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Geochronology and geochemistry of a Mesozoic magmatic arc system, Fiordland, New Zealand

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Abstract: The Median Tectonic Zone in Eastern Fiordland, SW New Zealand, comprises a tectonically disrupted belt of Mesozoic magmatic arc rocks related to subduction along the palaeo-Pacific margin of Gondwana. New ion microprobe (SHRIMP) U–Pb zircon ages confirm that the bulk of the plutonic rocks in eastern Fiordland range from Mid- Jurassic to Early Cretaceous (168–137 Ma) in age. Carboniferous age granitoids occur in SW Fiordland, along the western side of, and within the zone. Triassic plutonic rocks appear to be restricted to the eastern side of the zone. The Mid-Jurassic–Early Cretaceous igneous rocks (collectively referred to as the Darran Suite) are cut by several plutons of Na-rich granitoid (Separation Point Suite) that give ages of *c.* 124 Ma, slightly older than equivalent rocks in the NW part of the South Island. Early Cretaceous granulite facies orthogneisses (126–119 Ma) in western Fiordland (Western Fiordland Orthogneiss) are considered to be the lower crustal equivalent of the Separation Point plutons.

The majority of the Darran Suite rocks are I-type, hornblende-bearing calc-alkaline igneous rocks, most likely derived from melting in the mantle wedge above a subducting slab of oceanic lithosphere. In contrast, the Separation Point-type plutons are Na-rich, alkali-calcic granitoids with high concentrations of Sr (typically >500 ppm and up to 1000 ppm) and low concentrations of Y (≤ 5 ppm) and heavy REE (<10 times chondritic). Isotopic compositions are primitive, with $^{87}Sr/^{86}Sr$ initial ratios of *c.* 0.7038, and ε_{Nd} values of *c.* $+3$ at 120 Ma. Their geochemistry is consistent with melting of a mafic protolith of garnet amphibolite mineralogy. Mafic Darran Suite rocks have the appropriate chemical and isotopic compositions to generate the Western Fiordland Orthogneiss and the higher level Separation Point type plutons. We suggest that the sudden appearance of large volumes of Na-rich magma during the Early Cretaceous was triggered tectonically, perhaps by thrusting of the Median Tectonic Zone arc beneath western New Zealand. Melting of basal arc underplate at depths of >40 km would then have generated Na-rich granitoids, leaving residues of garnet + clinopyroxene + amphibole.

Keywords: New Zealand, Fiordland, absolute age, geochemistry.

Introduction

Fiordland, in the remote and rugged southwest corner of the South Island, New Zealand (Fig. 1), is well known for the occurrence of high-grade metamorphic rocks (e.g. Oliver & Coggan 1979; Gibson & Ireland 1995 and references therein). Cretaceous granulites are surrounded by a high-grade Palaeozoic cover sequence including orthogneisses and paragneisses which were metamorphosed predominantly during the Devonian–Carboniferous period (Ireland & Gibson in press). On the eastern margin of Fiordland is a belt of plutonic rocks, referred to here as the Eastern Fiordland Igneous Belt, whose affinities and relationships to the high-grade metamorphic rocks have been little examined. The Eastern Fiordland Igneous Belt forms part of the Median Tectonic Zone (e.g. Bradshaw 1993; Kimbrough *et al.* 1993, 1994) which separates the Western and Eastern Provinces of New Zealand (Landis & Coombs 1967). The Western Province is a fragment of Gondwana and comprises mainly Lower Palaeozoic metasedimentary rocks cut by Palaeozoic and Mesozoic granitoids (e.g. Oliver & Coggan 1979; Cooper 1989; Muir *et al.* 1994, 1995, 1996*a*, *b*, 1997). The Eastern Province consists of arc volcanic rocks, arc-derived sedimentary sequences and accretionary complexes of Permian and Mesozoic age that represent the products of convergent margin tectonics (e.g. Bradshaw 1989).

The Median Tectonic Zone is a narrow (10–35 km), generally north-trending belt of Mesozoic plutonic, volcanic and metasedimentary rocks. The zone has been interpreted as the disrupted remnants of a long-lived magmatic arc system that developed along the palaeo-Pacific margin of Gondwana (Fig. 2). According to Kimbrough *et al.* (1994), the Zone consists of a Carboniferous basement with magmatism ranging in age from Early Triassic to Early Cretaceous (247–131 Ma), with a pronounced gap in the Mid-Jurassic (190–160 Ma). The magmatic history of the Median Tectonic Zone in New Zealand closely matches events recorded in the Thurston Island area of

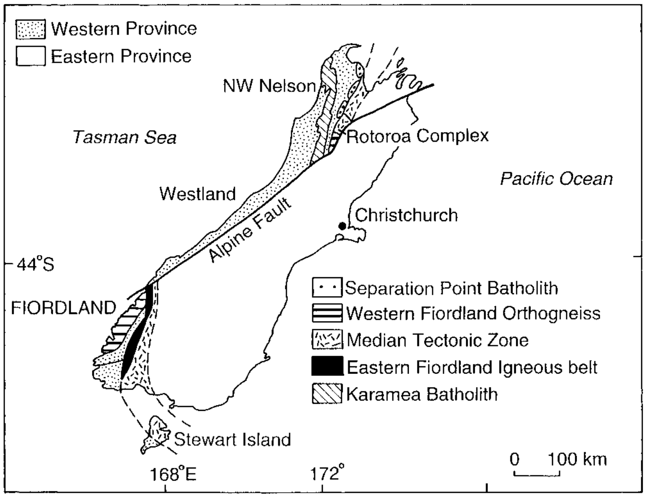

Fig. 1. Geological map of the South Island, New Zealand showing the location of the Median Tectonic Zone and Western and Eastern Provinces. The Eastern Fiordland Igenous Belt (term modified from Oliver & Coggan 1979) forms part of the Median Tectonic Zone.

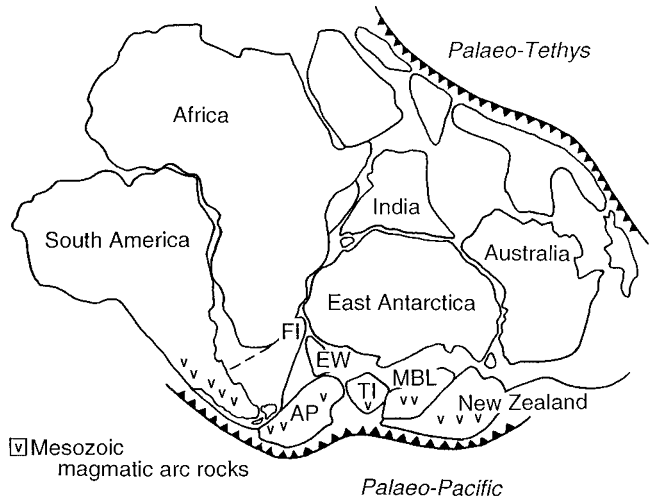

Fig. 2. Late Triassic–Early Jurassic Gondwana reconstruction showing the location of Mesozoic magmatic arc systems along the palaeo-Pacific margin (after Storey *et al.* 1992). AP, Antarctic Peninsula; FI, Falkland Islands; EW, Ellsworth–Whitmore mountains; TI, Thurston Island; MBL, Marie Byrd Land.

West Antarctica (Pankhurst *et al.* 1993) and the Amundsen Province of eastern Marie Byrd Land (Bradshaw *et al.* 1997; Pankhurst *et al.* in press).

This paper presents new U–Pb zircon SHRIMP (sensitive high-resolution ion microprobe) ages and geochemical data for plutonic rocks from the Median Tectonic Zone in eastern Fiordland. The data provide important constraints on the timing of magmatism and on the nature of the petrogenetic processes operating in the deeper levels of a continental arc system over a period of several tens of millions of years. The timescale used throughout this work is that of Harland *et al.* (1990).

Geological setting

Following Oliver & Coggan (1979) and Bradshaw (1985), three distinct geological regions are recognized in Fiordland (Fig. 3); the western and eastern belts and the SW Fiordland block. The latter comprises mainly low-grade metasedimentary rocks and granitoids that are generally regarded as an extension of the Lower Palaeozoic Western Province terranes of NW Nelson-Westland (e.g. Cooper 1989). The western belt has been interpreted as the deep crustal levels of a metamorphic core complex (e.g. Gibson *et al.* 1988; Gibson 1990; Gibson & Ireland 1995), in which Early Cretaceous granulite-facies rocks are overlain structurally by Palaeozoic (Western Province) amphibolite-facies metasedimentary and metaplutonic rocks. The Cretaceous granulites, collectively termed the Western Fiordland Orthogneiss by Bradshaw (1990), were derived from plutonic protoliths of dioritic composition and are comparable to Archaean rocks of the trondhjemite–tonalite–dacite association. Muir *et al.* (1995) regarded the Western Fiordland Orthogneiss as the lower crustal equivalent of the Early Cretaceous Separation Point Batholith of NW Nelson (Fig. 1).

To the east of the western belt, occupying a structural position known as the Median Tectonic Zone (Bradshaw 1993; Kimbrough *et al.* 1993, 1994), lies the Eastern Fiordland Igneous Belt (term modified after Oliver & Coggan 1979). The belt is dominated by Mesozoic plutonic rocks with minor (<5% by area) metamorphosed volcanic and sedimentary units. In Eastern Fiordland, the contact between these rocks and the Western Province appears to be tectonic—the Wall Mountain Fault system of Gibson (1992). However, there are several lines of evidence which suggest that the Median Tectonic Zone arc system developed in close proximity to the eastern margin of the Western Province. In the context of the present study, the intrusion of the 137 Ma Crow Granite in NW Nelson (Muir *et al.* 1997) suggests that at least in the latter part of the history of the Zone, the arc and the Western Province continental margin were contiguous. For a more detailed discussion of the nature of the boundaries of the Median Tectonic Zone, see Bradshaw (1993) and Kimbrough *et al.* (1993, 1994).

During the course of the present study, four distinct groups of plutonic rocks have been identified in eastern Fiordland on the basis of geochronology, petrography and geochemistry. These are:

(1) Palaeozoic plutonic rocks
(2) Triassic plutonic rocks
(3) Mid-Jurassic–Early Cretaceous plutonic rocks
(4) Early Cretaceous Separation Point-type granitoids

U–Pb zircon SHRIMP geochronology

U–Pb zircon crystallization ages for representative samples from each of the above groups were determined using the SHRIMP I and SHRIMP II instruments at the Australian National University, Canberra. The zircon data are presented in the form of Tera–Wasserburg concordia diagrams (Tera & Wasserburg 1972) in Figs 4, 5 and 6, and are compiled in Table 1. Results are presented in more detail in Supplemenatary Publication No. SUP 18121 (23pp.), which is available from the Society Library or from the British Library Document Supply Centre, Boston Spa, Wetherby, W. Yorks, LS23 7BQ. A full list of the analyses is available from the authors on request. The results obtained are presented in chronological order and are discussed with respect to the distribution of rock types in eastern Fiordland.

