for a solid inclusion enclosed in a liquid granitoid melt, where diffusion within the liquid is rapid compared to that within the solid. Taking the 20cm inclusion used previously and the same diffusing species (ie $D=1*10^{-12}$), Crank (1975, p.91) has developed the equation:

$$\frac{C-C_1}{C_0-C_1} = \frac{a}{r}\sum_{n=0}^{n=\infty}\left[\operatorname{erf}\frac{(2n+1)a-r}{2\sqrt{(Dt)}} - \operatorname{erf}\frac{(2n+1)a+r}{2\sqrt{(Dt)}}\right] \qquad -(6)$$

for short diffusion periods (or rapidly diffusing atoms),

and

$$\frac{C-C_1}{C_0-C_1} = 1+\frac{2a}{\pi r}\sum_{n=1}^{\infty}\frac{(-1)^n}{n}\,\mathrm{Sin}\,\frac{n\pi r}{a}\,\exp\left(-Dn^2\pi^2\frac{t}{a^2}\right) \qquad -(7)$$

for long time periods (or slowly diffusing species).

In practice the summation on the right hand side of equation (7) is only affected in the 10th decimal place when n>10. Fig 6.11 is a graphical representation of the equation (7) with the time for diffusion annotated on the curves. Substantial equilibration will occur in the order of 2 to $3*10^5$ yrs for species with geologically plausible D values, in an inclusion of 20cm diameter remembering that the volume subtended is proportional to the radius cubed, ie the greater volume of the inclusion lies in the outer 5cm rather than the inner 5cm.

This is a simplified model and would probably represent the upper limit of equilibration time as the species is assumed to diffuse at a constant rate of 10^{-12} cm^2/sec. In practice, since the inclusion is not an homogeneous mono-mineral, the time for substantial equilibration may be lower as grain boundary diffusion is somewhat faster than that of volume diffusion. However, the time for substantial equilibration is the same order as in the previous model ($1-3*10^5$ yrs).

Metasomatism

Metasomatism may occur as either diffusion metasomatism, infiltration metasomatism or both, the difference being defined by whether the pore fluid moves in relation to the solid matrix. Diffusion metasomatism of an ionic species through a static pore fluid may be treated in a similar fashion to that of crystalline diffusion. However, since the diffusing species within the pore fluid has to travel around the matrix crystals a factor to account for this increased path length (the tortuosity) must be included (Korzhinskii 1970, Hofmann 1972, Fletcher & Hofmann 1974).

$$D_e = \tau \beta D_i \qquad -(8)$$

Where τ=tortuosity, β =porosity and Di is the intrinsic diffusion coefficient for the species under study in the pore fluid. Taking values of tortuosity to be 0.5 (i.e. twice the path length), porosity as 10^{-3} (Fletcher & Hofmann 1974) and Di as 10^{-8} cm^2/sec, the effective diffusion coefficient will be $5*10^{-12}$ cm^2/sec.

Literature values for diffusion coefficients in silicate fluids at around 1000°C show relatively small variation (10^{-7} to 10^{-9} cm^2/sec) for all the elements studied (Hofmann 1980), compared with the large range in volume diffusivities in the solid state from 10^{-9} to 10^{-17} cm^2/sec (Hofmann 1980, Henderson 1983, Sneeringer et al. 1984 and Maaloe 1985). As a consequence elements whose rate of diffusion is slow in the solid state may have an effective diffusion coefficient a few to several orders of magnitude greater depending upon the porosity, tortuosity and grainsize of the mineral aggregate.

Infiltration metasomatism of a fluid species occurs in response to pressure differences across a porous medium. In the case of crystalline aggregates a pressure difference promotes the passage of pore fluids which may be of a different composition to that from which the minerals originally crystallised. This may cause exchange between the minerals and the pore fluid as it passes through the rock. Such a mechanism may be important for mineral reactions in regional metamorphic terrains (Fletcher & Hofmann 1974). For inclusions surrounded by semi-liquid magma however, there is no driving force by which the fluids may be forced through the crystalline aggregate, convective currents within the host granitoid tending to travel around the inclusion rather than through it.

A modified type of infiltration metasomatism may become important to exchange mechanisms at the periphery of the granitoid body. Here, inclusions become disc shaped in response to flattening stresses generated by the granitoid as the magma chamber inflates. The lack of rheological contrast between the inclusions and the granitoid provide strong evidence for the presence of lubricating pore fluid within the inclusion during the deformation event (Hutton 1982). As flattening proceeded, fluids from within the inclusion would be expelled being replaced by fluids from the host granitoids and promoting infiltration metasomatism.

6.4 APPLICATION TO INCLUSIONS

It has already been shown that elements with diffusivities greater than 10^{-12} cm^2/sec would equilibrate in the order of a few hundred thousand years (at constant temperature) depending on the size of the inclusion. Elements whose diffusivities in the solid state are 10^{-12} or less would be expected to equilibrate between a solid inclusion (of typically 20cm diameter) and its liquid host after 10^5 years at a reaction temperature of 1000°C even if volume diffusion through the solid minerals were the only diffusion mechanism. Elements whose volume diffusivities are slow ($<10^{-12}$ cm^2/sec), would not be expected to equilibrate for the same inclusion over the same period by volume diffusion alone. Differing volume diffusivities would thus lead to the observation that some elements were well correlated with their host (rapid diffusers) and others which would have no systematic relationship to their enclosing magma (slow diffusers).

For the slow diffusing species a second mechanism may be invoked to improve the potential for equilibration with the host granitoid, that is diffusion metasomatism. It has been shown that for elements with diffusivities of around 10^{-8} cm^2/sec in the (free) fluid state, the effective diffusivity within an inclusion of pore fluid will be in the order of 10^{-12} cm^2/sec depending upon the porosity and tortuosity of the inclusion. For elements whose diffusivities are 10^{-8} cm^2/sec in the free fluid and 10^{-12} cm^2/sec in the solid, neither mechanism will be preferred during equilibration of inclusions, both being of equal diffusive potential (10^{-12} cm^2/sec).

For elements with diffusivities of 10^{-8} in the free liquid state but less than 10^{-12} cm^2/sec in the solid state, diffusion metasomatism will become the preferred transport mechanism. Equilibrium will thus be approached in two stages:

a) diffusion in the solid state to the mineral grain boundary

b) diffusion out of the inclusion via the pore fluid.

Clearly for elements travelling from the host into the inclusion the two stages will be reversed.

Elements whose volume diffusion coefficient is slow precluding equilibration with the surrounding host in geologically reasonable time scales, may have a significantly faster effective diffusion coefficient and attain equilibrium in two stages.

Consider a mineral whose aspect is similar to that in Fig 6.12. For linear diffusion in one dimension Crank (1975) has developed the equation:

$$C_{(x,t)} = \frac{C_1}{2}\left[\operatorname{erf}\left(\frac{h+x}{2\sqrt{(Dt)}}\right) + \operatorname{erf}\left(\frac{h-x}{2\sqrt{(Dt)}}\right)\right] \qquad -(9)$$

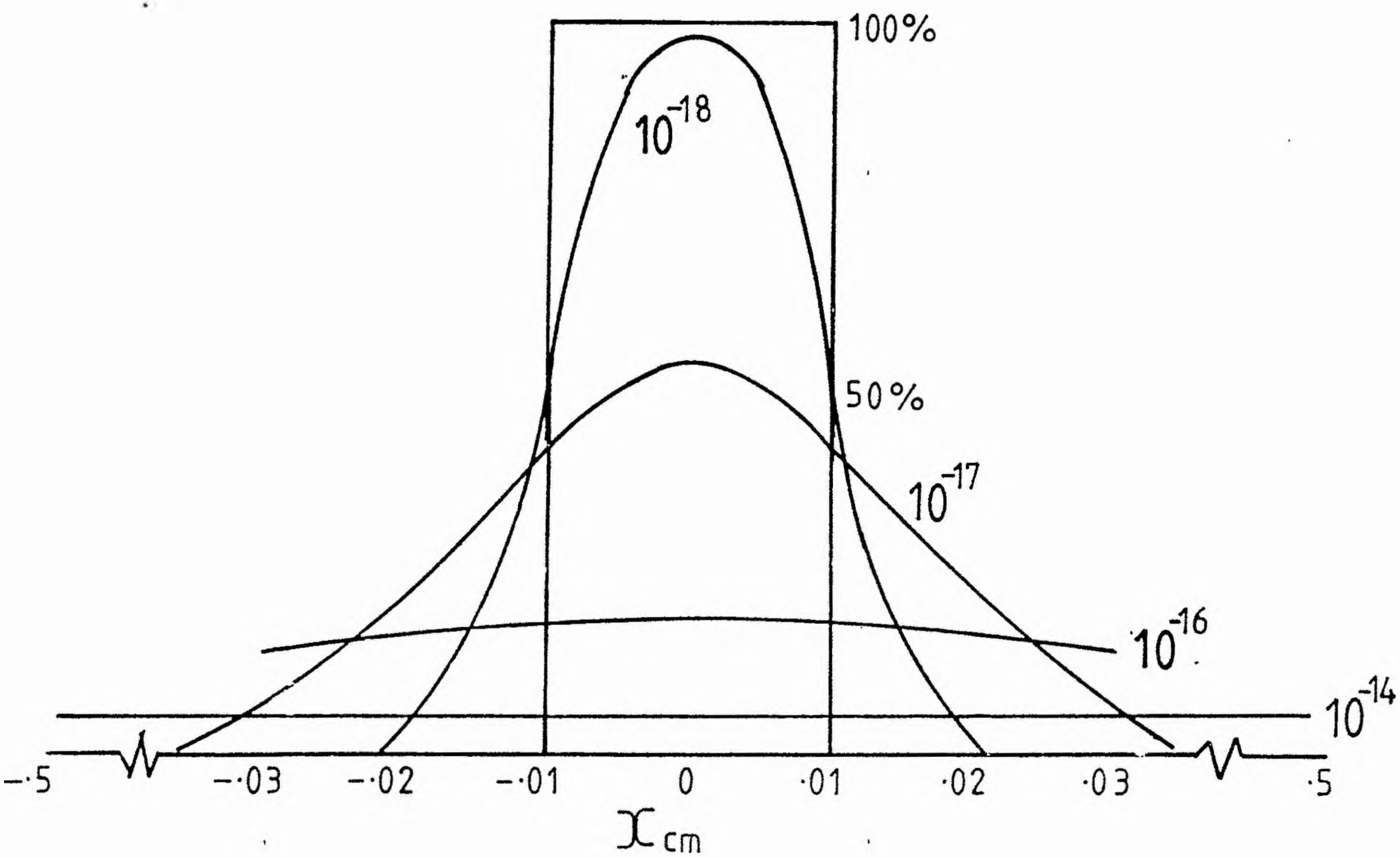
where C(xt) is the concentration of diffusing species at a distance x from the centre of the mineral after time t. h is the thickness of the mineral in the x direction and erf is the error function (see equation (5)) where:

$$\operatorname{erf}(-x) = -\operatorname{erf}(x) \qquad -(10)$$

Using the equation with values of $t=10^5$ yrs, $h=0.1$mm, C1

Fig 6·13 Diffusion from an acicular mineral

$t = 10^5$ yrs. D anotated



(the concentration of the diffusing species within the mineral) = 100%, then Fig 6.13 is generated. This shows clearly that elements within tabular minerals of thickness 0.2mm will equilibrate with their immediate surroundings if the diffusivities are greater than 10^{-17} cm^2/sec. For volume diffusion coefficients lower than 10^{-17} cm^2/sec, no equilibration will be observed regardless of any other transport mechanism operating outwith the mineral (provided all the assumptions above hold).

This leads to the following groups of elements:

1) For elements whose crystalline diffusivities are greater than 10^{-12} cm^2/sec and whose fluid diffusivities are in the order of 10^{-8} cm^2/sec, no preference for the transport mechanism will be shown during the equilibration process.

2) For elements whose crystalline diffusivities are less than 10^{-12} cm^2/sec but greater than 10^{-17} cm^2/sec, and whose silicate fluid diffusivities are in the order of 10^{-8} cm^2/sec, two stage transport will occur during equilibration. That is, volume diffusion to the mineral grain boundary whereupon diffusion metasomatism through the fluid becomes the dominant transport mechanism.

3) Elements whose volume diffusion coefficients are less than 10^{-17} cm^2/sec will not equilibrate unless the grainsize is extremely fine.

For an inclusion of 20 cm diameter, group 1 elements (above) will readily equilibrate with their host magma. Group 2 elements

IONIC DIFFUSIVITIES

Ionic Species	Charge	Diffusivity (cm^{2} /sec^{-1})	Medium	Temperature ^{o}C
Na	+1	$1*10^{-5}$	Obsidian	900
Na	+1	$1*10^{-7}$	Basalt	900
Na	+1	$8*10^{-5}$	Basalt	1300
Sr	+2	$1*10^{-10}$	Obsidian	900
Sr	+2	$5*10^{-9}$	Basalt	900
Sr	+2	$8*10^{-6}$	Basalt	1300
Eu	+3	$9*10^{-11}$	Obsidian	900
Eu	+3	$1*10^{-7}$	Basalt	1300
Ce	+3	$9.5*10^{-13}$	Obsidian	1100
Sr	+2	$1.6*10^{-15}$	Diopside	1250
Sr	+2	$7.3*10^{-17}$	Diopside	1100
Sm	+3	$4.0*10^{-14}$	Diopside	1250
Sm	+3	$9.1*10^{-17}$	Diopside	1100

Table 6.1 Typical diffusivities of ions of interest. Data sources are Hofmann (1980), Jambon (1983) and Sneeringer et al. (1985). Note experimental diffusivities on single mineral grains are several orders of magnitude slower than those obtained for fine grained basalts and obsidians.

will partially equilibrate, the degree of equilibrium depending upon the temperature, the time spent at high temperature, the porosity and tortuosity, and the grainsize of the minerals in which the species of interest resides. Group 3 elements will only equilibrate if the grainsize of the inclusion is very small (<.01mm).

Literature values for volume diffusivities are in the range 10^{-10} cm^2/sec to 10^{-17} cm^2/sec (Hofmann 1980, Henderson 1983, Sneeringer et al. 1984, Maaloe 1985), at reasonable temperatures for intermediate granitoids (950-1050°C Wyllie et al. 1976). However most of the values, particularly the higher ones, were determined on silicate glasses whose fine grainsize and high defect content will give erroneously high diffusion coefficients. Values determined on natural samples tend to be somewhat lower.

Elemental equilibration between host and inclusions.

In considering the application of the theoretical models above, the petrographic data (Chapter 5) and published diffusion data (Hofmann 1980, Henderson 1983, Sneeringer et al. 1984 and Maaloe 1985 and Table 6.1) may be combined to elicit the following general predictions:

1) Alkali metals, which have relatively rapid diffusivities in silicate structures because of their low charge, would be expected to be well equilibrated between host and inclusion. As the ionic size of the alkali metal increases then the degree of equilibration should decrease.

2) Transition metals occurring within ferromagnesian

minerals should be moderately well correlated reflecting the partial conversion of cpx to hornblende. This correlation should decrease as the ionic size of these elements increases.

3) The behaviour of the alkali earths are difficult to predict due to the large range of published diffusivities (Hofmann 1980, Sneeringer et al. 1984). Taking an average, it is likely that Sr should be moderately well correlated if the grainsize is less than 5mm diameter.

4) Elements with high charge, despite their small ionic size, have slow diffusivities and are not expected to have equilibrated.

5) Inclusions which do not contain a pore fluid are not expected to equilibrate with their host for any but the fastest diffusing species.

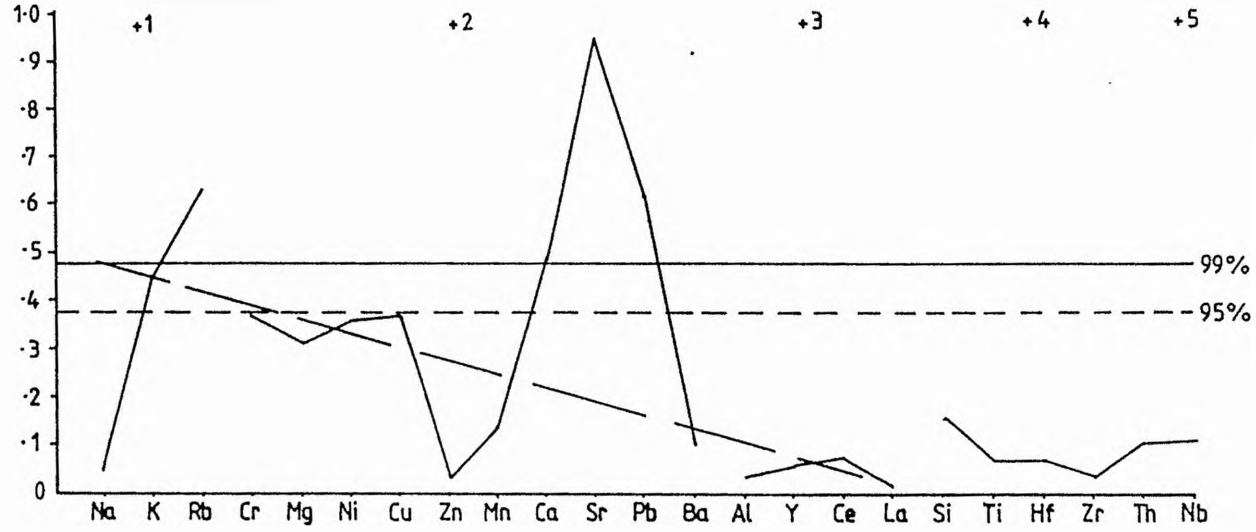
6.5 HOST-INCLUSION ELEMENT EQUILIBRATION

Correlation

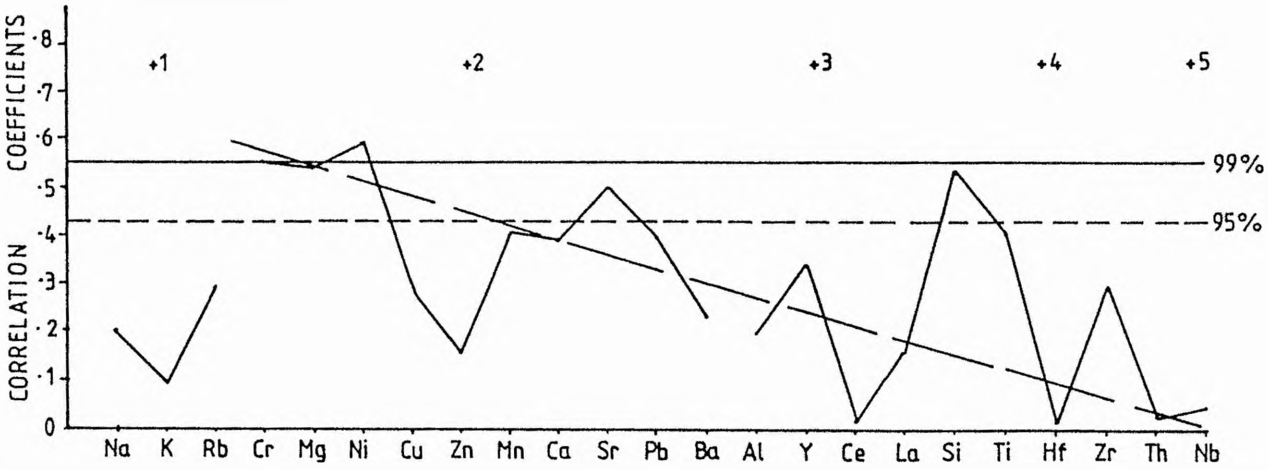
In order to test which elements have equilibrated with their host granitoids a statistical study of the data set was undertaken. Based upon the premise that ions will diffuse down concentration gradients, element distributions between host and inclusions should equalise as equilibration proceeds. In this respect a correlation coefficient (measuring the concentration of an element in the host relative to its concentration in the inclusion, in this case the Spearman Rank Order (Rs) correlation coefficient (Till 1974)) should provide at least a qualitative

Fig 6.14 RANK ORDER CORRELATION COEFFICIENTS

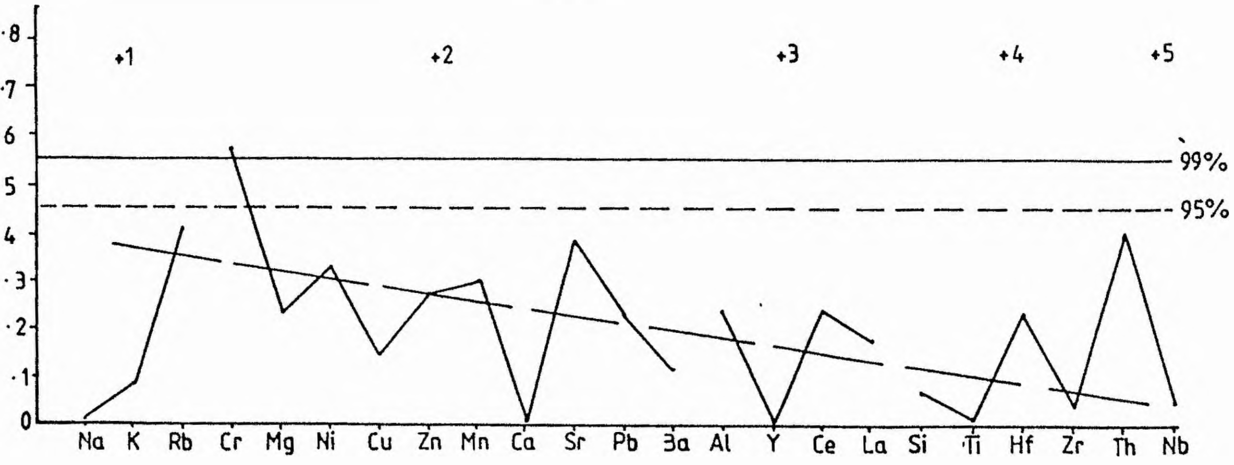
CRIFFELL



STRONTIAN



BALLACHULISH



estimate of the degree of equilibration. Host-inclusion pairs with correlation coefficients (Rs values; see Till 1974 for derivation) close to unity being well correlated and therefore equilibrated and corresponding Rs values close to zero being uncorrelated and therefore disequilibrated. Constructing a diagram which combines both charge and size parameters, as well as a correlation coefficient for each element, is useful in this respect.

In Fig 6.14 the elements are grouped along the abscissa in terms of their charge, and within each group are arranged in increasing ionic radius (data from Henderson 1983, p310-314). Thus within a given group the smallest ion is on the left and the largest on the right. The Rs values are plotted along the ordinate.

According to the theoretical predictions based on diffusion considerations, a discontinuous line with the highest correlation coefficients on the left (low charge) and lowest on the right (high charge) is expected (Fig 6.14). Within each group we would again expect a slope from highest on the left (small size) to lowest on the right (large size).

Whilst it may be seen by inspection of Fig 6.14 that there is a gross tendency for the Rs value to decrease to the right, there are several notable discrepancies, both in anomalously high correlations in (e.g. Sr and Pb in Criffell) and anomalously low correlations (e.g. Na in all three plutons).

Elemental concentrations in inclusions against the corresponding element concentration in the immediately adjacent

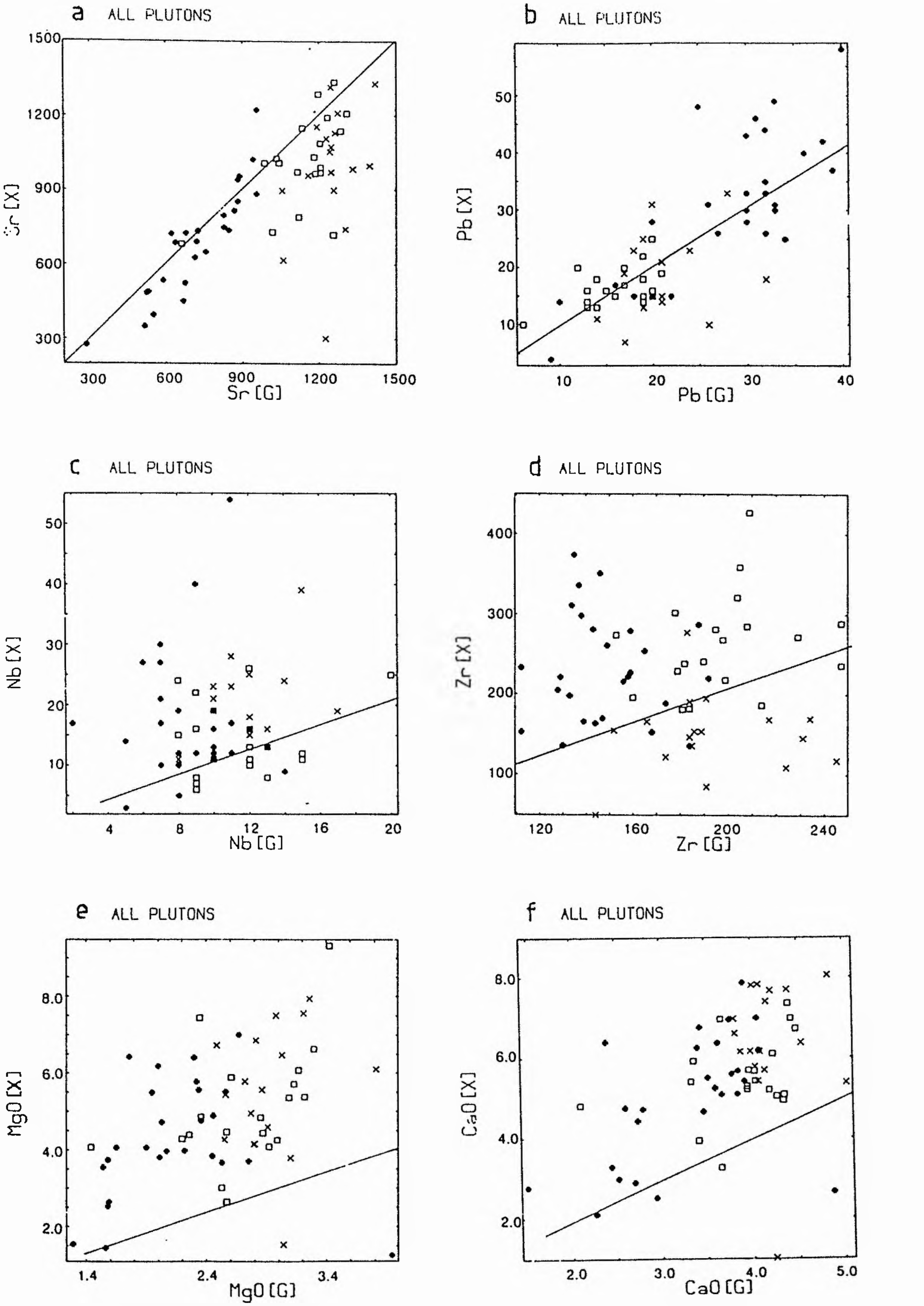


Fig 6.15 Element concentration in the host against element concentration in the inclusion. Symbols BOX-Strontian, X-Ballachulish, MALTESE CROSS-Criffell.

host (i.e. graphical representatives of the correlation coefficient), are given in Figs 6.15a-f. It appears from inspection of the above graphs, that the trace elements attempt to approach a 1:1 concentration relationship with their host, whilst the non-correlated major elements cluster about a line of slope greater than 1:1 (eg Ca and Mg; Figs 6.15 e & f). This non-unity slope for the major elements is taken to be a function of the higher modal percentages of cpx and hornblende in the inclusion as compared to their hosts. Thus whilst trace elements may be relatively free to diffuse (especially those that are held in interstitial lattice sites) and equilibrate, major lattice forming elements are more tightly held reducing the rate of mass transport.

Inspection of Fig 6.15a shows that Sr is extremely well equilibrated within the Criffell pluton (i.e. lying close to the 1:1 line), whilst it is less well correlated for the Strontian and Ballachulish plutons. It may be noticed, however, that the Strontian and Ballachulish host granitoids have a much more restricted concentration range for Sr than does the Criffell pluton. Deviations from the 1:1 line will have a larger effect on the correlation coefficient between these compositionally restricted host-inclusion pairs than is observed for the Criffell pluton.

The fact that those inclusions which deviate from a 1:1 correlation with their host almost without exception contain Sr concentration lower than their host rocks suggests that equilibration is approached by Sr being input from the host into the inclusion. This presents something of a paradox if all the inclusions were acquired by the host magma simultaneously, as the inclusions residing within the more evolved magma (lower Sr content), would have had a longer period of time in which to equilibrate. However, if the inclusions within the more evolved portions were 'second generation' and incorporated at a later stage, then there may have been insufficient time for equilibration to have been effective.

Similar arguments may be put forward for Pb except in the case of Criffell inclusions deviating from the 1:1 line tend to have Pb concentrations greater than their host (Fig 6.15b) suggesting that the direction of Pb transport is opposite to that of Sr.

For elements whose charge is >+2 good correlations on a 1:1 basis between host and inclusions are rare. Exceptions occur for Si and Ti within Strontian and Th at Ballachulish with no significant correlations for the high charge elements in Criffell. Nb and Zr are used to typify the lack of correlation for the high charge groups in Figs 6.15c & d.

Summary of observations.

The important general points may be summarised:

1) Ions with a charge greater than +2 have low correlation coefficients (with the exception of Si and Ti at Strontian and Th at Ballachulish).

2) Na and K show unexpectedly low Rs values.

3) Correlation coefficients generally decrease with

increasing ionic size for transition elements.

4) An anomaly occurs within the +2 group for the elements Ca, Sr and Pb (except for Ca at Ballachulish). The peak corresponds to an ionic radius of around 1.1 to 1.2 A.

5) Elements may be grouped as follows:

a) Well correlated elements, i.e. those with Rs values >99% significant.

b) Moderately correlated elements i.e. those which lie between the 95 and 99% significance levels.

c) Non-correlated elements whose Rs values lie below the 95% confidence level.

Discussion.

Ignoring for the moment Na and K, Ballachulish adheres most closely to a predicted line sloping from left to right. Strontian and especially Criffell have large peaks and troughs deviating from the expected line. In order to explain these large anomalies it is necessary to study each group more closely.

Elements with a charge > +2.

In general the high field strength elements (HFSE) behave in a manner similar to that predicted earlier, in that their correlation coefficients are low, with the exception of major elements Si and Ti at Strontian. Why these two elements should be relatively well correlated at Strontian is not easily reconciled with their low Rs values at Criffell and Ballachulish, and may represent a chance correlation. Recrystallisation of major Ti bearing minerals (e.g. sphene) would not be expected to significantly alter the Si budget as their modal percentage is low, but could explain the high Rs value for Ti.

Most HFSE reside in accessory phases (Henderson 1980) and it is noticeable that no matter how severely deformation has affected other minerals within the inclusions, minor phases such as zircon and apatite have remained pristine throughout.

In general therefore HFSE have low correlation coefficients as predicted. This suggests that their high charge precludes significant chemical diffusion in the solid state at temperatures associated with intermediate magmas.

Elements with a +2 charge.

For convenience the +2 charge group is split up into the two geochemically similar subgroups namely a) the transition elements, and b) the alkali earths and Pb.

a) Transition elements: The group Cr to Mn behave exactly as predicted in that that smallest ion (Cr) has consistently the highest correlation coefficient of the group. The Rs values decrease systematically with increasing ionic size (with the occasional reversal which is possibly a statistical result of the sample size) and give low correlation coefficients for Zn and Mn (Fig 6.14).

Petrographic evidence for the partial conversion of clinopyroxene to hornblende (Chapter 5) for inclusions from all three plutons, is evidence for the on-going equilibration of the inclusions. The fact that the reaction has not gone to completion is indicated by the correlation coefficients of the major ferromagnesian elements Fe, Mn, Ni, Cr being in the region of 0.3 to 0.5, that is 90-95% significant.

b) Alkali earths + Pb: All three plutons (against the general prediction) show sharp peaks in the correlation coefficient for the alkali earths, especially for the Criffell pluton (Fig 6.14). A more detailed look reveals that the peak corresponds to an ionic radius between 1.0-1.2 angstroms (Ca=0.99, Sr=1.12, Pb=1.2, Ba=1.34) close to the approximate radius for vacancies within plagioclase feldspar (1.18 angstroms, Henderson 1983, p.133). The correlation coefficients fit extremely well to the notion that equilibration of these elements was controlled by the ease of diffusion through the solid plagioclase matrix.

The petrographic observations support this view (Chapter 5), that is, plagioclase megacrysts contain a core of relict oscillatory zoned patches and an outer mantle which is normally zoned and compositionally more albitic. These observations suggest that the outer mantling, while plagioclase megacrysts were in equilibrium with their host, failed to equilibrate totally due to low diffusion rates. Smaller plagioclase grains within the matrix of the inclusions would therefore be expected to have more completely equilibrated,and this may be the case as they are of oligoclase composition and only weakly zoned. This does not however, explain why the correlation coefficient for Sr at Criffell is 0.9, at Strontian 0.5 and at Ballachulish 0.38.

The observation of such differing correlation coefficients for similar elements within similar inclusions equilibrating with similar magmas requires explanation, especially in view of the identical (and predicted) behaviour of the +2 transition elements for all three plutons.

Effects of inclusion size, grainsize and nature of diffusing medium which were identified as important rate controlling parameters, may be eliminated as differences between the three plutons as host-inclusion pairs are physically similar. The striking correlations for Sr & Pb (Fig 6.15 a & b) in the Criffell pluton also rule out temperature as a major difference. The low Sr host rocks (more evolved and lower temperature) contain well equilibrated inclusions as with the less evolved, higher temperature portions. If temperature were the rate controlling factor, then the more evolved rocks would be expected to contain inclusions which progressively deviate from a 1:1 relationship.

We must envisage other processes therefore in order to explain these data, such as:

1) Statistical effect of a greater compositional range of the host rocks in Criffell than for Strontian or Ballachulish.

2) Increase in diffusion rates due to deformation of the host rock and inclusion during emplacement.

3) Crystallisation sequence of the feldspars.

4) Residence time of the inclusions within the host magma

Fig 6.15 a & b show that the concentration ranges of Sr and Pb are greater for Criffell hosts than for Strontian or

Ballachulish hosts. Deviations from the 1:1 correlation line have, therefore, a far lower statistical effect on the Rs value than for the compositionally restricted Strontian and Ballachulish plutons. Although the extended compositional range of the Criffell pluton does contribute substantially to the high correlation coefficient for Sr and Pb, it cannot alone explain the adherence of the elements to a 1:1 concentration dependence between the host and inclusions.

Deformation of a solid mineral has the effect of promoting dislocations and vacancies within the lattice (Vernon 1976). Diffusion rates, which are dependent on the availability of suitable vacancies, will of course increase, the amount of increase depending upon the magnitude of deformation. Thus inclusions which have been substantially deformed are most likely to have feldspars which are in equilibrium with their host . It must be remembered that the total volume of plagioclase within the cores of the megacrysts is small compared to the total modal volume of plagioclase within an inclusion. An inclusion which contains a few plagioclase megacrysts which are not in equilibrium may well lie upon the 1:1 correlation line with its host because the vast majority of plagioclase minerals within the matrix have equilibrated.

If deformation is an important mechanism controlling equilibration of feldspars (the dominant rock forming mineral), we would predict - that Criffell (with the most highly deformed inclusions) would show the highest correlation coefficients. Strontian with inclusions close to the periphery being deformed, should lie on the 1:1 correlation line for the more basic hosts (highest Sr content) deviating from a 1:1 relationship for more evolved members. Ballachulish, which does not contain a deformation fabric, would be expected to have a relatively poor correlation coefficient for Ca, Sr and Pb compared to the other two plutons. However Criffell inclusions which lie within the lower temperature and undeformed member should, like Ballachulish, yield substantial deviations from the 1:1 line. The lack of such deviations suggests that whilst deformation will undoubtedly aid equilibration, it does not answer the question of why Criffell is so much better correlated for Sr and Pb than Strontian or Ballachulish.

The clustering of high correlation coefficients for host-inclusion pairs around the 1.2 angstrom ionic radius suggests that diffusion within the feldspar structure controls the rate of equilibration. If feldspars grew within the inclusions at a late stage in the presence of a migrating pore fluid, then the feldspars within the inclusions would be expected to grow in equilibrium with those in the surrounding granitoid melt. The fact that clinopyroxene is converted to hornblende and biotite in the inclusions suggests that water had infiltrated the inclusion from the granitoid magma. Such infiltration metasomatism is also evident from the high K values for inclusions (Fig 6.5). However, such a process, whilst explaining why elements with high partition coefficients for feldspars show good correlations, does not explain why the correlation is so much better for Criffell than elsewhere.

The residence time of an inclusion within its host at high temperature is an unknown quantity. Those inclusions at Strontian and Ballachulish which lie on or close to the 1:1 correlation line (Fig 6.15) tend to have been sampled from the outermost portion (least evolved and highest temperature) of each pluton. Significant deviation from the 1:1 line occur for only three samples at Strontian (S210X, S331X, S352X) which lowers the correlation coefficient. At Ballachulish, however, only a few samples achieve a 1:1 correlation with their host, the others having a somewhat lower Sr concentration within the inclusion.

Syn-plutonic intrusives were observed in the Strontian pluton which includes the appinite bodies (Chapters 3). The inference is that syn-plutonic injection of mantle-derived magma has occurred throughout the pluton's history. Inclusions incorporated at a late stage in the crystallisation sequence will not have had sufficient time for equilibration to reach completion.

The evidence from Figs 6.14 & 6.15 is that inclusions which are not equilibrated lie to lower Sr concentrations for the inclusions and that these concentrations, given time, migrate towards the 1:1 line. At Criffell the close adherence of the host-inclusion pairs to this line suggests that its inclusions were incorporated at an early stage of pluton development and sufficient time has elapsed to allow the equilibration of these elements between host and inclusions. Strontian and Ballachulish on the other hand must have attained their inclusions over the full spread of their emplacement histories as some inclusions are more equilibrated than others.

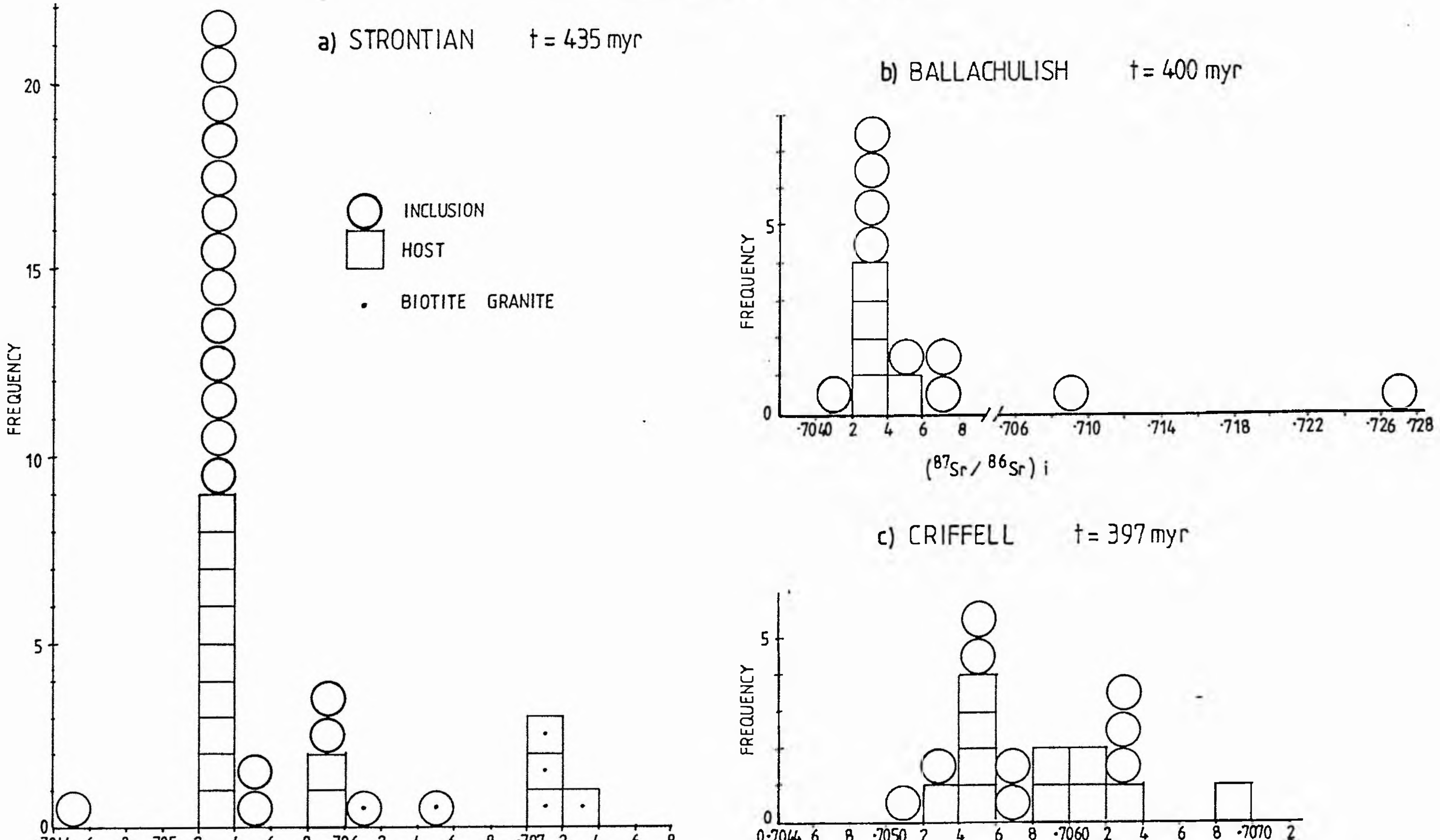
In essence, therefore, given sufficient time the elements Ca, Sr & Pb will equilibrate in a rational manner as predicted. It is also noted that in all three plutons Ba which has a significantly larger ionic size than Ca, Sr or Pb does not equilibrate as readily.

Alkali metals

Na shows the greatest deviation from the predicted values of all the elements analysed. Its small charge and size, and its substantial presence within plagioclase feldspar, together with the high correlation coefficients shown by other feldspar components (Ca, Sr and Pb), indicate that Na should be well correlated with its host. Further, due to the increasing ionic size (Na < K < Rb), Rb would be expected to have the lowest correlation coefficient, exactly opposite of what is observed (Fig 6.14).

This paradox is not easily reconcilable with known diffusion rates (Hofmann 1980). At Criffell, however, detailed geochemical surveys of the pluton (Stephens 1972, Stephens & Halliday 1980, Stephens et al. 1985) have presented composition surfaces for major and trace elements. All produced (roughly) concentric trend surfaces, with the notable exception of Na which behaved randomly. Thus in the purely magmatic sense Na was either decoupled from other elements during the plutonic evolution or that Na continued to be mobile after crystallisation was

Fig 6·16 Strontium isotope ratios



complete. Large scale differences in Na concentration may well be easily explained by fluid pathways, but on the centimetre scale, one would expect host and inclusion to be 'flushed' by the same post-magmatic fluid. Failure to maintain equilibrium may mean that fluids pass preferentially through the host, bypassing resistant, finer grained inclusions.

Potassium is thought to behave in a similar manner but to a much lesser degree, whilst Rb behaves more or less as predicted showing good correlation. In conclusion therefore, the mobility of Na is such that it continues to move in post-magmatic fluids. K and Rb also remain mobile but to a much lesser extent.

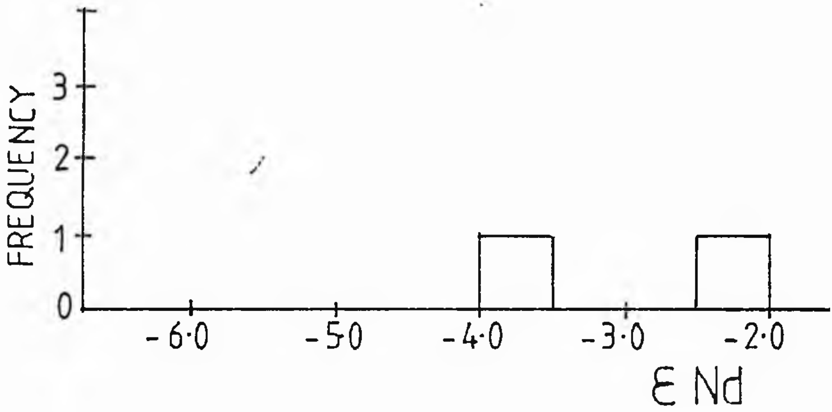
Isotopic constraints.

In simple terms it is clear that whilst the majority of host-inclusion pairs have similar Sr isotope compositions (Fig 6.16), Nd isotopes remain separate and distinct (Fig 6.17). This is not an entirely surprising result in view of the earlier findings on the distribution of elemental abundances between host and inclusions. However, whilst Nd isotopes remain distinctly more radiogenic in comparison with their host granitoids, it cannot be assumed that no equilibration has taken place despite the low correlation coefficients exhibited by the REE.

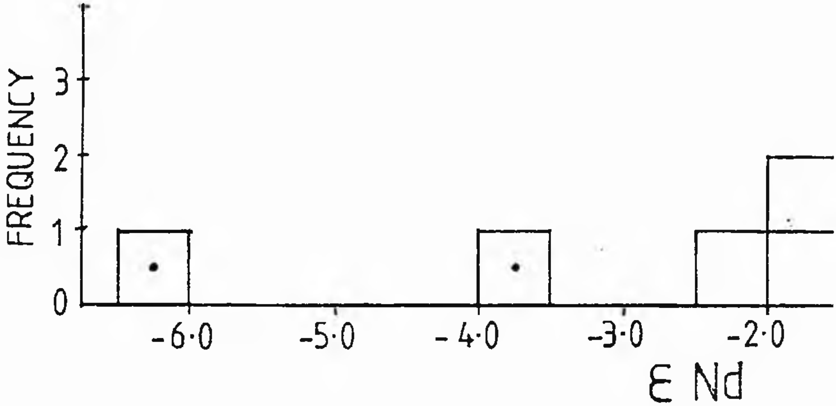
To test independently the degree to which Sr and Nd isotopes equilibrate with their hosts, three clearly identifiable country rock xenoliths from within the Ballachulish pluton (samples B205X, B311X & B320X) were analysed together with two Ballachulish hosts (B205G & B311G). Knowing what the approximate isotopic composition of local country rock sources were (i.e.

