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Stable and Transient Isotopic Trends in the Crustal Evolution of Zealandia Cordillera

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ABSTRACT

We present >500 zircon $\delta^{18}O$ and Lu-Hf isotope analyses on previously dated zircons to explore the interplay between spatial and temporal magmatic signals in Zealandia Cordillera. Our data cover ~8,500 km^2 of middle and lower crust in the Median Batholith (Fiordland segment of Zealandia Cordillera) where Mesozoic arc magmatism along the paleo-Pacific margin of Gondwana was focused along an ~100 km-wide, arc-parallel zone. Our data reveal three spatially distinct isotope domains which we term the eastern, central and western isotope domains. These domains parallel the Mesozoic arc-axis and these boundaries are associated with major lithospheric-scale faults that were reactivated as ductile shear zones during the Early Cretaceous. The western isotope domain has homogenous, mantle-like $\delta^{18}O$ (Zrn) values of 5.8 ± 0.3‰ (2SD) and initial ε_{Hf} (Zrn) values of +4.2 ± 1.0 (2SD). The eastern isotope domain is defined by isotopically low and homogenous $\delta^{18}O$ (Zrn) values of 3.9 ± 0.2‰ and initial ε_{Hf} values of +7.8 ± 0.6. The central isotope domain is characterized by transitional isotope values that display a strong E-W gradient with $\delta^{18}O$ (Zrn) values rising from 4.6‰ to 5.9‰ and initial ε_{Hf} values decreasing from +5.5 to +3.7. We find that the isotope architecture of the Median Batholith was in place before initiation of Mesozoic arc magmatism and its formation pre-dates Early Cretaceous contractional deformation and transpression. Our data show that Mesozoic pluton chemistry was controlled in part by long-lived, spatially distinct isotope domains that extend from the crust through to the upper mantle. Isotope differences between these domains are the result of the crustal architecture (an underthrusted low-$\delta^{18}O$ source terrane) and a transient event beginning at ca. 129 Ma that introduced a strong 'enriched' mantle signature. When data showing the temporal and spatial patterns of magmatism are integrated, we observe a pattern of decreasing crustal recycling of the low-$\delta^{18}O$ source over time, which ultimately culminated in a mantle-controlled flare-up with little influence from pre-existing crustal sources. Our data demonstrate that spatial and temporal signals are intimately linked and when evaluated together they provide important insights into crustal architecture and the role of both stable and transient arc magmatic trends in Cordilleran batholiths.

Keywords: Cordilleran magmatism, Zealandia, zircon, O isotopes, Hf isotopes

Introduction

The crustal architecture of continental margins plays an important role in influencing the location of Cordilleran-arc magmatism and the geochemical and isotope evolution of arc magmas from their source to emplacement (e.g., Ducea et al., 2015a). Geochemical and isotope data from arc magmas are often used as important features in evaluating source regions and differentiation processes that ultimately lead to the generation of continental crust through time (Rudnick, 1995; Taylor and McLennan, 1995; Ducea and Barton, 2007; Scholl and von Huene, 2007; Hawkesworth et al., 2010; Voice et al., 2011; Ducea et al., 2017). However, the record of pre-existing crustal sources and their relationship to terrane and intra-terrane faults is commonly highly disrupted by a variety of factors including voluminous magmatic intrusions, polyphase metamorphism, and various phases of brittle and ductile faulting. The end result is that surficial exposures of long-lived Cordilleran arcs preserve an incomplete record of crustal sources and the pre-batholithic architecture of the arc which were once key factors in its temporal and spatial magmatic evolution.

One of the problems in understanding isotope variations in arc magmas is that isotope signals can be influenced by a number of competing factors, including spatially controlled features such as the crustal and upper mantle architecture and composition of the arc (Armstrong, 1988; Ducea and Barton, 2007), versus various transient tectonic and non-tectonic processes that can introduce new sources. The latter may include processes such as delamination and arc root foundering (Kay et al., 1994; Ducea, 2002; Ducea et al., 2013), fore-arc underplating (Chapman et al., 2013), subduction erosion (Kay et al., 2005), retro-arc underthrusting of continental crust (DeCelles et al., 2009; DeCelles and Graham, 2015), relamination of subducted sediment (Hacker et al., 2011), and/or slab tears and slab windows (Thorkelson, 1996; Dickenson, 1997).

Understanding which of these mechanisms controls geochemical and isotope changes in arc magma chemistry is critical in evaluating continental crustal growth processes, including triggering mechanisms for voluminous arc magmatic surges (Paterson and Ducea, 2015; Ducea et al., 2015b; de Silva et al., 2015).

The Mesozoic Median Batholith that comprises much of Fiordland New Zealand is a prime location to explore the interplay between spatial- and temporal-isotope signals because it comprises ~10,000 km^2 of lower, middle and shallow arc crust built on the southeast margin of Gondwana from the Devonian to Early Cretaceous (**Fig. 1A-F**) (Landis and Coombs, 1967; Mortimer et al., 1999a; Tulloch and Kimbrough, 2003). The importance of the Median Batholith in understanding Cordilleran-arc magmatic processes is underscored by competing models for the Early Cretaceous surge of high-Sr/Y magmatism. In one model, Muir et al. (1995) and Milan et al. (2017) used bulk-rock and zircon radiogenic isotope data (Sr-Nd-Hf) to argue for increasing contributions of ancient (radiogenic) continental crustal sources during the continentward advance of the arc. In contrast, Decker et al. (2017) used both stable and radiogenic zircon isotope data (O and Lu-Hf) from the lower crust of the Median Batholith to propose a mantle (rather than crustal) trigger for the Early Cretaceous flare-up stage. In neither study were spatial isotope trends investigated nor was the role of the pre-existing crustal architecture considered. Consequently, an outstanding question is whether geochemical and isotope shifts observed in magmatic chemistry in the Mesozoic portion of the Median Batholith reflect temporally transient arc processes (c.f., increased coupling, underthrusting of continental crust, and changes to the lower plate), or temporally stable processes influenced by long-lived pre-batholithic crustal and/or upper lithospheric mantle architecture of the Cordilleran arc system. In the Median Batholith, this problem is compounded by the fact that there is little consensus about the pre-batholithic crustal architecture, the nature and location of isotope boundaries, nor the timing of terrane juxtaposition prior to voluminous arc-magmatic activity in the Early Cretaceous (Kimbrough et al., 1994; Adams et al., 1998; Muir et al., 1998; Mortimer et al., 1999a; Scott et al., 2009; McCoy-West et al., 2014).

Zircon isotope studies in plutonic rocks can improve our understanding of the crustal architecture and spatial isotope trends prior to batholith emplacement because they can reveal differences in source regions from which melts were derived (e.g., Valley, 2003; Lackey et al., 2005; Lackey et al., 2008; Cecil et al., 2011; Lackey et al., 2012). For example, oxygen isotopes are particularly sensitive indicators of melt-rock interaction and differentiate low-temperature hydrothermally altered sources, such as marine sediments ($\delta^{18}0$ >>6‰), from high-temperature hydrothermally altered sources or those altered at high paleo-latitude or -altitude conditions ($\delta^{18}O$ <<6‰). Similarly, hafnium isotopes differentiate depleted mantle-derived melts (ε_{Hf} = +16 to +18) from crustal sources (initial ε_{Hf} << +16). Moreover, variations in zircon isotope values within samples also provide information about whether isotope signatures were acquired from deep-crustal source regions versus during ascent or at the depth of emplacement. Isotope signatures acquired in the deep crust or upper mantle often display homogenous values with low intrasample standard deviations reflecting efficient isotope homogenization in high-temperature melt-rich systems, whereas large intrasample variations can be caused by assimilation of crustal sources during ascent or emplacement in melt-poor, crystal mushes (Valley et al., 1998; Miller et al., 2007; Bindeman, 2008). In this study, we use a series ~60-km long, arc-perpendicular zircon isotope transects from Jurassic to Early Cretaceous plutons to investigate the isotope characteristics of the Median Batholith with the goal of understanding the pre-batholithic crustal architecture and evaluating temporal and spatial isotope variations in Cordilleran crust construction in Zealandia Cordillera (**Fig. 1**).

Geologic Background

Regional Geology of the Median Batholith

Currently exposed Pre-Late Cretaceous Zealandia is divided into two lithologic provinces: the Eastern Province and the Western Province (**Fig. 1A-B, F**). The Eastern Province consists of dominantly Permian to Early Cretaceous accreted terranes composed of sedimentary and metasedimentary terranes (Murihiku, Caples, Torlesse, Waipapa Terranes, and Haast Schist), an ophiolite belt (Dun Mountain/Maitai Terrane), and an intraoceanic island arc terrane (Brook Street Terrane) (Landis and Coombs, 1967; Frost and Coombs, 1989; Bradshaw, 1993; Mortimer et al., 1999a; Tulloch et al., 1999; Mortimer, 2004; Campbell et al., 2020).

The Western Province is comprised of Early Paleozoic Gondwana-like affinity metasedimentary and metavolcanic rocks comprising Buller and Takaka terranes that were accreted to the Gondwana margin in the Early Paleozoic (Cooper and Tulloch, 1992, Jongens, 1997; 2006) and intruded by Ordovician to Cretaceous plutons (Mortimer et al. 2004; Tulloch et al. 2009) (**Fig. 1A-B**). While the boundary between the Eastern Province and Western Provinces is generally thought to lie between the Brook Street Terrane and the Takaka/Buller terranes, the region is overprinted by Cenozoic brittle faults and Mesozoic plutons of the Median Batholith (**Fig. 1A-B**). Thus, there is no consensus on the location of the boundary between the Eastern Province and Western Provinces nor the timing of its formation. We investigate this boundary in this contribution because it forms the background for understanding spatial and temporal isotope trends in the Median Batholith.

Mortimer et al. (1999) recognized that the boundary between the Eastern and Western Provinces is essentially defined by variably deformed batholithic rocks and they coined the term 'Median Batholith' to describe the region where batholithic rocks intrude both Western Province and Eastern Province terranes (**Fig. 1**). Tulloch and Kimbrough (2003) subsequently subdivided the Median Batholith into two overlapping plutonic belts (the inboard and outboard belts) which contain the Darran and the Separation Point Suites (**Fig. 1**). Recognition of correlative rocks in off-shore South Zealandia indicates that the Median Batholith extends along at least 2600km of the SE Gondwana margin (Tulloch et al. 2019). In the deeply exhumed 200km long segment of the Median Batholith in Fiordland, the inboard belt is dominated by the monzodioritic Western Fiordland Orthogneiss (WFO) phase of the Separation Point Suite (Oliver 1977; Mattinson et al 1986; Bradshaw 1990). The WFO was emplaced in the lower crust and metamorphosed to granulite facies, in marked contrast to the upper/mid-crustal plutonic and rare volcanic rocks of the Darran Suite that dominate the outboard belt (**Fig. 1; Table 1**).

The boundary between the inboard and outboard belts was defined by Allibone et al. (2009) by the distribution of metasedimentary rocks, whereby those of the inboard belt have Gondwana-affinities, whereas those in the outboard Median Batholith have no apparent association with cratonic Gondwana. In northern Fiordland, Marcotte et al. (2005) suggested that the Indecision Creek Shear Zone represented the boundary between the inboard and outboard Median Batholith, and Scott et al. (2009) suggested that this boundary continued to the south to the subvertical Grebe Shear Zone in Lake Manapouri in central Fiordland (**Fig. 1A**). Buriticá et al. (2019) extended the Grebe Shear Zone into South Fiord, Lake Te Anau and noted that deformation is partitioned into a diffuse network of high- and low-strain mylonitic shear zones whose core deformation zone is located within the Darran Suite. We use the location of the

Grebe and Indecision Creek Shear Zones as defined in these prior studies as the boundary between inboard and outboard plutons rocks in this study.

In a re-evaluation of the inboard/outboard concept, Scott (2013) subdivided the Median Batholith by terranes whereby rocks east of the Grebe Shear Zone-Indecision Creek Shear Zone are considered part of the Drumduan Terrane, and rocks west of the Grebe Shear Zone-Indecision Creek Shear Zone are considered part of the Takaka Terrane (**Fig. 1A, F**). These subdivisions are illustrated in **Table 1** and in **Fig. 1A** along with the isotope domains that we introduce later. For the sake of simplicity, we continue to use the historical terms 'inboard' and 'outboard' Median Batholith to describe plutonic rocks relative to the Grebe Shear Zone-Indecision Creek Shear Zone because these terms are ingrained in the literature.

Previous Isotope Studies of the Median Batholith

The inboard Median Batholith consists of Mesozoic plutons including the Western Fiordland Orthogneiss, and parts of Separation Point Suite and Darran Suite (**Fig. 1A**) Kimbrough et al., 1994; Gibson and Ireland, 1996; Scott and Palin, 2008; Allibone et al., 2009a, 2009b; Milan et al., 2016; Milan et al., 2017; Decker et al., 2017; Schwartz et al., 2017; Buriticá et al., 2019). Only the Western Fiordland Orthogneiss in the inboard Median Batholith has been investigated isotopically in detail. It has $\delta^{18}O$ (Zrn) values ranging from 5.2 to 6.3‰ and initial ε_{Hf} (Zrn) values ranging from -2.0 to +11.2 (Bolhar et al., 2008; Milan et al., 2016; Decker et al., 2017). Western Fiordland Orthogneiss plutonic rocks have bulk-rock initial $^{87}Sr/^{86}Sr$ values of 0.70391 ± 4, and initial ε_{Nd} values ranging from -0.4 to +2.7 (McCulloch et al., 1987).

The outboard Median Batholith was defined by Tulloch and Kimbrough (2003) by the presence of Triassic to Cretaceous Darran Suite (235-132 Ma) and also includes some Separation

Point Suite rocks (125-122 Ma) (Kimbrough et al., 1994; Muir et al., 1998; Tulloch and Kimbrough, 2003; Bolhar et al., 2008; Scott and Palin, 2008; Allibone et al., 2009b; Scott et al., 2009; Buriticá et al., 2019). There are no $\delta^{18}O$ (Zrn) data for the outboard Darran Suite; however, Scott et al., (2009) report initial ε_{Hf} (Zrn) values of +5.9 to +10.0 from 2 samples, and McCulloch et al. (1987) report bulk-rock initial $^{87}Sr/^{86}Sr$ values from 0.70373 and 0.70387 and initial ε_{Nd} values between +3.9 to +4.6. From the Separation Point Suite, Bolhar et al. (2008) report 3 samples that have $\delta^{18}O$ (Zrn) values of 3.1 to 4.4‰ and initial ε_{Hf} (Zrn) values of +7.4 to +8.3. Muir et al. (1998) also report two bulk-rock initial $^{87}Sr/^{86}Sr$ values from the same samples which range from 0.70375-0.70377 and have initial ε_{Nd} values of +3.2.