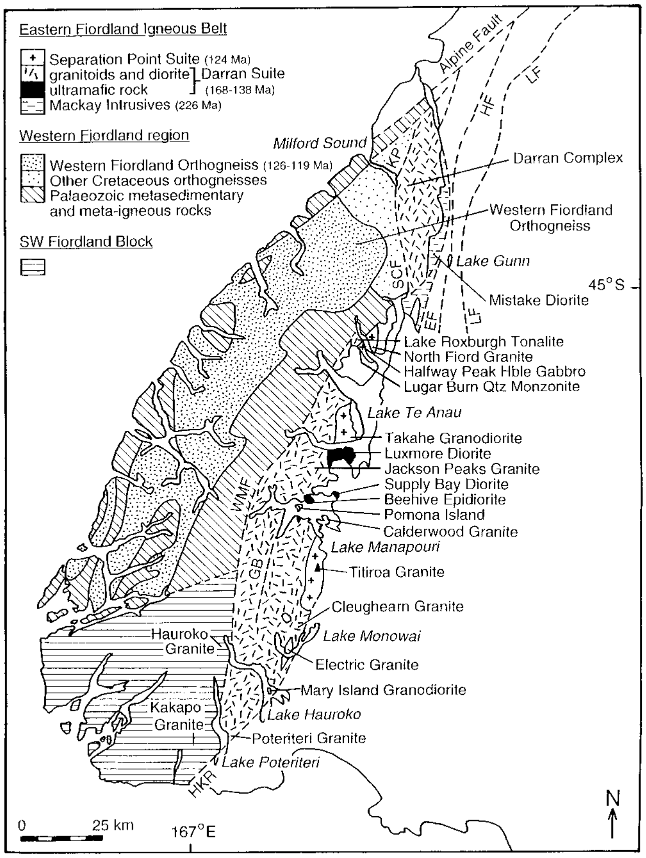

Fig. 3. Geological map of Fiordland (modified after Oliver & Coggan (1979) and Bradshaw (1990)) showing location of SHRIMP samples and features mentioned in the text. E, Electric Granite; B, Loch Burn Formation; WMF, Wall Mountain Fault; GB, Grebe Fault; HKR, Hauroko Fault; LF, Livingstone Fault; HF, Hollyford Fault; EF, Eglinton Fault; GF, Glade Fault; SCF, Surprise Creek Fault; KP, Kaipo Fault/Shear Zone.

Chronological framework

Palaeozoic plutonic rocks

Palaeozoic rocks occur mainly to the western side of the Median Tectonic Zone. In SW Fiordland (Fig. 3), they include the Hauroko Granite (new name) on the NW side of Lake Hauroko, the Kakapo Granite (Wood 1960; Bishop 1986) west of Lake Poteriteri and the Poteriteri pluton (Turnbull & Uruski 1995) on the eastern shore of Lake Poteriteri.

Hauroko Granite. The Hauroko Granite is a leucocratic medium- to coarse-grained biotite granite that intrudes schistose, semi-pelitic, metasedimentary rocks of uncertain age in the Princess Mountains. During the course of this study, an intrusive contact between granite and schist was found on the NW shore of Lake Hauroko at grid reference [S157/316 613]. A sample of granite (HKO14) from this locality was analysed, along with another sample from the lakeshore below Caroline Peak (HKO16). Both samples have relatively simple zircon populations and yield well-constrained ages of 358 ± 4 Ma and 359 ± 5 Ma (Early Carboniferous) respectively (Fig. 4a and b). Both granites contain a small proportion of 370 Ma inheritance as well as older material. The crystallization age of the Hauroko Granite is *c.* 15 Ma younger than the Paleozoic granitoids that make up the bulk of the Karamea Batholith in NW Nelson (Muir *et al.* 1994, 1996*a*).

Kakapo Granite. The Kakapo Granite (Wood 1960) intrudes Ordovician metasedimentary rocks (Preservation Formation), and is regarded by many workers (e.g. Tulloch 1983; Bishop 1986) as being equivalent in age to the Palaeozoic granitoids forming the Karamea Batholith in NW Nelson (Fig. 1). The U–Pb data show a high degree of dispersion and there is no indication of clustering around a magmatic age from which Pb-loss or inheritance could be inferred (Fig. 4c). Upon rejection of several high U analyses, the distribution becomes broadly bimodal with a broad peak at 390–370 Ma and a tighter cluster at *c.* 340 Ma. Palaeozoic ages of *c.* 340 Ma have been obtained from foliated granitoid enclaves within the Western Fiordland Orthogneiss (Bradshaw & Kimbrough 1991) and from the structurally overlying cover sequence in the Doubtful Sound area (Oliver 1980; Gibson *et al.* 1988). Ages of around 340 Ma have also been obtained from M2 (kyanite-grade) metamorphic monazites from the cover sequence (Ireland & Gibson in press). The best interpretation of the U–Pb data is that the Kakapo Granite is *c.* 340 Ma old with 370–390 Ma inheritance. An alternative would be that the Kakapo Granite is 370 Ma with 390 Ma inheritance and substantial Pb loss occurred at 340 Ma. We prefer the former scenario but more analyses of different samples of Kakapo Granite will probably be required before a conclusive answer is forthcoming.

Poteriteri Pluton. The Poteriteri Pluton (Turnbull & Uruski 1995) on the eastern shore of Lake Poteriteri (Fig. 3) is a pink, medium- to coarse-grained, equigranular biotite granite. Turnbull & Uruski (1995) inferred that the Poteriteri pluton is the youngest major intrusion in SW Fiordland. However, from the available SHRIMP data a Carboniferous age of *c.* 330 Ma is indicated (Fig. 4d). The zircons in this sample give scattered ages and a more definitive statement of its age is not warranted from the data. Nevertheless, it would appear to be much older than the relative age assignment offered by Turnbull & Uruski (1995).

Lake Roxburgh Tonalite. In addition to the Carboniferous granitoids in SW Fiordland, Palaeozoic plutonic rocks occur along the western side of, and within the Median Tectonic Zone in eastern Fiordland. Mid-Palaeozoic plutonic rocks which crop out between the Middle and North Fiords on Lake Te Anau have been collectively named the Lake Roxburgh Tonalite (Turnbull 1985). Kimbrough *et al.* (1994) reported U–Pb zircon TIMS (thermal ionization mass spectrometry) ages of 337^{+3}_{-1} Ma and 343^{+5}_{-5} Ma for two samples of the tonalite. These Early Carboniferous ages are in good agreement with a SHRIMP age of 344 ± 4 Ma (Fig. 4f) for a sample of the tonalite (NF11) from North Fiord, Lake Te Anau, immediately below Lake Roxburgh. Further south, dioritic rocks on the north side of Pomona Island, Lake Manapouri (LM15) gave a similar Early Carboniferous age of 345 ± 4 Ma (Fig. 4e).

Pomona Island Granite. Late Carboniferous rocks crop out extensively on Pomona Island and along the south shore of Lake Manapouri. Aronson (1968) reported a Rb–Sr whole

Table *Summary of SHRIMP ages from Fiordland*

Sample no.	Rock type	Grid reference	Location	Age	MSWD	Comments
TT6	Titiroa Granite	S158-612 888	Mount Titiroa	120.9 ± 1.8 Ma	0.86	13/15 (Inh)
SF2A	Takahe Granodiorite	S140-728 298	Honeymoon Beach, South Fiord, Lake Te Anau	123.3 ± 1.8 Ma	0.53	13/16 (Inh)
NF3	North Fiord Granite	S130-789 605	North Fiord, Lake Te Anau	123.7 ± 1.8 Ma	0.60	13/16 (Inh)
WFO1	Western Fiordland Orthogneiss	S120/1-723 942	West of Lake Quill, Franklin Mountains	125.9 ± 1.9 Ma	0.55	20/20*
MW4	Electric Granite	S158-523 615	Tangney Bend, Lake Monowai	136.7 ± 1.6 Ma	1.88	7/12 (Inh, PbL, hiPbc)
DN1	Felsic sheet (Darran Complex)	S122-935 095	Big Slip from Mount Underwood, Milford Road	136.8 ± 1.9 Ma	1.87	Bimodal (Inh = 148 ± 3)
DN2	Darran Diorite (Darran Complex)	S122-935 095	Big Slip from Mount Underwood, Milford Road	138.0 ± 2.9 Ma	0.67	12/12*
NF15	Lugar Burn Quartz Monzonite	S130-722 631	Lugar Burn, North Fiord, Lake Te Anau	138.3 ± 2.1 Ma	1.19	11/14 (Inh, hiPbc)
NF16	Halfway Peak Hornblende Gabbro	S130-754 632	North Fiord, Lake Te Anau	146.0 ± 2.2 Ma	0.54	16/16*
LM25	Calderwood Granite	S149-575 042	Calderwood Peninsula, Lake Manapouri	145.6 ± 2.6 Ma	0.65	Bimodal (Inh = 154 ± 3)
LM4	Beehive Epidiorite	S149-601 098	East side of the Beehive, Lake Manapouri	148.6 ± 2.3 Ma	1.16	12/12*
HKO5	Mary Island Granodiorite	S167-466 495	East side of Mary Island, Lake Hauroko	152.5 ± 2.4 Ma	1.69	10/12 (PbL)
GL5	Cleughearn Granite (coarse)	S158-513 693	Cleughearn Peak, south of Green Lake	153.8 ± 2.3 Ma	1.11	13/13*
MW5	Cleughearn Granite (fine)	S158-524 615	Tangney Bend, Lake Monowai	154.5 ± 2.9 Ma	1.45	14/16 (PbL)
KP6	Luxmore Diorite	S140-640 230	Kepler Track, North of Mount Luxmore	158.8 ± 2.3 Ma	1.89	14/14*
KP3	Jackson Peaks Granite	S140-622 222	Jackson Peaks, Kepler Mountains	161.9 ± 2.6 Ma	1.40	12/12*
LM27	Supply Bay Diorite	S149-680 062	Supply Bay, Lake Manapouri	168.4 ± 1.8 Ma	1.58	11/12 (PbL)
MST1	Mistake Diorite	S122-037 882	Melita Stream, west of Lake Gunn	226.4 ± 3.3 Ma	0.77	16/16*
LM17	Pomona Island Granite sheet	S149-560 087	North side of Pomona Island (cutting LM15)	291.5 ± 4.6 Ma	1.73	7/12 (hiU, PbL)
LM23	Manapouri Granite	S149-551 048	South shore of Lake Manapouri	296.6 ± 4.4 Ma	1.89	9/12 (hiU, PbL)
LM9A	Pomona Island Granite	S149-552 075	Southwest side of Pomona Island	305.1 ± 3.9 Ma	1.32	10/14 (PbL, hiPbc)
NF11	Lake Roxburgh Tonalite	S130-724 686	North Fiord, Lake Te Anau	343.9 ± 4.0 Ma	1.55	13/15 (PbL)
LM15	Pomona Island Diorite	S149-560 087	North side of Pomona Island	345.0 ± 3.6 Ma	1.64	10/12 (PbL)
POT2	Poteriteri pluton	S166-323 329	Southeast shore of Lake Poteriteri	*c.* 330 Ma	—	Scattered analyses
KAK1	Kakapo Granite	S166-193 339	Corrie, south of Solitary Peak, Cameron Mountains	*c.* 340 Ma	—	Bimodal (Inh, PbL)
HKO16	Hauroko Granite	S166-361 564	Northwest shore of Lake Hauroko	358.5 ± 5.3 Ma	1.24	11/15 (Inh)
HKO14	Hauroko Granite	S157-315 625	Lake Hauroko, below Caroline Peak	358.0 ± 3.7 Ma	0.98	10/13 (Inh)

Grid references refer to the New Zealand Topographical Map (NZMS 1) 1:63 360 (1 inch to 1 mile) series.

MSWD: Mean square of weighted deviates.

Abbreviations used in comments: 12/13, proportion of analyses used in calculating a mean age. Error on mean age is 2σ and includes error on standard.

*All data included; Inh, inherited components (analyses older than the mean); PbL, Pb loss (analyses younger than mean).

Analyses compromised because of: hiPbc, high common lead, hiU, high Uranium.