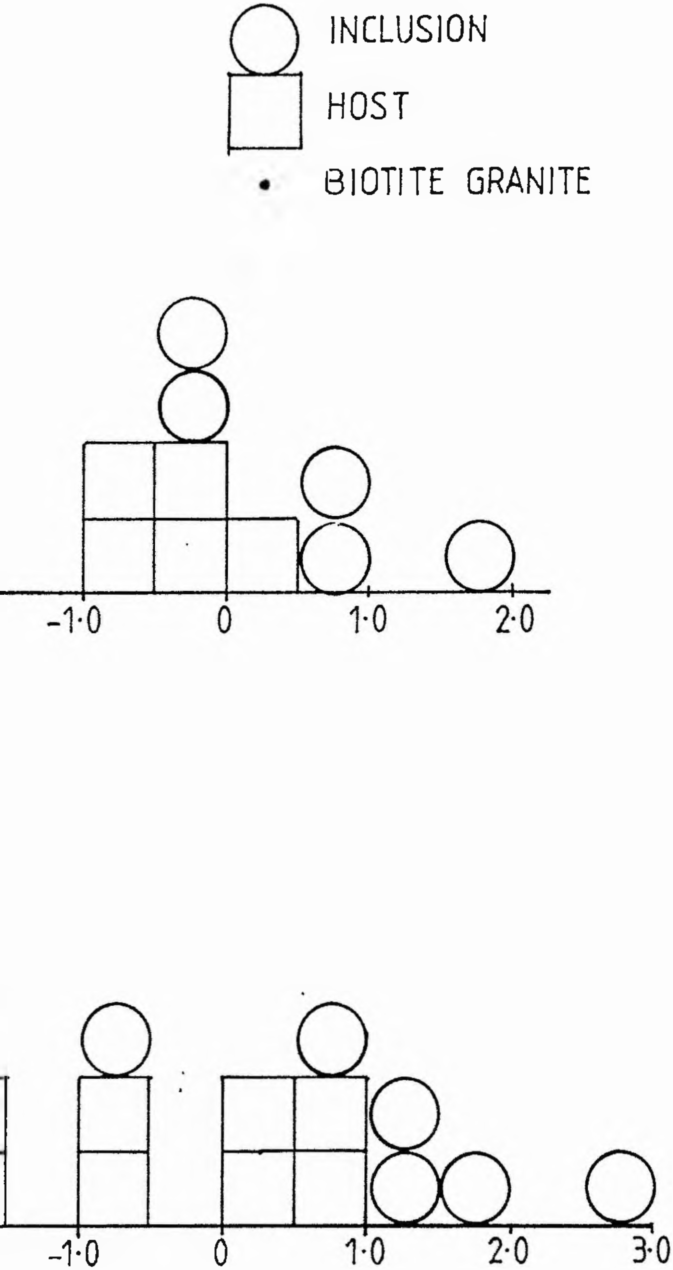
Fig 6·17 Nd ISOTOPE COMPOSITIONS

CRIFFELL t = 397 myr



STRONTIAN t = 435 myr




INCLUSION
HOST
BIOTITE GRANITE
-1·0
0
1·0
2·0
-1·0
0
1·0
2·0
3·0

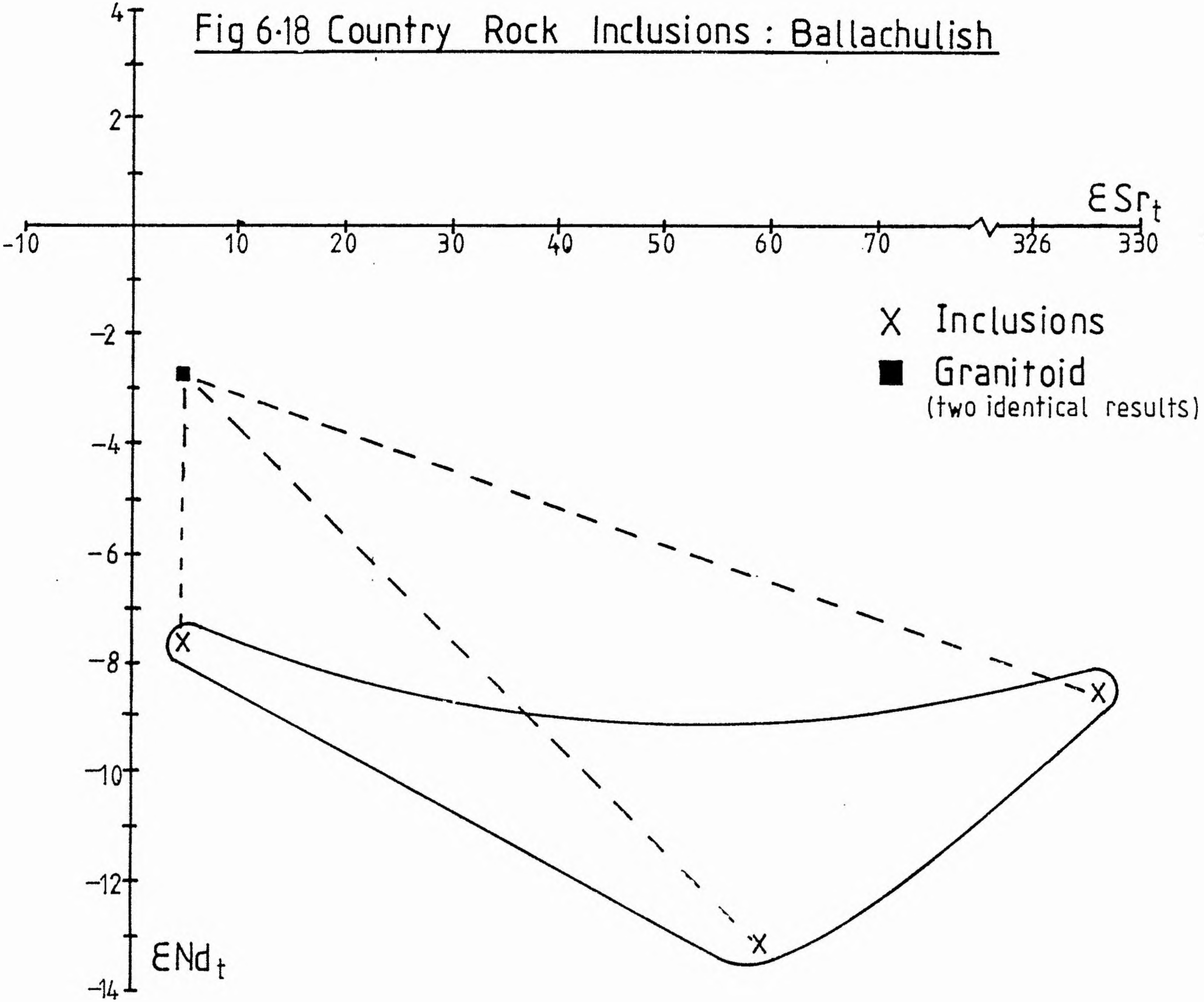


Fig 6·18 Country Rock Inclusions : Ballachulish

Dalradian metasediments), it is possible to estimate the degree to which these rocks have equilibrated.

Fig 6.18 gives the analytical results of this test and shows quite clearly that Sr isotopes (as expected) more towards equilibration.It further demonstrates that whilst assimilation of the inclusions proceeds, Nd isotopic compositions remain largely unaffected at pristine metasedimentary compositions.

An inclusion of Strontian porphyritic granodiorite (S402X) entrained within the central biotite granite retains its primary Nd isotopic signature ($\epsilon Nd_t = -0.6$ compared to central biotite granite of -3.2 to -6.3) whilst it s Sr isotopic signature is a little higher than average for granodiorite at 0.70605.

It is likely that the less evolved micrograniroid inclusions will not equilibrate with granitoid magmas as readily as metasedimentary ones, as the former have no readily fusible fraction (Bowen 1922, Tindle & Pearce 1983). The expectation, is therefore, that microdiorite inclusions will retain pristine or near pristine Nd isotopic compositions provided the inclusions have not been excessively flattened in the emplacement process, for periods longer than are normally associated with the cooling of granitoids.

Conclusions

From the theoretical and experimental study of host and inclusion pairs from the Strontian, Criffell and Ballachulish plutons, the following points may be raised.

1) Volume diffusion of an element in a mineral is the rate

controlling factor determining the degree to which an element will equilibrate between the host and inclusion. Mass transport in any pore fluid either by infiltration or diffusion metasomatism is likely to be relatively rapid by comparison. Hence, the rate at which an element reaches the mineral grain boundary from within the mineral is slow compared to the rate at which it travels through the pore fluid.

2) Elements with a charge greater than +2 do not diffuse easily through the solid silicate matrix. As a consequence, the amount to which they equilibrate between host and inclusion is small. Isotope systematics based upon high ionic charge elements are likely to reflect better the original isotopic signature of the inclusions prior to incorporation into the granitoid, i.e. Nd isotopic ratio is potentially a good indicator of source.

3) Elements with an ionic charge of +2 equilibrate to varying degrees with their host. The variation is a function of the ionic size of each of the species and equilibration decreases as the ionic size increases throughout the first transition series. For the alkali earths, the degree of equilibration increases as an ionic radius of 1.18 Angstroms is approached peaking at Sr and Pb (ionic radius 1.12 and 1.20 Angstroms respectively). Isotopic systematics based upon these elements would be reset for the inclusions and would not be expected to retain any memory of their original isotopic composition.

4) Deformation of inclusions promotes disorder within the constituent minerals and increases the rates of volume diffusion. Plutons where significant deformation has occurred yield higher

correlation coefficients for the +2 elements than for plutons which remain undeformed.

5) Alkali metals behave in an irrational manner. Experimentally determined diffusivities indicate that Na is the most mobile of the elements analysed (Hofmann 1980). However it remains consistently uncorrelated between inclusion and host in the three study plutons, and may remain mobile after the crystallisation of the host is complete. K and Rb become progressively more equilibrated with increasing ionic radius (against theoretical expectation) and may reflect a decreasing mobility in post crystallisation fluids.

6) Equilibration rates by diffusion, without the aid of significant deformation, are slow compared to the crystallisation time of the host. It is unlikely, therefore, that the lack of inclusions within the central portions of the study plutons is a function of their increasing assimilation, as microdioritic inclusions are unlikely to be assimilated by melting.

7) Sr isotopes readily equilibrate between host-inclusion pairs, whereas Nd isotopes retain pristine or near pristine isotopic compositions.

6.6 DIVERSITY OF INCLUSIONS

Having established those elements which were unlikely to have been mobilised during inclusion assimilation, it is now possible to investigate the earlier observation of inclusion clustering on geochemical diagrams in terms of 'real' source differences.

STRONTIAN

S(1)	S(3a)	S(3b)	S(3c)	S(3d)
S438	S203 S210 S212 S341 S343 S402	S205 S208 S211 S213 S340 S344 S345 S346 S347 S349 S351 S353 S354	S202 S209 S348 S352 S355	S338

Table 6.2 Strontian samples grouped according to their inclusion type both host and inclusions share the first four digits of the sample number, samples with suffix-X are inclusions, samples with suffix-G are their immediately adjacent host.

BALLACHULISH

B(1)	B(3a:1)	B(3a:2)
B201 B205 B306 B311 B319 B320	B207 B208 B219 B315 B317	B202 B204 B301 B307 B308 B309 B312 B314 B316 B318

Table 6.3 Ballachulish samples grouped according to their inclusion type.

CRIFFELL

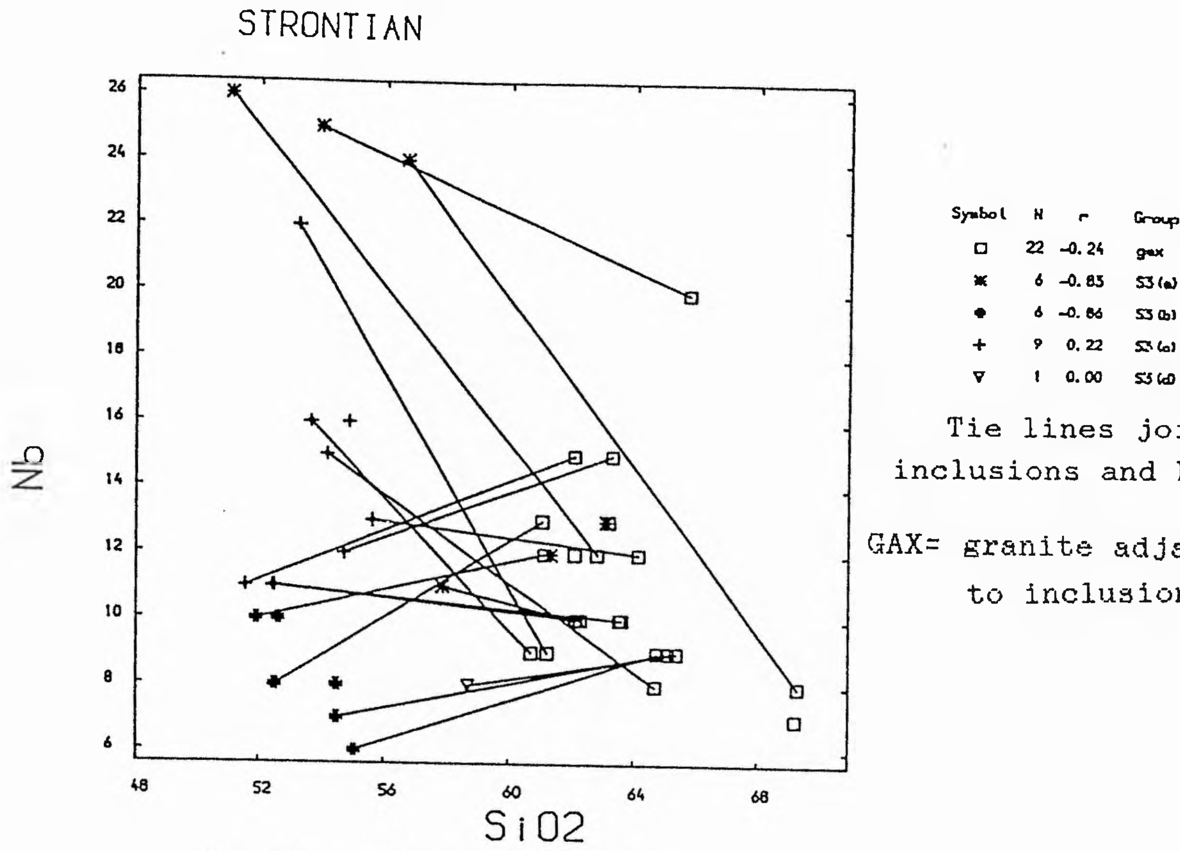
C(3a:1)	C(3a:2)	C(3b:1)	C(3B:2)
C206 C208 C210 C333 C335	C209 C301 C328 C331 C354	C201 C202 C204 C205 C319 C322 C324 C325 C326 C341 C343 C345 C346 C350	C304 C318 C320 C321 C323 C336 C337

Table 6.4 Samples grouped according to their inclusion type.

Inclusion varieties within the Strontian pluton

Nb is a good inclusion-type discriminator and all three major sub-groupings within the Strontian pluton may be readily identified on the Nb vs SiO_2 diagram (Fig 6.19, inclusions of similar petrographic type are listed in Table 6.2). The low Nb content (<11ppm) of the sub-type S3(b) is characteristic and suggests a slightly different source for this type with respect to other inclusion sub-types, as a genetic link via fractional crystallisation seems unlikely. The Nd isotope ratio of a representative of the sub-type (S210X) further confirms this view as it records the most radiogenic ϵNd_t value of any igneous rock in the Scottish Caledonides so far recorded (+3.0), and clearly rules out any cogenetic relationship with other inclusion sub-types or the host granitoid itself.

Nb can be further used in identifying two subsets of the fine grained type (S(3a)), namely high Nb-low SiO_2 and low Nb-high SiO_2 (Fig 6.19), suggestive of two separate groups. Nd isotopic evidence supports this view in that sample S207X (low Nb-high SiO_2) has an ϵNd value of -0.8 (the lowest of the Strontian inclusions) whilst sample S341X has an ϵNd value of +1.3. Furthermore since sample S207X falls within the host granitoid field in the ϵNd-$^{87}Sr/^{86}Sr$ plot, it is reasonable to assume that the sub-set containing low Nb abundances are genetically related to their host granitoid. Members of the low Nb sub-set plot consistently with the host granitoids on all trace element diagrams, supporting this conclusion.



Tie lines join inclusions and hosts.

GAX= granite adjacent to inclusion.

Fig 6.19 Nb against SiO_2.

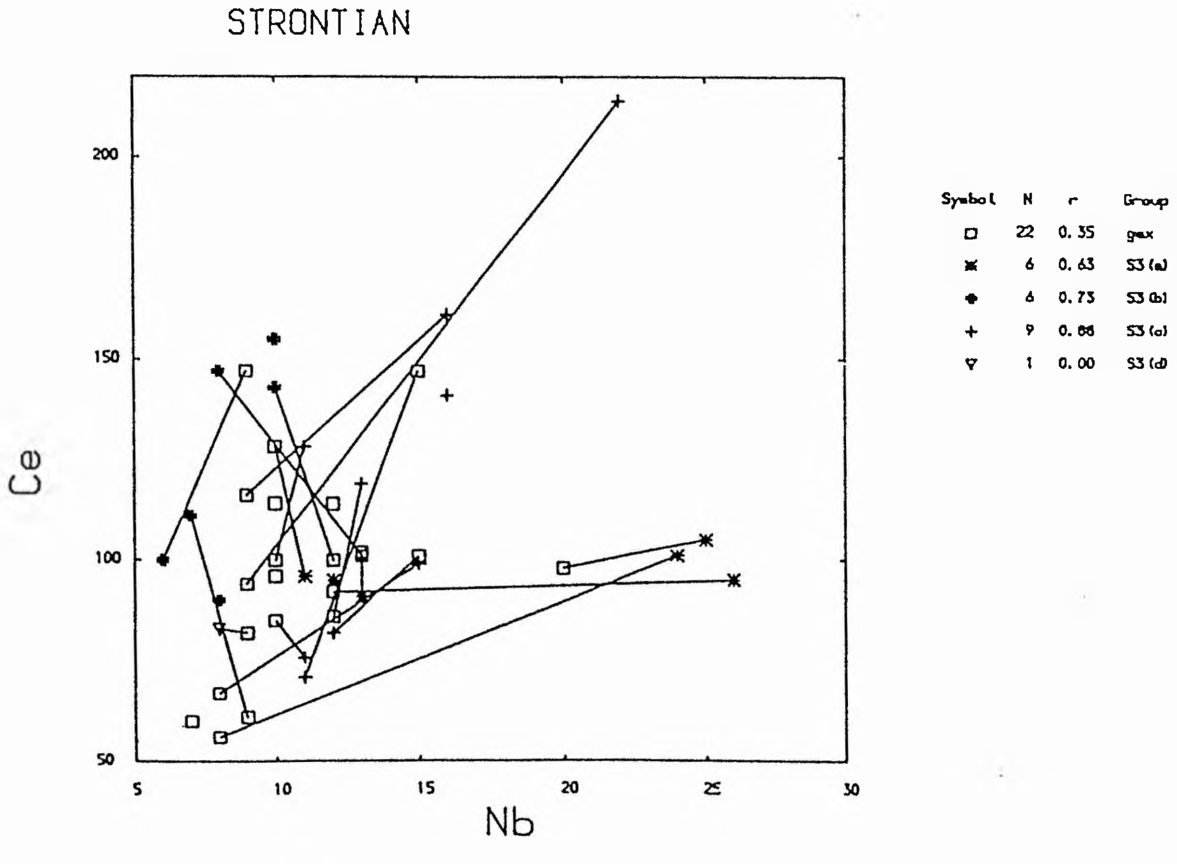


Fig 6.20 Ce against Nb.

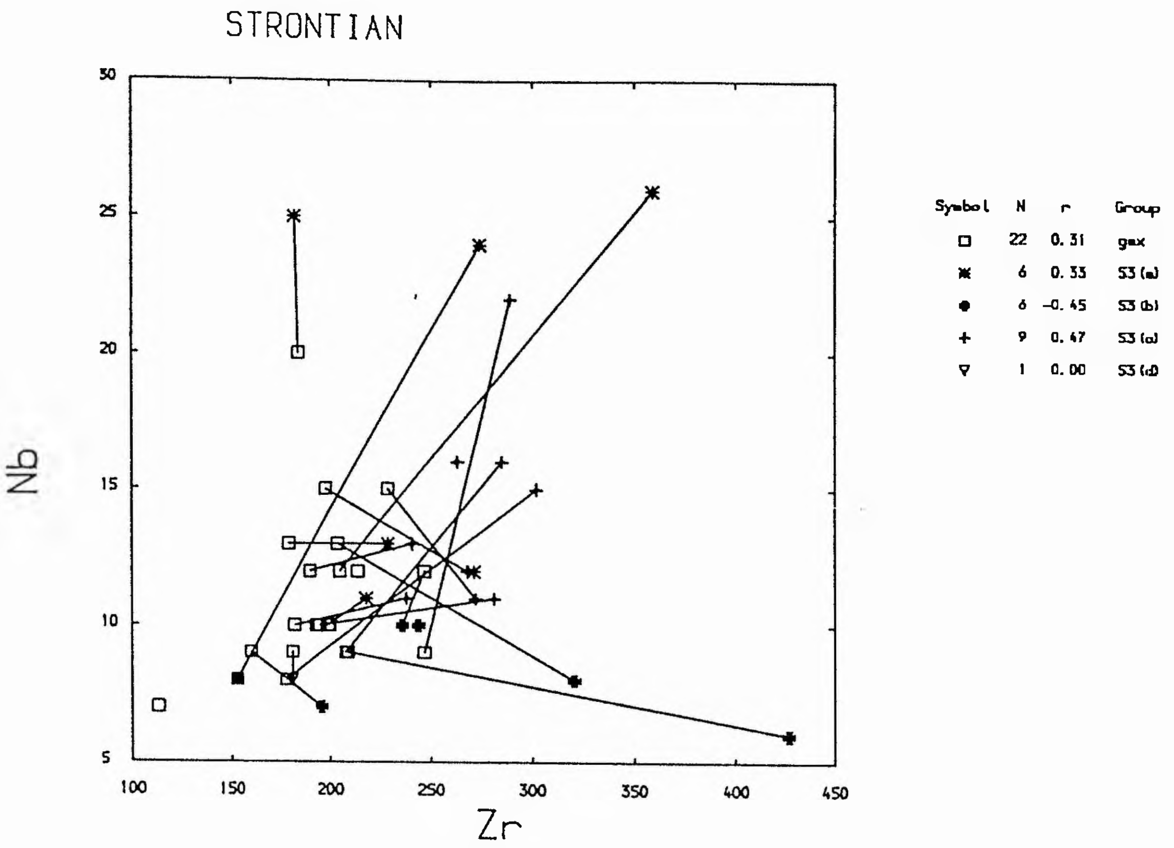


Fig 6.21 Nb against Zr.

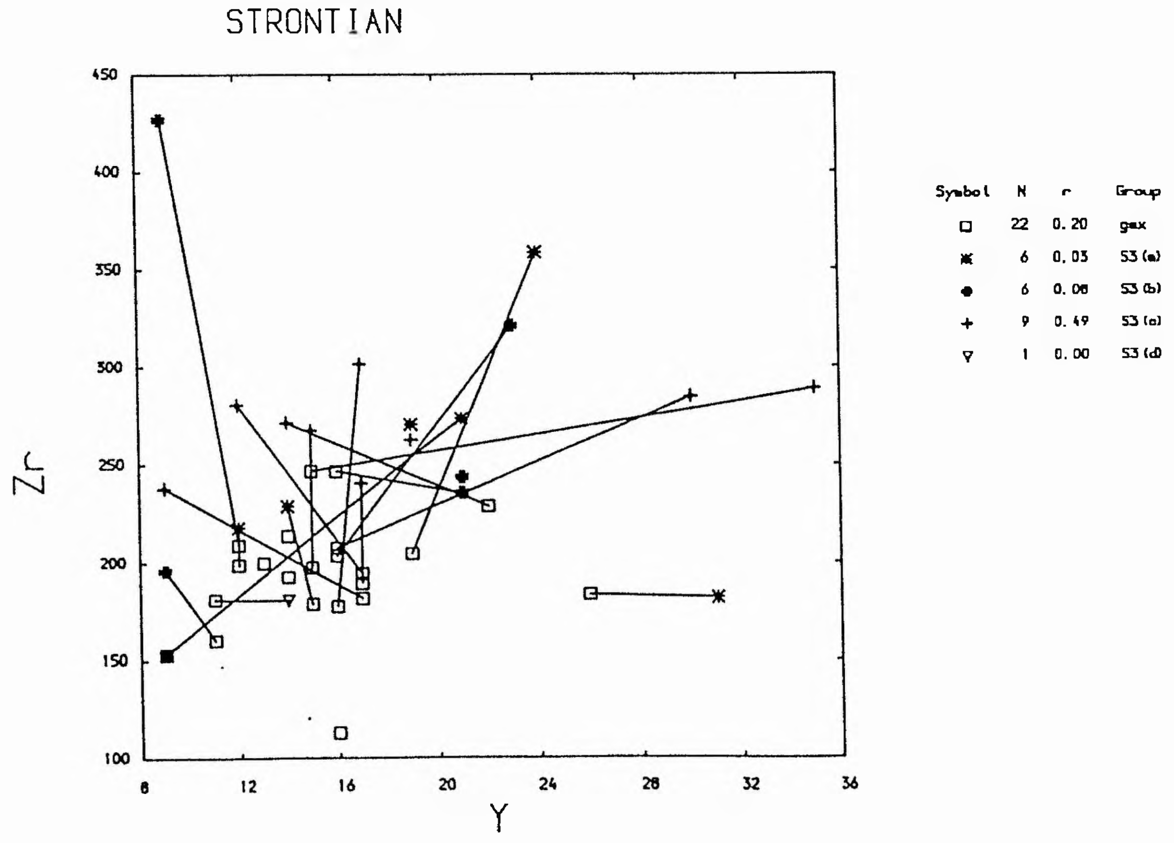


Fig 6.22 Zr against Y.

Inclusions containing plagioclase megacrysts form a coherent group with a tendency towards high LREE contents (Ce up to 215 ppm, see Fig 6.20). ϵNd values for a representative sample (S345X) was +0.5, slightly above the norm for the host granitoids but lower than the other inclusions (except low Nb sub-set of Class S(3a)).

The medium-coarse grained 'felsic' inclusion (S338X) is not as siliceous as it appeared in the field (Fig 6.19). It plots consistently with the ferromagnesian aggregate group (Figs 6.19 to 6.22) and has an ϵNd value of +1.1, suggesting no simple relationship with the hosts. However, the fact that it plots consistently with Type S(3c) suggests that it is a member of that suite despite the relatively low ϵNd value. If this is so, the coarse grainsize (contrasting with that of the inclusions rich in ferromagnesian aggregates) suggests fractionation prior to incorporation, yielding the two contrasting inclusion types. Conversely, the apparent association of sample S338X with Type S(3c) may be coincidental and only one representative was found; it may not be typical of the group.

Inclusion varieties within the Ballachulish pluton.

Although Nd isotopic analysis was not performed upon microdiorite inclusions from the Ballachulish pluton, it is instructive to show how easily country rock inclusions may be identified geochemically with respect to their micrognanitoid equivalents.

Fusion of a minimum melt composition from country rock

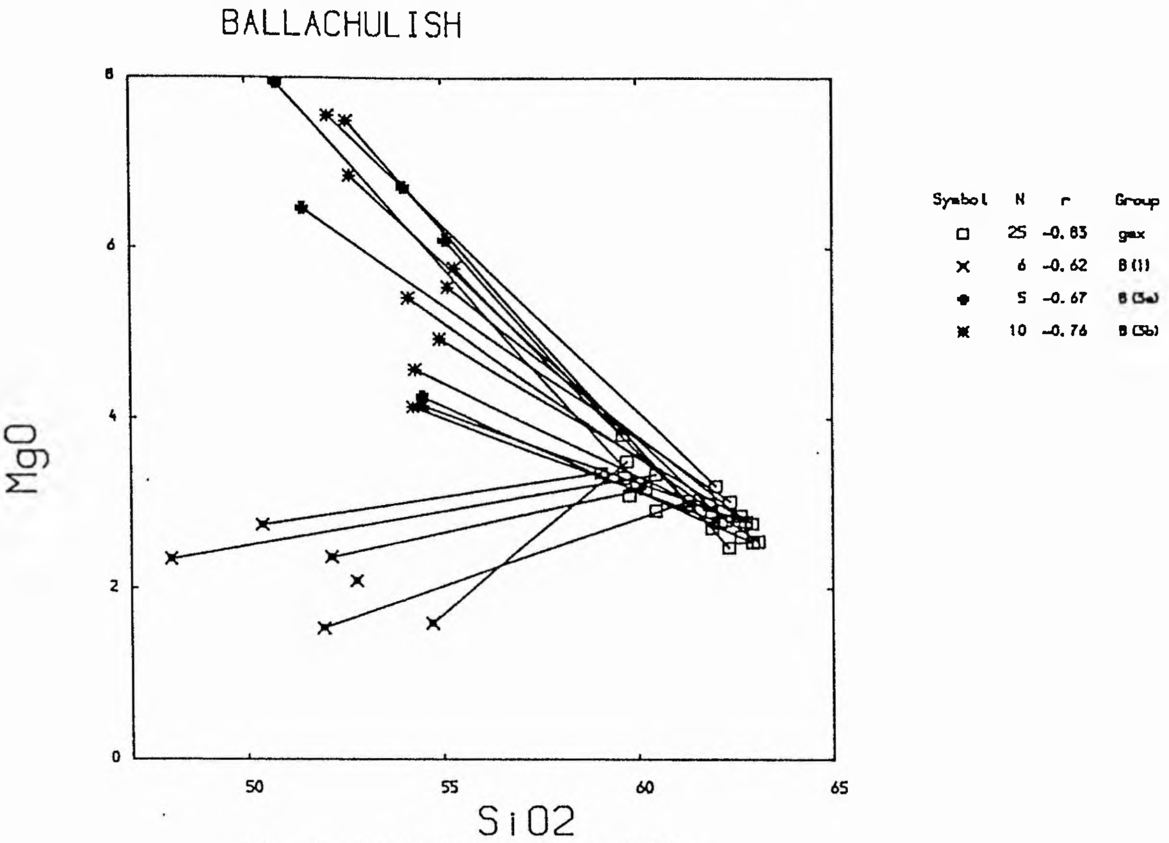


Fig 6.23 MgO against SiO_2.

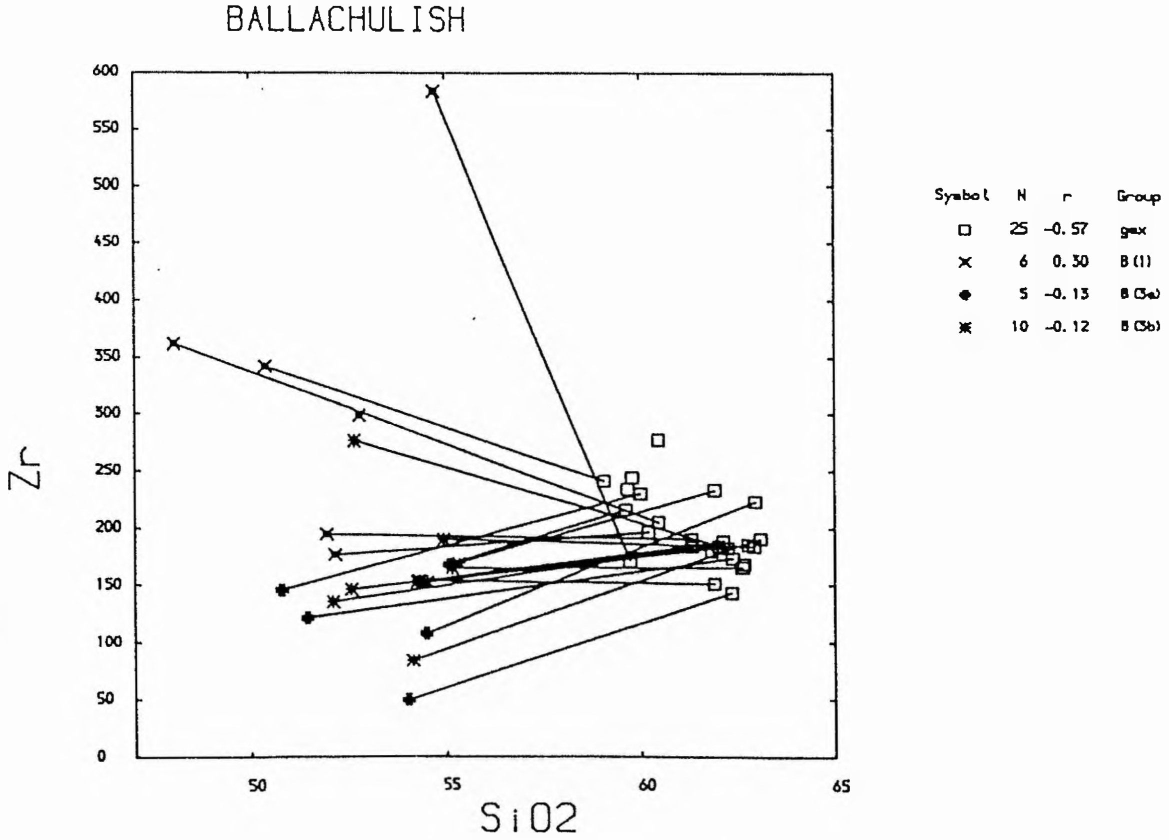


Fig 6.24 Zr against SiO_2.

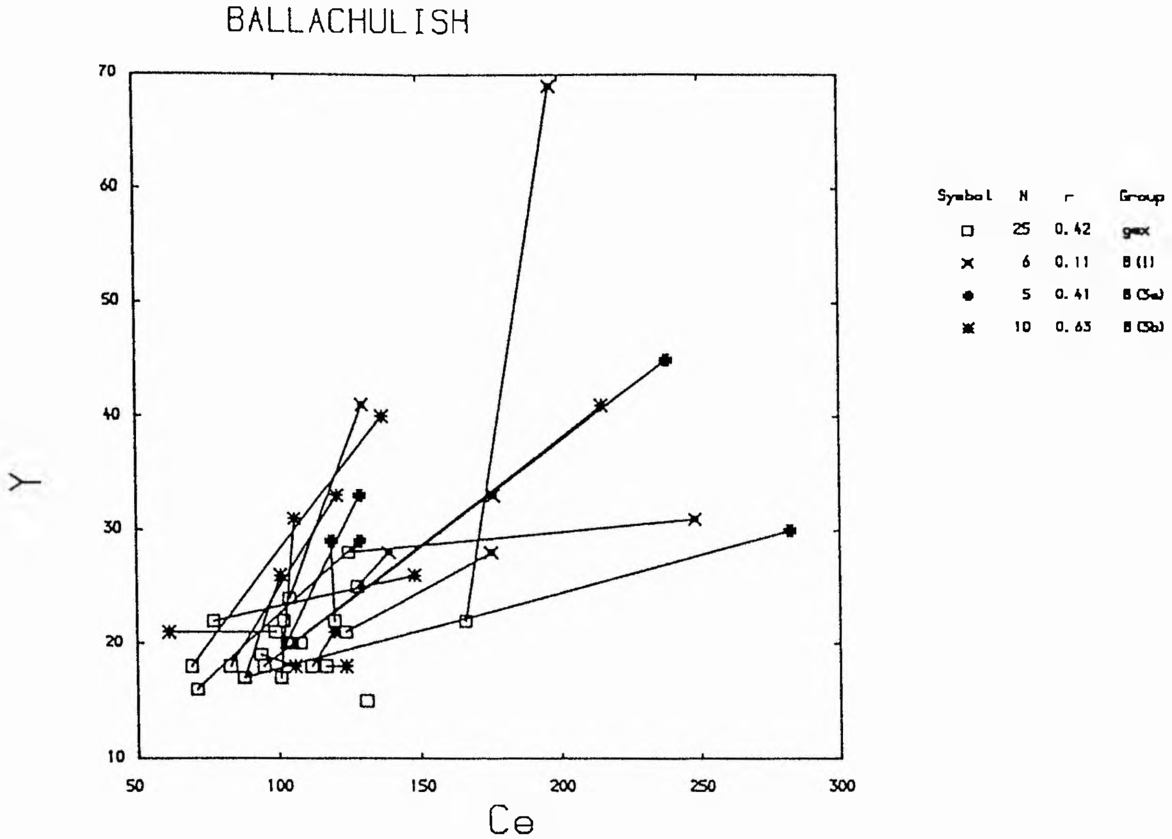


Fig 6.25 Y against Ce.

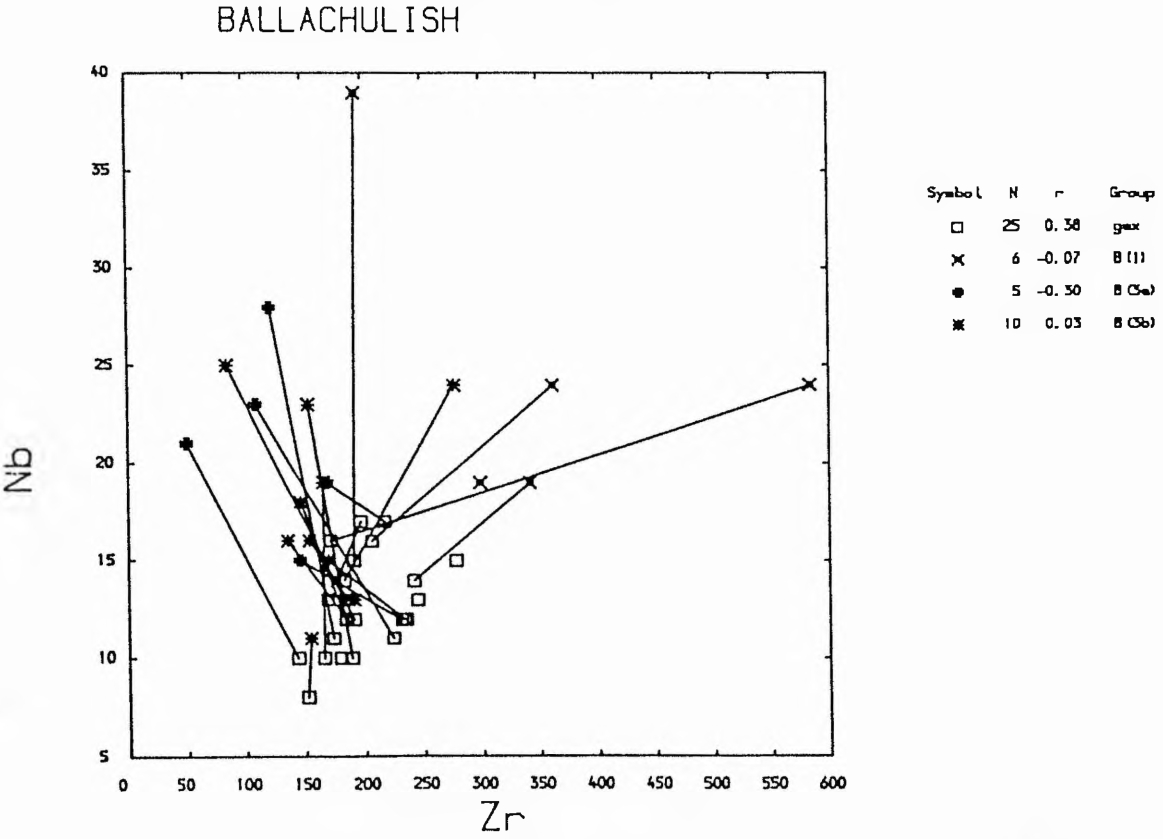


Fig 6.26 Nb against Zr.

fragments is unlikely to yield a discrete inclusion rich in refractory components, as the inclusions themselves would be expected to disintegrate due to the wetting effect of the melt (Bowen 1922). Furthermore there is no tendency for the country rock group (Fig 6.23), which are at differing degrees of equilibration, to trend towards Mg richer compositions. It may be concluded that the microgranitoid inclusions within the Ballachulish pluton are not derived from the partial digestion of country rock.

Three groups of microdiorite inclusions were identified on the basis of petrographic and field observations, and samples representative of each group are given in Table 6.3. Two of these groups were deemed subsets of the same primary group (Type B(3a:1) and B(3a:2)) whilst the other (B(3c)) contains glomeroporphyroblasts of ferromagnesian minerals. One medium to coarse-grained sample was also collected (Table 6.3).

Geochemically the two subsets of type B(3a) could not be resolved and have been amalgamated on Fig 6.23 for simplification. In all plots (Figs 6.23 to 6.26) type B(3a) and B(3c) substantially overlap each other, particularly for the least mobile elements (Si, Zr and Nb). It is reasonable to assume, therefore, that all the microgranitoid inclusions sampled within the Ballachulish pluton are related and could have evolved from the same parent.

Consideration of Fig 6.24 shows a clear host pluton evolutionary trend of decreasing Zr with increasing SiO_2. Microgranitoid inclusions however have low Zr contents (50-200 ppm) at low silica (50-55%) perhaps indicating a source different from the host. This is by no means conclusive as fractionation of zircon beginning at the 57-58% silica level may well produce the observed trend. However no inclusions were found whose silica contents were between 55-59% SiO_2 and furthermore, no coherent trend towards the quartz diorite host is observed on the Nb vs. Zr plot (Fig 6.26) arguing against a host-inclusion relationship. It is concluded (tentatively in the absence of Nd isotopic data) that the sources for the microgranitoid inclusions and for the host were different.

Inclusion varieties within the Criffell pluton.

It was emphasised in Chapter 5 that the petrographic criteria for classifying inclusions within the Criffell pluton are not as clear as those within the Strontian or Ballachulish plutons, due to the excessive flattening of the inclusions. As a consequence it was expected that these types would overlap considerably on geochemical diagrams (Figs 6.27 to 6.30) due to structural promotion of equilibration. All, however, may be shown to define discrete fields on 'immobile' element plots.

Fine-grained inclusions may be sub-divided into two sets, C(3a:1) and C(3a:2), the discriminating petrographic feature being the tendency of hornblende to form aggregates in the latter. Geochemically the sub-groups have similar Zr (240-380 ppm) but have differing La (60-120 group 3a:1 and 100-295 for 3a:2), Ce (140-310 for 3a:1 and 200-530 for 3a:2), Nb (28-55 for 3a:1 against 10-22 for 3a:2) and Y (23-36 for 3a:1 against 7-23

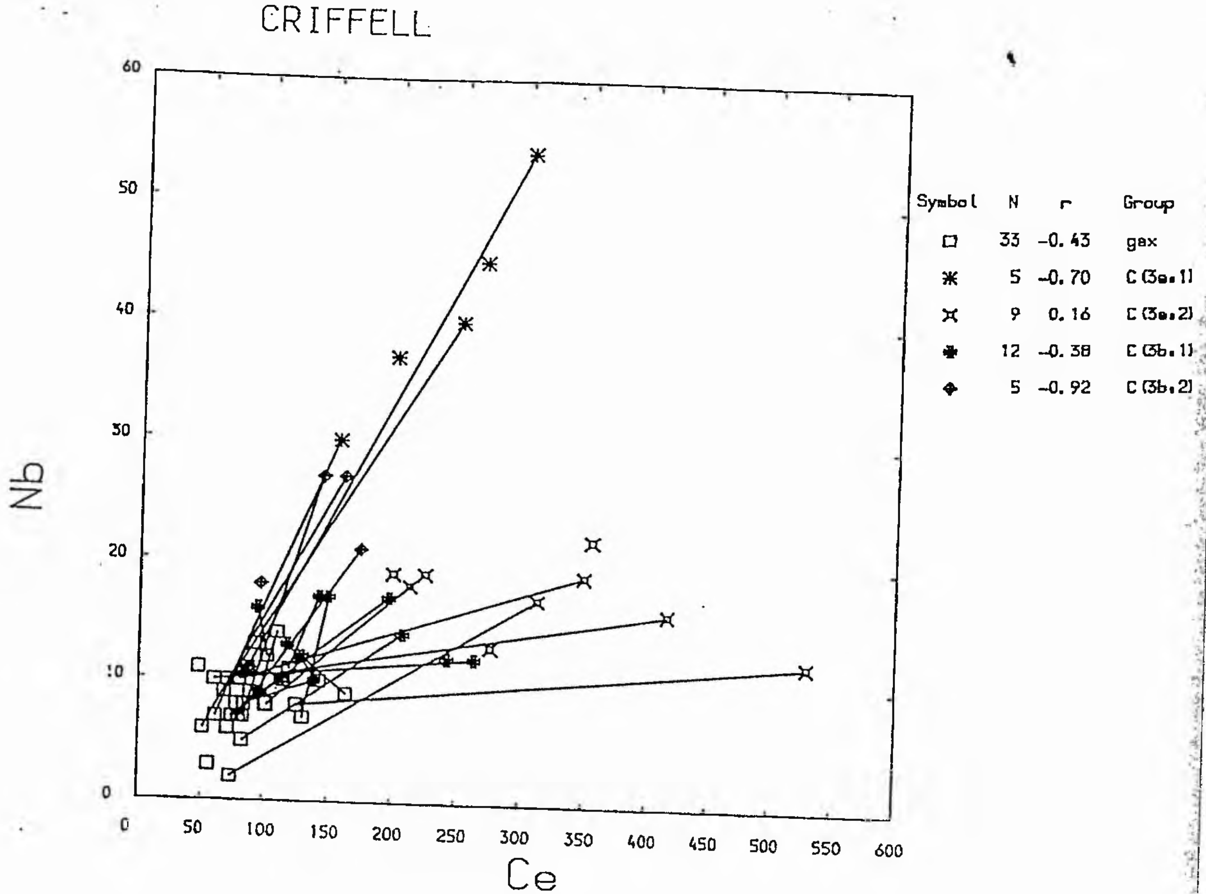


Fig 6.27 Nb against Ce.

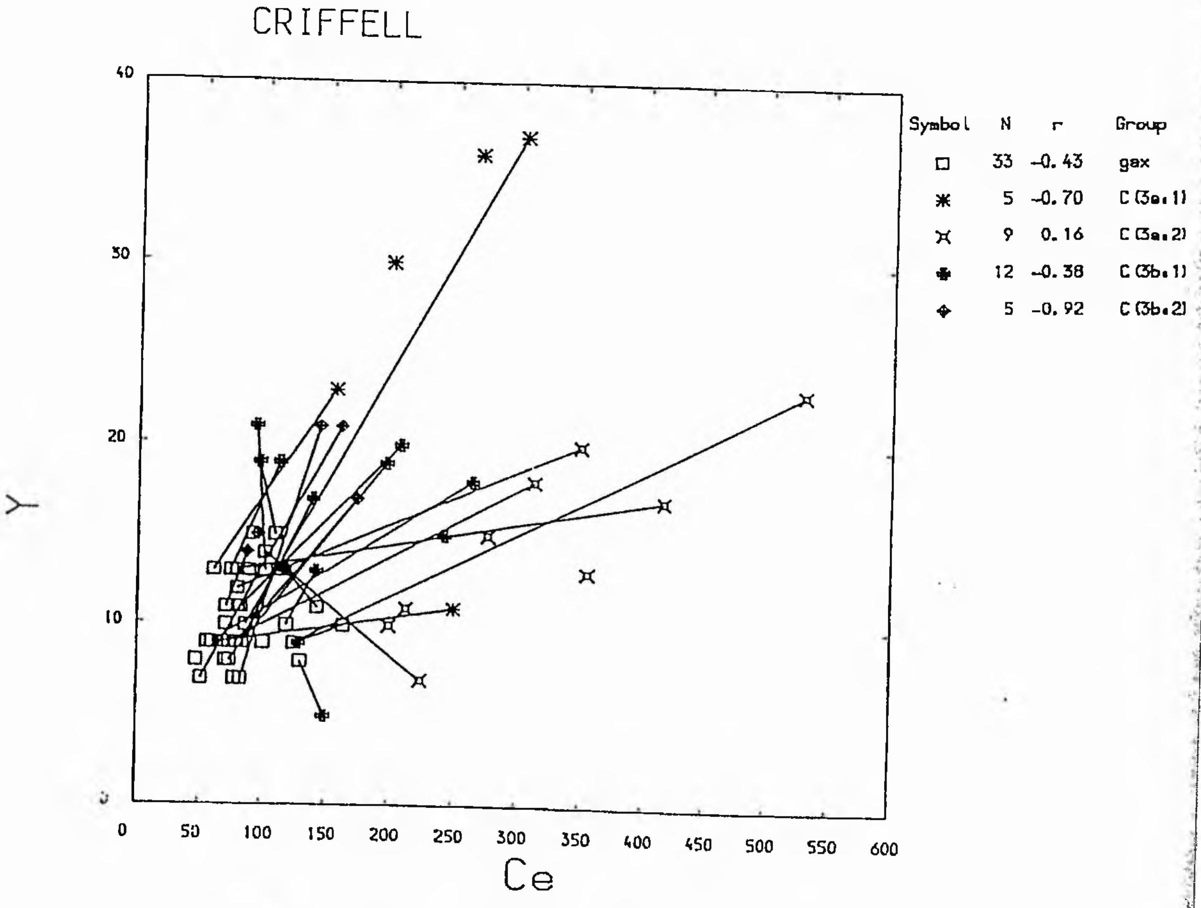


Fig 6.28 Y against Ce.

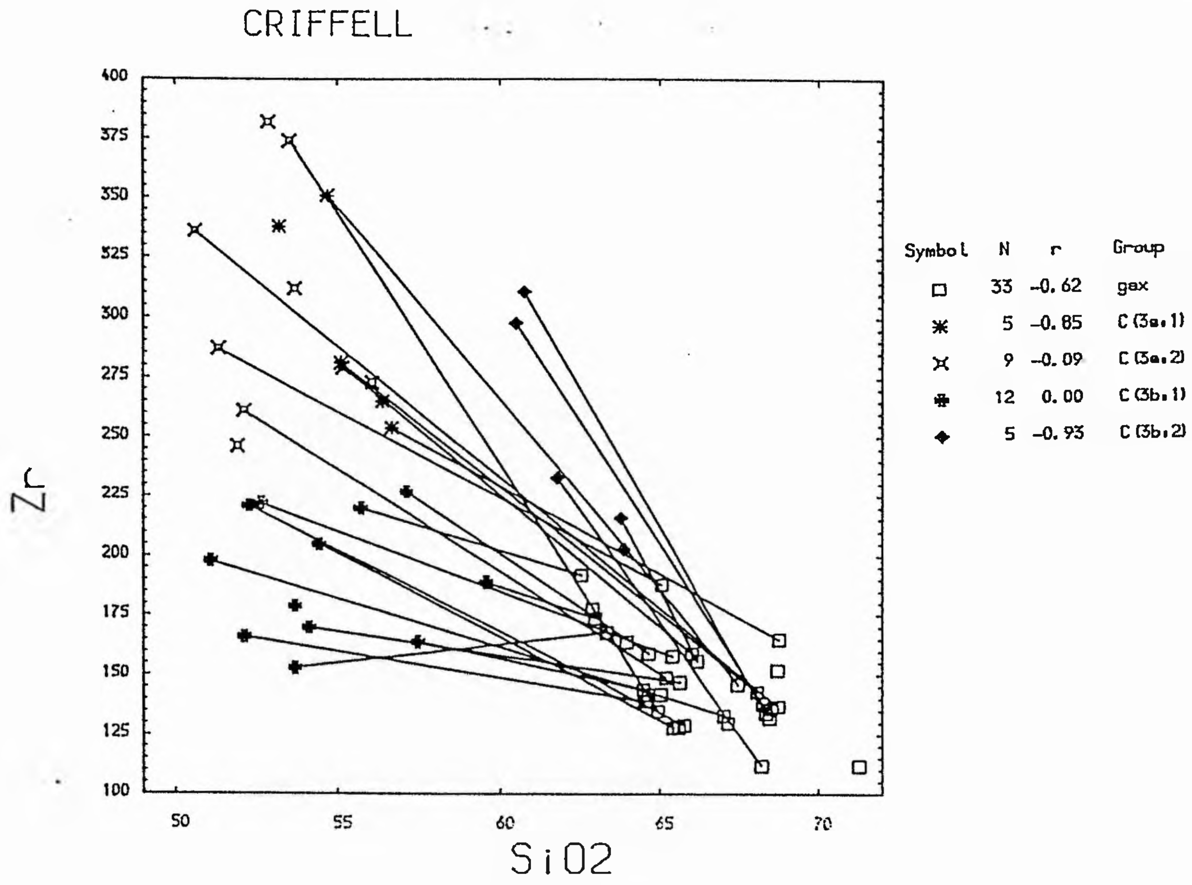


Fig 6.29 Zr against SiO_2.

