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Ion microprobe dating of Paleozoic granitoids: Devonian magmatism in New Zealand and correlations with Australia and Antarctica

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Abstract

Precise ion microprobe U–Pb zircon ages have been obtained from a representative set of Paleozoic igneous rocks from the Western Province of the South Island, New Zealand. Granitoid rocks forming the Karamea Batholith and related plutons in the Buller terrane all yield crystallisation ages that are indistinguishable within a ± 5 Ma uncertainty at 375 Ma (Middle–Late Devonian). Previous workers have suggested that the batholith was emplaced over a long time interval and comprises rocks ranging in age from 370 to 310 Ma, with the bulk in the Early Carboniferous. Granitoid rocks with Carboniferous ages (~ 330 Ma) do occur further to the west and younger Cretaceous granitoids occur within the Karamea Batholith, but these are not volumetrically significant. A sample from the ultramafic–mafic Riwaka Complex in the adjacent Takaka terrane gave a crystallisation age of 376.9 ± 5.6 Ma (2σ), indicating emplacement coeval with the Karamea Batholith.

The Paleozoic granitoids contain a large amount of inherited zircon, with distinct 390-, 500–600- and 1000-Ma components. The 390-Ma age corresponds to widespread plutonism in the Lachlan Fold Belt in SE Australia. The 500–600-Ma (Ross–Delamerian age) and 1000-Ma (Grenville age) age components have also been observed in granitoids from the Lachlan Fold Belt and in Ordovician metasedimentary rocks from SE Australia and New Zealand. The inherited zircons in the Karamea Batholith could be derived from continental basement at depth, from the incorporation of upper-crustal material into the granitoid magmas, or both.

The Devonian granitoids in New Zealand can now be correlated with rocks of similar age in northern Victoria Land (Admiralty Intrusives) and Marie Byrd Land (Ford Granodiorite) in West Antarctica, in NE Tasmania, and in the central part of the Lachlan Fold Belt in SE Australia. On a reconstruction of the SW Pacific margin of Gondwana, prior to later break-up, it is possible to trace out a semi-continuous magmatic belt in excess of 2000 km in length.

1. Introduction

The Western Province of New Zealand (Fig. 1) represents a disrupted fragment of the Gondwana supercontinent; geological terranes formerly contiguous with those in New Zealand are found in Antarctica, SE Australia and Tasmania (e.g., Cooper and Grindley, 1982; Grindley and Davey, 1982; J.D. Bradshaw et al., 1985; Stump et al., 1986; Cooper and Tulloch, 1992; Gibson, 1992). There are close similarities in the nature and timing of Paleozoic events in all of these regions. However, attempts to match the main features of the New Zealand Paleozoic, in particular the voluminous

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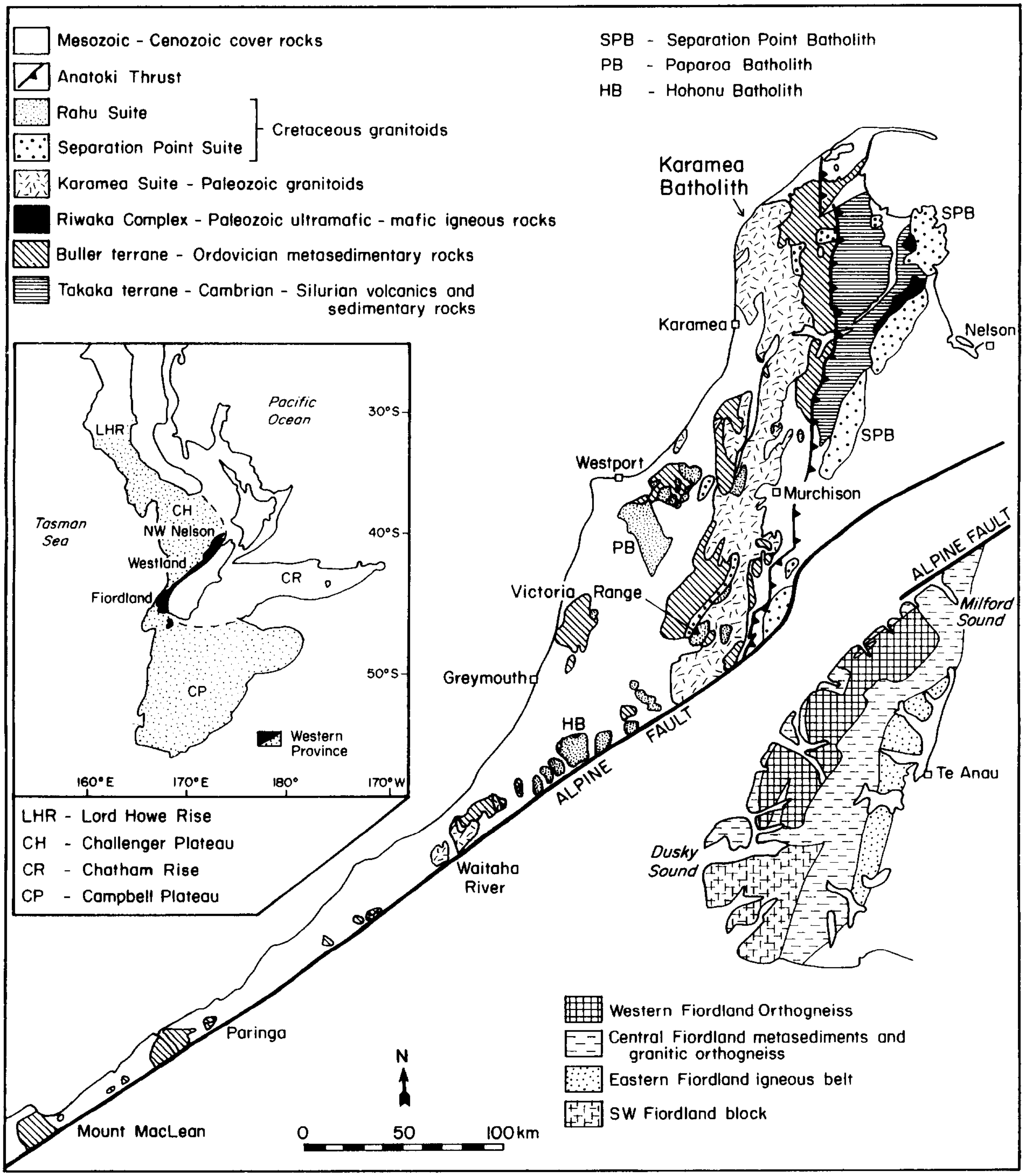


Fig. 1. Geological map of NW Nelson–Westland and Fiordland showing the distribution of granitoid rocks in New Zealand and the location of the Karamea Batholith (after Tulloch, 1988; Cooper, 1989; J.Y. Bradshaw, 1990). *Inset* shows the offshore extent of the Western Province.

granitoid intrusions, with equivalents in SE Australia or Antarctica have been hampered by the lack of precise age control.

Granitoid rocks and related diorites (nomenclature follows Le Maitre, 1989) make up ~50% of the total area of the Western Province of the South Island (Fig. 1). Those in NW Nelson–Westland have been reviewed briefly by Tulloch (1983), who has identified the major batholiths and compositional suites (Tulloch, 1988). Paleozoic granitoids make up the bulk of the Karamea Batholith, the largest contiguous tract of granitoid rocks in the Western Province, but also occur as isolated plutons further to the SW of the batholith.

In a recent review of the early Paleozoic terranes of the Western Province, Cooper and Tulloch (1992) concluded that the age and chemistry of the predominantly S-type Paleozoic granites make it unlikely that they are the direct equivalents of the Paleozoic granitoid terranes in SE Australia and Antarctica. These authors concluded that the Paleozoic granitoids in NW Nelson–Westland range in age from 370–310 Ma (Late Devonian–Carboniferous), with the bulk in the Early Carboniferous. Most of these granitoids have been assigned to the S-type Karamea Suite (Tulloch, 1983, 1988), but rare A- and I-types have also been proposed (Cooper and Tulloch, 1992).

In this paper we present new U–Pb zircon ages for the main intrusive phases forming the Karamea Batholith in the Buller terrane (Fig. 1). We have also dated zircons from two isolated plutons of suspected Paleozoic age lying to the west of the main batholith, and a sample of diorite from the Riwaka Igneous Complex to the east in the Takaka terrane (Fig. 1). The new dates establish, for the first time, a precise chronology for Paleozoic magmatism in the Western Province and enable more rigorous correlations to be made with other Gondwana elements.

2. The Karamea Batholith

The Karamea Batholith extends south from Kahurangi Point for nearly 200 km to the Alpine Fault east of Greymouth, and has an average width of 20 km (Fig. 1 and Fig. 2). It intrudes Ordovician metasedimentary rocks of the Greenland Group to the west and the Golden Bay Group to the east (Cooper, 1989). Large, stoped blocks of country rock are common near the steeply dipping margins of the batholith and biotite, andalusite, sillimanite and cordierite have been observed in the contact aureole, which is generally < 1 km wide (Roder and Suggate, 1990). The batholith is associated with weak tin–tungsten mineralisation (Braithwaite and Pirajno, 1993) and is cut by several Cretaceous (Separation Point Suite) Mo-bearing granodiorite porphyry plutons (Tulloch and Rabone, 1993). Granitoid plutons in the Victoria Range (Rahu Suite), SW of Murchison (Fig. 1), are also thought to be Cretaceous in age (Tulloch, 1983).

Most of the Karamea Batholith is poorly exposed and relatively inaccessible, and thus has been little studied. Apart from two 1:50,000 scale geological maps of the Buller Gorge area (Nathan, 1978; Roder and Suggate, 1990), there is little published information on the age, petrography or distribution of different rock types. The 1:250,000 scale regional geological maps of New Zealand (Grindley, 1961; Bowen, 1964) show only the outline of the batholith. Following a recent field mapping and geochemical sampling programme, covering the batholith north of the Victoria Range, we have identified five major intrusive phases (Fig. 2). Their distribution and petrography are outlined below. As far as possible, the original nomenclature proposed by Grindley (1961), Nathan (1978), Laird (1988) and Roder and Suggate (1990) has been retained. The order of emplacement of the different intrusive phases is only partly known from cross-cutting field relationships.