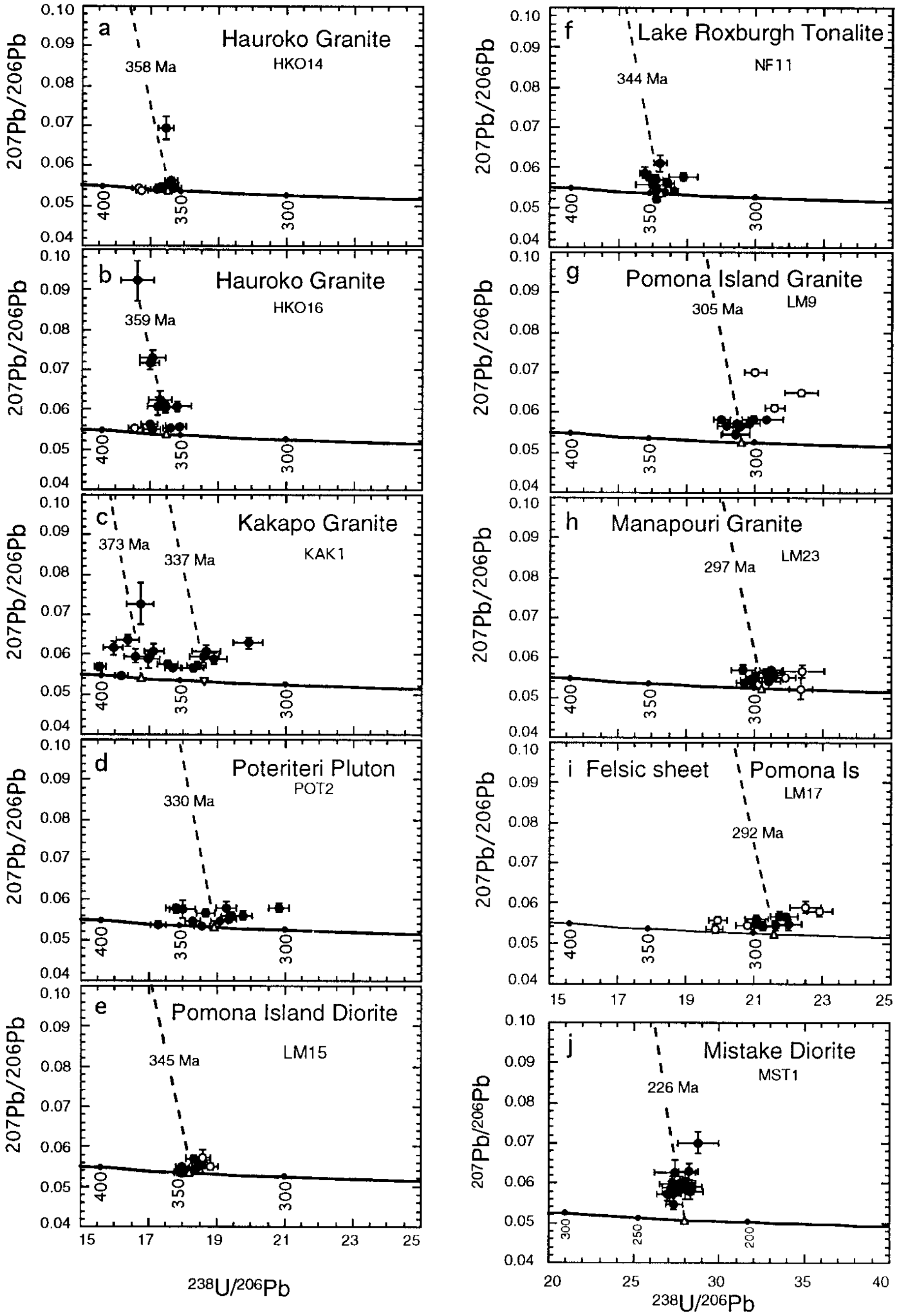

Fig. 4. U-Pb zircon concordia plots (Tera & Wasserburg 1972) for Palaeozoic intrusives from Eastern Fiordland. These plots use the measured Pb isotopic composition (i.e. not corrected for common Pb) and so an analysis must have no common Pb and have coincident $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ages to be concordant in this representation. Common Pb is assessed through the $^{207}Pb/^{206}Pb$ ratio on the basis that any point is a mixture between common Pb and radiogenic Pb, ie analyses are regressed from the estimated common Pb through the analysis to concordia.

Dashed line represents common Pb regression line and age of mean. Filled circles are data points included in the regresion. Open circles are data points omitted from the regression. Error bars are 1σ.

rock-mineral age of *c.* 110 Ma from Pomona Island, and Kimbrough *et al.* (1994) obtained U–Pb zircon ages of*c.* 155 Ma from the NE corner of the island and from the shores of Lake Manapouri. Jurassic and Cretaceous plutonic rocks are widespread in the Manapouri area (see below). However, we find that the bulk of the pink granite on Pomona Island and on the lakeshore immediately to the south is Carboniferous in age. A sample of pink granite from the SW side of the island (LM9A) has a SHRIMP age of 305 ± 4 Ma (Fig. 4g), and a sample from the south shore of the lake (LM23) is 297 ± 4 Ma (Fig. 4h). A sheet of pink granite (LM17) cutting the Early Carboniferous diorite (see previous section) on the north side of Pomona Island is 292 ± 5 Ma (Fig. 4i). Notwithstanding some problems with Pb loss affecting these data sets, all of these rocks are much older than the widespread Jurassic–Cretaceous rocks in this region.

Triassic plutonic rocks

Triassic plutonic rocks occur mainly along the eastern side of the Median Tectonic Zone (Kimbrough *et al.* 1994). They include the Mistake Diorite and related rocks in NE Fiordland (Williams & Harper 1978), and the Longwood Complex near the south coast of the mainland (Challis & Lauder 1977). Kimbrough *et al.* (1994) obtained a U–Pb zircon TIMS age of 224^{+6}_{-1} Ma from the Mistake Diorite and ages of 247–207 Ma from the Longwoods Complex. A sample of Mistake Diorite

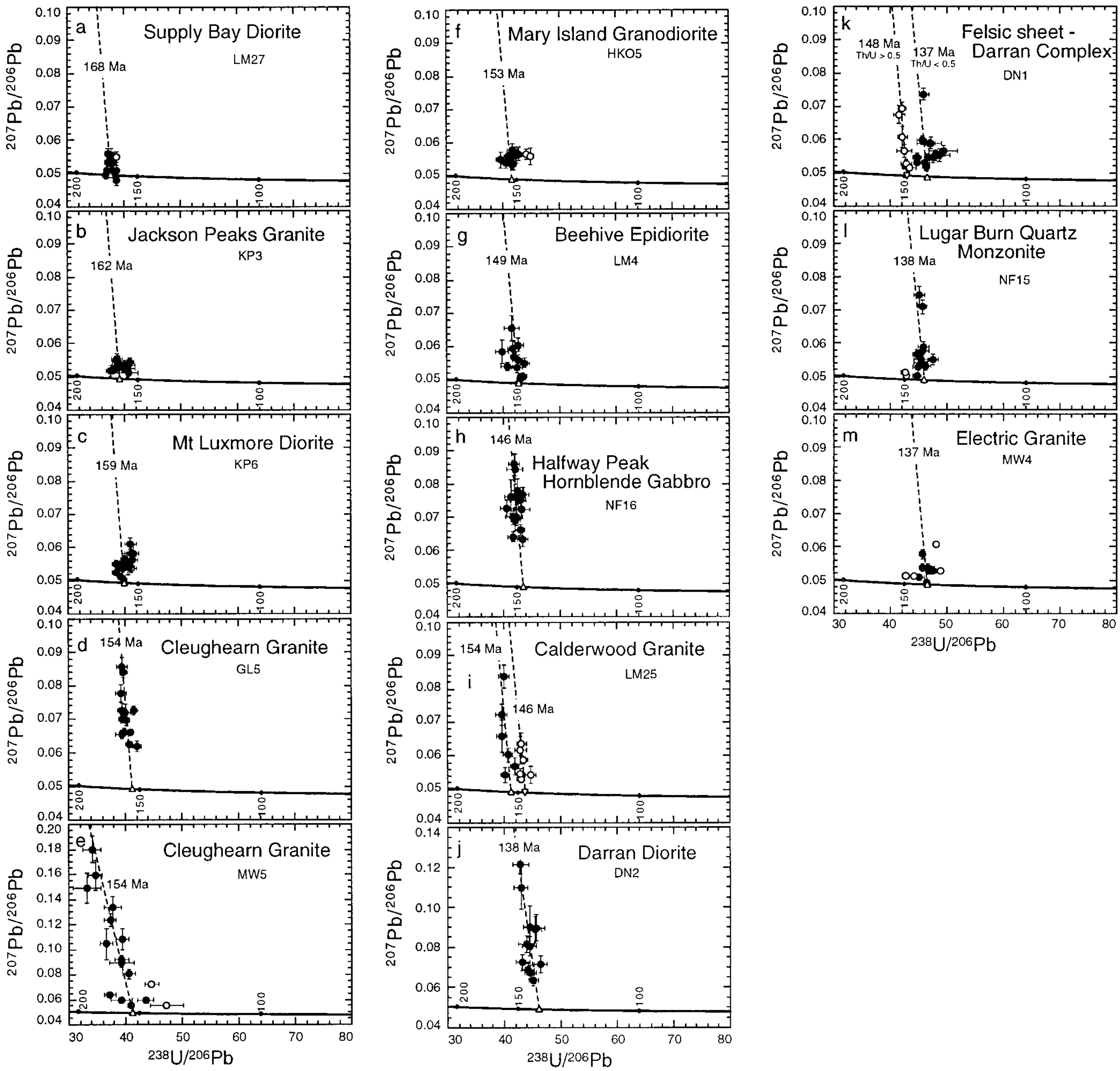

Fig. 5. U–Pb zircon concordia plots for the Mid-Jurassic–Early Cretaceous plutonic rocks (Darran Suite) from eastern Fiordland. See caption to Fig. 4 for details.

from Melita Stream near Lake Gunn (MST1) gives a SHRIMP U–Pb zircon age of 226 ± 3 Ma (Fig. 4j), in good agreement with the TIMS age.

Jurassic–Early Cretaceous plutonic rocks

Eastern Fiordland is dominated volumetrically by Mid-Jurassic–Early Cretaceous mafic and felsic plutonic rocks in approximately equal proportions. U–Pb zircon ages from these rocks range from 168 to 137 Ma (Kimbrough *et al.* 1994; this study). We regard all of these rocks as products of the same Mesozoic magmatic arc system and for historical reasons (see Bradshaw 1990, p. 479) refer to them collectively as the Darran Suite.

The oldest part of the Darran Suite that we have identified so far is a diorite from Supply Bay on the eastern side of Lake Manapouri (Fig. 3). The zircons from this sample yield a crystallization age of 168 ± 2 Ma (Fig. 5a). The Jackson Peaks Granite (new name) in the Kepler Mountains, and diorite from the ultramafic–mafic complex on Mount Luxmore (Fig. 3) give SHRIMP ages of 160 ± 3 and 159 ± 2 Ma respectively (Fig. 5b and c). Within error limits are the Cleughearn Granite (new name), north of Lake Monowai at 154 ± 2 Ma (Fig. 5d), fine-grained Cleughearn Granite on the shores of Lake Monowai at 154 ± 3 Ma (Fig. 5e), and foliated granodiorite from Mary Island on Lake Hauroko at 153 ± 2 Ma (Fig. 5f). The Beehive Epidiorite (Turner 1937) on Lake Manapouri gives a slightly younger SHRIMP age of 149 ± 2 Ma (Fig. 5g). The Calderwood Granite (new name) on Lake Manapouri has a bimodal distribution of 154 ± 3 Ma and 146 ± 3 Ma zircons (Fig. 5i). Either age is consistent with magmatic activity in the area. We prefer the younger age since a coincident predominance at the younger age through Pb loss is unlikely. In North Fiord, Lake Te Anau, the Halfway Peak Hornblende Gabbro (Turnbull 1985) gave a SHRIMP age of 146 ± 2 Ma (Fig. 5h), coincident with the younger (magmatic) age in the Calderwood Granite and a TIMS age of 146^{+4}_{-1} Ma (Kimbrough *et al.* 1994).