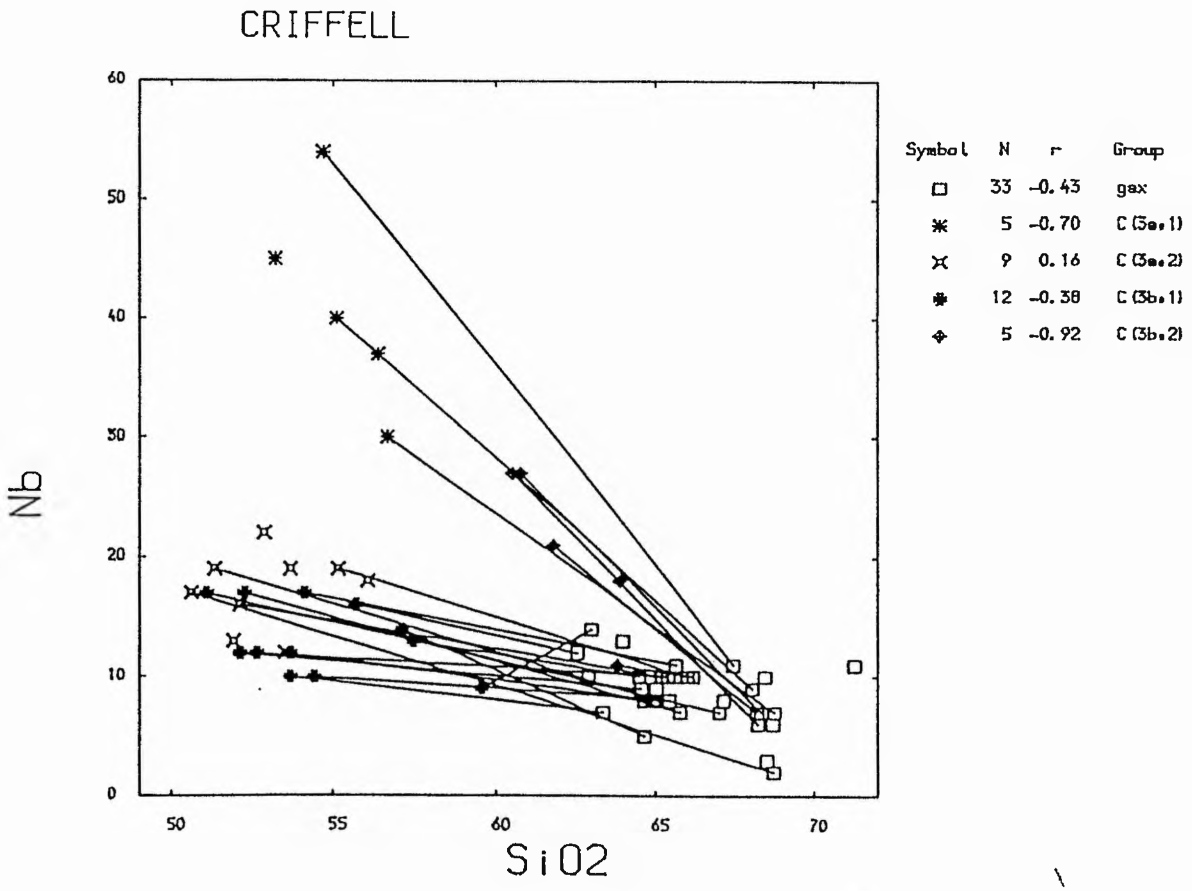


Fig 6.30 Nb against SiO_2.

for 3a:2) with the latter grouping being slightly poorer in SiO_2 (50-56% for 3a:2 and 53-57% for 3a:1).

Nd isotopes are distinctly different, ϵNd_t=+0.7 for 3a:1 and +1.9 for 3a:2, ruling out any genetic link by fractionation only. Further consideration of Fig 6.27 (Ce against Nb) confirms this view. Therefore, although the petrographic differences between the two sub-groups are small, geochemically they may be treated as the products of differing sources or as different degrees of crustal contamination.

Plagioclase megacryst bearing inclusions have also been divided into two sub-groups, that is C(3b:1) inclusions containing a significant fabric being found within the outer granodiorite and C(3b:2), unfoliated inclusions bearing distinctive flakes of biotite and having no appreciable fabric, the latter type being found exclusively within the inner pulse (2).

These two subsets have similar trace element concentrations (Figs 6.27 & 6.28) and overlap on trace element plots. However, the latter group (3b:2) has significantly higher SiO_2 concentrations (60-64% against 50-59% for group (3b:1) and tend towards higher Zr and Nb levels yielding entirely separate fields on Harker plots (Fig 6.29 & 6.30). Furthermore, trends for sub-group 3b:2 point away from the field of 3b:1 which argues against these sub-groups being related simply by processes of fractional crystallisation. Nd isotopic evidence is inconclusive with ϵNd_t values for 3b:1 and 3b:2 of -0.3 and +0.6 respectively, but does further suggest sources of slightly differing character.

However a representative of Class C(3b:1) (inclusion C323X), is the only inclusion analysed within the pluton which has equilibrated Nd isotopes. Whether this is because the inclusion is cogenetic with its host is unclear as it was collected from an area containing evidence of considerable flattening. It is tentatively suggested, therefore, that the two subsets were derived from sources which differ slightly in their isotopic and trace element signatures.

The graphical relationship between types C(3a:1) and C(3b:2) remains consistent on most plots, that is the type C(3b:2) always lies between type C(3a;1) inclusions and the host granitoids (Figs 6.27 to 6.30). In some instances a line drawn through the pluton evolution trend passes straight through 3b:2 and into 3a:1 without deviation suggesting a geochemical relationship (eg SiO_2 against Nb, Fig 6.30). Nd isotopic ratios for the two inclusion sets supports a fractionation relationship between inclusion type C(3a:1) and inclusion type C(3b:2); the ϵNd_t values are +0.7 and +0.6 respectively.

Careful consideration of Fig 6.31 (Harker plots of Ce, La, Y, Zr, TiO_2 and P_2O_5) indicates that an origin for the C(3b:2) inclusions by fractionation of a C(3a:1)-type magma composition is unlikely. In detail the trends for C(3a:1) inclusions follow a distinctly separate evolutionary path, particularly on the Zr and P_2O_5 Harker plot (Fig 6.31). Although the Nd isotopic evidence for these inclusion types indicates the possibility of a genetic link via fractional crystallisation, careful consideration of 'key' geochemical criteria (unaffected by equilibration

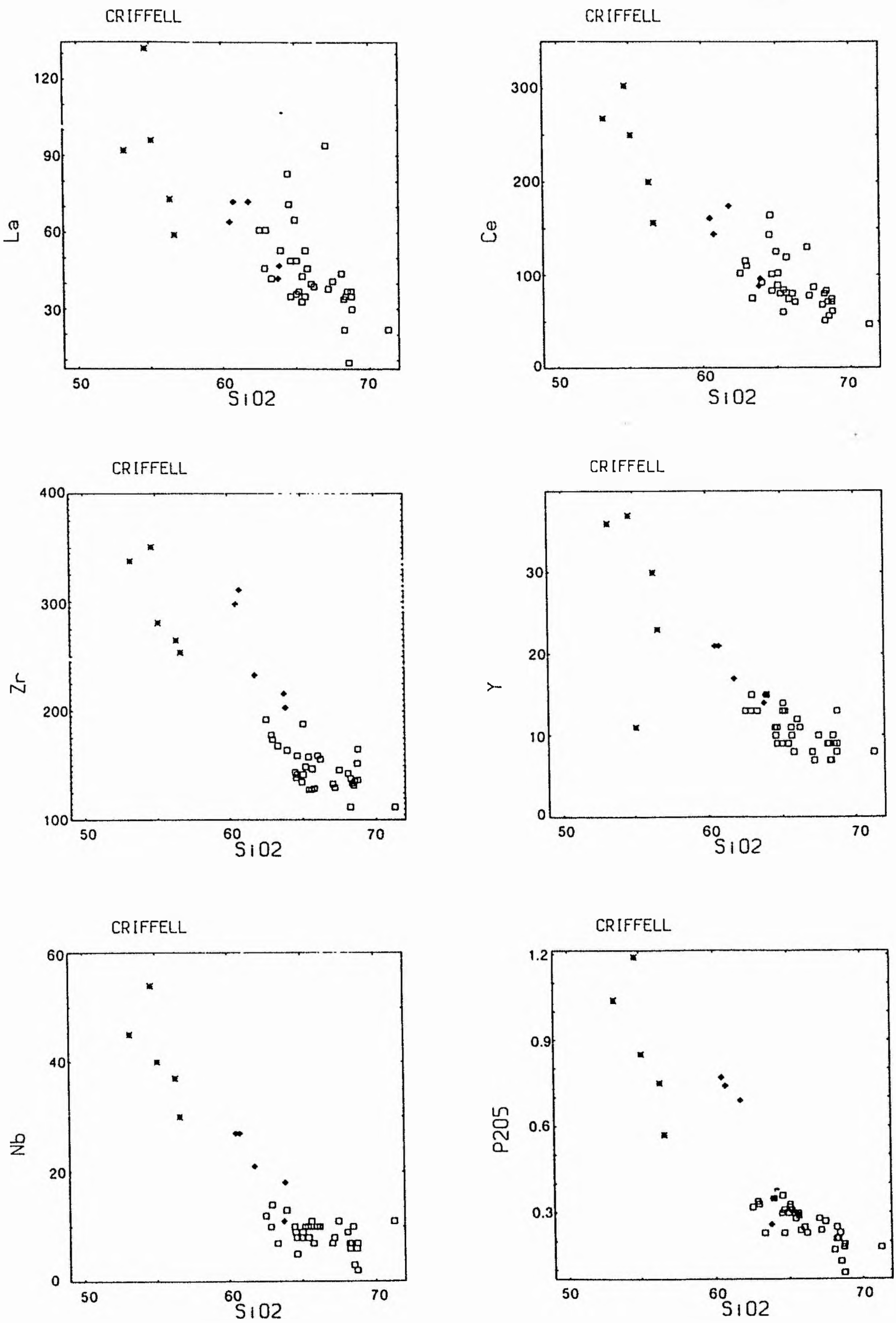


Fig 6.31 Composite Harker plots for Criffell inclusions C(3a:1) and C(3b:2) symbols as for Fig 6.27

processes) nullify this hypothesis. The conclusion to be drawn from this is that whilst Nd isotopes may retain a near pristine signature, the origin and diversity of inclusions cannot be determined by isotopic analysis alone.

The relationship of the C(3b:2) inclusions to their host remains clear, however, as a fractionation trend drawn through the host granitoid invariably passes through these inclusions (Fig 6.31). Petrographically they are similar to a suite of porphyrite dykes which cut pulse-1 but are seen to disappear in pulse-2 (Phillips 1956). Furthermore Nd isotopes for C(3b:2) inclusions are not dissimilar to the less evolved end of the host granitoid spectrum ($\epsilon Nd_t = +0.6$ for C(3b:2) inclusions and +0.4 for the most primitive granitoids). This subgrouping of inclusions may, therefore, be cogenetic with the host pluton but not simply by processes of fractional crystallisation.

In their analysis of geochemical and isotopic trends within the Criffell pluton, Halliday et al. (1980), Stephens & Halliday (1980) and Stephens et al. (1985) consider that the source for granitoid magma changed continually with time, later magmas having a greater proportion of crustal material (and hence less radiogenic Nd isotopes). The more primitive inclusions derived from the disaggregation of the porphyrite dykes would be expected to have Nd isotopes slightly more radiogenic than their more evolved granitoid counterparts. It is concluded, therefore, that the C(3b:2) inclusions are the remnants of disrupted porphyrite dykes intruded after the emplacement of pulse-1 and prior to pulse-2, and that these represent the early granitoid magma composition for pulse-2. Increases in crustal input combined with fractional crystallisation processes changed the isotopic and geochemical composition of the later more evolved granitoid magmas (Stephens et al. 1985).

Class C(3d) (felsic) inclusions are not common within the Criffell pluton. Geochemically these plot consistently with the host granitoids for all 'immobile' elements and it is suggested that the lack of acicular apatites within these inclusions and their generally coarse grainsize indicates that these are co-magmatic remnants entrained by the granitoid on its upward passage. However since no Nd isotopic determinations were carried out on these inclusions, this hypothesis remains unproven.

In summary

Inclusions within the high-K, calc-alkaline, Caledonian granitoid plutons of Strontian, Ballachulish and Criffell contain inclusion suites which show a greater range of geochemical and isotopic variability than their hosts. However whilst HFSE, REE and Nd isotopes remain disequilibrated between host-inclusion pairs, alkali-metals, alkali-earths (+ Pb) and Sr isotopes have substantially exchanged between host and inclusions (as was predicted earlier). The determination of the origin and diversity of inclusions within granitoid plutons cannot be reliably undertaken by the use of these latter 'mobile' elements and must be investigated using the relatively more 'immobile' HFSE, REE and Nd isotopes.

In detail inclusions may be subdivided into several petrographic groupings which cluster together on element-element diagrams. The existence of different clusters strongly suggests that inclusions were derived from more than one source. A few cogenetic autoliths have been identified from both the Strontian and Criffell plutons but for the majority the observation of distinctly more radiogenic Nd isotopes for microgranitoid inclusions with respect to their granitoid hosts is indicative of an exotic inclusion origin(s), in some cases indicative of a depleted mantle source.

CHAPTER 7

GEOCHEMISTRY AND SIGNIFICANCE OF THE APPINITES

7.1 INTERACTION OF APPINITES WITH HOST GRANITOID

In order to evaluate the relative importance of magma co-mingling, fractional crystallisation, country rock assimilation and volatile transfer in the generation of the reverse zonation pattern, a selection of samples representing the full range of textural variation was collected, the analyses of which are given in Appendix 2.

Interaction of appinite-BG with the biotite granite.

Compatible element variation.

Variation diagrams may be used to infer the relative importance of competing petrogenetic processes during the injection and cooling of magmas within the crust. Mixing on Harker diagrams yields straight lines connecting the two end-members, whilst the process of crystal-liquid fractionation may yield curves (Langmuir et al. 1977, Cox et al. 1979, Henderson 1983).

Consideration of Harker plots of elements Mg, Ni and Zn (Figs 7.1 to 7.3) show that with the exception of two core region samples (S444 & S327) the remaining appinite samples lie on a straight line terminating at high SiO_2 biotite granite compositions. Such a correlation could be achieved either by mixing of the most basic appinite with biotite granite or by fractional crystallisation of ferromagnesian minerals, notably amphibole (Henderson 1983). Compatible element concentrations within samples from the appinite core region are high compared

Symbols for all appinites.

□ Host pluton

× Marginal zone

▽ Intermediate zone

✳ Core region

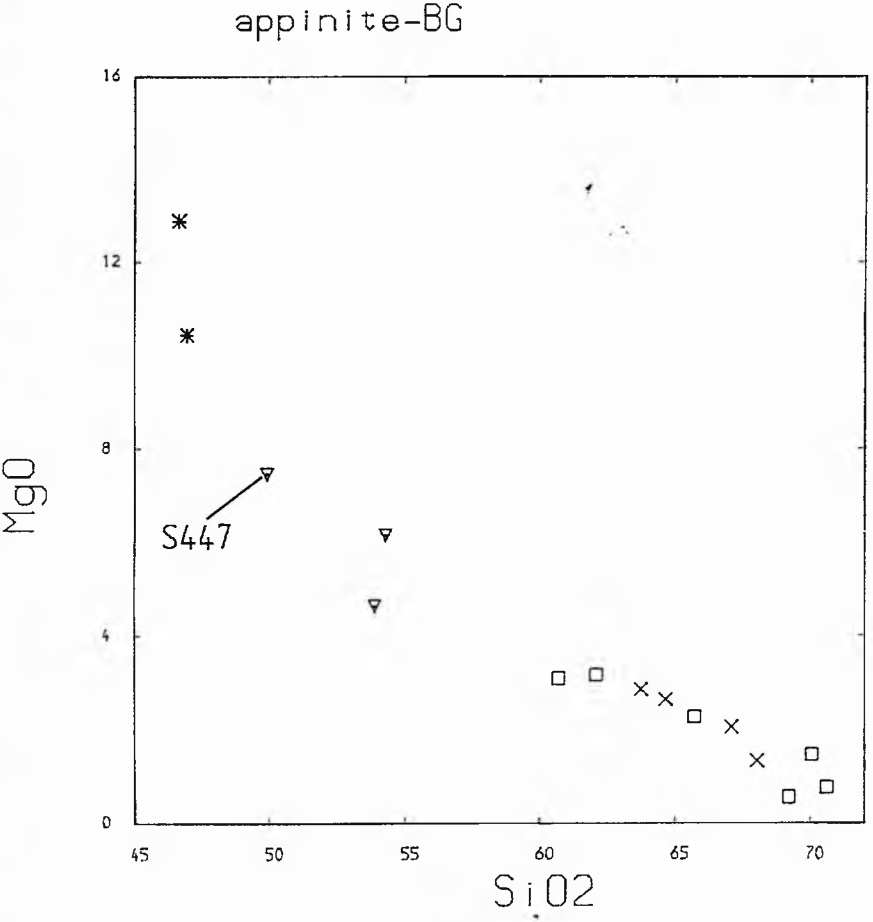


Fig 7.1 MgO against SiO_2.

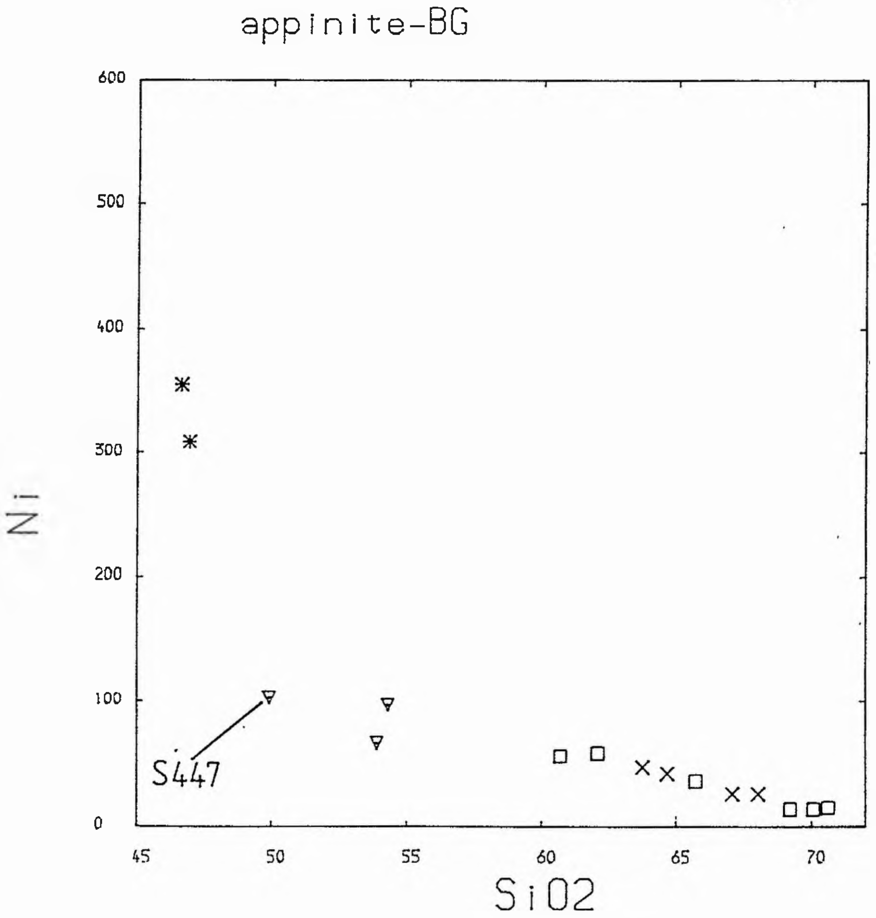


Fig 7.2 Ni against SiO_2.

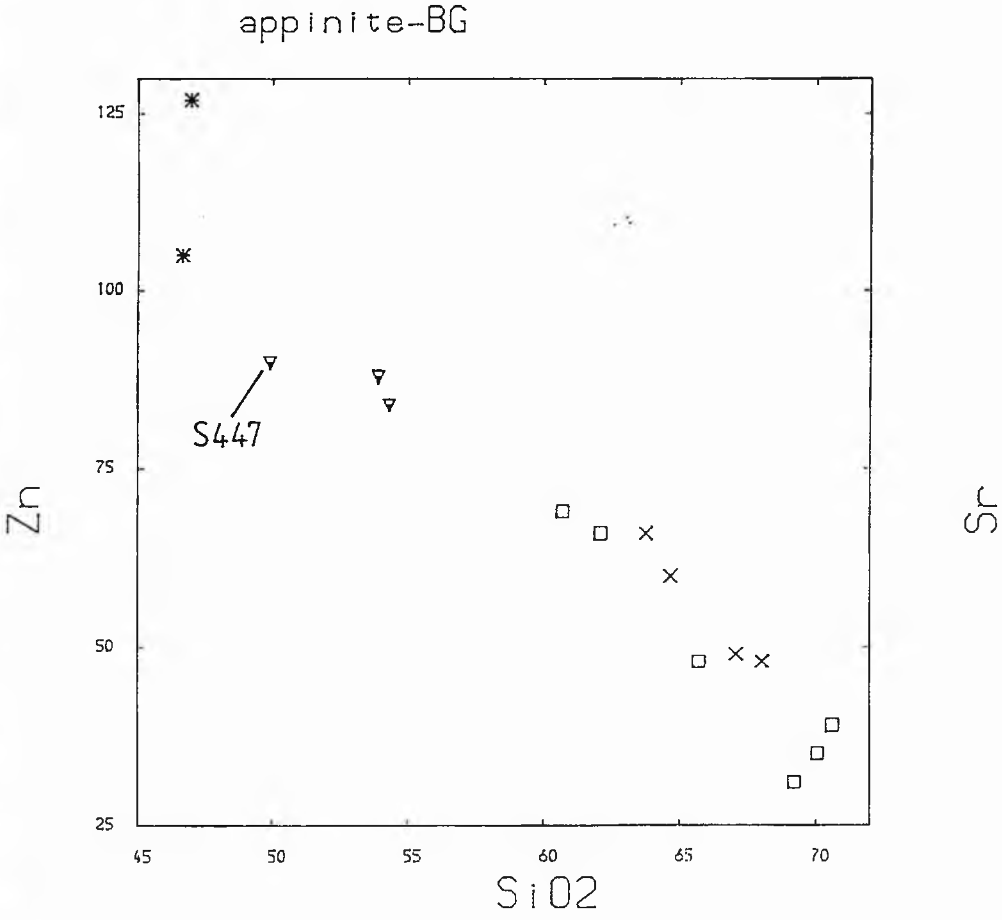


Fig 7.3 Zn against SiO_2.

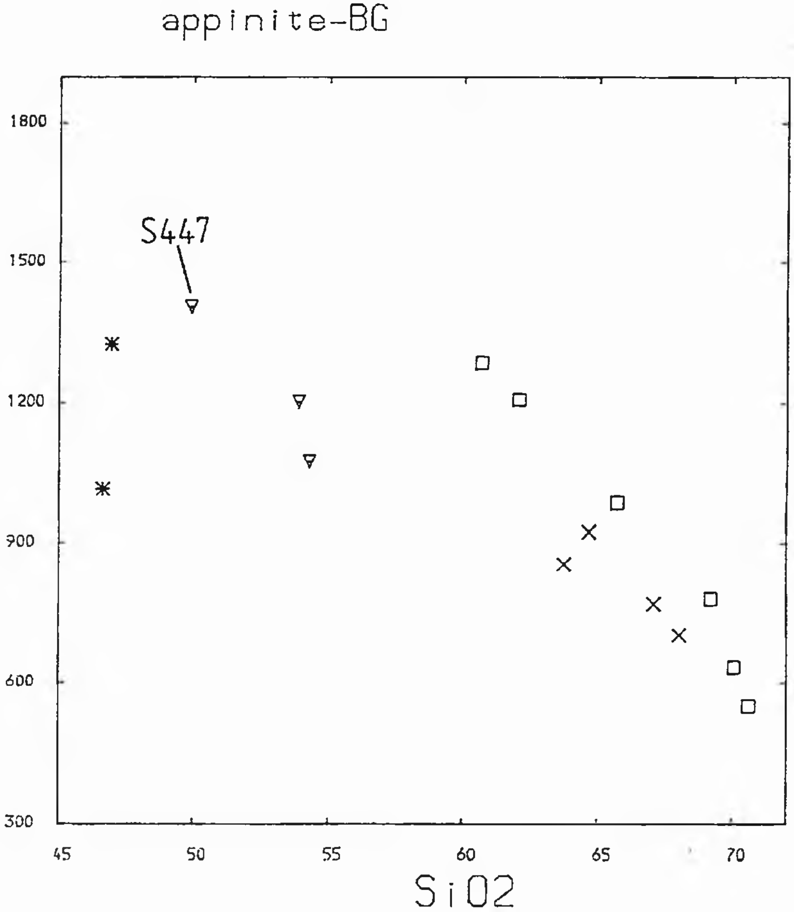


Fig 7.4 Sr against SiO_2.

with contemporaneous basic volcanics (e.g. the Lorne lavas; Thirlwall 1982). Nockolds (1941) considered that similar highly compatible enriched rocks from the Garabal Hill-Glen Fyne complex were the accumulated crystalline product of fractional crystallisation of a more acid magma.

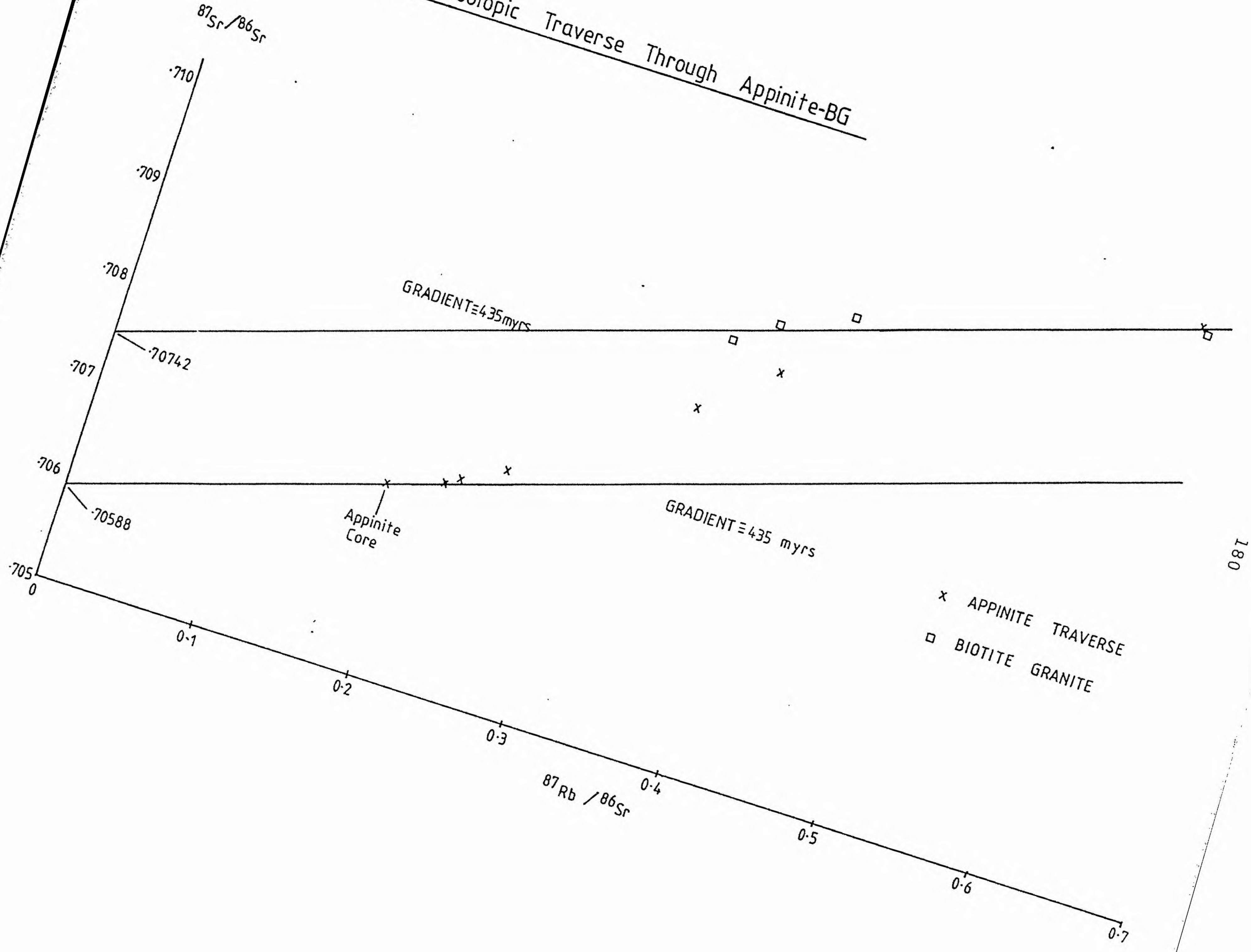
The presence of rapid growth textures in the vicinity of the appinite-biotite granite contact indicates that the appinite was in a mostly liquid state when emplaced. In identifying the core region (i.e. the most basic rocks) as having a chemistry dominated by ferromagnesian mineral accumulation, the most likely appinite liquid composition is that of the least evolved sample lying on the appinite-biotite granite trend (i.e. sample S447).

Incompatible element variation.

Like the ferromagnesian mineral compatible elements, the ferromagnesian mineral incompatible elements (e.g. Sr, Fig 7.4) do not yield definitive evidence for the dominant zonation mechanism. The clear break of slope at S447 (Fig 7.4) does indicate that at least two assemblages of minerals must have crystallised in order to produce the kinked curve. Alternatively, a combined mechanism of fractionation/accumulation of an assemblage rich in ferromagnesian minerals producing the relatively low SiO_2 core samples, followed by mixing of the resultant magma composition with Biotite Granite magma, could have produced a trend similar to that in Fig 7.4.

Nature of mixing on the Sr isotope diagram.

Fig 7.5 Isotopic Traverse Through Appinite-BG



On a ratio plot using the same denominator (eg X/Y vs Z/Y) simple mixing yields a straight line between two end-members (Langmuir et al. 1977). However, consideration of the strontium 'isochron' diagram (Fig 7.5) shows that a straight line cannot be drawn through the two end-members, indicating either that mixing was complex or that the selected end-members were unrepresentative.

Two parallel lines may be constructed through the upper and lower data points, whose slopes correspond to an 'age' of 435Myr (the U-Pb age of the pluton, Pidgeon & Aftalion 1979), with initial ratios of .7057 and .7071 respectively (Fig 7.5). Samples analysed from along the traverse define a straight line between the two bounding lines yielding a pseudo-isochron 'age' of 840Myr. This 'age' may be interpreted as a result of mixing between two distinct isotopic reservoirs and has no absolute time significance (Faure 1977).

Samples S444 and S327 (from the core region) maintain a separate Sr evolutionary trend from the other samples and cannot contain appreciable amounts of strontium from the biotite granite. S443 deviates only slightly from the appinite evolution line and as such can contain only minor Sr from the biotite granite. The mixing curve is dominated by samples taken from just within the biotite granite which contain hornblendes similar to those within the appinite. Thus 'mixing' with biotite granite can account for much of the variation at the periphery of the appinite, but this does not account for all the variation observed within the appinite body.

In summary, mineralogical and chemical variation within appinite-BG may be described in a two-fold manner. Firstly variation within the heterogeneous peripheral zones of the body are dominated by 'mixing' processes with the surrounding biotite granite. Secondly, variation within the largely homogeneous core region was dominated by crystal accumulation processes from the fractionation of the appinite liquid.

The fact that the cumulitic material lies at the centre of the pipe suggests that fractionation took place prior to emplacement as in-situ crystal-liquid separation would be expected to generate a normally zoned body. A unifying model is presented later in the chapter.

Interaction of appinites-PG, -T2 and -T1 with their hosts.

Appinite-PG.

Compatible element variation.

Harker plots for compatible elements Mg, Ca, Ni and Zn , show great scatter (Figs 7.6 to 7.9) and do not conform to simple co-mingling or crystal fractionation criteria. High compatible element concentrations are generally retained as silica increases yielding a sub-horizontal trend. The melt composition S427 has generally lower MgO and Ni than the rest of the appinites but has similar CaO and Zn concentrations. Direct mixing of S427 with granodiorite compositions cannot generate the observed trends. Fractional crystallisation, on the other hand, may explain the low silica-high compatible element group, but cannot explain the

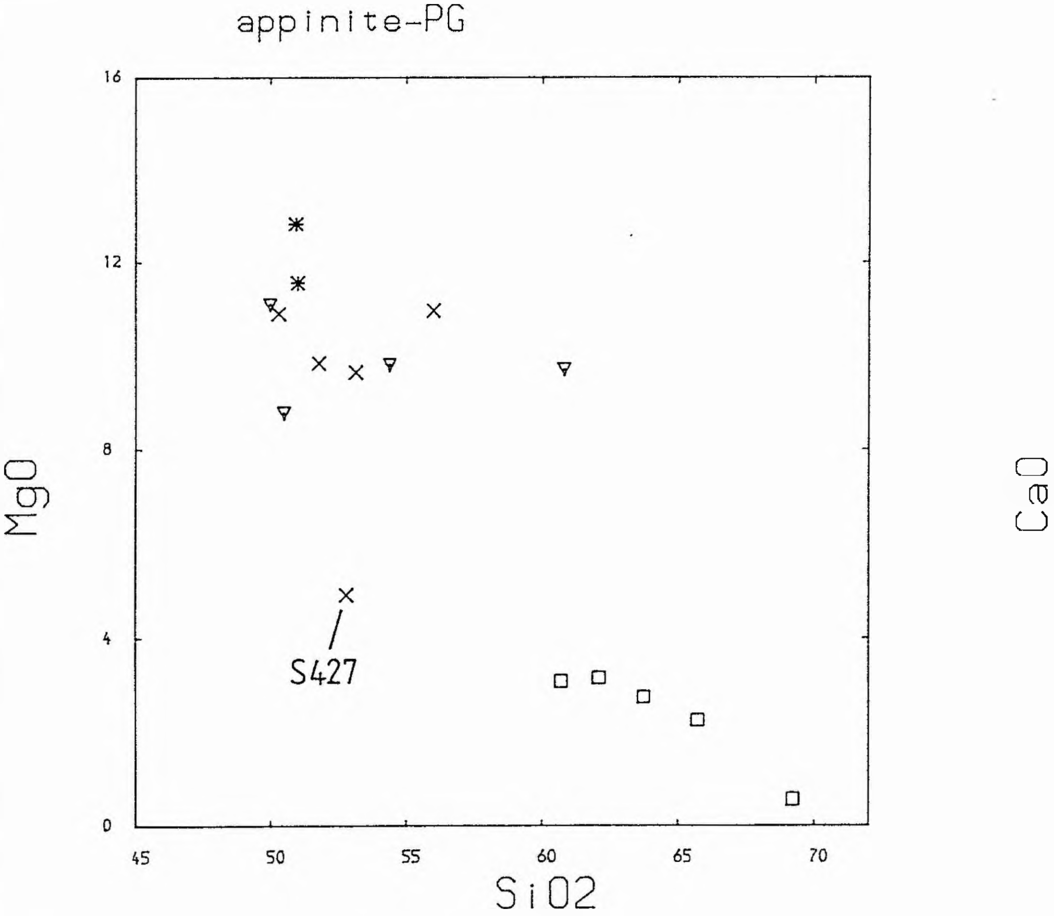


Fig 7.6 MgO against SiO_2.

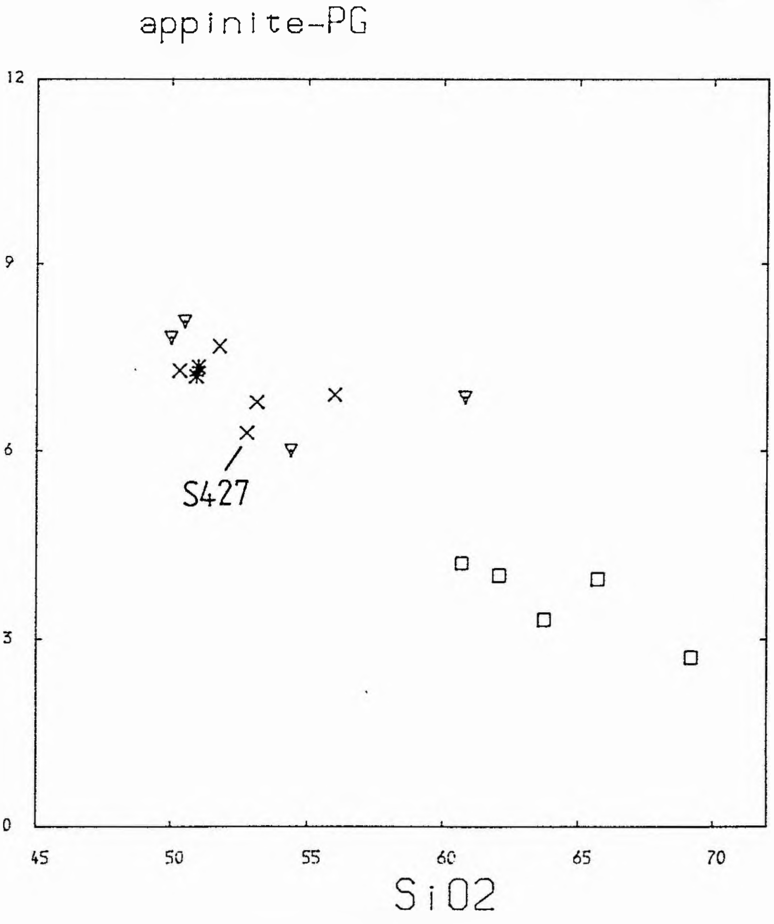


Fig 7.7. CaO against SiO_2.

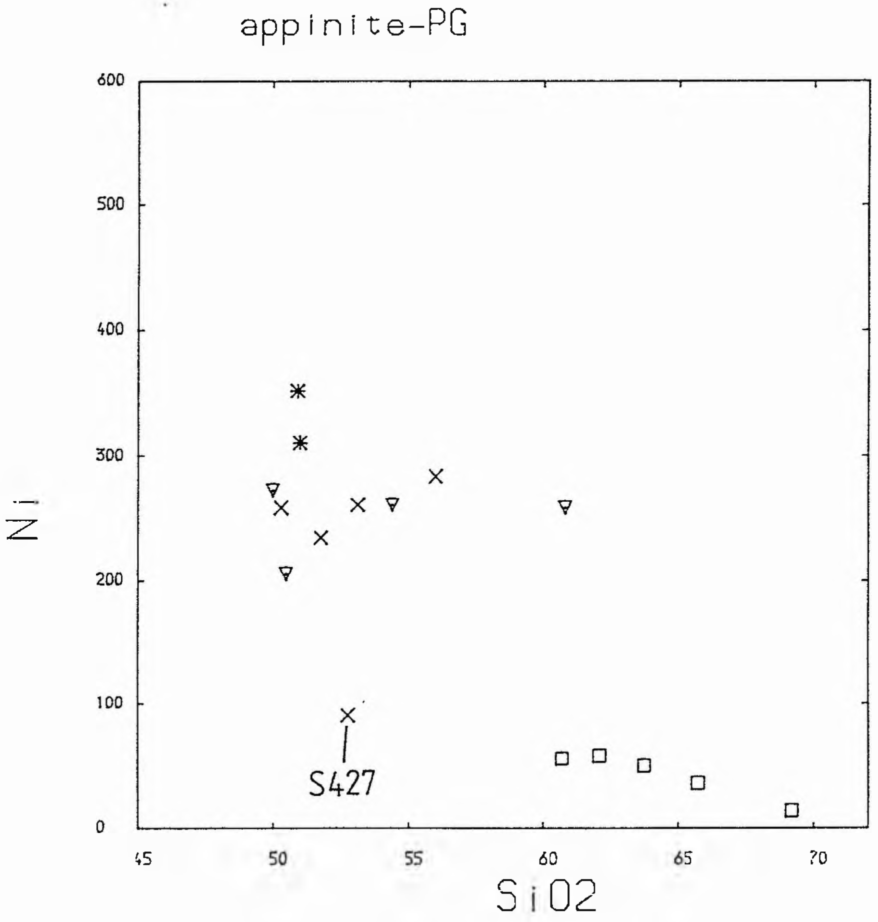


Fig 7.8 Ni against SiO_2.

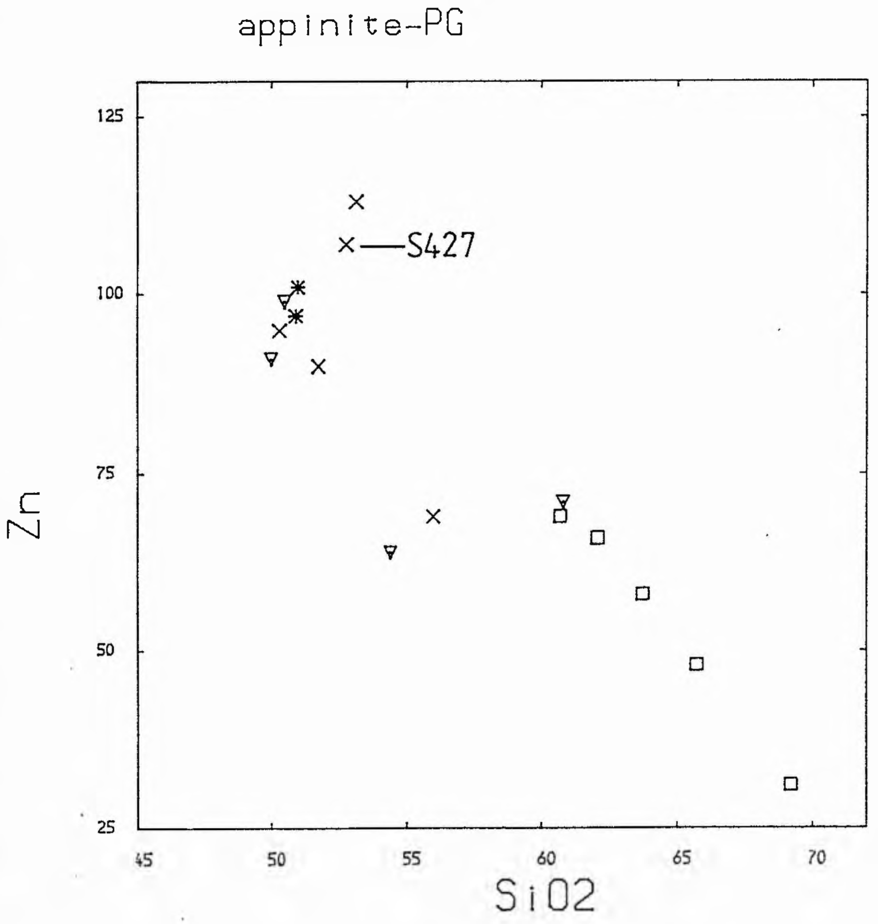


Fig 7.9 Zn against SiO_2.

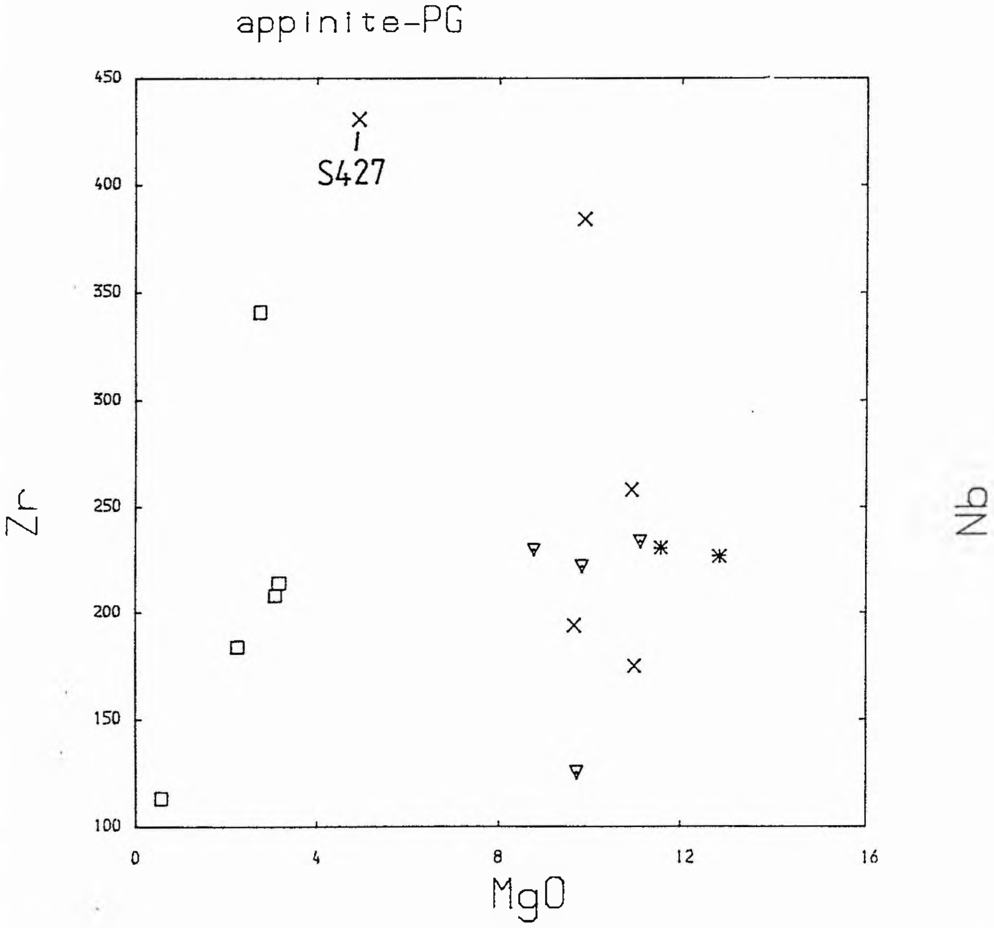


Fig 7.10 Zr against SiO_2.