2.1. Karamea Granite

The dominant lithology, forming the northern two-thirds of the batholith and a smaller body south of Murchison (Fig. 2), is a coarse-grained, porphyritic biotite granite (syenogranite–monzogranite), termed the Karamea Granite after Grindley (1961). It is a distinctive unit, containing large, pink, prismatic alkali-feldspar megacrysts (30–40 mm in length) in a groundmass of quartz, oligoclase, microcline, biotite and muscovite with accessory apatite, zircon and iron oxide. The alkali feldspar megacrysts — orthoclase, microcline, or microperthite — are generally kaolinised and bright pink in colour, although white alkali feldspar or plagioclase phenocrysts are also common locally (e.g., Karamea Gorge and Wangapeka Track areas). The megacrysts often show a strong preferred orientation, probably as a result of magmatic flow. The Kar-

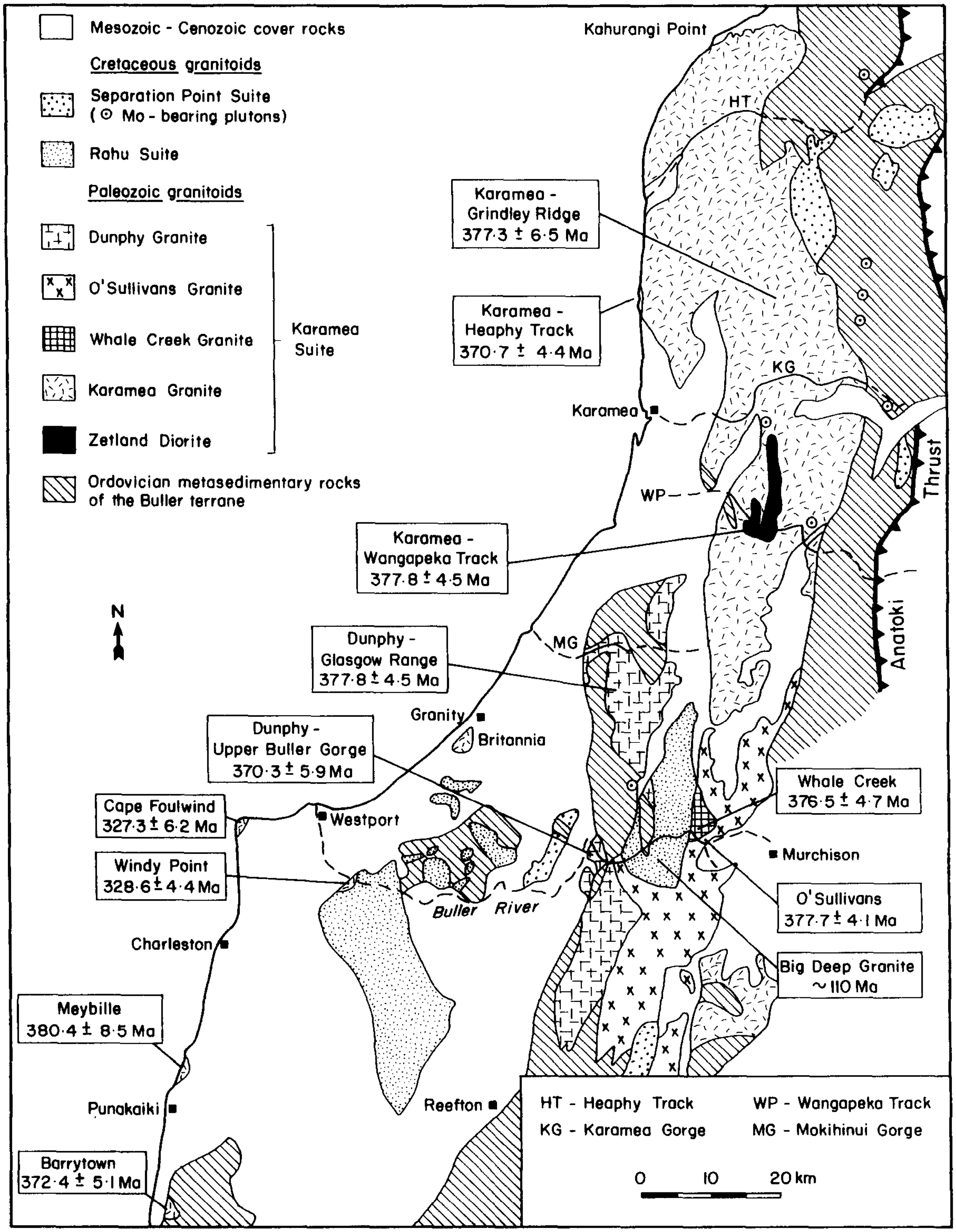


Fig. 2. Geological map of the northern part of the Karamea Batholith.

amea Granite exhibits comagmatic relationships with dioritic rocks in the Wangapeka Track area, but its position in the overall order of emplacement is not known. The Brittania Granite pluton, south of Granity (Fig. 2), has been correlated with the Karamea Granite by Nathan (1978) and Kutsukake (1988).

2.2. *Whale Creek Granite*

The Whale Creek Granite (Roder and Suggate, 1990) forms a small pluton (~ 20 km^2) at the eastern end of the Upper Buller Gorge (Fig. 2). It is a coarse-grained, dark grey, porphyritic biotite-granite, with large, white, tabular alkali feldspar megacrysts (typically 30–40 mm in length). In places, a well-developed foliation is defined by aligned megacrysts and flakes of dark-brown to reddish-brown biotite. Roder and Suggate (1990) have correlated the Whale Creek Granite with the Karamea Granite further north on the basis of petrographic similarities. However, the Whale Creek Granite is considerably more mafic and can also be distinguished from the Karamea Granite on geochemical grounds, having a lower SiO_2 and higher total Fe_2O_3 content (R.J. Muir, unpublished data, 1994). The Whale Creek Granite is cut by sheets of O'Sullivans Granite in the Upper Buller Gorge (Roder and Suggate, 1990).

2.3. *O'Sullivans Granite*

The O'Sullivans Granite (Roder and Suggate, 1990) is the dominant phase along the eastern side of the batholith in the vicinity of the Upper Buller Gorge (Fig. 2), extending south to the Victoria Range and north to the Matiri Range. It is a fine- to medium-grained, pale-grey, muscovite–biotite granite with sparse phenocrysts of plagioclase or alkali feldspar up to 15 mm in length. Irregular sheets and patches of finer-grained aplitic material within the granite have been termed the Fern Flat Granite Aplite by Roder and Suggate (1990).

2.4. *Dunphy Granite*

The Dunphy Granite (Nathan, 1978) occurs in the western part of the Upper Buller Gorge (Fig. 2), extending south along the Brunner Range, and north to the Glasgow Range and the Mohikinui Gorge. It is a coarse-grained leucocratic muscovite–biotite granite with abundant phenocrysts of white alkali feldspar (typically 25×10 mm). The phenocrysts are commonly aligned, probably as a result of magmatic flow. The Dunphy Granite has not been observed in contact with any other intrusive phases.

2.5. *Zetland diorite*

Fine- to coarse-grained biotite and biotite–hornblende quartz diorites crop out extensively around Mount Zetland and along the Herbert Range north of the Wangapeka Track (Fig. 2). Irregular contact relationships with the enclosing Karamea Granite suggest that mingling of the felsic and mafic magmas occurred. Minor cross-cutting intrusions of dioritic material occur in the granites in the Upper Buller Gorge, but are thought to be Cretaceous in age (Roder and Suggate, 1990).

3. Isolated plutons

3.1. *Meybille Granite*

The Meybille Granite (Laird, 1988) is exposed along a 3-km coastal strip centred on Meybille Bay, north of Punakaiki (Fig. 2), and is similar to the Karamea Granite further north. It is a light-grey, porphyritic, medium-grained biotite–muscovite granite containing large white or pink tabular megacrysts of alkali feldspar (30×15 mm). The megacrysts are commonly aligned. No intrusive relationships are seen.

3.2. *Barrytown Granite*

The Barrytown Granite (Tulloch, 1983) crops out over an area of ~ 4 km^2 near the settlement of Barrytown, south of Punakaiki (Fig. 2). It consists of light-grey, equigranular medium-grained biotite granodiorite and two-mica granite. The pluton was emplaced into the Greenland Group at a relatively high level and the central roof region is associated with tungsten mineralisation.

4. Riwaka Complex

The Riwaka Complex (Gill and Johnston, 1970), which lies to the east of the Karamea Batholith

(Fig. 1), intrudes Lower Paleozoic metasedimentary rocks of the Takaka terrane. The complex occupies an area of ~100 km^2 and consists of elongate sheets of ultramafic–gabbroic rocks and pyroxene-bearing diorite–monzodiorite. Grindley (1980) suggested that the complex was intruded as a layered lopolith along the axis of a major synclinal structure. The southern part of the intrusion is a narrow (1 km) concordant sheet of olivine pyroxenite and peridotite that has been interpreted as a feeder dike. The rocks become more felsic and the complex widens to the NE. The most felsic rock, known as the Brooklyn Diorite (Grindley, 1980), represents the upper part of the layered intrusion. The rock contains clinopyroxene, brown hornblende, plagioclase (andesine), interstitial biotite, orthopyroxene, quartz, alkali feldspar, apatite, iron oxides and titanite. Igneous textures are dominant, but recrystallisation to amphibolite is common.

5. Previous age determinations

Aronson (1968) obtained Rb–Sr whole-rock–mineral ages of ~350 Ma from the megacrystic granite at Oparara Quarry near Karamea in NW Nelson, and from isolated plutons near Mount MacLean and the Waitaha River in south Westland (Fig. 1). Roder and Suggate (1990) attempted to date a number of intrusive phases from the Karamea Batholith in the Upper Buller Gorge using the Rb–Sr and K–Ar methods. However, they were unable to obtain statistically acceptable isochron ages and their best estimates are in conflict with cross-cutting relationships in the field. The O'Sullivans Granite, the Fern Flat Granite and the Dunphy Granite all gave Rb–Sr whole-rock ages of ~300 ± 50 Ma (with mean square of weighted deviates, MSWD > 25). The Whale Creek Granite, which is cut by the O'Sullivans Granite, gave a considerably younger age of 186 ± 39 Ma (MSWD = 58). Clearly the Rb–Sr isotopic systems have not remained closed and probably reflect reheating and/or cooling associated with Cretaceous magmatism and tectonics. In support of this interpretation, K–Ar ages are all < 240 Ma, with many clustering around 120–100 Ma (Roder and Suggate, 1990).

Small plutons and stocks (generally < 1 km^2) belonging to the Early Cretaceous Separation Point Suite (Fig. 2) occur within and adjacent to the main Karamea Batholith (Tulloch and Rabone, 1993). Several of these bodies have yielded K–Ar ages in the range 116–104 Ma (Eggers and Adams, 1979). In the Victoria Range (Fig. 1), plutons assigned to the Rahu Suite have also yielded Cretaceous ages (Aronson, 1968; Tulloch, 1983). From the Upper Buller Gorge, Aronson (1968) obtained a Rb–Sr isochron age of ~110 Ma on a granitoid body subsequently mapped as Whale Creek Granite (Paleozoic) by Roder and Suggate (1990). Pickett and Wasserburg (1989) analysed a sample from this body and concluded that it would have had an unreasonably low $^{87}Sr/^{86}Sr$ initial ratio (< 0.702) if it were a Paleozoic rock. A Cretaceous age therefore seems likely. The extent of this body, here termed the Big Deep Granite, is shown in Fig. 2. The rock is a medium-grained, red or pink equigranular biotite–muscovite granite and shows chemical affinities with the Rahu Suite.

In a pilot SHRIMP (sensitive high-resolution ion microprobe) study of New Zealand granitoids, Muir et al. (1994) obtained the first precise Paleozoic crystallisation age from the Karamea Batholith. A sample of O'Sullivans Granite from the Upper Buller Gorge (Fig. 2) gave a U–Pb zircon age of 377.7 ± 4.1 Ma (2σ) (Middle–Late Devonian). Further west, granites at Cape Foulwind and Windy Point (Fig. 2) gave younger ages of 327.3 ± 6.2 and 328.6 ± 4.4 Ma, respectively. Carboniferous magmatism, represented by the Cape Foulwind and Windy Point Granites, has also been recognised (Tulloch et al., 1991) from offshore western New Zealand. Tulloch et al. (1991) reported a conventional U–Pb zircon age of 335 ± 7 Ma from a sample of granite dredged from the Challenger Plateau (Fig. 1), ~180 km SW of Cape Foulwind.

Harrison and McDougall (1980) obtained K–Ar hornblende ages of ~370 Ma from the Rameka Gabbro within the Riwaka Complex, that suggest emplacement coeval with the Karamea Batholith. These ages are probably more reliable than the Rb–Sr mineral ages of ~290 Ma obtained from boulders of float by Aronson (1968).

The more detailed analysis of the Karamea Batholith and related plutons undertaken in this study permits: (1) determination of the crystallisation ages of the main intrusive phases; and (2) examination of the patterns of crustal inheritance.