Methods

Bulk-rock geochemistry

Bulk-rock samples were powdered in an alumina ceramic shatter-box. Powders were mixed with a 2:1 ratio of SpectroMelt A10 lithium tetra borate flux and melted at 1000°C for approximately 20 minutes to create glass beads at California State University, Northridge. Beads were repowdered, refused following the initial melting parameters, and polished to remove carbon from the flat bottom where analysis occurs. Following procedures outlined in Lackey et al. (2012), glass beads were analyzed at Pomona College for major (SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MnO, MgO, CaO, Na_2O, K_2O, P_2O_5) and trace (Rb, Sr, Ba, Zr, Y, Nb, Cs, Sc, V, Cr, Ni, Cu, Zn, Ga, La, Ce, Pr, Nd, Hf, Pb, Th, U) elements by X-ray fluorescence (XRF). Beads were analyzed with a 3.0kW Panalytical Axios wavelength-dispersive XRF spectrometer with PX1, GE, LiF 220, LiF 200, and PE analyzer crystals. Bulk-rock geochemistry values are shown in **Figure 2** and reported in Supplemental Table 1.

Zircon separation

Zircons were extracted from rock samples at the CSUN rock lab following methods in Schwartz et al. (2017). Zircons without visible inclusions were hand-picked and placed onto double sided tape with zircon standards. The KIM-5 zircon standard ($\delta^{18}O$ = 5.09 +- 0.06‰ VSMOW, Valley 2003) was mounted near the center of each mount for oxygen isotope analysis. Zircons were imaged in epoxy mounts using a Gatan MiniCL cathodoluminescence detector on a FEI Quanta 600 Scanning Electron Microscope at CSUN (**Fig. 3**). U-Pb zircon geochronology data were collected by Secondary Ion Mass Spectrometry at the Stanford-USGS Sensitive High Resolution Ion Microprobe – Reverse Geometry (SHRIMP-RG) facility. These age data are reported in Buriticá et al. (2019) and geochronology data are summarized in **Table 2**.

Zircon Oxygen Isotopes

Zircon oxygen isotope analyses were conducted at the University of Wisconsin-Madison WiscSIMS lab using a CAMECA IMS 1280 ion microprobe, following procedures from Kita et al. (2009). CL and reflected light imaging were conducted on mounts after U-Pb analyses to select locations (from the same magmatic domain) for O analysis. Zircon mounts were polished with a 6, 3, and 1 μm diamond lapping film to remove U-Pb spots collected prior to O-isotope analyses. Where zircons maintained a U-Pb pit, oxygen isotopes were collected in a different location with the same igneous domain to avoid contamination from the oxygen beam used during U-Pb analyses. Mounts were cleaned in ethanol and deionized water baths using an ultrasonic cleaner, dried in a vacuum oven, gold-coated, and stored in a de-gassing vacuum prior to secondary ion mass spectrometry (SIMS) analysis. A 10kV $^{133}Cs^{+}$ primary beam was used for analysis of ~10 μm spots. Oxygen isotopes ($^{18}O^-$, $^{16}O^-$) and $^{16}O^1H^-$ were collected in three Faraday cups. For each session, four KIM-5 zircon standards (Valley et al., 2003) were measured before and after analyzing 8 – 15 unknowns. Individual spot analysis errors for KIM-5 ranged from 0.09 – 0.25‰ (ave. = 0.17‰) and were used to determine 2SD error in our samples (Table 2). Values of $^{16}O^1H$ /^{16}O are corrected for background and referred to as OH/O hereafter (Wang et al. 2014). After oxygen isotope analyses, mounts were imaged to verify igneous domains were analyzed at the University of California, Los Angeles using a Tescan Vega-3 XMU variable-pressure (VP) Scanning Electron Microscope (SEM) at 20kV and working distance of 20 mm with a cathodoluminescence detector. Values of $\delta^{18}O$ are reported in permil notation relative to V-SMOW (Supplementary Table 2) and summarized in **Table 2**. Representative zircons and sample spots are shown in **Figure 3**. Zircon oxygen isotope ratios relative to distance from the Grebe Shear Zone (GSZ) are plotted in **Figures 4 and 5**.

Zircon Lu-Hf Isotopes

Hafnium isotope analyses were collected at the Arizona LaserChron Center using a Nu Plasma multicollector ICPMS on individual zircon grains. Prior to analysis, gold coating on mounts and in SIMS pits from previous oxygen-isotope analysis was removed by polishing with 6, 3, and 1 μm diamond lapping film. Mounts were then immersed in a potassium iodine solution. When possible, Hf isotope analyses occurred on existing oxygen pits or similar igneous domains. Using a 7 Hz ArF laser ablation for analysis of 40μm spots, isotopes were collected in 10 Faraday cups that measured Hf, Yb, and Lu between masses 171 – 180. Analysis started by measuring $^{176}Hf/^{177}Hf$ of a 10 ppb solution of standard JMC475 (Vervoort et al., 2004) and one zircon of each standard MT, 91500, Temora (Woodhead and Hergt, 2005), PLES (Slama et al., 2008), and R33 (Fisher et al., 2014), followed by 10 unknowns. For each analytical session, standards and unknowns were reduced together (Supplementary Table 3). A summary of isotope values is reported in **Table 2**. Representative zircons and sample spots are shown in **Figure 3**. Zircon initial εHf values relative to distance from the Grebe Shear Zone are plotted in **Figures 4 and 5**.

Isotope Contour Plots

Isotope contour plots were created by importing latitude, longitude, and isotope values of each sample into Surfer™ 11 using a minimum curvature gridding method. Oxygen and hafnium isotope contour plots of Fiordland were constructed with a latitude range of S 45°8'00.0" to E 44°4'00.0" and longitude range of 166°4'00.0" to 168°2'00.0", and plots were clipped to the shape of Fiordland using Adobe Illustrator™ (**Fig. 6**). The oxygen isotope contour plot has a contour interval of 0.3‰ to encompass the average 2SD uncertainty (0.17‰) for the entire data set, and the hafnium isotope contour plot has a contour interval of 1.0 ε unit to encompass the average 2SD of 1.2 ε units. The oxygen contour plot of samples along Lake Te Anau and Lake Manapouri (locations shown in **Fig. 1**) are plotted with a contour interval of 0.4‰ in Figure 7A to encompass the average 2SD for these samples.

RESULTS

Bulk-Rock Geochemistry

The Western Fiordland Orthogneiss is dominantly monzodioritic in composition, metaluminous, alkali-calcic to calc-alkalic, and magnesian (red circles in **Fig. 2A-D**). It is also characterized by high-Sr/Y bulk-rock values (generally >40) and has moderate depletions in heavy rare earth elements in chondrite-normalized rare earth element plots (e.g., Decker et al., 2017). The Western Fiordland Orthogneiss is part of the more widely distributed high-Sr/Y Separation Point Suite that is characterized by a broad range in SiO_2 (54-76%). Other inboard Separation Point Suite plutons are classified as monzodiorite to granite, and are dominantly metaluminous to peraluminous, alkali-calcic to calc-alkalic, and magnesian when SiO_2 wt.% is less than ~74 wt.% and ferroan when SiO_2 wt.% is greater than ~74 wt.% (orange circles in **Fig. 2A-D**). Inboard Darran Suite samples display a broad range in SiO_2 wt.% from 49 to 69 wt.% (grey circles in **Fig. 2A-D**). Darran Suite rocks are classified as monzodiorite to granodiorite, metaluminous to peraluminous, alkali-calcic to calcic, and magnesian (**Fig. 2A-D**). In contrast to the inboard Separation Point Suite, rocks of the inboard Darran Suite have low Sr/Y values (generally <40; Tulloch and Kimbrough, 2003).

Outboard Separation Point Suite plutons are also classified as high Sr/Y but are distinguished from inboard Separation Point Suite plutonic rocks by a much more restricted range in SiO_2 (70-76 wt.%) particularly for the three large plutons shown in **Fig. 1A**. They are generally two-mica granites and are strongly peraluminous. They are also alkali-calcic to calc-alkalic, and magnesian (orange squares in **Fig. 2A-D**). Outboard Darran Suite plutonic rocks display broad range of SiO_2 wt.% from 50% to 76 wt.%. They are also classified as gabbro to granite, and they are metaluminous to peraluminous, alkali-calcic to calc-alkalic, and dominantly magnesian with several ferroan outliers at high SiO_2 values (grey squares in **Fig. 2A-D**). Like their inboard equivalent, outboard Darran Suite plutonic rocks are also low Sr/Y.

Zircon Oxygen Isotopes

Oxygen isotopes were analyzed from 34 samples collected along four ~60-km long, arc-perpendicular transects across the Median Batholith (**Fig. 1**). Samples include 10 samples from Western Fiordland Orthogneiss, 18 samples from Separation Point Suite, and 6 samples from Darran Suite. Zircon isotope data from all four transects shown in **Figures 4 and 5** include data from Decker et al. (2017) and display a strong isotope gradient with $\delta^{18}O$ (Zrn) values increasing from east to west. This gradient exists both within Darran Suite and Separation Point Suite plutons and across the Median Batholith (**Fig. 6**). Samples from Western Fiordland Orthogneiss located west of the George Sound Shear Zone have homogeneous, mantle-like values of 5.8 ± 0.3‰ (n=24; **Fig. 5A**), a feature also observed by Decker et al. (2017). Inboard Separation Point Suite and Darran Suite plutons located east of the George Sound Shear Zone and west of the Grebe Shear Zone–Indecision Creek Shear Zone show decreasing $\delta^{18}O$ values from west to east with $\delta^{18}O$ (Zrn) values decreasing from 5.9 to 4.6‰ and we observe a pronounced decrease in $\delta^{18}O$ (Zrn) values within ~10 km of the Grebe Shear Zone–Indecision Creek Shear Zone (**Fig. 5A**). Outboard Separation Point Suite and Darran Suite samples located east of the Grebe Shear Zone–Indecision Creek Shear Zone also have homogeneous values with a combined average value of 3.9 ± 0.2‰ (n=7; **Fig. 5A**). In the discussion section, we refer to $\delta^{18}O$ (Zrn) as 'mantle-like' if they fall within the average high-temperature SIMS mantle zircon value (5.3± 0.8‰). In contrast, we refer to values as low-$\delta^{18}O$ if they fall below the lower limit of the average high-temperature SIMS mantle zircon value of 4.5‰.

The geographic distribution of zircon ^{18}O values define three isotope domains: the Western Isotope Domain (WID) consisting of Western Fiordland Orthogneiss plutons with $\delta^{18}O$ (Zrn) values ranging from 5.3‰ to 6.1‰ (average = 5.8 ± 0.3‰); the Central Isotope Domain (CID) defined by inboard Separation Point Suite and Darran Suite plutons with $\delta^{18}O$ (Zrn) values increasing from 4.6‰ (east) to 5.9‰ (west); and the Eastern Isotope Domain (EID) defined by outboard Separation Point Suite and Darran Suite plutons with $\delta^{18}O$ (Zrn) values ranging from 3.7‰ to 4.1‰ (average = 3.9 ± 0.2‰). The WID and EID are defined by $\delta^{18}O$ (Zrn) values with low internal 2SD, whereas the CID is characterized by increasing $\delta^{18}O$ (Zrn) values from east to west. Geographically, the EID roughly corresponds to the outboard Median Batholith and the Drumduan Terrane, whereas the CID and WID are located within the inboard Median Batholith and the Takaka Terrane (Allibone et al., 2009a; Scott et al., 2009; Scott, 2013).

A characteristic feature of all zircons from a single hand sample is their isotope homogeneity, indicated by low intra-sample standard deviations ranging from 0.12‰ (OU75782) to 0.44‰ (**Table 2**). For all zircons sampled from Darran Suite, Separation Point Suite, and Western Fiordland Orthogneiss, the average intra-sample 2SD uncertainty is 0.28‰. The average two standard deviation uncertainty for inboard and outboard plutonic suites from this study are as follows: 0.22‰ for Western Fiordland Orthogneiss, 0.17‰ for inboard Separation Point Suite, 0.20‰ for outboard Separation Point Suite, 0.13‰ for inboard Darran Suite plutons, and 0.12‰ for outboard Darran Suite plutons. Individual zircon oxygen standard deviation values can be found in Supplemental Table 3.

Zircon Lu-Hf Isotopes

Zircon hafnium isotopes were analyzed for 19 samples and compiled with data from Decker et al. (2017) (**Figs. 4 and 5**). New analyses include two samples from Western Fiordland Orthogneiss, 13 samples from the Separation Point Suite, and 4 samples from the Darran Suite. Zircon initial ε_{Hf} data also display an isotope gradient increasing from west to east on both an intrapluton and regional scale (**Figs. 4-6**). Western Fiordland Orthogneiss plutons (WID) have nearly homogenous values of +4.2 ± 1.0 (2SD), inboard Separation Point Suite and Darran Suite plutons have an increasing west-east initial ε_{Hf} gradient of +4.0 to +5.5 (CID), and outboard Separation Point Suite and Darran Suite plutons have homogenous initial ε_{Hf} values of +7.8 ± 0.6 (EID) (**Fig. 5B**). Similar to $\delta^{18}O$ (Zrn) data above, zircon Hf isotope data also show a strong inflection within 10 km of the Grebe Shear Zone–Indecision Creek Shear Zone (**Fig. 5B**).

Isotope homogeneity is also prevalent in all samples across the Median Batholith, defined by low intra-sample standard deviations ranging from 0.5 (17NZ124A) to 2.5 ε units (17NZ140) (**Table 2**). The average 2SD precision for all zircons from inboard and outboard Darran Suite, Separation Point Suite, and Western Fiordland Orthogneiss is 1.2 ε units. From each plutonic suite, the average 2SD is as follows: 1.2 for Western Fiordland Orthogneiss, 1.2 for inboard Separation Point Suite, 1.2 for outboard Separation Point Suite, 1.6 for inboard Darran Suite, and 1.6 for outboard Darran Suite. Individual zircon standard deviation values can be found in Supplemental Table 3.