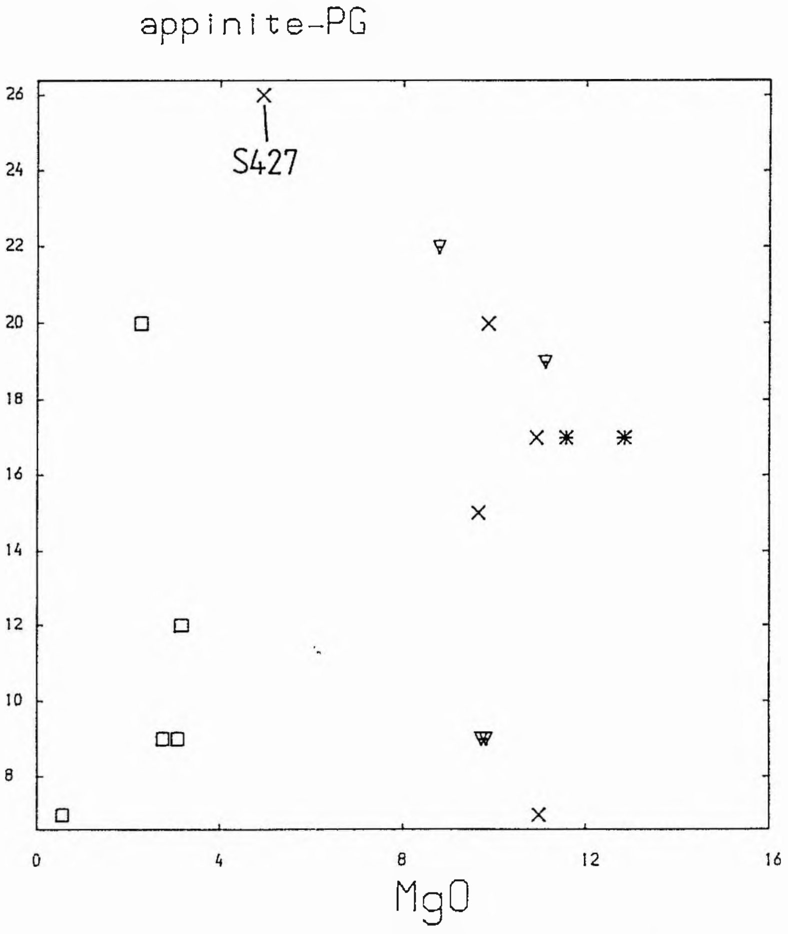


Fig 7.11 Nb against MgO.

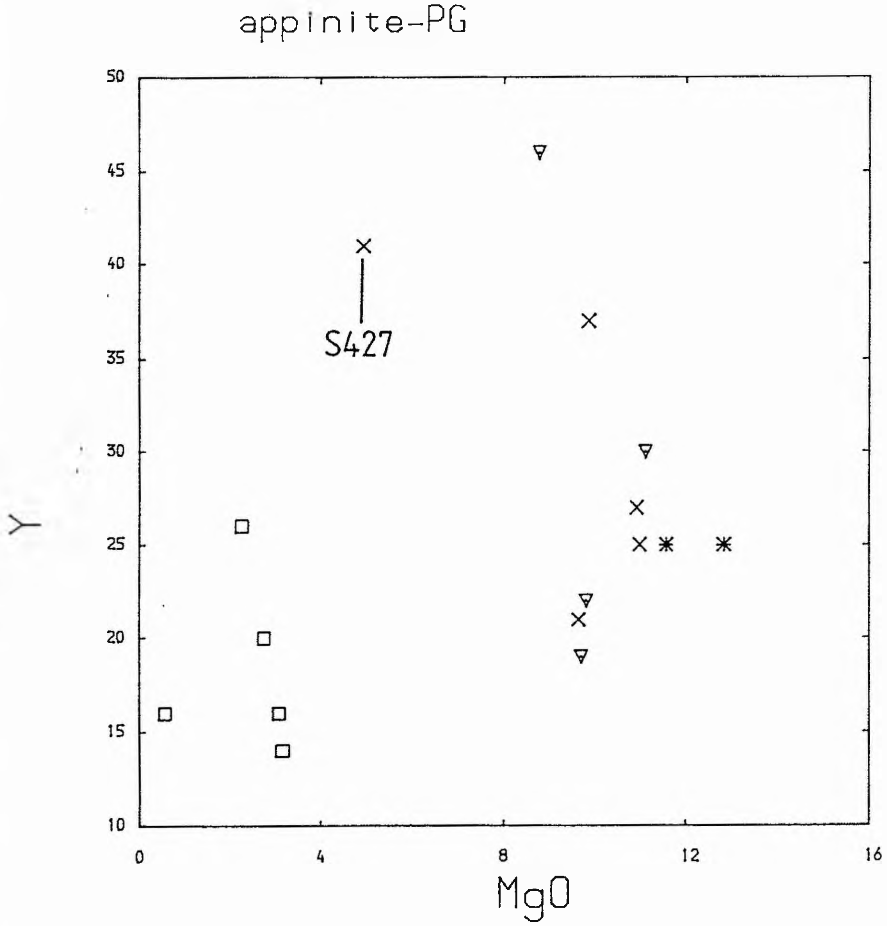


Fig 7.12 Y against MgO.

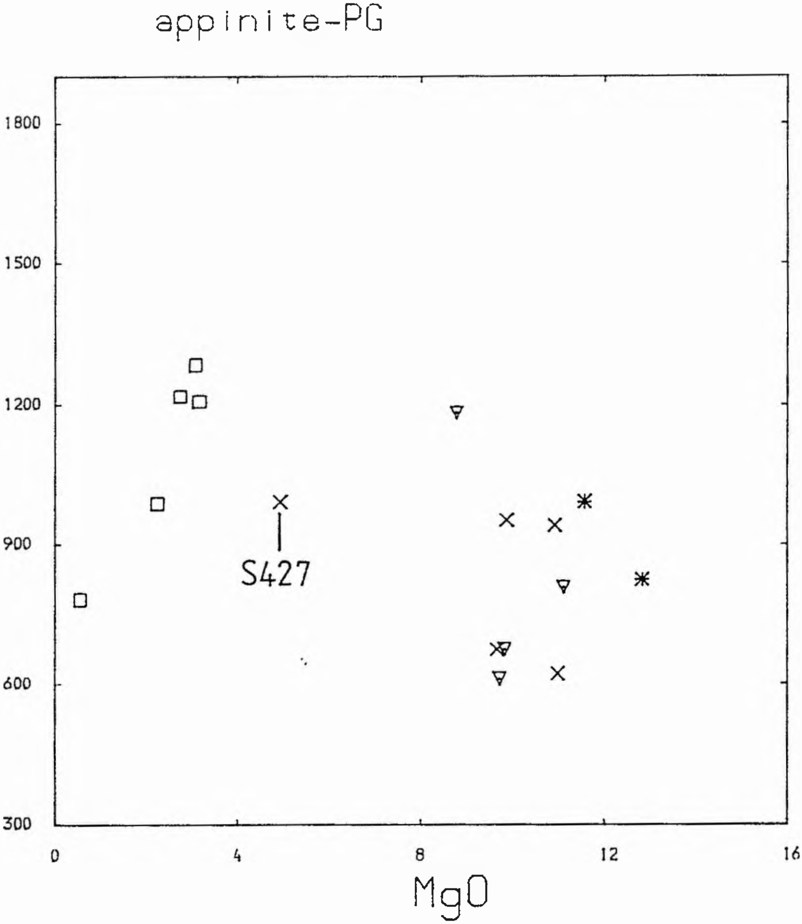


Fig 7.13 Sr against MgO.

high compatible element-high silica group, as fractional crystallisation would be expected to drive the residual liquid to low compatible-high silica values. Thus the compatible element variation diagrams yield no clear evidence for the major processes for zonation of appinite-PG.

Incompatible element variation.

Diagrams of incompatible elements Zr, Nb, Y and Sr plotted against MgO show even greater scatter than the compatible elements (Figs 7.10 to 7.13), with large variations occurring at near constant MgO values. Clearly co-mingling of granodioritic and appinitic liquids cannot alone be responsible for the large scatter. Furthermore, fractionation of ferromagnesian minerals from a liquid composition similar to S427 cannot explain the high variability of the incompatible elements. Such variations could have been the product of either fractionation/accumulation of incompatible rich accessory phases or pervasive metasomatism by late, incompatible element rich fluids.

Appinite-T2

Compatible element variation.

Harker diagrams of MgO, CaO and Ni (Figs 7.14 to 7.16) show that the coarse grained samples collected from within the core region plot off to higher compatible element contents than would be predicted by their SiO_2 levels. In this respect appinite-T2 is similar to appinite-BG and the central core material is likewise considered to be cumulitic. Furthermore a liquid composition for

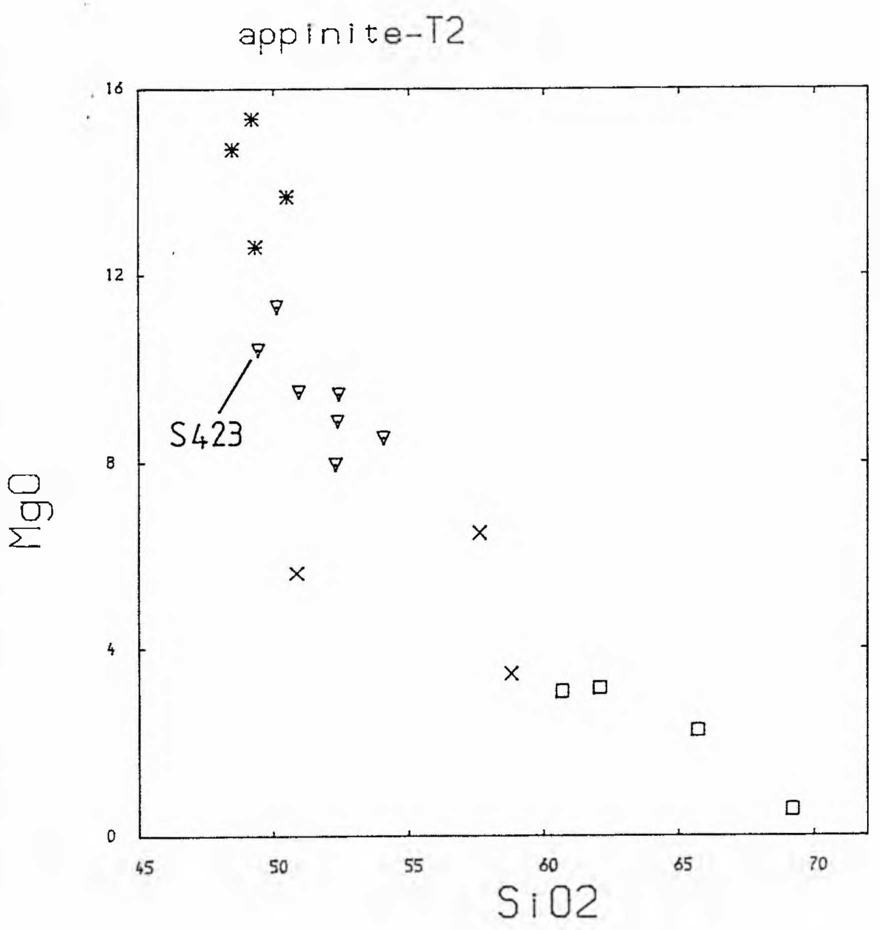


Fig 7.14 MgO against SiO_2.

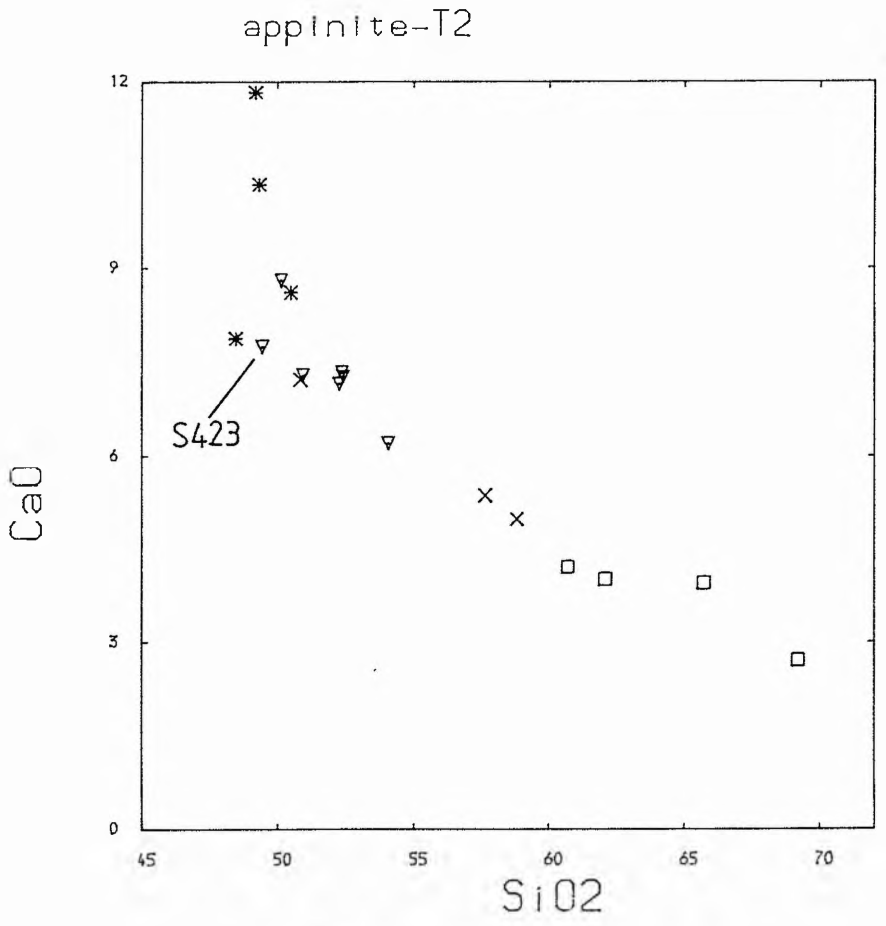


Fig 7.15 CaO against SiO_2.

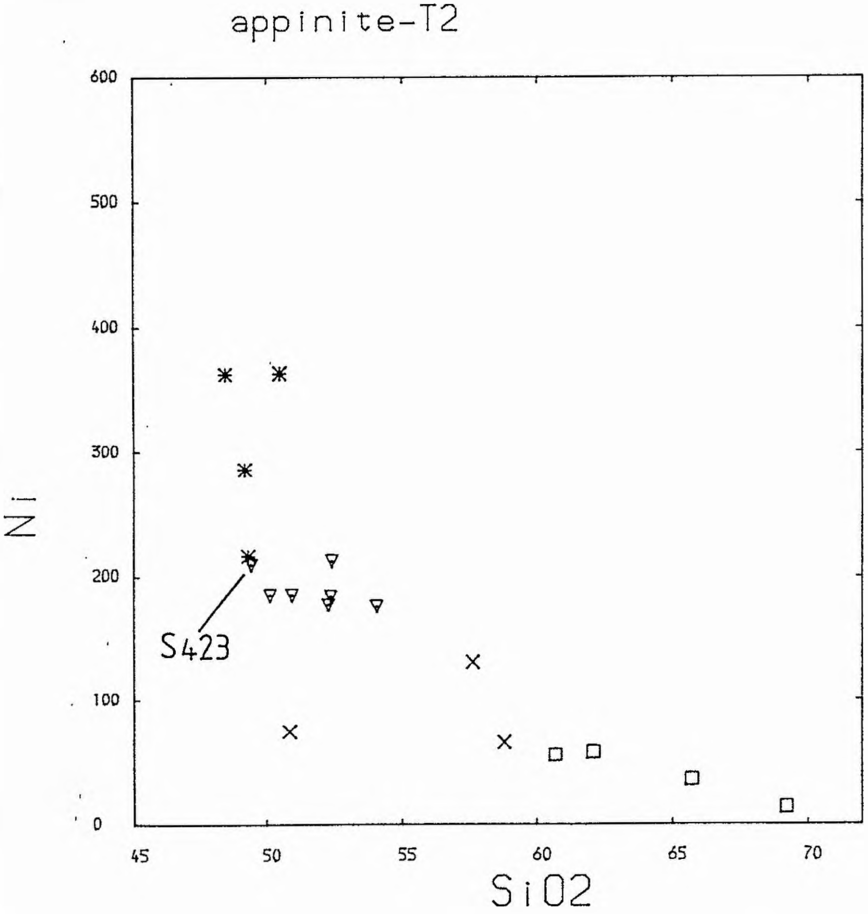


Fig 7.16 Ni against SiO_2.

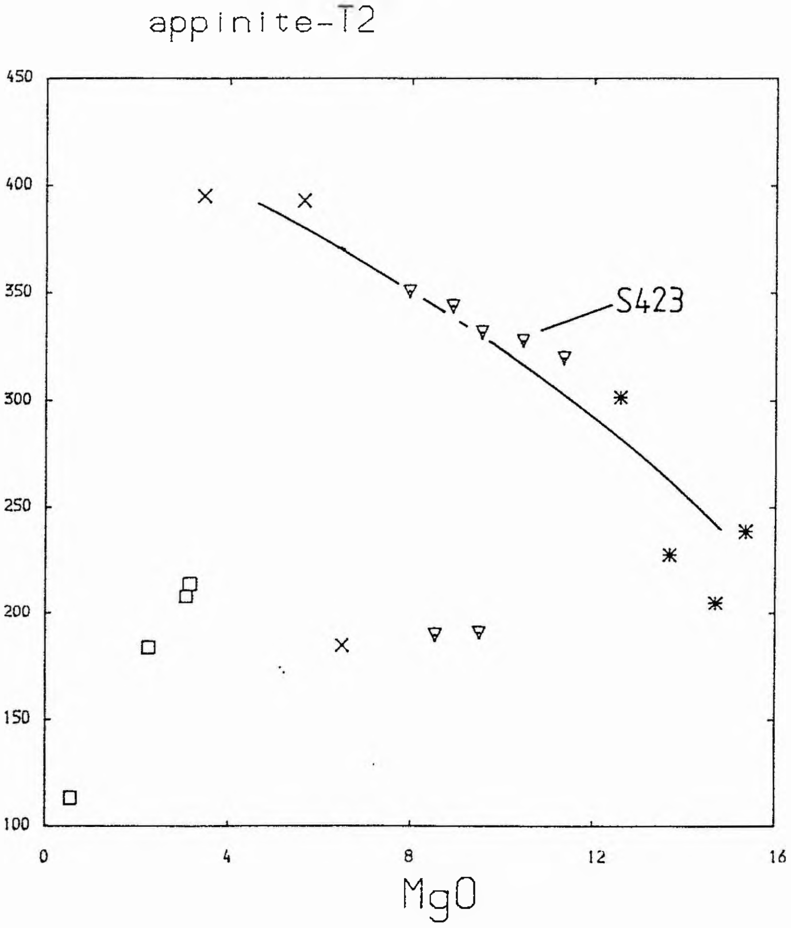


Fig 7.17 Zr against MgO.

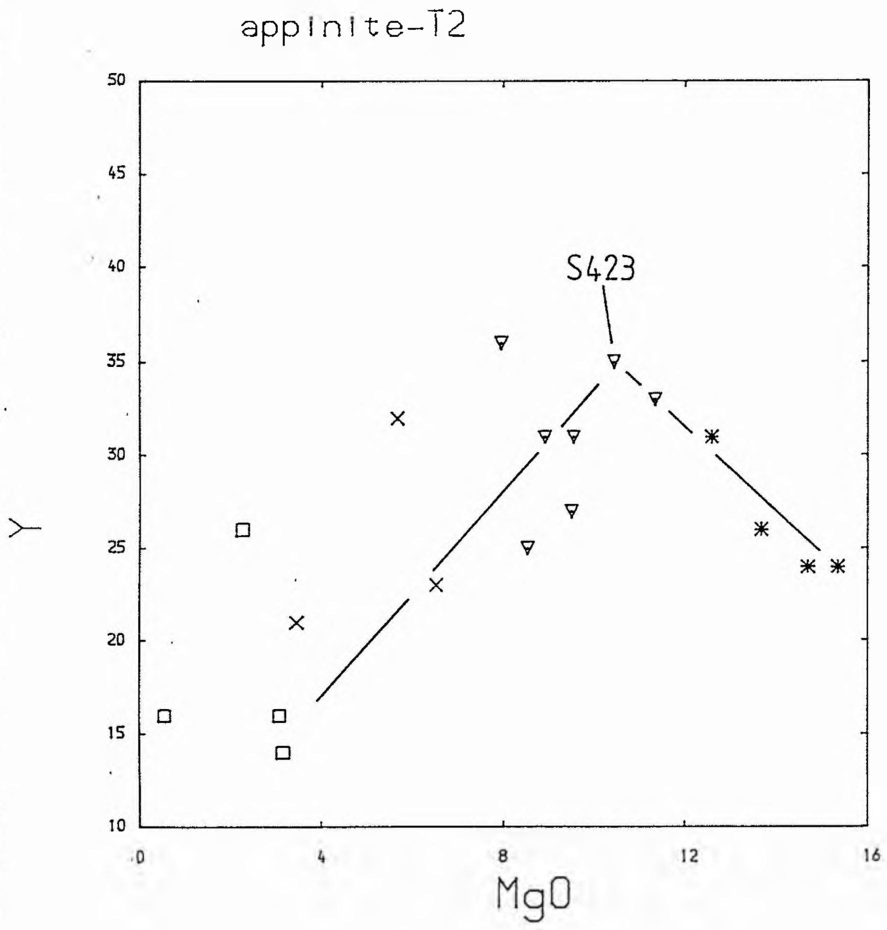


Fig 7.18 Y against MgO.

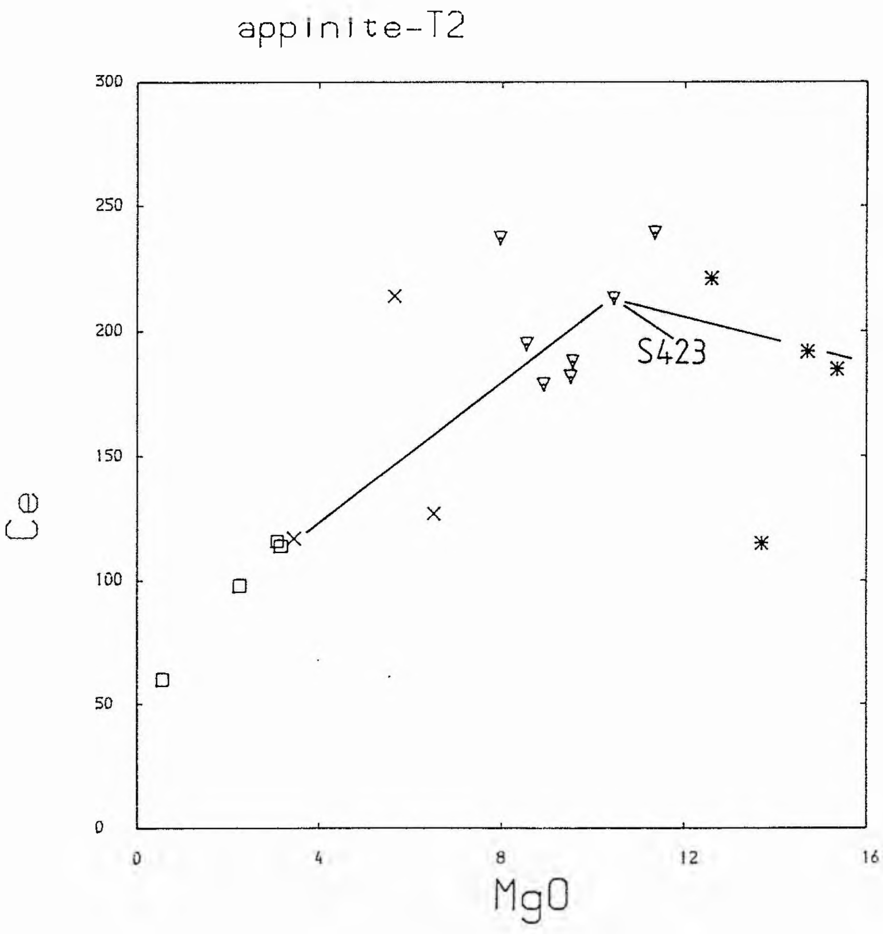


Fig 7.19 Ce against MgO.

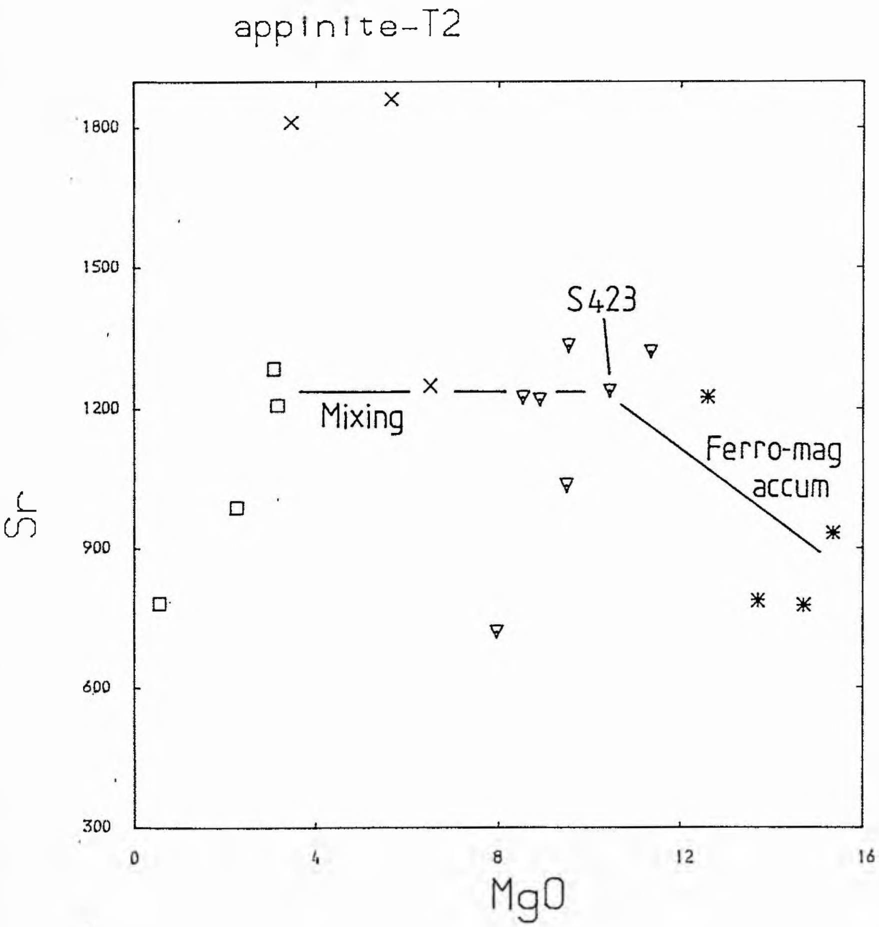


Fig 7.20 Sr against MgO.

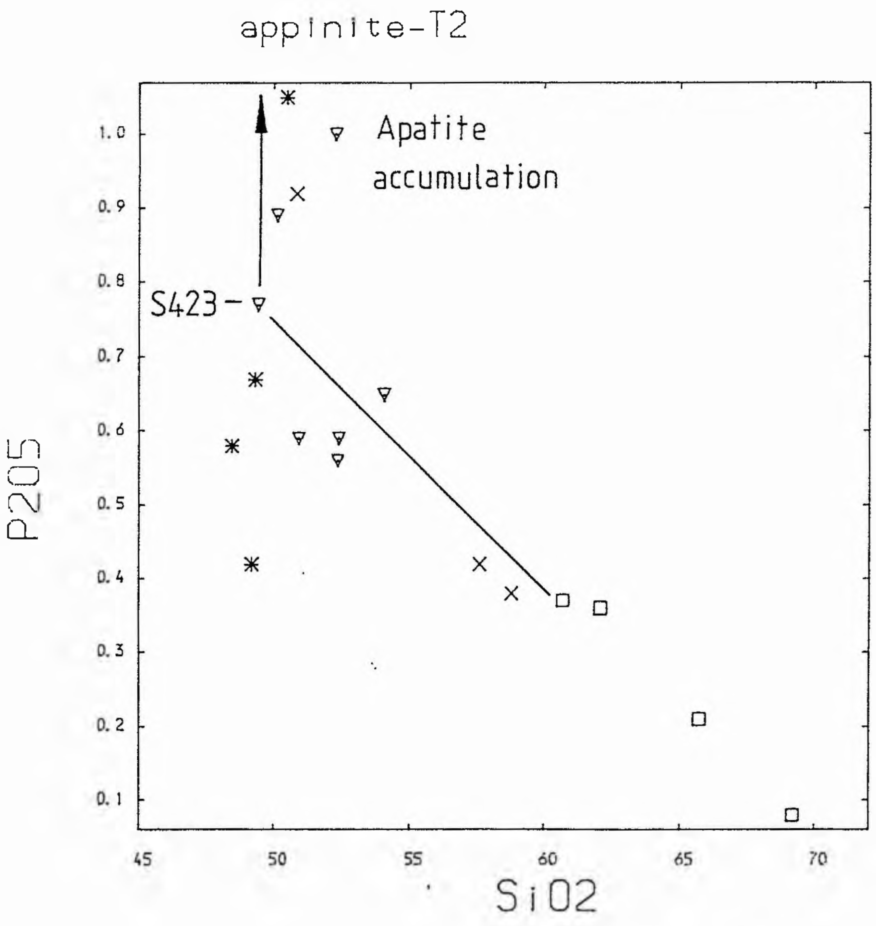


Fig 7.21 P_2O_5 against SiO_2.

appinite-T2 at the time of its emplacement may be estimated and in this this case is sample S423. Consideration of the appinite-tonalite trend for compatible elements (Fig 7.14 to 7.16), suggests more than one process was in operation around the periphery of the body.

Ferromagnesian incompatible elements.

Plots of incompatible elements Zr, Y, Ce and Sr against MgO (Figs 7.17 to 7.20) do not yield consistent trends. For instance Y against MgO (Fig 7.18) shows that intermediate and marginal appinite samples trend towards tonalitic compositions on a rough mixing line. Zr against MgO (Fig 7.17) is better described in terms of ferromagnesian mineral separation with no tendency for this trend to point towards host tonalite. Plots of Sr and Ce against MgO lie somewhere between with seeming evidence for both ferromagnesian mineral separation and appinite tonalite co-mingling.

It is worthy of note that there is a great deal of scatter on these diagrams, suggesting that more than one process worked simultaneously during appinite evolution. Furthermore Zr, Y and Ce concentrations are strongly controlled by minor phases such as zircon, sphene and apatite (Henderson 1980, but see also Gromet & Silver 1983). A plot of P_2O_5 against SiO_2 demonstrates the scatter brought about by minor phases (Fig 7.21) as P_2O_5 concentrations are strongly controlled by apatite.

Appinite-T1

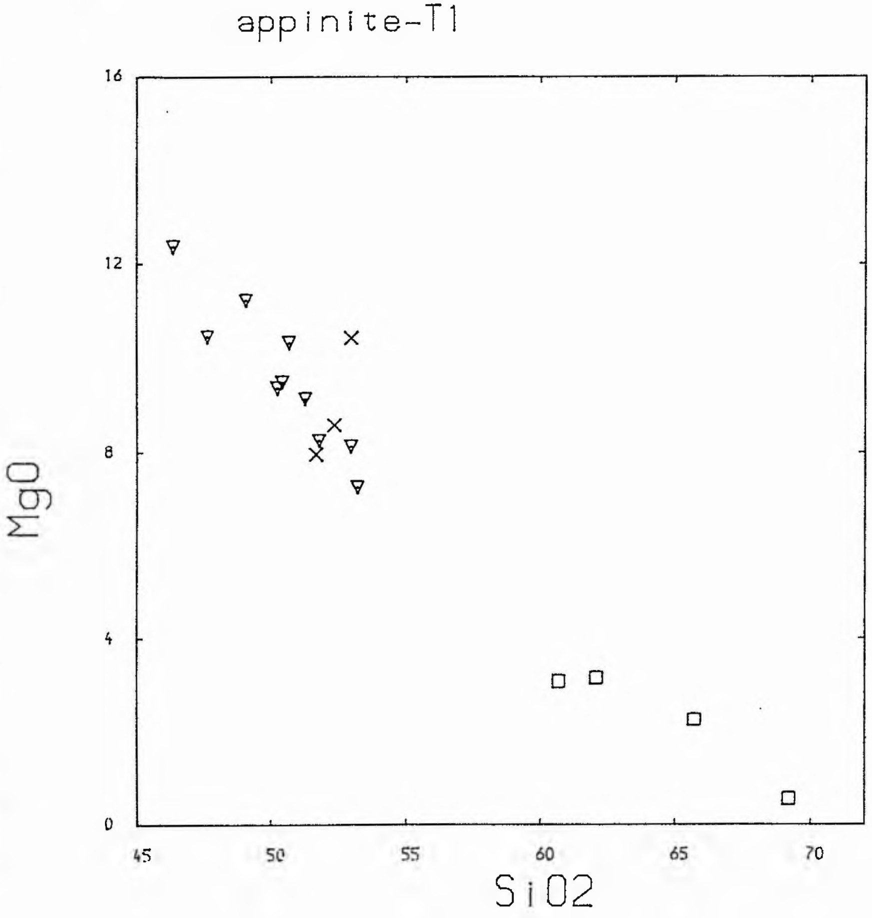


Fig 7.22 MgO against SiO_2.

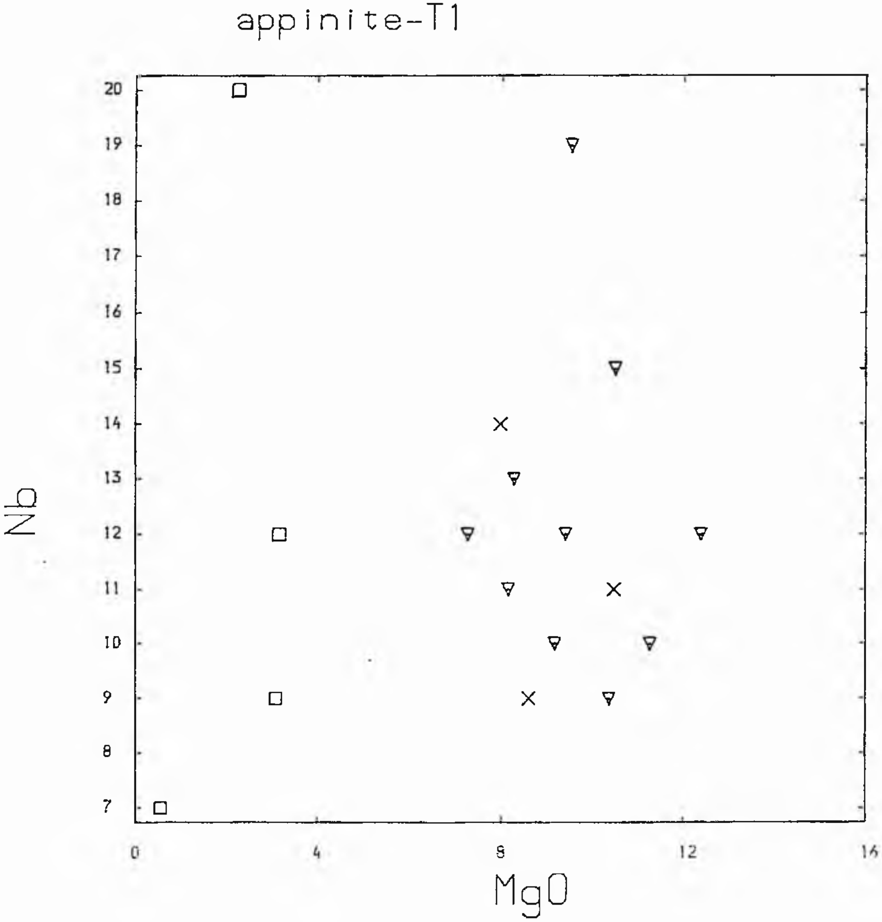


Fig 7.23 Nb against MgO.

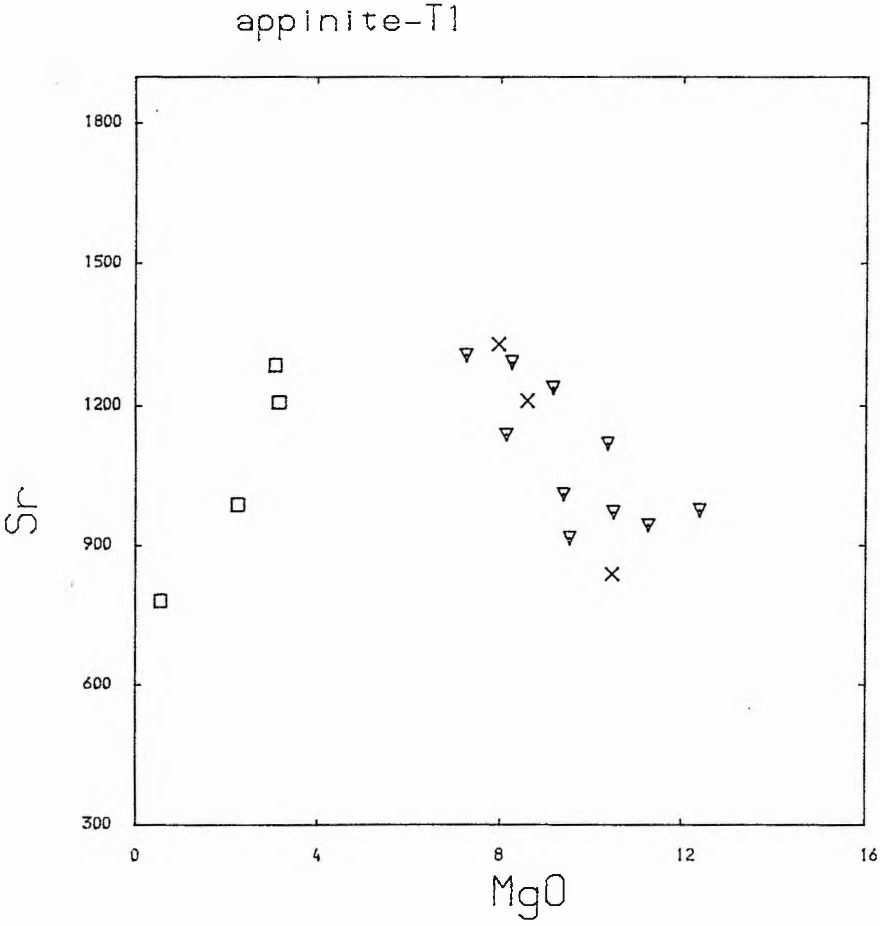


Fig 7.24 Sr against MgO.

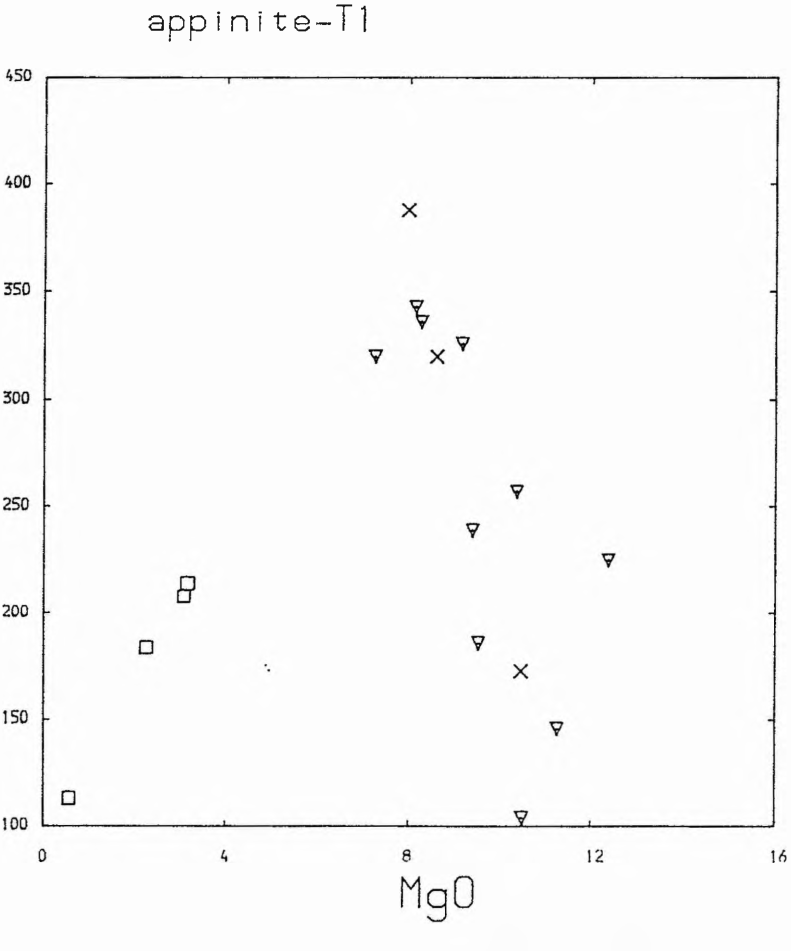


Fig 7.25 Zr against MgO.

Compatible element variation.

No core region containing coarse rocks rich in ferromagnesian minerals was found during mapping appinite-T1 (see Map 1). Compatible element oxide variation typified by MgO against SiO2 (Fig 7.22) produces linear trends which probably indicate either magma fractionation or magma co-mingling as the principal process. It is difficult to delimit any original magma composition for the body as no break of slope is observed on Harker diagrams for compatible elements. The usefulness of these Harker plots is limited due to the narrow range in SiO_2 exhibited by the appinite. Trace element variations plotted against MgO extend the the spread of the data and may be more informative.

Incompatible element variation.

Figs 7.22 to 7.25 display the variation of Nb, Sr and Zr with MgO. In each diagram appinite samples define linear trends (albeit with much scatter) which show no tendency to project towards tonalite compositions. Indeed the linear trend defined in Fig 7.25 (Zr vs. MgO) is normal to that expected for co-mingling.

Co-variation of Sr and MgO (Fig 7.24) shows that Sr increases as MgO decreases, consistent with crystal fractionation of ferromagnesian minerals. However, co-variations of Zr and Nb with MgO (Figs 7.25 & 7.23) are not so straightforward as both exhibit near vertical trends. Although Zr and Nb are incompatible elements within the amphibole or clinopyroxene structure, fractionation of these minerals alone cannot account for the doubling (Nb) or tripling (Zr) of concentrations over a 2% MgO drop. Such co-variation might be explained by the marginal accumulation of accessory phases (such as zircon and apatite) which have high concentrations of high field strength elements (Henderson 1980).

The chemical variation of appinite-T1 is best described by the fractional crystallisation of ferromagnesian minerals with concomitant sidewall accumulation of accessory zircon. No evidence was found to support the notion of marginal co-mingling of the appinite and tonalite during emplacement of the former.

7.2 DISCUSSION

The distribution of appinites within the Strontian pluton, and the chilled nature of the appinite contacts with their host suggest that the injection of the appinites was episodic and long-lived (at least the duration of plutonism, see Chapter 4). Repeated tapping of a large body of appinite magma throughout the duration of plutonism seems unlikely. This suggests that the appinite pipes represent discrete magma batches.

Nd-Sr isotopic constraints.

Regional studies.

Recent geochemical and isotopic studies on supposed mantle nodules (Menzies and Halliday 1985, Halliday et al. 1986) and Siluro-Devonian mantle-derived basalts and basaltic andesites from around Scotland (Thirlwall 1982), are summarised in Fig 7.26. Such studies have shown the mantle beneath northern Scotland to be extremely heterogeneous with different segments

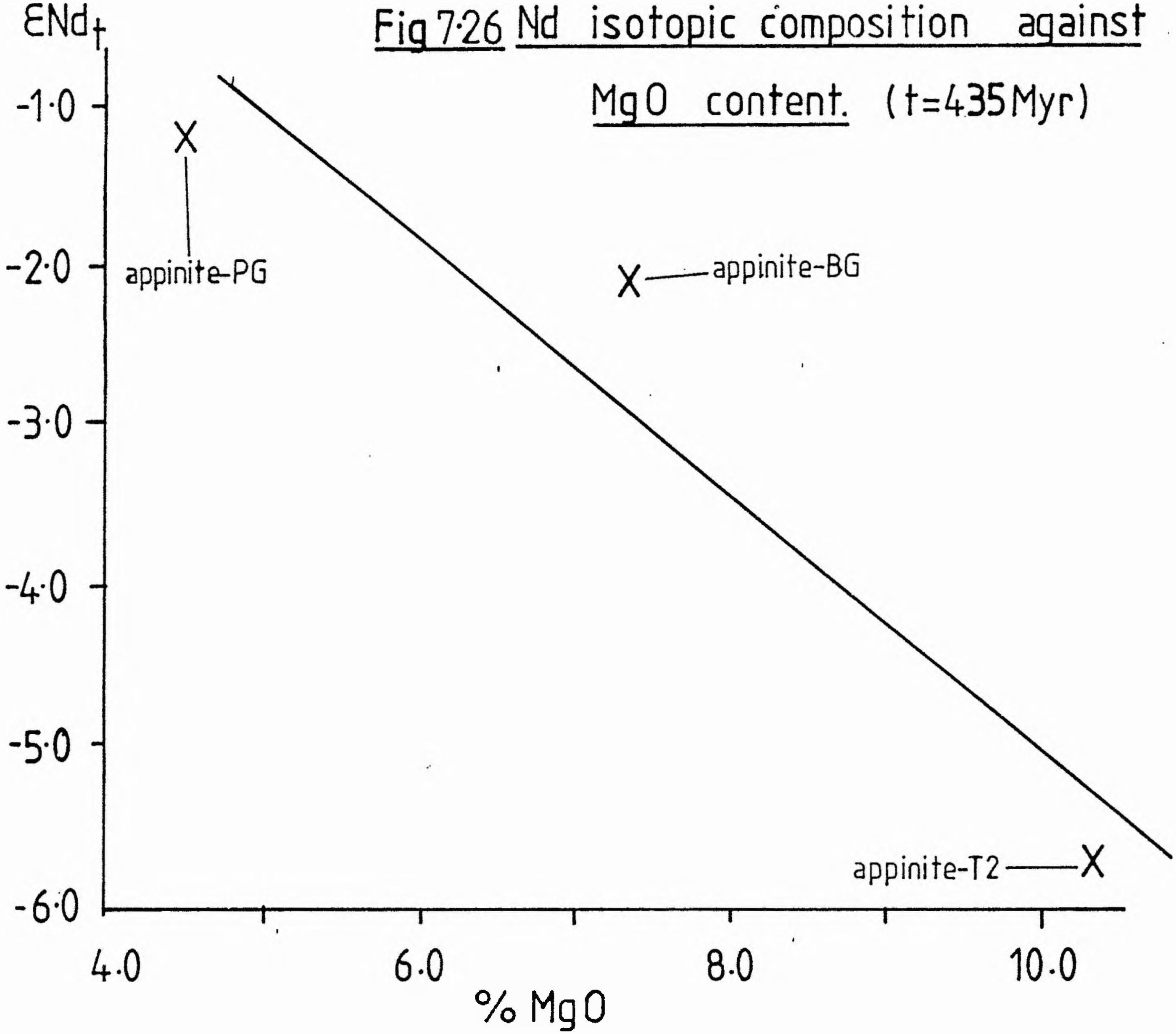


Fig 7·26 Nd isotopic composition against MgO content. (t=435Myr) having been affected by different processes. However, the mantle as apparently sampled by nodules closest to the Strontian pluton (i.e. from Streap Cumlaidh), defines an horizontal trend on the ϵNd_t-ϵSr_t diagram (Fig 7.27) and it is not unreasonable to expect mantle melts from this region to have a similar isotopic signature.

It is clear from this diagram that Strontian appinites do not contain an isotopic signature which is similar to any of the known Scottish mantle or crustal reservoirs. However there is considerable overlap with the field of Strontian tonalites and granodiorites which may or may not be significant (see later).

Origin of the Strontian appinites

Isotopic determinations were carried out on samples from the core of appinite-T2, -T2 & -BG (Fig 7.27). These samples, although probably cumulates, were chosen to avoid the effects of in-situ contamination from the host granitoid. Sample S316 has a high $^{147}Sm/^{143}Nd$ ratio due to the effects of high amphibole abundance which has a high partition coefficient for the HREE, and is not thought to represent the magma or mantle source conditions. It must be emphasised that the initial Nd isotope composition magma remains unaffected by the processes of fractional crystallisation over a short time period. Thus the initial Nd isotope signature of cumulate samples should be representative of appinite magma at the time of intrusion.

Strontian appinites define a vertical vector similar to that described by Loch Roag nodules but displaced to higher initial Sr isotope compositions (Fig 7.26). Five arguments may be advanced to explain this observation.

1) As appinites overlap the field for Strontian tonalites and granodiorites the appinites may represent re-mobilised cumulitic material from the host granitoids.

2) The appinites retain a primary mantle signature. Fractional crystallisation and cumulate formation occur in a closed system.

3) Parental magma compositions were derived from a mantle source which shows the time integrated effects of LREE and Rb enrichment similar to point A in Fig 7.27. Contamination of magma A with tonalitic material results in a vertical distribution trending towards less radiogenic Nd isotope compositions.

4) Appinite parental magmas are derived from a source similar to that for Lorne lavas (point B on Fig 7.27). Mixing of this composition with Moine or Dalradian metasediments could well explain the trend.

5) Contamination of an appinite parental magma derived from a source mantle similar to that described by mantle nodules from Streap (point A on Fig 7.27), by Lewisian amphibolites +/- granulites would define a trend towards unradiogenic Nd isotope compositions.

Each model is now discussed more fully.