6. Experimental techniques

Zircons were separated from the rock samples using standard crushing and heavy-liquid techniques, mounted in epoxy, and polished to reveal their mid-sections. Using SHRIMP I, U–Pb isotopes were analysed in situ on small areas within single zircon crystals. Standard operating techniques for the SHRIMP have been described by I.S. Williams and Claesson (1987). The advantage of this method over conventional U–Pb zircon dating is that petrographically selected areas of zircon can be analysed enabling the determination of the crystallisation age of a rock, as well as identification and dating of inherited zircon crystals, and older cores within magmatic crystals (e.g., I.S. Williams, 1992; Muir et al., 1994).

The granitoids from the Western Province contain zircon with a number of different morphologies (Fig. 3), including euhedral and rounded crystals and fragments that range from colourless to dark brown. Given the large amount of inheritance observed during our earlier SHRIMP study (Muir et al., 1994), the rationale behind crystal selection was to first find the magmatic age. Structurally homogeneous, clear euhedral crystals with simple terminations and length/breadth ratios of 4:1 tend to give good magmatic crystallisation ages as do more tabular crystals. Inherited ages come from cores and individual crystals that are often darker in colour and have more complex terminations. Euhedral crystals with length/breadth ratios of 2:1 commonly give ages around 550–600 Ma. However, such simple subdivisions do not always result in the selection of a magmatic crystal. Some granites, e.g., Dunphy, contain clear euhedral crystals that range in age from the magmatic age through to 1000 Ma. Moreover, the oldest crystal analysed in this study (~ 1700 Ma) is in fact euhedral.

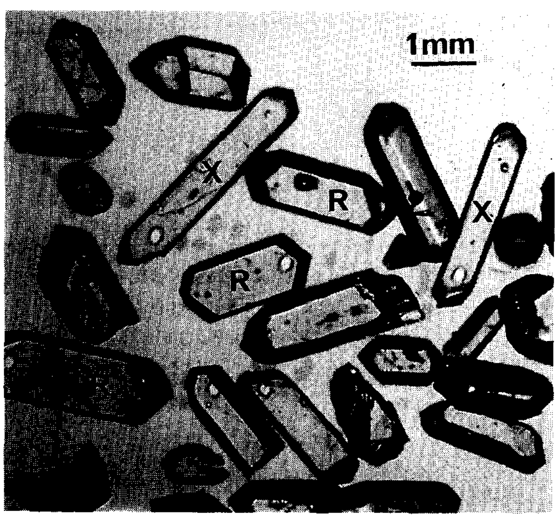
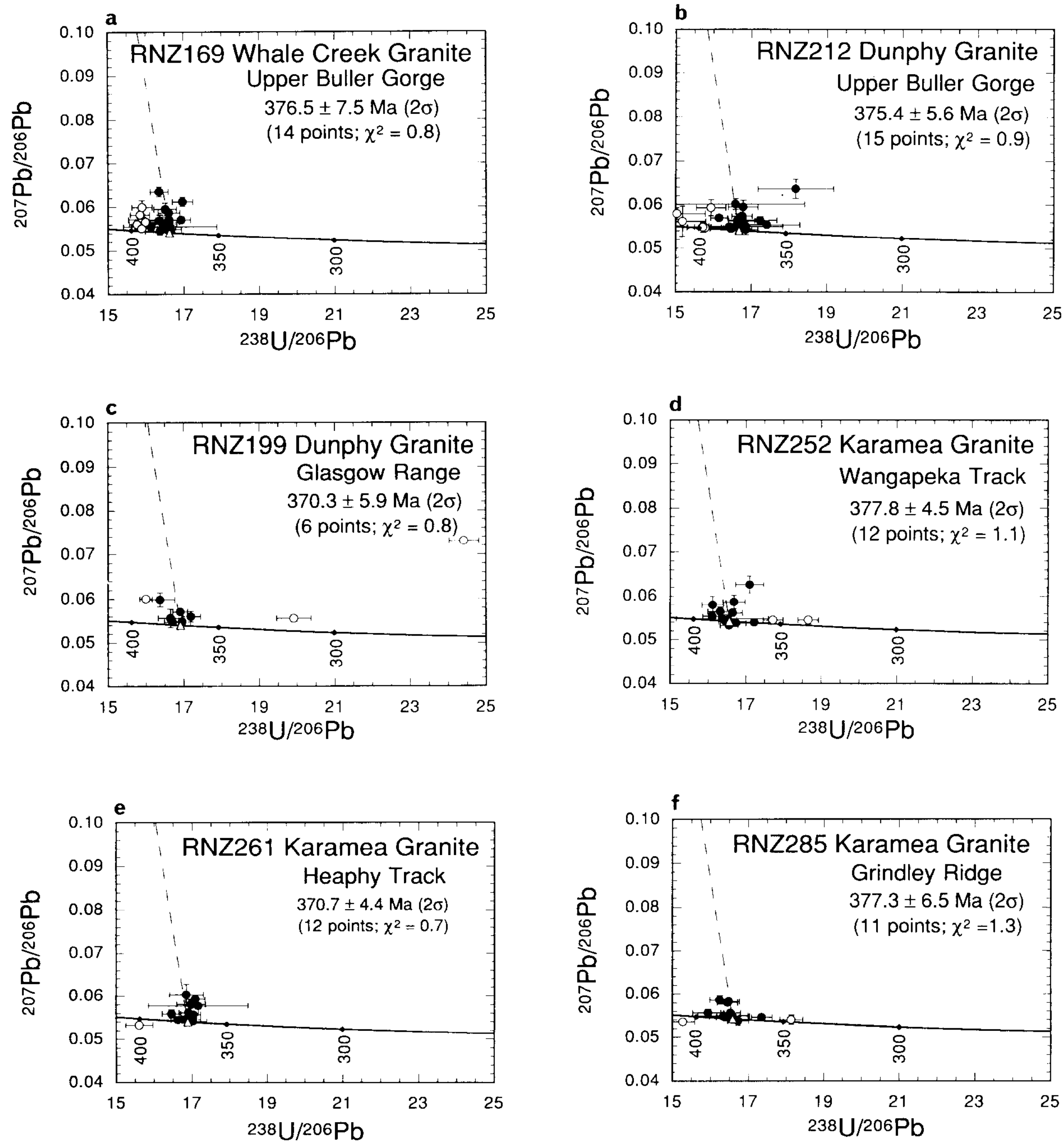


Fig. 3. Zircon crystals from the Dunphy Granite. Elongate crystals (×) tend to give the magmatic crystallisation age (375 Ma). Stubby crystals (*R*) are inherited and give ages around 500–600 Ma.

A 3-nA mass-filtered O_2^- primary beam was focussed to sputter a 30-μm-diameter area with positive ions extracted. The magnet was cyclically peak-stepped through a series of mass stations ranging from mass 196 ($^{90}Zr_2^{16}O^+$) to mass 254($^{238}U^{16}O^+$), and including the Pb isotopic mass stations at masses 204, 206, 207 and 208, as well as ^{238}U and $^{232}Th^{16}O$ peak at mass 248. Pb isotopic ratios were taken as measured; no corrections have been applied for isotopic mass fractionation or Pb hydride interferences; any corrections would have a negligible effect on this data set because the ages determined are primarily from the $^{238}U/^{206}Pb$ ratios.

While ^{204}Pb was monitored and can be used to correct the Pb isotopic composition for non-radiogenic lead, the errors propagated from the $^{204}Pb/^{206}Pb$ measurement to the corrected $^{207}Pb/^{206}Pb$ ratio can be substantial, particularly for young zircons with low ^{207}Pb concentrations. The ^{208}Pb signal can also be used to correct for common Pb based on the expected radiogenic Pb from the age and the measured Th/U ratio. However, this method is rather model dependent and can be problematic, especially for high-Th/U zircons. For this study we have used the ^{207}Pb correction which uses the $^{207}Pb/^{206}Pb$ ratio to monitor common Pb, and assumes that the analyses are simple mixtures of radiogenic and common Pb and the radiogenic $^{238}U/^{206}Pb$ is the extrapolation from the common Pb through the measured point to concordia. In this case the independent age estimate from the $^{207}Pb/^{206}Pb$ ratio is lost and the ages are solely determined from the $^{238}U/^{206}Pb$ ratio. This method is very reliable for analyses that are close to concordia, which indicates low inherent common Pb levels. The $^{238}U/^{206}Pb$ ages are pooled and are examined using standard statistical procedures. The χ^2 statistic is used as an overall indicator of the reliability of the mean. For a large number of estimates, the χ^2 should be close to unity. For smaller samples, the statistic should not exceed given values that can be found in standard statistical tables. If the χ^2 does exceed this value then outliers are searched for and rejected if such rejection is valid. While the rejection criteria are statistical, it should be noted that outliers are generally a function of the samples and not of the technique. The most common cause of outliers is inheritance but Pb loss is also apparent in some of the samples.

The $^{204}Pb/^{206}Pb$ correction method has been applied to inherited crystals and cores which are substantially older than the magmatic age; for zircons younger than 750 Ma, the $^{238}U/^{206}Pb$ age has been used. For older



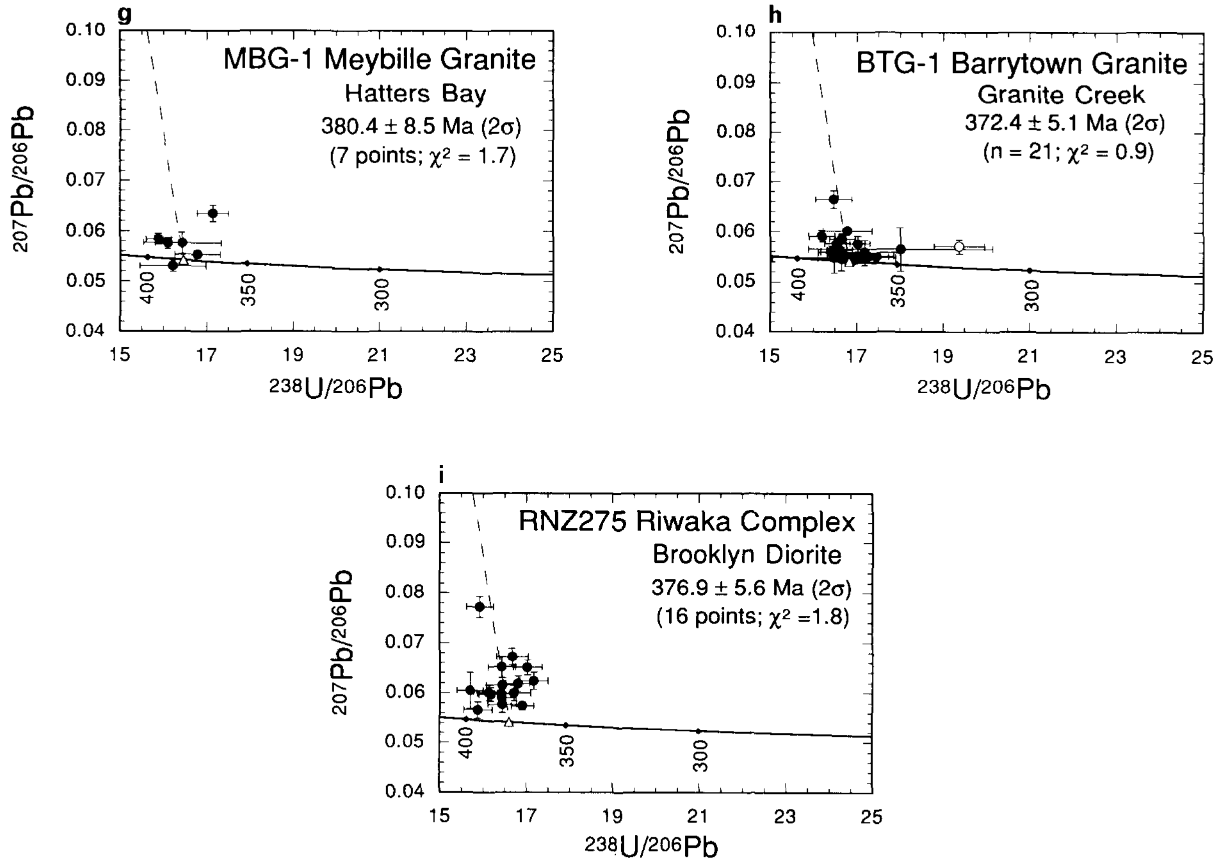


Fig. 4. Tera–Wasserburg U–Pb zircon concordia diagrams. Concordia are labelled in Ma and the *dashed line* represents the mixing line between Broken Hill common Pb and the inferred concordant age (*open triangle*) of the analyses in *solid circles*. *Open circles* are outliers and interpreted as either xenocrysts to the left of the mixing line or Pb loss to the right of the mixing line.

zircons the $^{207}Pb/^{206}Pb$ ratio provides a more accurate indicator of age. For both common Pb correction methods, the Pb is assumed to be of Broken Hill Pb isotopic composition. The use of another Pb isotopic composition would have a negligible effect on the magmatic ages since most of the data points are close to concordia.