Discussion

Crustal and Isotope Architecture of the Median Batholith

Understanding the crustal and mantle structure of a Cordilleran margin is the first step in evaluating spatial and temporal isotope trends through time because the underlying crust/upper mantle plays a key role in influencing pluton chemistry. In the case of the Median Batholith, much of the pre-Mesozoic architecture has been intruded by large Triassic to Cretaceous plutons, and this makes reconstructing the crustal architecture challenging. Ambiguities in the geology of this region have led to various attempts to understand this complex region and this has produced complex and oftentimes confusing terminology (**Table 1**). Our zircon O- and Hf-isotope data shed light on this problem by revealing the presence of three distinct arc-parallel isotope domains (EID, CID, WID). This information allows us to resolve a long-standing debate about the crustal structure of the Median Batholith including the relationship between the Eastern and Western provinces and the significance of shear zones as long-lived zones of lithospheric weakness and reactivation (Klepeis et al., 2019a,b).

A key finding in our data is that zircons in both the EID and WID have uniform O- and Hf- isotope values; however, zircons in the CID are characterized by transitional isotope values that lie between EID and WID end member values (**Fig. 5A-B**). This is particularly the case for $\delta^{18}O$ (Zrn) which appears to be a sensitive indicator of isotope differences within the Median Batholith (**Fig. 5A**). The transitional isotope domain (CID) lies within the inboard Median Batholith and the Takaka Terrane (Scott et al., 2009; Scott, 2013), and shares characteristics of both the EID and WID sources. The CID has not been previously recognized as a unique component of the Median Batholith; however, our data indicate that it is isotopically distinct from other parts of the batholith to the east and west. The CID is also distinguished the presence of a strong low-$\delta^{18}O$ signal which is strongest in the east and progressively decreases to the west where it becomes non-existent in the WID (**Fig. 6A**). The E to W decrease in the low-$\delta^{18}O$ signal implies that the source of the low-$\delta^{18}O$ signature is located predominantly to the east (trenchward) and towards the boundary with accreted terranes of the Eastern Province. We propose that the low-$\delta^{18}O$ signal and the transitional isotope signature of the CID and EID can be explained by partial melting of a west-dipping, low-$\delta^{18}O$ terrane that underlies the Median Batholith. The inflection of the isotope gradient near the Grebe-Indecision Shear Zone further suggests that the low-$\delta^{18}O$ terrane is a steeply dipping and possibly listric feature suggesting a possible underthrusted relationship to the Gondwana margin. This observation is consistent with multichannel seismic images which display a thin, lower-crustal terrane that extends from the Eastern Province and continues below the Median Batholith (Davey, 2005).

Although previous workers have recognized that the Mesozoic Median Batholith has undergone a complex history of polymetamorphism and collisional/transpressional deformation, the precise timing of these events remains unclear particularly in the context of juxtaposition of the Eastern and Western province and the origin of the low-$\delta^{18}O$ source in Fiordland. Models for the juxtaposition of the Eastern and Western provinces generally fall into two tectonic scenarios: 1) Late Jurassic to Early Cretaceous collision involving the Western Province and a fringing-arc Eastern Province terrane such as the outboard Median Batholith (McCulloch et al., 1987; Kimbrough et al., 1994; Muir et al., 1995; Adams et al., 1998; Mortimer et al., 1999a; Scott et al., 2009, 2011; Scott, 2013), and/or 2) Permian collision of the Brook Street Terrane with the Western Province (Mortimer et al., 1999; McCoy-West et al., 2014). Examination of spatial-isotope data with $^{206}Pb/^{238}U$ zircon ages from Darran Suite samples in the Lake Manapouri area demonstrates that the prominent east-west isotope gradient in the CID/EID was in place by at least 160 Ma (**Fig. 7A-B**). In addition, Late Triassic plutonic rocks and older xenoliths of unknown age in northern Fiordland also record low-$\delta^{18}O$ bulk-rock values as low as -12.4‰, which also suggests that the low-$\delta^{18}O$ source in Fiordland predates the Jurassic (Blattner and Williams, 1991). Thus, we conclude that the development of the spatial-isotope gradient in Fiordland cannot be attributed to Late Jurassic or Early Cretaceous contraction and must have been produced by an earlier event/process. Our data do not permit us to directly determine the timing of amalgamation between the Eastern and Western provinces; however, a Permian amalgamation event proposed by McCoy-West et al. (2014) is consistent with our data.

The three arc-parallel isotope domains in the Median Batholith are also bounded by major Cretaceous ductile transpressional shear zones, and this observation suggests that they are in some way related to the development and/or modification of the isotope domains in Fiordland (**Figs. 1 and 6**). Previous studies have documented that these shear zones are long-lived lithospheric-scale features that were periodically reactivated during various tectonic events from the Cretaceous to the Miocene (Marcotte et al., 2005; Scott et al., 2009; Buriticá et al., 2019; Klepeis et al., 2019a,b). In particular, the Grebe–Indecision Creek Shear Zone system has been postulated to be a paleo-suture zone between the Eastern and Western provinces (Marcotte et al., 2005; Scott et al., 2009). The following features of our data support this interpretation: (1) the Grebe–Indecision Creek Shear Zone system delineates the easternmost distribution of the Takaka Terrane (and thus the Western Province) and marks a change in $\delta^{18}O$ and Hf (Zrn) isotope values between the EID and the WID/CID, ((**Fig. 1**), (2) Late Cambrian and mid-Paleozoic plutons are primarily located west of the Grebe–Indecision Creek Shear Zone system, and (4) the Grebe Shear Zone–Indecision Creek Shear Zone system extends along the entire length of the Fiordland region and parallels the Mesozoic paleo-arc axis. The significance of the George Sound Shear Zone is less understood because it does not coincide with a recognized terrane boundary at the surface and εHf (Zrn) values do not appear to show a significant change across the WID-CID boundary. In contrast, $\delta^{18}O$ (Zrn) does shows a weak decrease at the WID and CID boundary and the fact that the George Sound Shear Zone has been reactivated by Miocene thrust faults supports the notion that it may also be a structural discontinuity (Klepeis et al., 2019a,b).

Although Early Cretaceous transpression was not responsible for the formation of the spatial-isotope gradient in Fiordland, it did have an effect on the present-day spatial isotope pattern. Because transpression involves simultaneous horizontal shortening, lateral translation and, possibly, vertical extrusion, the net result is an apparent steepening of the isotope gradient in regions affected by transpression (Harland, 1971; Sanderson and Marchini, 1984; Fossen et al., 1994; Dewey et al., 1998). The effects of transpression are observed primarily at the boundary between the CID and EID where shortening and vertical extrusion functioned to shorten the isotope gradient and produce a pronounced deflection of the isotope trend where transpression/extrusion was localized (**Fig. 5**). Similar shortening of isotope ratios is observed in the Western Idaho Shear Zone, where the $^{87}Sr/^{86}Sr$ isotope gradient across the Idaho batholith was steepened by syn-magmatic transpressional movement along the shear zone (Giorgis et al., 2005).

The geologic relationships we observe in the Median Batholith are shown in a block model in **Figure 8** which illustrates present-day relationships between plutons, shear zones and their spatial-isotope geochemistry. The diagram highlights the three isotope domains that are bounded by lithospheric-scale structural features, namely the George Sound Shear Zone and the Grebe-Indecision Creek Shear Zone system. As described above, our data support the assertion presented in Scott (2013) that the Grebe Shear Zone represents a fundamental lithospheric-scale fault zone that separates the Eastern and Western provinces and accreted terranes of the Eastern Province from the ancient Gondwana margin. In this interpretation, Median Batholith plutons in the EID intruded Eastern Province crust which is only exposed at the surface as isolated metasedimentary rocks (Scott, 2013). In our interpretation of the data, the EID is underlain by a previously unrecognized, mafic low-$\delta^{18}O$ terrane which now extends beneath the EID and CID and thins to the west (continentward).

Temporal Isotope Trends in Mesozoic Magmatism

The spatial isotope zonation that we observe forms the backdrop for understanding temporal variation in arc magmatism and continental crust production in the Median Batholith. Zircon age data confirm that Mesozoic arc construction involved two distinct magmatic trends (Tulloch and Kimbrough, 2003) which include: i) a prolonged, >100 Myr period of Late Triassic to Early Cretaceous, low-Sr/Y, Darran Suite magmatism which was spatially focused within a ~20 km-wide zone centered on the Grebe Shear Zone–Indecision Creek Shear Zone in Fiordland(blue band in **Fig. 9**), and ii) a brief flare-up event, ~20 Myr in duration, consisting of high-Sr/Y magmatism (Separation Point Suite, including Western Fiordland Orthogneiss) that occurred in association with abrupt widening and continentward migration of the Early Cretaceous arc axis (green band in **Fig. 9**). The focusing of Darran Suite magmatism in the Grebe Shear Zone–Indecision Creek Shear Zone and the CID/EID boundary for >100 Myr further substantiates the notion that this zone is a major lithospheric zone of weakness that formed prior to Mesozoic magmatic activity. Moreover, the location of Darran Suite magmatism at the CID/EID boundary is also not random, but instead reflects magmatic exploitation of a deep-seated, lithospheric scale fault system. This focusing of magmatism along an arc-parallel, lithospheric scale shear zone is also observed in other Cordilleran arc systems such as the Western Idaho Shear Zone (Giorgis et al., 2005) and the southern Sierra Nevada Batholith (e.g., Saleeby et al., 2008).

In contrast to the largely static nature of the Darran Suite, the Separation Point Suite (including the Western Fiordland Orthogneiss) is characterized by sweeping migration of high-Sr/Y magmas across all three isotope domains from 129-110 Ma (**Fig. 9**). Buriticá et al. (2019) noted that during this time, the width of the Mesozoic arc reached at least 70 km in Fiordland which is twice the average width of modern arcs (Ducea et al., 2015b; 2017). This is a minimum value since the western margin of the Median Batholith is truncated by the Alpine Fault, and Cretaceous and Miocene contraction has shortened the region after magmatic emplacement. This transition from largely spatially fixed magmatism at the EID/CID boundary to abrupt continentward arc migration across the EID-WID signifies an important change in arc dynamics, which we explore later.

Figure 10 shows temporal isotope variations in the Median Batholith and illustrates several important isotope trends in our data. Temporal variations in O-isotopes show that the migration of the arc axis through time resulted in variable crustal recycling of the low-δ^{18}O source, and this recycling decreased dramatically during Separation Point Suite magmatism (**Fig. 10A**). In the latter case, zircon O-isotope data overlap the high-temperature SIMS mantle field (Valley, 2003) and we see no significant involvement of high-δ^{18}O sources like those that characterize supracrustal rocks in the region (e.g., Deep Cove Gneiss: Decker et al., 2017).

In the CID and EID, some zircons lie well below the high-temperature SIMS mantle field, and such low-δ^{18}O zircons are rare in arc-related environments. Most zircons formed in arcs typically have δ^{18}O values between 5 and 10‰ consistent with mantle influences and/or interaction with rocks altered under low-temperature conditions (<250°C) (Valley, 2003; Cavosie et al., 2011). In contrast, low-δ^{18}O zircons (<5‰) are typically restricted to extensional or rift-related environments where rock-water interactions occur at temperatures exceeding 300°C (Bindeman and Valley, 2001; Zheng et al., 2004; Blum et al., 2016). In our study, low-δ^{18}O zircons also occur over a broad region and their distribution parallels both the Mesozoic arc axis and the paleo-Pacific Gondwana margin (**Fig. 6**). Unlike low-δ^{18}O zircons formed in rift-related environments, Median Batholith zircons formed in a continental arc setting where the arc axis was either stable for long-periods of time (Darran Suite) or was advancing continentward

(Separation Point Suite). We hypothesize that low-$\delta^{18}O$ zircons in the Median Batholith formed by partial melting and recycling of a deeply buried, hydrothermally altered, low-$\delta^{18}O$ terrane which predominately underlies the EID and decreases in abundance westward beneath the CID.

Zircon Hf-isotope data provide further evidence into the nature of the crust beneath the Median Batholith and illustrate temporal changes in magma chemistry through time (**Fig. 10B**). Our data show that Darran Suite zircons have strongly positive initial ε_{Hf} (Zrn) values (>+7), though these values are significantly less radiogenic than expected for direct partial melting of Cretaceous depleted mantle (~+15: Vervoort and Blichert-Toft, 1999) or average modern island-arcs (~+13: Dhuime et al., 2011). Initial ε_{Hf} (Zrn) values significantly decreased during Separation Point Suite magmatism, similar to observations of Milan et al. (2017). Our data also show that zircon O and Hf isotopes are decoupled for Separation Point Suite rocks, whereby low positive εHf values correspond to mantle-like-$\delta^{18}O$ values. This observation makes the Separation Point Suite flare-up distinct from those observed in other low-latitude Cordilleran arcs (**Fig. 8A-C**) (Chapman et al., 2017).

To investigate magma sources and quantify the amount of crustal recycling in the Median Batholith, we conducted a series of assimilation fractional crystallization (AFC) models (**Fig. 11**). We use the same initial mantle parameters as Decker et al. (2017), and for the low-$\delta^{18}O$ source we use average bulk-rock values from 29 samples of the Largs Terrane and related xenoliths in Northern Fiordland reported in Blattner and Williams (1991). The values we use for the low-$\delta^{18}O$ source are $\delta^{18}O$ = -1.2‰, εHf = 10, and Hf concentration = 6 ppm. The average value of the low-$\delta^{18}O$ source is quite low globally compared to other arc terranes, but these values are supported by the presence of low-$\delta^{18}O$ values in Largs Terrane which extend to -12.4 and low-$\delta^{18}O$ values in the Western Fiordland Orthogneiss where Decker (2016) reported xenocrystic Paleozoic $\delta^{18}O$ zircon values as low as -7.0‰. Thus, low-$\delta^{18}O$ rocks extend beyond the footprint of the Largs Terrane in Northern Fiordland and document that very low-$\delta^{18}O$ values are present in the region.