In northern Fiordland, the Darran Complex (Blattner 1978) comprises mainly two-pyroxene and hornblende diorites that lie between the Early Cretaceous Western Fiordland Orthogneiss and the Triassic Mistake Diorite (Fig. 3). U–Pb zircon TIMS ages from the Darran Complex indicate emplacement at *c.* 138 Ma (Mattinson *et al.* 1986; Kimbrough *et al.* 1994). A sample of the Darran Diorite (DN2) from the Milford Road below Mount Underwood (Fig. 2) gave a U–Pb zircon SHRIMP age of 138 ± 3 Ma (Fig. 5j). A felsic sheet (DN1) cutting the diorite at this locality also gave a similar age (137 ± 2 Ma), but approximately one third of the zircons in the vein are *c.* 10 Ma older, with a mean age of 148 ± 3 Ma (Fig. 5k). The Lugar Burn Quartz Monzonite (Turnbull 1985) gives a similar age of 138 ± 2 Ma and also has *c.* 150 Ma inheritance (Fig. 5l). The SHRIMP age of the Lugar Burn Quartz Monzonite is within error of a TIMS age of 141^{+7}_{-1} Ma obtained by Kimbrough *et al.* (1994).

The distinctive aegirine- and arfvedsonite-bearing peralkaline granite which crops out around Lake Monowai was named the Electric granite by Turnbull & Uruski (1995). Kimbrough *et al.* (1994) obtained a U–Pb zircon TIMS age of 134 ± 2 Ma from the Electric Granite at Tangney Bend. A sample from the same locality (MW4) has given a slightly older SHRIMP age of 137 ± 2 Ma and is complicated through the presence of both Pb loss and inheritance in this sample (Fig. 5m), The SHRIMP age is similar to that of the Darran Complex in northern Fiordland (cf. Fig. 5j).

Early Cretaceous Separation Point-type plutons

Na-rich granitoids, similar to the rocks forming the Early Cretaceous Separation Point Batholith in NW Nelson (Muir *et al.* 1995), occur on the west side of Lake Te Anau (Turnbull 1985) and farther south on Mount Titiroa (Fig. 3). Higgins & Kawachi (1977) referred to the belt of granitoid rocks between Lake Monowai and Mount Titiroa (including the area occupied by the Cleughearn Granite) as the Green Lake Granodiorite. However, the Cleughearn Granite, north of Lake Monowai is significantly older than the granite at Mount Titiroa (see above) and is not included in the Separation Point Suite plutons. Kimbrough *et al.* (1994) obtained a U–Pb zircon TIMS age of 118^{+5}_{-1} Ma from the North Fiord Granite, but SHRIMP ages for the North Fiord Granite, Takahe Granodiorite (Turnbull 1985) and Titiroa Granite (new name) are all slightly older at *c.* 121–124 Ma (Fig. 6). The systematics of all these granites are complicated by the presence of small components of inheritance as old as *c.* 180 Ma (in the Takahe granodiorite).

A sample of the Western Fiordland Orthogneiss from the type area near Lake Quill in the Franklin Mountains (Bradshaw 1990) gives an Early Cretaceous age of 126 ± 2 Ma (Fig. 6d). This lies within the range 126–119 Ma obtained by Mattinson *et al.* (1986), McCulloch *et al.* (1987), Gibson *et al.* (1988), and Gibson & Ireland (1995) for the orthogneiss protoliths by independent methods.

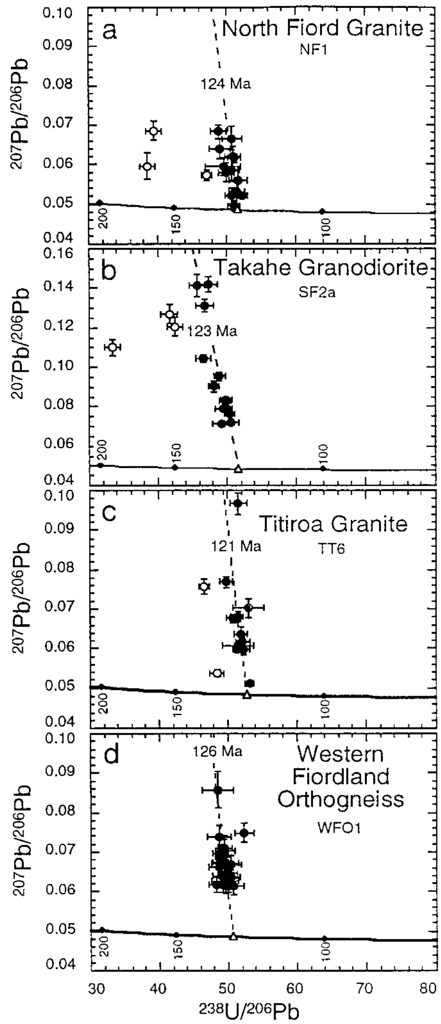

Fig. 6. U–Pb zircon concordia plots for the Western Fiordland Orthogneiss and Separation Point type plutons from eastern Fiordland. See caption to Fig. 4 for details.

Summary of geochronological data and regional correlation

The new SHRIMP ages confirm that the bulk of the plutonic rocks within the Median Tectonic Zone in Eastern Fiordland are Mid-Jurassic–Early Cretaceous (168–137 Ma) in age. Palaeozoic granitoids occur in SW Fiordland and Carboniferous age rocks occur immediately west of (e.g. Lake Roxburgh Tonalite), and within (e.g. Pomona Island) the Median Tectonic Zone. Triassic plutonic rocks (e.g. Mistake Diorite) appear to be restricted to the eastern side of the zone (Kimbrough *et al.* 1994). The Early Carboniferous Hauroko and Kakapo Granites are clearly part of the Western Province (both intrude Palaeozoic metasedimentary rocks) and the Early Carboniferous Pomona Island Diorite and Granite are an integral part of the Median Tectonic Zone. However, it is unclear whether the Early Carboniferous Lake Roxburgh Tonalite and Poteriteri plutons are part of the Western Province or Median Tectonic Zone. Further work on the boundaries of these units will be required to establish their field relationships and affinities.

It can be seen from the time–rock plot in Fig. 7 that the chronology of magmatic events in Fiordland bears a close

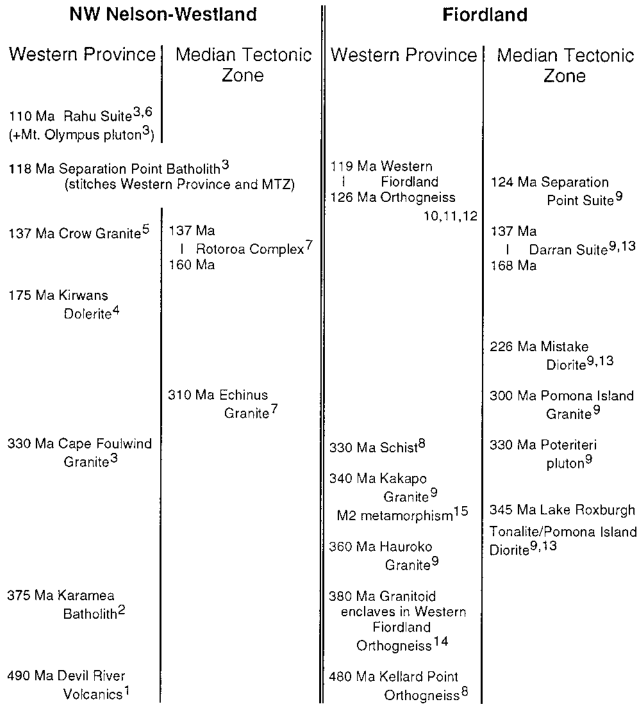

Fig. 7. Time–rock plot for the Western Province and Median Tectonic Zone in NW Nelson-Westland and Fiordland. Data sources: 1, Munker & Cooper (1995); 2, Muir *et al.* (1996*a*); 3, Muir *et al.* (1994); 4, Mortimer *et al.* (1995); 5, Muir *et al.* (1997); 6, Waight *et al.* (1997); 7, Kimbrough *et al.* (1993); 8, Gibson & Ireland (1996); 9, this study; 10, McCulloch *et al.* (1987); 11, Gibson *et al.* (1988); 12, Gibson & Ireland (1995); 13, Kimbrough *et al.* (1994); 14, Bradshaw & Kimbrough (1991); 15, Ireland & Gibson in press.

resemblance to events in the NW Nelson–Westland region. In detail, there appears to be a more extensive range of Palaeozoic magmatism in Fiordland, but overall, a broad similarity of igneous events is apparent.

The Early Cretaceous Crow Granite (Muir *et al.* 1997) and Late Jurassic–Early Cretaceous plutonic rocks forming the Rotoroa Complex in NW Nelson (Challis *et al.* 1994; Kimbrough *et al.* 1993) are probably equivalent to the voluminous 168–137 Ma rocks in Eastern Fiordland. The Rotoroa Complex is stitched to the Lower Palaeozoic rocks of the Takaka terrane by the Early Cretaceous (118 Ma) Separation Point Batholith (Muir *et al.* 1995). Separation Point magmatism in NW Nelson appears to be 4–5 Ma younger than the Separation Point plutons in Fiordland, indicating that there was along strike diachroneity in the arc system. Equivalents of the mid-Cretaceous Rahu Suite/Hohonu Supersuite granitoids in Westland (Waight *et al.* 1997) have yet to be identified in Fiordland.

No intrusive relationships between the Mid-Jurassic–Early Cretaceous (168–137 Ma) Darran Suite plutonic rocks and older Palaeozoic basement have been observed in the Fiordland region. However, the Crow Granite, which intrudes the Buller terrane in NW Nelson, has an age of 137 Ma (Muir *et al.* 1997), characteristic of the Median Tectonic Zone. This raises the possibility that the Median Tectonic Zone arc was tectonically linked, or at least in close proximity to the Western Province at this time.

Crustal inheritance

The new SHRIMP ages are in good agreement with previous U–Pb zircon TIMS age determinations by Kimbrough *et al.* (1994) on the same intrusions. Any differences in age can generally be ascribed to a small component of inheritance or Pb loss which renders any discordant TIMS ages difficult to interpret because of the model dependence of the Pb loss and/or inheritance trajectory. Most of the samples have only limited inheritance and Pb loss is also minimal for the Jurassic–Cretaceous samples analysed, and hence good agreement is to be expected. However, Kimbrough *et al.* (1994) reported significant crustal inheritance in samples from the Lake Manapouri region. Zircon fractions from one particular sample (undeformed biotite granite) gave $^{206}Pb^{*}/^{238}U$ ages of *c.* 150 Ma, $^{207}Pb^{*}/^{206}Pb^{*}$ ages of *c.* 300 Ma, and an upper intercept age of *c.* 1000 Ma. Apart from inherited Jurassic components in the Darran and Separation Point Suite rocks, we detected no Precambrian or Palaeozoic inherited material. Sr and Nd isotope data (see below) confirm that the Mid-Jurassic–Early Cretaceous magmas have undergone little, or no crustal interaction.