The Pb/U in the samples was calibrated using an empirical quadratic relationship between $^{206}Pb^+/U^+$ and UO^+/U^+ and normalised to the $^{206}Pb^{\star}/U$ measured for the standard. In this study we have used two standards since the first standard SL13 is gradually being consumed and it would be advantageous to have more standards available. SL13 is a homogeneous zircon crystal with around 240 ppm U and an age of 572 Ma. The second standard, AS3, is derived from a gabbroic anorthosite from the Duluth Complex, Minnesota, U.S.A. Zircons from this rock have been shown to be consistently concordant with an age of 1099 ± 1 Ma (Paces and Miller, 1989) and are up to 500 μm in size allowing for a single crystal to be used as a standard in a given epoxy mount. However, since the zircons crystallised individually, each crystal can have a different U concentration. The U and Th concentrations in the AS3 crystals, as well as unknowns, are determined relative to that measured from SL13, which is homogeneous to within 20%. The U concentrations in AS3 crystals range from ~ 100 to > 1000 ppm, but the range is much smaller in an individual crystal. AS3 chips with concentrations around 200–400 ppm were used for Pb/U normalisation. A detailed description of the comparison between AS3 and SL13 will not be given here, suffice to say that the internal comparisons

Table 1
U–Pb analyses of zircons

Sample	U (ppm)	Th (ppm)	Th/U	$^{204}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{238}U/^{206}Pb$	Age (Ma)
RNZ169 *Whale Creek Granite*:							
1.1	294	96	0.33	0.00081 ± 0.00024	0.0569 ± 0.0013	16.36 ± 0.35	381.4 ± 8.0
2.1	648	97	0.15	0.00034 ± 0.00010	0.0549 ± 0.0014	16.64 ± 0.30	376.0 ± 6.6
3.1	473	111	0.23	0.00054 ± 0.00014	0.0571 ± 0.0008	16.94 ± 0.24	368.7 ± 5.1
4.1	485	53	0.11	0.00056 ± 0.00013	0.0557 ± 0.0006	15.87 ± 0.26	393.3 ± 6.2
5.1	336	38	0.11	0.00060 ± 0.00016	0.0599 ± 0.0016	15.90 ± 0.27	390.9 ± 6.4
6.1	584	118	0.20	0.00032 ± 0.00012	0.0566 ± 0.0007	16.58 ± 0.31	376.6 ± 6.8
7.1	563	98	0.17	0.00017 ± 0.00007	0.0571 ± 0.0004	15.90 ± 0.36	392.0 ± 8.7
8.1	741	131	0.18	0.00016 ± 0.00009	0.0552 ± 0.0005	16.55 ± 0.24	377.9 ± 5.4
8.2	185	82	0.44	0.00120 ± 0.00038	0.0634 ± 0.0011	16.35 ± 0.24	379.0 ± 5.4
9.1	345	51	0.15	0.00064 ± 0.00018	0.0570 ± 0.0008	16.61 ± 0.25	375.7 ± 5.6
10.1	404	72	0.18	0.00038 ± 0.00015	0.0582 ± 0.0007	15.84 ± 0.25	392.9 ± 6.1
11.1	235	44	0.19	0.00108 ± 0.00032	0.0612 ± 0.0010	16.97 ± 0.27	366.4 ± 5.7
12.1	586	115	0.20	0.00028 ± 0.00012	0.0556 ± 0.0009	16.14 ± 0.24	387.0 ± 5.5
13.1	840	109	0.13	0.00016 ± 0.00006	0.0561 ± 0.0005	15.77 ± 0.21	395.5 ± 5.1
14.1	370	39	0.11	0.00060 ± 0.00016	0.0562 ± 0.0015	16.46 ± 0.36	379.4 ± 8.1
15.1	1,148	145	0.13	0.00009 ± 0.00003	0.0555 ± 0.0001	16.64 ± 1.23	375.7 ± 27.1
16.1	201	106	0.53	0.00059 ± 0.00025	0.0595 ± 0.0014	16.51 ± 0.30	377.0 ± 6.6
17.1	533	47	0.09	0.00024 ± 0.00009	0.0566 ± 0.0006	15.99 ± 0.26	390.1 ± 6.2
18.1	284	46	0.16	0.00049 ± 0.00012	0.0606 ± 0.0027	11.17 ± 0.27	549.2 ± 12.6
18.2	639	78	0.12	0.00034 ± 0.00011	0.0545 ± 0.0009	16.37 ± 0.39	382.1 ± 8.8
20.1	355	90	0.25	0.00038 ± 0.00018	0.0587 ± 0.0010	16.61 ± 0.29	375.2 ± 6.3
21.1	2,078	311	0.15	0.00008 ± 0.00002	0.0551 ± 0.0004	15.89 ± 0.18	393.2 ± 4.3
RNZ212 *Dunphy Granite*:							
1.1	176	71	0.40	0.00095 ± 0.00033	0.0595 ± 0.0015	16.79 ± 0.39	370.9 ± 8.5
1.2	199	72	0.36	0.00019 ± 0.00020	0.0563 ± 0.0015	16.75 ± 0.35	373.1 ± 7.6
2.1	354	61	0.17	0.00075 ± 0.00015	0.0570 ± 0.0007	16.15 ± 0.22	386.1 ± 5.2
3.1	146	81	0.55	0.00129 ± 0.00028	0.0616 ± 0.0010	10.53 ± 0.19	580.0 ± 10.1
4.1	557	93	0.17	0.00009 ± 0.00004	0.0728 ± 0.0027	6.22 ± 0.14	942.9 ± 20.2
4.2	952	49	0.05	0.00010 ± 0.00003	0.0698 ± 0.0006	7.32 ± 0.09	811.8 ± 9.8
5.1	767	12	0.02	0.00010 ± 0.00003	0.0712 ± 0.0043	7.00 ± 0.10	845.8 ± 12.0
6.1	432	50	0.12	< 0.00005	0.0638 ± 0.0037	5.99 ± 0.22	985.5 ± 34.2
7.1	276	97	0.35	0.00027 ± 0.00015	0.0605 ± 0.0008	11.30 ± 0.15	542.8 ± 6.9
8.1	114	64	0.56	< 0.00006	0.0593 ± 0.0018	15.94 ± 0.39	390.2 ± 9.3
9.1	162	45	0.27	0.00048 ± 0.00015	0.0768 ± 0.0089	6.40 ± 0.22	914.8 ± 30.6
10.1	109	77	0.70	0.00016 ± 0.00012	0.0719 ± 0.0049	5.88 ± 0.27	994.7 ± 42.0
11.1	622	25	0.04	< 0.00006	0.1008 ± 0.0047	3.26 ± 0.13	1,638.3 ± 88.3
12.1	146	49	0.33	0.00168 ± 0.00058	0.0636 ± 0.0022	18.18 ± 1.00	341.8 ± 18.4
12.2	424	43	0.10	0.00019 ± 0.00008	0.0562 ± 0.0034	15.18 ± 0.61	410.4 ± 16.2
13.1	2,714	66	0.02	0.00005 ± 0.00002	0.0543 ± 0.0011	16.86 ± 0.31	371.5 ± 6.7
14.1	5,301	232	0.04	0.00006 ± 0.00002	0.0550 ± 0.0003	16.66 ± 0.20	375.5 ± 4.3
15.1	4,288	148	0.03	0.00006 ± 0.00002	0.0547 ± 0.0004	15.78 ± 0.49	395.8 ± 11.9
16.1	2,754	29	0.01	0.00005 ± 0.00002	0.0548 ± 0.0011	15.72 ± 0.22	397.3 ± 5.5
17.1	1,378	452	0.33	0.00022 ± 0.00005	0.0549 ± 0.0005	16.42 ± 0.49	380.8 ± 11.1
18.1	625	127	0.20	0.00013 ± 0.00006	0.0648 ± 0.0012	9.74 ± 0.14	622.8 ± 8.9
19.1	6,556	189	0.03	0.00006 ± 0.00002	0.0550 ± 0.0003	14.43 ± 0.89	431.6 ± 25.8
20.1	1,519	134	0.09	0.00019 ± 0.00006	0.0570 ± 0.0005	13.89 ± 0.16	446.8 ± 4.9
21.1	184	22	0.12	0.00042 ± 0.00018	0.0602 ± 0.0014	10.64 ± 0.22	575.5 ± 11.7
22.1	862	40	0.05	0.00030 ± 0.00009	0.0554 ± 0.0006	17.42 ± 0.87	359.5 ± 17.5
22.2	388	521	1.34	0.00005 ± 0.00003	0.0804 ± 0.0005	5.36 ± 0.08	1,072.8 ± 14.0
23.1	225	87	0.39	0.00012 ± 0.00029	0.0575 ± 0.0013	13.10 ± 0.26	472.4 ± 9.2
24.1	149	77	0.52	0.00107 ± 0.00034	0.0601 ± 0.0012	16.59 ± 1.83	375.0 ± 40.4
25.1	2,850	114	0.04	0.00014 ± 0.00002	0.0579 ± 0.0013	15.03 ± 0.76	413.6 ± 20.2
26.1	428	97	0.23	0.00015 ± 0.00008	0.0564 ± 0.0008	17.24 ± 0.44	362.4 ± 9.1
27.1	676	352	0.52	0.00031 ± 0.00008	0.0545 ± 0.0006	16.47 ± 0.55	379.7 ± 12.4
28.1	372	161	0.43	0.00044 ± 0.00015	0.0549 ± 0.0007	16.47 ± 0.26	379.5 ± 5.8
29.1	477	336	0.70	0.00069 ± 0.00024	0.0551 ± 0.0012	16.85 ± 0.98	371.0 ± 21.0
30.2	212	90	0.42	0.00084 ± 0.00028	0.0564 ± 0.0015	16.63 ± 0.23	375.2 ± 5.1
31.1	202	89	0.44	0.00132 ± 0.00043	0.0574 ± 0.0006	16.76 ± 0.27	372.1 ± 5.9
32.1	346	362	1.05	0.00053 ± 0.00013	0.0587 ± 0.0011	12.12 ± 0.31	508.4 ± 12.5
33.1	392	258	0.66	0.00011 ± 0.00005	0.0730 ± 0.0014	5.54 ± 0.09	1,049.3 ± 15.4

Table 1 (*continued*)