Results from our calculations show that all EID and CID rocks can be modelled by 10-30% and 0-20% assimilation, respectively, of a low-$\delta^{18}O$ source with a similar enriched mantle-like source as observed in the WID (c.f., **Fig. 11A-C**). We propose that EID and CID are therefore hybrid melts produced by a two-stage process involving partial melting of an enriched mantle-like source followed by mixing/assimilation with partial melts of a lower-crustal arc root consisting of low-$\delta^{18}O$, hydrothermally altered mafic crust. High-heat flow in the lower crust would facilitate this second stage of melting, mixing and hybridization of the mantle-derived and crustally derived melts as postulated in the 'MASH' zone and hot zone concepts (Hildreth and Moorbath, 1988; Annen et al., 2006; Solano et al., 2012). Remelting of mafic lower-arc crust is also consistent with bulk-rock chemistry of the Separation Point Suite rocks in the EID which have high-average bulk-rock SiO_2 values (>70 wt.%), and trend toward peraluminous values. Separation Point Suite granitic rocks are dominated by high-Sr/Y values (>90) and depletions in heavy rare earth elements are consistent with involvement of garnet as a residual or fractionating phase in the lower crust. Remelting of low-$\delta^{18}O$ sediments is also a possibility, though positive initial eHf (+7) values in these rocks preclude significant sediment contributions. Direct partial melting of a less extreme low-$\delta^{18}O$ source (~3-4‰) is also possible, but this scenario overlooks the presence of strongly negative $\delta^{18}O$ values observed in the Largs Terrane or in the Western Fiordland Orthogneiss (Blattner and Williams, 1991; Decker, 2016).

The modeling results shown in **Figure 11** also imply two important features: 1) the low-$\delta^{18}O$ source is strongest in the EID, diminishes in the CID and is absent in the WID, and 2) an enriched mantle-like source with εHf values of ~+1 TO +5 was also present in all isotope domains and magmatic suites in the Mesozoic Median Batholith (i.e., both Darran and Separation Point Suites) irrespective of age or trace-element chemistry. The presence of the enriched mantle-like signal in both suites indicates a stable, and long-lived (>100 Myr), enriched mantle-like source component that was present in Mesozoic magmas. Notably, the enriched mantle signal becomes strongest in the lower crust of the WID in the Early Cretaceous (**Fig. 11A**). Below we explore implications for stable isotope domains and transient processes in the Median Batholith.

Evaluating Temporally Stable versus Transient Petrogenetic Processes

In a global study of spatial isotope trends in Cordilleran arcs, Chapman et al. (2017) noted that some arcs are characterized by a temporal persistence of consistent radiogenic isotope signatures in a given geographic region. They suggested that this temporal persistence indicates a stable petrogenetic mechanism such as long-term contamination of the melt region and/or assimilation in a lower-crustal 'MASH' or hot zone (Hildreth and Moorbath, 1988; Annen et al., 2006). In other cases where isotope signatures change through time in the same geographic region, they suggested that temporally transient processes may have been active such as relamination, forearc erosion, slab tears, and continental underthrusting. They note that temporally transient processes are distinguished by discrete excursions in temporal isotope trends resulting in melts of contrasting isotope compositions within the same geographic region (Chapman et al., 2017).

In the Median Batholith, we observe a combination of both stable and transient temporal isotope trends (**Fig. 10A-B**). Stable temporal isotope trends are best illustrated by $\delta^{18}O$ (Zrn) in the EID and in the CID. In these geographically controlled isotope domains, we observe that from ca. 170 to 120 Ma $\delta^{18}O$ (Zrn) values remained nearly unchanged for 40-50 Myr despite geochemical transitions from low-Sr/Y (Darran Suite) to high-Sr/Y (Separation Point Suite) (see 'EID and CID Isotope trend' lines in **Fig. 10A**). The consistency of $\delta^{18}O$ (Zrn) values in the EID and CID through time can be explained by a stable petrogenetic mechanism such as melting and assimilation in a lower-crustal MASH zone (Muir et al., 1995). These results are consistent with our isotope modeling which indicates that crustal recycling of mafic, low-$\delta^{18}O$ crust occurred in relatively fixed proportions over at least 50 Myr (**Fig. 10A; Fig. 11B-C**). In addition, the temporal transition from low-Sr/Y to high-Sr/Y values observed in both the EID and CID implies the involvement of a garnet-bearing source by ca. 129 Ma as suggested by previous workers (Muir et al., 1995).

Radiogenic isotope data show more complexity and evidence for both temporally stable and punctuated petrogenetic processes. As observed in oxygen-isotope data above, evidence for temporally stable petrogenetic processes is observed in the EID where initial ε_{Hf} (Zrn) also remained consistently positive (ε_{Hf} =+7 to +8) for ~40 Myr (see 'EID isotope trend' in **Fig. 10B**). In contrast, the CID displays evidence for temporal changes in isotope composition with initial ε_{Hf} (Zrn) decreasing from +8 at ca. 165 Ma to +6 to +3 Ma at ca. 129-110 Ma. These features are also observed in a compilation of bulk-rock Sr- and Nd- isotope data (Milan et al., 2017). While ε_{Hf} (Zrn) values decrease through time, $\delta^{18}O$ (Zrn) values increase from low-values to 'mantle-like' values during the terminal arc-magmatic flare-up in Early Cretaceous. The CID zircon Hf- and O-isotope data also converge on WID values where the Western Fiordland Orthogneiss represents melts in the lower crust of the Median Batholith. Isotope modeling of zircon Hf and O isotopes in this study and in Decker et al. (2017) for the Western Fiordland Orthogneiss demonstrate that there was little involvement from crustal sources, precluding significant involvement of either high-$\delta^{18}O$ metasedimentary rocks such as the Deep Cove Gneiss or low-$\delta^{18}O$ mafic crust (**Fig. 11A**). Instead, these data provide evidence for an 'enriched' mantle-like source which became increasingly volumetrically significant during the terminal Separation Point Suite flare-up starting at ca 129 Ma (see inflection in blue and red trends in **Figs. 10A and B**). The presence of the strong enriched mantle-like signal also distinguishes the Median Batholith from other Cordilleran batholiths which typically show an increase in continental crustal recycling during continentward arc migration (Chapman et al., 2017). These features support a temporally transient process in the Early Cretaceous such as the propagation of a slab tear or slab window as suggested by Decker et al. (2017) and Schwartz et al. (2017). In addition, the enriched mantle signal is also associated with the widening of the arc axis to >70 km and the abrupt change from a geographically stable magmatic arc axis in the Jurassic to rapid continentward migration in the Early Cretaceous (**Fig. 9**). Collectively, these features document a dynamic change in the arc prior to the cessation of Mesozoic magmatism at ca. 110 Ma in Fiordland (Muir et al., 1998; Tulloch and Kimbrough, 2003; Schwartz et al., 2017; Buritica et al., 2019).

Implications

Coupled zircon oxygen and hafnium isotope analyses provide a powerful tool to understand the crustal architecture of Cordilleran batholiths and to evaluate spatial and temporal arc magmatic trends. Our zircon Hf and O data show that the isotope architecture of the Median Batholith is partitioned into three isotope domains that reflect deep-seated and spatially controlled source regions that do not directly correlate with the surficial geology. Superimposed on these isotope domains, we confirm that Mesozoic magmatism involved two distinct spatio-temporal trends including: a) a prolonged, >100 Myr period of Late Triassic to Early Cretaceous (Darren Suite), low-Sr/Y magmatism spatially focused within a ~20 km-wide zone centered on the Grebe Shear Zone–Indecision Creek Shear Zone, and b) a brief, ca. 20 Myr long flare-up event, consisting of high-Sr/Y magmatism (Separation Point Suite) that occurred in association with abrupt widening and continentward migration of the Early Cretaceous arc axis. Temporal trends in stable and radiogenic zircon isotope values show evidence for both temporally stable and transient petrogenetic processes that led to the production of Mesozoic continental crust in the Median Batholith. Isotope modeling shows that arc magmatism involved significant production of new continental crust with 0-30% recycling of a low-$\delta^{18}O$ source. The terminal Early Cretaceous arc flare-up primarily involved partial melting of an enriched-mantle source from 129-110 Ma and signified the end of Mesozoic continental crust production in the Median Batholith.

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Figure Captions

Figure 1. **A.** Simplified geologic map of Fiordland, New Zealand showing sample locations (white dots). Shear zones referenced in this study are shown in yellow, and include George Sound Shear Zone, Grebe Shear Zone, and Indecision Creek Shear Zone. These latter two faults divide the inboard and outboard Median Batholith. Map is modified from Ramezani and Tulloch (2009). **B.** Inset map shows underlying basement terranes of present-day New Zealand. Dashed lines are extrapolations of terrane contacts. Line a-a' refers to cross-section in Fig. 1F. Figure adapted from Coombs et al. (1976). **C and D**. Simplified geologic map of Lake Te Anau (C) and Lake Manapouri area (D) and sample locations. **F**. Simplified reconstructed cross-section of Zealandia Cordillera prior to termination of arc magmatism. Modified after Mortimer et al. (2014). EP = Eastern Province; WP = Western Province; T. = Terrane; GSZ = Grebe Shear Zone; ICSZ = Indecision Creek Shear Zone; OB = Outboard Median Batholith; IB = Inboard Median Batholith.

Figure 2. Geochemical plots of Darran Suite, Separation Point Suite (SPS), and Western Fiordland Orthogneiss (WFO). The green field represents WFO samples from Decker et al. (2017). **A.** Classification of plutonic rocks based on SiO_2 and Na_2O+K_2O content, following Delavari et al. (2014); **B.** Shand's Index (Maniar and Piccoli, 1989); **C.** Modified alkali lime index versus SiO_2 (Frost et al., 2001); **D.** Fe-number versus SiO_2 (Frost et al., 2001).

Figure 3. Cathodoluminescence (CL) images of representative zircon and their analytical spots. Zircons display oscillatory and sector zoning consistent with magmatic growth. U-Pb spots (white circle), O spots (blue circle), and Hf spots (red circle) are shown where analysis occurred.

For each sample, a white 100-micron bar is shown for scale. Data for all spots can be found in Supplementary Tables 2-3.

Figure 4. Individual transect isotope data. $\delta^{18}O$ and initial ε_{Hf} zircon data from each arc-perpendicular transect are plotted against distance from the Grebe Shear Zone (GSZ)–Indecision Creek Shear Zone (ICSZ) and extension thereof (ext.). The western yellow bar marks the George Sound Shear Zone (GSSZ) and southern projection (proj.). Shear zone thicknesses are demarcated by thickness of the yellow bars. Transect locations are outlined on **Figure 1**. WID = Western Isotope Domain; CID = Central Isotope Domain; EID = Eastern Isotope Domain.

Figure 5. Combined isotope data relative to the Grebe Shear Zone–Indecision Creek Shear Zone. Yellow vertical bars show thicknesses of shear zone widths. **A.** $\delta^{18}O$ (Zrn) data vs. distance from the Grebe Shear Zone (GSZ). **B.** Initial ε_{Hf} (Zrn) data vs distance from the GSZ. WID = Western Isotope Domain; CID = Central Isotope Domain; EID = Eastern Isotope Domain.

Figure 6. Isotope Contour Plots for the Median Batholith. Sample locations are plotted as black dots. Ductile shear zones, outlined in a dark grey field, separate the three isotope domains. The George Sound Shear Zone and southern projection separate the WID and CID, and the Grebe Shear Zone–Indecision Creek Shear Zone separate the CID from the EID. Sample transects from **Figure 1** are outlined in black lines with associated transect letters. Oxygen isotope values increase from east to west (left), and hafnium isotope values increase from west to east (right). GSSZ = George Sound Shear Zone; GSSZ Proj. = Southern projection of the George Sound

Shear Zone; GSZ = Grebe Shear Zone; ICSZ = Indecision Creek Shear Zone; WID = Western Isotope Domain; CID = Central Isotope Domain; EID = Eastern Isotope Domain.

Figure 7. A. Zircon O-isotope contour plot in Lake Te Anau and Lake Manapouri showing pre-Cretaceous zircon sample locations and $^{206}Pb/^{238}U$ zircon ages. **B.** Zircon age vs. $\delta^{18}O$ (Zrn) for pre-Cretaceous samples. Existence of the isotope gradient in pre-Cretaceous samples indicates that the isotope in both the EID and CID gradient pre-dates Early Cretaceous transpression and contraction in the region and was therefore not caused by an Early Cretaceous tectonic event.

Figure 8. A. Block model of present-day locations of plutons from the Median Batholith and crustal architecture of the Eastern and Western provinces at depth. The George Sound Shear Zone separates the WID and CID, and demarcates the boundary of the underlying Eastern Province terrane. The Grebe Shear Zone – Indecision Creek Shear Zone separates the CID and EID, divides the inboard and outboard Median Batholith, and is the suture that separates the Eastern and Western Provinces. Graphs of oxygen (**B**) and hafnium (**C**) isotope trends are plotted against distance. WAL= West Arm Leucogranite, RO= Refrigerator Orthogneiss, Ptk=Puteketeke Pluton, ICC=Indecision Creek Complex, NF=North Fork Pluton, SZ=shear zone.

Figure 9. A. Zircon age vs. distance from GSZ (km). Low-Sr/Y, Darran Suite magmatism was focused within a 10-15 km zone relative to the Grebe Shear Zone–Indecision Creek Shear Zone system for >100 Myr from ca. 240 to 130 Ma. Emplacement of high-Sr/Y, Separation Point Suite

magmas is shown to have migrated continentward from 129-114 Ma at a rate of ~4-5 km/Myr. Distance from the GSZ (km) is measured from location shown in **Figure 1**.

Figure 10. A. Zircon oxygen-isotope values vs. zircon age (Ma). B. Zircon initial εHf values vs. zircon age (Ma).

Figure 11. Assimilation Fractional Crystallization (AFC) models for the WID (A), CID (B) and EID (C). Data illustrate increasing influence of low-$\delta^{18}O$ crust from west to east (A to C).