Geochemistry

Representative geochemical analyses of the plutonic rocks in Fiordland are presented in Table 2 and Sr and Nd isotope data are presented in Table 3. The bulk of this section deals with the geochemistry of the Mesozoic plutonic rocks. The limited amount of data available for the Palaeozoic and Triassic rocks precludes a rigorous discussion of their geochemistry and petrogenesis. The Hauroko Granite, Pomona Island Diorite, Lake Roxburgh Tonalite, Pomona Island Granite and Mistake Diorite are all Na-rich, metaluminous to weakly peraluminous, I-type, calc-alkaline igneous rocks (Table 2). In contrast, the Kakapo and Poteriteri Granites are closely similar to the high-K calc-alkaline granitoids forming the Karamea Batholith of NW Nelson, in terms of their petrography and chemical composition (Muir *et al.* 1996*a, b*).

Jurassic–Early Cretaceous plutonic rocks

Forty-four samples from the Darran Suite were analysed for major and trace element concentrations. On a Peacock diagram (Fig. 8a), the rocks have a Peacock Index of *c.* 58, corresponding to a calc-alkaline designation. Classification by the aluminium saturation index (ASI) of Zen (1986) indicates that the samples are metaluminous to weakly peraluminous (ASI $<$ 1.1) (Fig. 8b).

Major and trace element variations are illustrated on selected Harker diagrams in Fig. 9. Taken collectively, the samples exhibit a wide range in SiO_2 content from 43 to 76 wt%. However, a distinct compositional gap between 60 and 65 wt% SiO_2 separates the more mafic diorites and gabbroic rocks (e.g. Mount Luxmore Intrusives, Beehive Epidiorite, Halfway Peak Hornblende Gabbro, Darran Diorite) from the more felsic granites and granodiorites (e.g. Jackson Peaks Granite, Cleughearn Granite, Mary Island Granodiorite, Calderwood Granite, Lugar Burn Quartz Monzonite)

Harker diagrams exhibit generally continuous trends; TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO and P_2O_5 abundances decrease with increasing SiO_2, whereas Na_2O and K_2O increase. The trace elements exhibit considerably more scatter than the major elements; Sr, Cr, Ni, V and Zn decrease generally with increasing SiO_2, Rb and Ba increase with SiO_2, Zr increases initially, but decreases in samples having $>$65 wt% SiO_2. The

Table 2. *Representative analyses of plutonic rocks from Eastern Fiordland and adjacent regions*

	HKO16	KAK1	POT2	LM15	NF11	LM9A	MST1	KP3	KP6	ELX1	GL2	GL5	MW5	HKO5	LM25	LM4	NF16	NF15	DN2	DN1	MW4	WFO1	WFO7	TT6	NF3	SF2A
SiO_2	70.35	73.83	75.98	53.28	64.68	75.43	59.68	71.26	65.69	43.04	72.05	74.19	74.34	69.96	73.33	47.86	50.19	66.84	52.27	73.30	74.97	56.43	57.95	72.96	71.18	69.88
TiO_2	0.40	0.23	0.10	1.11	0.57	0.19	1.02	0.33	0.58	0.69	0.31	0.22	0.22	0.38	0.17	1.53	1.16	0.56	1.19	0.14	0.27	1.22	1.07	0.12	0.16	0.20
Al_2O_3	14.88	13.85	12.99	17.51	16.99	12.86	16.53	14.59	16.44	20.29	14.32	13.66	13.51	15.25	13.93	17.79	18.39	15.99	18.31	14.93	11.41	17.74	16.83	16.06	16.00	17.12
Fe_2O_3	3.25	2.14	1.36	9.21	4.64	1.62	6.59	2.13	4.69	11.08	2.39	1.85	1.48	3.09	1.38	12.75	10.00	3.24	10.03	1.27	3.10	6.96	6.14	1.08	1.30	1.37
MnO	0.07	0.05	0.03	0.24	0.11	0.04	0.10	0.08	0.08	0.12	0.07	0.06	0.02	0.07	0.05	0.20	0.17	0.11	0.16	0.03	0.19	0.10	0.08	0.05	0.04	0.04
MgO	0.93	0.54	0.18	4.01	1.39	0.34	3.21	0.62	1.31	8.94	0.64	0.45	0.32	0.80	0.30	5.22	5.53	0.95	4.33	0.44	0.15	3.74	3.22	0.27	0.40	0.44
CaO	2.67	1.45	0.52	6.55	4.53	0.79	5.66	1.35	3.31	13.82	2.05	1.60	1.26	2.65	0.87	9.93	9.21	2.35	8.82	3.06	0.16	5.95	5.04	1.92	1.87	2.56
Na_2O	3.40	3.33	3.34	4.08	4.93	4.08	4.29	4.57	4.12	1.17	4.19	3.92	3.35	4.14	4.15	3.18	3.80	5.02	3.70	5.05	5.14	5.06	4.95	5.16	5.23	6.22
K_2O	3.13	4.04	4.89	1.46	1.21	4.23	2.23	4.42	3.17	0.08	3.82	3.81	4.87	2.96	4.45	0.55	0.84	3.48	0.75	0.72	4.23	2.28	2.53	2.89	2.97	1.67
P_2O_5	0.15	0.06	0.03	0.32	0.20	0.06	0.28	0.09	0.18	0.02	0.09	0.06	0.05	0.12	0.05	0.72	0.25	0.19	0.26	0.06	0.02	0.41	0.37	0.06	0.08	0.08
LOI	1.18	0.89	0.95	2.60	1.17	0.47	0.73	0.0	− 0.37	0.83	0.20	0.40	0.33	0.63	1.79	− 0.62	0.75	0.93	0.56	0.76	0.30	0.59	0.51	− 0.13	0.58	0.62
Total	100.42	100.42	100.37	100.36	100.43	100.12	100.33	99.45	99.20	100.07	100.14	100.21	99.75	100.05	100.47	99.10	100.31	99.66	100.40	99.75	99.94	100.47	98.69	100.44	99.79	100.20
Ga	18	15	14	22	21	16	24	17	15	15	19	13	14	13	14	17	15	16	18	16	29	20	19	16	19	21
Pb	15	28	18	11	7	19	19	16	29	<1	27	26	21	23	21	10	15	31	15	28	45	17	22	26	37	24
Rb	124	167	153	57	31	90	70	101	75	1	75	115	183	66	99	5	18	102	12	22	124	49	82	71	71	36
Sr	572	149	55	418	545	67	682	224	406	882	237	202	178	371	171	1008	1002	333	757	579	3	1240	1087	630	687	1015
Th	13	18	11	4	4	21	8	18	9	2	21	15	36	6	19	3	<1	15	<1	2	31	2	5	6	3	1
Y	19	29	19	34	12	29	16	13	18	3	13	9	21	6	20	15	19	25	17	<1	78	12	14	5	4	3
V	43	26	14	220	52	19	146	26	53	348	26	21	20	41	19	367	261	46	274	34	13	121	122	19	28	32
Cr	14	12	<3	40	3	<3	63	<3	6	78	<3	<3	<3	3	<3	22	75	<3	26	<3	<3	64	61	<3	<3	<3
Ni	8	7	4	14	4	3	26	5	6	59	4	4	4	4	4	24	33	4	21	4	<3	33	34	3	4	4
Zn	52	39	29	118	72	29	82	32	50	61	29	29	20	44	19	133	107	66	93	22	62	67	83	35	45	38
Zr	182	101	82	198	219	139	122	203	262	11	112	82	123	141	112	55	46	234	62	11	1029	140	203	53	56	59
Nb	12	10	7	8	6	26	7	12	7	4	8	7	10	<2	17	5	4	10	6	4	37	8	9	5	4	4
Ba	746	631	270	467	546	189	500	780	713	21	588	800	869	1422	900	229	359	817	326	363	41	841	708	891	946	667
Sc	5.9	5.4	4.0	—	8.2	2.9	14.2	4.1	9.5	32.1	3.5	3.0	3.7	5.6	3.3	29.0	33.2	6.1	33.7	3.9	9.9	13.5	10.5	1.6	2.4	2.3
Cs	5.8	8.5	2.5	—	1.9	0.6	6.1	2.2	2.9	—	1.4	4.3	8.4	2.8	1.9	—	1.4	2.9	—	1.3	0.7	1.8	2.7	2.9	2.7	1.7
La	38.0	32.0	10.2	—	21.8	48.2	23.4	35.0	33.1	1.5	21.1	21.9	30.6	18.0	35.9	14.4	13.1	33.0	11.3	6.8	94.2	28.3	29.2	9.3	11.6	5.5
Ce	80.5	69.3	23.4	—	45.0	93.8	52.2	64.0	64.5	3.1	35.6	37.8	58.4	32.3	73.9	31.4	31.8	67.1	25.4	15.5	219.1	58.3	73.4	17.4	22.0	10.9
Nd	32.4	28.0	14.6	—	21.4	37.7	28.5	26.9	29.3	—	12.7	12.4	22.6	13.1	21.6	17.7	20.6	33.8	17.0	8.3	—	31.8	28.9	7.6	9.3	5.8
Sm	5.67	6.51	2.45	—	3.56	6.61	5.61	3.30	4.95	0.69	2.50	2.11	3.71	2.06	3.15	4.52	5.67	5.71	4.10	1.98	19.79	6.03	5.54	1.22	1.57	0.98
Eu	1.35	0.77	0.56	—	1.40	0.44	1.39	0.61	0.97	0.37	0.51	0.47	0.56	0.75	0.53	1.37	1.54	1.31	1.33	0.39	2.23	1.54	1.32	0.33	0.42	0.35
Gd	4.5	6.8	2.4	—	2.9	6.4	4.5	2.5	4.2	1.2	—	1.6	3.1	1.8	—	4.6	5.0	4.2	4.5	1.6	16.9	3.9	4.2	0.9	1.2	—
Tb	0.72	1.11	0.51	—	0.48	1.03	0.68	0.39	0.65	0.19	0.32	0.20	0.52	0.29	0.59	0.66	0.77	0.71	0.63	0.22	3.05	0.60	0.57	0.14	0.16	0.07
Tm	0.32	0.49	0.35	—	0.25	0.65	0.25	0.22	0.29	0.06	0.22	0.16	0.33	0.13	0.50	0.21	0.33	0.43	0.34	0.06	2.02	0.21	0.19	0.06	—	—
Yb	2.16	3.35	2.49	—	1.74	4.41	1.66	1.53	1.90	0.38	1.63	1.15	2.33	0.90	4.04	1.44	2.20	3.17	2.36	0.37	14.49	1.35	1.30	0.40	0.35	0.32
Lu	0.28	0.48	0.36	—	0.26	0.67	0.25	0.23	0.27	0.05	0.24	0.17	0.35	0.12	0.63	0.21	0.32	0.48	0.33	0.05	2.09	0.20	0.18	0.06	0.05	0.05
Hf	5.3	4.2	3.7	—	7.4	6.5	4.8	5.7	7.6	0.3	3.7	2.5	4.9	4.1	3.8	1.3	1.7	7.3	2.2	2.0	30.0	6.6	6.57	2.0	2.3	2.2
Ta	1.26	1.38	1.08	—	0.36	3.30	0.59	0.72	0.48	0.03	0.63	0.55	1.04	0.30	3.28	0.17	0.17	0.82	0.25	0.25	3.16	0.54	0.58	0.35	0.29	0.25
U	2.38	4.28	1.82	—	1.70	3.85	1.95	2.34	1.28	—	2.80	2.38	8.08	1.58	14.12	0.17	0.20	3.98	—	0.65	4.69	0.96	1.36	0.62	0.56	0.53

Major and selected trace elements (Ga–Ba) by XRF analysis at University of Canterbury (Loss on ignition (LOI) at 1000°C, Fe as total Fe_2O_3). Sc–U by INAA at University of Massachusetts, Lowell.
See Tables 1 and 3 for explanation of sample numbers and locality information.