Sample	U (ppm)	Th (ppm)	Th/U	$^{204}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{238}U/^{206}Pb$	Age (Ma)
RNZ261 *Karamea Granite*:							
1.1	1,474	262	0.18	0.00006 ± 0.00005	0.0532 ± 0.0004	15.61 ± 0.36	400.7 ± 8.9
2.1	350	60	0.17	0.00045 ± 0.00021	0.0587 ± 0.0028	14.31 ± 0.43	433.2 ± 12.6
3.1	113	69	0.61	0.00069 ± 0.00025	0.0763 ± 0.0011	6.52 ± 0.20	898.8 ± 25.4
4.1	1,009	410	0.41	0.00036 ± 0.00007	0.0552 ± 0.0005	16.92 ± 0.20	369.9 ± 4.3
5.1	513	63	0.12	0.00053 ± 0.00017	0.0559 ± 0.0009	16.46 ± 0.25	379.5 ± 5.5
6.1	1,511	412	0.27	< 0.0001	0.0545 ± 0.0005	16.63 ± 0.18	376.4 ± 3.9
7.1	1,988	173	0.09	0.00018 ± 0.00005	0.0557 ± 0.0003	17.04 ± 0.19	367.0 ± 4.0
8.1	633	70	0.11	0.00026 ± 0.00008	0.0609 ± 0.0010	12.07 ± 0.39	509.4 ± 15.7
9.1	1,130	223	0.20	0.00017 ± 0.00005	0.0541 ± 0.0001	17.03 ± 0.37	368.0 ± 7.7
10.1	129	77	0.60	0.00202 ± 0.00052	0.0603 ± 0.0025	16.85 ± 0.45	369.2 ± 9.6
12.1	595	125	0.21	0.00024 ± 0.00012	0.0561 ± 0.0010	16.92 ± 0.25	369.5 ± 5.4
13.1	1,134	252	0.22	0.00031 ± 0.00008	0.0546 ± 0.0005	16.76 ± 0.23	373.4 ± 4.9
14.1	1,790	256	0.14	0.00010 ± 0.00004	0.0546 ± 0.0005	17.00 ± 0.23	368.3 ± 4.9
15.1	263	68	0.26	0.00064 ± 0.00025	0.0581 ± 0.0009	16.98 ± 0.38	367.4 ± 8.0
16.1	326	535	1.64	0.00092 ± 0.00021	0.0592 ± 0.0008	17.09 ± 0.26	364.6 ± 5.3
17.1	782	104	0.13	0.00035 ± 0.00009	0.0577 ± 0.0008	17.17 ± 1.32	363.6 ± 27.2
RNZ275 *Brooklyn Diorite*:							
1.1	111	99	0.89	0.00294 ± 0.00065	0.0616 ± 0.0014	16.44 ± 0.37	377.6 ± 8.4
2.1	103	108	1.05	0.00080 ± 0.00040	0.0597 ± 0.0014	16.18 ± 0.28	384.4 ± 6.4
3.1	86	84	0.98	0.00391 ± 0.00083	0.0651 ± 0.0021	16.43 ± 0.32	376.5 ± 7.1
3.2	117	103	0.88	0.00081 ± 0.00044	0.0618 ± 0.0015	16.80 ± 0.28	369.7 ± 6.0
4.1	144	178	1.24	0.00250 ± 0.00049	0.0600 ± 0.0016	16.13 ± 0.28	385.3 ± 6.5
5.1	98	106	1.08	0.00461 ± 0.00101	0.0651 ± 0.0015	17.02 ± 0.34	363.8 ± 7.0
6.1	167	280	1.68	0.00158 ± 0.00041	0.0577 ± 0.0017	16.44 ± 0.33	379.3 ± 7.4
7.1	107	82	0.77	0.00273 ± 0.00070	0.0599 ± 0.0015	16.71 ± 0.38	372.3 ± 8.3
8.1	93	101	1.09	0.00395 ± 0.00083	0.0672 ± 0.0017	16.67 ± 0.37	370.2 ± 7.9
9.1	73	57	0.79	0.00344 ± 0.00096	0.0604 ± 0.0036	15.70 ± 0.30	395.5 ± 7.5
10.1	79	55	0.70	0.00414 ± 0.00107	0.0772 ± 0.0021	15.93 ± 0.31	382.8 ± 7.3
11.1	104	111	1.07	0.00081 ± 0.00037	0.0598 ± 0.0018	16.41 ± 0.30	379.0 ± 6.8
12.1	106	117	1.10	0.00085 ± 0.00072	0.0590 ± 0.0020	16.43 ± 0.33	378.9 ± 7.4
12.2	114	122	1.08	0.00005 ± 0.00005	0.0566 ± 0.0016	15.88 ± 0.32	392.8 ± 7.8
13.1	97	80	0.82	0.00151 ± 0.00050	0.0624 ± 0.0017	17.17 ± 0.33	361.8 ± 6.8
14.1	514	410	0.80	0.00044 ± 0.00012	0.0574 ± 0.0008	16.91 ± 0.26	369.2 ± 5.6

between AS3 and SI13 were consistent within $\pm 2\%$ (2σ).

In assessing the uncertainty in the measurement of Pb/U, the limiting factor is the external reproducibility in measuring SL13 ($\sim \pm 2\%$). However, the reproducibility of some unknowns is better than this (sometimes $< \pm 1\%$), indicating that Pb/U is heterogeneous at that level in the SL13 standard and the reproducibility is not an analytical problem related to sputtering or secondary ion extraction. Therefore the $\pm 2\%$ external reproducibility is in fact an overestimate of the external precision for some unknowns. If this estimate were propagated through to the individual analyses it could mask the presence of two age components that are not separated by more than 2%.

The precise estimate of temporal variations (e.g., from sample heterogeneity, primary beam stability) form the basis of the determination of ratios from the given ion currents of numerator and denominator isotopes. The current philosophy has been that, in order to minimise analytical time, only five to seven cycles are used to determine the isotopic ratios. For this admittedly limited sample, a linear fit to the individual isotope count rates is used to correct for temporal variations. Ratios are formed from the count rates at the analysis midpoint and initially a counting statistic error is assumed. Counting statistics are clearly an underestimate since some fits are obviously not linear and so the counting statistic error is augmented by a factor related to the goodness of fit for each line fit.

Table 2
Ion probe zircon ages from the Karamea Batholith and related rocks

Whale Creek Granite	RNZ 169	Upper Buller Gorge	[L29/438 360]	376.5 ± 4.7 Ma
O'Sullivans Granite	RNZ 120	Upper Buller Gorge	[L29/438 359]	377.7 ± 4.1 Ma
Dunphy Granite	RNZ 199	Upper Buller Gorge	[L29/286 326]	370.3 ± 5.9 Ma
Dunphy Granite	RNZ 212	Glasgow Range	[L28/311 544]	375.4 ± 5.6 Ma
Karamea Granite	RNZ 252	Wangapeka Track	[M28/507 790]	377.7 ± 4.5 Ma
Karamea Granite	RNZ 261	Heaphy Track	[L26/347 124]	370.7 ± 4.4 Ma
Karamea Granite	RNZ 285	Grindley Ridge	[M27/518 092]	377.3 ± 6.5 Ma
Meybille Granite	MBG-1	Hatters Bay	[K30/745 037]	380.4 ± 8.5 Ma
Barrytown Granite	BTG-1	Granite Creek	[K31/727 834]	372.4 ± 5.1 Ma
Riwaka Complex	RNZ275	Brooklyn Diorite	[N26/023 120]	376.9 ± 5.6 Ma

Grid references in square brackets refer to the NZ 1:50,000 scale Topomap series

However, well-behaved zircons with high U concentrations can dominate the population and so for this study we have used a minimum error of ± 1% for the Pb/U ratio (summed in quadrature with the measurement error) in order to examine whether the scatter in a given population is consistent with the internal errors. The data selection/rejection criteria are examined in more detail below for the Whale Creek pluton, which is an example of a granite which has inheritance very close in age to the magmatic population. Finally, in determining the group mean of the analyses, the error in determining the age of the standard is summed in quadrature since we cannot obtain an age of an unknown with a precision better than that for the current standard.

7. Results

The zircon data for the granitoid samples are presented in the form of Tera–Wasserburg concordia diagrams in Fig. 4 and are compiled in Table 1. All of the granitoid samples give crystallisation ages around 375 Ma (Middle to Late Devonian). The Brooklyn Diorite from the Riwaka Complex gave an age of 376.9 ± 5.6 Ma (2σ), indistinguishable from the Paleozoic granitoids forming the Karamea Batholith and related plutons

Representative analyses for four samples are shown in Table 2 and a full list of the analyses is available from the authors on request. For cogenetic samples that have had no later lead loss, the data points should lie, within measurement error, on a mixing line between purely radiogenic lead at one extreme, and common lead at the other (Tera and Wasserburg, 1972). The intersection of the regression line with the concordia determines the age. For samples that have lost lead at any later time, the data will be displaced to the right-hand side of the mixing line and any older inheritance will be to the left of the mixing line. In either case, data points that do not lie within error of the mixing line will lead to higher values of the χ^2 statistic and may be recognised as "outliers". The procedures used for assessing outliers are described below with the example of the Whale Creek Granite.

The Whale Creek Granite (RNZ169) gives a rather tight cluster of data with corrected ages ranging from 370 to 397 Ma; only one data point at 553 Ma is not plotted on the concordia in Fig. 3a. The weighted mean of the 19 analyses on the plot is 382.7 ± 2.7 Ma with a χ^2 of 2.3 for 21 points. The critical value of χ^2 for 21 points is < 1.6, indicating there is excess scatter in this data set. The distribution of the data suggests that there is an inherited component slightly older than the magmatic age. This is clearly seen in the probability distribution diagram (Fig. 5) which shows a distinct step at

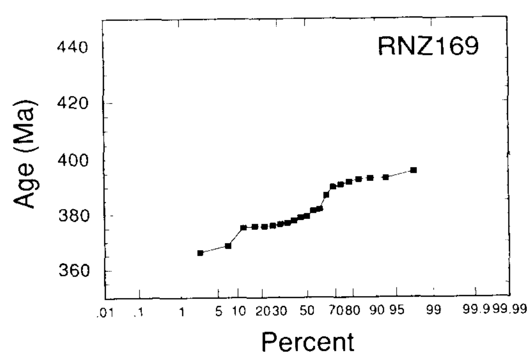


Fig. 5. Probability plot of Whale Creek analyses showing bimodal distribution with ages around 375 Ma (magmatic) and 390 Ma (inherited).

~ 385 Ma. Dividing the data set at this point initially gives two populations with ages of 376.5 ± 3.4 Ma (2σ; $\chi^2 = 1.1$; $n = 14$) and 393.7 ± 2.9 Ma (2σ; $\chi^2 = 0.5$; $n = 7$). This then is a satisfactory solution and appears consistent with the data points plotted. However, the dividing line between the two populations is rather subjective and so a spectral deconvolution program designed for such a task (Sambridge and Compston, 1994) was applied to this data set. This program models the distribution of the data with their assigned errors through a Gaussian deconvolution. The degree of convergence for a given number of components to be modelled is given by a misfit parameter which falls as the number of components is increased until no further fall in misfit is found for an increase in the number of components. The degree of misfit assuming a single population was 22 which fell to 15.2 for two components with ages of 377.1 ± 4.8 and 391.6 ± 4.4 Ma. For a three-component mixture, the degree of misfit did not fall, indicating no further subdivision is warranted. The spectral deconvolution therefore gives the same subdivision to the data as found in assessing the χ^2 statistic and the probability plot. The main point is that for a data set with a reasonable number of points (> 10), the addition or deletion of one data point does not make a significant difference to the mean of the population.

8. Crustal inheritance pattern

The main aim of the present study was to determine precise crystallisation ages for the Paleozoic granitoids in the Western Province. However, the SHRIMP data also reveal the presence of a considerable amount of old inherited zircon. Pickett and Wasserburg (1989) have previously used Sr and Nd isotopic data to demonstrate the involvement of older crustal material, most likely of Proterozoic age, in both the Paleozoic and Mesozoic granitoids. In our pilot SHRIMP study (Muir et al., 1994), we identified 500–600- and 1000-Ma and a number of older components contributing to the crustal inheritance pattern in the Western Province.