List of Tables

Appendix Files

Table 1. Terminology and Subdivisions of the Median Batholith

W GSSZ GSZ/ICSZ E

Geographic Subdivision	Western Fiordland	Central Fiordland	Eastern Fiordland
Batholith Subdivision[1]	Inboard Median Batholith	Inboard Median Batholith	Outboard Median Batholith
Terrane Subdivision[2]	Takaka Terrane	Takaka Terrane	Drumduan Terrane
Spatial Isotopic Domain[3]	Western (WID)	Central (CID)	Eastern (EID)
Defining Plutons	Separation Point Suite -Western Fiordland Orthogneiss Darran Suite -None recognized	Separation Point Suite -Puteketeke Pluton -Refrigerator Orthogneiss -West Arm Leucogranite Darran Suite -Hunter Intrusives -Murchison Intrusives	Separation Point Suite -Takahe Granodiorite -Titiroa Granodiorite -North Fiord Granite Darran Suite -Hunter Intrusives -Murchison Intrusives

Table 2. Summary of zircon error-weighted average $^{206}Pb/^{238}U$ ages, O and Lu-Hf isotope data, Fiordland, New Zealand

Pluton	Sample Number	Age	$\delta^{18}O$ (Zrn) Mean ± 2SD	δ18O (Zrn) Range (‰)	n	Zrn Initial εHf ± 2SD	Zrn Initial εHf range	n	Reference
Pembroke diorite	05NZ12P	134.2 ± 2.9	4.45 ± 0.28	4.2 - 4.5	8	8.2 ± 3.1	10.3 - 6.5	20	3
Malaspina Pluton	12DC41C	115.9 ± 1.2	5.74 ± 0.19	5.6 - 5.9	8	n.d.	n.d.	n.d.	1
Resolution Orthogneiss	12NZ12B	115.1 ± 2.1	5.85 ± 0.25	5.3 - 6.1	7	4.0 ± 3.2	7.9 - 2.2	20	2
Misty Pluton	12NZ22A	114.7 ± 1.7	5.68 ± 0.17	5.5 - 5.8	7	4.7 ± 3.4	10.8 - 1.3	20	2
Misty Pluton	12NZ24	115.8 ± 1.9	5.75 ± 0.12	5.6 - 6.0	6	3.9 ± 3.3	6.0 - 2.9	20	2
Misty Pluton	12NZ33	114.5 ± 2.3	5.56 ± 0.23	5.4 - 5.6	8	4 ± 3.6	5.7 - 2.0	20	2
Misty Pluton	12NZ36B	117.1 ± 2.4	5.78 ± 0.20	5.5 - 5.9	5	3.9 ± 3.4	4.9 - 2.8	20	2
Malaspina Pluton	13NZ11	n.d.	5.80 ± 0.35	6.0 - 5.6	5	n.d.	n.d.	n.d.	1
Malaspina Pluton	13NZ14	n.d.	5.82 ± 0.17	5.7 - 5.9	6	n.d.	n.d.	n.d.	1
Malaspina Pluton	13NZ16B	118.0 ± 1.7	5.74 ± 0.27	5.5 - 5.9	7	4.2 ± 3.1	5.6 - 2.1	20	2
Malaspina Pluton	13NZ22	116.9 ± 1.8	5.67 ± 0.37	5.5 - 5.9	5	4.2 ± 3.4	6.2 - 3.0	17	2
Breaksea Orthogneiss	13NZ33E	123.2 ± 1.3	5.30 ± 0.23	5.2 - 5.4	6	4.8 ± 3.5	6.5 - 2.7	20	2
Malaspina Pluton	13NZ34A	118.0 ± 1.6	5.74 ± 0.39	5.5 - 6.0	7	2.9 ± 3.3	4.6 - 1.2	20	2
Malaspina Pluton	13NZ40D1	116.4 ± 1.6	5.74 ± 0.37	5.4 - 5.9	9	3.6 ± 3.3	5.3 - 1.9	17	2
Misty Pluton	13NZ46	116.9 ± 1.8	5.87 ± 0.17	5.6 - 6.0	8	4.4 ± 3.1	11.2 - 2.6	20	2,4
Misty Pluton	13NZ52A	116.8 ± 2.1	6.05 ± 0.38	5.8 - 6.2	5	3.9 ± 2.9	5.4 - 2.5	20	2
Misty Pluton	13NZ55A	115.2 ± 1.9	5.87 ± 0.40	5.7 - 6.1	7	4.4 ± 2.9	7.7 - 2.1	20	2
Misty Pluton	13NZ58	116.8 ± 2.1	6.06 ± 0.27	5.7 - 6.2	7	4.3 ± 2.9	5.4 - 1.4	20	2
Malaspina Pluton	13NZ59	117.5 ± 1.0	5.75 ± 0.27	5.7 - 5.8	6	4.3 ± 3.1	6.5 - 1.9	20	2
Malaspina Pluton	14NZ82	n.d.	5.72 ± 0.17	5.9 - 5.7	6	n.d.	n.d.	n.d.	1
Worsley Pluton	15NZ02	120.3 ± 1.2	5.46 ± 0.38	5.2 - 5.6	8	4.9 ± 3.3	6.9 - 3.5	20	2
Eastern McKerr Intrusives	15NZ12	128.3 ± 2	-5.83 ± 0.30	(-)7.4 - 0.4	8	2.6 ± 3.1	5.2 - (-)0.2	17	2,3
Eastern McKerr Intrusives	15NZ20	118.8 ± 1.7	5.77 ± 0.27	5.5 - 6.0	6	3.8 ± 3.1	5.7 - -2.0	20	2
Worsley Pluton	15NZ27	123.4 ± 1.3	5.95 ± 0.45	5.6 - 6.2	10	5.0 ± 3.1	6.1 - 2.6	20	2
Devils Armchair Pluton	15NZ66	133.4 ± 2.1	4.63 ± 0.35	4.0 - 5.3	8	5.4 ± 3.3	8.5 - (-)4.9	19	3
Puteketeke Pluton	17NZ100	122.6 ± 1.4	4.98 ± 0.28	5.2 - 4.8	8	4.3 ± 1.4	5.2 - 3.2	10	1
Puteketeke Pluton	17NZ104	n.d.	4.88 ± 0.30	5.1 - 4.7	6	5.23 ± 2.0	7.0 - 3.8	8	1
West Arm Leucogranite	17NZ114	123.9 ± 1.5	5.39 ± 0.29	5.6 - 5.2	8	4.3 ± 0.6	4.7 - 3.8	7	1
West Arm Leucogranite	17NZ117	113.3 ± 1.6	5.88 ± 0.14	5.9 - 5.8	4	3.0 ± 8.4	8.8 - (-)0.8	4	1
Refrigerator Orthogneiss	17NZ12	128.1 ± 2.3	4.67 ± 0.22	4.8 - 4.5	6	5.5 ± 1.6	6.6 - 4.2	9	1
Refrigerator Orthogneiss	17NZ124A	127.4 ± 1.3	5.10 ± 0.26	5.2 - 4.9	7	3.7 ± 0.5	3.9 - 3.4	7	1
West Arm Leucogranite	17NZ134	122.5 ± 1.8	5.26 ± 0.39	5.5 - 5.0	4	n.d.	n.d.	n.d.	1
West Arm Leucogranite	17NZ140	116.9 ± 1.8	5.68 ± 0.38	6.0 - 5.5	5	4.1 ± 2.5	5.5 - 0.1	11	1
Refrigerator Orthogneiss	17NZ25A	129.1 ± 1.8	4.74 ± 0.25	4.6 - 4.9	6	5.9 ± 1.0	6.8 - 5.5	8	1
Murchinson Intrusives	17NZ31A	133.5 ± 2.8	4.81 ± 0.44	5.0 - 4.4	5	n.d.	n.d.	n.d.	1
Murchinson Intrusives	17NZ40	170.0 ± 1.5	5.13 ± 0.24	5.0 - 5.3	10	7.9 ± 0.8	8.3 - 7.4	8	1
Hunter Intrusives	17NZ46	351.5 ± 9.9	4.83 ± 0.27	5.0 - 4.7	7	n.d.	n.d.	n.d.	1
Hunter Intrusives	17NZ65A	156.0 ± 1.7	4.18 ± 0.12	4.2 - 4.1	7	9.6 ± 1.2	10.5 - 8.2	10	1
Hunter Intrusives	17NZ72A	143.4 ± 4.4	3.81 ± 0.44	4.2 - 3.6	5	7.5 ± 1.3	8.4 - 6.4	6	1
Takahe Granodiorite	17NZ78B	125.1 ± 1.7	3.86 ± 0.21	4.0 - 3.8	6	7.6 ± 0.8	8.4 - 7.2	8	1
Takahe Granodiorite	17NZ81	123.1 ± 1.8	3.74 ± 0.26	4.0 - 3.6	7	7.1 ± 1.7	8.3 - 5.9	9	1
Hunter Intrusives	17NZ87	156.0 ± 1.7	4.80 ± 0.18	5.0 - 4.7	7	7.9 ± 1.7	9.0 - 6.7	12	1
Puteketeke Pluton	17NZ98	122.0 ± 1.6	4.79 ± 0.23	5.0 - 4.7	6	4.4 ± 1.3	5.6 - 3.6	10	1
Puteketeke Pluton	OU75705	120.8 ± 0.9	4.89 ± 0.30	5.1 - 4.7	10	4.5 ± 1.2	5.3 - 3.8	8	1
Refrigerator Orthogneiss	OU75782	120.7 ± 1.1	5.31 ± 0.12	5.4 - 5.2	6	n.d.	n.d.	n.d.	1
Hunter Intrusives	OU49100	169.0 ± 3.1	4.91 ± 0.22	4.1 - 5.2	12	7.9 ± 3.3	11.1 - 5.2	20	3
Hunter Intrusives	OU49102	149.6 ± 1.6	4.36 ± 0.30	4.2 - 4.5	8	7.8 ± 3.3	10.1 - 3.7	20	3
Pomona Island granite	OU49120	163.1 ± 1.8	3.90 ± 0.19	3.6 - 4.1	10	11.2 ± 3.4	17.9 - 5.7	20	3
Darran Leucogranite	OU49127	135.8 ± 2.3	3.97 ± 0.32	3.7 - 4.0	9	8.4 ± 3.2	9.4 - 6.0	20	3
Nurse Suite	OU49128	140.8 ± 1.6	3.80 ± 0.28	3.4 - 4.1	6	n.d.	n.d.	n.d.	3
Glade Suite	OU49129	140.6 ± 1.5	4.06 ± 0.30	3.8 - 4.1	7	7.9 ± 3.1	10.0 - 5.9	20	3
Worsley Pluton	OU49144	119.3 ± 1.8	4.57 ± 0.17	4.8 - 4.5	6	5.5 ± 1.3	6.1 - 4.6	7	1
Titiroa Pluton	P69040	n.d.	3.83 ± 0.35	4.0 - 3.6	8	8.0 ± 0.8	8.5 - 7.5	8	1
Puteketeke Pluton	P73900	n.d.	4.67 ± 0.15	4.8 - 4.6	7	4.4 ± 0.7	4.9 - 3.9	9	1
Omaki Orthogneiss	P75785	124.91 ± 0.17	4.75 ± 0.2	4.9 - 4.5	7	4.3 ± 0.6	4.7 - 3.9	7	1
Misty Pluton	P76640	117.89 ± 0.13	5.73 ± 0.19	5.9 - 5.6	8	n.d.	n.d.	n.d.	1
Misty Pluton	P76705	n.d.	5.72 ± 0.22	5.9 - 5.6	6	4.4 ± 1.2	5.2 - 3.7	7	1
Misty Pluton	P76709	ca. 116.4	5.89 ± 0.20	6.0 - 5.8	7	n.d.	n.d.	n.d.	1
Misty Pluton	P77630	118.42 ± 0.06	5.30 ± 0.32	5.2 - 5.6	5	n.d.	n.d.	n.d.	1,4
Misty Pluton	P77844	n.d.	4.76 ± 0.23	4.6 - 4.9	6	n.d.	n.d.	n.d.	1

[1]this study

[2]Decker et al. (2017)

[3]Decker (2016)

[4]Tulloch, personal communication, 2019

[5]Ringwood and Schwartz. Unpublished data

[6]Buritica et al. (2019)