Table 3. *Rb–Sr and Sm–Nd isotope data*

Sample	Age (Ma)	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	$^{87}Sr/^{86}Sr_i$	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	ε_{Nd}	T_{DM} (Ma)
TT6	121	71	630	0.326	0.704400	0.70383	1.22	7.60	0.0966	0.512728	+3.3	430
SF2A	123	36	1015	0.103	0.703955	0.70377	0.98	5.80	0.1017	0.512727	+3.2	450
NF3	123	71	687	0.299	0.704272	0.70375	1.35	7.78	0.1049	0.512728	+3.2	462
WFO1	126	49	1240	0.114	0.703995	0.70379	6.03	31.80	0.1141	0.512695	+2.4	551
WFO7	126	82	1087	0.218	0.704275	0.70388	5.54	28.90	0.1154	0.512688	+2.3	568
MW4	137	124	3	126.0	1.298177	1.05103	11.76	46.97	0.1513	0.512708	+2.2	831
DN2	138	12	757	0.046	0.703810	0.70372	4.10	16.99	0.1453	0.512782	+3.7	605
NF15	138	102	333	0.886	0.705502	0.70373	5.71	33.80	0.1017	0.512782	+4.5	378
NF16	146	18	1002	0.052	0.703801	0.70369	5.67	20.60	0.1657	0.512797	+3.7	800
LM25	146	99	171	1.67	0.707384	0.70378	5.47	33.12	0.0999	0.512714	+3.4	460
HKO5	153	66	371	0.515	0.704999	0.70387	2.48	14.93	0.1005	0.512746	+4.0	421
GL2	154	75	237	0.915	0.705858	0.70380	2.36	12.30	0.1155	0.512732	+3.5	503
GL5	154	115	202	1.65	0.707412	0.70371	1.691	0.01	0.1020	0.512745	+4.0	428
MW5	155	183	178	2.98	0.710018	0.70332	3.02	17.46	0.1045	0.512730	+3.6	457
KP6	159	75	406	0.534	0.704980	0.70375	4.22	23.42	0.1088	0.512771	+4.4	418
ELX1	159	1	882	0.003	0.703795	0.70379	0.49	1.82	0.1628	0.512817	+4.2	701
KP3	162	101	224	1.300	0.706638	0.70364	7.93	54.52	0.0879	0.512706	+3.6	427
LM9B	305	90	67	3.89	0.721106	0.70450	5.23	28.63	0.1105	0.512722	+4.9	494
NF10	344	40	548	0.211	0.705035	0.70400	2.95	15.83	0.1127	0.512518	+1.4	802
NF11	344	31	545	0.165	0.704997	0.70419	3.49	21.98	0.0959	0.512506	+1.9	706
KAK1	340	167	149	3.25	0.722727	0.70537	2.62	11.61	0.1365	0.512376	−2.2	1315

Rb and Sr concentrations and $^{87}Rb/^{86}Sr$ ratios are derived from x-ray fluorescence data and are generally considered accurate to ±5% and 0.5% respectively. The $^{87}Sr/^{86}Sr$ composition was measured on a Finnigan MAT 262 instrument and is accurate to better than 0.01%. Average $^{87}Sr/^{86}Sr$ determined for NBS987 Sr isotope standard was 0.710192 ± 26 (2σ, *n*=10). Decay constant used for $^{87}Rb = 1.42 \times 10^{-11}\ a^{-1}$.
Sm and Nd concentrations were determined by INAA. The $^{143}Nd/^{144}Nd$ composition was measured on a VG354 fully automated, multiple collector mass spectrometer and is accurate to <20 ppm (2σ). Average $^{143}Nd/^{144}Nd$ determined for the La Jolla Nd isotope standard was 0.511864 ± 12 (2σ, *n*=10). Decay constant of $^{147}Sm = 6.54 \times 10^{-12}\ a^{-1}$ and $^{143}Nd/^{144}Nd_{CHUR} = 0.512638$ and $^{147}Sm/^{144}Nd_{CHUR} = 0.1967$. These values were used to calculate initial epsilon values (Jacobsen & Wasserburg 1980). Nd model ages (T_{DM}) are based on the depleted mantle model of DePaolo (1981). See Table 1 for an explanation of sample numbers and locality information. Additional samples: WFO7, Western Fiordland Orthogneiss, South of Lake Quill, Franklin Mountains [S120/121-735 925]; ELX1, Diorite, Mount Luxmore, Kepler Mountains [S140-668 232] ; GL2, same as GL5; NF10, same as NF11.

remaining trace elements exhibit little systematic variation throughout the whole compositional range.

Rare earth element (REE) data for representative samples are plotted on a chondrite normalized diagram in Fig. 10a. The majority of the felsic plutonic rocks (e.g. KP6, Jackson Peaks Granite) show light REE-enriched patterns ($La_N/Yb_N \approx 12$), with relatively flat heavy REE and small negative Eu anomalies ($Eu/Eu^* \approx 0.75$). The mafic plutonic rocks (e.g. DN2, Darran Diorite) have light REE-enriched patterns with La_N/Yb_N ratios of *c.* 3, but no Eu anomalies. Concentrations of the heavy REE of around 10 times chondritic values in both the mafic and felsic plutonic rocks (apart from HKO5 and DN1) suggest that garnet was absent from the magma source region. In contrast, concentrations of the heavy REE in the Mary Island Granodiorite (HKO5) and in felsic sheets from the Darran Complex (DN1) are much lower (less than 5 times chondritic values) suggesting retention of these elements by residual garnet in the source region. The Mary Island Granodiorite and felsic sheets from the Darran Complex are similar in composition to the Early Cretaceous Separation Point Suite granitoids (see below).

Mantle-normalized multi-element patterns for the Darran Suite rocks show enrichment in the large-ion lithophile (LIL) elements (e.g. Rb, Ba, Th, K) and light REE, and large negative Nb anomalies (Fig. 10b); features that are generally considered characteristic of subduction related magmas (e.g. Saunders *et al.* 1991). The mafic plutonic rocks show large positive Sr anomalies indicative of feldspar accumulation. On a plot of Rb v. Y+Nb (Fig. 11), all of the samples lie in the field of Volcanic Arc Granites (Pearce *et al.* 1984).

The primitive isotopic composition of the Jurassic–Early Cretaceous plutonic rocks and lack of inherited zircon suggests that the magmas have undergone little or no interaction with continental crust. Initial $^{87}Sr/^{86}Sr$ ratios (Table 3) range from 0.7037 to 0.7049 and ε_{Nd} values (Table 3) are all positive (+3 to +4), indicating derivation of the magmas from a source isotopically depleted relative to Bulk Earth. The felsic plutonic rocks have slightly higher $^{87}Sr/^{86}Sr$ ratios than the mafic rocks. On a graph of $^{87}Sr/^{86}Sr$ (i) v. ε_{Nd} (Fig. 12), the samples lie in the NW quadrant within the mantle array.

Electric Granite

The peralkaline Electric Granite is characterized by high SiO_2 (>75 wt%), high alkalis ($Na_2O + K_2O = 8$–9 wt%; with $Na_2O/K_2O = 1.2$), high Ga, Y, Zn, Zr, Nb and the REE (except Eu) (Fig. 10a), and low Sr. These geochemical features, together with the occurrence of alkali-rich amphibole are, in general, characteristic of A-type granites (e.g. Eby 1990). On a graph of Rb v. Y+Nb (Fig. 11) samples of the Electric Granite plot in the Within-Plate Granite field (Pearce *et al.* 1984). A-type granites are commonly associated with continental rifting, however, there is often a close spatial and temporal association of peralkaline felsic igneous rocks with calc-alkaline magmatic arc systems (e.g. Smith *et al.* 1977; Whalen *et al.* 1987; Houghton *et al.* 1992).

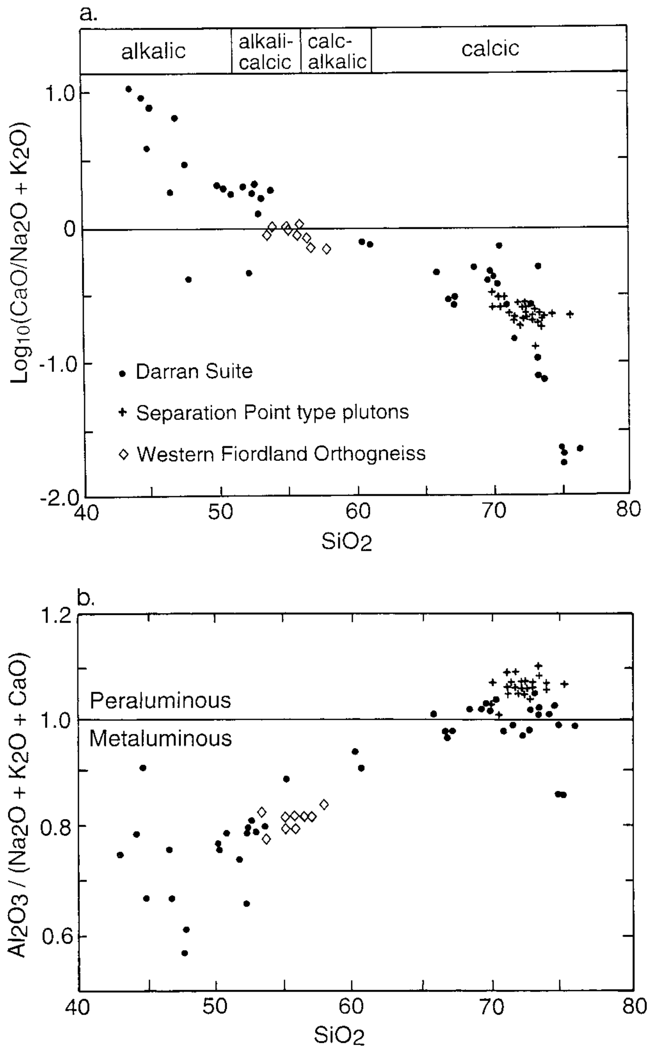

Fig. 8. (**a**) Peacock Index diagram (after Brown 1982). The Peacock Index for a particular rock suite is calculated from the point where the data trend crosses the zero $\log_{10}(CaO/Na_2O+K_2O)$ value. (**b**) Aluminium Saturation Index (Zen 1986). Plot of mol. $Al_2O_3/(CaO+Na_2O+K_2O)$ versus SiO_2 showing the metaluminous to weakly peraluminous composition of the Darran Suite. Samples of the Western Fiordland Orthogneiss are metaluminous (ASI<1) and the Separation Point type plutons are weakly peraluminous (ASI = 1.0–1.1).

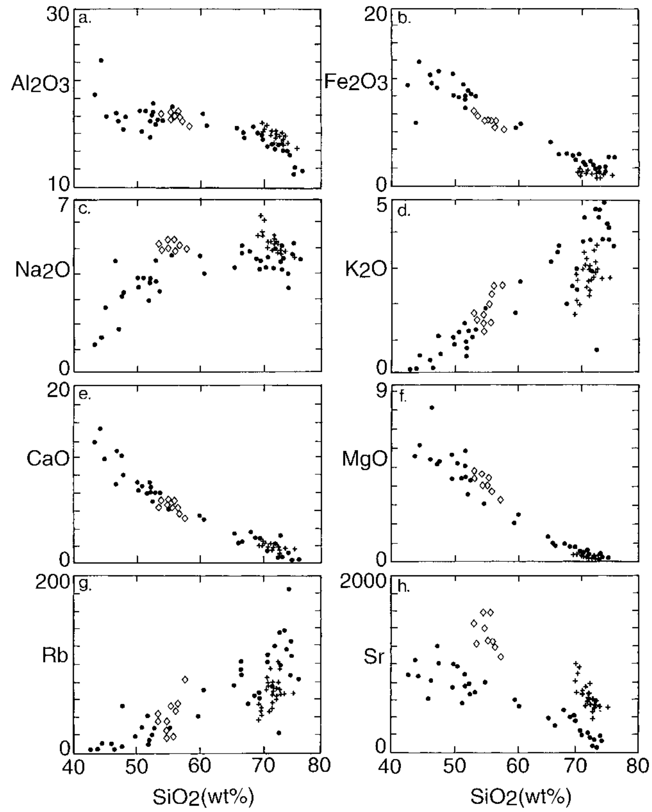

Fig. 9. Selected Harker plots to illustrate the variations of major and trace elements in the Darran Suite, Western Fiordland Orthogneiss and Separation Point type plutons.