Zircon ages from the Paleozoic granitoids have been combined on a frequency histogram in Fig. 6. In addition to the magmatic age peak at 375 Ma, distinct peaks are present at 500–600 and ~ 1000 Ma and a single zircon at ~ 1700 Ma has been identified. This pattern is remarkably similar to that of inherited zircons in granites from the Lachlan Fold Belt (e.g., Chen and Williams, 1990; I.S. Williams et al., 1992) and to detrital zircons from Ordovician sediments in SE Australia and New Zealand (e.g., I.S. Williams et al., 1992; Ireland et al., 1994). Zircons from all of these rocks show age distributions with peaks at 500–600 and ~ 1000 Ma and a scatter with ages ranging up to ~ 3400 Ma.

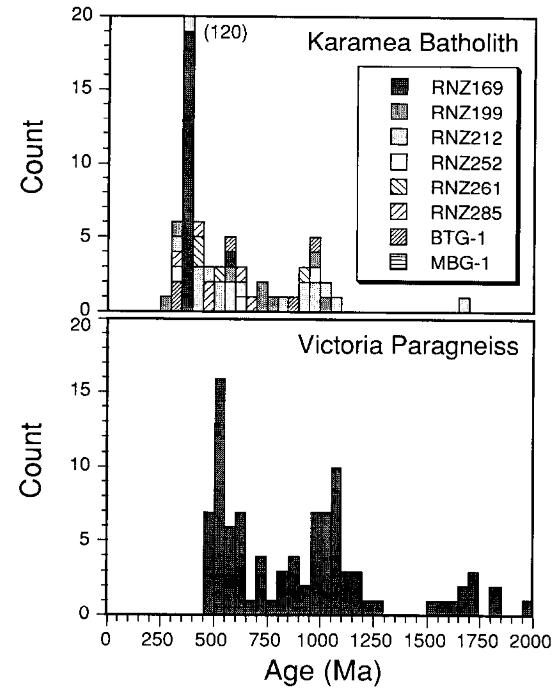


Fig. 6. Age distributions of zircons from the Paleozoic granitoids and a sample of paragneiss, thought to be metamorphosed Ordovician Greenland Group (Ireland, 1992; Ireland et al., 1994).

Precambrian rocks are not exposed in SE Australia or New Zealand, and the origin of the old zircons has been the subject of much speculation. One possibility is that igneous activity at 500–600 and ~ 1000 Ma, generated much of the continental basement beneath the Western Province (both igneous material and sedimentary detritus), and it is these rocks that were sampled during granitoid genesis. The 1000-Ma zircons are equivalent in age to the Grenville Province of North America, which has now been traced to East Antarctica (e.g., Moores, 1991; Dalziel, 1991). The period 500–600 Ma corresponds to the Ross–Delamerian Orogeny of the Transantarctic Mountains and SE Australia (Stump et al., 1986). This represents a major period of crustal growth involving subduction-related magmatism and although there appears to be no intact Precambrian basement in New Zealand (Kimbrough and Tulloch, 1989), there is certainly evidence for igneous activity around 500–600 Ma. The oldest igneous rocks in New Zealand are the Middle Cambrian (~ 530 Ma) Heath Creek beds, Devil River Volcanics and the ultramafic Cobb Igneous Complex (Cooper, 1989) in the Takaka terrane of NW Nelson (Fig. 1). Further evidence for Cambrian or perhaps older igneous activity comes from the occurrence of clasts of granitic material in the Lockett conglomerate, a thick marine fanglomerate of Late Cambrian age, that overlies the Devil River Volcanics (Grindley, 1980).

Alternatively, the inherited zircons could be derived from the country rocks enclosing the granitoids. Metasedimentary enclaves resembling the Ordovician Greenland Group greywacke have been observed throughout the Karamea Batholith. The age distribution of detrital zircons from a sample of paragneiss, thought to be metamorphosed Greenland Group (Ireland, 1992), closely matches the inheritance pattern in the granitoids (Fig. 6). Moreover, the Sr and Nd isotopic compositions of the granitoids are indicative of mixing between a mantle-derived component and large amounts of crustal material (Pickett and Wasserburg, 1989). However, it is still not clear whether the inherited zircons are derived by partial melting of the lower crust, from the melting and/or assimilation of upper-crustal material, or a combination of these factors.

In deriving the magmatic ages, it was found that the Dunphy Granites and the Whale Creek Granite, in particular, have a distinct component around 390 Ma. This is much younger than the youngest of the detrital zircons found in the Greenland Group and therefore requires another source. The grouping of these 390-Ma zircons into the magmatic population gives unsatisfactory statistics to the population (as shown in the example for the Whale Creek Granite above). These 390-Ma zircons are euhedral crystals with simple terminations and high length/breadth ratios, the same characteristics as expected for the magmatic population. Thus, there is a component in the granitoids which is only 15 Ma or so older than the magmatic age.

Granitoid rocks with crystallisation ages around 390 Ma are widespread in the Lachlan Fold Belt in SE Australia, but are not common in New Zealand. Ireland (1992) has reported a 400-Ma zircon analysed in situ (polished thin-section) in a sample of paragneiss from the Western Province and Gibson et al. (1988) obtained an age of 386 ± 10 Ma for a gneissic tonalite from the upper plate of the Fiordland Metamorphic Core Complex. The limited data suggest that localised igneous activity may have occurred in the Western Province a short time before the emplacement of the Karamea Batholith.

9. Regional correlation

9.1. NW Nelson–Westland

The Western Province in NW Nelson–Westland is divided into two Early Paleozoic terranes (Cooper, 1989) along the Anatoki Thrust (Fig. 1). The western Buller terrane comprises mainly quartz-rich turbidites (Greenland Group) and minor black shales in which Ordovician graptolites are preserved. The eastern Takaka terrane is lithologically much more diverse, comprising Cambrian volcanic and volcaniclastic rocks, Ordovician carbonate sequences and Silurian quartzites. Both terranes have small outliers of Lower Devonian rocks, although these differ in character.

Devonian granitoids appear to be restricted to the Buller terrane, although ultramafic–mafic igneous rocks with similar ages are represented in the Takaka terrane by the Riwaka Complex. The Paleozoic granitoids have many of the petrographic and geochemical characteristics of S-type granites (Tulloch, 1983, 1988; Cooper and Tulloch, 1992). In contrast, isotopic data from a sample of the Riwaka Complex (Pickett and Wasserburg, 1989) indicate a more primitive, mantle-derived origin [$(^{87}Sr/^{86}Sr)_i$ at 375 Ma is 0.704 and ϵ_{Nd} is + 1.3].

Cooper and Tulloch (1992) tentatively identified a chemical trend or compositional polarity within the Paleozoic granitoids, with decreasing crustal involvement to the SW. This polarity is based on the presence of a single I-type granite at Paringa in South Westland (Fig. 1), that has given K–Ar ages around 300 Ma (Aronson, 1968). However, further to the SW, Paleozoic two-mica granite exposed near Mount MacLean (Aronson, 1968) has an isotopic composition (Pickett and Wasserburg, 1989) that points to substantial crustal involvement [$(^{87}Sr/^{86}Sr)_i$ at 375 Ma is 0.722, ϵ_{Nd} is − 13]. Clearly more data are required before we can firmly identify any compositional polarity within the Paleozoic igneous rocks of the Western Province.

No Carboniferous ages (~330 Ma), similar to those from the Challenger Plateau (Tulloch et al., 1991), Windy Point or Cape Foulwind Granites (Muir et al., 1994), were obtained from the Karamea Batholith. The Cape Foulwind and Windy Point Granites have many of the characteristics of A-type granites (e.g., high FeO*/MgO, Ga/Al, Zr; Eby, 1992), and form a separate suite distinct from the Devonian granitoids further east.

9.2. *Fiordland*

The remote Fiordland region in the SW corner of the South Island (Fig. 1) is generally thought to represent deeper levels of exposure than the NW Nelson–Westland area (Cooper, 1989; Gibson, 1992). The two areas were probably in close proximity prior to being offset by dextral movement along the Alpine Fault during the late Cenozoic. Western Fiordland consists mainly of complexly deformed metaplutonic and metasedimentary rocks, whereas in the east the intrusives are relatively undeformed and appear to have been emplaced at higher levels (Oliver and Coggan, 1979). The western area is dominated by a belt of granulite-facies, dioritic orthogneiss and retrogressed amphibolite-facies equivalents, collectively termed the Western Fiordland Orthogneiss (J.Y. Bradshaw, 1990). Early Cretaceous ages from the orthogneiss (Mattinson et al., 1986; McCulloch et al., 1987; Gibson et al., 1988) are thought to represent the time of emplacement of the igneous protolith. Older ages of ~380 Ma have been obtained from foliated granitoid enclaves within the orthogneiss (J.Y. Bradshaw and Kimbrough, 1991) and from the structurally overlying cover sequence (Oliver, 1980; Gibson et al., 1988). In addition, a poorly constrained U–Pb zircon age of ~320 Ma has been reported from the Doubtful Sound area (Aronson, 1968). Hence, on the basis of the available age data, Late Devonian and Early Carboniferous magmatism can be identified in both NW Nelson–Westland and in Fiordland.

10. Gondwana

The Western Province of New Zealand (Landis and Coombs, 1967) is confined to the NW and SW parts of the South Island. It extends offshore, probably at least to the Campbell Plateau in the SE and the Lord Howe Rise in the NW (Fig. 1), and represents a fragment of Gondwana that has been separated from SE Australia and Antarctica as a result of late Mesozoic crustal extension and sea-floor spreading (e.g., Mayes et al., 1990). A pre-drift reconstruction of the SW Pacific margin of Gondwana during the Mid-Cretaceous (Weaver et al., 1994) is shown in Fig. 7.

The main sedimentary, tectonic and magmatic events that occurred in the Western Province of New Zealand during the Paleozoic can be matched with similar events recorded in SE Australia, Tasmania and Antarctica (e.g., Cooper and Grindley, 1982; Findlay, 1988; Cooper and Tulloch, 1992). For example, the Ordovician Greenland Group in the Buller terrane has been correlated with the Swanson Formation in Marie Byrd Land (Adams, 1986), the Robertson Bay Group of Northern Victoria Land (J.D. Bradshaw et al., 1985), the Mathinna Beds in NE Tasmania (Powell and Baillie, 1992), and the Stawell and Bendigo terranes in SE Australia (Stump et al., 1986; Cooper and Tulloch, 1992). Equivalents of the Cambrian–Silurian rocks of the Takaka terrane occur in the Bowers terrane of Northern Victoria Land (J.D. Bradshaw et al., 1985), in the Dundas Trough and Mount Read Belt in western Tasmania (Corbett and Solomon, 1989), and in the Grampians–Stavely terrane in SE Australia (Stump et al., 1986). Late Ordovician/Early Silurian and Early Devonian tectonism in the Western Province (Cooper, 1989) can also be identified in most of these areas (Findlay, 1988).

Middle–Late Devonian age granitoids are widespread in SE Australia, Tasmania and Antarctica. However, Cooper and Tulloch (1992) concluded that the Paleozoic granitoids in New Zealand were predominantly Carboniferous in age and had no direct equivalents in the Australian–Antarctic segment of Gondwana. We have demonstrated in this study that the majority of the Paleozoic granitoids in the Western Province are Middle–Late Devonian in age and possible correlations should therefore be discussed.