1) <u>Appinites as granitoid cumulates.</u>

As was outlined in the introduction, Nockolds (1941) and Deer (1938 & 1950) suggested that the close spatial and temporal

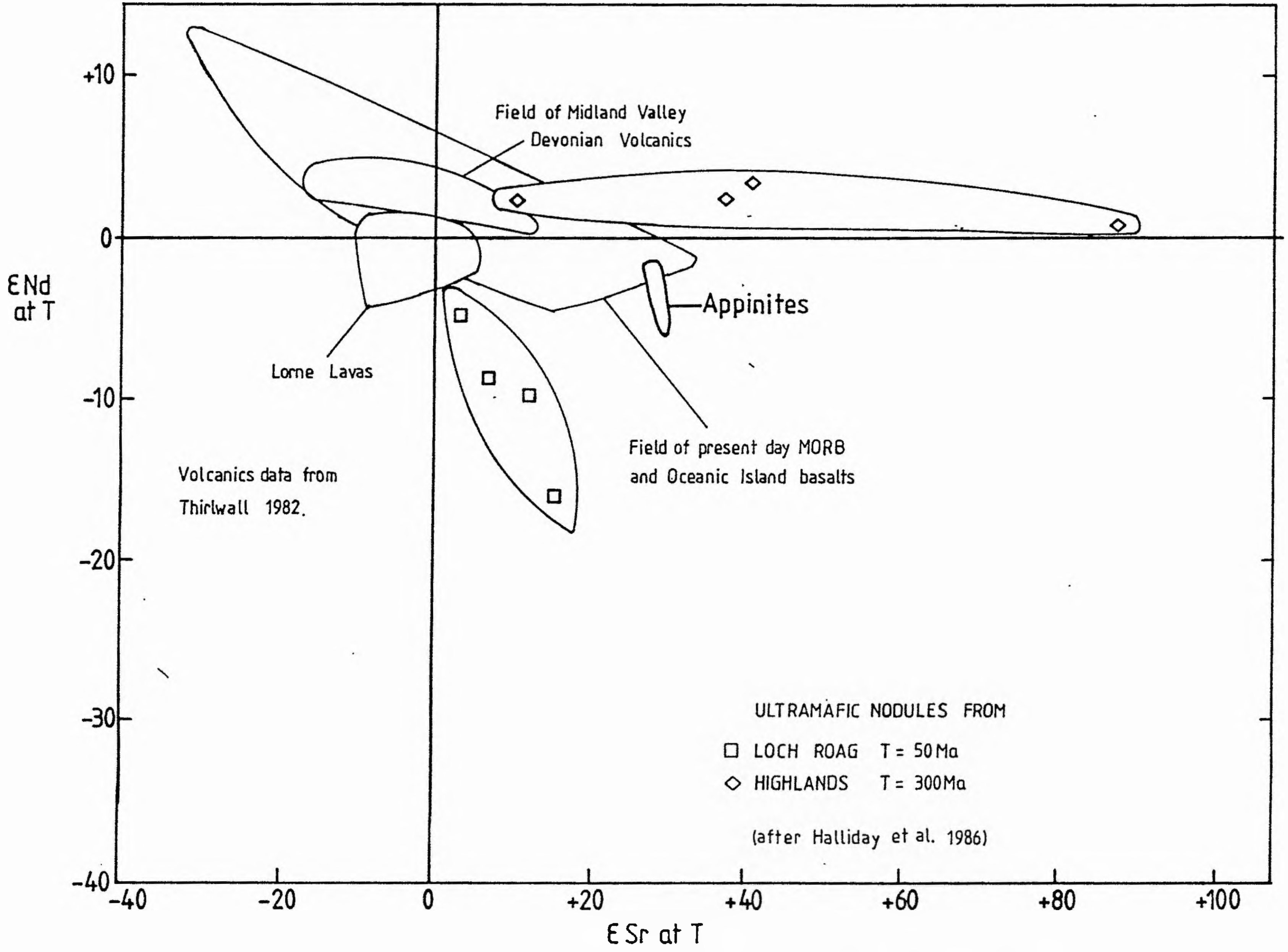


Fig 7.27 Regional isotopic variation of Scottish lithospheric mantle (after Halliday et al. 1986).

correlation between outcrops of appinite and Newer Caledonian granitoids could be explained if the appinites were the accumulated products of those granitoids. The fact that the appinites intrude the host granitoid as a magma, as is evident from the rapid growth textures in the contact zones, would seem to preclude this hypothesis at Strontian. Remelting and mobilisation of granitoid cumulates within a crustal setting would require exceptional circumstances as the temperature required for such melting would be expected to promote large scale melting of the surrounding crust generating an acidic magma.

Isotopic data do not support such an hypothesis either as ϵNd values for appinites are seen to be less radiogenic than the tonalite or granodiorites (Fig 7.26). It is most likely therefore, that appinites represent mantle melts whose Nd-Sr isotope characteristics are controlled by interaction with crustal materials.

(2) Appinites as uncontaminated mantle melts.

The Strontian appinites define on Fig 7.27 a trend similar to that observed for nodules within the Loch Roag dyke (Outer Hebrides, Menzies & Halliday 1984, Halliday et al. 1985). Whilst the Loch Roag samples plot to low Sr isotope compositions, the Strontian appinites are clearly displaced to more radiogenic values which requires explanation.

In their analysis of the North of England lamprophyre suite, MacDonald et al. (1985) invoked the enrichment of the mantle source region prior to magma generation. They described the high incompatible element concentration found in terms of enrichment of the source via CO_2-rich and H_2O-rich 'fluids'. Each type of enrichment process would be expected to define trends at right angles to each other on a Nd-Sr isotope diagram (Menzies & Wass 1983). That is, hydrous phase transport would be expected to enrich Rb relative to LREE defining an horizontal trend (like that observed for Streap nodules), whilst CO_2-rich fluids would be expected to transport LREE in preference to HREE and Rb, defining a vertical vector as seen at Loch Roag (Menzies & Wass 1983). Clearly if both processes are to operate simultaneously then an intermediate trend would be expected, the gradient of which would depend on the relative contributions of each process.

Enrichment of the mantle by hydrous and CO_2 'fluids' cannot explain the vertical nature of the appinite vector, especially the high radiogenic Sr compositions. Combined enrichment processes would be expected to define a trajectory similar to those described by enriched sources for ocean islands such as Kerguelen (White & Hofmann 1982) or potassic provinces similar to the Roman province (Hawkesworth & Vollmer 1979, Rogers et al. 1985). A solely mantle origin therefore seems unlikely.

(3) Mixing of an appinite magma with a tonalite.

Appinite magma generated from a mantle source similar to point A (Fig 7.27) within the Streap nodule field would be expected to generate a vertical mixing line with Strontian tonalite +/- granodiorite composition. However, the large degree

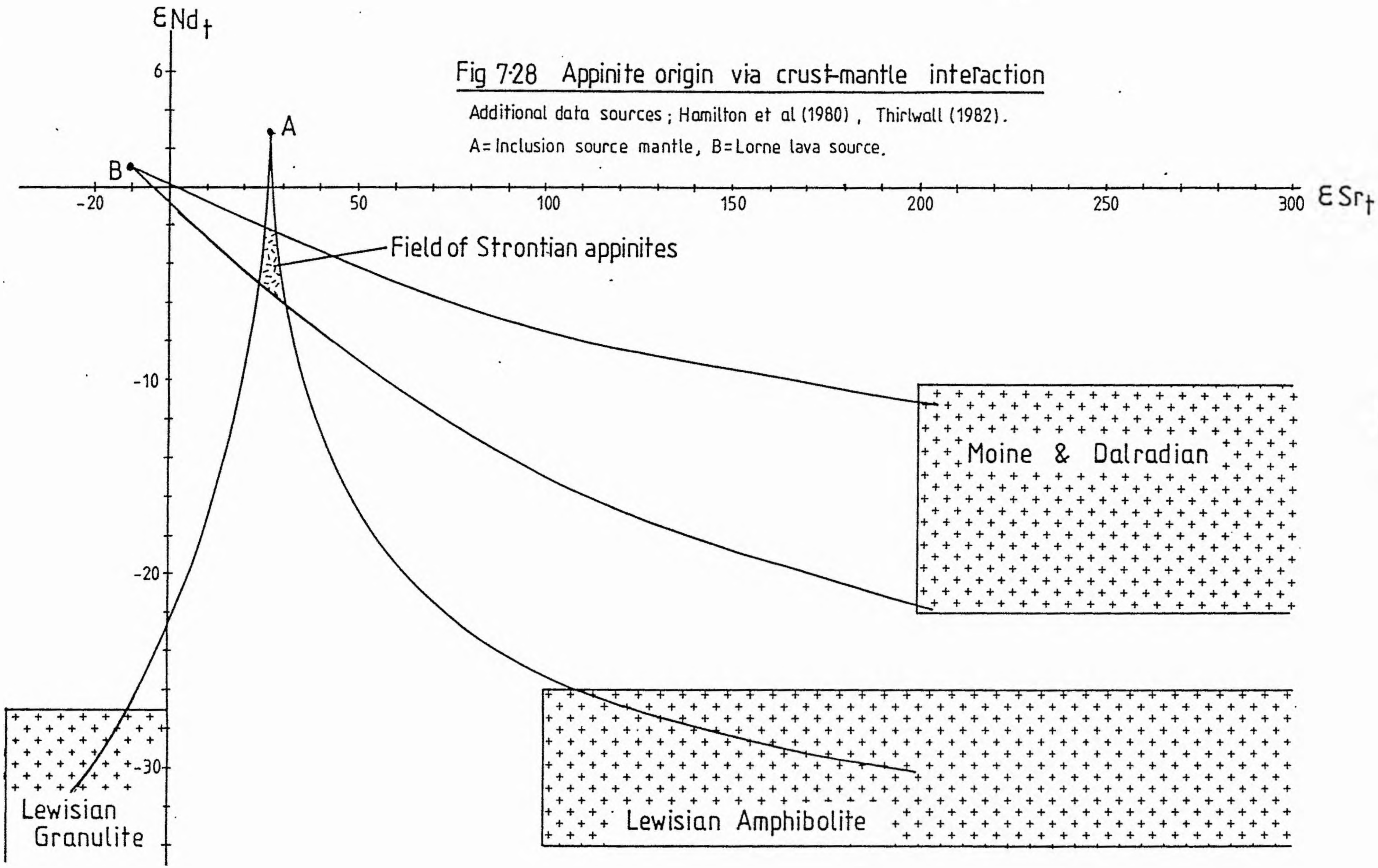


Fig 7·28 Appinite origin via crust-mantle interaction

Additional data sources; Hamilton et al (1980), Thirlwall (1982).

A = Inclusion source mantle, B = Lorne lava source.

of overlap of the two fields on the ϵNd-^{87}Sr/^{86}Sr (Fig 7.27), requires that the majority of the mixed magmas be of tonalite/granodiorite composition (i.e. >90% granitoid). Even so, such mixing cannot explain the isotopic composition of sample S316 which has a less radiogenic Nd isotope composition than Strontian granitoids.

A Sr isotope traverse through appinite-BG has shown that very little Sr derived from the biotite granite is permissible in the core region of the appinite indicating that mixing was not an important process. Furthermore, major and trace element studies on appinites-T1 to -BG have failed to establish any relationship between appinite core regions and surrounding granitoid. It appears unlikely therefore that Strontian granitoid had any role to play in chemical evolution of the appinites prior to their emplacement.

(4) Contamination of a magma derived from a source similar to that of the Lorne lavas with upper crustal material.

The mantle source for the geographically close Lorne lavas is not similar to that of Streap nodules (Thirlwall 1982). Contamination of a Lorne lava-like magma with Moine/Dalradian materials could under special circumstances yield a vertical trend. Magma derived from an homogeneous mantle source (point B on Fig 7.27) when contaminated by heterogeneous upper-crustal material would produce a family of curves the curvature of which would depend on the Nd and Sr elemental concentrations and isotopic compositions of the end-members (Langmuir et al. 1977).

Under fortuitous circumstances similar proportions of mantle and crustal end-members could produce a series of mixed rocks which would apparently yield a trend normal to the direction of the mixing curve.

Whilst such mixing is perfectly feasible for the origin of Strontian appinites, it requires special circumstances to prevail in order to explain the observed isotopic trends, and is not considered to be a principal process.

(5) Contamination of a Rb enriched, Nd depleted magma.

A magma derived from point A on Fig 7.28 could be contaminated by a number of crustal materials. However, contamination with a crustal component with an isotopic signature similar to Moine or Dalradian would not be expected to generate the observed trends and are as such unlikely candidates (see above). Lewisian material (especially amphibolites) would define a vertical mixing vector to less radiogenic Nd isotopic compositions.

Due to the uncertainty of the age of the nodule host, the exact isotopic composition of the Streap mantle at 410Myr is uncertain, as back-projection is unsound beyond the age of the host. However, it is clear that the Streap mantle at some time in the past has been enriched in Rb relative to Sr, the time integrated effects of which are to increase the ^{87}Sr production, defining an horizontal vector on the Nd-Sr isotope diagram.

Appinite magmas derived from within the Streap field (point A on Fig 7.28) when mixed with small amounts of Lewisian amphibolite material will define a vertical trend similar to that observed at Strontian. Of necessity this contamination must occur prior to the emplacement of the appinite. This combined with the cumulitic nature of the core regions of Strontian appinites suggests that they have resided for some period in a crustal magma chamber where processes of fractional crystallisation and crustal contamination may have occurred simultaneously.

A generalised model.

It is clear that the mantle beneath the Northern Highlands and Outer Hebrides has a long history of enrichment and depletion events (Halliday et al. 1986). Whether or not the same segment of the mantle has remained attached to the crust beneath the Lewisian since its formation around 2.9Gyr is unclear (Hamilton et al. 1979). As with most geological processes, the best fit answer to the appinite problem does not preclude alternative explanations. Thus the following model is proposed for the generation of the appinites at Strontian and is summarised in cartoon form in Fig 9.1. Its general applicability to other appinites will have to await detailed isotopic inspection of other bodies.

1) Magma is generated in a source mantle which shows the time integrated effects of Rb enrichment and LREE depletion. The degree of melting may locally be high, and may be due to local pockets rich in volatiles. Widespread mantle melting may not have taken place.

2) Batches of magma rise into the crust where they fractionate an assemblage of hornblende +/- clinopyroxene and biotite in small magma chambers. The nature of the crust is similar to that of Lewisian amphibolite. Contamination of the appinites may occur either whilst within the chamber or more likely, in the conduits that supply fresh magma to the chamber. It is worthy of note that magmas must have been contaminated prior to crystallisation, as they record the isotopic conditions at the time of their crystallisation from an unradiogenic magma.

3) Continued crystallisation caused a build up of volatile pressure within the chamber. A fresh influx of hot appinite material at the base of the chamber may be sufficient to trigger violent intrusive activity (Sparks et al. 1977). Although initially violent, later phases of activity are characterised by less energetic laminar flow regimes (Dawson 1979 for the kimberlitic analogue).

4) Pipes from the appinite chamber intrude randomly and some (fortuitously) invade still unconsolidated granitoid magma. Here, chilling of the appinite against the host is observed. Peripheral mixing may occur although this probably depends upon the 'liquidity' of the host granitoid at the time of intrusion.

5) At some stage the amount of liquid within the appinite chamber becomes exhausted, with increasingly more cumulitic material being drawn up as a crystal-liquid mush, eventually clogging the system. Laminar flow through the pipe allows the maintenance of geochemical and isotopic gradients with the wall rocks.

6) The process is repeated in other small magma chambers.

The volume of lamprophyric material (thought to be the fine-grained equivalents of appinites (Pitcher & Berger 1972)) within the British segment of the Caledonides, attests to the large overall volume of material intruded during the latter stages of closure of the Iapetus Ocean. It has long been observed that the appinite-lamprophyre suite has close spatial links with Caledonian calc-alkaline plutons (Bailey et al. 1960, Hall 1967, Pitcher & Berger 1972, Wright & Bowes 1979, Watson 1984). The relationships of appinites to their host granitoid plutons are further explored in Chapters 8 & 9.

CHAPTER 8

PETROGENESIS OF PLUTONS AND THEIR INCLUSIONS

8.1 INTRODUCTION

For the purposes of this discussion the term 'mantle-derived' will be used for those igneous rocks which have formed directly from mantle melts without recycling. Crustal-derived melts are thus those formed from the melting of source rocks which have had a residence period within the continental crust.

When considering the processes by which the continental crust segregated from the mantle, it is important to appreciate the role of granitoids. Granitoids wholly or partly derived from the mantle are an integral part of crustal growth increasing the continental mass, whereas granitoids which are crustally derived involve only intracontinental differentiation with no concomitant change in crustal mass.

With the advent of high-precision analytical techniques many authors have sought to use chemical and isotopic data to constrain the provenance of granitoids (e.g. Allegre & Ben Othman 1980). Chappell and co-workers, based upon various chemical criteria, have sub-divided the granitoids of the Lachlan fold belt (SE Australia) into two major types, the I-type and S-type granites (Chappell & White 1974). They suggested that most orogenic granitoids were the product of partial melting of the continental crust and that granitoid magmas were intruded as a mush of melt plus residual refractory material (restite) (White & Chappell 1977). Progressive separation of basic restite (e.g. inclusions) would ultimately yield a low temperature minimum melt or near minimum melt granitoid, as is often observed in the centres of calc-alkaline, normally-zoned, granitoid plutons throughout the world.

The simple concept of I-type and S-type granitoids has been applied to other granitoid provinces with varying degrees of success(Didier et al. 1982, Pitcher 1983, Stephens & Halliday 1984). Whilst a general consensus on the origin of S-type granitoids has been achieved (that is they represent the partial melting of a sedimentary source which has been through a weathering cycle and are by definition crustally derived) no such consensus has been achieved for the I-type plutons (see Vernon 1983 and Grey 1984 for discussion of Australian I-type granitoids).

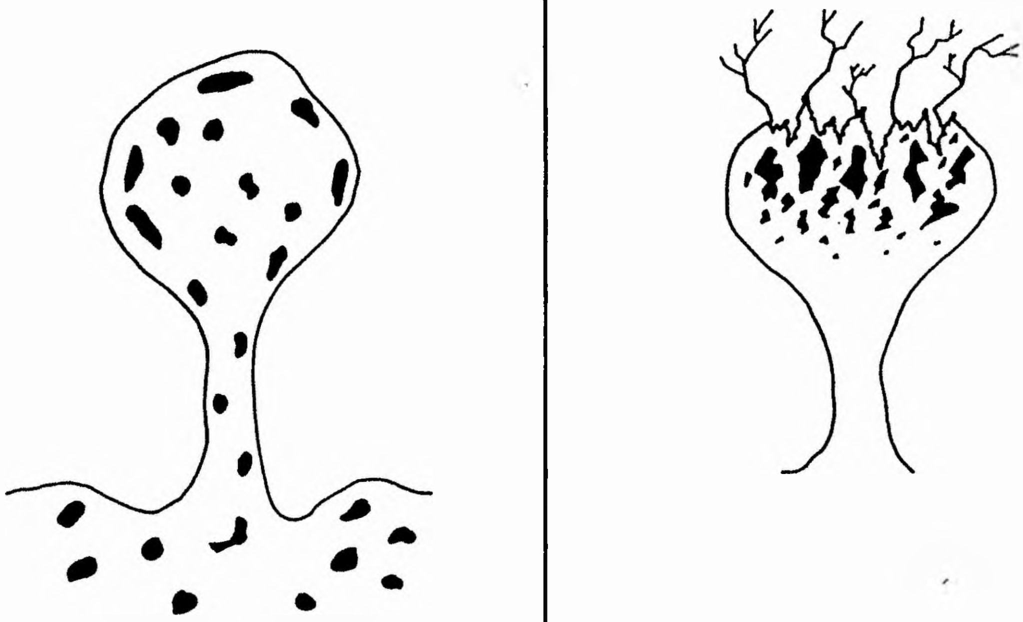
In destructive plate environments such as the eastern Pacific, intermediate magmas (of I-type) have been shown to have similar isotopic compositions to that of local mantle (e.g. Cameron & Cameron 1985) and these essentially mantle-derived melts have a much smaller proportion of crustal contamination (McNutt et al. 1975). Moreover, gabbroic (and other basic mantle-derived magmas) make up around 15% of the total igneous rocks of the area (Pitcher 1979) indicative of active mantle participation in magmatic activity. Eichelberger (1978, 1980) favours a two-stage mechanism for intermediate magma production, namely (1) the intrusion of hot basaltic material into the base of the crust followed by (2) melting of crustal rocks to give acid magmas and mixing of the two magma types. Inclusions in this instance represent the disrupted remnants of the early basaltic influx, a view supported by Gill (1981), Hildreth (1981), Wyllie

(1983b), Halliday (1983) and Cantagrel et al. (1985). These models contrast with the Australian case where basic bodies are a rare rock-type suggesting little mantle input, intermediate magma production being solely confined to intra-crustal melting.

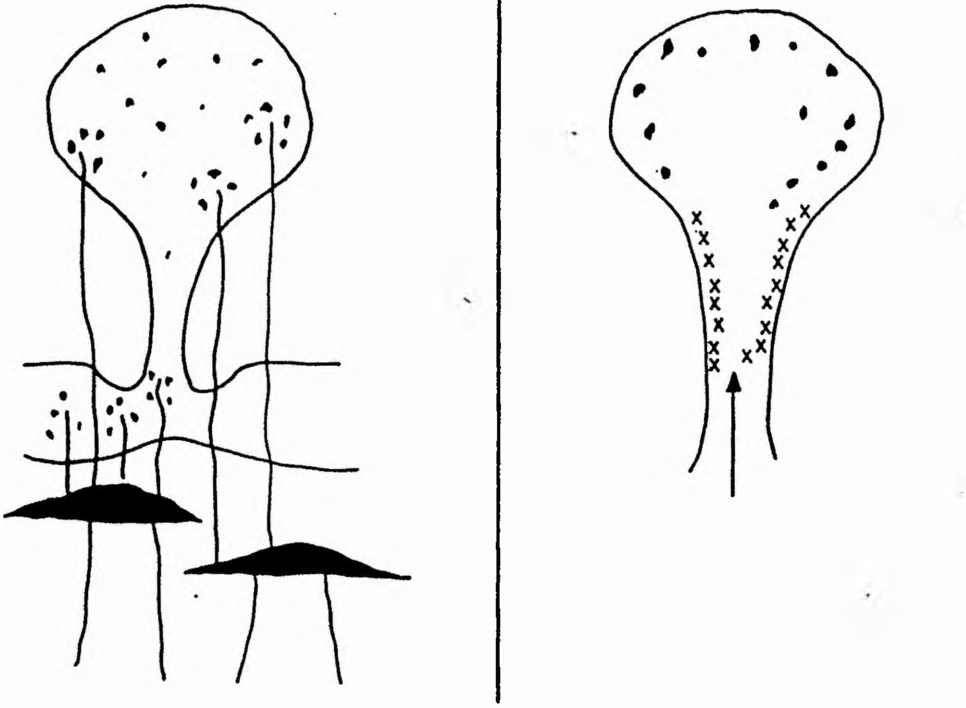
The interpretation of the role of inclusions is therefore central to the understanding the origin and compositional diversity of the host granitoids. This chapter examines the origin(s) of inclusions and in so doing adds to the constraints on the petrogenesis of the granitoids. It has been found that inclusions from both the Strontian and Criffell plutons contain more radiogenic Nd isotopes values than their hosts. Such radiogenic values (which are typical of local lithospheric mantle (Halliday et al. 1985) can only be explained by the presence of mantle derived melt either in the granitoid source region or at emplacement.

Role of inclusions

Microgranitoid enclaves have been observed throughout the world (Pabst 1928, Grout 1937, Didier 1973, White & Chappell 1977, Ferstater & Borodina 1977, Eichelberger 1978, Vernon 1983, Reid et al. 1983, Cantagrel et al. 1984). Typically these inclusions are spherical to discoid in shape with a well-defined size range seldom exceeding 1m in diameter. Generally these inclusions have igneous textures (Vernon 1983, Cantagrel et al. 1984), are of fine to medium grain size, and are compositionally similar to quartz diorites (Didier 1973, White & Chappell 1977, Fig 8·1 POTENTIAL INCLUSION SOURCES


A. Restite

B. Country Rock


C. Dyke Injection

D. Cummulate Remobilisation

Tindle & Pearce 1983). As discussed in Chapter 1 inclusions of microgranitoid type have been variously interpreted as the residue after partial digestion of country rock fragments (Grout 1937), restite brought up from an igneous source region (Chappell & White 1974, White & Chappell 1977, Chappell 1984), remobilised basic precursors or cumulates (Pabst 1928, Grout 1937, Ferstater & Borodina 1977), congealed pillows of basic liquid intruded into the granitic magma (Sparks et al. 1977b, Eichelberger 1978, 1980; Hildreth 1981; Domenick et al. 1983; Grey 1984) and segregated immiscible liquids (Bender et al. 1982).

There is no a priori reason why inclusions should be derived solely from a single source, indeed Didier (1973, p227) in considering the Aiggoual pluton (Massif Central) concluded that microgranitoid inclusions could be sub-divided into "one pale and two dark" types. Inclusions therefore must be classified and grouped before a suitable origin may be assigned to them.

8.2 ORIGIN OF INCLUSIONS

It is clear from the preceding Chapter (6) that each of the study plutons contains groups of inclusions derived from several sources (see also Didier 1973). Fig 8.1 is a schematic representation of the possible sources for inclusions outlined in Chapters 1 and 6. For the most part inclusions within the Strontian and Criffell plutons contain Nd isotopes which are in disequilibrium between host and inclusion by from 1 to 3 ϵ units. Exceptions at Strontian are one subgroup of three samples (S207X, S331X and S340X) with low ϵNdt (-0.8) which lies within the Nd

Fig 8.2 Nd-Sr isotopic data for Strontian and Criffell host-inclusion pairs (joined by tie lines). Additional data from Hamilton et al. (1980) and Halliday et al. (1980).

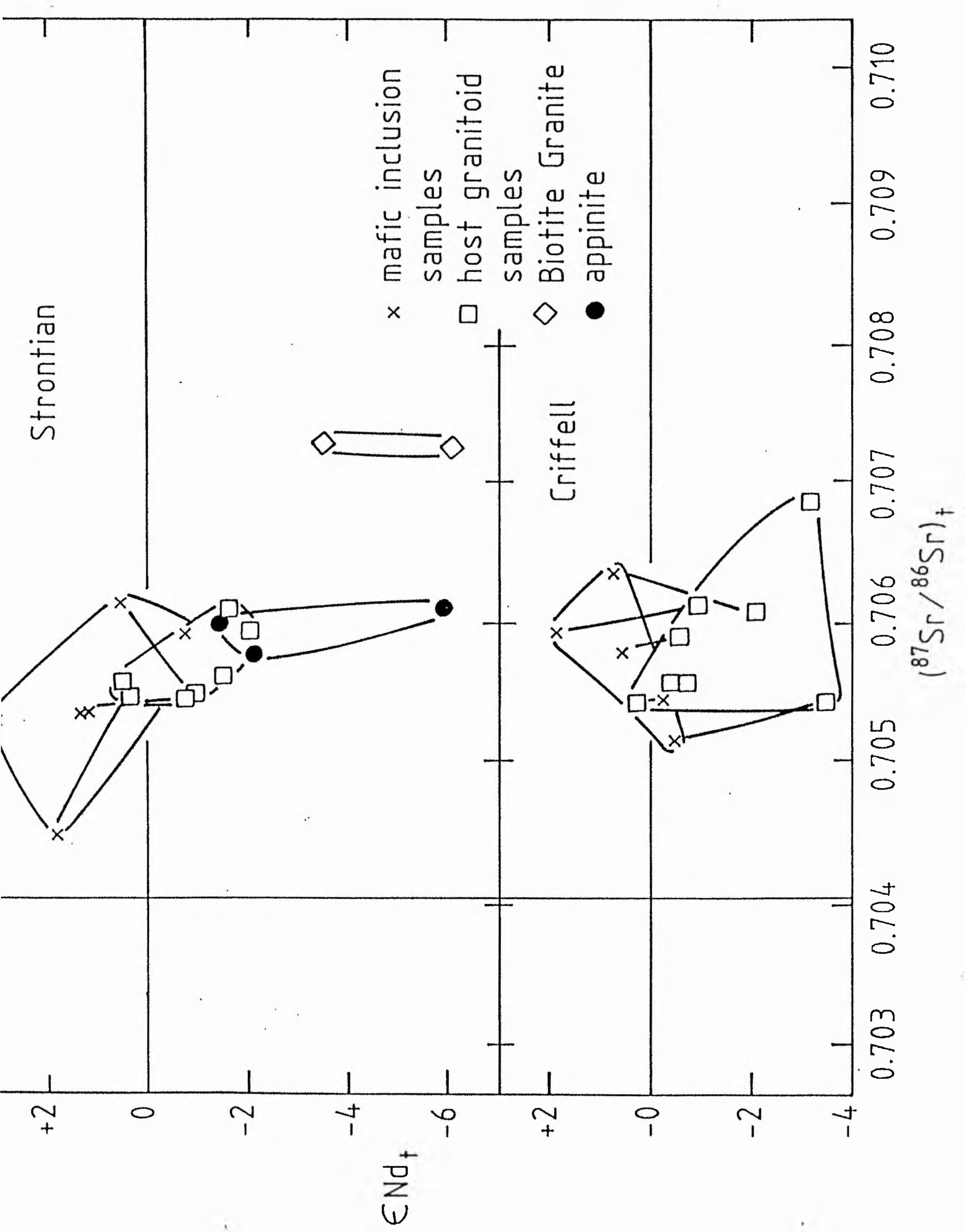
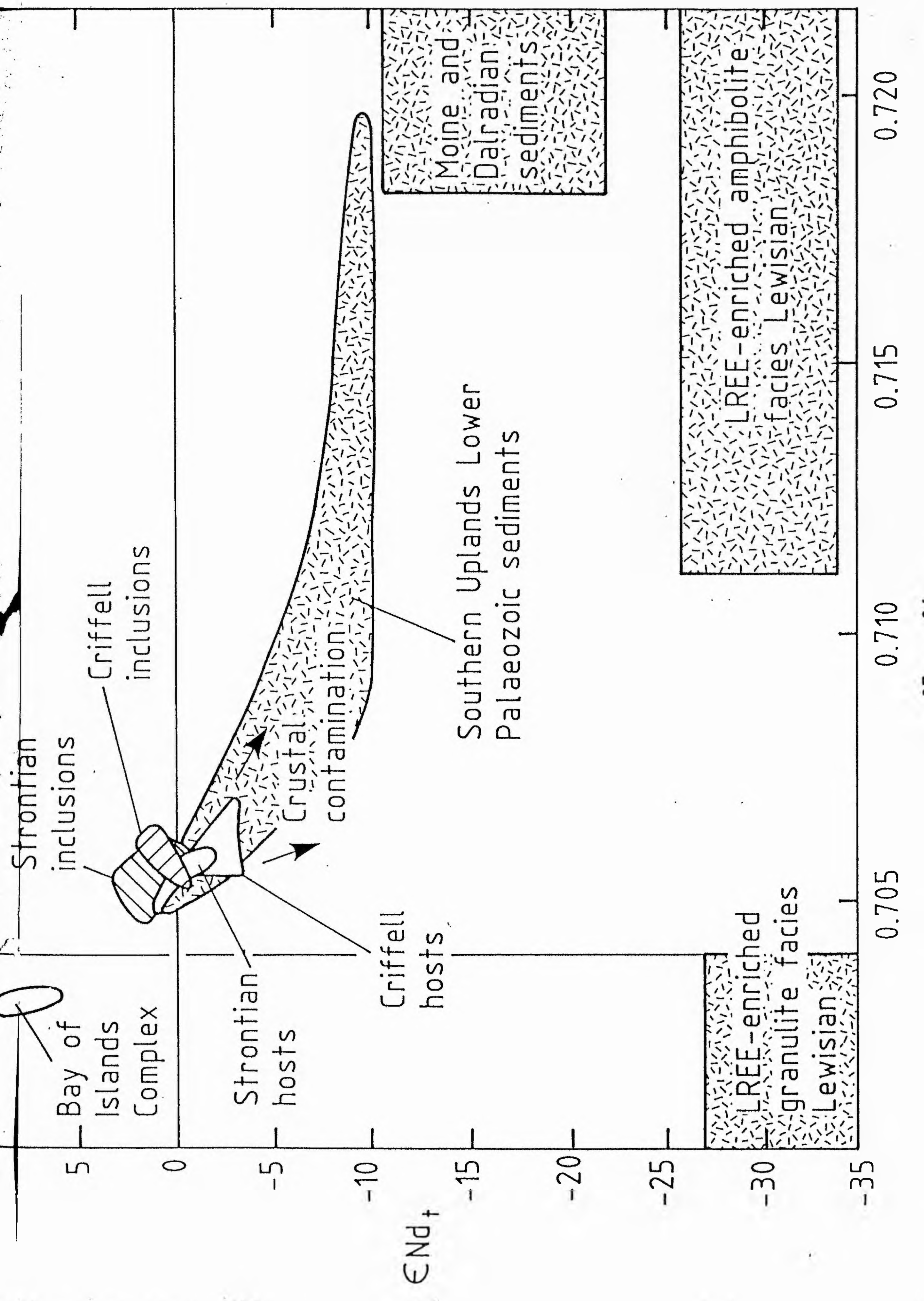


Fig 8.3 Nd-Sr isotopic data for Scottish crustal sources. Additional data from Hamilton et al. (1980) and Halliday et al. (1980).



isotopic field of the hosts. Since geochemical parameters such as SiO_2, Nb, Zr, Ce, and Y are similar to those for the host granitoid, it is assumed that this small group of inclusions is cogenetic with the host pluton and as such may be labelled as autoliths (Ferstater & Borodina 1977).

Based on similar criteria, a similar minor group of autoliths forming a basic extension to the geochemical trends of the Criffell pluton was also found. However without the benefit of Nd isotopic data inclusions within the Ballachulish pluton could not be confidently assigned to any genetic grouping.

Apart from these exceptions, the majority of inclusions (90% of those analysed) within the Criffell and Strontian granitoids contain a Nd isotopic signature which is <u>more</u> radiogenic than their immediate host granitoids (Figs 8.2 & 8.3). This immediately rules out a co-genetic origin for inclusions unless the host had suffered preferential and severe contamination from an unradiogenic Nd source (namely the crust). This appears unlikely at Strontian as analysis of zircon separates from the tonalite and granodiorite portions (Pidgeon & Aftalion 1979) do not record a significant Pb isotopic ‘memory’. Consequently major assimilation of local crust (Moine) or inferred substrata (Lewisian) is unlikely as both metasedimentary rock-types contain "old" zircons, and therefore inclusions are probably not cogenetic. Hence the majority of inclusions are not disrupted chilled margins to the pluton, the remnants of an earlier basic phase or relicts from the spontaneous segregation of a single melt due to immiscibility. It is wrong, therefore, to refer to these inclusions as autoliths or as cognate inclusions unless they can be proven to be genetically related to their host (e.g. S207X, see Chapter 6). Neither can these inclusions be the representatives of residual unmelted source material brought up with the host as both would be expected to have the same Nd isotopic signatures (White & Chappell 1977, Chappell 1984, but see also Grey 1984 for an alternative interpretation).

Thus models a) and d) from Fig 8.1 are discounted as viable sources for the majority of inclusions. Consideration of Fig 8.3 shows clearly that digested local country rock fragments can also be discounted as viable sources for microgranitoid inclusions at Strontian. Equilibration of country rocks (Frost & O'Nions 1985) such as Moinian metasediments cannot produce residual inclusions with ϵNd values greater than Bulk Earth at the time of emplacement. The same is true for the deep crust if it had ϵNd<0. Similarly the Southern Uplands sediments are not viable sources for Criffell inclusions since all those analysed, with the exception of one basic clast greywacke (Halliday et al. 1980) from the northern end of the accretionary prism, have ϵNd<0 (O'Nions et al. 1983, Halliday 1984).

Oscillatory zoned plagioclase megacrysts within the inclusions are taken to indicate an original magmatic origin (Vance 1962, Sibley 1973, Hibbard 1979), whilst hollow and acicular apatites suggest that the inclusions were the product of quenching (Wyllie et al. 1962) of a magma hotter and more basic than the host, probably at high level. Nd isotope ratios also indicate a more primitive source for the inclusions.

At Strontian, syn-plutonic appinite bodies provide clear evidence for the dual existence of mantle-derived ‘appinitic’ magma and a still-molten granitoid at a high structural level. The low Mg numbers (53-57) of the inferred appinite liquid compositions suggest that fractionation of the appinite took place prior to its injection to its present position. If this were the case, it might be expected to have been accompanied by crustal contamination which would explain the low ϵNd values (Huppert & Sparks 1985, Campbell 1985 and Fig 2.33). The vertical vector defined by the data for appinites (Fig 8.4) suggests that Lewisian crust would be the most likely contaminant. Such a contaminated ‘appinite’ magma cannot have been the source for microgranitoid inclusions since the appinites (like the country rock) have $\epsilon Nd_t < 0$. Nevertheless the possibility exists that both the appinite and inclusion parental magmas were derived from a similar mantle source, the appinites ponding within the crust and suffering crustal contamination whilst the inclusions are the product of direct magma injection into the host granitoid, quenching precluding significant modification of the Nd isotope ratio.

An alternative explanation, therefore, is that the mantle in the vicinity of Strontian is grossly heterogeneous including both enriched and depleted portions, the inclusions being partial melts of the depleted portion the ‘appinite’ magmas being melts from the enriched zone.

Since potential sources a), b) and d) of Fig 8.1 are precluded, microgranitoid inclusions from both Strontian and

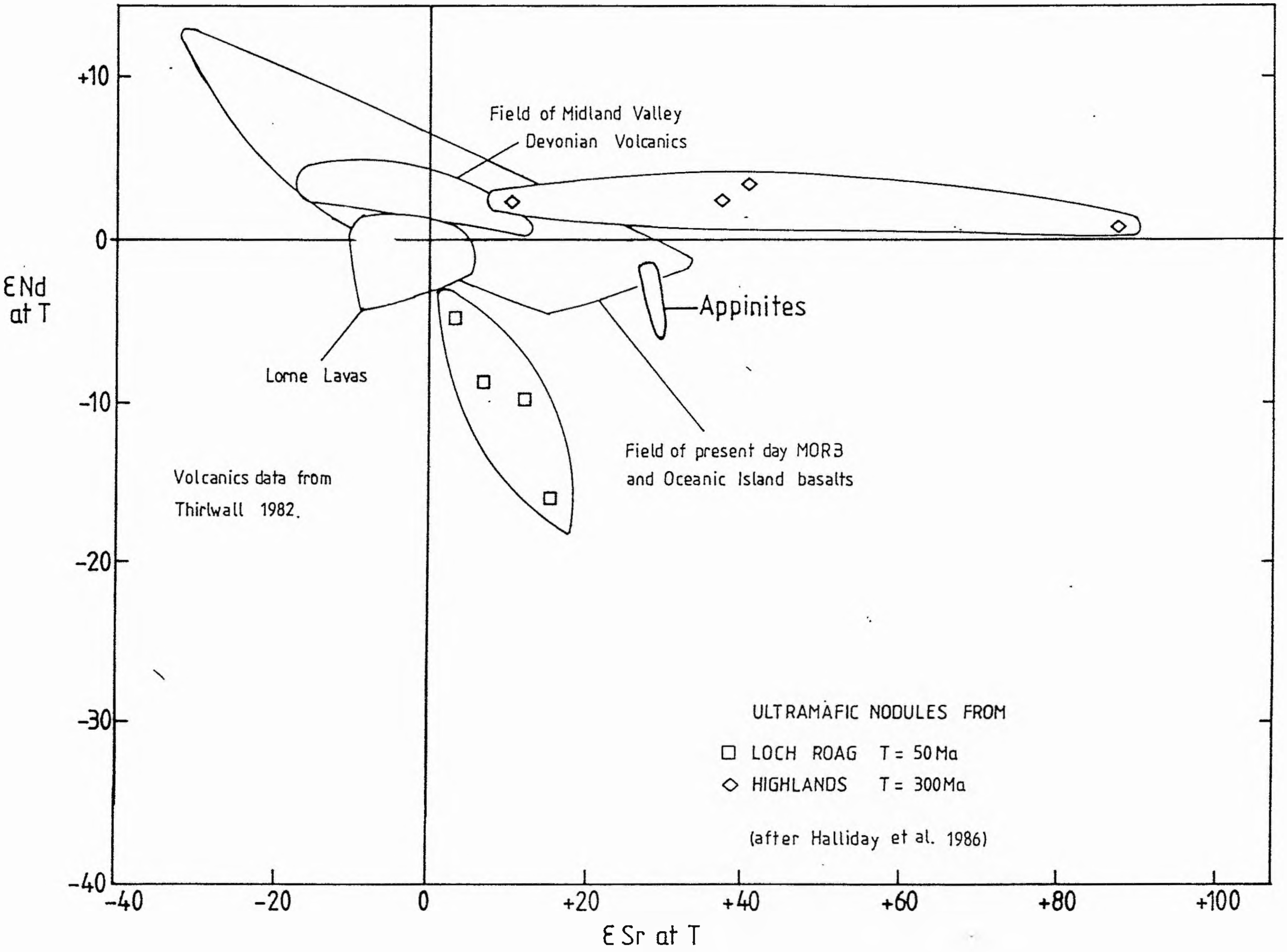


Fig 8.5 Mantle isotope characteristics in Scotland.

Criffell may be thought of in terms of syn-plutonic mantle-derived magmas injected into their host at some stage during the evolution of the granitoid. The fact that microgranitoid inclusions within all three plutons can be subdivided into discrete types suggests that the mantle source for inclusion precursors was repeatedly tapped throughout the host pluton's formation and emplacement.

Comparison with other mantle-derived magmas.

Having constrained the inclusions within the study plutons (with the exception of a small number of autoliths) to a sub-crustal lithospheric source, it is interesting to compare their geochemistry with that of other Scottish Caledonian mantle-derived magmas.

Recent isotopic and trace element analyses of Caledonian mantle-derived volcanics (Thirlwall 1979,1982 & 1986) and of mantle and lower crustal xenoliths (Menzies & Halliday 1984, Halliday et al. 1986) are summarised in Fig 8.5. It is clear from this diagram that the lower crust and upper mantle (as apparently sampled by nodule suites) beneath Scotland were extremely variable even in Nd isotopic ratios. A few general features may be noted:

1) Basalts, mantle nodules and lower crustal nodules from Midland Valley localities lie close to the field for present day MORB, defining a consistent group.

2) Lower crustal samples from both the Highlands and the Hebrides

are extremely variable in both Nd and Sr isotopes. One granulite sample from Streap Cumlaidh (Argyllshire) had +ve ∈Nd in Paleozoic times.

3) Mantle nodules from Streap Cumlaidh have a narrow range in ∈Nd but extremely variable $^{87}Sr/^{86}Sr$, whereas similar samples from Loch Roag (Hebrides) have a narrow range in $^{87}Sr/^{86}Sr$ and a large range in ∈Nd.

In terms of Nd isotopes, Strontian inclusions compare favourably with the Streap mantle nodules (+0.5 to +3.0 (excluding autoliths) and 0.0 to +4.0 respectively), whilst Criffell inclusions are similar to the less radiogenic end of the Midland Valley lavas (-0.4 to +1.9 and 0 to +10 respectively) (Fig 8.5). ϵNd_t values for Strontian inclusions overlap those of the Lorne lava series (particularly the Ni rich primitive basaltic andesites) but are not completely encompassed by them (∈Nd=-3.7 to +1.0 at 410; Myr Thirlwall 1982), whilst inclusions from the Criffell pluton are not typical of Midland Valley mantle nodules (Fig 8.5).

In view of the extremely variable isotopic compositions of the Scottish lithosphere, a lack of perfect overlap with local lavas or mantle nodule suites might be anticipated (not withstanding the effect of other processes such as crustal contamination). Indeed at Strontian basic mantle-derived appinites have ∈Ndt values of -6.1 to -1.3 at the time of plutonic activity, indicating either processes of contamination or considerable mantle heterogeneity.

However, despite the general Sr isotope equilibration which occurs between host-inclusion pairs, one key observation may be made. Tie-lines joining adjacent inclusion and host pairs do not follow parallel lines. Perfect equilibration would result in vertical tie-lines (all parallel) on the Nd-Sr diagram whilst quasi-equilibration of an homogeneous source would yield a fan of tie-lines (approximately sub-parallel) back towards the original starting composition. Equilibration of an inclusion with initially less radiogenic Sr isotopes than the host, cannot, as the result of exchange processes, end up with a composition more radiogenic than the host.

At Strontian, three inclusions pairs have distinctly different tie-line trajectories, one vertical, one to less radiogenic compositions than the host and one to more radiogenic compositions than the host (Fig 8.2), such that equilibration of an homogeneous source (in terms of strontium isotopes) is untenable. Such an exotic source for Strontian inclusions would have to have highly variable Sr isotope compositions (both more and less radiogenic than Strontian granitoids) whilst having a roughly constant and more radiogenic Nd isotopic signature (ϵNd_t +0.5 to +3.0). In short, inclusions from within the Strontian pluton are perfectly consistent with a derivation from a source similar to that beneath Streap Cumlaidh.