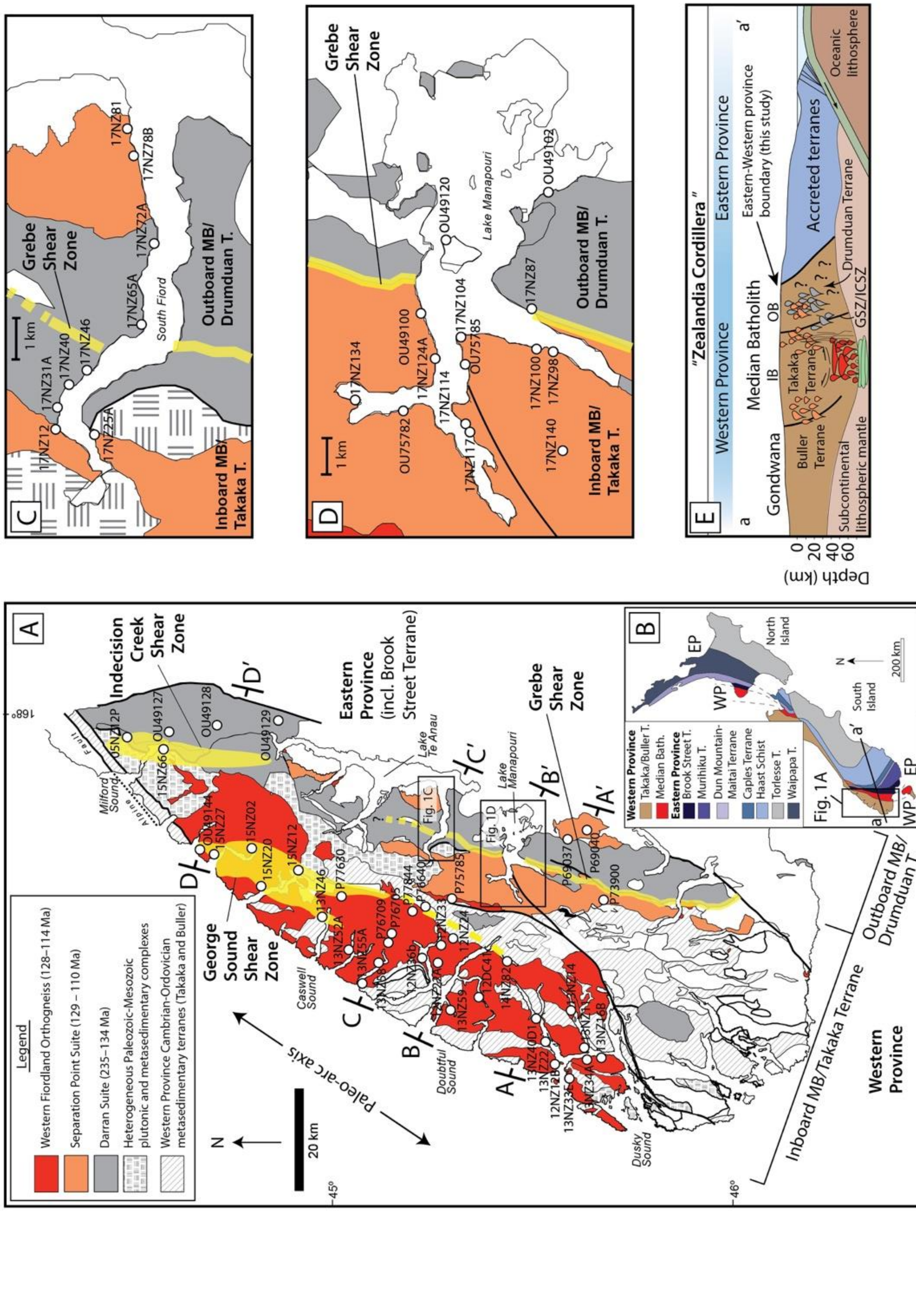

A
Legend
Western Fiordland Orthogneiss (128–114 Ma)
Separation Point Suite (129 – 110 Ma)
Darran Suite (235–134 Ma)
Heterogeneous Paleozoic-Mesozoic plutonic and metasedimentary complexes
Western Province Cambrian-Ordovician metasedimentary terranes (Takaka and Buller)
N
20 km
Paleo-arc axis
George Sound Shear Zone
Indecision Creek Shear Zone
Grebe Shear Zone
Eastern Province (incl. Brook Street Terrane)
Western Province
Inboard MB/Takaka Terrane
Outboard MB/ Drumduan T.
Milford Sound
Alpine Fault
Caswell Sound
Doubtful Sound
Dusky Sound
Lake Te Anau
Lake Manapouri
Fig. 1C
Fig. 1D
-45°
-46°
168°
B
Western Province
Takaka/Buller T.
Median Bath.
Eastern Province
Brook Street T.
Murihiku T.
Dun Mountain-Maitai Terrane
Caples Terrane
Haast Schist
Torlesse T.
Waipapa T.
Fig. 1A
WP
EP
North Island
South Island
200 km
C
1 km
Grebe Shear Zone
South Fiord
Inboard MB/ Takaka T.
Outboard MB/ Drumduan T.
D
1 km
Grebe Shear Zone
Lake Manapouri
Inboard MB/ Takaka T.
Outboard MB/ Drumduan T.
E
"Zealandia Cordillera"
Western Province
Eastern Province
Gondwana
Median Batholith
IB
OB
Eastern-Western province boundary (this study)
Buller Terrane
Takaka Terrane
GSZ/ICSZ
Drumduan Terrane
Accreted terranes
Subcontinental lithospheric mantle
Oceanic lithosphere
Depth (km)

Figure 1

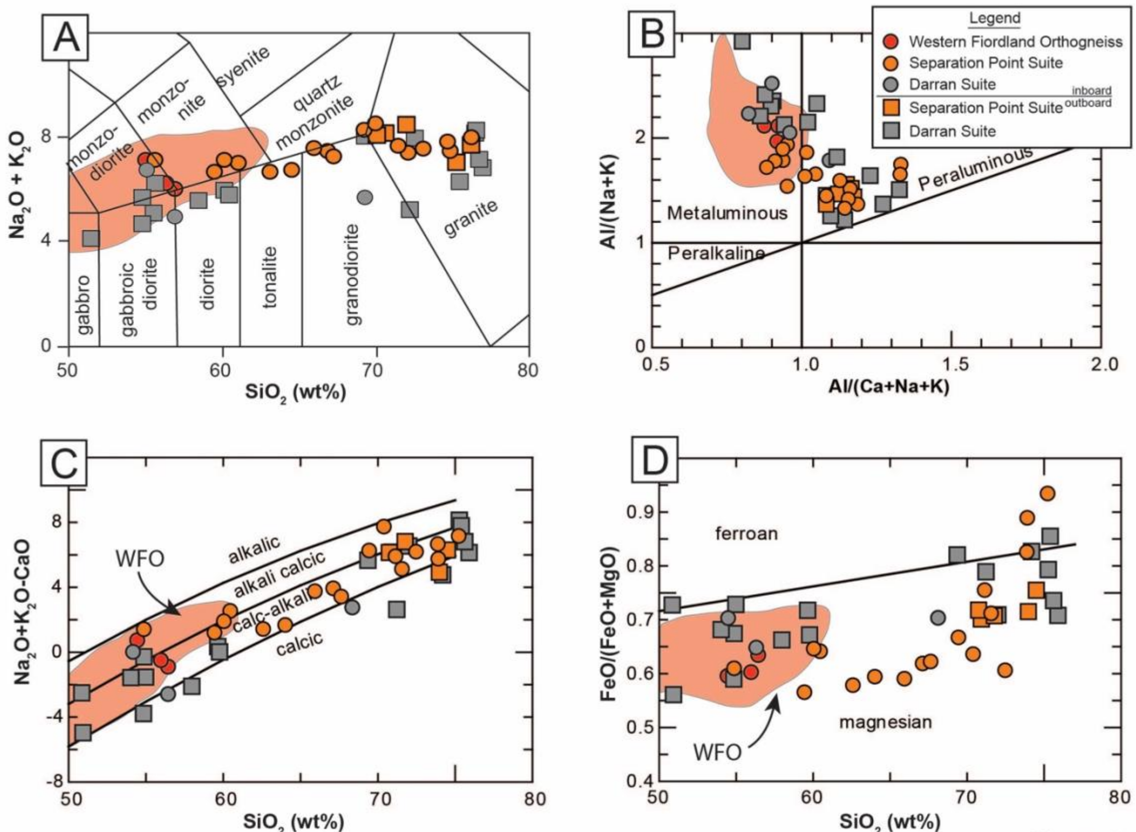
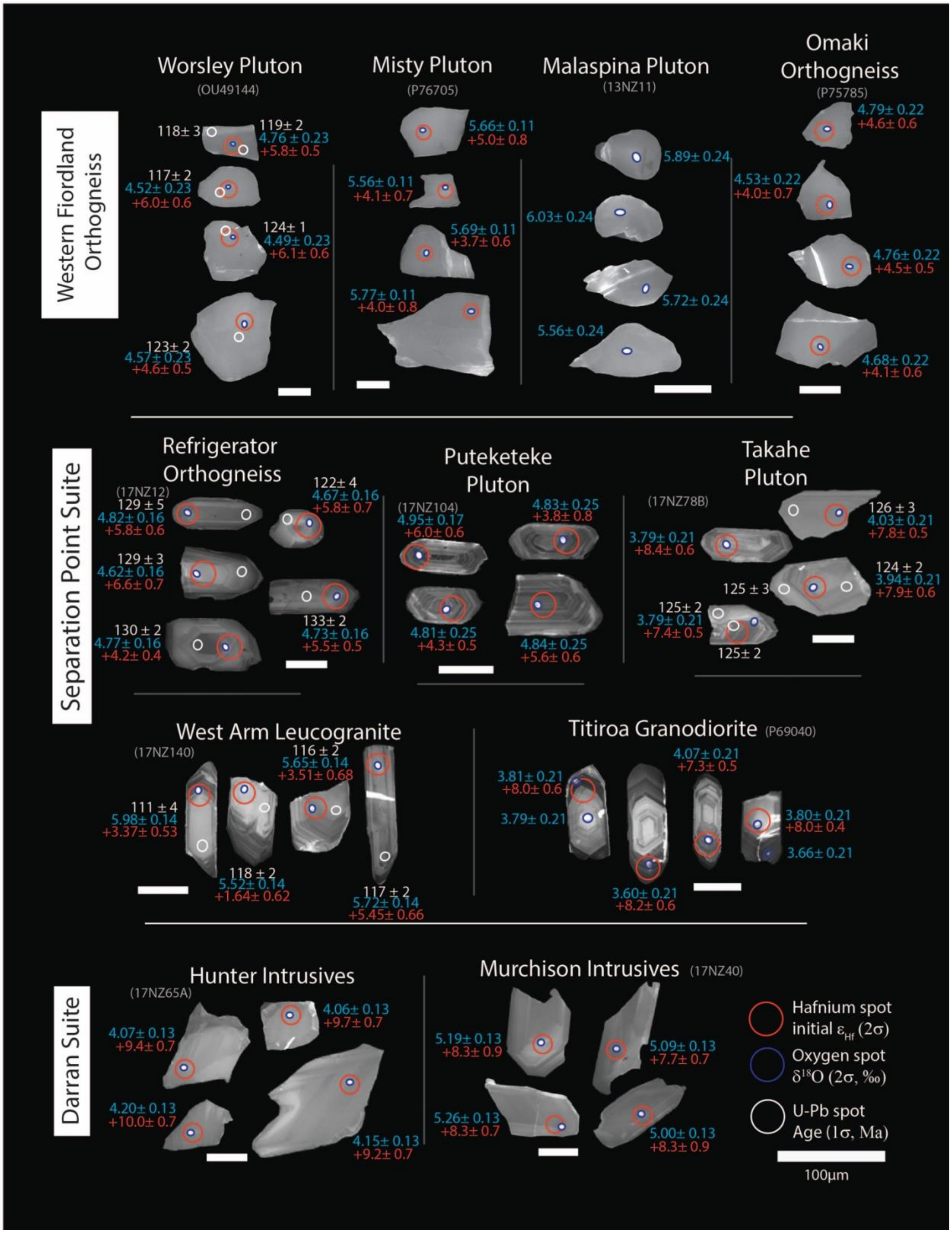

A
Na2O + K2O
SiO2 (wt%)
monzo-diorite
monzo-nite
syenite
quartz monzonite
gabbro
gabbroic diorite
diorite
tonalite
granodiorite
granite
B
Legend
Western Fiordland Orthogneiss
Separation Point Suite
Darran Suite
inboard
Separation Point Suite
outboard
Darran Suite
Al/(Na+K)
Al/(Ca+Na+K)
Metaluminous
Peralkaline
Peraluminous
C
Na2O+K2O-CaO
SiO2 (wt%)
WFO
alkalic
alkali calcic
calc-alkali
calcic
D
FeO/(FeO+MgO)
SiO2 (wt%)
ferroan
magnesian
WFO

Figure 2



Western Fiordland Orthogneiss
Worsley Pluton
(OU49144)
Misty Pluton
(P76705)
Malaspina Pluton
(13NZ11)
Omaki Orthogneiss
(P75785)
Separation Point Suite
Refrigerator Orthogneiss
(17NZ12)
Puteketeke Pluton
(17NZ104)
Takahe Pluton
(17NZ78B)
West Arm Leucogranite
(17NZ140)
Titiroa Granodiorite (P69040)
Darran Suite
Hunter Intrusives
(17NZ65A)
Murchison Intrusives (17NZ40)
Hafnium spot initial εHf (2σ)
Oxygen spot δ18O (2σ, ‰)
U-Pb spot Age (1σ, Ma)
100μm

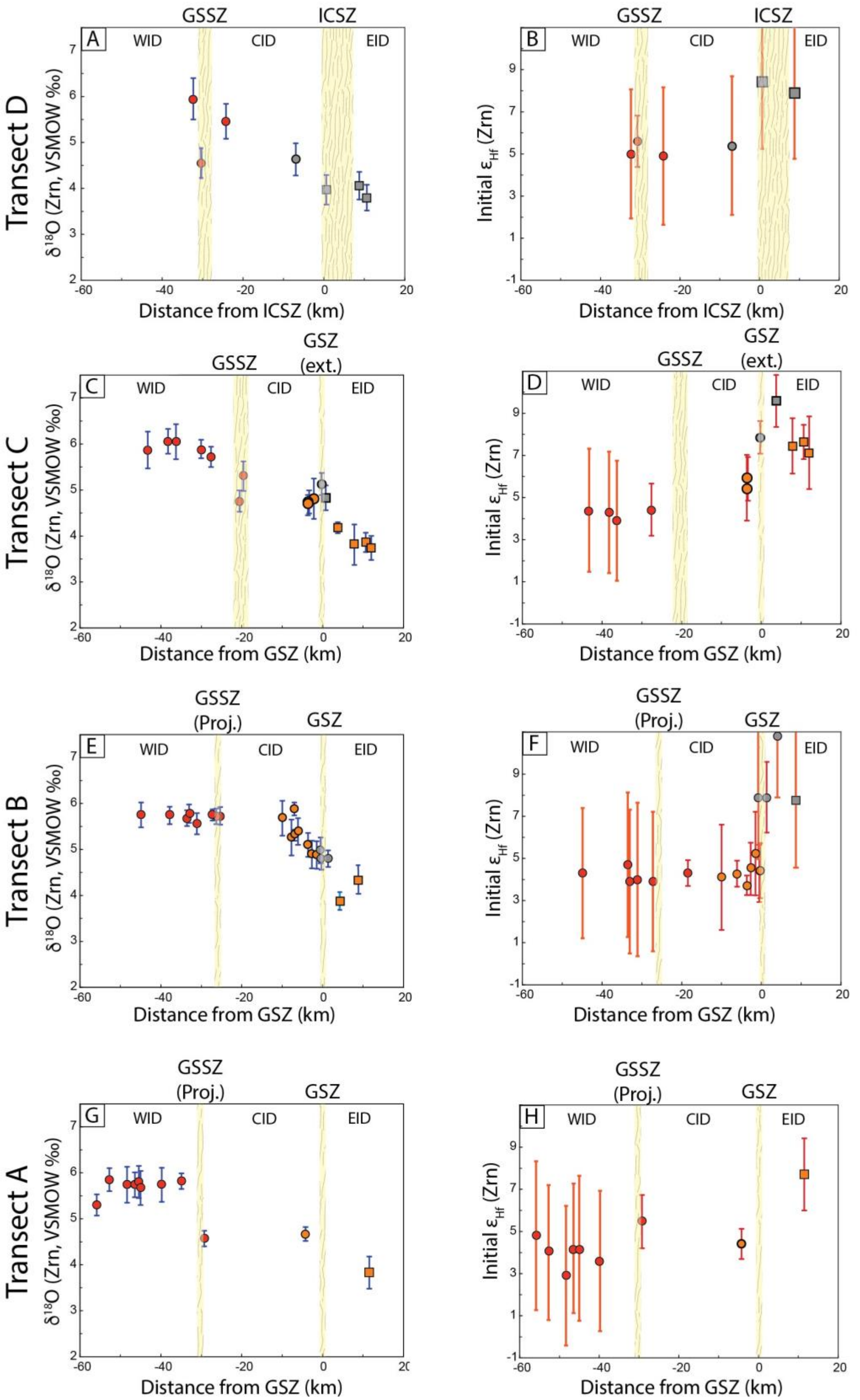

Transect D
Transect C
Transect B
Transect A
GSSZ
ICSZ
GSZ (ext.)
GSSZ (Proj.)
GSZ
A
B
C
D
E
F
G
H
WID
CID
EID
δ18O (Zrn, VSMOW ‰)
Initial εHf (Zrn)
Distance from ICSZ (km)
Distance from GSZ (km)