Granitoid plutonism of mid-Paleozoic age occurred in both the Bowers and Robertson Bay terranes in northern Victoria Land, and in Marie Byrd Land in West Antarctica (e.g., Adams, 1987; Borg et al., 1987; Weaver et al., 1991). In northern Victoria Land, the Admiralty Intrusives (Fig. 7) are a suite of tonalites,

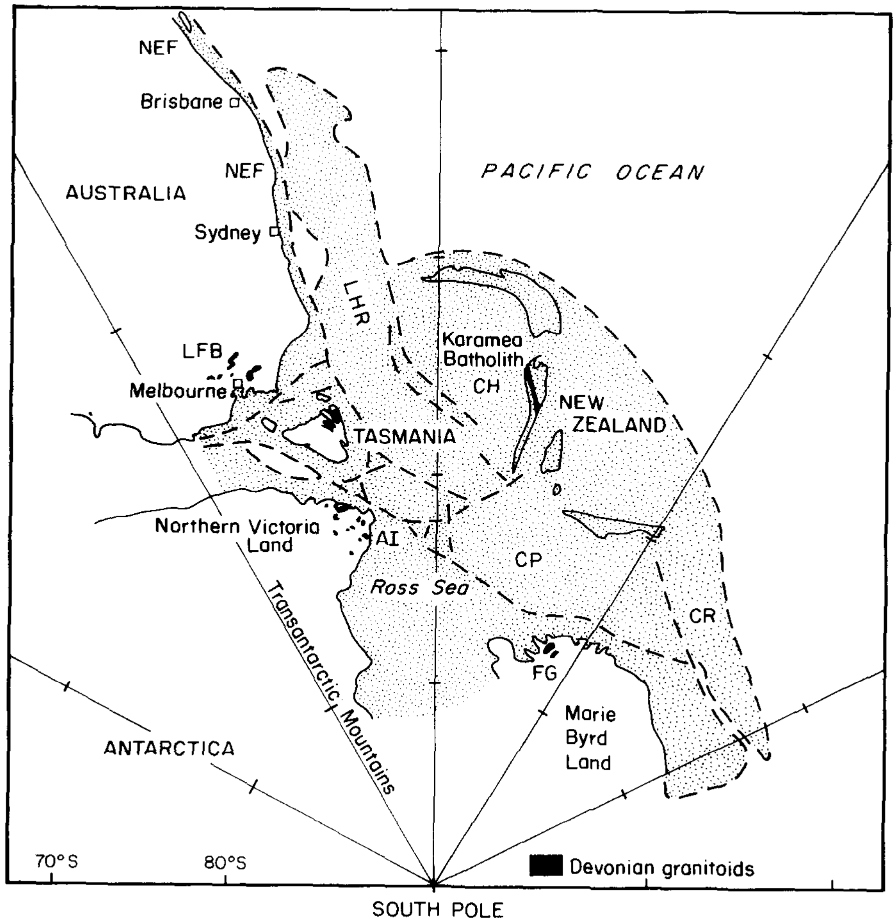


Fig. 7. Distribution of continental masses around the SW Pacific at 100 Ma (after Weaver et al., 1994). *Stipple* indicates thinned continental crust. *FG* = Ford Granodiorite; *AI* = Admiralty Intrusives; *LFB* = Lachlan Fold Belt; *NEF* = New England Fold Belt; *LHR* = Lord Howe Rise; *CH* = Challenger Plateau; *CP* = Campbell Plateau; *CR* = Chatham Rise.

granodiorites and monzogranites that give Rb–Sr whole-rock ages in the range 390–360 Ma. The rocks are predominantly metaluminous/I-type, but according to Borg et al. (1987) there is a strong compositional polarity, indicating increasing involvement of old crustal material from SW to NE. I-type granitoids, collectively known as the Ford Granodiorite, are also extensively exposed in Marie Byrd Land (Fig. 7). The petrography, geochemistry and Rb–Sr age range of the Ford Granodiorite (380–350 Ma) closely match the southern and central Admiralty Intrusives (Weaver et al., 1991).

The crystallisation ages of the Paleozoic granitoids forming the Karamea Batholith and related plutons (375 Ma) clearly lie within the age range of both the Admiralty Intrusives and the Ford Granodiorite. Weaver et al. (1991) suggested that the S-type Karamea Suite in the South Island might represent an extension of the compositional trend established for the Admiralty Intrusives. Time-equivalents of the Early Carboniferous Cape Foulwind and Windy Point Granites also occur in northern Victoria Land (Borg et al., 1987) and in the plutonic terranes of SE Australia and Tasmania.

Paleozoic granitoids occur extensively in both the Lachlan and New England Fold Belts in SE Australia (Shaw and Flood, 1981; Chappell and White, 1992). The ages of the granitoids in the New England Fold

Belt range from Late Carboniferous to Triassic, and are therefore younger than the Paleozoic granitoids in the Western Province. Most of the granitoids in the Lachlan Fold Belt are somewhat older (I.S. Williams et al., 1992), and on the basis of reliable U–Pb zircon and Rb–Sr whole-rock data, appear to have been emplaced within a restricted period from 440 to 380 Ma (Silurian–Devonian). However, both I- and S-type granitoids with ages down to ~360 Ma occur in the terranes in the central part of the belt (Fig. 7) north of Melbourne (Richards and Singleton, 1981; Gray, 1990). In the most easterly part of the Lachlan Fold Belt even younger ages of ~320 Ma have also been reported (Chappell and Stephens, 1988). Hence it appears that the two pulses of Paleozoic magmatism identified in New Zealand can also be recognised in the Lachlan Fold Belt. In addition, granitoid rocks of Late Devonian age (~380 Ma) have been recovered from offshore eastern Australia (Hubble et al., 1992) and Devonian and Carboniferous granitoids are abundant in Tasmania. I- and S-type granitoids in the northeastern part of Tasmania range in age from ~395 to 348 Ma (Rb–Sr and ^{40}Ar–^{39}Ar data), and are regarded as significantly older than those of western Tasmania which range from ~367 to 332 Ma (McDougall and Leggo, 1965; Cocker, 1982; E. Williams et al., 1989; Sawka et al., 1990).

In summary, we believe that the similar ages of the Paleozoic igneous rocks from the matched terranes in New Zealand, SE Australia, Tasmania and Antarctica are strong evidence for their correlation. In the continental reconstruction shown in Fig. 7, the Middle–Late Devonian belt of magmatic activity along the Paleo-Pacific margin of Gondwana appears to have been in excess of 2000 km in length. The Antarctic granitoids have many of the geochemical characteristics of subduction-related magmas (Borg et al., 1987; Vetter and Tessensohn, 1987; Weaver et al., 1991). However, at present, there is no geological evidence for a contemporary active subduction zone. According to Chappell and Stephens (1988), the Silurian–Devonian granitoids in SE Australia and Tasmania do not have subduction-related geochemical signatures, although many workers have attempted to relate these rocks to subduction processes (e.g., Fergusson and Glen, 1992 and references therein). In the absence of a clear link between subduction and magmatism, alternative tectonic models for the Lachlan Fold Belt have therefore been proposed. Wyborn (1992) and Collins (1994) have suggested that the granitoids in SE Australia were derived by partial melting of the lower crust in response to crustal thickening and lithospheric delamination. As more geochemical and isotopic data become available from the granitoids in New Zealand, it should prove possible to evaluate these models more fully.

11. Conclusions

Individual zircon crystals from the main intrusive phases forming the Karamea Batholith and related plutons in NW Nelson–Westland, New Zealand, have been dated by ion microprobe (SHRIMP) mass spectrometry. The new dates presented here establish, for the first time, a precise chronology for this composite batholith, and indicate that the duration of plutonic activity spanned a remarkably short period of time. Apart from minor Cretaceous granitoids, the batholith is dominated by rocks that yield crystallisation ages that are unresolvable at 375 Ma (Middle–Late Devonian). The Riwaka Complex, lying to the east of the Karamea Batholith, gave a similar age of 376.9 ± 5.6 Ma.

The granitoids contain a significant amount of inherited zircon, with distinct components at 390 Ma (Lachlan Fold Belt age), 500–600 Ma (Ross–Delamerian age) and 1000 Ma (Grenville age). However, it is not clear whether the old zircons are derived from the lower crust, from the incorporation of Ordovician metasedimentary rocks enclosing the granitoids, or both.

Time-equivalents of the Devonian granitoids in New Zealand can be found in SE Australia, Tasmania and Antarctica. These plutonic rocks are presently scattered around the borderlands of the Southern Ocean, but were almost certainly contiguous prior to the breakup of Gondwana during the late Mesozoic. The Karamea Batholith can now be seen as part of an extensive magmatic belt close to the Paleo-Pacific margin of Gondwana during the Middle–Late Devonian.

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References

Adams, C.J.D., 1986. Geochronological studies of the Swanson Formation of Marie Byrd Land, West Antarctica, and correlation with northern Victoria Land, East Antarctica and the South Island, New Zealand. N.Z. J. Geol. Geophys., 29: 345–358.

Adams, C.J.D., 1987. Geochronology of granite terranes in the Ford Ranges, Marie Byrd Land, West Antarctica. N.Z. J. Geol. Geophys., 30: 51–72.

Aronson, J.L., 1968. Regional geochronology of New Zealand. Geochim. Cosmochim. Acta, 32: 669–697.

Borg, S.G., Stump, E., Chappell, B.W., McCulloch, M.T., Wyborn, D., Armstrong, R.C. and Holloway, J.R., 1987. Granitoids of northern Victoria Land, Antarctica: Implications of chemical and isotopic variations to regional crustal structure and tectonics. Am. J. Sci., 287: 127–165.

Bowen, F.E., 1964. Sheet 15, Buller 1st ed., scale 1:250,000, Geological Map of New Zealand. DSIR (Dep. Sci. Ind. Res.), Wellington.

Bradshaw, J.D., Weaver, S.D. and Laird, M.G., 1985. Suspect Terranes and Cambrian Tectonics in Northern Victoria Land, Antarctica. In: D.G. Howell (Editor), Tectonostratigraphic Terranes of the Circum-Pacific Region. Circum-Pac. Counc. Energy Miner. Resour., Earth Sci. Ser., pp. 467–479.

Bradshaw, J.Y., 1990. Geology of crystalline rocks of northern Fiordland: details of granulite facies Western Fiordland Orthogneiss and associated rock units. N.Z. J. Geol. Geophys., 33: 465–484.

Bradshaw, J.Y. and Kimbrough, D.L., 1991. Mid-Paleozoic age of granitoids in enclaves in Early Cretaceous granulites, Fiordland, southwest New Zealand. N.Z. J. Geol. Geophys., 34: 455–469.

Braithwaite, R.L. and Pirajno, F., 1993. Metallogenic Map of New Zealand. Inst. Geol. Nucl. Sci., Wellington, Monogr. 3, 215 pp.

Chappell, B.W. and Stephens, W.E., 1988. Origin of infracrustal (I-type) granite magmas. Trans. R. Soc. Edinburgh, Earth Sci., 79: 71–86.

Chappell, B.W. and White, A.J.R., 1992. I- and S-type granites in the Lachlan Fold Belt. Trans. R. Soc. Edinburgh, Earth Sci., 83: 1–26.

Chen, Y.D. and Williams, I.S., 1990. Zircon inheritance in mafic inclusions from Bega batholith granites, southeastern Australia: An ion microprobe study. J. Geophys. Res., 95(B11): 17787–17796.

Cocker, J.D., 1982. Rb–Sr geochronology and Sr isotopic composition of Devonian granitoids, eastern Tasmania. J. Geol. Soc. Aust., 29: 139–158.

Collins, W.J., 1994. Upper- and middle-crustal response to delamination: and example from the Lachlan Fold Belt, eastern Australia. Geology, 22: 143–146.

Cooper, R.A., 1989. Early Paleozoic terranes of New Zealand. J.R. Soc. N.Z., 19: 73–112.

Cooper, R.A. and Grindley, G.W., 1982. Late Proterozoic to Devonian sequences of southeastern Australia, Antarctica and New Zealand and their correlation. Geol. Soc. Aust., Spec. Publ. No. 9, 103 pp.