Early Cretaceous Separation Point-type granitoids

The granitoids with affinity to the Separation Point Suite (North Fiord Granite, Takahe Granodiorite, Titiroa Granite) are weakly peraluminous (ASI = 1.0–1.1) (Fig. 8) and are characterized by high SiO_2 contents (69–75 wt%); high Al_2O_3 (14–17 wt%); high alkalis (Na_2O + K_2O = 7–8 wt%; with Na_2O/K_2O = 1.8); high Sr (typically >500 ppm and up to 1000 ppm) and Ba (up to 1400 ppm) and low Y (≤5 ppm). Many of these features can be seen on the Harker plots in Fig. 9 and on the multi-element diagram in Fig. 10. On a plot of Rb v. Y + Nb, all of the samples lie in the field of Volcanic Arc Granites (Fig. 11).

Chondrite-normalized REE patterns for these rocks (e.g. Titiroa Granite (TT6); Fig. 10c) are strongly fractionated (La_N/Yb_N>15) with no Eu anomalies. Low concentrations of the heavy REE (down to two times chondritic abundances) suggest that garnet was present in the source region.

Three samples of the Separation Point-type granitoids from eastern Fiordland show a small range of positive ε_{Nd} values (+3) and low $^{87}Sr/^{86}Sr$ initial ratios (*c.* 0.7038), indicating derivation of the magmas from a source isotopically depleted relative to Bulk Earth. On a graph of $^{87}Sr/^{86}Sr$ (i) v. ε_{Nd} (Fig. 12) the samples cluster tightly within the mantle arrray, just below the field for the Darran Suite and above the field for the Separation Point Batholith of NW Nelson (Muir *et al.* 1995)

Western Fiordland Orthogneiss

The dioritic rocks forming the Western Fiordland Orthogneiss are metaluminous (ASI ≈ 0.8) (Fig. 8) and alkali-calcic, with a Peacock Index of 54–57 (McCulloch *et al.* 1987; this study). The rocks are characterized by moderate SiO_2 contents (54–57 wt%), high Al_2O_3 (17–23 wt%), high alkalis (Na_2O+K_2O = 6–9 wt%; with Na_2O/K_2O = 1.5–2.0), high Sr (>1000 ppm) and low Y (<14 ppm). A mantle-normalized multi-element pattern for a typical sample of the orthogneiss shows a marked positive Sr anomaly and negative Rb, Th, Nb, Ti and Y anomalies (Fig. 10d). Chondrite normalized REE patterns (e.g. Fig. 10c) show steep slopes similar to the Separation Point-type granitoids, although total REE concentrations are higher in the Western Fiordland Orthogneiss.

Two samples of the Western Fiordland Orthogneiss that were analysed during the present study have low $^{87}Sr/^{86}Sr$ initial ratios (*c.* 0.70385) and positive ε_{Nd} values (+2.3). On a graph of $^{87}Sr/^{86}Sr(i)$ vs ε_{Nd} (Fig. 12), the two samples lie at the

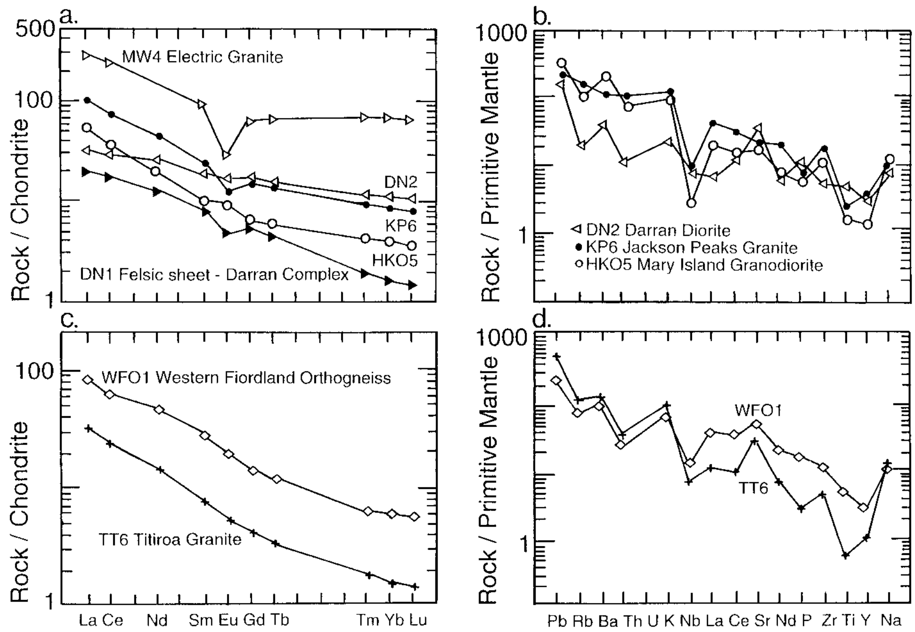

Fig. 10. Mantle normalized multi-element patterns (Sun & McDonough 1989) and chondrite normalized (Nakamura 1974) REE patterns for the Darran Suite, Western Fiordland Orthogneiss and Separation Point Suite.

top end of the Western Fiordland Orthogneiss field of McCulloch *et al.* (1987), and below the fields for the Separation Point-type granitoids and the Darran Suite.

Petrogenetic model

The majority of the Jurassic–Early Cretaceous plutonic rocks (Darran Suite) in Eastern Fiordland have a typical I-type, subduction-related, calc-alkaline chemistry. In contrast, the Early Cretaceous Separation Point plutons are alkali-calcic, Na-rich granitoids, with high concentrations of Sr and low concentrations of Y and the heavy REE. Similar Na-rich rocks have been variably compared with Archaean tonalite–trondhjemite–dacite suites (e.g. Martin 1987), melts of subducted oceanic lithosphere known as adakites (e.g. Defant & Drummond 1990; Drummond *et al.* 1996), and melts of mafic lithosphere underplated beneath a thickened continental arc system (e.g. Atherton & Petford 1993; Petford & Atherton 1996).

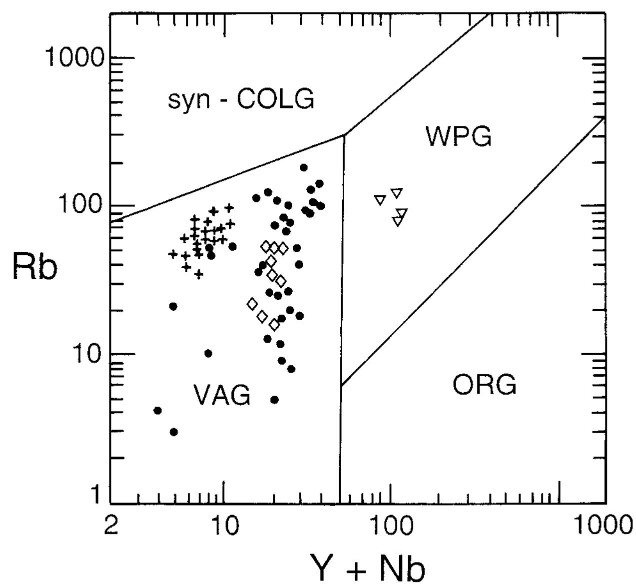

Fig. 11. Rb (ppm) versus Y + Nb (ppm) tectonic discrimination plot of Pearce *et al.* (1984): ORG, ocean ridge granites; VAG, volcanic arc granites; syn-COLG, syn-collision granites; WPG, within-plate granites.

Some Darran Suite rocks, such as the Mary Island Granodiorite (153 Ma) and felsic sheets in the Darran Complex (137 Ma), do possess the distinctive chemistry of the Separation Point Suite. However, they are not volumetrically significant. Any petrogenetic model for the Jurassic–Early Cretaceous plutonic rocks of the Median Tectonic Zone arc system must therefore be capable of explaining the generation of typical calc-alkaline subduction-related magmas from 168 to 137 Ma, and the production and emplacement of large volumes of Na-rich granitoid at around 124 Ma.

Muir *et al.* (1995) presented a petrogenetic model for the granitoid rocks forming the Separation Point Batholith in NW Nelson, similar to that proposed by Atherton & Petford (1993) for the Cordillera Blanca Batholith in Peru. The distinctive geochemical and isotopic features of the batholiths, including high Na_2O and Al_2O_3, low heavy REE and Y contents, together with high Sr/Y (Fig. 13), are those obtained from experimental work on the partial melting of mafic lithologies within the amphibolite–eclogite transition (e.g. Rapp *et al.* 1991; Peacock *et al.* 1994). Atherton & Petford (1993) argued that in the case of the Cordillera Blanca batholith, the Na-rich granitoids were generated by melting of newly underplated basaltic material beneath a thickened continental arc at depths of >40 km. Melting of subducted oceanic lithosphere was precluded by the age of the subducted slab (>60 Ma) and the location of the batholith >300 km inboard from the trench.

Muir *et al.* (1995) demonstrated on the basis of geochemical and tectonic evidence that melting of mafic underplate was also the most plausible origin for the Early Cretaceous Na-rich granitoids in New Zealand. Developing this model further, we propose that mafic rocks currently exposed within the Eastern Fiordland Igneous Belt (e.g. Darran Diorite, Beehive Epidiorite, Halfway Peak Hornblende Gabbro) have the appropriate chemical composition to represent the underplated mafic component.

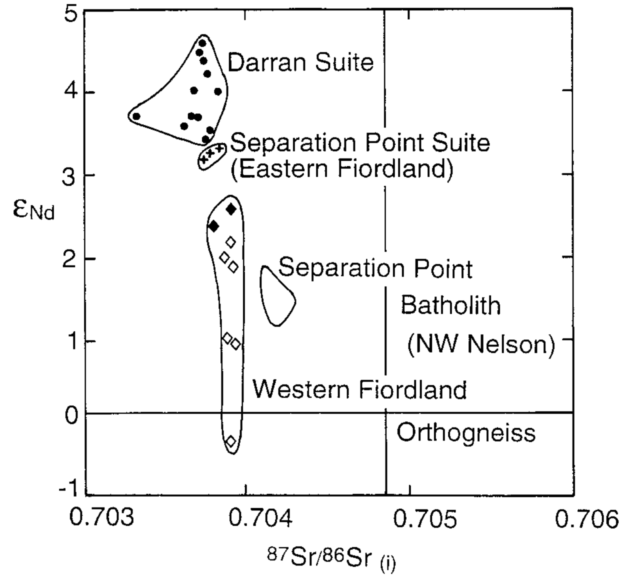

Fig. 12. Plot of $^{87}Sr/^{86}Sr$ (i) versus ε_{Nd}. Fields for Western Fiordland Orthogneiss (WFO) and Darran Suite include data from McCulloch *et al.* (1987). Filled diamonds are data points from this study. Field for Separation Point Batholith from Muir et al. (1995). The vertical array shown by the WFO has been attibuted by McCulloch *et al.* (1987) to the involvement of a light REE-enriched, low Rb/Sr protolith of mid-late Palaeozoic age (*c.* 300 Ma).