Figure 4

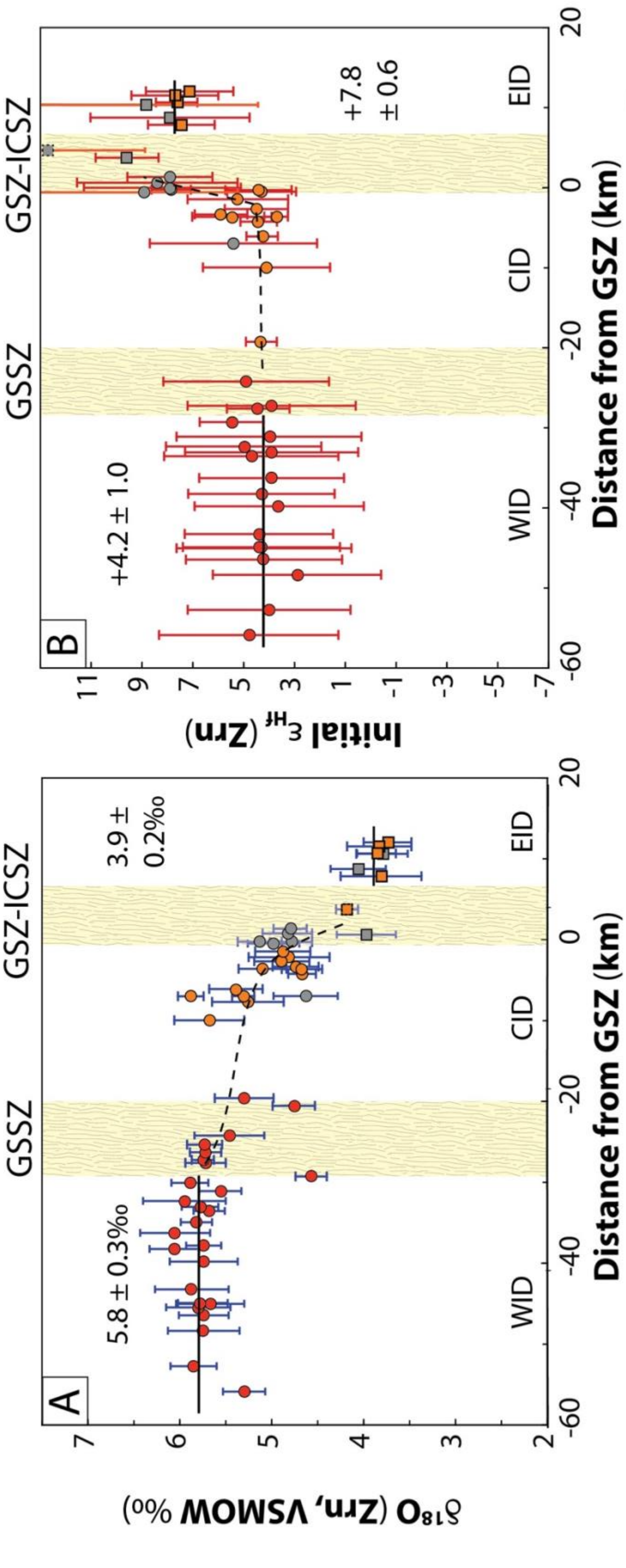
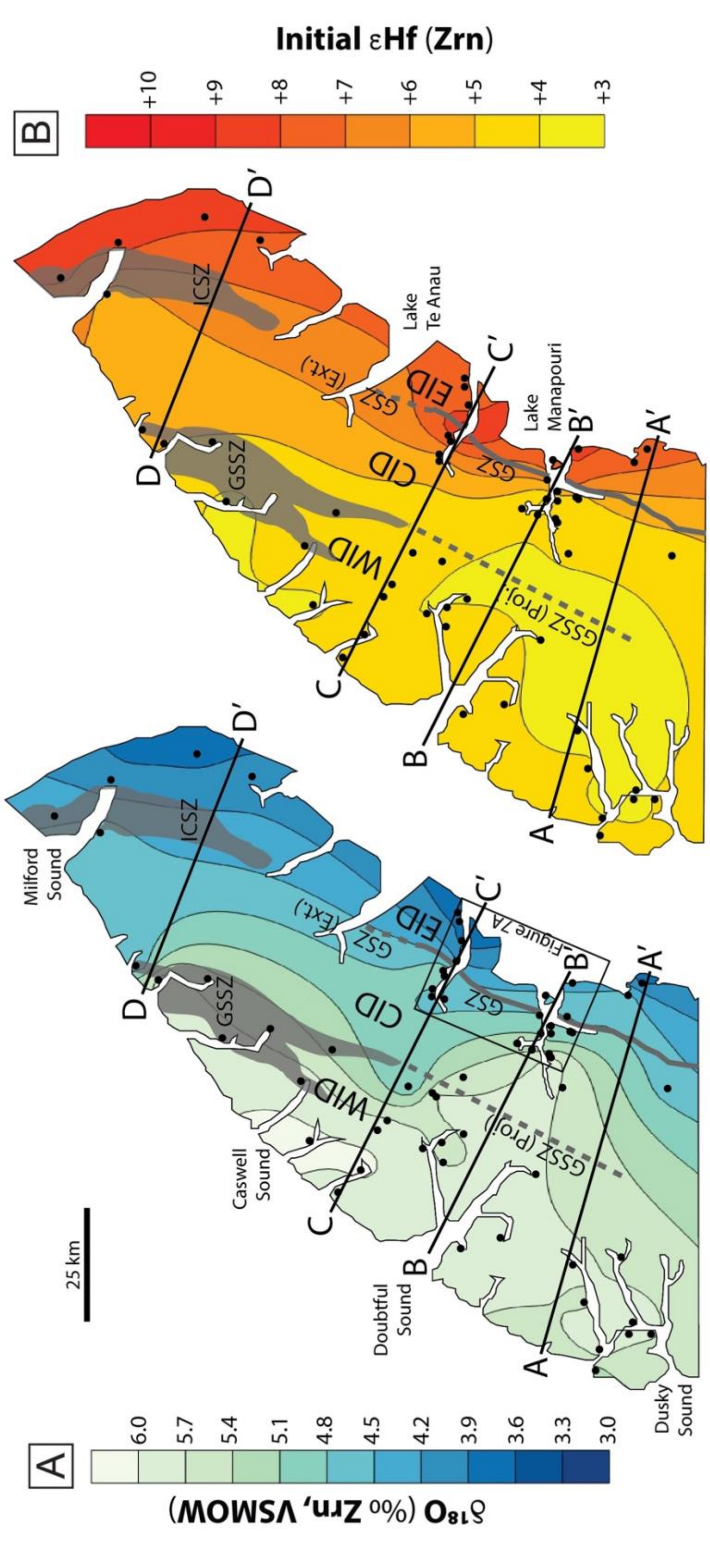

A
B
GSSZ
GSZ-ICSZ
WID
CID
EID
5.8 ± 0.3‰
3.9 ± 0.2‰
+4.2 ± 1.0
+7.8 ± 0.6
δ18O (Zrn, VSMOW ‰)
Initial εHf (Zrn)
Distance from GSZ (km)

Figure 5



A
δ18O (‰ Zrn, VSMOW)
6.0
5.7
5.4
5.1
4.8
4.5
4.2
3.9
3.6
3.3
3.0
25 km
Milford Sound
Caswell Sound
Doubtful Sound
Dusky Sound
ICSZ
GSSZ
GSZ (Ext.)
EID
CID
WID
GSZ
GSSZ (Proj.)
Figure 7A
A
A′
B
B′
C
C′
D
D′
B
Initial εHf (Zrn)
+10
+9
+8
+7
+6
+5
+4
+3
Lake Te Anau
Lake Manapouri

Figure 6

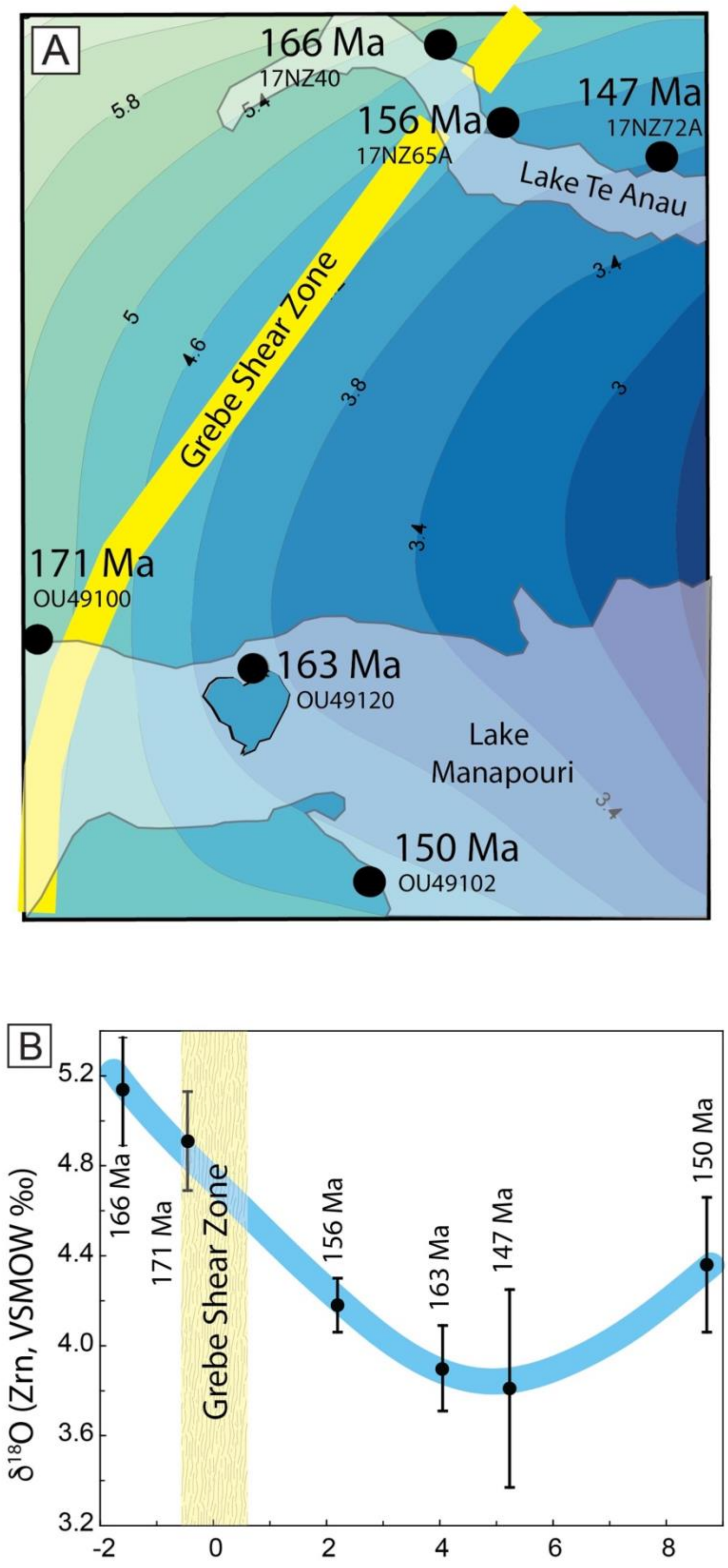

A
166 Ma
17NZ40
5.8
5.4
156 Ma
17NZ65A
147 Ma
17NZ72A
Lake Te Anau
Grebe Shear Zone
5
4.6
3.8
3.4
3
171 Ma
OU49100
163 Ma
OU49120
Lake
Manapouri
150 Ma
OU49102
B
5.2
4.8
4.4
4.0
3.6
3.2
δ18O (Zrn, VSMOW ‰)
166 Ma
171 Ma
156 Ma
163 Ma
147 Ma
150 Ma
-2
0
2
4
6
8
Distance from GSZ (km)