At Strontian, comparison with a best estimate local mantle in terms of Nd isotopes appears to show that inclusions are close representatives of lithospheric mantle conditions. Comparison of Tdm ‘ages’ (Farmer & DePaolo 1983) for inclusions and mantle samples (Fig 8.6) also gives close correlation, although there is

Fig 8·6 Histograms of Tdm 'ages'

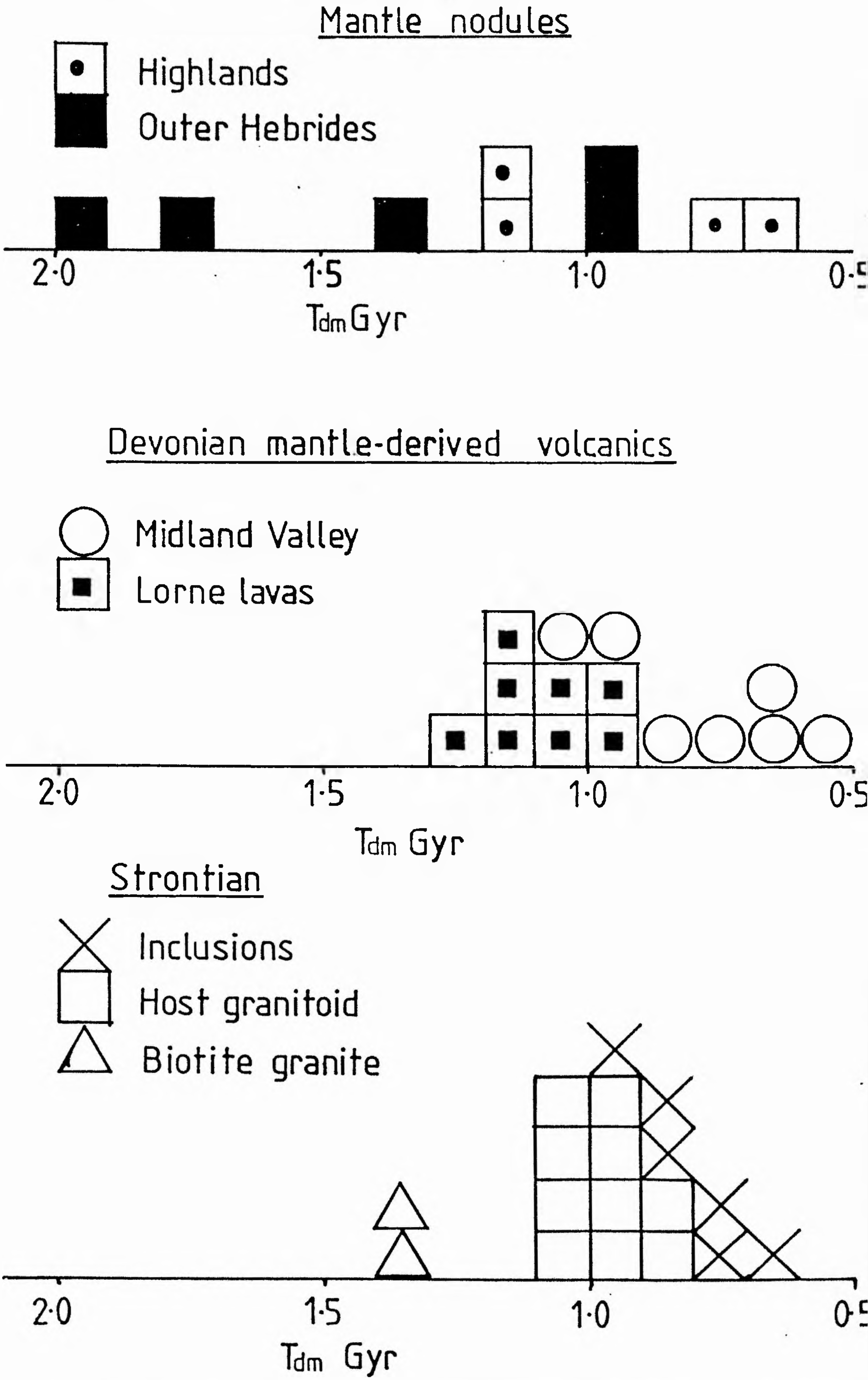
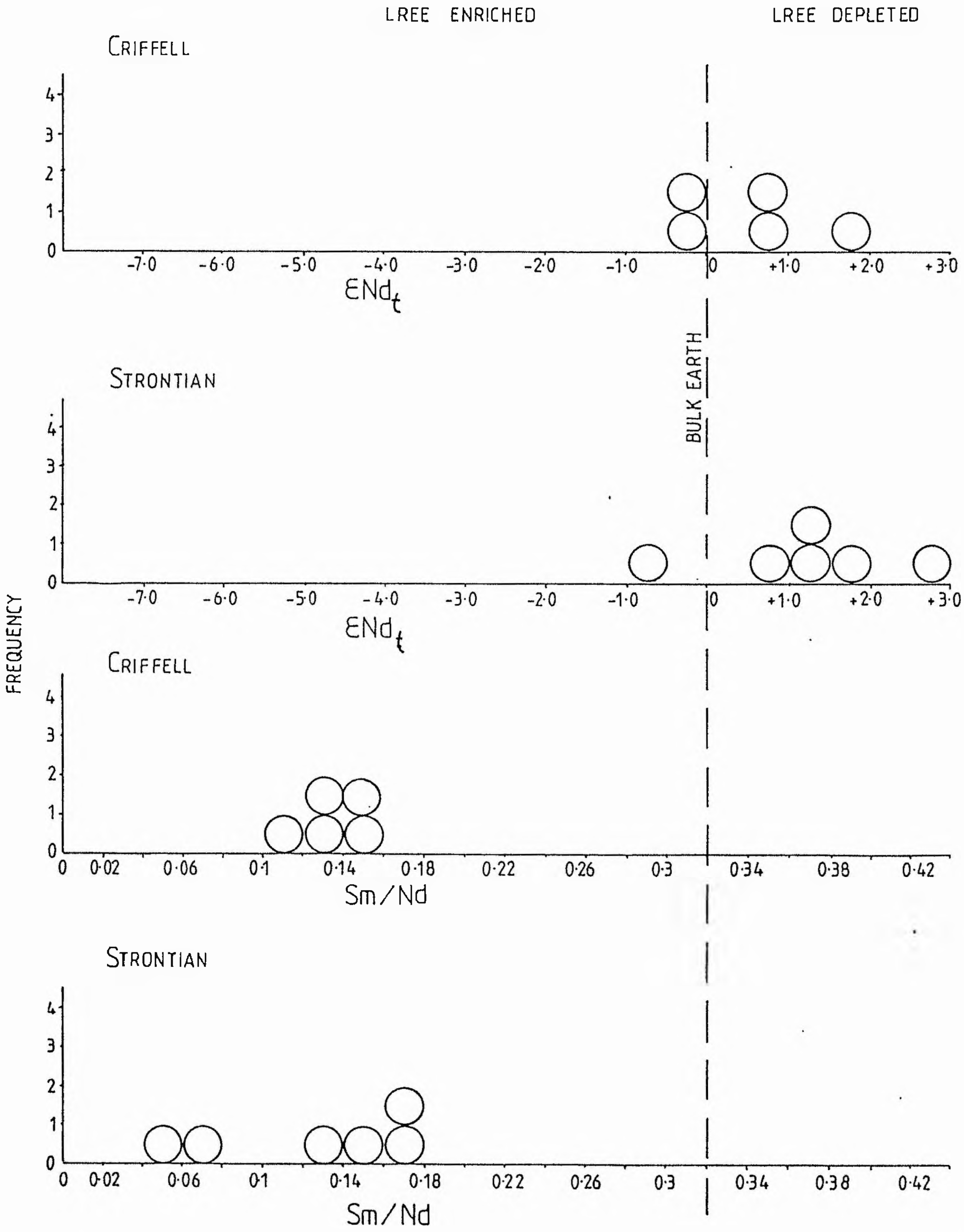


Fig 8·7 Histograms of εNd_t and Sm/Nd for inclusions.

Bulk Earth data from Menzies and Wass (1983)



little overlap between Strontian inclusions and the Lorne lavas. This may be due to mantle heterogeneity and/or crustal contamination within the lavas as Tdm ages for inclusions are similar to those of local mantle nodules (670-950Myr and 600-1200 Myr respectively). The Strontian granitoid inclusions actually are a closer approximation to local mantle nodules than the local contemporaneous mantle-derived lava sequence, a feature which may have significance for determining mantle character in areas where direct samples are not available.

In terms of determining other mantle parameters such as elemental abundance or Sr (or Pb) isotope systematics, inclusions are of limited use due to extensive equilibration of the low ionic charge species. Species which do not readily equilibrate may have been affected by fractionation processes prior to incorporation within the granitoid (see previous section) and it is difficult to predict whether differences within trace element patterns are the product of fractionation or primary features of the mantle source.

Consideration of Sm/Nd and ϵNd histograms (Fig 8.7) for Strontian and Criffell inclusions show clearly that elemental abundance is unrelated to the isotopic ratio. Low Sm/Nd ratios of the inclusions (as low as 0.033) should reflect a LREE enriched mantle source, whilst the isotope systematics reflect a source region which shows the time-integrated effects of LREE depletion. This contradiction can be resolved by considering a mantle enrichment event having taken place ‘just prior’ to magmatism, the fractionation of a Sm-rich phase such as garnet or the disequilibrium melting of accessory phases within the mantle source. The high Nd content of inclusions (up to 160ppm) is not consistent with major garnet removal or residual garnet within the source region. It is likely, therefore, that either an enrichment event took place prior to melt production or disequilibrium melting occurred.

A similar observation has been made for the South-East Australian alkali basalt suite (Menzies & Wass 1983) in which enrichment by CO_2 degassing of the lower mantle was envisaged just prior to the generation of the alkali basalt magma, leading to a decoupling of the isotopes from the trace elements. Alternatively Campbell & Gorton (1980) contend that 'flash melting' of accessory phases within the mantle source will have the same apparent effect of decoupling the trace element and isotopic data. Each process is equally viable with respect to Strontian and Criffell inclusions and cannot be effectively tested here as detailed REE determinations are required (Campbell & Gorton 1980).

High Nd contents for mantle-derived inclusions from the Criffell pluton were not predicted by Halliday et al. (1986) in their appraisal of the Scottish lithosphere, where Nd contents are seen to rise towards the north-west. Inclusions appear to have been derived from a similar isotopic source at Strontian and Criffell against expectation and one explanation is that the lithospheric mantle beneath the two plutons suffered enrichment after closure of the Iapetus Ocean and just prior to or during magmatism, which concurs with the views of MacDonald et al.

(1985) for the north of England lamprophyre dyke suite. Whether this enrichment event triggered mantle melting, or indeed whether both inclusions sets were generated by 'flash melting' events is not clear, but the similarity of the Strontian and Criffell inclusion suites suggests that comparable processes have acted upon the source regions of both plutons prior to and during magma genesis.

Summary

Detailed petrographic, isotopic and chemical analysis has shown that inclusions from the Strontian, Criffell and Ballachulish plutons may be placed into one of several groupings. The presence of disequilibrated Nd isotopes to more radiogenic ∈Nd values in the inclusions, the presence of oscillatory zoned feldspars and the acicular nature of apatites indicates that the vast majority of inclusions within the study plutons were the product of crystallisation of evolving basic mantle-derived magmas. These inclusions are not the relicts after country rock digestion, immiscible liquids or restite from the source region. However a small number of autoliths were identified.

Despite their immersion/injection within hot granitoid liquid, inclusions appear to retain their primary Nd isotope signature, and seem accurately to reflect their mantle source. If this is shown to be correct then Nd isotopic analysis of granitoid inclusions opens up the possibility of geochemically mapping large areas of the lithospheric mantle at a given time in areas where inclusion-rich granitoids are common.

8.3 DISCUSSION.

The origin of inclusions within the Strontian and Criffell granitoids has been shown to originate in the lithospheric mantle. The implication is therefore that mantle melts took an active part in the generation of the host granitoid and have also added to the continental mass. Although the host Strontian, Criffell (and most probably Ballachulish) plutons are not solely the product of intracrustal melting, the amount (if any) and the provenance of the crustal component of these intermediate magmas remain to be determined.

Implications for the Strontian pluton

From field observations the pluton may be described in terms of three units, an outer hornblende and inclusion-bearing tonalite, a porphyritic granodiorite of generally similar aspect, and a central portion of biotite granite of more evolved type (MacGregor & Kennedy 1932, Sabine 1963). Isotopically the outer tonalite and granodiorite portions have restricted initial $^{87}Sr/^{86}Sr$ (0.7053 to 0.7059) and ϵNd (+0.5 to -1.5) whereas the central more evolved portion has higher initial $^{87}Sr/^{86}Sr$ (0.7071 to 0.7073) and lower ϵNd (-3.2 to -6.3).

Zircons extracted from the central portion reveal a U-Pb upper concordia intercept "age" of 1420 Myr +/- 10 Myr (Pidgeon & Aftalion 1979) whilst similar zircons from the outer portions yield no, or at best, insignificantly small Pb isotopic memory.

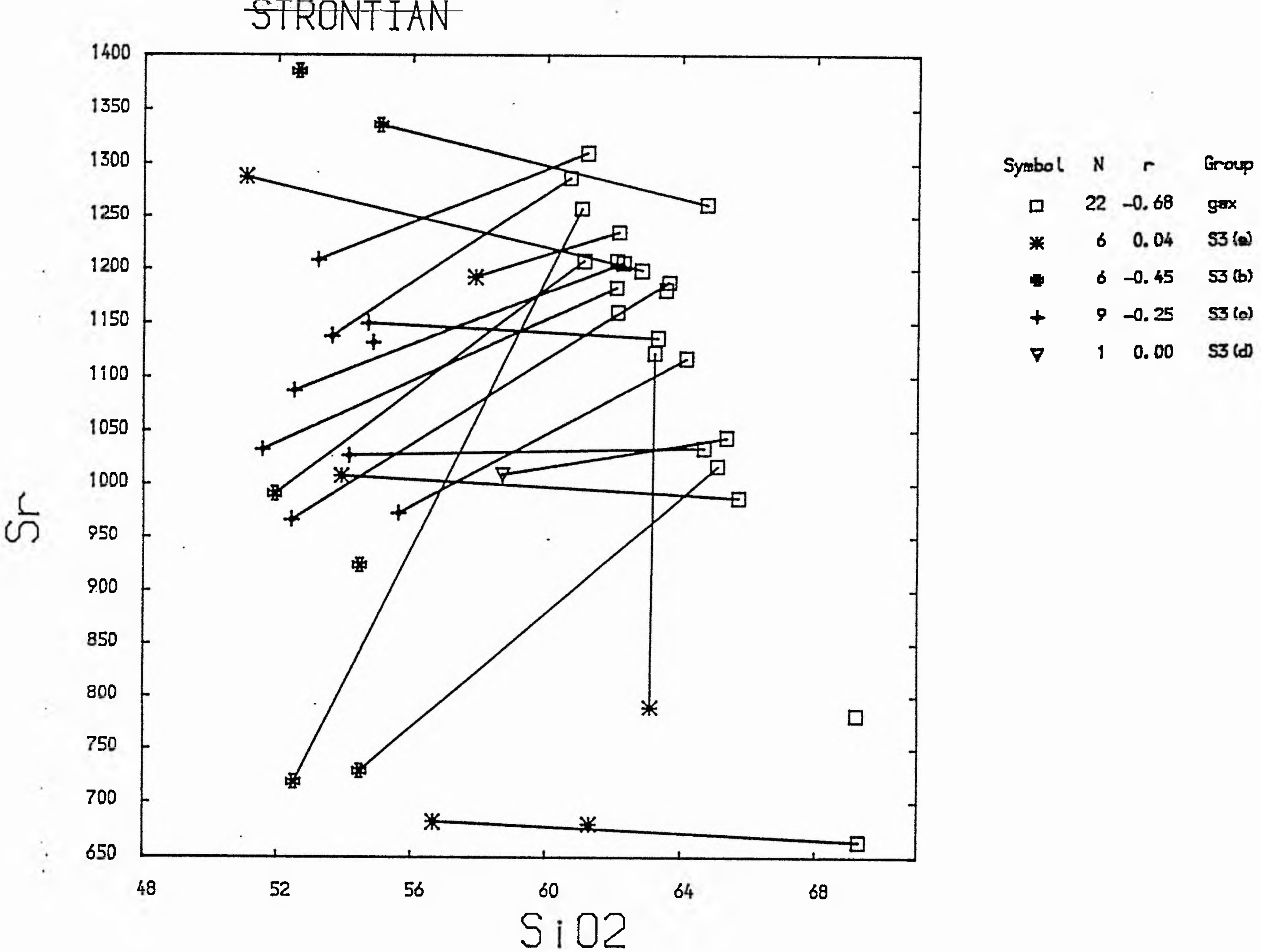


Fig 8.8 Sr against SiO_2; Strontian.

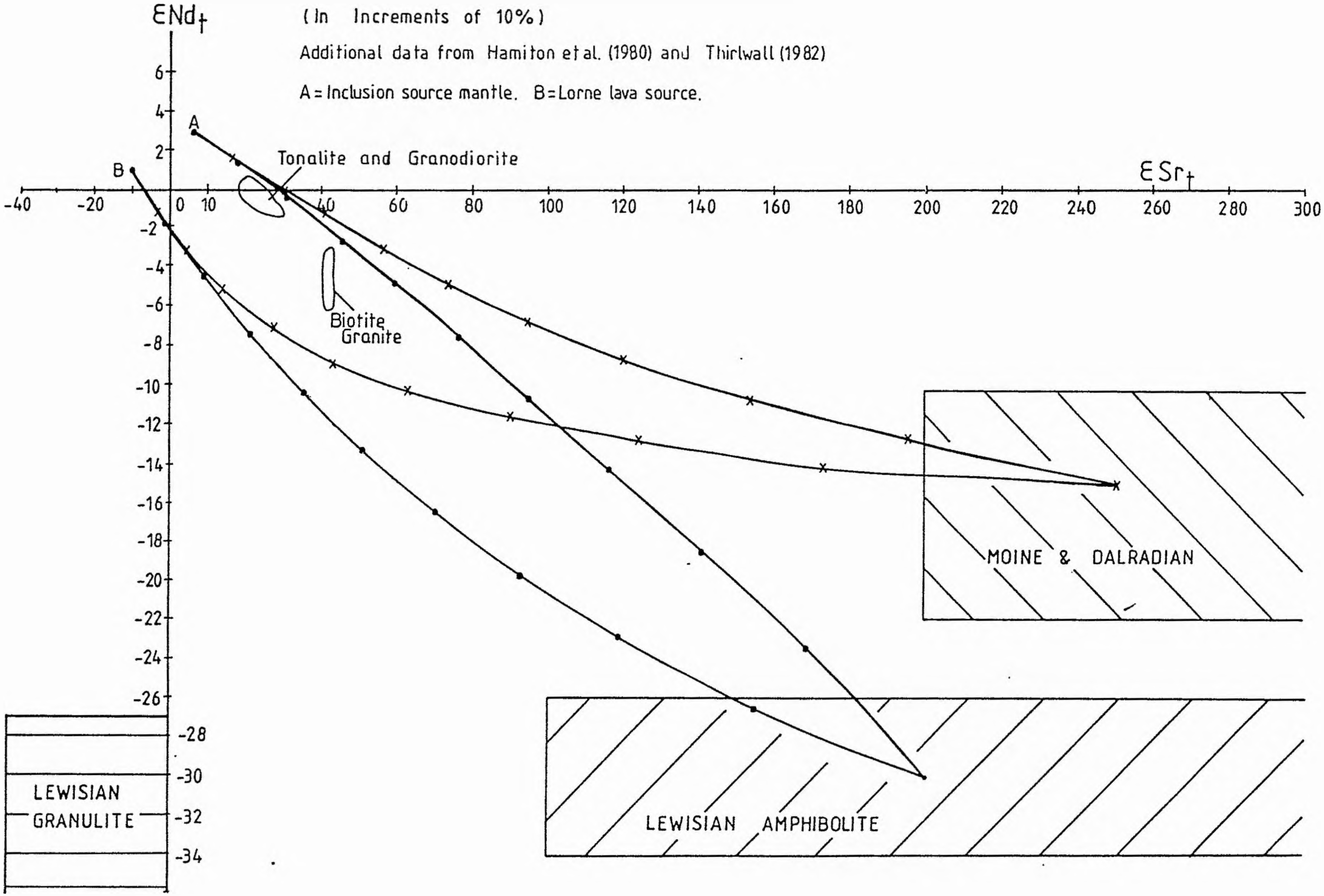
Since zircons are minerals essentially confined to crustal rocks (Halliday et al. 1985), a zircon "age" in excess of that of the pluton can only be attained by the wholesale incorporation of old crustal material, the zircons remaining unresorbed during the assimilation process. The central biotite granite thus has a significant input of old crustal material reflected by the inherited zircon component, higher $^{87}Sr/^{86}Sr$ and lower ϵNd than the outer members. Evolution of the whole complex from direct mantle melting and subsequent isochemical fractionation may be discounted, and a more complex origin involving mantle-derived (inclusions) and crustal-derived (old zircons) magmas must be invoked.

Consideration of the SiO_2-Sr diagram (Fig 8.8) indicates a well-defined trend for the outer tonalite-granodiorite portions, indicative of a continuous zonation pattern. Fractional crystallisation, mixing or restite unmixing processes may be put forward in order to explain the trend. High concentrations of lithophile elements have been cited as powerful arguments for significant crustal involvement in the generation of intermediate plutons but the great abundance of Sr in the Strontian pluton is not easily explained by crustal input as wholesale melting of Moinian or Lewisian material would not be expected to yield such enrichment (Halliday et al. 1985). Partial melting of a crustal source would also yield low Sr magmas as residual plagioclase would retain Sr until significant plagioclase breakdown occurred. It is worthy of note that within the geographically-close Devonian Lorne lava sequence, Mg-rich basalts and basaltic-andesites have high (>1000 ppm) Sr contents indicating that high concentrations of lithophile elements are a mantle as well as a crustal feature (Thirlwall 1982 & 1986, Halliday et al. 1985 for a review of the Scottish lithosphere). Furthermore the lack of inherited zircons within the outer portions of the Strontian pluton constrains the amount of crust within these rocks to be small. Unless selective contamination of Sr from the wall-rock was an important process, the majority of Sr is confined to the mantle or young (< 500 Myrs before magmatism) sub-Lewisian lower crust.

An attempt to constrain the amount of crustal contamination permissible within the Strontian granitoids is given in Fig 8.9. End-member Lewisian amphibolite and Moine metasediment data are the approximate numerical average of several published isotopic analysis, mantle end-members being a magnesium rich basalt from the Devonian Lorne lava sequence (L23 of Thirlwall 1982) and a projected inclusion source composition (see later). Whilst the selection of average end-members is not an entirely desirable method for quantitative analysis, it shows in a qualitative manner the source and possible contamination paths. It is likely that since inclusions have substantially exchanged Sr with their host the original Sr isotope composition was lower than that now observed and was possibly similar to that of the Lorne lavas. Consequently a mean Lorne lavas Sr isotope composition is used, the Nd isotopic signature being that of S210X.

Melts derived from Lewisian granulites played little or no part in the development of the Strontian pluton. A best fit

Fig 8·9 MIXING OF MANTLE & CRUSTAL MATERIAL



(In Increments of 10%)

Additional data from Hamiton et al. (1980) and Thirlwall (1982)

A = Inclusion source mantle. B = Lorne lava source.

binary mixing solution is that between the theoretical inclusion source and Lewisian amphibolite as vectors drawn through the two major Strontian units are parallel to this mixing trajectory. However material derived from Moine metasediments cannot be ruled out as a potential contaminant. A mantle-derived end-member similar to that proposed for the Lorne lavas (Thirlwall 1982) appears an unlikely starting point for mixing with known crustal material as mixing trajectories for Strontian biotite granite are at right angles to those observed.

The amount of Lewisian amphibolite permissible by isotopic modelling in the simplest case (binary mixing) is constrained to 10-20% for the tonalite and granodiorite, and 35-50% for the biotite granite, the granitoids being essentially contaminated mantle-derived melts. The major flaw in such an exercise is the possibility of a contribution to the isotopic inventory from a different sub-Lewisian source. It is worthwhile noting that if a simple mass-balance calculation is performed, using a typical inclusion as the mantle end-member and a ternary minimum melt composition for the crustal end-member, then for the resulting tonalite of 80-90% mantle derivation the SiO_2 content should be between 52-55%. This is some 5-9% lower than that observed for the Strontian tonalite suggesting that significant fractionation must have taken place prior to emplacement, even if the only crustal contaminant was a leucocratic minimum melt. Cumulate piles to calc-alkaline granitoids are seldom found (Pitcher 1978, 1979), an observation which (amongst others) has led many authors to conclude that granitoids intrude through the crust as a crystal mush and are the product of intra-crustal melting (White & Chappell 1977, Holder 1979, Pitcher 1979, Bateman 1985 a & b)).If calc-alkaline granitoids do not yield cumulate piles (and none was observed at Strontian), then the parental magma to the Strontian granitoid may have been of intermediate composition. The low Sr isotopic composition of calc-alkaline volcanics and plutonics of the circum-Pacific belt have been cited as evidence for direct mantle melting to yield intermediate magmas (McNutt et al. 1975, Brown 1977). Cameron & Cameron (1985) using Sr and Nd isotopic evidence on the Cenozoic dacites of California and Mexico found little difference between these and mantle-derived basaltic rocks. Variation within the rocks was attributed to the sub-crustal source, the implication being that these intermediate melts were derived by direct melting of the mantle.

In assessing possible and impossible sources of voluminous intermediate plutonic magmas from an experimental standpoint, Wyllie and co-workers (1976, 1983a & b) considered that tonalitic material within the circum-Pacific batholiths may have been derived from the mantle only under special conditions of high volatile content and low confining pressure (<35Km). They further considered that these tonalites were the product of lower crustal melting, the heat required for melting being supplied by mantle-derived basaltic melts. The Strontian pluton was intruded into a thickened crustal sequence (Ashworth & Tyler 1983) following the closure of the Iapetus Ocean, which suggests that the majority of tonalite was not derived directly from the mantle. If the experimental interpretations are correct then the

Strontian tonalite must contain a significant proportion of lower crustal material.

The lack of an inherited zircon component within the Strontian tonalite and the ϵNd_t values indicate that the lower crustal source was not similar to any of the Lewisian-like rocks observed at the surface. Clayburn et al. (1983) in investigating the origin of the Etive pluton, appealed to an unseen late-Proterozoic (Grenvillian) lower crustal contamination of an original mantle-derived melt to explain their isotopic data, whilst Davies (1984) argued for a substantial Grenvillian basement beneath Scotland. Halliday et al. (1985) have shown that mixing in a three component system (in their case MORB, Lewisian amphibolite and Moine/Dalradian) could adequately explain most of the isotopic variation observed within the Newer Caledonian granites of Scotland and dismisses the idea of a widespread Grenvillian lower crust. Similar three component mixing at Strontian is untenable for the outer tonalite due to the lack of an inherited zircon component unless the crustal zircons were completely assimilated. Similar granitoids to the north of the Mid-Grampian Line (MGL) retain their inherited zircon memory which suggests that zircon inheritance is a robust property during assimilation.

Regional geochemical surveys of the Scottish Caledonian granitoids (Pidgeon & Aftalion 1979, Halliday 1984, Halliday & Stephens 1984, Stephens & Halliday 1984, Thirlwall 1986) have revealed a change in chemistry about the Grampian Highlands (the MGL of Halliday, 1984). In the north granitoids tend to have low

∈Ndt (<-6) and contain inherited zircons, whilst granitoids to the south of the MGL have ∈Ndt >-6 and no inherited zircons. Pidgeon & Aftalion (1979) interpreted the inherited zircon component as unequivocal evidence for an old crustal input, whereas granitoids without such zircons had assimilated (or consisted entirely of) young lower crust, with correspondingly young zircons.

The inference from these observations is that the lower crust to the north of the MGL is old whilst to the south it is young (i.e. Paleozoic) and that granitoids contain a significant amount of lower crust. Three notable exceptions occur to the north of the MGL, namely Cluanie (∈Ndt=+1.3, Halliday 1984), Strontian tonalite and granodiorite (∈Ndt=-1.5 to +0.5) and Ballachulish (∈Ndt=-2.1); none of the above has an inherited zircon component (Halliday et al. 1979) and all three occur within a geographically restricted zone close to the western end of the Great Glen Fault (GGF). If these calc-alkaline plutons do contain a significant portion of lower crust, as suggested by inherited zircons from other contemporaneous Scottish plutons and experimental data, then as has been stated earlier this must be young.

A basic granulite sample (presumed of lower crustal origin) from the Streap Cumlaidh nodule locality (Menzies & Halliday 1984, Halliday et al. 1985) has yielded a young Tdm 'age' (< 1500Myr) and with a +∈Nd in Paleozoic times. Halliday et al. (1986) cite such crust as the possible origin of the Cluanie and Strontian tonalite bodies. A young lower crust of this type is not thought to be widespread, only subcropping within a restricted area around the western end of the GGF (Halliday et al. 1986). Magmas derived by partial melting of such a young lower crust could intrude the mid to upper crust with little assimilation if fault-aided along the GGF. The proportion of mantle-derived component within such a melt would be significantly lower than that indicated by the earlier modelling. Determination of this proportion in a quantitative sense is not possible given the poorly constrained nature of the lower crust and is further hampered by the equilibrated nature of the Sr isotope systematics of inclusions and hosts.

Implications for melt dynamics and crustal growth.

Melting of a (refractory?) basic lower crust presents many thermal problems (Thompson 1985) in that large quantities of additional heat are required. Fusion may be aided by the lowering of the melting point by adiabatic decompression during post closure uplift (Winkler 1976). In regions of rapid uplift and erosion that follow continent-continent collisions, the dominant granitoids produced are of initially peraluminous composition, (often with high initial $^{87}Sr/^{86}Sr$ ratios) and fitting the S-type classification (e.g. Pitcher 1983). Heat necessary for I-type granitoids must be derived from the mantle or by crustal thickening (England & Thompson 1986), which may or may not be aided by secondary factors such as decompression or volatile influx.

The ponding of a large reservoir of basic-ultrabasic magma at the base of the crust would be expected to produce large degrees of partial melting over a small area, but would not be expected to produce batholitic volumes. A plexus of dykes and sheets with great surface area could transfer heat much more effectively, enabling partial melting on a regional scale similar to that envisaged by Thompson et al. (1982) for the contamination of the Skye Main Lava Series. Hildreth (1981) also gives a similar explanation for the production of the intermediate to acid volcanic rocks of the Western U.S.

The observation of inclusion groups with distinctive petrographic, geochemical and isotopic features is consistent with inclusions being the fragmented remains of mantle-derived dykes. Such diversity would not be predicted for a single magma pool. The implication is therefore that inclusions record the very beginnings of granitoid formation and provide the clearest possible link between mantle melting and granitoid production.

Aggregation and upwelling of these lower crustal melts would take place during continued magma input from the mantle. Where basaltic mantle-derived magmas intersect crustal magma pools then the lower temperatures of intermediate magmas would be expected to quench the basalt, possibly producing pillow-like structures. These synplutonic intrusives would further disaggregate as the magma rose upwards to produce the sub-spherical inclusions.

The facts that mantle-derived appinites and lamprophyres cut all three parts of the Strontian pluton and that inclusions are primarily mantle-derived indicate the longevity of mantle input to the crust. This evidence supports a mantle supply of heat for lower crustal melting.

This interpretation of the inclusion data for the Strontian pluton implies that both the distillation of the mantle to form new crust and the differentiation of the crust itself occurred simultaneously. At Strontian in particular, the lower crust must have been young and underplated beneath the Lewisian gneisses. The implication is that this lower crust must itself have been recently derived from the mantle and as such provides a mechanism of crustal growth. The production of the Strontian granitoid has accompanied a significant amount of local crustal growth, although there is no evidence that this has been widespread in the Caledonian of Scotland (but see Thirlwall 1986) by direct melting of the mantle, but by a two stage process of lower crustal accretion followed by intracrustal melting initiated by intrusion of mantle-derived basaltic melts.

Mineralogical and chemical zonation.

Plutons of I-type are commonly zoned often becoming mineralogically and chemically more evolved towards the core (normal zonation). To explain this a variety of hypotheses have been proposed, including restite separation (White & Chappell 1977, Chappell 1984), vapour phase transfer (Vance 1962), crystal-liquid fractionation (Batemann & Chappell 1979, Phillips et al. 1981), melting of a vertically zoned source region (Michard-Vitrac 1979), source melting control (Halliday et al. 1979) or combinations of assimilation and fractional crystallisation (DePaolo 1981a, Halliday 1983, Stephens et al. 1985).

In the cases of Strontian and Criffell, it has already been shown that both plutons are normally zoned, with both continuous and discontinuous changes in mineralogical and chemical compositions. For instance, at Strontian the transition from tonalite to granodiorite is everywhere continuous and gradational (albeit with gradients of compositional change being steeper in some parts rather than others), whilst the change from granodiorite to biotite granite is everywhere sharp and intrusive. It has also been shown earlier in this chapter that the Sr and Nd isotopic compositions of the central portions become more and less radiogenic respectively, a process which cannot be the result of fractional crystallisation alone.

In a series of papers, Stephens, Halliday and co-workers argued that the origin of compositional zonation within the Criffell pluton was twofold:

Firstly, the source of the granitoid changed with time such that the later magmas contained a greater proportion of evolved sources from the mid to upper crust than earlier magmas. The implication is that the region of melting rose in the crust with time.

Secondly, fractional crystallisation occurred within each magma batch to produce complex geochemical variations. These variations have been successfully modelled by Stephens & Halliday (1980) and Stephens et al (1985) using the AFC model.

The outer portions of Strontian do not contain an inherited zircon component, as well as having Sr and Nd isotopes less and more radiogenic respectively than their counterparts within the biotite granite. The presence of inherited zircons resulted from assimilation of old crust (Lewisian) or reworked old crust (Moine). The conclusion is that the zone of melting migrated upward with time, the later melts containing a larger percentage (~50%) of mid to upper crustal material.

This model produces something of a problem. Heat for granitoid production has been shown to be derived from the mantle by the input of basic melts which are recorded as plentiful inclusions. As the more chemically evolved magmas intrude the crust, inclusions are notably lacking indicative of the decreasing importance of mantle melts. The question is, therefore; in the absence of of such basic melts higher in the crust, what drives the melt zone upwards? Crystalline rocks are notably bad conductors of heat, and it is worthy of note that granitoids in general have larger thermal aureoles than their gabbroic counterparts, a function of the aqueous solutions driven out of the granitoid (or recycled meteoric water) during crystallisation. It is likely that the upward migration of the melt zone as clearly demonstrated for the Strontian and Criffell granitoids, is associated with volatile mobilisation, the volatiles being driven off the lower crustal rocks by the input of mantle melts.

The lack of inclusions within the more evolved portions of calc-alkaline plutons is a common observation (Didier 1973, White and Chappell 1977, Pitcher 1978). This paucity is not a function of their increasing assimilation or of having greater elapse time with which to settle out but is a function of the decreased importance of direct mantle melt injection into the granitoid source region. The implication is that the primary control on the zonation of the study plutons is the migration of the melt zone upwards yielding more evolved (isotopically more crustal) magmas (Halliday et al.1979, Halliday 1983).

8.4 SUMMARY

Based on the Nd and Sr isotopic evidence, inclusion suites from the Strontian and Criffell plutons have been compared with potential lithospheric sources in Scotland. At Strontian inclusion isotope ratios closely resemble those of presumed mantle xenoliths exumed at Streap Cumlaidh and it is concluded that microgranitoid inclusions (both at Strontian and Criffell) are derived from a sub-crustal lithospheric source. Inferences about the dynamics and location of crustal melting associated with the Strontian pluton have also been discussed.

Inclusions improve and extend our understanding of the petrogenetic evolution of major calc-alkaline granitoids and provide crucial evidence in the discrimination of the most plausible magma generation model(s) (viz intra-crustal melting restite separation, direct mantle melting plus fractional crystallisation or mixed mantle-derived and crustal-derived magmas)..

CHAPTER 9

DISCUSSION AND CONCLUSIONS

9.1 THE INITIATION AND EVOLUTION OF NORMALLY ZONED CALC-ALKALINE GRANITOIDS: A UNIFYING MODEL

Field, petrographic, geochemical and isotopic studies based on the Strontian, Ballachulish and Criffell plutons have demonstrated that important petrogenetic information may be gleaned from the study of granitoid inclusions. In particular the Nd isotopic signatures of Strontian inclusions are similar to those of supposed local mantle nodules (the isotopic composition of the mantle beneath Criffell is poorly constrained), a strong indication that these inclusions are indeed mantle-derived.

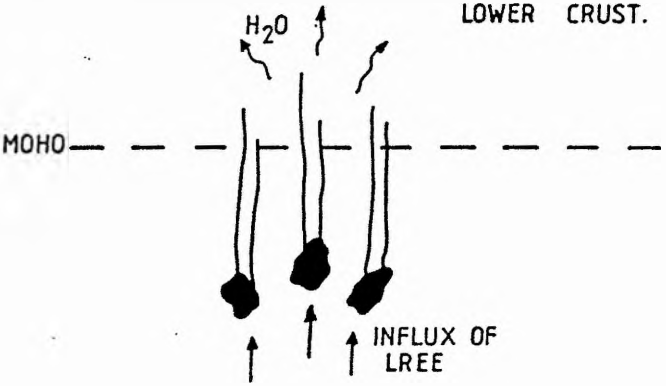
The evolution of the plutons is now presented in the form of an idealised model, represented in cartoon fashion in Fig 9.1.

1) Initiation: Melting took place within the depleted lithospheric mantle soon after an influx of LREE enriched material. The melts (the precursors to inclusions) were injected into the lower crust as a plexus of minor intrusives. Advection of heat from the mantle in this way heated the lower crust allowing the generation of a volatile phase, perhaps through the dehydration of mica and amphibole.

2) Generation of the granitoid: As yet more heat was released into the lower crust, partial melting (possibly non-minimum) may have occurred on a large scale. Pools of crustal melt may have interacted with mantle-derived magmas to form a magma of

Fig 9.1 ORIGIN AND DEVELOPMENT OF THE

1. INFLUX OF LREE INTO A MANTLE SOURCE. MELTING AND INJECTION OF BASALTIC MELTS INTO THE LOWER CRUST.



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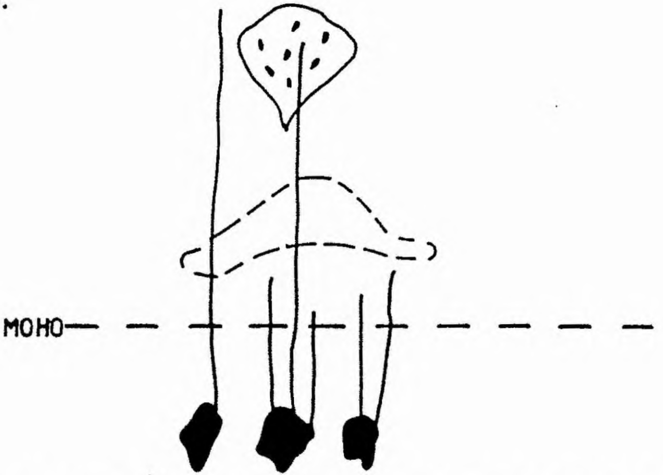
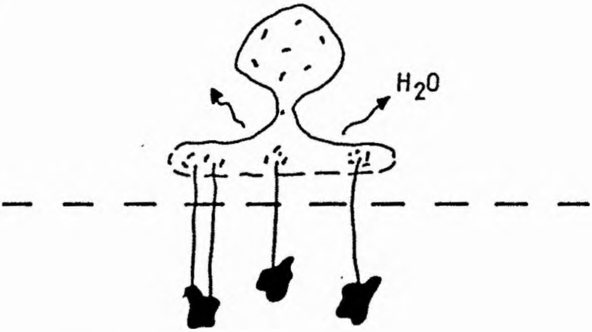


VOLATILES FORCED OFF FROM THE LOWER CRUST PROMOTE MELTING HIGHER IN THE CRUST.

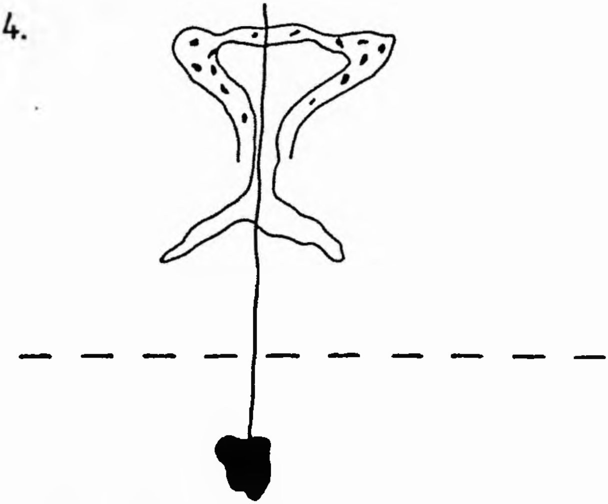
STRONTIAN AND CRIFFELL PLUTONS

2.



THE LOWER CRUST MELTS. FURTHER INJECTION OF BASALTIC MATERIAL YIELDS INCLUSIONS.

4.



MANTLE MELTS WAIN. ONLY OCCASIONAL DYKE (OR APPINITE) EMPLACEMENT OCCURS. CRUSTAL MELTING CEASES.

intermediate composition. Further mantle-melt injection may have been quenched as pillow-like globules within this lower temperature intermediate magma.

The intermediate melt aggregated until it reached plutonic volumes (10's Km^3) when it escaped upwards from its source region. Fractional crystallisation processes began as the intermediate magma intruded colder country rock causing the pulse to become compositionally zoned. The passage of the first pulse warmed the crustal rocks adjacent to the conduit promoting later melting at higher levels.

3) Zonation of the pluton: Continued input of heat into the base of the crust caused temperatures at shallower levels to rise. Volatiles driven off from the lower crust combined with higher temperatures promoted partial melting at higher (mid?) crustal levels. At these shallower depths the direct input of mantle-derived melts was less voluminous than at lower crustal levels and the presence of quenched basic magmas within these essentially crustal melts is rare. Such magmas however may consist of materials derived from three geochemically distinct reservoirs namely mantle, lower crust and (?)mid crust.

Upward passage of these magmas was aided by the existence of a prewarmed conduit presenting a course of easy passage. Compared with the earlier pulse the second pulse was more evolved and its emplacement into the former resulted in the formation of a discontinuously zoned pluton, the constituent pulses being compositionally zoned by processes of fractional crystallisation and/or assimilation.

4) End of plutonism: Mantle melting waned and the fusible portion of the lower crust had essentially been removed, melting higher in the crust also ceased. Dyke emplacement continued but not in volumes sufficient to maintain crustal melting on a large scale.

Implications

The origins of I-type granitoids has centred on essentially three types of model:

1) Granitoids are the mirrors of their crustal source region, the compositional diversity being a function of the type of source rock and degree of restite separation (Chappell & White 1974, White & Chappell 1977, Chappell 1984).

2) Granitoids are essentially mantle-derived melts, the compositional diversity being a function of fractional crystallisation and/or amount of assimilation of country rock (McNutt et al. 1975, Brown 1977, Foland 1985).

3) A mixed origin for granitoids. Initiation of the body within the lower crust by the injection of mantle-derived melts which provide the heat energy necessary for crustal melting. The compositional diversity is a function of the percentage and character of the original melts +/- fractional crystallisation +/- assimilation of country rock materials at high levels within

the crust (Halliday et al. 1980, Stephens & Halliday 1980, Halliday 1983, Wyllie 1983b, Whalen & Currie 1984, Stephens et al. 1985).

Models based on the thermal regimes within thickened orogenic belts have demonstrated that I-type granitoids similar to the Strontian, Ballachulish and Criffell plutons may be generated entirely by intra-crustal melting processes (England & Thompson 1986). In their quantitative modelling, England & Thompson (1986) suggested that tectonically thickened orogenic belts may, after a period of thermal relaxation, give rise to water-saturated melts of S-type which are followed some tens of millions of years later by melts of I-type derived by mica and amphibole dehydration melting of the lowermost portion of the crust. This model might account for the Caledonian granites of Scotland where early S-type plutons in the N and NE Highlands were followed some 60Myr later by the voluminous upwelling of the I-type Newer Granites. However, data derived from the Strontian and Criffell inclusions provide a clear link between mantle melting and granitoid production suggesting that intra-crustal melting required a thermal trigger for plutonism to be initiated. Lower crustal temperatures may well have been high prior to mantle-magma injection and a combination of mantle magmatism, volatile liberation and a high transient geotherm may well explain the sudden burst of plutonism within the Caledonian of Scotland at circa 400Myr mark.

The Strontian and Criffell inclusions and their hosts support model 3 above as being the best approximation to the observations, i.e. that these bodies are the products of crustal melting promoted by an influx of mantle-magma. Inclusion data have shed light on the early stages of pluton evolution and in part constrain their subsequent development. One inference from this study is that the investigation of calc-alkaline granitoids is incomplete without the careful study of their constituent inclusions and minor intrusives. Very few complete studies are to be found in the literature.

9.2 IMPLICATIONS FOR THE CALEDONIAN OF SCOTLAND

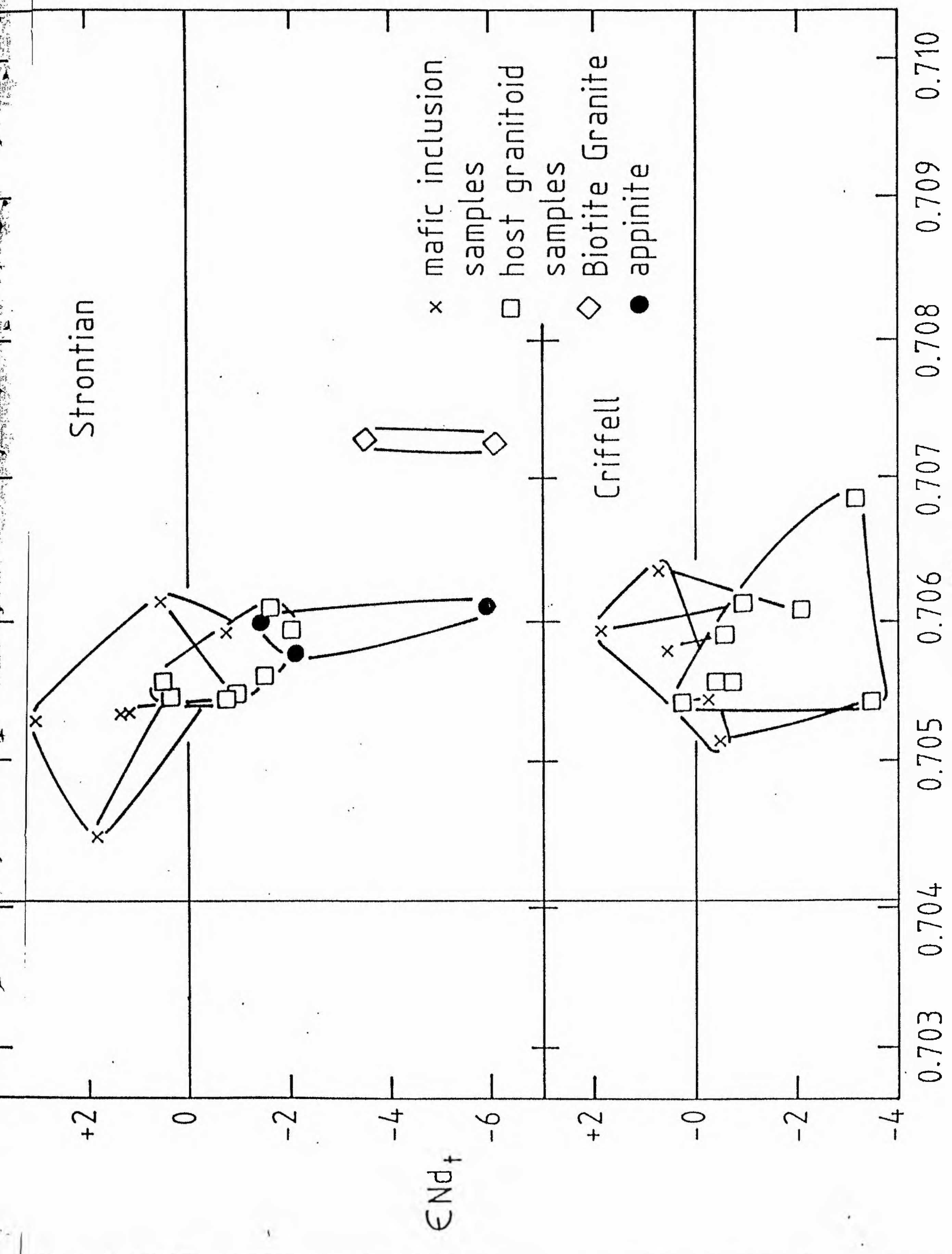
Three major findings as to the origin of calc-alkaline granitoids were put forward in the previous section indicating the importance of different sources in determining the compositional diversity of such intermediate magmas. How these reservoirs identified as end-members fit into the overall Caledonian framework is now explored.

Strontian and Criffell and the evidence for a Grenvillian basement.

Recent regional geochemical and isotopic surveys on the Caledonian granitoids of Scotland (Halliday 1984, Stephens & Halliday 1984, Halliday et al. 1986, Thirlwall 1986) have concentrated on the apparent differences between granitoids north and south of the Mid-Grampian Line (MGL, Halliday 1984). This line is thought to represent a major tectonic suture at depth but has no surface expression, the granitoids to the north consisting of old recycled basement whilst to the south granitoids consist of young crustal components which have had insufficient time to develop comparatively unradiogenic Nd isotopes. In detail however, exceptions do occur particularly to the north where the outer portions of the Cluanie (+1.3), Strontian (+0.6), Ballachulish (-2.3) and Etive (-4.8) plutons all yield ϵNd_t values >-6.

Studies of the Etive complex in detail by Clayburn et al. (1983) and Frost & O'Nions (1985) have indicated the possibility of contributions from an enriched mantle source +/- country rock assimilation or derivation of the complex from a largely (unseen) Grenvillian basement. However, Halliday et al. (1986) suggest isotopic variations within the Etive pluton could be explained in terms of three component mixing (MORB, Lewisian amphibolites and Moine/Dalradian material). Such mixing relationships are unlikely within the Strontian tonalite as it does not contain an inherited zircon component which it surely would if significant amounts of Lewisian +/- Moine/Dalradian material had been assimilated. Furthermore there is no evidence for a MORB-like mantle source beneath northern Scotland (Thirlwall 1982 & 1986, Menzies & Halliday 1984, Halliday et al. 1986). The implication is therefore that the lower crustal contribution to the Strontian granitoid was not Lewisian-like but had a T_{dm} 'age' of 800-1200Myr. The further implication is that this unseen crust is similar to that envisaged by Clayburn et al. (1983) for the Etive

Fig 9.2 Nd-Sr isotopic data for Strontian and Criffell host-inclusion pairs (joined by tie lines). Additional data from Hamilton et al. (1980) and Halliday et al. (1980).