Cooper, R.A. and Tulloch, A.J., 1992. Early Palaeozoic terranes in New Zealand and their relationship to the Lachlan Fold Belt. Tectonophysics, 214: 129–144.

Corbett, K.D. and Solomon, M., 1989. Cambrian Mt Read Volcanics and Associated Mineral Deposits. In: C.F. Burnett and E.L. Martin (Editors), Geology and Mineral Resources of Tasmania. Geol. Soc. Aust., Spec. Publ., 15: 84–153.

Dalziel, I.W.D., 1991. Pacific margins of Laurentia and East Antarctica–Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent. Geology, 19: 598–601.

Eby, G.N., 1992. Chemical subdivision of the A-type granitoids: Petrogenetic and tectonic implications. Geology, 20: 641–644.

Eggers, A.J. and Adams, C.J., 1979. Potassium–argon ages of molybdenum mineralisation and associated granites at Bald Hill, and correlation with other molybdenum occurrences in the South Island, New Zealand. Econ. Geol., 74: 628–637.

Fergusson, C.L. and Glen, R.A., 1992. The Palaeozoic eastern margin of Gondwanaland: Tectonics of the Lachlan Fold Belt, southeastern Australia and related orogens. Tectonophysics, 214: 461 pp.

Findlay, R.H., 1988. Silurian and Devonian events in the Tasman Orogenic Zone, New Zealand and Marie Byrd Land and their comparison with northern Victoria Land. Mem. Soc. Geol. Ital., 43: 9–32.

Gibson, G.M., 1992. Medium–high-pressure metamorphic rocks of the Tuhua Orogen, western New Zealand, as lower crustal analogues of the Lachlan Fold Belt, SE Australia. Tectonophysics, 214: 145–157.

Gibson, G.M., McDougall, I. and Ireland, T.R., 1988. Age constraints on metamorphism and the development of a metamorphic core complex in Fiordland, southern New Zealand. Geology, 16: 405–408.

Gill, K.R. and Johnston, M.R., 1970. The geology and nickel–copper sulphide mineralisation in the Graham Valley, north-west Nelson. N.Z. J. Geol. Geophys., 13: 477–494.

Gray, C.M., 1990. A strontium isotopic traverse across the granitic rocks of southeastern Australia: Petrogenetic and tectonic implications. Aust. J. Earth Sci., 37: 331–349.

Grindley, G.W., 1961. Sheet 13, Golden Bay 1st ed., scale 1:250,000, Geological Map of New Zealand. DSIR (Dep. Sci. Ind. Res.), Wellington.

Grindley, G.W., 1980. Sheet S13, Cobb 1st ed., scale 1:63,360, Geological Map of New Zealand. DSIR (Dep. Sci. Ind. Res.), Wellington.

Grindley, G.W. and Davey, F.J., 1982. The reconstruction of Australia, New Zealand and Antarctica. In: C. Craddock (Editor), Antarctic Geoscience. University of Wisconsin Press, Madison, Wis., pp. 15–29.

Harrison, T.M. and McDougall, I., 1980. Investigations of an intrusive contact, northwest Nelson, New Zealand, 1. Thermal, chronological and isotopic constraints. Geochim. Cosmochim. Acta, 44: 1985–2003.

Hubble, T.C.T., Packham, G.H., Hendry, D.A.F. and McDougall, I., 1992. Granitic and monzonitic rocks dredged from the southeast Australian continental margin. Aust. J. Earth. Sci., 39: 619–630.

Ireland, T.R., 1992. Crustal evolution of New Zealand: Evidence from age distributions of detrital zircons in Western Province paragneisses and Torlesse Greywacke. Geochim. Cosmochim. Acta, 56: 911–920.

Ireland, T.R., Bradshaw, J., Muir, R., Weaver, S. and Adams, C., 1994. Zircon age distributions in granites, greywackes and gneisses from the southwest Pacific Gondwana region. Geol. Soc. Aust. Abstr., 37: 192–193.

Kimbrough, D.L. and Tulloch, A.J., 1989. Early Cretaceous age of orthogneiss from the Charleston Metamorphic Group, New Zealand. Earth Planet. Sci. Lett., 95: 130–140.

Kutsukake, T., 1988. The Britannia Granite pluton in the Karamea Batholith, South Island, New Zealand. N.Z. J. Geol. Geophys., 31: 275–286.

Laird, M.G., 1988. Sheet S37, Punakaiki 1st ed., scale 1:63,360, Geological Map of New Zealand. DSIR (Dep. Sci. Ind. Res.), Wellington.

Landis, C.A. and Coombs, D.S., 1967. Metamorphic belts and orogenesis in southern New Zealand. Tectonophysics, 4: 501–518.

Le Maitre, R.W., 1989. A Classification of Igneous Rocks and Glossary of Terms: Recommendations of the international Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks. Blackwell, Oxford, 193 pp.

Mayes, C.L., Lawver, L.A. and Sandwell, D.T., 1990. Tectonic history and new isochron chart of the South Pacific. J. Geophys. Res., 95B: 8543–8567.

Mattinson, J.M., Kimbrough, D.L. and Bradshaw, J.Y., 1986. Western Fiordland orthogneiss: Early Cretaceous arc magmatism and granulite facies metamorphism, New Zealand. Contrib. Mineral. Petrol., 92: 383–392.

McCulloch, M.T., Bradshaw, J.Y. and Taylor, S.R., 1987. Sm–Nd and Rb–Sr isotopic and geochemical systematics in Phanerozoic granulites from Fiordland, southwest New Zealand. Contrib. Mineral. Petrol., 97: 183–195.

McDougall, I. and Leggo, P.J., 1965. Isotopic age determinations on granitic rocks from Tasmania. J. Geol. Soc. Aust., 12: 295–332.

Moores, E.M., 1991. Southwest U.S.–East Antarctic (SWEAT) connection: A hypothesis. Geology, 19: 425–428.

Muir, R.J., Ireland, T.R., Weaver, S.D. and Bradshaw, J.D., 1994. Ion microprobe U–Pb zircon geochronology of granitic magmatism in the Western Province of the South Island, New Zealand. Chem. Geol. (Isot. Geosci. Sect.), 113: 171–189.

Nathan, S., 1978. Sheets S31 and part S32, Buller–Lyell 1st ed., scale 1:63,360, Geological Map of New Zealand. DSIR (Dep. Sci. Ind. Res.), Wellington.

Oliver, G.J.H., 1980. Geology of the granulite and amphibolite gneisses of Doubtful Sound, Fiordland, New Zealand. N.Z. J. Geol. Geophys., 23: 27–41.

Oliver, G.J.H. and Coggan, J.H., 1979. Crustal structure of Fiordland, New Zealand. Tectonophysics, 54: 253–292.

Paces, J.B. and Miller, J.D., 1989. Precise U–Pb ages of Duluth Complex and related mafic intrusions, Northeastern Minnesota: Geochronological insights to physical, petrogenetic, paleomagnetic and tectonomagmatic processes associated with the 1.1 Ga midcontinent rift system. J. Geophys. Res., 98B: 13997–14013.

Pickett, D.A. and Wasserburg, G.J., 1989. Neodymium and strontium isotopic characteristics of New Zealand granitoids and related rocks. Contrib. Mineral. Petrol. 103: 131–142.

Powell, C.McA. and Baillie, P.W., 1992. Tectonic affinity of the Mathinna Group in the Lachlan Fold Belt. Tectonophysics, 214: 193–210.

Richards, J.R. and Singleton, O.P., 1981. Palaeozoic Victoria, Australia: igneous rocks, ages and their interpretation. J. Geol. Soc. Aust., 28: 395–421.

Roder, G.H. and Suggate, P.R., 1990. Sheet L29BD —Upper Buller Gorge, Geological Map of New Zealand 1:50 000. DSIR (Dep. Sci. Ind. Res.), Wellington.

Sambridge, M.S. and Compston, W., 1994. Mixture modelling of multi-component data sets with application to ion probe zircon ages. Earth Planet. Sci. Lett., 128: 373–390.

Sawka, W.N., Heizler, M.T., Kistler, R.W. and Chappell, B.W., 1990. Geochemistry of highly fractionated I- and S-type granites from the tin–tungsten province of western Tasmania. In: Ore-bearing Granite Systems: Petrogenesis and Mineralizing Processes. Geol. Soc. Am., Spec. Pap., 246: 161–179.

Shaw, S.E. and Flood, R.H., 1981. The New England Batholith, eastern Australia: Geochemical variations in time and space. J. Geophys. Res., 86B: 10530–10544.

Stump, E., White, A.J.R. and Borg, S.G., 1986. Reconstruction of Australia and Antarctica: evidence from granites and recent mapping. Earth Planet. Sci. Lett., 79: 348–360.

Tera, F. and Wasserburg, G.J., 1972. U–Th–Pb Systematics in three Apollo 14 basalts and the problem of initial Pb in lunar rocks. Earth Planet. Sci. Lett., 14: 281–304.

Tulloch, A.J., 1983. Granitoid rocks of New Zealand — a brief review. Geol. Soc. Am. Mem., 159: 5–20.

Tulloch, A.J., 1988. Batholiths, plutons and suites: nomenclature for granitoid rocks of Westland–Nelson, New Zealand. N.Z. J. Geol. Geophys., 31: 505–509.

Tulloch, A.J., Kimbrough, D.L. and Wood, R.A., 1991. Carboniferous granite basement dredged from a site on the Challenger Plateau, Tasman Sea. N.Z. J. Geol. Geophys., 34: 121–126.

Tulloch, A.J. and Rabone, S.D., 1993. Mo-bearing granodiorite porphyry plutons of the Early Cretaceous Separation Point Suite, west Nelson, New Zealand. N.Z. J. Geol. Geophys., 36: 401–408.

Vetter, U. and Tessensohn, F., 1987. S- and I-type granitoids of north Victoria Land, Antarctica, and their inferred geotectonic setting. Geol. Rundsch., 76(1): 233–243.

Weaver, S.D., Bradshaw, J.D. and Adams, C.J., 1991. Granitoids of the Ford Ranges, Marie Byrd Land, Antarctica. In: M.R.A. Thomson, J.A. Crame and J.W. Thomson (Editors), Geological

Evolution of Antarctica. Cambridge University Press, Cambridge, pp. 345–351.

Weaver, S.D., Storey, B.C., Pankhurst, R.J., Mukasa, S.B., Di-Venere, V.J. and Bradshaw, J.D., 1994. Antarctica–New Zealand rifting and Marie Byrd Land lithospheric magmatism linked to ridge subduction and mantle plume activity. Geology, 22: 811–814.

Williams, E., McClenaghan, M.P. and Collins, P.L.F., 1989. Mid-Palaeozoic Deformation, Granitoids and Ore Deposits. In: C.F. Burnett and E.L. Martin (Editors), Geology and Mineral Resources of Tasmania. Geol. Soc. Aust., Spec. Publ., 15: 238–292.

Williams, I.S., 1992. Some observations on the use of zircon U–Pb geochronology in the study of granitic rocks. Trans. R. Soc. Edinburgh, Earth Sci., 83: 447–458.

Williams, I.S. and Claesson, S., 1987. Isotopic evidence for the Precambrian provenance and Caledonian metamorphism of high grade paragneisses from the Seve Nappes, Scandinavian Caledonides, II. Ion microprobe zircon U–Th–Pb. Contrib. Mineral. Petrol., 97: 205–217.

Williams, I.S., Chappell, B.W., Chen, Y.D. and Crook, K.A.W., 1992. Inherited and detrital zircons —vital clues to the granite protoliths and early igneous history of southeast Australia. Trans. R. Soc. Edinburgh, Earth Sci., 83: 503.

Wyborn, D., 1992. The tectonic significance of Ordovician magmatism in the eastern Lachlan Fold Belt. Tectonophysics, 214: 177–192.