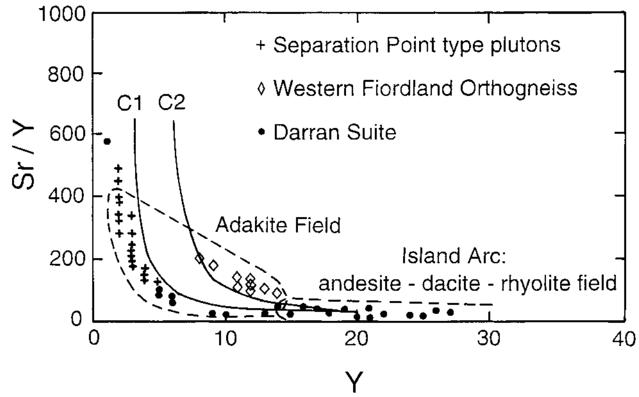

Fig. 13. Plot of Sr/Y versus Y for the Darran Suite (felsic plutonic rocks (>50% SiO_2) only), Western Fiordland Orthogneiss and Separation Point plutons from eastern Fiordland. The adakite and island arc fields are from Defant & Drummond (1990). Partial melting curves for basalt leaving residues of eclogite (C1) and 10% garnet amphibolite (C2) are from Drummond & Defant (1990). Partial melts of a deep (>40 km) basaltic source will have high Sr/Y ratios and low concentrations of Y (i.e. Western Fiordland Orthogneiss and Separation Point type plutons). Apart from the Mary Island Granodiorite and felsic sheets within the Darran Complex, all of the felsic plutonic rocks from the Darran Suite plot within the island arc field.

Petford & Atherton (1996) presented batch melting models of garnet amphibolite to produce the distinctive Na-rich rocks forming the Cordillera Blanca Batholith. One such model for generating heavy REE depleted tonalite from mafic underplate is illustrated in Fig. 14a. Relatively high degrees of partial melting of underplated basalt (33–40%) are required to produce tonalitic melts with steep REE patterns and low concentrations of the heavy REE. For comparative purposes, REE patterns from two Fiordland rocks are also shown (Fig. 14b). The Darran Diorite (DN2) and the Western Fiordland Orthogneis (WFO1) have REE patterns similar to the Peruvian mafic underplate and a modelled partial melt of tonalitic composition, respectively.

We have attempted to reproduce the REE pattern of the Western Fiordland Orthogneiss from the Darran Diorite using a variety of batch melting models. Notwithstanding uncertainties in the source mineralogy and mineral melting rates, reasonably good fits can be obtained using both modal and non-modal batch melting conditions. In all cases, a residue of *c.* 10% garnet and *c.* 30% partial melting appear to be critical. The results of one such model are shown in Fig. 14b. The model indicates that the Western Fiordland Orthogneiss, which is the deep crustal equivalent of the Separation Point Suite (Muir *et al.* 1995), could represent a 30% melt of Darran Diorite, from a garnet amphibolite source with 5% plagioclase feldspar, 11% garnet, 15% amphibole and 63% clinopyroxene. The Separation Point type granitoids have much lower total REE concentrations than the Western Fiordland Orthogneiss, which could reflect smaller degrees of partial melting (<20%), or a greater amount of residual garnet (>10%) in the source region.

The Darran Complex in northern Fiordland is considered to be a high-level intrusive complex (Blattner 1978; Williams & Harper 1978) and has not been buried to lower crustal levels. However, the results of our geochemical modelling demonstrate that a compositional equivalent of the Darran Diorite could have partially melted within the roots of the magmatic arc system to form the Western Fiordland Orthogneiss and Separation Point-type granitoids.

Tectonic model

Many aspects of the Mesozoic geology of New Zealand are poorly understood, including the position and polarity of the subduction zone that fed the Median Tectonic Zone arc, and the location of the corresponding fore-arc, back-arc and accretionary complexes. Throughout the Mesozoic, there was active subduction along the Pacific margin of the Eastern Province (Bradshaw 1989). A minimum estimate for the distance between the Median Tectonic Zone arc and the Pacific margin trench is 600 km (Muir *et al.* 1995, fig. 11). Clearly, the Median Tectonic Zone arc cannot be related directly to subduction at this margin. One possible position for the Median Tectonic Zone trench is the Livingstone Fault (Fig. 3).

Muir *et al.* (1995, fig. 11) presented schematic sections across the Pacific margin of Gondwana in the New Zealand region in the Mid-Jurassic and Early Cretaceous. During the Mesozoic until Early Cretaceous times (*c.* 130 Ma), partial melting of mantle wedge material gave rise to hydrous calc-alkaline magmas represented by the Mid-Jurassic–Early Cretaceous plutonic rocks (Darran Suite). It is possible that the arc may have become thickened by magma loading (cf. Oliver 1990; Brown 1996) and the addition of mafic underplate (cf. Atherton & Petford 1993; Petford & Atherton 1996). Melting of basal arc material at depths of >40 km (required for garnet stability) then gave rise to Na-rich magmas represented by the Western Fiordland Orthogneiss and the Separation Point Suite. P–T conditions of 10–12 kbar and 700–800°C have been recorded from the Western Fiordland Orthogneiss (Gibson & Ireland 1995 and references therein), indicating a crustal thickness in excess of 35–40 km beneath Fiordland during the Early Cretaceous.

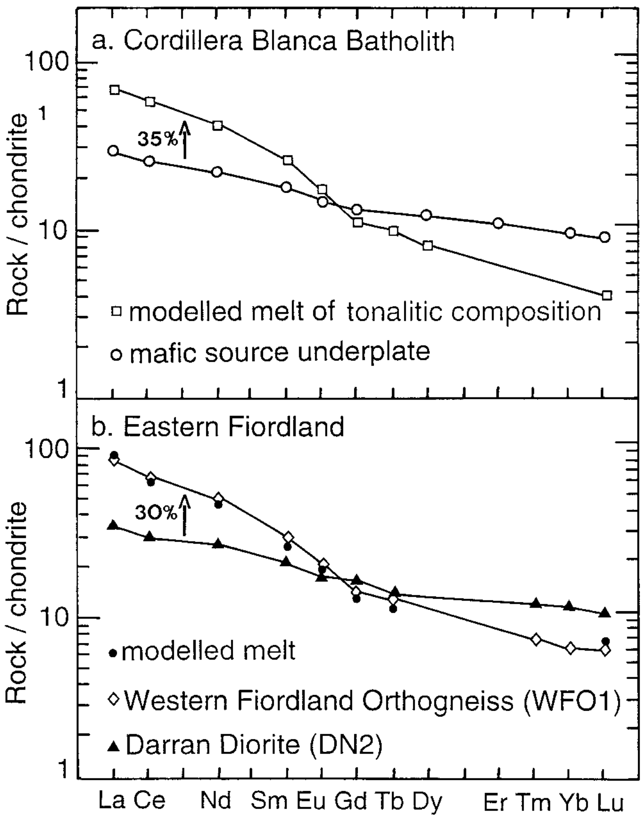

Fig. 14. (**a**) Batch melting model (35% melt) of garnet amphibolite basaltic underplate to produce a typical Na-rich, heavy REE depleted tonalite from the Cordillera Blanca Batholith, Peru (after Petford & Atherton 1996, fig. 18). Residual mineralogy: 40% amphibole, 36% clinopyroxene, 8% garnet, 8% alkali feldspar, 6% quartz, 1% ilmenite, 0.5% magnetite. (**b**) Chondrite normalized REE patterns for a sample of Darran Diorite and the Western Fiordland Orthogneiss. Darran Diorite (DN2) has the appropriate chemical composition to represent the mafic underplate beneath the Median Tectonic Zone arc.

The modelled melt, which is similar to WFO1, was generated from DN2 using a modal batch melting model (30% melting) leaving a residue of 11% garnet, 15% amphibole, 63% clinopyroxene, 5% plagioclase feldspar, 2% alkali feldspar, 3% ilmenite and 1% magnetite. References to mineral-melt distribution coefficients can be found in Petford & Atherton (1996).

The occurrence of small volumes of Na-rich granitoid during the early history of the arc (e.g. Mary Island Granodiorite and felsic sheets within the Darran Complex) suggests that these rocks are a minor, but important part of the spectrum of subuction-related magmatism. However, Muir *et al.* (1995) suggested that the sudden appearance of large volumes of Na-rich granitoid during the Early Cretaceous was triggered tectonically, perhaps by closure of a back-arc basin, and thrusting of the Median Tectonic Zone arc beneath the Western Province. Recent deep crustal seismic studies which show the Median Tectonic Zone extending for a considerable distance beneath the Western Province (F. Davey pers. comm), lend support to this model. Inherited zircons with Darran Suite ages (160–130 Ma) in the Separation Point type plutons (Fig. 6) and inherited Nd isotope signatures in the Western Fiordland Orthogneiss (Fig. 11) also point to the presence of Median Tectonic Zone type basement at depth. There appear to be no Separation Point type plutons to the north of Lake Te Anau (Fig. 3), where the Darran Complex is still at a high crustal level. Further south, the lack of voluminous 137 Ma plutonic rocks (Darran Complex equivalents) and the occurrence of Separation Point type plutons may be a consequence of underthrusting of the Darran Complex to deeper crustal levels. Median Tectonic Zone rocks were strongly deformed prior to the emplacement of the Separation Point Suite rocks (Kimbrough *et al.* 1993) and it is possible that this deformation represents an attempt to subduct the Median Tectonic Zone arc beneath the Western Province.

Conclusions

The Eastern Fiordland Igneous Belt, SW New Zealand forms part of the Median Tectonic Zone between the Western Province of Gondwana continental affinity and the Eastern Province of accreted sedimentary and volcanic rocks. Palaeozoic granitoids and dioritic rocks occur to the west of, and within the Median Tectonic Zone, but the zone is dominated by Jurassic–Early Cretaceous plutonic rocks. These igneous rocks are products of a Mesozoic magmatic arc system that developed along the palaeo-Pacific margin of Gondwana.

The following tectonic scenario is proposed.

(1) Pre-168 Ma. Development of a subduction system to the east of the Western Province. The location of the trench associated with the Mesozoic arc is uncertain, but may be located in the vicinity of the Livingstone Fault (Fig. 3). There is some evidence to suggest that the Median Tectonic Zone arc developed on, or close to the eastern edge of the Western Province (Kimbrough *et al.* 1994; Muir *et al.* 1997), but Muir *et al.* (1995) have suggested that the Western Province and Median Tectonic Zone arc were separated by a back-arc basin.

(2) 168–137 Ma. Generation and emplacement of large volumes of calc-alkaline igneous rocks, most likely derived from melting in the mantle wedge, induced by hydrous fluids released from the subducted slab. Primitive Sr and Nd isotopic ratios and the absence of inherited zircon indicates that these magmas experienced little or no crustal interaction.

(3) Approximately 130 Ma. Major period of transcurrent faulting, possibly related to oblique collision and docking of the Western and Eastern Provinces (Kimbrough *et al.* 1993, 1994). Subduction/thrusting of the Median Tectonic Zone arc beneath the Western Province may have occurred at the same time.

(4) 126–110 Ma. Partial melting of underplated mafic arc material at depths of >40 km results in the generation of Na-rich granitoids leaving residues of garnet+clinopyroxene+amphibole. The Western Fiordland Orthogneiss was emplaced at deeper levels in the crust than the more felsic Separation Point granitoids.

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