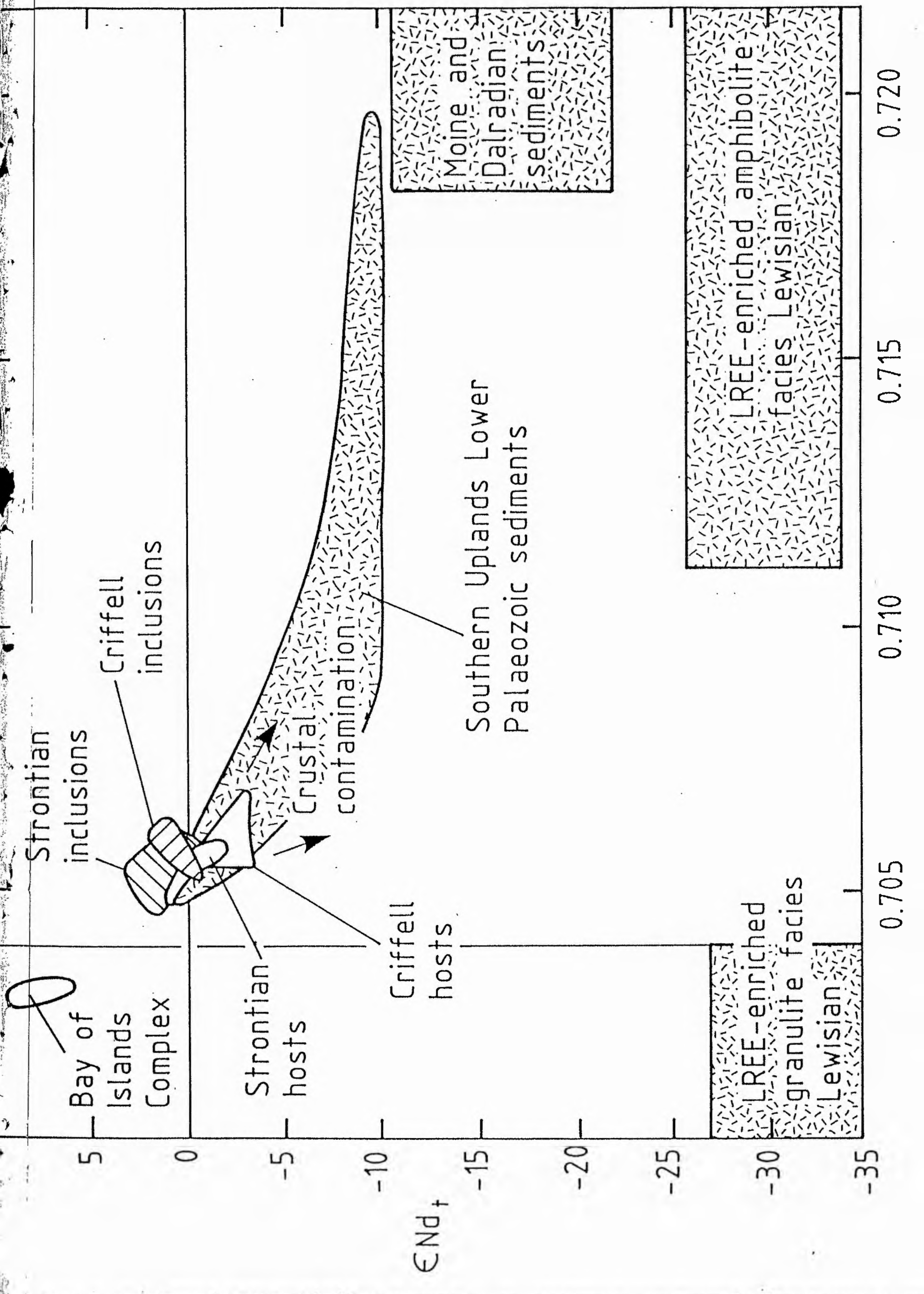


pluton (see also Harmon et al. 1984).

Halliday et al. (1986) report a lower crustal granulite facies xenolith from Streap Cumlaidh with a 'young' age and +ve ϵNd_t in Palaeozoic times. Clearly such a lower crustal source would fit the isotopic requirements for the anomalous plutons without the necessity for a Lewisian contribution.

Comparison of the Strontian and Criffell plutons on the ϵNd_t-$^{87}Sr/^{86}Sr$ diagram (Fig 9.2) is useful especially when considering the age and compositional differences between the crustal blocks into which the plutons were emplaced. Consideration of depleted mantle ages shows the following: inclusions from both Strontian and Criffell have T_{dm} 'ages' of 0.84+/-0.12Gyr and 0.76+/-0.09Gyr respectively, whilst host granitoids yield T_{dm}'s of 0.99+/-0.1Gyr for Strontian and 0.98+/-0.23Gyr for Criffell. Applying the Student t-test, statistically significant differences in the T_{dm} 'ages' are observed between Strontian inclusions and Strontian hosts and between Criffell inclusions and Criffell hosts, whilst no significant T_{dm} 'age' differences are observed between Strontian inclusions and Criffell inclusions or between Strontian hosts and Criffell hosts (at the 95% confidence level). Sm-Nd model ages for granitoids should be interpreted with caution because of REE fractionation during partial melting and because of the possibility of crystallisation and separation of minerals (e.g. monazite) which strongly affect the REE budget and distribution. Nevertheless the most likely explanation is that a substantial

Fig 9.3 Nd-Sr isotopic data for Scottish crustal sources. Additional data from Hamilton et al. (1980) and Halliday et al. (1980).



portion of the Nd in the granitoids was derived from the mantle less than 500Myr before granitoid genesis and that the Nd in the inclusion precursors was derived even more recently. It is intriguing that these two granitoids in such different settings should yield such similar characteristics.

Inspection of Fig 9.3 shows that the isotopic requirements of the Criffell pluton could be fulfilled by the wholesale melting of Southern Uplands Lower Palaeozoic Sediments (SULPS), presumably at depth. However, Halliday et al. (1980), Stephens & Halliday (1980), and Halliday (1983) consider initiation of the pluton was aided by an influx of mantle derived material entering a lower crust similar to SULPS, mixing of the two reservoirs forming an essentially linear array on the $\epsilon Nd_t - {}^{87}Sr/{}^{86}Sr$ diagram (Fig 9.2). These isotopic requirements are equally satisfied by an origin similar to that of the Strontian body, i.e. mantle melts derived from a lithospheric source similar to Streap Cumlaidh mantle nodules, injected into a young (Grenvillian?) lower crust with a T_{dm} 'age' of 800-1200Myr. The similarity between these two plutons means that they could have been derived by the interaction of melts from similar sources, followed by assimilation of mid-crustal rocks of different type. It must be emphasised that whilst the mantle end-member may be constrained to be broadly similar to that of Streap nodules (a MORB source would not satisfy the mixing requirements) as inclusions have a similar range in Nd isotope ratios, the nature of the crustal component cannot be uniquely determined. Either SULPS material or proposed Grenvillian lower crust satisfy these requirements.

Making the assumption that the lower crust beneath Criffell was a young crust of approximately Grenville age raises several interesting points; Firstly, whether Grenvillian basement was widespread beneath Scotland during the Caledonian (Clayburn et al. 1983, Davies et al. 1984, Harmon et al. 1984) or only locally present below Strontian (Halliday et al. 1985) and Criffell. Secondly if such a Grenvillian lower crust is widespread, there are important implications for the MGL. Thirdly, if such a lower crust is widespread why are there no significant outcrops of this material. Fourthly, there are implications for the growth of the Scottish lithosphere.

Grenvillian basement has been postulated beneath Etive, Strontian and Cluanie (Clayburn et al. 1983, Halliday et al. 1986) and below the Midland Valley (Davies et al. 1984). If the Criffell pluton was derived (at least in part) from Grenvillian material, then there is evidence of a young lower crust from the Northern Highlands to the Iapetus suture line. Granitoids derived from the north and particularly the northeast of Scotland have extremely unradiogenic Nd isotopes and inherited zircon components (Halliday 1984), interpreted as reflecting mass assimilation of old (Lewisian) continental crust. Interestingly the same is true of the Strontian central biotite granite where there is strong support for a sub-Lewisian young lower crust. It is possible that major crustal melting has taken place at different depths, as at Strontian and Criffell (Halliday 1983). Unradiogenic Nd isotopes do not preclude a Grenvillian basement but merely indicate that a large proportion of the Nd budget was derived from old crust, perhaps swamping a Grenvillian contribution. If this is the case then the MGL may reflect granitoids to the north being largely derived from melting at shallower mid-crustal levels and their counterparts to the south being derived at depth from a young crustal source.

The implication of a widespread young basement is that the Grenvillian was a major crust-forming event throughout Scotland. Caledonian granitoids, whilst also adding to the crustal mass by the injection of mantle-derived melts (represented by inclusions and basic bodies), were largely recyling pre-existing (Grenvillian(?) and Lewisian) crust.

It must be emphasised again that the evidence for widespread Grenvillian basement is far from conclusive. A strong case may be put forward for a contribution from young lower crust for the Strontian, Etive, Cluanie and Ballachulish plutons suggesting a restricted subcrop of Grenvillian-type around the northwestern end of the Great Glen fault, elsewhere evidence is tentative.

Nappe formation to the north of the Midland Valley suggests that the lithospheric keel was irregular throughout Scotland at the time of granitoid formation and further speculation as to the regional implications of a Grenvillian basement are unwarranted until detailed isotopic studies have revealed the extent of variations within individual plutons from the northeast. It is noteworthy that few outcrops of Grenvillian-aged material are observed but in their vicinity (e.g. Ardgour, Brook et al. 1976), the local granitoid plutons contain evidence of a young lower crustal source.

The implication from the above discussion is that although regional geochemical and isotopic studies provide a unifying overview of the building of the Scottish lithosphere (Stephens & Halliday 1984, Halliday et al. 1986) our understanding is far from complete and detailed investigations of calc-alkaline granitoids and their inclusions should help to constrain regional models.

9.3 FURTHER RESEARCH

It has been established in this thesis that detailed investigations of host-inclusion relationships yield valuable evidence as to the initiation and compositional diversity of calc-alkaline granitoids. However areas of uncertainty remain which with continued study could be further constrained. These areas include:

(1) The degree to which Nd isotopes may equilibrate between host-inclusion pairs.

(2) Whether all calc-alkaline granitoids contain an element of contemporanous mantle-derived magma.

Evidence from the Strontian and Criffell plutons indicates that inclusions retain their primary Nd isotopic signature. However due to the uncertainty of the isotopic composition of the local mantle beneath Scotland, partial equilibration of Nd isotopes cannot be ruled out. Confirmation of this relationship may be sought from the investigation of calc-alkaline granitoids whose local mantle is well characterised and known to be different from that of the host granitoid. Substantial equilibration would be expected to yield inclusions with similar isotopic compositions to their host and not to local mantle.

Nd, Sr and O isotopic studies on the Penninsular Ranges and Sierra Nevada batholiths have demonstrated the increasingly 'crustal' isotopic signature of the granitoids to the east of the province (reviewed by Farmer & DePaolo 1983 and Domenick et al. 1983). Such shifts have been correlated with changes in the nature of the crustal component within the granitoids, magmas to the east assimilating older unradiogenic Nd, the mantle component remaining relatively homogeneous and similar to that of island arcs. If the major conclusions of this thesis are broadly applicable elsewhere then isotopic studies on host inclusion-pairs from a traverse through the Penninsular Ranges and Sierra Nevada batholiths would be expected to reveal inclusions that have near constant Nd isotopes (which are similar to local mantle) whilst host Nd isotopes should become progressively less radiogenic. Furthermore, this would directly confirm the relationship between mantle magmatism as a trigger for crustal plutonism.

The concept of I-type and S-type granitoids were developed to describe the intermediate plutonics of the Lachlan Fold belt SE Australia (reviewed by Chappell 1984). Unlike Scotland or the Sierra Nevada, granitoids have seemingly evolved without major mantle magma input, inclusions being interpreted as restite. Detailed study of Scottish inclusions suites have identified only a small percentage of inclusions as restitic or cognate, the majority having an exotic origin. The simple concept of intracrustal melting of differing sources to yield compositionally different granitoids may be critically examined by detailed study of Australian inclusion suites. Such a study would confirm or reject the simple restite separation hypothesis for granitoid development and would more closely constrain the nature and development of the southeast Australian granitoids.

Detailed study of inclusions present the opportunity to test directly evolutionary models of granitoid production whilst data from inclusions may be used to infer local mantle conditions.

9.4 CONCLUSIONS

1) Inclusions are derived from several sources:

Based on field and petrographic evidence (Chapters 2 & 5), inclusions may be grouped into one of three major inclusion categories: type 1 metasedimentary inclusions, type 2 synplutonic (autolithic) inclusions or type 3 inclusions of exotic origin. Within the study plutons type 3 inclusions dominate and this category may be further sub-divided depending on the nature of the megacrysts. Type 1 inclusions were only abundant within the outer portions of the passively emplaced Ballachulish pluton.

2) Microgranitoid inclusions as quenched magma globules:

The presence of oscillatory zoned plagioclase, acicular apatites and the interlocking texture leads to the conclusion that these inclusions are the quenched remains of an evolving basic-intermediate magma. Subsequent reaction with the host granitoid may cause grain boundary readjustment (which close to the pluton periphery may be considerable) causing a clouding of the feldspars and uralitisation of pyroxenes, but does not affect the morphology of the apatites. Where flattened, the inclusions have not behaved in a manner rheologically different from that of the host granitoid suggesting that a pore fluid existed well after the initial quenching.

3) Appinites as mantle-derived melts:

Appinites within the Strontian pluton are the remnants of pipes transferring mantle-derived material to higher levels. They are not remobilised cumulates from the host granitoid, but are separate mantle-derived magmas which fractionated prior to injection into the Strontian pluton. Since appinites are injected into all three portions of the pluton and since those on the periphery of the body have a strong foliation due to the incoming of the later granitoid pulses, it appears that mantle magma injection took place over the entire plutonic emplacement history of the pluton.

The geochemistry of these appinites may be explained in terms of fractional crystallisation of a mantle-derived melt at depth followed by marginal interaction with the host granitoid during emplacement. Flow through the conduit was probably laminar after an initial volatile-charged chaotic injection.

4) Element equilibration between host-inclusion pairs:

Theoretical modelling and geochemical analyses (Chapter 6), demonstrate that the exchange of elements between host and inclusions (and vice-versa) depends on the size of the inclusion, its grain size, the presence or absence of a pore fluid, the charge and size of the species of interest and the character of the constituent minerals within the inclusion. Elements of low charge (<+3) will readily equilibrate for inclusions of 20cm diameter in around 10^5 years, whilst elements of charge +3 or greater may equilibrate in the presence of a pore fluid if the grainsize is small enough (i.e. <.05mm) otherwise only partial equilibration (at best) will occur. For a given charge smaller ions will equilibrate more readily than larger ones.

Geochemical investigation agrees with theory in that +2 ions (e.g. Sr and Pb) are largely equilibrated between host-inclusions pairs whilst +3 ions (e.g. La & Ce) are not. Those elements that have a charge greater than +2 are considered immobile.

5) Isotopic equilibration between host-inclusion pairs:

Isotopic systems based on elements whose ions have a charge of +2 or less (e.g. Rb & Sr) have equilibrated between host-inclusion pairs (as predicted) whilst isotope systematics based on higher charge species (e.g. Sm & Nd, +3) show a consistent disequilibrium of 0-3 ϵNd units to more radiogenic Nd isotope values. Despite the small minority of autoliths which have similar 'immobile' element and isotopic characteristics to their host rocks, microgranitoid inclusions and supposed mantle-nodules from Streap Cumlaidh (Menzies & Halliday 1984) have similar Nd isotopes. It is concluded that type 3 inclusions are derived from the quenching of lithospheric mantle melts. Such inclusions are not restites, autoliths, partially digested country rock or immiscible magma fragments.

6) Source of inclusions:

Strontian inclusions (other than minor autoliths) were derived from a lithospheric mantle source similar to Streap Cumlaidh nodules. Microgranitoid inclusions form clusters on 'immobile' element-element diagrams suggesting either an heterogeneous source region or separate sources. Geochemical data are consistent with an origin for inclusions by the repeated tapping of a heterogeneous mantle source showing the time integrated effects of LREE depletion. Ponding of such magmas may have occurred at depth and fractionation processes may have allowed the formation of megacryst phases prior to injection into the host granitoid. The LREE enriched nature of the nodule suite suggests that the mantle source was flushed by LREE rich fluids just prior to melting.

7) Inclusions as petrogenetic tracers.:

Field, petrographic, geochemical and isotopic studies on inclusions suites within the Strontian, Ballachulish and Criffell plutons have demonstrated that useful petrogenetic information may be provided by inclusions and additional constraints placed on the initiation, evolution and emplacement of calc-alkaline granitoids. Geochemical and isotopic analysis has shown the complex contributions made by the upper mantle in the triggering of granitoid production. Furthermore, those studies of host and inclusions or acid-basic magma complexes which depend on $^{87}Sr/^{86}Sr$ isotopic as critical evidence (e.g. Bender et al. 1982) must, in view of the findings within Chapters 6 be treated with caution.

Inclusions from within the Strontian and Criffell plutons have Nd isotope ratios similar to local mantle nodules (although the mantle beneath Criffell is not well constrained) apparently remaining unaffected by interaction with the host. Inclusions may therefore be used (with caution) to estimate mantle conditions beneath calc-alkaline plutons and extend the range of available mantle-derived samples. Inclusion studies may facilitate the mapping of the subcrustal lithosphere over large areas where direct samples are not available.

APPENDIX 1

SAMPLE LOCALITIES AND MODAL ANALYSES.

Sample localities.

Samples were collected as host-inclusion pairs from all plutons. Sample first digits refer to the pluton (e.g. S= Strontian etc), whilst the next three define the sample number. The last digit (not given in the tables below) defines whether the sample is an inclusion (-X) or its immediately adjacent host (-G).

STRONTIAN

Sample	Host	Inclusion Type	Locality	Grid Ref (NN-)
S203	Gd	S(3a)	Road cut A861	846610
S210	T	S(3a)	N. Shore Loch Sunart	786612
S212	T	S(3a)	N. Shore Loch Sunart	786612
S341	Gd	S(3a)	Road cut A884	806603
S343	Gd	S(3a)	Road cut A884	818603
S205	T	S(3b)	N. Shore Loch Sunart	846610
S208	Gd	S(3b)	N. Shore Loch Sunart	786612
S211	Gd	S(3b)	N. Shore Loch Sunart	786612
S213	Gd	S(3b)	Road cut A884	781595
S340	Gd	S(3b)	Road cut A884	806603
S344	Gd	S(3b)	Road cut A884	818603
S345	Gd	S(3b)	Road cut A884	818603
S346	Gd	S(3b)	Road cut A884	818603
S347	Gd	S(3b)	Road cut A884	796602
S349	Gd	S(3b)	Road cut A884	796602
S351	T	S(3b)	Road cut A884	743507
S353	Gd	S(3b)	Road cut A884	802609
S354	Gd	S(3b)	Road cut A884	802609
S202	Gd	S(3c)	Road cut A861	846610
S209	T	S(3c)	N. Shore Loch Sunart	786612
S348	Gd	S(3c)	S. Shore Loch Sunart	818603
S338	Gd	S(3c)	Hill Top quarry	756771
S352	Gd	S(3c)	Road cut A884	802609
S355	Gd	S(3c)	Road cut A884	802609
S401	BG	S(3a)	Waterfall	751448
S402	BG	Granodiorite	Waterfall	751448
S338	BG	S(1)	E. of Loch Tearnait	756462

Host granitoid nomenclature; T= tonalite, Gd= porphyritic granodiorite, BG= biotite granite. Inclusion group definitions and descriptions are given in Tables 5.1 to 5.4.

CRIFFELL

Sample	Host	Inclusion type	Locality	Grid Ref (NX-)
C206	Gd	C(3a:1)	Quarry, Craignair	817608
C208	Gd	C(3a:1)	Quarry, Craignair	817608
C210	Gd	C(3a:1)	Quarry, Craignair	817608
C333	Gd	C(3a:1)	Quarry, Craignair	817608
C335	Gd	C(3a:1)	Quarry, Craignair	817608
C209	Gd	C(3a:2)	Quarry, Craignair	817608
C301	HGd	C(3a:2)	Shore section	825496
C328	Gd	C(3a:2)	Quarry, Craignair	817608
C331	Gd	C(3a:2)	Quarry, Craignair	817608
C354	HGd	C(3a:2)	Old quarry	841554
C201	HGd	C(3b:1)	Blocky outcrop	952569
C202	HGd	C(3b:1)	Blocky outcrop	952569
C204	HGd	C(3b:1)	Blocky outcrop	955591
C205	HGd	C(3b:1)	Blocky outcrop	955591
C319	Gd	C(3b:1)	Quarry, Cowpark	827618
C322	HGd	C(3b:1)	Quarry, New Abbey	961647
C324	HGd	C(3b:1)	Quarry, New Abbey	961647
C325	HGd	C(3b:1)	Quarry, New Abbey	961647
C326	HGd	C(3b:1)	Quarry, New Abbey	961647
C341	HGd	C(3b:1)	Quarry, Lotus	897685
C343	HGd	C(3b:1)	Quarry, Lotus	897685
C345	HGd	C(3b:1)	Quarry, New Abbey	961647
C346	HGd	C(3b:1)	Quarry, New Abbey	961647
C350	HGd	C(3b:1)	Quarry, New Abbey	961647
C304	HGd	C(3b:2)	Shore section	818505
C318	HGd	C(3b:2)	Road cut A710	910559
C320	HGd	C(3b:2)	Quarry, New Abbey	961647
C321	HGd	C(3b:2)	Quarry, New Abbey	961647
C323	HGd	C(3b:2)	Quarry, New Abbey	961647
C336	Gd	C(3b:2)	Quarry, Craignair	817608
C337	Gd	C(3b:2)	Quarry, Criagnair	817608

Host granitoid nomenclature; Gd= hornblende-biotite granodiorite HGd= clinopyroxene-hornblende-biotite granodiorite. Inclusions are as described in Tables 5.1 to 5.4.

BALLACHULISH

All samples analysed from the Ballachulish pluton were collected from one of three disused quarries just off the main Ballachulish-Kentallen road grid reference NN022591.

Strontian modal analyses.

	S355X	S355G	S210X	S210G	S343X	S354X	S338X
Rock type	S(3c)	Gd	S(3a)	T	S(3a)	S(3b)	S(3d)
plagioclase	37.3	48.1	21.1	31.1	50.8	32.7	62.1
hornblende	33.0	12.1	46.9	24.6	11.2	29.0	14.7
cl-pyroxene	-	-	-	-	-	-	-
quartz	1.0	13.2	1.8	13.4	4.1	7.8	7.7
biotite	23.0	9.4	17.6	14.8	14.8	13.9	8.4
K-feldspar	3.3	15.1	11.0	13.8	11.0	12.8	5.1
opaques	1.0	1.4	<0.2	0.3	5.1	<0.2	0.7
accessories	1.4	0.5	1.2	2.1	2.8	1.9	1.0

	S338G	S401X	S401G	S438G
Rock type	Gd	S(3a)	BG	BG
plagioclase	49.0	50.2	47.4	51.0
hornblende	10.5	-	-	-
cl-pyroxene	-	-	-	-
quartz	18.6	26.2	25.0	21.4
biotite	6.6	10.6	5.8	4.5
K-feldspar	13.1	11.8	21.2	22.2
opaques	0.9	0.3	0.2	0.3
accessories	0.8	0.6	0.3	0.3

Samples with suffix -G are host granitoids whilst those with suffix -X are inclusions. Host sub-types are; T= tonalite, Gd= porphyritic granodiorite, whilst inclusion sub-types are as described in Tables 5.1 to 5.4.

Ballachulish modal analyses.

	B201X	B201G	B315X	B314X	B321X
Rock type	B(1)	QD	B(3a:1)	B(3a:2)	B(1)
plagioclase	52.6	55.4	50.4	40.6	36.8
hornblende	15.6	8.0	29.8	28.7	-
cl-pyroxene	0.9	0.4	1.0	7.7	-
quartz	2.8	10.6	5.5	4.1	24.6
biotite	9.6	13.6	9.7	13.6	7.6
K-feldspar	16.5	10.4	1.1	2.3	27.8
opaques	0.4	1.2	0.4	0.9	0.2
accessories	1.6	0.4	2.1	2.1	1.1

Sample nomenclature is similar to that above, where QD= quartz diorite, and inclusion sub-types are as described in Tables 5.1 to 5.4.

Criffell modal analyses.

	C201X	C340X	C324X	C328X
Rock Type	C(3b:1)	C(3B:2)	C(3a:1)	C(3a:2)
plagioclase	51.1	38.8	48.9	47.5
hornblende	21.8	27.0	22.7	3.1
cl-pyroxene	1.5	4.1	0.8	-
quartz	9.5	7.1	8.0	10.5
biotite	10.2	10.3	11.1	16.6
K-feldspar	1.5	8.8	6.6	16.8
opaques	1.6	1.2	0.4	1.8
accessories	2.8	2.3	1.4	3.8

Sample nomenclature as per Strontian; inclusion sub-types are described in Tables 5.1 to 5.4.

Appinites, modal analyses.

	S404	S406	S313	S421	S318	S336	S427	S424
Appinite-	T1	T1	T1	T2	T2	PG	PG	BG
Position	ma	co	co	co	co	co	ma	ma
plagioclase	14.0	26.6	31.8	12.3	1.5	23.3	25.6	11.0
hornblende	45.0	41.0	50.8	65.0	31.2	47.3	32.6	21.3
cl-pyroxene	-	-	-	-	-	-	-	-
quartz	2.0	2.6	0.3	0.2	1.2	0.4	1.1	0.6
biotite	4.0	18.8	14.0	14.0	48.4	22.4	22.8	40.8
K-feldspar	18.0	6.7	1.8	1.8	12.7	4.0	13.6	14.7
opaques	0.9	0.4	<0.2	1.4	0.6	0.3	0.9	0.6
accessories	2.3	3.6	1.0	1.4	4.4	2.0	3.4	5.3

Modal analysis of Strontian appinites (see Fig 4.2 for localities), where co= core of appinite and ma= margin.

APPENDIX 2

ANALYTICAL DATA FOR HOST-INCLUSION PAIRS.

Strontian granitoid and inclusions

Sample	S205G	S206G	S208G	S209G	S210G	S211G	S213G	S329G	S331G	S338G	S340G	S341G	S343G	S344G	S345G	S346G
SiO2	69.03	62.11	61.19	61.09	61.02	60.69	64.68	69.22	63.20	65.36	62.13	62.81	64.75	63.53	62.28	63.30
TiO2	0.46	0.82	0.80	0.87	0.91	0.84	0.60	0.30	0.74	0.56	0.74	0.81	0.78	0.75	0.81	0.81
Al2O3	15.46	16.46	16.83	16.30	16.77	17.02	15.74	16.99	16.67	15.86	16.75	16.12	15.78	16.41	16.49	16.22
Fe2O3	2.63	4.51	4.76	4.85	4.94	4.89	3.93	1.80	4.00	3.68	4.53	4.41	4.09	4.12	4.42	4.34
MnO	0.06	0.07	0.07	0.07	0.06	0.07	0.07	0.03	0.06	0.06	0.07	0.07	0.06	0.07	0.07	0.07
MgO	1.44	2.86	3.22	3.30	3.43	3.09	2.57	0.56	2.53	2.20	2.92	2.99	2.57	2.67	2.85	2.87
CaO	2.10	3.99	4.48	4.43	4.40	4.23	3.37	2.72	3.66	3.34	4.27	4.04	3.42	4.14	4.35	4.19
Na2O	4.49	4.57	4.88	4.47	4.79	5.18	3.54	4.29	4.58	4.52	4.93	4.68	4.12	5.49	4.60	4.96
K2O	3.44	2.91	2.72	2.97	2.60	2.60	4.32	3.02	3.17	3.52	2.60	2.88	3.41	1.83	2.80	2.79
P2O5	0.16	0.33	0.31	0.34	0.37	0.37	0.32	0.08	0.31	0.26	0.34	0.38	0.25	0.34	0.35	0.33
loss	0.40	1.00	0.40	0.60	0.40	0.80	0.60	0.80	0.80	0.40	0.80	0.60	0.60	0.60	0.80	0.40
Total	99.87	99.92	99.97	99.60	99.99	100.08	100.05	100.01	100.00	100.04	100.37	100.08	100.08	100.19	100.11	100.56
Nb	8	10	9	12	13	9	8	7	13	9	10	12	9	10	10	15
Zr	153	200	247	247	204	208	178	113	179	181	199	205	209	193	195	198
Y	9	13	15	16	16	16	16	16	15	11	12	19	12	14	17	15
Sr	664	1160	1310	1208	1258	1287	1034	782	1122	1044	1235	1199	1262	1181	1206	1136
Rb	92	52	48	53	49	44	65	81	63	58	45	49	54	36	42	44
Th	6	2	13	9	6	8	5	8	8	8	4	9	5	0	7	8
Pb	20	17	21	20	17	12	19	16	19	19	15	16	14	17	17	13
Zn	47	61	66	70	71	69	55	31	59	55	66	59	14	55	15	12
Cu	18	13	9	12	11	13	17	5	11	13	16	10	59	20	62	62
Ni	19	53	54	60	60	56	43	14	45	42	52	53	44	48	49	49
Cr	25	78	81	91	87	93	49	10	61	53	75	69	62	63	67	67
V	44	82	99	95	113	95	79	29	74	70	83	87	143	74	83	85
Ba	800	1048	981	1061	983	908	1386	748	965	1053	894	974	416	503	975	963
Hf	5	7	6	5	6	8	6	5	5	5	6	8	8	8	7	7
Ce	56	96	94	100	101	116	67	60	102	82	128	92	147	114	100	101
La	35	53	50	50	48	46	32	33	60	52	42	58	87	38	57	58

Sample	S347G	S348G	S351G	S352G	S353G	S354G	S205X	S207X	S208X	S209X	S210X	S211X	S212X	S213X	S331X	S338X
SiO2	62.06	62.08	65.72	65.02	63.61	64.15	56.67	61.29	53.20	51.92	52.51	53.60	52.60	54.13	63.10	58.72
TiO2	0.85	0.87	0.76	0.65	0.70	0.69	1.12	0.97	1.48	1.21	1.21	1.33	1.45	1.38	0.80	0.09
Al2O3	16.41	16.13	15.93	15.98	16.71	15.90	17.68	16.26	17.22	16.13	12.69	17.47	18.09	17.63	15.69	15.20
Fe2O3	4.67	4.47	3.67	3.50	3.94	3.83	6.13	4.70	8.11	8.77	7.66	7.72	8.62	7.66	4.37	6.50
MnO	0.08	0.07	0.06	0.05	0.06	0.06	0.14	0.09	0.13	0.15	0.15	0.12	0.12	0.11	0.06	0.12
MgO	3.13	3.17	2.26	2.35	2.61	2.36	4.07	2.99	5.37	6.63	9.33	5.35	5.17	4.46	3.01	4.29
CaO	4.34	4.03	3.97	3.67	3.95	3.95	4.81	4.06	6.73	6.99	7.36	6.10	6.67	5.92	3.27	5.40
Na2O	4.94	5.10	4.21	4.75	4.85	4.78	4.96	3.66	5.13	4.58	2.67	5.12	4.72	5.18	5.23	5.12
K2O	2.57	2.70	2.77	2.88	2.90	2.75	2.55	3.97	2.13	2.21	3.96	2.23	2.01	2.09	3.02	3.02
P2O5	0.35	0.36	0.21	0.29	0.25	0.27	0.36	0.34	0.51	0.60	0.63	0.51	0.51	0.39	0.27	0.52
loss	0.40	0.80	0.00	0.60	0.20	1.40	1.20	1.60	0.20	0.80	1.80	0.60	0.20	0.80	1.20	1.40
Total	100.09	100.09	99.84	99.99	100.06	100.41	99.90	100.16	100.50	100.26	100.39	100.44	100.48	99.98	100.26	100.61
Nb	15	12	20	9	10	12	24	12	22	10	8	16	10	15	13	8
Zr	229	214	184	160	182	190	274	271	289	236	321	285	244	302	229	181
Y	22	14	26	11	17	17	21	19	35	21	23	30	21	17	14	14
Sr	1183	1208	987	1017	1188	1117	682	680	1209	991	719	1138	1386	1027	790	1009
Rb	47	52	50	45	46	47	114	119	53	64	72	60	51	68	103	81
Th	11	0	5	3	6	10	13	17	12	4	14	9	3	16	15	0
Pb	13	13	6	14	19	20	25	33	19	15	17	20	22	18	22	15
Zn	12	66	48	48	56	53	111	71	105	125	86	107	100	94	70	95
Cu	66	21	28	17	16	16	31	22	52	47	14	48	370	107	30	85
Ni	56	58	36	50	46	47	68	42	84	133	263	107	37	54	35	68
Cr	85	84	52	52	65	65	136	77	116	274	749	203	30	55	92	60
V	93	78	70	72	83	76	100	101	166	166	166	148	184	145	91	122
Ba	863	1094	1200	877	931	844	360	688	427	426	1528	455	475	197	747	391
Hf	7	8	6	7	7	5	6	6	6	4	7	5	9	6	7	4
Ce	147	114	98	61	85	86	101	95	214	143	147	161	155	99	91	83
La	60	54	33	37	38	55	43	50	97	68	71	80	66	59	54	57

Sample	S340X	S341X	S342X	S343X	S345X	S346X	S347X
SiO2	57.87	51.06	54.85	55.04	52.50	54.69	51.56
TiO2	1.04	1.99	1.36	1.43	1.66	1.30	1.46
Al2O3	17.11	17.99	17.30	19.47	17.95	18.22	18.13
Fe2O3	5.93	9.60	7.10	6.57	8.11	6.73	8.57
MnO	0.10	0.07	0.12	0.14	0.10	0.10	0.12
MgO	4.08	4.25	4.68	2.64	4.83	4.43	5.70
CaO	5.04	5.42	5.71	3.94	5.08	5.20	4.94
Na2O	4.90	5.29	5.26	6.23	5.04	5.78	5.22
K2O	2.29	2.59	2.59	3.15	2.81	2.16	3.10
P2O5	0.39	0.90	0.64	0.63	0.89	0.49	0.54
loss	1.00	0.60	0.40	0.40	0.80	1.00	0.40
Total	100.00	100.13	100.29	99.91	100.08	100.34	99.99
Nb	11	26	16	6	11	12	11
Zr	218	359	263	427	281	268	272
Y	12	24	19	9	12	15	14
Sr	1193	1288	1132	1336	1088	1150	1033
Rb	72	86	56	65	90	64	103
Th	6	3	7	1	3	1	10
Pb	16	15	14	18	20	16	14
Zn	87	104	88	73	160	91	40
Cu	94	253	90	73	101	33	126
Ni	66	24	74	30	65	60	90
Cr	114	28	133	17	70	89	117
V	120	91	137	151	83	132	166
Ba	362	1304	589	345	894	365	358
Hf	6	5	7	5	6	7	5
Ce	96	95	141	100	128	82	71
La	49	42	57	34	42	42	49

S348X	S351X	S352X	S353X	S354X	S355X	S214S	S215S
52.23	53.90	54.47	52.43	55.62	54.45	63.14	61.74
1.35	1.45	0.92	1.39	1.10	1.13	0.87	0.92
16.81	18.02	14.51	17.51	16.61	16.44	16.45	16.48
8.58	7.13	8.72	8.33	6.50	8.05	6.49	6.78
0.06	0.12	0.17	0.12	0.11	0.14	0.11	0.11
6.07	4.39	7.44	5.87	4.85	5.94	2.13	2.46
5.67	5.69	6.96	5.21	5.28	5.73	2.79	3.50
4.09	5.22	4.30	4.65	5.42	4.24	3.49	3.14
2.79	2.12	1.92	3.06	2.39	2.69	2.93	3.48
0.42	0.38	0.38	0.59	0.52	0.49	0.19	0.28
1.60	1.40	0.40	0.60	1.40	0.60	1.60	1.40
100.00	100.04	100.42	100.00	100.04	100.13	100.37	100.49
11	25	7	11	13	8	16	13
187	182	196	238	241	153	227	236
11	31	9	9	17	9	37	39
970	1008	729	967	973	924	339	452
87	76	50	107	67	89	116	137
8	5	3	8	6	1	13	10
13	10	13	14	16	14	22	16
125	95	127	121	94	118	86	88
221	69	90	83	16	125	25	0
100	49	200	85	107	127	29	32
199	64	429	141	164	125	79	86
119	154	134	144	114	132	117	110
925	314	207	350	394	309	589	612
12	5	3	7	6	4	4	5
191	105	111	76	119	90	95	137
89	30	31	58	66	42	42	55

Ballachulish granitoid and inclusions

Sample	B201G	B202G	B203G	B204G	B205G	B207G	B208G
SiO2	59.08	62.27	60.46	61.93	59.74	59.63	60.00
TiO2	0.89	0.75	0.82	0.84	0.86	0.85	0.84
Al2O3	16.34	16.42	17.05	16.25	16.35	16.33	16.56
Fe2O3	5.74	4.91	5.28	4.98	5.38	5.25	5.32
MnO	0.10	0.08	0.08	0.08	0.09	0.10	0.09
MgO	3.37	2.81	2.92	2.72	3.50	3.81	3.26
CaO	4.74	4.09	4.42	3.81	4.70	4.54	4.84
Na2O	4.35	3.90	3.64	4.07	4.43	3.56	4.29
K2O	3.25	3.32	3.61	4.01	3.27	3.56	3.13
P2O5	0.33	0.33	0.29	0.30	0.41	0.37	0.33
loss	1.40	0.80	1.20	0.60	1.00	1.20	1.20
Total	99.94	100.02	100.13	99.97	100.09	99.57	100.23
Nb	14	14	15	12	16	17	12
Zr	242	183	278	234	172	217	231
Y	28	20	20	17	25	22	20
Sr	1288	1259	1252	1159	1308	1243	1397
Rb	77	75	82	83	72	78	73
Th	14	9	15	12	7	9	10
Pb	26	19	26	19	13	17	21
Zn	74	65	69	64	69	70	69
Cu	55	21	29	28	34	12	21
Ni	39	34	32	31	36	44	38
Cr	79	65	61	59	81	97	85
V	107	98	102	97	107	106	107
Ba	1268	1392	1473	1794	1464	1535	1438
Hf	6	7	6	8	6	5	4
Ce	125	105	108	101	128	120	103
La	56	52	55	49	64	62	49

B209G	B210G	B219G	B301G	B306G	B307G	B308G	B309G	B311G
59.67	59.79	62.95	62.77	61.33	62.63	63.09	62.03	60.46
0.81	0.82	0.73	0.70	0.80	0.68	0.69	0.77	0.87
16.85	17.01	16.36	16.61	16.56	16.13	16.04	16.47	17.40
5.39	5.36	4.53	4.54	4.86	4.77	4.37	4.99	5.39
0.10	0.09	0.07	0.07	0.08	0.08	0.07	0.08	0.08
3.31	3.10	2.55	2.80	3.04	2.86	2.56	3.21	3.35
5.04	5.03	4.04	4.07	4.24	4.01	3.82	4.21	4.22
3.97	3.58	3.71	4.03	3.78	3.61	3.85	3.29	3.70
3.25	3.47	3.32	3.15	3.28	3.50	4.37	3.24	3.28
0.33	0.33	0.23	0.26	0.33	0.31	0.22	0.34	0.31
1.00	1.00	1.20	0.60	1.20	1.20	0.60	1.20	0.60
100.09	99.96	100.00	99.93	99.84	100.11	100.02	100.17	100.01
12	13	11	13	15	10	12	12	16
235	245	224	186	191	166	191	185	206
22	20	16	18	24	18	18	19	22
1355	1418	1193	1275	1225	1230	1264	1304	1271
70	72	75	71	72	73	70	72	90
15	5	13	14	19	16	15	12	21
22	24	28	21	20	21	18	17	23
70	68	60	51	60	57	53	75	66
24	15	16	27	26	15	21	26	12
34	40	27	28	32	34	33	35	37
82	79	50	47	35	63	52	66	66
111	110	91	89	105	91	84	108	92
1487	1563	1209	1319	1393	1352	1461	1355	1290
4	6	4	8	5	8	6	6	8
102	104	71	117	104	69	83	94	166
52	48	37	38	67	31	50	34	98

Sample	B312G	B313G	B314G	B315G	B316G	B317G	B318G	B320G	B323G	B201X	B202X	B204X	B205X	B207X	B208X	B210X
SiO2	62.92	62.67	61.35	62.38	61.92	62.36	62.14	60.20	61.85	50.38	52.64	55.34	54.72	55.09	50.78	59.01
TiO2	0.71	0.70	0.77	0.67	0.69	0.76	0.74	0.84	0.80	1.09	1.84	1.28	1.27	1.11	1.37	0.96
Al2O3	16.41	16.32	16.63	16.59	16.15	16.71	16.46	16.71	16.83	21.80	14.67	15.38	22.93	15.25	14.65	16.59
Fe2O3	4.49	4.79	4.95	4.68	4.76	4.96	4.85	5.40	5.07	9.60	9.02	7.22	5.00	7.40	9.30	5.70
MnO	0.07	0.07	0.08	0.07	0.07	0.08	0.08	0.09	0.08	0.09	0.17	0.15	0.05	0.17	0.21	0.10
MgO	2.77	2.61	2.98	3.03	2.91	2.49	2.79	3.18	3.00	2.75	6.85	5.77	1.60	6.10	7.94	3.79
CaO	3.88	3.86	4.39	4.08	3.99	4.16	4.14	4.24	4.36	4.43	6.16	6.97	5.44	6.38	8.05	5.37
Na2O	3.40	3.67	4.22	3.47	4.00	4.10	4.54	4.40	3.44	4.66	2.71	3.56	5.64	3.32	3.23	4.45
K2O	3.62	3.57	3.34	3.63	3.69	3.43	2.55	3.26	3.34	3.68	4.41	2.85	2.44	3.92	2.47	2.35
P2O5	0.30	0.30	0.32	0.26	0.27	0.23	0.43	0.37	0.29	0.19	0.91	0.49	0.26	0.36	0.47	0.34
loss	1.00	1.20	0.80	1.00	1.20	0.40	0.80	0.60	0.60	1.20	0.20	1.00	0.40	0.60	1.40	1.00
Total	99.93	100.13	100.17	100.20	99.97	99.97	99.82	99.64	99.99	100.37	100.00	100.31	100.23	100.08	100.21	99.96
Nb	13	13	12	11	8	10	10	17	10	19	24	15	24	19	15	13
Zr	184	169	184	174	152	144	189	197	180	342	277	170	584	169	146	118
Y	17	15	21	18	18	17	22	21	17	31	41	31	28	29	33	16
Sr	1250	1271	1332	1249	1057	1061	1247	1238	1303	1582	899	959	1933	1055	998	1329
Rb	77	78	68	74	72	66	70	81	71	106	124	72	66	82	65	64
Th	9	14	10	14	13	13	12	17	16	28	8	12	23	5	7	14
Pb	32	27	21	26	20	14	21	21	22	37	13	25	27	19	21	23
Zn	56	57	58	61	60	49	63	66	59	99	113	90	32	95	119	74
Cu	22	35	40	37	13	5	40	27	42	497	2	4	3	12	9	4
Ni	33	30	36	30	33	30	35	39	36	40	145	89	17	117	139	38
Cr	53	52	71	55	58	53	63	78	65	124	292	265	126	320	488	91
V	88	91	92	93	86	78	85	101	88	123	199	159	102	132	177	126
Ba	1599	1713	1350	1448	1421	1244	1014	1362	1248	1584	1751	960	1589	1615	995	965
Hf	8	8	7	7	6	7	6	8	9	9	5	6	15	3	3	3
Ce	88	131	99	95	112	88	77	124	88	248	215	106	139	119	129	98
La	39	35	38	32	70	38	34	60	31	108	95	52	67	54	50	50

Sample	B219X	B301X	B306X	B307X	B308X	B309X	B311X	B312X	B314X	B315X	B316X	B317X	B318X	B319X	B320X
SiO2	54.49	54.52	51.93	55.17	54.14	52.09	48.04	54.95	52.56	51.42	54.31	54.02	54.25	52.78	52.14
TiO2	1.19	1.10	1.18	1.23	1.21	1.51	1.02	1.43	1.53	1.35	1.08	1.32	1.23	1.04	0.85
Al2O3	18.16	18.03	24.36	15.31	17.08	15.02	26.83	17.07	14.61	15.62	17.15	13.98	18.57	21.04	23.02
Fe2O3	7.02	7.37	8.00	7.61	7.78	8.25	8.88	7.59	8.17	8.92	7.44	8.81	7.88	9.03	8.11
MnO	0.12	0.12	0.05	0.16	0.15	0.15	0.10	0.11	0.15	0.17	0.13	0.19	0.12	0.09	0.08
MgO	4.26	4.15	1.54	5.55	5.42	7.56	2.35	4.94	7.50	6.47	4.58	6.72	4.14	2.09	2.37
CaO	5.78	5.42	1.02	7.80	6.61	7.67	1.12	6.15	7.70	7.82	6.16	7.40	5.69	2.56	1.68
Na2O	4.74	5.24	3.13	3.82	3.92	3.19	2.18	4.78	3.69	3.37	5.12	2.53	4.61	3.70	3.94
K2O	2.34	2.19	7.74	2.22	2.14	2.62	7.55	2.36	2.82	2.32	2.41	2.79	2.32	6.30	6.86
P2O5	0.32	0.41	0.08	0.77	0.47	0.58	0.15	0.41	0.55	1.23	0.41	0.73	0.54	0.12	0.15
loss	1.20	1.20	0.60	0.60	0.80	1.00	1.40	0.20	1.00	1.00	1.00	1.20	0.40	0.80	0.40
Total	99.85	99.99	99.92	100.54	100.00	99.93	99.91	100.25	100.58	99.98	100.03	100.00	100.02	100.06	99.98
Nb	23	16	39	19	25	16	24	13	18	28	11	21	23	19	14
Zr	109	154	195	166	85	136	362	191	147	122	155	50	154	299	177
Y	29	18	41	40	33	18	69	26	21	45	21	30	26	33	28
Sr	1155	1210	301	1107	1130	742	268	1074	984	972	897	616	1316	1188	661
Rb	79	69	58	40	62	55	88	54	21	75	56	58	76	122	80
Th	13	2	27	12	14	2	40	8	2	15	6	15	11	20	15
Pb	33	14	31	15	23	7	45	18	21	10	15	11	21	39	30
Zn	90	89	66	83	81	85	123	72	72	107	88	116	93	116	101
Cu	42	55	14	4	23	1	1	114	18	7	1	1	1	12	641
Ni	39	28	43	53	131	132	31	28	134	81	41	99	22	30	37
Cr	38	32	66	155	281	339	62	39	331	187	122	325	24	55	90
V	131	134	104	176	177	202	95	142	173	162	145	150	170	109	81
Ba	363	426	1711	973	571	995	1344	665	954	769	636	1184	579	2836	1595
Hf	4	7	5	4	4	4	7	5	6	4	7	3	5	6	6
Ce	129	124	130	137	121	106	197	101	61	238	120	282	148	176	175
La	55	45	39	61	54	44	95	49	38	102	56	124	63	85	73

Criffell granitoid and inclusions

Sample	C201G	C202G	C204G	C205G	C206G	C211G	C301G	C304G	C312G	C319G	C320G	C322G	C323G	C325G	C326G	C328G
SiO2	63.33	64.66	62.97	62.53	68.79	68.75	66.22	68.77	61.69	71.28	66.04	65.41	65.20	65.79	67.07	68.39
TiO2	0.61	0.63	0.78	0.74	0.44	0.41	0.50	0.34	0.65	0.30	0.46	0.46	0.45	0.40	0.45	0.38
Al2O3	16.05	16.07	16.34	16.35	15.51	15.85	15.95	16.23	17.01	14.96	15.81	16.25	16.07	15.85	15.09	15.60
Fe2O3	3.84	3.96	4.58	4.39	2.59	2.41	2.98	2.70	4.71	2.09	3.43	3.38	3.41	3.12	3.09	2.60
MnO	0.07	0.07	0.08	0.09	0.05	0.05	0.05	0.05	0.06	0.05	0.06	0.06	0.06	0.05	0.05	0.05
MgO	2.46	2.53	2.75	2.45	1.58	1.50	2.01	1.76	3.94	1.29	2.07	2.32	2.30	2.00	1.95	1.58
CaO	4.06	3.47	3.84	4.08	2.80	2.32	2.70	2.39	4.88	1.49	3.67	3.41	3.91	3.63	3.60	2.44
Na2O	3.97	4.16	3.62	4.72	3.66	4.08	4.44	4.18	3.64	2.79	4.20	4.13	4.07	4.17	4.54	4.05
K2O	2.81	3.16	3.50	3.31	3.80	3.83	3.67	3.43	2.41	4.85	3.57	3.61	3.11	3.07	2.47	3.58
P2O5	0.23	0.23	0.33	0.32	0.19	0.18	0.23	0.09	0.26	0.18	0.25	0.30	0.31	0.24	0.28	0.21
loss	2.00	0.80	1.00	0.80	0.20	0.80	0.60	0.20	0.60	0.60	0.20	0.40	1.00	1.00	1.60	1.00
Total	99.65	99.96	100.02	100.02	99.81	100.36	99.56	100.35	100.08	100.04	100.02	99.98	100.12	99.56	100.41	100.07
Nb	7	5	14	12	7	6	10	2	5	11	10	10	10	7	7	7
Zr	168	159	174	192	165	152	156	137	184	112	159	158	149	129	133	134
Y	13	11	15	13	13	8	11	9	14	8	12	9	13	8	8	7
Sr	680	638	722	727	669	479	623	676	757	285	884	889	829	848	829	525
Rb	112	127	115	105	110	168	108	109	57	226	122	137	108	110	94	161
Th	14	10	8	12	14	8	18	7	3	7	12	8	5	1	12	10
Pb	32	26	18	20	32	37	16	9	10	38	40	39	30	30	30	31
Zn	55	61	59	62	49	56	32	36	57	42	46	61	52	47	43	50
Cu	52	99	13	15	12	14	20	21	20	27	15	3	7	20	19	21
Ni	34	33	34	34	19	16	29	18	88	21	35	38	40	36	25	24
Cr	59	56	54	58	24	19	14	123	129	9	49	41	63	13	9	3
V	73	84	94	93	50	48	61	55	100	43	69	73	68	77	66	58
Ba	738	810	844	873	722	701	874	773	748	687	985	984	796	920	656	791
Hf	5	5	6	7	5	5	5	6	6	5	5	6	6	5	6	3
Ce	75	83	110	102	61	71	71	74	46	47	80	60	80	74	130	83
La	42	35	61	61	30	37	39	35	41	22	40	33	37	46	94	35