Figure 7

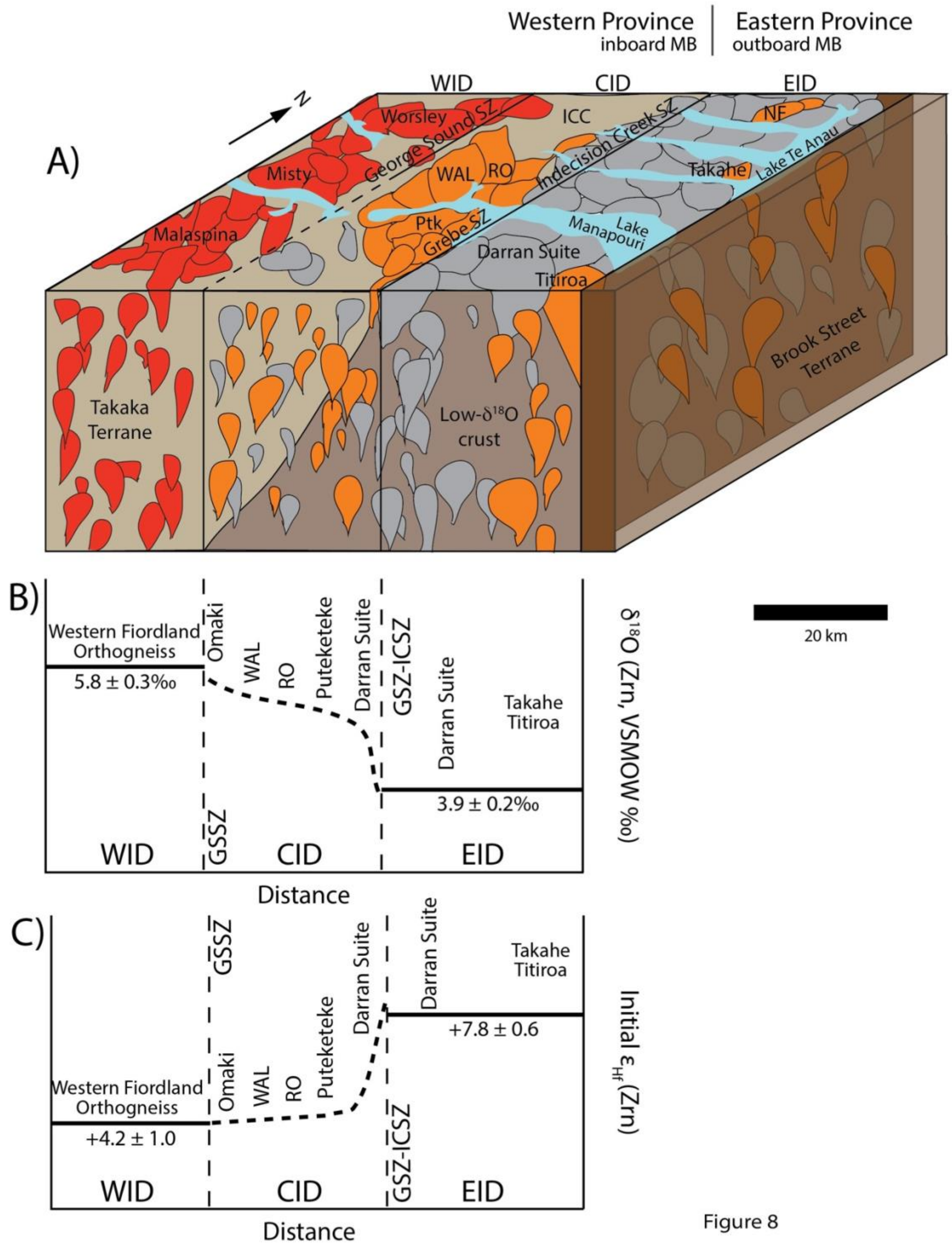

Western Province
inboard MB
Eastern Province
outboard MB
WID
CID
EID
N
A)
Worsley
George Sound SZ
ICC
Indecision Creek SZ
NF
Lake Te Anau
Misty
WAL
RO
Takahe
Malaspina
Ptk
Grebe SZ
Lake Manapouri
Darran Suite
Titiroa
Takaka Terrane
Low-δ18O crust
Brook Street Terrane
B)
Western Fiordland Orthogneiss
5.8 ± 0.3‰
Omaki
WAL
RO
Puteketeke
Darran Suite
GSZ-ICSZ
Darran Suite
Takahe
Titiroa
3.9 ± 0.2‰
GSSZ
WID
CID
EID
Distance
δ18O (Zrn, VSMOW ‰)
20 km
C)
GSSZ
Darran Suite
Darran Suite
Takahe
Titiroa
+7.8 ± 0.6
Western Fiordland Orthogneiss
+4.2 ± 1.0
Omaki
WAL
RO
Puteketeke
GSZ-ICSZ
WID
CID
EID
Distance
Initial εHf (Zrn)

Figure 8

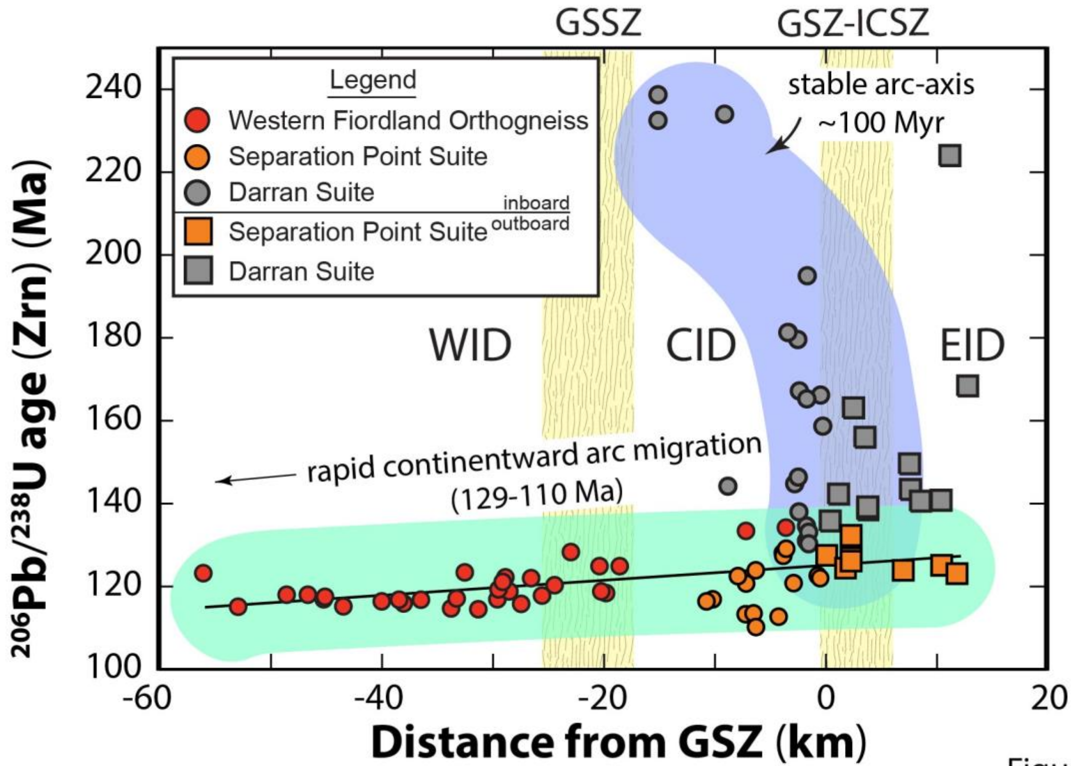

GSSZ
GSZ-ICSZ
Legend
Western Fiordland Orthogneiss
Separation Point Suite
Darran Suite
inboard
outboard
Separation Point Suite
Darran Suite
stable arc-axis
~100 Myr
WID
CID
EID
rapid continentward arc migration
(129-110 Ma)
206Pb/238U age (Zrn) (Ma)
240
220
200
180
160
140
120
100
-60
-40
-20
0
20
Distance from GSZ (km)

Figure 9

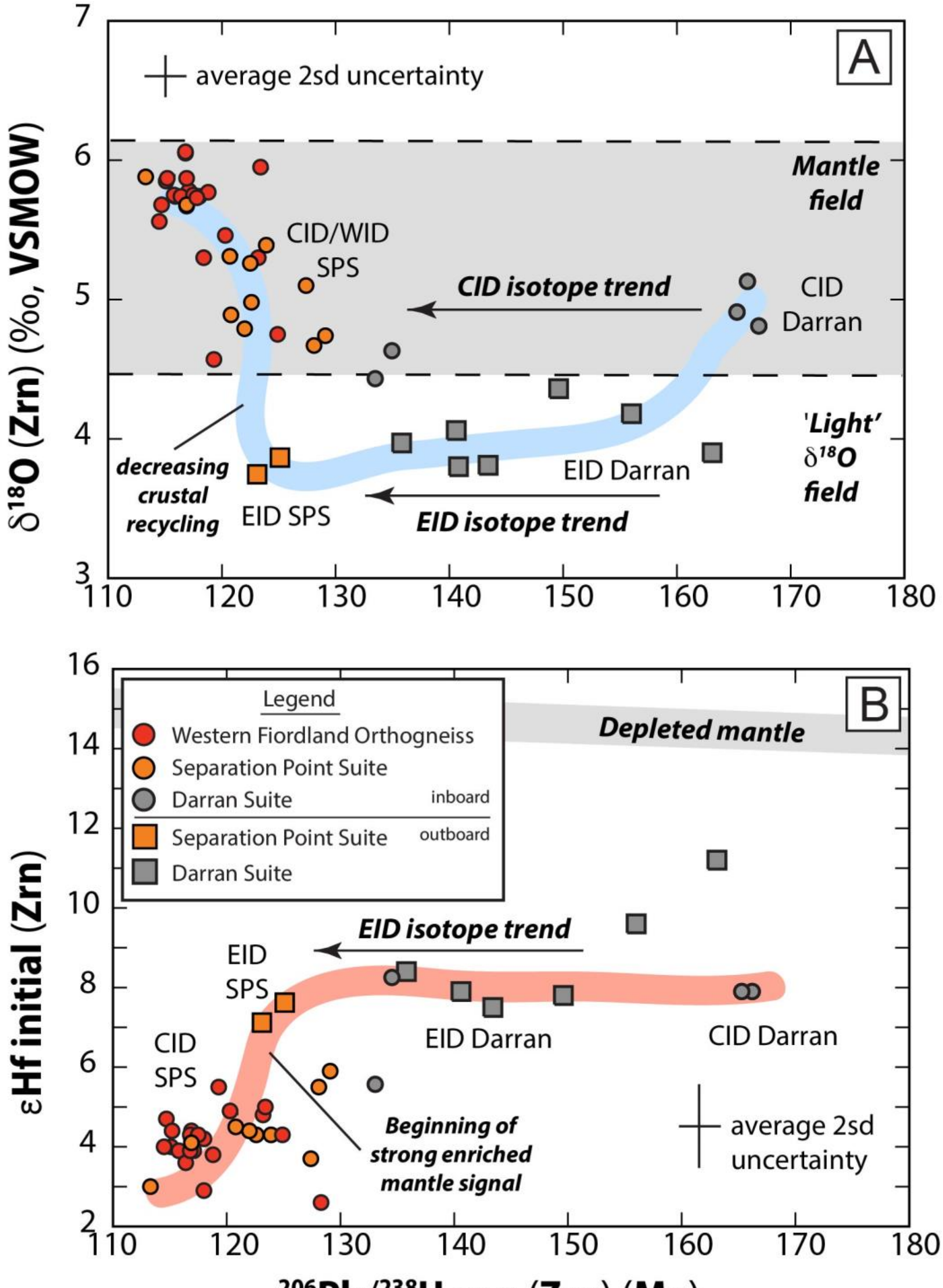

A
average 2sd uncertainty
Mantle field
CID/WID SPS
CID isotope trend
CID Darran
'Light' δ18O field
decreasing crustal recycling
EID SPS
EID isotope trend
EID Darran
δ18O (Zrn) (‰, VSMOW)
B
Legend
Western Fiordland Orthogneiss
Separation Point Suite
Darran Suite
inboard
Separation Point Suite
outboard
Darran Suite
Depleted mantle
EID isotope trend
EID SPS
CID SPS
EID Darran
CID Darran
Beginning of strong enriched mantle signal
average 2sd uncertainty
εHf initial (Zrn)
206Pb/238U age (Zrn) (Ma)

Figure 10

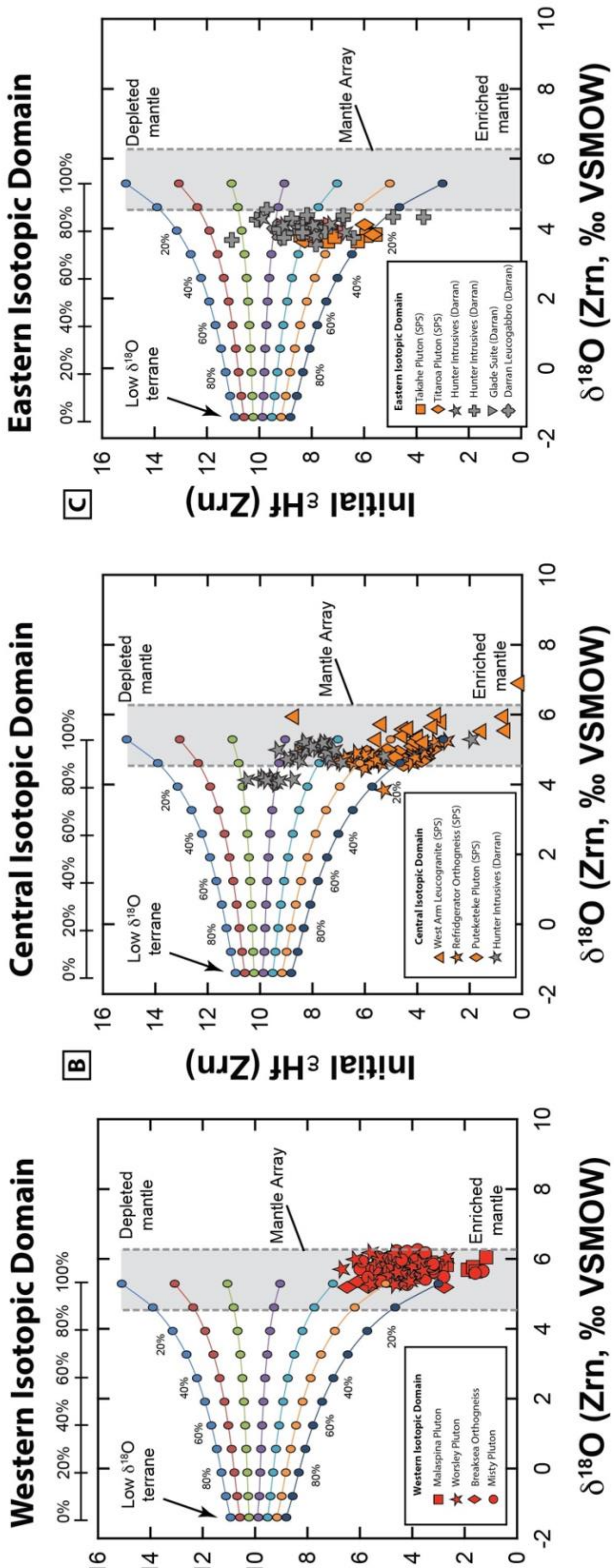

A
Western Isotopic Domain
B
Central Isotopic Domain
C
Eastern Isotopic Domain
Initial εHf (Zrn)
δ18O (Zrn, ‰ VSMOW)
Low δ18O terrane
Depleted mantle
Mantle Array
Enriched mantle
0%
20%
40%
60%
80%
100%
Western Isotopic Domain
Malaspina Pluton
Worsley Pluton
Breaksea Orthogneiss
Misty Pluton
Central Isotopic Domain
West Arm Leucogranite (SPS)
Refridgerator Orthogneiss (SPS)
Puteketeke Pluton (SPS)
Hunter Intrusives (Darran)
Eastern Isotopic Domain
Takahe Pluton (SPS)
Titiroa Pluton (SPS)
Hunter Intrusives (Darran)
Hunter Intrusives (Darran)
Glade Suite (Darran)
Darran Leucogabbro (Darran)

Figure 11