Sample	C331G	C332G	C333G	C335G	C336G	C339G	C340G	C341G	C342G	C343G	C345G	C346G	C347G	C348G	C349G	C352G	C354G
SiO2	68.28	68.50	68.11	67.50	65.07	62.88	63.96	64.56	64.95	65.44	64.49	65.66	65.04	65.62	64.66	67.21	68.25
TiO2	0.37	0.41	0.41	0.44	0.69	0.72	0.68	0.54	0.49	0.47	0.42	0.50	0.46	0.46	0.44	0.39	0.47
Al2O3	15.78	15.94	15.80	15.90	15.89	16.49	16.26	16.21	16.29	15.88	16.08	15.68	15.93	16.42	16.23	16.14	15.65
Fe2O3	2.63	2.78	2.92	2.93	4.40	4.37	4.38	3.66	3.56	3.41	3.50	3.18	3.57	3.44	3.57	2.85	2.67
MnO	0.05	0.05	0.05	0.05	0.08	0.07	0.07	0.06	0.06	0.06	0.06	0.05	0.06	0.05	0.06	0.05	0.05
MgO	1.59	1.53	1.65	1.90	2.67	2.76	2.59	2.56	2.22	2.34	2.36	2.03	2.43	2.17	2.25	1.56	1.54
CaO	2.52	2.73	2.60	2.74	3.44	3.86	4.05	3.92	3.85	3.76	3.78	3.52	3.77	3.55	3.62	2.94	2.26
Na2O	3.62	3.98	3.92	3.54	3.77	3.95	3.83	4.04	4.23	4.77	4.00	4.60	4.05	4.24	4.22	4.39	4.36
K2O	3.58	3.77	4.15	3.77	3.46	3.65	3.47	3.50	3.36	3.09	3.28	3.39	3.52	3.53	3.57	3.49	3.73
P2O5	0.25	0.23	0.17	0.27	0.33	0.34	0.35	0.36	0.30	0.28	0.30	0.29	0.32	0.30	0.31	0.24	0.21
loss	1.60	0.20	0.20	0.80	0.20	0.60	0.40	0.40	0.60	0.40	1.20	1.00	0.60	0.20	0.80	0.60	0.80
Total	100.46	100.32	100.21	100.03	100.25	99.94	100.27	100.14	100.18	100.15	99.74	100.15	100.01	100.24	99.97	100.05	100.17
Nb	6	10	9	11	8	10	13	9	8	8	10	11	9	10	8	8	7
Zr	138	132	143	146	188	178	164	139	135	128	144	147	142	128	142	130	112
Y	7	10	9	10	14	13	15	10	9	9	11	10	13	11	9	7	9
Sr	550	537	532	525	715	711	721	955	956	942	882	869	904	916	881	589	517
Rb	155	172	169	163	119	121	113	129	136	113	119	110	118	127	128	151	159
Th	9	14	8	19	8	8	10	21	19	23	16	14	14	12	12	22	15
Pb	33	33	32	32	22	13	16	33	34	27	36	25	41	31	29	33	30
Zn	47	51	59	55	67	65	55	53	53	47	51	45	49	51	51	53	48
Cu	30	22	103	25	25	99	44	29	13	24	8	36	9	21	11	13	7
Ni	20	24	40	26	39	52	38	42	37	40	36	31	36	36	36	23	24
Cr	4	28	65	9	33	84	63	55	53	23	36	12	50	15	47	35	19
V	53	52	54	63	100	85	87	151	71	81	74	62	72	77	81	61	52
Ba	773	767	934	732	1030	887	824	1402	940	936	998	944	973	1026	765	691	634
Hf	4	4	5	6	5	6	7	4	5	5	5	3	6	6	5	6	4
Ce	51	71	68	87	102	115	92	164	125	84	143	119	89	81	101	78	80
La	22	37	44	41	49	46	53	71	65	43	83	53	36	35	49	38	34

Sample	C206X	C208X	C210X	C333X	C335X	C209X	C301X	C328X
SiO2	56.65	53.22	56.36	55.11	54.71	63.87	63.79	60.77
TiO2	1.26	1.76	1.59	1.39	1.63	0.82	0.57	1.34
Al2O3	17.40	15.72	17.00	18.09	17.75	16.08	16.56	16.45
Fe2O3	5.94	7.51	6.37	6.75	6.70	3.74	3.98	4.98
MnO	0.13	0.17	0.11	0.13	0.12	0.08	0.05	0.08
MgO	3.74	4.97	3.16	4.06	4.06	2.78	3.81	2.53
CaO	4.72	5.52	4.26	4.75	4.43	2.74	2.90	3.29
Na2O	4.72	3.86	4.26	4.53	4.74	3.37	4.15	3.47
K2O	3.38	4.24	4.65	3.10	3.55	4.75	3.17	5.91
P2O5	0.57	1.04	0.75	0.85	1.19	0.35	0.26	0.74
loss	1.20	1.40	1.20	1.00	1.00	1.20	0.60	0.40
Total	99.91	99.69	99.95	100.02	100.15	100.00	100.08	100.29
Nb	30	45	37	40	54	18	11	27
Zr	254	338	265	281	351	203	216	311
Y	23	36	30	11	37	15	14	21
Sr	450	305	373	489	486	476	723	485
Rb	248	269	279	256	256	213	92	312
Th	22	19	20	10	18	14	4	11
Pb	44	45	44	33	26	37	17	46
Zn	114	139	114	134	122	73	45	76
Cu	27	52	43	246	64	16	23	106
Ni	44	83	26	61	36	35	53	30
Cr	61	145	27	69	4	49	65	6
V	125	155	139	147	172	73	103	128
Ba	373	819	738	473	676	825	891	1513
Hf	5	6	5	7	8	5	8	8
Ce	156	268	200	250	303	96	88	144
La	59	92	73	96	132	47	42	72

C331X	C354X	C201X	C202X	C204X	C205X	C319X	C322X
60.50	61.78	53.70	57.11	59.57	55.70	68.61	52.66
1.40	0.90	1.22	1.25	0.88	1.22	0.43	1.04
16.31	15.62	17.60	19.00	17.22	17.29	16.50	17.03
5.13	4.37	7.00	6.11	5.42	6.70	3.26	7.25
0.08	0.07	0.13	0.09	0.11	0.13	0.08	0.13
2.64	3.55	4.88	3.67	3.70	3.84	1.55	5.77
2.99	2.11	6.99	4.66	5.09	6.18	2.77	6.25
3.13	3.61	4.29	4.48	4.01	4.47	4.74	4.01
6.08	5.14	2.00	2.92	2.69	2.03	1.71	3.59
0.77	0.69	0.37	0.63	0.25	0.36	0.20	0.94
0.80	1.80	1.80	0.20	0.80	1.40	0.20	1.00
100.13	99.92	100.23	100.41	99.96	99.52	100.18	99.97
27	21	10	14	9	16	12	12
298	233	153	227	189	220	153	222
21	17	19	20	19	21	11	18
395	350	725	687	691	734	276	955
317	226	97	153	109	84	154	191
17	20	5	27	7	10	10	24
49	43	35	31	15	28	42	37
81	116	76	87	66	80	71	98
97	44	550	620	70	225	119	34
30	51	40	27	46	22	17	41
7	95	52	6	100	37	20	15
119	92	155	150	117	144	68	151
1331	1229	436	538	600	261	246	773
8	6	7	9	5	5	6	7
161	174	113	208	97	94	79	264
64	72	54	107	49	48	36	156

Sample	C324X	C325X	C326X	C341X	C343X	C345X	C346X	C350X
SiO_2	56.05	52.29	51.10	52.14	54.45	57.47	54.14	53.71
TiO_2	0.93	1.18	1.59	1.24	0.90	0.10	1.06	1.13
Al_2O_3	16.41	17.23	18.29	17.46	17.89	17.49	17.61	17.70
Fe_2O_3	6.42	7.60	8.29	7.36	6.30	6.46	7.96	7.62
MnO	0.12	0.15	0.12	0.12	0.13	0.12	0.13	0.13
MgO	4.63	6.18	5.49	5.50	5.55	4.75	4.72	4.93
CaO	6.15	6.37	5.24	5.42	6.95	5.59	5.50	5.84
Na_2O	4.61	4.46	4.36	4.15	4.60	4.61	5.20	5.11
K_2O	2.00	2.62	3.29	3.29	1.94	2.19	2.36	2.60
P_2O_5	0.81	0.76	0.82	0.68	0.55	0.46	0.47	0.77
loss	1.80	1.00	1.40	2.40	0.60	0.60	0.60	0.40
Total	100.16	100.08	100.22	100.02	100.10	100.09	100.01	100.23
Nb	18	17	17	12	10	13	17	12
Zr	273	221	198	166	205	164	170	179
Y	11	19	10	9	17	13	13	15
Sr	745	736	796	884	1023	853	815	866
Rb	120	175	244	235	114	140	159	181
Th	42	33	13	12	9	19	16	24
Pb	27	30	33	31	26	40	48	38
Zn	91	123	47	104	84	84	116	111
Cu	8	25	27	220	12	304	422	381
Ni	55	85	30	92	73	51	71	60
Cr	76	91	44	113	70	108	76	86
V	126	180	190	181	151	154	167	177
Ba	346	356	438	370	389	365	304	415
Hf	7	6	6	5	6	6	7	5
Ce	213	197	149	129	139	118	142	243
La	108	133	64	65	78	58	77	139

C304X	C318X	C320X	C321X	C323X	C336X	C337X	C342X	C312X	C352X
50.63	52.89	55.17	51.95	52.14	51.36	53.70	53.55	68.56	68.49
1.06	0.94	0.90	1.28	1.11	1.40	1.50	0.87	0.32	0.37
15.80	15.11	17.83	17.90	16.28	16.70	17.89	18.37	16.65	15.61
7.57	5.96	6.62	7.84	7.47	8.02	7.46	7.06	2.80	2.51
0.20	0.11	0.10	0.13	0.17	0.13	0.16	0.10	0.04	0.04
6.43	6.27	3.97	4.94	6.40	6.99	4.23	3.98	1.26	1.45
6.40	6.34	5.08	5.98	7.86	6.76	5.75	5.67	2.65	2.51
3.83	2.62	4.53	5.16	4.21	3.17	3.87	4.64	4.04	3.28
3.55	4.56	3.28	2.28	1.69	3.45	3.79	3.34	2.70	4.01
0.91	0.91	0.74	0.62	1.14	0.73	0.80	1.09	0.13	0.20
3.40	3.80	1.00	1.60	1.60	1.20	0.60	0.80	1.00	1.40
100.13	99.96	99.62	99.97	100.35	100.22	100.13	99.97	100.35	100.08
17	22	19	13	16	19	19	12	3	5
336	382	279	246	261	287	312	374	136	136
18	13	20	15	17	7	10	23	9	6
523	898	941	840	748	627	945	1224	649	535
115	126	168	161	105	168	120	199	88	173
17	17	38	9	26	9	12	37	4	4
4	8	58	30	28	15	22	25	14	30
123	104	86	106	112	108	8	81	67	44
32	28	283	22	15	53	345	611	28	67
169	205	76	54	65	120	17	31	26	20
277	422	99	58	156	154	10	32	37	22
170	150	174	135	180	222	194	160	65	52
1173	1544	1211	757	384	921	1442	1319	842	895
7	7	6	5	6	5	9	9	6	4
317	355	349	277	416	225	200	530	56	58
179	193	210	128	218	114	101	295	9	28

Sr ISOTOPIC DATA

Strontian hosts and inclusions.

Sample	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
S207X	117	662	.5119	.70902(4)	.70585	26.5
S208X	53	1157	.1333	.70618(4)	.70535	19.5
S208G	49	1286	.1095	.70603(4)	.70535	19.5
S209X	64	945	.1974	.70672(4)	.70550	21.6
S209G	52	1185	.1276	.70617(3)	.70538	19.8
S210X	73	682	.3082	.70713(3)	.70523	17.7
S211X	62	1098	.1623	.70633(3)	.70533	19.1
S212X	58	1336	.1260	.70612(3)	.70534	19.3
S213X	68	973	.2033	.70652(3)	.70526	18.2
S338X	85	996	.2480	.70684(4)	.70530	18.8
S338G	61	1016	.1746	.70637(3)	.70529	18.6
S340X	79	1302	.1766	.70646(4)	.70537	20.0
S340G	48	1218	.1136	.70603(4)	.70532	19.1
S341X	100	1397	.2063	.70656(3)	.70529	18.5
S341G	53	1187	.1284	.70613(4)	.70534	19.3
S342X	62	1117	.1598	.70632(3)	.70533	19.2
S343X	79	1394	.1647	.70633(3)	.70531	18.9
S344X	75	1135	.1900	.70642(3)	.70525	18.0
S344G	40	1159	.1007	.70591(4)	.70529	18.6
S345X	94	1271	.2134	.70737(3)	.70605	29.4
S345G	50	1235	.1170	.70607(4)	.70535	19.4
S346X	67	1130	.1768	.70641(3)	.70535	19.4
S349X	83	763	.3122	.70794(3)	.70601	28.8
S349G	59	821	.2063	.70724(3)	.70596	28.2
S353X	114	788	.4189	.70706(4)	.70444	6.5
S353G	53	1202	.1271	.70613(3)	.70534	19.3
S354G	48	1179	.1180	.70612(4)	.70539	20.0
S355X	92	925	.2889	.70724(3)	.70545	20.8

Strontian biotite granite and inclusions.

S401X	35	782	.3282	.70834(4)	.70649	35.0
S401G	92	732	.3617	.70923(3)	.70719	44.9
S402X	95	845	.3245	.70788(4)	.70605	28.8
S402G	98	740	.3849	.70948(4)	.70731	46.7
S438X	317	285	3.222	.72612(4)	.70799	56.3
S438G	116	536	.6310	.71065(4)	.70710	43.6

Ballachulish hosts and inclusions.

Sample	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
B205X	69	1927	.1035	.70535(3)	.70475	10.5
B205G	77	1363	.1634	.70532(4)	.70439	5.3
B207X	89	1013	.2445	.70604(4)	.70461	8.6
B311X	92	261	1.0211	.73295(4)	.72699	326.4
B311G	97	1297	.2170	.70563(5)	.70439	5.3
B312X	72	1277	.1632	.70522(4)	.70424	3.6
B312G	83	1241	.1930	.70537(5)	.70424	3.3
B313X	83	1286	.1864	.70548(4)	.70439	5.4
B314X	83	1051	.2297	.70539(5)	.70405	0.5
B314G	71	1329	.1552	.70520(4)	.70430	4.1
B315X	80	1007	.2305	.70559(4)	.70424	3.3
B317X	79	874	.2621	.70596(4)	.70443	6.0
B318X	78	1286	.1800	.70539(4)	.70434	4.7
B320X	97	721	.3883	.71043(4)	.70816	59.0
B320G	85	1251	.2026	.70565(5)	.70449	6.7

Criffell hosts and inclusions.

Sample	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
C208X	268	302	2.5631	.72087(4)	.70638	33.5
C208G	160	572	.8110	.71059(4)	.70601	28.0
C209X	213	454	1.3607	.71346(3)	.70576	24.7
C209G	150	475	.9116	.71104(4)	.70588	26.4
C210X	227	360	2.2292	.71881(4)	.70620	31.0
C211X	292	422	2.0053	.71755(3)	.70621	31.0
C311G	168	461	1.0577	.71235(4)	.70637	33.2
C312X	87	629	.4001	.70798(4)	.70572	24.0
C312G	61	802	.2207	.70647(4)	.70522	16.9
C321X	167	882	.5494	.70829(4)	.70518	16.4
C321G	128	884	.4189	.70775(4)	.70538	19.3
C323X	108	759	.4108	.70778(4)	.70546	20.4
C323G	105	849	.3601	.70741(3)	.70537	19.0
C324X	124	783	.4584	.70813(4)	.70554	21.5
C324G	128	870	.4255	.70824(4)	.70584	25.7
C330X	368	354	3.0139	.72295(4)	.70591	26.8
C330G	150	539	.8062	.71065(3)	.70601	29.4
C336X	168	627	.7731	.70971(4)	.70534	18.7
C336G	116	729	.4631	.70820(4)	.70558	22.0

Details of intial Sr isotope ratios and ϵ calculations are given in Appendix 4.

Southern Uplands Lower Palaeozoic sediments (Nd isotopic analyses and sample details are given by O'Nions et al. 1983).

Sample	t (Myr)	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
SU19A	(430)	33	42	2.2779	.73361(8)	.71966	222
SU21	(450)	49	370	.3849	.70779(6)	.70532	19
SU22A	(450)	113	95	3.4606	.73168(7)	.70950	79
SU37A	(450)	70	81	2.4987	.72387(5)	.70785	55
SU37B	(450)	39	100	1.1401	.71547(10)	.70816	60
SU38	(450)	49	59	1.9841	.72010(6)	.70738	49
SU42	(470)	57	40	.5481	.71163(7)	.70796	57
SU50B	(430)	91	118	2.2266	.72420(7)	.71056	93
SU50C	(430)	57	228	.7223	.71390(9)	.70948	78

Nd ISOTOPIC DATA

Strontian hosts and inclusions (t=427Myr).

Sample	Sm (ppm)	Nd (ppm)	$(^{147}Sm/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_t$	ϵNd_t	T_{dm} (Gyr)
S207X	7.526	45.140	.10003	.512319(20)	.51204	-0.8	0.97
S210X	8.003	63.414	.07553	.512446(20)	.51223	3.0	0.67
S210G	7.226	46.524	.09314	.512372(20)	.51211	0.6	0.99
S338X	5.963	34.667	.10218	.512426(21)	.51216	1.1	0.99
S341X	6.921	44.103	.09503	.512417(20)	.51215	1.3	0.81
S341G	7.084	44.262	.09502	.512309(20)	.51204	-0.8	0.9
S345X	1.711	37.837	.09224	.512364(19)	.51210	0.5	0.86
S345G	8.829	43.910	.12130	.512381(19)	.51204	-0.8	1.09
S353X	2.104	32.866	.06561	.512353(21)	.51217	1.7	0.72
S353G	5.421	32.832	.09926	.512382(30)	.51210	0.4	0.89

Ballachulish hosts and inclusions (t=400Myr).

Sample	Sm (ppm)	Nd (ppm)	$(^{147}Sm/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_t$	ϵNd_t	T_{dm} (Gyr)
B205X	7.912	47.305	.10096	.511999(20)	.51173	-7.6	1.41
B205G	9.283	56.514	.09932	.512244(19)	.51198	-2.7	1.07
B311X	14.611	77.241	.11406	.511999(21)	.51170	-8.3	1.59
B311G	8.205	51.922	.09550	.512234(20)	.51198	-2.7	1.04
B320X	11.518	68.031	.10189	.511719(20)	.51145	-13.1	1.80

Criffell host and inclusions (t=397Myr).

Sample	Sm (ppm)	Nd (ppm)	$(^{147}Sm/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_t$	ϵNd_t	T_{dm} (Gyr)
C208X	17.970	166.410	.06497	.512330(20)	.51216	0.7	0.74
C208G	4.461	26.260	.10150	.512288(20)	.51202	-2.0	1.03
C209X	7.905	64.931	.07387	.512349(20)	.51216	0.6	0.76
C209G	3.451	25.207	.08051	.512311(21)	.51210	-0.5	0.84
C321X	7.517	50.310	.09030	.512340(20)	.51211	-0.4	0.87
C321G	6.020	25.980	.14090	.512312(20)	.51195	-3.5	1.52
C323X	18.699	145.918	.07693	.512314(21)	.51212	-0.2	0.82
C323G	5.361	36.009	.09096	.512389(19)	.51215	0.4	0.82
C330X	4.370	129.431	.02008	.512274(20)	.51222	1.9	0.61
C330G	4.301	28.760	.09619	.512328(21)	.51208	-0.9	0.93

Bulk Earth parameters for ϵ calaculations are given in Appendix 4.

APPENDIX 3

ANALYTICAL DATA FOR STRONTIAN APPINITES.

Sample	S324A	S327A	S443A	S444A	S447A	S442A	S439A	S440A
SiO2	46.97	46.66	54.29	53.90	49.91	63.74	68.02	67.08
TiO2	1.75	1.74	1.63	1.61	2.25	0.86	0.54	0.59
Al2O3	12.36	11.78	14.15	15.00	14.04	15.29	15.50	15.71
Fe2O3	10.03	10.15	7.45	7.16	9.17	4.36	3.08	3.08
MnO	0.15	0.16	0.09	0.09	0.10	0.05	0.05	0.05
MgO	10.47	12.90	6.15	4.64	7.47	2.85	1.31	2.03
CaO	7.79	8.12	6.47	5.91	7.49	3.06	2.36	2.41
Na2O	1.82	2.16	3.73	3.39	3.60	4.38	4.41	4.65
K2O	4.23	2.81	3.71	4.22	3.05	3.82	3.98	3.75
P2O5	1.16	0.88	0.81	0.76	0.97	0.42	0.22	0.24
loss	2.40	1.80	1.00	2.60	1.60	1.20	0.60	0.40
Total	99.70	99.61	99.87	99.68	100.20	100.30	100.32	100.25
Nb	23	27	21	22	28	12	8	10
Zr	291	290	388	428	503	256	254	271
Y	25	25	35	39	40	29	29	29
Sr	1326	1015	1075	1204	1406	856	705	771
Rb	153	60	84	106	80	95	114	106
Th	6	2	9	8	8	6	7	10
Pb	0	5	11	14	10	24	22	28
Zn	127	105	84	88	90	66	48	49
Cu	0	29	18	16	65	20	21	21
Ni	309	355	97	66	103	47	26	26
Cr	568	871	227	140	372	85	15	23
V	200	188	171	160	220	79	54	58
Ba	2292	1281	1490	1495	2254	1048	1079	1047
Hf	5	4	6	6	8	5	5	6
Ce	239	147	132	146	184	75	51	72
La	107	82	62	77	87	47	36	50

Sample	S441A
SiO2	64.66
TiO2	0.08
Al2O3	15.27
Fe2O3	4.08
MnO	0.06
MgO	2.63
CaO	2.90
Na2O	4.24
K2O	3.77
P2O5	0.39
loss	1.20
Total	99.56
Nb	11
Zr	205
Y	33
Sr	995
Rb	113
Th	8
Pb	23
Zn	60
Cu	22
Ni	42
Cr	62
V	85
Ba	1033
Hf	3
Ce	86
La	61

Appinite-PG

Sample	S335A	S336A	S431A	S423A	S312A
SiO2	50.98	50.90	51.74	56.00	50.30
TiO2	1.19	1.11	1.20	0.96	1.21
Al2O3	12.47	11.99	12.48	9.97	12.49
Fe2O3	9.49	9.04	8.36	7.60	9.06
MnO	0.15	0.14	0.10	0.08	0.14
MgO	11.57	12.83	9.86	10.98	10.91
CaO	7.36	7.21	7.69	6.91	7.30
Na2O	2.49	2.44	3.13	2.64	2.55
K2O	3.31	3.16	3.08	2.29	3.16
P2O5	0.76	0.70	0.73	0.30	0.70
loss	0.00	0.00	0.80	2.40	0.00
Total	100.18	99.92	99.57	100.44	98.22
Nb	17	17	20	7	17
Zr	231	227	384	175	258
Y	25	25	37	25	27
Sr	991	825	952	623	941
Rb	87	81	74	54	82
Th	16	11	22	12	15
Pb	13	10	12	4	1
Zn	101	97	90	69	95
Cu	21	18	66	28	22
Ni	311	352	235	284	259
Cr	772	937	597	809	728
V	171	160	157	110	169
Ba	1018	945	1001	780	989
Hf	4	2	6	2	5
Ce	239	233	211	77	248
La	106	108	96	40	112

S333A	S427A	S310A	S311A	S428A	S429A
53.11	52.75	49.97	59.79	50.47	60.39
1.10	1.65	1.15	0.56	1.58	0.62
12.61	16.30	12.55	10.22	12.92	9.97
8.52	8.87	8.88	6.77	8.94	6.19
0.16	0.10	0.15	0.13	0.12	0.07
9.66	4.93	11.11	9.72	8.79	9.82
6.79	6.30	7.83	6.87	8.09	6.02
2.57	4.80	2.40	1.82	3.06	2.61
3.33	2.35	3.05	3.47	2.96	3.43
0.67	1.03	0.69	0.37	0.94	0.24
0.00	0.20	1.60	0.20	2.00	0.60
98.90	99.59	99.76	100.28	100.26	100.26
15	26	19	9	22	9
194	431	234	125	230	222
21	41	30	19	46	22
674	991	809	612	1184	675
70	67	80	41	92	55
10	25	16	8	11	13
5	15	11	4	10	14
113	107	91	71	99	64
296	98	28	25	45	72
261	91	273	259	206	261
653	217	827	684	538	688
149	212	160	94	240	81
992	392	882	1495	839	689
6	7	3	2	3	4
202	236	225	131	251	127
94	120	107	60	109	48

Appinite-Tl

Sample	S304A	S409A	S412A	S305A	S307A	S308A	S405A	S406A	S407A	S408A	S410A	S411A	S413A
SiO2	52.96	51.65	52.33	47.62	50.40	49.06	50.66	50.23	53.19	52.96	51.76	51.25	46.35
TiO2	1.33	1.49	1.44	1.92	1.78	1.64	1.48	1.71	1.33	1.33	1.43	1.47	2.17
Al2O3	12.53	14.41	14.02	12.60	12.83	12.67	13.21	13.08	15.63	14.07	14.83	14.11	13.03
Fe2O3	8.70	8.53	8.11	9.75	9.12	10.26	9.10	9.48	8.06	8.19	8.06	8.40	10.26
MnO	0.15	0.10	0.09	0.12	0.12	0.15	0.11	0.11	0.09	0.10	0.09	0.10	0.11
MgO	10.46	7.97	8.61	10.49	9.53	11.26	10.36	9.40	7.27	8.15	8.28	9.17	12.38
CaO	7.73	7.33	7.46	8.95	7.55	8.67	8.12	8.36	7.31	7.00	7.14	7.65	9.50
Na2O	2.69	2.69	2.73	2.95	2.56	1.62	2.67	2.60	2.79	3.25	3.00	2.77	1.91
K2O	2.16	2.77	2.46	2.07	2.89	2.53	2.45	2.15	2.16	2.40	2.43	2.22	2.14
P2O5	0.35	0.78	0.67	1.03	0.91	0.56	0.65	0.67	0.38	0.60	0.68	0.57	0.62
loss	1.00	1.80	1.40	2.00	2.00	1.20	1.20	1.80	1.60	1.80	1.60	1.80	1.40
Total	100.36	99.91	99.68	99.86	100.08	99.97	100.38	99.93	100.14	100.18	99.67	99.86	100.21
Nb	11	14	9	15	19	10	9	12	12	11	13	10	12
Zr	173	388	320	104	186	146	257	239	320	343	336	326	225
Y	28	38	33	37	42	32	32	35	29	33	34	33	38
Sr	839	1330	1211	972	917	944	1119	1010	1308	1139	1292	1239	976
Rb	46	59	54	33	78	47	53	42	47	51	51	44	40
Th	3	12	6	7	0	0	1	5	10	14	10	7	4
Pb	8	8	11	6	0	3	2	7	14	11	15	6	6
Zn	83	89	79	98	96	103	89	81	71	82	83	82	84
Cu	25	35	37	26	28	27	35	41	37	33	26	36	52
Ni	205	142	173	187	165	222	231	203	96	138	136	148	243
Cr	649	360	453	502	551	710	626	569	293	373	370	407	535
V	197	206	207	244	227	227	189	254	183	187	205	215	302
Ba	603	1004	833	1039	1329	827	896	732	724	718	867	796	661
Hf	3	5	6	3	4	2	4	5	7	4	5	5	3
Ce	122	160	115	210	207	143	157	148	108	132	159	131	189
La	53	75	58	90	84	49	45	59	51	73	74	62	71

Appinite-T2

Sample	S421A	S316A	S422A	S418A	S425A	S426A	S320A
SiO2	49.19	48.46	50.49	49.31	58.78	50.83	57.61
TiO2	0.84	1.01	1.20	1.10	0.99	1.58	1.11
Al2O3	8.36	10.71	10.14	10.21	17.90	16.97	14.54
Fe2O3	9.82	9.36	9.56	9.46	4.97	8.90	6.61
MnO	0.12	0.12	0.10	0.12	0.04	0.10	0.10
MgO	15.35	14.70	13.69	12.60	3.46	5.65	6.51
CaO	11.83	7.90	8.63	10.35	5.00	7.25	5.38
Na2O	2.07	2.49	2.46	2.52	5.31	4.51	4.02
K2O	1.21	2.62	1.64	1.78	1.82	2.31	2.56
P2O5	0.42	0.58	1.05	0.67	0.38	0.92	0.42
loss	0.40	1.60	1.00	1.60	0.80	0.40	0.80
Total	99.96	99.95	100.29	100.11	99.82	99.88	99.97
Nb	6	14	14	10	14	16	17
Zr	239	205	225	302	395	393	185
Y	24	24	26	31	21	32	23
Sr	933	778	788	1226	1812	1863	1251
Rb	24	66	30	43	35	44	70
Th	7	13	10	2	7	6	10
Pb	6	11	1	8	7	7	12
Zn	84	94	84	84	58	92	78
Cu	36	21	32	31	20	27	19
Ni	286	363	364	217	66	75	131
Cr	885	1142	789	695	69	166	271
V	147	153	171	179	120	207	145
Ba	523	798	566	765	894	1328	670
Hf	3	4	4	4	8	6	5
Ce	185	192	115	221	117	214	127
La	78	93	59	92	54	102	60

S321A	S417A	S317A	S414A	S415A	S423A	S424A
54.06	50.13	52.40	52.35	52.28	49.44	50.94
1.11	1.22	1.12	1.10	1.67	1.64	1.37
14.25	11.94	13.02	13.48	12.37	13.27	13.84
8.08	9.63	9.02	8.32	9.27	9.87	9.09
0.13	0.12	0.15	0.10	0.13	0.12	0.10
8.55	11.34	9.50	8.91	7.97	10.44	9.55
6.22	8.82	7.29	7.36	7.18	7.77	7.32
3.44	2.58	3.00	2.87	2.72	3.14	3.93
3.12	1.99	2.59	2.48	3.06	2.42	2.37
0.65	0.89	0.59	0.56	1.00	0.77	0.59
0.20	1.20	0.80	2.00	2.00	0.60	0.20
100.18	100.25	99.84	99.91	100.00	99.86	99.69
20	14	14	19	21	19	18
190	320	191	344	351	328	332
25	33	27	31	36	35	31
1225	1322	1035	1220	721	1239	1335
84	49	51	56	67	51	52
6	6	11	8	2	10	8
4	5	17	8	3	10	8
88	94	105	80	112	90	83
32	36	11	24	25	30	25
176	185	213	184	177	210	185
440	535	540	438	403	448	480
161	184	174	167	164	210	200
920	713	942	950	1108	807	908
5	4	3	5	5	4	5
195	239	182	179	237	213	188
94	118	82	87	96	89	76

Nd and Sr isotopic analyses on appinite core region samples.

Sample	Sm (ppm)	Nd (ppm)	$(^{147}Sm/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_0$	$(^{143}Nd/^{144}Nd)_t$	ϵNd_t	T_{dm} (Gyr)
S316A	24.33	75.01	.19659	.512345(21)	.51179	-5.7	-
S327A	9.58	59.41	.09746	.512246(21)	.51197	-2.1	1.05
S336A	13.22	80.91	.09886	.512290(21)	.51201	-1.4	1.00

Sample	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
S316A	68	791	.2486	.70750(3)	.70595	28.0
S327A	67	1022	.1891	.70684(3)	.70567	24.0
S336A	91	817	.3206	.70790(3)	.70592	27.5

Sr isotopic traverse through appinite-BG

Sample	Rb ppm	Sr ppm	$(^{87}Rb/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_0$	$(^{87}Sr/^{86}Sr)_t$	ϵSr_t
S439A	113	515	.6374	.71061(4)	.70707	43.1
S440A	108	795	.3965	.70901(4)	.70678	39.0
S441A	115	941	.3545	.70842(4)	.70643	34.1
S442A	111	967	.3345	.70836(4)	.70648	34.8
S443A	86	1065	.2336	.70711(4)	.70579	25.1
S444A	107	1212	.2555	.70729(4)	.70586	26.0

Details of Bulk Earth parameters used in T_{dm} and ϵ calculations are given in Appendix 4.

APPENDIX 4

ANALYTICAL TECHNIQUES AND PRECISION.

ANALYSIS BY X.R.F.

Method of analysis.

1) The sample is prepared from the solid rock specimen by first removing all obvious weathered portions using a grinding wheel. It is then split using a Lake and Elliot hydraulic splitter with stainless steel jaws to reduce the rock to small chips. The chips are then ground in a Tema swing mill with tungston carbide discs until the powder is much less than 200 mesh.

2) Major elements are determined using a fused glass bead prepared from 0.5g of rock powder and 2.5g Spectroflux 105 and ammonium nitrate as oxidant. X-ray analysis is performed on a Phillips PW1212 using a Rh tube for primary excitation. Calibration is done by reference to a monitor (H12) supplied by K. Norrish, and using 25 international rock standards. Matrix corrections are applied on an iterative basis using an on-line microcomputer and using coefficients for the Rh tube also supplied by K. Norrish.

3) Trace elements are analysed on pressed powder discs, prepared for the better international standards by pressing to 5 tons a powder mixed with Moviol binder. The elements are excited with a Rh tube and analysed with either LiF200 or LiF220 crystals. The apparent fluorescence values (AFV) are calculated for the better international standards by inverting the normal matrix correction procedure. These AFV values are then regressed against their intensities (already corrected for line overlap or

blank values) and the coefficients of the line used in the analysis. All intensities are calculated in terms of ratios to a monitor in the first sample position to eliminate the effects of machine drift. For consistency all calculations are performed by an on-line microcomputer.

ISOTOPIC ANALYSIS.

Powder samples of rock is dissolved in concentrated HF in clean Teflon beakers. Chromatographic separation techniques for Rb from Sr and for Sm from Nd are similar to those described by Richard et al. (Earth Planetary Science Letters 1977). Rb & Sr isoptopic analysis were carried out on a V.G. Micromass 30B mass spectrometer whilst Sm & Nd isoptopic analysis were carried out on a V.G. Micromass 54E fully automatic machine. 143Nd/144Nd and 87Sr/86Sr are normalised to 146Nd/144Nd=0.7219 and 88Sr/86Sr=8.73521 respectively. Determination of standard isotopic ratios over the duration of the analysis gave 143Nd/144Nd=0.512633+/-12 for BCR-1 and 87Sr/86Sr=0.710275+/-7 for NBS-987 on the two machines respectively. Calculation of ϵNd, ϵSr and Tdm values used the following Bulk Earth parameters 143Nd/144Nd=0.512638, 147Sm/144Nd=0.1966, 87Sr/86Sr=0.7045, 87Rb/86Sr=0.0839.

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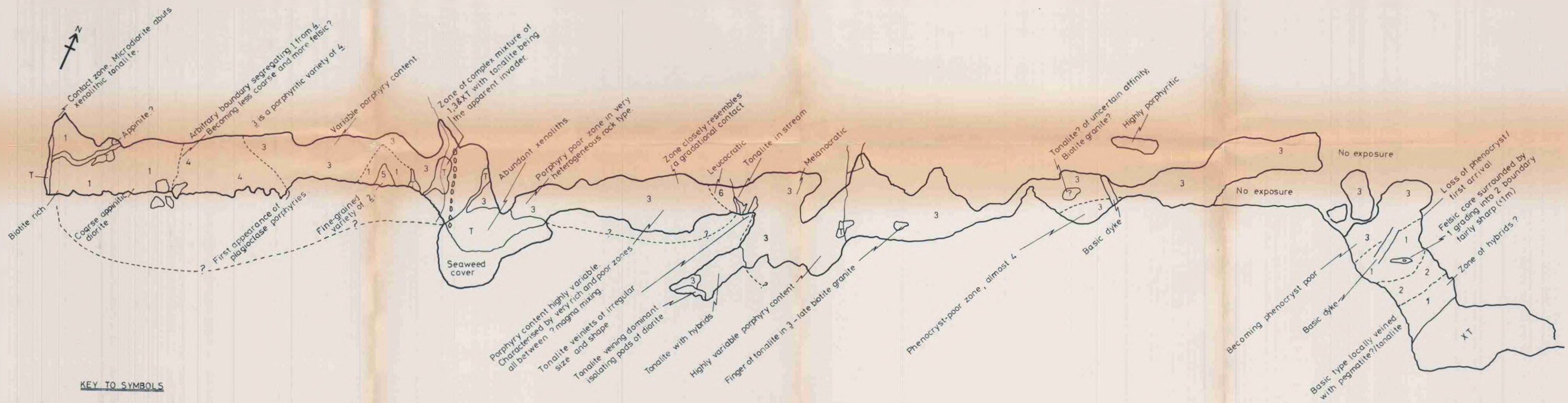
APPINITE - T1



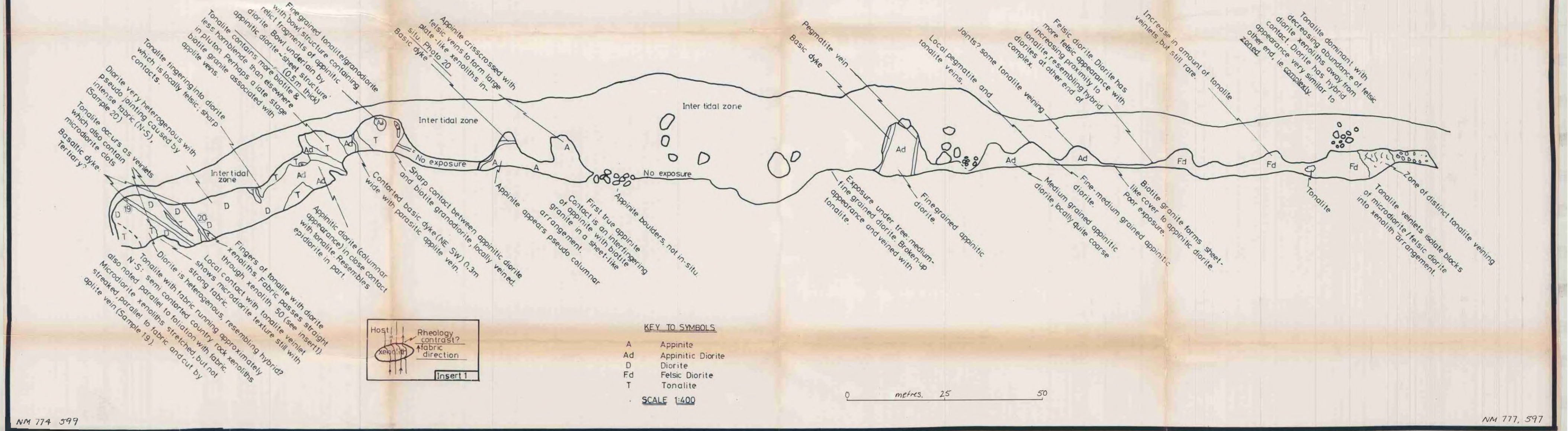
KEY TO SYMBOLS

T	Tonalite
XT	Xenolithic Tonalite
1	Appinitic Diorite
2	Dioritic Appinite
3	Diorite (porphyritic?)
4	Diorite
5	Dark, fine grained Diorite
6	Felsic Diorite

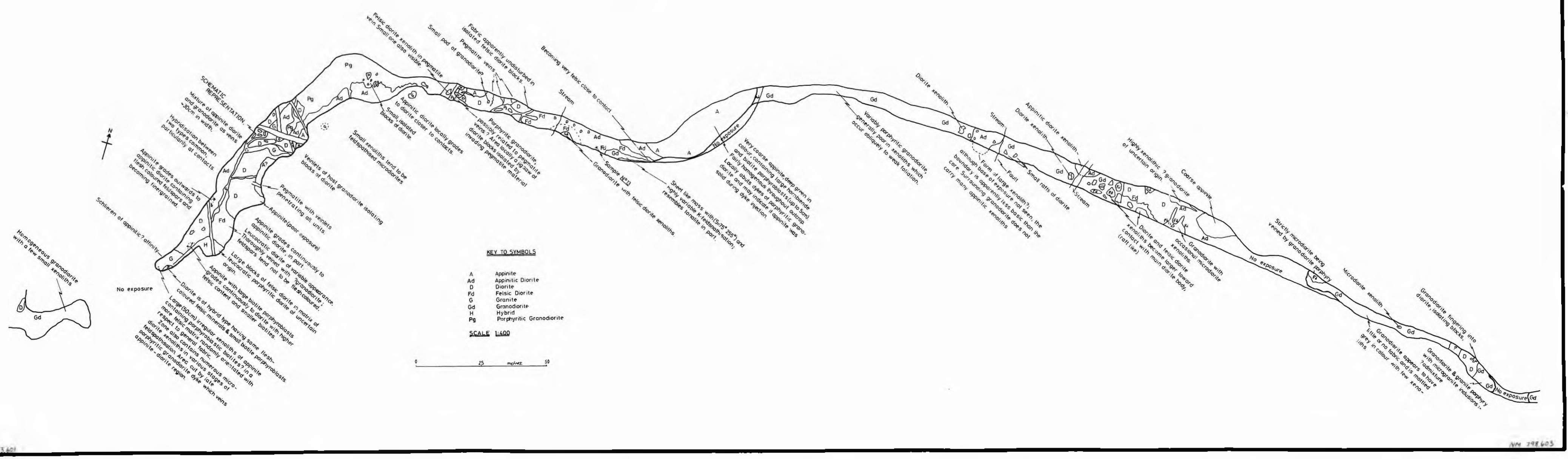
SCALE 1:400

0 metres 25 50

APPINITE - T2


Finegrained tonalite/granodiorite with bowl structure containing relict fragments of appinitic diorite. Bowl underlain by appinitic diorite-'sheet structure'. (0.5m thick)
Tonalite contains more biotite & less hornblende than elsewhere in pluton. Perhaps a late stage biotite granite associated with aplite veins.
Tonalite tingering into diorite which is locally felsic; sharp contacts.
Diorite very heterogenous with pseudo jointing caused by intense fabric (N-S). (Sample 20)
Tonalite occurs as veinlets which also contain microdiorite clots
Basaltic dyke. Tertiary?
Appinite crisscrossed with felsic veins to form large plate-like xenoliths in-situ. Photo 20
Basic dyke
Inter tidal zone
Intertidal zone
No exposure
Sharp contact between appinitic diorite and biotite granodiorite, locally veined.
Contorted basic dyke (NE-SW) 0.3m wide with parasitic aplite vein.
Appinitic diorite (columnar appearance) in close contact with tonalite. Resembles epidiorite in part.
Fingers of tonalite with diorite xenoliths. Fabric passes straight through xenolith 50 (see insert1)
Local contact with tonalite veinlet shows microdiorite texture still with strong fabric.
Diorite is heterogenous, resembling hybrid?
Tonalite with fabric running approximately N-S: semi contorted country rock xenoliths also noted parallel to foliation with fabric.
Microdiorite xenoliths stretched, but not streaked, parallel to fabric and cut by aplite vein (Sample 19).
Appinite appears pseudo columnar
First true appinite Contact is an interfingering of appinite with biotite granite in a sheet-like arrangement.
Appinite boulders, not in-situ.
Pegmatite vein
Basic dyke
Exposure under tree: medium-fine grained diorite. Broken-up appearance and veined with tonalite.
Finegrained appinitic diorite.
Local pegmatite and tonalite veins.
Joints? some tonalite veining
Medium grained appinitic diorite, locally quite coarse
Fine-medium grained appinitic diorite.
Felsic diorite. Diorite has more felsic appearance with increasing proximity to tonalite, resembling hybrid diorites at other end of complex.
Biotite granite forms sheet-like cover to appinitic diorite. Poor exposure.
Increase in amount of tonalite veinlets, but still rare.
Tonalite dominant with decreasing abundance of felsic diorite xenoliths away from contact. Diorite has hybrid appearance very similar to other end, ie complexly zoned
Tonalite
Zone of distinct tonalite veining
Tonalite veinlets isolate blocks of microdiorite/felsic diorite into xenolith arrangement.
A
Ad
D
Fd
T
19
20
Host
xenolith
Rheology contrast?
fabric direction
Insert 1
KEY TO SYMBOLS
A Appinite
Ad Appinitic Diorite
D Diorite
Fd Felsic Diorite
T Tonalite
SCALE 1:400
0 metres 25 50
NM 774 599
NM 777, 597

APPINITE - P.G.


N
Homogeneous granodiorite with a few small xenoliths
D
Gd
No exposure
Schlieren of appinitic? affinity
Appinite grades outwards to appinitic diorite containing flesh coloured feldspars and becoming finergrained.
Hybridisation between two types common, particularly at contacts.
Mixture of appinite diorite and granodiorite as veins ~30cm in width.
SCHEMATIC REPRESENTATION.
Pg
Ad
D
G
A
A?
Fd
H
Diorite is of hybrid type having some flesh-coloured felsic minerals & small biotite porphyroblasts
Large (50cm) irregular xenoliths of appinite containing porphyroblastic biotites? in a more felsic matrix randomly orientated with respect to general fabric.
Zone also contains numerous micro-diorite xenoliths in various stages of feldspathisation. Area cut by late porphyritic granodiorite dyke which veins appinite-diorite region.
Appinite with large biotite porphyroblasts grades continuously to diorite with higher felsic content and smaller biotites.
Large blocks of felsic diorite in matrix of leucocratic porphyritic diorite of uncertain origin.
Appinite grades continuously to appinitic diorite, in part.
Leucocratic diorite of variable appearance. Thoroughly veined with ?granodiorite; feldspars tend not to be flesh coloured.
Appinite (poor exposure)
Pegmatite with veinlets penetrating all units.
Veinlets of host granodiorite isolating blocks of diorite
Small xenoliths tend to be feldspathised microdiorites
Small, isolated blocks of diorite.
Appinitic diorite locally grades to diorite closer to contacts.
Felsic diorite xenolith in pegmatite vein. Small one also visible.
Small pod of granodiorite?
Fabric apparently undisturbed in isolated felsic diorite blocks.
Pegmatite veins
Porphyritic granodiorite, possibly related to pegmatite veins ? Area locally a jigsaw of diorite blocks isolated by invading pegmatite material
Becoming very felsic close to contact
Stream
Sample No 11
Granodiorite with felsic diorite xenoliths.
Sheet like mass with (S=75° 255°) and highly variable K-feldspathisation; resembles tonalite in part.
No exposure
Very coarse appinite, deep green in colour, containing large hornblende and biotite porphyroblasts (up to 5cm). Fairly homogenous throughout outcrop. Locally abuts dykes of porphyritic grano-diorite and may indicate if appinite was solid during dyke injection.
Gd
Variably porphyritic granodiorite, generally poor in xenoliths which occur obliquely to weak foliation.
Diorite xenolith
Stream
Form of large xenolith?; although base of exposure not seen, the boundary is apparently less basic than the core. Surrounding granodiorite does not carry many appinitic xenoliths
Fault
Small rafts of diorite
Appinitic diorite xenolith
Diorite xenolith
Stream
Highly xenolithic ?granodiorite of uncertain origin.
Coarse appinite
Diorite and felsic diorite xenoliths become larger toward contact with main diorite body, (raft like)
Granodiorite with occasional microdiorite xenoliths.
No exposure
Strictly microdiorite being veined by granodiorite porphyry
Microdiorite xenolith
Granodiorite appears to have little or no fabric and is mottled grey in colour with few xeno-liths
Granodiorite fingering into diorite, isolating blocks.
Granodiorite & granite porphyry with microgranite inclusions: ?admixture
?
No exposure
KEY TO SYMBOLS
A Appinite
Ad Appinitic Diorite
D Diorite
Fd Felsic Diorite
G Granite
Gd Granodiorite
H Hybrid
Pg Porphyritic Granodiorite
SCALE 1:400
0
25
metres
50
NM 793 601
NM 798 603